5.3 SITE HYDROGEOLOGY

This subsection describes the hydrogeologic framework of the site area, the testing and monitoring activities that have provided information about the hydrology of the site, and the hydraulic properties of the rocks and other associated characteristics. Then, individually for the unsaturated and saturated zones, hydrologic; geochemical; geothermal; and other pertinent results are applied in descriptive and modeling syntheses of the subsurface hydrology. Finally, the paleohydrology of the site area is discussed.

5.3.1 Hydrogeologic Framework of the Flow System

The Yucca Mountain Site occupies an intermediate position between the areas of recharge and discharge of the regional system, as described in Subsection 5.2. The aridity and small average rate of recharge produce a correspondingly small groundwater flux, which is readily transmitted through the well fractured rocks that compose most of the region. Consequently, the average regional hydraulic gradient is also small, resulting in a water table (the upper surface of the regional saturated zone) that commonly is deep beneath the land surface, particularly beneath prominent ridges such as Yucca Mountain. The combination of aridity, large topographic relief, and transmissive rocks results in a thick unsaturated zone that is a principal hydrologic attribute of the site and that is proposed as the host environment for the repository.

The areas of investigation and the methods of study differ significantly for the unsaturated and saturated portions of the Yucca Mountain hydrologic system. In the unsaturated zone, infiltration of precipitation and subsequent percolation to the water table involve dominantly vertical flow, so the area of principal interest is essentially that of the potential repository and possible expansion areas, with a small contiguous area to accommodate lateral dispersion and deflection of flow by geologic features. Figure 5.3-1 (reprint of Figure 3.7-1) shows the topography of this area, the potential repository and expansion areas, and the locations of the exploratory tunnel and boreholes. A geologic map and cross section of the site area are shown in Figures 5.3-2 and 5.3-3, respectively.

Although unsaturated percolation to the water table contributes recharge to the saturated zone at Yucca Mountain, most of the groundwater flux in the saturated zone beneath the vicinity of the site probably is throughflow which, in a regional context, is from generally north to generally south (Figure 5.3-4). At the site itself, the local saturated-zone flow direction is southeast toward Jackass Flats but is believed to turn southward within a few kilometers. To characterize this throughflow, the area of interest for saturated-zone studies historically has been large enough to include hydrogeologic features that influence flow and potential contaminant transport to the downgradient boundary of the "postclosure controlled area," beyond which total releases of radioactivity to the accessible environment have been limited by federal regulation.

Probable changes to the regulations will specify a dose-based standard to be applied at the nearest point of likely persistent human consumption of water and of food products using that water. Historical patterns of settlement, which are influenced by topography and a supply of potable water at an economically reasonable depth, indicate that the revised standard would be applied at the northern Amargosa Desert, including the community of Amargosa Valley (formerly Lathrop Wells)

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and the nearby agricultural area known locally as Amargosa Farms. Consequently, the area currently to be considered for site-scale saturated-zone studies extends southward to the northern Amargosa Desert.

Figure 5.3-5, from Czarnecki, Faunt et al. (1997), shows the major geographic features of the site saturated-zone study area and the surrounding region. The area extends from central Jackass Flats (longitude 116°18') on the east to western Crater Flat (longitude 116°37'30") on the west, and from the headwaters of Beatty Wash (latitude 36°56') on the north to northern Amargosa Desert (latitude 36°34') on the south. The southern limit of this area is about 30 km south of the potential repository. As thus defined, the site saturated-zone study area coincides with the area of the site saturated-zone flow model that is in preparation (Subsection 5.3.5.5). The boreholes used as hydraulic-head observation wells in the northern half of the study area were, with few exceptions, drilled within the anticipated controlled area by the Yucca Mountain Site characterization program, whereas those in the southern part were drilled for other purposes, principally as private water supply wells. Between these clusters of wells, hydrologic control is sparse.

The potential repository site at Yucca Mountain is a north-trending ridge formed by an eastward-tilted, fault-bounded block of layered volcanic rocks (Figure 5.3-4), near the southern limit of an extensive terrane of Miocene volcanics that blanket much of southern Nevada. The volcanic rocks, mainly ashflow tuffs and lavas, thin southward from their source areas to about 2 km thick beneath Yucca Mountain.

Southward from the site area toward the northern Amargosa Desert, depositional thinning and post-Miocene erosion combine to produce an increasingly discontinuous distribution of the volcanics, allowing exposure of the underlying Paleozoic and Proterozoic rocks at the surface or beneath surficial valley fill sediments. Figures 5.3-6 and 5.3-7 show a generalized hydrogeologic map and a highly schematic, north-south cross section that demonstrate these relationships.

Numerous faults occur at and near the site. Only the more prominent faults exposed at the surface are shown ("Major structural features") in Figure 5.3-6. The north-striking faults that border and intersect the site are part of a more extensive set that forms additional ridges to the east and west of the site and that underlies the intervening valleys. Northwest-striking faults occur just to the north of the primary site, as well as south and southeast of the site. Northeast-striking faults are regionally important to the east and south of the site, projecting into the southern part of the saturated-zone study area. These faults probably influence the movement of ground water significantly, whether they serve as preferential pathways or impediments for flow.

The upper part of the Miocene volcanic sequence and the surficial Quaternary-Tertiary sediments compose an unsaturated zone that is 400 to 750 m thick in the site area, owing principally to topographic irregularity above a relatively flat water table. From west to east between block-bounding faults, the generally eastward dip of the layered volcanic rocks causes successively younger strata to be submerged beneath the more gently sloping water table. Therefore, part of the suite of rocks described in the context of the unsaturated zone in the potential repository area also forms part of the saturated-zone suite in the larger site area.

Table 5.3-1 provides the principal stratigraphic units, correlative hydrogeologic units, and their general distributions in the unsaturated and saturated regimes in the vicinity of Yucca Mountain. A comprehensive description of the stratigraphy and lithologies is contained in Section 3 (Table 3.6-2). The hydrostratigraphy and hydrologic properties are described in greater detail in subsequent sections on unsaturated-zone and saturated-zone hydrology.

Quaternary-Tertiary Deposits–The surficial Quaternary-Tertiary deposits comprise mostly alluvial and colluvial sediments that fill the washes and broader valleys. These sediments consist of fluvial sand and gravel, as well as poorly sorted flood and mudflow debris. At and near Yucca Mountain, the Quaternary-Tertiary deposits form a discontinuous, generally thin surficial layer. Nonetheless, the alluvium and colluvium within the washes that cross the potential repository site have an influence on infiltration to the unsaturated bedrock. Along Fortymile Wash in western Jackass Flats, Quaternary-Tertiary alluvium is extensive and possibly hundreds of meters thick. However, the water table beneath most of Jackass Flats, except in the extreme south, is probably within the underlying Tertiary rocks.

In the Amargosa Desert, 25 km and more south of the repository site, the Quaternary-Tertiary is a voluminous basin-filling aquifer that stores vast quantities of water. Fluvial sand and gravel dominate in the northern part and, in the central part, along the modern and ancestral channels of the Amargosa River, Fortymile Wash, and Topopah Wash. At depth and southward in the Amargosa Basin, lacustrine deposits probably dominate, consisting of dense limestone and compacted silt and clay, as well as basin-margin sand, gravel, and travertine spring deposits. At the modern land surface, playas occur in the southern and eastern parts of the basin. The subsurface extent of the associated fine-grained sediments (QTc) is not well established, and other playa deposits could exist at depth elsewhere in the basin.

Quaternary-Tertiary basalt lava flows occur at the surface in Crater Flat and locally in the subsurface in both Crater Flat and northern Amargosa Desert. These flows are classed as part of the QTa because of their probable moderate to large fracture permeability, but they are too small to be hydrologically important except at very local scales.

Tertiary Volcanic Rocks–The thick sequence of Tertiary (Miocene) volcanic rocks that forms Yucca Mountain comprises principally ashflow tuffs with lesser thicknesses of interstratified ashfall tuffs, lavas, and other lithologies of volcanic origin. Because the heat and pressure varied widely with position in the flows, ashflow tuffs range from porous and poorly lithified rocks to very dense, highly lithified rocks.

Welded tuffs are the highly compacted rocks composed mainly of glassy (noncrystalline) particles, with lesser amounts of pre-existing rock fragments ("lithic fragments"), all bonded by heat and pressure within the interiors of ashflow sheets. As a function of welding intensity, moderately to densely welded tuffs have progressively smaller porosities, matrix permeabilities, and ductilities. Cooling fractures, often columnal, have formed within individual flow units. Because of their brittle nature, densely welded tuffs fracture extensively under tectonic stresses, thus developing moderate to large secondary permeability. Low permeability matrix is resistant to secondary mineral alteration, but fractures are locally coated by secondary minerals, and smaller fractures may become

plugged. In the unsaturated zone, the small matrix permeability can accept only a very small flux, and water potential is highly sensitive to the degree of saturation in the small pore spaces. Exchange of water between matrix and fracture pathways is not accomplished readily, but the fracture system provides permeable, interconnected pathways for the movement of water and gases. Similar conditions prevail in the saturated zone; matrix permeability is very small, but bulk (effective) permeability may be very large so that the welded tuffs are generally aquifers. Matrix and fracture porosity of welded tuffs vary widely because of their sensitivities, respectively, to degree of welding and structural setting.

Nonwelded tuffs include ashflow units, which were subjected to lower temperature or less pressure (or both) than welded units, and ashfall tuffs, which are composed of particles that settled from the atmosphere. Commonly, ashfall tuffs are distinctly stratified and are designated as "bedded tuffs." Thin, unnamed bedded tuffs occur at several horizons within the volcanic section. Unaltered nonwelded tuffs typically are composed principally of vitric (glassy or noncrystalline) particles and lesser amounts of lithic fragments, whereas secondary alteration in the presence of water produces zeolite and clay minerals. Primary properties of nonwelded and bedded tuffs include high porosity, moderate permeability, and sufficient ductility to deform under stress rather than fracture readily. Secondary alteration, which becomes more intense and more pervasive at greater depths, has further increased the porosity but decreased the permeability. In the unsaturated zone, nonwelded tuffs retain water in their small pore spaces by capillary tension (low water potential) and may imbibe water from contiguous sources of higher water potential.

Montazer and Wilson (1984) used the categories of welded and nonwelded tuffs as the principal determinants of unsaturated hydrologic properties for those units occurring above the water table at the site, that is, those units within 500 to 750 m of the surface. Their terminology is given in Table 5.3-1. Luckey, Tucci et al. (1996) defined a series of volcanic aquifers and confining units (also called aquitards) for the entire sequence of volcanic rocks in the saturated zone at the site, and they adopted the earlier definitions of Winograd and Thordarson (1975) for the pre-Tertiary rocks. Within the volcanic suite at the site, Luckey et al. recognized an upper volcanic aquifer, an upper volcanic confining unit, a lower volcanic aquifer, and a lower volcanic confining unit. Elsewhere in the region, Laczniak et al. (1996) and Czarnecki, Faunt et al. (1997) recognize a lava flow and welded tuff aquifer that is stratigraphically equivalent to the central units of the Luckey et al. lower volcanic confining unit. To facilitate correlation between site scale and regional models, separation of the Luckey et al. lower volcanic confining unit into middle and lower volcanic confining units is adopted herein (Table 5.3-1), despite that the "lower volcanic aquifer" has not been identified in the subsurface of the site area.

The uppermost 500 m of the volcanic section at Yucca Mountain is dominated by the Paintbrush Group, which comprises two thick welded tuffs (Tiva Canyon Tuff and Topopah Spring Tuff), separated by a lesser thickness of nonwelded Yucca Mountain Tuff and Pah Canyon Tuff. Hydrogeologically, from the surface downward, these units are termed the Tiva Canyon welded unit, the Paintbrush nonwelded unit, and the Topopah Spring welded unit. The Tiva Canyon Tuff and the Topopah Spring Tuff have nonwelded tuffs and vitrophyres near their lower and upper margins, respectively, and unnamed bedded tuffs also occur between them. For classification of the

unsaturated hydrogeologic units, these nonwelded and bedded tuffs are grouped with the PTn, as shown on Table 5.3-1.

The TCw unit occurs only in the unsaturated zone in the site area, forming the conspicuous, tilted caprock of Yucca Mountain. The PTn probably occurs only in the unsaturated zone, although it possibly dips beneath the water table near Fortymile Wash if it is present. Beneath the potential repository site, it serves to deflect, attenuate, and disperse water that infiltrates rapidly through the fractures of the overlying TCw unit. The thicker and deeper TSw unit occurs discontinuously below the water table at several places in the vicinity of the site, mainly to the southeast, where it is designated the upper volcanic aquifer. However, the TSw unit is best known as the dense, welded tuff that dominates the unsaturated sequence at the site and which is the host rock for the potential repository.

The Calico Hills Formation in the site area occurs partly above the water table and partly below, becoming progressively more submerged as it dips eastward. The thick nonwelded tuffs of the Calico Hills Formation, together with the vitrophyre and bedded and nonwelded tuffs at the base of the overlying Topopah Spring Tuff, were designated as the upper part of the CHn unit by Montazer and Wilson (1984). (The lower part of the CHn unit is the Prow Pass Tuff, described below.) The Calico Hills Formation contains several relatively thin lava flows in the northeastern part of the site area. In the southwestern part of the site area, where it is structurally elevated well above the water table, the tuff remains unaltered or vitric. Elsewhere, it is variable zeolitized. Respectively, these two facies have moderately large and very small permeabilities; the difference is recognized in the further refined designations, CHnv and CHnz. Where the equivalent rocks occur beneath the water table, they compose the upper volcanic confining unit, although they may be quite permeable locally near faults.

The Crater Flat Group, which comprises in descending order the Prow Pass Tuff, the Bullfrog Tuff, and the Tram Tuff, occurs principally in the saturated zone. Locally within the site area, mainly nonwelded Prow Pass Tuff occurs above the water table and is treated as part of the CHn unit for analyses of the unsaturated zone. In its more extensive saturated-zone occurrence, the Crater Flat Group is designated the middle volcanic aquifer. It includes interstratified welded tuffs, partly welded tuffs, nonwelded tuffs, and rhyolite lavas. Consequently, its primary hydraulic properties vary widely both vertically and laterally. Although pervasive alteration has formed zeolites and clays, principally affecting the nonwelded rocks, the composite rock mass has developed and maintained significant fracture permeability in places, probably controlled by proximity to faults.

The deeper part of the Tertiary volcanic suite, generally 1,500 m or more beneath the surface, includes tuffs, lava flows, volcanic breccias, and sediments composed of volcanic clasts, as was discussed in Section 3. Because of their deep burial and intense alteration, these "older tuffs," where tested, have very small matrix and fracture permeability. In the Yucca Mountain area, where the middle volcanic aquifer is not recognized, the composite older tuffs comprise both the middle volcanic confining unit and the lower volcanic confining unit.

Paleozoic and Precambrian Rocks–A thick section (thousands of meters) of lower to middle Paleozoic rocks, mainly limestone and dolomite, occurs beneath the erosional unconformity at the

base of the volcanic rocks beneath the contiguous region. This section is the regionally important "lower carbonate aquifer" of Winograd and Thordarson (1975). The unit has been extensively fractured during many episodes of tectonism, and dissolution features are evident near discharge areas and in some boreholes. As was discussed in Subsection 5.2, this highly permeable aquifer accumulates flow from a large region of the southern Great Basin and transports it to discharge areas such as Ash Meadows and Death Valley. Based on its occurrence in borehole UE-25p #1 and its regional distribution, the lower carbonate aquifer is inferred (Robinson, G.D. 1985) to underlie the Yucca Mountain Site area and to dip regionally to the north (Figure 5.3-7).

In the Calico Hills, about 10 km east of Yucca Mountain, the Devonian-Mississippian Eleana Formation occurs in outcrop and in the subsurface in borehole UE-25a #3. This unit was termed the "upper clastic aquitard" by Winograd and Thordarson (1975) and more recently has been designated the "Eleana confining unit." It stratigraphically separates the lower carbonate aquifer from the "upper carbonate aquifer," which comprises upper Paleozoic (Pennsylvanian-Permian) carbonates that occur mainly to the east and northeast of Yucca Mountain. The Eleana Formation primarily consists of argillite with lesser thicknesses of carbonate rocks and quartzite. In the Calico Hills, it contains sufficient amounts of secondary magnetite to cause a prominent magnetic anomaly, which extends westward beneath Yucca Mountain, just to the north of the potential repository site area. Based on the magnetic anomaly and the interpreted northward dip of the Paleozoic section beneath the erosional unconformity, G.D. Robinson (1985) infers that the Eleana Formation occurs in stratigraphic juxtaposition over the lower carbonate aquifer and beneath the volcanic rocks of northern Yucca Mountain, as is shown in Figure 5.3-7. Where present, the Eleana Formation functions hydraulically as a confining unit, impeding flow upward from or downward into the lower carbonate aquifer. Whether it is important to the northern part of the Yucca Mountain saturated flow system remains speculative. However, the Eleana Formation and Upper Paleozoic rocks are absent between Yucca Mountain and Death Valley.

Beneath the lower carbonate aquifer but above the Archean crystalline basement, there is a sequence of dominantly clastic, now metamorphosed Proterozoic and lower Cambrian sediments (quartzite, slate, and marbleized dolomite) that were named the "lower clastic aquitard" by Winograd and Thordarson (1975). The abbreviation "qcu" is shown in Table 5.3-1, following the usage of Czarnecki, Faunt et al. (1997) in the site saturated-zone model. These rocks have very small permeability except where they are disrupted by unhealed fractures. Active, gravity-driven groundwater flow is generally restricted to the rocks above this effective hydraulic basement. Southward from the site area, or in the updip direction relative to the structural attitude of the pre-Tertiary rocks, the Proterozoic-Cambrian section rises toward surface exposures in the Specter Range and Skeleton Hills near Amargosa Valley. Beneath southwestern Jackass Flats and the northern Amargosa Desert, these metasediments probably subcrop at the erosional unconformities beneath the Tertiary volcanics or the younger surficial alluvium (Figure 5.3-7).

Superimposed on this hydrostratigraphic framework is a complex structural regime. Paleozoic and older rocks were strongly deformed by thrust faulting and folding. Before the thick Tertiary volcanics were extruded over the region, a long period of erosion removed much of the Paleozoic rocks. Regional crustal extension in Tertiary time first was associated with the extensive volcanism and, subsequently, with faulting that segmented the crust. Major faults, generally north-striking,

separate large blocks that have been tilted eastward in the site area. A less dominant but nonetheless important set of faults strikes northwest into the site area from the east. Northeast-striking faults occur south of the site and are probably concealed beneath alluvium in southwestern Jackass Flats and the contiguous Amargosa Desert. The most obvious hydrogeologic effects of faults are disruption and displacement of stratigraphic continuity and tilting of the units. The faults themselves, from place to place, may act as conduits for unsaturated downward percolation or saturated flow through highly fractured rock, may be open impediments to unsaturated capillary flow, or may contain gouge that would shun saturated flow but attract capillary flow. Perhaps more important volumetrically are the pervasively fractured rocks near the faults. The results of field investigations at Yucca Mountain have shown that the volcanic section can be quite impermeable where it is intact, but also that large volumes of intact rock are unexpected.

The Yucca Mountain groundwater system can be considered within a simplified, conceptual context to consist of generally downward percolation through the unsaturated zone at the site, becoming recharge to the shallow saturated zone. The saturated-zone throughflow, which originates in the highlands to the north and northeast, is initially southeasterly from the site but changes to the regional southward direction within a few kilometers, partly because of the influence of the dominantly north-striking faults. Groundwater flows from the volcanic aquifers, where they pinch out south of Yucca Mountain, into a thick sequence of valley fill sediments in the Amargosa Desert. However, the transition from the volcanics to the valley fill remains speculative because of the lack of hydrogeologic information. The alluvial sediments of Fortymile Wash in the western part of Jackass Flats provide a potential pathway that is probably continuous into the valley fill sediments, but flow along fault controlled pathways from the volcanics, through structurally high Paleozoic rocks, and then into the Amargosa valley fill is a likely possibility.

5.3.2 Testing and Baseline Monitoring

Boreholes drilled, sampled, tested, and monitored at the site have provided information on the vertical distribution of hydrogeology, hydrologic properties of the rocks, thermal and other geophysical conditions and properties, chemistry of the contained fluids, and fluid potential. Additional information for some of these parameters in the unsaturated zone has been obtained from the excavations for the Exploratory Studies Facility, and from boreholes drilled from the drifts or alcoves of the Exploratory Studies Facility.

This section explains the basis for the borehole designation system and describes the selection of borehole locations. It specifies the sources of field data and samples for unsaturated zone and saturated zone investigations. Although separate maps and tables are given below in the unsaturated zone and saturated zone sections, many of the same boreholes are addressed in both sections; for example, the water table series of boreholes provides information on the top of the saturated zone but also provides samples and access for geophysical logs to characterize the unsaturated zone.

Geographic features that are referred to in the following sections are shown in Figure 5.3-8. Discussions of the hydrology of the site area commonly address the influence of faults, which are shown in Figure 5.3-9.

The locations of surface based deep boreholes at and near the potential repository site are shown in Figure 5.3-10. Not shown are the numerous shallow boreholes drilled to characterize near-surface infiltration of water, the tunnels of the Exploratory Studies Facility, the boreholes drilled from the Exploratory Studies Facility, and certain special purpose holes that have not provided hydrologic information.

Borehole Designations–Each borehole used in the study of the Yucca Mountain area has a unique name or number. Surface-based boreholes on the Nevada Test Site use a Nevada Test Site designation, whereas those off the Nevada Test Site use a slightly different designation. Most boreholes on the Nevada Test Site begin with UE (underground exploratory), followed by the Nevada Test Site area number (usually area 25). This designation, "UE-25," commonly is followed by one or more letters signifying the purpose of the hole or simply a site or locality, followed by a sequence number. Thus, the designation "UE-25 UZ #4" specifies a hole on the Nevada Test Site in Area 25 and that it is the fourth hole planned (not necessarily drilled) in a series of UZ (unsaturated zone) holes. The designation "UE-25c #3" specifies the third hole planned at a site named "c" in Area 25, but it reveals nothing about the purpose; the UE-25c site, commonly referred to as the "C-Well Complex," is the location of multiple hole aquifer and tracer tests.

Boreholes off the Nevada Test Site, on adjacent land under the control of the U. S. Bureau of Land Management or Nellis Air Force Range, begin with the letters USW (for underground, Southern Nevada, waste). The designation "USW" is followed by one or more letters signifying the series (the primary purpose) of the borehole followed by a sequence number. For USW boreholes, the series identifier and sequence number are separated by a hyphen rather than the number symbol, #. Thus, the first hole drilled (1983) in the unsaturated-zone series was off Nevada Test Site and therefore was designated "USW UZ-1."

Because sequence numbers were assigned to all proposed well sites, gaps in the sequence numbers of wells monitored indicate that not all proposed wells were drilled. For example, borehole USW H-2 was proposed for drilling near the northern edge of Yucca Mountain, but it was not drilled.

The letters signifying primary purpose that are used in this section of the site description are G (collection of geologic data), H (collection of hydrologic data), N (neutron-logging holes for infiltration data), NRG (collection of geologic data along the Exploratory Studies Facility North Ramp alignment), p (collection of data on rocks of Paleozoic age), SD (collection of geotechnical and hydrologic data from a statistically located Systematic Drilling program), SRG (collection of geologic data along the Exploratory Studies Facility south ramp alignment), UZ (collection of unsaturated-zone hydrologic data), VH (collection of geologic and hydrologic data pertaining to young volcanic rocks of Crater Flat), and WT (collection of water table data).

The only surface based boreholes not using this designation system and referred to in this report are Well J-11, Well J-12, and Well J-13, which were drilled in the 1960s as water supply wells for the Jackass Flats facilities. In some reports specific to Yucca Mountain studies, however, the original Nevada Test Site nomenclature has been changed to include the UE-25 prefix, as in "UE-25 J#11." Boreholes that were drilled (or, in some cases converted) to test or monitor the saturated zone are frequently referred to as "wells" rather than boreholes in project reports and in this document.

Boreholes that have been drilled from the underground Exploratory Studies Facility workings for hydrologic and pneumatic tests are designated by the abbreviation for the alcove in which they were spudded; a code for the principal purpose of the hole or test; and a sequence number. For example, upper Paintbrush Contact alcove RBT#2 designates the second hole for the radial borehole test in the upper Tiva Canyon alcove, whereas Bow Ridge fault alcove HPF#1 designates the first hole for testing hydrologic properties of faults in the Bow Ridge fault alcove.

Some boreholes were intended to serve both geologic and hydrologic purposes and, thus, they were constructed and tested approximately in the same way as H-holes. These include UE-25b #1 (sometimes designated UE-25b#1h), UE-25p#1, USW G-4, and USW VH-1. Other boreholes have provided incidental potentiometric data; these include some of the G-series, NRG-series, and SD-series holes. Temperature logs for site characterization studies (obtained after the effects of drilling, geophysical logging, and borehole testing had subsided) were obtained in most of the holes in the following series: G, H, UE-25a, UE-25b, UE-29a, VH, and WT. Temperatures were also logged in USW UZ-1 and in some holes drilled from alcoves in the Exploratory Studies Facility.

Historical Considerations–Early exploration (1979 to 1982) at Yucca Mountain targeted deep rock units (Crater Flat Group) within the saturated zone as candidate host rocks for a repository. As compared with other low permeability rocks, such as salt and granite, that were also being evaluated elsewhere, the deeper tuffs were found to be quite permeable, and subsequent exploration was redirected to the unsaturated zone. The earlier holes at and closely contiguous to the potential repository site include the UE-25a series, the UE-25b series, G-series holes preceding USW G-4, and H-series holes preceding USW H-6. Drilling fluid losses in the unsaturated zone were large in several of these holes. Later drilling to explore the unsaturated zone encountered lost drilling fluids from USW G-1 in Drill Hole Wash, near the northern boundary of the potential repository site (Whitfield 1985). Some hydrologic disturbance is likely near other early boreholes. However, these boreholes occur along or beyond the periphery of the site.

5.3.2.1 Unsaturated-Zone Data Collection

To support development of conceptual and numerical models of fluid flow in the unsaturated zone at Yucca Mountain, various types of hydrologic data have been collected. These include:

- Infiltration data, including neutron moisture profiles in shallow boreholes, meteorological and climatic data, hydrologic properties of soils, and soil water potential data
- Small-scale matrix hydrologic properties of core samples from boreholes
- Bulk properties of the rock mass from air-injection tests conducted in deep, surface based boreholes
- In situ pneumatic pressure, water potential, and temperature data from multiple depths in deep boreholes

- Hydrochemical and isotopic characteristics of gas and water samples from surface based boreholes

- Results of hydrochemical sampling and air-injection testing conducted in test alcoves and the primary tunnels of the Exploratory Studies Facility

For each type of data, the methods used to collect the data and the scope of the data collection are discussed in the subsections below. Locations of surface based boreholes and the Exploratory Studies Facility are shown in Figure 5.3-11.

5.3.2.1.1 Infiltration Data

Various field and laboratory methods were used to collect data needed to construct conceptual, analytical, statistical, and numerical models of infiltration (see Subsection 5.3.4.1, "Site Infiltration"). These methods included:

- Neutron logging of boreholes to determine volumetric water content

- Laboratory analysis of matrix properties and hydrochemistry of core samples from shallow boreholes

- Monitoring of meteorological and climatic conditions

- Evaporation-pan measurements

- Measurement of soil hydrologic properties

- Measurement of water potential of soils and rocks using heat dissipation probes

These methods are described in more detail below.

All data for the infiltration investigations were collected, compiled, and processed in accordance with a fully compliant quality assurance program, and all data were submitted to the project data system as indicated in Table 5.3-2.

5.3.2.1.1.1 Neutron Logging of Boreholes

A network of 99 shallow boreholes was established for measurement of subsurface volumetric water content by neutron geophysical logging using a CPN model 501 hand held neutron-moisture meter (Blout et al. 1994). These boreholes are referred to as the unsaturated-zone neutron (N) boreholes (Figure 5.3-11). The neutron-moisture meter uses a weak neutron radiation source with an ionizing chamber to detect epithermal neutrons reflected back towards the meter following collisions with hydrogen atoms in the surrounding medium. Epithermal neutrons are slower than neutrons emitted from the source following collisions with hydrogen because the mass of a neutron is equivalent to the mass of the hydrogen atom. The epithermal neutrons thus provided a measure of the volume

fraction of hydrogen in the water present in the surrounding soils and volcanic rocks at the time the borehole was logged. Neutron logging of the boreholes was performed monthly from the time of installation through the end of water year (fiscal year) 1995. Following major storm events (precipitation depth greater than approximately 25 mm), selected boreholes also were logged weekly and daily. The logs provided a measure of volumetric water content at borehole intervals of 0.1 and 0.3 m, and were used to develop vertical water content profiles that indicate changes as a function of time and depth. These records were used to analyze the movement and redistribution of wetting fronts in response to precipitation and run off, provide information on ambient conditions deeper in the profile and during periods of drought, and provide data for the development and calibration of a numerical model of infiltration for Yucca Mountain. The neutron boreholes also provided core samples for the measurement of matrix hydraulic properties and for hydrochemical analysis.

The neutron boreholes were drilled using the ODEX-115 drilling system with air as the drilling fluid (Hammermeister et al. 1985). Core samples were obtained from selected boreholes by drive core methods in unconsolidated alluvium/colluvium and by rotary core methods in tuffaceous bedrock. Core samples obtained from the boreholes using these methods were approximately 63.5 mm in diameter. The diameter of the resulting hole was approximately 152 mm. The drilling method uses simultaneous advancement of the steel casing with the deepening of the hole to minimize the effects of air-drying on the borehole walls. The steel casing has an inside diameter of 130 mm, and is approximately 13 mm thick. The ODEX drilling system produces dry cuttings which tend to fill the small annular spaces between the casing and formation. Boreholes with a high degree of rugosity and void spaces between the casing and formation were grouted at the ground surface to reduce the potential of enhanced infiltration of water through the disturbed annular space. The grouting filled the annular space to a depth of approximately 0.3 m.

The initial set of 74 neutron boreholes was drilled between May 15, 1984 and February 26, 1986 ranged in depth from 4.6 to 36.6 m (Fenix & Scisson, Inc. 1987c). An additional borehole, USW UZ7, was drilled to 63.1 m, but the casing was pulled back up to 6.7 m. Boreholes were located on ridgetops, sideslopes, and terraces and in active channels based on the hypothesis that infiltration rates would vary in relation to topographic and geomorphic conditions. Drilling was halted from March 1986 to September 1991 while a fully compliant quality assurance program was developed for the YMP. Subsequently, 24 additional neutron boreholes, ranging in depth from 18.3 to 78.0 m, were drilled in late 1991 and early 1992. In addition to increasing the spatial coverage for the four infiltration zones, boreholes in this second set were located based on geologic features such as lithology of the underlying bedrock and fault zones (Flint, L.E. and Flint 1995). For example, several boreholes were located to penetrate through and collect core samples from the Paintbrush nonwelded hydrogeologic unit. Locations of neutron boreholes in the site area are indicated in Figure 5.3-12. More detailed information on borehole location, altitude, drilling date, topographic position, infiltration zone, total depth, depth of casing, depth to bedrock, and lithology of uppermost bedrock for all 99 boreholes is contained in L.E. Flint and Flint (1995). The use of the volumetric water content data obtained from the neutron boreholes in the development of an infiltration model for the Yucca Mountain Site is described in Subsection 5.3.4.1.3.1 of this report.

5.3.2.1.1.2 Matrix Properties and Geochemical Analysis of Core Samples

The primary purpose of obtaining core samples from the neutron boreholes was to measure matrix hydrologic properties and determine lithology. Matrix properties measured in the laboratory were density, porosity, volumetric water content, water potential, and permeability. (Note: In the context of matrix properties, "water content" always means volumetric water content.) These measurements were used in conjunction with data obtained from other surface based boreholes, surface outcrop samples, and Exploratory Studies Facility samples to help define the hydrogeologic framework of Yucca Mountain (see Section 3 of this report. The measured properties also provided calibration data for the neutron-moisture meters (see Subsection 5.3.2.1.1.1) and defined hydrogeologic parameters for neutron boreholes for use in analyzing and modeling water content profiles measured in the boreholes. To obtain accurate measurements of water content (and is some cases, geochemistry), retrieved core was quickly packaged in Lexan or sealed cans at the drilling site to minimize moisture losses. Prior to core packaging, a video record of all core retrieved was obtained. The processing and analysis of core samples is discussed in greater detail in Subsection 5.3.2.1.2.

Core samples obtained from the neutron boreholes also were subjected to geochemical and isotopic analysis including ^{36}Cl and tritium methods. The methodology of ^{36}Cl analysis is described in Fabryka-Martin, Wightman et al. (1994) and the application of ^{36}Cl analysis to infiltration studies at Yucca Mountain is described in Subsection 5.3.4.1.5.2 of this report. Interpretation of vertical profiles of tritium detected in pore water extracted from core samples of the neutron boreholes was used in a qualitative manner to develop conceptual models of infiltration at Yucca Mountain, as described in the study plan for infiltration studies (DOE 1991). However, tritium profiling was not used in a quantitative manner to estimate infiltration at Yucca Mountain.

5.3.2.1.1.3 Meteorological and Climatic Data Collection

Regional Precipitation and Climate Monitoring–To provide regional data on present-day climatic conditions to support hydrologic studies of unsaturated-zone infiltration and saturated-zone recharge, daily records of 24-hour precipitation amounts were available from non-YMP sources from a regional network of 156 precipitation stations located throughout the Yucca Mountain region. Precipitation records for varying periods of time between 1901 and 1995 were used, including data from stations located in parts of Southern and Central Nevada, Southern California, southwestern Utah, and northwestern Arizona (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). For many of these sites, daily measurements of maximum and minimum air temperature, snow depth, and (or) pan evaporation amounts also were available. These regional precipitation data, collected by various federal, state, and private agencies, were not collected under an approved YMP quality assurance program. However, these data were needed for determining how conditions measured at Yucca Mountain compare to the spatial and temporal distribution of precipitation on a regional scale, and for developing statistical models of the spatial and temporal distribution of precipitation that could be used as input for infiltration and recharge modeling. Climate and meteorologic data from regional sites are especially important because records for Yucca Mountain are short (less than 10 years) compared to records of 30 years and longer which were available at the regional sites. A description of the regional network and the daily precipitation data used for development of spatial and temporal precipitation models for infiltration and recharge studies is provided by Hevesi and

Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). An analysis of the spatial and temporal distribution of precipitation amounts measured at regional stations for two wetter-than-average water years (1992 and 1993) is presented by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). Precipitation records and analysis of data obtained from a similar regional network of precipitation stations for the purpose of defining present climate and site meteorology for the Yucca Mountain environmental program are presented in Subsection 4.1 (Site Climate and Meteorology) of this report.

Site Precipitation and Meteorological Monitoring–To obtain site-specific meteorological data for water balance studies at Yucca Mountain, five automated Penman-type weather stations were installed and operated from water years 1988 through 1995. Measurements obtained at the weather stations included precipitation timing, intensity, and duration; air temperature; wind speed and direction; humidity; and solar radiation. Detailed descriptions of these monitoring sites and their locations and a summary of the data collected during water years 1988 through 1994 are contained in A.L. Flint and Davies (1997). An expanded network of tipping bucket precipitation gauges was installed and operated for water years 1995 and 1996 to improve the spatial coverage of precipitation monitoring at Yucca Mountain. Additional site-specific data were collected for water years 1992 through 1995 from a dense network of approximately 130 nonrecording storage type rain gauges located mostly over the area of the potential repository site (Ambos et al. 1995). Measurements of total precipitation depth accumulated from a single storm event or from a series of closely spaced storm events were obtained using the storage type (nonrecording) gauge network. Although the data from the storage type gauges do not provide an accurate measure of the timing of precipitation or of the temporal distribution of precipitation intensity and duration, the data were used to characterize the spatial distribution of accumulated precipitation amounts over the area of the potential repository. Total precipitation amounts were measured by the nonrecording rain gauge network for all major storm events from October 1991 through October 1993 (Ambos et al. 1995). A geostatistical analysis was performed of the spatial variability of total precipitation for these storm events, including a comparison of winter versus summer storms (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). The relationships between precipitation and infiltration at the site are described in Subsection 5.3.4.1.3.4 of this report.

Site meteorological data also were collected by a network of nine automated weather stations by the YMP environmental program. These stations provided precipitation data from both automated tipping bucket and storage type (nonrecording) precipitation gauges, wind speed and wind direction data, air temperature, relative humidity, barometric pressure, and solar radiation data. Locations, instrumentation, data collection and analysis, and summary of measurements for these meteorological stations are described in detail in Subsection 4.1 (Site Climate and Meteorology) of this report. Although the primary purpose of the this network was to obtain meteorological data for atmospheric transport modeling in support of the Environmental Impact Statement for Yucca Mountain, the data also were used for infiltration and water balance studies.

5.3.2.1.1.4 Measurement of Pan Evaporation

Pan evaporation was measured from 1989 through 1995 using a Class A evaporation pan at a site in Area 25 of the Nevada Test Site near one of the site characterization weather stations located

approximately 10 km east of the Yucca Mountain Site (Flint, A.L. and Davies 1997). Measurements of pan evaporation were obtained periodically at intervals of one day to several weeks and daily and average daily pan evaporation rates were calculated. Although the data were collected in support of potential evapotranspiration studies, the data cannot be used as a direct measure of potential evapotranspiration because of strong advective effects on such measurements in arid environments. The data were obtained to estimate the daily, seasonal, and annual variability in potential evapotranspiration, and to establish the upper limit of available energy for potential evapotranspiration modeling given that modeled potential evapotranspiration should not exceed pan evaporation.

5.3.2.1.1.5 Measurement of Soil Hydrologic Properties

Field and laboratory analyses of the various surficial soil types present at Yucca Mountain were conducted to define the spatial distribution of soil hydraulic properties (Guertal and Davies 1996). Large volume, field bulk density samples were collected from the surface to a depth of 0.3 m using the irregular hole, bulk density, bead-cone device described by A.L. Flint and Childs (1984). Bulk density, grain density, porosity, rock fragment content, and sand, silt, and clay percentages were determined. Saturated hydraulic conductivity was measured using a double-ring infiltrometer (Hofmann et al. 1993) on soils in measurable locations and estimated analytically using textural data for the fine soil fraction (less 2 mm) and equation 6.12 of Campbell (1985). Log-log water characteristic curves were determined using equations 2.15, 2.16, 2.17, 2.18, 5.10, and 5.11 of Campbell (1985) and were converted to van Genuchten curves using the RETC computer code (van Genuchten et al. 1991).

5.3.2.1.1.6 Monitoring Soil Water Potential Using Heat Dissipation Probes

Heat dissipation probes were used to obtain continuous water potential measurements in four depth intervals in the soil profile, including the soil-bedrock interface, near neutron borehole N15 in upper Pagany Wash on Bleach Bone Ridge for water year 1995. Water potential monitoring sites also were established in Split Wash during water year 1996. Heat dissipation probes measure water potential by recording how rapidly a pulse of heat moves away from the probe (Reece 1996). In general, the wetter the soil, the more rapidly heat is dissipated from the probe. The temperature of the probe is recorded at the time of the heat pulse and several milliseconds after the pulse. The probe is attached to a ceramic medium that is in contact with the soil or bedrock medium and is assumed to be in equilibrium with the surrounding medium. The thermal "characteristic" of the ceramic medium is known or calibrated so that a water potential can be derived from the measured temperature difference and corrected for the ambient temperature of the surrounding medium. The continuous records of soil water potential were analyzed to determine the arrival and subsequent redistribution of a wetting front caused by periodic infiltration of water as a result of rainfall or snow melt. These water potential data are interpreted in Subsection 5.3.4.1.3.2.

5.3.2.1.2 Matrix Hydrologic Properties Data

To determine the physical and hydrologic properties of the rock matrix of the hydrogeologic units that comprise the unsaturated zone at Yucca Mountain, 4,892 rock core samples from 31 boreholes

were analyzed in the laboratory and the measurements were compiled into a database. For each sample, the database includes measurements of porosity, bulk density, particle density, volumetric water content, saturated hydraulic conductivity, moisture retention characteristics, and the associated lithostratigraphic unit. (Note: In the context of matrix properties, "water content" always means volumetric water content.) To provide initial conditions and calibration data for the site unsaturated-zone flow model (Subsection 5.3.4.4), present-day field moisture and water potential conditions also were included in the database. Selected samples also were analyzed for unsaturated hydraulic conductivity. Comprehensive descriptions of the methods used to measure matrix properties and the results of analysis are contained in L.E. Flint (1998).

Core samples were collected and processed in accordance with a fully compliant quality assurance program and all data were submitted to the Project data system (Table 5.3-3). All samples were analyzed to determine porosity, bulk density, particle density, and water content. Subsets of samples were analyzed to determine water potential, saturated hydraulic conductivity, particle density by helium pycnometry, and moisture retention characteristics. The designations, locations, altitudes, drilling dates, and depths for all boreholes from which cores were collected for matrix property measurements are listed in Table 5.3-4. (Note: In this report section, borehole designations are abbreviated by indicating only the borehole series identifier, "N," "UZ," "SD," or "NRG," and the associated number, omitting the prefixes "UE-25" and "USW.") The borehole database includes 23 of the most recently drilled neutron (N) boreholes, which range in depth from 18 to 83 m. (Long-term, volumetric water-content measurements in the neutron boreholes are described in detail in Flint, L.E. and Flint (1995) and in Subsection 5.3.4.1.3.1 of this report.) The database also includes eight deep boreholes, five of which penetrated the water table (three SD boreholes, UZ-14, and UZ#16) and three of which did not penetrate the water table (UZa and two NRG boreholes). The lithostratigraphic units sampled for matrix properties analysis are indicated for the neutron and deep boreholes in Tables 5.3-5 and 5.3-6, respectively.

5.3.2.1.2.1 Sample Collection Methods

Boreholes from which core samples were collected for matrix property measurements were drilled using dry drilling technology. The neutron (N) boreholes were drilled using the ODEX 115 drilling and casing system (Hammermeister et al. 1985) that employs simultaneous advancement of casing with deepening of the hole to minimize the effects of drilling fluids (in this case, air) on the borehole walls. Formation and core moisture conditions are minimally affected, and some of the cuttings produced from drilling fill the small annular spaces between the casing and formation, minimizing the void space. The remaining boreholes were drilled using reverse circulation dual-wall methodology that does not leave a casing in the borehole but also is designed to minimally affect formation and core moisture conditions. However, it has been demonstrated that despite rigorous sample preservation precautions, this drilling method does significantly dry out core samples of welded tuff, impacting the validity of laboratory water potential measurements. Both types of drilling are designed to provide continuous core sampling. Core samples approximately 7 cm in diameter and 10 cm and 20 cm in length were collected approximately every 0.8 m during drilling of the boreholes. Occasionally, gaps occurred in the core record because core samples were not recovered due to drilling or formation conditions. Typically, samples were placed in steel cans at

the drill site within 5 minutes of being recovered from the borehole to maintain field moisture conditions.

5.3.2.1.2.2 Core Processing in the Laboratory

Cores were processed following the steps indicated in Figure 5.3-13; details of laboratory sample processing and associated errors are described in L.E. Flint (1998). Cores were removed from the steel cans and divided into approximately two equal parts. One part was placed immediately on a balance to obtain field weight for moisture content, and then labeled with permanent ink. The other part was crushed and immediately sealed in a labeled, taped, glass jar, and then analyzed within 4 days for water potential using a laboratory-chilled mirror psychrometer. The intact core subsample then was evacuated of air under a vacuum and saturated with CO_2 to enable water saturation of small internal pores. The subsample then was submersed in distilled, de-aired water and left overnight. Samples were removed from the water, dried with a damp towel (ASTM R97-47), and weighed to determine saturated weight. The sample then was suspended in a beaker of water in a wire basket to determine volume displacement, dried for 48 hours in a relative humidity oven at 60°C and 65 percent relative humidity, and reweighed. Relative humidity drying removes water from the pores but retains water in the crystal or mineral structure (Bush and Jenkins 1970), or within very small pores or channels, providing an estimate of "residual water content," and relating total pore space to effective space available for water flow (see Subsection 5.3.3.1.1.1). The samples were then dried at 105°C for at least 48 hours to obtain a standard dry weight (ASTM R97-47). Porosity, bulk density, particle density, volumetric water content, and saturation were calculated for both relative humidity and 105°C oven dry weights. (Note: In this report, the term "porosity" refers to that determined from the 105°C oven dry weights unless otherwise noted.)

Borehole core samples 15 to 20 cm long were collected adjacent to the shorter 5 to 10 cm samples preserved in steel cans. These longer core samples were preserved in Lexan® tubing for measurements of saturated hydraulic conductivity, particle density using helium pycnometry, and moisture retention (Figure 5.3-13). Subsamples 2.5 cm in diameter and approximately 6 cm long were prepared and subsamples approximately 1 cm long were cut off the end of these samples for determination of moisture retention curves. The rest of the subsample was used for conductivity and helium-pycnometry measurement of particle density. Hydrologic flow properties were measured before conventional oven drying (105°C) because structural damage may occur in certain samples with delicate clay structures or zeolites. Finally, the samples were dried at 105°C to obtain a standard dry weight.

Saturated hydraulic conductivity (K_s) was determined on subsamples from several boreholes. Cores were vacuum saturated, and K_s was measured using a steady-state permeameter that forces water through the core at a measured pressure while weighing the outflow over time. K_s was calculated using Darcy's law. This measurement differs from the typical saturated hydraulic conductivity value that is measured under field conditions when rocks have entrapped air, and provides an upper bound to the prediction of unsaturated hydraulic conductivity necessary for the numerical modeling of water flow through rocks. Unsaturated hydraulic conductivity, $K_{rel}(\psi)$, was measured on several samples using a steady-state ultra-centrifuge (Conca and Wright 1992).

5.3.2.1.2.3 Moisture Retention Characterization

At sites where unsaturated hydraulic conductivity was not measured directly, moisture retention curves were used with saturated hydraulic conductivity to predict unsaturated hydraulic conductivity for use in numerical flow models. Moisture retention curves were determined in the laboratory by using the chilled mirror psychrometer to determine water potential at various saturations. Samples were saturated in a vacuum and successive water potential measurements were made after allowing samples to evaporate to a new water content and equilibrate. The final measurement point, which was used to approximate the residual water content, was determined following the drying of the sample in an oven set at 60°C and 65 percent relative humidity. For each hydrogeologic unit, the data sets for individual samples were compiled to include variability throughout the unit. These data sets represent the desorption process and are the upper bound of the moisture retention characteristic curve, neglecting hysteretic behavior. For rock units in the shallow unsaturated zone where hysteretic behavior might occur due to periodic wetting and drying and entrapment of air in pores, measured data from borehole samples were included with the laboratory data set to produce a composite data set. This was done for hydrogeologic units of the Tiva Canyon Tuff (except for the Tpcrn and Tpcpul, because no field water potential measurements were available), the PTn, and the upper rocks of the Topopah Spring Tuff (Tptrv and Tptrn). Each of the moisture retention data sets was fitted with curve functions from van Genuchten (1980). $K_{rel}(\psi)$ was also predicted from moisture retention fit parameters using the model of van Genuchten/Mualem (van Genuchten 1980) for comparison with unsaturated hydraulic conductivity measurements. Unsaturated hydraulic conductivity, $K_{rel}(\psi)$, was measured on several samples using a steady-state ultra-centrifuge (Conca and Wright 1992).

5.3.2.1.3 Air-Injection Testing in Surface Based Boreholes

Air-injection tests were conducted in surface based vertical boreholes at Yucca Mountain to quantify the in situ air permeability of the unsaturated, fractured and unfractured, tuffaceous volcanic rocks (LeCain 1997). The permeability of these rocks controls the movement of fluids in Yucca Mountain, including the transmission of water from the surface downward to the repository horizon and the movement of gases from the repository horizon to the surface. Variations in the tuff permeability can result in perched water zones, fast pathways, and capillary barriers. These variations may occur between stratigraphic units or within individual stratigraphic units. Knowledge of the spatial and directional variability of permeability is needed to formulate conceptual models and is required input to flow and transport models that attempt to represent the flow system at Yucca Mountain.

Air-permeability values were determined from 194 single hole, air-injection tests conducted in four surface based, vertical boreholes at Yucca Mountain from 1993 through 1995 using a straddle-packer air-injection system (LeCain 1997). The four boreholes were UZ#16, SD-12, NRG-6, and NRG-7a (Figures 5.3-10 and 5.3-11).

The geologic units tested were the Tiva Canyon Tuff, Yucca Mountain Tuff, Pah Canyon Tuff, Topopah Spring Tuff, Calico Hills Formation, and three bedded tuff units that are parts of the Paintbrush nonwelded hydrogeologic unit. Air-injection permeability values were calculated and analyzed by borehole and by geologic unit. Air-injection permeability values were compared to

permeability values derived from laboratory tests and pneumatic monitoring. Regression analysis was conducted between air-injection permeability values and rock characteristics such as fracture density and presence of lithophysal cavities. The results of these analyses are presented in Subsection 5.3.3.1.2.1 of this report.

5.3.2.1.3.1 Field Test Method

The surface based air-permeability testing was conducted at Yucca Mountain using state-of-the-art hydraulic, pneumatic, and electrical systems that allow the installation of pneumatic packers in vertical boreholes (LeCain 1997). Tests began with a review of the borehole geophysical logs (caliper, natural gamma, and gamma-gamma) and selection of the downhole test intervals. After a test interval was selected the pneumatic packers were installed in the vertical borehole, lowered to the selected test interval, and inflated with compressed air. Inflation of the packers isolated the selected test interval from the open borehole. After packer inflation the compressed air, tagged with sulfur hexafluoride tracer gas, was injected into the isolated test interval at a constant rate until the pressure in the test interval stabilized. An average test lasted approximately 10 minutes. Tests conducted in borehole SD-12 were followed by recovery tests.

The downhole test equipment consisted of four pneumatic packers connected end-to-end by aluminum pipe, which forms a packer assembly with three intervals (LeCain 1997). The middle interval was the test interval and ranged in length from 3.5 to 4.9 m. The two end intervals, each 1.2 m in length were guard intervals. The purpose of the guard intervals was to monitor for injection-air leakage around the inner two packers, which would invalidate the test. In boreholes that required a longer test interval because of poor wall conditions or other problems with the packers, only the outside packers were inflated. When only the outside packers were inflated the test interval length was 11.3 m. Each of the intervals contained a pressure transducer to measure absolute pressure and a thermistor to measure temperature. The downhole packer assembly was connected to the surface by a 700 m long tubing bundle that contained:

- Four 3/8-inch diameter electrical cables to power and monitor the pressure transducers and thermistors
- Two 3/8-inch diameter nylon tubes to inflate the packers
- One 1-inch diameter tube for air injection
- A 3/8-inch diameter steel cable to support the weight of the packers and tubing bundle

The packer assembly was lowered and raised in the borehole with a surface mounted hydraulic winch. The instruments were powered and monitored and the data recorded by data loggers at the surface. When a test interval was selected, the packer assembly was lowered to the selected interval and all four packers were inflated. Once the packers were inflated mass flow controllers were used to inject a constant mass flow of compressed air with a sulfur hexafluoride tracer. The mass flow rate ranged from 1.0 to 1,500.0 standard liters per minute. The air flow into the test interval resulted in an increase in pressure and a change in temperature. The electrical outputs of the pressure

transducers and thermistors were converted into engineering units of pressure and temperature. The test interval pressure and temperature responses were used to calculate an air-injection permeability value for the test interval, according to the approach described by LeCain (1997).

5.3.2.1.3.2 Air-Injection Test Analysis

The method used to analyze the air-injection tests originally was developed for incompressible fluids (LeCain 1997). Therefore, modifications for compressible fluids and flow geometry require the assumptions listed below:

- The ideal gas law applies, and therefore, the compressibility of the gas is inversely related to the pressure.
- The system is isothermal, and therefore, the gas density and viscosity, which are both temperature dependent, remain constant.
- Gravitational effects can be excluded, which is reasonable because the increased pressure from the weight of the higher density gas is small compared to the gas-injection pressure.
- Flow is laminar. Calculation of air-injection permeability values assumes that the flux of a gas through a medium is proportional to the differential pressure across which the flow occurs, which is true as long as Darcy's Law is valid.
- The medium is homogeneous, isotropic, and incompressible.

The effects of these assumptions on the air-injection tests are described in detail in LeCain (1997).

Because the air permeability of a rock changes with water content, a given permeability also has an associated capillary pressure. Capillary pressure is the pressure difference across the interface between the gas and liquid phases. Capillary pressure increases when this interface is confined to smaller pores or smaller fractures and decreases as this interface moves to larger pores or larger fractures. The larger pores and fractures are potentially the most conductive features and are dry at all but the wettest conditions (lowest capillary pressures). The relationship between air-permeability and capillary pressure was evaluated as described in LeCain (1997).

A solution for steady-state elliptical flow was used to evaluate the test interval pressure response, as described in LeCain (1997). The air-injection tests used a packer string with a test interval length of approximately 4 m. Based on the assumption that the rock is isotropic and the flow geometry is elliptical, the zone of investigation is an ellipsoid that extends approximately 4 m in all directions from the test interval. Analysis is based on the assumption that the pressure in the injection interval is at steady state.

5.3.2.1.4 Instrumentation and Monitoring of Surface Based Boreholes

Prior to 1997, nine deep boreholes were instrumented in the unsaturated zone at Yucca Mountain to monitor in situ pneumatic pressure, water potential, and temperature, at multiple depths in each borehole in accordance with a fully compliant quality assurance program. These boreholes were NRG#5, NRG-6, NRG-7a, SD-7, SD-9, SD-12, UZ#4, UZ#5, and UZ-7a (Figure 5.3-8). Two of these boreholes (UZ#4 and UZ#5) are located in Pagany Wash; four boreholes (NRG#5, NRG-6, NRG-7a, and SD-9) are located in Drill Hole Wash; UZ-7a is located in WT-2 Wash; SD-12 is located in Antler Wash; and SD-7 is located on the flank of Highway Ridge. Boreholes NRG#5, NRG-6, NRG-7a, and SD-9 are located along the Exploratory Studies Facility North Ramp (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164), whereas boreholes SD-12, UZ-7a, and SD-7 are located along the Exploratory Studies Facility Main Drift (Rousseau et al. 1997). The specific types of instrumentation, which vary from borehole to borehole, are described in the subsections below. The locations of instrument stations with respect to lithostratigraphy in each borehole are shown in Figure 5.3-14; the symbols depicting lithostratigraphic units are defined in Figure 5.3-15.

5.3.2.1.4.1 In Situ Pneumatic Pressure Measurements

In situ pneumatic pressure responses to atmospheric pressure changes have been monitored at multiple depths in the nine boreholes described above for various time periods since late 1994, when the first of these boreholes was instrumented. The periods of record for pneumatic pressure measurements in these boreholes are indicated in Table 5.3-7. In addition, boreholes UZ-6 and UZ-6s (located on Yucca Crest at Highway Ridge) and various neutron holes were used to collect composite borehole pressure measurements, gas-flow measurements, and gas-chemistry samples.

Boreholes NRG-6, NRG-7a, UZ#4, UZ#5, SD-12, and UZ-7a were instrumented with downhole pressure transducers that measured absolute pneumatic pressure (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997). Pressure transducers were calibrated over a range of 70 to 115 kPa and over a temperature range of 5 to 50°C (Kume and Rousseau 1994). Minimum calculated accuracy was ±0.035 kPa (95 percent confidence), although accuracies of ±0.020 kPa (95 percent confidence) have been obtained. Individual monitoring stations in these boreholes were backfilled or packed with porous stemming materials (polyethylene beads or coarse sand) to couple the monitoring zone to the formation rock of interest. Isolation of individual instrument stations in these boreholes was achieved with intervening lifts of stemming materials consisting of grout and sand. For these boreholes, the absolute pressure data was converted to pressure potential by subtracting the long-term mean from the absolute reading, effectively removing the gravity component of total pneumatic pressure and allowing pressure potential to be compared directly with pressure data collected from other boreholes (Patterson et al. 1996). Therefore, throughout this report the term "pressure" should be taken to mean "pressure potential." In the few instances where absolute measurements are referred to, the term "absolute pressure" is used to differentiate those measurements from the pressure potentials.

In boreholes NRG#5 and SD-7, downhole pressure potentials were measured at the land surface through individual tubes that were connected to open ports located along the length of an impermeable membrane liner (Patterson et al. 1996). The liner was inflated to seal the formation wallrock from direct communication with the atmosphere. Pressure potentials were measured using a differential pressure transducer at the ground surface that was referenced to atmospheric pressure at the time measurements were taken. A second pressure transducer was used to measure atmospheric pressure, which was added back to the differential pressure measurement to compute the pneumatic pressure potential at the location of the port. It was assumed that the gas in the tubes extending from depth to land surface was of the same composition and at the same temperature as gas in the adjacent rock. Therefore, because the weight of the hanging column of air in the tube compensated exactly for the weight of gas in the unsaturated zone profile, the readings represented the differences in pressure potential produced by barometric fluctuations. In borehole SD-9 and during the initial monitoring in borehole SD-7, pneumatic pressures were measured at the land surface in a manner similar to that for boreholes NRG#5 and SD-7. However, downhole pressures were measured through access tubes inserted into the annular spaces between a pair of nested casings and between the outer casing and formation wallrock, as described in Subsection 5.3.4.2.1.

In addition to visual inspection, the downhole pressure data were analyzed using cross-spectral estimation techniques (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Patterson et al. 1996) to describe the character of the time varying response of downhole pressure to changes in the synoptic barometric signal from the atmosphere. The power spectral density of barometric pressure and the transfer function from barometric pressure to downhole pressure were estimated using Fourier transform algorithms that detects narrowband frequency signals buried in wideband noise. The barometric signal is a broad frequency band pressure fluctuation that contains predominant semidiurnal, and longer period (synoptic) frequencies. The synoptic frequencies, associated with the passage of storm fronts, provided a longer term pressure fluctuation that generally resulted in a well developed downhole pressure response. For these analyses, the synoptic barometric frequency having the greatest power spectral density was selected and the residual amplitude and phase lag of that signal at each downhole sensor were calculated.

These cross-spectral analyses are helpful in comparing pressure responses between stations within individual boreholes. However, because the period of the synoptic pressure signal varies over time, some of the differences indicated by the analyses among different boreholes may be the result of differing time periods, data collection frequency, and lengths of the data record. Ideally these analyses would be performed on long continuous data sets representing one or two years of monitoring. However, because data sets of that length are not available for most of the boreholes discussed in this report, the use of these analyses is limited to events and conditions that can be supported by visual inspection of the record or other supporting information. The results of these analyses are discussed in Subsection 5.3.4.2.1.

5.3.2.1.4.2 In Situ Water Potential Measurements

In situ water potential has been monitored at multiple depths in instrumented boreholes NRG-6, NRG-7a, UZ#4, UZ#5, UZ-7a, and SD-12 (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) using thermocouple psychrometers. Water potential data were collected from these boreholes under a fully compliant quality assurance program for the same periods of time indicated in Table 5.3-7 for the pneumatic pressure data. In addition, nonqualified water potential data were collected from borehole UZ-1 from October 1983 through October 1985 using a prototype borehole instrumentation system. The thermocouple psychrometers measure the combined osmotic and matric potential, but not the gravitational potential, of liquid water contained within the pore spaces of the rock matrix and openings of fractures. Psychrometers measure liquid water potential in the vapor phase by sensing the relative humidity, or vapor pressure, of the formation rock gas that is in contact with water filled pores and fracture openings. The vapor concentration of the rock gas is a function of the capillary tension exerted on liquid water contained within the largest filled openings of the media. As saturation increases, larger and larger pores become filled, capillary tension is reduced, and the vapor concentration of the air in contact with the liquid water increases. Because a water potential of zero indicates that the medium is saturated, water potential is always less than or equal to zero. In this report, water potential is expressed in MPa; a megapascal is equal to 10 bars.

The thermocouple psychrometers were calibrated over a water potential range of −0.1 to −7.0 MPa (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164), which is equivalent to a relative humidity range of 99.9+ to 93 percent. Calibration accuracy of these sensors was on the order of ±0.2 MPa (95 percent confidence level) over the full range of calibration. The sensitivity, or resolution, of these sensors is on the order of 0.005 MPa. Psychrometer instrument stations in boreholes NRG-6, NRG-7a, UZ#4, UZ#5, UZ-7a, and SD-12 were backfilled with polyethylene beads and capped with a thin layer of 20/40 silica sand before placement of intervening grout isolation plugs. At UZ-1, instrument stations were backfilled with 12/16 silica sand and capped with dry bentonite and intervening layers of silica flour. Grout isolation plugs were placed at various locations throughout borehole UZ-1.

Polyethylene beads were used in all boreholes instrumented after UZ-1 to minimize the time needed for the water potential of the dry stemming materials in the instrument station to equilibrate with the ambient water potential of the surrounding formation (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Although the use of polyethylene beads greatly reduced equilibration times, it was apparent that other factors also influenced equilibration rates. Very rapid changes in water potential occurred early in the recovery cycle when diffusion gradients were large and small changes in vapor concentration caused large changes in relative humidity. Later in the recovery cycle, very slow, asymptotic changes in water potential occurred when diffusion gradients were small and small changes in vapor concentration produced progressively smaller changes in relative humidity at very

high vapor concentrations. Differences in recovery rates for various boreholes were attributed to:

- The volume of dry materials needed to backfill an instrument station
- The moisture retention and adsorption characteristics of the dry stemming materials
- The volumetric water content of the surrounding rocks for a given saturation and water potential
- The extent to which the borehole may have been disturbed by drilling air or by in-hole air circulation while the borehole was left open prior to instrumentation
- The presence or absence of fractures in communication with the instrument station

Further analyses and interpretations of borehole water potential data are discussed in Subsection 5.3.4.2.2 of this report.

5.3.2.1.4.3 In Situ Temperature Measurements

In situ temperature has been monitored at multiple depths in instrumented boreholes NRG-6, NRG-7a, UZ#4, UZ#5, UZ-7a, and SD-12 for the same periods of time indicated in Table 5.3-7 for the pneumatic pressure data (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997). Temperatures have been measured with thermistors originally calibrated against primary temperature standards to an accuracy of ±0.005 °C (95 percent confidence; Kume and Rousseau 1994), and with uncalibrated thermocouples (backup system) using thermocouple reference tables to convert voltage output to temperature). Measurement precision of the thermistors, with all data acquisition system components functioning properly, ranges between 0.001 to 0.003 °C. Thermocouple accuracy using the voltage output reference tables is estimated to be on the order of 0.1 to 0.5 °C. Measurement precision of the thermocouples ranges between 0.05 to 0.10 °C. In addition, temperatures were monitored with thermocouples in borehole UZ-1, which was instrumented in 1983, and in borehole NRG#4, which was instrumented with a packer system by Nye County in 1994. Temperature data from boreholes NRG-6, NRG-7a, UZ#4, and UZ#5 were collected under a fully compliant quality assurance program. Although useful as corroborative and supporting information, temperature data from UZ-1 and NRG#4 were not collected under a qualified quality assurance program.

Additional unsaturated-zone temperature data are available from 10 open boreholes in the vicinity of Yucca Mountain (Sass, Lachenbruch et al. 1988). As part of a regional heat flow study, multiple temperature profiles were measured between 1980 and 1984 for boreholes a#1, #4, #5, #6, #7, G-1, G-4, H-1, WT#4, and WT#18. Because temperature data from these boreholes were not collected under a compliant quality assurance program, the data are used as supporting information in this report. Analysis and interpretation of borehole temperature data are discussed in Subsection 5.3.4.2.3 of this report.

5.3.2.1.5 Collection of Water and Gas Samples From Surface Based Boreholes for Chemical and Isotopic Analysis

Water samples were collected from the unsaturated zone by extracting water from rock cores of selected boreholes and by bailing or pumping water from perched water zones encountered in boreholes at Yucca Mountain (Yang et al. 1996, 1998). Water samples were analyzed for both major ion concentrations and isotopic compositions. Boreholes from which water samples were collected are listed in Table 5.3-8. Borehole locations are shown in Figure 5.3-8.

Pore water was extracted from cores by using a high pressure, one-dimensional (uniaxial) compression procedure (Mower et al. 1991, 1994; Higgins et al. 1997). Briefly, compression operations start with applying pressure to a first stress level of 103.4 MPa at a rate of 69 kPa/s. Loading continues in eight increments of 103.4 MPa (at the same loading rate) until the final stress level of 827 MPa is reached. Additional pore water can be extracted by injecting dry nitrogen gas (greater than 99.99 percent pure). The representativeness and accuracy of water recovered by this method was tested by obtaining several aliquots of pore waters as a function of increasing pressure. This test showed that concentrations did not vary as a function of pressure, and therefore, compression does not appear to have altered the original pore water compositions through water-rock interaction during the extraction processes. An alternative method of water extraction using a high speed centrifuge was used to validate the compression method. A further test for reproducibility was accomplished by extracting water from adjacent samples of drill core from the same stratigraphic unit by both compression and centrifugation methods. Both methods yielded similar pore-water compositions in series of analyses (Yang et al. 1990).

Water samples were analyzed for isotopic content including tritium, ^{36}Cl, carbon isotopes, and stable isotopes of deuterium and oxygen-18. Over 700 ^{36}Cl measurements have been conducted thus far for the YMP (Fabryka-Martin, Wolfsberg et al. 1997; Fabryka-Martin, Turin et al. 1998; Levy et al. 1997), including samples of surface soils and soil profiles; rock samples collected from boreholes and within the Exploratory Studies Facility; water samples including surface runoff, unsaturated-zone pore water, perched water, and saturated-zone water; and fossilized urine from pack-rat middens.

Two methods of pore water extraction for stable isotope analysis of deuterium and oxygen-8 were used in this investigation: the vacuum distillation and the compression-extraction methods (Yang et al. 1998). The vacuum distillation technique employs the use of a distillation line, nitrogen gas, dry ice, a set of heat elements, and a vacuum pump. Vacuum distillation of pore waters for stable isotope analysis produced reliable results from tuffs free of clay and zeolite minerals but not for tuffs with large contents of hydrated minerals (Yu 1996). The compression-extractionmethod was similar to the procedures described previously for pore water extraction for chemical compositions and ^{14}C analyses (Mower et al. 1991, 1994; Higgins et al. 1997). Water samples were collected in glass syringes attached to the compression cell. After extraction, the water was filtered through a 0.45 μm filter into glass bottles for stable isotope analysis by mass spectrometer. The compression-extraction process only extracts pore water whereas vacuum distillation extracts both pore water and water held by zeolites or clays.

Gas samples for analysis of CO_2 concentration and carbon isotope content have been collected from permanently instrumented boreholes (UZ-1, SD-12), from open boreholes (NRG-6, NRG-7a, SD-9), and from boreholes instrumented temporarily with borehole liners (SD-7, UZ-14). The CO_2 gas borehole was collected after overnight pumping to remove atmospheric air. In addition, gases have been collected from radial bore holes drilled from alcoves driven from the Main Drift of the Exploratory Studies Facility. Sulfur hexafluoride, SF_6, was used in each of these bore holes as a gaseous tracer in order to identify drilling-air contamination of the rock gas. Based on 10 years of records and the stabilized value in bore hole UZ-1, the tracer (SF_6) concentration of less than 0.1 ppm in the borehole air is used to define uncontaminated rock gas. Gas samples were collected using a 500 cm^3 per-minute peristaltic pump connected through short silicone tubing within hole gas-sampling tubes, as described by Yang et al. (1996). Sampling tubes were pumped overnight in order to purge any atmospheric air. Two different methods were used to collect CO_2 for isotopic analysis: a molecular sieve method and a whole-gas method as described by Yang et al. (1996). Carbon-13 was measured by mass spectrometry with a precision of about ±0.2 per mil. Carbon-14 analysis was accomplished by counting in a gas proportional counter with an uncertainty of ±0.7 percent modern carbon.

Interpretations of results of chemical and isotopic analysis of unsaturated-zone water and gas are described in Subsection 5.3.4.2.4 of this report.

5.3.2.1.6 Testing and Monitoring in the Exploratory Studies Facility

Pneumatic monitoring, air-injection testing, and hydrochemical sampling were conducted in test alcoves and the primary tunnels of the Exploratory Studies Facility (Figure 5.3-16) as the primary Exploratory Studies Facility loop was constructed and following its completion. Although initial studies in some of these alcoves are complete, studies in several other alcoves are continuing. Therefore, only results from studies in the first three alcoves are included in this report. These alcoves include the upper Tiva Canyon alcove, the Bow Ridge fault alcove, and the upper Paintbrush contact alcove. Results of studies in other alcoves, including those constructed across the Ghost Dance fault in the Topopah Spring Tuff, will be included in future updates of this report.

5.3.2.1.6.1 Pneumatic Monitoring and Air-Injection and Gaseous Tracer Tests

Pneumatic monitoring and air-injection and gaseous tracer tests have been conducted in horizontal boreholes drilled from alcoves of the Exploratory Studies Facility. The tests in the Exploratory Studies Facility differ from the surface based borehole tests in that both single-hole and cross-hole air-injection tests were conducted. In addition, because clusters of closely spaced boreholes were drilled from the Exploratory Studies Facility alcoves, gaseous tracer tests also were conducted. To date, air-injection tests have been completed in boreholes located in the upper Tiva Canyon alcove, the Bow Ridge fault alcove, and the upper Paintbrush contact alcove. These tests are described in Subsection 5.3.3.1.2.2 of this report. Air-injection testing is ongoing in the Northern and Southern Ghost Dance fault alcoves and is planned for future alcoves and drifts to be constructed from the Exploratory Studies Facility. These tests will be described in future updates of this report.

Air-injection tests in the Exploratory Studies Facility were conducted in a manner similar to the surface based tests except that the Exploratory Studies Facility boreholes were smaller in diameter (about 6 cm), shallower (about 30 m), and horizontal (LeCain et al. 1997; LeCain 1998). After test intervals were selected, pneumatic packers connected by plastic pipe were installed in the horizontal borehole. Because the Exploratory Studies Facility tests were intended to characterize the rock mass at a smaller scale (5 to 30 m) than the surface based tests, the Exploratory Studies Facility testing equipment allowed for isolation of as many as 10 intervals in each horizontal borehole. In the single-hole tests, one interval at a time was isolated by inflating the packers with compressed air and then injecting compressed air tagged with sulfur hexafluoride tracer gas into the interval at a constant rate until the pressure stabilized. In the cross-hole tests, air was injected into one interval of one hole and pressure changes were monitored in up to 10 intervals in a nearby hole.

Pneumatic monitoring and air-injection and gaseous tracer tests were conducted in the upper Tiva Canyon alcove, Bow Ridge fault alcove, and upper Paintbrush contact alcove of the Exploratory Studies Facility (Figure 5.3-16) from November 1994 through January 1997 to determine air permeability, pneumatic porosity, and effective porosity of rock units (LeCain 1998). Single-hole and cross-hole air-injection tests were conducted using a straddle-packer assembly in horizontal boreholes drilled from the Exploratory Studies Facility alcoves. Test interval lengths ranged from 1 to 3 m and gas injection rates ranged from 10 to 800 standard liters per minute. Testing was conducted at incrementally increasing injection pressures to identify water redistribution. Injection interval differential pressures were up to 110.0 kPa.

In the upper Tiva Canyon alcove, pneumatic monitoring was conducted in the Tiva Canyon Tuff crystal-poor, upper lithophysal zone to evaluate the pneumatic pressure responses to barometric pressure changes and to estimate bulk pneumatic diffusivity of the overlying rocks (LeCain et al. 1997). Analysis of the pneumatic pressure responses also was used to assess the effectiveness of the packer system and borehole seals and to qualitatively determine the presence of zones of high-permeability or high-fracture connectivity. Pneumatic monitoring was conducted in three, subhorizontal radial boreholes (RBT#1, RBT#2, and RBT#3) during November and December 1994 and during April 1995. The 1994 monitoring measured borehole air pressures throughout the open length of the individual boreholes. This monitoring revealed no differential air pressure between boreholes nor between the boreholes and barometric pressure. The 1995 monitoring used downhole pneumatic packers to isolate up to nine separate intervals per borehole, but also indicated that borehole pressures were essentially indistinguishable from the barometric pressure. Single-hole air-injection tests were conducted during August 1995 in the same boreholes that penetrate the Tiva Canyon crystal-poor, upper lithophysal lithostratigraphic unit. Single-hole tests were conducted in 11 intervals of borehole RBT#1, 18 intervals of RBT#2, and 12 intervals of RBT#3. Cross-hole air-injection tests were conducted in the same boreholes in the upper Tiva Canyon alcove during May through July 1995.

In the Bow Ridge fault alcove, pneumatic monitoring was conducted in borehole HPF#1 during November 1995 and in HPF#2 during May 1996 (LeCain et al. 1997). The pressure data indicated no significant, analyzable differential between barometric pressure and pressure in the boreholes, indicating that the Tiva Canyon Tuff, Bow Ridge fault breccia, and pre-Rainier Mesa Tuff all have sufficient air permeability to allow essentially instantaneous propagation of the barometric pressure,

either from the land surface or from the walls of the alcove or Exploratory Studies Facility. Air-injection tests were conducted in boreholes HPF#1 and HPF#2, which are located in the Tiva Canyon crystal-poor, middle nonlithophysal zone on the east side (footwall) of the Bow Ridge fault. Single-hole tests were conducted in 19 intervals of borehole HPF#1 in December 1995 through January 1996 and cross-hole tests were conducted between boreholes HPF#1 and HPF#2 during June and July 1996. A total of 13 cross-hole tests were conducted with HPF #1 as the injection borehole and HPF #2 as the monitor borehole. Six convergent, cross-hole, gaseous tracer tests were conducted during June and July 1996 in the Bow Ridge fault alcove between boreholes HPF#1 and HPF#2 using SF_6 tracer gas.

In the upper Paintbrush contact alcove, pneumatic monitoring was conducted in subhorizontal boreholes RBT#1 and RBT#4 (LeCain and Patterson et al. 1997). The results indicated that there was virtually no attenuation or lag of the barometric pressure signal and that the rocks penetrated by both boreholes had sufficient permeability to air to allow the instantaneous response, but it was impossible to determine whether the pressure propagation was from the surface or from the wall of the alcove or from Exploratory Studies Facility North Ramp. Single-hole air-injection tests were conducted during April through June 1996 in boreholes RBT#1 and RBT#4 that penetrate the moderately welded Tiva Canyon crystal-poor, lower nonlithophysal columnar unit, which is very near the base of the Tiva Canyon welded hydrogeologic unit and just above the top of the PTn. Single-hole tests were conducted in 14 intervals of borehole RBT#1 and 15 intervals of RBT#4. Analysis and interpretation of all of the Exploratory Studies Facility air-injection and gaseous-tracer tests are discussed in Subsection 5.3.3.1.2 of this report.

5.3.2.1.6.2 Hydrochemical Data Collection

Hydrochemical investigations involved collection of gas and pore water samples and analysis for gas composition, ^{36}Cl, tritium, carbon isotopes, chloride, and major ions. The hydrochemical investigations in the first three alcoves of the Exploratory Studies Facility were intended to provide a better understanding of gas flow in the upper part of the unsaturated zone; provide evidence of gas flow direction, gas flux, and travel time; support conceptual models of fluid flow in the unsaturated zone. In the upper Tiva Canyon alcove, gas samples were collected from boreholes RBT#1, RBT#2, and RBT#3 and analyzed for CO_2 concentration and carbon isotopes to determine the source and age of gas in the Tiva Canyon Tuff, help assess pneumatic pathways and gas phase circulation, and determine the degree of fracture connectivity in the rocks (LeCain et al. 1997). Prior sample collection, the boreholes were pumped with peristaltic pumps for several days to remove drilling air and atmospheric air that may have entered the borehole. During pumping, concentrations of CO_2 and the sulfur hexafluoride (SF_6) tracer gas injected during drilling were monitored until they stabilized, indicating that the gas being pumped from the borehole contained little or no drilling air and was composed primarily of rock gas. In general, stable CO_2 concentrations, at levels significantly higher than atmospheric (350 ppm) or alcove (450 ppm) concentrations, were an indication that the gas being pumped from the borehole was representative of rock gas. Gas samples for CO_2 and SF_6 concentrations were collected in glass syringes with three-way stopcocks so that the syringes could be filled and purged a minimum of three times before the actual sample was obtained. Three duplicate samples were collected from each borehole interval and then transported to the laboratory for analysis by gas chromatograph. Gas samples for ^{14}C analysis were collected

by pumping the borehole gas through molecular sieves designed to collect the CO_2. Gas samples for ^{13}C analysis were collected in Mylar balloons.

In the Bow Ridge fault alcove, gas samples were collected from borehole HPF#1 during November and December 1995 and from borehole HPF#2 during May 1996 (LeCain et al. 1997). Gas samples were obtained using the same methods as described above for the Upper Paintbrush Contact alcove. Gas samples from the fault zone were analyzed for CO_2 concentration and ^{14}C to provide estimates of the age of the gas in the fault zone. In addition, samples of pore water were obtained from the cores of boreholes HPF#1 and HPF#2 by vacuum distillation and analyzed for tritium to assess the rate of infiltration of water from the land surface.

In the upper Paintbrush contact alcove, gas samples were obtained from boreholes RBT#1 and RBT#4 after the completion of pneumatic pressure monitoring (LeCain and Patterson et al. 1997). Concentrations of CO_2 and the sulfur hexafluoride tracer gas were monitored while gas was pumped from the borehole until they stabilized. Gas samples then were collected for ^{13}C and ^{14}C analysis from six intervals of borehole RBT#1 and five intervals of borehole RBT#4. Three samples of pore water were obtained by vacuum distillation from the core of borehole RBT#1 and seven were obtained from the core of RBT#4. These water samples were analyzed for tritium to provide an indication of the spatial distribution of percolation flux near the PTn-TSw contact. To date, only two rock samples from the Exploratory Studies Facility North Ramp have provided sufficient water for ^{14}C age estimates, both from nonwelded tuff in the upper part of the Paintbrush nonwelded hydrogeologic unit. One sample was obtained from core from the 21.9 to 22.6 m interval of borehole RBT#4, and the other was obtained from a 1 cubic foot block of rock excavated from the wall of the Exploratory Studies Facility North Ramp in a moist zone near the eventual location of the upper Paintbrush contact alcove (LeCain and Patterson et al. 1997).

The results of these hydrochemical investigations are discussed in Subsection 5.3.4.2.4 of this report.

5.3.2.2 Saturated-Zone Data Collection

5.3.2.2.1 Water Level Monitoring Program

Drilling for site characterization at Yucca Mountain began in 1978, but the first hydrologic test well was completed in 1981 (J.H. Robison [U.S. Geological Survey], written communication to D.L. Vieth [U.S. Department of Energy/Nevada Operations Office], September 17, 1986 [regarding revisions of Yucca Mountain water levels reported in Robison, 1984], TIC 219488). Although water levels were measured as each well was completed, water level monitoring, in terms of a long-term water level network, did not begin until the initiation of regularly scheduled periodic water level measurements during 1983 (J.H. Robison [U.S. Geological Survey], written communication to D.L. Vieth [U.S. Department of Energy/Nevada Operations Office], September 17, 1986 [regarding revisions of Yucca Mountain water levels reported in Robison, 1984], TIC 219488, p. 4). Periodic measuring of water levels continues through 1997. Measuring of water levels hourly (continuous measuring) in support of the water level network began in 1985 (Luckey, Lobmeyer et al. 1993, p. 4), but was discontinued during 1996.

Periodic measurement of water levels was generally conducted monthly, using either calibrated steel tapes or electronic cable units (J.H. Robison [U.S. Geological Survey], written communication to D.L. Vieth [U.S. Department of Energy/Nevada Operations Office], September 17, 1986 [regarding revisions of Yucca Mountain water levels reported in Robison, 1984], TIC 219488, pp. 9-11). The continuous measurement of water levels was accomplished using pressure transducers, with water level data collected every hour and stored into and retrieved from electronic data loggers (Lobmeyer, Luckey et al. 1995, p. 12). Though referred to as continuous measuring of water level data, data loggers were programmed to receive transducer pressure only hourly. This frequency of data collection is sufficient to document daily, monthly, and yearly water level changes and trends. However, to detect water level fluctuations due to seismic events, it is necessary to instrument wells with data loggers that record all transducer pressure changes. Subsequently, several of the wells instrumented with a transducer were equipped with analog chart recorders so continuous pressure change could be recorded on an analog strip chart. Instrumenting of wells with the analog chart recorders began in 1992 and was discontinued in 1996.

Since 1981, water-level data in the Yucca Mountain area has been collected and reported for as many as 33 wells monitoring 41 depth intervals (Figure 5.3-17, Table 5.3-9) (several wells have more than one depth interval monitored, these intervals are isolated by either an inflatable packer or by cement plugs). The frequency and methods of monitoring each well and the water-level data collected for these wells is reported in Robison et al. (J.H. Robison [U.S. Geological Survey], written communication to D.L. Vieth [U.S. Department of Energy/Nevada Operations Office], September 17, 1986 [regarding revisions of Yucca Mountain water levels reported in Robison, 1984], TIC 219488), Gemmell (1990), O'Brien (1991), Luckey, Lobmeyer et al. (1993), Boucher (1994a, 1994b), O'Brien et al. (1995), Lobmeyer, Luckey et al. (1995), Tucci, O'Brien et al. (1996), Tucci, Goemaat et al. (1996), Graves et al. (1996, 1997), and Graves and Goemaat (1998). These reports also contain detailed descriptions of well location, wellhead elevation, well construction and completion, geologic units each well is completed in, and quality assurance considerations for water-level data collection

5.3.2.2.2 Saturated-Zone Hydraulic Tests and Analytical Procedures

Knowledge of hydraulic properties, such as hydraulic conductivity, transmissivity, and storativity, is critical to understanding the hydrogeology of Yucca Mountain. Several methods have been used to conduct and analyze tests to determine the hydraulic properties of the saturated zone in the Yucca Mountain area. Field tests conducted to determine transmissivity and storativity include single-well, constant rate discharge tests (commonly known as pumping tests); multiple well pumping tests, in which observation wells are used; constant rate injection tests in both single-well and multiple-well configurations; single-well, variable rate discharge tests; single-well, slug-injection and slug-withdrawal tests; and single-well, pressure-injection tests. Borehole flow and temperature surveys during pumping, described below, provided information on the vertical distribution of the hydraulic properties. Table 5.3-10 lists the types of hydraulic tests that have been conducted in the vicinity of Yucca Mountain, the analytical methods used to evaluate the tests, and references that describe the analytical methods. The actual applications of these methods to investigations at Yucca Mountain are described in Craig and Robison (1984), Craig and Reed (1991), Geldon (1996), Geldon et al. (1997, 1998), Lahoud et al. (1984), Lobmeyer (1986), Moench (1984), O'Brien (1997, 1998),

Robison and Craig (1991), Rush et al. (1984), Thordarson (1983), Thordarson et al. (1985), and Whitfield, Eshom et al. (1985). The saturated-zone tests discussed in those reports have been completed in 17 boreholes, and 3 additional holes have been used as observation wells during multiple-well tests. The locations and results of the field tests are discussed in Subsection 5.3.3.2.2.

Because of the great depth that must be drilled to test the saturated zone at Yucca Mountain, almost all of the hydraulic tests were single-well tests of the entire open interval or of specific depth intervals in a borehole. Multiple-well tests have been conducted mainly at one site, the "C-hole" or UE-25c complex, but tests of well pairs were conducted also in Drill Hole Wash and Jackass Flats. A single-well pumping test provides an estimate of the transmissivity, the integrated water conducting capability of the entire saturated section to which the borehole is open. The principal limitations are that:

- No information regarding the vertical distribution of permeability or the degree of anisotropy are obtained directly, although a flow survey mitigates the former.
- Head losses close to the pumping well, due to flow convergence and plugging of openings near the borehole during drilling, are a significant part, sometimes the major part, of the drawdown, causing transmissivity to be underestimated and diminishing the sensitivity of drawdown to hydraulic properties of the more distant reaches of the affected rock mass.
- Storativity cannot be determined reliably from pumped-well drawdown data.

An observation well eliminates the well-loss factor, providing a more reliable drawdown value, though only in the direction connecting the pumping and observation wells. Multiple observation wells located on intersecting axes give a more accurate record of the directional aspects of drawdown, allowing evaluation of anisotropy as well as the transmissivity and storativity where hydrogeologic conditions are favorable.

Slug tests are single-well tests that involve injection or withdrawal of small volumes of water at decreasing rates in relatively short periods of time. Consequently, concurrent flow surveys are impractical, no directional information is gained, and a very limited volume around the borehole is tested. Their usefulness is in revealing limited information on the transmissive and storage properties of less permeable intervals, to which pumping tests are insensitive, even when supplemented with flow surveys.

Pumping Tests–A pumping test, whether single-well or multiple-well, requires a submersible pump set far enough below the static water level to keep the pump submerged during the test, a flow meter to measure pump discharge rate and/or volume, and pressure transducers set at one or more intervals within the borehole and connected to digital recorders at the surface to record changes in hydraulic head during pumping. An initial test usually is required to select an appropriate pumping rate and, commonly, the first pump must be removed and adjusted or replaced. If a flow or temperature survey is to be conducted during pumping, tubing of sufficient diameter to provide access of the survey tool around the pump is usually provided. Boreholes that are planned for pumping tests,

therefore, are drilled and completed at larger diameters (about 25 cm or larger) than are geologic core holes.

Flow and Temperature Surveys–Borehole flow surveys are commonly conducted in conjunction with constant rate pumping tests at Yucca Mountain to detect transmissive intervals that provide inflow to the borehole and to determine the relative contributions of each interval to the total production. The procedure is merely to conduct at various depths a fluid velocity measurement in the borehole, which then is combined with the cross sectional area of flow to provide uphole (or downhole) discharge at the depth of measurement. Most flow surveys used a tracer injection and detection technique (commonly known as a "tracejector" survey), but heat-pulse flowmeter surveys, spinner surveys, and oxygen activation surveys also have been used. The percentage of flow contributed by each test interval is calculated by dividing the test-interval flow rate by the measured pumping rate. Static or non-pumping flow surveys have also been conducted to determine intraborehole flow driven by ambient differences of hydraulic head at different depths in the hole. More complete descriptions of the techniques for flow surveys are provided by Blankennagel (1967), Keys (1988), Hess (1990), and McKeon et al. (1990).

Temperature logs were obtained in many of the boreholes at the same time as the borehole flow surveys. A temperature sensor measures fluid temperature as the probe is moved vertically along the borehole. Inflections in the temperature profile indicate flow of water into or out of the borehole, providing additional information concerning transmissive zones. Blankennagel (1967) and Keys (1988) describe temperature logging and interpretive techniques. Sass, Lachenbruch et al. (1988) describe precise temperature logging after boreholes have been allowed to equilibrate for weeks, months, or more after construction and testing disturbances. These logs are used to establish the natural thermal gradient and conductive heat flow as well as perturbations of the groundwater system by the continued presence of the borehole.

Packer Tests–Inflatable packers are used to isolate specific intervals within a borehole for testing and to measure head at various depths. Usually two packers ("straddle packers") were used to isolate intervals during slug injection tests at Yucca Mountain. In some constant discharge tests one packer was used above or below the pump to isolate the pumping interval from other parts of the borehole. Transducers generally are set within the isolated interval(s) to monitor the hydraulic-head response of the interval to the applied stress. Measurements of static head in packer-isolated intervals, which help to define the potential for natural flow in the groundwater system, are made with a steel tape or wireline probe.

Packers for conducting slug injection tests are set at a fixed interval on subs within a tubing string. They are inflated by injecting water through a valve system and, once they are inflated, the tubing can be manipulated to open ports into the borehole above, between, or below the packers. For slug tests, the tubing is filled to the desired initial head from a water source at the surface while the injection ports are closed. Upon opening a port to the borehole, usually between packers, the head in the tubing is instantaneously applied to the borehole wall and is allowed to decline as water is injected into the test interval. The change in hydraulic head is sensed with transducers and recorded at the surface. Except for the option of running a test below the straddle packers, the packer interval

is fixed but can be moved to various depths. Caliper logs are examined to select in-gauge seats for the packers to avoid leakage as well as to prevent rupturing the packers by overinflation.

The falling head, slug injection test can be analyzed by type-curve methods to obtain both transmissivity and storativity of the interval (Table 5.3-10). However, the volume of water actually injected is limited to the tubing storage between the level of the initial applied head and the static head in the interval tested, so the volume of rock that responds to the stress is small. In addition, the test should not be applied in transmissive intervals because the resistance to flow in the valve system and tubing becomes limiting. Consequently, flow surveys during earlier, open hole pumping are very useful for delineating the intervals that can be tested effectively by slug tests. An additional consideration in using slug tests where the water table is deep is that fractures can be dilated by the application of large differential heads, and, if the initial head is large enough, hydraulic fracturing may also be accomplished. Whether new fractures were induced in tests at Yucca Mountain is in doubt, but fracture dilation and subsequent closure is evident in the recovery curves for many of the tests (see, for example, Thordarson et al. 1985).

Pressure injection tests were conducted in formations that yielded very little water in response to stresses. In these tests, water pressure is applied suddenly to the tested interval, as with slug tests, but a valve is closed in the injection line, isolating the test interval from atmospheric pressure. The decay of head in the pressurized system is monitored using transducers. The cycle may be repeated several times.

Swabbing Tests–Water can be withdrawn from deep boreholes without a pump by swabbing. Swabbing tests are conducted by lowering swab cups below the water level inside internal-flush tubing or drill pipe, either on rods or on wireline with a weighted sinker bar. The swab cups, which expand to fit the drill stem or tubing, and the water column in the tube above the swab are then raised. The process is repeated until the desired volume of water has been lifted to discharge at the surface. Changes in hydraulic head in response to the removal of water are monitored using pressure transducers. Because the withdrawal rate is irregular, hydraulic analyses are usually based on specific capacity or on the recovery curve. Swabbing can be done from an open hole or from packer-isolated intervals, in which case the tubing used to set the packers is also used for swabbing. A rising head, slug withdrawal test can be initiated by pulling a single swab and is analyzed by the same methods as a slug injection test.

Multiple Well Tests–At the C-hole Complex, multiple well, constant discharge (pumping) tests were conducted in March, May, and October to December 1984 (Geldon 1996); May and June 1995; February 1996, and from May 1996 to November 1997. In these tests, a borehole or an isolated interval within a borehole was pumped, and hydraulic responses in nearby observation wells were monitored. In preliminary hydraulic tests conducted at the C-hole Complex in 1984, shortly after completion of drilling, wells UE-25c #2 and UE-25c #3 were pumped and monitored in a variety of configurations, using several different testing methods. A detailed description of each of the 1984 tests, as well as interpretations of test results, is provided by Geldon (1996). The hydraulic test conducted in May 1995 involved pumping UE-25c #3 and observing drawdown and recovery in UE-25c #1 and UE-25c #2; as no packers were used, the pumping and observation wells were open to the composite saturated sections. The hydraulic tests conducted in June 1995 involved pumping

UE-25c #3, again without packers installed, and observing drawdown and recovery in six isolated intervals in UE-25c #1 and UE-25c #2. Geldon et al. (1998) describe the analysis of the May and June 1995 tests.

In February 1996, an interval including the lower Bullfrog Tuff and upper Tram Tuff was isolated in UE-25c #3 and pumped 5 days to establish a gradient for a tracer test, and the Calico Hills, Prow Pass, upper Bullfrog, and Bullfrog-Tram intervals in UE-25c #1 and UE-25c #2 were monitored. In the 1996-1997 testing the lower Bullfrog Tuff interval was isolated and pumped in UE-25c #3, and the Calico Hills, Prow Pass, and upper and lower Bullfrog intervals in UE-25c #1 and UE-25c #2 were monitored. Five additional wells (UE-25 ONC-1, USW H-4, UE-25 WT#3, UE-25 WT#14, and UE-25p #1) were monitored during this 18-month long test to evaluate heterogeneity and scale effects in the volcanic rocks and to determine any possible hydraulic connection between the volcanics and the underlying Paleozoic carbonate rocks. Additional details of the hydraulic tests conducted at the C-hole Complex during 1995-1997 are provided by Geldon et al. (1997).

Assumptions made concerning the hydraulics of the aquifer system at and in the vicinity of the test borehole, and the availability and configuration of any nearby observation wells, determine the methods used to analyze the tests. The several methods used to analyze the multiple-well tests are listed in Table 5.3-10.

5.3.2.3 Geophysical Logging

Throughout the YMP activities, geophysical logging has been performed in conjunction with borehole development and geologic and hydrologic analysis. The primary use of these logs has been for correlation and stratigraphic studies, but they have also been used for evaluation of hydrologic and physical rock properties. A compilation of the geophysical surveys performed in the site boreholes and the logging tools used is included as Tables 5.3-11 and 5.3-12.

Detailed histories for geophysical logging at each well are reported by Fenix & Scisson, Inc. (1986a, 1986b, 1986c, 1986d, 1987a, 1987b, 1987c). Nelson et al. (1991) present geophysical log data, obtained prior to 1990, for 40 boreholes at Yucca Mountain. Nelson (1996a, 1996b) interpreted these logs to re-evaluate hydraulic and physical rock properties for stratigraphic units at Yucca Mountain, discussed in Subsection 5.3.3.2.1. Spengler (1995) compiled a comprehensive database for fractures identified by acoustic televiewer logs. The results of the fracture attribute studies are discussed in Subsection 5.6.3.3. Flow surveys are often performed with geophysical logging tools (such as spinner, trace ejector, or heat pulse flow meter probes), and the results of those surveys are summarized in Subsection 5.6.5.

Geophysical logging at Yucca Mountain has been conducted by various private well-logging service contractors and by YMP participants. The geophysical logging tools were used to evaluate borehole construction and integrity, lithology, stratigraphic unit contacts, geologic structure, porosity, thermal properties, fluid flow, and water bearing zones. Detailed descriptions of most of the logging tools

used at Yucca Mountain are found in Nelson et al. (1991) and Muller and Kibler (1986). The most commonly used logging tools at Yucca Mountain include:

- Caliper tools, used to measure drill hole diameter
- Dielectric tools, used to measure dielectric permittivity, and indirectly to estimate porosity and water content
- Neutron tools, used primarily to estimate porosity and water content
- Density tools (including borehole gravity), used to measure rock density, and indirectly to estimate porosity and water content
- Velocity tools, used for determination of formation velocity, and indirectly porosity, water content, and thermal conductivity
- Resistivity tools, used to measure formation electrical resistivity
- Gamma-ray tools, used to measure gamma radiation emitted by a formation (primarily used for stratigraphic correlation)
- Flow meter tools, used to determine the distribution of the water flow rate within a borehole
- Temperature tools, used to measure ambient borehole temperature
- Acoustic televiewer tools, used to identify the location, orientation, and relative size of fractures, and the location and orientation of bedding
- Television cameras that are also used to identify the location, orientation, and relative size of fractures
- Borehole deviation tools, used to measure deviation from vertical of the borehole and to provide a depth-correction factor for borehole measurements

Quality Assurance Considerations–Because the majority of the test wells were drilled prior to implementation of an approved quality assurance program, the results and interpretations of these logs were originally considered non-qualified. By letter dated August 29, 1995, the DOE/YMSCO Assistant Manager for Licensing authorized qualification of 766 logs from forty boreholes, based on the recommendation of the TRW Environmental Safety Systems, Inc. Technical Assessment Team for Borehole Geophysical Data. The Technical Assessment Team report to the Technical Project Office provides a qualification checklist for all geophysical log information included in this subsection.

5.3.3 Properties of Hydrogeologic Units

The hydrogeologic system at the Yucca Mountain Site consists of a sequence of Tertiary and older rocks that includes an unsaturated zone as much as 750 m (2,500 feet) thick, underlain by a saturated zone that includes as much as 1,500 m (4,570 feet) of Tertiary rocks and probably a greater thickness of transmissive Paleozoic rocks. In general, the unsaturated zone may include Quaternary and Tertiary alluvium and colluvium deposits that are underlain by Tertiary welded and nonwelded volcanic rocks of the Timber Mountain, Paintbrush, and Crater Flat Groups, including the Rainier Mesa Tuff, the Tiva Canyon Tuff, the Topopah Springs Tuff, the Calico Hills Formation, and the Prow Pass Tuff (Buesch, Nelson et al. 1996; Buesch, Spengler et al. 1996; Moyer and Geslin 1995). Based on hydrogeologic properties, the unsaturated zone at Yucca Mountain has been defined as consisting of six principal hydrogeologic units: unconsolidated alluvium, TCw, Paintbrush nonwelded, Topopah Spring welded, Calico Hills nonwelded, and Crater Flat (Montazer and Wilson 1984).

The dip of the Cenozoic strata and the transitional nature in hydraulic and fracture properties of stratigraphic units are evidenced by the fact that many of the strata included in the unsaturated zone also comprise the saturated-zone system. Nomenclature for the rocks comprising the saturated-zone groundwater flow system is based largely on geologic work by Scott and Bonk (1984). Thermal-mechanical and some hydraulic properties have been described by Ortiz et al. (1985). The saturated-zone hydrologic units at Yucca Mountain described by Luckey, Tucci et al. (1996) include the lower, densely welded member of the Topopah Spring Tuff and the underlying Calico Hills Formation and Crater Flat Group. The Tertiary volcanic sequence is subdivided into two aquifers and two confining units, and an underlying Paleozoic sequence comprises an upper confining unit and a lower carbonate aquifer. These saturated-zone designations vary somewhat, but are correlative to the hydrologic units first described by Winograd and Thordarson (1975) who characterized the hydrology and hydrochemistry of the groundwater flow system in the Nevada Test Site region, which encompasses Yucca Mountain. More recently, Faunt et al. (1997) have determined the definition of the Tertiary saturated zone system described by Laczniak et al. (1996) may more accurately describe the site saturated zone. Laczniak et al. (1996) have characterized the Tertiary saturated-zone system at the Nevada Test Site and vicinity, including Yucca Mountain, to include three volcanic aquifers. As in the work cited above, the lowermost Topopah Springs Tuff comprises the upper volcanic aquifer. The Crater Flat Group is considered a middle volcanic aquifer. As in preceding work, the underlying flow breccia and Lithic Ridge Tuff act as a confining unit. However, the sequence of variably welded ashflow tuffs and rhyolite lavas underlying this confining strata comprise a Tertiary-age lower volcanic aquifer.

Recent work (Buesch, Nelson et al. 1996; Moyer and Geslin 1995) provides new, detailed analyses of the Tertiary-age Paintbrush Group, Calico Hills Formation, and the Prow Pass Formation of the Crater Flat Group strata that may enhance interpretation of geologic controls on the site hydrologic system. Notwithstanding, hydrogeologic unit properties will be discussed in the context of the principal stratigraphic formations except where reference to the lithostratigraphic unit terminology developed by Buesch, Nelson et al. (1996) and Moyer and Geslin (1995) is necessary to identify unique or anomalous field and core data and enhance interpretation of flux through the site

hydrologic system. Correlation between the geologic, hydrologic, thermal-mechanical, and lithostratigraphic unit designations is shown in Figure 5.3-18.

5.3.3.1 Unsaturated Zone

For regional investigations in the vicinity of Yucca Mountain, definition of hydrogeologic units at a level roughly equivalent to geologic formations (Figure 5.3-18) generally is adequate. However, at the site scale it is necessary to define hydrogeologic units in much greater detail to account for important variations in properties. (Site-scale investigations of the unsaturated zone are those that focus on the rock units at and above the water table within the boundaries of the site-scale unsaturated-zone flow model (see Subsection 5.3.4.4). The vertical stratification and specific sequencing of units, as well as transitions between units, have large effects on the spatial variation of percolation flux and distribution of water through the units. Therefore, the manner in which rocks are designated as hydrogeologic units can greatly influence numerical-model representations of hydrologic conditions in the unsaturated zone.

In the early years of the YMP, hydrogeologic unit designations for the unsaturated zone encompassed large rock-unit divisions based primarily on the degree of welding (Scott et al. 1983; Montazer and Wilson 1984). In addition to welding, this system of identification also recognized other characteristics potentially important to hydrologic properties, such as occurrence of lithophysal cavities, fracturing, bedding, and mineral alteration. For example, the Tiva Canyon and Topopah Spring Tuffs range from nonwelded at the very top and bottom of each unit to densely welded in the interior with alternating lithophysal and nonlithophysal zones. Although the densely welded tuffs are highly fractured, the matrix generally has low porosity and permeability. Nonwelded tuffs have higher porosity and permeability but are sparsely fractured. Bedded ashfall tuffs are interlayered between the major pyroclastic flow tuffs and have the highest porosity and permeability of any of the rock units. Although the Calico Hills Formation is nonwelded and generally has high porosity, it has low permeability due to zeolitic alteration over most of the site area. The Prow Pass Tuff of the Crater Flat Group ranges from welded to nonwelded and exhibits various degrees of mineral alteration. The Bullfrog Tuff is welded everywhere it has been sampled in boreholes.

5.3.3.1.1 Characterization of Hydrogeologic Units Using Core-Scale Matrix Properties

To define detailed hydrogeologic units (and associated physical and hydrologic properties) adequate for 3-D flow modeling of the unsaturated zone at Yucca Mountain, an intensive study of matrix properties was performed (Flint, L.E. 1998) using core samples from eight deep and 23 shallow boreholes (see Tables 5.3-5 and 5.3-6 in Subsection 5.3.2.1.2). The resulting detailed hydrogeologic units are listed in Table 5.3-13 along with formal and informal lithostratigraphic nomenclature and the major hydrogeologic units. A major goal in this study was defining detailed hydrogeologic unit boundaries so that they corresponded to lithostratigraphic boundaries. This would facilitate the spatial distribution of properties within the layering scheme for any numerical modeling effort that used the 3-D lithostratigraphic framework model developed by Buesch, Spengler et al. (1996). To provide reasonable detail and predictive ability within the magnitude and heterogeneity of the Yucca Mountain Site, and to allow for detailed process modeling as well as large-scale 3-D modeling, the number of distinct vertical layers was limited to 30. The relations between modeling parameters and

porosity also were described to provide the means by which even more detailed process models could be constructed, for example, through scaling of parameters in transitional zones where properties change dramatically over short distances, or in clay-altered or vapor-phase corroded zones. (Note: In this report, the term "porosity" refers to that determined from the 105°C oven-dry weights unless otherwise noted.) The approach of Rautman (1995) for correlating porosity with lithologic features and other hydrologic properties was used for a site-scale model that spatially distributed porosity, saturated hydraulic conductivity, and thermal conductivity throughout the modeling domain. This approach is discussed in detail later in this subsection.

The distributions of physical properties and hydrologic parameters at the Yucca Mountain Site are complex because of the depositional, cooling, and alteration history of the rocks. This history results in differences in porosity, connectivity and tortuosity of flowpaths, water-retention character, vertical heterogeneities, and scales of features that all influence the resulting hydrology of the unsaturated zone at the site. These characteristics are discussed in detail in various subsections of this report. For example, there are various rock types that result from combining the type of deposition, amount of welding, presence of glass, and composition of grains, which may have crystallized at high temperature (quartz, feldspar, and vapor-phase minerals), or diagenetically altered (zeolites and clays). Data are presented and described within the context of these rock (lithologic) types. In several locations, horizontal boundaries between lithostratigraphic and hydrogeologic units are very abrupt. These contacts need to be represented distinctly in numerical flow models as capillary or permeability barriers so that the hydrologic response to these conditions (such as lateral flow) are simulated accurately (Montazer and Wilson 1984). In other locations, the transitional zones apparent in the gradual change in matrix properties also need to be modeled appropriately so that numerical flow models do not simulate lateral flow or ponding of water where they do not actually exist (Moyer et al. 1996). At several locations, glass has been altered to clays and zeolites where lithologic transitions caused high saturations, ancient water tables or perched water caused complete saturation, weathering followed deposition, or thick-welded tuffs cooled slowly (Levy 1984; Broxton et al. 1987; Bish and Aronson 1993; Buesch, Spengler et al. 1995; Levy et al. 1996). In the upper, crystal-rich zones of both of the major pyroclastic flow units (Tiva Canyon Tuff and Topopah Spring Tuff), varying degrees of vapor-phase corrosion have occurred and resulted in changes in properties with depth, particularly porosity. (Vapor-phase corrosion or alteration is the dissolution of pumice fragments by hot vapors before moderately welded tuffs cool.) Vapor-phase corrosion also occurs in the rocks of the Crater Flat Group, and to a lesser degree in the Yucca Mountain and Pah Canyon Tuffs (Table 5.3-13). The relations between lithostratigraphic features (porosity, alteration, micro-fracturing, connectivity, capillary barriers, and welding transitions) and hydrologic conditions and how these characteristics influence the flow of water through the rocks are discussed in detail in L.E. Flint (1998).

Numerical hydrologic flow modeling of the rocks at Yucca Mountain must account for both centimeter-scale features evidenced in matrix-properties characterization and larger-scale properties such as bulk permeability, which is greatly effected by fracturing. Fractures range in length from a few centimeters to several meters and in aperture from less than a micrometer to hundreds of micrometers. In addition, fractures may or may not extend across lithostratigraphic unit boundaries, depending on how and when they formed. Although characterization of matrix properties involves the measurement and description of both small-scale heterogeneities of the rocks and larger scale

features of the hydrogeologic units as a whole, the definition of detailed hydrogeologic units is based, nevertheless, on measurement of rock properties from 7 cm diameter core samples collected during the drilling of boreholes. Thus, matrix-propertiesanalysis is limited to features and properties that can be captured on a relatively small scale. Many matrix properties of the rock units are adequately represented by analyses at this scale (McKenna and Rautman 1996), but even core samples of this size contain very small-scale heterogeneities that may not be accounted for in the sampling or measurement phases. Mineral alteration, micro fractures, and even porosity occur variably within the rocks at the core scale. Although some matrix properties are indicative of larger scale features (e.g., the relatively high matrix porosity of lithophysal zones), many larger scale features (bulk properties of lithophysal cavities, fractures, and faults) will not be adequately represented. However, these inadequacies are critical in numerical flow modeling only when hydrologic conditions are present that result in flow of water in fractures or faults.

5.3.3.1.1.1 Porosity and Saturation

Data collected according to procedures and methods outlined in Subsection 5.3.2.1.2 include measurements of porosity, bulk density, volumetric water content, saturated hydraulic conductivity, and moisture-retentioncharacteristicsfor all lithostratigraphicunits. (Note: In the context of matrix properties, "water content" always means volumetric water content.) To provide initial conditions and calibration data for numerical flow models, present-day field moisture and water potential also were measured. Table 5.3-13 indicates the currently accepted lithostratigraphic nomenclature as it is associated with the major hydrogeologic units historically used and the detailed hydrogeologic units developed for hydrologic modeling of the unsaturated zone at Yucca Mountain. Because water is stored in and flows through the interconnected pores of the rock matrix, porosity is useful for characterizing the hydrologic character of the various rock types. In general, the porosity can be related primarily to the depositional features and amount of welding of a rock, with the lowest porosities occurring in the most densely welded rocks and the highest porosities occurring in the nonwelded and bedded rocks. Examples of borehole profiles illustrating porosity, saturation, and particle density of all lithostratigraphicunits are shown in Figures 5.3-19 and 5.3-20. These profiles clearly illustrate that distinct vertical changes in porosity and particle density correspond with various lithostratigraphic boundaries. (Note: In this report the term "porosity" refers to that determined from the 105°C oven-dry weights unless otherwise noted.)

Borehole SD-9 (Figure 5.3-19) was drilled in Wren Wash to the north of the potential repository footprint. At the land surface, the borehole penetrates the Paintbrush nonwelded hydrogeologic unit, which is relatively thick at this location. The borehole bottoms out within unit 1 of the Prow Pass Tuff. Because infiltration is high (Flint, A.L. et al. 1996) at this location, the TCw hydrogeologic unit is absent, and the Calico Hills Formation is composed only of zeolitic rocks, saturations are very high throughout the profile below the Paintbrush nonwelded hydrogeologic unit, especially in the Topopah Spring Tuff middle nonlithophysal zone, which is the potential repository horizon. In contrast, borehole SD-7 (Figure 5.3-20) is located on Highway Ridge to the south of the repository footprint. At this location, the Tiva Canyon Tuff is about 80 m thick, the Paintbrush nonwelded hydrogeologic unit is very thin, infiltration is lower, and the upper several meters of the Calico Hills Formation are not zeolitic. This combination of conditions in the rocks at borehole SD-7 results in abrupt decreases in saturation below the repository horizon. The unaltered rocks below the basal

vitrophyre of the Topopah Spring Tuff are very low in saturation, as are the rocks in the upper part of the Prow Pass Tuff unit 3. This results in large decreases in unsaturated hydraulic conductivity and, therefore, reductions in transport pathways and flux through the matrix or fractures from the repository horizon to the water table.

The relations between porosity and saturation were used to subdivide the major hydrogeologic units into the detailed hydrogeologic units as described below.

Tiva Canyon Welded Hydrogeologic Unit–The TCw hydrogeologic unit was described by Montazer and Wilson (1984) as the densely to moderately welded uppermost stratigraphic layer that underlies much of Yucca Mountain. It dips 5 to 10° eastward, is absent in some washes and is about 150 m thick beneath Yucca Crest.

The transition from densely welded to vapor-phase corroded rocks in the uppermost part of the Tiva Canyon Tuff is sparsely sampled and is observable in its entirety only in surface-outcrop transects (Flint, L.E. et al. 1996b). Therefore, properties of the uppermost units are based on the analogous rocks in the uppermost welded units of the Topopah Spring Tuff. The upper part of Tpcrn4 (hydrogeologic unit CCR) is densely welded, whereas the lower part of Tpcrn4 and the underlying Tpcrn3 are increasingly vapor-phase corroded with depth. Although the Tpcrn4 is not present at this location, borehole N27 is located fairly high within the section, beginning in the base of Tpcrn3 where the CUC is highest in porosity. Tpcrn2 and Tpcrn1 have lower porosity and less vapor-phase corrosion than the overlying units, but higher porosity and more vapor-phase corrosion than the Tpcpul. Therefore, boundaries of the hydrogeologic units were selected primarily on the basis of porosity. The boundary between CCR and CUC (Flint, L.E. 1998) is located at the depth at which the porosity increases to greater than 9 percent, and the boundary between CUC and CUL (Figure 5.3-21) is located at the depth at which the porosity decreases to less than 20 percent. Because the CUL has somewhat higher porosity than the underlying densely welded rocks of the Tiva Canyon Tuff, the lower boundary of CUL was defined as the lithostratigraphic contact between the Tpcpul and the Tpcpmn.

The middle and lower parts of the crystallized rocks of the Tiva Canyon Tuff include the Tpcpmn, Tpcpll, Tpcplnh, and Tpcplnc (Table 5.3-13). As indicated by the profiles for boreholes SD-7 and N55 (Figure 5.3-21), because porosity is very uniform through these lithostratigraphic units, together they comprise the CW. Toward the base of the lower nonlithophysal columnar subzone (Tpcplnc), porosity increases because of an increased amount of vapor-phase corrosion and the transition from welded to nonwelded in the vitric rocks below (Buesch, Spengler et al. 1996). The base of the CW is where the porosity increases in the vapor-phase corroded rocks to greater than 14 percent.

Paintbrush Nonwelded Hydrogeologic Unit–The PTn hydrogeologic unit of Montazer and Wilson (1984) is described as consisting of the nonwelded and partially welded base of the Tiva Canyon Tuff, the Yucca Mountain and Pah Canyon Tuffs, the upper nonwelded and partially welded upper part of the Topopah Spring Tuff and the associated bedded tuffs. It consists of thin, nonwelded ashflow sheets and bedded tuffs that attain a maximum thickness of 100 m and thin to the southeast to about 20 m. This unit outcrops in several locations, particularly along Solitario Canyon and Yucca Wash.

To define the detailed character of these rocks, they are separated into several hydrogeologic units. The rocks near the base of the Tiva Canyon Tuff exhibit transitions in porosity and mineral alteration and are divided into the CMW and CNW hydrogeologic units. The CMW hydrogeologic unit consists of moderately welded rocks near the base of the Tpcplnc and usually most of Tpcpv2 (Table 5.3-13). As indicated in the profiles for boreholes N31 and N32 (Figure 5.3-22), porosity of the CMW ranges from greater than 15 percent at the top of the unit to greater than 28 percent at the bottom of the unit. The CNW consists of nonwelded to partially welded rocks in Tpcpv1, and locally includes the base of Tpcpv2.

Although the lithostratigraphic units of the Paintbrush nonwelded hydrogeologic unit generally are relatively thin, their properties are distinct enough to warrant delineation as separate hydrogeologic units. A numerical modeling exercise was performed by L.E. Flint et al. (1996b) to assess the hydrologic significance of these units and to determine whether or not their properties were sufficiently different to warrant delineation as separate units. The modeling results indicated that abrupt and linear contacts along with the contrasts in properties were instrumental in creating lateral flow of water along the sloping contacts. However, in reality this exercise simplified the linear character of the contacts because the contacts are only locally linear and are less likely to divert water laterally due to heterogeneities, which cause increases in saturation and local instabilities that result in vertical flow across the contact. The Yucca Mountain Tuff has properties very similar to the Tpbt4 and Tpbt3 to the south of Drillhole Wash but has lower porosity and becomes moderately welded to the north. In the modeling exercise (Flint, A.L. et al. 1996; Moyer et al. 1996), this unit was represented by the properties of the lower porosity rocks, and it caused water to be diverted laterally. As a result, TPY represents rocks of less than or equal to 30 percent porosity, and the bedded tuff units Tpbt4 and Tpbt3 represent the Yucca Mountain Tuff where there is greater than 30 percent porosity (Table 5.3-13). To the south of Drillhole Wash, the hydrogeologic unit BT3 represents both Yucca Mountain Tuff and Tpbt3. The Pah Canyon Tuff is welded far to the north in Yucca Wash, but no welded or even moderately welded rocks were found in the samples of the Pah Canyon from boreholes at the site. Therefore, the nonwelded to partially welded Pah Canyon Tuff is hydrogeologic unit TPP. Because the modeling results indicated that the properties of the Tpbt2 and the nonwelded top of the Topopah Spring Tuff (Tptrv3) were not sufficiently different to warrant delineation of two separate units (Flint, A.L. et al. 1996; Moyer et al. 1996), the Tpbt2 and the Tptrv3 are represented by hydrogeologic unit BT2 (Table 5.3-13). Because the Tptrv2 is a thin unit that marks the sharp transition from nonwelded to densely welded rocks, it is included as the lower part of BT2.

Mineral Alteration–Mineral alteration has a significant influence on the hydrologic properties of rock units. Porosity, for example, is directly related to the alteration history of the rocks at Yucca Mountain because clays, zeolites, opal, and calcite form in situ or are deposited in pore spaces, causing a reduction in porosity. However, the occurrence of clay or zeolite zones reduces the measured porosity only slightly because water is stored in clay and zeolite mineral structures rather than in pore spaces. An exception involves smectites, which swell with the incorporation of water to as much as 300 times their original size, thus reducing the porosity. Because clays and zeolites influence pore-size distribution, their presence also affects moisture-retention characteristics and permeability. Rocks containing clay or zeolites typically contain a relatively large volume of water,

but because the water is held tightly within clay structures and very small pores, the permeability is reduced.

A laboratory measurement that is used for identifying rocks containing altered minerals is "residual water content." Residual water content has been defined, historically, as the water content in a sample for which the change in water content divided by the change in water potential is zero (van Genuchten 1980). It may also be thought of as the degree of saturation at which liquid flow stops, due to discontinuous columns of water, and vapor flow begins (Flint, L.E. 1998). Studies of soils indicate that residual water content can be determined as the amount of water left in the pores after equilibrating a rock sample in an environment of approximately −700 bars water potential, or about 65 percent relative humidity (Jackson 1964; Rose, D.A. 1963). Residual water content generally is used as the dry-end fit parameter for moisture-retention curves and its magnitude greatly affects the prediction of unsaturated hydraulic conductivity at low water contents. "Residual water" is the water remaining in rock at very low saturation after complete drainage, residing in the smallest pores and channels, within the structure of minerals such as clays or zeolites, and adsorbed on the surfaces of these minerals. Residual water, whether contained in mineral structures or in very small pores, remains relatively immobile in the liquid phase and is removed from the rock primarily in the vapor phase.

However, the definition of "residual water content" is critically dependent on the method or methods used to dry rock samples. Bush and Jenkins (1970) determined that drying rock samples in an oven at a temperature of 60°C and 40 percent relative humidity maintained water within the clay structures, as well as one to two molecular layers of water adsorbed to their surface, while removing it from the pore channels. Accordingly, after saturation (see Subsection 5.3.2.1.2.2), rock samples from Yucca Mountain were dried at 60°C and 65 percent relative humidity and at 105°C and ambient relative humidity (< 20 percent), and porosity was calculated using both dry weights. The difference between the porosities calculated from the two different drying methods is equivalent to the volumetric water content of the sample after it has been dried at 65 percent relative humidity and is defined herein as the "residual water content." A residual water content of 5 percent or greater was determined to be an appropriate threshold indicative of the presence of altered minerals (Rautman and McKenna 1997; Flint, L.E. 1998). For example, a residual water content of 5 percent or greater was determined for nearly saturated rock samples from the Tpcpv unit near the base of the Tiva Canyon Tuff (borehole N31; Figure 5.3-22). This zone of significant residual water content also correlates with the presence of up to 35 percent smectite, which is pervasive in varying quantities across the site (Bish and Chipera 1989). The extent of mineral alteration seems to be related to the topographic location of borehole N31 in that it is located in a narrow up-wash channel with thin alluvium. In such a setting, the subsurface rocks have been subjected to more frequent infiltration events that have sustained relatively high water contents in the rocks for longer periods of time (Flint, L.E. and Flint 1995). The adjacent borehole (N32; Figure 5.3-22), located only 15 m away but not in the channel, has a thinner altered zone in the base of the Tiva Canyon Tuff. This zone of mineral alteration at the base of the Tiva Canyon Tuff, which occurs within the CMW unit, is an important hydrologic feature in that it may accentuate the capillary barrier caused by small pores overlying large pores at this transition from welded to nonwelded tuffs. Although the altered zone continues downward through the base of the BT3 unit, high saturations are not maintained despite

the presence of clay. This is because the combination of very large porosities and the particular pore-size distributions of the units below the CMW result in drainage.

Topopah Spring Welded Hydrogeologic Unit–The TSw hydrogeologic unit consists of a very thin upper vitrophyre, a thick central zone consisting of several densely welded, devitrified ashflow sheets, and a thin lower vitrophyre (Montazer and Wilson 1984). The unit contains several lithophysal-cavity zones of varying thickness, and the rocks are intensely fractured. This unit is the thickest and most extensive of the Paintbrush Group, and contains the central and lower densely welded and devitrified zones being considered for the potential repository.

The vitric, densely welded subzone (Tptrv1), which contains the upper vitrophyre, is typically less than 0.5 m thick, but varies from 0 to 2 m thick across Yucca Mountain (Flint, L.E. 1998). The Tptrv1 typically has a porosity less than 5 percent (Figure 5.3-23, boreholes UZ-14 and UZ-16) and has properties that sharply contrast with the rocks overlying it. The underlying dense subzone of the crystal-rich nonlithophysal zone also has very low porosity (less than 9 percent) and has been combined with the Tptrv1 as the hydrogeologic unit TC. Tptrn3 is typically thicker to the north and very thin in the southern parts of the site area.

Most of the crystallized, moderately to densely welded Topopah Spring Tuff is divided into five hydrogeologic units that closely correspond to lithostratigraphic units. The vapor-phase corroded and crystal-transition subzones of the crystal-rich nonlithophysal zone (Tptrn2) have porosities similar to the underlying lithophysal zone (Figure 5.3-23), but moisture-retention characteristics (discussed below) differ significantly. Therefore, Tptrn2 is represented by the single hydrogeologic unit TR and has porosities greater than 9 percent. Porosities of the lithophysal zones (Tptrl, Tptpul and Tptpll) and nonlithophysal zones (Tptpmn, Tptpln) differ significantly, with the lithophysal zones having porosities of about 14 percent and the nonlithophysal zones having porosities of about 11 percent (Figure 5.3-23). In addition, because they differ significantly in moisture-retention characteristics, each zone is represented by individual hydrogeologic units: TUL (upper lithophysal zone), TMN (middle nonlithophysal zone), TLL (lower lithophysal zone), and TM2 and TM1 (lower nonlithophysal zone). The lower nonlithophysal zone of the Topopah Spring Tuff is divided into two hydrogeologic units because porosities decrease with greater depth to about 9 percent and as locally the rocks become more vitric. The result is that the rocks at the base that have different moisture-retention characteristics due to an increase in the amount of smectite with greater depth (Bish and Chipera 1989). Porosities of the vitric rocks at the base of the Topopah Spring Tuff contrast greatly with the rocks above them (Figure 5.3-23). Therefore, the very low-porosity, densely welded subzone (Tptpv3) that contains the vitrophyre has been designated as a discrete hydrogeologic unit, PV3.

Calico Hills Nonwelded Hydrogeologic Unit–The CHn hydrogeologic unit of Montazer and Wilson (1984) includes the nonwelded to partially welded vitric zone, which is locally zeolitic, that is the lowermost part of the Topopah Spring Tuff. It also includes the pre-Topopah Spring Tuff bedded tuff, the beds of the Calico Hills Formation, a bedded tuff, a bedded sandstone, the Prow Pass Tuff, which is nonwelded to welded, the pre-Prow Pass bedded tuff, and the upper part of the Bullfrog Tuff, where it is above the water table. The presence of zeolites and other altered minerals in the rock units below the Topopah Spring Tuff is an important criteria for the classification of

hydrogeologic units. Within the unsaturated zone the thickness of the CHn hydrogeologic unit ranges from zero to more than 500 m.

The moderately welded subzone of the Topopah Spring Tuff (Tptpv2) is PV2. Because the nonwelded base of the Topopah Spring Tuff (Tptpv1) is similar in properties to the underlying bedded tuff (Tpbt1), they have been combined in hydrogeologic unit BT1 (Figure 5.3-24). The Calico Hills Formation consists of a nonwelded pyroclastic flow, thin fallout deposits, and, locally, a basal bedded sandstone. These rocks are either vitric or zeolitic and this difference strongly controls the flow of water through them. The part of the formation that is primarily pyroclastic flow deposits, which includes units 1 through 4, is divided into hydrogeologic units CHV and CHZ, depending on whether it is vitric or zeolitic, respectively. In this analysis, the vitric-zeolitic boundary was determined based on measurements of residual water content rather than on lithologic descriptions that include estimates of mineral percentages. The presence of zeolites was defined as a threshold of 5 percent or greater residual water content in altered rocks that were suspected of being zeolitic (Rautman and McKenna 1997). Subsequent measurements of mineral percentages indicated a strong correlation between the occurrence of the zeolite mineral clinoptilolite and rocks with high- saturation and low-saturated hydraulic conductivity (Figure 5.3-25), confirming the effect of mineral alteration on hydrologic properties. These rocks do, however, have relatively high porosity, as shown in Figure 5.3-24 for borehole SD-7 and in Figure 5.3-26 for borehole UZ-16. The bedded tuff near the base of the Calico Hills Formation (Tacbt) has relatively high porosity, but is partially silicified and contains opal, strongly affecting the moisture-retention characteristics and permeability. Tacbt, along with the thin basal sandstone (Tacbs), which is similar in properties to the bedded tuff, comprise the hydrogeologic unit BT.

Prow Pass and Bullfrog Tuffs–The Prow Pass Tuff has been sampled in only four deep boreholes (SD-7, SD-9, SD-12, and UZ-16; Table 5.3-6) and shows significant variation in properties among them. The Prow Pass Tuff is composed of four pyroclastic flow units and an underlying interval of bedded tuff (Table 5.3-13; Figure 5.3-24). Lithostratigraphic units Tcp4, Tcp2, and Tcp1 are nonwelded to partially welded pyroclastic flow deposits, and Tcp3 is a partially to moderately welded pyroclastic flow deposit that displays a compound cooling history. The alteration of the initially vitric rocks to zeolites is pervasive throughout the Prow Pass Tuff with the exception of Tcp3, which is devitrified and vapor-phase crystallized where the welding is most intense. Consequently, the Prow Pass Tuff is divided into hydrogeologic units based primarily on lithostratigraphic-unitboundaries and secondarily on the residual water content as the indicator for mineral alteration. In this case, the 5 percent and greater residual water content correlates well to high saturations (Figures 5.3-24 and 5.3-26). The detailed hydrogeologic units within the Prow Pass Tuff are: PP4, which is equivalent to lithostratigraphic unit Tcp4, is zeolitic, and has high porosity; PP3, which is the upper part of Tcp3, has no zeolites but contains vapor-phase altered minerals that influence the residual water content, and is high in porosity; PP2, which is the lower half of Tcp3 and is welded, devitrified, unaltered, and lower in porosity; and PP1, which is composed of Tcp2 and Tcp1, both of which are zeolitic. The boundary between PP3 and PP2 varies among the four boreholes penetrating Tcp3, but dividing the unit into two halves seems to adequately minimize the statistical variance in properties.

The Bullfrog and Tram Tuffs were penetrated only by borehole SD-7 (Table 5.3-6, Figure 5.3-24) and have been divided only on the basis of mineral alteration. The hydrogeologic units are BF3, which is composed of the unaltered units 4 and 3 of the Bullfrog Tuff, and BF2, which is composed of the slightly altered units 2 and 1 of the Bullfrog Tuff, the underlying basal sandstone, and the altered Tram Tuff.

5.3.3.1.1.2 Effects of Mineral Alteration and Estimation of Saturated-Hydraulic Conductivity

The altered rocks of the CHn unit in boreholes SD-7 and UZ-16, such as BT (consisting of Tacbt and Tacbs), PP4 (Tcp4), and PP1 (Tcp1), have relatively low porosities using the relative humidity drying method (Figures 5.3-24 and 5.3-26) but 5-percent-or-greater residual water content because a significant quantity of water is held in the altered minerals. Chipera et al. (1995) reported significant percentages of clinoptilolite in these units in borehole UZ-16 (Figure 5.3-26, right-most column), confirming that these units are zeolitic. However, other intervals in borehole SD-7 (e.g., PP3 and parts of PP2) have significant residual water content even though no zeolites were found (Figure 5.3-24). This is because these intervals contain significant quantities of the vapor-phase alteration minerals tridymite and cristobalite (Rautman and Engstrom 1996a, Appendix H). (Cristobalite is present from depths 502.3 to 569.4 m (2 to 35 percent) and tridymite is present from 502.3 to 538.4 m (1 to 6 percent).) These results indicate that although 5-percent-or-greaterresidual water content always is present in rocks where zeolites are found, the converse is not true because the presence of other alteration minerals like tridymite and cristobalite can produce the same effect. However, it is apparent that the presence of cristobalite alone does not produce this effect but, rather, that tridymite is necessary, possibly in combination with cristobalite. This conclusion is based on the fact that although cristobalite alone was present in the interval from 538.4 to 569.4 m in unit PP2, significant residual water content was not found throughout the interval. Overall, these findings indicate that certain vapor-phase alteration minerals have significant effects on the structure and geometry of the pore spaces in the rock, resulting in greater moisture-retention capacity at low saturations. The greater tortuosity of the flow channels caused by the presence of clay, zeolites, or vapor-phase minerals also reduces the permeability, as discussed below. (There is evidence that the very high percentages of opal-CT (a disordered silica phase containing both cristobalite- and tridymite-like structural units) often found in the Tacbt (Bish and Chipera 1989) may influence the hydrologic character of rocks by reducing the permeability even more than the presence of zeolites.)

The effects of alteration minerals on the hydrologic properties of the unsaturated rocks at Yucca Mountain were evaluated further by comparing the porosity calculated from relative humidity oven drying with the measured saturated hydraulic conductivity (K_s) for 593 samples collectively representing all lithostratigraphic units (Figure 5.3-27). The upper figure shows the relation of the log of saturated hydraulic conductivity (K_s) to porosity for samples grouped by hydrogeologic unit. The location on the plot of the points representing the zeolitic rocks of the Calico Hills Formation (CHZ) is distinct in that they indicate relatively high porosities (> 0.2) but very low saturated conductivities (most less than 1×10^{-10} m/s). Samples from the crystal-rich and crystal-poor vitrophyres of the Topopah Spring Tuff (TC and PV3) have very low-matrix porosities (< 0.1) yet several have relatively high conductivities (1×10^{-8} to 1×10^{-7} m/s) because of the presence of microfractures. Similarly, three crystallized and welded samples from the Topopah Springs Tuff

(Tpt) with fairly low porosity and high conductivity contained visible microfractures. It is possible that some of the samples from the TC and the PV3 with low values of K_s had microfractures, but they were filled with alteration minerals.

The correlation between porosity and saturated hydraulic conductivity is indicated clearly on Figure 5.3-27b where the porosity and saturated conductivity are plotted by more generalized rock type: vitric/crystallized,altered, and microfractured. The microfractured category includes TC and PV3, plus the three welded samples from the Tpt with the visible microfractures that do not truly represent matrix permeability. The "altered" category includes all lithostratigraphic units below the vitric/zeolitic boundary in the CHn unit (BT1, CHZ, BT, PP4, PP3, PP2, PP1, BF3, and BF2) and the CMW unit in the Tiva Canyon Tuff. The vitric/crystallized category includes all of the other hydrogeologic units. The "altered" category is only crudely defined because there are crystallized and unaltered zones within the hydrogeologic units that comprise this category. Nevertheless, the category was defined as indicated for simplicity in the application of predictive equations. Relations between these simplified rock-type categories and porosity and saturated conductivity for the 593 samples were described by regression models developed by L.E. Flint (1998, Table 6 and Figure 12). Regression models were used to estimate K_s for the 4,300 samples for which only porosity had been measured. Some of these samples had saturated conductivities too low to measure ($<1\times10^{-12}$ m/s) with the current laboratory technique. The estimates provide a representation of K_s based on porosity for the densely welded tuffs, as well as providing estimates of matrix permeability for samples with microfractures for prediction of unsaturated hydraulic conductivity. These microfractured samples are likely to have high conductivities only when saturated and are likely to have relatively low conductivity after the fractures drain.

The hydrogeologic units present at the vitric/zeolitic boundary vary spatially in the vicinity of Yucca Mountain. In the southern part of the site where the boundary is as much as 140 m below the Tptpv3, the upper parts of the Calico Hills Tuff, Tpbt1, Tptpv1, and Tptpv2 are unaltered. This condition was observed in the samples from only two boreholes, SD-7 and SD-12. All other boreholes are located to the north and east where the vitric/zeolitic boundary is typically defined at the base of Tptpv3, with zeolitization along the fractures in that unit. Properties for the BT1 hydrogeologic unit, which includes Tptpv1 and Tpbt1, have been characterized for both altered (BT1a) and unaltered (BT1) samples.

The location of the vitric/zeolitic boundary seems to be an important factor contributing to formation of perched water bodies (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The high saturations in boreholes SD-7 and UZ-16 in the rocks below the Topopah Spring Tuff (Figures 5.3-24 and 5.3-26) strongly indicate that the presence of alteration minerals has reduced the permeability and the capacity of these rocks to transmit water downward. In zones where there is little or no alteration, such as in the vitric rocks of the Calico Hills Formation (CHV) and some of the devitrified and vapor-phase corroded rocks of the Prow Pass Tuff (PP3), saturations are extremely low because the pores are larger and drain more easily. In the crystallized and minimally vapor-phase-corroded Prow Pass Tuff (PP4, PP2, and PP1), saturations are high because the rocks are welded and have small pores. There is also some evidence that high-temperature alteration products, such as the vapor-

phase crystallization minerals tridymite and cristobalite, have influenced the retention of water in this zone.

5.3.3.1.1.3 Mean Matrix-Property Values for Hydrogeologic Units

To facilitate the use of matrix hydrologic properties data in numerical flow models, mean values and standard deviations were calculated for all measured properties and for estimated K_s values from the regression analysis for each hydrogeologic unit (Table 5.3-14). In general, the variation in porosity within each hydrogeologic unit was relatively small, with the exception of BT4, BT2, and PV2 (Flint, L.E. 1998). Because BT4 is very thin and few samples were collected, substantial variation in properties was expected. The relatively large variation in properties in BT2 resulted from the fact that it is composed of lithostratigraphic units that vary significantly in the amount of mineral alteration, and it includes the moderately welded Tptrv2 that rapidly grades downward into low porosity rocks. The same is true for the PV2, which grades relatively sharply from very low-porosity to relatively high-porosity rocks.

Several important results are apparent from comparison of mean values of porosity for the hydrogeologic units (Flint, L.E. 1998). In the Tiva Canyon Tuff, the lithophysal CUL has a mean porosity (16 percent) that is twice that of the mostly nonlithophysal CW (8 percent). In contrast, within the TSw unit, mean porosities of lithophysal and nonlithophysal units do not differ quite as much as in the Tiva Canyon Tuff in that the TUL and TLL have mean porosities of 15 and 13 percent, respectively, and the TMN, TM2, and TM1 have mean porosities of 11, 11, and 9 percent, respectively. The vitric bedded tuffs and nonwelded ashflow tuffs of the Paintbrush nonwelded hydrogeologic unit, included in CNW, BT4, BT3, and BT2, vary in mean porosity from 39 to 49 percent. These porosities are significantly higher than similarly deposited vitric tuffaceous rocks in BT1 and CHV that have mean porosities of 27 and 34 percent, respectively. All units below the Topopah Spring Tuff have moderate to high mean porosity (26 to 34 percent) except for BF3, which has 11 percent. Frequency distributions of porosity developed for each detailed hydrogeologic unit (Flint, L.E. 1998; Appendix I) indicate normal distributions for most units.

Mean water content, saturation, and water potential also were calculated (Table 5.3-14), but the meaningfulness of these values must be considered in the context of the entire unsaturated-zone system (Flint, L.E. 1998). Actually, spatial averaging of water content, saturation, and water potential values is not realistic because these characteristics are dependent on the spatial distribution of net infiltration. Spatially variable net infiltration results from spatial and temporal distribution of precipitation, varying thickness of alluvial cover, topographic positions of boreholes, and variable thickness of shallow rock units with different properties (Flint, A.L. and Flint 1994; Flint, L.E. and Flint 1995; Hudson and Flint 1996).

Examination of the means and standard deviations for saturated hydraulic conductivity (K_s) reveals large statistical variances for many hydrogeologic units (Flint, L.E. 1998). In particular, unit CMW exhibits large variances because the degree of mineral alteration likely is variable vertically within the unit and laterally between boreholes. The large variance in K_s values for unit TC likely is due to the random sampling of microfractures. For some units, geometric means of the K_s values estimated from porosity using the regression analyzes are lower than the geometric means of the

measured values of K_s. This is particularly true for units that contain microfractures because K_s was estimated using the regression equation for the matrix only (discounting the K_s contributed by the microfractures), resulting in estimated mean K_s that is four orders of magnitude lower than the measured K_s for TC and two orders of magnitude lower for PV3. Significant differences between estimated and measured mean K_s also resulted for units below the Topopah Spring Tuff because all K_s values were estimated using the regression equation for altered rocks even though the individual units have significant variable degrees of alteration and crystallization.

5.3.3.1.1.4 Moisture-Retention Characteristics and Unsaturated-Hydraulic Conductivity

Moisture-retention curves (water potential versus saturation) are shown for six representative hydrogeologic units in Figure 5.3-28. For the first three hydrogeologic units (CW, BT3, and TR), the data sets include both laboratory desorption curves and composite curves combining laboratory data and all field data that were collected from cores. Standard errors of the curve-fit parameters (a and n) were used to include the 66 percent confidence intervals about the curve-fit models. These intervals were used for graphical purposes because the lower limits of the 95 percent intervals were undefined for a number of hydrogeologic units. Two models were developed: one model is associated with the laboratory desorption data (thick line on Figure 5.3-28a-c) and the second model is associated with the composite data set. In all three units, the composite model plotted lower on the graph because of inclusion of field samples that had lower water contents for given water potentials than the laboratory desorption data. This results in slightly wetter, air entry values (lower a), which increases modeled fluxes for the same matrix saturation. For the other three hydrogeologic units (TUL, PV3 and CHZ), the data sets consist of laboratory desorption measurements only and include the error incorporated into the laboratory measurement due to the instrument resolution.

Notable differences in the moisture-retention curves are chiefly a function of the pore size distributions of the rock types. These different distributions are primarily due to degree of welding, and are exemplified by comparing the curve for the bedded tuff (BT3) to those of the welded tuffs CW and TUL. The welded tuff units all have high air entry pressures, approximately equivalent to 1/a, and describe the water potential at which the pores initially drain, and thus represent the largest pores. Rocks in the hydrogeologic unit TR are vapor-phase corroded, which increases the size of the pores and results in pore structures that drain at lower water potential than non-corroded welded rocks. Another contrasting characteristic is the residual saturation, which is approximately represented by the saturation at which the dry end of the curves becomes asymptotic. Residual saturation is caused by the abundance of very small pores that retain water at approximately 65 percent relative humidity (−700 bars). Altered rocks such as CHZ, or vitrophyres such as PV3, have the largest residual saturations, although for altered rocks some of this is due to residual water (see Subsections 5.3.3.1.1.1 and 5.3.3.1.1.2 for discussions of mineral alteration.)

Moisture-retention van Genuchten curve-fit parameters (a, n, and m) for each hydrogeologic unit are listed in Table 5.3-15. The log of alpha is well correlated to porosity for the unaltered vitric, crystallized, and microfractured samples (Flint, L.E. 1998), but is not as well correlated for the altered samples. Regression models were made for the prediction of a log that was determined for all hydrogeologic units for the purposes of estimating parameters from porosity. These models are useful for the representation of more gradual transition zones in numerical flow models (Glass et al.

1994) to eliminate unwarranted lateral flow of water that results from flow-model layers that unrealistically simulate sharply contrasting properties (Flint, L.E. et al. 1996a; Moyer et al. 1996). The form of the model equation for the relation of a log to porosity is the same as for K_s for both altered and unaltered hydrogeologic units (Flint, L.E. 1998).

Moisture-retention curve-fit parameters were used to predict unsaturated hydraulic conductivity based on values of saturated conductivity (van Genuchten 1980). These predictions were compared with direct measurements of unsaturated hydraulic conductivity (Figure 5.3-29) made using a steady-state ultra-centrifuge, also called an unsaturated flow apparatus. Moisture-retention curves are shown with data obtained using two laboratory techniques: the chilled-mirror psychrometer (Model CX2, Decagon, Pullman, WA), which was used to compile moisture-retention data for all hydrogeologic units, and the unsaturated flow apparatus, which has higher resolution at the wet end of the curve and also uses the entire plug sample that is used to measure conductivity, rather than a subsample (see Figure 5.3-14). The driest point for both samples was obtained using the estimated residual water content, as discussed in previous subsections on mineral alteration (Subsection 5.3.3.1.1.1). Data from the two laboratory methods match relatively well and the points from the unsaturated flow apparatus fill in the wet-end gap left by the chilled-mirror psychrometer measurements (Flint, L.E. 1998). Measured unsaturated conductivity values for two replicates from two samples also are shown in Figure 5.3-29 along with estimated unsaturated hydraulic conductivity from the moisture-retention data using three methods. Method 1 predicts unsaturated hydraulic conductivity from moisture-retention parameters obtained from fitting the measured moisture-retention data (van Genuchten 1980) collected using the CX2 chilled-mirror psychrometer. Method 2 fits the conductivity data measured on the centrifuge, then predicts the moisture retention. Method 3 simultaneously fits both the moisture-retention and conductivity data. Method 1 results in a poor prediction of the conductivity data. The driest point, collected from volumetric water content at relative humidity dryness, is not predicted from either methods 2 or 3, suggesting that it is an inappropriate definition of residual water content when considered for unsaturated flow processes. It is possible that there are two residual water contents, one representing physical and static properties and another representing flow processes. If the moisture retention-data are fit without including the "residual water content" point, the unsaturated-hydraulic-conductivitydata can adequately be predicted by moisture-retention data for these two samples.

5.3.3.1.1.5 Lithologic Stratification: Transitions and Capillary and Permeability Barriers

Lithologic features, such as degree of welding, layering in ashfall tuffs, fracture distributions, and subsequent mineralogic alteration, are critical to characterize in order to determine the distribution of hydrologic conditions within the vertically stratified system at Yucca Mountain. The primary large-scale feature controlling the moisture distribution and flux at Yucca Mountain is the occurrence of nonwelded and bedded tuffs of the PTn hydrogeologic unit between two welded and fractured pyroclastic flow units, the Tiva Canyon and Topopah Spring Tuffs (Flint, L.E. 1998). This stratigraphic orientation results in transition zones that exhibit pronounced vertical variations in matrix properties, typically caused by variations in porosity due to welding, mineral alteration, or vapor-phase corrosion, or by abrupt contrasts in properties due to layering of ashfall tuffs. Corresponding to the degree of welding, fracture characteristics (density, geometry, and continuity) typically are poorly developed in nonwelded tuffs and well developed in the welded rocks. The

transition zone from fracture-dominated flow in welded rocks (where saturation is high) to matrix-dominated flow in nonwelded rocks, can significantly affect the overall nature of large-scale flow in the unsaturated zone. Sharp decrease or increase in the magnitude of hydrologic properties with depth (that also corresponds to changes in the degree of fracturing) in these transition zones may result in localized concentration of water due to capillary or permeability barriers or effects, or depletion of water due to drainage (Flint, L.E. 1998).

The concept of a natural capillary barrier is summarized by Montazer and Wilson (1984) as a fine-grained layer overlying a coarse-grained layer, where water cannot flow from the smaller pores into the larger pores until the height of water in the overlying layer exceeds a critical height, equivalent to the difference in the capillary rise of the two pore sizes. At Yucca Mountain the potential for this condition occurs when a porous medium, such as the nonwelded rocks of the Paintbrush nonwelded hydrogeologic unit, is bounded above and below by double-porosity media such as the welded and fractured rocks of the Tiva Canyon and Topopah Spring Tuffs (Montazer and Wilson 1984). A capillary barrier could theoretically be present at the top of the nonwelded rocks where fine-grained welded rocks overlie coarse-grained nonwelded rocks. A barrier could also be present at the base of the nonwelded rocks where the coarse-grained rocks overlie the larger-aperture fractures present in the welded rocks. If these conditions exist in the layered, dipping beds at Yucca Mountain, they could potentially result in lateral flow of water (Flint, L.E. 1998). The theoretical bases for and behavior of capillary and permeability barriers are discussed further in L.E. Flint (1998).

A detailed study of the contrasting and transitional hydrologic properties of the lithostratigraphic units that comprise the PTn hydrogeologic unit was undertaken to evaluate the degree of contrast necessary to support lateral flow of water above the potential repository. At borehole SD-9 (Figure 5.3-30) the PTn hydrogeologic unit is relatively thick and the TPY exhibits a higher degree of welding than elsewhere (Flint, L.E. 1998). Similar to boreholes N31 and N32 (Figure 5.3-22), the altered zones in borehole SD-9 (indicated by residual water content of 5 percent or greater) seem to occur in the transition zone at the base of the Tiva Canyon Tuff, in the BT3 and in layers within the BT2 and BT3. These zones indicate either contrasts in properties that have resulted in long-term preferential flow pathways, high-saturation zones that allowed development of clays, or high-temperature depositional features that influence the flow pathways (Levy et al. 1996).

A 2-D, cross-sectional model, tilted 6.5°, was used to investigate the influence of detailed unit divisions on the lateral flow of water in the PTn hydrogeologic unit (Figure 5.3-31) (Moyer et al. 1996). All layers are identified to correlate with hydrogeologic units in Table 5.3-13 but in the model had different properties. With a simulated vertical percolation flux at the upper boundary flux of only 1 mm/year, significant lateral flow occurs, particularly at the TPP-BT2 boundary because of a large contrast in permeability. Later flow also occurs in the simulation at CMW-CNW transitional zone at the base of the Tiva Canyon Tuff and at the TPY- BT3 contact. Saturations within the PTn hydrogeologic unit correlate relatively well with saturated hydraulic conductivity but are not high enough to initiate fracture flow at the sharp contact of the PTn hydrogeologic unit and the TSw unit (Flint, L.E. 1998). An ongoing study is attempting to quantify the contrasts in properties of the hydrogeologic units within the PTn hydrogeologic unit that would be necessary to cause significant lateral flow of water under existing and future climatic conditions.

5.3.3.1.1.6 Properties of Fault Zones

Analysis of small-scale core samples from fault zones indicates subtle changes in the physical properties and saturations associated with fault zones. The Ghost Dance fault zone has been penetrated by core holes at several locations including borehole N35, which penetrates the fault in the Tiva Canyon Tuff, and in horizontal boreholes in the Northern and Southern Ghost Dance fault alcoves of the Exploratory Studies Facility, which penetrate the fault at the potential repository horizon. Borehole N35 is located in Broken Limb Wash near SD-12 and penetrates the fault zone at a depth of approximately 30 m. The porosity of the brecciated zone (Figure 5.3-32) is not significantly different than the surrounding rock units and the saturation of the breccia is only slightly higher. Rocks in the foot wall of the fault (below the breccia zone) have properties appropriate for the unit but exhibited saturations 15 to 20 percent higher than expected at the time of drilling. Systematic neutron-moisture logging was done in N35 from the time of drilling in early 1993 until the fall of 1995 to monitor changes in volumetric water content with depth and time (Figure 5.3-33). Significant precipitation occurred during the winter of 1993, resulting in an increase in water content at the surface. Coincidentally, one to two months later, rocks in the foot wall of the fault showed increase in water content from 2.5 to 7.5 percent and, over the next few months, to over 10 percent. While the water did not travel directly through the soil profile, as apparent in the figure, the fault is exposed at the surface several meters upslope from the borehole, providing a fast pathway for precipitation to penetrate the surface (Flint, L.E. 1998).

Analysis of rock core from the horizontal borehole in the Exploratory Studies Facility Northern Ghost Dance fault alcove indicates that the porosity of the TMN changes abruptly at the main trace of the fault in that the porosity of the rock on the foot wall is about 2 percent lower than the porosity of the rock in the hanging wall (Figure 5.3-34). The change in porosity likely was not caused by the faulting itself but is simply a reflection of the spatial variability of porosity within the TMN that is accentuated by the offset and by the brecciated fault zone. Although the saturation profile of the borehole indicates preferential drying of the most highly fractured zones by drilling air, a core sample collected at the deepest penetration of the hanging wall (about 49 m) was completely saturated and all of the other samples had a saturation of about 90 percent.

Based on saturation data from these two boreholes, the first penetrating the Ghost Dance fault above the PTn hydrogeologic unit and the second penetrating the fault in the TSw hydrogeologic unit, it is likely that the fault zone is a fast pathway for water percolating through the Tiva Canyon Tuff, causing increases in saturation in the rock matrix of the foot wall. Although the fault probably also is a fast pathway for water percolating through the PTn hydrogeologic unit, there is little indication based on core samples that the fault serves as a preferential pathway for water through the TSw hydrogeologic unit. Rather, it is likely that water moves through the surrounding highly fractured rock of the TSw hydrogeologic unit just as readily as it moves through the fault zone. Core samples from a horizontal borehole in the Southern Ghost Dance fault alcove are being analyzed to determine if similar conditions exist there.

5.3.3.1.1.7 Spatial Variability of Matrix Properties

Rock properties vary vertically and laterally as the result of original depositional, cooling processes, and subsequent post-depositional alteration (Rautman and Flint 1992; Rautman 1995; Flint, L.E. et al. 1996b; Istok et al. 1994; Buesch, Nelson et al. 1996). At certain scales and at certain positions within the volcanic rocks at Yucca Mountain, the influence of deterministic geologic processes is so strong that the changes in material properties can be predicted with a high degree of confidence (Flint, L.E. 1998). For example, porosities are relatively high in the rocks at the base of the ashflow sheets (Tiva Canyon and TSw hydrogeologic units) where cooling occurred quickly, with progressively lower porosities occurring at higher stratigraphic levels and decreasing to very low porosities in the interior of the deposit where temperatures remained high long enough for compaction and welding to occur. Such trends in matrix hydrologic properties are readily observable in rock samples collected from surface transects of outcrops and from systematic sampling of the rocks exposed n the Exploratory Studies Facility.

Vertical Variability From Surface Transects–Because of the importance of the transitional zone at the base of the Tiva Canyon Tuff, and because of the need to predict properties and conditions at unmeasured points, a study was conducted to evaluate the 2-D spatial variability of the physical and hydrologic properties of that unit (Rautman et al. 1995). Samples of the Tiva Canyon Tuff outcrop on the west-facing slope of Solitario Canyon were collected along a series of 26 vertical transects in the vicinity of borehole UZ-6 (Figure 5.3-8). The sampling area extended northward from near borehole UZ-6 along a 1000 m exposure of the Tiva Canyon Tuff that encompasses a stratigraphic sequence beginning in the moderately welded rocks of the Tpcplnc, extending downward through the Tpcpv3, Tpcpv2, and Tpcpv1, and terminating in the Tpbt4. Porosities of all samples are plotted in Figure 5.3-35 in relation to the scaled thickness from the base of each transect, expressed as a dimensionless proportion. Porosity ranges from about 10 percent at the top of the sequence to as high as 60 percent at the bottom. The composite plots of porosity and saturated-hydraulic conductivity (Figure 5.3-36) indicate strong vertical trends in porosity and saturated-hydraulic conductivity. At the top of the sequence, porosity is relatively low for the moderately welded Tpcplnc but then increases gradually through the Tpcpv2, is consistently about 42 percent in the Tpcpv1, and increases with depth in the Tpbt4 at the bottom of the sequence.

The nonlinear vertical trends in porosity (and in the correlated property saturated-hydraulic conductivity) are attributed to the compaction and cooling history of the rocks. Because minimal compaction occurred in the lower subunits, where heat loss to the underlying cooler units caused the glass shards to lose their plasticity quickly, these subunits remained nonwelded and retained higher porosities. Analysis of the horizontal trends in the properties indicated a lack of substantial spatial correlation. However, because of the obvious vertical trends, vertical models were fit to the data sets in order to predict properties of the corresponding stratigraphic elevation in other locations. Prediction errors (measured minus predicted values) for three surface transects and five boreholes (N11, N37, N53, N54, and N55) are shown in Figure 5.3-37. The model for porosity was able to predict within 95 percent confidence intervals the porosity of samples from surface outcrops at locations within the potential repository block, south to Busted Butte, and north to Pagany Wash. The model poorly predicted porosities in the upper part of the vertical transect to the north in Yucca Wash, which is closer to the source caldera of the tuffaceous flows. Using properties of outcrop

samples was less successful in predicting properties measured on borehole samples, at least for the Tpcplnc and Tpcpv2, which comprise the welding transition zone. The error is the greatest for the borehole (N11) located closest to the source caldera, with the next highest error for the other borehole located to the north of the potential repository block. Apparently, the parts of the large-volume ashflow sheets closest to the source are more affected by complex depositional processes than the more distant parts.

The vertical trends apparent in the hydrologic properties of rocks near the base of the Tiva Canyon Tuff seem to be the result of deterministic processes, at least to the extent that regression equations based on measured physical and hydrologic properties can reasonably predict properties of this zone in other locations at Yucca Mountain. The results of this study support the use of properties such as porosity, that display deterministic trends, to correlate with parameters that are more difficult to measure or predict for spatially distributing properties for site-scale numerical flow models. The regression models describing trends in hydrologic properties provide important insights for process modeling that are not readily discernable based solely on the sets of measured values.

Vertical Variability From Exploratory Studies Facility Sampling–To evaluate the variability of matrix properties of the Topopah Spring Tuff at the potential repository horizon, a lithostratigraphic section encompassing the Tptpmn (TMN) and upper part of the Tptpll (TLL) was sampled systematically in the Exploratory Studies Facility Main Drift. From north to south, this section encompasses a gradual transition from the upper lithophysal zone (Tptpul) down through a series of vapor-phase partings or stringers, through a highly fractured zone, through a pumice swarm at the base of the Tptpmn, and then through a relatively sharp transition zone into the lower lithophysal zone (Figure 5.3-38). At this location in the Exploratory Studies Facility, the highly fractured zone does not exhibit the lithophysal-bearing subzone that exists farther to the north and partially exists to the south. For comparison, porosity profiles (from relative humidity drying) of the Tptpmn for three nearby boreholes (SD-7, SD-12, and SD-9) are shown in Figure 5.3-39a. Although the porosity profile for the northern most borehole (SD-9) shows considerable scatter of values in the lithophysal-transition zone, the lithophysal-bearing subzone is readily apparent in the greater porosity values near the midpoint of the profile (Figure 5.3-39a). However, farther south at borehole SD-12, porosities within this same subzone of the Tptpmn are lower and fairly uniform (about 11 percent) throughout the subzone. In the Exploratory Studies Facility Main Drift, 142 rock samples were collected systematically every 20 m and analyzed for physical properties, including porosity (Figure 5.3-39b). Thus far, saturated hydraulic conductivity has been measured on 60 of the 142 samples. The same trends in porosity observed in the boreholes are apparent in the Exploratory Studies Facility transect in that there is substantial scatter in the lithophysal transition zone (Exploratory Studies Facility station 2800 to 2500) and then an increase in porosities (from relative humidity drying) from about 9 percent to about 12 percent midway through the transect. Saturated-hydraulic-conductivity values from a preliminary data set (60 samples) measured using a steady-state, high-pressure permeameter correlate relatively well with porosity calculated with an r^2 value of 0.39. Mean porosity for the 60-sample subset from the Exploratory Studies Facility Main Drift systematic sampling transect is 7.9 percent (standard porosity is 9.9 percent) and mean saturated hydraulic conductivity is 4.7×10^{-12} m/s. A geostatistical analysis of the porosity data with distance provided an omnidirectional semivariogram, indicating that porosity changes distinctly with changes in elevation downward through the lithostratigraphic section (Figure 5.3-40). Based on this analysis,

porosity can be predicted accurately between measured points within a lithostratigraphic unit at subhorizontal distances up to about 800 m.

Areal Distribution of Porosity–Few of the hydrogeologic units present at Yucca Mountain were penetrated by a sufficient number of surface-based boreholes to determine whether or not areal trends in porosity are present. However, a large-scale analysis of areal distribution of porosity was done using kriging for the two hydrogeologic units that were penetrated by the largest number of boreholes, the welded Tiva Canyon Tuff (CW) and Pre-Pah Canyon/Topopah Spring bedded tuff (BT2). In the vicinity of the potential repository, CW has a relatively smooth increase in porosity from south (about 7 percent) to the north (about 12 percent) based on data from 15 boreholes (Figure 5.3-41a). In contrast, BT2 has a smaller areal distribution and a range in porosity from 37 percent at borehole SD-7 to 52 percent at the northern end of the repository block, based on data from 13 boreholes. The calculated areal variation in porosity for these two units is in contrast to the lack of lateral trends in the 2-D surface transect study of the base of the Tiva Canyon Tuff discussed above.

Vertical variability of matrix properties within hydrogeologic units can be reasonably described for most of the units as indicated in previous sections. Problematic units are very thin or layered units, or those with features larger than core-size samples that dominate the flow of water at the site scale. However, there is a large degree of uncertainty associated with the representations of areal variability of hydrologic properties because of the high degree of heterogeneity of the rocks and because the bulk of the data is from core-scale sized samples from 1-D boreholes sparsely distributed over the study area. The areal distribution of core-scale properties might adequately be represented for large-scale numerical flow models by the correlation of flow properties with surrogates such as porosity, which can be modeled using lithostratigraphic distributions and which has been shown to be related to flow properties (Istok et al. 1994; Rautman et al. 1995; Flint, L.E. et al. 1996b; Flint, L.E. 1998). The correlation of porosity with lithology provides a much larger database with which to calculate spatial distributions. This relationship is discussed in detail in the description of the property-modeling exercises contained in the following subsection.

5.3.3.1.1.8 Three-Dimensional Modeling of Rock-Property Heterogeneity

A somewhat different, alternative approach to the understanding of variability and spatial heterogeneity of rock material properties involved 3-D property modeling of the Yucca Mountain Site and was described by Rautman (1996) and Rautman and McKenna (1997). Rather than subdividing the volcanic tuffs at Yucca Mountain into a large number of slightly differing geologic units, the spatial positions and boundaries between which must be inferred from widely spaced drill holes, these authors examined and modeled the complete spatial distribution of material properties corresponding to several major rock-type groupings. The modeling of spatial variability described by Rautman and McKenna (1997) involved the assessment of the uncertainty associated with predictions of spatially varying material properties at unsampled locations within the subsurface of Yucca Mountain. In effect, the approach of Rautman and McKenna (1997) involves the generation of a suite of spatially variable material-property models, each of which is consistent with the observed data at the spatial locations of those data, and which vary stochastically away from data locations in a manner that is compatible with the quantitative spatial continuity description of the data. The concept is borrowed from more conventional uncertainty assessment in engineering

through Monte Carlo analysis. Recognizing that it is impossible to know with certainty the exact hydrologic properties of a sparsely sampled spatially heterogeneous domain, the performance consequences of a set of geologically plausible alternative models, all of which are compatible with knowledge of the site, was evaluated. Variation among a set of statistically indistinguishable heterogeneous models thus represents the uncertainty that results from less-than-exhaustive observation and description. Because material properties in geologic materials (including at Yucca Mountain) are generally spatially correlated, simulation of rock properties effectively amounts to drawing entire material-property models from some hypothetical "distribution" of alternative "realities." This conceptual framework for hydrogeologic description is represented in Figure 5.3-42.

A fundamental premise of the Monte Carlo simulation approach is that each individual realization is a plausible model of the unknown real world and that variation between the different stochastic realizations represents a variety of outcomes consistent with all that is known. Presumably, the only meaningful difference between realization 1 and realization N is that a different random number "seed" was used to initiate the simulation process definition of the random path and the generation of the various uniform random numbers that produce the output value at each node from the locally conditioned probability density function. Therefore, it is possible to test the validity of individual models in terms of statistical similarity to the data.

Recognizing that the term "validation" has a number of frequently controversial meanings within the modeling community, Rautman and McKenna (1997) restricted the meaning of this word to the following empirical tests of agreement between a simulated exhaustive model of reality and the underlying partially known (i.e., sampled) vision of reality obtained from site data.

A. Does the model reproduce the measured values of the particular attribute that were used to generate the model at the proper spatial position of those values?

B. Does the model as a whole reproduce the statistical character of the data ensemble used to condition the model?

C. Do the individual modeled attribute values exhibit spatial correlation consistent with the spatial continuity patterns exhibited by the sample data and with the geologic processes that produced the real world?

Drawing upon abundant evidence that the presence or absence of welding is the single greatest determinant of material properties at Yucca Mountain, Rautman and McKenna (1997) divided the near-surface volcanic section comprising the unsaturated and shallow saturated zones into dominantly welded and dominantly nonwelded categories. This binary classification scheme thus closely resembles older material-property-based stratigraphic and hydrologic subdivisions listed in Table 5.3-13 and the thermal/mechanical units shown in Figure 5.3-18. The three aggregated geologic units modeled by Rautman and McKenna (1997) consist of the PTn unit, the welded portion of the Topopah Spring Tuff, and the CHn unit including the Prow Pass Tuff. Each of the three aggregated geologic units consists of combinations of the detailed hydrogeologic units listed in Table 5.3-13. Matrix properties of the TCw unit were not modeled because water flow through this unit occurs dominantly through fractures.

The data used by Rautman and McKenna (1997) included both matrix-property measurements of core samples (described previously) and down-hole petrophysical measurements of in situ rocks (Nelson 1996a, 1996b; CRWMS M&O 1996a). In the latter case, geophysical measurements of density, water content, and water resistivity on a foot-by-foot basis were converted to porosity through accepted petrophysical relationships. The petrophysical measurements represent a substantially larger volume of rock (roughly a spherical volume 0.5 to 0.6 m in diameter) than that of most core samples (roughly a sphere 0.05 to 0.06 m in diameter). However, with the exception of measurements of the principal lithophysae-bearing intervals of the Topopah Spring Tuff, the measurements obtained by the two methods are quite similar and match one another quite closely where both are available for the same borehole interval. Core porosity values (from relative humidity drying) within lithophysae-bearing intervals were observed by Rautman and McKenna (1997) to be approximated quite closely in non-cored boreholes by a related petrophysical measurement, the so-called "water-filled" porosity of Nelson (1996a, 1996b). The difference between the water-filled porosity and the more conventional total petrophysical porosity in these lithophysae-bearing intervals is attributed to the presence of lithophysal cavities varying up to several tens of centimeters across. These cavities act to increase the (total) porosity of the bulk rock; however, under unsaturated hydrologic conditions, moisture cannot be held by capillary forces within these macro pores, and thus this petrophysical property can be used as a surrogate for core "matrix" porosity. The inclusion of petrophysical data increased the number of boreholes used to characterize material properties to 51, of which 37 provide deep penetrations of the volcanic section.

Statistical and Spatial Continuity Description–The statistical description of the modeling units defined by Rautman and McKenna (1997) is quite straightforward, and relies principally on construction of histograms and the corresponding cumulative distribution functions as probabilistic descriptions of the likelihood of obtaining differing material property values when sampling from a spatially variable hydrogeologic unit. Note that within this conceptual (probabilistic; Monte Carlo) framework, sampling variability is not equated with measurement "error" associated with some "true" property value. The true properties of the "real" rocks vary with the spatial position at which one observes those rocks.

Spatial continuity analysis builds on the geologic intuition that pairs of samples located relatively close to one another will resemble one another statistically more than pairs of samples located farther apart. Quantitative description of spatial continuity patterns (also frequently referred to as spatial correlation) employed standard variography techniques (Journel and Huijbregts 1978; Isaaks and Srivastava 1989), and was conducted separately for each of the three hydrogeologic modeling units. Experimental (semi) variograms were computed in three stratigraphic dimensions, corrected for post-depositional tilting and faulting (Rautman and McKenna 1997). The variogram estimator used by Rautman and McKenna (1997) was the semivariogram, g, which is traditionally defined as half the average squared difference between two attribute values approximately separated by a vector, h:

$$\gamma(\boldsymbol{h}) = \frac{1}{2N(\boldsymbol{h})} \sum_{i=1}^{N(\boldsymbol{h})} (Z_x - Z_{(x+h)})^2, \qquad \text{(Eq. 5.3-1)}$$

where Z_x is the value of a variable at a spatial location, x, and Z(x+h) is the value of that same variable located the vector distance, h, away. Note that with the exception of the factor of 2 and the fact that the comparison is between two sample values rather than between a sample value and a mean value, Equation 5.3-1 is identical to that used to estimate a variance. Thus, it is no coincidence that for many earth-science applications, the variogram estimator, g, is observed to converge on the variance of the data, s^2, as the separation distance between the pairs of samples being compared becomes very large. A sketch showing the various terms used to describe variogram features is shown in Figure 5.3-43a.

Anisotropy in the spatial-correlationpatterns of various material properties is expected in a layered sequence of rocks, such as those at Yucca Mountain. Continuity of rock properties in the stratigraphically horizontal direction(s) generally should be greater than in the vertical direction. Although standard geostatistical practice is to fit model variograms to the sample data individually in each desired direction, these separate models cannot be used directly as input to the simulation algorithm. Rather than attempt to compute the necessary spatial covariance values differently in different directions, it is typically the coordinate system describing samples and grid nodes that is modified to an isotropic system in which computation of the spatial covariance values is straightforward. This process can be conceptualized by envisioning an ellipsoid in three dimensions, or a simple ellipse in two dimensions, for which all points on the surface are at the same structural (geologic) distance, as opposed to the same Euclidean distance (Figure 5.3-43b). Anisotropy is then described in terms of the stretching and rotation that would be necessary to transform a sphere (circle) into an ellipsoid with specified major, intermediate, and minor axes. The coordinate system is then transformed using this information, the model values are generated, and the coordinate transformations are reversed.

Porosity data (105°C oven drying) obtained for samples of the PTn model unit are portrayed in histogram and cumulative-distribution-functionformat in Figure 5.3-44. A similar portrayal of the distribution of values for the two different types of porosity ("matrix" and "lithophysal") from the TSw unit is presented in Figure 5.3-45. Recall that "lithophysal porosity," as used by Rautman and McKenna (1997), means the porosity of volumes of rock many tens of centimeters in diameter, such that the porosity effect of large (centimeter scale and larger) lithophysal cavities is included. In contrast, the "matrix porosity" means the porosity equivalent to that measured for laboratory core samples using a relative humidity oven in which the size of the matrix pores is small enough that water is held in them under unsaturated (negative pressure) conditions. The distributions of both relative humidity oven-dried porosity and 105°C oven-dried porosity from the combined Calico Hills-Prow Pass unit are shown in Figures 5.3-46. Rautman and McKenna (1997) also were able to distinguish altered from unaltered rocks within the combined Calico Hills-Prow Pass unit on the basis of differences in porosity measured by relative humidity oven and 105°C oven methods (see Figure 5.3-26). A similar distinction between altered and unaltered rocks was made by Rautman and McKenna (1997) using the difference between the "water-filled" and "total-porosity" petrophysical measurements applied to non-lithophysae-bearing rocks. Histograms and cumulative frequency distributions of porosity for 105°C dried samples of altered and unaltered rocks from the Calico Hills-Prow Pass model unit are presented in Figure 5.3-47. Statistical summaries of all these porosity data is given in Table 5.3-16.

Spatial Continuity in the Paintbrush Nonwelded Hydrogeologic Model Unit–Sample variograms computed in the three directions of minimum, maximum, and intermediate continuity for total porosity in the PTn hydrogeologic model unit are presented in Figure 5.3-48, together with the fitted model computed in those same three directions. The vertical variogram (Figure 5.3-48a) is the best defined, and it exhibits modest "hole effect" for intersample distances of about 60 to 120 feet (Rautman and McKenna 1997). A hole-effect variogram is defined by a variance that increases and then decreases with increasing separation between pairs of samples; typically there is some sort of approximately periodic relationship between the peaks and troughs. The hole effect is so named because it is most frequently observed when computing sample variograms down vertical drill holes in a horizontally layered lithologic sequence. When variograms are computed for such a layered sequence, two relatively similar layers may be separated by a layer of somewhat different character. At small intersample distances, one is comparing essentially only samples from within the same unit. As the distance between sample pairs increases, one begins to compare across the different units, leading to higher computed variability. However, as the separation increases further, one begins to compare samples located in the more similar units straddling the intervening layer and the computed variability thus decreases.

The horizontal variograms of total porosity in the PTn hydrogeologic model unit (Figure 5.3-48b) as presented by Rautman and McKenna (1997) are less well defined than the vertical variogram. The maximum direction of spatial continuity is observed in the S 45° E (azimuth = 135°) direction, and these authors have inferred a modest degree of anisotropy. The orientation of the direction of maximum spatial correlation is attributed to the relative location of the source vents for the Yucca Mountain and Pah Canyon Tuffs, which are principal components of the PTn hydrogeologic model unit, to the northwest of the repository site itself. A set of three nested spherical variogram models plus a small nugget effect has been fitted to the sample data using the parameters specified in Table 5.3-17. The composite spatial correlation model appears to fit the sample variograms rather well, and the model is appropriately consistent with sample variograms computed in other directions (not shown).

Spatial Continuity in the Topopah Spring Welded Model Unit–Sample variograms computed using matrix-porosity data from the TSw model unit for the minimum, maximum, and intermediate directions of spatial continuity are presented in Figure 5.3-49 along with the fitted model variograms. Part (a) of Figure 5.3-49 is the variogram computed in the minimum continuity, stratigraphically vertical direction with a sample lag of 3 feet (~1 m), whereas part (b) shows sample variograms computed horizontally at azimuths of 0 and 90 degrees (north-south and east-west), using a lag-class interval of 2,000 feet. Maximum spatial correlation is observed in the north-south direction. The vertical variogram is particularly well defined because of the closely spaced sampling along borehole traces, whereas the horizontal variograms are somewhat more irregular, largely because of the irregular pattern of surface-based boreholes.

Sample variograms, with the appropriate fitted model values, for the minimum, maximum, and intermediate directions of spatial continuity for lithophysal porosity in the TSw model unit are presented in Figure 5.3-50. Again, the experimental plot in the minimum continuity, vertical direction is very well defined because of the close down-hole sample spacings, nominally 3 feet (1 m), and the figure indicates a very small nugget value as sample spacings decrease toward zero.

The horizontal variograms are more irregular because of the non-systematic spacing of the available boreholes; a mild anisotropy is observable between the sample values computed in the north-south and east-west directions.

The vertical variogram exhibits a very strong hole effect with an apparent period of about 200 feet. This phenomenon is easily attributable to the marked differences in porosity values between the lithophysal and nonlithophysal lithologic intervals. The significant lithophysae-bearing intervals typically have a stratigraphic thickness of this order of magnitude.

The sample variograms of lithophysal porosity values have been fitted with a series of three nested spherical variogram models using the parameters in Table 5.3-18. Only modest anisotropy has been modeled in the horizontal dimensions, but the vertical-to-horizontal anisotropy is quite strong because of the relatively stratiform lithophysal intervals. The search in the vertical direction has been limited to 220 feet because of the hole effect. The third nested structure, with a principal range of 50,000 feet is mostly a dummy structure, as mandated by the simulation requirement that the total sill be equal to 1.0. As can be observed in Figure 5.3-50, the contribution of this structure is virtually nil for separation distances less than about 5,000 feet (the variance contribution of the two shorter range structures totals 0.6), and the impact of this dummy structure becomes important only for separations greater than 10,000 to 12,000 feet.

The approximately 200 foot vertical range (= 7000 ± 0.0286 anisotropy ratio 2) for nested structure 2 is clearly related to the lithophysal-nonlithophysal layering. The origin of the approximately 30 foot vertical range (= 2,000 ± 0.015; prominent knick-point in Figure 5.3-50a) is less obvious, but because a structure with a range of this magnitude was also identified for the TSw matrix porosity data (Figure 5.3-49), it probably involves the intrinsic variability of porosity in welded tuff.

Spatial Continuity in the Combined Calico Hills-Prow Pass Model Unit–Sample variograms for total porosity from the combined Calico Hills-Prow Pass model unit are presented in Figure 5.3-51 for the directions of minimum, maximum, and intermediate spatial continuity. Continuity in the stratigraphically vertical minimum-continuity direction is shown in part (a) of the figure whereas part (b) shows spatial correlation in the north-south and east-west directions. Even though the Calico Hills-Prow Pass combined unit comprises unaltered (vitric) and altered (zeolitic) rock types, the presence of zeolitic alteration appears not to have affected the spatial distribution of total porosity in any meaningful manner. All data for porosity have been combined in Figure 5.3-51, thus achieving greater statistical mass. The coherent spatial continuity patterns also are evidence that this aggregation of porosity data from the different rock types is appropriate.

The vertical variogram of Figure 5.3-51a exhibits many of the same characteristics observed for vertical variograms in the other modeling units. There is a moderate hole-effect structure with an approximate periodicity of 200 feet, and the observed variability at all lags, up to somewhat more than half the total unit (stratigraphic) thickness of 800 feet, never reaches the expected sill value of 1.0. Much of the hole-effect character of the vertical variogram is attributed to the inclusion of a number of somewhat different lithologic units within the combined Calico Hills-Prow Pass model unit. The principal contributor to the hole-effect phenomenon is most likely the so-called Prow Pass welded thermal/mechanical unit of Ortiz et al. (1985). This unit appears to constitute "ashflow unit

3" of the Prow Pass Tuff (Table 5.3-13), as described by Moyer and Geslin (1995; see also Engstrom and Rautman 1996; Rautman and Engstrom 1996a, 1996b). Although Prow Pass unit 3 does appear to constitute an identifiable lithostratigraphic unit that is typically welded, the degree of welding is markedly less than that of the welded units within the overlying Paintbrush Group, and the effect of the welding process on porosity is entirely different. This distinction in welding based on material properties has been noted previously by Schenker et al. (1995, Appendix III). In most boreholes, the porosity values through much of the Prow Pass unit 3 interval approximate those of the nonwelded tuffs enclosing this unit.

The horizontal variograms of Figure 5.3-51b are quite well defined and exhibit a less irregular pattern than the horizontal variograms from the other two model units. Spatial continuity appears greatest in the azimuth = 0° direction (north-south) and the degree of anisotropy at longer lag spacings is quite pronounced. An interesting phenomenon apparent in the horizontal variograms is that although variability in the north-south direction appears to increase relatively steadily toward the sill value over separation distances of up to 25,000 feet, the variance in the east-west direction increases to maximum values at lags of about 5,000 and 10,000 feet and then decreases markedly and remains low.

The variograms shown in Figure 5.3-51 for the Calico Hills-Prow Pass model unit have been fitted with three nested spherical structures using the parameters given in Table 5.3-19. Three structures are clearly required by the vertical variogram of Figure 5.3-51a: the knick-point at a lag of about 50 feet, the prominent intermediate sill beginning with lags of about 175 to 200 feet, and a third to make the total variance equal to 1.0 at very long separations. Two nested structures would appear to suffice for modeling the horizontal variograms of Figure 5.3-51b. However, use of the three-structure model in the vertical direction mandates maintenance of this model in the horizontal plane as well, and there is little practical difference between the two longer range components. We have dealt with the hole effect in the vertical dimension by restricting the search in this direction during modeling to a maximum of about 180 feet, as noted in Figure 5.3-51a.

Results of Modeling–Rautman and McKenna (1997) generated 100 simulated models for each of the three model units using the sequential gaussian conditional simulation algorithm, SGSIM (Deutsch and Journel 1992). Stochastic porosity models were created by direct conditioning using virtually all available measured-porosity data. Simulated models of secondary rock properties demonstrated to be strongly correlated with porosity, specifically saturated hydraulic conductivity and thermal conductivity, were created using methods of linear co-regionalization (Journel and Huijbregts 1978; Altman et al. 1996; Rautman and McKenna 1997). The overall statistical and spatial continuity properties of the secondary variables are thus preserved, even though direct conditioning by measured secondary-property values is not possible with this technique. Rautman and McKenna (1997) also created summary models for each of their material properties by post-processing the replicate stochastic simulations in a manner similar to that illustrated conceptually in Figure 5.3-42. A summary cut-away perspective view showing the expected value of total (lithophysal) porosity for the Yucca Mountain extended site area is shown in Figure 5.3-52.

Because the material-property models of Rautman and McKenna (1997) are stochastic in nature, it is important to "validate" the underlying property values statistically, rather than relying solely upon

visual inspection of images, such as Figure 5.3-52. Figures 5.3-53 through 5.3-55 present selected elements of such a validation exercise, fully recognizing that "validity" of any particular model is dependent upon the use of that model in downstream activities.

Figure 5.3-53 illustrates that simulated vertical porosity profiles extracted from the model at the grid location closest to borehole USW SD-12 closely reproduce the measured porosity data, subject to limits imposed by the grid discretization. Figure 5.3-54 compares histograms for one particular simulated model (run number 34) with the histogram of all measured porosity data from the TSw model unit. Finally, Figure 5.3-55 compares exhaustive variograms for several simulated models with both the original experimental (data) variogram and its modeled representation provided as input to the simulation algorithm. Figures 5.3-53 through 5.3-55 also illustrate the smoothing and reduction of observed variability that is characteristic of summary-type expected-value models, such as those produced through post-processing of simulations or by standard kriging methodologies. If extreme values (high or low) from the overall measured material property distribution are important in the further modeling of physical processes, this reduction of variance may distort the results of those modeled physical processes.

5.3.3.1.2 Field-Scale Characterization of Hydrogeologic Units Using Air-Injection Testing

Air-injection testing has been and continues to be conducted in the unsaturated zone at Yucca Mountain to determine the field-scale bulk permeability, porosity, and anisotropy of the rock units that comprise, overly, and underlie the potential repository horizon. Permeability, along with gradient, controls the flux of fluids (gas and water) through Yucca Mountain; pneumatic porosity controls the velocity of pressure fronts; and anisotropy controls the direction of fluid flow through Yucca Mountain (LeCain et al. 1997). These are critical parameters required for all analytical and numerical models that attempt to predict fluid flow through the unsaturated zone at Yucca Mountain. For the nonwelded and bedded units (such as the PTn unit), air-injection tests provide field-scale measurements of the intergranular (or matrix) permeability and porosity. For the highly fractured units (such as the TSw unit), air-injection tests provide measurements of the bulk permeability and porosity of the fracture network. As described briefly in Subsections 5.3.2.1.3 and 5.3.2.1.6, air-injection test were conducted in surface-based, vertical boreholes and in horizontal boreholes drilled from alcoves of the Exploratory Studies Facility. Only single-hole tests were conducted in the surfaced-based boreholes because of the relatively large distances between them. However, in the Exploratory Studies Facility alcoves, the horizontal boreholes were drilled close enough together to have allowed cross-hole air-injection tests in addition to single-hole tests. In the Exploratory Studies Facility alcove boreholes, cross-hole gaseous tracer tests also were conducted to estimate effective porosity and tortuosity. Effective porosity controls the velocity of mass transport through Yucca Mountain and tortuosity controls the actual pathway of transport. Accurate effective porosity values are required to estimate transport times for fluid flow through the unsaturated zone at Yucca Mountain.

5.3.3.1.2.1 Surface-Based Air-Injection Testing

Air-injection testing using a downhole straddle-packer assembly and a surface-based data acquisition system was conducted in surface-based boreholes UZ-16, SD-12, NRG-6, and NRG-7a

(Figure 5.3-11). Test interval lengths were approximately 4 m and gas-injection rates ranged from 10 to 1,500 standard liters per minute. To minimize turbulence, injection-interval differential pressures were generally limited to less than 10 kPa (LeCain 1997).

Air-Injection Permeability Values by Borehole–Air-injection permeability values with depth for the test intervals in boreholes UZ-16, SD-12, NRG-6, and NRG-7a (LeCain 1997) are shown on Figures 5.3-56 through 5.3-59. Also shown on the figures are the lithostratigraphic units corresponding to the test intervals (Geslin et al. 1995; Moyer et al. 1995; Geslin and Moyer 1995). (For definition of lithostratigraphic and hydrogeologic units, see Table 5.3-13)

Borehole UZ-16–Borehole UZ-16 was tested November 1993 through May 1994. The borehole diameter was 0.31 m and total depth was 514.1 m. The test interval length was generally 4.0 m. The borehole penetrated the water table00 at approximately 489 m below land surface. The lithostratigraphic units tested extended from the crystal-poor lower-lithophysal zone of the Tiva Canyon Tuff down to the Calico Hills Formation. The PTn hydrogeologic unit was not tested because of caving of the borehole wall. Air-injection permeability values (Figure 5.3-56) ranged from 2.3×10^{-14} m^2 in the crystal-poor middle nonlithophysal zone of the Topopah Spring Tuff to 2.7×10^{-11} m^2 in the crystal-poor lower nonlithophysal zone of the Tiva Canyon Tuff. Most air-injection permeability values are between 10^{-13} m^2 and 10^{-11} m^2. The two highest air-injection permeability values, 2.7×10^{-11} m^2 and 1.5×10^{-11} m^2, were located in the shallow Tiva Canyon crystal-poor lower nonlithophysal zone. Air-injection permeability values of the Topopah Spring Tuff range from 2.3×10^{-14} m^2 to 9.5×10^{-12} m^2. The Topopah Spring crystal-poor upper lithophysal and middle nonlithophysal zones show a distinct decrease in permeability with increased depth. The Topopah Spring lower lithophysal and lower nonlithophysal zones indicate a decrease in permeability with increased depth. Figure 5.3-56 also shows a distinct shift in the air-injection permeability values at the Topopah Spring middle nonlithophysal, lower lithophysal contact.

Borehole UZ-16 was the only borehole tested that penetrated the Calico Hills Tuff. The air-permeability value of the Calico Hills was 1.7×10^{-14} m^2. The Calico Hills test showed a breakover pressure response indicating water redistribution. Based on the indication of water redistribution, the capillary pressure of the Calico Hills test interval was less than 43.3 kPa (LeCain 1997).

Borehole SD-12–Air-injection testings of borehole SD-12 were conducted from February 1995 through May 1995. The borehole diameter was 0.31 m, and total depth was 335.4 m. The test-interval length ranged from 4.6 to 4.9 m. The lithostratigraphic units extended from the crystal-poor lower lithophysal zone of the Tiva Canyon Tuff down to the crystal-poor lower nonlithophysal zone of the Topopah Spring Tuff. The PTn hydrogeologic unit and the lower lithophysal zone of the Topopah Spring Tuff were not tested because of caving of the borehole wall. Air-injection permeability values (Figure 5.3-57) ranged from 3.8×10^{-11} m^2 in the shallow Tiva Canyon crystal-poor lower lithophysal zone to 1.2×10^{-13} m^2 in the Topopah Spring crystal-rich nonlithophysal zone. Most air-injection permeability values were between 10^{-11} m^2 and 10^{-13} m^2. The Topopah Spring crystal-rich nonlithophysal and the crystal-poor upper lithophysal zones showed decreases in permeability with depth.

The results of several of the Tiva Canyon Tuff tests indicated water redistribution because pressure responses showed peaks and then subsequent decline in pressure. The tests indicated that water was present in the fractures of the Tiva Canyon Tuff at depths of 39.6, 42.4, and 77.1 m.

Borehole NRG-6–Air-injection testing was conducted in borehole NRG-6 during October and November 1994. The borehole diameter was 0.20 m and the total depth was 332.3 m. The test-interval lengths generally were 4.3 m. The lithostratigraphic units tested extended from the crystal-poor lower lithophysal zone of the Tiva Canyon Tuff down to the crystal-poor middle nonlithophysal zone of the Topopah Spring Tuff. The PTn hydrogeologic tuffs and much of the lower Topopah Spring crystal-poor upper lithophysal zone were not tested because of caving of the borehole wall. Air-injection permeability values (Figure 5.3-58) ranged from 2.8×10^{-11} m^2 in the Tiva Canyon crystal-poor lower lithophysal zone to 8.2×10^{-14} m^2 in the Topopah Spring crystal-rich nonlithophysal zone. Most air-injection permeability values were between 10^{-11} m^2 and 10^{-13} m^2. Sixteen of the test intervals had air-injection permeability values larger than 10^{-12} m^2, and four test intervals had values larger than 10^{-11} m^2. Of the four test intervals with values larger than 10^{-11} m^2 two were within 26 m of the ground surface. The Topopah Spring crystal-rich nonlithophysal zone showed a distinct decrease in permeability with depth. The Topopah Spring crystal-poor middle nonlithophysal zone showed an increase in permeability with depth.

Borehole NRG-7a–Air-injection testing was conducted in NRG-7a from July through September 1994. The borehole diameter varied with depth, measuring 0.20 m from the ground surface to 91.5 m and 0.15 m from 91.5 m to 379.9 m. The test-interval lengths of the 0.20 and 0.15 m diameter sections were 4.3 and 3.5 m, respectively. The lithostratigraphic units tested extended from the crystal-poor lower nonlithophysal zone of the Tiva Canyon Tuff down to the crystal-poor lower lithophysal zone of the Topopah Spring Tuff. The borehole wall was in good condition, and most of the 0.20 and 0.15 m diameter sections of the borehole were tested. Air-injection permeability values (Figure 5.3-59) ranged from 3.5×10^{-14} m^2 in the Topopah Spring crystal-rich nonlithophysal zone to 5.4×10^{-11} m^2 in the Tiva Canyon crystal-poor lower nonlithophysal zone. Most air-injection permeability values were between 10^{-13} m^2 and 10^{-12} m$_2$. Seven of the test intervals had values greater than 10^{-12} m^2. Of the seven, five were within 60 m of the ground surface. The three test intervals with air-injection permeability values larger than 10^{-11} m^2 were within 20 m of the ground surface. The Topopah Spring crystal-poor middle nonlithophysal zone shows a decrease in permeability with depth.

Borehole NRG-7a was the only borehole where the borehole wall was in sufficiently good condition to allow tests of the PTn hydrogeologic unit (Tpbt4, Tpy, Tpbt3, Tpp, and Tpbt2 in Figure 5.3-59). The nonwelded tuff air-injection permeability values ranged from 1.2×10^{-13} m^2 at the bottom of the Pah Canyon Tuff and Tiva Canyon crystal-poor vitric 1 subzone to 3.0×10^{-12} m^2 in bedded tuff number 3. The Yucca Mountain and Pah Canyon Tuffs showed decreases in permeability with depth.

Air-Injection Permeability Values by Lithostratigraphic Unit

Tiva Canyon Tuff–A statistical summary of the Tiva Canyon Tuff air-injection permeability values by borehole is presented in Table 5.3-21 (LeCain 1997). All test intervals were located in zones of

the crystal-poor member of the Tiva Canyon Tuff; including the lower lithophysal, the lower nonlithophysal, and crystal-poor vitric. The database is small with only four test intervals in each of boreholes UZ-16, NRG-6, and NRG-7a. The air-injection permeability values ranged from 0.24 $\times 10^{-12}$ m^2 in NRG-7a to 54.0 $\times 10^{-12}$ m^2 in NRG-7a. The mean air-injection permeability value for all boreholes was 12.0 $\times 10^{-12}$ m^2.

Based on the assumption that the borehole samples are log normally, distributed, and have similar variances (two assumptions that cannot be supported by the small data set) an analysis of variance between samples of the natural log air-injection permeability values gives a p-value of 0.71. The p-value indicates that the mean permeability values of the four boreholes are not statistically different. The absence of a statistically significant difference between the means may be because there really is no difference or may be the result of the small data set.

A statistical summary of the Tiva Canyon Tuff air-injection permeability values by lithostratigraphic unit and borehole is presented in Table 5.3-20. The total number of test intervals in Table 5.3-21 does not equal the number in Table 5.3-21 because test intervals that straddled two lithostratigraphic units are not included in Table 5.3-20.

Paintbrush Nonwelded Hydrogeologic Unit–A statistical summary of the air-injection permeability values of the PTn hydrogeologic unit is presented in Table 5.3-22 (LeCain 1997). All values are from borehole NRG-7a. The summary includes test intervals in the Yucca Mountain Tuff, Pah Canyon Tuff, the three bedded tuffs, and two test intervals in the crystal-poor nonwelded vitric zone of the Tiva Canyon Tuff. The nonwelded tuff air-injection permeability values ranged from 1.2 $\times 10^{-13}$ m^2 at the bottom of the Pah Canyon Tuff and the crystal-poor vitric zone of the Tiva Canyon Tuff to 3.0 $\times 10^{-12}$ m^2 in the Pre-Yucca Mountain Tuff bedded tuff (Tpbt3). The mean air-injection permeability value was 0.54 $\times 10^{-12}$ m^2.

A statistical summary of the air-injection permeability values for individual lithostratigraphic units within the PTn is presented in Table 5.3-23. Test intervals that straddle lithostratigraphic units are not included. Exceptions were made for the Pre-Pah Canyon Tuff bedded tuff (Tpbt2) and the Pre-Tiva Canyon Tuff bedded tuff (Tpbt4) test intervals, which were thinner than the length of the test interval.

Air-injection permeability values for the PTn unit with depth are shown on Figure 5.3-60. The largest air-injection permeability value (3.0 $\times 10^{-12}$ m^2) was in Tpbt3, and the lowest value (0.12 $\times 10^{-12}$ m^2) was in the crystal-poor vitric zone of the Tiva Canyon Tuff and in the lower Pah Canyon Tuff. The geology and rock structure log (SNL 1995b) describes the Tpbt3 as "course grained and weakly consolidated" with no fractures, indicating that flow in Tpbt3 is through the matrix. The log describes the Tiva Canyon crystal-poor vitric test interval as a "welding transition zone." The low permeability probably is the result of partial welding. The Yucca Mountain and Pah Canyon Tuffs both showed decreased permeability with increased depth, consistent with a decrease in welding with depth, as discussed in Subsection 5.3.3.1.1.7 above. The log describes the Yucca Mountain Tuff as "weak to partially welded." The logs do not report welding in the Pah Canyon Tuff.

Statistical analysis of the air-injection permeability values for the PTn unit indicates that, although slightly skewed by the relatively larger values for the Pre-Yucca Mountain Tuff bedded tuff (Tpbt3), the distribution is reasonably log normal (LeCain 1997).

Topopah Spring Tuff–A statistical summary of air-injection permeability values by borehole for the entire Topopah Spring Tuff is presented in Table 5.3-24. Permeability values ranged from 0.02 $\times 10^{-12}$ m^2 in the crystal-poor middle nonlithophysal zone of borehole UZ-16 to 33.0 $\times 10^{-12}$ m^2 in the crystal-rich lithophysal zone of borehole SD-12. The borehole mean permeability values ranged an order of magnitude from 4.7 $\times 10^{-12}$ m^2 in borehole SD-12, to 0.4 $\times 10^{-12}$ m^2 in borehole NRG-7a.

More detailed information on air-injection permeability values for the Topopah Spring Tuff is shown in Table 5.3-25 where the data are presented by lithostratigraphic unit. Test intervals that straddled lithostratigraphic units are not included in Table 5.3-25.

Because the data set of air-injection permeability values for the Topopah Spring Tuff is larger than that of either the Tiva Canyon Tuff or the PTn unit, more rigorous statistical analysis was possible, including comparisons of permeability values among the boreholes. Figure 5.3-61 presents histograms of air-injection permeability values and basic statistics by borehole for the Topopah Spring Tuff. The histograms show that the distributions are not normal because most of the air-injection permeability values plot on the left. The distributions are skewed to the right by a small number of higher permeability values.

However, histograms of the natural log air-injection permeability values (Figure 5.3-62) indicate that the natural log permeability values could be considered normally distributed. Analysis of variance between the mean values of the boreholes gives a p-value less than 0.01, indicating that at a 99 percent confidence level, at least one of the natural log permeability-value means is statistically different. Examination of Figures 5.3-61 and 5.3-62 indicates that the borehole NRG-7a permeability values generally are smaller than those for the other boreholes. The borehole NRG-7a histogram is shifted to the left and has a smaller variance. Analysis of variance between the means of the natural-log values from boreholes UZ-16, SD-12, and NRG-6 gives a p-value of 0.10, indicating that there is no statistical difference among the geometric-mean permeability values for these boreholes. Therefore, it seems that the permeability values from borehole NRG-7a represent a different population, at least from a statistical standpoint. Coincidently, borehole NRG-7a is the northernmost of the four boreholes and is closest to the boundary between the Yucca Crest and Azreal Ridge structural subdomains (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164, section titled "Structural Setting of the North Ramp of the Exploratory Studies Facility"). In the Azreal Ridge subdomain older cooling joints are more predominant than younger tectonic fractures.

Relation of Fracture Density, Lithophysal Cavities, and Core Recovery to Air-Injection Permeability–The average number of natural fractures per test interval, by lithostratigraphic unit and borehole, are shown in Table 5.3-26. The fracture data for boreholes UZ-16 and SD-12 are from the Yucca Mountain Project Sample Management Facility structural logs (CRWMS M&O 1995b,

1996b) and the fracture data for boreholes NRG-6 and NRG-7a are from the Agapito and Associates geology and rock structure logs (SNL 1995a, 1995b).

Although the number of fractures per test interval is known to be inaccurate because of missing core, the data can be used for comparisons, if it is assumed that any potential bias is independent of rock unit. Table 5.3-26 indicates that boreholes UZ-16 and SD-12 have similar fracture densities in both the Tiva Canyon and Topopah Spring Tuffs, whereas in boreholes NRG-6 and NRG-7a fracture densities in Tiva Canyon Tuff are about four times larger than in the Topopah Spring Tuff. The NRG boreholes are both located on the edge of Drill Hole Wash near the boundary between the Yucca Crest and Azreal Ridge structural subdomains (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164, section titled "Structural Setting of the North Ramp of the Exploratory Studies Facility"), whereas boreholes UZ-16 and SD-12 are located farther south, well within the Yucca Crest subdomain. The differences in fracture density and air-injection permeability between boreholes may be related to their location with respect to these structural subdomains and the structural styles that distinguish them.

A series of univariate regression analyses were performed between air-injection permeability values and six explanatory variables: total fractures, natural fractures, indeterminate fractures, and percentages of lithophysal cavities, core rubble, and core lost (LeCain 1997). The analyses resulted in no correlation between permeability and the number of indeterminate fractures, percentage of lithophysal cavities, core rubble, nor core lost. However, the regression analyses indicated a 0.22 and 0.24 goodness-of-fit between the number of natural fractures and permeability for boreholes NRG-6 and NRG-7a. This correlation is evident in the clustering of data points in Figures 5.3-63 and 5.3-64, which are semilog plots of air-injection permeability values and the number of natural fractures per test interval for boreholes NRG-6 and NRG-7a.

5.3.3.1.2.2 Exploratory Studies Facility Air-Injection Testing

As described in Subsection 5.3.2.1.6, air-injection and gaseous tracer testing has been conducted in test alcoves constructed off of the North Ramp of the Exploratory Studies Facility (Figure 5.3-16). Although initial studies in some of these alcoves are complete, studies in several other alcoves off of the Exploratory Studies Facility North Ramp and Main Drift are continuing. Therefore, only results from studies in the first three alcoves are included in this report. These alcoves include the upper Tiva Canyon alcove, the Bow Ridge fault alcove, and the upper Paintbrush contact alcove. Testing was conducted in the upper Tiva Canyon alcove to confirm the gaseous-phase properties of the Tiva Canyon Tuff determined through investigations in surface-based boreholes. Testing in the Bow Ridge fault alcove was intended to determine the properties of the Bow Ridge fault zone and the extent to which it may be a fast pathway for the flow of fluids (gas and water). Testing in the upper Paintbrush contact alcove was intended to determine the properties of the transition zone between the TCw hydrogeologic unit and the PTn hydrogeologic unit. The properties of the TCw hydrogeologic unit-PTn hydrogeologic unit contact are important because of the potential for lateral flow of water at this horizon above the potential repository (see Subsection 5.3.3.1.1.5).

Air-Injection Permeability, Porosity, and Water-Redistribution Pressures by Alcove–Air-injection testing using a downhole straddle packer assembly was conducted in boreholes drilled from Exploratory Studies Facility alcoves, upper Tiva Canyon alcove, Bow ridge fault alcove, and upper Paintbrush contact alcove (LeCain 1998; LeCain et al. 1997; LeCain and Patterson 1997). The test interval lengths ranged from 1 to 5 m. Gas-injection rates ranged from 10 to 800 standard liters per minute. Testing was conducted at increasing injection pressures to identify water redistribution. Injection-interval differential pressures were up to 165 kPa in the upper Tiva Canyon alcove boreholes, 64 kPa in the Bow Ridge fault alcove, and 111 kPa in the upper Paintbrush contact alcove. Turbulence was identified as a decrease in calculated permeability with increasing flow rate and corrected. Single-hole and cross-hole air-injection tests were conducted in the upper Tiva Canyon alcove and the Bow Ridge fault alcove to determine field-scale air permeability and, in the Bow Ridge fault alcove, porosity. In addition, cross-hole gaseous tracer tests were conducted in the Bow Ridge fault alcove to determine field-scale effective porosity. Because not all of the planned boreholes have been drilled to date, only single-hole air-injection testing has been conducted in the upper Paintbrush contact alcove.

Upper Tiva Canyon Alcove–The upper Tiva Canyon alcove was constructed in the Tiva Canyon crystal-poor upper lithophysal unit (Tpcpul), which is moderately to densely welded and fractured (LeCain et al. 1997). Three subhorizontal boreholes were drilled in the upper Tiva Canyon alcove for the radial borehole test: RBT#1, RBT#2, and RBT#3. Borehole RBT#1 penetrates an ashflow tuff, pale reddish-brown to pale red, densely welded, and devitrified with 10 to 20 percent pumice, 2 to 5 percent phenocrysts, and 10 to 12 percent lithophysal cavities. Borehole RBT#2 penetrates an ashflow tuff, pale reddish-brown, grayish-red, and brownish-gray, moderately to densely welded, and devitrified with 10 to 20 percent pumice, 2 to 5 percent phenocrysts, and 10 percent lithophysal cavities. Borehole RBT#2 penetrates an ashflow tuff, pale reddish-brown, grayish-red, and brownish-grey, moderately to densely welded, and devitrified with 5 to 10 percent pumice, 2 to 10 percent phenocrysts, and 5 percent lithophysal cavities. At a depth of 13.7 m the phenocrysts content increases to 10 percent. Borehole RBT#3 penetrates an ashflow tuff, pale red to grayish-red, moderately to densely welded, and devitrified with 3 to 5 percent pumice, decreasing with depth, 3 to 7 percent phenocrysts, and 8 to 15 percent lithophysal cavities (LeCain et al. 1997).

Single-hole air-injection testing was conducted in the upper Tiva Canyon alcove during August 1995 in horizontal boreholes RBT#1, RBT#2, and RBT#3, all of which are about 30 m deep (LeCain et al. 1997). Water-redistribution pressures were monitored during the tests to estimate near-field capillary pressure, which is discussed in more detail below. Because the boreholes had very high rugosity due to the lithophysal cavities and caving, the ability to isolate test intervals with pneumatic packers was limited and resulted in a limited number of test intervals. Successful air-injection tests were conducted on 25 percent of RBT#1, 55 percent of RBT#2, and 21 percent of RBT#3.

Air-permeability values of the three boreholes range from 0.2×10^{-12} m^2 to 85.0×10^{-12} m^2 with an arithmetic mean of 28.6×10^{-12} m^2 and geometric mean of 16.0×10^{-12} m^2. Nine test intervals indicated water redistribution at pressures less than 62.9 kPa and one test interval in borehole RBT#2 had a water-redistribution pressure less than 4.5 kPa. Composite histograms and statistics for air-permeability values and their natural logarithms for the three boreholes in the upper Paintbrush contact alcove are shown is Figure 5.3-65. The histograms are inconclusive due to the small

database and indicate that the values are not normally distributed. The apparent lack of a log-normal distribution disagrees with the conclusions of LeCain (1997) which indicate that the distribution of the Tiva Canyon permeability values from surface-based boreholes are log normal. The disagreement may be due to the limited number of intervals tested in the upper Paintbrush contact alcove boreholes due to the poor condition of borehole walls.

Table 5.3-27 presents a statistical summary of the permeability values of the upper Paintbrush contact alcove radial boreholes.

Although cross-hole air-injection testing was conducted between the three radial boreholes in the upper Paintbrush contact alcove from May to July 1995, these tests found no pneumatic connections between the air-injection intervals and the monitor intervals, thus precluding gaseous tracer tests. Analytical predictions based on an equivalent porous medium using the single-hole air-permeability values indicated that the gas-injection rates and pressure transducer sensitivity should have been sufficient to provide cross-hole pressure responses. Thus the failure of the cross-hole tests was attributed to the complex nature of the fracture network. To help understand these complexities, Anna (1998) used the Exploratory Studies Facility fracture mapping data to develop a stochastic fracture network of the upper Paintbrush contact alcove. An average of 422 fractures were generated and boreholes RBT#1, RBT#2, and RBT#3 had an average of 20, 22, and 21 single-fracture intersections, respectively. An average of 53 fracture networks were connected to at least one borehole. However, the connections between boreholes were sparse and composed of fractures connected in series. The simulations indicated that few fractures are connected from source to sink, implying that only a small percentage of the fractures comprise the flowpath from source to sink. Twenty realizations were simulated to examine the fracture pneumatic connections between the radial boreholes. Six realizations showed no fracture pneumatic connections, four showed a single pathway connection from source to sink, and ten showed two pathway connections from source to sink.

The fracture-network modeling indicates that at the scale of the radial boreholes (3 to 10 m), cross-hole testing must be analyzed using discrete fracture methods and that monitor intervals must isolate individual fractures (LeCain et al. 1997). Further, the absence of any cross-hole pressure responses were due to the combination of a limited number of cross-hole fracture connections and the inability to isolate discrete fractures in the monitor and air-injection boreholes. The caving and high rugosity of the boreholes severely limited the packer location and spacing and made it impossible to isolate individual fractures. The non-ideal packer placement resulted in the lengths of the monitor intervals being expanded to up to 4 m and therefore the intervals intersected numerous fractures. The pressure responses in the long monitor intervals represented a composite pressure response of several fractures as opposed to a discrete pressure response representing a single fracture. Even if a fracture did connect a monitor interval and injection interval, the other fractures would act as constant-head boundaries preventing any pressure increase in the monitor interval. The absence of any cross-hole pneumatic connections meant the cross-hole tracer testing could not be conducted. Identification of connected pneumatic flowpaths is required before successful tracer tests can be conducted in the upper Paintbrush contact alcove boreholes.

Bow Ridge Fault Alcove–The Bow Ridge fault alcove was constructed on the east side (footwall) of the Bow Ridge fault in the Tiva Canyon crystal-poor middle nonlithophysal zone (Tpcpmn), which is densely welded and fractured (LeCain et al. 1997). The Bow Ridge fault, which is a normal fault with approximately 128 m offset, is located 199 m into the Exploratory Studies Facility from the North Portal and has a strike of 180° and a dip of 75°. On the west side (hanging wall) of the fault, the younger, nonwelded, Tertiary pre-Rainier Mesa bedded tuff #1 (Tmbt1) has been dropped down to the same elevation as the Tpcpmn and the underlying Tiva Canyon crystal-poor lower lithophysal zone (Tpcpll). Two boreholes were drilled for the Hydrologic Properties of Major Faults Test: HPF#1 and HPF#2. Figure 5.3-66 presents a cross section of the alcove, fault, geologic contacts, and borehole HPF#1. The fault zone is approximately 2.7 m wide and contains three breccia zones. The first 10.7 m of borehole HPF#1 is in the Tpcpmn, which is an ashflow tuff, pale red to grayish-red, densely welded, and devitrified with 5 percent pumice, 5 percent phenocrysts, and one percent lithophysae. The contact between the Tpcpmn and the Tiva Canyon crystal-poor lower lithophysal zone (Tpcpll) was encountered at a depth of 10.7 m. From 10.7 to 15.6 m the lithology changes to an ashflow tuff, pale red to grayish-red, densely welded, devitrified with 3 to 5 percent pumice, 5 percent phenocrysts and 10 to 15 percent lithophysae. The contact between the Tpcpll and the Bow Ridge fault zone was encountered at a depth of 15.5 m and the fault zone extended from 15.6 to 17.3 m. The contact between the Bow Ridge fault zone and the Tmbt1 was encountered at a depth of 17.3 m. The Tmbt1 is a bedded/reworked tuff, white to very light gray, unconsolidated, vitric, with 75 percent pumice and 10 percent phenocrysts. The lithology and depth to contacts in borehole HPF#2 were almost identical to borehole HPF#1 (LeCain et al. 1997).

Single-hole air-injection testing was conducted in the Bow Ridge fault alcove in borehole HPF#1 in December 1995 and January 1996. Water-redistributionpressures were monitored during the tests to estimate near-field capillary pressure, which is discussed in more detail below. Testing was conducted in the Tpcpmn and Tpcpll, the Bow Ridge fault zone, and the Tmbt1. Permeability values of the eight test intervals in the Tpcpmn ranged from 6.0×10^{-12} m^2 to 26.4×10^{-12} m^2. Permeability values of the five test intervals in the Tpcpll ranged from 0.6×10^{-12} m^2 to 2.0×10^{-12} m^2. Permeability values of the three intervals in the fault zone ranged from 8.0×10^{-12} m^2 to 15.8×10^{-12} m^2. Permeability values of the two Tmbt1 test intervals were 41.3×10^{-12} m^2 and 22.0×10^{-12} m^2. Table 5.3-28 presents a statistical summary of the HPF#1 Tpcpmn and Tpcpll air-permeability values. The results indicate that the permeability of the Tpcpmn is about an order of magnitude greater than that of the Tpcpll. Assuming that the populations are log normally distributed, an analysis of variance between the two zones gives a p-value of 0.006, confirming that the mean values are statistically different. Comparing the mean permeability values of the Tpcpmn and the Tpcpll with the permeabilities of the Bow Ridge fault zone and the Tmbt1 indicates that the permeability of the Tpcpll is 5 to 10 times smaller than the other units. This may be due to less intense fracturing of the Tpcpll associated with the Bow Ridge fault because the lithophysal cavities of the Tpcpll may have absorbed some of the tectonic stress.

Eleven test intervals indicated water redistribution at pressures less than 63.6 kPa and one test interval had a water-redistribution pressure of less than 5.6 kPa. Testing in the fault breccia and Tmbt1 did not indicate water redistribution. The absence of water redistribution may be because of higher capillary pressures associated with the increased porosity of the nonwelded Tmbt1, or it may simply be because the high-permeability values limited the maximum injection pressures.

Cross-hole air-injection tests were conducted in the Bow Ridge fault alcove between boreholes HPF#1 and HPF#2 during June and July 1996. A total of 13 cross-hole tests were conducted with HPF#1 as the injection borehole and HPF#2 as the monitor borehole, but several of these tests were not valid because the injection intervals used did not indicate pneumatic connections to the monitor intervals in HPF#2. Cross-hole testing was conducted in the Tpcpmn, Tpcpll, Bow Ridge fault zone and Tmbt1. Table 5.3-29 presents the permeability and porosity values from cross-hole tests 4, 5, 6, 8, and 11 for both the spherical-flow type curve analysis and the steady-state analysis. Tests conducted in the fractured Tiva Canyon Tuff (Tpcpmn and Tpcpll) had transient periods of less than 15 seconds, which were too short to provide a reasonable type-curve match. Overall, the cross-hole tests indicate that although the air permeability of the Bow Ridge fault zone, the pre-Rainier Mesa bedded tuff, and the middle nonlithophysal are all about the same, their permeability is 5 to 10 times greater than that of the lower lithophysal unit. Furthermore, the cross-hole air-injection permeability values agree with the single-hole results and indicate that the scale differences between the two sets of tests (1 versus 3 m) did not effect the test results.

Convergent cross-hole tracer testing was conducted in the Bow Ridge fault alcove between boreholes HPF#1 and HPF#2 during June and July 1996 using sulfur hexafluoride (SF_6) as the tracer gas. Although six cross-hole tracer tests were conducted, only four tests yielded usable analytical results (Table 5.3-30). (In test 1, the system for releasing the slug of tracer gas malfunctioned and the results were not usable. In test 4, in which the tracer slug was released into the Tmbt1, no tracer had arrived after a duration more than twice the estimated tracer travel time and the test was terminated.) In Table 5.3-30, the tracer velocity is an average velocity based on the peak arrival time and the darcy velocity is based on the pneumatic permeability and pneumatic gradient.

The effective porosity values of the fault zone determined from tests 2 and 3 are larger than their corresponding pneumatic porosity values from cross-hole air-injection tests 4 and 5 (see Table 5.3-29). The larger effective porosity values indicate increased tortuosity and/or adsorption of the tracer. The darcy velocity values and subsequent effective porosity values of cross-hole tracer tests 5 and 6 are based on the steady-state permeability values obtained from cross-hole air-injection tests 8 and 11 (see Table 5.3-29). Because type curve analysis of cross-hole air-injection tests 8 and 11 was not possible, there are no analytical pneumatic porosity values. However, the 0.12 and 0.04 effective porosity values of cross-hole tracer tests 5 and 6 are several orders of magnitude larger than the general range of 10^{-3} to 10^{-5} fracture porosity of fractured rock (LeCain et al. 1997). The large effective porosity values indicate tortuosity and/or adsorption of the tracer gas. Because of its low cation exchange capacity, it is unlikely that adsorption of SF_6 occurred within the Tiva Canyon Tuff (LeCain et al. 1997). However, the clay and sand-size Tmbt1 material that comprises the fault-zone breccia matrix probably is responsible for retarding the tracer gas, resulting in the relatively large effective porosity of the fault zone. Adsorption may also explain the loss of all tracer gas during cross-hole tracer test 4, conducted in the Tmbt1, and the identical first arrival times yet different peak arrival times of tests 2 and 3. The tracer released in test 2 may have occupied adsorption sites so that during test 3 less adsorption occurred, resulting in faster average velocity. The low SF_6 adsorption properties of the Tiva Canyon Tuff indicate that the high-effective porosity values in test 5 and 6 may be due to increased tortuosity associated with gaseous flow through the fracture network.

Upper Paintbrush Contact Alcove–The upper Paintbrush contact alcove was constructed in the moderately welded Tiva Canyon crystal-poor lower nonlithophysal columnar subzone (Tpcplnc) in the vicinity of the contact between the TCw and PTn (LeCain and Patterson 1997). To date, only two of the boreholes planned for the upper Paintbrush contact alcove radial boreholes test have been drilled: RBT#1 and RBT#4. Borehole RBT#1 was drilled generally westward from the upper Paintbrush contact alcove in a "down-section" direction so as to penetrate the upper subunits of the PTn (Figure 5.3-67). Borehole RBT #4 was drilled generally eastward from the upper Paintbrush contact alcove in an "up-section" direction so as to penetrate the lower subunits of the TSw. The first 4.8 m RBT#1 are in the Tpcplnc, which is an ashflow tuff, pale red to light brown, devitrified, with 8 to 15 percent pumice, and 3 to 5 percent phenocrysts. The welding decreases and the porosity increases with depth. At a depth of 4.2 m, RBT#1 intersects the Tiva Canyon crystal-poor vitric subzone 2 (Tpcpv2) and at a depth of 12.2 m the borehole intersects the Tiva Canyon crystal-poor, vitric subzone 1 (Tpcpv1). The Tpcpv1 and Tpcpv2 are pyroclastic-flow deposits, grayish orange to dark gray, vitric, with 3 to 12 percent pumice, 3 to 6 percent phenocrysts, within an argillic altered matrix. Welding decreases with depth, becoming nonwelded at the Tpcpv2-Tpcpv1 contact at a depth of 25.3 m. The Tpcpv1 and Tpcpv2 are the uppermost subunits of the PTn (see Table 5.3-13), although the Tpcpv2 is a transition zone at many locations and generally is considered part of TSw. The first several meters of borehole RBT#4 are in the Tpcplnc, which is an ashflow tuff, grayish-pink, devitrified, with 3 to 15 percent pumice, and 3 to 5 percent phenocrysts and increasing welding with depth. At a depth of approximately 9 m borehole RBT#4 intersects the Tiva Canyon crystal-poor lower nonlithophysal hackly subzone (Tpcplnh) and the fracture pattern changes from columnar to hackly (LeCain and Patterson 1997).

Single-hole air-injection testing was conducted in the upper Paintbrush contact alcove boreholes RBT#1 and RBT#4 from April through June 1996 (LeCain and Patterson 1997; LeCain 1998). The lithostratigraphic units tested were the Tpcplnh, Tpcplnc, Tpcpv2, and Tpcv1. All test intervals were one meter in length except for bottom-hole test intervals, which were 3.2 and 5.0 m in boreholes RBT#1 and RBT#4, respectively. Water-redistribution pressures were monitored during the tests to estimate near-field capillary pressure, which is discussed in more detail below. Permeability values of the Tpcplnh ranged from 0.1 to 12.0 × 10^{-12} m^2 with an arithmetic mean of 3.69 × 10^{-12} m^2 and a geometric mean of 2.07 × 10^{-12} m^2 (Table 5.3-31). Permeability values of the Tpcplnc ranged from 0.02 to 2.0 × 10^{-12} m^2 with an arithmetic mean of 0.71 × 10^{-12} m^2 and a geometric mean of 0.26 × 10^{-12} m^2. Permeability values of the Tpcpv2 and Tpcpv1 ranged from 0.40 to 57.0 × 10^{-12} m^2 with an arithmetic mean of 16.5 × 10^{-12} m^2 and a geometric mean of 7.01 × 10^{-12} m^2. The relatively large permeability values of the Tpcpv subzones, compared to the Tpcpln subzones, indicate significant fracturing in the less welded Tpcpv. Twenty-two test intervals in boreholes RBT#1 and RBT#4 indicated water redistribution at pressures less than 95.2 kPa and one test interval in borehole RBT#1 had a water-redistribution pressure of less than 5.2 kPa.

Water-Redistribution Pressures as an Indication of Capillary Pressures–During air-injection tests, occurrence of a "breakover" pressure response, or a decrease in pressure with time, indicates near-field redistribution of water. Water redistribution occurs when the gas-injection pressure exceeds the capillary forces that hold water in the matrix pores and fractures (LeCain 1997). The differential air-injection pressure at which water redistribution is identified can be interpreted to represent an upper limit of the near-field capillary pressure.

Water redistribution was not identified in many of the high-permeability test intervals in Exploratory Studies Facility alcove boreholes because their high permeability limited the maximum injection-interval differential pressures to a few kPa. This is especially true in the upper Paintbrush contact alcove boreholes where 29 of the 41 test intervals had maximum differential air-injection pressures of 0.0 or less than about 5.0 kPa, and showed no water redistribution (LeCain et al. 1997). However, the water-redistribution pressures for the remaining 12 intervals provide some insight on the in situ capillary pressures. One test interval indicated a capillary pressure of less than 10 kPa, six indicated capillary pressures between 10 and 50 kPa, and two intervals indicated capillary pressures between 50 and 100 kPa. Two intervals had water-redistribution pressures greater than 130.8 and 164.9 kPa. The larger capillary pressures are significant because they indicate that water is being held in the rock matrix and very small fractures at pressures exceeding one atmosphere (about 100 kPa); therefore, the rocks are drier than those with smaller capillary pressures.

In the Bow Ridge fault alcove, capillary pressures appeared to be significantly lower than in the upper Paintbrush contact alcove. Of the 19 test intervals of borehole HPF#1 in the Bow Ridge fault alcove, four had capillary pressures of less than 10 kPa, six had capillary pressures between 10 and 50 kPa, one had a capillary pressure between 50 and 100 kPa, and eight intervals showed no water redistribution. The maximum differential pressure in the eight test intervals that did not show water redistribution was 12.5 kPa, indicating that the limited gas-injection rate and low test-interval pressure due to high permeability was the reason no water redistribution was identified.

However, capillary pressures in the upper Tiva Canyon alcove were as large as in the upper Paintbrush contact alcove. Of the 29 upper Paintbrush contact alcove test intervals, six had capillary pressures of less than 10 kPa, 10 had capillary pressures between 10 and 50 kPa, six had capillary pressures between 50 and 100 kPa, and seven intervals showed no water redistribution, three of which had differential pressures near or greater than 100 kPa. Once again, the larger capillary pressures are significant because they indicate that water is being held in the rock matrix and very small fractures at pressures exceeding one atmosphere.

In contrast, the rocks that show water redistribution at relatively low differential pressures also exhibit low capillary pressures. Overall, the water-redistribution data indicate that about 20 percent of test intervals had capillary pressures less than 20 kPa, indicating that these intervals have higher saturations than previously documented. (A capillary pressure of 20 kPa is equivalent to a water potential of about −0.2 bar, which is somewhat wetter than the −3.0 to −1.0 bar water potentials measured in surface-based boreholes (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) in the TSw and the PTn.) Based on the pore size distributions from Anderson (1991), the matrix characteristics curves of Klavetter and Peters (1987), and those presented in Subsection 5.3.3.1.1.4 of this report, the lower capillary pressures indicate that all the matrix pores and the smaller fractures (aperture less than 14 μm) of these intervals are saturated. These wetter capillary pressures could indicate a potential for significant microfracture (apertures of 1 to 14 μm) water flow through the unsaturated zone. The potential for water flow in microfractures through the unsaturated zone is consistent with recent estimates of an average percolation flux at Yucca Mountain of about 5 mm per year (Flint, A.L. et al. 1996).

5.3.3.1.2.3 Comparison of Laboratory Permeability Values with Surface-Based and Exploratory Studies Facility Air-Injection Permeability Values

Saturated hydraulic-conductivity values determined from laboratory analysis of core samples of various hydrogeologic units (Table 5.3-14; Flint, L.E. 1998) were compared to air-injection permeability values from surface-based boreholes and the Exploratory Studies Facility. (Saturated hydraulic-conductivity in m/s can be converted to "water permeability" in m^2 by dividing by 10^7.) The water-permeability geometric-mean values of individual welded units of the Tiva Canyon Tuff and the Topopah Spring Tuff ranged from 4.0×10^{-18} m^2 (TMN) to 3.8×10^{-15} m^2 (CUC) with overall geometric means of 6.6×10^{-17} m^2 for the TCw and 3.8×10^{-17} m^2 for the TSw. The surface-based and Exploratory Studies Facility air-injection permeability values for the TCw and TSw are 3 to 6 orders of magnitude greater than the laboratory water-permeability values (LeCain 1997). Negating the difference in the viscosities of water and air, the air-injection permeability values from the surface-based boreholes and the Exploratory Studies Facility are so much larger because they dominantly represent the bulk permeability of the fracture network in the TCw and the TSw. Because of their scale, the permeability values derived from laboratory analysis of core samples dominantly represent the permeability of the rock matrix and, therefore, it is not surprising that the field air-injection and laboratory core measurements differ so much for the welded units.

The water-permeability geometric-mean values from core analysis of individual lithostratigraphic units within the PTn ranged from 3.1×10^{-15} m^2 (CNW) to 3.2×10^{-13} m^2 (BT2) with an overall geometric mean of 6.0×10^{-14} m^2. Although greater, the surface-based and Exploratory Studies Facility air-permeability values for the PTn are much closer to the laboratory core values, falling in the upper range of the core values or differing by no more than an order of magnitude. This similarity between the field air-injection and laboratory core permeability values indicates that although the permeability of the PTn is much more homogeneous at the two scales than the welded units, the PTn nevertheless has some fracture permeability (LeCain 1997). An alternative explanation is that the laboratory permeability values are skewed toward the lower range because the higher-permeability core samples were destroyed during drilling because they were so friable.

5.3.3.1.2.4 Comparison of the Surface-Based, Exploratory Studies Facility, and Pneumatic-Monitoring Air-Permeability Values and Estimation of Anisotropy

Comparisons of mean air-permeability values determined from air-injection testing in surface-based boreholes and the Exploratory Studies Facility and determined by pneumatic monitoring are shown in Table 5.3-32. The air-permeability values are the arithmetic means times 10^{-12} m^2; the number of test intervals is indicated in parenthesis.

Air-permeability values for the Tpcpul in the upper Paintbrush contact alcove fall in the upper range of the surface-based air-permeability values for the Tpcpll and Tpcpln (Table 5.3-32), which may be associated with increased fracturing or reduced overburden pressure because of the shallow depth of the upper Paintbrush contact alcove (about 24.0 m below ground surface). Unfortunately, the surface-based permeability values are for the Tpcpll and Tpcpln, making direct comparison with the Tpcpul values from the upper Paintbrush contact alcove questionable. Similarly, air-permeability values for the Tpcpmn in the Bow Ridge fault alcove are at the upper range of the surface-based

values for the Tpcpll and Tpcpln. Air-permeability values for the Tpcpll in the Bow Ridge fault alcove are about an order of magnitude smaller than the surface-based Tpcpll air-permeability values.

Air-permeability values for the Tpcpln in the upper Paintbrush contact alcove are at the lower range of the surface-based values for the Tpcpln (Table 5.3-32). In contrast, air-permeability values of the Tpcpv in the upper Paintbrush contact alcove are almost two orders of magnitude larger than the surface-based values for the Tpcpv. However, any comparison of the Exploratory Studies Facility permeability values with the surface-based permeability values should take into account that analysis of the surface-based testing data found statistically significant differences between the four surface-based boreholes (LeCain 1997). Therefore, it is not surprising that air-permeability values from a single Exploratory Studies Facility alcove match some of the surface-based air-permeability values and not others.

In other studies, air permeability of major hydrogeologic units also was determined through analysis of vertical pneumatic-pressure profiles obtained from pneumatic monitoring of instrumented borehole. (Pneumatic monitoring of boreholes is discussed in Subsection 5.3.4.2.1 of this report.) The determination of pneumatic diffusivity and air permeability from pneumatic monitoring of boreholes NRG-6 and NRG-7a is described in J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Similar analysis of pneumatic-monitoring data from borehole SD-12 is described in Rousseau et al. (1997). The pneumatic-monitoring permeability values for the Tpcpll and Tpcpln subunits of the Tiva Canyon Tuff derived from boreholes NRG-6 and NRG-7a are similar to the Exploratory Studies Facility values but generally fall at the lower range or are smaller than the air-injection values for surface-based boreholes (Table 5.3-32). However, for the Tpcpv subunit, both the pneumatic monitoring and the surface-based permeability values are significantly smaller than that obtained for the Tpcpv from the Exploratory Studies Facility's upper Paintbrush contact alcove. However, the larger permeability value for the Tiva Canyon Tuff obtained from pneumatic monitoring of borehole SD-12 is closer to the air-injection values obtained from the Exploratory Studies Facility and the surface-based boreholes. The larger air-permeability values for the Tiva Canyon Tuff obtained from pneumatic monitoring of boreholes SD-12 compared to those obtained from boreholes NRG-6 and NRG-7a may be due to the difference in the methods used (see J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997).

The air-permeability values for the PTn from pneumatic monitoring of boreholes NRG-6, NRG-7a, and SD-12 agree well with the mean air-injection permeability value from borehole NRG-7a (Table 5.3-32). For the Topopah Spring Tuff, the pneumatic-monitoring permeability values from boreholes NRG-6, NRG-7a, and SD-12 are about an order of magnitude larger than the surface-based air-injection permeability values.

Comparison of the pneumatic-monitoring and air-injection permeability values may provide some insight on how scale influences the results of these different methods for determining air permeability (LeCain 1997). The pneumatic-monitoring permeability values for the Tiva Canyon Tuff and PTn generally are within the range of the surface-based and Exploratory Studies Facility air-injection permeability values (Table 5.3-32), with the exception of the larger pneumatic-monitoring values obtained for the Tiva Canyon Tuff from borehole SD-12. This is expected because at the scale of the air-injection tests (an ellipsoid with a zone of influence of approximately 4 m), the Tiva Canyon Tuff and the PTn are heterogeneous. At the scale of the pneumatic-monitoring (40 to 200 m vertically), individual lithostratigraphic intervals must be combined and the heterogeneity indicated by the range in air-infection values is lost in the pneumatic-monitoring average permeability value.

In contrast, the pneumatic-monitoring permeability values for the Topopah Spring Tuff are an order of magnitude larger than the surface-based air-injection permeability values (Table 5.3-32). One possible explanation is that the permeability of the Topopah Spring Tuff is anisotropic (LeCain 1997). Consequently, a vertical borehole that penetrates the Topopah Spring Tuff, which is dominated by vertical fractures, may intersect few fractures and, therefore, may not provide a representative permeability value from in situ air-injection testing. Because the air-injection test intervals may have poor pneumatic connections to the vertical fracture system, the test-interval permeability value will be dominated by the poor connection, not by the vertical fracture permeability. Therefore, the air-injection permeability values may be more representative of the horizontal permeability of the Topopah Spring Tuff than the vertical permeability. In the pneumatic monitoring, pressure responses are controlled by transmission through the vertical fractures. Even if a pneumatic-monitor interval itself has a poor pneumatic connection to the vertical fracture system, the pneumatic-monitor interval pressure response still is controlled by the vertical fracture system because of the relatively large scale (40 to 200 m vertically) of the pressure measurements. Thus, the air-injection values most closely represent the horizontal permeability and the pneumatic-monitoring values most closely represent the vertical permeability, resulting in a significant vertical-to-horizontal anisotropy ratio. The apparent anisotropy of the Topopah Spring Tuff has been noted in other studies including Rousseau et al. (1997), which concluded that the ratio of vertical-to-horizontal permeability in the Topopah Spring Tuff is 3:1 to 10:1.

Like the Topopah Spring Tuff, the Tiva Canyon Tuff contains an abundance of vertical fractures. However, comparison of the air-injection and pneumatic-monitoring permeability values (Table 5.3-32, with the exception of borehole SD-12) suggests that the Tiva Canyon Tuff may have a larger horizontal air permeability than vertical, at least in certain locations such as in the vicinity of Drill Hole Wash. The larger horizontal permeability of the Tiva Canyon Tuff may be because of a higher density of horizontal fractures. As indicated in Table 5.3-26, boreholes NRG-6 and NRG-7a have fracture densities in the Tiva Canyon Tuff that are about four times larger than in the Topopah Spring

Tuff. A possible explanation for this apparent higher density of horizontal fractures could be the thinner overburden and associated stress-relief opening of the horizontal fractures (LeCain 1997).

When comparing air-injection and pneumatic-monitoring permeability values, it is important to keep in mind that the pneumatic-monitoring permeability values are derived from a pneumatic-diffusivity model (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Weeks1978). Because the pneumatic-diffusivity term has permeability in the numerator and porosity in the denominator, the calculated permeability value will be different for a different assumed value of porosity while the pneumatic-diffusivity term remains constant. Thus, if the effective air-filled porosity is actually one-half the estimated value, then the calculated permeability values also will be half (LeCain 1997).

Assuming that air-injection permeability values obtained from vertical boreholes are more representative of the formation horizontal permeability, and that air-injection permeability values from horizontal boreholes in the Exploratory Studies Facility are more representative of the formation vertical permeability, comparison of the air-permeability values for the Tpcpll obtained from the Bow Ridge fault alcove and those obtained from the surface-based boreholes (Table 5.3-32) indicates that the permeability of the Tpcpll is anisotropic with an average horizontal-to-vertical ratio of approximately 10:1 (LeCain et al. 1997). The anisotropy is not a function of depth because the boreholes in the Bow Ridge fault alcove and test intervals in the surface-based boreholes are located at about the same depths below ground surface (30 to 40 m). However, the apparent anisotropy of the Tpcpll may be due to the limited spatial distribution of testing locations and the small databases for the Bow Ridge fault alcove and surface-based permeability values. Anisotropy of the Tpcpln in the upper Paintbrush contact alcove is dependent on which of the surface-based boreholes is used for comparison. Comparisons with SD-12 and NRG-6 indicate isotropy while comparisons with UZ-16 and NRG-7a indicate anisotropy with a vertical to horizontal ratio of up to 10:1. The 17 test intervals of the Tpcpln in the upper Paintbrush contact alcove have air-permeability values that range from 0.02 to 12.0×10^{-12} m^2 (LeCain and Patterson 1997; LeCain 1998), indicating that it probably is not valid to compare an average of 17 tests to the single value of UZ-16 or the two values of NRG-7a. Therefore, based on the limited database, the Tpcpln appears to be isotropic. Significantly, considering the predominance of vertical fracturing in the Tiva Canyon Tuff (Beason et al. 1996; Barr, D.L. et al. 1996), there is no indication that the vertical permeability of the Tpcpln is greater than the horizontal permeability. The small database of surface-based values for the Tpcpv (two test intervals) indicates that the air permeability of the Tpcpv is anisotropic with a vertical-to-horizontal ratio of about 80:1 (LeCain and Patterson 1997; LeCain 1998).

5.3.3.2 Saturated Zone

Hydrologic properties determined most commonly by the saturated-zone investigations are porosity and hydraulic conductivity at the rock-matrix scale and transmissivity at the rock-mass scale. Respectively, the two scales are characterized by tests on core samples and in boreholes. A limited number of multiple-borehole tests provided storage and dispersion coefficients for part of the saturated system.

Matrix properties–Porosity is simply the ratio of the void space in the rock to its total volume (voids plus solids), expressed either as a decimal or as a percent. The intrinsic permeability of a geologic material is a measure of the ease with which it transmits fluids and is usually expressed in the units of darcies or millidarcies (md) in Yucca Mountain reports. Values of intrinsic permeability reported in Subsection 5.3.3.2.1 were determined by laboratories that analyzed samples primarily for geophysical properties or for characterizing unsaturated flow of both water and air. For hydrologic investigations, hydraulic conductivity is the more common parameter that describes the ease of water movement through water-saturated rock or soil. Hydraulic conductivity has the units of volume per unit time per unit area under a unit gradient (expressed as height of water head divided by length of flow path), $L^3/T/L^2/L/L$, which reduces to L/T, the units of velocity. On the YMP, meters per day is the usual expression. The "velocity" obtained is not an actual water-particle velocity but, rather, is the Darcian velocity, the volumetric rate of flow per unit area. Hydraulic conductivities determined by laboratory tests on core are reported for the density and viscosity of dilute water at 20°C.

Rock-Mass Properties–The capability of a large volume of geologic material ("rock") to transmit or store water includes the properties of its included matrix, along with the properties of fractures (joints or faults). Ideally, rock-mass hydraulic characteristics are determined for a hydrogeologic unit (usually an aquifer) of stratigraphically defined thickness, but in reality they commonly are determined for the thickness actually tested, whether that is the saturated-zone depth penetrated by a borehole or the thickness of an interval of a borehole that is mechanically isolated by inflatable packers. The results of tests to determine rock-mass properties are given in Subsection 5.3.3.2.2.

Transmissivity is the parameter used to describe the capability of a specified rock mass to transmit water. It is the volumetric flux (L^3/T) that would be transmitted by the entire thickness of the interval under a unit hydraulic gradient. It has units of L^2/T, dimensionally equivalent to those of hydraulic conductivity times the unit thickness. Networks of interconnected fractures usually are orders of magnitude more permeable than the rock matrix, so a fractured rock mass normally has a much larger transmissivity than would be predicted by multiplying its matrix hydraulic conductivity by the thickness. "Apparent hydraulic conductivity" is used in many Yucca Mountain reports to normalize tested thicknesses for qualitatively comparing the results of hydraulic tests. It is calculated by dividing the primary test result, transmissivity, by the tested thickness. It may include the matrix properties of multiple rock types as well as the typically dominant properties of fractures. Therefore, comparisons with matrix hydraulic conductivity determined for cores can be very misleading.

Storativity, or the storage coefficient, relates the change in the volume of water in a rock unit (ideally, in a defined aquifer) to the change in hydraulic head. It is the change in the volume (L^3) of water in a unit square (L^2) of the full thickness of the tested unit when the head changes one unit (L); therefore, it is dimensionless. Determination of the storage coefficient requires measurements in an observation well during the transient drawdown due to pumping another well, and the results are commonly ambiguous. It is not required to describe or model flow in steady-state systems, but at least reasonable estimates are needed to assess the effects of changed groundwater flux through the system, such as might result from climate change.

Solute transport in both steady-state and transient groundwater systems is sensitive to effective porosity, which is conceptually simple but elusive to determine. Conceptually, it is the part of the porosity that preferentially permits the flow of water. The Darcian flow velocity divided by the effective porosity ideally would provide the seepage velocity. However, in a fractured, faulted rock mass, there is a wide distribution of permeability that defies precise measurement and description. In the practical sense, effective porosity is incorporated into a set of dispersion coefficients that are determined in tracer tests and large-scale geochemical investigations. Consequently, it is not considered further in the hydrogeology section but, rather, is deferred to later discussions of transport characteristics of Yucca Mountain.

5.3.3.2.1 Core Scale Hydrologic Properties

Compilation and evaluation of data from core samples recovered from exploratory and test holes at and in the vicinity of Yucca Mountain since the early 1970s are presented here to characterize hydrologic properties of the stratigraphic units that compose the site-scale hydrogeologic system. By definition, data to be included would be for analyses of core recovered from the area bounded approximately by latitude 36°34'N and 37°N, and longitude 116°18'W and 116°38'W, and lying within the Alkali Flat-Furnace Creek sub-basin. The location of the drillholes from which core samples considered in this subsection have been recovered are listed in Table 5.3-33.

Most of the hydrologic data considered here are for core from drillholes that penetrated the saturated zone beneath and in the immediate vicinity of Yucca Mountain. Seven of the hydrogeologic units named in the site-scale modeling of the saturated zone groundwater system are not commonly recognized in the immediate vicinity of Yucca Mountain but have been identified in core recovered from drillholes that penetrated the saturated zone in the nearby region. These other hydrogeologic units may be stratigraphically equivalent to units local to Yucca Mountain. They also could correlate with disconformities in local strata. Other strata somewhat distal to Yucca Mountain but included in the site-scale groundwater flow system may not have been penetrated, or may not occur in the immediate vicinity. These units may occur in saturated zone strata to the north (Pahute Mesa), east (Nevada Test Site), and south (Franklin Lake Playa) of Yucca Mountain in areas that are recognized as part of the Alkali Flat-Furnace Creek sub-basin, and from areas within the Ash Meadows sub-basin (Winograd and Thordarson 1975; Czarnecki 1990).

Syntheses of the analytical data compiled and presented in this subsection provide parameter values for hydraulic conductivity, permeability, and matrix porosity. Core data for characterization of each hydrogeologic unit in the immediate vicinity of Yucca Mountain are presented (Table 5.3-34) using

first-order statistics (when appropriate), including calculations of the mean and standard deviation of the sample population considered.

Based on available information, evaluation of parameters using the Shapiro-Wilk W test (SAS Institute 1996; NRC 1994) is included to determine whether parameters may be treated as normally distributed (W = to 1.0). Correlation of strata property estimates was considered by visual inspection of two-parameter plots.

Methods–All cited reports (Table 5.3-33) include a discussion or reference to the purpose and procedure for conducting core recovery programs. Examples of common core programs may be found in work such as that of Winograd and Thordarson (1975, pp. C17 to C19), Anderson (1981, 1994), Lobmeyer, Whitfield et al. (1983, p.18), and Lahoud et al. (1984, p.11). Laboratories where the analyses of hydraulic or physical rock properties for saturated media in the Yucca Mountain vicinity have been performed are cited in the drillhole investigation reports that provide results of core analyses, and include the Denver laboratories of the USGS, Sandia National Laboratories, and Holmes and Narver Material Testing Laboratory. Commonly, hydraulic and physical properties that have been determined by laboratory measurements of core from the saturated zone include porosity, bulk density, hydraulic conductivity (ambient water properties), and permeability (intrinsic). Saturated hydraulic conductivity, K, is related to intrinsic permeability, k, by the relationship,

$$K = kC\rho g/\eta, \qquad \text{Eq. 5.3-2}$$

where ρ and η are density and dynamic viscosity of water at the specified (or ambient laboratory) conditions, g is the acceleration of gravity, and the constant C converts between units of expression. To obtain the equivalent hydraulic conductivity in m/day under typical laboratory conditions, intrinsic permeability in millidarcies is multiplied by 0.00083. Values for compressional sonic velocity, electrical resistivity, induced polarization, remnant magnetization, and magnetic susceptibility were also estimated but are not discussed here.

Discussions of the techniques for sample collection and preservation can be found in work by Spengler, Muller et al. (1979, p. 39), Anderson (1981, 1991), and Rautman and Engstrom (1996b, pp. 19-26). Laboratory techniques, errors that could be expected or encountered, and the association of analytical results and rock type properties are also described in Anderson (1981, pp. 3-34; 1991, pp. 13-26).

Methods for core recovery to analyze structural features including joints, shear fractures, and faults are described by Spengler, Muller et al. (1979, p. 24) and Spengler, Byers et al. (1981, p. 39). Criteria for determining the origin and type of structural discontinuity observed may be found in work by Scott and Castellanos (1984, p. 57) and Spengler, Byers et al. (1981, p. 40). Methods for assigning attributes such as frequency, density, orientation, and coating of fractures are reviewed by Scott and Castellanos (1984, pp. 57-67).

Hydraulic and Physical Rock Properties Observed in Core

Topopah Spring Tuff–The upper volcanic aquifer in the site area saturated zone comprises the Topopah Spring Tuff. The tuff primarily occurs in the unsaturated zone (see Subsection 5.3.2), but is saturated to the east and south of Yucca Mountain and in Crater Flat (Luckey, Tucci et al. 1996). The degree of welding reported for core samples indicates nonwelded tuffs at shallow depths (less than 152 m (500 feet)), changing to moderately and densely welded tuff (about 152 to 427 m (500 to 1400 feet) depth), and then returning to a nonwelded tuff in the deepest intervals of recovered core that are considered here (427 to 549 m (1,400 to 1,800 feet)). Luckey, Tucci et al. (1996, Figure 7) include the lowermost, nonwelded tuffs in the upper volcanic confining unit.

The W Test for measurements of matrix porosity indicates this parameter is not normally distributed. Also, it appears that changes in matrix porosity are only weakly related to depth. Available data indicate effective porosity, calculated at UE-25a #4, UE-25a #5, and UE-25 #7 (Anderson 1991), and J-13 (Thordarson 1983), decreases with an increase in depth. As could be expected, matrix porosity decreases with an increase in welding. For nonwelded tuffs, the matrix porosity is between 10 and 50 percent and for densely welded tuffs, porosity ranges between 5 and 10 percent.

Horizontal hydraulic conductivities for the 33 samples provided by four drillholes indicate a moderately normal distribution (W = 0.96). Values generally decrease from a high of 2e-4 m/day at the southernmost drillhole, J-13 (Thordarson 1983), to a low of 2e-12 at drillhole SD-12 (Rautman and Engstrom 1996b), and then increase to 8.3e-7 at the northernmost drillhole that provides data, UE-25b #1 (Lahoud et al. 1984). As expected, an increase in welding is associated with a decrease in hydraulic conductivity. There is no apparent trend in hydraulic conductivity values with an increase in depth or in a lateral, south to north direction (Figure 5.3-68).

Vertical permeability was determined by Anderson (1994) for core samples from USW G-3 and USW G-4. Horizontal permeability also was determined for the Topopah Spring Tuff penetrated at UE-25a #1 (Anderson 1981), and at UE-25a #4, UE-25a #5, -25a #6, and -25a #7 (Anderson 1991). Based on these data, parameters display a moderately normal distribution, with W for the vertical and horizontal permeability equal to 0.97 and 0.95, respectively. Neither parameter appears to exhibit changes correlative to the degree of welding, and neither parameter displays a trend in values that is strongly correlative to depth (Figure 5.3-68). The relatively large range in vertical permeability, 0.0001 to 5.28 md, was determined for the moderately to densely welded tuff at USW G-3. A large range for horizontal permeability, 0.0009 to 38.8 md, occurs in densely welded tuffs at UE-25a #4 and UE-25a #7, respectively. Anderson (1991, p.1) indicated the large range in estimates of permeability is a consequence of pore structure rather than porosity. A schematic of the log of horizontal permeability suggests that estimates for permeability do not trend in a south to north direction (Figure 5.3-68).

Calico Hills Tuff–At Yucca Mountain and the nearby vicinity, the Calico Hills Tuff is part of the upper volcanic confining unit. Based on core sample descriptions, the unit is commonly nonwelded and vitric in the upper intervals, and transcends to zeolitic nonwelded tuff, bedded tuff, or basal sandstone in the lower part (Rush et al. 1984; Lahoud et al. 1984; Thordarson 1983; Flint, L.E. and Flint 1990; Geldon 1993; Rautman and Engstrom 1996a, 1996b).

Matrix porosity data have been analyzed for intervals from 11 drillholes that penetrate the Calico Hills Formation beneath, and in areas proximal to Yucca Mountain to the north, east, south, and southwest. For the total sample population the matrix porosity ranges between 11.8 and 47 percent, and data appear to be normally distributed (W equals 0.99). There is only a general correlation of changes in porosity with depth of cored strata (Figure 5.3-69). A comparison of porosity data and sample locations based on the northern coordinates of drillholes indicates the mean value for the matrix porosity in the Calico Hills Formation is comparable across the Yucca Mountain vicinity in a north-south sense of direction (Figure 5.3-69).

The range for saturated hydraulic conductivity in the Calico Hills Formation for samples from drillholes, UE-25b #1 (Lahoud et al. 1984), SD-7 and SD-12 (Rautman and Engstrom 1996a, 1996b), and SD-9 (Engstrom and Rautman 1996) is 6.1 x 10-1 to 6.3 x 10-7 m/day. These data when compared to depth indicate a general decrease in the conductivity with depth. A plot of the northern coordinate of the drillhole locations versus hydraulic conductivity suggests the conductivity also increases in a north to south direction (Figure 5.3-70). Sufficient data for evaluating changes in saturated hydraulic conductivity in the vertical direction were not available. Neither the untransformed or common log values for hydraulic conductivity nor the permeability parameters are normally distributed.

Although both the vertical and horizontal permeability values vary by as much as four to five orders of magnitude, a statistical evaluation using the Mahalanobish distance to determine outlier data (SAS Institute Inc. 1996) indicates there are no significant outliers. For vertical permeability, approximately 79 percent of the reviewed values are between 0.01 to 0.06 md. Large vertical permeabilities, 27.9 and 15.6 md, were determined in separate investigations by Anderson (1994) and L.E. Flint and Flint (1990) for samples of vitric tuff at depths of about 458 and 457 m (1,503 and 1,499 feet), respectively, at USW G-3. For horizontal permeability, approximately 78 percent of the values range from zero to 1.0 md, 11 percent are between 1.0 and 6.0 md, and the remaining 10 percent range from 11 to 32.7 md. The largest reported horizontal permeability, 32.7 md, also occurs at 458 m depth in drillhole USW G-3 (Anderson 1994). Although the vertical permeability generally decreases with an increase in depth, the distribution for values of horizontal permeability is fairly scattered.

Crater Flat Group–Core analyses for the Crater Flat Group at and near Yucca Mountain are derived from core programs for the three formations composing the Crater Flat Group, the Prow Pass, Bullfrog, and Tram Tuff(s), in order of increasing depth. Descriptions for the Prow Pass Tuff indicate it is mostly (about 90 percent) nonwelded to partly welded tuff, whereas about 10 percent is moderately to densely welded. The Bullfrog apparently is more welded, being 60 percent moderately to densely welded, and only 40 percent nonwelded to partly welded. In the lowermost formation, the Tram Tuff, the degree of welding is less and, again, the tuffs are generally (80 percent) nonwelded to partly welded; 20 percent are partly to densely welded.

Prow Pass Tuff–Available data for matrix porosity in the Prow Pass Tuff are based on core analyses for samples from 10 drillholes that are distributed at and near the northern, eastern, and southern circumference of Yucca Mountain, and positioned in a manner that encompasses the potential repository. Data for core analyses from two of the drillholes, UE-25a #1 (Anderson 1981) and SD-7

(Rautman and Engstrom 1996a) are for intervals of strata intermittently dispersed through most of the unit including 146 and 149 m (480 and 490 feet) of section, respectively. Core from two more drillholes USW G-4 (Anderson 1994) and SD-9 (Engstom and Rautman 1996), provide data from intermittently spaced intervals with less complete coverage of the sequence, 105 and 99 m (345 and 325 feet) of section, respectively. Data for these and for samples from smaller sections of the Prow Pass unit at six other drillholes, UE-25b #1(Lahoud et al. 1984), USW G-1 (Flint, L.E. and Flint 1990), USW G-3 (Anderson 1994), USW H-1 (Rush et al. 1984), J-13 (Thordarson 1983), and USW SD-12 (Rautman and Engstrom 1996b) show the range in matrix porosity is 19.0 to almost 39 percent, with a mean value of 24 percent. These data are fairly normally distributed (W equal to 0.96). There is only a general correlation of porosity and depth, and a general trend of an increase in the mean porosity value at drillholes, in a north to south direction (Figure 5.3-71).

Horizontal hydraulic conductivities for the Prow Pass Tuff were provided for unit intervals at five drillhole locations. At USW H-1(Rush et al. 1984) and UE-25b #1 (Lahoud et al. 1984), only two samples each are provided. These samples are from intervals at approximately 74 and 75 m (243 and 246 feet) below the upper unit contact at USW H-1, and 140 and 145 m (459 and 476 feet) below the unit top at UE-25b #1. Drillholes SD-7 (Rautman and Engstrom 1996a) and SD-12 (Rautman and Engstrom 1996b) provide core from intervals distributed through the majority of the Prow Pass section (78 and 90 percent, respectively), but SD-9 penetrated only the upper 73 m (241 feet) (Engstrom and Rautman 1996). Values for horizontal hydraulic conductivity range between zero and 0.013 m/day, with a mean of 1 x 10-5 m/day; neither the untransformed nor natural log of estimates are normally distributed. The largest conductivity in the Prow Pass Tuff, 0.013 m/day, occurs at about 30 m (100 feet) below the upper contact of the unit at SD-12, at approximately 530 m (1,740 feet) below the surface (Rautman 1996). The geologic core log for drillhole SD-12 notes slickensides were reported in an interval about one meter below the core sample location. Comparisons of data and associated depths of core recovery from single drillholes, as well as comparison of the compiled data and associated depth of core recovery that is considered here, indicate there are no trends in horizontal hydraulic conductivity associated with depth for the Prow Pass Tuff. Review of values at each drillhole location versus the northern coordinate for drillholes from which data are derived indicates a relatively large distribution of hydraulic conductivity in the Prow Pass Tuff at each drillhole location and also indicates there is no notable change in the parameter with latitude across the Yucca Mountain vicinity. The mean for all values composing the sample population compared to the mean value at each drillhole indicates a variance from the population mean of about one order of magnitude also would suggest relatively consistent hydraulic conductivity in the Prow Pass unit.

Hydraulic conductivities in the vertical direction were reported for two samples each from USW H-1 and UE-25b #1. At UE-25b #1 (Lahoud et al. 1984, p.13), the analyses for core recovered from 626 m (2,053 feet) depth indicated a vertical hydraulic conductivity of 1.8 x 10-5 m/day. The other analysis was for core taken 53 m (175 feet) deeper and was an order of magnitude smaller, 1.7 x 10-6 m/day. At USW H-1 (Rush et al. 1984), the analyses were for samples within one meter of one another at about 640 m (2,100 feet) depth, and resulted in comparable values of approximately 6.0 x 10-5 m/day. The four values exhibit a log normal distribution. There are insufficient data to evaluate the lateral distribution of parameter values or to determine the correlation between depth and vertical conductivity.

Horizontal permeability values for the Prow Pass Tuff were provided by analyses of core from UE-25a #1, UE-25c #1, UE-25c #2, and UE-25c #3 and USW G-1, -3, and -4. At the C-hole Complex, the core analyses were performed for samples distributed intermittently throughout the entire sequence, ranging between about 131 and 143 m (430 to 470 feet) thick (Geldon 1993, 1996). Intermittently spaced core samples also were recovered from an interval about 27 m (90 feet) thick in the lower part of the unit at USW G-1 (Flint, L.E. and Flint 1990), an upper through middle interval about 70 m (229 feet) thick and spanning approximately 70 percent of the Prow Pass at USW G-3 (Anderson 1994), and intermittently throughout the 122 m (400 feet) thick interval at USW G-4 (Anderson 1994). At UE-25a #1, core samples were recovered for analyses of the upper 109 m (358 feet) of the 148 m (486 feet) unit (Anderson 1981). Based on these data, permeabilities in the horizontal direction have a log normal distribution (W = 0.97). There is no apparent correlation between permeability and the degree of welding (nonwelded to partly welded).

Estimates for horizontal permeability for 89 samples ranged between 0.002 and 43.9 md, with a mean for these data of 1.89 md. The majority of values were between 0.1 and 10 md. Data for core from UE-25a #1 (Anderson 1981), USW G-4 (Anderson 1994), and the C-hole Complex (Geldon 1993, 1996) did not indicate a trend in horizontal permeability that can be associated with increased depth. A plot of the log of data points versus the drillhole northern coordinate suggests the range in horizontal permeability are comparable for the Prow Pass Tuff evaluated at most locations (Figure 5.3-72). However, a line connecting the mean of the log of permeability estimates at each drillhole indicates a general increase in horizontal permeability to the south (Figure 5.3-72).

A mean vertical permeability for the Prow Pass was determined from core collected from intervals of the Prow Pass at UE-25c #1 (Geldon 1993), USW G-1 (Flint, L.E. and Flint 1990) and USW G-4 (Anderson 1994), and for core from USW G-3 where samples were intermittently spaced throughout the entire Prow Pass Tuff. Values reported for vertical permeability ranged between 0.003 and 217md. The mean values for data from each drillhole location indicate vertical permeability in the Prow Pass Tuff also increases to the south (Figure 5.3-72). There was no strong correlation of vertical permeability with depth. The maximum value, 217 md, is from analyses of core recovered at the most southerly drillhole (752,779 North) providing data, USW G-3. Because this large value occurs as a statistical outlier, the mean vertical permeability for the sample population both with and without this data point is presented here, but the sample population mean reported in Table 5.3-34 was determined by omitting the outlier. The mean vertical permeability calculated by including the large value is 11.3 md. Omission of the outlier value results in use of the next largest permeability, 50.5 md, as a maximum and a calculated mean vertical permeability of 3.07 md. The alternative, next largest permeability, 50.5 md, also was calculated for core recovered from USW G-3. This maximum value is from the same stratigraphic interval that provided the highest reported horizontal permeability in the Prow Pass Tuff; there was no horizontal permeability value calculated for the same depth as the outlier. As for horizontal permeability, there is no apparent correlation between permeability and the degree of welding (nonwelded to partly welded). Vertical permeability estimates do not fit a log normal distribution (W = 0.94).

Bullfrog Member Tuff–Eleven drillholes provide a laterally distributed sample population for some properties to characterize the Bullfrog Tuff of the Crater Flat Group. Drillholes UE-25c #1, -25c #2, -c25#3 (Geldon 1993, 1996); USW G-3 and USW G-4 (Anderson 1994); and USW H-1 (Rush et

al. 1984) were cored at intervals through all or nearly all of the Bullfrog Tuff thickness. Core from intervals spanning approximately 90 percent of the reported unit thickness were collected for analyses from drillholes UE-25b #1 (Lahoud et al. 1984) and SD-7 (Rautman and Engstrom 1996a). Core from drillhole J-13 (Thordarson 1983) represents sampling from about 82 percent of the total penetrated thickness. The Bullfrog Member was not completely penetrated at UE-25a #1 (Anderson 1981), and it isn't certain how representative this sampling is. However, this drillhole is proximal to USW H-1 and UE-25b #1 where the Bullfrog Tuff is 113 m (371 feet) and 99 m (325 feet thick), respectively, allowing one to estimate that the upper 35 m (115 feet) interval from which core was recovered at UE-25a #1 would represent less than half the entire unit. Finally, one sample was recovered from drillhole SD-12 (Rautman and Engstrom 1996b). Descriptions of welding indicate the tuff is variably welded, alternating between nonwelded and moderately to densely welded throughout the sections at each drillhole.

Matrix porosity data for the Bullfrog Tuff are available from all noted drillholes, except the C-hole Complex. These data indicate that the mean porosity value determined for drillholes generally decreases to the south across the Yucca Mountain vicinity. The mean for the compiled data was 19 percent, and data are not normally distributed (Shapiro-Wilk W Test number of 0.93). Comparison of data suggests an increase in porosity for the Bullfrog Tuff only generally correlates to depth. An increase in welding may be associated with a decrease in the percent porosity. Notwithstanding, porosities between 15 and 20 percent are not restricted to any particular class of welding and commonly occur in tuff samples varying from the least to the most welded.

Vertical hydraulic conductivity was determined for the Bullfrog Tuff penetrated at two drillholes. Ten estimates were from the Bullfrog Tuff penetrated at USW H-1 (Rush et al. 1984) and four were for samples recovered from UE-25b #1 (Lahoud et al. 1984). Estimates for hydraulic conductivity in the horizontal direction have been reported for core samples of the Bullfrog Tuff penetrated at these drillholes, as well as for analyses of seven samples from drillhole SD-7 (Rautman and Engstrom 1996a) and one sample from SD-12 (Rautman and Engstrom 1996b). Vertical hydraulic conductivity ranges between zero and 5 x 10-4 m/day, with a mean of 1.7 x 10-4 m/day. Horizontal hydraulic conductivity ranges between zero and 1 x 10-3 m/day, with a mean of 2 x 10-4 m/day. Based on available data, neither parameter may be treated as normally distributed. Furthermore, neither parameter appears to correlate well with either welding or depth (Figure 5.3-73). Hydraulic conductivity in the horizontal direction appears to decrease to the south (Figure 5.3-74). There are insufficient data to evaluate trends in the lateral distribution of vertical hydraulic conductivity.

Based on a total of 108 core analyses for samples recovered from UE-25c #1, -25c #2, -25c #3 (Geldon 1993, 1996), UE-25a #1 (Anderson 1981) and USW G-3 and -4 (Anderson 1994), the mean value for horizontal permeability in the Bullfrog is 6.1 x 10-1 md. A relatively high standard deviation for the sample population exemplifies the scatter in data (Table 5.3-34). First-order statistics for the 26 samples recovered from USW G-3 and -4 (Anderson 1994) to determine vertical permeability, also indicate a scatter of parameter values (Table 5.3-34). Values of horizontal permeability exhibit a fairly good log normal distribution (W equals 0.96), but the vertical permeability estimates do not. Plots of compiled permeability estimates versus depth reveal horizontal permeability tends to increase with depth; vertical permeability data are inconclusive (Figure 5.3-75). An increase in the degree of welding appears to correlate to a decrease in

horizontal permeability (Figure 5.3-75). Based on the estimates for permeability in the horizontal and vertical direction, permeability in the Bullfrog Tuff tends to decrease to the south (Figure 5.3-76).

Relation of Hydraulic Conductivity in the Bullfrog Tuff to Depth and the Degree of Welding

Tram Tuff–The reported thickness of the Tram Tuff varies between about 265 m (870 feet) at drillhole J-13 (Thordarson 1983), to 375 m (1,230 feet) at USW G-3 (Anderson 1994). At drillhole USW H-1 (Rush et al. 1984) and UE-25b #1 (Lahoud et al. 1984), the reported total section thickness was 271 and 329 m (889 feet and 1,079 feet), respectively. Of these four drillholes, only USW G-3 was sampled intermittently through almost the entire (97 percent) unit thickness. At USW H-1, UE-25b #1, and J-13, core was recovered intermittently within intervals spanning between 72 to 85 percent of the total reported unit thickness. At the C-hole Complex, the drillholes bottomed in the upper zone of the Tram Tuff (Geldon 1993); core samples were from the upper 83 to 101 m (272 to 330 feet). At USW G-4 (Anderson 1994) and SD-7 (Rautman and Engstrom 1996a) core also were only from the uppermost interval of the unit. A summary of welding descriptions for core samples from the noted drillholes indicates 64 percent of the Tram Tuff is nonwelded to partly welded. At drillhole USW G-3 and UE-25b #1, most of the strata were described as partly to moderately welded, while core from the upper interval at the other drillholes was fairly consistently logged as nonwelded to partly welded, and the lower, deeper interval was usually described as partly to moderately welded.

Matrix porosity is characterized based on reports for five drillholes, USW H-1 (Rush et al. 1984), USW G-3 and -4 (Anderson 1984), UE-25b #1 (Lahoud et al. 1984), and J-13 (Thordarson 1983). It is a normally distributed parameter, with a W Test value of 0.98. Evaluation of porosity versus depth of sample reveals that the matrix porosity generally decreases with depth (Figure 5.3-77).

The largest matrix porosity, 39.4 percent, occurs in nonwelded tuff, and the smallest matrix porosity, 6 percent, occurs in moderately welded tuff; however, porosities between about 15 and 25 percent occur in tuffs ranging from the least to most welded. Compiled data indicate matrix porosity values from the different sample locations are comparable.

Hydraulic conductivities have been reported for core recovered from USW H-1 (Rush et al. 1984), UE-25b #1 (Lahoud et al. 1984), and SD-7 (Rautman and Engstrom 1996a). Horizontal hydraulic conductivity ranges between zero and 4.0 x 10-4 m/day. Vertical hydraulic conductivity ranges between zero and 5.0 x 10-4 m/day. At drillhole USW H-1, where the greatest number of samples were analyzed, three were recovered in the upper 11 m (36 feet), and the remaining samples were recovered from the lowermost 107 m (350 feet) of the Tram Tuff. A comparison of depth to vertical and horizontal hydraulic conductivity values at this one drillhole affirms the character of the whole sample population, for which changes in neither property are obviously correlative with depth.

Horizontal permeability analyses are available from UE-25c #1, -25c #2, and -25c #3 (Geldon 1996), and USW G-3 and USW G-4 (Anderson 1994). Vertical permeability analyses were reported for only USW G-3 and USW G-4 (Anderson 1994). The degree of welding does not noticeably influence the permeability, and there are no consistent trends in these parameters associated with an

increase in depth. Neither parameter displays a normal distribution. Values for horizontal permeability range between zero and 5.2 md, with a mean of 0.13 md. Based on statistical analysis using the Mahalanobish distance (SAS Institute, Inc. 1996), the maximum value of 5.2 md, which was calculated for a sample from USW G-3, is an outlier. Anderson (1994, p. 20) suggested that some of the larger discrepancies in permeability between sample pairs, where a pair includes vertical and horizontal permeability calculations for samples collected at the same interval within individual tuff members and for various stratigraphic units, could be caused by the occurrence of an unequal orientation or distribution of microfractures. Evaluation of the sample population without this datum results in a sample mean of 0.05 md. In the C-hole Complex, the horizontal permeability in the Tram Tuff is typically about 1.0 x 10-2 md; the exception is in the uppermost 30 feet interval at UE-25c #1 where it is 3.0 x 10-1 md. Horizontal permeability in the Tram Tuff is often less at the C-hole Complex than at USW G-3, where the maximum sample permeability, 5.2 md, occurs at 1,070 m (3,511 feet) depth. Analyses of core samples recovered at USW G-4 provide values for horizontal permeability in the Tram Tuff that range between 1.0 x 10-1 and 2.6 x 10-1 md and also indicate a commonly larger horizontal permeability there than at the C-hole Complex. Vertical permeability in the Tram Tuff ranged between a minimum of zero and maximum of 33.7 md, with a median of 9.0 x 10-3 md. Similar to horizontal permeability, the maximum vertical permeability was measured for core recovered from USW G-3. Except at USW G-3, both vertical and horizontal permeability decrease in a north-to-south direction.

Flow Breccia and Lithic Ridge Tuff–At Yucca Mountain and the nearby areas, a flow breccia unit and the underlying Lithic Ridge Tuff compose the middle volcanic confining unit. At USW H-1, where the flow breccia was estimated to be 119 m (390 feet) thick, one core sample was recovered at 1,201 m (3,940 feet) depth, located 34 m (111 feet) above the lower contact. A relatively small hydraulic conductivity of 8 x 10^{-7} m/day was reported by Rush et al. (1984).

Seventeen core samples were recovered at depths intermittently spaced through approximately 93 percent (288 m (945.8 feet)) of the estimated total thickness of the Lithic Ridge Tuff, where it is penetrated by USW G-3 (Anderson 1994). The only other available core analyses for the Lithic Ridge strata are for single samples from both drillholes J-13 (Thordarson 1983) and UE-25p #1 (Anderson 1991). At UE-25p #1, where the Lithic Ridge is 200 m (656 feet) thick, the core sample was taken at 1,052 m (3,453) feet depth, 10 m (34 feet) above the lower unit contact. At drillhole J-13, where the Lithic Ridge is 81 m (267 feet) thick, a sample was collected from the tuff at its lowermost contact at 324 m (1,063 feet) depth. At USW G-3, the Lithic Ridge was reported as nonwelded (Schenker et al. 1995). The sample from drillhole J-13 was reported as partly welded (Thordarson 1983). At both locations, the Lithic Ridge Tuff was reported to be zeolitized, which may be the reason for its exceptionally small permeability (Table 5.3-34).

Matrix porosity estimates for the Lithic Ridge Tuff were provided based on analysis of 18 core samples. The majority (16) of these are for core collected at USW G-3 (Anderson 1994). The other estimates are for core recovered at UE-25p #1 (Anderson 1991) and drillhole J-13 (Thordarson 1983). Values range between 15.2 and 27.4, and they display a normal distribution. Changes in the porosity do not appear to be a function of depth.

All but one estimate for both the vertical and horizontal permeability in the Lithic Ridge Tuff are from USW G-3 (Anderson 1994). Core analyses for the sample recovered from drillhole UE-25p #1 (Anderson 1991) provide the one other estimate for both parameters. Neither parameter is normally distributed, nor shows a trend in values with an increase in depth. The maximum and minimum estimated vertical permeability are 1.0 x 10-3 and 3.0 x 10-1 md. Horizontal permeability ranged between zero and 1.6 x 10-1 md.

Older Tuffs–Two reports have included core analyses each for the stratigraphic sequence referred to as the Older Tuffs, which are considered as the lower volcanic confining unit in the Yucca Mountain vicinity. These reports are for drillholes USW H-1, where 320 m (1,049 feet) of the Older Tuffs were penetrated (Rush et al. 1984), and USW G-3, where only 47.5 m (156 feet) of the Older tuffs were penetrated (Anderson 1994). At USW H-1, two core samples from approximately 61 m (200 feet) below the upper unit contact were recovered for analyses. At USW G-3, core samples were recovered at approximately 17, 37, and 47 m (55, 122, and 154 feet) below the uppermost contact depth. Although core analyses were not performed, the Older Tuffs also were penetrated at USW G-1 and UE-25p #1. At USW G-1, where 323 m (1,060 feet) of the Older Tuffs were drilled, Spengler, Byers et al. (1981) reported the drillhole bottomed in the tuffs. At UE-25p #1, where the base of the Older Tuffs was drilled through, the total thickness was reported as about 75 m (246 feet). The variance in thickness could be due to a thinning or faulting of strata. It may also be due to the depths being estimated to include other Tertiary Tuffs lying between the Older Tuffs and the Paleozoic carbonate strata (Craig and Robison 1984). At USW G-3, the Older Tuffs are described as zeolitized ashflow and bedded tuff that is nonwelded (Anderson 1994) and at USW H-1, they are described as partly to moderately welded (Rush et al. 1984). The few available data did not allow for detection of trends in hydraulic parameters with an increase in depth, or to determine lateral trends in properties.

Evaluation of porosity values for the Older Tuff indicates they are moderately normally distributed, with a W Test number of 0.95. Based on the three core analyses for the unit at USW G-3 (Anderson 1994) and two from USW H-1 (Rush et al. 1984) porosity ranges between 10.0 and 19.3 percent. Based on the same two samples from USW H-1, an average vertical hydraulic conductivity of 2.0 x 10-4 m/day and an average horizontal hydraulic conductivity of 1.8 x 10-4 m/day were determined for the Older tuffs.

The three core analyses from USW G-3 (Anderson 1994) also provided estimates for vertical and horizontal permeabilities. The minimum and maximum vertical permeability were 7.0 x 10-3 and 3.8 x 10-2 md, respectively. Although the sample population was relatively small, vertical-permeability values were determined to have a normal distribution, for which the W Test number of 0.99; the horizontal permeability with a similar range of 9.0 x 10-3 md to 4.1 x 10-2 md, was determined to not be normally distributed.

Lower Carbonate Aquifer–Hydraulic properties including matrix porosity, vertical permeability, and horizontal permeability for the lower carbonate aquifer were determined for core recovered at UE-25p #1 (Anderson 1991). Data from the Lone Mountain Dolomite (10 to 14 core analyses) reveal horizontal permeability ranges between 164 and 0.003 md. Horizontal permeability estimated for

the one sample from the Robert's Mountain Formation was 0.002 md. None of the estimated parameters were normally distributed, and none correlate significantly with depth (Figure 5.3-78).

Summary–The compilation and synthesis of core data presented here may be used to describe the apparent attributes of the stratigraphic units. However, summaries of parameter distributions do not incorporate evaluation of the relation of the areal lithologic variance of units or the impacts of structural deformation or geochemical alteration, so that descriptions of vertical or lateral trends in the hydraulic or physical rock properties are considered preliminary. The relation of these features to the distribution and transition in parameter values for stratigraphic units is considered significant (Luckey, Tucci et al. 1996, p.17) and should be determined for a comprehensive description of the saturated zone groundwater flow system at Yucca Mountain. For example, Rautman and Engstrom (1996a, p.13) and Ortiz et al. (1985, p.8) recognize that rock units at the Yucca Mountain Site are more clearly defined based on porosity, grain density, and thermal properties rather than on the basis of permeability, which probably is more sensitive to geochemical processes and mechanical strain.

Although the Topopah Spring Tuff is part of the saturated zone system in the more southerly extent of the site area, it is mostly included in the unsaturated-zone system and the reader is referred to that subsection of this document for detailed evaluation of this unit. The core data generally confirm that porosity, permeability, and hydraulic conductivity for the Topopah Spring Tuff are commonly less that those for the stratigraphically lower middle volcanic aquifer strata.

Based on referenced data, the largest matrix porosity is in the Calico Hills Formation, for which estimates range between 11.8 and 47 percent, with a mean of about 28 percent. In the volcanic tuffs that compose both the Crater Flat Group and the Lithic Ridge Tuff, the mean matrix porosity is 21 percent. In the deeper tuffs of the lower volcanic confining unit, the mean porosity was estimated as approximately 15 percent. The lowest horizontal porosity estimates were for the dolomites composing the lower carbonate aquifer, where values ranged between about 5 percent in the Lone Mountain Dolomite and 0.3 percent in the Roberts Mountain Formation. The trend through the volcanic strata in a descending order below the Topopah Spring Tuff, then, is to transition from the most porous unit in the upper interval in the saturated zone, through the less porous tuffs including the Crater Flat Group and Lithic Ridge Tuff. The least matrix porosity occurs in the underlying lower carbonate aquifer (Figure 5.3-79).

The likeness between the Calico Hills Formation and Prow Pass Tuff has been characterized by Rautman and McKenna (1997, p. 6) in work to model 3-D thermal hydrological and thermal properties at Yucca Mountain. Based on material properties including porosity the author treated the Calico Hills Formation and Prow Pass Tuff as a distinct unit. The high degree of normality (the Shapiro-Wilks test resulted in 0.99) for the distribution of the combined porosity data would confirm the likeness of the two strata. Notwithstanding, a comparison of the estimates for the horizontal hydraulic conductivity indicates that with a mean of 0.61 m per day, the horizontal hydraulic conductivity for the Calico Hills Formation is three orders of magnitude larger than that for the Crater Flat Group, having a mean of 1.4 x 10-4 to 1.8 x 10-4 m per day (Figure 5.3-80).

With respect to the mean permeabilities estimated for the saturated zone units, except for the Lone Mountain Dolomite, plots for the log of both vertical and horizontal permeability from the

uppermost through and including the lowermost units in the saturated zone strata indicate permeability decreases with depth (Figure 5.3-81). The relatively large means for permeability in the Lone Mountain Dolomite are from Anderson (1991, p. 32), and may reflect the author's observation that the orientation of open fractures could mean that some core samples were not suitable for measurement. Consequently, these values are queried on Figure 5.3-81.

The matrix hydrologic properties for the saturated zone strata do not exhibit consistent lateral trends across the site area. Data for Topopah Spring Tuff parameters indicate that overall, values do not significantly change. Based on available data, horizontal hydraulic conductivity increases southward in the Calico Hills Formation and Tram Tuff and decreases in the Bullfrog Tuff; in the Prow Pass Tuff, there is no noticeable trend. Lateral trends in permeability appear to be exist, where estimated values increase southward for the Calico Hills Formation and Prow Pass Tuff, but decrease within the lower members of the Crater Flat Group. Values for matrix porosity indicate there is no clear lateral trend for the Calico Hills Formation, Prow Pass Tuff, or Tram Tuff. The mean for porosities determined for the Bullfrog Tuff indicates the mean porosity at drillholes decreases approximately 10 percent (from about 22 to 12 percent) in a southward direction.

The comparability of the data for hydraulic conductivity (K) and those for permeability (k) was tested for corresponding sets of data means, each of which was based on data for 14 or more samples. For the same stratigraphic unit and data type (horizontal versus vertical), the ratio k/K (both in m/day) ranged from 0.0003 to 210 K. The range of values for both parameters are very large, indicating that the differences between the data sets could be valid. However, the range of k/K over almost six orders of magnitude indicates that the data should be applied with caution until the differences have been evaluated more thoroughly.

5.3.3.2.2 Rock-Mass Hydraulic Properties

More than 150 individual hydraulic tests have been conducted on boreholes (Figure 5.3-82) on and around Yucca Mountain. Almost all the tests were single-borehole tests in specific depth intervals and may have included single-well, constant-discharge pumping tests; slug-injection (falling-head) tests, pressure-injection tests, temperature surveys, and tracer injection flow surveys. Multiple-well pumping tests were conducted only at the C-hole Complex and, in limited fashion, between two boreholes in Drill Hole Wash. The various hydraulic parameters discussed in the following subsection were determined mostly from single-borehole tests, although some results from the C-hole Complex are included. Because multiple-well tests provide both more information and more reliable hydraulic data, these tests are discussed in greater detail in a subsequent subsection.

5.3.3.2.2.1 Single-Borehole Tests

Transmissivity and Apparent Hydraulic Conductivity–Estimated transmissivity values determined from hydraulic tests for the hydrogeologic units (upper volcanic aquifer, upper volcanic confining unit, lower volcanic aquifer, lower volcanic confining unit, and carbonate aquifer) in the vicinity of Yucca Mountain are listed in Table 5.3-35. Apparent hydraulic-conductivity values for the hydrogeologic units are listed in Table 5.3-36. The apparent hydraulic-conductivity values were based on reported single-borehole hydraulic tests and generally were calculated by dividing the total

reported transmissivity of the tested interval by its thickness in the borehole. Hydraulic-conductivity values for individual intervals in a borehole could vary by several orders of magnitude, depending on whether or not water-bearing fractures were present. In intervals that contained no open fractures, hydraulic conductivity tended to be low–generally less than 0.01 m/d–reflecting the hydraulic conductivity of the rock matrix or of very small fractures. In tested intervals that contained water-bearing fractures, the apparent hydraulic-conductivity values listed in Table 5.3-36 may be somewhat misleading. Most, if not all, of the water produced in such an interval could have been produced by a few thin, highly conductive fractures in an otherwise thick, essentially nonproductive rock matrix. Therefore, hydraulic conductivity is termed "apparent" (Luckey, Tucci et al. 1996, pp. 32, 35).

Upper Volcanic Aquifer–Hydraulic properties for the upper volcanic aquifer were obtained from tests in boreholes USW WT-10, UE-25 WT#12, and UE-25 J-13. In borehole USW WT-10, O'Brien (1997, p. 14) reported a mean transmissivity of 1,600 m^2/d (Table 5.3-35). Assuming an aquifer thickness of 83 m (saturated thickness tested), the apparent hydraulic conductivity for borehole USW WT-10 was 19 m^2/d (Table 5.3-36). In borehole UE-25 WT#12, O'Brien (1997, p. 22), estimated a mean transmissivity of 7 m^2/d for the saturated interval tested. Of the interval tested, 41 m were completed in the upper volcanic aquifer and 12 m were completed in the upper volcanic confining unit. The low transmissivity could indicate the very lowest limit of transmissivity values for the upper volcanic aquifer, or, that the degree of fracturing was very low in the interval of the upper volcanic aquifer tested, and the transmissivity is more representative of the upper volcanic confining unit (O'Brien 1997, pp. 22-23). Assuming a saturated thickness of 53 m, an apparent hydraulic conductivity of 0.13 m/d was estimated for borehole UE-25 WT#12 (Table 5.3-36). Thordarson (1983) reported a transmissivity of 120 m^2/d (Table 5.3-35) for geologic units that correspond to the upper volcanic aquifer in borehole UE-25 J-13. The apparent hydraulic conductivity estimated for borehole UE-25 J-13 was 1.0 m/d (Table 5.3-36) (Luckey, Tucci et al. 1996, p. 35).

The upper volcanic aquifer also was tested at borehole USW VH-1 (Thordarson and Howells 1987); however, the tests were conducted over the entire saturated interval, which includes the lower volcanic aquifer. Most of the water produced by the borehole may be from a small part of the lower volcanic aquifer (Thordarson and Howells 1987, p. 9). Reported transmissivity values ranged from 450 to 2,400 m^2/d (Thordarson and Howells 1987, pp. 14-18), and apparent hydraulic-conductivity values range from 0.8 to 4.2 m/d for five tests at borehole USW VH-1 (Luckey, Tucci et al. 1996, p. 35). Because of the uncertainty of which aquifer was tested, the hydraulic properties for borehole USW VH-1 are not shown in Tables 5.3-35 or 5.3-36.

Upper Volcanic Confining Unit–Hydraulic properties for the upper volcanic confining unit were determined from tests conducted at the C-hole Complex and boreholes UE-25b #1, UE-25 J-13, and USW G-2. Reported transmissivity values for the upper volcanic confining unit ranged from 2.0 to 26 m^2/d (Table 5.3-35), and apparent hydraulic-conductivity values from tests at these boreholes ranged from 0.02 to 0.26 m/d (Table 5.3-36) (Luckey, Tucci et al. 1996, p. 35; O'Brien 1998, p. 7).

Lower Volcanic Aquifer–The lower volcanic aquifer was tested in boreholes USW H-1, USW H-3, USW H-4, USW H-5, USW H-6, USW G-4, UE-25b #1, UE-25p #1, and UE-25 J-13 (Tables 5.3-35 and 5.3-36). For these wells, reported transmissivity values of the lower volcanic aquifer ranged

from less than 1.1 to 589 m^2/d. The arithmetic mean of the transmissivity values was about 152 m^2/d, and the geometric mean was about 43 m^2/d. Apparent hydraulic-conductivity values for the lower volcanic aquifer for these boreholes ranged from less than 3.7×10^{-3} to 2.0 m/d. The arithmetic mean of the apparent hydraulic conductivity values was 0.49 m/d and the geometric mean was 0.15 m/d. In borehole USW H-1, apparent hydraulic conductivity of individual intervals tested in the lower volcanic aquifer ranged 4×10^{-5} to 18 m/d in single borehole tests (Rush et al. 1984; Luckey, Tucci et al. 1996, p. 35).

Multiple-well aquifer tests conducted at the C-hole Complex during May to June 1995 gave a range of transmissivities from 1,600 to 3,200 m^2/d, and a range of hydraulic conductivities of 6.5 to 13 m/d (Tables 5.3-35 and 5.3-36). It is assumed that these values of transmissivity and hydraulic conductivity are for the lower volcanic aquifer. However, pumping and monitoring wells were completed in saturated sections of both the lower volcanic aquifer and lower volcanic confining unit (Geldon et al. 1998, p. 1). More information on the multiple-well testing at the C-hole Complex is given in Subsection 5.3.3.2.2.2.

Lower Volcanic Confining Unit–Hydraulic properties for the lower volcanic aquifer were collected from boreholes USW H-1, USW H-3, USW H-4, USW H-6, UE-25b #1, UE-25p #1, and UE-25 J-13 (Tables 5.3-35 and 5.3-36). Reported transmissivity values for the lower volcanic confining unit ranged from less than 3.0×10^{-3} to 23 m^2/d. The arithmetic mean of the transmissivity values was 3.7 m^2/d and the geometric mean was 0.20 m^2/d. Apparent hydraulic-conductivity values for the lower volcanic confining unit ranged from 5.5×10^{-6} to 0.11 m/d. The arithmetic mean of the apparent hydraulic conductivity values was 1.7×10^{-2} m/ d, and the geometric mean was 1.3×10^{-3} m/d (Luckey, Tucci et al. 1996, p. 35).

Lower Carbonate Aquifer–Data were available for the lower carbonate aquifer in the vicinity of Yucca Mountain only for borehole UE-25p #1. Hydraulic data presented by Craig and Robison (1984) indicated a transmissivity of 118 m^2/d (Table 5.3-35) and apparent hydraulic-conductivity values for specific tested intervals that ranged from 0.1 to 0.2 m/d and averaged 0.19 m/d (Table 5.3-36) (Luckey, Tucci et al. 1996, p. 36).

Anisotropic Conditions–In fractured-rock aquifers, such as those at Yucca Mountain, hydraulic conductivity probably is anisotropic–that is, greater in a direction parallel to the predominant fracture orientation than in other directions. Although this information is critical to determining the true groundwater flowpaths from the potential repository, only one test has been conducted at Yucca Mountain to determine whether or not the aquifers actually are anisotropic. Erickson and Waddell (1985, pp. 24-29) reported that at borehole USW H-4, the maximum hydraulic conductivity was from 5 to 7 times greater than the minimum hydraulic conductivity; the maximum hydraulic conductivity was oriented in a direction 23° east of north. Czarnecki and Waddell (1984, pp. 27-28), however, reported that their subregional model duplicated measured water levels more accurately when the aquifer was simulated as isotropic than anisotropic. Therefore, the question of anisotropy remains unresolved. Recent long-term hydraulic testing at the C-hole Complex resulted in an apparent elliptical pattern in drawdown, oriented northwest-southeast (Geldon et al. 1997, pp. 42-44). This pattern may indicate anisotropic conditions in that orientation.

Vertical Hydraulic Conductivity–Vertical hydraulic conductivity values were reported only for borehole USW H-5 and the C-hole Complex. Robison and Craig (1991, p. 26) reported an apparent vertical hydraulic conductivity value of 62 m/d for the lower volcanic aquifer at borehole USW H-5. This value was about 100 times greater than the reported value for horizontal hydraulic conductivity of 0.6 m/d (Table 5.3-36). Data for the C-hole Complex were in direct contrast with the data from borehole USW H-5. The vertical hydraulic conductivity values for the lower volcanic aquifer were estimated to range from 1 to 2 m/d (Geldon 1996), and were from 0.01 to 0.17 times the values for the horizontal hydraulic conductivity (6 to 100 m/d) obtained over the same intervals using multiple-borehole aquifer test analyses (Luckey, Tucci et al. 1996, p. 36).

Aquifer Storage–The storativity reported for tests at the C-hole Complex was 0.0005 (Geldon 1996) and 0.001 to 0.003 (Geldon et al. 1998), and specific yields ranged from 0.003 to 0.07 (Geldon 1996) and 0.01 to 0.2 (Geldon et al. 1998). The boreholes at the C-hole Complex penetrate the upper volcanic confining unit, the lower volcanic aquifer, and the upper part of the lower volcanic confining unit. Storativity for borehole USW H-1 was estimated by Rush et al. (1984, Table 12) to range from 6×10^{-6} to 6×10^{-3} for the lower volcanic aquifer and from 6×10^{-4} to 2×10^{-3} for the lower volcanic confining unit. G.E. Barr (1985) used the same data and obtained similar results. Thordarson et al. (1985, Table 2) reported storativity for borehole USW H-3 that ranged from 4×10^{-6} to 7×10^{-6} for the entire saturated interval of the borehole. Specific yields for borehole USW H-5 ranged 0.15 to 0.28 (Robison and Craig 1991, p. 24) for the entire saturated interval of the borehole; however, more than 95 percent of the water in the borehole was produced from fractures in the upper part of the lower volcanic aquifer (Robison and Craig 1991, Figure 6). G.E. Barr (1991, Table 3) estimated storativity for nine intervals in the lower volcanic aquifer in borehole USW G-4. The values ranged from 1×10^{-5} to 8×10^{-6} (Luckey, Tucci et al. 1996, p. 37).

Flow Surveys–Flow surveys were conducted in most of the deeper boreholes drilled at Yucca Mountain. Flow surveys are useful to determine which intervals of the borehole, and possibly which fractures, produced water. Most of the flow surveys were conducted using a tool developed for oil-field use that releases small quantities of radioactive iodine-131 (Blankennagel 1967, pp. 15-19). As the iodine moves up or down the borehole, it is sensed by gamma-ray detectors. Most surveys are done while water is being pumped from or injected into the borehole. Static tests also were occasionally conducted (Luckey, Tucci et al. 1996, p. 37)

Flow surveys are useful in determining which parts of the borehole produce (or accept) most of the flow. The information is useful when subdividing a system into aquifers and confining units. Minor producing intervals are not readily detected by this method. Flow surveys generally report velocities to about 0.03 m/s, but the variability between successive measurements is considerably larger than 0.03 m/s. Therefore, for borehole diameters generally found at Yucca Mountain, intervals producing less than 0.1 L/s can be easily missed. Flow surveys for seven boreholes are shown in Figure 5.3-83. For comparison purposes, the depths for all surveys were adjusted to a standard thickness of 524.3 m for the lower volcanic aquifer, which is the estimated thickness of this aquifer in borehole USW G-1, but it is not necessarily representative of the thickness of the aquifer elsewhere (Luckey, Tucci et al. 1996, p. 37).

The flow surveys generally indicated that production (or acceptance) of water occurred in a few intervals. Boreholes USW H-1 and USW H-5 are the most dramatic examples of this generalization. In borehole USW H-1, 64 percent of the production came from a 1-m interval, whereas in borehole USW H-5, 73 percent of the production came from a 42-m interval. Boreholes USW H-4 and UE-25b #1 had production from larger intervals. All of the boreholes had little production (or acceptance) of water from their lower intervals (Luckey, Tucci et al. 1996, p. 37).

None of the boreholes for which flow surveys were conducted had major production from the lower volcanic confining unit, but borehole USW H-4 produced about 12 percent of its flow from the uppermost part of that unit. The upper volcanic confining unit is saturated only in boreholes UE-25b #1 and USW G-2 (Figure 5.3-83). The upper confining unit produced approximately one-third of the water in borehole UE-25b #1 and accepted virtually all the water in the test at borehole USW G-2. The lower volcanic aquifer did not accept any water in the test at borehole USW G-2 despite being fully penetrated. At the C-hole Complex, flow surveys indicated that the lower volcanic aquifer produced virtually all the flow, but temperature surveys indicated that the upper volcanic confining unit produced some water in borehole UE-25c #3 and possibly in borehole UE-25c #2 (Geldon 1993, Figures 31-33). The upper volcanic aquifer is above the saturated zone in all of the boreholes shown in Figure 5.3-83. No flow surveys were available from boreholes that penetrated the upper volcanic aquifer in the saturated zone, but hydraulic tests at borehole UE-25 J-13 indicated that most of the production was from the upper volcanic aquifer (Thordarson 1983, Table 12). At borehole USW VH-1 in Crater Flat, most of the production came from the lower volcanic aquifer (Thordarson and Howells 1987, p. 9), whereas the upper volcanic aquifer produced little water. The upper volcanic confining unit is absent at this borehole (Luckey, Tucci et al. 1996, pp. 37, 39).

5.3.3.2.2.2 Multiple-Well Tests

C-hole Complex–Boreholes UE-25c #1, UE-25c #2, and UE-25c #3 (collectively called the C-hole Complex) were drilled each to a depth of 914.4 m at Yucca Mountain, on the Nevada Test Site, in 1983 and 1984 for the purpose of conducting aquifer and tracer tests. The C-hole Complex consists of three orthogonally spaced boreholes (Figure 5.3-84) that are 30.4 to 76.6 m apart at the land surface. The C-holes are located at the northern end of Bow Ridge, on the west side of Midway Valley (Figure 5.3-8). Below the water table, which is 400 to 402 m deep at the site, the C-holes penetrate the Calico Hills Formation, and the Prow Pass, Bullfrog, and Tram Tuffs of the Crater Flat Group. A northerly trending fault, believed to be either the Paintbrush Canyon or Midway Valley fault (Figure 5.3-9), and an offsetting northwesterly trending fault intersect the C-holes at the bottom of the Bullfrog and top of the Tram Tuffs.

In 1983 and 1984, 16 falling-head slug tests and 9 pressure-injection tests were conducted in UE-25c #1; a constant-head injection test was conducted in UE-25c #2, which was converted into a constant-flux injection test; two unsuccessful pumping tests were conducted in UE-25c #1; one pumping test was conducted in UE-25c #2; and two pumping tests were conducted in UE-25c #3 (Geldon 1996). Results from these tests, combined with results from subsequent tests, are summarized below.

Testing resumed in 1995, and hydraulic tests were conducted in May 1995, June 1995, February 1996, and May 1996 to November 1997. In all of these tests, borehole UE-25c #3 was used as the

pumping well, and boreholes UE-25c #1 and UE-25c #2 were used as observation wells. The May 1995 test is described, and results and interpretations presented, in Geldon et al. (1998). Results from the June 1995, February 1996, and May 1996 to November 1997 tests are presented in Geldon et al. (1997).

During the first and last hydraulic tests listed above, boreholes UE-25 ONC-1, USW H-4, UE-25 WT #14, UE-25 WT#3, and UE-25p #1, also, were used as observation wells. Borehole UE-25 ONC-1, which is 842.8 m from UE-25c #3, is completed in the Prow Pass Tuff. Borehole USW H-4, which is 2,245 m from UE-25c #3, is completed in the Prow Pass, Bullfrog, Tram, and Lithic Ridge Tuffs. Borehole UE-25 WT#14, which is 2,249 m from UE-25c #3, is completed in the Topopah Spring Tuff and Calico Hills Formation. Borehole UE-25 WT#3, which is 3,526 m from UE-25c #3, is completed in the Bullfrog Tuff. Borehole UE-25p #1, which is 630 m from UE-2c #3, is completed in Paleozoic carbonate rocks. The May 1996 to November 1997 test lasted 553.2 days, which allowed full responses to develop in the distant observation wells, UE-25 ONC-1, USW H-4, UE-25 WT #14, and UE-25 WT#3. Because of the lengthy records used for analysis, hydraulic properties calculated from these responses are considered very reliable, within the constraints of the assumptions made in the analytical methods used to obtain these properties.

Hydraulic tests conducted at the C-hole Complex from 1995 to 1997 were designed to:

- Confirm the results of the 1983 to 1984 tests
- Determine hydraulic properties of the composite saturated-zone section in the C-holes
- Determine hydraulic properties of the six hydrogeologic intervals in the C-holes
- Determine heterogeneity in the Miocene tuffaceous rocks, including the influence of faults, in the area encompassed by observation wells

Additionally, it was hoped that monitoring UE-25p #1 might establish whether the tuffaceous rocks are connected hydraulically to the Paleozoic carbonate rocks, a regional aquifer. The Paleozoic rocks are estimated to be about 455 m below the bottom of the C-holes.

In the hydraulic tests conducted from 1983 to 1997 that are listed above, the C-holes were either open or contained packers to isolate one or more intervals. Intervals to be packed off were determined from flow surveys, geophysical logs, and aquifer tests conducted between 1983 and 1995.

Six hydrogeologic intervals exist in the C-holes: Calico Hills, Prow Pass, Upper Bullfrog, Lower Bullfrog, Upper Tram, and Lower Tram. Flow within these intervals comes from diversely oriented fractures and the interstices of variably welded ashflow, ashfall, and reworked tuff. During hydraulic tests, the Lower Bullfrog interval contributes about 70 percent of the flow, the Upper Tram interval contributes about 20 percent of the flow, and all other intervals combined contribute about 10 percent of the flow. Because these hydrogeologic intervals are defined by spatially related faults and fracture zones, their existence and hydraulic properties cannot be extended beyond the immediate

vicinity of the C-holes. However, the results of hydraulic tests in the C-holes imply that the transmissivity of a rock mass is relatively large close to faults and associated fracture zones, and decreases with distance from these features. Although this result is based on observations at the C-hole Complex in the vertical direction, it is believed that the same principle can be applied areally.

In all hydraulic tests at the C-hole Complex, large, rapid drawdown (and recovery) in the pumping well far exceeded what could have been predicted from hydraulic properties calculated from observation-well drawdown in the same tests. Most of this pumping-well drawdown probably can be attributed to frictional head loss or "borehole skin." Because the pumping-well drawdown largely is independent of aquifer properties, analyses of this drawdown result in misleadingly small values of transmissivity and hydraulic conductivity. Hydraulic tests at the C-hole Complex imply that analyses of pumping-well drawdown throughout the Yucca Mountain area may be one to two orders of magnitude too low.

Hydrogeologic intervals in the C-holes have layered heterogeneity. The Calico Hills interval is unconfined, the Prow Pass and Upper Bullfrog intervals are confined, the Lower Bullfrog interval is a fissure-block aquifer, and the Upper Tram interval receives flow from crosscutting faults in response to pumping. Transmissivity increases downhole from 4-10 m^2/d in the Calico Hills interval to 1,300 to 1,600 m^2/d in the Lower Bullfrog interval, although this trend is reversed near the bottom of the C-holes. In the Upper Tram interval, transmissivity is 800 to 900 m^2/d. Hydraulic conductivity increases downhole from 0.1 to 0.2 m/d in the Calico Hills interval to 20 to 50 m/d in the Lower Bullfrog and Upper Tram intervals. Storativity in UE-25c #2 generally increases downhole from 0.0002 to 0.0004 in the Calico Hills and Prow Pass intervals to 0.001 to 0.002 in the Lower Bullfrog and Upper Tram intervals, although the storativity of the Upper Bullfrog interval is an order of magnitude smaller than in other intervals above the Lower Bullfrog. Storativity in the Prow Pass and Upper Bullfrog intervals of UE-25c #1 is similar to that in UE-25c #2, but the storativity of the Lower Bullfrog and Upper Tram intervals in UE-25c #1 is about one-tenth of that in UE-25c #2. Vertical distributions of hydraulic properties in the C-holes are believed to represent the increasing influence of faults and related fractures toward the bottom of the boreholes.

Distributions of drawdown in the hydraulic tests conducted in May 1995 and from May 1996 to November 1997 were influenced strongly by northwesterly and northerly trending faults. Drawdown in UE-25 ONC-1 in the latter test showed a fissure-block aquifer response, possibly because of fractures related to a northwesterly trending zone of discontinuous faults that extends between Bow Ridge and Antler Wash. Drawdown in UE-25 WT#14 and USW H-4 during the same test reached steady state after 72,000 minutes of pumping because of recharge from the Paintbrush Canyon fault and a zone of northerly trending faults that transects Boundary Ridge.

Drawdown data from four wells monitored during the hydraulic test conducted from May 1996 to November 1997, matched to the type curve for a confined aquifer, indicated transmissivity of 2,200 m^2/d and storativity of 0.002 for the Miocene tuffaceous rocks within a 21 km^2 area surrounding the C-holes. Plots of drawdown in observation wells as a function of their distance from the pumping well at various times during the same test indicated transmissivity of 2,100 to 2,600 m^2/d and storativity of 0.0005 to 0.002. Analyses of drawdown in the composite saturated-zone section of the C-holes in May 1995 and in outlying observation wells from May 1996 to November

1997 indicated that the transmissivity of the tuffaceous rocks decreases northwesterly from 2,600 m^2/d in UE-25 WT#3, to about 2,000 m^2/d in the C-holes, to 700 m^2/d in USW H-4.

The hydraulic conductivity of the Miocene tuffaceous rocks, areally, ranges from less than 2 to more than 10 m/d. It is largest where prominent, northerly trending faults are closely spaced or intersected by northwesterly trending faults. Relatively large hydraulic conductivity occurs beneath Fran Ridge, Bow Ridge, and Boundary Ridge. Hydraulic conductivity is smallest toward the crest of Yucca Mountain and Jackass Flats.

During the four hydraulic tests conducted at the C-hole Complex from 1995 to 1997 drawdown occurred in all monitored intervals of the C-holes and other observation wells, regardless of the geologic interval being pumped. This hydraulic connection across geologic and lithostratigraphic contacts is believed to result from interconnected faults, fractures, and intervals with large matrix permeability. Because flow is not strata bound, the Miocene tuffaceous rocks appear to be a single aquifer in the structural block bounded on the east by the Paintbrush Canyon fault, on the south by the Dune Wash fault, and on the west by faults cutting Boundary Ridge, at least as far north as lower Midway Valley. This conclusion is supported by the ability to combine the drawdown from observation wells completed in different Miocene geologic units in the same analytical solution of drawdown as a function of time, or of drawdown as a function of distance. Hydraulic tests at the C-hole Complex from 1995 to 1997 indicate that formal designation of aquifers and confining units within the Miocene tuffaceous sequence at Yucca Mountain may not be justified in areas containing interconnected faults and fractures.

Following the hydraulic test in February 1996 discussed above, during which quasi-steady-state conditions were reached, iodide, in the form of sodium iodide, was injected into borehole UE-25c #2 while UE-25c #3 continued to be pumped, in what was the first conservative tracer test at the C-hole Complex (Fahy 1997). During the hydraulic test from May 1996 to November 1997 discussed above, additional conservative tracer tests and a multi-constituent reactive-tracer test were conducted. The conservative tracer tests began concurrently in January 1997 and concluded when pumping stopped in November 1997. While UE-25c #3 was being pumped to maintain a quasi-steady-state hydraulic gradient, a benzoic acid was injected into UE-25c #2, and a pyridone was injected into UE-25c #1. The average discharge from UE-25c #3 during this test was about 8.9 liters per second.

The breakthrough curves from conservative tracer tests were fitted to an analytical solution of the advection-dispersion equation for a homogeneous, isotropic, dual-porosity medium (Moench 1995), and values of porosity and longitudinal dispersivity were calculated (Geldon et al. 1997). Two porosity values were calculated for each tracer test. The first is a "flow porosity," which is attributable to a network of both continuous fractures, and discontinuous fractures connected by segments of matrix, that forms the flow pathway through which tracers move between the injection and recovery wells. This flow pathway has high hydraulic conductivity, but low storage attributes. The second porosity value determined from the breakthrough curve is a "storage porosity," which is associated with the rock matrix surrounding the tracer flow pathway. Tracers are assumed to enter and leave the rock matrix beyond the flow pathway by molecular diffusion according to Fick's first law (Freeze and Cherry 1979).

Only the two conservative tracer tests from UE-25c #2 to UE-25c #3, representing a scale of approximately 30 m, have been analyzed definitively. The values of flow porosity obtained range from 7 to 10 percent; the values of storage porosity obtained range from 9 to 19 percent, and the values of longitudinal dispersivity obtained range from 1.9 to 2.6 m. The values of storage porosity above are consistent with porosity values of 12 to 43 percent determined by laboratory analyses of core from the C-holes and gamma-gamma logging in UE-25c #1 (Geldon 1993).

Drill Hole Wash Complex–In 1981 and 1982, prior to the C-hole tests, a series of multiple-well hydraulic tests was conducted in Drill Hole Wash, about 2.6 km northwest of the C-hole Complex and 0.8 km west of Exile Hill. In these tests, which are not very well documented, the pumping well was UE-25b #1, and the observation well was UE-25a #1.

Borehole UE-25b #1, which is 1,220 m deep, penetrates the Calico Hills Formation, and the Prow Pass, Bullfrog, Tram, and Lithic Ridge Tuffs below the water table (Lobmeyer, Whitfield et al. 1983). The depth to water in UE-25b #1 was about 471 m in 1981. A tracejector flow survey conducted in UE-25b #1 during the first hydraulic test indicated that the Bullfrog Tuff produces 49 percent of the flow, the Calico Hills Formation produces 32 percent of the flow, and the Prow Pass Tuff produces 19 percent of the flow in this borehole. Total transmissive thickness in the borehole is 110 m.

Borehole UE-25a #1, which is about 110 m south-southwest of UE-25b #1, is 762 m deep and penetrates the Calico Hills Formation, Prow Pass Tuff, and Bullfrog Tuff below the water table (Spengler, Muller et al. 1979). Splays of the Drill Hole Wash fault intersect the borehole at depths of 67 to 100, 125 to 130, 203 to 221, 372 to 415, and 739 to 762 m. The borehole is open in the Bullfrog Tuff below a depth of 747 m. Because the entire open interval is faulted and fractured, it is assumed to be transmissive. No flow surveys have been run in the borehole to confirm this assumption.

Three open-hole hydraulic tests were conducted in borehole UE-25b #1 and monitored in borehole UE-25a #1 from August 3 to September 1, 1981 (Lahoud et al. 1984). Discharge rates during these tests ranged from 13.8 to 37.0 liters per second, and the length of the tests ranged from 3 to 9 days. Using a technique developed for a dual-porosity aquifer with fracture skin, Moench (1984) analyzed the combined drawdown data from the pumping and observation wells obtained during the third test and determined fracture transmissivity of 340 m^2/d, matrix transmissivity of 14 m^2/d, fracture hydraulic conductivity of 0.9 m/d, matrix hydraulic conductivity of 0.2 m/d, fracture storativity of 0.0006, and matrix storativity of 0.02.

A convergent tracer test was conducted in the Bullfrog Tuff at the Drill Hole Wash complex from June 23 to July 20, 1982 (Ogard et al. 1983). UE-25a #1 was used as the injection well, and UE-25 b#1 was used as the recovery well during the test. Assuming no collapse at the bottom of the well, the injection interval in UE-25a #1 was 15 m thick. The recovery interval in UE-25 b#1, in part determined by emplaced packers, was 60 m thick. Pumping to establish a steady-state hydraulic gradient for the tracer test began on June 22, 1982. On June 23, about 20.5 hours later, bromide tracer, in the form of sodium bromide, was injected into UE-25a #1. When pumping stopped on July 20, 1992, the tracer test ended. Average discharge during the tracer test was 12.8 liters per second.

According to Ogard et al. (1983), the bromide concentration in water recovered from borehole UE-25 b#1 increased from about 90 parts per billion on June 25 to about 280 parts per billion on June 27 and then decreased to the detection limit of 80 parts per billion on July 17.

5.3.3.2.2.3 Uncertainty of Hydraulic Properties

No matter how well the saturated-zone flow system at Yucca Mountain and vicinity is described, uncertainties of the data collected will always remain. A wide variety of methods have been used to estimate transmissivity, hydraulic conductivity, and aquifer storativity with mixed success and results (Luckey, Tucci et al. 1996, p. 32). For the determination of transmissivity and hydraulic conductivity, analysis of hydraulic tests require assumptions that frequently cannot be demonstrated to be correct. For aquifer storage properties, a detailed discussion is not possible because so little data exists for storage properties.

Transmissivity or hydraulic conductivity estimates based on hydraulic tests in boreholes are available from a variety of sources, which used a variety of methods for the Yucca Mountain area, but the analysis of the tests require assumptions that frequently cannot be demonstrated to be even approximately correct. The reliability of the commonly used slug tests for estimating values of transmissivity in fractured tuff at Yucca Mountain is suspect in many cases, in part because the testing could have caused either elastic fracture dilatation or fracture-deformation by hydraulic fracturing of the rock mass (Thordarson et al. 1985, p. 19; Stock and Healy 1988). Additionally, the results of many of these slug tests could not be analyzed reliably because conditions in the boreholes were not appropriate for the porous-medium type of analysis required of slug tests (Luckey, Tucci et al. 1996, p. 53).

The geometric mean value of the transmissivities listed in Table 5.3-35 was about 70 m^2/d, but the range was from 1.5 to 3,250 m^2/d. Transmissivity probably varies considerably around Yucca Mountain, but part of the range in values probably is due to uncertainty in the estimates. There is uncertainty in how to interpret the results of individual aquifer tests. Craig and Reed (1991, pp. 9-13) analyzed one aquifer test in borehole USW H-6 four different ways and got values of 240, 480, 210, and 370 m^2/d. They chose 240 m^2/d as the most reasonable value. The need to rely principally on single-borehole tests is a major cause of uncertainty in estimating both transmissive and storage properties.

Geldon (1996) reported that transmissivity and apparent hydraulic-conductivity values determined using multiple-borehole hydraulic tests tend to be much higher–1 to 2 orders of magnitude–than values reported for single-borehole tests conducted at the same borehole. For example, a single-borehole test at borehole UE-25c #3 resulted in a transmissivity value of 27.9 m^2/d, but results of a multiple-borehole test for the same interval in the borehole indicated a transmissivity value of 1,860 m^2/d. Geldon (1996) concluded that multiple-borehole tests generally sample a much larger volume of the aquifer material and incorporate a larger number of water-bearing fractures than single-borehole tests. Because most transmissivity and hydraulic-conductivity values at Yucca Mountain were obtained from single-borehole tests, the values listed in Tables 5.3-35 and 5.3-36 may not be representative values appropriate for scales of tens to hundreds of meters (Luckey, Tucci et al. 1996, pp. 36-37).

The hydraulic characteristics of a particular interval at a borehole are highly dependent on the fracture characteristics of that interval. A large conductive feature, such as the fault penetrated in the lower part of borehole UE-25c #1, can contribute a large percentage of the total flow to a borehole. The tested interval containing such a feature can have a disproportionately large transmissivity and apparent hydraulic conductivity compared to other intervals in the same borehole. Intervals that have a few conductive fractures generally contribute little water to the borehole and may have small transmissivity compared to other intervals in the borehole (Luckey, Tucci et al. 1996, p. 37).

Groundwater flow models have been used to estimate transmissivity over large areas. Waddell (1982), in his groundwater flow model of the Death Valley flow system, had two zones that included the Yucca Mountain area. Zone 1, which included the western and central part of the Yucca Mountain area, had an estimated transmissivity of about 700 m^2/d, whereas Zone 8, which included the eastern part of the Yucca Mountain area, had an estimated transmissivity of about 7 m^2/d. The zones were quite large, and the model was not intended to be used to estimate transmissivity at Yucca Mountain, so these values need to be used with discretion. Czarnecki and Waddell (1984) modeled the subbasin of the Death Valley flow system that includes Yucca Mountain. Their Zone 10, which included the western part of Yucca Mountain, had an estimated transmissivity of about 80 m^2/d, whereas Zone 6, which included the eastern part of Yucca Mountain, had an estimated transmissivity of more than 3,000 m^2/d. Transmissivity estimates from field tests and simulations seem to be generally consistent, both in the order of magnitude of the values and the range in values. However, the field tests and simulations have a large uncertainty associated with them (Luckey, Tucci et al. 1996, pp. 53-54).

Vertical hydraulic conductivity and anisotropy are parameters that need to be incorporated into 3-D flow models, but frequently are not simply because no data on them are available. Tests to estimate vertical hydraulic conductivity were conducted only at borehole USW H-5 (Robison and Craig 1991, p. 26) and at the C-hole Complex (Geldon 1996). Data from borehole USW H-5 should be used with discretion because it may not be representative of the area. As was discussed in the "Moderate hydraulic gradient" section, USW H-5 has a potentiometric level consistent with boreholes west of the Solitario Canyon fault, even though the borehole is east of the fault. Anisotropy was estimated only for borehole USW H-4 (Erikson and Waddell 1985, pp. 24-29). These parameters are uncertain, both because of the few estimates available and because the estimates themselves are uncertain.

Storage properties of the saturated-zone flow system at Yucca Mountain are not discussed in great detail because so little data exist for storage properties. Estimates for storage properties are uncertain even when aquifer tests are conducted with nearby observation holes; storage properties are particularly unreliable for single-hole tests, which is the usual type for Yucca Mountain. Therefore, the storage properties discussed in this subsection are estimated or assumed only for a few boreholes in the Yucca Mountain area, and these estimates are quite different from borehole to borehole.

Storage properties are not required for steady-state simulations of groundwater flow, but they are required for time-dependent simulations. The range of values for storativity for water table conditions (unconfined flow) may be large because effective porosity may be as small as 0.001 for

fractured tuff and may exceed 0.20 for alluvium or nonwelded tuff. For confined flow or deep unconfined flow within the saturated zone, the storativity could be an order of magnitude or more smaller (Luckey, Tucci et al. 1996, p. 53).

Tests to determine effective (or flow) porosity and dispersivity, which are critical in transport simulations, also require multiple-well test sites. In addition, they require data from long-term conservative tracer experiments. As discussed in the preceding Subsection (5.3.3.2.2.2), tracer tests at the C-hole Complex provided effective-porosity estimates of .07 to 0.10, whereas a test in Drill Hole Wash provided an estimate of 0.0003. Longitudinal dispersivities were estimated as 1.9 to 2.6 m at the C-hole Complex and 4.6 m at Drill Hole Wash. A very tentative explanation might suggest that the larger effective porosity but smaller dispersivity at the C-hole Complex indicate more pervasive fracturing relative to a more dominant, fracture-controlled flowpath in Drill Hole Wash. Whether this tentative characterization is reasonable and, particularly, whether these ranges of effective porosity and longitudinal dispersivity are representative cannot be assessed on the basis of this small data set.

5.3.4 Unsaturated-Zone System

This section consists of four major subsections that describe observations, data, interpretations, and conclusions about the physical setting and important processes that characterize the 500 to 750-m thick unsaturated zone at Yucca Mountain. The first major subsection (5.3.4.1, Site Infiltration) describes the physical, climatic, and hydrologic factors that control infiltration in the relatively shallow part of the unsaturated zone. The data and interpretations that were used to develop a conceptual and numerical model of infiltration for the site also are described. The distribution of infiltration under both present-day and possible future wetter climatic conditions is also described. The second major subsection (5.3.4.2, Characterization of the Deep Unsaturated Zone) describes the apparent behavior of gas and water that are present in and moving through the deeper part of the unsaturated zone. Observations of pneumatic pressure and water potential are described in detail, as well as their implications for overall behavior of the unsaturated-zone system at the site. Occurrences of perched water are described and the chemical and isotopic content of gas and water are used to infer the nature of fluid movement within the unsaturated zone. The third major subsection (5.3.4.3, Movement of Water in the Unsaturated Zone) integrates a wide range of unsaturated-zone data and insights into a consistent conceptual model of movement of water through Yucca Mountain. The fourth and final major subsection (5.3.4.4, Three-Dimensional Unsaturated-Zone Modeling) describes how the data and interpretations from the earlier three major subsections were represented in the site-scale unsaturated-zone flow model of Yucca Mountain. This subsection also describes how the numerical flow model was calibrated using site data and applied to predict the overall behavior of the unsaturated zone under both present-day and possible future climatic conditions.

5.3.4.1 Site Infiltration

A quantitative description of net infiltration at the Yucca Mountain Site is needed for defining the upper boundary condition of unsaturated-zone flow models that are being to evaluate the performance of the potential repository. Infiltration is defined as the flow of surface water downward across the atmosphere-soil or the atmosphere-bedrock interface. Net infiltration is the flow of water downward below the zone of evapotranspiration. Recharge is the flow of water from the unsaturated zone downward to the water table where it enters the saturated zone. Percolation is the flow of net infiltration through the deep unsaturated zone and is discussed in later subsections of this report. Estimates of net infiltration may not be directly relatable to recharge because:

- The length of time being evaluated or considered
- The extent to which the thick unsaturated zone at Yucca Mountain attenuates or smooths the temporal and spatial variability of net infiltration
- The combined effects of lateral flow and vapor flow (or atmospheric pumping) on percolation

The estimates of net infiltration are indicative only of a potential recharge estimate for current climatic conditions.

The general area of interest for infiltration and recharge studies is defined as the Yucca Mountain region and includes the southern part of the Great Basin Desert and the northern part of the Mojave Desert (Figure 5.3-85). The Yucca Mountain region includes several study locations for regional saturated-zone recharge, analog sites for estimation of infiltration and recharge under potential future climates, the area of the regional 3-D saturated-zone flow model, and the regional precipitation network used for infiltration and water-balance studies. The area for estimates of site net infiltration is defined by a 228.24 km^2 rectangular region bounded by the UTM (zone 11) coordinates of 544661 to 555641 m easting and 4067133 to 4087833 m northing (Figure 5.3-86). This "site area" encompasses the area of the site-scale 3-D unsaturated-zone flow model (Bodvarsson et al. 1997) that is centered over the area of the potential repository. Estimates of presentday and potential future net infiltration rates are provided as detailed mappings of temporally and spatially varying time-averaged fluxes. The estimates are used as input for defining the upper boundary condition of the site-scale unsaturated-zone flow model described in Subsection 5.3.4.4.

Subsection 5.3.4.1.1 presents an historical overview of studies of infiltration and recharge in the Southern Basin and Range Province relevant to the Yucca Mountain Site. The overview provides a summary of the various methods for quantifying infiltration and recharge. Subsection 5.3.4.1.2 presents a summary of the conceptual model of infiltration for Yucca Mountain as described by A.L. Flint et al. (1996). The conceptual model includes a description of the physical processes controlling infiltration and net infiltration at the site, as well as the importance of the spatial and temporal variability of these processes. Subsection 5.3.4.1.3 presents a summary of the results of infiltration studies and monitoring conducted at the site using the methods described in Subsection 5.3.2.1.1, and described how these results support or weaken the conceptual model. Subsection 5.3.4.1.4

describes the development of the numerical infiltration model for the site as developed by A.L. Flint et al. (1996), including model calibration using the data from infiltration studies and monitoring. Subsection 5.3.4.1.5 presents estimates of the spatial and temporal distribution of numerically modeled net infiltration for the site under current and possible future climatic conditions. Subsection 5.3.4.1.6 concludes with a summary of infiltration modeling results, the limitations of those results, and work in progress to address some of these limitations.

5.3.4.1.1 Overview of Studies and Methods for Determining Net Infiltration and Recharge in the Vicinity of the Site

Studies of recharge have been conducted at various locations throughout the Yucca Mountain region from the 1940s through the present for the purpose of evaluating groundwater resources in the southwestern United States. Studies of infiltration and recharge conducted for the purpose of evaluating potential radioactive waste repository sites (both high level and low level) in the vicinity of Yucca Mountain began in about 1980 (Winograd 1981). These studies include a broad range of investigative techniques and methods including:

- Flux calculations using measured hydraulic properties
- Simple water-balance models
- Statistical models
- Geochemical methods
- Temperature models
- Deterministic water-balance models

The following subsections summarize these methods, their results, and their advantages and limitations.

5.3.4.1.1.1 Flux Calculations Using Measured Hydraulic Properties

Net infiltration can be estimated based on measurements of water contents and hydraulic properties of soil and/or bedrock materials and assumptions concerning the hydraulic gradient. This method for direct calculation of unsaturated flux involves use of a measurement or estimate of the vertical water-potential gradient for a selected depth profile and calculation of the flux based on the known or estimated moisture-characteristic curves and the assumption that lateral flow components in the profile are negligible. In cases where only the water-content profile is known, the moisture-characteristic curve can be used to calculate both the water-potential gradient and the flux, provided both moisture-characteristic curves are defined. This method requires sufficient hydrologic-properties data for the application of an appropriate or fitted moisture-characteristic curve. For estimating flux through soils, soil texture data and empirical equations can be used to estimate moisture-characteristic curves (Campbell 1985). If time-continuous measurements of water contents or water potentials are not made, flux estimates require an assumption of steady-state conditions, which may not be valid for shallow depths and in arid environments characterized by episodic infiltration events. Finally, the method is problematic for conditions dominated by fracture flow.

Steady-state flux estimates of 2 mm/year net infiltration were made by Winograd (1981) for Sedan Crater located on the Nevada Test Site approximately 50 km northwest of the potential repository site. This site is at an elevation of 1,400 m, receives approximately 200 mm/year mean annual precipitation (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press), and consists of thick alluvium deposits. Net infiltration was estimated at 0.04 mm/year by Nichols (1987) using a steady-state method for a study site in thick alluvial deposits in the Amargosa Desert approximately 30 km west of Yucca Mountain. This site is at a lower elevation (about 800 m) than the Sedan Crater site and receives about 130 mm/year mean annual precipitation. The estimate made by Nichols was based on moisture-characteristic data and water-potential measurements for 9 m deep caisson installed at the site.

Spatially distributed estimates of net infiltration for the initial area of the 3-D site-scale unsaturated-zone flow model (Wittwer et al. 1995) were made by A.L. Flint and Flint (1994) using hydraulic property measurements of the bedrock matrix underlying the soil, an 11-year record of monthly water-content measurements at depths of 10 to 20 m, and an assumed unit gradient. The matrix properties and water-content data were obtained from the neutron-series boreholes used for monitoring infiltration (see Subsection 5.3.2.1.1.). A maximum net infiltration of 13.4 mm/year was obtained for the northern part of the study area and a minimum of 0.02 mm/year were obtained for most of the area overlying the potential repository block. The approach yielded a spatially averaged net infiltration of 1.4 mm/year for the entire study area. The advantage of the approach used by A.L. Flint and Flint (1994) is that it indicated the high degree of spatial variability in net infiltration due to the variability of soil and bedrock properties. The weaknesses of this approach were:

- Episodic pulses of infiltration through fractures could not be accounted for.
- Assumptions had to be made concerning water contents at locations between monitoring sites.
- The method could not be readily applied to provide predictions of net infiltration rates in response to possible future climatic conditions.

More recent measurements of water potentials for several depths in the soil profile have been obtained at the site using heat dissipation probes (Flint, A.L. et al. 1996). These measurements were applied to the moisture-characteristiccurves measured for the soils to obtain estimates of infiltration into bedrock. The estimates of infiltration were as high as 150 mm for a 1-month period following the storm of March 10-11, 1995, and were in agreement with estimates obtained from measured changes in water-content profiles at nearby boreholes (Hudson and Flint 1996). These results were important in providing supporting data for developing the conceptual model of infiltration and are discussed in more detail below.

5.3.4.1.1.2 Water Balance Models for Estimating Net Infiltration and Recharge

Estimates of net infiltration and recharge can be made using measurements or estimates of the components of the water-balance equation. The most direct application of this approach is the use of lysimeters that allow for continuous measurement of soil-water content changes in a closed

system. This method works well in agriculture where lysimeters can be constructed which are representative of the field environment. Unfortunately, the construction of representative lysimeters becomes very problematic for characterizing the water balance over large heterogenous areas such as the Yucca Mountain Site that consists of shallow soils and rugged terrain. In general, a direct application of the water-balance approach to estimating net infiltration in arid environments such as Yucca Mountain is problematic because the net infiltration term is small compared to the other components and is usually well below the magnitude of the measurement or estimation error of the other components.

Basin-wide water balance studies have been conducted in the Yucca Mountain region to evaluate groundwater resources. Using a water balance approach and information on climate, vegetation, geology, streamflow, and groundwater levels and withdrawal rates, recharge in the Yucca Mountain region was estimated by Maxey and Eakin (1949) for the White River Valley, a 4,196 km^2 basin east of the Grant Range in Nevada (Figure 5.3-85). Their results indicated an average annual recharge volume of approximately 49.3 m^3, or 11.8 mm basin-wide, which is 3.7 percent of a basin-wide average annual precipitation estimate of approximately 320 mm. Their water-balance method used an empirical linear correlation between recharge and precipitation ranges or zones developed using the results from various study basins in Eastern and Central Nevada as well as study sites in the Las Vegas area, including the Spring Mountains of Southern Nevada (Figure 5.3-85). The empirical correlation is defined by a linear stepfunction that estimates average annual recharge to be zero for average annual precipitation less than 203 mm, 3 percent of average annual precipitation for a range of 203 to 305 mm average annual precipitation, 7 percent of average annual precipitation for a range of 305 to 381 mm average annual precipitation, 15 percent of average annual precipitation for a range of 381 to 508 mm average annual precipitation, and 25 percent of average annual precipitation for average annual precipitation greater than 508 mm.

Application of the Maxey-Eakin step function to basins or locations where the water balance can not be directly verified using available data assumes that the empirical relation developed at the study sites analyzed for the Maxey-Eakin model is representative of the hydrologic characteristics of the location or area where estimates are desired. Using estimates of mean annual precipitation, Rush (1970) applied the Maxey-Eakin empirical relation to obtain a recharge estimate of 1.5 mm/year for Jackass Flats, which is approximately 10 km east of the potential repository site. Subsequent recharge estimates for locations at Yucca Mountain included 6 mm/year by Scott et al. (1983), 0 to 2 mm/year by Czarnecki (1985), and 4.5 mm/year by Montazer and Wilson (1984). The variability in the estimates primarily reflects the spacial variability in estimates of mean annual precipitation.

Recent estimates of recharge in the Death Valley region were obtained by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press) a modified version of the Maxey-Eakin model and a geostatistical model for estimating mean annual precipitation using digital elevation data. Estimates for the Yucca Mountain Site using the geostatistical model ranged from 0.1 to 5 mm/year, depending on elevation. A more elaborate modification of the Maxey-Eakin model was defined by D'Agnese et al. (1997) using a geographic information system model for incorporating vegetation data, topographic data for slope and aspect calculations, bedrock permeability data, and the geostatistical estimates of mean annual precipitation provided by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). This recharge model was used to help define the

boundary conditions for the regional 3-D saturated-zone flow model developed by D'Agnese et al. (1997) for the Death Valley Basin, and is discussed in detail in Subsection 5.2.3. The model was calibrated using inverse modeling, regional discharge data, and data defining the regional potentiometric surface, and thus represents a combination of water balance and empirical modeling methods. The estimate of recharge obtained for the 1,500 m by 1,500 m model element overlying the potential repository block was approximately 2 mm/year.

Although the Maxey-Eakin method is efficient to apply and can be used to predict recharge if average annual precipitation for potential future climates is known, the original method does not account for site-specific differences in topography, vegetation, soil type, soil thickness, and geology. The model presented by D'Agnese et al. (1997) incorporates spatial data using geographic information system applications to account for much of this site-specific variability, but groups the data into a limited number of discrete recharge zones that diminishes the uniqueness of individual locations and results in a loss of spatial detail on the scale of the net infiltration study area for Yucca Mountain (Figure 5.3-86). In addition, recharge estimates based on mean annual precipitation cannot account for the timing, frequency, intensity, and duration of precipitation or precipitation occurring as snow, all of which affect infiltration and recharge.

Measurements of stream discharge, changes in vertical water-content profiles, and the height of the water table in Fortymile Wash on the Nevada Test Site during the wetter-than-normal winter of 1991 to 1992 provide evidence of recharge in response to precipitation from large storms (Savard 1994, 1995b). Estimates of recharge by Osterkamp et al. (1994) for a 20,000 km^2 area of the Amargosa River basin indicate a basin-wide average annual recharge rate of 1.88 mm, which is approximately 1.6 percent of an estimated basin-wide average annual precipitation of 180 mm. The average annual recharge estimates for the Amargosa River basin and several sub-basins were obtained using a combination of geomorphic field techniques and streamflow simulation models for calculating channel transmission losses and upland recharge. The study by Osterkamp et al. (1994) also indicates that approximately 90 percent of the total average annual recharge volume of 20.5 m^3 occurs as channel transmission losses following high magnitude, low frequency precipitation events, while the remaining 10 percent occurs as net infiltration in interchannel areas. The highest average annual recharge values, which were approximately 3 percent of average annual precipitation, were estimated for the headwater areas of Fortymile Wash and the upper Amargosa River, reflecting an increase in precipitation and runoff rates for the higher elevations of Pahute Mesa in the northern part of the Nevada Test Site. The lowest average annual recharge values, which were approximately 0.5 percent of average annual precipitation, were estimated for the southern part of the Amargosa River basin where topographic relief is low, elevation and precipitation are the lowest, and evapotranspiration rates are the highest. The results obtained by Osterkamp et al. (1994) for the headwater areas of the upper Amargosa River and Fortymile Wash, where precipitation records and geostatistical estimates indicate average annual precipitation to be from 200 to 300 mm (Hevesi, Flint et al. 1991), also are consistent with the Maxey-Eakin step function. As indicated by Osterkamp et al. (1994), the variability of recharge along channel sections of alluvial fans and basins is likely to be high, depending on the variability of average discharge volumes being routed from upstream drainages. Average recharge rates along alluvial channels are expected to be much higher than for interchannel areas of alluvial fans and basins, and are likely to be better correlated to the frequency of storms

generating runoff in the upstream drainages than to average annual precipitation at the location of the channel.

5.3.4.1.1.3 Statistical Models for Estimating Net Infiltration at the Site

Measured changes in water-content profiles at 60 neutron boreholes located over the area of the potential repository were used to estimate net infiltration by assuming that flow is vertical and that changes in water content below a depth of 2 m into bedrock are indicative of transient occurrences of net infiltration (Hudson and Flint 1996). The measurement of water-content profiles for the purpose of monitoring infiltration is described in Subsection 5.3.2.1.1, and selected examples of changes in measured water-content profiles through time are presented in Subsection 5.3.4.1.5. To reduce the sensitivity of the method applied by Hudson and Flint (1996) to apparent changes in water content caused by measurement error, a numeric filter was applied to the calculated relative water-content changes. The developed net infiltration data were modeled using multiple linear regression and three variables: total water year precipitation, soil thickness, and topographic position, which are all directly or indirectly dependent on elevation. The spatial distribution of precipitation was defined using a geostatistical model developed by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press) and the 30 m digital elevation data provided by U.S. Geological Survey 7.5-minute digital evaluation model. Soil thickness was defined using a combination of ground slope and mapped soil types available as a geographic information system coverage (Guertal and Davies 1996; S.C. Lundstrom et al., *Preliminary Surficial Deposits Map of the Northeast Quarter of the Busted Butte 7.5' Quadrangle, Nye County, Nevada*, U.S. Geological Survey, Open File Report, in press, GS940108315142.004;S.C. Lundstrom et al., *Preliminary Surficial Deposits Map of the Northwest Quarter of the Busted Butte 7.5-minute Quadrangle*, U.S. Geological Survey, Open File Report, in press, GS940708315142.008;S.C. Lundstrom and E.M. Taylor, *Preliminary Surficial Deposits Map of the Southern Half of the Topopah Spring NW 7.5-minute Quadrangle*, USGS-OFR-95-132,in press, GS940108315142.005, MOL.19960403.0218; Lundstrom, Whitney et al. 1995). Topographic position was defined using the available channel network geographic information system coverage and the 30 m elevation grid. The advantage of this approach is that it is well conditioned to the measured water-content profiles and it can be efficiently applied using geographic information system techniques to provide a spatially detailed estimate of net infiltration for the potential repository site. Estimates of net infiltration obtained for the site infiltration study area (Figure 5.3-86) ranged from 0 to 45 mm/year, with an areally averaged estimate of 11.9 mm/year for the site. The disadvantage of the method applied by Hudson and Flint (1996) is that the three variables used in the model account for only 40 percent of the variability observed in the developed net infiltration data. This is primarily due to the model's inability to account for annual and seasonal variability in the timing, frequency, and duration of precipitation, the effects of spatially varying soil and bedrock properties, or the affects of topography on potential evapotranspiration.

5.3.4.1.1.4 Geochemical Methods

Estimates of recharge using the chloride-balance technique for 16 basins in Nevada were found to be in fairly good agreement with estimates obtained by the Maxey-Eakin linear step function (Dettinger 1989b). (An application of the chloride-balance technique for estimating infiltration at Yucca Mountain itself is described in Subsection 5.3.4.1.5.2.) In a 6-year study of two upland basins

selected as "analog recharge" sites and representing the effects of possible wetter climate at Yucca Mountain, Lichty and McKinley (1995) show the chloride-balance method to be more robust than a water-balance model that used a deterministic watershed model Precipitation Runoff Modeling System for estimating basin-wide recharge. They attribute the robustness of the chloride-balance method to the small number of measured parameters required as compared to the number of parameters needed for defining a deterministic watershed model. For the 0.9 km^2 study basin (East Stewart Creek) located in the Toiyabe Range of Central Nevada (Figure 5.3-85) at an average elevation of approximately 3,100 m, their results indicate average annual recharge to be 300 to 320 mm, or about 50 percent of the estimated average annual precipitation of 639 mm. For the 4.2 km^2 study basin (Three Springs) located in the Kawich Range of South-Central Nevada (Figure 5.3-85) at an average elevation of 2,500 m, Lichty and McKinley (1995) estimate average annual recharge to be 11 to 33 mm, or 3.3 to 9.8 percent of an average annual precipitation value of 336 mm. The results obtained for the basin in the Kawich Range are consistent with the estimate obtained by the Maxey-Eakin step-function. The results obtained for the basin in the Toiyabe Range are approximately 100 percent greater than the Maxey-Eakin estimate of 160 mm average annual recharge. The apparent inconsistency in results may be attributed to localized differences in factors such as climate, physiography, and geology for smaller basins.

Point estimates of net infiltration or recharge using the chloride-balance technique tend to be less robust than basin-wide estimates because of additional assumptions concerning vertical flow and surface-water flow. Values of net infiltration estimated at Yucca Mountain using the chloride-balance technique range from 0 to 5.4 mm/year (Fabryka-Martin, Wightman et al. 1994).

5.3.4.1.1.5 Temperature Models

Recent estimates of recharge at the site have been made using thermal models and borehole temperature data (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997). The method uses the known geothermal gradient, estimates of the average temperature of water entering the top of the unsaturated zone as net infiltration, and an assumption of vertical flux through the unsaturated zone. Estimates of percolation flux based on thermal modeling and borehole temperature data ranged from 0.5 to 20 mm for various locations within the site. The results of these analyses are discussed in detail in Subsection 5.3.4.2.3 of this report.

5.3.4.1.1.6 Deterministic Water-Balance Modeling of Net Infiltration and Recharge

Deterministic numerical models are an extension of the water-balance approach in that they allow estimates of infiltration and recharge to be calculated based on mathematical representations of the physical processes linking the components of the water balance. In conjunction with the chloride-balance modeling approach used by Lichty and McKinley (1995) for estimating recharge at the two wetter climate analog-recharge sites discussed above, their study included the application of the U.S. Geological Survey's Precipitation Runoff Modeling System (Leavesley et al. 1983) watershed model. The application of this method required measurement and estimation of basin-wide precipitation, snow-pack depth and density, stream discharge, and meteorological parameters for

defining potential evapotranspiration. Spatially varying parameters such as canopy cover, snow-pack cover and density, and soil characteristics needed to be defined, as well as various additional parameters needed for defining empirical relations such as actual evapotranspiration. The model was calibrated against the measured streamflow hydrographs for each basin. The model results compared reasonably well with results obtained using chloride-balance estimates. However, it was concluded that the reliability of the numerical model was less than the chloride-balance model because of uncertainties in measured and estimated parameters needed for the model.

A numerical model of net infiltration has been developed for the Yucca Mountain Site using detailed spatially variable data provided by digital elevation data and geographic information system coverages for soil type, soil depth, bedrock geology, and channel networks (Flint, A.L. et al. 1996). The model requires an initial estimate of soil-water content and a record of daily precipitation amounts for the intended period of simulation. The spatial parameters are defined uniquely for all 253,597 30 m^2 grid cells of the digital elevation model using geographic information system applications to provide a high degree of spatial detail in model results. The model was calibrated using daily precipitation data from 1980 through 1995 and water-content profiles measured from 1984 to 1995 in 99 neutron-access boreholes (N series) located at Yucca Mountain. Results obtained from this model provide an areally averaged estimate of 3.25 mm/year average annual net infiltration for the entire study area, and an estimate of 7.1 mm/year for the smaller 11.61 km^2 area of the potential repository site. The model was used to provide a prediction of future net infiltration at the site using stochastically generated 100-year records of daily precipitation representative of possible future wetter climatic conditions in the Yucca Mountain region. The results obtained using the numerical model are in good general agreement with the Maxey-Eakin estimates of recharge for analog climate sites in the Yucca Mountain region, with the Lichty and McKinley (1995) estimate of recharge for Three-Springs Basin, and with results obtained at Yucca Mountain using temperature-profile modeling. The development and calibration of the numerical net infiltration model, along with the spatially and temporally varying estimates of net infiltration provided by the model, are discussed in detail below.

5.3.4.1.2 Conceptual Model of Net Infiltration

Prior to development of a numerical infiltration model, a conceptual model of net infiltration for Yucca Mountain was developed (Flint, A.L. et al. 1996). The conceptual model defines the physical processes determining infiltration and net infiltration, and is based on evidence provided from field studies at Yucca Mountain, combined with an established understanding of hydrologic processes documented in the literature. The hydrologic cycle provides a framework for the processes on the surface and in the shallow subsurface that affect net infiltration and thus potential recharge. The processes include precipitation, infiltration, runoff and runon, evapotranspiration, and the redistribution of moisture in the shallow subsurface (Figure 5.3-87). Precipitation is the dominant hydrologic process at the site because it is the source of all moisture for the surface and shallow subsurface (less than 6 m) at Yucca Mountain, excluding water introduced to the site by human activity for dust control, drilling, and waste disposal. There are no permanent streams or surface-water bodies affecting the site. Precipitation is dependent mostly on meteorological factors, but geographic location, elevation, and physiography also are important. Evapotranspiration is the second most dominant hydrologic process (in terms of the total volume of water involved) at Yucca

Mountain and is dependent on vegetation, the distribution of available moisture stored in the shallow subsurface, and potential evapotranspiration, which is determined by the energy balance. Redistribution of moisture in the shallow subsurface occurs in response to gravity and matrix potentials and is strongly dependent on soil and bedrock properties. The removal of water through evapotranspiration is dynamically integrated with the redistribution of moisture in the shallow subsurface. Surface water runoff occurs when the storage capacity of the shallow subsurface is exceeded or when the precipitation rate exceeds the infiltration capacity of surficial materials. Runon and the routing of surface flow is dependent mostly on surficial material properties, topography, and channel geometry. Net infiltration is the percolation of infiltrated water below the zone of evapotranspiration, which is both temporally and spatially variable.

5.3.4.1.2.1 Field Water Balance

The field water balance is a fundamental aspect of conceptual and numerical infiltration models (Flint, A.L. et al. 1996). The water balance is based on the principle of the conservation of mass for water:

$$P+A+U+\Delta W_s+\Delta S_s+\Delta B_s+L_i+R_{on}-R_{off}-D-E-T-L_o-E_x=0 \qquad \text{(Eq. 5.3-3)}$$

where P is precipitation, A is applied water (man induced), U is upward flow, ΔW_s is change in soil-water storage, ΔS_s is change is surface storage, ΔB_s is change in above-ground biomass storage, L_i is lateral flow in, R_{on} is surface runon, R_{off} is surface runoff, D is deep drainage or percolation, E is evaporation, T is transpiration, L_o is lateral flow out, and E_x is extraction of water (man induced). Equation 5.3-3 states that the sum of all inputs, outputs, and changes in storage in the hydrologic system must equal zero. To be applied, Equation 5.3-3 must be defined over some arbitrary time interval (i.e., day, year) and over some arbitrary volume or depth in the soil.

In most cases, the general form of the water-balance equation can be simplified by assuming one or more of the terms to be zero or negligible in magnitude. For example, the terms B_s, R_{on}, and R_{off} can often be set to zero in cases involving arid and semi-arid sites where overland flow occurs only in direct response to relatively infrequent, larger magnitude precipitation events. In the case of 1-D flow with zero or negligible runon and runoff under current climatic conditions, negligible upward flow, man-induced fluxes, and surface storage, the water-balance equation becomes:

$$P+\Delta W_s-D-E-T=0 \qquad \text{(Eq. 5.3-4)}$$

This equation provides a more practical starting point for analyzing net infiltration (D). For some applications, Equation 5.3-4 is further simplified by setting P to zero (for periods having no precipitation), assuming D to be negligible compared to the other terms, and combining E and T into a single term of evapotranspiration. In this form, the water balance can provide a method of using measurements of ΔW_s (change in soil-water storage) to estimate evapotranspiration (Nichols and Cuenca 1993).

Determination of net infiltration rates using water-balance calculations requires very precise and accurate measurements of the components of the water-balance equation (Equation 5.3-3). In the absence of such measurements, the water-balance equation can still be applied in analyzing the spatial and temporal distribution of infiltration rates using the following input (Flint, A.L. et al. 1996):

- A conceptual understanding of the physical processes involved
- Deterministic or stochastic numerical models for approximating each of the physical processes
- An adequate coupling of the physical processes
- Direct or indirect measurements of as many of the components of the water balance as possible to provide some means of calibrating the numerical model

The dominant physical processes in the general field water-balance equation are precipitation, evapotranspiration, overland flow (runoff and runon), infiltration, and redistribution of the soil moisture. It is also important to consider the variability of these processes in space and time. A calibrated model can correct, to a certain extent, for a lack of detailed measurements if it can bound the solution within the range of robust field observations.

Because net infiltration is the upper boundary condition for flux in the unsaturated zone, it is not necessarily equivalent to recharge, which is the flux across the water table (Flint, A.L. et al. 1996). The surface and shallow subsurface environments are distinct from deeper zones of percolation (depths greater than 50 m) because conditions tend to be transient and because non-equilibrium conditions are likely to exist between soils and bedrock and between bedrock matrix and fractures. Annual, seasonal, and diurnal variations in meteorology and vegetation cause temporal variability in net infiltration, while topography, vegetation, soil depth, soil properties, and properties of the underlying bedrock cause spatial variability in net infiltration. Because of the thickness of the unsaturated zone at Yucca Mountain, net infiltration is strongly affected by short-term variability in climatic conditions, whereas recharge is more likely to be affected by longer term, average climatic conditions. However, if flux through the unsaturated zone is at steady-state and has no horizontal components, then time-averaged net infiltration should be representative of recharge at the site. The potential ability (or inability) of the thick unsaturated zone to dampen and smooth out spatially and temporally variable net infiltration to a more uniformly distributed flux at the potentially repository level will be discussed in detail in later subsections.

5.3.4.1.2.2 Climatic Factors

Of all the factors affecting net infiltration, changes in climate will probably have the greatest impact on net infiltration and the surface hydrology at Yucca Mountain over the next 10,000 years (Flint, A.L. et al. 1996). Water transport and retention processes in surficial materials (soils and bedrock) are affected by high variability in solar radiation fluxes, diurnal and seasonal temperature cycles, relative humidity, and periodic inputs from precipitation, in the form of either rain or snow followed

by extended periods of drought. Hydrologic responses at the surface may or may not be translated to a change in flux across the potential repository horizon or a change in recharge, depending on the persistence or the intensity of a climatic change. The evaluation of past climate is needed to evaluate the control of climate on the hydrologic conditions at depth. The evaluation of current climate is necessary to relate changes in climate to near-surface processes, such as infiltration, runoff, redistribution, and evapotranspiration. Current climatic conditions for the site and the Yucca Mountain region are discussed in detail in Section 4. Predictions of potential future climates are critical to developing predictions of future net infiltration and potential recharge at the site.

The primary climatic parameter affecting surface hydrology and net infiltration is precipitation, including timing, frequency, duration, and intensity of precipitation (Flint, A.L. et al. 1996). The dominant controlling factor determining the occurrence of precipitation at the site is the synoptic-scale weather circulation pattern in relation to fixed-geographic moisture sources and local physiographic features (Hevesi, Ambos et al. 1996). The moisture sources for the Yucca Mountain region are the Pacific Ocean, the Gulf of California, and the Gulf of Mexico. Most precipitation at the site occurs in response to low pressure systems which are steered eastward by the jet stream and bring moisture inland from the Pacific Ocean. These weather patterns generally occur during the winter, but can also occur during the winter-summer and summer-winter transition periods. Winter storms tend to result in low intensity (light) precipitation and may occasionally last up to several days. Orographic influences tend to increase precipitation frequency and intensity for higher elevations, and precipitation often occurs as snow. On a regional scale, the orographic effects are more pronounced, particularly in the vicinity of major mountain ranges such as the Spring Mountains and the Panimint Range (Hevesi, Ambos et al. 1996). A significant positive correlation between calculated average annual precipitation and station elevation was indicated by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press) for 114 precipitation stations in the Yucca Mountain region, having at least 8 years of complete record (Figure 5.3-88). The higher elevations of the mountain ranges correspond to the locations of maximum precipitation and also to the locations of maximum recharge. During the summer, the jet stream migrates to the north, and weather circulation in the Yucca Mountain region tends to be dominated by high-pressure cells that cause a reverse circulation, or monsoonal flow. Surface heating and convection of air masses carrying moisture from the Gulf of Mexico, the Gulf of California, and the Pacific Ocean results in sporadic but occasionally intense thunderstorms during the southwestern summer monsoon (Hevesi, Ambos et al. 1996). These storms tend to be of short duration, but can produce high intensity (heavy) precipitation which often results in flash flooding.

The combined effect of the average winter and summer weather circulation patterns is to create zones of precipitation excess and deficit relative to the regional mean, for a given elevation, from west to east across the Southern Nevada region, with Yucca Mountain being located in an approximate transition zone between the deficit zone to the west and the excess zone to the east (French 1983). The precipitation deficit (relative to the regional mean) is caused by the Sierra Nevada, which causes a regional rain-shadow during winter and a westward diminishing of moisture from the monsoon during the summer. Generally, precipitation stations east of longitude 115°45' W receive from 1.5 to 2.5 times more precipitation than stations at similar elevations located west of longitude 116°45' W (Winograd and Thordarson 1975). The combined effects of seasonal weather patterns and orographic influences causes considerable spatial variability in average annual precipitation

throughout the Yucca Mountain region. Average annual precipitation estimates provided by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123,in press) over the area of the Nevada Test Site range from a maximum of 370 mm in the Belted Range to a minimum of 110 mm in the Amargosa Desert (Figure 5.3-89). Annual precipitation within the Yucca Mountain infiltration study area averages 170 mm and ranges from a minimum of 130 mm at low elevations along the southern boundary to a maximum of 250 mm at high elevations along the northern boundary.

In addition to the seasonal variability in precipitation, the arid climate at Yucca Mountain is characterized by considerable annual variability. Measured total annual precipitation at the site ranged from a minimum of 12 mm for water year 1989 to a maximum of 312 mm for water 1993 at one location (Flint, A.L. and Davies 1997). For the precipitation station 4JA located approximately 10 km east of Yucca Mountain and having a 36-year record of daily precipitation, total annual precipitation ranged from a minimum of 35 mm for calendar year 1975 to a maximum of 290 mm for calendar year 1978 (Figure 5.3-90), as compared to the calculated mean of 133 mm (Figure 5.3-91). The 5-year sliding annual mean calculated using this record indicates a possible trend of increasing precipitation from 1960 to 1995, with the wettest 5-year sliding mean occurring from 1991 through 1995. The wettest average 3-month total is 60 mm for January through March and the driest average 3-month total is 16 mm for April through June. The 1991 to 1995 period contained the wettest average January through March in the entire record, while the 1983 through 1987 period contained the wettest July through September in the entire record. The record indicates that, on average, most precipitation tends to occur during the winter (October through March) at station 4JA, which is the closest station to the potential repository site with more than 30 years of complete record.

Much of the annual variability observed in precipitation records in the Yucca Mountain region can be attributed to the effects of global circulation patterns on the position of the jet stream during the winter. In particular, the El Niño Southern Oscillation is a perturbation in the global circulation pattern that is strongly related to sea surface temperatures in the Pacific Ocean and tends to steer the jet stream south of its average winter position in the Western United States, bringing a higher frequency of storms into the Great Basin (Hevesi, Ambos et al. 1996). The wettest 5-year sliding means for January through March totals recorded at station 4JA correspond to periods in the record during which the strongest El Niño Southern Oscillation events were documented. Although the El Niño Southern Oscillation is known to significantly impact precipitation patterns across the southwestern United States, this annual-to-decade-scalevariability in climate is still not completely understood, and the consistency (or lack of consistency) in its occurrence is still being evaluated. A detailed discussion of the El Niño Southern Oscillation and its effect on synoptic scale weather patterns in the Yucca Mountain region, along with a thorough treatment of global circulation patterns and climatic variability, is provided in Section 4.

In addition to precipitation, net infiltration is dependent on incoming solar radiation, relative humidity, air temperature, and to a lesser degree atmospheric turbidity, ozone, and wind. Along with site-specific parameters such as slope, aspect, and albedo, these climate variables control the energy balance which determines potential evapotranspiration. Air temperature also determines the occurrence of precipitation as snow, affects the ability of vegetation to transpire available soil

moisture, and can affect vapor flow in the shallow subsurface. Barometric pressure also may influence vapor flow in the upper part of the unsaturated zone (Flint, A.L. et al. 1996).

5.3.4.1.2.3 Physiographic and Topographic Factors

Although topography has an important impact on the spatial distribution of precipitation and potential evapotranspiration, topography also channels surface flow when runoff occurs and causes the concentration of large volumes of water in channels during short periods. A complete discussion of surface-water hydrology is provided in Subsection 5.1, along with descriptions of drainages and channel networks for the area of Yucca Mountain. In addition to topography, other important site-specific physiographic variables include the water storage capacity of the soil, the permeability of the soil and underlying bedrock, and vegetation (Flint, A.L. et al. 1996). Soil storage capacity is determined mostly by soil thickness and porosity. Soil thickness tends to be well correlated to local topography, with the upland areas generally having thin soils and the lower washes and alluvial fans having thicker soils (surficial materials). Soil porosity and permeability, along with various other hydrologic properties such as residual water content and field capacity, can be correlated to soil type. Bedrock permeability can also be correlated to lithology. A complete description of site geology, including soils (Quaternary geology) and geomorphology is provided in Section 3. A brief summary of site characteristics, with an emphasis on characteristics affecting net infiltration, is provided below.

The topography at Yucca Mountain consists of generalized topographic positions that represent infiltration zones: ridgetops, sideslopes, terraces, and channels (Flint, L.E. and Flint 1995). The ridgetops generally are flat to gently sloping and are higher in elevation than the other topographic positions. They have thin (less than a meter) to no surficial deposits (soils) but are relatively stable morphologically. Soils are fairly well developed, commonly containing thin calcium-carbonate layers. The underlying bedrock on ridgetops is moderately to densely welded (Flint, L.E. et al. 1996b) and moderately to highly fractured. The higher elevations of the ridgetops and upper slopes (above 4,500 feet) have a greater potential of sustaining a thin snow cover for up to several weeks during the winter. Sideslopes are distinguished from the terraces and channels by depth of soils and slope. Soil cover on sideslopes is thin to nonexistent, and in most locations, bedrock is densely welded and highly fractured. The sideslopes are approximately north or south facing in the southern part of the site and, therefore, have different seasonal solar radiation loads. In the northern washes, although the sideslopes face more southwestward and northeastward, the steepness of the slopes accentuates seasonal radiation differences. Terraces and channels are located at lower elevations of primary washes and have thin soil cover in the upper washes and thick soils farther down. Very little bedrock is exposed in the washes. The soil in washes has varying degrees of calcium-carbonate cementation that commonly is quite extensive. The porosity of the soil ranges from 15 to 50 percent with an average of about 30 percent. The surface of terraces and washes is relatively flat and dissected by old soil channels and active channels. Channels differ from terraces in that the periodic runoff that occurs in the channels in response to extreme precipitation conditions can rework the channel materials. Although channels occupy a very small surface area of the wash, they may contribute significantly to net infiltration during runoff events. Over the area of the site-scale unsaturated-zone flow model (Bodvarsson and Bandurraga 1996), ridgetops (including the

shoulders) encompasses about 7 percent of the model area, sideslopes (including footslopes) about 47 percent, terraces about 44 percent, and active channels about 2 percent (Flint, A.L. et al. 1996).

5.3.4.1.2.4 Soils

The dominant soil deposits in the general vicinity of Yucca Mountain are fluvial sediments and fluvial debris-flow deposits that are found in the basins, washes, and alluvial fans (Flint, A.L. et al. 1996). These deposits have varying degrees of soil development and thickness and have a gravelly texture with rock fragments constituting between 20 and 80 percent of the total volume. The soils range from 100 m thick in the valleys to less than 30 m thick in the mouths of the washes. Halfway up the washes, soil fill generally is less than 15 m deep in the center of the channels and well developed, cemented calcium-carbonate layers are common. More stable surfaces, generally on the ridgetops, have developed soils 0.5 to 2 m thick with high clay contents. Overall, soil thickness was classified into four categories: 0 to <0.5, 0.5 to <3.0, 3.0 to <6.0, and ≥6.0, which encompass 48 percent, 7 percent, 5 percent, and 40 percent of the site area, respectively. The spatial distribution of soil thickness is determined primarily by topography: the deeper soils are present on the shallower slopes of the ridgetops, washes, and alluvial fans, whereas the soils on the steeper sideslopes are thin to absent. Exceptions to this are the relatively deeper soils (0.5 to 3.0 m) which occur locally as slumps and debris slides at the base of sideslopes.

Mapped surficial, unconsolidated geologic units were reconfigured into eight surficial soil units based on classical soil-taxonomy descriptions and distinctly different hydrologic properties (Flint, A.L. et al. 1996). Two additional surficial units were added to the eight soil units to account for bare rock and disturbed ground, consistent with the geologic mapping. Physical and hydrologic properties of the upper 0.3 m of the surficial soil units were determined by field measurements or by modeling of soil properties. These properties included porosity, percent rock fragments, bulk density, saturated hydraulic conductivity, water content at −0.1 bar and −60 bars water potential, and the moisture-retention curve-fit parameters *a* and *n*. Porosity of the soil units ranged from 0.21 to 0.37 and saturated hydraulic conductivity ranged from 5.6×10^{-6} m/s to 3.8×10^{-5} m/s (Flint, A.L. et al. 1996). The distribution and properties of the surficial soil units are described in detail in A.L. Flint et al. (1996).

5.3.4.1.2.5 Bedrock Hydrologic Properties

Matrix hydrologic properties of the bedrock volcanic units in the unsaturated zone at Yucca Mountain have been determined for the rocks of the Tiva Canyon, Yucca Mountain, Pah Canyon, and Topopah Spring Tuffs, the interlayered bedded tuffs, the Calico Hills Formation, and the Prow Pass Tuff (see Subsection 5.3.3.1.1). These properties include porosity, bulk density, particle density, saturated hydraulic conductivity (K_s), and moisture retention characteristics. Although the matrix properties of the bedrock units have been characterized in great detail, the hydrologic properties of the fracture network are not nearly as well defined. The density and properties of fractures are critical to understanding and modeling net infiltration because of the potential for significant infiltration to occur as fracture flow in the shallow subsurface under certain moisture conditions (Flint, A.L. et al. 1996). Fractures are prevalent in the welded rocks and vary significantly in density in the nonwelded rocks. Fracture densities and apertures are not well

characterized for all lithostratigraphic units, but estimates have been made from borehole core logs. Properties of fractures are dependent on fracture aperture and whether the fractures are open or filled with calcium carbonate or siliceous materials. Estimates of porosity and K_s for fractures can be made using assumed or estimated density and aperture. Under the assumption that significant net infiltration in fractures occurs only under saturated or near-saturated conditions, unsaturated flow parameters for fractures are not required to model infiltration by way of fracture flow. Average K_s values of fractures were estimated based on data from fluid-flow experiments conducted in the laboratory on a block of fractured tuff (E.M. Kwicklis et al., *Numerical Simulation of Water- and Air-Flow Experiments in a Block of Variably Saturated, Fractured Tuff*, USGS-WRIR-97-4274, Water-Resources Investigations Report, U.S. Geological Survey, in press) for a range of apertures as follows: for 2.5 μm aperture–410 mm/day; for 25 μm–15,700 mm/day, and for 250 μm–745,000 mm/day. K_s values of fracture-fill materials also were measured and averaged 43.2 mm/day. Estimates of fracture K_s were calculated for each of the 73 lithostratigraphic units identified by Scott and Bonk (1984) by assigning the appropriate conductivity values for fracture aperture and integrating over the percent of area per square meter of rock, given the fracture density and aperture available for water flow. These values of fracture conductivity then were added to the K_s values of the rock matrix (from Table 5.3-14), yielding combined matrix-fracture effective saturated hydraulic conductivity values for each lithostratigraphic unit for each of six fracture types: open or filled for each of the three aperture classes. A weighted average effective conductivity also was calculated for each lithostratigraphic unit based on the proportions of fracture apertures present. The effective conductivity values for the six fracture-aperture classes and the weighted average for each of the 73 lithostratigraphic units are tabulated in Table 2 of A.L. Flint et al. (1996). For the numerical infiltration model, only the effective conductivity values for filled 250-μm aperture fractures were used (Figure 5.3-92) because this resulted in the best overall match between numerical model results and field observations for the entire Yucca Mountain Site area, as is described in Subsections 5.3.4.1.4 and 5.3.4.1.5. The importance of the effective conductivity of the bedrock in the calculation of infiltration flux at the soil-bedrock interface is discussed further in Subsections 5.3.4.1.2.7 and 5.3.4.1.5.

5.3.4.1.2.6 Vegetation

Yucca Mountain is located within a broad ecological transition zone, referred to as Transition Desert, between the northern boundary of the Mojave Desert and the southern boundary of the Great Basin Desert. The northern boundary of the Mojave Desert is usually delimited by the northern limits of Larrea Tridentata (Beatley 1976). On Yucca Mountain, Mojave Desert vegetation occurs at lower elevations on bajadas (broad, continuous alluvial slopes) and in washes (Flint, A.L. et al. 1996). Transition Desert associations are situated topographically above the larrea ecotone. Over broad areas, coleogyne ramosissima is the species most identified with the Transition Desert (Beatley 1976). These associations are located primarily on ridges and canyons. Associations on Yucca Mountain are very heterogeneous and are composed of a large diversity of shrub associations.

Vegetation characterization research has been conducted on the Nevada Test Site and at Yucca Mountain periodically for the past 30 years (Flint, A.L. et al. 1996), most recently by EG&G Energy Measurements for the Environmental Assessment for the YMP. Three dominant vegetation associations occur in the EG&G Energy Measurements study area, which extends from Solitario

Canyon in the west to Fortymile Canyon in the east and from Yucca Wash to the north to Busted Butte in the south (EG&G Energy Measurements 1991). The most common vegetation associations are Larrea-Lycium-Grayia, which covers approximately 35 percent of the surface area at Yucca Mountain, Coleogyne (30 percent), and Lycium-Grayia (26 percent). Larrea-Lycium-Grayia predominates on the eastern bajadas of Central Yucca Mountain, occurring at intermediate elevations in the study area ranging from 1,000 to 1,500 m. Coleogyne occurs across the northern third of the study area from the valley floors at elevations of approximately 1,030 m to the flat ridge tops at roughly 1,710 m. However, Coleogyne generally does not occur on the steep slopes connecting these two areas. The dominant species is Blackbrush (Coleogyne ramosissima), and because this species tends to exclude other species, this association is the least diverse. Lycium-Grayia occurs on steep slopes throughout the study area and is dominant on the ridgetop of Yucca Mountain south of Antler Ridge. Lycium-Grayia is the most complex and diverse of the associations, with several species (Mormon teas, Ephedra spp. and buckwheats, Eriogonum spp.) forming subassociations and dominating in local areas. An additional association, Larrea-Ambrosia, although covering only 9 percent of the area, occurs in the south and southeastern sections of the study area at elevations between 900 and 1,050 m. The dominant plant species in the study area is Bursage (Ambrosia dumosa), accounting for almost half of the total vegetation cover. The percent cover of these species has changed from year to year, sometimes dramatically, notably following the drought years of 1987 to 1990. More detailed information on the characteristics of these plants and their distributions at Yucca Mountain is contained in A.L. Flint et al. (1996).

Additional information from Hansen et al. (CRWMS M&O 1998) has provided insight into details of plant habits and distributions that support estimates of evapotranspiration based on energy balance calculations and modeling of radiation loads over the surface of Yucca Mountain. North- and south-facing slopes receive different amounts of solar radiation, resulting in differences in temperature and water availability in the habitats located on these aspects. South-facing slopes tend to be warmer and drier while north-facing slopes tend to be cooler and wetter. Plant species exhibit differential preference or tolerance for these growing conditions. Preliminary data (CRWMS M&O 1998) indicates that some species, often typical of the flora of the Great Basin area, such as yellow rabbitbrush, green ephedra, big sagebrush, and burrobrush are commonly found on north-facing slopes, while bursage and range rhatany, often typical of the flora of the warmer Mojave Desert, are commonly found on south-facing slopes.

Distribution, minimum xylem water potential, and rooting depths were the vegetation characteristics most important to development of the evapotranspiration module of the numerical infiltration model (Flint, A.L. et al. 1996). Minimum xylem water potential was used to calculate the −60 bar lower limit of plant-available water, which amounts to residual water content in terms of evapotranspiration and is equivalent to the wilting point of vegetation. Rooting depths were used to estimate extraction depths for transpiration.

5.3.4.1.2.7 Summary of the Conceptual Model of Infiltration and the Episodic Occurrence of Fracture Flow

For the conceptual model, net infiltration at Yucca Mountain is defined as an episodic, transient process which depends primarily on the length of time saturated or near-saturated conditions are maintained at the soil-bedrock interface and the effective conductivity of the underlying bedrock (Flint, A.L. et al. 1996). Saturated soils are primarily established during the winter in response to a series of large storms that tend to occur more frequently during years when the El Niño Southern Oscillation is active. The timing, intensity, and duration of precipitation, the storage capacity of the soil, and evapotranspiration determine the availability of water for net infiltration. The lower the effective conductivity of the bedrock, the greater the portion of water in the soil available for evapotranspiration. During winter when potential evapotranspiration is at a minimum, saturated conditions at the soil-bedrock interface can develop and be maintained in response to smaller amounts of precipitation than during the warmer months of the year. When the storage capacity of the soil and the effective conductivity of the underlying bedrock is exceeded, or when precipitation intensity exceeds the infiltration capacity of the soil, runoff is generated and water is available for routing down slopes and into channels. The significance of net infiltration beneath channels in washes relative to sideslopes and ridgetops depends on the frequency and magnitude of runoff events.

Differences in the response of the bedrock to the degree of saturation and available moisture in the overlying soil is illustrated in Figure 5.3-93. The graph indicates the measured change through time of the water content of the soil layer above the soil-bedrock interface in neutron access borehole N52 at Yucca Mountain from 1984 through 1994 (Flint, A.L. et al. 1996). The air-entry water potentials needed to initiate and maintain fracture flow for both an open and filled fracture are plotted along with the water potential of the soil calculated using the known moisture characteristic of the soil. The increases in water content of the soil profile correspond to the wetter-than-average winters of 1988, 1992, and 1993. Net infiltration for both the filled and open fractures occurs only during 1992 and 1993 when the water potential exceeds about −0.1 MPa (−1 bar), although the filled fracture comes close to initiating infiltration during 1988. During most of the record, the soil above the bedrock interface remains too dry for the infiltration of water into bedrock.

5.3.4.1.3 Results of Infiltration Monitoring and Water Balance Studies

To test and attempt to validate various aspects of the conceptual model of net infiltration, a series of field monitoring activities were conducted at Yucca Mountain over a period of several years (Flint, A.L. et al. 1996). These field activities included:

- Periodic measurement of water-content profiles in neutron-access boreholes
- Continuous monitoring of water potential above and at the soil-bedrock interface
- Geochemical analysis of soil and rock samples
- Monitoring of precipitation and runoff

5.3.4.1.3.1 Measurement of Water-Content Profiles in Neutron-Access Boreholes

Volumetric water-content profiles were measured at least monthly in 99 neutron-access boreholes (N series) ranging in depth from 4.6 to 78.0 m for various periods of time from 1984 to 1995 (Flint, A.L. et al. 1996). The relative changes in water content through time were used to study infiltration and moisture redistribution in surficial materials at Yucca Mountain. A quantitative analysis of the water-content profiles in the time domain was used to develop the conceptual model of infiltration (Flint, A.L. and Flint 1994; Flint, L.E. and Flint 1995). Although water-content profiles are not a direct measure of infiltration flux, relative changes in profiles were used to estimate flux given assumptions concerning the hydraulic gradient and the depth of evapotranspiration (Hudson and Flint 1996). The water-content profiles also were used to calibrate numerical models of infiltration (Hevesi, Flint et al. 1994; Flint, A.L. et al. 1996).

Redistribution of subsurface moisture in relatively thick alluvium (about 11 m) following channel runoff in Pagany Wash during the summer of 1984 is illustrated by the record of water-content changes with time for borehole N1 (Figure 5.3-94). The initially high water contents in the soil profile (far left side of figure) are attributed to deep infiltration that occurred in the channel in response to the wetter than normal conditions during 1983 and a runoff event in July 1984 (see Subsection 5.1) prior to drilling of the borehole (Flint, A.L. et al. 1996). The record suggests that about 6 years was required for the profile to return to equilibrium following the deep infiltration pulse. Runoff at this location in Pagany Wash did not occur again until the winter of 1995. Following the wettest January ever recorded at Yucca Mountain (see Subsection 3.4.1), heavy precipitation during March 10 through 11, 1995 exceeded the combined storage capacity of the soil and the infiltration capacity of the underlying bedrock, causing runoff to occur in most of the drainages on Yucca Mountain. The hydrographs developed from the stream flow records obtained from stations in Pagany and Wren Washes are shown and discussed in Subsection 5.1. The 1995 runoff event in Pagany Wash had a duration of approximately 10 hours and resulted in infiltration along the channel which returned the profile to the 1984 conditions, and resulted in the occurrence of net infiltration as shown by the increase in moisture contents across the alluvium bedrock interface at a depth of 11 m (far right side of Figure 5.3-94).

Infiltration pulses caused by runoff during 1993 and 1995 and subsequent moisture redistribution in bedrock directly under a channel with this alluvium (0.7 m deep) are illustrated by the record of water-content changes with time for borehole N15 (Figure 5.3-95), located on Bleachbone Ridge in the head-water part of Pagany Wash (Flint, A.L. et al. 1996). The depth versus time profile shows a wetting front moving rapidly to a depth of 7.5 m in 1993 and 10 m in 1995, and then advancing more slowly to a maximum infiltration depth of over 15 m. Most of the increases in water content depicted by the depth-time profile are assumed to be representative of net infiltration volumes because of the depth of penetration into the bedrock. The total increase in water in storage below 3 m was equivalent to 500 mm of water in 1995. If roots fail to penetrate to greater than 3 m, then this water would be considered net infiltration. Two nearby boreholes in the same geologic setting (N36 and N17), indicated similar patterns of redistribution but with a smaller volume of infiltrating water because no surface runoff occurred at these sites. The total increases in water in storage below 3 m were equivalent to 150 mm and 110 mm, respectively, in 1995. These values are considerably lower, but likely more typical of the soil thickness and underlying rock type for the general

topographic area (away from the small channel). Because these estimates of net infiltration are based on volumetric water content using a neutron probe, they represent water that has imbibed into the rock matrix. Additional water may have penetrated to greater depths through fractures but is not indicated in the neutron-logging water-content profile.

Establishment saturated or near-saturated conditions at the soil-bedrock interface at a depth of approximately 1.9 m on a terrace is illustrated by the depth-versus-time water-content profile for borehole N63 (Figure 5.3-96), located close to the edge of Pagany Wash (Flint, A.L. et al. 1996). The shallow soil profile is wetted following winter storm events in 1993 and 1995, 2 years during which the El Niño Southern Oscillation was active. In 1993, the soil-bedrock interface may not have been sufficiently saturated to induce infiltration into bedrock because there is no detectable increase in water content beneath the interface. Therefore, it is assumed that the moisture increase in the profile during 1993 was removed by evapotranspiration. In 1995, the soil above the interface was wetter, and a well defined wetting front penetrated into the bedrock, but not so deep as to be unaffected by evapotranspiration. At this location, evapotranspiration and net infiltration both compete for the available moisture in the soil profile because the soil is relatively thin and the bedrock is not very fractured. Although some net infiltration occurred dominantly as matrix flow, most of the moisture in the soil profile was removed by evapotranspiration above the interface because the effective conductivity of the bedrock is relatively low.

The analysis of moisture profiles in the 99 neutron-access boreholes distributed over a large area of Yucca Mountain, and representing 11 years of climatic record resulted in both qualitative and quantitative assessments of net infiltration within the four topographically based zones (Flint, L.E. and Flint 1995; Flint, A.L. et al. 1996). Shallow infiltration processes within these four zones were categorized generally on the basis of the manner in which volumetric water content changed with depth and time. The ridgetops are flat to gently sloping, higher in elevation, have thin soils mostly developed in places with clays, and a higher water-holding capacity that reduces rapid evaporation and drainage. The ridgetops generally are located where the bedrock is moderately to densely welded and, therefore, fractured. These conditions lead to deeper penetration of infiltration pulses than in the other topographic positions, but relatively smaller volumes of water. In some locations, however, where runoff is channeled high in small watershed, large volumes of water can infiltrate. The sideslopes are steep, commonly have very shallow to no soil cover, and are usually developed in welded, fractured tuff, resulting in rapid runoff. The low-storage capacity at the surface of the sideslopes and the exposure of fractures at the surface can allow small volumes of water to infiltrate to greater depths, especially on slopes with north-facing exposures, resulting in relatively lower evapotranspiration rates. Shallow soil at the bases of the sideslopes can become saturated and initiate flow into the underlying fractures. Soil terraces are flat, broad deposits of layered rock fragments and fine soil with a large storage capacity, resulting in little runoff and only shallow penetration of infiltrating water, which subsequently is almost entirely removed from the soil profile by evapotranspiration. Consequently, terraces contribute the least to net infiltration in the watershed. Active channels differ little from the terraces except that they collect and concentrate runoff which, although it occurs infrequently, can then penetrate deeply. However, this mechanism is not considered to be a major contributor to areally distributed net infiltration because of the infrequency of precipitation resulting in runoff and because the channels encompass only about 2 percent of the

Yucca Mountain Site area. Nevertheless, significant, localized pulses of net infiltration may occur in active channels.

Overall, analyses of changes in water-content profiles indicated deeper penetration and higher net infiltration rates at sites with thin soil cover (Flint, A.L. et al. 1996), which tended to be on ridgetops and sideslopes, but also included some channels in the upper part of watersheds (Figure 5.3-97). The deepest penetration of wetting fronts and the highest rates of net infiltration (greater than 60 mm/year) were observed at sites where surface flow was concentrated or channeled toward locations underlain by fractured bedrock. Such sites include the headwater zones of washes and the lower sections of sideslopes. Furthermore, more precipitation infiltrates into the ridgetops during the winter because lower intensity storms result in little or no runoff and slowly melting snow results in slow but steady infiltration over a period of several weeks each winter. This slow rate of input over long time periods allows for larger volumes of water to penetrate below the root zone and thereby escape the high evapotranspiration rates of the following summer. When surface flow is negligible, the deepest infiltration was observed on the ridgetops and the shallowest infiltration was observed in the washes, although there was little difference in infiltration characteristics between terraces and channels. Exceptions were noted, however, following significant runoff events when relatively large volumes of water often infiltrated more than 5 to 6 m into the alluvium under washes, which is below the estimated root zone. However, conditions causing significant channel runoff occurred infrequently and only in a few washes during any single runoff event.

5.3.4.1.3.2 Continuous Monitoring of Water Potential Using Heat-Dissipation Probes

Heat-dissipation probes were used to obtain continuous water-potential measurements for one year at several depths at a location on Bleach Bone Ridge near a cluster of three neutron-access boreholes, N15, N16, and N17 (Flint, A.L. et al. 1996). Water-potential measurements were made at depths of 7, 15, 36.5 and 73.7 cm, with the deepest measurement being at the soil-bedrock interface (Figure 5.3-98). Within 2 weeks of installation (February 1995), heavy winter precipitation saturated the soil-bedrock interface to within 36 cm of the ground surface. The probe at the soil-bedrock interface remained saturated until the end of March and then dried out to less than -10 bars by September. The probes closer to the ground surface dried out faster with the near-surface probes wetting up periodically due to summer precipitation events. The absolute value of water potential beyond −100 bars is questionable but provided a relative indication of drying conditions. The water-retention curve for the soil at this location was used to convert water potential to water content, which also is shown in Figure 5.3-98. A series of selected data were chosen and the rate of water loss was calculated between the dates using the change in water content. In early March, the profile was changed at a rate of over 10 mm/day but dropped to less than 2 mm/day within 30 days. The evaporation rate was estimated to be no more than 2 mm/day based on potential evapotranspiration calculations using the Priestley-Taylor equation, yielding a maximum infiltration flux into the bedrock of 8 mm/day. The average flux for 30 days was 5 mm/day, equaling a total flux into the bedrock of 150 mm. The estimate of net infiltration from analysis of the volumetric water-content profile for the nearby borehole N17 was 110 mm for the same time period (Hudson and Flint 1996). The combined effective saturated hydraulic conductivity (rock matrix and filled 250 μm fractures) estimated for this location was 0.322 mm/day (Tiva Canyon Tuff caprock, Figure 5.3-92), which is much less than the 8 mm/day indicated by the combined heat-dissipation probe/evapotranspiration

estimate. Fractures in a nearby exposure were found to be mostly unfilled and thus the assumption of filled 250 μm fractures to calculate effective conductivities was not consistent with actual field conditions at this site. Overall, however, the assumption of filled 250 μm fractures to calculate effective conductivities of the bedrock resulted in the best match between numerical-model results and field observations for the entire Yucca Mountain Site area, as described in Subsections 5.3.4.1.4 and 5.3.4.1.5.

5.3.4.1.3.3 Precipitation and Runoff Monitoring

Development of the numerical infiltration model was highly dependent on successful measurement and simulation of the spatial distribution of precipitation in the vicinity of Yucca Mountain (Flint, A.L. et al. 1996). All available precipitation data were analyzed to determine relationships between average annual precipitation and both elevation and geographic location within the infiltration study area (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). A summary of precipitation data collected at weather station sites in support of water balance and infiltration studies at Yucca Mountain for 1988 to 1994 is provided by A.L. Flint and Davies (1997). Additional summaries of measured precipitation at the site are provided by Ambos et al. (1995), Hevesi, Ambos et al. (1994), and CRWMS M&O (1997a). Values of total annual precipitation for calendar year 1995 at nine precipitation monitoring sites over the area of the potential repository site range from 140 mm to 343 mm (CRWMS M&O 1997a).

A comparison of average annual precipitation calculated at monitoring stations with estimates of average annual precipitation as a function of elevation using both geostatistical (co-kriging) and regression models (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press) indicated a good overall match between measured and modeled values (Figure 5.3-99). The apparent station-to-station variability observed in average annual precipitation can be attributed to a variety of topographic effects. These include localized terrain effects on wind, which can decrease the amount of precipitation reaching the ground surface, and precipitation-gauge catch deficits that tend to increase with increasing wind speed.

A geostatistical analysis of the spatial variability of precipitation amount indicated much higher spatial variability over the area of the site area for summer storms than for winter storms. Precipitation amounts resulting from summer storms were observed to range from more than 25 mm over one drainage on the site to approximately 0 mm for another nearby drainage (Hevesi, Ambos et al. 1994). Precipitation rates measured for a storm on August 31, 1991, having a total duration of approximately 30 minutes, were as high as 175 mm/hour (Figure 5.3-100). Such high precipitation rates can exceed the infiltration capacity of the soils, and if maintained for a long enough duration, can result in significant runoff. In contrast, maximum precipitation rates measured for the March 10-11 storm of 1995, which resulted in the occurrence of runoff and channel flow throughout the study area, were only as high as 14 mm/hour. However, wetter than normal antecedent soil conditions caused by frequent storms in January 1995 enabled the 2-day precipitation event in March to exceed the storage capacity of the soil and also the effective bedrock infiltration capacity at most locations, causing widespread runoff into channels. A summary of calculated flow volumes and descriptions of the runoff hydrographs measured in Pagany and Wren Washes are provided in Subsection 5.1.

5.3.4.1.4 Development of the Numerical Net-Infiltration Model

The numerical model of infiltration was developed in accordance with the conceptual model, which qualitatively describes the processes of precipitation, runoff, evapotranspiration, and redistribution of water within the shallow-unsaturated zone (Flint, A.L. et al. 1996). The conceptual model predicts that the most important factor controlling net infiltration is the initiation of fracture flow at the soil-bedrock interface. In most cases, net infiltration is not limited by the effective conductivity of the bedrock, but rather by the availability of water percolating beneath the zone of evapotranspiration, which is mostly dependent on precipitation and soil thickness. According to the conceptual model, the properties and processes most important in determining if net infiltration occurs are the timing and amount of precipitation, the storage capacity of the soil (which includes soil depth), the seasonality and amount of evapotranspiration, and the hydrologic properties of the underlying bedrock. The numerical infiltration model was developed to correctly account for these properties and processes.

The numerical model of net infiltration (Flint, A.L. et al. 1996) was designed to produce an estimate of net infiltration at any location within the infiltration study area (Figure 5.3-86), which includes the area of the 3-D site-scale unsaturated-zone flow model (Bodvarsson and Bandurraga 1996). The net infiltration model requires two external data sets specific to the location or subarea being modeled. One data set contains physiographic and hydrologic information (physical data set) for the site and the other data set contains estimates or measurements of daily precipitation (precipitation data set). The physical data set requires location identification, northing, easting, slope, aspect, elevation, latitude, longitude, soil type, soil depth, underlying geologic formation, geomorphic position, and the location of the surrounding topography that blocks the site for diffuse or direct-beam solar radiation. Using this information and a daily precipitation record as input, the model calculates daily values of net infiltration using a water-balance approach, first to calculate the amount of water available for net infiltration, and then to calculate the actual net infiltration using the effective conductivity of the bedrock. Thus, the model uses the water balance to establish a potential for net infiltration, but actual net infiltration is ultimately limited by the effective conductivity of the bedrock. In this formulation, runoff is calculated as the solution to the water balance (Flint, A.L. et al. 1996).

The physical data set input file for the infiltration model is defined using the 30 m by 30 m U.S. Geological Survey 7.5-minute digital elevation models available for the Yucca Mountain area. The regular grid defined by the 30 m cells also defines the grid for the infiltration model. The model grid consists of a total of 253,597 cells in the rectangular region bounded by 547751 to 548921UTM (zone 22) easting and 4077783 to 4078563 UTM northing. Model parameters were assigned to each cell using geographic information system applications and digital map coverages of bedrock geology, soil type, and soil depth. The geology was based on the Scott and Bonk geologic map (1984), which was the only geographic information system coverage available for site geology at the time of model development (Flint, A.L. et al. 1996). The channel network for the model was defined using an available geographic information system coverage of channel nodes that were combined with the model grid cells such that cells containing channel nodes were defined as channel grid cells.

5.3.4.1.4.1 Daily Precipitation Input

A record of daily precipitation for 1980 through 1995 for the Yucca Mountain Site area, corresponding to the record of water content in the 99 neutron boreholes, was needed for calibrating the numerical infiltration model (Flint, A.L. et al. 1996). Site-specific data that were available for various locations on Yucca Mountain were combined with the regional data to develop a record of estimated total daily precipitation amounts from January 1, 1980 through September 30, 1995. The daily totals were adjusted for elevation to provide a spatial distribution of precipitation that is consistent with the correlation between average annual precipitation and elevation (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). Table 5.3-37 lists selected statistics for the developed record of daily precipitation for 1980 to 1995 at an elevation of 1,400 m for the area of the potential repository, including the maximum daily precipitation, the number of days of measurable precipitation, and the number of days for which precipitation exceeded 10 mm. Especially noteworthy is the very few number of days for which precipitation exceeds 10 mm, indicating that a very few number of infrequent storms greatly effect total annual precipitation. Figure 5.3-101 shows total annual precipitation for the same developed record for the area of the potential repository. For 1980 through 1987, the developed precipitation record is somewhat uncertain because site-specific data were not available for this period and the daily and annual totals were interpolated using records from Nevada Test Site stations and nearby regional stations. Beginning in 1988, however, site-specific data from two weather stations at Yucca Mountain (Flint, A.L. and Davies 1997, stations 1 and 3) were included in the interpolation model, greatly reducing the uncertainty in the developed record.

Records from the Nevada Test Site stations and the regional network were used also to analyze the temporal distribution of daily precipitation and to define the parameters of a stochastic model of daily precipitation that was used for 100-year simulations of infiltration (Flint, A.L. et al. 1996). The stochastic precipitation model consists of a pseudo-random number generator that provides a normalized uniform deviate for a process to simulate daily precipitation (Flint, A.L. et al. 1996). The 32 transition probabilities for a third-order, two-state Markov chain model of the occurrence of measurable daily precipitation and the cumulative-probability distribution for daily precipitation amounts were estimated for each month using the daily precipitation records available at four selected regional precipitation stations; 4JA, Nevada Test Site Area 12, Lake Valley, and South Lake (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). Station 4JA, with 133 mm average annual precipitation, is the closest station to Yucca Mountain having a daily precipitation record of 30 years or more and was selected to provide a 100-year representative climate simulation for Yucca Mountain by scaling the simulated daily totals to Yucca Mountain. The other three stations were selected as analog sites for potential wetter climates, with Nevada Test Site Area 12 receiving 320 mm average annual precipitation, Lake Valley receiving 360 mm average annual precipitation, and South Lake receiving approximately 440 mm. Precipitation-intensity data measured at the site were used for defining approximate average storm durations of 2 hours for summer storms and 12 hours for winter storms as input to the infiltration model (Flint, A.L. et al. 1996).

Daily precipitation amounts input to the infiltration model are spatially distributed over the modeling area as a function of grid-cell elevation using an empirical function calibrated to the geostatistical estimates of average annual precipitation obtained by Hevesi and Flint (Hevesi, J.A. and Flint, A.L.,

USGS-WRIR-96-4123, in press). This method of spatially distributing daily precipitation is considered to be a reasonable representation of winter precipitation but is not considered to be an accurate representation of summer thunderstorms. Using the empirical spatial scaling function and the developed daily precipitation record for 1993, spatially distributed total annual precipitation ranges from a minimum of 175 mm to a maximum of 350 mm over the area of the modeling grid (Figure 5.3-102).

5.3.4.1.4.2 Field Moisture Capacity

The infiltration component of the model uses a daily mass-balance approach driven by the input of precipitation (Flint, A.L. et al. 1996). All precipitation is assumed to infiltrate into the soil provided the available storage capacity of the soil is greater than the precipitation input. Precipitation input in excess of the available soil-storage capacity is calculated as a daily runoff term for each grid cell, and is available for surface-flow routing. The available storage capacity of each cell is calculated as the difference between soil porosity and soil-water content. The maximum available storage capacity for a given soil type is the soil porosity minus the residual water content. All water in excess of the residual water content is available for evapotranspiration. Actual evapotranspiration is calculated for each cell as a function of soil-water content and potential evapotranspiration. Water in excess of the field capacity for a given soil type is available for both evapotranspiration and net infiltration. Net infiltration occurs only when the field capacity is exceeded, and is calculated as a daily rate equal to the effective saturated-hydraulic conductivity of the bedrock type assigned to each cell (Subsection 5.3.4.1.2.5). If the daily effective saturated-hydraulic conductivity exceeds the soil-water content minus the field capacity, then the soil-water content minus the field capacity is calculated as net infiltration, and the new soil-water content is set equal to the field capacity minus evapotranspiration. In this model, all drainage from the soil profile into the bedrock is defined as net infiltration. The soil infiltration storage capacity above which any additional water causes net infiltration to occur is the field capacity minus the water content multiplied by the soil depth. The spatial distribution of the soil-infiltration storage capacity (Figure 5.3-103) is dominated by the spatial distribution of soil depth, which is determined primarily by physiography.

5.3.4.1.4.3 Evapotranspiration

Evapotranspiration rates calculated by the infiltration model are based on available soil moisture and on modeled potential evapotranspiration, which is estimated using modeled solar radiation, air temperature, and ground heat flux (Flint, A.L. et al. 1996). The radiation load is calculated based on slope, aspect, elevation, latitude, longitude, and surrounding topography (blocking ridges). Blocking ridges are defined by the angle above a horizontal surface intersecting the ridges that block the sky from direct-beam or diffuse radiation. The atmospheric inputs are ozone, precipitable water, atmospheric turbidity, circumsolar-diffuse radiation, and ground albedo. These values are calculated as monthly averages for the entire model grid. The position of the sun is calculated for each grid cell every hour starting at sunrise on each day. Direct-beam and diffuse sky radiation are then calculated based on the atmospheric input parameters and applied to the surface based on the slope and aspect of each cell. Diffuse sky radiation is reduced by that amount of sky that would be blocked by the surrounding topography. The direct-beam radiation also is blocked when the sun is positioned behind the surrounding topography. Ground-reflected radiation is added to the site based on the area

of the surrounding topography, the ground albedo, and the direct-beam and diffuse sky radiation that reflects from the surrounding topography. Actual evapotranspiration is modeled as a function of potential evapotranspirationand soil-water content using a modified Priestley-Taylorequation (Flint, A.L. et al. 1996). The modified equation uses an empirical function dependent on water content to scale actual evapotranspiration as a variable percentage of potential evapotranspiration. The empirical function must be calibrated or defined using calibrations available from analog sites. This modification of the Priestley-Taylor equation allows for the soil-water content to limit evapotranspiration. Evapotranspiration a very important component of the infiltration model because potential evapotranspiration in arid environments can exceed precipitation on an annual basis by an order of magnitude. Evapotranspiration is solved on an hourly basis and summed for each day, and then the change in water content is updated at the beginning of the next day. The spatial distribution of modeled potential evapotranspirationis dependent mostly on the topographic controls of slope and aspect, and ranges from a minimum of 500 mm/year to a maximum of 900 mm/year for the area of the site (Figure 5.3-104). Minimum values of evapotranspiration occur at locations on steep, northerly and northeasterly facing slopes, particularly at locations surrounded by blocking topography.

5.3.4.1.4.4 Surface Water Runoff

For the initial version of the surface-flow routing model, stream discharge from the model grid is calculated as a fraction of total runoff, then the remaining runoff volume is distributed evenly across grid elements containing channel nodes where it is added to net infiltration occurring in direct response to precipitation (Flint, A.L. et al. 1996). In this application, the water balance is satisfied, but runoff is not modeled as an interactive component of the system, only as an excess precipitation term for each grid cell. This method does not correct for potential input volumes which may exceed the effective conductivity of the underlying bedrock. The determination of the percentage of runoff that becomes outflow is subjective. Using a geographic information system coverage of the stream channel network for Yucca Mountain consisting of 21,832 channel nodes, a total of 14,321 grid cells were identified as containing one or more channel nodes (Figure 5.3-105). Due to the characteristics of the geographic information system coverage, some of the stream channels are represented as discontinuous segments in the model.

In the interactive flow-routing application of the model, runoff is routed laterally from element to element using a simple routing algorithm that uses the grid-cell elevations. The basic, coupled infiltration-flow routing algorithm performs an instantaneous daily routing of runoff. The algorithm does not require prior identification of a channel network and thus avoids difficulties associated with segmented stream channels resulting from the transposition of channel nodes onto a regular grid. The routing module is interactive with the infiltration module and multiple iterations are performed on the entire grid until all runoff generated has either infiltrated or has been routed off the grid as stream discharge on a daily basis. The amount of routed surface flow allowed to infiltrate into the soil is limited by the saturated-hydraulic conductivity of the soil. Using an assumed runoff duration of 2 hours for July through September and 12 hours for October through June, the surface runon rate for each grid cell is compared against the soil saturated-hydraulic conductivity. If the runon rate exceeds the soil conductivity, infiltration is set equal to the soil conductivity up to the soil-storage capacity. When the soil-storage capacity is exceeded, infiltration is limited to the net infiltration rate

as defined by the effective bedrock conductivity. The instantaneous daily routing method allows for a much faster simulation time as compared to routing flow in the time domain and does not require estimates for parameters defining surface resistance and channel geometries. The daily routing method assumes all runoff events have durations less than 24 hours, and can only provide an estimate of total daily flow volumes for each grid cell. A more rigorous application of the flow-routing module that allows real-time simulations of overland flow using the kinematic-wave approximation currently is under development. Real-time simulations of overland flow are used to provide simulated hydrographs that can be compared to hydrographs recorded at stream flow gauging sites, such as the hydrographs obtained at Wren and Pagany Washes during the March 11, 1995 runoff events described in Subsection 5.1. A comparison of simulated and observed hydrographs can be used as a part of overall model calibration, provided the uncertainty of the observed hydrograph is not excessive. The daily flow-routing algorithm still can be applied to model calibration using a comparison of simulated and measured total daily flow volumes, or simply by comparing the simulated and observed occurrences of runoff events.

5.3.4.1.4.5 Model Calibration

Calibration of the numerical infiltration model was achieved in several phases by several different model-calibration techniques (Flint, A.L. et al. 1996), including comparison of model-calculated solar radiation and air temperature with field data. However, the primary model-calibrationtechnique involved comparison evapotranspiration calculated by the Preistley-Taylor equation to water loss from the water-content profiles of the 99 neutron boreholes. The infiltration model was calibrated using the neutron-probe logging data by visually comparing the total water-content change in the soil profile through time against the model simulation results for the same period using the estimated or measured record of daily precipitation for each borehole site. The visually calibrated data set then was used to simulate the parameters defining the modified Priestley-Taylor equation (in particular, the alpha coefficient) using an inverse- modeling technique. An example of model-calibration results for borehole N63 (Figure 5.3-106) indicates that a reasonable match between modeled and measured water content was obtained.

A more rigorous calibration of the numerical infiltration model is in progress using time-continuous soil-water potential and water-content changes measured for water years 1996 and 1997 using heat-dissipation probes at several locations on Yucca Mountain. In conjunction with the heat-dissipation probe data, measurements of evapotranspiration obtained using eddy-correlation stations also will be incorporated into the ongoing calibration process. As described in the previous subsection, efforts also are continuing to apply available streamflow records to a more rigorous model calibration.

Because model results indicated a satisfactory preliminary calibration, the model was used subsequently to produce 100-year simulations of infiltration that allowed for an evaluation of the temporal and spatial distribution of net infiltration throughout the area of Yucca Mountain under both current climatic conditions and possible future climatic conditions.

5.3.4.1.5 Application of the Numerical Infiltration Model

The numerical infiltration model was used to obtain estimates of net infiltration for the entire Yucca Mountain Site area (Area 1) covered by the 30 m grid (Figure 5.3-107), and also for smaller areas (Area 3 and Area 5) centered over the potential repository site. Descriptive information for all three model areas is listed in Table 5.3-38. The smaller areas were analyzed to obtain more detailed information on infiltration and runoff over the area of the potential repository block. Model simulation results consist of total-annual or average-annual net infiltration for each 30 m grid cell and the mean daily infiltration for all grid cells. In addition to simulated infiltration, results also were obtained for simulated evapotranspiration, generated runoff, and surface-water outflow (stream discharge). The model was applied to provide estimates of net infiltration rates at the site for current climate as well as potential future climates. Model results were compared with estimates of net infiltration and recharge obtained using alternative methods discussed at the beginning of this subsection. Model results obtained using the initial method of estimating net infiltration for channels were compared with results obtained using the coupled infiltration, runoff routing version of the model. A summary of model results along with a discussion of model limitations and uncertainty is provided at the end of this subsection.

5.3.4.1.5.1 Estimates of Present-Day Net Infiltration Rates for Yucca Mountain

An initial estimate of average annual net infiltration was obtained using results from a 1-year numerical simulation scaled to the average annual precipitation value from a regression model fitted to the 1980 to 1995 results obtained at neutron borehole locations (Flint, A.L. et al. 1996). For the initial estimate, average annual net infiltration for grid cells containing channel nodes was estimated to be a uniform value of 10 mm/year. The mean average annual net infiltration rate for the site area (Area 1) obtained using the initial estimate was 3.0 mm/year, the maximum rate was about 81.9 mm/year, and the minimum was 0.0 mm/year (Table 5.3-38). Estimates of present-day average annual net infiltration rates for Yucca Mountain also were obtained using a scaled 100-year stochastic simulation of precipitation for station 4JA (Flint, A.L. et al. 1996). Average annual precipitation for the area of the entire grid (Area 1) for the scaled 100-year simulation was 150 mm/year and resulted in the distribution of infiltration rates shown in Figure 5.3-108. Maximum net infiltration rates of 63 mm/year were obtained for relatively high elevations of Yucca Mountain north of the potential repository site that have surface exposures of higher permeability, nonwelded tuffs. Minimum values of 0.0 mm/year net infiltration were obtained for all locations having a soil depth of 6 m or greater. The mean average annual net infiltration rate for the entire site area (Area 1) was 3.2 mm/year (Table 5.3-38). The total net infiltration for all channel cells was estimated to be 90 percent of the generated runoff volume added to net infiltration resulting from precipitation on channel cells. (According to model results, only 10 percent of generated runoff leaves the modeled area as stream discharge because of evapotranspiration and net infiltration in the channels.) For the 100-year stochastic simulation, runoff ranged from 0.0 for much of the site area to about 10 mm/year in some small areas at higher elevations north of the potential repository site (Figure 5.3-109). The mean annual runoff for entire model area obtained for the 100-year simulation was 0.58 mm, and this results in a uniform channel net infiltration rate of 9.2 mm/year added to the average annual net infiltration rate resulting from precipitation onto channel cells. For Area 5 centered over the potential repository site, net infiltration ranged from 0.0 mm/year for locations with thick soils

to over 25 mm/year at a location close to the crest of Yucca Mountain (Figure 5.3-110). The mean average net infiltration rate for all of Area 5 was about 6.0 mm/year (Table 5.3-38).

As indicated in Table 5.3-38, model results obtained using the 100-year scaled stochastic simulation of precipitation for station 4JA are similar to the initial estimates of net infiltration obtained using the scaled 1-year simulation. However, the 100-year simulation is considered to be the more representative of present-day climatic conditions than the scaled 1-year simulation. In addition, the 100-year simulation more accurately represents the seasonality of precipitation, which is an equally important factor as total annual precipitation in determining annual net infiltration than (Flint, A.L. et al. 1996). Simulation results for each 5-year period of the 100-year simulation, calculated as 5-year averages, are listed in Table 5.3-39. The maximum 5-year average infiltration rate is 10.95 mm/year, and the minimum 5-year rate is 0.17 mm/year. To compare results obtained using longer term averages, the sliding 15-year average rates calculated from the 5-year averages also are listed in Table 5.3-39. The maximum 15-year average net infiltration rate is 6.49 mm/year and the minimum rate is 1.31 year. The sliding 15-year averages indicate that considerable variability in estimated net infiltration values can be expected for 5-year and even 15-year simulations.

Estimates of average annual net infiltration rates for the Yucca Mountain Site also were obtained using the developed 1980 to 1995 daily-precipitation record that was used for model calibration (Flint, A.L. et al. 1996). The proportion of generated runoff contributing to channel infiltration was again assumed to be 90 percent, with only 10 percent of the runoff volume generated being discharged from the model area. The results obtained for the 1980 to 1995 simulation indicated higher net infiltration rates and runoff amounts than for the 100-year stochastic simulation. Average annual precipitation for the area of the entire grid (Area 1) was 157.5 mm/year for the 1980 to 1995 simulation (Table 5.3-40), as compared to 150.19 mm/year for the scaled 100-year simulation. For 1995, model results indicated an average net infiltration for the entire site area (Area 1) of 43.58 mm and an average runoff of 23.77 mm. Simulated runoff magnitudes from the 1980 to 1995 simulation allow for a crude measure of model calibration by a comparison with streamflow records for Yucca Mountain, which are discussed in detail in Subsection 5.1. Runoff was generated during the years of 1983, 1984, 1985, 1992, 1993, and 1995, but average runoff amounts greater than 1 mm were generated only during 1992, 1993, and 1995 (Table 5.3-40). This result is generally consistent with records of streamflow at Yucca Mountain, although the amount of runoff simulated for 1983 and 1984 was not as high as expected. The current version of the net infiltration model does not accurately simulate runoff magnitudes resulting from short-duration, high-intensity precipitation events common during the summer, such as the August 1991 and March 1995 storms (Figure 5.3-100). This is because the model does not account for precipitation intensities exceeding the soil-infiltration capacity as determined by the soil permeability and water content (Flint, A.L. et al. 1996). The streamflow records for Yucca Mountain show that much of the runoff that occurred during 1983 and 1984 was the result of severe summer storms. In contrast, runoff generated during 1992, 1993, and 1995 resulted from a loss of soil-storage capacity because of a higher frequency of longer duration winter precipitation, conditions for which the infiltration model provides a more accurate numerical representation. Ongoing refinement of the infiltration model includes a better representation of conditions involving short-duration, high-intensity precipitation events by limiting the infiltration rate to the infiltration capacity of the soil. Such a modification should result in

slightly lower estimates of net infiltration rates for interchannel locations and slightly higher infiltration rates for channel locations.

The average net infiltration rate of 7.64 mm/year obtained using the 1980 to 1995 record (Table 5.3-40) is greater than the maximum 15-year sliding-average net infiltration rate of 6.49 mm/year obtained using the scaled 100-year stochastic simulation for Yucca Mountain (Table 5.3-39). This suggests, but does not prove, that the 1980 to 1995 record represents a wetter than normal period compared to average conditions as represented by the 36-year record of precipitation available at station 4JA (Flint, A.L. et al. 1996). The higher frequency of years affected by an active El Niño Southern Oscillation, particularly the 1991 to 1995 period, which resulted in some of the wettest winters on record for the Southern Great Basin, is not well reproduced by the stochastic simulation. In addition, the stochastic simulation does not account for climatic trends, such as the predicted trend towards wetter conditions caused by an increase in atmospheric carbon dioxide. The onset of such "greenhouse" climates, which may be analogous to an increased frequency in El Niño Southern Oscillation events, might already be causing bias in the 1980 to 1995 climate record relative to longer-term records in the Yucca Mountain region, such as the station 4JA record. A thorough discussion of potential future climates, including the likely onset of a "greenhouse" climate, the eventual onset of a glacial climate, and the results of general circulation models being applied to predict potential future climates, is provided in Subsection 4.1 of this report.

5.3.4.1.5.2 Comparison of Numerical-Model-Derived Net Infiltration Rates With Flux Estimates From a Chloride Mass-Balance Method

Infiltration flux estimates using a chloride mass-balance method for analysis of soil and rock samples are indicated in Table 5.3-41 (Fabryka-Martin, Wightman et al. 1994), along with point and spatially averaged net infiltration estimates derived from the numerical infiltration model (Flint, A.L. et al. 1996). The chloride mass-balance flux estimates vary considerably at the same location depending on the depth of soil or rock samples and the type of material. For example, under a soil terrace at borehole UZ#16, the flux was 0.02 mm/year for the soil samples but 3.0 mm/year and 3.5 mm/year for PTn and CHn samples. Interestingly, the numerical model indicated no infiltration at borehole UZ#16 because of the thick alluvium (12 m) under the terrace. Similarly, under an alluvium-filled channel at borehole N54, the chloride mass-balance flux was 0.02 mm/year for the soil samples but 3.3 mm/year for the PTn samples. These differences between chloride mass-balance fluxes for the soil and the PTn may indicate that significant lateral redistribution of higher fluxes into the nearby sideslopes occurs by the time the water reaches the PTn. In support of this, the chloride mass-balance fluxes for the PTn at borehole N55 (on a south-facing slope) and at borehole N53 (a north-facing slope), which are on opposite sides of the alluvium-filled channel in which N54 is located, were 2.4 mm/year and 3.3 mm/year, respectively. Further, under a soil terrace at borehole UZ-14, the chloride mass-balance flux was 1.2 mm/year for the PTn and 5.9 for the CHn, supporting the hypothesis that lateral redistribution of higher fluxes into the nearby sideslopes occurs as water penetrates deeper into the subsurface. Finally, the consistency of estimated fluxes between the chloride mass-balance method and the numerical model are noteworthy for the channel at borehole N37, the ridgetop at borehole N11, and the channel and sideslope at boreholes UZ#4 and UZ#5, respectively. Overall, the chloride mass-balance fluxes are consistent with the conceptual model of relative flux rates for the four topographically based infiltration zones. In addition, although there

are some differences, the chloride mass-balance fluxes and those derived from the infiltration model are quite similar, and in nearly all cases are the same order of magnitude. More detailed comparison between chloride mass-balance estimates and modeled net infiltration rates at borehole locations is discussed in Fabryka-Martin, Flint et al. (1997).

Estimates of infiltration also have made in alluvial soils based on $^{36}Cl/Cl$ ratios by evaluating the shape and magnitude of the of the ^{36}Cl bomb-pulse profile and the depth to the center of mass of the ^{36}Cl bomb pulse in soils (Fabryka-Martin, Wightman et al. 1994). However, this method yielded infiltration fluxes in channels at boreholes N37 and N54 that were one and two orders of magnitude greater than the fluxes from either the chloride mass-balance method or the numerical model (Flint, A.L. et al. 1996). This difference may be due to the recycling of infiltration water within the root zone, which is expected in the alluvial channels where evapotranspiration is high and little or no infiltration penetrates beyond the root zone (Flint, A.L. et al. 1996).

5.3.4.1.5.3 Estimates of Net Infiltration Rates for Potential Future Climates

To evaluate the effects on net infiltration of potential future climates, daily precipitation records from three locations in the South-Central Basin and Range were used: Nevada Test Site Area 12, South Lake, and Lake Valley-Steward. All three locations are at higher latitudes and elevations than Yucca Mountain and were considered to be reasonable analogs for potential future climates at Yucca Mountain. The Nevada Test Site Area 12 and Lake Valley-Steward sites are considered to be approximately representative of long-term average future climatic conditions that can be expected at Yucca Mountain (Schelling and Thompson 1997; Forester et al. 1996; P. Wigand et al., *Evidence of Late Quaternary Wet/Dry Climate Episodes Derived from Paleoclimatic Proxy Data Recovered from the Paleoenvironmental Record of the Great Basin of Western North America: Paleobotanical Studies*, unpublished Milestone Report 36213D5, Desert Research Institute, University and Community College System of Nevada). The spatially distributed average annual precipitation rate for the Yucca Mountain Site area using the 100-year stochastic simulation for the Nevada Test Site Area 12 analog was 289 mm/year, resulting in estimated average-annual net infiltration for the site area (Area 1) of 19.73 mm/year (Table 5.3-42). Net infiltration rates of about 200 mm/year and greater were estimated for the upper headwaters of Drill Hole Wash and along the northeastern-facing slope of the Prow of Yucca Mountain north of the potential repository site (Figure 5.3-111). Values of net infiltration are greater than 0 for the alluvial fan in the upper part of Yucca Wash, whereas estimates of 0 mm/year are obtained for these locations using current climate simulations. The mean net infiltration of 19.73 mm/year is in good agreement with estimates of recharge obtained for study sites located close to the Nevada Test Site Area 12 precipitation station in the northern part of the Nevada Test Site (Flint, A.L. et al. 1996). For the potential repository area (Area 5), the average annual net infiltration resulting from application of the Nevada Test Site Area 12 analog was about 33 mm/year, with a maximum estimate of more than 150 mm/year for several locations in the upper headwaters of drainages. Average net infiltration rates for Split and WT-2 Washes (Area 3) was 29 mm/year and the maximum rate was 169 mm/year for a channel. Net infiltration along stream channels was estimated to be about 77 mm/year, assuming that under wetter climatic conditions 50 percent of the runoff generated contributes to net infiltration along channels and the other 50 percent is runoff. Average annual runoff for the Nevada Test Site Area 12 analog simulation was 8.69 mm/year for the site area (Table 5.3-42), with runoff amounts of about 100

mm/year and greater at locations along the northern boundary of the site area(Figure 5.3-112). Minimum runoff estimates of 0.0 mm/year were obtained for most, but not all locations with deep soils.

Considerable temporal variability in average net infiltration and runoff rates for the Nevada Test Site Area 12 analog simulation is apparent in the 5-year and 15-year sliding means shown in Table 5.3-42. The statistical information indicates that 5-year and 15-year periods are not adequate for defining long-term average net infiltration rates because maximum and minimum values differ by a factor of 10 for the 5-year averages and even by a factor of 3 for the 15-year averages. Runoff is even more temporally variable in that the maximum and minimum 5-year and 15-year averages differ by factors of 100 and 7, respectively. The statistical information also indicates that 5-year and 15-year average runoff is not as well correlated to average precipitation as net infiltration because runoff is more sensitive to daily precipitation magnitudes. Therefore, a 5-year average for annual precipitation is likely to be an acceptable indicator of average net infiltration rates, but not as good an indicator of average runoff rates.

The spatial distribution of simulated net infiltration using the Nevada Test Site Area 12 analog (Figure 5.3-111) were similar, but not identical, to the spatial distribution obtained using the 100-year simulation for current climate (Figure 5.3-108). In Nevada Test Site Area 12 analog simulation, the relatively high net infiltration rates of 100 to 200 mm/year occurred at locations along the Prow of Yucca Mountain and along the steep, northeast-facing escarpment. At these locations, nonwelded tuff outcrops and relatively high-precipitation amounts and low evapotranspiration rates were simulated because of high elevation and the aspect of the slopes. However, maximum net infiltration greater than 200 mm/year resulted in the headwaters of channels because of more than an order of magnitude increase in the runoff volume generated for the wetter climate analog. The shift in the locations of maximum net infiltration from ridgetops and upper sideslopes to channels represents an important transition in the hydrologic system. The comparative results of these simulations were predicted by the conceptual model and demonstrate that the patterns of net infiltration vary both temporally and spatially. Although net infiltration rates may be very high in the upper channels under wetter climatic conditions (because the total area of the channels is very small), they still would not dominate the overall infiltration pattern. Rather, the general spatial distribution of net infiltration under wetter climatic conditions would remain quite similar to the current pattern in that increased net infiltration caused by increased precipitation at higher elevations and the mitigation of net infiltration by deeper soils in the alluvial fans and washes still would dominate the overall spatial pattern.

The South Lake analog site is considered to be approximately representative of climatic conditions at Yucca Mountain under possible future "super pluvial" conditions (Schelling and Thompson 1997; Forester et al. 1996; P. Wigand et al., *Evidence of Late Quaternary Wet/Dry Climate Episodes Derived from Paleoclimatic Proxy Data Recovered from the Paleoenvironmental Record of the Great Basin of Western North America: Paleobotanical Studies*, unpublished Milestone Report 36213D5, Desert Research Institute, University and Community College System of Nevada). The spatially distributed average annual precipitation rate for the Yucca Mountain Site area using the 100-year stochastic simulation for the South Lake analog was 427 mm/year, resulting in estimated average annual net infiltration for the site area (Area 1) of 81.0 mm/year (Table 5.3-43). Net

infiltration exceeded 800 mm/year in channels in upper parts of most washes and exceeded 500 mm/year in all parts of virtually all washes (Figure 5.3-113). The estimated average annual runoff resulting from the South Lake analog simulation was 52.3 mm/year. Assuming that 50 percent of the total runoff volume contributes to net infiltration along channels, an average net infiltration rate of 663 mm/year was calculated for the channels and added to net infiltration from precipitation in the channels, resulting in a maximum rate of 848 mm/year. The aerially averaged net infiltration rate obtained for the area of the potential repository (Area 5) was 108 mm/year, with a maximum rate of 706 mm/year occurring in several channels. Average net infiltration for the area of Split and WT-2 Washes (Area 3) was 106 mm/year.

In contrast to the model simulations of the current climate (Table 5.3-39; Figures 5.3-108 through -110) and the Nevada Test Site Area 12 wetter-climate analog (Table 5.3-42; Figures 5.3-111 and -112), simulation of the South Lake "super pluvial" analog resulted in net infiltration over the entire Yucca Mountain Site area of the grid (Figure 5.3-113), including penetration of deep soils of alluvial fans. Under these semi-arid conditions (427 mm/year precipitation), the primary control on estimated net infiltration would be the effective hydraulic conductivity of the bedrock, rather than soil thickness or even the spatial distribution of precipitation. Under current climatic conditions, and under even the wetter climatic conditions of the Nevada Test Site Area 12 analog, the relative control of bedrock conductivity was secondary to soil thickness and distribution of precipitation. In the South Lake "super pluvial" analog simulation, the much greater net infiltration rates throughout the channel network (500 mm/year and greater) resulted in the channels accounting for a relatively larger proportion of total areal net infiltration despite their relatively small area.

Although the 5-year and 15-year sliding means of net infiltration obtained for the South Lake "super pluvial" analog simulation indicate considerable year to year variability (Table 5.3-43), the coefficient of variation for the 15-year means is only 0.15, compared to 0.30 for the Nevada Test Site Area 12 analog results (Table 5.3-42) and 0.39 for the Yucca Mountain current climate results (Table 5.3-39). These statistical data strongly suggest that as the climate becomes wetter, the temporal distributions of net infiltration and runoff from year to year become less variable. It should be noted that the South Lake "super pluvial" analog probably represents the wettest conditions possible in the vicinity of Yucca Mountain during an interglacial-to-glacial transition. Unfortunately, these simulations do not take into account the dramatic changes in vegetation, soil cover, changes in climatic parameters such as air temperature and cloud cover that would be expected in conjunction with the wetter conditions. Therefore, these simulations of net infiltration under wetter climatic conditions probably are in the upper range of what might actually occur. Further, depending on the height of the regional water table or any local perched water tables, some locations in the study area, particularly channels, might become points of groundwater discharge rather than areas of net infiltration or recharge.

5.3.4.1.5.4 Estimates of Net Infiltration Rates Using a Coupled Infiltration-SurfaceFlow-Routing Model

As discussed in the previous subsection, contributions to net infiltration and recharge along channels due to the concentration of surface flow become more significant for wetter climates. Therefore, a more rigorous modeling approach that accounts for surface flow is needed for simulation of net

infiltration under climatic conditions that might be considerably wetter than current conditions. The subjective estimates of the percentages of runoff contributing to channel net infiltration that were made using the simplified runoff-routing method (90 percent for current conditions; 50 percent for wetter climatic conditions) were evaluated using a coupled infiltration–daily runoff-routing model (J.A. Hevesi and A.L. Flint, *Coupled Infiltration - Surface Flow Model for Estimating Spatially and Temporally Distributed Net Infiltration at Yucca Mountain, Nevada*, Milestone Report SPH253M4/SPH225M4, U.S. Geological Survey, in press). Because simulations of net infiltration using the daily surface-flow routing method for the entire infiltration-modeling area (Figure 5.3-107) are still in progress, results available at present are limited to the smaller modeling areas over the potential repository site (Area 5 and Area 3).

The coupled infiltration-daily runoff-routing model was applied to the area of the potential repository (Area 5) using the Yucca Mountain 1980 to 1995 precipitation record to estimate total daily surface runoff into channels (J.A. Hevesi and A.L. Flint, *Coupled Infiltration - Surface Flow Model for Estimating Spatially and Temporally Distributed Net Infiltration at Yucca Mountain, Nevada*, Milestone Report SPH253M4/SPH225M4, U.S. Geological Survey, in press). Simulation results for the runoff event on March 11, 1995, indicated total daily runon (same as cumulative runoff) of more than 10,000 mm (or flow volumes of more than 9,000 ^{M3}) for the lower segments of Drill Hole Wash, Wren Wash, and Split Wash (Figure 5.3-114). Because these flow volumes exceeded the hydraulic conductivity of the soil along the channels, most of the flow ran off the surface of Area 5 without infiltrating. Unfortunately, this simulation does not account for channel flow along Drill Hole Wash originating as runoff from the headwater areas because the northern part of the drainage is outside of the model area. Although considerable amounts of surface flow infiltrated into the 6-m deep soil profile along channels, the field capacity was exceeded only for the shallower soils in the headwater areas of the washes. Model-calculated total net infiltration for March 11, 1995, was more than 20 mm/day for the upper parts of channels and for bottom sideslopes having intermediate soil thicknesses that intercepted surface flow from the higher sideslopes (Figure 5.3-115). In addition, exposures of nonwelded tuffs (PTn) along the west-facing sideslope of Solitario Canyon intercepted runoff from the crest of Yucca Mountain and into Solitario Canyon, causing large amounts of infiltration into the nonwelded tuffs. Although fairly high rates of net infiltration (5 to 7 mm/day) occur along the crest and ridgetops, the maximum infiltration rates simulated for this event were caused by the concentration of surface flow.

Application of the coupled infiltration, daily runoff-routing model to the developed 1980 to 1995 daily precipitation record (J.A. Hevesi and A.L. Flint, *Coupled Infiltration - Surface Flow Model for Estimating Spatially and Temporally Distributed Net Infiltration at Yucca Mountain, Nevada*, Milestone Report SPH253M4/SPH225M4, U.S. Geological Survey, in press) resulted in an areally averaged net infiltration rate of 11.5 mm/year for the area of the potential repository site (Area 5), corresponding to an aerially averaged precipitation rate of 171 mm/year. Although greater, these values are quite similar to the 6.0 mm/year average net infiltration and 163 mm/year average precipitation obtained for Area 5 using the 100-year stochastic precipitation simulation (Table 5.3-38). The maximum net infiltration for this simulation of 77 mm/year occurred in a channel in upper Solitario Canyon where the soil is thin and the bedrock is nonwelded tuff. Net infiltration rates greater than 50 mm/year were calculated for locations overlying the nonwelded tuffs along the western facing slope of Solitario Canyon, and for channels in the upper part of drainages on the

eastern slope of Yucca Mountain (Figure 5.3-116). The areally averaged mean runoff for this simulation was 4.6 mm/year, while the mean amount of stream discharge was 1.5 mm/year, or 33 percent of the runoff volume generated.

An important difference between the simulations using the daily flow-routing model and the empirical runoff-redistribution approach is that the flow-routing model allowed and accounted for net infiltration into the sideslopes due to overland flow (J.A. Hevesi and A.L. Flint, *Coupled Infiltration - Surface Flow Model for Estimating Spatially and Temporally Distributed Net Infiltration at Yucca Mountain, Nevada*, Milestone Report SPH253M4/SPH225M4, U.S. Geological Survey, in press). This resulted in a significant concentration of net infiltration (greater than 20 mm/year) at the base of many sideslopes immediately adjacent to channels in which net infiltration is near zero because of thick soils. In addition, results of the daily flow-routing simulation suggests that net infiltration rates are highly variable along channels, with relatively high rates restricted to the upper headwaters sections, while downstream along the same channel, infiltration rates were still near zero. This result is consistent with changes in water-content profiles measured in neutron boreholes located in upper channels (such as N15 in Pagany Wash; Figures 5.3-95 and -98) which tended to indicate high rates of net infiltration. It should be noted, however, that actual channel geometries have not been represented by this daily flow-routing simulation. Channel segments represented by the model grid are 30 m wide, which is unrealistic for most of the drainages on Yucca Mountain. Narrowing the channel width would increase the cumulative runoff in the channel and force a greater portion of the total flow volume to continue downstream. In the current model, the 30 m wide channel segments spread the total flow volume out across a larger area, causing most of the flow to infiltrate upstream.

Application of the coupled infiltration–daily runoff-routing model to the 100-year stochastic simulation for the South Lake "super pluvial" analog (J.A. Hevesi and A.L. Flint, *Coupled Infiltration - Surface Flow Model for Estimating Spatially and Temporally Distributed Net Infiltration at Yucca Mountain, Nevada*, Milestone Report SPH253M4/SPH225M4, U.S. Geological Survey, in press) resulted in an areally averaged net infiltration rate for the area of Split and WT-2 Washes (Area 3) of 116 mm/year. In this simulation, net infiltration rates greater than 700 mm/year occurred in relatively long sections of the channels, but the maximum rates of more than 1,000 mm/year still occurred only in the upper parts of the channel (Figure 5.3-117). Interestingly, the mean runoff calculated for Area 3 was 67 mm/year and the mean simulated outflow was 29 mm/year, or 43 percent of the runoff generated. This percentage is fairly close to the estimated 50 percent assumed for the simpler, empirical runoff-redistributionmethod, and the values of mean and maximum net infiltration rates also were fairly consistent between the two methods. The mean runoff amount generated using the daily runoff-routing model was greater than the mean amount obtained using the simpler method because the flow-routing model allowed downstream cells to saturate and contribute to runoff, whereas this process was neglected by the simpler model.

5.3.4.1.6 Summary of Net Infiltration Modeling Results

Results of numerical-model simulations of net infiltration at Yucca Mountain indicate that the average annual rate representative of the long-term (100 year) current climate conditions is 3.2 mm/year for the site area. A maximum estimate of 63.2 mm/year was obtained for a location

on the Prow to the northwest of the potential repository area, and estimates of 0 mm/year were obtained for all locations having 6 m or more soil cover. Net infiltration along stream channels was estimated to be 9.2 mm/year, assuming that 90 percent of the runoff generated contributes to net infiltration along channels. Average annual net infiltration for the vicinity of the potential repository was estimated to be 6.0 mm/year, with a maximum net infiltration of 29.7 mm/year occurring near the crest of Yucca Mountain. These estimates are in good agreement with an initial estimate obtained for the site area using a scaled 1-year simulation. Average net infiltration rates simulated using the 1980 to 1995 precipitation record for the site (7.6 mm/year for the site area) are higher than the estimated long-term averages, but are consistent with the variability in the 15-year sliding means resulting from the 100-year stochastic precipitation simulations. The high frequency of wetter-than-normal winters for 1992, 1993, and 1995 result in the 1980 to 1995 period being wetter than the average of the previous 100 years. However, this period may be representative of inherent climatic change in response to greenhouse warming, and thus may be a better estimate of conditions at the site for the near future.

For simulation of the Nevada Test Site Area 12 wetter climate analog, which is considered to be potentially similar to conditions that can be expected at Yucca Mountain during the longer term future, average annual net infiltration using a 100-year stochastic simulation was 19.7 mm/year for the site area. For this simulation, infiltration rates of 200 mm/year and greater were estimated for the northeast-facing slideslope near the Prow of Yucca Mountain, and for a channel in the headwaters of Drill Hole Wash. The mean estimate of 19.7 mm/year is in good agreement with estimates of recharge obtained for study sites located close to the Nevada Test Site Area 12 precipitation station in the northern part of the Nevada Test Site. For the area of the potential repository, average annual net infiltration was estimated to be 32.7 mm/year, with a maximum estimate of more than 150 mm/year for several locations in the upper headwaters of drainages. Net infiltration along stream channels is estimated to be 77 mm/year, assuming that 50 percent of the runoff generated contributes to net infiltration along channels. For simulation of the Nevada Test Site Area 12 wetter climate analog, average annual runoff was 8.7 mm/year.

To compare results from the numerical infiltration model with previous estimates of net infiltration and recharge (Subsection 5.3.4.1.1), the 15-year sliding means of annual net infiltration obtained for the various 100-year stochastic simulations (assuming 90 percent of generated runoff contributes to net infiltration in the channels) were plotted against the corresponding 15-year sliding means of average annual precipitation (Figure 5.3-118). For this comparison, modeled net infiltration at Yucca Mountain was assumed to be representative of average annual recharge to the saturated zone at the site. The 15-year sliding means include results obtained for Yucca Mountain and three wetter climate analog sites (Nevada Test Site Area 12, Lake Valley, and South Lake). The comparison also includes:

- The modified Maxey-Eakin model used by Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press)

- The original Maxey-Eakin (1949) step function

- The estimates obtained by Lichty and McKinley (1995) using both the Precipitation Runoff Modeling System water balance model and the chloride-balance model

- The average annual net infiltration estimate for Yucca Mountain obtained by Hudson and Flint (1996) using a statistical model based on the 1984 to 1995 neutron logging record

- The estimate of average annual recharge for the upper Amargosa River basin obtained by Osterkamp et al. (1994) using streamflow simulation models.

In general, results obtained using the numerical infiltration model are found to be fairly consistent with previous estimates of recharge in the Yucca Mountain region, although the model results tend to be higher, on average, than previous estimates. The results obtained using the wetter-climate analog simulations are in very good agreement with the modified Maxey-Eakin model developed using a completely independent approach (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press). The results for the Lake Valley and Nevada Test Site Area 12 analogs match well with the modified Maxey-Eakin model, while the results for the South Lake "super pluvial" analog are slightly higher than the modified Maxey-Eakin model. Results obtained for the Nevada Test Site Area 12 and Lake Valley analogs also are in good agreement with the higher estimate of recharge obtained by Lichty and McKinley (1995) for the Three-Springs Basin in the Kawich Range. Unfortunately, even the South Lake "super pluvial" analog is too dry to provide an adequate comparison to the East Stewart Creek site (recharge of about 300 mm/year) studied by Lichty and McKinley, although extrapolation of the model results suggests good agreement. Numerical-model estimates of net infiltration for average annual precipitation representative of the current climate at Yucca Mountain, including the 100-year mean of 3.25 mm/year, tend to be higher than recharge estimated by the Maxey-Eakin method, the modified Maxey-Eakin method, and by Osterkamp et al. (1994). However, the numerical-model 100-year estimates are lower than the estimate of 11.9 mm/year obtained by Hudson and Flint (1996) and the modeled estimate of 7.6 mm/year obtained using the 1980 to 1995 daily precipitation record.

The 15-year sliding mean net infiltration rates obtained using the numerical infiltration model are well correlated with the corresponding average annual precipitation values, suggesting that models of recharge based primarily on average annual precipitation, such as the Maxey-Eakin model and its various modified versions, are likely to provide reasonable estimates of average annual recharge in the Yucca Mountain region, at least on a regional scale. In particular, the results of the numerical infiltration model obtained using the wetter climate analog sites are consistent with recent methods used to model recharge for the Death Valley Basin which were applied in developing the recharge boundary condition for the regional groundwater flow model described in an earlier subsection (D'Agnese et al. 1997). Discrepancies in modeling results can be attributed to a combination of model uncertainty, differences in the hydrologic characteristics of the various study sites, and differences in climatic characteristics, such as the timing and frequency of precipitation, which are not accounted for using a comparison based on average annual precipitation.

Sources of uncertainty in the numerical model include the uncertainty in the geospatial input parameters, uncertainty in the data and developed data used to calibrate the model, and limitations concerning the model's representation of physical processes. Soil depth, soil hydrologic properties,

and the effective hydraulic conductivity of the underlying bedrock are all critical in the net infiltration model. The assumption of a uniform soil depth of 0.5 m for all upland areas mapped as having thin to no soil cover may not be an adequate representation of the site. Values for effective bedrock conductivity are strongly dependent on a limited amount of data concerning fracture properties, particularly the spatial distribution of fracture properties in the near surface. Evapotranspiration of water stored in the bedrock fractures and matrix is not currently included in the version of the model used to estimate net infiltration for the entire study area. Daily precipitation amounts used as input to drive the model do not correctly account for short duration, high-intensity precipitation events common in the summer, which are likely to produce significant runoff without causing a large increase in the water content of the soil profile. The modeling of the spatial distribution of daily precipitation based the distribution of average annual precipitation probably is a reasonable representation of winter precipitation, but clearly does not adequately represent the distribution of summer precipitation. The effect of vegetation on evapotranspiration in response to wetter climates is not accounted for, nor are the effects of air temperature and cloud cover on precipitation, runoff, and evapotranspiration. The failure of the current numerical infiltration model to account for these factors introduces significant uncertainty into the estimates of infiltration under possible future wetter climatic conditions. The representation of stream channels using 30 m grid cells is a significant simplification of stream-channel geometries and is apt to strongly affect simulated quantities of stream discharge and net infiltration along channels. The use of a rectangular model grid that does not correspond to watershed boundaries precludes the accounting for surface-water inflows from upstream drainages, which is especially problematic for estimating net infiltration along Fortymile Wash.

Efforts are underway to improve the numerical infiltration model, particularly with respect to the representation of the physical processes of evapotranspirationand runoff accumulation. Field studies involving the time-continuous monitoring of soil-water potentials and water contents using heat dissipation probes, along with time-continuous monitoring of the near-surface vapor flux density using eddy correlation stations, are providing results that should allow for a more rigorous calibration of the evapotranspiration component of the model. Current progress in modifying the model to allow runoff routing in the time domain using a kinematic-wave approximation for overland flow will provide a more accurate representation of net infiltration along channels, and may allow for a more elaborate calibration of the overall model using both the neutron-moisture profiles of boreholes and hydrographs of channelized runoff available for the Yucca Mountain Site. This work includes redefining the modeling domain using watershed boundaries rather than arbitrary rectangular grids. Also in progress is an analysis of the sensitivity of modeling results to parameters such as soil depth and the effective hydraulic conductivity of the underlying bedrock. Despite the current uncertainties and limitations of the model, estimates of net infiltration for the Yucca Mountain Site area and the area of the potential repository are considered reasonable, although still preliminary. Confidence in the net infiltration rates derived from the numerical model is derived from approximate agreement with results obtained using the chloride mass-balance and temperature models, calibration against 1980 to 1995 daily precipitation and 1984 to 1995 water-content profiles, and consistency with the large body of filed data used to develop the conceptual model of infiltration.

5.3.4.2 Characterization of the Deep Unsaturated Zone

Prior to 1997, nine deep boreholes were instrumented in the unsaturated zone at Yucca Mountain to measure in situ pneumatic pressure, water potential, and temperature at multiple depths in each borehole in accordance with a fully compliant quality assurance program (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997). These boreholes were NRG#5, NRG-6, NRG-7a, SD-7, SD-9, SD-12, UZ#4, UZ#5, and UZ-7a (Figure 5.3-8). Two of these boreholes (UZ#4 and UZ#5) are located in Pagany Wash; four boreholes (NRG#5, NRG-6, NRG-7a, and SD-9) are located in or near Drill Hole Wash; UZ-7a is located in WT-2 Wash; SD-12 is located in Antler Wash; and SD-7 is located on the flank of Highway Ridge. Boreholes NRG#5, NRG-6, NRG-7a, and SD-9 are located along the Exploratory Studies Facility North Ramp, whereas SD-12, UZ-7a, and SD-7 are located along the Exploratory Studies Facility Main Drift. The methods used to instrument, sample, and monitor these boreholes are described in Subsections 5.3.2.1.4 and 5.3.2.1.5. Locations of instrument stations with respect to lithostratigraphy in each of these boreholes are shown in Figure 5.3-14; the symbols depicting lithostratigraphic units are defined in Figure 5.3-15. In addition, two other boreholes (UZ-1 and NRG#4) were instrumented and monitored, but most of the data were not obtained in accordance with a fully compliant quality assurance program. Borehole UZ-1, located in Drill Hole Wash, was instrumented in October 1983 and monitored through October 1987. Because of sensor drift problems, only the early-time pneumatic-pressure and water-potential data from this borehole are considered reliable (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Borehole NRG#4 was instrumented by Nye County to monitor pneumatic pressure near the Exploratory Studies Facility North Ramp (Multimedia Environmental Technology, Inc. 1995). Despite their non-quality status, data from boreholes UZ-1 and NRG#4 were used to characterize the unsaturated-zone system in the vicinity of the Exploratory Studies Facility North Ramp because data from both UZ-1 and NRG#4 tend to corroborate and support observations made in other boreholes, especially with regard to the pneumatic characteristics of the PTn.

Data obtained from the monitoring and sampling of the instrumented boreholes were used to characterize various aspects of the deep unsaturated-zone system. Pneumatic-pressure fluctuations in the deep unsaturated zone were analyzed to determine general pneumatic characteristics and vertical permeability of major hydrogeologic units, flow of gas and water vapor through the deep unsaturated zone, and the effects of Exploratory Studies Facility construction on the gaseous-phase system (Subsection 5.3.4.2.1). In situ water potentials and the occurrence of perched water were analyzed to describe the overall liquid-water flow regime in the deep unsaturated zone (Subsection 5.3.4.2.2). Selected temperature data from previous studies and the instrumented boreholes were analyzed to estimate heat flow in the unsaturated zone (Subsection 5.3.4.2.3). In addition, geochemical data were collected from most of the instrumented boreholes to help characterize the flow of gas and water through the unsaturated zone (Subsection 5.3.4.2.4). Geochemical data include the chemistry and isotopic content of pore water extracted from rock cores, bulk composition and isotopic content of gas pumped from the boreholes, and chemical and isotopic content of perched water encountered in some of the boreholes.

5.3.4.2.1 Pneumatic Pressure and Gaseous-Phase Movement

Gas and water vapor movement through the unsaturated zone is driven by changes in barometric pressure, temperature-induced density differences, and wind effects (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Changes in barometric pressure at the land surface result in corresponding changes in pneumatic pressure in the unsaturated zone, which commonly are amplitude attenuated and time lagged relative to the pressure change at the surface. The attenuation and lag are functions of the air permeability and the associated pneumatic diffusivity of the rocks. Temperature-induced density effects are significant only in hilly terrain and are dominated by drier, cooler air entering along hillsides and moving through the subsurface. This air becomes geothermally heated as it moves through the rock and eventually can move vertically upward to exhaust along hillcrests. This is referred to as the "thermosyphon" effect by mechanical engineers and is referred to as such in this report (Patterson et al. 1996).

The subsurface-pressure response to surface barometric changes is dominated by the bulk pneumatic diffusivity of the layers of rock present at a given location. Consequently, the subsurface responses are controlled by the distribution and interconnectedness of fractures, the presence of faults and their ability to conduct gas and vapor, and the moisture content and matrix permeability of the rock units themselves. Specific aspects of the pneumatic system discussed in this report include the ability of the PTn, vitric units, and perched-water zones to act as impediments to the propagation of surface barometric-pressure changes and thus act as impediments to gas and vapor movement (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164).

As part of the characterization of Yucca Mountain, several studies have been conducted to develop an understanding of the movement of gas and water vapor through the thick unsaturated zone. As part of these studies, several boreholes have been instrumented in the unsaturated zone and continuously monitored to record changes in pneumatic pressure at depth. The purposes of the pneumatic monitoring were threefold:

- Measure in situ pneumatic pressure response to atmospheric pressure changes to determine the influence of lithostratigraphy, bulk properties, and major structural features on the natural gaseous-phase circulation system

- Use the response of the system at depth to changes in atmospheric pressure to estimate pneumatic diffusivities and, ultimately, effective gas permeabilities of the lithostratigraphic layers

- Document and quantify the effects of excavation of the Exploratory Studies Facility on in situ pneumatic pressure and on the overall gaseous-phase-circulation system (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory*

Studies Facility, Yucca Mountain, Nevada, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164)

5.3.4.2.1.1 Synthesis of Subsurface Pneumatic Pressure Prior to the Influence of Exploratory Studies Facility Excavation

The computed phase lags and residual amplitudes of the in situ pressure data indicate that individual lithostratigraphic units can be grouped conveniently into four distinct pneumatic systems:

- The TCw tuffs
- The PTn hydrogeologic unit that includes the crystal-poor vitric base of the Tiva Canyon Tuff (Tpcpv1, 2), the Yucca Mountain and Pah Canyon Tuffs (with associated bedded tuffs), and the crystal-rich top of the Topopah Spring Tuff (Tptrv2, 3)
- The Topopah Spring lithophysal and nonlithophysal welded units
- The pre-Topopah Spring bedded tuff, the nonwelded tuffs of the Calico Hills Formation, and the pre-Calico Hills Formation bedded tuff (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164).

This grouping conforms to the definition of the major hydrogeologic units presented in Figure 5.3-18. The responses observed in each of these pneumatic systems to the propagation of atmospheric pressure changes from land surface to depth are discussed in the following subsections.

Tiva Canyon Welded Hydrogeologic Unit–For the most part, pneumatic pressure records for the TCw display very little attenuation and lagging of the synoptic pressure signal (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Patterson et al. 1996), as exemplified by the pressure record from borehole NRG#5 (Figure 5.3-119). Spectral analysis of the subsurface synoptic pressure signal indicates very rapid transmission of atmospheric pressure changes throughout the TCw (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). In many cases it is not possible to distinguish the atmospheric signal from the downhole pressure signal. Because of the topographic position of the unit and its direct exposure to the atmosphere throughout most of the site area, gas flow within the TCw can be expected to be strongly influenced by topographic and barometric-pumping effects. The only significant exception to this behavior is the pneumatic pressure record from the Tiva Canyon crystal poor columnar subzone (instrument station G) in borehole NRG-6 (Figure 5.3-120). The data from this instrument station for May and June 1995 indicate that the residual amplitude of the pressure signal is only 56 percent of the synoptic signal and that its phase lag is 1.8 hours. These spectral parameters are significantly different from those of all other pressure records from TCw monitoring instrument

stations where residual amplitudes of the synoptic signal are on the order of 96 percent or greater, and phase lags are less than one hour. Shallow soil cover and temporary pneumatic sealing of near-surface fractures by localized infiltrating water associated with unusually high precipitation during late 1994 and early 1995 seems to be a plausible explanation for the small residual amplitudes in the TCw at borehole NRG-6 (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Spectral analysis of earlier data from borehole NRG-6 (beginning in February 1995) indicates that the residual amplitude was only 43 percent with a phase lag of 9.6 hours (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Both of these analyses are for data collected prior to the effects of Exploratory Studies Facility excavation, thus the increasing residual amplitude and decreasing phase lag are likely the result of drying out of surface materials following transient infiltration events. It is likely that transient infiltration occurred at other boreholes but the effects are not readily apparent in the pneumatic records (Patterson et al. 1996).

Paintbrush Nonwelded Hydrogeologic Unit–The PTn includes the crystal-poor vitric zones (Tpcpv1, 2) at the base of the Tiva Canyon, the crystal-rich vitric zones (Tptrv2, 3) at the top of the Topopah Spring Tuff, as well as the intervening Yucca Mountain, Pah Canyon, and associated bedded tuffs (Table 5.3-13). Across the Yucca Mountain Site area, the PTn becomes thinner from north to south (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) and in the monitored boreholes this unit is thickest at borehole UZ#4 (84 m) and thinnest at UZ-7a (24 m) (Patterson et al. 1996).

Pneumatic pressure records from monitoring boreholes at different locations within the site area indicate that the attenuation of the synoptic pressure signal across the PTn is greater in boreholes in the northern part of the site area (NRG-7a, UZ#4, and UZ#5) than in the southern part of the site area (SD-7, SD-12, and UZ-7a) (Patterson et al. 1996). For example, in borehole UZ#4, substantial attenuation of the synoptic pressure signal across the PTn is evidenced by a residual amplitude at instrument station A of 16 percent and a phase lag of 22.9 hours (Figure 5.3-121). In borehole UZ-7a, however, attenuation of the synoptic pressure signal across the PTn is substantially less, as evidenced by a residual amplitude at instrument station D of 73 percent and a phase lag of 6 hours and the relatively high residual amplitudes at all the TSw instrument stations (Figure 5.3-122). Evaluation of the pneumatic pressure data from various boreholes indicates that the attenuation characteristics of different units of the PTn vary from one location to another, and that the composite thickness of the PTn is insufficient to uniquely account for these variations (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Patterson et al. 1996). Differences in the pneumatic diffusivities of individual units within the PTn are obviously present. These differences probably are a reflection of local differences in the saturation of individual units and/or the presence or absence of open fractures within these units. In the absence of localized fracturing, the very low matrix porosities of the crystal-rich vitric and crystal-poor vitric zones render the pneumatic permeabilities of these units

particularly sensitive to small changes in moisture content. Locally the pneumatic characteristics of the crystal-rich vitric zone appear to be dominated by fracturing. Detailed descriptions of these borehole-to-borehole variations are contained in J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) and Patterson et al. (1996). Although the values of residual amplitude and phase lag from these two studies are different for some borehole, primarily because different period of pneumatic record were analyzed, the overall conclusions of the two studies were the same, as summarized above.

Topopah Spring Lithophysal and Nonlithophysal Welded Tuffs–Regardless of location within the site area, borehole pneumatic pressure data indicate negligible attenuation of pneumatic pressure signals across the Topopah Spring crystal-rich vitric zone (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Patterson et al. 1996), as exemplified by boreholes UZ#4 and UZ-7a (Figures 5.3-121 and 5.3-122). Further, pressure signals appear to be transmitted nearly instantaneously throughout most of the entire vertical section of the TSw, as exemplified by borehole NRG-6 (Figure 5.3-120) and borehole SD-12 (Figure 5.3-123). Nearly all the pressure data from the TSw indicate that the fractures within the TSw apparently are very permeable and highly interconnected within both the lithophysal and nonlithophysal units. The only exceptions to this are the lowermost instrument stations of boreholes NRG#5 (Figure 5.3-124) and SD-12 (Figure 5.3-123). In borehole NRG-5, the residual amplitude of pneumatic pressure increases from 29 percent at instrument station 9 to 39 percent at station 10, and the phase lag decreases by over 10 hours from station 9 to station 10. This reversal in the normal trend of increasing attenuation with depth probably is due to the barometric pressure signal bypassing the PTn by traveling preferentially down the Drill Hole Wash fault (Patterson et al. 1996). A similar situation exists in borehole SD-12 where station C exhibits a larger residual amplitude and smaller phase lag than the overlying stations. In the vicinity of borehole SD-12, the likely preferential pathway for the pneumatic signal is through the Ghost Dance fault.

Calico Hills Nonwelded Hydrogeologic Unit–Pneumatic pressures in boreholes SD-9 and SD-7 were monitored from inside the annular space of nested casings that were open between the bottom of a perched-water zone and the top of the water table (Patterson et al. 1996). Perched water was encountered during drilling near the base of the Topopah Spring Tuff in borehole SD-9, and near the base of the Calico Hills Formation in borehole SD-7 (see Subsection 5.3.4.2.2.2). Although no perched-water reservoir of sufficient magnitude to cause standing water in the borehole was detected during the drilling of SD-12 (Rautman and Engstrom 1996b), the video-camera log of this borehole indicates that a perched-water zone of limited extent probably is present in the densely welded vitric subunit (Tptpv3) of the Topopah Spring Tuff (Rousseau et al. 1997). In borehole SD-12, instrument stations A and B are located within the Calico Hills Formation and within the crystal-poor vitric unit at the base of the Topopah Spring Tuff, respectively, both of which are below the perched-water zone in the densely welded vitric subunit (Tptpv3). Figure 5.3-125 shows pressure records measured below the perched-water zones in boreholes SD-7, SD-9, and SD-12. The pressure records at all three boreholes indicate that essentially all of the synoptic barometric signal is attenuated. This is consistent with the concept that the presence of a perched-water zone with extremely low permeability to air would effectively impede the downward propagation of the surface barometric

signal. The small fluctuations seen in these records likely are the result of horizontal propagation of the barometric signal from outside the margins of the perched-water zones.

5.3.4.2.1.2 Determination of Pneumatic Diffusivity and Vertical Permeability Using Borehole Pneumatic-Pressure Data

The pneumatic-pressure data from the monitoring boreholes show that barometric pressures at the ground surface experience amplitude reductions and phase lags as they propagate through the subsurface. The magnitudes of the amplitude reductions and phase lags in pneumatic pressure are a reflection of the pneumatic diffusivities of the layers through which the pneumatic-pressure signal has passed, and can be described by the transient gas-diffusion equation (Rousseau et al. 1997). The time-varying pneumatic pressures were used to estimate the pneumatic diffusivities, and ultimately, the effective gas permeabilities of single and combined stratigraphic layers bounded by monitoring stations. These analyses were conducted using the AIRK program developed by Weeks (1978) and the TOUGH2 program developed by Pruess (1987, 1991). Both the AIRK and TOUGH2 codes solve the gas diffusion equation numerically using either finite-difference (AIRK) or integrated finite-difference (TOUGH2) methods to estimate the time-dependent vertical distribution of gas pressure, given initial and boundary conditions provided by the data from the instrumented boreholes. Beginning at the bottom or top of a stack of layers, the pneumatic diffusivity of each layer was estimated one layer at a time, using the pressures at the upper or lower stations as the appropriate boundary condition for that layer. Estimates of pneumatic diffusivity of the layer were adjusted, either manually by the user or automatically by the program, until a satisfactory match between the measured and simulated pressures was achieved. Because pneumatic diffusivity is a function of both the effective permeability to air of the medium and the air-filled porosity, unique or even reasonable solutions for permeability are possible only if air-filled porosity is known and the flow domain is truly one-dimensional. Given that air flow in the unsaturated zone at Yucca Mountain occurs predominantly through fractures, the estimates of permeability are subject to uncertainty because independent estimates of secondary fracture porosity were not available at the time these analyses were performed.

Pneumatic diffusivities for lithostratigraphic intervals in boreholes NRG-6 and NRG-7a in Drill Hole Wash and boreholes UZ#4 and UZ#5 in Pagany Wash were determined using the AIRK program by manually adjusting estimated values for each layer until a satisfactory match between the measured and simulated pressures was achieved (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Pneumatic diffusivities for hydrogeologic units in boreholes NRG#5, NRG-6, NRG-7a, SD-7, and SD-9 were determined using the AIRK program by allowing the program to automatically adjust estimated values for each layer until a satisfactory match between measured and simulated pressures was achieved (Patterson et al. 1996). In addition, pneumatic diffusivities for lithostratigraphic intervals in boreholes SD-7, SD-12, and UZ-7a were determined using the TOUGH2 program (Rousseau et al. 1997).

A comparison of the results of these three sets of analyses are shown in Table 5.3-44. For the TCw, the means of the air-permeability values from Patterson et al. (1996) and Rousseau et al. (1997),

seem in agreement, but both are two orders of magnitude higher than the mean of the values obtained by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). However, the mean of the air-permeability values for the TCw estimated by Patterson et al. (1996) was greatly affected by a very large value $3{,}000\times10^{-12}$ m^2 from borehole NRG#5, which is very near the Drill Hole Wash fault. The means of the air-permeability values for the PTn are in much closer agreement, especially considering that the mean of the values from Rousseau et al. (1997) was greatly affected by a very large value 150×10^{-12} m^2 from borehole UZ-7a, which was drilled into the Ghost Dance fault zone. For the TSw, the means of the air-permeability values from J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) seem in agreement, with the mean of the values from Patterson et al. (1996) being an order of magnitude higher. Interestingly, both J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) and Patterson et al. (1996) derived the highest value of air permeability for the TSw at borehole NRG-6, 50×10^{-12} m^2 and 500×10^{-12} m^2, respectively.

The differences in the estimates of permeability derived by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) and Patterson et al. (1996) are related to differences in the estimated values of air-filled porosity within given depth intervals, and to the fact that, in the TCw and the TSw, J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) manually identified the minimum permeability that would result in no additional amplitude dampening or phase lag in the pneumatic-pressure signal measured at adjacent stations. J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) reasoned that when there was no offset or amplitude changes with depth, only a minimum permeability could be identified. All permeability values above this threshold permeability would result in simulated pneumatic-pressure signals that were nearly identical such that the optimization criteria was relatively insensitive to changes in permeability above this threshold. Patterson et al. (1996), on the other hand, used the automatic search procedure in the AIRK program to identify the permeability that minimized the optimization criteria, even though relatively large changes in permeability probably resulted in only small changes in this value. Therefore, permeability values for the TCw and TSw estimated by Patterson et al. (1996) tend to be significantly larger than those of J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) because of the manner in which the optimization criteria was applied.

Comparison of air-permeability values on a borehole-to-borehole basis probably is more meaningful than comparison of mean values for hydrogeologic units because of significant spatial variability in

bulk properties due to faulting and associated fracturing (see Subsection 5.3.3.1.2). For example, air-permeability values estimated by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) and Patterson et al. (1996) for the same hydrogeologic units in boreholes NRG-6 and NRG-7a compare very favorably, with both sets of values falling in the same order-of-magnitude range except for the values for the TSw in borehole NRG-6. Borehole-to-boreholecomparisons between the air-permeability values derived from the pneumatic-diffusivity determinations of J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; 1997) and air-permeability values determined by air-injection testing are discussed in Subsection 5.3.3.1.2.4. In general, air-permeability values for the TCw and the PTn compare very well, whereas values from the diffusivity determinations are about an order of magnitude higher than air-injection values for the TSw.

Similar analyses of pneumatic pressure at Yucca Mountain have been reported previously for borehole NRG#4 (Multimedia Environmental Technology, Inc. 1995) and a subset of the monitoring stations at boreholes NRG-6 and NRG-7a (Ahlers, Bandurraga et al. 1995b).

5.3.4.2.1.3 Effects of Excavation of the Exploratory Studies Facility Tunnel on In Situ Pneumatic-Pressure Measurements in Boreholes

Pneumatic-pressure interference effects associated with excavation of the Exploratory Studies Facility by the tunnel boring machine have been observed in all monitored boreholes (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997; Patterson et al. 1996), with the possible exception of UZ-7a. Because borehole UZ-7a was drilled within the zone of the Ghost Dance fault, all zones within the borehole exhibit only partially attenuated responses to the surface barometric signal, which may effectively mask the detection of any possible effects of Exploratory Studies Facility excavation. The effects of Exploratory Studies Facility excavation on in situ pneumatic pressure were carefully monitored because such effects offer some insight on how the overall gaseous-phase system in the unsaturated zone has been effected by the direct exposure of deeply buried rock units to atmospheric pressure by way of the Exploratory Studies Facility tunnel. Further, monitoring of such effects also presented an opportunity to use the pneumatic stress caused by Exploratory Studies Facility excavation to determine large-scale pneumatic properties of the rock units and to identify pneumatic fast pathways within the subsurface. Table 5.3-45 indicates the event and first occurrence of pneumatic-interference effects for each affected borehole at Yucca Mountain. The tunnel boring machine began excavation of the Exploratory Studies Facility tunnel at the North Portal, proceeded northwestward subparallel to Drill Hole Wash to complete the North Ramp, crossed under Drill Hole Wash, and then turned straight south to complete the Main Drift. Monitored boreholes were located along the North Ramp and the Main Drift; no monitored boreholes were located along the South Ramp. Locations of monitored boreholes in relation to the Exploratory Studies Facility are shown in Figure 5.3-8.

Detection of the first occurrence of a pneumatic disturbance effect may not coincide exactly in time with key excavation events in the Exploratory Studies Facility tunnel, such as penetration of a fault or removal of a confining or pneumatic-impeding layer (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Many of the recorded dates for pneumatic-interferenceevents appear in the pneumatic record several days after the excavation event. The apparent lag effect can be attributed to two primary causes. First, changes in the frequency and amplitude of the synoptic pressure signal as registered downhole were used to determine when a tunnel interference effect had occurred. The wavelength of this signal is on the order of 96+ hours, depending on the time of year (longer in the summer, shorter in the winter). To observe a change in in situ pressure using this signal requires at least a half wavelength of record (2 days or more) before changes in the amplitude and frequency become apparent. Second, the rate of advance of the tunnel boring machine tends to obscure detection of the first arrival of the pneumatic interference because this interference is initially propagated as a weak spherical wavefront that gradually evolves into a stronger cylindrical wavefront as the tunnel boring machine advances closer and closer to the affected borehole. Obviously, if the tunnel boring machine was advancing very rapidly the weak spherical interference may not be readily discernible given the half wavelength delay time (2+ days) needed to confirm a change in the frequency and/or amplitude of the in situ pressure signal. Conversely, pressure interferences at boreholes UZ#4 and UZ#5 that clearly were propagated along a fault plane were detected in advance of fault penetration by the Exploratory Studies Facility tunnel, but only because the tunnel boring machine was shut down for 10 days and did not actually penetrate the fault until 8 days after tunneling operations resumed. At borehole NRG-6, the pneumatic interference was not observed until 2 days after the Drill Hole Wash fault structure was penetrated for reasons cited above.

Penetration of the crystal-rich vitric zone at the base of the PTn by the tunnel boring machine occurred in the Exploratory Studies Facility North Ramp on June 20 1995 (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164), but was first evidenced by pneumatic interference to the ambient pressure system in borehole NRG#4 on June 16, 1995. The pneumatic interference first occurred when the tunnel boring machine was excavating in the pre-Pah Canyon Tuff (Tpbt2) and the crystal-rich vitric zone at the base of the PTn was still undisturbed. The pneumatic-interference effects observed at NRG#4 are the result of the removal of the PTn pneumatic impeding layer, thus exposing the TSw units directly to the atmospheric pressure signal.

Pneumatic-interferenceeffects at UZ#4 and UZ#5 were first observed on August 12, 1995 when the face of the Exploratory Studies Facility North Ramp tunnel was approximately 18 m east of the intersection with a north-to-northwest-trending, west dipping, normal fault (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The pneumatic interference observed at UZ#4 and UZ#5 on August 12, 1995 was transmitted along this fault zone over a distance of 377 to 415 m (see Figure 5.3-8). The pneumatic-pressure responses in all of the PTn and Topopah Spring Tuff instrument stations in these two boreholes were effected. Pressure records and results of cross-spectral analyses after the effects

of Exploratory Studies Facility excavation for borehole UZ#4 (Figure 5.3-126) indicate that residual amplitudes at stations within the Yucca Mountain and Pah Canyon Tuffs increased by about 40 percentage points while those from stations in the Topopah Spring crystal-rich vitrophyre and crystal-rich nonlithophysal units increased by about 60 percentage points (compare Figure 5.3-121 with Figure 5.3-126). The lower stations displayed larger residual amplitudes and shorter phase lags after the Exploratory Studies Facility excavation effects because the barometric signal was being transmitted from below the PTn by way of the open Exploratory Studies Facility tunnel.

The first indication of a pneumatic interference in the pressure record for borehole NRG#5 occurred on September 14, 1995 when the tunnel boring machine had advanced to approximately 93 m from the borehole and in the lower section of the Topopah Spring crystal-rich nonlithophysal unit (Patterson et al. 1996). Pneumatic-interference effects were recorded in all of the PTn and TSw instrument stations in this borehole. Pressure records and the results of cross-spectral analyses after the effects of Exploratory Studies Facility excavation for borehole NRG#5 indicate that residual amplitudes in zones above the PTn were not significantly effected by Exploratory Studies Facility excavation. However, residual amplitudes at stations 7 through 9 in the TSw (Figure 5.3-127) increased by 40 to 60 percentage points after the effects of Exploratory Studies Facility excavation (compare Figure 5.3-124 with Figure 5.3-127). Interestingly, the residual amplitude at station 10 near the bottom of the upper lithophysal unit, previously discussed as being effected by the Drill Hole Wash fault, increased only 9 percentage points. The smaller increase in residual amplitude at station 10 after the effects of Exploratory Studies Facility excavation corroborates the earlier conclusion that the barometric pressure signal was bypassing the PTn by traveling preferentially down the Drill Hole Wash fault. Further, the fact that the effects of Exploratory Studies Facility excavation propagated downward to station 10 suggests that the overlying stations (6 through 9) in the TSw also probably were effected by the Drill Hole Wash fault and without its influence may have had residual amplitudes about 10 percentage points lower.

Although not recognized immediately, pneumatic-interference effects began to occur at borehole NRG-6 on October 1, 1995, approximately 2 days after the tunnel boring machine intersected the northern boundary of the Drill Hole Wash fault structure (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Several months of pressure data were required to confirm that there was indeed a change in the amplitude and phase lag of the synoptic pressure signal in the borehole NRG-6 pressure records for the TSw instrument stations. Numerical model simulations using the site-scale unsaturated-zone flow model (Bodvarsson and Bandurraga 1996) were able to help confirm and isolate the approximate date that these changes first occurred. The numerical model simulation results corroborated the results of the post-tunnel boring machine interference cross-spectral analysis of the pneumatic-pressure record at borehole NRG-6 (Figure 5.3-128), which showed that residual amplitudes in all stations in the TSw increased by about 20 percentage points (compare Figure 5.3-120 with Figure 5.3-128). Borehole NRG-6 is located south of the inferred boundaries of the Drill Hole Wash fault. It is likely that the weak pressure interference seen at borehole NRG-6 is a reflection of pressure changes that are propagated both along, and perpendicular to, the alignment of this fault. It is clear from these data that faults in Drill Hole Wash are open in the Topopah Spring Tuff, and the relatively high-residual amplitudes in the Topopah Spring Tuff stations prior to

Exploratory Studies Facility effects probably were the result of the surface barometric signal being transmitted preferentially down through the Drill Hole Wash fault.

At borehole NRG-7a, pneumatic-interferenceeffects were first observed on October 21, 1995, 5 days after the tunnel boring machine crossed the southern boundary of the northwest-trending Drill Hole Wash fault (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The pneumatic interference was not detected until the tunnel boring machine had advanced to within 26 m of the borehole. The sequence of events leading up to the first-observed pneumatic interference at NRG-7a indicate that secondary, near-field fracturing associated with the Drill Hole Wash fault apparently is very limited immediately south of where the Exploratory Studies Facility North Ramp tunnel crosses the Drill Hole Wash fault. However, it also should be noted that the rate of advance of the tunnel boring machine, approximately 22 m/day, may have obscured earlier detection of the first arrival of the pneumatic-interference effect. Pressure records and the results of cross-spectral analysis for borehole NRG-7a after the effects of Exploratory Studies Facility excavation (not shown) indicate that residual amplitudes of all zones within the Topopah Spring Tuff increased 40 to 60 percentage points and the phase lags decreased by about 50 hours, very similar to the changes in attenuation observed at borehole NRG#5.

At borehole SD-9, pneumatic-interference effects first occurred on November 7, 1995, when the residual amplitude in the upper zone increased about 20 percentage points and the phase lag decreased from about 100 hours to about 30 hours. The tunnel boring machine was approximately 184 m distant from SD-9 in the Topopah Spring middle nonlithophysal unit (Tptpmn), but no known fault or fracture-zone features have been associated with this location. Therefore, the first occurrence of pneumatic-interference effects at borehole SD-9 was a function only of the bulk horizontal permeability of the Tptpmn. The lower monitored zone in borehole SD-9 (below the perched-water zone) continued to be unaffected by Exploratory Studies Facility excavation.

At borehole SD-12, pneumatic-interference effects were recognizable in the pressure records beginning on March 23, 1996 when the tunnel boring machine was only 49 m from the borehole, which is surprising given that the tunnel boring machine was within a zone of steeply dipping, intense fractures (Patterson et al. 1996). This fracture zone is coincident with the location of maximum displacement and deformation of the nearby Ghost Dance fault, but is not considered to be part of the Ghost Dance fault. Because of the high-presumed air permeability of the fracture zone, interference effects were expected to occur much sooner and, in fact, there is evidence that the earliest effects were subtle, and not visually identifiable. The fact that the large fracture zone would have large storage capability may have caused the pneumatic effects from the Exploratory Studies Facility excavation to occur gradually in borehole SD-12 and mask the initial onset until the effects were large enough to be visually identified in the record. Pressure records and the results of cross-spectral analysis for borehole SD-12 after the effects of Exploratory Studies Facility excavation (Figure 5.3-129) indicate that residual amplitudes of all stations in the TSw increased about 20 percentage points (compare Figure 5.3-123 with Figure 5.3-129). The pneumatic character of the two deepest stations (A and B) in borehole SD-12 were discussed in Subsection 5.3.4.2.1.1 in relation to the presence of a perched-water zone above these stations.

More recent analysis of pneumatic-pressure records suggests that interference effects from Exploratory Studies Facility excavation may have occurred at borehole SD-12 as early as February 26, 1996, when the tunnel boring machine was about 340 m from the borehole (Rousseau et al. 1997). Identification of this earlier date is based on differencing the pressure record for borehole SD-12 station G (Tptpmn), from the pressure records for borehole NRG-7a station A (Tptpul) and borehole UZ-7a station A (Tptpll). In using this cross-hole pressure-differencing approach, the assumption was made that the pressure responses to atmospheric-pressure changes in both UZ-7a and NRG-7a should be stable. Pressures at NRG-7a had already affected Exploratory Studies Facility interference, and pressures in UZ-7a were assumed to be immune to interference effects (at least prior to SD-12 being affected) because of the borehole's location with respect to SD-12 and its completion in the Ghost Dance fault zone where downhole pressures are dominated by the atmospheric pressure signal (see Subsection 5.3.4.2.1.1 and Figure 5.3-122). Differencing of the pressure data between the Tptpmn instrument station in SD-12 and the Tptpul instrument station in NRG-7a indicates progressive deterioration in the pressure-amplitude and phase-lag relationships beginning on or about February 26, reaching a maximum shortly before March 23, 1996 for borehole NRG-7a and shortly after that date for borehole UZ-7a. Prior to March 23, the signal at borehole SD-12 lagged that in NRG-7a. After March 23, the signal at SD-12 began to lead that in NRG-7a, producing a differential pressure-decay envelope (Rousseau et al. 1997).

At borehole SD-7, located about 500 m north of the intersection of the Exploratory Studies Facility Main Drift and the South Ramp, pneumatic-interference effects became apparent on June 5, 1996 and were evidenced by simultaneous initiation of downhole responses to the diurnal barometric signal (Patterson et al. 1996). Unlike any other borehole, the tunnel boring machine had already gone 30 to 40 m beyond the nearest approach of the Exploratory Studies Facility to borehole SD-7 when the pneumatic-interference effects first occurred. This may reflect a somewhat lower horizontal permeability in the rock mass between the Exploratory Studies Facility and the borehole such that the interference did not propagate horizontally until a more permeable zone was encountered. Pressure records and cross-spectral analysis for borehole SD-7 after the effects of Exploratory Studies Facility excavation (not shown) indicate that residual amplitudes of all zones within the TSw increased about 30 percentage points and the phase lags decreased by about 10 hours, very similar to the changes in attenuation observed at borehole NRG-6. Because all the instrument stations in the TSw responded simultaneously to barometric signals both prior to and after Exploratory Studies Facility interference effects, the vertical permeability of the TSw at the location of borehole SD-7 must be very large.

5.3.4.2.1.4 Determination of Horizontal Permeability and the Vertical Anisotropy of the Topopah Spring Tuff by Numerical Simulation of Pressure Responses to Exploratory Studies Facility Excavation

Pneumatic pressure responses to Exploratory Studies Facility excavation observed at boreholes NRG#5, NRG-6, and NRG-7a were used in the formulation of a Modflowp-based (Hill, M.C. 1992) numerical model to estimate the permeability to air of the PTn and TSw (Patterson et al. 1996). Particular emphasis was placed on determining the horizontal permeability and the ratio of horizontal to vertical permeability of the TSw because these properties cannot be estimated reliably from the monitoring data in the absence of Exploratory Studies Facility interference effects. Further, although

these properties were determined by air-injection testing (Subsection 5.3.3.1.2), the values obtained are for a much smaller scale than those that can be determined based on the larger scale pneumatic-interference effects caused by Exploratory Studies Facility excavation. Permeabilities determined from these simulations may be useful in more complete and larger scale models of the repository that would account for heat and moisture movement that would accompany waste emplacement.

Use of a groundwater-based flow code to simulate air flow through unsaturated, fractured porous media requires some explanation and much qualification (Patterson et al. 1996). Groundwater flow codes typically are formulated assuming that the porous medium is saturated with a single slightly compressible homogeneous fluid. The unsaturated rocks being simulated, on the other hand, are partly saturated with water, and the fluid of interest is highly compressible, resulting in the governing flow equation being nonlinear. However, water movement through the rocks comprising Yucca Mountain generally is slow and diffuse, resulting in liquid saturations that vary little on the time scale of the pneumatic-interference effects being simulated. Therefore, the air-filled porosity can be considered constant with time. Water movement also occurs as intermittent, rapid-focused percolation that bypasses the main volume (rock matrix) of the unsaturated zone. Such recharge may affect saturations and air-filled porosity briefly immediately after infiltration events, but probably would not affect air-permeability determinations based on longer term (monthly) data. Airflow simulations may be made using the linear groundwater flow equation if the pressure variations about the mean pressure are no more than about 10 percent (Weeks 1978). For naturally induced pressure-potential differences arising from atmospheric-pressure fluctuations, temperature-induced density effects, and wind pumping, pressure fluctuations typically are no more than about one percent of the mean pressure, ensuring that the effects of nonlinearity in the gas-flow equation arising from high gas compressibility will be quite minimal.

The appropriateness of using a porous-media-based flow model to simulate fluid movement through fractured porous media also must be considered, particularly in regard to whether the unit is sufficiently fractured that it may be modeled as an equivalent porous medium (Patterson et al. 1996). Such appropriateness depends on whether the model scale is large relative to the representative elementary volume of the fracture-matrix continuum. In a system dominated by flow in discrete fractures, it would be anticipated that monitoring stations that did not intersect the fractures would be bypassed by barometrically induced flow from the Exploratory Studies Facility tunnel, and stations near the tunnel might show less response than those more distant but within the zone of influence of the fracture. Based on examination of data from monitoring boreholes NRG-6, NRG-7a, and NRG#5, deeper stations at borehole NRG#5 do show greater response than the shallower stations prior to tunnel excavation, illustrating this effect. However, most observed pressure responses, which have been made at distances of tens to hundreds of meters from the Exploratory Studies Facility, both vertically and horizontally, indicate that, in general, the responses decrease monotonically with radial distance from the Exploratory Studies Facility. Thus, the TSw must be extensively fractured, and an equivalent-porous-media model probably is adequate to examine pressure responses at a scale of tens to hundreds of meters. However, the TSw is variably fractured, as can be seen by exposures in the Exploratory Studies Facility, and the initial response of various boreholes appears highly dependent on the timing of the intersection of faults or of more broken rock by the Exploratory Studies Facility.

Faults at Yucca Mountain seem to have properties that cause them to function both as highly permeable conduits and as low-permeability barriers (see Subsections 5.3.4.2.1.1 and 5.3.4.2.1.3). Presumably, the highly fractured breccia zones are very permeable, whereas fault gouge along the base of the faults seems to create low-permeability barriers to flow (Patterson et al. 1996). These properties perhaps are illustrated best by the residual amplitudes for borehole UZ-7a prior to the effects of Exploratory Studies Facility excavation (Figure 5.3-122). Monitoring stations D through F, extending through the PTn and well into the Topopah Spring Tuff, exhibit large and nearly uniform residual amplitudes, indicating that even the PTn is highly fractured at this location. However, monitoring stations A through C exhibit much reduced residual amplitudes that are nearly constant among these stations. Thus, the lower stations are separated from the highly permeable breccia zone by a low-permeability layer. These features likely are nearly vertical and extend through the entire thickness of the TSw. However, the preliminary conceptual model for equivalent porous medium applied here assumes that the media properties of both the PTn and the Topopah Spring Tuff are areally uniform, and neither the conduit effects of breccia zones nor the barrier effects of gouge planes are simulated. Thus, the identified permeability of fractured rock surrounding the monitoring boreholes incorporates the effects of the breccia zones as larger block permeability and the effects of gouge as smaller block permeability. Therefore, use of the bulk permeability values determined below must be used with caution and the full recognition that simulations based on these values might poorly represent the system.

Model Structure–The Modflow model of pneumatic-pressure interference effects for the Exploratory Studies Facility North Ramp area was formulated to include only the PTn and the TSw (Patterson et al. 1996). Based on the pneumatic-pressure data collected in several instrumented boreholes (Subsection 5.3.4.2.1.1), it was concluded that the TCw is so permeable that atmospheric-pressure changes were assumed to be transmitted instantaneously to the top of the PTn throughout the model area. This assumption allows substantial simplification of the model representation, as the complex topography representing the Tiva Canyon Tuff outcrop can be ignored. The Calico Hills nonwelded hydrogeologic unit (CHn), which immediately underlies the TSw, was assumed to be impermeable to air, based on the occurrence of perched water in either the CHn or in the basal vitrophyre of the TSw throughout much of the model area. The modeled region was bordered on the northeast by a 1,780-m section of the Exploratory Studies Facility North Ramp extending northwestward from where the Exploratory Studies Facility intersects the Bow Ridge fault. The model area extended 2,100 m southwestward from the Exploratory Studies Facility and was 385 m thick. The upper boundary of the model was a specified-pressure boundary to simulate time-variant atmospheric pressure. All other boundaries were no-flow boundaries, except the boundary representing the Exploratory Studies Facility tunnel, which was a general-head boundary where pressure potential always was specified the same as that of the upper boundary of the model. Based on evaluation of pressure and temperature data collected in the Exploratory Studies Facility, pressure effects from the Exploratory Studies Facility mechanical ventilation system were determined to be negligible and, therefore, were ignored in the model. Additional details of model construction are given in Patterson et al. (1996).

Model Input Parameters and Initial Conditions–Properties equivalent to transmissivity and storage coefficient for a saturated groundwater system were assigned to the gas-flow simulation using intrinsic permeability, thickness, relative permeability to air, dynamic viscosity, air-filled

porosity, and mean atmospheric pressure (Patterson et al. 1996). Because matrix porosity generally is large relative to fracture porosity, matrix porosity was assumed equal to total porosity for this model. Data available for parameter estimation in the model included time series of pressure potentials observed for various monitoring stations in boreholes NRG-6, NRG-7a, and NRG#5. To simplify these records, data used in the model were based on the sine waves that best fit the synoptic atmospheric changes. Pressures at the top of the PTn and within the Exploratory Studies Facility tunnel were represented as a sine wave of unit amplitude and no phase lag. Pressure potentials at each monitoring station are represented by sine waves of equal period, but having an amplitude and phase lag equal to that determined for the borehole based on cross-spectral analysis (see Subsection 5.3.4.2.1.1). Simulation of air flow due to atmospheric pressure change required frequent updating of pressure potentials along the specified-pressure and general head boundaries. Initial conditions representing effects of atmospheric pressure fluctuations were complicated by the fact that atmospheric pressure was never constant for a long-enough period that pressure differences with depth disappear. Therefore, a simulated cyclic steady state was established prior to the time interval over which bulk-rock properties were determined using the parameter-estimation simulation.

Parameter Estimation for Boreholes NRG#5, NRG-6, and NRG-7a–Several attempts were made at parameter estimation using the Modflow gas-flow model of the Exploratory Studies Facility North Ramp area before convergence to stable parameters was obtained (Patterson et al. 1996). Early simulations produced satisfactory results for the permeability of the PTn, which was determined to be isotropic with respect to permeability, but failed to produce satisfactory results for TSw due to problems associated with the fact that the three boreholes (NRG#5, NRG-6, and NRG-7a) only partially penetrate the TSw. This problem was overcome by recognizing that borehole NRG-6 was far enough away from the Exploratory Studies Facility such that the simulation of pneumatic-interference effects at NRG-6 would not be affected by partial penetration. Once this problem was recognized and solved, a series of simulations were made in which the horizontal and vertical permeability of the TSw and the isotropic permeability of the PTn were determined, using data from borehole NRG-6 alone and in conjunction with data from borehole NRG#5 and borehole NRG-7a separately, and boreholes NRG#5 and NRG-7a together (Table 5.3-46).

These simulations resulted in permeability values for the PTn that were about one-half the average value obtained from the one-dimensional analyses of diffusivity by Patterson et al. (1996) (see Subsection 5.3.4.2.1.2). For the TSw, vertical-permeability values obtained from the parameter-estimation simulations also were about one-half the average value obtained from the one-dimensional analyses of diffusivity by Patterson et al. (1996) and were closer to the values obtained using other methods. The simulations also confirmed that the ratio of vertical to horizontal permeability in the TSw is at least 2:1.

To help validate the Modflow gas-flow model of the Exploratory Studies Facility North Ramp area, a series of model simulations were made to test whether the model could match the times at which pneumatic-interference effects were first observed in various boreholes (Patterson et al. 1996). In these simulations, the Exploratory Studies Facility tunnel was progressively advanced on a weekly basis from July 5 to November 15, 1995. Model results indicate that simulated responses for boreholes NRG#5 and NRG-6 began to exhibit subtle interference effects on about September 5, 1995. A significant increase in residual amplitude of the synoptic-pressure signal was simulated for

borehole NRG#5 between September 12 and 15, which is in agreement with the first occurrence of interference effects based on evaluation of the pressure records (Table 5.3-45). For borehole NRG-6, model results indicated that the residual amplitude of the synoptic-pressure signal increased quite slowly throughout the simulation period, with no clear-cut increase attributable to a given date. Nonetheless, the increase in amplitude between September 26 and 30 is somewhat larger than those for previous periods, and agrees reasonably well with the first occurrence of interference effects based on evaluation of the pressure records (Table 5.3-45). The simulated residual amplitudes in borehole NRG-7a exhibited dramatic increases between October 17 and 20, slightly sooner than the October 21 date identified based on the pressure records (Table 5.3-45). Overall, the first occurrence of pneumatic-interferenceeffects identified in Table 5.3-45 agree well with simulated results based on model-identified properties.

5.3.4.2.1.5 Parameter Estimation for Boreholes UZ#4 and UZ#5

A numerical model very similar to the one described in Subsection 5.3.4.2.1.4 was constructed for the vicinity of boreholes UZ#4 and UZ#5 in Pagany Wash about 0.5 km northeast of the Exploratory Studies Facility North Ramp (Patterson et al. 1996). The observed pneumatic-interference effects in boreholes UZ#4 (Figures 5.3-121 and 5.3-126) and UZ#5 were simulated for each borehole separately and for the two boreholes combined, resulting in the permeability values listed in Table 5.3-47. Permeability values for the TSw are very large, and likely represent the effects of the fault zone extending from the Exploratory Studies Facility northward to the near vicinity of boreholes UZ#4 and UZ#5 (see Subsection 5.3.4.2.1.3). In regard to whether the model reasonably represents the fault, it can be argued that, with the tunnel extending the length of the model, the model is in essence a cross-sectional model, particularly if the breccia zone is highly permeable and the slip plane of the fault is a low-permeability barrier. In these circumstances, the pressure response propagated along the highly permeable zone would be isolated from the less fractured TSw to the west of the fault trace. Hence, the model probably does reasonably represent the system. The large difference between permeabilities determined for the two boreholes could be accounted for by relatively small differences in width of more typical TSw rock separating the fault zone from the boreholes, and/or in differences in the effectiveness of the small-permeability barrier created by the fault.

These model results for borehole UZ#4 and UZ#5 are contradictory to the separate simulations for boreholes NRG-5, NRG-6, and NRG-7a (Subsection 5.3.4.2.1.4) and illustrate some of the problems encountered in using a porous-media model to simulate effects of major discrete features that have quite different permeabilities than those of the rest of the system (Patterson et al. 1996). The results indicate that a local portion of the repository block might be satisfactorily simulated by a porous-media model when data are available for calibration, but the use of parameters identified for the local system might provide quite erroneous results if applied to the repository block as a whole. For example, the good agreement between simulated and observed results for data from boreholes NRG#5, NRG-6, and NRG-7a indicates that the presence of highly conductive faults and of partial flow barriers that divide the Topopah Spring Tuff into more permeable compartments are not necessarily important for describing the overall behavior of air movement through that part of Yucca Mountain immediately south of the Exploratory Studies Facility North Ramp. However, the permeability values identified for these rocks would completely fail to simulate the Exploratory

Studies Facility-induced pneumatic-interference effects determined for boreholes UZ#4 and UZ#5. Conversely, results of parameter identification for these boreholes would be totally inappropriate for the area south of the Exploratory Studies Facility North Ramp.

5.3.4.2.2 Distribution of In Situ Water-Potential and Occurrences of Perched Water in the Deep Unsaturated Zone

A conceptual model of fluid flow in the unsaturated zone at Yucca Mountain was developed during the early stages of the project (Montazer and Wilson 1984) based on very limited hydrologic data (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The conceptual model suggested that the vertical matrix-flux capacity of the porous and sparsely fractured rocks of the PTn was much greater than the vertical matrix-to-matrix flux capacity of the underlying, low-porosity, densely welded, but intensely fractured rocks of the TSw. Vertical flux through the CHn matrix was postulated to be very low. In the 1984 conceptual model, redistribution of vertical flux in the PTn as downdip lateral flow within and/or above the upper contact of this unit with the overlying TCw, with subsequent drainage into bounding structural features, was presented as the most likely mechanism to account for mass-balance deficits in the vertical matrix-flux profile. Dipping beds, permeability layering, and capillary-barrier effects were envisioned as the combined hydrogeologic features and hydrologic processes needed to promote lateral flow in the PTn. Thus, most of the flux through the PTn would find its way into the deeper subsurface along major structural pathways through downdip lateral flow, interception, and diversion. One-dimensional, vertical, matrix-to-fracture flow between the PTn and the TSw was presumed to be limited because of the capillary-barrier retardation effects between the smaller pores of the high-porosity matrix of the PTn overlying the larger fracture apertures of the densely welded TSw. According to the 1984 model, the extent of deep fracture flow within the TSw also was presumed to be limited because of imbibition of fracture water into the surrounding matrix. A maximum upper limit of 1 mm/year was proposed for saturated-matrix flux in the Topopah Spring Tuff under an assumed net infiltration rate of 4.5 mm/year. Net infiltration in excess of the 1 mm/year estimated to flow through the TSw matrix was presumed to move laterally either within or directly above the PTn toward structural features that would then intercept and transmit this flow downward to the base of the TSw. Here, the water either would enter the CHn along the same but less permeable structural flowpaths, or would accumulate as perched water and then move downward into the CHn as matrix percolation. The 1984 conceptual model heavily emphasized lateral flow in the PTn, with interception and transmission of this flow along major structural pathways.

Investigations conducted since 1984 tend to support many of the key attributes of the 1984 conceptual model for unsaturated flow at Yucca Mountain including low-net infiltration due to the arid climate, rapid infiltration into the fractures of the TCw, unit hydraulic gradients within the PTn and TSw, and perched water at the base of the TSw (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). However, direct evidence to document extensive lateral flow in the PTn and diversion of flow into structural pathways has been elusive. Nevertheless, there is very strong evidence for significant fracture flow

within the PTn locally where secondary fracture permeability, in association with mapped faults, is evident. Fracture flow also is evident below the PTn in the highly fractured rocks of the TSw, a flow mechanism that was not strongly emphasized in the 1984 conceptual model. Furthermore, perched water has been encountered at the base of the TSw or in the top of the CHn in nearly every borehole that was dry-drilled deep enough to penetrate the TSw-CHn contact in the site area.

5.3.4.2.2.1 In Situ Water Potentials

In situ water potential has been monitored at multiple depths in instrumented boreholes NRG-6, NRG-7a, UZ#4, UZ#5, UZ-7a, and SD-12 (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) using thermocouple psychrometers. Water-potential data were collected from these boreholes under a fully compliant quality assurance program for various periods of time from November 1994 through the present (see Subsection 5.3.2.1.4). In addition, non-qualified water-potential data were collected from borehole UZ-1 from October 1983 through October 1985 using a prototype borehole-instrumentation system.

Although the water-potential data from borehole UZ-1 are non-qualified, these data provide important insights regarding the changes and stabilization of in situ water potential following instrumentation and stemming of deep boreholes in the unsaturated zone (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The water-potential profiles for borehole UZ-1 for elapsed times following instrumentation of 30, 90, 230, 365, and 730 days (Figure 5.3-130) indicated that the rock surrounding most instrument stations still was becoming wetter even after 2 years. The recovery time for water potentials after the borehole dried out prior to stemming was much longer than for any other boreholes instrumented at Yucca Mountain, probably because of the large borehole diameter (92 cm from 12 m to 30 m in depth, 45 cm from 30 m to total depth) and the type of stemming materials used.

Water potentials at instrument stations 12 and 13 located in the lower lithophysal unit of the Topopah Spring Tuff may not have achieved equilibrium even after 2 years of monitoring (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Core fracture data from borehole UZ-14, located about 27 m northwest of UZ-1, indicate that the fracture density of the lower lithophysal unit at this location is very low (see Figure 5.3-143 in Subsection 5.3.4.2.2.2), which may account for the very slow rates of recovery of water potentials at these two instrument stations. Relatively slow rates of recovery also are apparent in the time-lapse profiles for instrument stations 3, 5, 6, 9, and 10, which also are located in lithostratigraphic units with low-fracture densities. The more rapid recovery of water potentials at instrument stations 4, 7, 11, 14, and 15 correlate with zones of high-fracture density in nearby borehole UZ-14. Water-potential data from instrument stations 1 and 2 may have been affected by water that was used to drill the first 18 m of the borehole and may not be reliable. Slow recovery at instrument station 3 probably is a reflection of the large diameter of the borehole at this

location, which required a large volume of dry stemming material to back-fill. The resultant water-potential profile for October 25, 1985, indicates that water-potential gradients in all TSw units, with the possible exception of the lower lithophysal unit, are very close to zero. Extrapolation of the observed trend in water potentials at stations 12 and 13 from October 1984 to October 1985 indicates that the water-potential gradient throughout the Topopah Spring Tuff is close to zero. Thus, the gravitational potential may be the dominant component of the total hydraulic gradient, which is very close to unity.

Boreholes NRG-7a and NRG-6 were instrumented in October and November 1994, respectively (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The diameters of these boreholes averaged approximately 25 cm, with large breakouts of 30 to 45 cm occurring across the Pah Canyon Tuff and crystal-rich vitric subunits at the base of the PTn. Selected temperature and water-potential profiles for these two boreholes are shown in Figures 5.3-131 and 5.3-132. For borehole NRG-7a (Figure 5.3-131), the elapsed time since instrumentation are about 4, 7, and 13 months. For borehole NRG-6 (Figure 5.3-132), the elapsed time since instrumentation are about 3, 6, and 12 months. Although temperatures at all but the shallowest stations reach equilibrium in 3 to 4 months, water potentials do not appear to approach equilibrium until at least a year after instrumentation. To investigate water-potential equilibrium further, Figures 5.3-133 and 5.3-134 show 100-day, time-series records of water potential in these two boreholes from November 1, 1995 to February 8, 1996, which is a period that begins about a year after the boreholes were instrumented. For station B in NRG-7a (Figure 5.3-133) and station A in NRG-6 (Figure 5.3-134), water potentials appear to be slightly above zero because the calibration accuracy of these sensors was on the order of ±0.2 MPa.

The time-series water-potential data for borehole NRG-6 for November 1, 1995 to February 8, 1996 (Figure 5.3-134) indicate that water potentials in nearly all the instrument stations had equilibrated after about a year of monitoring (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). During this period, however, very small increases in water potential continued at instrument stations B, D, and E (less than about 0.0005 MPa per day based on cumulative changes over a 2,400-hour period). The profile of water potential with depth for November 14, 1995 (Figure 5.3-132) indicates a water-potential gradient that is very close to zero across the PTn and through the upper portion of the TSw.

The time-series water-potential data for borehole NRG-7a (Figure 5.3-133) indicate that water potentials after November 1, 1995 were still increasing at a rate slightly greater than 0.0005 MPa per day at all but one of the deeper instrument stations (station B) in this borehole (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The water potential at instrument station E, which is located at a depth of about 5 m, is strongly affected by seasonal temperature changes at the land surface and the movement of air through the TCw. The November 1, 1995 water-potential profile with depth for NRG-7a (Figure 5.3-131) indicates a water-potential gradient that is very close to zero across the entire instrumented section of this borehole, including the TCw, PTn, and TSw.

Despite the apparent stability and uniformity of the water-potential profile in borehole NRG-7a, transient moisture-redistribution events apparently do occur sporadically. For example, on May 9, 1996 water potentials and temperatures at station D, located in the Yucca Mountain Tuff (Tpy) of the PTn, began to oscillate in a manner that was completely unlike any previous activity observed at this instrument station (Rousseau et al. 1997). These oscillations were manifested by large shifts in water potential from −0.2 MPa (−2 bars) to less than −7.5 MPa (−75 bars), with concurrent shifts in temperature. These oscillations followed a prolonged period (18 months) of slow, asymptotic, and uneventful recovery in water potential and temperature. The largest water-potential change was a decrease greater than 7.5 MPa (75 bars) and was accompanied by a decrease in temperature of 0.68°C. These oscillations lasted for over a month, with at least 13 separate episodes recorded between May 9 and June 14, 1996 (Rousseau et al. 1997, Figure 4.4-1). Each temperature and water-potential excursion was followed by a short but truncated period of recovery. This unusual activity ceased almost as abruptly as it started. From June 14, 1996 to February 14, 1997 water-potentials and temperatures in station D have been slowly recovering to their pre-disturbed state. On February 14, 1997 and March 4, 1997 two less severe temperature and water-potential excursions were recorded at this instrument station.

Although barometric pumping had been observed in the water potential record for this station in borehole NGR-7a (similar to that observed in borehole UZ-7a), during the period of extreme excursions in May and June 1996 there was no obvious correlation between surface-pressure changes and downhole temperature and water potentials changes. In fact, falling temperatures were accompanied by decreasing water potentials, which is a complete reversal of the normal response. Instrument malfunction and station leakage have been completely ruled out as possible explanations for these excursions. Further, there appears to be no correlation between the anomalous behavior of station D and construction activities in the Exploratory Studies Facility. Although analysis of the pneumatic-pressure record indicated very significant changes in the residual amplitude and phase lag of the pressure signal from station D after excavation of the Exploratory Studies Facility (see Subsection 5.3.4.2.1), those changes occurred in October 1995. Overall, the data indicate that the water-potential and temperature disturbances recorded in station D originated from above the instrument station and probably were due to enhanced communication between this station and the atmosphere. The nearly instantaneous nature of the episodic disturbances and the slow, but truncated recovery following each episode suggest nearly complete replacement of the gas in the instrument station. Moisture redistribution in the fractures of the overlying TCw, accompanied by topographically induced density-driven gas flow, may be the cause of the unusual water-potential responses observed at station D in borehole NRG-7a.

Water-potential profiles for September 26 and December 25, 1995 for borehole UZ#4 and UZ#5 are shown in Figure 5.3-135 for elapsed-time periods of about 3 and 6 months following instrumentation (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Plotted with the two sets of in situ water-potential measurements are measurements of water potential made on core samples shortly after the boreholes were drilled in 1984 (Loskot and Hammermeister 1992). Shortly before being instrumented in June and July 1995, these boreholes were reamed to a diameter of 22 cm. Borehole

UZ#4 is located in the active channel of Pagany Wash, and UZ#5 is located on a sideslope adjacent to Pagany Wash, approximately 38 m south of UZ#4.

At several stations in UZ#4, there is little or no difference between water potentials measured on September 26, 1995 and those measured on December 25, 1995 (Figure 5.3-135) (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). This indicates that water potentials equilibrated fairly rapidly at instrument station A, located at the top of the crystal-rich nonlithophysal unit of the Topopah Spring Tuff, and at instrument stations E, F, and G, located in the upper units of the PTn where water potentials are between -0.1 and −0.25 MPa (−1 to −2.5 bars). Time-series water-potential records for UZ#4 (Figure 5.3-136) indicate that water potentials at three instrument stations in the lower units of the PTn (B, C, and D) were still increasing slightly after December 25, 1995. The increasing trend at these stations indicates that the equilibrated profile will be very close to that derived from the core samples and that the resulting profile will very closely approximate a zero water-potential gradient across most of the PTn section at this borehole. This conclusion is consistent with the results of the comparison of field-scale and core-scale water potentials and water-potential gradients described in J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). There is some indication that the December 25, 1995 water-potential gradient (Figure 5.3-135) across the pre-Yucca Mountain Tuff unit (Tpbt3) in UZ#4 may be preserved after all instrument stations have fully equilibrated. However, accuracy limitations of the psychrometers (± 0.2 MPa) render this conclusion somewhat uncertain.

The time-series water-potential records for borehole UZ#5 (Figure 5.3-137) indicate that water potentials at all instrument stations were still increasing slightly after December 25, 1995. Overall, therefore, equilibration rates in borehole UZ#5 are slower than for UZ#4. The water-potential profile for UZ#5 as of December 25, 1995 (Figure 5.3-135) indicates that the water-potential gradient at this borehole also is very nearly zero across the entire PTn section. Water potentials in the upper PTn units (crystal-poor vitric, Yucca Mountain Tuff, and the pre-Yucca Mountain Tuff bedded tuff) are 0.2 to 0.3 MPa (2 to 3 bars) lower than water potentials of the same units at UZ#4. These data may indicate lateral flow from the vicinity of borehole UZ#4 toward borehole UZ#5, which would be from the channel toward the sideslope. Although this interpretation is subject to limitations in accuracy (±0.2 MPa) and uncertainty over final equilibration values, it is consistent with the distribution of tritium found in pore water from the cores of boreholes UZ#4 and UZ#5 (Yang et al. 1996; J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) and with heat-flow data from the same boreholes (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164).

Borehole SD-12, located along the Exploratory Studies Facility Main Drift, is the deepest borehole instrumented in the unsaturated zone at Yucca Mountain and is the only borehole that has instrument stations in all four major hydrogeologic units: TCw, PTn, TSw, and CHn. (See Figures 5.3-14 and

-15 for locations of instrument stations.) In borehole SD-12, water-potential recovery trends across the TSw and CHn tend to follow distinctive patterns during the first 6 months after instrumentation (Rousseau et al. 1997). Water potentials at instrument stations located at the base (station C) and below (stations A and B) the lower nonlithophysal unit (Tptpln) tend to follow a pattern of monotonically increasing water potentials (Figure 5.3-138). In contrast, water potentials at stations located at the top and above the lower nonlithophysal unit (Tptpln; stations D, E, F, G, H, and J) tend to follow a pattern of initially decreasing followed by increasing water potentials (Figures 5.3-138 and -139). Water potentials at stations G and H, located in the intensely fractured middle nonlithophysal unit of the Topopah Spring Tuff (Tptpmn), recovered very quickly, suggesting enhanced recovery attributable to a dense network of interconnected fractures. Very early-time decline in water potentials probably was an artifact of the effects of the dry stemming materials used to backfill the instrument stations (polyethylene beads and 20/40 silica sand). In contrast to the other stations in the TSw, water-potential recovery for station K, located in the TSw crystal-rich nonlithophysal unit (Tptrn), was somewhat slower and this interval started out much drier than the other stations with a water potential of less than −2.0 MPa (−20 bars). The initial dryness at this station and the slower water-potential recovery probably reflects the effects of high-matrix porosity (12 to 20 percent) and low-matrix-saturation(35 to 60 percent) of this unit (see Rousseau et al. 1997, Figure 4.2.1-9). Despite differences in water-potential recovery patterns during the first 6 months after instrumentation, all of these stations in borehole SD-12 (A-K) appear to be close to equilibrium and have recovered to water potentials within the fairly narrow range of −0.5 to −0.3 MPa (−5 to -3 bars). Overall, the recovery of water potentials in the TSw at borehole SD-12 resulted in a moisture regime quite similar to that in other boreholes such as UZ-1, NRG-6, and NRG-7a, although possibly a little drier.

Time-series water-potential records for borehole SD-12 at instrument stations within the PTn and TCw (Figure 5.3-140) contrast markedly with the water-potential records for deeper stations (Rousseau et al. 1997). The dynamics of barometric pumping are readily apparent in the water-potential record of station L, located near the base of the nonwelded PTn (upper Tptrv3, Tpbt2, and lower Tpp). The small-amplitude oscillations superimposed on the water-potential recovery curve of this station reflect barometric pressure changes that generate downward and upward air movement across sections where there is a shift in the temperature gradient and/or a significant difference in the vapor-concentration gradient (as determined from temperature, pressure, and water-potential measurements). Typically, rising pressures are accompanied by decreasing temperatures (flow from above) and increasing water potentials (effect of the lower temperatures on relative humidity even though the vapor concentration of the inflow gas is lower). Conversely, falling pressures are accompanied by increasing temperatures and decreasing water potentials. This barometric-pumping effect is strongest early during the recovery period and diminishes over time, but never dissipates completely as water-potential equilibrium is reached at about −0.5 MPa (−5 bars). The relative dryness of the TCw is reflected in the water-potential recovery trend for station P, which recovers to near equilibrium with a water potential of less than −1.5 MPa (−15 bars).

Borehole UZ-7a was drilled in the Ghost Dance fault zone about 200 m east of the Exploratory Studies Facility Main Drift near the southern end of the potential repository area (Figure 5.3-8). The borehole intersected the easternmost trace (main trace) of the Ghost Dance fault zone at a depth of about 120 m within the TSw crystal-poor, upper lithophysal unit (Tptpul) very near the contact of

this unit with the TSw crystal-rich, upper lithophysal unit (Tptprl) (Rousseau et al. 1997). Total displacement across the fault zone has been estimated at about 24 m (80 feet).

The time-series water-potential records for borehole UZ-7a (Figures 5.3-141 and 5.3-142) indicate that the recovery period for the deepest station (A) is considerably longer than that of the other two instrument stations (B and C) located below the easternmost trace of the Ghost Dance fault zone. The initial water potential at station A (−2 MPa, −20 bars) also is lower than the initial water potentials at stations B and C (about −0.5 MPa, −5 bars). The lower initial water potential and longer recovery period probably reflects the influence of higher bulk porosity (lithophysal porosity) and lower fracture density of the TSw lower lithophysal unit (Tptpln). Very little change in water potentials occurred at stations B and C during the first 10 months after the borehole was instrumented in October 1995. Much lower initial water potentials and much slower water-potential recovery rates are apparent in the time-series records for all instrument stations located above the easternmost trace of the Ghost Dance fault zone (stations E, F, G, H, I, and J). In addition, many of these stations exhibit water-potential oscillations that are directly related to surface barometric-pressure changes as described above for station L in borehole SD-12. Water-potential oscillations in station J, located at a depth of 13.5 m in the TCw lower lithophysal unit (Tpcpll) and within the zone of influence of the annual temperature wave may increase or decrease in response to rising barometric-pressure changes depending on the time of the year and the phase of the temperature cycle.

Water-potential oscillations are conspicuously absent in the water-potential records of the three lower stations (A, B, and C) located below the easternmost trace of the Ghost Dance fault zone but are very pronounced in station D, located immediately above the fault trace in borehole UZ-7a (Rousseau et al. 1997). These observations and the analysis of pneumatic pressures in borehole UZ-7a prior to Exploratory Studies Facility excavation (Figure 5.3-122) indicate that the pneumatic characteristics of the TSw in the fault zone are much different than elsewhere at Yucca Mountain. Specifically, within the Ghost Dance fault zone, the TSw is much less isolated from the atmosphere than elsewhere, probably due to higher fracture densities in the PTn. This enhanced connection to the atmosphere and the effects of more deeply penetrating barometric pumping result a moisture regime that has the potential to be more dynamic than elsewhere because it may be subject to both rapid, focused percolation of liquid water and the drying effects of barometric pumping.

Overall, the time-lapse in situ water-potential profiles indicate that water potentials are generally high, greater than about -0.3 MPa (-3 bars), and are nearly depth-invariant across the entire TSw (Rousseau et al. 1997). Although water potentials in the overlying PTn are slightly lower than in the TSw, the measured values in these two hydrogeologic units are considerably greater than those that would result if water potentials in the unsaturated zone were in a state of static equilibrium with the water table, implying that the downward percolation flux is non-zero. The relatively large water potentials also indicate that the rock matrix of the TSw is sufficiently wet, at least near fractures, that should water flow through the fractures, it would undergo little imbibition by the rock matrix, due to small matrix permeability and small water-potential gradients across the fracture-matrix interface. Across the TCw, water potentials are progressively lower with decreasing depth, as exemplified by the in situ water-potential profiles for boreholes UZ-7a and SD-12. These drier conditions indicate greater imbibition capacity associated with the TCw and upward matrix flow across this unit.

However, these observations do not exclude the possibility of episodic, downward fracture flow across this unit.

Water potentials measured for rock cores by L.E. Flint (1998) exhibit much greater variability than the in situ water-potential data from instrumented boreholes (Rousseau et al. 1997). Core-scale water-potential profiles for boreholes UZ-7a and SD-12 do not exhibit the depth-invariant characteristics of the in situ data. This is readily apparent in the core water-potential data for the Tiva Canyon lower lithophysal (Tpcpll), lower nonlithophysal (Tpcpln), and the Topopah Spring middle nonlithophysal (Tptpmn) units at UZ-7a and SD-12; and the Topopah Spring lower nonlithophysal (Tptpln) and crystal-poor densely welded vitric (Tptpv3) subunits at SD-12. Although corrections were applied to account for the effects of moisture losses during coring, processing, and handling operations (Flint, L.E. 1998), these corrections were not sufficient to reverse the apparent inversions in the water-potential profile across these units. Hydrogeologic units that display the most significant departure from the in situ measurements are those whose moisture-retention curves are the most sensitive to small changes in saturation (e.g., Tpcpll, Tpcpln, Tptpmn, Tptpln, and Tptpv3; Flint, L.E. 1998).

Effects of Exploratory Studies Facility Excavation on Pneumatic Properties and the Moisture Regime of the TSw–Exploratory Studies Facility-excavation-induced pneumatic-pressure changes in the TSw in borehole SD-12 were described in Subsection 5.3.4.2.1. Further analysis of the Exploratory Studies Facility interference effects suggests that they have been propagated differentially across the TSw (Rousseau et al. 1997). Residual amplitudes are larger and phase lags shorter below station G in borehole SD-12, the closest instrument station to the Main Drift, and are smaller and longer above station G. Increasing phase lags accompanied by decreasing amplitudes are consistent with a radially propagated disturbance. However, the data also imply that there may be differences in the pneumatic diffusivities of individual units comprising TSw. These differences appear to follow a trend that is correlated, at least qualitatively, with a trend of decreasing matrix porosity and increasing matrix saturation with increasing depth across the entire TSw. Matrix porosities and matrix saturations are higher above the middle nonlithophysal unit (Tptpmn) and lower below this unit. In addition, fracture densities are very high within both the middle nonlithophysal unit (Tptpmn) and the lower nonlithophysal unit (Tptpln) in borehole SD-12 (see Figure 5.3-146 in Subsection 5.3.4.2.2.2). Therefore, the pneumatic diffusivity of the TSw should be larger below the middle nonlithophysal unit (Tptpmn) because drained porosities are lower and fracture densities (and presumably air permeability) are higher.

Additional analysis of the pneumatic-pressure records for borehole SD-12 since the onset of Exploratory Studies Facility-excavation interference effects reveals that significant changes in the pneumatic properties have occurred in the TSw since the first quarter of 1996. Specifically, from April 1996 to January 1997, residual amplitudes decreased between 7 and 12 percentage points and phase lags increased between 3.7 an 4.1 hours with the larger adjustments occurring above the middle nonlithophysal unit (Tptpmn). These changes probably are directly related to drying out in the rock mass due to Exploratory Studies Facility excavation resulting in an increase in the drained porosities and a decrease in the pneumatic diffusivities of the affected units. The slow, prolonged adjustment of in situ pneumatic pressures to Exploratory Studies Facility interference effects indicates that in the vicinity of borehole SD-12 a considerable volume of gas was needed to stabilize

underground pressures. The pneumatic disturbances observed at borehole SD-12 suggest that the intensely fractured zone exposed in the Exploratory Studies Facility Main Drift may not be confined exclusively to the TSw middle nonlithophysal unit (Tptpmn). Unlike other monitored boreholes along the Exploratory Studies Facility Main Drift, pneumatic pressures at boreholes NRG-7a and SD-7 appear to have re-equilibrated nearly instantaneously in response to Exploratory Studies Facility excavation. In addition, at NRG-7a, which is closer to the Exploratory Studies Facility than SD-12, no reversals in water potential or temperature-recovery trends had yet occurred as of the first quarter of 1997. However, it is likely that further adjustments in the pneumatic characteristics of the system will occur in response to gas-driven moisture removal.

Recently, additional evidence of the effects of Exploratory Studies Facility excavation on the moisture regime in the TSw became apparent in the monitoring records from borehole NRG-7a. A significant change in water potential was observed in borehole NRG-7a (26 m horizontal distance from the Exploratory Studies Facility North Ramp) at station A in the TSw crystal-poor, upper lithophysal unit (Tptpul), which is close to the level of the potential repository. Water potentials at this station indicated a slow but steady wetting of the volume surrounding the borehole interval since the borehole was instrumented in October 1994 until about June 27, 1997 when the trend was reversed and the volume surrounding the borehole interval began to dry out (USGS 1997a, 1997b). Prior to June 27, 1997, water potentials were in the range of −0.1 to −0.2 MPa (−1 to −2 bars) but since June 27 have been in the range of −0.2 to −0.3 MPa (−2 to −3 bars). Temperature data indicate that the drying may be accompanied by a slight warming at several stations in borehole NRG-7a. Tentatively, this change in water potential has been attributed to drying out of the rock-fracture network between the borehole and the Exploratory Studies Facility due to the removal of moisture from the Exploratory Studies Facility tunnel by the mechanical Perche ventilation system.

5.3.4.2.2.2 Perched-Water Occurrences in the Site Area

By analogy with the definition of a water table, which is a surface along which water pressure is equal to atmospheric pressure, perched water may be defined as an accumulation of water above the regional water table within which water pressure is greater than atmospheric, and outside of which water pressure is less than atmospheric (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). According to this definition, perched water is not merely a volume of high saturation but a reservoir from which water will flow freely when intersected by boreholes or tunnels that are under constant atmospheric pressure. The fact that perched water exists in the vicinity of the potential repository has several implications. The very presence of perched water implies that, at least at some time in the past, the percolation rate through the unsaturated zone has exceeded the saturated-hydraulic conductivity of the perching layer. The perched-water reservoir may be a remnant from a time during which percolation rates were higher or may reflect long-term steady-state conditions. Perched-water reservoirs of large volume could indicate that structural or stratigraphic traps are present that allow percolation to accumulate. The mechanical stability of these trapping mechanisms becomes an important issue if perched water is discovered above or updip from the potential repository. Perched water in close proximity to the waste-emplacement tunnels in a repository is an additional potential source of water that may become mobilized as vapor resulting from waste-generated heat, a fact that

needs to be considered when attempting to analyze the impact of that mobilized water on repository performance.

Perched water has been identified below the potential repository horizon in six boreholes in the Yucca Mountain Site area: UZ-1, UZ-14, and NRG-7a in Drill Hole Wash, and SD-9, SD-12, and SD-7 along the Exploratory Studies Facility Main Drift. The perched-water bodies identified are at elevations significantly below (100 to 200 m) the potential repository horizon and do not represent obstacles to repository design or construction. Although other boreholes in the site area did not detect perched water, this probably is because they were not drilled to sufficient depth to intercept the geologic units where perched water has been identified or because they were drilled with water or foam that obscured the perched water when it was encountered. Although not detected in all boreholes, based on field observations and the apparent prerequisite conditions, perched water beneath the site area seems to be a common occurrence. In all cases, accumulation of perched water seems to be caused by either the basal vitrophyre of the Topopah Spring Tuff or the vitric-zeolitic boundary in the Calico Hills Formation acting in concert with a lateral structural barrier. Site-scale numerical-model simulations of the unsaturated zone flow system (Subsection 5.3.4.4) indicate that perched water is probably more extensive beneath the site area than might be suggested by the occurrences in the six boreholes (Bodvarsson et al. 1997, Chapter 13).

Perched-Water at Boreholes UZ-1 and UZ-14–Borehole UZ-1 is located in Drill Hole Wash at an elevation of about 1,349 m above sea level. Drilling stopped at a depth of 387.1 m when a possible perched-water reservoir was detected (Whitfield, Thordarson et al. 1990). The water encountered was at an elevation of 966.7 m, which is about 190 m above the water table at nearby UZ-14. Although no geologic samples were collected below 370 m in UZ-1, based on stratigraphy from nearby UZ-14, the top of the perched water is within the lower nonlithophysal zone of the Topopah Spring Tuff (Tptpln) about 8 m above the contact with the crystal-poor vitric zone (Tptpv). Chemical analysis of the water showed that it was contaminated by water used to drill G-1, a borehole located about 330 m to the southeast (Whitfield, Thordarson et al. 1990). During the drilling of G-1, approximately 8.7 million liters of drilling fluid was lost into the rock (Whitfield 1985).

Borehole UZ-14, located about 26 m from borehole UZ-1 in Drill Hole Wash at an elevation of 1,349 m, was drilled to a depth of 677.8 m. Perched water was detected at a depth of about 380 m (elevation of 967 m), which is about 9 m above the contact between the lower nonlithophysal zone (Tptpln) and the crystal-poor vitric zone (Tptpv) of the Topopah Spring Tuff. The top of the perched water is about 190 m above the water table. After the water was detected, several pumping tests were conducted to determine if the reservoir could be dewatered (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The perched water was pumped for about 3 days at an average rate of about 0.059 L/s but fully recovered about 5.6 days after the pumping stopped. This indicates that this perched-water reservoir may be extensive.

The perched-water reservoir at boreholes in UZ-1 and UZ-14 occurs near the upper contact of the basal vitrophyre (Tptpv), the lowermost unit in a sequence of highly fractured pyroclastic rocks

above relatively unfractured tuffs (Figure 5.3-143). The matrix permeability of the overlying lower nonlithophysal (Tptpln) unit is almost 30 times higher than the matrix permeability of the vitrophyre (Flint, L.E. 1998). The matrix permeability of the vitrophyre is an order of magnitude lower than the underlying partially welded unit and up to six orders of magnitude lower than the nonwelded Calico Hills Tuff. The difference in matrix permeabilities alone would provide for perching on top of the vitrophyre and not at the base, as is indicated by the occurrence of perched water. For water to perch on top of the vitrophyre, the vitrophyre would have to be relatively unfractured. However, limited data from borehole UZ#16 (LeCain 1997) show the air permeability of the vitrophyre is within an order of magnitude of the permeability of the Topopah Spring Tuff lower nonlithophysal zone. The air permeability of the nonwelded Calico Hills Tuff is two orders of magnitude less than the vitrophyre. If a correlation is assumed between air and fracture permeability, the fracture permeability in the nonwelded Calico Hills Tuff may be much lower than that of the vitrophyre. Water might be impeded on its way to the water table by the lack of fracture pathways in the nonwelded Calico Hills Tuff, and the perched water encountered at a depth of 381 m in the Topopah Spring Tuff may simply represent the upper part of the perched water reservoir whose base is within the Calico Hills Tuff.

Another possible explanation for the formation of the perched water is a hypothesis that invokes the presence of a lateral barrier to flow that effectively ponds the perched water in the vicinity of UZ-1 and UZ-14 (Barr, G.E. et al. 1996; J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). In this hypothesis, the lateral barrier may be formed by a northeast trending fault that is a splay off the Solitario Canyon fault. This fault has been interpreted to be a growth fault, thus by definition it has greater offset and may be more laterally extensive at depth than at the surface. The structural trap that allows the perched water to accumulate is created by the juxtaposition of a more permeable layer (Tptpln or upper Tptpv) west of the fault against a less permeable layer (the lower Tptpv or Tpbt1) east of the fault. This fault may intercept the water flowing downdip along the crystal-poor vitric zone (Tptpv) or the zeolitic alteration boundary, and may not have sufficient transmissivity to allow the water to drain as rapidly as it accumulates. This combination could explain the presence of the thick perched-water reservoir at boreholes UZ-1 and UZ-14.

Pumping Tests in Borehole UZ-14–Because approximately 9.8 m of a water-filled borehole was available, several drawdown/recovery tests were conducted in the perched-water reservoir at borehole UZ-14 (Striffler et al. 1996). Pump test #3 was run for approximately 9.3 hours with an average output of 0.118 l/s and a maximum drawdown of about 6.1 m. The water level was still recovering when pump test #4 was started. Pump test #4 was run for 66.75 hours at an average pump rate of 0.059 l/s. Maximum drawdown was 3.5 m and the water level recovered completely within 72 hours. Because the water level recovered completely following the pump tests, the perched-water reservoir was considered infinite in extent for the test analysis. Transmissivity was calculated to be 0.55 m^2/day for test #3, and 0.62 m^2/day for test #4. In discreet fracture flow regimes, the majority of the permeability and flow is from a unique zone, with little contribution from the rock matrix. The analytical solutions provide reasonable estimates of average permeability in porous-media flow conditions even if the assumptions are not strictly met. However, in this

situation, with predominate fracture flow, violation of the assumptions most likely produces errors in the calculation of hydraulic conductivity and transmissivity.

Perched Water at Borehole NRG-7A–Borehole NRG-7a is located in Drill Hole Wash at an elevation of 1,282.2 m. The borehole was drilled to a depth of 461.3 m into the top of the Calico Hills Formation. Water was first noticed in the borehole at a depth of about 460 m (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164), below the contact between the bedded tuff (Tpbt1) and the crystal-poor vitric zone (Tptpv) at the base of the Topopah Spring Tuff. The water level in the borehole then rose about 30 m to stand at a depth of 428 m (elevation of 855 m), which is 4 m above the base of the lower nonlithophysal zone (Tptpln). The perched water was initially detected about 91 m above the predicted water table at this borehole (Ervin et al. 1994). Because perched water was not a high priority at the time NRG-7a was drilled, no pumping tests or television logs were conducted to determine the exact nature and location of the water influx. Based on information from other boreholes, it is probable that water entered the borehole prior to the initial detection during drilling. The perched water was encountered near the contact of a series of highly fractured welded tuffs overlying relatively unfractured, nonwelded tuffs (Figure 5.3-144). This is similar to the situation at boreholes UZ-1 and UZ-14 where perched water may be entrapped in fractures while slowly imbibing into the matrix of the less fractured, underlying rock unit.

Perched Water at Borehole SD-9–Borehole SD-9 is located adjacent to Drill Hole Wash at an elevation of 1,302.2 m. The borehole was drilled to a total depth of 677.6 m and standing water was first noticed at a depth of about 449 m (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Television logs revealed that water was seeping through a fracture into borehole SD-9 at a depth of 413 m (elevation of 889 m), which is 3 m above the contact between the TSw lower nonlithophysal zone (Tptpln) and the TSw crystal-poor vitric zone (Tptpv) and about 157 m above the predicted regional water table (Ervin et al. 1994). The contact between the welded and fractured TSw crystal-poor vitric zone (Tptpv) and the pre-Topopah Spring bedded tuff (Tpby1) is at a depth of about 446 m (Engstrom and Rautman 1996). The perched-water reservoir is in fractured welded tuff (Tptpv) underlain by less-fractured nonwelded and bedded tuffs that comprise the uppermost part of the CHn (Figure 5.3-145). Because no pumping tests have been conducted on this perched-water reservoir, its areal extent is uncertain.

Perched Water in Borehole SD-12–Although no perched-water reservoir of sufficient magnitude to cause standing water in the borehole was detected during the drilling of SD-12 (Rautman and Engstrom 1996b), the video-camera log of this borehole indicates that a perched-water zone of limited extent probably is present in the densely welded vitric subunit (Tptpv3) of the Topopah Spring Tuff (Rousseau et al. 1997). The video-camera log indicates increased light reflectance and shimmering effect in the depth interval between 388.1 and 403.8 m, or from the top of the TSw densely welded vitric unit (Tptpv3) down to the base of the moderately welded vitric unit of the CHn. Perched water over this interval is also indicated by the in situ pneumatic-pressure responses observed in the instrument station immediately below this zone (see Subsection 5.3.4.2.1.1). The

fracture histogram for this borehole (Figure 5.3-146) indicates intense fracturing within the TSw lower nonlithophysal unit (Tptpln) and sparse fracturing in the underlying TSw crystal-poor vitric zone (Tptpv).

Perched Water in Borehole SD-7–Borehole SD-7 is located on the eastern slope of Yucca Mountain near the Exploratory Studies Facility Main Drift and near the southern extent of the potential repository area. The borehole was completed at a total depth of 815 m and water was first observed during coring at a depth of 488 m in the bedded tuffs (Tacbt) at the base of the Calico Hills Tuff (O'Brien 1997). This level is 4.5 m above the top of the Prow Pass Tuff (Figure 5.3-147) and about 143 m above the regional water table (Ervin et al. 1994). The perched-water level subsequently rose 8 m to a drilling depth of 480 m. The stratigraphically complex bedded-tuff zone is a well sorted volcanic sandstone layer with argillically altered pumice in all layers, predominantly horizontal fractures, and some lamination below 487 m (Rautman and Engstrom 1996a). The occurrence of perched water at this depth may be directly related to the bedding layers in the bedded tuff, Tacbt (Striffler et al. 1996). The matrix permeability of the nonwelded vitric horizon is greater than that of the zeolitized horizon beneath it and the permeability differences alone may be sufficient to cause perched water. Lateral flow may contribute to form a significant reservoir of water, and the horizontally fractured sandstone layer may also accommodate flow, and could be a water bearing stratum.

Similar to the perched-water reservoir at boreholes UZ-1 and UZ-14, a structural trap may be present in the vicinity of borehole SD-7 (Striffler et al. 1996). Water flowing laterally along fractured bedding layers from west to east may encounter a fault to the east of borehole SD-7. The permeable bedding layers abut the nonwelded to partially welded Prow Pass Tuff at the fault offset. Presumably, the fault does not have sufficient transmissivity to allow the water to drain as rapidly as it accumulates, forming the perched water reservoir.

Hydraulic Tests at Borehole SD-7–Several hydraulic tests were conducted in borehole SD-7 during March 1995 (O'Brien 1997). After repeatable drawdown curves were obtained, a 30-hour hydraulic test was conducted with a mean discharge rate of 0.21 l/s. The mean transmissivity determined from drawdown data using two analytical solutions was 8 m^2/day. Recovery data were not analyzable because of borehole-storage effects dominating the response. The apparent permanent lowering of the water level (residual drawdown was 2.3 m) indicates that the reservoir is of limited extent and not hydraulically connected to the regional water table. Water levels appeared to slowly recover after the completion of the March tests, but they never returned to prepumping levels. Following deepening of the borehole by 9.1 m to a depth of 497 m, additional hydraulic testing was conducted in August 1995. Because deepening of the hole allowed for more drawdown, a 64.6-hour hydraulic test at a mean discharge rate of 0.16 l/s was conducted. The response to pumping was similar to the March 1995 testing with the exception of the water level being lowered below the interval that was apparently producing water. After about 60 hours of pumping the water level in the borehole reached a depth of 488 m and the time-rate of drawdown increased. The sudden change in the drawdown curve indicated that this interval contained the primary water-producing fracture(s) or layer of porous rock matrix. Analysis of the drawdown data resulted in a mean transmissivity estimate of 4 m^2/day. Similar to the March tests, most of the recovery occurred during the first 20 minutes after the termination of pumping. Slow recovery, at the rate of about 1 m per 6 days,

continued until water-level monitoring was terminated, at which time the residual drawdown was 2.9 m. Long-term recovery could not be monitored due to the perched reservoir in the borehole being cased to allow the drilling operations to resume.

Perched-Water Reservoir Volume at Borehole SD-7–Using the results of the March and August 1995 hydraulic tests, the reservoir volume for the perched water at borehole SD-7 was estimated to be 97,000 liters prior to pumping (O'Brien 1997). This reservoir-volume estimate does not include any water that might be downdip from the borehole or otherwise hydraulically inaccessible to the borehole.

A Conceptual Model of Perched-Water Flow at Borehole SD-7–As previously discussed, the occurrence of perched water at borehole SD-7 could be due to a hydraulic-conductivity contrast between layers, the presence of a fractured layer overlying an unfractured layer, or a stratigraphic situation involving layered heterogeneity. Although the physical orientation of the perched-water reservoir at SD-7 can only be hypothesized based on the available information, a favored conceptual model involves a north-south trending fault that forms a low-permeability boundary that prevents the water from draining downdip, as illustrated in Figure 5.3-148 (Striffler et al. 1996). This conceptual model provides a consistent set of explanations for the behavior of the system during hydraulic testing, as described below.

There is evidence of a discrete fracture zone at borehole SD-7 that follows the dip of the bedding (~6° east) (Striffler et al. 1996). Although there could be a network of interconnected fractures at random or high-angle orientations, for simplicity a single zone of subhorizontal fractures following bedding is considered. Fractures probably are the most important factor for significant saturated flow to occur in perched-water reservoirs at Yucca Mountain. Boreholes that do not intersect significant fractures probably will not produce significant volumes of water. Most of the perched-water reservoirs encountered at Yucca Mountain yielded insufficient quantities to the borehole to permit hydraulic testing. However, this does not necessarily mean that the perched reservoir was smaller than that intersected by borehole SD-7; it could merely be due to the borehole not intersecting a zone that is sufficiently fractured to produce water. At borehole SD-7, the hydraulic test data suggest that the lower permeability rock matrix contributed little to the water pumped from the borehole and that most of the flow probably was from discrete fractures in a small interval of the borehole. Otherwise the time rate of drawdown probably would not have increased so dramatically when the water level was lowered below the producing interval.

The initial occurrence of water in borehole SD-7 indicated that the perched-water reservoir was under confined conditions, because the water rose above the level at which it was encountered (Striffler et al. 1996). However, given the dip of the bedding and the limited height of the water column, the system probably was in contact with the atmosphere and under water table conditions at some distance from the borehole (Figure 5.3-148). Under these conditions, early-time pumping would appear to be under confined conditions, with a rapidly expanding cone of depression. A decrease in water level in the borehole would be quickly transmitted to the edges of the reservoir. As the drawdown reached the reservoir boundary, where water table conditions exist, an unconfined response would dominate the drawdown data. An actual dewatering of the fracture zone would occur at the interface between the saturated and unsaturated portions of the fractured zone. As the

reservoir volume was depleted, the driving hydraulic head on the system would be reduced, which would lead to lower rates of inflow to the borehole. Therefore, the apparent transmissivity would decrease with increased depletion of the reservoir due to pumping. This is a possible explanation for the lower transmissivity obtained from the August 1995 test (4 m^2/day) than that obtained from the March 1995 test (8 m^2/day). A perched-water system is effectively altered following dewatering due to pumping, so different systems were tested during March and August 1995 in borehole (O'Brien 1997).

After termination of pumping, the fracture zone near the borehole quickly refilled with water and reached equilibrium within several minutes. A relatively small amount of the total recovery occurred after the initial surge of water into the borehole. This slow recovery could be a delayed yield-type of response, where water was draining from the interval dewatered during pumping (Striffler et al. 1996). The drainage resulted in rising water levels after the fracture network had reached its initial equilibrium level. However, depth-to-water measurements between the March and August hydraulic tests indicated that the water level was continuing to slowly rise (O'Brien 1997). If drainage from the dewatered portion of the reservoir was contributing to the recovery, its influence would be expected to only last for a few days. Long-term recovery of the water levels may be indicating that there are other perched-water reservoirs adjacent to the reservoir intersected in borehole SD-7. The reservoirs probably are separated by low-permeability boundaries (Figure 5.3-149). The water level in the block intersected by borehole SD-7 was artificially lowered due to pumping, which created a hydraulic-head difference between the adjacent blocks and induced flow into the SD-7 block. Over a long period of time, the perched-water level in these blocks would equilibrate to the same level. This conceptual model implies that there is potentially a much larger reservoir that is compartmentalized by low-permeability boundaries (Striffler et al. 1996). The possibility of compartmentalization of fractured aquifers in Southern Nevada has been suggested in previous studies at the Nevada Test Site (Young 1972).

This conceptual model of the perched-water reservoir at borehole SD-7 is one of several possible models that could reasonably explain the responses observed in the hydraulic tests (Striffler et al. 1996). Other occurrences of perched water at Yucca Mountain potentially could be explained by this model. The fact that several perched-water reservoirs have been identified at generally the same stratigraphic level suggest that some of them may represent compartmentalized flow systems separated by low-permeability boundaries, in support of this conceptual model.

5.3.4.2.3 Estimation of Heat Flux Using Borehole Temperature Data

Temperature measurements were made in numerous boreholes within the central block of Yucca Mountain and in the surrounding area in the early 1980s (Sass and Lachenbruch 1982; Sass, Lachenbruch et al. 1988) as part of a regional heat-flow study and, more recently, as part of the instrumented borehole monitoring program which measured in situ temperatures, pneumatic pressures, and water potentials (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997). The measured temperature profiles provided a means for calibrating field-scale values of thermal conductivity (Bodvarsson and Banduragga 1996) and for calculating heat fluxes, from which

estimates of percolation flux were made (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Bodvarsson and Banduragga 1996; Bodvarsson et al. 1997; E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation). (See Subsection 5.3.4.3.1.2 for estimates of percolation flux based on heat flux.)

The early regional heat-flow studies were conducted in 18 boreholes drilled either for geologic sampling (G-series, and a- and b-series boreholes) or hydrologic testing (H-series boreholes), and 17 boreholes used to monitor water table elevations (WT-series boreholes), as well as one unsaturated zone test hole. Except for a few boreholes (G-1, G-2, G-4, and a#1) which had continuous temperature measurements made in water-filled standpipes in both the saturated and unsaturated zones, Sass, Lachenbruch et al. (1988) temperature data from the unsaturated zone generally consisted of point measurements made from air-filled standpipes. In the air-filled standpipes, measurements in the unsaturated zone were made every 100 or 200 feet, and equilibrium temperatures at particular depths were estimated from plots of temperature versus 1/time rather than measured directly, due to the relatively long thermal response time of the thermistors in the air-filled boreholes. The calibration accuracy of the thermistors used to measure the temperature data contained in Sass, Lachenbruch et al. (1988) was estimated to be a few thousandths of a degree Celsius (J. Sass, USGS, Flagstaff, Arizona, to E.M. Kwicklis, USGS, Denver, Colorado personal communication, March 1998). However, as described in Sass, Lachenbruch et al. (1988), some data were affected by the thermal instability of air in the standpipe and/or barometric pressure changes (Sass, Lachenbruch et al. 1988), so that, in some cases, the calibration accuracy is not representative of the true accuracy of the measurements.

Additional temperature data in J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) were obtained from stemmed boreholes using thermistors that had been calibrated against primary standards to an accuracy of 0.005°C (at a 95 percent confidence interval). Temperatures reported by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) for instrumented borehole UZ-1, in which temperature measurements were made with thermocouples, were estimated to have an accuracy of 0.1 to 0.5°C, based on the overall accuracy of thermocouples in general (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Published temperature data (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) appear to be in approximate thermal equilibrium with the surrounding wall rock, based on the exceedingly small rate of change in temperature, with time observed, since the results were first reported. After the drilling and stemming of UZ-1, thermal equilibration times for later instrumented boreholes were greatly reduced through the use of stemming material (polybeads) with a small mass (J.P. Rousseau

et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). In spite of the complications involved in measuring temperature in air-filled standpipes, the unsaturated zone temperature data in Sass, Lachenbruch et al. (1988) appear to be considerably more reliable, in general, than temperature data collected from the saturated zone, where intra- or interformational flow within the borehole, or between the casing and the borehole wall, appear to have significantly perturbed the ambient temperature profiles.

The study conducted by Sass, Lachenbruch et al. (1988) included 204 thermal conductivity measurements. Fifty-seven measurements of thermal conductivity on cores from the unsaturated zone with minimally disturbed in situ saturations were found to be bimodally distributed, with a mode of 1.0 W m^{-1} °C^{-1} for nonwelded tuff and a mode of 2.1 W m^{-1} °C^{-1} for welded tuff. The 134 measurements of thermal conductivity on tuffs from the saturated zone were approximately normally distributed, with a mode of 1.72 W m^{-1} °C^{-1}. Thirteen measurements of thermal conductivity were also reported for Paleozoic carbonates from the saturated zone, with a mean thermal conductivity of 5.0 W m^{-1} °C^{-1}.

Recently, Rautman (1995) developed a relationship between thermal conductivity and temperature, saturation, and porosity, based on tuff samples obtained from borehole NRG-6:

$$\lambda = C_0 + C_1T + C_2S + C_3\phi \qquad \text{(Eq. 5.3-5)}$$

where λ is the thermal conductivity, in J s^{-1} °C^{-1} m^{-1}
T is temperature, in °C
S is saturation, dimensionless
ϕ is porosity, dimensionless

The coefficients in the regression were different for welded and nonwelded samples. The regression for the welded tuffs involved 38 samples and 78 separate analyses, and had a multiple r^2 of 0.58. For welded tuffs:

- C_0=1.748 J s^{-1} °C^{-1} m^{-1}
- C_1= 3.529 x 10^{-4} J s^{-1} °C^{-2}
- C_2=0.4347 J s^{-1} °C^{-1} m^{-1}
- C_3 = -3.210 J s^{-1} °C^{-1} m^{-1}

For nonwelded tuffs:

- C_0=1.250 J s^{-1} °C^{-1} m^{-1}
- C_1= 2.096 x 10^{-3} J s^{-1} °C^{-2}
- C_2=0.5853 J s^{-1} °C^{-1} m^{-1}
- C_3 = -2.238 J s^{-1} °C^{-1} m^{-1}

The regression for the nonwelded tuffs involved only seven tuff samples, but included 204 separate analyses, and had a multiple r^2 of 0.95. A graphical representation of the relation of λ to saturation

and porosity for a temperature of 20°C is shown in Figure 5.3-150. The temperature dependence of thermal conductivity was examined to account for the expansion of the rock under the extreme heat loads imposed by the radioactive decay of emplaced nuclear waste. For the range of temperatures that exist in the unsaturated zone under ambient conditions (15 to 35°C), estimates of λ produced by the regression relationship do not vary by more than a few percent and the effects of temperature on thermal conductivity are ignored in the analyses that follow.

Formation thermal conductivities at individual boreholes were calculated by Sass, Lachenbruch et al. (1988) using a variety of estimation and weighting methods. These calculations were then combined with data on thermal gradients to produce estimates of heat flux. As mentioned above, calculations of heat flux for the saturated zone were complicated by the occurrence of cross-formational water flow in the annular space between the casing and borehole wall that occurred as a result of vertical-head differences between formations. The problem of convective heat-flow within the borehole resulted in several instances, notably at H-5, G-3, and a#1/b#1h (drilled on the same pad), where saturated zone conductive heat flux was estimated to be less than the unsaturated zone conductive heat flux at the same borehole. Within individual boreholes, the calculated heat fluxes estimated by Sass, Lachenbruch et al. (1988) for discrete intervals within the unsaturated zone also showed considerable variability without discernible vertical trends that might indicate advective liquid or gas movement. Vertical variability in estimates of unsaturated zone heat flow was most pronounced at boreholes WT-3, WT-4, WT-5, WT-10, WT-12, WT-13, and WT-16 (Sass, Lachenbruch et al. 1988, Table 6).

Heat-flux estimates from the saturated zone were made using both least-squares estimates of the temperature gradients for the entire saturated zone temperature profiles and using only linear segments that appeared to be unaffected by convective flow within that interval of the borehole. Using only short, linear segments, the saturated zone heat flux was estimated to average 49 ± -8 mW m^{-2}. Using temperature gradients for the entire saturated zone thickness estimated from least-squares linear fits to the temperature data; the heat flux of the saturated zone averaged 40 ± 9 mW m^{-2}. Both estimates are considerably smaller than the regional heat flux value of approximately 85 mW m^{-2}. The average unsaturated zone heat flux was 41 mW m^{-2}, a value which was not significantly different at a 95 percent confidence interval from the saturated zone heat flux calculated with either method.

The unsaturated zone heat fluxes calculated by Sass, Lachenbruch et al. (1988) at individual boreholes were averages that considered the heat fluxes estimated across multiple intervals of the borehole. In some cases, the heat fluxes estimated at individual boreholes showed considerable variability between intervals. Some of this variability was undoubtedly caused by poor definition of the temperature gradient across the thinner, nonwelded units that occurred because of the widely spaced measurements in the unsaturated zone. In addition, thermal conductivity is a function of saturation, which is highly variable in the nonwelded tuffs, so that the use of formation averages cannot reflect the local saturation conditions at a particular borehole. In contrast, temperature gradients in the Topopah Spring Tuff are better defined because they are based on a larger number of measurement points, and saturation is less variable in the welded Topopah Spring Tuff, so the use of the average formation thermal conductivity is more appropriate. Therefore, unsaturated zone heat fluxes were reevaluated using only data from the Topopah Spring Tuff. The average thermal

conductivity for this unit is 1.93 W m^{-1} °C^{-1}, with a standard deviation of 0.29 W m^{-1} °C^{-1}, based on data for the welded Topopah Spring Tuff contained in Sass, Lachenbruch et al. (1988). Temperature gradients were determined from linear fits to the temperature data for the Topopah Spring Tuff. Table 5.3-48 lists the estimated temperature gradients, heat fluxes, and 95 percent confidence limits on the temperature gradients and estimated heat fluxes for data from the Topopah Spring Tuff. The table includes data from both Sass, Lachenbruch et al. (1988), as well as the recently drilled and instrumented boreholes SD-12, NRG-6, UZ-7a, and NRG-7a.

As noted by Sass, Lachenbruch et al. (1988), average heat flux in the unsaturated zone varies both geographically (Figure 5.3-151) and as a function of unsaturated zone thickness (Figure 5.3-152). Some of the apparent correlation between unsaturated zone thickness and heat flux may, in fact, be related to the geographic distribution of rock types with different thermal conductivities in both the unsaturated and saturated zone, fluid movement in both the saturated and unsaturated zones, and deep-seated variations in heat flux related to magmatic activity, as well as to topography.

As shown in Figure 5.3-151, heat flux with the Topopah Spring Tuff increases toward the southern end of Yucca Mountain, and to a lesser extent, east of Yucca Mountain toward Fortymile Wash. Over most of the central block of Yucca Mountain and in areas of Yucca Mountain north of Drill Hole Wash, heat flux within the Topopah Spring Tuff varies between 30 and 40 mW m^{-2}, with values toward the lower end of this range concentrated in Drill Hole Wash. Higher values of heat flux occur northeast of G-2 in Yucca Wash. The relatively large heat fluxes observed near UE-25p #1 may be related to a structural high in the underlying carbonate rocks, which have thermally conductivities greater than twice those of welded tuffs (Sass, Lachenbruch et al. 1988).

The spatial variability in unsaturated zone heat flux, as originally depicted in Sass, Lachenbruch et al. (1988), has been qualitatively explained as the result of either the upwelling or downwelling of water within the saturated zone (Fridrich et al. 1994), or as the result of spatially variable rates of water movement within the unsaturated zone (Bodvarsson and Banduragga 1996). The explanation offered by Fridrich et al. (1994) is difficult to defend because it assumes that the effects of unsaturated-zone heat-transfer processes are negligible and that the average unsaturated-zone heat flux represents the heat flux at the water table, and a detailed examination of the curvature in individual borehole temperature profiles from the saturated zone does not indicate upward or downward head gradients where their model requires them to be. The explanation of the spatial pattern of heat flux presented in Bodvarsson and Banduragga (1996) is similarly difficult to defend because it assumes that spatial variations in heat flux arriving at the water table because of saturated-zone processes are negligible in comparison with the effects imposed by unsaturated-zone percolation. Also, because the thermal conductivity of nonwelded tuff is only half that of welded tuffs, geographic variations in formation thicknesses within the unsaturated zone can be expected to affect the thermal gradients within the unsaturated zone, and hence, estimates of unsaturated-zone heat flux. In northern Yucca Mountain, lava flows intermingled with the relatively thicker Crater Flat and Calico Hills Tuffs created a paleo-highland that caused the Topopah Spring Tuff to thin to the north (Buesch, Nelson et al. 1996). Similarly, relatively greater thicknesses of the Pah Canyon, Yucca Mountain, and associated bedded tuffs (the PTn) north of Drill Hole Wash reestablished a paleo-highland that resulted in a thinning of the Tiva Canyon Tuff in the north. Therefore, in the north, the fractional thickness of the unsaturated zone composed of nonwelded tuffs is greater, and

the thickness-weighted thermal conductivity smaller, compared with the southern end of Yucca Mountain where the nonwelded tuffs are thinner and comprise a smaller fractional thickness of the unsaturated zone. A comprehensive explanation of the spatial variability in unsaturated-zone heat flux would need to consider these geologic factors, as well.

5.3.4.2.4 Geochemistry and Isotopes

This subsection presents aqueous-phase hydrochemical data for water recovered from the unsaturated zone. The data include both major-ion concentrations and isotopic compositions. Interpretations of these data provide constraints for possible flow mechanisms and residence times. Methods used to collect and process water and gas samples for chemical and isotopic analysis are described in Subsection 5.3.2.1.5. Pertinent information about boreholes from which samples were collected is listed in Table 5.3-8. Borehole locations are shown in Figure 5.3-8.

5.3.4.2.4.1 Major-Ion Chemistry

Chemical composition of pore water samples extracted from the drill cores of selected boreholes at Yucca Mountain are presented in Table 5.3-49 (Yang et al. 1996, 1998). The pore waters extracted from bedded tuff of the PTn are calcium-chloride or calcium-sulfate-type waters that plot near the top part of the diamond in the Piper diagrams (Figure 5.3-153). This characteristic becomes more pronounced for samples collected deepest within the PTn. A single sample of pore water taken approximately 1.5 m below the top of the PTn within UZ#16, where the PTn is relatively thin, has not yet acquired the calcium-chloride or calcium-sulfate signature.

The pore waters extracted from Calico Hills nonwelded unit are sodium carbonate-bicarbonate-type waters that plot near the lower part of the diamonds (Figure 5.3-153). Again, waters become more strongly sodium carbonate-bicarbonate types with increasing depth within the Calico Hills Formation. In spite of large distances separating boreholes, chemical compositions of pore waters are generally similar within a given stratigraphic unit and markedly dissimilar between different host lithologies in any given borehole.

The total concentration of major ions in pore waters as shown by total dissolved solid contents (Table 5.3-49) is highly variable, and is often greater near contacts than in the middle of stratigraphic units. The larger concentrations suggest a greater degree of rock-water interaction which could indicate a prolonged contact of percolating water with silicate rocks. Note too that the unsaturated-zone pore water has significantly larger concentrations of total dissolved solids than either perched water (Table 5.3-50) or saturated-zone water (Benson and McKinley 1985).

The concentration of chloride in many pore water samples is one to two orders of magnitude greater than that of either perched water or saturated-zone water. Chloride concentrations of perched waters are between 6 and 15.5 mg/L (Table 5.3-50) with a mean of 6.8 mg/L, which is similar to that of the saturated zone water from the volcanic rocks beneath Yucca Mountain (Benson and McKinley 1985). In contrast, the chloride concentration of matrix pore water ranges from 10 to 245 mg/L with a mean of 49 mg/L. If matrix pore water had contributed significantly to the perched-water bodies, the chloride concentration of perched water should be similar to that of the pore water. The smaller

concentration of chloride in perched water means that there has been less interaction of these fluids with rock, and that the perched water probably was derived from flow through fractures or faults that fully penetrate the PTn.

The chemical composition of pore waters within the nonwelded units does not change in a simple or predictable fashion. Figures 5.3-154 through -156 show the variation in cation and anion concentrations in the pore water versus depth in the PTn and CHn units in three boreholes. Chloride and sulfate show large concentration changes within pore water of the CHn, even for samples separated by less than 0.5 m. Similar erratic changes are noted for sodium and total carbonate within the PTn. If flow within these units were dominated by vertical percolation of pore water, gradual increase in sodium and decrease in calcium should be noted in the CHn because of cation exchange by zeolites (replacement of Na with Ca or Mg). Pore waters for SD-9 do show a continual increase in the sodium content with depth, but there are only four samples. Data for UZ#16 show a decrease in calcium with depth, but changes are fairly erratic and the concentration of calcium rebounds sharply just below the CHn. The data, therefore, suggest at least some component of lateral flow within the nonwelded units.

The isotopic composition of strontium in pore waters varies less erratically than the major ion chemistry and provides further evidence of rock-water interaction (Paces, Marshall et al. 1997). Strontium was separated from coarsely-crushed whole rock samples by brief leaches with distilled water. The results show that water enters the surface at Yucca Mountain with an isotopic composition roughly equal to that of the wet and dry fall material and begins to become more radiogenic by reacting with strontium in the Tiva Canyon Tuff (Figures 5.3-157). Pore water within the PTn becomes markedly more radiogenic, possibly due to longer reaction times. Further rock-water interaction within the Topopah Springs Tuff increases the $\delta^{87}Sr$ slightly. Finally, cation exchange with zeolites of the CHn (and possibly complication from a component of lateral flow) leads to marked variations in the $\delta^{87}Sr$.

5.3.4.2.4.2 Tritium and Chlorine-36

Tritium and ^{36}Cl are both natural environmental isotopes and are used widely in hydrology to trace the movement of groundwater. The application here is limited to the tracing of recent water which incorporates these radionuclides produced by atmospheric nuclear tests between 1952 and 1963. The half life of ^{3}H is 12.4 years and ^{36}Cl is 301,000 years.

Tritium–Natural tritium is produced from cosmic radiation interacting with the nitrogen in the upper atmosphere. Historical concentrations in rainwater in the middle latitudes have been estimated to be on the order of 10 tritium units (TU) (Davis and Bentley 1982), one TU being equal to one tritium atom per 10^{18} atoms of ^{1}H. The nuclear tests in northern hemisphere between 1952 and 1963 produced significant amounts of tritium in the earth's atmosphere. A yearly average value of 2,700 TU in precipitation near Salt Lake City in 1963 was reported by Michael (1989). A nuclear-test ban in 1963 stopped further tests in the atmosphere, and the tritium concentration has gradually decreased to the present level of about 10 to 40 TU measured in precipitation at the Nevada Test Site (Milne et al. 1987).

The largest concentrations of ^{3}H found thus far within the unsaturated zone are associated with the Bow Ridge fault. Samples were collected from borehole HPF#1, which was drilled from the Bow Ridge fault alcove of the Exploratory Studies Facility into the fault (LeCain et al. 1997). Values as great as 155 TU were found in pore waters 3.4 m west of the fault within the pre-Rainier Mesa Tuff. Values of approximately 120 and 130 TU were found in pore waters within the fault breccia. The high values are not surprising given the width of the fault zone at the surface, the lack of alluvium up-slope from the fault, and the shallow depth of the fault where sampled (approximately 20 to 30 m).

Tritium data for the unsaturated-zone boreholes show several "inversions" (Yang et al. 1996, 1998), with larger tritium concentrations located below smaller tritium concentrations in a vertical profile (Figures 5.3-158 and -159). These inversions indicate that vertical water percolation through the matrix is not the only type of flow mechanism at Yucca Mountain. Post-bomb tritium concentrations were observed down to bedded tuff or Pah Canyon Tuff in many boreholes. The occurrence of bomb-generated tritium in waters below non-tritium-bearing water (and hence older water) in a vertical profile is a strong evidence of fracture and lateral flow occurring at Yucca Mountain.

The post-bomb (post-1952) ^{3}H in the deep unsaturated zone greater than 366 m (1,200 feet) was found only in the borehole of UZ#16 (Figure 5.3-159). The chemical composition of pore water at 501.5 m (1,645 feet) in Prow Pass Tuff of this borehole is similar to that at the shallower depth at 49.8 m (164 feet), that is it is a calcium-bicarbonate type water instead of the sodium-bicarbonate type commonly seen at this depth (Yang et al. 1996). This suggests that younger water from the shallow depth was transported through fractures with little cation exchange in the zeolitic unit of Calico Hills nonwelded unit, which is consistent with the observed post-bomb tritium.

Chlorine-36–Measurements of chloride (Cl^-) and bromide (Br^-) concentrations and $^{36}Cl/Cl$ for salts extracted from water, soil, and rocks are being used to provide information on characteristics of water movement and solute transport through the unsaturated zone at Yucca Mountain. Chlorine-36 is a radioactive isotope of chlorine, and in nature occurs primarily as the chloride anion. As such, it is relatively inert in the subsurface environment and behaves conservatively. This radionuclide is present in infiltrating waters as a natural tracer produced mainly by cosmic rays interacting with argon in the atmosphere. The relatively long half-life of ^{36}Cl theoretically permits the detection of travel times up to several hundred thousand years. In order to normalize the data for the variable effects of evapotranspiration, the ^{36}Cl concentration is generally reported relative to that of stable Cl^-. Expressed in this manner, the present-day background level is $^{36}Cl/Cl$ of 500×10^{-15} (Fabryka-Martin, Turin et al. 1998). Global fallout from thermonuclear tests conducted primarily in the Pacific Proving Grounds resulted in a ^{36}Cl "bomb-pulse" with maximum meteoric ratios in excess of $200{,}000 \times 10^{-15}$. These extremely high values were rapidly diluted by mixing processes in the soil zone and subsurface, and are not observable today. Nevertheless, high $^{36}Cl/Cl$ (those greater than about 1250×10^{-15}) indicate some bomb-pulse component, and their appearance in an environmental sample signals the presence of at least a small component of bomb-pulse ^{36}Cl. In groundwater anomalous ratios suggest travel times from the ground surface of 50 years or less (Fabryka-Martin, Wolfsberg et al. 1997; Levy, Sweetkind et al. 1997; Sweetkind, Fabryka-Martin et al. 1997).

Although the existence of bomb-pulse ^{36}Cl enables the study of solute transport with travel times of 50 years or less, the natural cosmogenic background permits analysis of much longer term transport processes. However, the use of $^{36}Cl/Cl$ for dating older waters is not as straightforward as the case for bomb-pulse signals due to the time-varying input signal. Both theoretical considerations as well as measurements of soil profiles and fossil urine from ancient pack-rat middens support the hypothesis that the present-day background ratio has remained relatively constant during the Holocene (last 10 ka). Expected lower rates of stable Cl deposition during the Pleistocene (10 ka to 2 Ma), in combination with southern shifts in the jet stream, would have led to higher local meteoric $^{36}Cl/Cl$, relative to those that were present throughout the Holocene (Plummer et al. 1997). Superimposed on this effect are varying atmospheric ^{36}Cl production rates caused by variations in the earth's geomagnetic field. Based on these two factors, the meteoric $^{36}Cl/Cl$ at Yucca Mountain has been reconstructed for the last 1.8 Ma (Fabryka-Martin, Turin et al. 1998). Analyses of $^{36}Cl/Cl$ in pack-rat midden samples dated by the ^{14}C technique are generally consistent with the reconstruction, although these data can only cover the last 50 ka due to the comparatively short half-life of ^{14}C.

Together the bomb-pulse record and the long-term reconstruction allow $^{36}Cl/Cl$ observations to be divided into four classes (Fabryka-Martin, Turin et al. 1998):

- Ratios over 1250×10^{-15} provide clear evidence of bomb-pulse influence, and indicate the presence of some rapid transport pathways.

- Ratios near the present-day value of 500×10^{-15} suggest either extremely young (post-1980) precipitation, or Holocene precipitation of pre-nuclear-age (from 1950, extending back to 10 ka before present).

- Ratios that are elevated above present-day background but less than 1250×10^{-15} cannot be interpreted unambiguously in terms of water travel time, in the absence of additional geochemical or isotopic information. These may be attributed to Pleistocene-aged water that is not so old as to unequivocally demonstrate radioactive decay. Alternatively, these ratios may reflect the presence of a small component of bomb-pulse ^{36}Cl.

- Ratios significantly less than 350×10^{-15} clearly show the effects of radioactive decay of ^{36}Cl. Actual estimates of water age will depend on the past meteoric ratios which must be considered highly speculative for the period before the earliest pack-rat data (40 ka) but, based on the current reconstruction, these ratios imply ages in excess of 200 ka.

Unfortunately, the interpretation of a given sample in one of the above categories is not always clear-cut. Only interpretation of the highest ratios (indicating bomb-pulse water) and the lowest ratios (indicating radioactively decayed ancient water) may usually be considered relatively unequivocal. Intermediate ratios may represent Holocene or Pleistocene input with corresponding travel times, or may represent mixtures of waters of different ages. Nonetheless, the intermediate ratios shed light on the interpretation of water ages when considered in conjunction with other independent lines of evidence, such as other isotopic measurements and infiltration studies and results of flow and transport simulations of alternative conceptual models.

Data Collection and Interpretation. Over 700 ^{36}Cl measurements have been conducted thus far for the YMP (Fabryka-Martin, Wolfsberg et al. 1997; Fabryka-Martin, Turin et al. 1998; Levy, Sweetkind et al. 1997; Sweetkind, Fabryka-Martin et al. 1997). Analyzed samples include surface soils and soil profiles; rock samples collected from boreholes and within the Exploratory Studies Facility; water samples including surface runoff, unsaturated-zone pore water, perched water, and saturated-zone water; and fossilized urine from pack-rat middens.

Analyses of bomb-pulse ^{36}Cl and Cl^- profiles in soil and alluvial profiles show that thick alluvium is effective at reducing net infiltration to levels less than 1 mm/year (Fabryka-Martin, Turin et al. 1998). Similarly, borehole samples collected from rock units underlying such thick alluvium do not show any evidence for the presence of bomb-pulse ^{36}Cl (Fabryka-Martin, Turin et al. 1998). In contrast, where alluvial cover is thin (e.g., less than a few meters thick) or missing, water is able to readily enter fractures of the bedrock, as shown by the detection of bomb-pulse ^{36}Cl measurements in samples from such locations. For example, fast transport through the welded Tiva Canyon is indicated by bomb-pulse $^{36}Cl/Cl$ measured in several boreholes that intersect the nonwelded Paintbrush Tuff which lies below the TCw. Other evidence of fast pathways that persist into the TSw is provided by bomb-pulse $^{36}Cl/Cl$ measured at locations in the Exploratory Studies Facility (Figure 5.3-160) (Fabryka-Martin, Wolfsberg et al. 1997; Levy, Sweetkind et al. 1997; Sweetkind, Fabryka-Martin et al. 1997). A suite of 247 samples has been analyzed from this 8 km tunnel, of which 31 had ^{36}Cl levels sufficiently elevated as to be interpreted as unambiguous evidence for the presence of bomb-pulse ^{36}Cl. The interpretation of the elevated ^{36}Cl signals as evidence for bomb-pulse is supported by the presence of bomb-pulse ^{99}Tc at two locations that also contained elevated ^{36}Cl: the Bow Ridge fault exposed at Exploratory Studies Facility station 2+00, and UZ-N55 cuttings from the base of the PTn. The correlation of the elevated measurements in the Exploratory Studies Facility with the surface expression of faulting indicates that the pathway and travel time may involve locally modified PTn fracture properties (Fabryka-Martin, Wolfsberg et al. 1997; Levy, Sweetkind et al. 1997; Sweetkind, Fabryka-Martin et al. 1997). These data support the hypothesis that faulting or other disturbances increase PTn fracture permeability, thereby generating a local environment in the PTn that supports fracture flow and hence rapid transport of solutes. Once through the PTn, flux distributions favor fracture flow in the welded Topopah Spring, thereby providing a continuous pathway to the Exploratory Studies Facility.

Most sample analyses in deep boreholes and in the Exploratory Studies Facility had ratios less than the threshold for indicating bomb-pulse Cl (Figure 5.3-160). In the southern part of the Exploratory Studies Facility, beyond station 45+00, most samples had $^{36}Cl/Cl$ typical of Holocene water, which could suggest travel times less than 10 ka to this depth. (Averages of water with greater and lesser values provides a less likely explanation because the mixing proportions would have to be the same for the large number of samples.) A number of samples from this part of the system also had ^{36}Cl signals that were significantly below the present-day background value, suggesting the possible presence of zones of relatively stagnant water. In contrast, many of the $^{36}Cl/Cl$ in samples from the northern part of the Exploratory Studies Facility, up to station 45+00, had ratios variably above the present-day background. Aside from the unambiguous bomb-pulse samples in this area, it is not presently possible to assess whether or not these intermediate signals indicate the presence of a very small component of bomb-pulse ^{36}Cl or whether the elevated signals reflect travel times exceeding

10 ka, the most recent time when the signal in infiltrating water would have been high enough to provide the $^{36}Cl/Cl$ signatures found in water from this area.

Measured $^{36}Cl/Cl$ for samples of moderately to densely welded tuff from boreholes are systematically lower than the values for Exploratory Studies Facility samples (e.g., compare results for the TSw unit in Figure 5.3-161 to those in Figure 5.3-160). This apparent discrepancy is probably attributable to differences in the sample collection techniques. Borehole samples are obtained from ream cuttings. Rock Cl^- is released to the cuttings during drilling, and subsequently is leached together with the pore water Cl^- during sample preparation for ^{36}Cl analysis. This source of Cl has a $^{36}Cl/Cl$ on the order of 40×10^{-15}. Correction of the measured $^{36}Cl/Cl$ for dilution by rock Cl^-, with its lower ^{36}Cl content, thus increases the $^{36}Cl/Cl$ value of the meteoric component of Cl in the borehole samples. The Exploratory Studies Facility samples are not as greatly affected by rock Cl as are borehole samples because the manual method of collection used in the Exploratory Studies Facility does not pulverize the rock as does the ream bit. Boreholes samples from nonwelded units (PTn, CHn, Prow Pass) also do not appear to be greatly affected by the release of rock Cl^-, probably because the concentration of Cl^- in these particular units is negligibly small. Differences between uncorrected $^{36}Cl/Cl$ for borehole samples from welded units and those from the Exploratory Studies Facility demonstrate the importance of recognizing the magnitude of the influence of rock Cl to the $^{36}Cl/Cl$ values measured for borehole samples. After correcting for this effect, the two sets of analyses are consistent (Fabryka-Martin, Wolfsberg et al. 1997). None of the few $^{36}Cl/Cl$ measurements available for samples collected below the potential repository horizon are sufficiently high as to indicate the unambiguous presence of bomb pulse ^{36}Cl (Figure 5.3-161).

Corroboration of ^{36}Cl analyses with other isotopic studies helps bound the uncertainty and enhance the utility of the method. Corroboration with other isotopic analyses is demonstrated in the analysis of perched water ages. A meteoric water curve relating ^{14}C activity and $^{36}Cl/Cl$ for water samples of increasing age is constructed based on the reconstructed input signals for both isotopes (Figure 5.3-162). Then, ^{14}C activity is plotted against $^{36}Cl/Cl$ for perched water samples on the same curve. The data match the meteoric water curve well and the ages of the perched water can be estimated by interpolation along the meteoric water curve. Chlorine-36 and other isotopic analyses of perched water provide a consistent estimate of the age of the perched water bodies. Ages of the perched water can be estimated by interpolation along a characteristic curve based on a predicted correlation between reconstructed ^{14}C and ^{36}Cl input functions for the past 25 ka. By this approach, the ^{36}Cl analyses indicate ages ranging from 2 to 12 ka for the different perched-water bodies sampled, which is in general agreement with ^{14}C-based ages.

Consistency of Data with Site-Scale Flow and Transport Models. A flow and transport model using the FEHM code was used to simulate transport of ^{36}Cl into the Exploratory Studies Facility tunnel (Fabryka-Martin, Wolfsberg et al. 1997; Robinson, B.A. et al. 1996). Modeling results show that observed ^{36}Cl signals are consistent with the above site conceptual model and with reasonable parameter estimates. The overall picture from these studies is that:

- Infiltration is spatially variable and, on the average, exceeds 1 mm/year over the potential repository block.

- Fracture transport can be critical, permitting rapid transport through otherwise low-conductivity materials.

- Isolated fast paths associated with faults and fractures may penetrate deep into the mountain.

5.3.4.2.4.3 Carbon Isotopes ($\delta^{13}C$ and ^{14}C)

Carbon-14 and $\delta^{13}C$ values of perched waters are shown in Table 5.3-51. The ^{14}C values range from 66.9 to 27.2 percent modern (pmc), corresponding to apparent ^{14}C residence times of about 3,500 to 11,000 years (Yang et al. 1998). Water ^{14}C ages are affected by the dissolution of older carbon in the rocks, which would result in anomalously old apparent ages. Reaction with or incorporation of CO_2 from deep in the unsaturated zone will also result in anomalously apparent age. Total chemistry can be used to model these reactions and a corresponding age correction. Modeling for age corrections was performed with NETPATH Version 2.0 (Plummer et al. 1994), and is discussed later in the geochemical modeling subsection. Any $^{14}CO_2$ introduced by drilling air is expected to be insignificant because the mass of carbon in water reservoirs is orders of magnitude larger than that in the drilling air. Recent input of ^{14}C is expected to be minor because all perched water samples contain background tritium concentration. If post-bomb water is present in the perched bodies, the component is too small to be detectable.

The $\delta^{13}C$ values for perched water are quite variable ranging from -9.2 to -16.6. There is a weak correlation between these values and the surface material in which the borehole is located. SD-7, which is essentially started in bedrock, has heavy values of about -9.5, whereas NRG-7, which was drilled through soil, has the lightest value of -16.6. This is similar to the observation made for strontium data which shows a less radiogenic input for bedrock and fracture coatings where there is no soil and for pore waters from a borehole that was started in bedrock.

Pore waters from both the PTn and CHn have significant amounts of modern carbon (Figure 5.3-159) such that most samples suggest an apparent age of less than 10,000 years (Yang et al.1996, 1998). The apparent young ^{14}C ages (high ^{14}C pmc values) in the pore water from all nine boreholes might be explained by contamination from $^{14}CO_2$ in the drilling air. However, the concentration differences between carbon in water and air make this unlikely. Furthermore, the $\delta^{13}C$ values, particularly for pore waters in the Calico Hills nonwelded unit, do not correlate with the ^{14}C activities (Figure 5.3-163). If modern air were an end member for ^{14}C, it would also be an end member for $\delta^{13}C$. Note that the apparent ages may represent either a single episode of inflow or a mixing of older and younger pore waters.

The CO_2 gas borehole was collected after overnight pumping to remove atmospheric air. Gas samples from UZ-14 had ^{14}C activities of 48 to 68 pmc, similar to the pore water ^{14}C activities in CHn. Boreholes SD-9 and -12 exhibit significantly smaller ^{14}C activities for most of the gas samples than in pore waters from the same zone (Yang et al.1998). If the CO_2 in the gas phase were in isotopic equilibrium with CO_2 in the aqueous phase of the pore water, the ^{14}C activities in the two phases should be comparable. The reason for the disequilibrium currently is unknown.

5.3.4.2.4.4 Stable Isotopes (δD and $\delta^{18}O$)

Two methods of pore water extraction for stable-isotope analysis of δD and $\delta^{18}O$ were used in this investigation. These were the vacuum-distillation and the compression-extraction methods, which are described in Subsection 5.3.2.1.5.

Testing for isotopic effects for the two different water-extraction procedures was accomplished by analyzing adjacent core samples (Yang et al.1998). The two methods yielded comparable results in the PTn and TSw (Figure 5.3-164). However, the isotopic compositions of water obtained by vacuum distillation are more depleted in δD and $\delta^{18}O$ than water obtained by compression method for cores containing clay or zeolite minerals (CHn, Prow Pass Tuff, and Bull Frog Tuff) (see Figure 5.3-164). A series of tests were conducted to determine which method was providing the correct values for pore water. Cores from various lithologies were dried in an oven at 250°C for approximately 2 days. Dried samples were then evacuated and cooled in a sealed jar. While still under vacuum, water with a known isotopic composition was imbibed into the cores. Variable amounts of water were used to provide cores representing saturated and variable degrees of unsaturated conditions. The imbibed cores were kept under vacuum at room temperature for 10 days to allow the sample and water to equilibrate. After equilibration, the water was extracted by using the two methods, and the stable isotope compositions of extracted water were measured for $\delta^{18}O$ and δD.

Vacuum distillation of pore waters for stable isotope analysis produced reliable results from tuffs free of clay and zeolite minerals but not for tuffs with large contents of hydrated minerals (Yu 1996). Zeolite minerals will cause isotope fractionation by uptake of light-water molecules thereby leaving isotopically heavy water molecules in the pore spaces. The compression-extraction process only extracts pore water whereas vacuum distillation extracts both pore water and water held by zeolites or clays. Percolating water can replace the water in the large pore space as well as exchange with water in the zeolite. Exchange rates range from days to months. Pore waters extracted from zeolitic core by compression may, therefore, reflect the most recently infiltrated water but may lose the record of past percolating water.

The oxygen and hydrogen isotopic composition of pore waters can yield significant information about infiltrating waters, provided that sampling has not disturbed the original compositions. Significant amounts of drilling air were injected into the formation during borehole drilling which could cause drying of the core samples. The potential magnitude of this effect can be examined by plotting the stable isotope compositions of the pore water on the δD versus $\delta^{18}O$ diagram (Figure 5.3-165). All unsaturated-zone data points were close to the local precipitation line (Benson and Klieforth 1989) indicating very little evaporation loss, although the slight displacement below and to the right of the line may indicate some evaporation. This could have occurred either prior to recharge or during drilling. The composition of pore waters is lighter than that of rain measured for four major storms that occurred at Yucca Mountain in July and August of 1984 (Figure 5.3-166). The rain waters yielded values ranging from -67 to -97 ppm for δD and from -10.5 to -13.0 ppm for $\delta^{18}O$ (Milne et al. 1987). The lighter values for pore waters suggest that much of the pore water originated as snow or during a time of colder climate (Yang et al.1998).

Comparison of isotopic compositions of the unsaturated-zone pore waters (Figure 5.3-165) with the saturated-zone groundwater (Figure 5.3-166) beneath Yucca Mountain shows a large overlap (Yang et al.1998). There is a slight tendency for the saturated-zone waters to be isotopically lighter, which would be consistent with recharge at a higher elevation or during colder conditions or both. Stable isotopic data of the perched water are generally between -12.8 and -13.8 ppm for $\delta^{18}O$ and -94 and -99.8 ppm for δD (Table 5.3-51). These values are slightly greater than those for the saturated-zone values and similar to values for pore waters. They are generally close to the Yucca Mountain precipitation line, indicating little evaporation before infiltration (Craig 1961; Dansgaard 1964). All stable-isotopic values are similar to modern precipitation and are, therefore, consistent with recharge during the Holocene as suggested by the ^{14}C data for the perched waters.

Causes of Large Dissolved Solids–Six samples have very large dissolved-solid concentrations (>800 ppm, Table 5.3-49) in their pore waters. The isotopic composition of water extracted by squeezing for three of these samples is shown in Figure 5.3-167. All of the pore waters plot to the right of the Yucca Mountain precipitation line, which is consistent with evaporation after precipitation (Yang et al.1998). The extent of evaporation for sample NRG-6 (256 feet), which has the largest apparent loss of moisture by evaporation, can be approximated assuming a temperature of 20°C for evaporation, equilibrium Rayleigh fractionation (*c.f.* Fontes and Garnier 1979, p. 8), and an initial starting composition equal to the average winter precipitation (Ingraham et al. 1991; Milne et al. 1987). The results suggest a maximum evaporation of 12 percent.

The data do not provide information on the timing of evaporation; however, field investigations suggest that core processing may cause approximately 6 percent loss of pore water. The remaining loss could have occurred prior to infiltration or within the subsurface. In either case, the total amount of evaporation is insufficient to account for the large amount of total dissolved solids, and thus, rock-water interaction provides the best explanation for the anomalous values. The rock-water interaction is not likely to have been surficial because this process would be expected to lead to a greater homogeneity for at least the shallower pore waters, which is not observed. Furthermore, there is abundant calcium carbonate at the surface of Yucca Mountain which would likely be a major contributor to dissolved solids prior to infiltration, but the samples with the largest total dissolved solids do not contain substantially larger amounts of calcium or total carbonate than other samples. Thus the rock-water interaction is most likely a subsurface process, and variability noted for pore waters of various units probably reflects varying amounts of time for reaction.

5.3.4.2.4.5 Interactions of Old Matrix Water with Younger Percolating Water

Water within the unsaturated zone can be thought of as a mixture of three chemical end members:

- Old infiltrations trapped in the pores of the rock matrix
- Modern infiltration moving primarily through fractures
- Water bound (to varying degrees) within minerals

Furthermore, water within tuffs at Yucca Mountain undergoes isotopic exchange (Yu 1996), particularly between pore water and percolating infiltration. Isotopic exchange appears to be diffusion controlled and the equilibration time is dependent on the mean travel distance and the

diffusion coefficient (Yu 1996). Zeolites contain loosely held water which is referred to as channel water, and as noted above, this can contribute to analytical results for pore water. The channel water of pure clinoptilolite powder exchanges with the water percolating through laboratory columns within several hours to days. For ground tuffs from Yucca Mountain, the exchange rate is also several hours to days, whereas exchange for zeolites within intact cores take longer, depending on the core size and grain size of the zeolites within the core. The exchange processes for zeolite-bearing core can be divided into two steps: the exchange between the pore water and the percolating water, and the exchange between the pore water and channel water. If the system reaches equilibrium, the stable isotopic compositions of pore water should equal those of percolating water, and those of channel water will differ from pore water by a fractionation factor. The second mechanism is still unknown, but the reaction is fast, as shown by column experiments. Therefore, the equilibration time is determined by the first process (Yu 1996). For Topopah Spring Tuff, which bears no zeolites, the stable isotope composition of pore water will be controlled by exchange occurring through diffusion mechanisms.

As shown in Figure 5.3-164, pore water from the unsaturated zone has stable-isotope values generally heavier than -100 per mil in δD; which is larger than the saturated-zone water δD values of -101 to -103 per mil (Yang et al.1998). Note that these waters had uncorrected ^{14}C ages between 12,000 and 18,000 years. If mixing of a significant amount of last ice-age waters is involved in the matrix water of the TSw, the δD values should shift to more negative values, which is not observed. However, in the vitrophyre zone near the bottom of the TSw, the δD values in all three boreholes (SD-7, -9, and -12) are consistently shifted to more negative values, implying that waters were infiltrated during the colder climate of the last ice-age. The δD values in the CHn in all three boreholes are larger than those of the vitrophyre zone and similar to those values above the vitrophyre zone, indicating post-ice-age water.

5.3.4.2.4.6 Gas-Phase Chemistry and Gas-Water Interactions

Methods used to collect and process gas samples for chemical and isotopic analysis are described in Subsection 5.3.2.1.5.

Rock-gas compositions in the unsaturated zone (Table 5.3-52) are similar to atmospheric air except that CO_2 concentrations are generally larger in the rock gas (about 0.1 percent) than in air (0.034 percent). The CO_2 concentration profiles as a function of time in UZ-1 show that relatively low CO_2 concentrations were measured in 1983 shortly after completion of the borehole (Yang et al. 1996), and that, except for probe 13, concentrations steadily increased until 1987 (Figure 5.3-168). Thus the early CO_2 samples, were probably diluted by drilling air. As the semiannual gas sampling process proceeded, most of the drilling air was removed from the hole and the more recent samples probably represent uncontaminated rock gas. This conclusion is supported by data collected from 1986 through 1994, which show CO_2 concentration from 100 to 360 m changed very little (Figure 5.3-168). The CO_2 concentrations on the upper portions of UZ-1 are elevated due to their proximity to the soil zone where biologic activity increases CO_2 concentrations. The increased CO_2 concentrations at the bottom of the borehole probably represent breakdown of organic polymers from drilling fluids. The consistently low values for CO_2 concentrations at probe #13 cannot be explained by currently available data. If probe 13 is sensing a fracture connected to the atmosphere, the

concentrations would indeed be anomalously low, but the isotopic composition would be anomalously heavy, which is not the case.

$\delta^{13}C$ values relative to depth in UZ-1 showed large variations in both space and time (Figure 5.3-169). Gas samples collected in 1984 are heavier than those collected in 1985, but as noted above, the 1984 samples are probably heavily contaminated with drilling air which would be enriched with ^{12}C. The light values near the bottom of the hole are likely caused by breakdown of organic polymers from drilling fluids. From 1986 through 1994, most values for $\delta^{13}C$ were in the range of -14 to -17. A similar range was obtained for gas samples from three radial boreholes drilled from the upper Tiva Canyon contact alcove of the Exploratory Studies Facility. Boreholes within the upper Paintbrush contact alcove tend to be slightly heavier ($\delta^{13}C$ of generally -12 to -13), but this tendency is not noted in the shallower reaches of UZ-1. In fact, the earliest gas samples tend to be anomalously light (<-20 per mil) at shallow depth, reflecting a biogenic (plant root respired) component. Strong departures from the average $\delta^{13}C$ values were observed in 1985, 1988, 1991, 1992, and 1994 at one or more depths. As noted above, the 1985 data may not represent equilibrium rock-gas data. Subsequent excursions from average values remain to be explained.

The isotopic composition of pore waters is highly variable, generally ranging from approximately -9 to -19 (Table 5.3-53) (Yang et al. 1996, 1998). The isotopic composition of rock gas is likewise highly variable, ranging from approximately -12 to -26 (Table 5.3-54). There is no evidence that the rock gas is in isotopic equilibrium with the pore waters equilibrium; carbon in bicarbonate in water should be about 8.5 per mil heavier than carbon in CO_2 in gas. There are no data for gas and pore water from the same sample, but Table 5.3-55 compares a few samples that were collected from the same general depth ranges. For these, the differences for $\delta^{13}C$ in gas and pore water are much less than 8.5. The larger range for $\delta^{13}C$ in pore water than for gas in a given borehole also argues against large-scale equilibrium of carbon isotopes in the two phases.

The ^{14}C data for borehole UZ-1, like the $\delta^{13}C$ values, also show large variations in 1984 and 1985. The values tend to lie towards more modern carbon than has been observed since 1986 (Figure 5.3-170). This is consistent with the explanation given previously of the drilling-air contamination in the test hole. Present-day ^{14}C activity in the atmospheric air is about 120 percent modern carbon (pmc), which would cause the ^{14}C activity of borehole CO_2 to shift toward larger values. The ^{14}C data have been very consistent for the last 7 years, with a gradual decrease in ^{14}C activity with depth to about 23 pmc at 368 m. The ^{14}C profile shows that there is an abrupt change in the slope within the PTn. The transport velocity (^{14}C concentration gradient over distance) within the PTn is smaller than the transport velocity in the TSw unit. The smaller transport velocity may be due to greater amount of water in this unit. An estimate of the minimum travel-time of gas in the TSw unit based on the apparent ^{14}C ages and depths in the borehole yields gas movement of 3.26 cm/year. This rate, as well as results of gas-transport modeling (Yang et al. 1996), is consistent with downward movement of atmospheric CO_2 by simple Fickian diffusion. Although diffusion may not be the only mechanism for gas movement in the TSw unit at UZ-1, it seems to be the dominant mechanism and can account for the observed distribution of gaseous ^{14}C with depth.

Although the data sets are less extensive, data for boreholes NRG-6, NRG-7a, and UZ#16 yield different results from those for UZ-1. In these boreholes, percent modern carbon does not appear

to decrease with depth (Figures 5.3-171 and -172), and in fact, all three analyses from NRG-7a show post-bomb carbon. All three have similar but measurably different $\delta^{13}C$ values, and in two of the boreholes there is a single outlier near the bottom of the hole.

The ^{14}C activity of gas in borehole SD-12 (Table 5.3-55) is similar to that of UZ-1 in that there is a decrease from surface to depth, but with some irregularly large deviations from the trend (Figure 5.3-172). Another striking difference is the markedly lighter carbon in SD-12 and the strong correlation between $\delta^{13}C$ and ^{14}C activity. Rousseau et al. (1997) noted a gas-pressure deficit in the two deepest stations of SD-12 (both within the CHn) which they postulated could be due to the presence of hydrocarbons. The oxidation of hydrocarbons, such as methane, could account for the anomalously light carbon in the borehole, and assuming that the methane is more than 50,000 years old, it would account for the low-percent modern carbon. Because the postulated methane would be affecting both carbon isotopes, the two isotopic compositions would tend to be correlated. In addition, tests conducted on all instrument stations in borehole SD-12 in January 1997 revealed that there is a leak between station C (third from the bottom in Figure 5.3-172) and the open fiberglass support pipe (Rousseau et al. 1997), which probably caused gas samples from this station to be contaminated with atmospheric air.

The smooth trends in ^{14}C activity within borehole SD-12 are consistent with the data obtained from borehole UZ-1, indicating gas transport by a diffusion mechanism (Yang et al.1998), as previously concluded for borehole UZ-1. However, the large fluctuations toward greater percentages of modern carbon may represent a fracture-flow component. In the bottom of the hole, that component could be modern atmospheric, given the sharp increase in both ^{13}C and ^{14}C. Alternatively, the gas-pressure deficit in the deepest two stations in borehole SD-12 (Rousseau et al. 1997) may have caused atmospheric air to enter the rock mass surrounding these two stations while the borehole was standing open just prior to instrumentation (Yang et al.1998).

5.3.4.3 Movement of Water in the Unsaturated Zone

This subsection summarizes estimates of percolation flux that have been made for Yucca Mountain and vicinity, and describes evidence concerning the flowpaths that water takes as it moves from the ground surface to the water table. Included in the discussion are: the inferred spatial patterns in percolation flux; the role that discrete pathways play in controlling the movement of water in the unsaturated zone; the implications of perched water for flow directions and magnitudes; and the extent of fracture-matrix interactions.

5.3.4.3.1 Percolation Flux

Percolation flux is defined as the volumetric rate of water movement per unit area at depths below which the rate of water removal by evapotranspiration is insignificant. Percolation flux is an important variable that directly impacts water seepage into and drainage from the potential waste emplacement drifts, and thereby influences waste package corrosion rates, and waste dissolution and migration rates. A knowledge of its magnitude, spatial and temporal variability, and mode of delivery as fracture- or matrix-derived flow is therefore important for the design of canisters and engineered barriers, repository layout, and assessments of waste mobility and migration. A number

of methods have been employed to assess percolation rates. The basis for these methods, associated uncertainties, and estimates of percolation rates derived by these methods are discussed below.

5.3.4.3.1.1 Infiltration Modeling

If water movement from the ground surface to the depth of the potential repository horizon is predominantly vertical, and the rate at which moisture is removed by processes such as shallow gas-phase circulation is relatively insignificant, then the magnitude and distribution of net infiltration becomes the dominant control on the magnitude and distribution of percolation at depth. The advantage of this approach, relative to others described below, is that geographic coverage of the mountain is possible through a relatively large number of shallow boreholes drilled specifically for the purpose of monitoring changes in the moisture state of the unconsolidated surficial material and shallow bedrock. The drawbacks of this approach include: the relatively large-temporal variability in processes affecting net infiltration; the potential for large-spatial variability in net infiltration due to focussed flow in large and small-scale surface depressions; and other factors such as variability in the depth of alluvial cover (see Subsection 5.3.4.1), and the short period (15 years) relative to the intended containment period (>10,000 years) over which time-dependent processes such as precipitation have been monitored. An additional source of uncertainty is that the distribution of percolation at depth must be estimated using a numerical model which employs the estimated infiltration distribution as input, and so is subject to uncertainty in parameters of that model (see Subsection 5.3.4.4.4).

As described in Subsection 5.3.4.1, several estimates of the spatial distribution of net infiltration have been made using a variety of approaches and assumptions. The earliest map of net infiltration (Flint, A.L. and Flint 1994) was based on an assumed unit gradient and the effective hydraulic conductivity of the rock at a depth at which flow was estimated to be steady, based on the absence of changes in moisture content over several years of monitoring with a neutron-moisture probe. Net infiltration ranged from 0.02 mm/year in areas underlain by the welded Tiva Canyon Tuff to 13.4 mm/year in areas underlain by the nonwelded Paintbrush Tuff. The higher infiltration areas occurred where nonwelded tuffs were exposed by erosion in the northwest-trending drainages between Drill Hole and Yucca Washes. Average infiltration was 1.4 mm/year over the model area, which coincided with the model domain of Wittwer et al. (1995). Because the existence of transient flow in the fractures in the shallow bedrock was not reflected by the matrix-effective hydraulic conductivity, the estimates of net infiltration were probably underestimates of the actual net infiltration, particularly in the areas underlain by welded tuffs.

Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press) used an empirical relationship developed by Maxey and Eakin (1949) to estimate infiltration over the model domain of D'Agnese et al. (1997). The empirical model of Maxey and Eakin (1949) relates average net infiltration of small watersheds to the average precipitation falling on the watershed, using estimates of the baseflow component of surface outflow and subsurface drainage from the watershed as a measure of the net infiltration. Hevesi and Flint (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press) applied this empirical relation developed from individual watersheds on a point-by-point basis to obtain a spatially distributed map of infiltration over an area approximately 100,000 square

km. Because precipitation is correlated with elevation, estimates of net infiltration also were strongly elevation dependent.

Hudson and Flint (1996) used neutron-moisture logs recorded at 69 boreholes over an 11-year period to develop estimates of shallow infiltration over an area of 230 km^2 centered around Yucca Mountain. At the boreholes from which the moisture logs were collected, a correlation was established between shallow infiltration and precipitation, depth of alluvial cover, and a variable that indicated the possible presence of the borehole in a channel. Net infiltration at individual boreholes was estimated from the neutron logs as the moisture content changes occurring below a 2-m depth in bedrock. Shallow infiltration in areas between the boreholes was determined from the correlation and estimates of the distribution of soil thickness and precipitation. The calculated average shallow infiltration over the entire modeled area was 11.6 mm/year, with the upland areas within the central block of Yucca Mountain receiving 20 to 30 mm/year and the lower elevation areas within the central block generally receiving 10 to 20 mm/year.

A water-balance approach was applied by A.L. Flint et al. (1996) to estimate net infiltration over an area of approximately 230 km^2. This approach attempted to compensate for the potential temporal bias introduced by the relatively short period when neutron moisture logs were made by developing a stochastic rainfall model that was applied for periods of up to 100 years. Net infiltration was calculated using the simulated precipitation and a moisture budget approach which calculated moisture gains due to precipitation, and moisture losses from the soil zone due to evaporation, root uptake, and drainage. The precipitation and moisture budgets were calculated on a daily basis. Runoff was not explicitly modeled, but was treated in an ad hoc manner by adding approximately 30 percent of annual average precipitation back into the channels. Average net infiltration was as little as 0 mm/year where alluvial thickness was 6 m or more, and over 80 mm/year at high elevations on north-facing slopes with thin alluvial cover and permeable bedrock. Average infiltration over the model area was 3.2 mm/year, although on a year-to-year basis, average net infiltration could be 0 mm/year in dry years to over 20 mm/year in those years where average precipitation exceeded 300 mm.

The average net-infiltration values determined by A.L. Flint et al. (1996) for different areas centered on boreholes at Yucca Mountain and vicinity are listed in Table 5.3-56. Based on 30 x 30 m estimates, infiltration rates are relatively high (>5 mm/year) for boreholes located along Yucca Crest (G-3, H-3, H-5), upper Drill Hole Wash (UZ-1, G-1), some smaller washes (H-4 in Antler Wash), and at WT-12, located on an alluvial fan blanketing the slope of East Ridge. Two additional boreholes (G-2, WT-18) located on ridgetops north of the central block have net-infiltration rates of approximately 3 mm/year. Other boreholes at Yucca Mountain have infiltration rates that are zero or a few mm/year, depending on the area for which the average net infiltration was calculated.

5.3.4.3.1.2 Temperature and Heat Flow

Because water consumes heat as it moves from cooler, shallow depths to warmer deeper environments, borehole temperature profiles are potentially a sensitive indicator of the percolation flux. The use of temperature data to estimate percolation flux has several advantages over other methods that have been used to estimate flux. Unlike isotope data collected from the unsaturated

zone, which reflect fluid velocities and require assumptions about the effective porosity along the flowpath to obtain fluid flux, temperature data are directly sensitive to the mass flux, irrespective of fluid velocity. Because temperature data reflect the total fluid flux, including both the fracture and matrix flow components, no assumptions about poorly known parameters such as those that describe fracture-matrix interaction are necessary, as has been the case where attempts have been made to determine percolation flux from matrix-saturation or matrix-isotope data (see, for example, Bodvarsson and Bandurraga 1996; Fabryka-Martin, Wolfsberg et al. 1997; Fabryka-Martin, Turin et al. 1998). Furthermore, thermal conductivity is a linear function of saturation, and therefore subject to less uncertainty than other saturation-dependentquantities, such as hydraulic conductivity, that vary logarithmically with saturation and are therefore extremely sensitive to measurement errors in saturation or in the saturation-dependent functional relationships, as well as the degree of mismatch between modeled and observed saturation profiles.

Early attempts to estimate percolation flux (Bodvarsson and Banduragga 1996; J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) using temperature data from Yucca Mountain applied a simplified version of a heat-balance equation which relates the conductive heat fluxes at two elevations z_2 and z_1 ($z_2>z_1$) to each other and to the liquid, gas and diffusive vapor fluxes by differencing the transport of sensible and latent heat flux into and out of a control volume. Assuming steady, one-dimensional vertical flow of gas, liquid, and heat, that relationship can be expressed as

$$(-\lambda\frac{\partial T}{\partial z})_2 = (-\lambda\frac{\partial T}{\partial z})_1 - q_l c_l \Delta T - q_g(c_g \Delta T - H_v(\chi_{v_2} - \chi_{v_1})) - H_v[(-D_v\frac{\partial \rho_v}{\partial T}\frac{\partial T}{\partial z})_2 - (-D_v\frac{\partial \rho_v}{\partial T}\frac{\partial T}{\partial z})_1]$$

(Eq. 5.3-6)

where:

λ = thermal conductivity, J s^{-1} °C^{-1} m^{-1};

T = temperature, °C;

z = elevation, with z positive upwards, m;

q_l = liquid flux, kg s^{-1};

c_l = the specific heat of the liquid water (4.187×10^3 J kg^{-1} °C^{-1});

ΔT = the temperature difference between elevations z_2 and z_1, °C;

q_g = gas flux, kg s^{-1};

c_g = the specific heat of the bulk gas (1.1×10^3 J kg^{-1} °C^{-1});

H_v = the heat of vaporization (2.45×10^6 J kg^{-1});

χ_{v2} and χ_{v1} = the vapor mass fractions of the gas at elevations z_2 and z_1, kg kg^{-1};

D_v = the effective diffusion coefficient for vapor that accounts for the effects of tortuosity, partial saturation and an empirical vapor enhancement factor, m^2 s^{-1};

ρ_v = the vapor density, kg m^{-3}; and subscripts 2 and 1 indicate that quantities are estimated at elevations z_2 and z_1.

The term on the left-hand side of Equation 5.3-6 is the conductive heat flux at elevation z_2. The terms on the right-hand side of Equation 5.3-6 are, in order of appearance:

- The conductive heat flux at z_1
- The change in the conductive heat flux between z_1 and z_2 that results from the movement of liquid water
- The change in conductive heat flux between z_1 and z_2 that results from the advective movement of gas, including a term that accounts for the transfer of latent heat by vapor transported advectively by the gas phase
- The change in the conductive heat flux that occurs as a result of the transfer of latent heat by vapor diffusion

Thus, Equation 5.3-6 relates the conductive heat flux, which is the quantity calculated from the temperature profiles, to other more difficult-to-measureterms in the heat-balance equation. Through comparison with numerical simulations, where all terms can be quantified through the model output, Equation 5.3-6 can be an accurate and efficient means of estimating the percolation flux. However, assuming that the gas and vapor heat transport terms can be ignored, the main difficulty in applying Equation 5.3-6 to estimate the percolation flux has been that reliable estimates of the conductive heat flux at individual boreholes are generally available only at a single stratigraphic elevation–within the Topopah Spring Tuff. As described in Subsection 5.3.4.2.3, heat fluxes from the saturated zone generally appear to be affected by flow within or along the borehole, as well as by horizontal flow within the saturated-zone formations. Additionally, temperature gradients across the nonwelded units within the unsaturated zone are poorly defined because of their small thickness compared with the relatively large spacing of the unsaturated-zone temperature measurements. And, temperature gradients within the Tiva Canyon Tuff are affected by convective gas movement and vapor-transport processes that make application of Equation 5.3-6 more difficult compared with deeper formations, where complexities related to gas and vapor transport processes can more safely be ignored.

To circumvent some of the uncertainties associated with applying Equation 5.3-6, percolation fluxes were estimated for a subset of the boreholes for which temperature data were available using TOUGH2 (Pruess 1991). The subset of boreholes selected for analysis, which included a#1, G-3, G4, H-1, H-3, H-5, WT-2, WT-12, WT-18, and UZ-1, were chosen on the basis of data density, apparent data quality, and their geographic distribution. The observed temperature profiles at these boreholes were simulated numerically and the percolation flux assumed in the model was varied

until a visual best-fit match between the observed and simulated profiles was obtained. Average annual ground-surface temperatures estimated for individual boreholes and measured water table temperatures formed the upper and lower boundary conditions for these models. A one-dimensional model with a 5 m vertical grid spacing was created for each borehole examined. Borehole stratigraphic information was obtained from stratigraphic logs compiled by Spengler (USGS) and Buesch (USGS) (written communication, August 1997), and additional information on the distribution of lithophysal and zeolitic intervals, as well as welding transition zones, was obtained from borehole geophysical logs developed by Nelson (1994, 1996a).

Thermal conductivities for individual stratigraphic units were estimated for completely saturated and completely dry conditions using the average porosity of the unit, as determined by L.E. Flint (1998), and the regression relation developed by Rautman (1995). Thermal conductivities at intermediate saturations were determined by linear interpolation of the dry and saturated thermal conductivities through an option available within TOUGH2. In some boreholes where lithophysal zones were exceptionally well developed (H-5, UZ-1, WT-18, a#1, WT-12), the saturated and dry thermal conductivities of the lithophysal zones that were estimated from matrix porosity and the Rautman (1995) thermal conductivity model were reduced by a factor equal to the lithophysal porosity, as estimated from geophysical logs. The adjustment in thermal conductivity of the lithophysal units was made to compensate for the reduced cross-sectional area available for thermal conduction due to the presence of the air-filled lithophysae.

Within the Topopah Spring Tuff, the non-fillable porosity associated with lithophysal cavities is typically 15 to 20 percent in the upper lithophysal zone and 10 to 15 percent in the lower lithophysal zone, but up to 40 percent in the lithophysal zones at borehole H-5. Matrix hydraulic parameters of individual units used in the model were those determined by L.E. Flint (1998) (Subsection 5.3.3.1.1). The enhancement of permeability due to fracturing was treated by manually altering the permeability of individual units in the TOUGH2 input in such a way that the assumed percolation flux was transmitted at a high-matrix saturation (>90 percent) through those units that required secondary permeability to transmit the flux.

Average annual ground surface temperatures at the selected boreholes were often poorly defined because temperature measurements above depths of 100 feet were lacking, and/or shallow temperature profiles may have been subject to seasonal variations that are not represented in one-time measurements. Shallow temperatures may also have been affected by gas-phase heat transport processes that were not included in the model, which assumed a passive-gas phase. For these reasons, ground surface temperature at individual boreholes were considered to be a fitted parameter identified as that temperature which allowed the optimum fit to the observed borehole temperature profile. Assumed surface temperatures were subject to certain qualitative constraints. Because ground surface temperature is affected by air temperature, vegetative cover, and incoming solar radiation, boreholes at similar elevations, with similar slope aspects and vegetative cover, can be expected to have similar ground surface temperatures. Similarly, a borehole at high elevation, with a north-facing slope aspect and extensive vegetation cover would be expected, at least qualitatively, to have a cooler surface temperature than a borehole at a lower elevation, located on a south-facing slope that had no vegetation cover.

Other model assumptions also require some additional comments. The model is one-dimensional and assumes the steady flow of both liquid water and heat. Because the model is one-dimensional, it cannot address the effects of topography on heat flow, nor can it directly consider the effects of lateral water flow. Two-dimensional simulations of heat flow presented in J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) showed that for conduction dominated systems, temperature contours tended to be a subdued reflection of topography. Heat flow was perpendicular to these temperature contours, and, therefore, heat flux increased with elevation under washes due to the convergent nature of flow and decreased with elevation under ridgetops due to the divergent nature of flow. However, these same simulations showed that the effects of topography were most pronounced in the Tiva Canyon Tuff, and that the effects of topography appeared to be negligible in the PTn and deeper units. With regard to the lateral movement of water, it is currently believed that although site-wide lateral flow in the PTn and along perching layers in the deep unsaturated zone is unlikely because faults would disrupt the lateral continuity of these units, lateral flow across distances comparable to fault spacing (tens to hundreds of meters) is possible.

Although the effects on the temperature profiles of a vertically variable percolation flux could be mimicked in the models through the inclusion of water source or sink terms at stratigraphic contacts, this was not done, except at UZ-1 (where the occurrence of perched water is well documented), to keep the number of adjustable parameters to a minimum. Therefore, the percolation fluxes identified in this study should be qualified as those depth-invariant fluxes that best reproduce the observed borehole temperature data. Similar comments pertain to the assumption that water flow is at steady-state. Evidence from the climate record indicates that average annual temperatures during past-pluvial periods may have been as cool as 5 to 7 ^{0}C and that up to 80 percent of the past 800,000 years may have been cooler and wetter than today's climate (Forester et al. 1996). Therefore, to examine the possibility that today's temperature profiles may be still responding to cooler surface temperatures of the past rather than present-day percolation rates, a simulation was performed using the model developed for borehole H-3 which examined the effect of a step change in surface temperature from 7.5 to 17.5°C on the steady- temperature profile developed under the cooler boundary temperature. The simulation showed that a new steady-state with the 17.5°C surface temperature was essentially established after approximately 8,000 years. Because it is generally believed that the relatively warm and dry Holocene climate was established at least by 8,000 years before present, it appears that any relic effect from past, cooler climates on present-day subsurface temperatures at Yucca Mountain probably is minimal.

The simulated temperature profiles and measured temperature data are shown in Figures 5.3-173a through 5.3-173j. After determining the ground surface temperature and percolation flux that allowed the best fit between the simulated and observed temperature profiles at a particular borehole, percolation flux was varied plus or minus a relatively small amount to determine the sensitivity of the temperature profile to changes in the flux. The robustness of the estimate can be qualitatively evaluated based on the deviations of the best-fit simulation profile from the measured data, and the sensitivity of this profile to changes in the flux.

The data and best-fit simulated profiles for G-4 are shown in Figure 5.3-173a. Borehole G-4 is located in upper Coyote Wash within the potential repository outline. G-4 is one of a few boreholes for which replicate and continuous data for the unsaturated zone are available. The variability in the upper part of the profile exhibited by data from different measurement dates indicates the possible effects of seasonality on shallow temperatures, and highlights the difficulty in choosing a meaningful ground surface temperature when data from only a single-measurement date are available, as is the case with most borehole temperature data from the unsaturated zone. The data between the water table (730 m elevation) and 1,140 m is well reproduced using a percolation flux of 10 mm/year; obvious departures from the profile are apparent using fluxes of 5 or 15 mm/year. The data above 1,140 m, which includes the uppermost Topopah Spring Tuff and higher units, are poorly matched by the model for any assumed percolation flux. However, the data between 1,180 m and the ground surface for different measurement dates are also quite variable.

Temperature data for borehole WT-2, located in WT-2 Wash within the potential repository boundary, is shown in Figure 5.3-173b along with the temperature profiles simulated using percolation fluxes of 0.1 to 2.0 mm/year. The simulated temperature profiles are relatively insensitive to percolation flux over this relatively narrow range, and all the simulated profiles provide a reasonable match to the data. From visual inspection alone, the optimal match appears to be provided by the simulated profile produced using 0.5 mm/year.

Temperature data for borehole H-3 are shown in Figure 5.3-173c along with the simulated temperature profiles for 5, 10 and 15 mm/year. Borehole H-3 is located on Yucca Crest approximately 0.5 km south of the potential repository boundary. The simulated profile using a percolation flux of 10 mm/year provided an excellent match to the data except for the shallowest data point. The obvious deviation between the simulated profiles generated with 5 and 15 mm/year from the measured profile indicates that the 10 mm/year estimate is fairly well constrained by the data.

Temperature data for borehole G-3 are shown in Figure 5.3-173d along with simulated temperatures using 5, 10 and 15 mm/year percolation fluxes. Borehole G-3 is also located on Yucca Crest, approximately 1.2 km south of H-3. The measured data at G-3 agrees well with the data at nearby H-3, except for the uppermost two data points, which are several degrees cooler than similarly located data at H-3. Because the topographic position and geologic conditions at the two boreholes are also very similar, and because a reasonable match is obtained between the deeper data and the simulated profile using the same ground surface temperature (17.5°C) and percolation flux (10 mm/year) as at H-3, the departure of the two uppermost points at G-3 from the simulated profile is attributed to measurement error. The data above 1,100 m appears to be fit best by the simulated profile that was generated using a 15 mm/year percolation flux, whereas the data below 1,100 m appears to be best fit using a 5 mm/year percolation flux. However, the temperature profile generated using 10 mm/year provides the best overall fit to the data.

The temperature data for borehole H-5, located on Yucca Crest about 3 km north of H-3 within the potential repository, is shown in Figure 5.3-173e along the simulated temperature profiles obtained using percolation rates of 15, 20, and 25 mm/year. The measured temperature profile within the Topopah Spring Tuff (approximately 962 to 1,315 m elevation) exhibited breaks in slope that were

not evident in the profiles presented previously. Examination of borehole geophysical logs indicated exceptionally well-developed lithophysal zones (with lithophysal porosity of approximately 40 percent) were present in the lower nonlithophysal zone between elevations of 1,098 to 1,138 m, and in the upper lithophysal zone between 1,174 and 1,266 m elevations. The simulated profiles shown in Figure 5.3-173e were generated by reducing the thermal conductivity estimated from matrix porosity and the Rautman (1995) regression by 40 percent within these intervals. The match between the measured profile and simulated profile generated using 20 mm/year is good, and indicates that the breaks in slope of the measurements within the Topopah Spring Tuff are probably reflecting stratigraphically controlled variations in thermal conductivity and are not simply measurement error. Simulations run with assumed percolation rates of 15 and 25 mm/year resulted in significant separation of the measured and simulated temperature profiles. The percolation flux of 20 mm/year estimated for H-5 is the highest of the 10 boreholes examined.

The temperature data for H-1 are shown in Figure 5.3-173f along with simulated temperature profiles obtained percolation fluxes of 5, 10, and 15 mm/year. H-1 is located in shallow alluvium on the edge of Drill Hole Wash at the toe of Tonsil Ridge and near the projected trace of the southwest splay of the Drill Hole Wash fault. The borehole is located approximately 100 m north of the potential repository boundary. Data from two measurement dates at H-1 generally agree with each other to within 0.1 to 0.3°C. Lithophysal porosity at H-1 was estimated from geophysical logs to be 18 percent within the upper lithophysal zone of the Topopah Spring and approximately 12 percent in the lower lithophysal zone. These values are typical of the values estimated at the boreholes that remain to be discussed. The shallow data appears to be best matched by the 15 mm/year flux and the deeper data by the 5 mm/year flux, but the temperature profile generated with the 10 mm/year flux again provides the best overall fit to the data.

Temperature data for borehole a#1 are shown in Figure 5.3-173g along with temperature profiles generated using percolation fluxes of 5, 7.5, and 10 mm/year. Borehole a#1 is located in lower Drill Hole Wash near the main channel and near the projected intersection of the southwest trace of the Drill Hole Wash fault with several north trending faults of the so-called imbricate fault zone. Borehole a#1 was one of several boreholes for which continuous temperature measurements were made in the unsaturated zone in a water-filled standpipe (Subsection 5.3.4.2.3). The data for all but the interval between the 1,160 and 1,130 m elevations (corresponding to the crystal-rich nonlithophysal zone of the Topopah Spring Tuff (Tptprn) and units of the PTn) is bounded by the temperature profiles generated with 5 and 10 mm/year. The temperature profile generated with a percolation flux of 7.5 mm/year appears to provide the best overall fit to the data.

Borehole WT-18 is located on Bleachbone Ridge, north of Pagany Wash and approximately 800 m northeast of H-1. The temperature data for this borehole are shown in Figure 5.3-173h along with the temperature profiles that were simulated using percolation fluxes of 10, 13, and 15 mm/year. The observed temperature gradient across the PTn (approximately 1,123 to 1,244 m) is not well matched by any of the these simulations, and appears to be more consistent with a lower flux than that required to match the deeper data. Below the elevation of the PTn, the data were well matched using a percolation flux of 13 mm/year and the simulations began to deviate from the measured profile when percolation was varied even a few mm/year.

Borehole WT-12 is located on an alluvial fan that blankets the toe of East Ridge near Ambush Pass, just west of Busted Butte. The temperature data for WT-12 are shown in Figure 5.3-173i along the temperature profile simulated using a percolation flux of 1.0 mm/year. The estimated lithophysal porosities for this borehole, based on geophysical logs, were 10 percent in the upper lithophysal zone of the Tiva Canyon Tuff (Tcpul), 20 percent in the upper lithophysal zone of the Topopah Spring Tuff (Tptul), and 5 percent in the lower lithophysal zone of the Topopah Spring Tuff (Tptpll). The thermal conductivities of these units were reduced a corresponding amount. The data are matched very well by the simulation below a 950 m elevation. As with some other boreholes, the temperature gradient across the PTn (approximately 960 to 975 m elevation) is poorly approximated by the model.

Both Sass, Lachenbruch et al. (1988) and J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164)reported temperatures for borehole UZ-1, which is located in Drill Hole Wash approximately 200 m north of the potential repository outline and near the projected intersection of short north-south trending and east-west trending faults. The temperature data agree with each other reasonably well, but the temperature data reported by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) are approximately 0.5°C warmer than the data of Sass, Lachenbruch et al. (1988) throughout the Topopah Spring Tuff (Figure 5.3-173j). Because drilling was halted after perched water was encountered at the base of the Topopah Spring Tuff, water table temperatures were not available. The water table temperature at UZ-1 was estimated from a map of water table temperatures in Fridrich et al. (1994) to be 31°C. The reductions in thermal conductivity due to the presence of lithophysal cavities were assumed to be the same as those estimated for nearby H-1. Attempts to match the temperature profile at UZ-1 using a depth-invariant percolation flux failed to produce an acceptable match. The presence of perched water at UZ-1 indicates that the percolation flux may not be constant with depth. The simplest scenario that explains the perched water is one in which the vertical flux through the Topopah Spring Tuff exceeds that seeping through the basal vitrophyre of the Topopah Spring Tuff and underlying layers, with the excess flowing laterally across the top of the vitrophyre. From chloride mass-balance arguments, the vertical seepage through the perching layers was estimated to be approximately 5 mm/year (see Subsection 5.3.4.3.1.4), based on the assumption that all water in the Calico Hills Tuff arrived as a result of downward seepage of the matrix pore water and perched water from the overlying Topopah Spring Tuff, and not as a result of lateral flow. The best match of the Sass, Lachenbruch et al. (1988) data for UZ-1 was obtained using a 15 mm/year flux from the ground surface to the basal vitrophyre, and a 5 mm/year flux between the basal vitrophyre to the water table (Figure 5.3-173j). The reduction in flux in the model was achieved by imposing a sink term with a magnitude equivalent to 10 mm/year at the node corresponding to the top of the basal vitrophyre. Due to the larger number of adjustable parameters involved in matching the UZ-1 temperature data, and the larger number of assumptions involved, the percolation fluxes estimated for UZ-1 clearly have a larger degree of uncertainty that the fluxes estimated for other boreholes.

5.3.4.3.1.3 Isotopic Evidence of Fluid Flow and Percolation

Weeks (written communication, presentation at the Unsaturated Zone Flow Model Expert Elicitation Workshop, February 1997) applied a model presented in Cook and Solomon (1995) for the simultaneous diffusive transport in the gas phase and convective transport in the liquid phase to selected ^{14}C data collected from boreholes at Yucca Mountain (see Subsection 5.3.4.2.4). The model assumes local equilibrium between the phases, so that coupled transport can be treated using the form of the standard transport equation for single-phase flow, with appropriate redefinition of the coefficients.

The equation used by Weeks (written communication, presentation at the Unsaturated Zone Flow Model Expert Elicitation Workshop, February 1997) is

$$\frac{\partial C_g}{\partial t} = \frac{D^*}{\theta^*}\frac{\partial^2 C_g}{\partial z^2} - \frac{q^*}{\theta^*}\frac{\partial C_g}{\partial z} - \lambda C_g \qquad \text{(Eq. 5.3-7)}$$

where: $\theta^* = \theta_g + \theta_l\rho_l K_w + (1-\theta_g-\theta_l)\rho_s K_w K_s$

$D^* = D_g + D_l\rho_l K_w$
$q^* = q_l\rho_l K_w + q_g$
$D_g = D_g^0\theta_g\tau_g$
$D_l = D_l^0\theta_l\tau_l + \alpha q_l/\theta$, and

θ is volumetric moisture content, dimensionless;
ρ is density, $M\,L^{-3}$;
q is Darcy flow velocity, LT^{-1};
λ is the radioactive decay constant, T^{-1};
C is concentration, ML^{-3};
K is the partitioning coefficient, with K_w having units of L^3M^{-1} and K_s dimensionless;
D is the diffusion coefficient, L^2T^{-1};
τ is tortuosity, the displacement distance divided by the path length, dimensionless;
Superscript 0 indicates the free-fluid diffusion coefficient; and
Subscripts g, l and s indicate gas, liquid and solid phases, respectively.

This model was used to match the observed ^{14}C activities in the gas phase or perched water at eight boreholes (and pore water at UZ-6) by adjusting the Darcy flux (q) until a satisfactory match was attained. The percolation rates were described by Weeks (written communication, presentation at the Unsaturated Zone Flow Model Expert Elicitation Workshop, February 1997) as diffuse because rapid, highly channelized liquid flow could conceivably not exchange with the gas phase rapidly enough to maintain equilibrium assumed by the model, and therefore, the gas phase ^{14}C might reflect only the slower moving (diffuse) component of recharge. It was also assumed that no correction to the ^{14}C ages was necessary to account for gas-water-rock interactions, based on the comparison of the ^{14}C and $^{36}Cl/Cl$ data from perched water with relationships between ^{14}C and $^{36}Cl/Cl$ derived from pack-rat midden data and theoretical reconstructions of past variations in the production and

deposition of ^{36}Cl (see Figure 5.3-162). Estimated percolation rates calculated by Weeks (written communication, presentation at the Unsaturated Zone Flow Model Expert Elicitation Workshop, February 1997) are shown in Table 5.3-57. These rates might be more appropriately characterized as apparent percolation rates because they assume that the flowpath was vertical. Some of the deeper ^{14}C data, particularly those associated with perched water, may have flowed laterally along the perching layer, so that the flowpath was longer than that assumed by the one-dimensional model and the actual fluxes correspondingly higher. Also, where lateral flow has occurred, the calculated percolation fluxes may be reflecting the flux in locations updip from the borehole, rather than at the borehole itself. The calculated percolation fluxes are typically in the range of from 4 to 9 mm/year.

Chloride-36 has been used as an indicator of rapid transport within the unsaturated zone, and to a lesser extent, as an indicator of the flux (Fabryka-Martin, Wightman et al. 1994; Fabryka-Martin, Wolfsberg et al 1997; Fabryka-Martin, Turin et al. 1998). Chloride-36 is produced naturally in the atmosphere as a result of the interaction of argon isotopes and cosmic-ray secondaries, and by neutron activation of Cl in sea salt during testing of nuclear weapons at the Pacific Proving Grounds between 1952 and 1958. The present-day meteoric deposition rate of ^{36}Cl at Yucca Mountain is about 493 (±34) x 10^{-15} $^{36}Cl/Cl$ atoms (Fabryka-Martin, Wightman et al. 1994; Table 6-1). Chloride-36 production related to weapons tests resulted in peak rates of deposition that are estimated to be about 2 to 3 orders of magnitude greater than the present-day meteoric deposition rates. It has also been recognized from pack-rat midden studies that natural production of ^{36}Cl has varied in the past, primarily as a result of fluctuations in the intensity of the earth's geomagnetic field. Decreases in the natural production of ^{36}Cl between 15,000 and 10,000 years ago resulted in declines in $^{36}Cl/Cl$ deposition in the area of the Nevada Test Site from about 1,000 x 10^{-15} to the present-day levels of 493 x 10^{-15} $^{36}Cl/Cl$ (Fabryka-Martin, Turin et al. 1998, Figure 2-2). Higher rates of production may have resulted in elevated deposition rates over most of the last 40,000 years. Because ^{36}Cl has a half-life of 301,000 years, it was originally believed to be suitable for measuring the ages of water between about 50,000 and 1,000,000 years. However, the recent realization that past-deposition rates have been variable has imparted considerable uncertainty to the use of ^{36}Cl as an age-dating tool.

The center of mass of the bomb-related ^{36}Cl in the atmosphere occurred about 1958 to 1959. The elapsed time between these years and the time of sampling, and the depth of the center of mass of the bomb-pulse ^{36}Cl peak in the alluvial profiles, were used by Fabryka-Martin, Wightman et al. (1994) to estimate infiltration rates. The rates calculated from the ^{36}Cl distribution were considerably higher than those calculated from the chloride mass-balance method at the same boreholes. However, because most of the bomb-related ^{36}Cl remained in the zone of evapotranspiration, these flux rates were considered by Fabryka-Martin, Wightman et al. (1994) to be unrepresentative of the deep infiltration rates, which are more accurately estimated by the chloride mass-balance method using data from below the root zone.

Fabryka-Martin, Wightman et al. (1994) reported that bomb-pulse ^{36}Cl (then conceived of as >500 x 10^{-15} $^{36}Cl/Cl$) occurred in the PTn in at least three (and possibly four) of the five neutron boreholes investigated. Bomb-pulse levels of ^{36}Cl (now conservatively defined as >1250 x 10^{-15} $^{36}Cl/C$ by Sweetkind, Fabryka-Martin et al. 1997) occurred to depths of 25 m at UZ-N11, to a 57 m depth in UZ-N53, and through the PTn and into the top of the Topopah Spring Tuff at a depth of 77 m in

UZ-N55 (see Figure 5.3-174). The occurrences of bomb-pulse levels of ^{36}Cl occurred throughout the total depth of boreholes UZ-N11 and UZ-N55. It was speculated that the extremely high values reported for cuttings collected during drilling of N55 might have been contaminated by equipment previously used elsewhere at the Nevada Test Site. However, subsequent analyses of core samples that had not been in contact with that equipment have confirmed the validity of the original measurements reported by Fabryka-Martin, Wightman et al. (1994) (see Flint, A.L. et al. 1996). Both UZ-N53 and UZ-N55 are located on bedrock sideslopes with thin to no alluvium, and UZ-N55 is cut by a fault. Borehole UZ-N37, which was located in a channel, had $^{36}Cl/Cl$ values in excess of 1,100 x 10^{-15} throughout the upper 7.7 m of alluvium, and several values of approximately 850 x 10^{-15} at a depths of 54 and 67 m in the PTn. Much of the chloride in the alluvial profile, as well as the bomb-pulse ^{36}Cl peak, had apparently been flushed by a recent infiltration event. The only neutron-logging borehole that did not show bomb-pulse levels of ^{36}Cl in the PTn was N54 which was located on an alluvial terrace between N53 and N55. Bomb-pulse ^{36}Cl was confined to the upper 4 m or so in the alluvium at this hole. Bomb-pulse ^{36}Cl was absent from the alluvium below a few meters depth at UZ#16, which also was located on an alluvial terrace in WT-2 Wash. A local peak in the $^{36}Cl/Cl$ values of 1,030 x 10^{-15} just below the alluvium-bedrock contact suggested the possible occurrence of lateral flow from sideslopes along the alluvium-bedrock contact.

Many of the samples collected from these boreholes had $^{36}Cl/Cl$ ratios well below the meteoric value of 493 x 10^{-15}. Attempts to correct these ratios for rock chloride incorporated into the analysis during sample preparation were sensitive to the Br/Cl ratio assumed for the meteoric end member. The discrepancy between the very old water ages implied by some of the deeper $^{36}Cl/Cl$ ratios obtained from UZ-16, and relatively young ^{14}C ages, was shown to be the possible consequence of mixing of water of different ages, the substantial difference in the half-lives of ^{14}C and ^{36}Cl, and the different sensitivities of ^{14}C and ^{36}Cl to the inclusion of small amounts of bomb-pulse water.

Fabryka-Martin, Wolfsberg et al. (1997) summarized the results of sampling and analysis of ^{36}Cl from the Exploratory Studies Facility. The results of 112 isotopic analyses were available for 98 sampling stations in the Exploratory Studies Facility between stations 2+00 and 46+00 (200 and 4,600 m into the Exploratory Studies Facility), of which systematic samples collected every 200 m constituted approximately 20 percent of the database and the remainder from "feature-based" samples collected from fractures, faults, lithophysal cavities, breccia zones, and broken rock. Results concluded that the distribution of "fast paths," as evidenced by the distribution of bomb-pulse levels of ^{36}Cl, was controlled by two factors: the availability of water, and the occurrence of continuous fracture or fault flow through the PTn. Their modeling showed that for even small (1 percent of the input flux) amounts of bomb-pulse ^{36}Cl to arrive at the Exploratory Studies Facility horizon in under 50 years generally required percolation fluxes of at least 1 to 10 mm/year, depending on the characteristics of the secondary permeability assumed for the model layers corresponding to the PTn (Fabryka-Martin, Wolfsberg et al. 1997, Tables 8-3 to 8-5). Fabryka-Martin et al. considered the stratigraphy at two locations within the Exploratory Studies Facility, one where the thickness of the PTn was large (Exploratory Studies Facility station 35) and one where the PTn was relatively thin (Exploratory Studies Facility station 59). Generally, fracture flow through the PTn could be sustained more readily with increases in percolation flux and secondary permeability, and when fractures contributing to the secondary permeability were assumed to be sparse (thereby minimizing fracture-matrix interactions) and have large apertures. Increases

to the assumed water retention capacity of the fractures by adjusting poorly constrained parameters that described the fracture's capillary properties also resulted in more sustained fracture or fault flow through the PTn at lower fluxes. They also concluded that fluxes insufficient to produce fracture flow in the densely welded Tiva Canyon and Topopah Spring Tuffs resulted in water samples at the Exploratory Studies Facility that were very old, with predicted $^{36}Cl/Cl$ ratios that indicated significant decay (approximately 300 to 600 x 10^{-15}) and which were generally too small relative to those obtained for what were considered to be non-bomb-pulse samples (500 to 1,500 x 10^{-15}). The $^{36}Cl/Cl$ ratios for these non-bomb-pulse samples were generally consistent with those predicted using assumed percolation fluxes of 1 to 10 mm/year. Just as fluxes on the order 0.1 mm/year led to $^{36}Cl/Cl$ ratios at the Exploratory Studies Facility horizon that displayed considerable decay of the meteoric signal, fluxes greater than 10 mm/year resulted in short travel times that led to introduction of Holocene water with $^{36}Cl/Cl$ ratios lower (500 x 10^{-15}) than observed for the non-bomb-pulse samples. The samples with $^{36}Cl/Cl$ ratios between 500 to 1,500 x 10^{-15} were considered to be water that infiltrated during the Pleistocene when ^{36}Cl production was greater and stable Cl deposition lower, rather than Holocene water with a small component of bomb-pulse ^{36}Cl, based on the belief that the fracture characteristics required for sustained fracture flow through the PTn were typical only of fault zones. However, the assumption that samples with $^{36}Cl/Cl$ ratios between 500 to 1,500 x 10^{-15} initially had ratios of 1,500 x 10^{-15} implied that many of these samples were old enough to have undergone significant decay, and resulted in a distribution of calculated travel times that were either very recent (<50 years) or very old (100,000 to 1,000,000 years) with very few samples between 50 and 100,000 years (Fabryka-Martin, Wolfsberg et al. 1997, Figures 7-1 and 7-2). Because it is improbable that there has been more water movement during the past 50 years than in the previous 100,000 (which included a major ice age), it is likely that samples with $^{36}Cl/Cl$ values of 500 to 1500 x 10^{-15} contained water that infiltrated during the time when meteoric $^{36}Cl/Cl$ values were declining from 1,500 to 500 x 10^{-15} (approximately 10,000 years ago), or contained a mixture of bomb-pulse water and older, Holocene-age water. The latter interpretation implies either that fast flowpaths through the PTn are not restricted to mapped faults, or that lateral dispersion within the Topopah Spring Tuff spreads bomb-pulse ^{36}Cl over distance of hundreds of meters away from mapped faults through the PTn.

5.3.4.3.1.4 Chloride Mass Balance

The Cl^- mass-balance method uses the chloride concentration of pore, perched, or saturate d-zone water, the concentration of chloride in precipitation, and the average annual precipitation, to estimate the fraction of precipitation that escapes evapotranspiration and becomes net infiltration. The method assumes that the flux of Cl^- along a one-dimensional column of rock is constant and equal to estimates of the present-day chloride flux. The method has traditionally been applied to shallow alluvial materials where runoff and run on do not occur and where piston displacement can reasonably be assumed. Application of the method to sites where there may be significant runoff and run on, subsurface lateral diversion, and partitioning of water flow between fractures (little concentration of chloride due to evapotranspiration) and matrix (potentially significant degree of concentration of chloride due to evapotranspiration) remains to be thoroughly evaluated.

The chloride mass-balance approach estimates net infiltration using the relationship

$$PC_p = IC_i \quad \text{(Eq. 5.3-6)}$$

where: P is the average annual precipitation rate (170 mm/year; Hevesi, Ambos et al. 1996);

I is the infiltration rate (mm/year); and

C_p is the equivalent chloride concentration in precipitation resulting from the deposition of chloride both in precipitation and as dry fallout (0.62 mg/l; Fabryka-Martin, Wightman et al. 1994); and

C_i is the either the concentration of chloride of perched water, pore water squeezed from the rock matrix, or for alluvium, the pore-water chloride concentration estimated using the mass of leachable chloride and the estimated water content of the alluvium (mg/l).

On the basis of chloride concentration data from both alluvium and the nonwelded PTn, along with supporting information from neutron moisture logging and ^{36}Cl analyses, Fabryka-Martin, Wightman et al. (1993, 1994) concluded that the alluvium is, in most cases, a significant barrier to water movement and that infiltration is indeed higher where alluvium is thin or absent. Some channel boreholes, as well as some soil samples collected from sideslopes, displayed chloride profiles that suggested recent flushing of chloride from the soil profile. Fabryka-Martin, Wightman et al. (1994, Table 4-1) estimated water flux through the alluvium to be 0.02 mm/year at UZ-16 and N-54, which are located on alluvial terraces, and 0.5 mm/year at N-37, which is located in a channel. Pore water chloride concentrations from the PTn and the Calico Hills Tuff at UZ-16 resulted in estimates of percolation flux of 3.0 and 3.5 mm/year. The higher percolation flux estimates made for the deeper units at UZ-16 were attributed to the lateral influx of water that moved along the alluvium-bedrock interface or along other stratigraphic contacts. Pore water chloride concentrations from the PTn and Calico Hills Tuff at UZ-14 resulted in estimates of percolation flux of 1.2 and 5.9 mm/year for the PTn and Calico Hills Formation. Estimates of percolation flux in the PTn using the chloride mass-balance equation were 1.1 mm/year at UZ-4 and approximately 2 mm/year at UZ-5 (Fabryka-Martin, Wightman et al. 1994).

The assumption that chloride in the pore water reservoir is well mixed is probably reasonable for alluvium but needs to be evaluated further for the tuffs. The complex nature of flow in the unsaturated zone at Yucca Mountain is shown schematically in Figure 5.3-175. It is possible that relatively dilute water that has infiltrated rapidly through fracture pathways may be inadequately represented by matrix pore water because of incomplete mixing. If this is the case, matrix pore water samples might be biased toward the slower moving, more concentrated matrix component of flow; and percolation estimates based on these samples would constitute lower bounds on the actual percolation rates. In the PTn, some component of the flux can bypass the matrix as fracture or fault flow, as evidenced by the presence of environmental tracers in the Exploratory Studies Facility at concentrations indicating an origin related to nuclear weapons testing (see Subsection 5.3.4.2.4). In the Calico Hills Tuff, the matrix pore water concentration may not reflect the fracture component

of flux for similar reasons, and because some component of the fracture flow through the Topopah Spring Tuff may be moving laterally as perched water above the Calico Hills Tuff or the Tptpv3. This water may ultimately recharge the saturated zone through faults or fracture zones, or where the perching layer intersects the water table. In areas where perched water is being diverted, chloride concentrations in the Calico Hills Tuff would underestimate the flux through the overlying Topopah Spring Tuff. However, in downdip areas toward which perched water is flowing laterally and accumulating, local drainage could result in percolation flux estimates that are high relative to the actual percolation in the Topopah Spring Tuff.

Another complicating factor not typically encountered in applications of the chloride mass-balance method involves runoff. Surface runoff and run on, including shallow subsurface flow along the alluvium-bedrock interface, would result in a spatially variable chloride source term, with higher chloride application rates at lower elevations where run on occurs. Data on the chemical composition of surface runoff from the Yucca Mountain area indicate chloride concentrations in runoff range from 1.3 to 11 mg/l (Subsection 5.1.6). For illustration, if it is assumed that the representative chloride concentration in run-off is 5 mg/l, areas in which run on is focussed would receive the equivalent of the chloride derived from precipitation if run on averaged 21 mm/year. Where run on occurs, such as in washes, the basic chloride mass-balance equation will underestimate net infiltration because the chloride applied at the soil surface is underestimated if only the contribution from precipitation is considered.

Since the work of Fabryka-Martin, Wightman et al. (1994), additional chloride concentration data has been collected from deep unsaturated-zone boreholes for both pore water and for perched and saturated-zone water (Yang et al. 1996, 1998). The chloride data for boreholes UZ-16, UZ-14, SD-7, SD-9, SD-12, and NRG-7a are shown in Figures 5.3-176a to 5.3-176f. The average chloride concentrations within individual hydrostratigraphic or lithostratigraphic units, and for perched and saturated-zone water, are listed in Table 5.3-58 for these boreholes and borehole NRG-6. Based on television logs, perched water also occurred near the top of the densely welded, crystal-poor vitrophyre of the Topopah Spring Tuff (Tptpv3) at borehole SD-12 (Rousseau et al. 1997). Unfortunately, the perched water at borehole SD-12 was not sampled.

Figure 5.3-176a shows that the chloride concentration in groundwater bailed from UZ#16 is substantially less than that in the pore water of the adjacent matrix. This difference may be real, or it may reflect the fact that saturated-zone groundwater did not enter the borehole until a fracture was encountered several meters below the depth at which the core samples were obtained, whereupon the groundwater rose to the sampling depth of the cores. A similar phenomena was observed in the zone of perched water at UZ#14 (Figure 5.3-176b), where the perched water was considerably more dilute than pore water in the adjacent matrix. Figure 5.3-176e shows a decrease in chloride concentration with depth in the Calico Hills Formation at SD-12 that may be indicating the occurrence of lateral flow within the base of the unit.

It is apparent from both Figure 5.3-176 and Table 5.3-58 that at boreholes UZ-14, SD-7, SD-9, and SD-12, the chloride concentration in the Calico Hills Formation (Tac) is substantially less than that in the overlying PTn. It is also evident that the chloride concentration in the perched water at UZ-14,

NRG-7a, SD-7, and SD-9 is much less than that of the pore water from either the PTn or the Calico Hills Formation. The more dilute water in the lower unsaturated zone implies that:

- Water is bypassing the matrix of the PTn along fractures or faults, or where the PTn is absent in Solitario Canyon
- Water is moving through the PTn updip from these boreholes in high-flux areas for which no chloride data are yet available
- Pore and perched water in the lower unsaturated zone is a remnant from the Pleistocene, when infiltration rates were higher and chloride concentrations correspondingly lower

There is evidence that each of these processes may be a factor contributing to differences in the chloride concentrations of the perched water and pore water samples from the PTn and Calico Hills Formation. The absence of significant mixing with the PTn because of rapid flow along fractures or faults seems to be a reasonable explanation for at least some dilute water in the deep unsaturated zone, based on the presence of bomb-pulse $^{36}Cl/Cl$ in the Exploratory Studies Facility, and ^{3}H in the Calico Hills Formation and possibly perched water at NRG-7a (Subsection 5.3.4.2.4). Recent analyses of pore water samples taken from the PTn in the North and South Ramps of the Exploratory Studies Facility (Fabryka-Martin, Flint et al. 1997, 1998) have resulted in chloride concentrations values that are considerably more dilute (geometric means of 20 and 42 mg/l for the North and South Ramps, respectively) than values listed for many of the boreholes in Table 5.3-58, indicating that the siting of most boreholes (UZ#16, UZ-14, SD-9, SD-12, NRG-6) in washes or alluvial terraces may have introduced a spatial bias, either because washes are genuine locations of small net infiltration, or because run on to the washes results in more chloride deposited in the wash than in upland areas. However, even boreholes located on sideslopes (NRG-7a) and ridgetops (SD-7) appear to have high chloride concentrations in the PTn (although the number of pore water samples at these boreholes is small), and the relatively dilute average concentrations found in the North Ramp PTn samples are still considerably more concentrated than most of the perched-water samples. Therefore, additional processes also must be contributing factors. Uncorrected ^{14}C ages indicate a significant portion of perched water pumped from UZ-14 and SD-7 was derived from water that infiltrated between 8,000 and 12,000 years ago, or approximately at the Pleistocene-Holocene boundary. This dilute perched water may, therefore, be unrepresentative of the water that infiltrated during the more arid conditions that existed throughout most of the Holocene. Sonnenthal and Bodvarsson (1997) simulated the chloride distribution in the unsaturated zone using a spatially and temporally variable chloride source term at the ground surface that reflected the expected infiltration pattern during the Pleistocene and under the present-day climate. They achieved good qualitative agreement between the model results and the observed chloride concentration values.

The uncertainty regarding this interpretation of the dilute perched-water chloride concentrations centers around the magnitude of the ^{14}C age corrections that need to be made to the perched-water data, and whether the corrected ages indicate an origin during the Pleistocene or Holocene climate. Yang et al. (1998) applied a reaction-based correction using NETPATH and concluded that the corrected ^{14}C ages for the perched water were 2,150 to 2,650 years for the NRG-7a sample, 4,400 to 5,370 years for the sample from SD-9, and 5,260 to 6,260 years for a pumped sample from the

perched water at UZ-14. On the other hand, as discussed in Subsection 5.3.4.3.1.3, it can be argued that no correction to the ^{14}C ages is necessary to account for gas-water-rock interactions, based on the comparison of the ^{14}C and $^{36}Cl/Cl$ data from perched water with relationships between ^{14}C and $^{36}Cl/Cl$ derived from pack-rat midden data and theoretical reconstructions of past variations in the production and deposition of ^{36}Cl (Figure 5.3-162). Although the magnitude of the corrections required for the perched water ^{14}C ages is uncertain, it appears that pore water concentration from the Calico Hills Formation are not a relic of wetter conditions that may have existed during the Pleistocene. The ^{14}C activities of the shallow and deep pore water samples both indicate uncorrected ages of a few thousand years (Subsection 5.4.2.4, Figure 5.3-159). The similarity in uncorrected ^{14}C ages of the shallow and deep pore water samples implies that the differences in chloride concentrations between the PTn and Calico Hills Formations are related to differences in the amounts of evapotranspiration undergone by water recharged by different infiltration mechanisms or in different topographic settings, rather than to climate change.

Table 5.3-59 lists the apparent vertical percolation fluxes based on these average chloride concentrations in Table 5.3-58 and Equation 5.3-8. As discussed earlier, application of Equation 5.3-8 assumes that surface run on and runoff could be ignored, subsurface lateral flow was insignificant, the pore and fracture water reservoirs were well-mixed, and the surface chloride flux is constant through time. The effects of surface run on and runoff and of subsurface lateral flow were qualitatively discussed in the preceding paragraphs. Tyler et al. (1996) argued that the chloride deposition rates in Frenchman Flat, approximately 45 km east of Yucca Mountain have been constant over the last 100,000 years or so, based on the good agreement between the chloride ages of recharge events dated using the assumption of constant chloride deposition with wet periods dated using regional data on paleolake levels, ages of spring discharge deposits, pack-rat midden data, and other paleoclimatic indicators. The effects of partitioning precipitation arriving at the ground surface between fractures and matrix can be evaluated quantitatively with the expression

$$\frac{I}{P} = C_p \frac{C_m - \frac{P_m}{P}(C_m - C_f)}{C_m C_f} \qquad \text{(Eq. 5.3-9)}$$

where: I is net infiltration (mm/year);

P is total precipitation (mm/year);

P_m is the precipitation that enters the matrix at the ground surface (mm/year);

C is chloride concentration (mg/l) in precipitation, matrix pore water and fracture water, designated with subscripts p, m and f, respectively.

The apparent percolation fluxes given in Table 5.3-59 were calculated under the assumption that all of the chloride in precipitation (PC_p) was incorporated into the water sample (i.e., the fracture and matrix waters are well-mixed). In contrast, Equation 5.3-9 assumes that there is no mixing between

fracture and matrix water. Although the complete absence of mixing between fractures and matrix is unlikely, it is worthwhile to consider this case for the purpose of making a bounding calculation to counterbalance the percolation rates calculated with Equation 5.3-8, which assumes complete mixing. As discussed in the preceding paragraphs, the disparity in the chloride concentrations for the perched and matrix water (Table 5.3-58), and the large differences in the resulting flux estimates (Table 5.3-59) may be indicating incomplete chemical mixing between fractures and matrix at Yucca Mountain. In Equation 5.3-9, it is assumed that precipitation (and the chloride in precipitation) that entered and became concentrated in the matrix as a result of evapotranspiration, is unavailable to the fracture network, and vice-versa. From this perspective, the use of P (rather the P_m or P_f, the precipitation entering the fracture system) would result in the overestimation of the actual matrix or fracture flux. Figure 5.3-177 shows how the total (I), fracture (I_f) and matrix (I_m) fluxes are sensitive to the fraction of precipitation that originally entered and became concentrated in the matrix. The concentrations used in the calculations were C_m = 74.8 mg/l (the average of PTn pore-water concentrations listed in Table 5.3-58) and C_f = 6.4 mg/l (the average of the perched-water concentrations listed in Table 5.3-58).

Figure 5.3-177 shows that the fracture-water chloride concentrations (as represented by the average chloride concentrations for the perched water) may indicate fracture fluxes as great as 16.3 mm/year, or as little as 0.16 mm/year, depending on whether 99 percent of the water falling on the mountain entered the fractures or one percent of the water falling on the mountain entered the fractures. Similarly, the PTn chloride concentrations may indicate a flux as large as 1.4 mm/year if all water falling on Yucca Mountain entered the matrix (or if all fracture flow was captured by the PTn), or as little as 0.014 mm/year if only one percent of the precipitation entered the matrix and no fracture flow mixed in the matrix of the PTn. The maximum total flux of 16.3 mm/year is attained if it is assumed that essentially all (99 percent) of the precipitation enters the fracture system.

The flux-weighted concentration of chloride moving through the unsaturated zone at Yucca Mountain is difficult to estimate with certainty from the resident concentrations of water samples obtained from core or from perched water in the unsaturated zone because of probable departures of the one-dimensional assumption on a local scale, incomplete mixing between fractures and matrix, and the difficulty of estimating P_m and P_f directly. However, departures from the one-dimensional assumption are likely to be minimized, and spatial variations in chloride deposition rates due to runoff and run on are more likely to be averaged out, if data from a sufficiently large area are considered. Similarly, it might be possible to calibrate the average value of P_m and P_f by considering chloride concentrations in the shallow groundwater under Yucca Mountain. Chloride concentrations in the shallow groundwater may better approximate the flux-weighted chloride concentration of water moving through the unsaturated zone in a variety of fracture and matrix pathways. Unfortunately, extremely limited data exist on the water chemistry of the shallow saturated zone under Yucca Mountain. No groundwater samples have been taken from the upper few meters of the saturated zone except for those taken from UZ#16 (Table 5.3-58), where there is also some uncertainty as to whether the fracture and matrix-chloride concentrations are well mixed (Figure 5.3-176a). The chloride concentration of shallow groundwater from UZ#16 is relevant to the analysis of percolation flux through the central block because the ^{14}C activity of the groundwater sample is 42 pmc, which is approximately twice the typical value obtained for deeper groundwater at Yucca Mountain (Benson and McKinley 1985). The young uncorrected ^{14}C age (7,200 years) of this

sample, relative to deeper samples, indicates that the shallow groundwater sample from UZ#16 was probably locally recharged through the overlying unsaturated zone during the Holocene. Corrections to the apparent ^{14}C age to account for the possible incorporation of isotopically depleted mineral carbonate may make the ^{14}C age even younger. Furthermore, UZ#16 is downgradient from the western half of the central block where most of the boreholes in Table 5.3-58 are located, and therefore may reflect the integrated effects of local recharge in this area.

The shallow groundwater sample from UZ#16 has an average chloride concentration of 10.8 mg/l, from which a flux of 9.8 mm/year was calculated using Equation 5.3-9. Assuming that the data from UZ#16 represents the integrated effects of local recharge in upgradient areas, an average flux value of 9.8 mm/year through the western half of the central block implies that about 45 percent of the precipitation in this area enters the matrix and 55 percent enters the fractures (Figure 5.3-177). The average fracture flux between the ground surface and the depth of the perched water is about 9.1 mm/year and matrix flux through the PTn is about 0.7 mm/year. The fluxes implied by the chloride concentration of the perched water may therefore be considerably smaller than the values listed in Table 5.3-59, after the effects of incomplete mixing have been considered. This analyses also indicates that, on average, only about 7 percent of the total flux through the unsaturated zone at Yucca Mountain occurs through pores of the rock matrix of the PTn, with the remaining 93 percent presumably flowing through fractures and faults of the PTn.

In Table 5.3-59, the percolation fluxes through the Calico Hills Formation are generally slightly higher than those through the PTn at each of the boreholes, possibly indicating the partial uptake of dilute perched water or a greater fraction of the dilute fracture flow component of the flux. Possibly as a result of this uptake of dilute water, the average chloride concentration of pore water in the Calico Hills Formation is 32 mg/l, which is considerably less than that of the PTn. The fraction of the total flux (9.8 mm/year) that passes through the matrix of the Calico Hills (I_m) can be calculated by equating the total chloride flux estimated from the shallow saturated zone groundwater sample (or equivalently, with PC_p) with the flux-weighted concentrations of the pore and perched water:

$$(10.8 \text{ mg/l})(9.8 \text{ mm/year}) = I_m\,(32 \text{ mg/l}) + I_f\,(6.4 \text{ mg/l})$$

Substituting (9.8 mm/year - I_m) for I_f gives I_m = 1.7 mm/year, which is roughly half of the apparent flux given in Table 5.3-59. This flux is nearly identical to the geometric mean hydraulic conductivity of 1.4 mm/year estimated for the zeolitic Calico Hills Formation by L.E. Flint (1998) (Subsection 5.3.3.1.1). These results imply that, on average, only about 17 percent of the 9.8 mm/year unsaturated-zone flux passes through the matrix of the Calico Hills Formation. The remainder is recharged to the saturated zone through fractures and faults. (Possible discrete pathways through and around the Calico Hills Formation are described in more detail in Subsections 5.3.4.3.2.2 and 5.3.4.4.4.3.5.)

The foregoing analysis has been based on extremely limited data and would benefit greatly from the collection of additional chloride data, particularly from the shallow saturated zone. The available chloride data can be interpreted to suggest an unsaturated-zone flow system that is dominated by fracture flow, and in which the fracture water has only limited interaction with water in the matrix pores. However, this conclusion ignores the possibilities that 1) average chloride concentrations in

the PTn have been biased toward more concentrated values by siting most of the boreholes in washes, and that elsewhere, along ridges and sideslopes, the PTn chloride concentrations are much lower (Fabryka-Martin, Flint et al. 1997), and 2) the differences in PTn and perched-water chloride concentrations reflect the effects of climate change rather than preferential flow through the PTn (Sonnenthal and Bodvarsson 1997). Definitive conclusions regarding the correct conceptual model must await the results of additional sampling of pore water chloride concentrations in upland areas, as well as further analyses of the chemical reactions that resulted in the perched-water chemistry, and the impact these reactions may have had on perched water ^{14}C ages.

The saturated-zone groundwater chemistry can be used to establish an upper limit on the areally averaged percolation rate through the unsaturated zone at Yucca Mountain under the present climate. Typical chloride concentrations of groundwater beneath Yucca Mountain are approximately 7 mg/l (Benson and McKinley 1985), whereas present-day chloride deposition rates from precipitation are estimated to be approximately 0.62 mg/l. Assuming the constancy of chloride deposition rates over time, and that all of the groundwater beneath Yucca Mountain resulted from percolation through Yucca Mountain itself, the increase in chloride concentration in groundwater by a factor of 11.7 relative to precipitation implies that a maximum of only 9 percent of the 170 mm annual precipitation, or 15.3 mm/year, can be recharged through Yucca Mountain. This is very close to the maximum possible flux estimated on the basis of the perched water chloride concentrations (16.3 mm/year). The likelihood that the saturated-zone groundwater beneath Yucca Mountain is a mixture of older water that recharged under past wetter climates and water that recharged beneath higher elevation areas to the north of Yucca Mountain, as well as present-day percolation through Yucca Mountain itself, indicates that the actual areally averaged percolation rate at Yucca Mountain is probably less than this value.

5.3.4.3.1.5 Effective Hydraulic Conductivity/Potential Gradient Methods

The method uses estimates of the measured or estimated effective hydraulic conductivity of the rock at its prevailing state of water potential or saturation, in combination with estimates of the hydraulic head gradient, to directly calculate the percolation flux through Darcy's law. This method is also the basis for estimates of percolation flux produced from model calibration against existing borehole saturation and water potential data sets, wherein percolation flux is adjusted until an optimal match between the observed data and the model output is achieved (Ahlers, Bandurraga et al. 1995b; Bodvarsson and Bandurraga 1996). Uncertainties in application of the method to Yucca Mountain have arisen primarily because few effective hydraulic-conductivity data have been measured for either welded or nonwelded tuffs, and because of uncertainty in the degree of fracture-matrix interaction, i.e., it has not been clear to what extent the saturation or water potential of the rock matrix reflects the passage of water in adjacent fractures. Attempts to model matrix saturation with dual-permeability models have demonstrated that it is possible to obtain similar simulated matrix saturations for a range of percolation fluxes by adjusting poorly constrained model parameters such as effective block size, that control fluid exchange between fractures and matrix (Ho et al. 1995).

Ahlers, Bandurraga et al. (1995b) used the parameter estimation code ITOUGH2 (Finsterle 1993) to simultaneously develop estimates of percolation flux and a set of rock hydrologic parameters for individual layers in the site-scale unsaturated-zone model for Yucca Mountain. These hydrologic

parameters include matrix permeability and parameters that describe the moisture characteristic function of van Genuchten (1980). A uniform steady-state percolation flux of 0.02 mm/year was determined to provide the best overall fit to the combined saturation and water potential data from boreholes UE-25a #1, UZ-4, UZ-5, UZ-6s, UZ-7, UZ-13, UZ-14, UZ-16, G-1, H-1, SD-7, SD-9, SD-12, and NRG-6 (Ahlers, Bandurraga et al. 1995b, Table 5.2).

Bodvarsson and Bandurraga (1996) continued the work of Ahlers, Bandurraga et al. (1995b) by incorporating a greater number of geologic layers and restricting the analyses to project-qualified ("Q") saturation and water-potential data collected from boreholes UZ-14, UZ-16, SD-7, SD-9, and SD-12. Using ITOUGH2 (Finsterle 1993), the best-fit match of the combined saturation and water-potential data from these boreholes using initial rock-property values developed from laboratory measurements was obtained for a percolation flux of 0.1 mm/year, based on the sum of the squared weighted residuals (Bodvarsson and Bandurraga 1996, Table 4.4-1). Additional calibration runs in which both rock properties and percolation flux were simultaneously optimized identified the best-fit percolation flux as 0.01 mm/year. Because the water-potential data determined from measurements made on core with a chilled-mirror psychrometer showed considerable scatter even within individual geologic units, all calibrated models resulted in equally marginal fits to the water-potential data, and identification of the optimal percolation flux was based largely on the fit to the measured saturation data (see Bodvarsson and Bandurraga 1996, Figures 4.3-1 to 4.3-4). Although the best fit to the saturation data was obtained for a uniform percolation rate of 0.01 mm/year, the small increases in the sum of the squared weighted residuals, residual standard deviation, and biases obtained for calibration runs in which higher (0.1, 1.0, and 5.0 mm/year) percolation rates were considered, may not have been statistically meaningful except in the 5.0 mm/year case. The authors concluded that if the percolation flux rate at Yucca Mountain was similar to the higher values investigated in the study, the flux in excess of 1 mm/year was probably moving as non-steady pulsed flow through local structural features. Attempts to match the data from saturation and water-potential data from SD-7 and SD-9, using spatially variable infiltration rates estimated from a soil moisture-budget model (Flint, A.L. et al. 1996), found that the best fit to the borehole saturation and water-potential data was achieved when the estimated infiltration rates were reduced by a constant factor of 0.025, which resulted in estimated percolation fluxes of 0.02 to 0.09 mm/year.

As described above, the percolation flux can also be estimated directly by multiplying the effective hydraulic conductivity of the rock at a given depth by the hydraulic gradient across that depth interval. In situ monitoring with thermocouple psychrometers at boreholes UZ-4, UZ-5, UZ-7a, SD-12, NRG-6, and NRG-7a has shown that, except for relatively dry intervals within the Tiva Canyon Tuff at UZ-7a and SD-12, water potentials within the unsaturated zone are generally between 0 and −4 bars (see Subsection 5.3.4.2.2). Thus, there is no evidence from these data that the water-potential gradient (the change in water potential with depth) within the 500 to 700 m thick unsaturated zone is substantially different than zero. In this case, the hydraulic gradient is approximately equal to the gravitational gradient of unity, and the percolation flux is numerically equal to the effective hydraulic conductivity. Effective hydraulic conductivities for the hydrogeologic units defined by L.E. Flint (1998) were estimated at boreholes UZ-16, UZ-14, SD-7, SD-9, SD-12, and UZ-7a where saturation data was available (Figure 5.3-178). These estimates were made using the average saturation values estimated for individual hydrogeologic units at the boreholes (Flint, L.E. 1998) (Subsection 5.3.3.1.1), and site-wide averages of hydraulic properties

for the hydrogeologic unit of interest. The relevant hydraulic properties are the geometric mean saturated hydraulic conductivity and average parameters of the moisture characteristic function of van Genuchten (1980) for each hydrogeologic unit (Flint, L.E. 1998) (Subsection 5.3.3.1.1). As listed in L.E. Flint (1998) and in Subsection 5.3.3.1.1: units 1 to 5 correspond to the welded Tiva Canyon Tuff; units 6 to 12 constitute the PTn; units 13 to 20 correspond to the welded units of the Topopah Spring Tuff; units 21 to 25 include the Calico Hills Formation and adjacent bedded tuffs; units 26 to 29 make up the Prow Pass Tuff; and units 30 to 31 are part of the Bullfrog Tuff. Plotting effective hydraulic conductivity versus the negative of the unit number therefore gives an idea of the variation of percolation flux with depth, although hydrogeologic unit thicknesses are distorted when represented in this manner.

The effective hydraulic-conductivity values shown in Figure 5.3-178 vary among boreholes within a given hydrogeologic unit because of differences in saturation and the assumed uniformity in the hydraulic properties. The variability in effective hydraulic conductivity among hydrogeologic units at an individual borehole may be reflecting nonvertical-flow components, differences in the relative amounts of fracture flow and matrix flow, uncertainty in the measured saturation and hydraulic-parameter values, and the uncertain applicability of the van Genuchten (1980) functions for estimating effective hydraulic conductivity in tuffs. As demonstrated in Figure 5.3-178, the calculated effective hydraulic conductivity within the PTn can be highly variable, even at a given borehole. However, the geometric mean effective hydraulic-conductivity values indicate a percolation flux through the PTn of less than 1 mm/year at UZ-16 and SD-9, on the order of 1 mm/year at SD-7 and perhaps UZ-7a, and between 1 to 2 mm/year at SD-12. The effective hydraulic conductivity decreases systematically at UZ-14 from greater than several hundred mm/year near the top of the PTn to less than 0.01 mm/year near its base. The effective hydraulic conductivity through the welded Topopah Spring Tuff at and above the potential repository horizon (units 13 to 15) is generally on the order of a few tenths mm/year or less, whereas the effective hydraulic conductivity values below the potential repository horizon (units 16 to 20) in the welded Topopah Spring Tuff suggest an increase in flux with depth at several boreholes (UZ-16 and UZ-14). The upper part of the lower nonlithophysal zone of the crystal-poor unit of the Topopah Spring Tuff (unit 17 or TM2 of Flint, L.E. 1998) has an effective hydraulic conductivity greater than a few mm/year for all boreholes in Figure 5.3-178 in which saturation data exist for the unit. The calculated effective hydraulic conductivities for the vitric part of the Calico Hills Formation (unit 23) are much greater than 1000 mm/year at UZ-14 and SD-9, perhaps reflecting perching on the underlying zeolitic rocks at these boreholes. In contrast, calculated effective hydraulic conductivities are a few tenths of a mm/year in the vitric Calico Hills Formation at SD-7 and SD-12. The zeolitic part of the Calico Hills Formation (unit 24) generally has an effective hydraulic conductivity on the order of a few tenths mm/year, and the pre-Calico Hills bedded tuff (unit 25) has an effective hydraulic conductivity of 1 to 2 mm/year. The Prow Pass Tuff (units 26 to 29) generally has an effective hydraulic conductivity of 2 mm/year or less except at UZ-14 and SD-9, where the effective hydraulic conductivity of the devitrified, partially welded tuff (unit 27) can be tens or even hundreds of mm/year.

As a qualitative check on the saturation data and parameters of the van Genuchten moisture-characteristic function used to calculate effective hydraulic conductivities, the same data were used to estimate water potentials in each of the hydrogeologic units for which estimates of

effective hydraulic conductivity were made (Figure 5.3-179). Most of the estimated water potentials plot within or near the 0 to -4 bar range indicated by the in situ psychrometer measurements at UZ-4, UZ-5, UZ-7a, SD-12, NRG-6, and NRG-7a. However, predicted water potential values are considerably more negative than 0 to -4 bars within two subzones of the crystal-poor lower nonlithophysal zone (Tptpln) of the Topopah Spring Tuff (units 17 to 18), the crystal-poor densely welded vitrophyre (Tptpv3) of the Topopah Spring Tuff (unit 19), the upper nonwelded, zeolitic subunit of the Prow Pass Tuff (unit 26), and, locally, the pre-Calico Hills bedded tuff (unit 25). The anomalously low-water potentials estimated for these units may be due to small amounts of drying of the core during drilling and sample handling, rather true variability in water potentials. In particular, units 17, 18, and 19 have average porosities of 0.09, 0.07, and 0.02, so that small amounts of evaporation could easily reduce saturations from near 100 percent to the approximately 90 percent measured on these samples. Water potentials are very sensitive to saturation changes in these rocks over this range of saturations. Similar phenomena could explain the more negative water potentials estimated for the densely and moderately welded rocks in the crystal-poor base of the Tiva Canyon Tuff (units 4 and 5, respectively), although at some boreholes (UZ-7a and SD-12) in situ water potentials were also lower in the Tiva Canyon Tuff, presumably because of evaporation due to natural air circulation or barometric pumping. Significant drying is not apparent in the pre-Calico Hills bedded tuff (unit 25) where saturations are near 100 percent at each of the boreholes, but may have occurred in the upper nonwelded zeolitic Prow Pass Tuff at SD-7 and UZ-16, where saturations are approximately 0.06 saturation units lower than in the other boreholes. Other, as-yet unidentified factors may explain the anomalously low-water potentials estimated for these units. In units where the estimated water potentials suggest that drying of core during drilling or sample handling has occurred, the calculated effective hydraulic conductivities may be underestimating the true in situ values.

5.3.4.3.1.6 Calcite Accumulation Rates

Paces et al. (Paces, Neymark et al. 1996; Paces, Marshall et al. 1997) presented calculations in which the mass of secondary minerals deposited within fractures was used to compute the temporally averaged flux over the last 10 to 12 million years. The later work by Paces, Marshall et al. (1997) provided revised estimates of the mass of accumulated minerals (principally calcite) and percolation flux presented in the earlier work, as well as estimates of percolation flux at multiple locations along the length of the Exploratory Studies Facility. Major uncertainties in estimating percolation flux from secondary mineral abundances involve:

- The assumption that all of the dissolved Ca^{2+} in percolating water (estimated to be 25 mg/L) is precipitated as calcite in the unsaturated zone as a result of evaporation (i.e., there is no recharge to the saturated zone)

- The assumption that, for a given location, the abundances of secondary minerals as determined from scanline surveys in the Exploratory Studies Facility represent the average volumetric abundance between the ground surface and water table

- The passage of water is recorded by the deposition of secondary minerals for all climate conditions and flux values.

The exact mechanisms required for calcite deposition (evaporation or degassing of CO_2) are not known at this time, nor is there presently sufficient data to assess the representativeness of the Exploratory Studies Facility mineral abundances for the entire thickness of the unsaturated zone. Incomplete evaporation of percolation in the unsaturated zone, calcite precipitation as a result of $CO_{2(g)}$ degassing from solution, and non- or incomplete deposition because percolation was undersaturated with respect to calcite, or because rates of deposition were slow relative to the velocity of water movement, would all result in underestimation of the actual percolation flux.

Paces, Marshall et al. (1997) estimated the percolation flux using the following relationship:

$$q = (\rho_{cc} f_{cc} f_{min} h x_{Ca})/(C_{Ca} f_{ppt} t) \qquad \text{(Eq. 5.3-10)}$$

where: q = the percolation flux (m year^{-1});

ρ_{cc} = the density of calcite, (2700 kg m^{-3});

f_{cc} = the fraction (by volume) of calcite in the secondary mineral coatings (0.9 m^3 m^{-3});

f_{min} = the fraction (by volume) of secondary mineral estimated from scanline surveys in the ESF (m^3 m^{-3});

h = the thickness of the unsaturated zone, (m);

x_{Ca} = the mass fraction of calcium in the secondary calcite (0.40 kg kg^{-1});

C_{Ca} = the assumed initial concentration of calcium in the water (0.025 kg m^{-3});

f_{ppt} = the fraction of calcium precipitated in the volume defined by hA (1.0 kg kg^{-1}); and

t = time (10^7 years).

The quantities f_{min} and h varied with position in the Exploratory Studies Facility. The values of other quantities were assumed to be constant throughout the Exploratory Studies Facility. The distribution of secondary minerals (f_{min}) for Exploratory Studies Facility stations 11 through 69 based on sampling the first 30 m of each 100 m interval is shown along with the fault and shear frequency in Figure F2 of Paces, Marshall et al. (1997). The volume percent of secondary minerals contained in both fractures and lithophysal cavities varies from near-zero values observed at multiple stations in the Exploratory Studies Facility to 0.43 percent measured at Exploratory Studies Facility station 30. Considering only the mineral accumulation in fractures, the highest volume percent of secondary minerals is 0.37 percent measured at Exploratory Studies Facility station 13. The volume percent of secondary minerals contained in fractures appears to correspond to the fault and shear frequency, which suggested to Paces, Marshall et al. (1997) that mineralization is associated with percolation along pathways having extensive vertical continuity resulting from structural features. A similar plot of secondary mineral volumes with fracture frequency data did not show any correspondence,

perhaps because fractures that lacked displacement were often related to cooling and lacked connections to the large-scale network.

The percolation fluxes estimated on the basis of the secondary mineral volumes are shown in Figure 5.3-180, along with infiltration estimates based on the infiltration models of Hudson and Flint (1996) and A.L. Flint et al. (1996). The percolation fluxes calculated on the basis of secondary mineral volumes are generally much lower than the infiltration rates predicted by the infiltration models, except for: Exploratory Studies Facility stations 13 and 16 underlying Azreal Ridge; Exploratory Studies Facility station 22 underlying Drill Hole Wash; Exploratory Studies Facility station 27 underlying Diabolus Ridge; and Exploratory Studies Facility station 30 under Dead Yucca Ridge. In these stations, percolation fluxes estimated from secondary mineral abundances were comparable or slightly greater than the infiltration rates estimated from the model of A.L. Flint et al. (1996). In general, however, there is no obvious correlation between the percolation rates calculated from secondary mineral abundances and the infiltration rates calculated from the soil water-budget model. Only 5 (Exploratory Studies Facility stations 13, 16, 22, 27, and 30) of 54 stations for which estimates were made from secondary mineral volumes had percolation fluxes greater than 5 mm/year, in contrast with 25 of the 78 stations for which estimates of infiltration are available using the model of A.L. Flint et al. (1996) and nearly all of the estimates of Hudson and Flint (1996).

5.3.4.3.1.7 Perched-Water Volumes/Residence Times

J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) estimated the areal extent and volume of the perched-water body intersected by UZ-14 from the elevation of the perched-water surface and a structural contour map of the densely welded, crystal-poor vitrophyre of the Topopah Spring Tuff (Tptpv3), which appears to be the layer responsible for perching in the vicinity of Drill Hole Wash. Residence time of water within the perched water body was calculated using uncorrected ^{14}C ages and a mass-balance equation for ^{14}C that assumed steady flow to and from a well-mixed reservoir in which chemical reactions involving carbon (if occurring) were also at steady-state:

$$Q\ {}^{14}A_0\,(HCO_3^-)_0 - Q\ {}^{14}A_{pw}\,(HCO_3^-)_{pw} - V\lambda\,{}^{14}A_{pw}\,(HCO_3^-)_{pw} = 0$$

(Eq. 5.3-11)

where:

Q = the steady-state flow rate to and from the reservoir (m^3/year);
^{14}A = the ^{14}C activity expressed in percent modern carbon (pmc);
(HCO_3^-) = the total carbon concentration, expressed as bicarbonate (moles/m^3);
0 and *pw* are subscripts that indicate the terms for the inflow and for the perched water;
V = the volume of the perched water reservoir (m^3); and
λ = the radioactive decay constant for ^{14}C (1.21×10^{-4} year^{-1}).

The term $V\lambda^{14}A_{pw}$ accounts for the change with time in the activity of ^{14}C due to radioactive decay of ^{14}C within the perched water reservoir. If there are no carbon reactions within the reservoir itself, the dissolved carbon concentrations of the inflowing and outflowing water are equal, and Equation 5.3-11 can be rearranged to solve for the residence time, t_{res}

$$t_{res} = V/Q = ({}^{14}A_0 - {}^{14}A_{pw})/\lambda^{14}A_{pw}$$

(Eq. 5.3-12)

Residence times that resulted from this equation and estimated perched water volumes were used to calculate percolation flux from the relationship

$$t_{res} = V/Q = Ab\phi_e/qA$$

(Eq. 5.3-13)

where:

A = the area of the perched water reservoir (m^2);
b = the thickness of the perched water reservoir (m);
ϕ_e = the effective porosity within the perched water reservoir; and
q = the apparent vertical percolation flux into and out of the reservoir (m/year).

Based on these equations, it was estimated that water fluxes to and from the perched-water reservoir intersected by UZ-14 ranged from 0.001 to 0.29 mm/year. The range in flux values resulted from variability in the ^{14}C values in multiple water samples, from uncertainty in the effective porosity of the rock within the perched water, and from uncertainty in the area actually contributing to the perched water. Additional uncertainties not incorporated in this range of flux values are related to the assumptions that the perched water is well-mixed, the initial ^{14}C activity ($^{14}A_0$) of water entering the perched water reservoir is 100 pmc, and there are no carbon reactions within the perched water reservoir itself.

Hydrochemical data from bailed samples of perched water from UZ-14 collected during drilling when the borehole bottom was at slightly different depths (384.6, 387.7, and 390.75 m) exhibited considerable variability (Yang et al. 1996, Tables 6 and 7), in contrast with later pumped samples obtained when the borehole was at 390.75 m, and with bailed samples obtained following pumping. Pumping had the apparent effect of homogenizing the perched water in the wellbore. The chemical heterogeneity observed in the early bailed samples suggests the existence of incompletely mixed pockets of water, at least in the upper part of the reservoir.

J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) argued that the dilute nature of the perched water, in comparison with more concentrated matrix pore water in the overlying PTn, suggested the perched water was recharged by fracture flow that had minimal interaction with pore water and that travel times were probably short. Therefore, in the absence of chemical reactions that may have reduced the ^{14}C activity of the dissolved carbon, the initial ^{14}C activity of the dissolved carbon recharging the perched water would be close to 100 percent.

Yang et al. (1998) presented analyses conducted with NETPATH (Plummer et al. 1994) which corrected for chemical reactions that may have affected the initial ^{14}C activity ($^{14}A_0$) of carbon dissolved in the water. If all dissolved carbon in the water were derived solely from interactions with the atmosphere and with biogenically generated $CO_{2(g)}$ in the shallow soil zone, the value of $^{14}A_0$ would be close to 100 pmc. Alternatively, the value of $^{14}A_0$ would be substantially less than 100 pmc if some of the carbon in the water was acquired through the dissolution of pedogenic or fracture-filling calcite in which the ^{14}C had been partially or completely decayed. The model presented in Yang et al. (1998) also included the effects of $CO_{2(g)}$ dissolution, silicate weathering, clay precipitation, glass dissolution, and Ca^{2+}/Na^{+} and Mg^{2+}/Na^{+} exchange, among other reactions. The model was used to analyze the chemical reactions that might have led to the evolution of perched water sample UZ-14D, which was the bailed sample obtained following extensive pumping of the perched water. This sample had a measured ^{14}C activity ($^{14}A_{pw}$) of 29.2 pmc. The values of $^{14}A_0$ ranged from 48.4 to 51.4 pmc (corresponding to adjusted ^{14}C ages of 4,170 to 4,669 years) for reactions models in which there were small amounts of glass dissolution and large amounts of cation exchange, to values for $^{14}A_0$ of 55.2 to 62.3 pmc (corresponding to adjusted ^{14}C ages of 4,932 to 6,261 years) for models which had large amounts of glass dissolution and small amounts of cation exchange. Substitution of values of $^{14}A_0$ from 48.4 to 62.3 pmc and the measured $^{14}A_{pw}$ value of 29.2 pmc into Equation 5.3-12 results in estimates of residence time ranging from 5,434 to 9,368 years, which falls within the range of values developed by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) (5,150 to 11,000 years) in a more ad hoc fashion based on qualitative corrections to the measured ^{14}C ages. Because sample UZ-14D was obtained by bailing after extensive pumping during which mixing of perched water in the wellbore presumably occurred, it may be representative of the water that would have existed had natural mixing been complete. However, recent analyses of two bailed samples obtained from UZ-14 prior to pumping (samples UZ-14A2 and UZ-14C) with a NETPATH reaction model similar to that described in Yang et al. (1998) resulted in adjusted ^{14}C ages of 1354 to 1974 years for sample UZ-14A2 and a modern age (0 years) for sample UZ-14C, indicating that flow to the perched water may be ongoing (E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation). Yang et al. (1996) reported that sample UZ-14C contained a polymer that had been used in the drilling of nearby borehole G-1, indicating partial contamination of this sample by drilling fluid. The high SO_4^{2-} concentration of this sample relative to Cl^- may also reflect the presence of lost drilling fluid from G-1, which contained large amounts of $BaSO_4$. However, isotopic evidence indicates this sample may simply be a pulse of recent infiltration that is unusually enriched in SO_4^{2-} for natural (although currently unexplainable) reasons. Chemical data of surface water for the Yucca Mountain area indicates that summer runoff has a much higher SO_4^{2-}/Cl^- ratio than winter runoff (Subsection 5.1.6). Sample UZ-14C had:

- δD and $\delta^{18}O$ values of -87.4 and -12.1 per mil (Yang et al. 1996), which plot on the present-day world meteoric water line and are consistent with the δD and $\delta^{18}O$ values of summer thunderstorms at Yucca Mountain (Yang et al.1998).

- A $\delta^{87}Sr$ value of 4.37 consistent with a pedogenic source for the Sr (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies*

Facility, Yucca Mountain, Nevada, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164)

- A $\delta^{13}C$ value of -9.2 per mil (Yang et al. 1996), which is consistent with $CO_{2(g)}$ and calcite dissolution, but not reaction with isotopically light polymers

- An ^{3}H value of 3.1 TU, indicating the possible inclusion of a small component of post-bomb water

Possible reactions that could alter either the amounts of total dissolved carbon in the perched water and/or the ^{14}C activity of the dissolved carbon are calcite dissolution or precipitation, dissolution or out gassing of $CO_{2(g)}$, and oxidation of organic carbon to produce $CO_{2(g)}$ and water. The effects of calcite and $CO_{2(g)}$ dissolution in the perched water reservoir were accounted for in the NETPATH adjustments to $^{14}A_0$, but not in $(HCO_3^-)_0$ and $(HCO_3^-)_{pw}$, which were assumed to be equal. NETPATH did not consider the effects of calcite precipitation or degassing of $CO_{2(g)}$ within the perched water reservoir because only the net changes in the water chemistry relative to pure water were calculated.

An estimate of percolation flux can be calculated for the perched water encountered at SD-7 in a similar manner. First, the initial ^{14}C activity of the pumped water sample SD-7(3/17) (Yang et al. 1996) was estimated with NETPATH by determining the amounts of various mineral phases that could have interacted with the water to produce the observed water chemistry. Application of NETPATH using reaction models similar to those described above resulted in estimates for adjusted ^{14}C ages of 3245 to 3917 years, and estimates of the initial ^{14}C activity of the water ($^{14}A_0$) of 42.1 to 45.6 pmc. Applying these estimates of $^{14}A_0$ in Equation 5.3-12 along with the measured value of $^{14}A_{pw}$ of 28.4 pmc for sample SD-7(3/17) resulted in calculated residence times of 3987 to 5005 years. The perched water volume at SD-7 was estimated to be 96,000 liters (96 m^3) based the pump discharge rate, the slope of the linear drawdown plot, the height of the water column (8.4 m) above the bottom of the reservoir in the borehole, and a lumped storage term composed of the product of the specific yield and reservoir area (O'Brien 1997). However, this estimated volume actually represents the volume of water that can be recovered through short-term (day-long) pumping tests and not the total reservoir volume, which is possibly much larger.

Using the estimated residence times calculated using Equation 5.3-12 (3,987 to 5,005 years), a value for ϕ_e of 0.35 determined from the porosity log for this depth interval (Rautman and Engstrom 1996a), and estimates of the saturated thickness of the perched water of 8.4, 16.8, and 4.2 m, percolation fluxes were estimated to range from 0.29 to 1.5 mm/year. These estimates of percolation flux are based on the assumption that all water in the reservoir arrived through a vertical pathway.

As with borehole UZ-14, pumping may have homogenized the perched water to a greater degree than had occurred naturally. All four pumped samples taken from SD-7 when the borehole was at a 488.3 m depth have nearly identical isotopic and chemical characteristics. However, as at UZ-14, a bailed sample taken from a shallower depth (479.8 m) prior to pumping (sample SD-7 (3/8)) displayed slightly different chemical and isotopic characteristics than the pumped samples, including a slightly higher ^{14}C value (34.4 pmc), slightly more depleted $\delta^{13}C$ values (-10.4 per mil), and an ^{3}H activity of 6.2 TU.

5.3.4.3.2 Flowpaths in the Unsaturated Zone

5.3.4.3.2.1 Distribution of Percolation Flux

In this subsection, estimates of percolation flux calculated at various borehole locations with the methods described in Subsection 5.3.4.3.1 are summarized and compared with estimates of average net infiltration for 30 m x 30 m and 90 m x 90 m areas centered around these boreholes. Table 5.3-60 summarizes the estimated percolation fluxes and infiltration rates at the various boreholes. A comparison between estimates of percolation flux based on secondary mineral abundances and infiltration rates estimated by Hudson and Flint (1996) and A.L. Flint et al. (1996) was shown previously for stations along the Exploratory Studies Facility (Figure 5.3-180). This comparison showed that percolation flux rates estimated from secondary mineral abundances were generally much lower than those estimated from the soil moisture-budget model described in A.L. Flint et al. (1996), except for: Exploratory Studies Facility stations 13 and 16 underlying Azreal Ridge; Exploratory Studies Facility station 22 underlying Drill Hole Wash; Exploratory Studies Facility station 27 underlying Diabolus Ridge; and Exploratory Studies Facility station 30 under Dead Yucca Ridge. In these stations, percolation fluxes estimated from secondary mineral abundances were comparable or slightly greater than the infiltration rates estimated from the model of A.L. Flint et al. (1996).

Figures 5.3-181a and b show there is only a weak correlation, at best, between the percolation fluxes estimated from the borehole temperature profiles and estimates of infiltration rate provided by the model of A.L. Flint et al. (1996). Two independent estimates of percolation flux based on borehole temperature profiles are shown, including those presented in Subsection 5.3.4.3.1 (labeled USGS) and those derived by Bodvarsson et al. (1997) (labeled LBNL). The USGS results include the estimates of flux calculated for UZ#4 (18 mm/year) and UZ#5 (5 mm/year) in J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The two estimates of percolation flux calculated on the basis of the temperature profiles agree with each other for some locations (WT-2, H-1, WT-18, UZ#4, and a#1/a#7) but are in substantial disagreement in other locations (H-3, H-5, G-3, G-4, UZ#5). The apparent disagreement between percolation fluxes for UZ-1 reflects the fact that the USGS results were derived using a vertically variable flux that decreased with depth due to the presence of low-permeability layers that resulted in perched water (15 mm/year above the perching layer to 5 mm/year below the perching layer), whereas the Lawrence Berkeley National Laboratory results were obtained using a vertically constant flux (8 mm/year).

The percolation fluxes estimated from the temperature data are in most obvious disagreement with the estimates of infiltration rate in some of the smaller washes draining the eastern slope of Yucca Mountain (boreholes SD-12 in upper Antler Wash and G-4 in Coyote Wash), the distal end of Drill Hole Wash (boreholes a#1 and a#7), lower Pagany Wash (UZ#4), and upper Dune Wash (WT-1). In contrast, the percolation fluxes estimated from the temperature data are lower than the infiltration estimates at H-4 (located in lower Antler Wash) and at WT-12 (located on an alluvial terrace to the west of Busted Butte). Both the temperature data and the soil water-budget model indicate relatively small flux rates at WT-2 (located in WT-2 Wash) and WT-17 (located on an alluvial terrace in lower

Dune Wash) and moderately high percolation fluxes (5 to 10 mm/year) in upper Drill Hole Wash (boreholes UZ-1 and G-1). The wide range of percolation fluxes estimated from the temperature data for boreholes located in alluvium (WT-12, WT-17, WT-1, WT-2, G-4, a#1, a#7, WT-1, SD-12, UZ-1, G-1, and UZ#4) suggest the variable effectiveness of alluvium for inhibiting net infiltration. The temperature data also indicate that areas in shallow alluvium immediately adjacent to sideslopes (borehole H-1 in Drill Hole Wash and borehole WT-4 in Midway Valley) may be locations where deep percolation is higher than presently predicted by the infiltration model. The USGS estimate of percolation flux results agrees with the estimate of infiltration at UZ#5, which is at the bottom of a north-facing slope in Pagany Wash, whereas the Lawrence Berkeley National Laboratory model predicts negligible percolation at this location. These conclusions are not significantly altered when infiltration rates are averaged over 90 x 90 m areas centered around the boreholes (Figure 5.3-181b). Estimates of the percolation flux based on borehole temperature data are either higher or lower than the estimated infiltration rates for the Yucca Crest boreholes (G-3, H-3, and H-5), depending on the analyst. Percolation rates estimated by Lawrence Berkeley National Laboratory range from 3 to 5 mm/year, whereas those estimated by the USGS range from 10 to 20 mm/year. Infiltration rates estimated with the soil water-budget model range from about 7 to 9 mm/year when averages are calculated over 30 x 30 m areas, and about 8 to 11 mm/year when 90 x 90 m areas around the boreholes are considered. Differences in the percolation fluxes estimated at the Yucca Crest boreholes by the USGS and Lawrence Berkeley National Laboratory models may be due to slightly different thermal conductivity-porosity models used in the numerical simulations, the greater stratigraphic detail in the USGS models, the use of slightly different assumed ground-surface temperatures, or a combination of these factors.

A comparison between the percolation fluxes estimated by Weeks (written communication, presentation at the Unsaturated Zone Flow Model Expert Elicitation Workshop, February 1997) with the infiltration fluxes calculated with the soil water-budget model of A.L. Flint et al. (1996) is shown in Figures 5.3-182a and b. Although there is not a correlation between the fluxes estimated by the two methods, the ranges of the flux estimates are similar and the differences in the estimates for particular wells (UZ-6, SD-9, SD-7, UZ-1, UZ-14) do not differ by more than just a few millimeters per year. The most significant disagreement occurs for boreholes NRG-7a (located on a north-facing sideslope bounding Drill Hole Wash), UZ-16 (located on an alluvial terrace near the junction of Antler and WT-2 Washes), and SD-12 (located in upper Antler Wash), where the estimates of flux calculated from the ^{14}C data are much larger than the estimated infiltration rates. Some of the differences in fluxes estimated for these boreholes may be attributed to the depth of the ^{14}C data used by Weeks (written communication, presentation at the Unsaturated Zone Flow Model Expert Elicitation Workshop, February 1997) to estimate flux. Because the ^{14}C data were taken from or below the depths at which perched water is frequently observed, it is possible that lateral flow associated with perching may have redistributed water in the deep subsurface so that any spatial correlations that may have existed between the two estimates have become obscured. The other borehole at which significant disagreement exists is UZ-13, located on Yucca Crest a few hundred meters south of G-3. At this borehole, the percolation flux estimated from the ^{14}C data is 1 mm/year, whereas the infiltration rate is estimated to be 9 mm/year. The ^{14}C data for UZ-13 are from gas samples collected from the upper part of the Topopah Spring Tuff, so that discrepancies related to subsurface lateral flow may be less severe than for UZ-16, SD-12, or NRG-7a. No explanation for the relatively large differences in the two estimates at UZ-13 is available at this time.

Figures 5.3-183a and b show a comparison between apparent percolation fluxes estimated from the chloride concentrations from perched and pore water samples (Table 5.3-59) with the infiltration fluxes estimated by A.L. Flint et al. (1996). Also shown are the site-averaged fluxes (horizontal lines) estimated for the PTn matrix, the Tac matrix, and PTn and Topopah Spring Tuff fractures based on the averages of the chloride concentrations listed for these units and perched water in Table 5.3-58 (see Subsection 5.3.4.3.1.4). The apparent percolation fluxes estimated from the pore water chloride concentrations appear to be uncorrelated with infiltration estimates and are generally lower than the infiltration rates where these rates are estimated to be non-zero (UZ-4, UZ-5, UZ-14, SD-7, and SD-9). Where infiltration rates are estimated to be zero (UZ#16, SD-12, NRG-7a, and NRG-6), the pore water chloride data indicate fluxes of between 1 to 4 mm/year. In contrast, the apparent percolation rates estimated on the basis of the perched water chloride data are much higher than the estimated infiltration rates, but appear to be positively correlated. Based on borehole video logs, perched water was encountered at boreholes SD-12 and UZ-16, but no water samples were taken. The much higher apparent percolation rates estimated from the perched water chloride concentrations may be, in part, an artifact of incomplete mixing of matrix and fracture water in the PTn and the Calico Hills Formation, and the assumption inherent in the equation used to produce the flux estimate that the water is well-mixed. The total average flux (9.8 mm/year) for the site (uppermost horizontal line), calculated on the assumption that fracture and matrix waters are incompletely mixed, is much lower than the apparent fluxes calculated on the basis of the chloride concentrations of the perched water, but higher than the average percolation rates calculated for the PTn or Tac matrix (0.7 and 1.7 mm/year). Based on this analysis, faults and fractures in the PTn are predicted to carry over 90 percent (9.1 mm/year) of the total average flux through the unit. However, as discussed in Subsection 5.3.4.3.1.4, other possible explanations for the relatively dilute perched water underlying more concentrated PTn pore water are that water is moving through the PTn updip from the perched water in high-flux areas for which no chloride data are yet available; and perched water in the lower unsaturated zone is a remnant from the Pleistocene, when infiltration rates were higher and chloride concentrations correspondingly lower.

Figures 5.3-184a and b show a comparison between the percolation fluxes estimated by Bodvarsson and Bandurraga (1996) and Ahlers, Bandurraga et al. (1995b) from the borehole saturation and water-potential data and the infiltration rates estimated from the soil water budget model of A.L. Flint et al. (1996). Bodvarsson and Bandurraga (1996) and Ahlers, Bandurraga et al. (1995b) identified the single-percolation rate that provided the best fit to the matrix-saturation and water-potential data from multiple boreholes. Based on matches to the matrix-saturation and water-potential data from UZ-16, SD-12, UZ-14, SD-9, and SD-7, Bodvarsson and Bandurraga (1996) estimated the percolation flux to be 0.1 mm/year, which was slightly higher than the value of 0.02 mm/year estimated by Ahlers, Bandurraga et al. (1995b) for these same boreholes plus a#1, NRG-6, H-1, UZ#4, UZ#5, UZ-13, UZ-7, and UZ#6s. These percolation fluxes were shown in Subsection 5.3.4.3.1.5 to be of comparable magnitude to the matrix fluxes estimated directly from the matrix-property data and saturations. The percolation rates estimated by Ahlers, Bandurraga et al. (1995b) and Bodvarsson and Bandurraga (1996) are much smaller than the infiltration rates estimated by A.L. Flint et al. (1996), which range to up to 10 mm/year at SD-9 (Figure 5.3-184a and

b). The degree of mismatch is particularly pronounced at boreholes SD-9, SD-7, UZ-13, UZ-14, UZ#4, and UZ#5. The discrepancy between the estimated percolation rates and the infiltration rate estimates suggests that:

- The results of Bodvarsson et al. (1997) and Ahlers, Bandurraga et al. (1995b) are being unduly weighted toward low values by including boreholes in the inversion for which percolation rates may be genuinely low.
- The data (including the matrix-property measurements and saturation and water potential measurements) are systematically in error.
- Matrix saturations and water potentials reflect only the matrix component of the flux and not the total flux through the unsaturated zone.

Figures 5.3-185a and b show a comparison between the percolation flux rates calculated from perched water thickness and residence times with infiltration rates estimated from the soil water budget model of A.L. Flint et al. (1996). A range of percolation fluxes is presented for both UZ-14 and SD-7 in an attempt to account for uncertainties in reservoir thickness, effective porosity, and/or residence time. The infiltration rates are larger than the calculated percolation rates at both UZ-14 and SD-7, even considering the uncertainty in the estimates of reservoir volume and residence time.

In summary, Table 5.3-60 and Figures 5.3-181 through 5.3-185 show that there is only marginally good agreement between estimates of net infiltration provided by the soil water budget model of A.L. Flint et al. (1996) and estimates of percolation rates provided by various methods that rely on subsurface data. Both the soil water budget method and subsurface methods indicate that approximately 6 percent or less of the average annual precipitation of 170 mm/year that falls on Yucca Mountain becomes net infiltration, but these surface and subsurface methods differ in the details as to how net infiltration and percolation are spatially distributed. However, the estimates of percolation flux based on subsurface data also vary among themselves, even at a given borehole location. Some of the discrepancies may be the result of the lack of sensitivity of certain types of data to the fracture component of the percolation flux. The much lower percolation fluxes estimated on the basis of matrix saturations and water potentials, and matrix-chloride concentrations, may be the result of relatively weak hydraulic and chemical interactions between the fracture and matrix components of the flux. However, discrepancies among other subsurface estimates of percolation flux that should be equally sensitive to the fracture flux are more difficult to explain. These include the estimates based on borehole temperature logs, gas, and aqueous ^{14}C measurements and perched water volumes and residence times. The percolation flux rates estimated on the basis of the perched water volumes and residence times are considerably lower than those estimated from other methods, including matrix-chloride concentrations. One possible explanation for the low rates estimated from perched water volumes and residence times is that the perched water reservoirs remain at near-full capacity, and recent influxes of water simply overflow the reservoir without mixing with more stagnant water already present. Alternatively, it is possible that perched water residence times were overestimated and percolation fluxes underestimated because travel times from the ground surface to the perched water were assumed to be negligibly short. Travel times of a few thousand years, in addition to the ^{14}C dilution that results from the interaction of percolating water with carbonate

minerals, would be sufficient to reduce the initial ^{14}C activity of the perched water to levels compatible with percolation fluxes in the 5 to 10 mm/year range. Estimates of percolation flux based on borehole temperature profiles derived independently by the USGS and Lawrence Berkeley National Laboratory agree with each other except for boreholes located along Yucca Crest. The USGS results are in qualitative agreement with the results of the soil water budget model that predicts that infiltration flux along Yucca Crest is 10 mm/year or more. Elsewhere, the USGS and Lawrence Berkeley National Laboratory results indicate that percolation fluxes in many of both the major and smaller washes are greater than presently calculated by the soil water budget model. The estimates of percolation flux calculated on the basis of the gaseous and aqueous ^{14}C data are in reasonable agreement with the infiltration estimates, and where differences exist, these may be due to subsurface lateral flow along perching layers. The combined chloride-concentration data from matrix pore water, perched water, and shallow groundwater samples could be interpreted to indicate a site-wide average percolation flux of approximately 10 mm/year, with up to 8 to 9 mm/year occurring as fracture or fault flow through the PTn and Calico Hills Formation. However, these conclusions are contingent upon the magnitude of the corrections applied to the ^{14}C ages of the dilute perched water, as well as the results of further sampling of pore water chloride concentrations in the PTn and Calico Hills Formation under ridgetop and sideslope locations.

The differences in the average values of percolation flux estimated by a particular method in Table 5.3-60 may reflect both true spatial variability (averages for each method were, in general, based on different subsets of boreholes) and the sensitivity of the method to the fracture component of the percolation flux. The estimates of percolation flux listed in Table 5.3-60 are too sparsely distributed to calculate an areally-weighted average percolation flux for any individual method. Combining different types of estimates would provide greater spatial coverage with which to calculate an areally-weighted average, but the inclusion of estimates from methods with potentially low sensitivity to the fracture component of the flux would introduce bias into the calculation. Calculation of the arithmetic average percolation flux from the values listed in Table 5.3-60 (excluding those that resulted from methods 3 and 4) resulted in an estimate of the average percolation flux of 7.0 ± 6.0 mm/year. The mean percolation flux is thus estimated to be approximately 2.2 times greater than the average present-day infiltration rate of 3.2 mm/yr estimated by A.L. Flint et al. (1996). However, given the complexity of the problem and the relatively large standard deviation of the percolation flux, the estimates for the average infiltration and percolation fluxes must be considered to be in relatively good agreement.

5.3.4.3.2.2 Role of Discrete Pathways

Discrete pathways exist at a variety of scales, ranging from channels that exist within the planes of individual fractures, heterogeneity in properties of the rock matrix resulting from depositional processes or later alteration, individual pathways through the network of interconnected fractures, large-scale heterogeneity related to differences among hydrogeological units in the degree of welding and/or intensity of later alteration to clay or zeolites, and faulting. The role of faults is discussed more completely in the following subsection. In general, when flow occurs along discrete pathways, much of the storage capacity for water or dissolved solids within the unsaturated zone may be bypassed and, for a given flux of water or solute, transport velocities may be higher and travel times shorter than would be anticipated on the basis of the total porosity.

Evidence that at least some of water moving through the unsaturated zone at Yucca Mountain is moving through faults and fractures is provided by the distribution of environmental isotopes such as ^{3}H and ^{36}Cl (Yang et al. 1996, 1998; Fabryka-Martin, Wolfsberg et al. 1997; Fabryka-Martin, Turin et al. 1998). These isotopes have been found in the PTn, in the Exploratory Studies Facility at the level of the potential repository horizon, and in the Calico Hills Formation below the potential repository horizon, at concentrations that indicate an origin related to nuclear-weapons testing (see Subsection 5.3.4.2.4). The presence of by-products of nuclear-weapons tests at these depths is incompatible with the slow transport velocities anticipated on the basis of the rock-matrix permeability and porosity, indicating that these isotopes probably arrived at their sampling depths as a result of flow through fractures and faults. Unfortunately, the presence of these isotopes does not in itself provide strong constraints on the quantity of flow associated with their transport. There is also some ambiguity as to whether samples with $^{36}Cl/Cl$ ratios between the 500×10^{-15} (the meteoric concentration prior to the onset of nuclear-weapons testing) and 1250×10^{-15} (the meteoric concentration during much of the Pleistocene) represent Holocene water with a bomb-pulse component, or a mixture of Holocene and Pleistocene water. If the former interpretation is correct, rapid flow through the PTn may be more pervasive than recent analyses have indicated, whereas if the latter interpretation is correct, rapid flow through the PTn appears to be associated with mapped faults (Levy, Sweetkind et al. 1997; Sweetkind, Fabryka-Martin et al. 1997). Although sampling for other isotopes associated with nuclear weapons tests could discriminate between these possibilities, this has not yet been done on a systematic basis.

The distribution of low-temperature secondary mineral coatings in fractures and lithophysal cavities also provides direct evidence that fractures and faults have been conduits for water movement in the past (Paces, Neymark et al. 1996; Paces, Marshall et al. 1997). From scanline surveys, it has been estimated that only about 8 percent of the fractures sampled are mineralized (Anna 1998) indicating either that only a minor percentage of the fractures observed in the Exploratory Studies Facility are connected to the mountain-scale fracture network, or that flowpaths have been remarkably stable throughout many changes in climate over the millions of years since the tuffs were deposited. These observations are compatible with the distribution of ^{3}H along several recently drilled boreholes which suggest that flow through the fracture system is spatially erratic (Yang et al. 1996, 1998). In dipping fractures, mineralization is confined to the rock surface on the footwall side of the fracture (Paces, Neymark et al. 1996). Interestingly, where the Ghost Dance fault has been intersected by the Exploratory Studies Facility and associated alcoves, it does not appear to have been the site of more abundant secondary mineral deposition than nearby unfaulted rock (Eatman et al. 1997). Therefore, if it can be assumed that secondary mineral abundance is related to the volume of past water movement, it does not appear that the volume of flow through the Ghost Dance fault at and above the level of the Exploratory Studies Facility has been greater than in the surrounding rock. Examination of fracture-filling secondary calcite sampled from depths below the present-day potentiometric surface from boreholes drilled into the saturated zone has shown that some of this calcite has characteristics similar to unsaturated-zone calcites (Whelan and Moscati 1997). This observation implies not only that the potentiometric surface was several hundreds of meters deeper at some time in the past, but that fracture flow existed at these depths under a climate as or more arid than the present one. Moreover, the calcite saturated water that produced the calcite-fracture coatings probably did not react significantly with the zeolitic-rock matrix, which appears to

exchange sodium for calcium in solution and causes pore water to become undersaturated with respect to calcite.

The hydrogeologic framework provides constraints on the distribution of permeability within the unsaturated zone and gives an indication of where recharge to the saturated zone is potentially the highest. Figures 5.3-186a through c show simplified geologic cross-sections for the central block along Nevada coordinates N770,000 feet, N765,000 feet, and N760,000 feet. These cross-sections were created from contour maps based on the lithostratigraphic model of Buesch, Nelson et al. (1996). Also shown on these cross-sections is the distribution of zeolites within the lower unsaturated-zone units. Site-wide zeolite development is believed to be related to hydrothermal activity associated with expulsion of the Rainier Mesa Tuff approximately 10.6 Ma (Bish 1989), and the zeolitic-vitric interface may define the position of the paleo-water table relative to the rocks at the time of alteration. However, as shown schematically in these figures, extensive zeolitic alteration in the Prow Pass and Bullfrog Tuffs was confined to the margins of these units. Partial welding and devitrification of the centers of these units relatively soon after deposition apparently inhibited later zeolite development. The zeolitic intervals in the Calico Hills Formation, and Prow Pass and Bullfrog Tuffs have a permeability that is approximately 2 to 4 orders of magnitude less than their unzeolitized vitric or devitrified equivalents (Loeven 1993). Using more recent measurements, the geometric mean hydraulic conductivity of the zeolitic Calico Hills Formation is estimated to be 1.4 mm/year, whereas the geometric mean hydraulic conductivity for the devitrified interval of the Prow Pass Tuff is about 600 mm/year (Flint, L.E. 1998, Subsection 5.3.3.1.1). Pump tests conducted where these devitrified zones are submerged below the water table (e.g., Rush et al. 1984) have shown that these intervals generally have permeabilities in excess of 10^{-12} m^2, which is 5 to 6 orders of magnitude greater than the matrix permeabilities of the surrounding zeolitized nonwelded tuffs. Percolation fluxes in excess of about 1.4 mm/year would therefore presumably be recharged to the saturated zone through faults and fractures, along the top of perching layers, or through the less altered intervals of the Prow Pass Tuff. Cross-sections along Nevada State Coordinates N765,000 and N760,000 feet show that in the southwestern part of the central block, the Calico Hills Formation and upper Prow Pass Tuff are unzeolitized and, in combination with the devitrified interval of the Prow Pass Tuff, provide an uninterrupted pathway of high-permeability, nonzeolitized rock between the potential repository horizon and the water table.

Because waste is more likely to be mobilized when percolation flux exceeds 1 mm/year (Chapters 9 and 7, CRWMS M&O 1995a), the distribution of these higher permeability pathways at the water table also indicates the probable entry points of dissolved radionuclides to the saturated zone. Maps of the geology at the water table based on the lithostratigraphic model of Buesch, Nelson et al. (1996) showed good agreement with a similar map developed by Fridrich et al. (1994) where these maps overlapped when the potentiometric surface of Robison (1984) was used. Because of uncertainties in the elevation of the true regional potentiometric surface in the apparent large gradient area north of Drill Hole Wash (see Czarnecki, O'Brien et al. 1994) and because of more limited borehole control in this area, the geology at the elevation of the potentiometric surface is more uncertain there than within central block.

Hydrochemical data could, in principle, either confirm or refute these hypotheses concerning the distribution of flowpaths and locations of recharge, which have been developed primarily on the

basis of the geologic framework and known hydraulic characteristics. Particularly useful would be ^{14}C data from the shallow saturated zone in the vicinity of faults and from the devitrified interval of the Prow Pass Tuff in the lower unsaturated zone. Young ages for shallow groundwater near faults would provide support to the hypothesis that faults are sites of active recharge. Similarly, young ages for water in the devitrified interval of the Prow Pass Tuff in the southwestern part of the central block would lend support to the hypothesis that recharge to the saturated zone occurs preferentially through this interval, particularly if this young water was located below older water in the Calico Hills Formation. Unfortunately, critical data that could be used to evaluate these hypotheses have yet to be fully developed. Samples from the shallow saturated zone directly beneath the central block are unavailable except for UZ-16, where the corrected ^{14}C age of the groundwater is about 600 years (E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation). UZ-16 is located in a heavily faulted area (faults not shown in Figure 5.3-186c), and the potentiometric surface is located within the devitrified interval of the Prow Pass Tuff. Core has been taken from the devitrified interval of the Prow Pass Tuff at boreholes SD-7, SD-9, and SD-12, but pore water has not yet been extracted and analyzed. At UZ-14, major ion chemistry is available for pore water from the devitrified Prow Pass Tuff, but no ^{14}C data are yet available for this interval. There is no clear indication from the major ion chemistry that lateral flow is occurring in the devitrified Prow Pass Tuff at UZ-14. The combined ^{14}C, Na^+ and HCO_3^- data from UZ-14 suggest that lateral flow is occurring in the lower Calico Hills Formation above the pre- Calico Hills bedded tuff (Tacbt). This unit was the apparent perching layer at SD-7. Cl^- data from SD-12 display a trend of decreasing concentration with depth in the Calico Hills Formation as the pre-Calico Hills bedded tuff (Tacbt) is approached, again suggesting the possibility of lateral flow above this unit (Figure 5.3-176e).

Role of Faults in the Unsaturated-Zone Hydrologic System–Because of their vertical extent, faults have the potential to intercept and divert both gas and liquid moving in all unsaturated zone units, or introduce fluids into those units from the ground surface. The liquid and vapor entering or leaving the unsaturated-zone system through or adjacent to faults may represent a significant fraction of the overall moisture balance in the unsaturated zone.

Chloride-36 and ^3H data from the Exploratory Studies Facility and surface-based boreholes indicates that faults allow some water to bypass the matrix of the PTn, with the result that relatively short travel times are required for these tracers to penetrate hundreds of meters deep within Yucca Mountain. Bomb-pulse concentrations ^{36}Cl appear to be associated with structural features that allow the matrix porosity of the PTn to be bypassed (Fabryka-Martin, Wolfsberg et al. 1997; Levy, Sweetkind et al. 1997; Sweetkind, Fabryka-Martin et al. 1997). Day et al. (1998) reported that fault zones tend to coalesce at depth, resulting in the potential for faults to funnel infiltration from a relatively wide capture area (tens to hundreds of meters) into a narrow zone, and by inference, create shadow zones that receive relatively little infiltration. This funneling mechanism may enhance the likelihood that flow may bypass the matrix of the PTn matrix even at low-average fluxes if the fault splays coalesce within or above the PTn.

The manner in which barometric pressure fluctuations propagate through the unsaturated zone allow estimates to be made of the rock pneumatic diffusivity. Assuming independent estimates of the effective air-filled porosity are available, subsurface pneumatic pressure measurements have also

enabled estimates to be made of the lumped-formation permeabilities. At faulted locations where such measurements are available (UZ#1, UZ#4, UZ#5, UZ#7a), it appears that the permeability of the PTn is significantly enhanced relative to the matrix permeabilities measured from core samples or estimated from the average permeability of the hydrogeologic unit (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation).

Much of the water recharging the saturated zone through Yucca Mountain, and perhaps elsewhere, may do so along fractures and faults. Hydraulic test data are available for the Calico Hills and Prow Pass Tuffs where these units are submerged beneath the water table. Loeven (1993) published log-probability plots of core-scale and field-scale permeability values that indicated the occurrence of secondary permeability within the Calico Hills Formation and the portions of the Prow Pass and Bullfrog Tuffs included in the CHn hydrogeologic unit of Montazer and Wilson (1984). Some of this secondary permeability is associated with faults. At boreholes a#1 and b#1, which are located at the projected intersection of the Drill Hole Wash and unnamed north-south trending faults of the imbricate fault zone, hydraulic test data indicated that secondary permeability was high in the Calico Hills Formation and Prow Pass Tuff, even through intervals that are pervasively zeolitized (Figure 5.3-187). Hydraulic test data from the Prow Pass Tuff at H-1, which is located near the Drill Hole Wash fault, also exhibited permeability values indicative of fracturing or faulting ($>10^{-12}$ m^2, or hydraulic conductivities $>10^{-5}$ m/s).

Some recharge to groundwater beneath Yucca Mountain by water moving through the overlying unsaturated zone is indicated by the scatterplot of SO_4^{2-} versus Cl^- concentration of groundwater samples from Yucca Mountain and vicinity (Figure 5.3-188) (E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation). This scatterplot shows that minimum concentrations in both anions occurs for water sampled from boreholes located within the central part of the repository block, with higher concentrations to be the west and east. This pattern is consistent with relatively dilute water being recharged through the overlying unsaturated zone. The pattern of corrected ^{14}C ages of the groundwater samples also suggest that recharge is occurring beneath Yucca Mountain, as well as to the east along Fortymile Wash (Figure 5.3-189) (E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation). The groundwater samples generally came from relatively deep, stratigraphically mixed samples obtained by pumping the entire borehole. However, a water sample bailed from the upper few meters of the saturated zone at UZ-16 had a ^{14}C activity that was 42 pmc, which is much larger than the ^{14}C activities of deeper groundwater samples from beneath Yucca Mountain (Benson and McKinley 1985). After applying corrections to the initial ^{14}C activity of the sample to account for water-rock reactions that introduced isotopically "dead" carbon into the water sample, it was determined that the corrected ^{14}C age may have been as little as 600 years. The very young age for this shallow groundwater, located 490 m below ground surface in the devitrified interval of the Prow Pass Tuff, as well as the elevated ^{3}H values and high ^{14}C activities in the overlying Calico Hills and Prow Pass Tuff (Yang et al. 1996;

also see Subsection 5.3.4.2.4), suggests that faults of the nearby imbricate fault zone may have played a role in the transport of percolation through the Calico Hills Tuff to the water table.

Average pore-water chloride concentrations of the Calico Hills Formation were presented earlier for boreholes UZ-16, UZ-14, NRG-7a, SD-7, SD-9, and SD-12 (Table 5.3-58). The average concentrations range from 23.8 to 40.7 mg/L and are considerably higher than those obtained from saturated-zone samples, which are typically about 7 mg/L (Benson and McKinley 1985), but up to about 10.6 mg/L for the shallow saturated-zone water at UZ-16 (Yang et al. 1996). If saturated-zone water beneath Yucca Mountain is being recharged through the overlying unsaturated zone, as indicated by the young ^{14}C age of the shallow UZ-16 sample, then recharge must include a significant component of more dilute water that bypasses the zeolitic matrix of the Calico Hills and Prow Pass Tuffs through faults or as a result of lateral flow.

Open borehole geochemistry samples acquired by lowering sampling tubes to various depths (Yang et al. 1996, Table 9) suggest that some boreholes located near faults, such as NRG-7a and SD-12, may have gases with ^{14}C ages that are young for their depths, based on comparisons with trends in ^{14}C activity versus depth defined by the majority of the samples from these and other boreholes. Borehole NRG-7a is located near the Drill Hole Wash fault and borehole SD-12 is located near the Ghost Dance fault. Although sampling conditions were not ideal, mixing of modern atmospheric gases with in situ rock gas down these open boreholes was not indicated to be significant for these samples, based on their CO_2 concentrations and $\delta^{13}C$ values. The young ^{14}C ages of some of the gas samples from these boreholes may have resulted from the degassing of CO_2 contained in percolation moving down nearby faults, or by barometric pumping of young gases through the faults, which may have short-circuited the PTn in a manner similar to that which has been attributed to boreholes at Yucca Crest (e.g., Weeks 1993). However, conclusions based on these samples should also be viewed within the context of the ^{14}C values of gases sampled from the Ghost Dance fault from isolated intervals of a horizontal borehole drilled from the Exploratory Studies Facility (LeCain and Patterson 1997). The uncorrected ^{14}C activities and ages (2600 to 4700 years old) of these gas samples did not indicate high rates of either water or gas movement down the fault. To the contrary, the gas samples from the fault zone with the youngest uncorrected ^{14}C ages occurred updip from the main trace of the fault in less intensely brecciated rock. In those locations where the downward transport of atmospheric gases has been accelerated as a result of barometric pumping, it might be expected that the exchange of vapor-laden formation gases with the atmosphere would be greater than in areas where gas exchange took place as a result of diffusion alone. In situ measurements of water potential presented in Subsection 5.3.4.2.2 for boreholes SD-12 and UZ-7a (located within the Ghost Dance fault zone) showed evidence that the rock became relatively dry (water potentials less than −12 bars) at shallow depths, compared with similar measurements at other boreholes (Rousseau et al. 1997). This observation suggests that gas exchange between the rock and the atmosphere may be enhanced by the presence of the fault near or at these boreholes. However, in contrast to UZ-7a and SD-12, borehole NRG-7a, which is located near the southwestern splay of the Drill Hole Wash fault, did not show a trend toward decreasing water potentials at shallower depths (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164).

5.3.4.3.2.3 Inferences From Perched Water

Flow directions in the lower part of the unsaturated zone may be inferred from the structural contours of the upper surfaces of units that have been identified as potential perching layers, based on field observations and measurements of their low-intrinsic permeability. These units include 1) the densely welded, crystal-poor vitrophyre of the Topopah Spring Tuff (Tptpv3) (Figure 5.3-190), which is suspected to be the uppermost perching layer in the northern part of Yucca Mountain at least as far south as SD-12, and 2) the surface marking the transition from vitric or devitrifed tuff to tuff with a significant fraction of zeolites (Figure 5.3-191). Based on these contours, the dominant directions of flow resulting from perching would be to the southeast in areas north of and coinciding with Drill Hole Wash, and in an eastwardly direction in areas south of Drill Hole Wash. Drill Hole Wash coincides roughly with a structural trough in these surfaces, and water perched on these surfaces would tend to flow toward Drill Hole Wash, unless intercepted by faults. Current understanding of the distribution of zeolitic material in the lower unsaturated zone is that it is to some degree stratabound. That is, there are intervals below the site vitric-zeolitic surface that have not been extensively zeolitized, such as the devitrified, partially welded zone of the Prow Pass Tuff, that remain as highly permeable conduits within an otherwise relatively impermeable zeolitic-tuff framework. Therefore, more rapid flow may occur beneath the perched water when water enters these intervals where they intersect the more regional vitric-zeolitic interface (see Subsection 5.3.4.3.2.2).

5.3.4.3.2.4 Fracture-Matrix Interaction

It is important to determine to what degree matrix saturations and water chemistry reflect the overall movement of water through the rock if these are to be considered reliable indicators of the rates and quantities of present-day water movement. In general, water movement within fractures may be attenuated by absorption of water into the rock matrix ("imbibition") across the wetted fracture surface, or by discontinuities in the fracture network, which require that the water continue its downward movement through the rock matrix. Solutes dissolved in the water moving within fractures may enter the matrix during imbibition and by diffusion. Diffusion may take place even if there is no net transfer of fluid. Matrix imbibition requires the presence of fillable pore space, and differences in water potential between fractures and matrix to drive the transfer of water. The transfer of fluid and dissolved solutes between fractures and matrix is facilitated when the surface area available for the exchange (the "wetted" surface area) is large, and inhibited when the surface area available for exchange is small. The wetted surface area depends on the whether many or relatively few fractures are conducting flow, which in the unsaturated zone is a function of not only the fracture density and connectivity, but also on the availability and distribution of water sources, and degree of channelization within the fractures. Arguments concerning the strength of fracture-matrix interactions are based largely on circumstantial evidence provided by geochemical data, water-potential measurements, and observations of the character of fracture coatings.

Some inferences regarding the strength of fracture-matrix interactions can be made on the basis of geochemical evidence, in particular the distribution of ^{3}H and ^{36}Cl in boreholes and the Exploratory Studies Facility (e.g., Yang et al. 1996, 1998; Fabryka-Martin, Wolfsberg et al. 1997; Sweetkind, Fabryka-Martin et al. 1997). The presence of ^{3}H and ^{36}Cl at concentrations implying an origin

related to nuclear-weapons testing deep within the unsaturated zone demonstrates that incomplete attenuation occurred between the ground surface and the sample depth as a result of matrix imbibition and diffusion. However, imbibition or diffusion occurred to some degree, at least at the sample depth, or these same environmental tracers would not have been detected in the rock matrix. The presence of bomb-pulse ^{36}Cl in the Exploratory Studies Facility appears to be correlated with faults that cross-cut the PTn, thereby reducing the effective porosity along the flowpath. However, not all of the samples in the Exploratory Studies Facility with $^{36}Cl/Cl$ ratios indicating an unambiguous bomb-pulse origin (> 1250 x 10^{-15}) are associated with faults (Sweetkind, Fabryka-Martin et al. 1997), and numerous samples with $^{36}Cl/Cl$ ratios between 500 to 1250 x 10^{-15} are present, which may be the result of either the mixing of smaller amounts of bomb-pulse water and Holocene water, or water that infiltrated when meteoric $^{36}Cl/Cl$ was decreasing from the high values that typified much of the late Pleistocene (1250 x 10^{-15}) to lower values typical of most of the Holocene (500 x 10^{-15}). If these samples with intermediate $^{36}Cl/Cl$ ratios are interpreted to be the result of mixing of a small amount of bomb-pulse water with Holocene water, it would imply that buffering of pulsed infiltration by the PTn outside the fault zones is less complete than some modeling studies have suggested (Fabryka-Martin, Wolfsberg et al. 1997). This conclusion would be consistent with the analysis of the chloride data from the unsaturated zone and shallow-saturated zone (see Subsection 5.3.4.3.1.4), in which it was estimated that, as a result of water movement down fractures and faults, 90 percent of the percolation flux bypassed the matrix of the PTn and approximately 80 percent of the flux bypassed the matrix of the Calico Hills Formation. These conclusions were based on the large differences in chloride concentrations of pore water from the PTn and the Calico Hills Formation, and more dilute, perched, and shallow saturated-zone groundwater samples. As discussed earlier in Subsection 5.3.4.3.1.4, the differences between the chloride concentrations in pore water from the PTn and the perched-water samples also might be attributable to changes in the infiltration rates and chloride concentrations over time (Sonnenthal and Bodvarsson 1997), as well as to inadequate characterization of PTn pore water chloride concentrations in areas where alluvium is thin or absent, such as beneath sideslopes and ridgetops (Fabryka-Martin, Flint et al. 1997).

In two locations, water samples have been collected from both fractures and matrix and the matrix water does not appear to be in chemical equilibrium with the fracture water (Yang et al. 1996). However, this apparent disequilibrium may have resulted because the fracture and matrix water were not in direct contact prior to the drilling of the borehole. At UZ-14, the perched water pumped and bailed from the borehole was considerably more dilute than water squeezed from the matrix of the adjacent non- to partially-welded subzone of the crystal-poor vitrophyre of the Topopah Spring Tuff. However, the perched water filled the borehole adjacent to the matrix sample depths only after open fractures were encountered during drilling. A similar situation existed where UZ-16 penetrated the regional potentiometric surface in the devitrified interval of the Prow Pass Tuff. Water rose in the borehole only after the borehole had penetrated an open fracture several meters below the potentiometric surface and the matrix sampling depths. In both these instances, the higher concentrations of most solutes in the matrix pores should have resulted in chemical diffusion of these species from the matrix into the more dilute water in the fractures.

The high water potentials (−4 to 0 bars) that have been measured throughout much of the unsaturated zone indicate that strong water potential gradients that would promote matrix imbibition of water

flowing within fractures are absent. High water potentials throughout the unsaturated zone are indicated by:

- Psychrometers in instrumented boreholes (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997)

- Heat-dissipation probes emplaced in shallow boreholes drilled into the walls of the Exploratory Studies Facility (YMP-USGS Data Package, "Heat Dissipation Probe Measurements in the South Ramp of the ESF at Stations 66+99, 67+21, and 67+33," September 1997, DTN GS970908312242.010, TBV)

- Air pressures at which water is displaced in the rock during air-injection tests (LeCain 1997; LeCain et al. 1997; LeCain and Patterson 1997; Subsection 5.3.3.1.2)

- The near-saturated rock matrix throughout much of the zeolitic Calico Hills Formation (which also lacks fillable pore space).

The absence of strong water-potential gradients to drive matrix imbibition suggests the minimal importance of imbibition as a mechanism for removing water and solutes from fractures in units where flow is primarily fracture controlled.

Finally, it has been observed that secondary mineral deposits in moderately dipping fractures are confined to the fracture surface along the footwall, and that the fracture surface defined by the hanging wall block generally lacks secondary mineral deposits (Paces, Neymark et al. 1996; Paces, Marshall et al. 1997). The same authors report that in near-vertical fractures, secondary mineral deposits alternate between fracture surfaces. In addition, Beason (Eatman et al. 1997) has reported that approximately 8 percent of the fractures sampled in the Exploratory Studies Facility by scanline surveys contain secondary mineral deposits. These observations suggest that fracture flow wets a very small fraction of the surface area of the matrix blocks defined by the mapped fracture distribution, and that fracture-matrix exchange processes may be similarly limited as a result of channelized flow within the fracture network and within the plane of individual fractures.

5.3.4.4 Three-Dimensional Unsaturated-Zone Modeling

5.3.4.4.1 Introduction

As described in the previous subsections, Yucca Mountain has been extensively studied by the DOE and other governmental agencies for more than 15 years, during which time many types of data have been collected. These data have been used in the development of conceptual and numerical models of the hydrological behavior of the site to simulate ambient conditions and perform predictive studies of changes in the mountain caused by climate-related, thermal, and geochemical perturbations. The comprehensive model that integrates all pertinent data from the unsaturated zone at Yucca Mountain is the 3-D site-scale unsaturated-zone model, developed over the past decade at the Lawrence

Berkeley National Laboratory (Wittwer et al. 1995; Bodvarsson and Bandurraga 1996; Bodvarsson et al. 1997, Chapter 11).

The primary objectives of the unsaturated-zone hydrologic model development are:

- To integrate the available data from the unsaturated zone into a single, comprehensive 3-D model.
- To quantify the flow of moisture, heat, and gas through the unsaturated zone, under present-day and hypothesized future climate scenarios.
- To evaluate the effects of thermal loading on moisture, gas, and heat flow within the mountain.
- To provide Performance Assessment and Repository Design with a defensible and credible model of all relevant unsaturated-zone processes.

The unsaturated-zone hydrologic model provides estimates for important parameters and processes such as:

- The spatially and temporally averaged values of the percolation flux at the potential repository horizon
- The components of fracture and matrix flow within and below the potential repository horizon
- The probable flowpaths from the potential repository to the water table

The unsaturated-zone flow model is an important process model for the YMP's Repository Safety Strategy and for support of the License Application. The Total System Performance Assessment (TSPA) will use the unsaturated-zone flow model to provide input to other models such as ambient and thermal drift-scale models, the mountain-scale thermohydrological model, and the unsaturated-zone transport model; the interrelationships between these various models are shown schematically in Figure 5.3-192.

Much emphasis is being placed on preparing a defensible and credible unsaturated-zone model for the 1998 Viability Assessment of Yucca Mountain as a suitable location for the underground radioactive waste repository. Major activities associated with the unsaturated-zone model include:

- Model calibrations
- Development of hydrologic properties data sets
- Integration and incorporation of all relevant data into the model

- Evaluation of the spatial distribution of the percolation flux
- Evaluation of the impact of future climate changes at Yucca Mountain on the behavior of the unsaturated-zone hydrologic system

The implementation of a quality assurance program in recent years by participants on the YMP is intended to ensure the integrity of scientific findings with regard to the site. Many of the data recently collected are considered quality assured; however, much of the earlier work at Yucca Mountain is not quality assured, especially in areas away from the potential repository block. The vast majority of recent unsaturated-zone analyses are based on qualified data and computer codes. As this subsection is basically a summary of the analyses performed with the unsaturated-zone model, the designation of the quality assured status of data and software will not be provided here. For detailed information about the quality assured status of the data used in the unsaturated-zone model and unsaturated-zone model conclusions, the reader is encouraged to refer to the referenced documents provided at the end of this subsection.

Other activities undertaken to enhance the unsaturated-zone flow model include investigation of conceptual models of fracture-matrix interaction, development of a comprehensive fracture property data set, development of various geochemical models to evaluate flow patterns and percolation fluxes, and assessment of various fracture flow modeling approaches. A preliminary surface hydrology model for the investigation of infiltration patterns has also been developed. Evaluation of postclosure moisture seepage into excavated drifts within the potential repository horizon is another important activity being conducted. One current modeling effort is the investigation of construction-water plume migration from the Exploratory Studies Facility, the 8 km long tunnel that has recently been completed through the upper rock formations of the mountain.

These various activities require the integration and utilization of numerous data collected at Yucca Mountain over the past 15 years. The data include surface-based measurements and outcrop sampling, borehole logs with lithostratigraphic descriptions and contact depths, core sample property measurements, geophysical logging data, and information obtained from the Exploratory Studies Facility. Of major importance are the data recently collected from the Exploratory Studies Facility, which include fracture mapping data, environmental isotope and hydrochemical data, fracture coating data, preliminary fault characteristics from testing of the Ghost Dance fault, data from the ambient characterization of the heater test area, and preliminary results of various hydrological studies currently underway in the Exploratory Studies Facility.

Hydrological testing is being conducted in the tunnel to determine the percolation flux and the rate of moisture seepage into drifts. More reliable estimates of percolation flux and seepage into drifts will be possible when the niche studies and the percolation flux tests have been completed. In addition to the various hydrological studies in the Exploratory Studies Facility, the construction of a cross drift (east-west drift) is expected to provide important confirmatory data for the unsaturated-zone model. Confidence in the model will be enhanced if and when predictions of the model can be compared to observations and test results in the cross drift. Therefore, use of the unsaturated-zone model for Viability Assessment activities represents an intermediate step towards developing a more realistic unsaturated-zone model to be used for the License Application. Enhancement of the model

will continue as new test data are collected and incorporated into the model. Figure 5.3-193 summarizes the growth and major development stages of the unsaturated-zone model, including the 1995 model (Wittwer et al. 1995), the 1996 Unsaturated-Zone Flow Model report (Bodvarsson and Bandurraga 1996), and the 1997 Unsaturated-Zone Model YMP milestone report (Bodvarsson et al. 1997).

5.3.4.4.2 Previous Work on Unsaturated-Zone Modeling of Yucca Mountain

The 3-D unsaturated-zone model of Yucca Mountain has evolved from simple 1-D and 2-D models that included little geological or hydrological detail, to a fully coupled 3-D model that incorporates much of the known geological and hydrological complexity of Yucca Mountain. These models consist of: discretized computational grids that represent the hydrogeological structure of the potential repository location and its vicinity; a collection of rock property data, such as porosities, permeabilities, etc., that represent the various hydrogeological units, faults, etc.; and a set of boundary conditions such as the location of the water table and surface-infiltration rates. The grid, rock properties data, and boundary conditions are used as input to a numerical code that simulates fluid flow processes in porous media. The earliest versions of the unsaturated-zone model were run with the TOUGH code (Transport of Unsaturated Groundwater and Heat; Pruess 1987), which is a numerical computer program that simulates the transport of water, air, and heat in an unsaturated porous medium. Later versions of the model used TOUGH2 (Pruess 1991), which is an enhanced version of TOUGH that contains additional capabilities such as solute transport, etc. All versions of the TOUGH family of codes use the method of integrated finite-differences (Narasimhan and Witherspoon 1976) to represent the equations of conservation of mass and conservation of energy in an algebraic form; these algebraic equations are solved by various numerical techniques, such as Gaussian elimination or the conjugate-gradient method (Moridis and Pruess 1995). The ability of the TOUGH2 code to accurately simulate unsaturated flow processes has been documented by Pruess and Antunez (1995).

An important early step in the development of an unsaturated-zone hydrological model for Yucca Mountain was the work of Rulon et al. (1986). They used 2-D cross-sectional models to investigate various important flow processes at Yucca Mountain such as lateral flow within the Paintbrush unit and the effect of flow partitioning between fractures and the rock matrix. In the early 1990s, Wittwer and coworkers at Lawrence Berkeley National Laboratory and the USGS (Wittwer et al. 1992, 1995) began the development of a complex 3-D hydrological model that incorporated many geological and hydrological complexities such as geological layering, degree of welding of the rock layers, fault offsets, and differing matrix and fracture properties. This work resulted in a report that addressed many processes and hydrological issues, and showed complex moisture flow patterns caused by the dipping of the hydrological units and the offsets of the faults. Ahlers, Bandurraga et al. (1995a, 1995b) continued the development of the 3-D unsaturated-zone model, adding increased spatial resolution. Their studies considered processes such as gas pressures and gas flow, and temperature and heat flow analyses. Inverse modeling, in many cases performed with the code ITOUGH (Finsterle 1993), was used to estimate rock property values by varying the physical properties until an optimal fit was achieved between computed and measured values of parameters such as liquid saturations and water potentials.

Much progress was made in the development of the unsaturated-zone model during fiscal year 1996 (Bodvarsson and Bandurraga 1996). Calibration exercises conducted with the model included matching the saturation and water potential data, gas pressures and gas flow rates, ambient temperature and heat flow data, locations of perched water bodies, and environmental isotope concentrations. Preliminary work began on model calibrations with respect to percolation fluxes inferred from fracture-coating data. Coupled calibrations of various state variables (such as liquid potentials, liquid saturations, etc.) were also started, providing tighter parameter constraints. In addition to model calibration, considerable effort was devoted to parameter uncertainties and sensitivity studies for fracture and matrix properties, infiltration patterns, fault properties, and percolation fluxes. Studies of the various numerical approaches (e.g., effective continuum method, dual-porosity method, dual-permeability method, etc.) and their appropriate use for different processes such as moisture flow, gas flow, heat flow, and chemical transport were also presented.

The most recent improvements and advances in the unsaturated-zone hydrologic model are described in the milestone report by Bodvarsson et al. (1997). This report describes the continued model calibrations, incorporation of Exploratory Studies Facility data, development of hydrologic properties data, evaluation of the spatial distribution of the percolation flux, and the evaluation of the impact of future climate changes at Yucca Mountain. Ongoing unsaturated-zone modeling activities include the investigation of various conceptual models for fracture-matrix interaction, development of a comprehensive fracture property set, evaluation of postclosure moisture seepage into drifts, evaluation of construction-water plume migration from the Exploratory Studies Facility, development of various geochemical models to evaluate flow patterns and percolation flux, and assessment of fracture flow modeling approaches. The conceptual ideas and results of the numerical modeling efforts contained in that report are emphasized in the following discussion of the unsaturated-zone flow model.

5.3.4.4.3 Conceptual Model of the Unsaturated Zone

The conceptual model consists of a framework to explain unsaturated-zone flow processes such as liquid, heat, and gas flow. Important components of the conceptual model include:

- Geological framework
- Hydrologic properties
- Spatially variable infiltration
- Percolation flux
- Capillary barriers and lateral diversion of flow
- Characterization of major faults and their effect on flow
- Fracture flow and fracture-matrix interaction
- Perched water and perching layers
- Environmental isotopic signatures deep within the unsaturated-zone
- Gas and fluid ages based on isotope measurements
- Geothermal gradients

The conceptual model of the unsaturated zone at Yucca Mountain has been developed through the evaluation of collected data, and has undergone continual revision based on the results of modeling

studies and calibration tests. This conceptual model is based to a great extent on the ideas originally presented by Montazer and Wilson (1984). Since the development of this initial conceptual model, most of the data collected have been in general agreement with its ideas and concepts. Figure 5.3-194 schematically depicts the important data collected at Yucca Mountain, relating them to various aspects of the conceptual model. A summary of the key conceptual issues is provided in the subsections that follow.

An important fundamental choice that is posed in any attempt to develop a hydrological model of the unsaturated zone at Yucca Mountain is whether to use a heterogeneous continuum model, or a discrete feature model for the dominant flowpaths. It is a simple matter to estimate that the total number of fractures (conceptualized, for the purposes of the present argument, as vertical sheet-like structures), and hence also matrix blocks, in Yucca Mountain is on the order of millions. This estimate follows from the observation that the mean fracture spacing observed in units such as the Tiva Canyon and Topopah Spring is about 50 cm, whereas the lateral extent of the potential repository region is at least 1,000 m in each direction. A basic question that must be answered before proceeding with the development of an unsaturated-zone hydrologic model is whether the flow takes place primarily through a small number of preferential paths, or whether flow occurs pervasively through a sufficiently dense network of flowpaths to justify the use of a continuum model. Although there is little hard data to allow one to distinguish conclusively between these two alternative flow models, there is much circumstantial evidence that flow does in fact occur through a very large number (at least thousands) of fractures. This conclusion, the evidence for which is discussed below, justifies the development of a numerical model based on a continuum representation of the fracture networks at Yucca Mountain.

Some evidence pointing to the quasi-uniformity of the flow field at Yucca Mountain can be found in the fact that matrix saturations, measured on cores from surface boreholes as well as from boreholes within the Exploratory Studies Facility, show nearly uniform saturations for most of the units within the mountain (Flint, L.E. 1998; Subsection 5.3.3.1.1). In situ measurements of moisture tension also show little variability; in fact, the moisture tension throughout the entire unsaturated zone seems to be in the relatively narrow range of 3 to 6 bars (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997; Subsection 5.3.4.2.2). Additional evidence for dispersed flow is supplied by the observation that calcite-filled fractures are distributed fairly evenly throughout the Topopah Spring unit. Accepted models of calcite deposition on fracture surfaces imply that these deposits could not occur as a result of occasional, rapid pulses of water moving through the fractures; slow, continuous flow would seem to be needed to cause such pervasive deposits.

Finally, recent measurements made within the Exploratory Studies Facility have shown that, at all locations, humidity returns to nearly 100 percent soon after ventilation is halted (Wang et al. 1996). As the adjacent matrix blocks are not fully saturated, and have very low permeabilities (on the order of microdarcies), it would seem that the rapid increase in the humidity within the Exploratory Studies Facility could only be fed by water that flows in from the more permeable fractures. All of the various general observations mentioned above support the conclusion that flow takes place

pervasively throughout the unsaturated zone at Yucca Mountain, rather than through a small number of discrete flow channels (Pruess, Faybishenko et al. 1997).

5.3.4.4.3.1 Geological Framework

Subsurface hydrologic processes at Yucca Mountain occur within heterogeneous-volcanicrocks that have been welded and fractured to varying degrees (see Subsection 3.3.6.3). These volcanic rocks consist of alternating layers of welded and nonwelded ashflow and airfall tuffs, deposited during the Miocene epoch (from 5 to 24 million years before present). The primary geologic formations found at Yucca Mountain are (beginning from the land surface) the Tiva Canyon, Yucca Mountain, Pah Canyon, and Topopah Spring members of the Paintbrush Group. Underlying these are the Calico Hills Formation and the Prow Pass, Bullfrog, and Tram Tuffs of the Crater Flat Group (Buesch, Spengler et al. 1995). The geologic formations have been divided into hydrogeologic units (Montazer and Wilson 1984), based roughly on the degree of welding. These are the TCw, the PTn (consisting primarily of the Yucca Mountain and Pah Canyon members and the interbedded tuffs), the TSw, the CHn, and the Crater Flat undifferentiated units. As shown in Table 5.3-61 (see also Subsection 3.3.6.4), the hydrogeological unit boundaries in all cases do not coincide exactly with the geological unit boundaries. This is because the margins of the major cooling units (such as the Topopah Spring Tuff) often display significantly different hydrologic characteristics than are found at the core of the unit, reflecting the rock's cooling history.

The welded units typically have low matrix porosities and high fracture densities, whereas the nonwelded and bedded tuffs have relatively higher matrix porosities and lower fractures densities (Montazer and Wilson 1984). At smaller scales, the fracture density is correlated with increases in the degree of welding of the volcanic rocks. Portions of these welded and nonwelded units are altered to zeolites or clays, depending on their cooling history and the presence of moisture. Secondary mineralization of the glassy component of the tuff is characterized by zeolitic alteration (Loeven 1993). Therefore, altered tuffs are not a significant component in those layers where welding, vapor-phase alteration, or devitrification has occurred. Differences in the amount of zeolites in the various pyroclastic and bedded tuff units may be due to variations in their content of primary glass (Bish 1989; Bish and Chipera 1989). Alteration does not affect porosities greatly but decreases the permeabilities of the formations where it occurs (Subsection 5.3.3.1.1). The presence and distribution of altered zones consequently has an important effect on flow patterns in the unsaturated zone.

The representation of geologic complexities and alteration history at Yucca Mountain in the unsaturated-zone model was described in detail by Bodvarsson and Bandurraga (1996). In general, model layers correspond explicitly to major geologic formations. Table 5.3-61 shows the relationships between the Lawrence Berkeley National Laboratory unsaturated-zone model layers and the major geological formations and hydrogeological units, as well as the Buesch, Spengler et al. (1995) and Moyer and Geslin (1995) nomenclature for the geological formations. In places where the geologic formations are relatively thin, such as in the bedded tuffs, model layers represent the combined thicknesses of two or more formations. Zones of zeolitic alteration are represented in the model where they are known to exist. Complexities found in the geological model, such as faults and their associated offsets, dipping beds, and eroded units, are captured by the unsaturated-zone

model numerical grid. An east-west geological cross-section through the unsaturated zone, with the 3-D unsaturated-zone model numerical grid superimposed on it, is shown in Figure 5.3-195.

5.3.4.4.3.2 Infiltration

Infiltration at Yucca Mountain is spatially and temporally variable due to the nature of storm events that supply precipitation (Subsection 5.3.4.1). The net infiltration to the bedrock geologic formations comprising the host rock of the potential repository (the Paintbrush Group) is also dependent on variations in soil cover and topography. The current conceptual model for infiltration is based on numerous measurements of water-content profiles in shallow boreholes at Yucca Mountain, verified by evaluations of other data sets, such as geothermal profiles and geochemical measurements. Any significant thickness of alluvium overlying the fractured volcanic rock will not allow an infiltration pulse to penetrate to the underlying bedrock because of attenuation of the pulse by storage in the alluvium. Therefore, infiltration is higher on sideslopes and ridgetops, where outcrops are exposed and fracture flow can be initiated directly into the bedrock (Flint, A.L. and Flint 1994).

Temporal variations in precipitation are important for modeling studies evaluating conditions in the TCw and PTn, respectively, near the land surface. Near-surface infiltration data (Subsection 5.3.4.1) suggest that significant infiltration into the mountain occurs only once every few years (Flint, A.L. et al. 1996). Therefore, in most years, little or no infiltration occurs. The amount of infiltration occurring during wet years varies greatly, since it is dependent on storm intensity, durations, and frequencies. In very wet years, infiltration pulses on the order of several tens mm/year may infiltrate into the mountain within relatively short periods of time. However, when the water enters the PTn, the relatively high matrix permeabilities and porosities, and low fracture densities, greatly attenuate infiltration pulses. In general, this spatial and temporal attenuation of infiltration pulses by the PTn leads to near steady-state conditions in the underlying TSw; this conceptual hypothesis is supported by the results of modeling studies (e.g., Wang and Narasimhan 1985; Finsterle et al. 1996), but field data to support this hypothesis are lacking. As many modeling studies focus on the potential repository horizon in the TSw, the assumption is usually made that flow in the potential repository horizon can be evaluated with maps that provide steady-state infiltration rates. This assumption is also believed to be valid for studies of the hydrogeologic units underlying the TSw (e.g., the CHn and CFu units).

Several different spatially variable steady-state infiltration maps have been developed by YMP scientists. The estimated average infiltration rates over the unsaturated-zone model domain from the recent maps are 1.22 mm/year (Flint, A.L. and Flint 1994), 0.35 mm/year (Hevesi, J.A. and Flint, A.L., USGS-WRIR-96-4123, in press), 4.9 mm/year (Flint, A.L. et al. 1996) and greater than 10 mm/year (Hudson and Flint 1996). However, the A.L. Flint et al. map (Figure 5.3-196) is presently considered to contain the most reliable estimates of infiltration, based on the judgment of the principal investigator for infiltration studies (A.L. Flint, USGS, January 1997, personal communication to G.S. Bodvarsson, LBNL). Other maps have been developed showing the predicted changes in precipitation due to possible climate changes (Schelling and Thompson 1997).

5.3.4.4.3.3 Percolation Flux

The fractured nature of the TCw unit is such that it has generally been assumed that infiltration pulses will move rapidly through the fracture system (relative to travel times in the matrix), and thus will not be significantly attenuated. This is supported by the observation of rapid changes in gas pressures (Subsection 5.3.4.2.1) at the bottom of the TCw at many borehole locations (Subsection 5.3.4.2.4). Pneumatic sensors in the TCw show little attenuation of the barometric signal in monitoring boreholes compared with the barometric signal observed at the land surface. The percolation flux through the TCw may be on the order of tens or even hundreds of mm/year at certain locations. Anomalously high levels of ^{36}Cl and ^{3}H (associated with nuclear weapons testing over the last 40 years, and consequently often referred to as bomb-pulse) measured in TCw and PTn samples at many borehole locations (see Subsection 5.3.4.2.4) also support the concept that flow through the TCw is dominantly through fractures.

However, when the water enters the PTn, radically different flow processes are hypothesized to occur. Few samples obtained from areas other than the top of the PTn rarely show anomalous signatures of ^{36}Cl and ^{3}H. The samples that do show anomalously high levels of these isotopes are generally thought to be associated with fast flow pathways through localized fault features (Fabryka-Martin, Wolfsberg et al. 1997). The PTn has relatively high-matrix permeabilities and porosities (about 0.30) and low-fracture densities. Pneumatic data confirm that the degree of fracturing within the PTn is rather limited and that, except for boreholes located near mapped faults such as UZ-7a, UZ#4, and UZ#5, the pneumatic signal is propagated predominantly through the high-porosity matrix (Ahlers, Shan et al. 1996; J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997; Subsection 5.3.4.2.1). Since a majority of the flow probably occurs in the relatively high-porosity matrix in this unit, the PTn greatly attenuates infiltration pulses. The time required for a typical water pulse to move through a 100 m thick portion of the PTn unit, as is found north of Drill Hole Wash, would be on the order of several thousand years, assuming that matrix flow is dominant, and using average calculated PTn permeabilities.

Depending on geologic conditions and the magnitude of the flux, flow concentrations leading to fast pathways across the PTn may occur. Relatively high fracture densities generally associated with fault features may provide such fast pathways. These pathways are possibly responsible for the presence of high levels of ^{36}Cl and tritium within the TSw at the potential repository horizon (Fairley and Sonnenthal 1996; Fabryka-Martin, Wolfsberg et al. 1997). In general, however, modeling results show that the infiltration pulses are spatially and temporally attenuated by the PTn unit, leading to near steady-state conditions in the TSw. The ability of the PTn to temporally buffer and spatially redistribute infiltration probably decreases south of Drill Hole Wash, in the potential repository area, where the unit thins to about 25 m (E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation).

The percolation flux at the level of the potential repository, the middle nonlithophysal formation of the TSw (Tptpmn), is one of the most important hydrological aspects of the site that can be

investigated through the use of the unsaturated-zone model. Data exist to suggest that the amount of water reaching the potential repository level is relatively small and sparsely distributed. Observations in the Exploratory Studies Facility have not found weeping fractures, even in areas above perched water bodies (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). One exception is the recent finding of a wet fracture in a niche located in the Main Drift of the Exploratory Studies Facility; the niche was partly excavated without the use of water. A wet area was observed in the PTn in the South Ramp of the Exploratory Studies Facility in an area crossed by numerous faults, but more recent data suggest that the water was a result of Exploratory Studies Facility tunneling operations. Work by Paces, Neymark et al. (1996) on the geochemistry of fracture coatings also supports the concept of low percolation flux through the TSw—perhaps as little as 2 mm/year.

However, other data suggest that fracture flow is significant, and that percolating water may reach the potential repository level through this mechanism (also see Subsection 5.3.4.3). An analogue site at Rainier Mesa, which has an average annual precipitation that is approximately double that of Yucca Mountain, is reported to have a liquid infiltration rate of about 23.7 mm/year infiltrating deep into the subsurface through fault or fracture flow (Wang et al. 1993). The fact that weeping fractures are usually not observed at Yucca Mountain may be due to the Exploratory Studies Facility tunnel ventilation system, which removes moisture from the faces of the tunnel before it can form observable seeps. Also, assuming a unit gradient and using average matrix permeabilities measured in the TSw geologic formations, the matrix percolation rate cannot be more than 1 to 2 mm/year, only a fraction of the average infiltration rate estimated from the spatially varying map developed by A.L. Flint et al. (1996). This conclusion also assumes that lateral diversion within the PTn does not significantly reduce the vertical percolation flux at the Exploratory Studies Facility level. Additional data supporting significant fracture flow in the TSw include observations and tests of perched water bodies of fairly young age located below the potential repository horizon (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). The young ages of the water indicate that fluid arrives at the perched water body much faster than could be accounted for solely through matrix flow.

An analysis of available temperature and heat flow data by Bodvarsson et al. (1997) suggests that these data may be helpful in bracketing current values of percolation flux across the potential repository area. The results of the analysis indicate that the average percolation flux in the potential repository area is 5 to 10 mm/year. Three-dimensional unsaturated-zone modeling results of the percolation flux at the potential repository horizon using the latest infiltration map by A.L. Flint et al. (1996) show very similar percolation rates. The temperature analysis suggests that this range of percolation flux is prevalent over most of the potential repository area, between the Solitario Canyon fault and the Ghost Dance fault. However, the available data are limited, and the method of analysis used to reach this conclusion has not been validated for this type of problem.

5.3.4.4.3.4 Capillary Barriers and Lateral Flow Diversion

The PTn has been hypothesized to create lateral flow diversion through a capillary barrier effect (Montazer and Wilson 1984; Bodvarsson and Bandurraga 1996). This well-known effect is due to the inability of water at a low pressure in a fine-pored medium to penetrate into a layer that has a larger pore size (Ross 1990; Oldenburg and Pruess 1993), despite the fact that the coarser material has a larger absolute permeability. If the interface between the finer and coarser media is not horizontal, the water will flow laterally down the slope until it encounters another vertical flowpath. Moisture is thought to be diverted laterally through the fractures in the TCw above the PTn, or possibly within the pores of the PTn matrix where they overlie the larger openings of the fractures of the TSw, following the dip of the geologic formations until intercepting a fault or a fracture that can provide an avenue for downward flow. However, this mechanism requires the existence of an abrupt and laterally continuous contact between the TCw and PTn or PTn and TSw. Recent observations in the Exploratory Studies Facility suggest that the TCw-PTn contact can be gradational, and that numerous small faults with minor offsets interrupt this and the PTn-TSw contact's lateral continuity (L.E. Flint, USGS, April 1996, personal communication to G.S. Bodvarsson, LBNL). The results of the fiscal year 1996 modeling studies reported in Bodvarsson and Bandurraga (1996) show that lateral diversion of flow caused by the capillary barrier effect of the PTn will decrease with increasing infiltration rate.

5.3.4.4.3.5 Characterization of Major Faults and Their Effect on Flow

The potential repository area is surrounded and crossed by numerous strike-slip and normal faults with varying amounts of offset and brecciation (Subsection 3.6.2.4). Figure 5.3-197 shows the location of many major faults at Yucca Mountain, as well as the potential location of the repository. The Ghost Dance fault along the eastern boundary of the potential repository location has offsets ranging from meters to tens of meters (Spengler, Braun et al. 1993), while the Solitario Canyon fault, located just west of the potential repository, has offsets as large as hundreds of meters. Modeling studies have been conducted under the assumption that the faults act either as capillary or permeability barriers (Wittwer et al. 1995), although the evidence for either situation is mixed, as is demonstrated below.

Pneumatic data (Subsection 5.3.4.2.1) indicate that most faults are permeable to gas, thereby suggesting that they are less than fully saturated with water. The pneumatic data from neighboring boreholes UZ#4 and UZ#5 indicate the presence of a fault with a very high lateral permeability connecting these boreholes to the Exploratory Studies Facility. However, the fault must have a relatively low vertical permeability from the TSw to the surface (Ahlers, Shan et al. 1996), since the surface gas-pressure signal is attenuated in the PTn to a similar extent as in boreholes that are not affected by faults.

Studies in the Exploratory Studies Facility indicate that the permeability of the Bow Ridge fault is approximately 20 darcies, about the same as the permeability inferred from air-permeability testing of highly permeable bedded tuff formations or highly fractured welded units (LeCain et al. 1997; LeCain 1998). A darcy is equal to approximately 1×10^{-12} m^2; a millidarcy is equal to approximately 1×10^{-15} m^2; and a microdarcy is equal to approximately 1×10^{-18} m^2.) Furthermore, data from Nye

County's Borehole ONC#1 (N. Stellavato, Nye County, August 1995, personal communication to G.S. Bodvarsson, LBNL) show that the geothermal temperature profile is offset by several degrees as the borehole passes through the Bow Ridge fault zone. This indicates that the fault may be very permeable to gas or moisture flow, thus decreasing the temperature in that region.

Borehole UZ-7a was drilled into the Ghost Dance fault zone, and exits the fault zone in the middle of the TSw, due to the dipping fault plane. Matrix saturations in the fault zone are not significantly different than the matrix saturations in other boreholes outside of the reported fault zones. It is unclear if this result indicates that the matrix properties in the fault are similar to matrix properties outside of reported fault zones, or rather that the saturations are dependent on a balance between flow that is laterally diverted by the PTn and moisture evaporation due to gas flow in the fault. However, none of the boreholes near the Ghost Dance fault show an increase in matrix saturations consistent with lateral diversion. Temperature data for borehole UZ-7a, which penetrates the Ghost Dance fault, suggest that the average percolation flux within the fault is similar (or slightly less) than percolation through the surrounding fractured porous rock (Bodvarsson et al. 1997). The possible presence of bomb-pulse tritium measured in the North Ghost Dance fault in the Exploratory Studies Facility (Subsection 5.3.4.2.4) does, however, provide evidence that some rapid flow takes place in the fault zone.

Pneumatic data for borehole UZ-7a show that the Ghost Dance fault zone is highly fractured (Ahlers, Shan et al. 1996). The gas pressure signal is propagated with little attenuation into the TSw, consistent with flow through a low-porosity, high-permeability fracture system. The UZ-7a sensors in the TSw below the fault zone show an attenuated signal, implying the existence of a lateral permeability barrier, i.e., a condition wherein the permeability in the direction perpendicular to the fault is much less than in the direction parallel to the fault. In fact, the data show a degree of anisotropy in the fault zone that is consistent with a lateral permeability barrier.

Additional evidence supports the hypothesis that the faults may serve as permeability barriers in those locations where zeolites are offset across the fault. For example, the water body observed at borehole SD-7 is thought to be perched over a zeolitic layer and prevented from lateral movement by the presence of the Ghost Dance fault (Striffler et al. 1996), which has offset the zeolitic layer downward to the west. J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) also argued that the perched water body intercepted by borehole UZ-14 is caused by a Solitario Canyon fault splay that has offset the zeolitic Calico Hills Formation, and thereby blocks lateral movement of the water. Additionally, the potentiometric surface elevation on the western side of the Solitario Canyon fault is approximately 40 m higher than the surface on the eastern side of the fault (Tucci and Burkhardt 1995). This gradient could only be maintained if the Solitario Canyon fault is somehow a permeability barrier to flow.

5.3.4.4.3.6 Perched Water

The presence of perched water (see Subsection 5.3.4.2.2) has implications for travel times and flowpaths of water through the unsaturated zone, and may affect the construction, operation, and performance of the potential repository. Perched water may occur where large permeability differences exist between adjacent geologic formations (Freeze and Cherry 1979). Perched water zones at Yucca Mountain were reported by Striffler et al. (1996) in the following boreholes: USW UZ-1, USW UZ-14, USW NRG-7/7a, USW SD-7, and USW SD-9. Additionally, wet zones were reported to be found in borehole USW SD-12 (Rousseau et al. 1997). However, the perched water bodies identified are at elevations significantly below (100 to 200 m) the potential repository horizon and do not represent obstacles to repository design or construction. Several pumping tests were conducted in borehole UZ-14, and water levels recovered to their initial values after about 5½ days (Striffler et al. 1996), suggesting that there is an extensive perched water body around this area. Wu et al. (1997) estimated the formational volume of the perched water body at UZ-14 to be about 50,000,000 cubic meters. This volume is equivalent to a disk-like region having a radius of 900 m and a thickness of 20 m. In contrast, pressure transient tests of the perched water zone around SD-7 indicate a perched water body that occupies only about 30,000 cubic meters of porous rock (R.R. Luckey, USGS, September 1995, personal communication to G.S. Bodvarsson, LBNL).

The majority of the perched water bodies or wet layers in these boreholes were observed in formations overlying relatively impermeable matrix material such as the TSw basal vitrophyre. Although the vitrophyre is extensively fractured, in many locations the fractures have been filled with clays and zeolitic material that impede vertical flow (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Additionally, in borehole SD-7, and possibly elsewhere, portions of the Calico Hills unit have been extensively altered to zeolites to create perched water bodies. Figure 5.3-198 shows the relationships between perched water elevations and low permeability horizons.

The implications of the perched water observations are that percolation fluxes are or have been higher than vertical flow rates through zeolitic layers or zeolite- and clay-blocked fractures in nonzeolitic layers. This is consistent with the relatively low zeolitic matrix permeabilities measured by L.E. Flint (1998). The perching indicates the potential for significant downdip lateral diversion above the zeolites or blocked fractures, with a significant portion of the flow occurring in the more permeable units above the perching layers. The possible existence of lateral flow along the tops of the perching layers has important implications for flow and transport in the lower unsaturated zone, as discussed in the following subsection.

Based on the conceptual model currently incorporated in the unsaturated-zone model (Bodvarsson et al. 1997), it appears that most of the perched water is created primarily through vertical flow from land-surface infiltration within the central block, rather than subsurface lateral flow of water that infiltrated outside of or on the perimeter of the central block. Infiltration along the west-facing sideslope in the northern part of Solitario Canyon may contribute to the perched water body at borehole UZ-14. However, sideslope infiltration probably cannot be the source of the perched water body at borehole SD-7, because the CHn unit under Yucca Crest to the west of borehole SD-7 is

unaltered and highly permeable, and therefore would be unlikely to divert any sideslope infiltration towards SD-7. Another source of perched water recharge could be high infiltration to the west of borehole G-2 migrating southeastward across the Solitario Canyon fault, which is downdropped to the east-north of Teacup Wash (see Subsection 3.6.2.3). However, data have not yet been collected to support this hypothesis.

Additionally, perched water is not thought to be recharged significantly from the high potentiometric surface measured in the vicinity of borehole G-2. Carbon-14 measurements at borehole UZ-14 found the perched water body to be younger in age than the water sampled at borehole G-2. If the G-2 potentiometric surface is concluded to be perched water, the fracture systems serving as flow pathways for that perched water body do not appear to be well-connected. This conclusion is supported by the fact that boreholes have been observed to act as short-circuit pathways. For instance, borehole NRG-7a was not observed to encounter significant volumes of perched water until a flowing fracture system was intercepted, after which it filled with more than 30 m of water. Drilling staff at borehole UZ#16 did not find the potentiometric surface at the expected elevation. However, upon drilling deeper, a fracture system that provided water was encountered, and the water level in the borehole rose to about the expected level (Thamir et al. 1997). These are all indications that fracture systems that could serve as flow pathways for recharge to the perched water bodies are not well-connected in the vicinity of boreholes G-2, NRG-7a, or UZ#16.

The flow predictions made by the unsaturated-zone model must be consistent with the age estimates of the perched water bodies. Carbon-14 data from the perched water bodies provide apparent age estimates ranging from approximately 2,000 to 6,000 years, after correction for the dissolution of carbonate (Yang et al. 1998). The ages are much younger than would be possible if flow reaching the perching layer occurred primarily in the low-permeability matrix. Therefore, the data indicate that a significant portion of the flow must occur in the fractures, which can permit more rapid fluid flow from the surface to the perched water body. Numerical simulations of the perched water zones discussed by Wu et al. (1996, 1997) found that the perching layers had to be of very low vertical matrix permeability (less than one microdarcy), with little or no fracture permeability, in order to produce perched water. In addition, there had to be lateral constraints on water flow by permeability barriers (with permeabilities no greater than a few hundred millidarcies) downdip of the expected perched water location. These permeabilities are about an order of magnitude lower than the global large-scale horizontal permeabilities inferred from model calibration against the pneumatic pressure data (Subsection 5.3.4.2.1), and somewhat lower than those inferred from the gas permeability packer tests (Subsection 5.3.3.1.2).

5.3.4.4.3.7 Fracture Flow and Fracture-Matrix Interactions

In general, fracture permeability and fracture connectivity in the welded rocks at Yucca Mountain are apparently very high, as most pneumatic (i.e., gas-phase) pressure sensors detect atmospheric pressure changes throughout the mountain (Subsection 5.3.4.2.1). The fracture permeabilities are generally in the range of 1 to 10 darcies (LeCain 1997), whereas fault permeabilities are on the order of tens to hundreds of darcies (Ahlers, Shan et al. 1996). In contrast, the matrix permeabilities of those units can be as small as several microdarcies (Subsection 5.3.3.1.1). The nonwelded units at Yucca Mountain, including the PTn and the CHn (vitric), have matrix permeabilities on the order

of several hundred millidarcies, and these units contain much less fracturing, as evidenced by field observations (Subsection 3.7.3) and permeability estimates determined by inversion of pneumatic pressure data collected from boreholes that do not intersect known faults (Ahlers, Shan et al. 1996).

The current estimates for average net infiltration are on the order of 5 to 10 mm/year (Flint, A.L. et al. 1996; Bodvarsson and Bandurraga 1996). Assuming a unit gradient in the liquid-phase hydraulic potential, matrix permeabilities in the welded units can only transmit water at rates of no more than several mm/year. The remainder of the water must therefore be flowing through fractures and major faults. The calcite deposition data imply that considerable fracture flow does in fact occur. Calcite fracture coating data show that most of the deposition is found within the fractures and lithophysal cavities (Paces, Neymark et al. 1996); very little calcite or other mineral deposition has been found within the matrix blocks. Surveys of fractures on the walls of the Exploratory Studies Facility show that approximately 8 percent of the sampled fractures are mineralized, indicating the relative stability of flowpaths over millions of years and many climate cycles (E.M. Kwicklis et al., *A Conceptual Model of Unsaturated Zone Flow and Transport, Yucca Mountain, Nevada*, Milestone Report 3GUM612M, U.S. Geological Survey, in preparation).

Chloride concentrations within the mountain vary considerably, and provide further evidence of extensive fracture flow. Figure 5.3-199 illustrates the major processes affecting chloride chemistry at Yucca Mountain. Chloride concentrations of subsurface water can be used to give an indication time-averaged infiltration rates. As precipitation reaches the ground surface and begins to infiltrate, the chloride concentration in the infiltrating water increases as the processes of evapotranspiration proceeds, removing water but leaving the dissolved salts behind. Eventually the infiltration reaches a depth beyond which evapotranspiration is no longer important. The ratio of the chloride concentration in precipitation relative to that in water below the zone of evapotranspiration represents the fraction of precipitation that escaped evapotranspiration to become net infiltration. Thus, low chloride concentrations found at depth under Yucca Mountain, such as those of perched water, indicate that the infiltration that led to recharge underwent only small amounts of evapotranspiration, possibly as a result of rapid infiltration down fractures. Another relevant process is that, over time, the chloride concentration in pore water will tend to equilibrate with the concentration in the host rock material. The initial chloride concentrations in the water as it arrives at the surface of the mountain are on the order of 0.6 mg/L, due to dissolution of wind-blown chloride in the rainwater (Fabryka-Martin and Liu 1995). Chloride concentrations within the PTn lie in the range of 30 to 80 mg/L. Chloride concentrations in pore water in the TSw rock matrix can be approximately 80 mg/L. However, within perched water bodies, chloride concentrations have been measured to be 5 to 10 mg/L (Yang et al. 1996). These data imply that the perched water bodies are fed by fracture flow from the surface (with low concentrations of chlorides), and continuing within the TCw to depth. The data also suggest that recharge to the perched water bodies occurs through localized channels of flow that transport fluid rapidly through the PTn without allowing sufficient residence time for the water to equilibrate with the high chloride concentration water found in the matrix rock. However, alternative explanations exist that do not require chemical disequilibrium between fractures and matrix in the PTn. These explanations are:

- Perched water was recharged during much wetter conditions that existed near the end of the Pleistocene. The more dilute pore water that existed in the PTn at that time was later

flushed during the more arid climate of the Holocene by recharge water more concentrated in chloride (Sonnenthal and Bodvarsson 1997).

- Perched water was created and sustained by infiltration in sideslope and ridgetops, with subsurface later redistribution of flow. Sideslopes and ridgetops are areas which are predicted to have higher infiltration rates than the wash locations from which most of the borehole chloride data has been obtained (Fabryka-Martin, Flint et al. 1997). Most likely, some combination of these reasons explains the differences between the PTn pore water and perched water chloride concentrations.

It is evident from the isotope age results that the contribution of matrix flow to the perched-water bodies is probably rather small. Corrected ^{14}C ages of the perched-water bodies range from approximately 2,000 to 6,000 years (Yang et al.1998), again suggesting that fracture-dominated flow creates and maintains the perched water bodies. Water-flow velocities through the matrix within Yucca Mountain are much slower, so that water flowing through the matrix rock would require tens of thousands of years or more to reach the perched water bodies. Carbon-14 age-dating of waters within the PTn yields apparent ages of a few thousand years, and similar (or younger) ages are obtained from measurements on samples from deep within the zeolitic CHn in boreholes UZ-14, SD-9, and SD-12 (Subsection 5.3.4.2.4). These findings imply that the water has arrived at those locations through fracture flow.

The average fracture spacings in the welded units such as Tiva Canyon and Topopah Spring are on the order of 0.5 m (Beason 1996; Sweetkind, Barr et al. 1997). Previous research on fracture flow strongly suggests that, typically, only a small fraction of fractures actually transmit groundwater. What has been unknown, however, is whether the fractures provide an effective continuum containing a large number of flow seeps, or if the preponderance of the water is only transmitted through a few fractures. Although there are insufficient data to conclusively eliminate either hypothesis, existing data supports the conclusion that the fractures behave as an equivalent continuum at Yucca Mountain, at least at a sufficiently large scale.

Calculations of the average and standard deviation of the water saturations in matrix samples suggest relatively uniform values for most of the units within the mountain. Specifically, the standard deviations are typically only about 15 saturation points (Flint, L.E.1998). These data have been measured on cores from surface boreholes as well as from short boreholes within the Exploratory Studies Facility. In situ measurements of water potential also show little variability between layers and between boreholes. Most of the data suggest that the water potential at Yucca Mountain is in the range of −3 to −6 bars (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997; Subsection 5.3.4.2.2). The measured saturations and water potentials are consistent with a conceptual model in which a sufficient number of fractures flow and interact with the matrix to create relatively uniform conditions within a given geologic formation.

Calcite coatings have been found in many fractures within the welded units. Calcite-deposition models suggest that the flow of water through fractures must have been sufficiently slow to allow

the calcite to be deposited. Furthermore, as observed in the Exploratory Studies Facility, the calcite-filled fractures are distributed fairly evenly throughout the TSw. Temperature data from boreholes show nearly uniform temperatures within the Topopah Spring unit within the potential repository area. All of the above data suggest that fracture flow is pervasive at Yucca Mountain, rather than occurring along only a few dominant flow pathways.

Although it has been concluded that there are numerous flowing fractures at Yucca Mountain, the fractures seem to have limited interaction with the rock matrix. Studies have shown (Thoma et al. 1992) that fracture coatings can reduce imbibition into the matrix significantly. Ho (1997) evaluated methods of incorporating the conceptual model of fracture-matrix interactions into a dual-permeability model, which is a model that allows fluid to flow through the interconnected fracture system, through an interconnected system of matrix blocks, or between the fractures and the matrix blocks. His results suggested that agreement with observed matrix saturation and water-potential data is improved by using techniques that significantly reduce the magnitude of the fracture-matrix flux. Although Ho attributed this decrease in the strength of the fracture-matrix interaction to a reduced fracture-matrix surface area, it could equally well be attributed to fracture coatings that reduce the effective permeability of the matrix blocks.

In accordance with these observations, a reduction factor has been used in the unsaturated-zone model simulations to reduce the fracture-matrix interconnection area by a factor equal to the liquid saturation of the upstream block (Bandurraga et al. 1997). However, Ho (1997) concluded from his studies that the multiplicative reduction factor should be much smaller, on the order of 10^{-4}. This is supported by the disequilibrium that has been observed between the chloride concentrations in the perched water and TSw matrix pore water (Yang et al. 1996). If the perched water bodies are recharged through vertical flow, then the low chloride concentrations found in the perched water must be due to fracture flow that has had relatively little interaction with the high-chloride matrix water.

The presence of fast pathways is confirmed through measurements of environmental isotopes. Bomb-pulse levels of ^{36}Cl have been found at the potential repository level in the Exploratory Studies Facility (Fabryka-Martin, Wolfsberg et al. 1997). Although systematic sampling within the Exploratory Studies Facility found elevated ^{36}Cl levels restricted to only a few locations associated with specific structural features (faults and fractures), mixing of fracture waters with matrix pore waters may be masking bomb-pulse ^{36}Cl in many places. Furthermore, the predominant sampling population came from the tuff matrix very near fractures, with relatively few samples taken from within fractures themselves. This sampling bias may, therefore, underestimate the potential for fast flow beneath Yucca Mountain. Within the Northern Ghost Dance Fault Alcove, tritium has been found, also associated with a major structural feature, indicating the existence of fast flowpaths. These isotopes suggest that water transport through the mountain occurs within fast pathways at velocities on the order of 10 m/year or more.

5.3.4.4.3.8 Fluid (Water and Gas) Flowpaths and Ages

A number of isotope measurements have been performed on samples from perched water zones, matrix pore water, and rock-matrix material from the unsaturated zone at Yucca Mountain

(Subsection 5.3.4.2.4). Some of these measurements have been discussed above to support and develop the conceptual model. In addition, fracture coatings have been tested for isotope concentrations (Paces, Neymark et al. 1996; Paces, Marshall et al. 1997). These isotope measurements provide age estimates of the water or, in the case of the fracture coatings, the water that precipitated them; however, there are many factors that cause uncertainties in these estimates. For example, variations in historic atmospheric isotopic concentrations, anomalous isotopic concentrations associated with nuclear weapons testing at the Nevada Test Site and elsewhere, and processes such as deposition, dilution, and molecular diffusion each lead to uncertainties in the use of isotope data to constrain the ages of pore water. Additionally, the potential for sample contamination during drilling and sampling needs to be considered.

Recent work on fracture coatings from the Exploratory Studies Facility have yielded age estimates of about 30,000 years (Paces, Neymark et al. 1996). However, the rates and distribution of mineral deposition over time are uncertain. There is some evidence to suggest that deposition may occur in tiny "islands" on the coating surface. Therefore, the sampling procedure may not be able to isolate the youngest material, and may be biasing the measurements on the side of older ages.

Anomalous ^{36}Cl and tritium data show that fast flowpaths exist at Yucca Mountain. An evaluation of these data suggest that in most cases major structures are needed to transport fluid and solute rapidly through the PTn unit, and that this is most efficiently done by intermittent water pulses. The hypothesis developed as a result of the data is that infiltration pulses carry the "bomb-pulse" signal down to the Exploratory Studies Facility below washes that have little alluvium, and which overlie fault features (Fairley and Sonnenthal 1996; Fairley and Wu 1997). Exploratory Studies Facility sampling and ^{36}Cl measurements imply that fast pathways are not ubiquitous; elsewhere, however, elevated ^{36}Cl values may be disguised through dilution or mixing with in situ chloride, thereby affecting the observed ^{36}Cl /Cl ratio (Sonnenthal and Bodvarsson 1997).

The anomalously high "bomb-pulse" ^{36}Cl levels at the Exploratory Studies Facility are associated with relatively few fast pathways, but with volumes of water that may or may not be relatively small. Consequently, they are of limited use in determining the global percolation flux. Analysis of the ^{36}Cl data in terms of occurrence and spreading can, at best, constrain the estimate of the pulse strength that is required to transport the isotope to the potential repository horizon. The percolation pulses associated with bomb-pulse isotope measurements may occur infrequently, in which case they may not affect the average behavior of the mountain, or its "quasi-steady-state" flow patterns. Evaluation of isotopic data is complicated by the fact that the background ^{36}Cl /Cl ratio can be attributed to high historic ^{36}Cl, as well as diluted bomb-pulse water. The interpretation of the ratio is also very sensitive to the amount of chloride that is assumed to be available in the rock before the arrival of the bomb pulse.

5.3.4.4.3.9 Geothermal Gradient

The geothermal gradient at Yucca Mountain is relatively well established and known through numerous temperature and heat-flow measurements (Subsection 5.3.4.2.3). The relevant data include borehole temperature profiles (Sass, Lachenbruch et al. 1988; J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain,*

Nevada, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997), and recent Exploratory Studies Facility temperature measurements. Deviations of the observed temperature profile from that predicted to occur solely from heat conduction have been used to infer gas and/or moisture flow patterns. J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164) incorporated an analysis of temperature data in their report that links variations in the geothermal gradient at several boreholes to infiltration rates. Bodvarsson and Bandurraga (1996) extended this analysis to include additional boreholes. The analyses used two models: a conduction-only model and a coupled conduction/convection model. The importance of temperature data in unsaturated-zone model calibration and percolation flux estimation is emphasized in the appropriate subsections below. Previous calibration efforts (Ahlers, Bandurraga et al. 1995a; Bandurraga et al. 1997) have determined the thermal conductivity values that are required in order to match the measured temperature profiles.

5.3.4.4.4 Summary of Unsaturated-Zone Modeling Studies and Results

Figure 5.3-200 shows some of the unsaturated-zone model components, in terms of the important input and output data, various calibration activities, and numerical representation of the conceptual model. The following subsections briefly describe the key components in the model.

The modeling studies performed thus far use the extensive data that have been collected during the investigations of Yucca Mountain. The available data include liquid saturations, in situ and core-sample water potentials, saturated conductivities and desaturation curves, rock-property measurements made on core samples, pneumatic monitoring, temperature data, air-permeability test results, geochemical analyses, and perched water body testing. The Exploratory Studies Facility data include fracture mapping data, environmental isotopes and hydrochemical data, fracture coating data, and rock properties from in situ and core-sample measurements. The incorporation of the data into modeling studies has provided a comprehensive and complex unsaturate-zone model that represents the important unsaturated-zone flow processes such as moisture flow, capillary pressure effects, gas flow, convective and conductive heat transfer, evaporation and condensation, moisture and gas flow travel times, and transport of conservative and reactive species in the mountain. The model grid is based on the best available geologic data, and captures the complex geologic features characterized by the nearly 60 boreholes that penetrate a significant portion of the mountain, in addition to Exploratory Studies Facility, pavement, trench, and section studies. Figure 5.3-201 shows a plan view of the current unsaturated-zone model grid. The model has been evaluated by comparing the predictions of the model to observations of liquid saturations, water potentials, temperatures, and pneumatic pressures in newly drilled boreholes, as well as gas-flow changes caused by the construction of the Exploratory Studies Facility. The validation process and extensive data set have helped to build confidence in the ability of the model to simulate ambient conditions, as well as to perform predictive studies.

5.3.4.4.4.1 Input Data

5.3.4.4.4.1.1 Geological Framework and Rock Properties

With the development of the Geological Framework Model (CRWMS M&O 1997b; Subsection 3.8), a standardized model has become available for the geologic layers incorporated into the TCw, PTn, and TSw hydrogeologic units. However, the vitric and zeolitic parts of the Calico Hills and the Prow Pass formations (the CHn unit), so crucial for flow and transport, are not represented in the Geological Framework Model. Simulation studies with numerical grids developed at Lawrence Berkeley National Laboratory using the Geologic Framework Model and a geologic model developed by Lawrence Berkeley National Laboratory, give nearly identical results for most of the important processes in the unsaturated zone above the CHn. However, as the Lawrence Berkeley National Laboratory geological model incorporates the hydrologically important vitric/zeolitic zones below the potential repository horizon, this model is assumed to be more appropriate for unsaturated-zone flow modeling.

The unsaturated-zone model uses matrix property data developed by L.E. Flint (1998) that characterize the important hydrogeologic layers at Yucca Mountain. The data show low matrix permeabilities (on the order of microdarcies) in the welded units (TCw and TSw), and much higher permeabilities (on the order of hundreds of millidarcies) in the units that are mainly nonwelded (PTn and CHn). A detailed fracture property model has been developed by Lawrence Berkeley National Laboratory (Sonnenthal, Ahlers et al. 1997), using fracture-frequency data from boreholes, as well as detailed fracture data from the Exploratory Studies Facility. Fracture permeability data are obtained from air permeability testing in boreholes, the in situ pneumatic monitoring program, and smaller scale air permeability testing in the heater test area. The fracture property model yields results for various fracture parameters (including anisotropic permeabilities within each of the geological units), fracture porosities, effective fracture spacings, and derived van Genuchten parameters for the capillary pressure and relative permeability functions of fractures in each hydrogeologic layer in the unsaturated-zone model. Fault properties, derived primarily from pneumatic data, show that faults have gas permeabilities on the order of hundreds of darcies. Lateral variations in permeability within fault zones are also significant, as shown by fault testing in the Ghost Dance fault alcoves of the Exploratory Studies Facility. Additional data on fault properties and processes are needed, some of which can be derived from the proposed testing in the Solitario Canyon fault.

5.3.4.4.4.1.2 Infiltration Rates

Most of the recent modeling studies conducted at Lawrence Berkeley National Laboratory make use of the spatially variable steady-state infiltration map by A.L. Flint et al. (1996) (Subsection 5.3.4.1), which estimates an average infiltration rate of 4.9 mm/year over the unsaturated-zone model domain, and 6.7 mm/year over the potential repository area (Figure 5.3-196). The Flint et al. model indicates that most of the infiltration occurs in sideslopes and ridgetops, where there is minimal alluvial cover. The expert elicitation panel evaluating the unsaturated-zone flow model and unsaturated-zone processes in December 1996 had reservations concerning some of the conclusions reached by A.L. Flint et al. (1996); for example, most of the experts felt that more infiltration should occur in the washes, rather than on ridgetops and sideslopes, because surface runoff during and after storm events tends to channelize in washes, creating ponded conditions that can potentially lead to high-infiltration

rates. At the experts' request, Lawrence Berkeley National Laboratory developed a 3-D surface hydrological module for the computer code TOUGH2 that accounts for precipitation, surface runoff, evapotranspiration, root uptake, lateral flow at alluvium-bedrock contacts, and 3-D flow within the washes (Ahlers, Wu et al. 1997). Demonstration of the use of this TOUGH2 module for a simplified 2-D cross-section through Wren Wash showed that the possibility exists for lateral flow at the alluvium-bedrock contact (Figure 5.3-202), and that multi-dimensional processes may be important in the characterization of the spatial variability of the infiltration. Studies of the additional infiltration into the washes, compared with that which occurs on the side slopes, showed little effect on the percolation flux at the potential repository horizon. Future development of the 3-D surface hydrology module may provide better estimates of the spatial and temporal variability of infiltration.

5.3.4.4.4.1.3 Pneumatic Data

Pneumatic (gas-phase) pressure measurements were taken from 11 pneumatic monitoring boreholes in the vicinity of the potential repository (Subsection 5.3.4.2.1). Ten of the boreholes (NRG#4, NRG#5, NRG-6, NRG-7a, UZ#4, UZ#5, UZ-7a, SD-7, SD-9, and SD-12) were located at various points near the North Ramp and Main Drift of the Exploratory Studies Facility. Borehole ONC#1 was located slightly west of the Bow Ridge fault, about 1 km north of the South Portal of the Exploratory Studies Facility. Data were collected continuously, over various time periods (for the different boreholes) from 1994 to 1997. In eight of the boreholes, the data were collected from in situ pressure transducers; grout or packers were used to seal off the borehole between the transducers. Data in the other three boreholes were collected with tubes that led to pressure transducers at the surface.

Pneumatic data can be used to estimate the gas-phase permeability of the various rock layers. The barometric pressure at the surface varies in time due to several effects, each of which has its own characteristic frequency. Heating and cooling of the atmosphere on a diurnal (daily) basis leads to pressure variations with a period of one day. Tidal effects cause semi-diurnal oscillations of the barometric pressure, with a period of one-half day. The passage of weather systems across the earth's surface produces pressure variations that have periods of a few days or weeks. Finally, there is also a yearly cycle of changes in the barometric pressure caused by seasonal heating and cooling of the atmosphere. The result is a complex pressure signal at the surface of the mountain that consists of a superposition of these various periodic pressure signals. As this pressure signal moves down into the mountain, it undergoes an attenuation in magnitude; also, each periodic component undergoes a phase-shift, which means that the pressure variations at depth will, in general, lag behind those at the surface. The amount of attenuation depends on the pneumatic diffusivity of the rock mass, which is a parameter that is proportional to the effective gas-phase permeability, and inversely proportional to the gas-filled porosity. Pneumatic data therefore lend themselves to being used to estimate these latter parameters.

An example of the pneumatic pressure signal measured over a 24-day period in 1995, in borehole NRG-7a, is shown in Figure 5.3-203 for transducers at three different depths. A more comprehensive presentation of the available data is contained in Subsection 5.3.4.2.1. As shown in Figure 5.3-203, the mean pressure increases with depth, but the magnitude of the pressure perturbations decreases with depth, due to the aforementioned attenuation. The figure also illustrates that the high-frequency components of the pressure variation are more strongly attenuated than the lower frequency

components. A mathematical analysis of this phenomenon, showing the relationship between pneumatic diffusivity, attenuation, and phase shift, is given by Ahlers, Bandurraga et al. (1997).

5.3.4.4.4.2 Calibration of the Model

Considerable effort has been devoted to calibration of the unsaturated-zone hydrological model. "Calibration" is the process in which various parameters in the model, such as permeabilities, are modified so as to achieve an optimum agreement between the predictions of the model and parameters that have been measured in the field. The purpose of the calibration process is to ensure that the model is able to faithfully represent all thermohydrological processes that are known to occur at the mountain, thereby giving some confidence that it will also be capable of predicting future behavior of the system during the lifetime of the potential repository.

One important aspect of the calibration process has been the development of a set of layer-averaged parameter values for the properties of the fractures and matrix blocks (Bandurraga and Bodvarsson 1997). The data sets that were used in this calibration process include:

- Core sample saturation and water potential data (Flint, L.E. 1998)
- Geophysical saturation data (CRWMS M&O 1996a)
- Borehole log data (Moyer et al. 1995)
- Sample porosity and rock grain density data (Flint, L.E. 1998)
- In situ water potential and temperature measurements (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997)
- Layer-averaged fracture permeabilities from in situ tests (LeCain 1997)
- Spatially-varying infiltration rates (Flint, A.L. et al. 1996)
- In situ pneumatic data (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997)

Core sample porosities, saturations, and densities were supplied by the USGS from the following boreholes that penetrated the TSw: SD-7, SD-9, SD-12, NRG-6, NRG-7a, UZ-14, and UZ#16; core sample water potential data were also available for SD-7, SD-9, and UZ-14. Water potential data were collected by J.P. Rousseau et al. (*Hydrogeology of the Unsaturated Zone, North Ramp Area of the*

Exploratory Studies Facility, Yucca Mountain, Nevada, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) using psychrometer sensors in boreholes SD-12, NRG-6, and NRG-7a; these data are thought to have an uncertainty of about one bar, based on calibration using vapor in equilibrium with salt solutions having known water potentials. The error in the water potential of the core samples is more difficult to determine, due to the evaporation of water from low-porosity rock during the handling of the cores (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Matrix permeability and van Genuchten parameters have been estimated by L.E. Flint (1998) for many of the rock layers at Yucca Mountain; however, there are large uncertainties associated with these data, due to the difficulties in performing permeability and water-retention tests on such relatively impermeable rocks. [The permeability of rock is a measure of the rock's ability to transmit fluid under a pressure gradient or gravitational gradient. This parameter is measured in units of square meters, or, more commonly, in units of darcies, which are roughly equivalent to 10^{-12} m^2. The van Genuchten parameters (van Genuchten 1980) are used to quantify the water retention curve, which relates the liquid saturation to the capillary pressure, and the relative permeability curve, which quantifies the decrease in permeability caused by partial liquid saturation.]

The calibration approach used by Bandurraga and Bodvarsson (1997) can be summarized as follows. A number of 1-D submodels are first extracted from the full 3-D model, corresponding to the locations of the boreholes used in the moisture flow calibration. The submodels are first used in conjunction with measured saturation and water potential data, to estimate the hydrologic properties of the matrix rock in each layer. The hydraulic properties of the fractures are estimated by adjusting them to fit the measured borehole gas pressure records. Finally, layer-averaged values are developed from the set of properties chosen from the individual inversions. Criteria that are used in arriving at the base-case fracture and matrix properties include:

- Are the saturation and water potential data matched by the model, with these choices of property values?
- Are the property values consistent with the range of values obtained from core measurements?
- Are the hydrological properties used in each layer consistent throughout the entire rock unit?
- Do the property values lead to travel times through a given layer that are consistent with geochemical data such as chloride concentrations in perched water bodies and the locations of anomalously high "bomb-pulse" ^{36}Cl occurrences?

The property values arrived at from the calibration effort did not necessarily agree with the measured values, although the measurements were used as constraints on the inversion process. The results of the inversions depend on factors such as the assumed surface infiltration rates, and the manner in which fracture/matrix flow is treated in the model. The base-case data set was developed using a dual-permeability model, which allows flow to occur either through the matrix or through the fractures, and also allows fracture/matrix flow (Gerke and van Genuchten 1993). The surface area available for

fracture-to-matrix flow was taken to be the nominal fracture/matrix interface area, as estimated by fracture spacing data, multiplied by the liquid saturation in the fracture, to account for the fact that only a fraction of the fracture plane is actually occupied by water. An example (for borehole SD-7) of the agreement that was achieved between simulated and measured liquid saturations and water potentials of the core samples, using the base-case properties, is shown in Figure 5.3-204; the agreement with borehole gas pressures is shown in Figure 5.3-205. The resulting parameter data sets, which are described in more detail in Chapter 6 of Bodvarsson et al. (1997), have been made available to all modeling groups within the YMP through an anonymous FTP site maintained at Lawrence Berkeley National Laboratory. Tables 5.3-62 through 5.3-65 summarize model parameter values for the base case, the infiltration-times-five case, the infiltration-divided-by-five case, and fault calibrated parameter sets, respectively. (The definitions of unsaturated-zone model layers were described previously in Table 5.3-61.) In each of the four cases, matrix properties are listed in part (a) of the tables and fracture properties are listed in part (b).

Borehole temperature measurements (Subsection 5.3.4.2.3) provide another set of data against which the unsaturated-zone model can, and should, be calibrated. Calibration of the model against temperature data is necessary because of the fact that many of the processes that occur within the mountain are temperature-dependent. These temperature-dependent processes include fluid, gas, and heat flow, as well as the kinetics of rock/water chemical interactions. Furthermore, calibration of the model against known temperature measurements provides the initial conditions needed for repository thermal-loading studies. Hence, it is important that the model be able to accurately represent known temperature profiles and thermal conditions at the mountain.

The model has been calibrated against temperature profiles (Sass, Lachenbruch et al. 1988; J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164; Rousseau et al. 1997) for 25 boreholes at Yucca Mountain. The basic idea behind the calibration process is that some portion of the total vertical heat flux through the mountain is transported by conduction, and the remainder is transported advectively by the flowing fluids (Bredehoeft and Papadopulos 1965). In the absence of fluid flow effects, the steady-state temperature profiles would be linear within each rock layer. Hence, any local curvatures in the temperature profiles can be correlated with the rate of fluid flow through the mountain. The calibration exercise was performed using the TOUGH2 code, using 1-D submodels taken from the full 3-D grid, and also using a 1-D multi-layer analytical model (Bodvarsson et al. 1997) that was an extension of the single-layer model of Bredehoeft and Papadopulos (1965). Good matches were obtained between the analytical and numerical models, and the measured data, for all boreholes. An example of the agreement between observed and modeled temperature profiles for borehole SD-12 is shown in Figure 5.3-206. In each case, the appropriate percolation flux needed to achieve a match was in the range of 0 to 15 mm/year. The temperature profiles are quite sensitive to the assumed percolation flux, which implies that only a relatively narrow range of percolation fluxes will lead to profiles that are at all consistent with the measured data. On the other hand, however, some uncertainty remains in these flux estimations, due to uncertainties and measurement errors in the temperature profiles and thermal conductivities, as well as small, unquantifiable effects due to gas flow.

The model has also been calibrated against measured concentrations of various environmental isotopes, such as ^{36}Cl, ^{87}Sr, and ^{14}C (Subsection 5.3.4.2.4). Although such data are limited in number, they yield important information about fluid flow patterns, evaporation and condensation processes, rock/water interaction, percolation flux, and groundwater ages. These data are also helpful in identifying fast paths, and in verifying conceptual models of unsaturated-zone flow.

Perched water data have proven to be extremely useful in the model calibration process, as the locations and extent of perched water bodies are related to percolation flux and the permeability of the underlying perching layers (i.e., the basal vitrophyre of the TSw, or the zeolitic CHn). The model has been calibrated against the assumed areal extent of perched water bodies, estimated volumes and ages of perched water, hydrological (pumping) test results, and geochemical signatures. With these calibration activities, the 3-D unsaturated-zone model is a complex and comprehensive model that contains spatially distributed parameter values that accurately reflect the various processes and observations.

5.3.4.4.4.3 Output Data

The remainder of Subsection 5.3.4.4 on 3-D site-scale unsaturated-zone modeling focuses on the recent simulation results contained in the Bodvarsson et al. (1997) report. Among the issues addressed are: flow above and below the potential repository horizon, percolation flux at the potential repository horizon, fracture flow, groundwater seepage into drifts, and the impact of future climate change scenarios on the unsaturated flow system.

5.3.4.4.4.3.1 Flow Above the Potential Repository

Water flow through the potential repository area at Yucca Mountain is affected by near-surface processes within the TCw, the PTn, and that portion of the TSw that lies above the potential repository. The net infiltration into the bedrock is episodic, and significant pulses probably occur only once every few years (Subsection 5.3.4.1). The pulses move rapidly through the TCw, as indicated by numerous measurements of anomalously high levels of the environmental isotopes ^{36}Cl and ^{3}H within the TCw and upper PTn (Subsection 5.3.4.2.4). The results from the surface hydrology model suggest that lateral flow at the alluvium-bedrock interface produces non-uniform infiltration, with focused flow close to the edges of washes.

Water flow within the TCw is also significantly affected by gas flow, because the TCw is very permeable to gas. The gas flow causes evaporation of water, and it is estimated that water can be removed from the TCw through evaporative processes at a rate of about 0.5 mm/year. Borehole temperature data from the TCw show relatively uniform temperatures, indicating that most of the heat transfer is by evaporation.

Upon reaching the PTn, the infiltration pulses are attenuated by the high storage capacity of the generally nonwelded matrix, and also through lateral flow. As a result, flow through the lower portion of the PTn is fairly uniform spatially, and relatively slow in comparison to the pulses through the TCw. This assertion is supported by the fact that anomalous levels of ^{36}Cl are primarily found near the top of the PTn. As fracture frequencies are much smaller in the PTn than in the TCw, it is expected that

PTn fractures carry no more than a small fraction of the total water flux. Most of the "fast-path" flow through the PTn unit is associated with structural features, such as faults or fault-associated fractures, that cross the various geologic formations comprising the PTn.

Because of attenuation of the infiltration pulses within the PTn unit, it is expected that percolation reaching the TSw is fairly uniform, both spatially and temporally. However, the matrix permeability of the topmost vitric caprock of the TSw is relatively low, thus creating lateral flow and flow concentration, which reinitiates fracture flow in the TSw. The TSw fracture permeability is fairly uniform (ranging from 1 to 5 darcies), as indicated by the pneumatic monitoring data analyses. More variability (ranging from a few millidarcies to several darcies) is obtained from air permeability testing of surface boreholes, and in Exploratory Studies Facility alcoves at smaller scales. The fracture permeabilities are anisotropic, with vertical permeability larger than horizontal permeability by a factor of 5 (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Matrix permeabilities are typically smaller than fracture permeabilities by approximately 5 to 6 orders of magnitude, and modeling results show that a majority of the flow in the TSw occurs through the fractures. Fracture-flow travel times from the PTn to the potential repository are small, on the order of a few to tens of years.

5.3.4.4.4.3.2 Percolation Flux at the Potential Repository Horizon

Bodvarsson and Bandurraga (1996) concluded that percolation flux in the potential repository horizon was on the order of 1 to 10 mm/year, which is consistent with infiltration estimates by A.L. Flint et al. (1996). Those results were based on using the spatially variable infiltration rates at the ground surface as a boundary condition, and calculating the flux magnitudes entering the potential repository horizon from the simulation results. The conclusion also was based on preliminary results from borehole temperature analyses. In their fiscal year 1997 milestone report, Bodvarsson et al. (1997) used a variety of methods to constrain the percolation flux from temperature and geochemical data analyses. Perhaps the most potentially useful data set used in the evaluation of percolation flux is the total chloride data. The study using these data indicates that the assumed mean infiltration rate (~4.9 mm/year) is probably a maximum value for the potential repository at Yucca Mountain (Sonnenthal and Bodvarsson 1997; see Figure 5.3-207). Depending on the total chloride flux at the surface, the mean infiltration may be as low as 1 mm/year. Strontium and ^{14}C data analyses of selected boreholes also are consistent with a percolation flux in the range of 1 to 10 mm/year (Sonnenthal, dePaolo et al. 1997; Moridis et al. 1997).

Analyses of the borehole temperature data provide percolation rate estimates. Overall, the results of temperature data analyses are consistent with the average rate from the A.L. Flint et al. (1996) map, and suggest average percolation rates of 5 to 10 mm/year in the potential repository horizon. The range of borehole percolation flux estimates is from zero to over 15 mm/year (Bodvarsson et al. 1997). Boreholes on Yucca Crest show fluxes on the order of 5 mm/year, with slightly higher values for some downslope boreholes away from the crest, such as Boreholes SD-12, UZ-1, H-1, and G-2. Many of the high percolation flux estimates are obtained from boreholes that are located near faults. An alternate explanation of the thermal gradients observed in the apparently high-flux boreholes is that significantly greater gas flow occurs in the fault zones. This gas flow causes evaporation and

condensation deep within the mountain, affecting the observed heat flux in the same manner as does a high percolation flux. This alternative explanation is consistent with the gas age data collected from some boreholes located near faults, such as NRG-7a. However, at other locations where gas from fault zones has been sampled, such as Exploratory Studies Facility Northern Ghost Dance Fault Alcove which intersects the Northern Ghost Dance fault, apparent ^{14}C ages of the gas are not younger than the ages of gases sampled from similar depths at boreholes not located near known faults.

In order to bound possible values of percolation flux, various modeling exercises using the unsaturated-zone model have been conducted. Model analyses with high percolation flux values indicate that both the temperature and chloride data are inconsistent with percolation flux values exceeding 20 mm/year. It would be difficult to obtain a calibrated unsaturated-zone model with an average percolation flux greater than 20 mm/year that is still consistent with measured matrix properties and with our current understanding of flow processes within the mountain. The lower limit could be close to zero, but is more likely to be several tenths mm/year. This estimate accounts for the relatively high saturations observed in some of the welded hydrogeologic units, and the departure of the measured water potential profiles from the much dryer static equilibrium profile that would exist in the absence of flow (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164).

As also noted in Subsection 5.3.4.3, percolation flux estimates have large uncertainties associated with them, as estimates of the flux are greatly affected by assumed infiltration rates, geological and conceptual models, distributed properties of fractures and matrix blocks, and the schemes used for treatment of fracture-matrix flow (Wu et al. 1997). More well-constrained estimates may be possible when the planned percolation flux tests and the drift seepage (niche) tests are completed in the Exploratory Studies Facility.

5.3.4.4.4.3.3 Fracture Flow

Fracture flow is believed to account for the majority of the mass flux in the TSw (approximately 80 percent), and thus in the potential repository horizon at Yucca Mountain. The low permeabilities of the rock matrix will only permit, at most, a few mm/year of flow, with the remainder occurring through the fractures. The unsaturated-zone model output shown in Figures 5.3-208 and 5.3-209 (with flux contours in mm/year) illustrate the importance of the relative partition of the flow into fracture and matrix components at the potential repository horizon. As discussed above under the conceptual model subsection, the data indicate that flux in the TSw is carried by numerous fractures, rather than by a few major flowpaths. It is likely that most of the perched water bodies are fed by vertical flow through fractures, because the ages of the perched water bodies, which are on the order of 2,000 to 6,000 years (Yang et al. 1998) are too young to be accounted for through matrix flow. The chloride contents of the perched water bodies are low, and in apparent disequilibrium with relatively high chloride concentrations measured in TSw pore water samples. These observations may indicate that the fracture flow does not interact greatly with the matrix. Therefore, when implementing the dual-permeability formulation of the unsaturated-zone model, the surface area available for fracture-matrix flow was reduced for the purposes of our most recent numerical studies carried out at Lawrence Berkeley National Laboratory (Bodvarsson et al. 1997).

Only a small fraction of the fracture flow is believed to be related to the fast-path flow responsible for the observations of anomalously high levels of ^{36}Cl in the Exploratory Studies Facility. This conclusion is based on the ratio of the number of samples having anomalous ^{36}Cl levels (obtained during feature-specific sampling) to the number that were found to have background (or only slightly elevated) ^{36}Cl levels, which were obtained through systematic and feature-specific sampling in the Exploratory Studies Facility (Subsection 5.3.4.2.4). The collection and analysis of anomalously high ^{36}Cl and tritium data are useful only in identifying localized fast-path locations, and in providing an upper bound on fracture-flow velocities in the calibration of the unsaturated-zone model. Data on geochemical components such as chloride, strontium, and ^{14}C yield additional information for the understanding of large-scale fluxes and flowpaths at Yucca Mountain.

5.3.4.4.4.3.4 Seepage into Drifts

The unsaturated-zone model provides data for both the ambient drift-scale model and the thermohydrological drift-scale model. It also provides the hydrological parameter values and the necessary boundary conditions for moisture, gas, and heat flow. To obtain boundary conditions for the drift-scale models, a detailed 2-D cross-sectional model was developed that explicitly accounts for each drift (Figure 5.3-210; also see Haukwa et al. 1997). The detailed (fine-grid) model simulations and the site-scale unsaturated-zone model simulations provide consistent estimates of the spatial variability in percolation fluxes at the potential repository horizon (Figure 5.3-211). The results also show that the dual-permeability and effective continuum method formulations of the detailed model give nearly identical results for steady-state flow. Furthermore, a case was investigated in which the total mass flux for the cross-section, based on the map of A.L. Flint et al. (1996), was concentrated at infiltration points that were spaced every 100 m. The results demonstrated the effectiveness of the PTn in dampening temporal variations in the strength of the infiltration pulses. The simulated percolation flux distributions above the drifts for "time-averaged" and "discrete" infiltration scenarios, showed excellent agreement. The detailed model showed that the variability in percolation flux above each of the drifts ranged from 0 to 20 mm/year. Figure 5.3-212 gives an example of the percolation flux distribution above the drifts, as computed using the dual-permeability model and a time-averaged (but spatially-variable) infiltration at the surface. The amount of seepage into the drifts, as estimated on the basis of the assumed percolation flux, should be resolved by the current niche studies being performed in the Exploratory Studies Facility. Lastly, preliminary modeling of construction water migration through fractures beneath the Exploratory Studies Facility shows that the water could have migrated significant distances since construction (James et al. 1997).

5.3.4.4.4.3.5 Flow Below the Potential Repository

The flow patterns below the potential repository are important for two reasons. First, understanding advective and dispersive processes in the unsaturated zone is necessary, and second, some of the zeolitic rocks below the potential repository provide retardation for radionuclides. Unfortunately, limited data are available for the geology or the hydrology below the potential repository, because only a small number of boreholes penetrate the CHn. However, data such as matrix permeability, fracture permeability, geochemical observations, and percolation flux estimates can be used in the modeling to provide some indication of likely flowpaths below the potential repository. Strontium concentrations in pore water samples and from perched water bodies are strongly affected by flow

through zeolitic layers, because the zeolites remove strontium through ion exchange (Vaniman and Chipera 1996). Therefore, strontium data, as well as data for other divalent cations such as calcium and magnesium, may be important for the evaluation of flowpaths to the water table.

The TSw basal vitrophyre is the perching layer for the perched water bodies found in the northern part of the potential repository near boreholes SD-9 and UZ-14. This layer has a very low matrix permeability and is reported to contain fractures that are often blocked with zeolites and clays (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Simulations using the 2-D site-scale model are able to reproduce these perched-water conditions. Figure 5.3-213 shows a saturation plot, along with composite-velocity streamlines (averaged over the fractures and matrix) along a north-south cross section through boreholes UZ-14 and G-2. Perched water is seen around UZ-14, and the time-steps (indicated in the figure by white squares) imply that 7,500 years are required for a particle of water to travel from the land surface to the approximate location of the perched water. This is consistent with current estimates using ^{14}C data collected from the perched water in UZ-14, which range from 4,170 to 6,260 years (Yang et al. 1998), or 5,150 to 11,000 years (J.P. Rousseau et al., Eds., *Hydrogeology of the Unsaturated Zone, North Ramp Area of the Exploratory Studies Facility, Yucca Mountain, Nevada*, Milestone Report 3GUP667M, U.S. Geological Survey, in press, MOL.19980220.0164). Particle travel times through an east-west cross section passing through borehole UZ-14 indicate a perched water age of about 6,000 years. The particle travel time results (7,500 or 6,000 years) along the two cross-sections are not inconsistent, because the particles did not travel through precisely the same paths. Travel times estimated using this particle tracking analysis are used only for approximate comparisons with residence times, because residence times derived from isotopic data reflect mixing of transient and steady-state pulses of water that percolated through fractures at different times, as well as a small amount of matrix flow.

The CHn underlying the TSw has both vitric and zeolitic portions, with differing hydraulic properties. The zeolitic portion of the CHn is reported to perch water at borehole SD-7 and possibly at borehole NRG-7a near the Exploratory Studies Facility. The average permeability of the vitric zones in the CHn is on the order of hundreds of millidarcies, whereas the average permeability of the zeolitic layers is only a few microdarcies. Flow below the potential repository that encounters a generally eastward-dipping perching layer will be laterally diverted. The diversion is expected to continue until the water table is reached, or until a fault or extensive fracture system that can reinitiate vertical flow is encountered (Ritcey et al. 1997). Figure 5.3-214 illustrates schematically the hypothesized flowpaths below the potential repository horizon at the approximate latitude of borehole UZ-14, whereas Figure 5.3-215 shows the simulated flowpaths below the potential repository, as predicted by the model.

For zeolitic layers that do not cause significant perching or diversion, percolation fluxes higher than several mm/year must be a result of fractures, because the measured matrix permeabilities are relatively small. The relatively fast fracture flow component would have a reduced contact time with the sorptive zeolites. In the southern portion of the potential repository, a significant portion of the Calico Hills Formation is vitric with a relatively high permeability. Flow below the potential repository in this area can travel vertically, until the lower altered portion of the Prow Pass Formation

is encountered near the water table. Because of these flow processes, the amount of water percolating from the potential repository to the water table that passes through the sorptive zeolites as matrix flow in the Calico Hills Formation is not known with certainty, but it is likely that this amount is a small part of the total flow (see Subsection 5.3.4.3). Figure 5.3-216 shows the combined fracture and matrix percolation flux at the water table. If the sorption mechanism is an important part of the radionuclide retardation conceptual model, the flow below the potential repository must be characterized more accurately than can be done with available data.

5.3.4.4.4.3.6 Impact of Future Climate-Change Scenarios

Because of the long half-lives associated with many high-level nuclear waste by-products, it is important to examine long-term changes in climate that could affect potential repository performance at Yucca Mountain. For example, the half-life of neptunium-237 (^{237}Np), is about 2.14 million years. Climatic changes could result from increased CO_2 production due to anthropogenic activities, from variations in the earth's distance from the sun, and from local effects such as volcanic eruptions. The climate that exists in the Southern Great Basin today is warmer and dryer than the climate that prevailed there over approximately 80 percent of the past million years (Forester et al. 1996). Therefore, climate change will most likely result in increased infiltration, and could affect the movement of water, gas, and heat in the unsaturated zone.

The 3-D site-scale unsaturated-zone model has been used to examine the effect of increased infiltration on the unsaturated-flow system due to increased CO_2 production, and due to the onset of pluvial conditions (Ritcey et al. 1997). These numerically modeled wetter climate scenarios were based on estimates of precipitation obtained from the National Center for Atmospheric Research, located in Boulder, Colorado. For the increased CO_2 "greenhouse" scenario, infiltration rates at the modeled land surface were increased by a factor of two relative to the most recent estimate of 4.9 mm/year (Flint, A.L. et al. 1996). For the pluvial scenario, corresponding approximately to the 21,000-year glacial maximum infiltration rate, an infiltration rate is used that is approximately 5 times the modern rate.

Figures 5.3-217 through 5.3-219 show the saturation contours along 2-D east-west cross-sections through borehole UZ-14 that were computed using the 3-D unsaturated-zone model, along with the effective continuum method formulation to account for fracture-matrix interactions. These figures also show the movement of individual tracer particles along streamlines moving through the velocity field. The velocity field uses a composite velocity (combining fracture and matrix components), based on the local equilibrium assumption inherent in the effective continuum method.

Results of the present-day baseline scenario (with an average infiltration rate of 4.9 mm/year) are shown in Figure 5.3-217. Approximately 7,000 years were required for a particle to travel from the ground surface to the location of perched water in this simulation. This estimate agrees with the perched water ages obtained through reaction-based corrections to the ^{14}C data from the perched- water (3,000 to 7,000 years;Yang et al.1998), but is also not significantly different from the uncorrected age of approximately 11,000 years determined from the ^{14}C data of the pumped perched water samples from UZ-14 (Yang et al. 1996). Again, as stated previously, travel times estimated using this particle tracking analysis are used only for approximate comparisons with residence times, because residence times derived from isotopic data reflect mixing of transient and steady-state pulses of water that

percolated through fractures at different times, as well as a small amount of matrix flow. Travel-time estimates by this particle-tracking approach, however, follow the movement of individual particles along distinct flowpaths.

Figures 5.3-218 and 5.3-219 show the results of the future climate simulations along the same east-west cross section through borehole UZ-14. Estimates of travel times for a single particle from the land surface to the perched water body at UZ-14 decreased to approximately 3,000 years in the greenhouse scenario, and to less than 1,500 years in the pluvial scenario. In all cases, water particles take the most time traveling through the PTn hydrogeologic unit. Simulation results for wetter future climates also show a decrease in the lateral diversion of water along the PTn (represented without fractures), and an increase in the lateral diversion along the CHn zeolitic interface. The decrease in lateral diversion along the PTn occurs because the relative permeability increases as the saturation increases, decreasing the capillary barrier effect, and thereby reducing the capability of this layer to divert moisture. The CHn zeolitic horizon, however, has a very low permeability, and as a result, increased saturation forces the water to be laterally diverted.

Increased saturation, resulting from higher infiltration rates, creates larger percolation fluxes at the potential repository horizon, but appears to have little impact on the spatial distribution of percolation, according to unsaturated-zone modeling efforts (Ritcey et al. 1997). The most significant changes occur in the northwest portion of the model domain, and between the Solitario Canyon and Ghost Dance faults. Fracture flow is expected to be the dominant flow mechanism affecting the distribution of moisture at the potential repository horizon, because of the low matrix permeabilities within the TSw hydrogeologic unit. Figures 5.3-220 and 5.3-221 show percolation flux in mm/year at the potential repository horizon, and at the water table, for the pluvial climate scenario. In general, where zeolitic CHn and other low permeability layers are present, water accumulates above these units and is diverted laterally downdip until it reaches structural features that allow vertical movement to occur. In some instances, these structural features block lateral flow, and perching occurs. In the central and southern portion of the potential repository area low permeability zeolites are not present in the CHn, and as a result water moves vertically from the potential repository horizon to the water table. In contrast, at borehole UZ-14, low permeability layers block vertical movement of water, resulting in a lower percolation flux at the water table. Percolation flux is significant in the northwest portion of the model domain where infiltration is relatively large and lateral flow is blocked by the Solitario Canyon fault. Drill Hole Wash, Sever Wash, and Pagany Wash faults are also associated with high percolation fluxes at the water table.

Transient 3-D simulations were conducted to determine the response of temperature and gas pressure to climate change. Higher percolation fluxes resulted in a decrease in temperature due to convective heat transfer. Simulation times on the order of 1,000 to 10,000 years were required for the system to establish new steady-state temperature profiles. Analysis of transient changes in gas pressure revealed that the system reaches a steady-state fairly rapidly (within about 50 to 100 years), and that increased infiltration had little long-term impact on gas pressures.

Lastly, wetter climate simulation results showed an increase in the elevations of perched water from 0.1 to 7.9 m, with the largest changes occurring near borehole SD-12, due to the focusing of water

moving laterally and downdip above the crystal-poor vitric zone at the base of the Topopah Spring welded hydrogeologic unit.

5.3.4.4.5 Summary and Conclusions

In summary, the 3-D site-scale unsaturated-zone model is a valuable tool for examining the key unsaturated flow processes at Yucca Mountain. The model's ability to simulate the ambient state of moisture, heat, and gas flow and to predict the response of the system to potential future perturbations reflects the incorporation of many types of data collected at the site and extensive model calibration efforts. In addition, considerable attention has been given to the sensitivity of model parameters, such as rock properties and boundary conditions. The alternative numerical methods available to model complex, coupled flow and the effect of grid discretization have also been explored, with the most appropriate and efficient method selected for use in the 3-D unsaturated-zone modeling studies described above.

One of the most important components of unsaturated-zone site characterization is the estimation of percolation flux through the mountain across the potential repository area. Results of unsaturated-zone modeling studies suggest a range of average percolation flux from 1 to 10 mm/year. Confidence in this estimation is derived by considering several different aspects of the unsaturated- zone, such as moisture-balance, geothermal conditions, environmental isotope ratios, hydrochemistry, and fracture coating data; each suggesting similar values for average percolation flux. These values are in qualitative agreement with the results presented in Subsection 5.3.4.3.

Another important issue addressed by the 3-D site-scale unsaturated-zone model is fracture-matrix flow partitioning. Several fracture-matrix interaction conceptual models have been used to develop different calibrated hydrological property sets for use in unsaturated-zone model sensitivity studies. Pneumatic and geochemical data imply fractures to be the dominant pathway for percolation flux in the welded tuffs, such as the TCw and TSw hydrogeologic units. The unsaturated-zone model estimates that approximately 80 percent of the flow in the TSw travels through fractures. In the predominantly nonwelded units (PTn and CHn) where fractures are few and often discontinuous, flow is attenuated and spatially redistributed through the typically high-porosity matrix. However, near major fault zones, fracture densities may be greater in the nonwelded units; thus, fracture flow may be more significant along faults in the PTn and CHn.

Where alteration of nonwelded vitric material has produced low-permeability clays and zeolites, such as in the Calico Hills Formation, lateral diversion and, in some instances, perched water are seen at the altered/unaltered-rock interface. The spatial distribution and hydrologic character of altered zones have important implications for flowpaths and travel times below the potential repository horizon. Unfortunately, due to the sparseness of data at this depth, flow patterns below the potential repository horizon are uncertain. However, based on available borehole data, it is unlikely that percolation flux through zeolitic material exceeds more than a few mm/year because measured matrix permeabilities are relatively small. Consequently, the remainder of flow must be diverted laterally above the low-permeability layers, or transmitted vertically through fractures or fault zones to the water table. Continued analyses of strontium and other divalent cation data collected from pore water samples and

perched water bodies should provide important insight into flowpaths below the potential repository horizon, since these cation concentrations are strongly affected by flow through zeolitic layers.

Enhancement of the 3-D site-scale unsaturated-zone model, particularly with regard to flow characterization below the TSw hydrogeologic unit, will continue as new data are collected and incorporated into the model. Also, continued testing within the Exploratory Studies Facility is expected to provide important confirmatory data for the unsaturated-zone model. Confidence in the model will be enhanced if and when model predictions can be favorably compared to observations and test results.

5.3.5 Saturated-Zone System

5.3.5.1 Potentiometric Levels and Gradients

Potentiometric levels (water levels) have been monitored in wells in the Yucca Mountain area since 1981. Generally groundwater levels in the Yucca Mountain area are stable with time, with water-level fluctuations being attributed primarily to barometric and earth-tide changes. However, regional earthquakes have been observed to cause simultaneous water-level fluctuation in several wells.

Significant variations in the gradients of the potentiometric surface provide the basis for subdividing the potentiometric surface of Yucca Mountain area into three zones:

- Large hydraulic gradient, where potentiometric levels change at least 300 m in a few kilometers
- Moderate hydraulic gradient, where potentiometric levels change about 45 m in a few kilometers
- Small hydraulic gradient, where potentiometric levels change only about 2 m in several kilometers (Figure 5.3-222)

5.3.5.1.1 Potentiometric Levels

Potentiometric levels (water levels) have been monitored in wells in the Yucca Mountain area since 1981. The methods and frequency of water-level monitoring are discussed in Subsection 5.3.2.2.1, Water-Level Monitoring Program, and are not repeated here. The water-level data collected have been used to determine the direction and rate of groundwater flow, to estimate hydraulic characteristics of the flow system, and to determine if there are groundwater trends that could affect the potential repository (Graves et al. 1997, p. 3).

Generally water levels in wells in the Yucca Mountain area seem to be stable with time. After water levels have equilibrated following drilling or borehole workovers, they generally change very little with time (Luckey et al. 1996, p. 29) (an exception to this is well USW H-3, lower interval, which, as of 1996, was still recovering, seeking equilibrium following a borehole workover during 1990).

Water-level data from 1985 to 1995 for 28 wells monitoring 36 depth intervals were analyzed by Graves et al. (1997, Table 2) (Table 5.3-66). Annual mean water-level altitudes ranged from 727.93 to 1,034.60 m. The maximum range in water-level change was 12.22 m and the minimum range was 0.31 m. The change of 12.22 m is for well USW H-3, lower interval, and is thought to be due to the water level gradually seeking equilibrium following the 1990 borehole workover. In 31 of the 36 depth intervals monitored, the range of water-level change was less than 1 m. Generally, water-level fluctuations could be attributed to barometric and earth-tide changes.

Potentiometric Trends–Trends in groundwater water-level changes in the Yucca Mountain area have been discussed by Lehman et al. (1990), Ervin et al. (1994), Tucci and Burkhardt (1995), and Graves et al. (1997).

Lehman et al. (1990) analyzed 1983 to 1987 periodically measured water-level data for eight wells to determine if the data contained periodic (cosine) components. In their report, they conclude that there may be periodic components in five of the eight wells analyzed of 2- to 3-year periods in the fluctuation of the levels. They also state that the periodic behavior had a spatial distribution, with wells west of Yucca Mountain having different periods and phase shifts from wells on and east of Yucca Mountain. This relation of spatial distribution was compared to the findings of Matuska (1988), who discussed a similar spatial distribution of chemistry. Making this comparison, Lehman et al. (1990) suggested that the spatial distribution may be due to physical causes and the periodic fluctuations may be due to periodic recharge related to a periodicity in precipitation.

Ervin et al. (1994, p. 1) published a potentiometric-surface map of 1988 average water levels and analyzed 1986 to 1989 water levels for possible trends. Using linear least squares-regression of time verses water level, water-level data from 23 wells were analyzed. Of these wells, it was concluded that water levels in five wells exhibited apparent trends that were both statistically and hydrologically significant (Ervin et al. 1994, p. 11). These five wells (UE-25 WT-2, UE-25 WT#6, UE-25 WT#16, UE-25p #1, and USW H-5) all had apparent positive water level trends. The remaining wells analyzed displayed no trend.

Tucci and Burkhardt (1995) published a potentiometric-surface map of 1993 water levels and analyzed water-level data collected from 1986 to 1993 for possible water-level trends. The linear least-squares regression of time versus water level was also used for the trend analyses. Tucci and Burkhardt (1995, p. 13) concluded that water-level data from 3 of 22 wells analyzed exhibited apparent trends that were statistically significant. These wells (UE-25 WT#3, USW WT-7, and USW WT-10) all had apparent positive trends. The remaining wells had no trend. Of note, neither Ervin et al. (1994) nor Tucci and Burkhardt (1995) found the same wells to have positive trends.

Graves et al. (1997) presented hydrographs of 1985 to 1995 water-level data and discussed periods of rising, declining, or stable water levels in 36 depth intervals of 28 wells. In comparing the 1985 to 1995 data, few uniform water-level trends were detected. Graves et al. (1997, pp. 58-68, Figures 39-45) pointed out that:

- Seasonal water-level trends were not detected in any of the wells.

- Regional groundwater withdrawals did not appear to cause water-level changes.

- Most water-level fluctuations can be attributed to barometric and earth-tide changes.

- With the exception of wells located in Crater Flat, no uniform water-level changes were observed based on wells located in the same general area, or on wells with the same general water-table altitude. The wells located in Crater Flat with comparable water-level fluctuation were wells USW VH-1, USW WT-7, and USW WT-10 (Figures 5.3-222, 5.3-223, and 5.3-224). These wells also have similar water-level altitudes in the 776 m range. Water-level fluctuations in these two wells have similar patterns from February 1989 through July 1992. Vague similarities can also be seen between wells USW VH-1 and USW WT-7 from September 1986 through February 1987 and from June 1993 through January 1994. However, these similarities are not as continuous or striking as those seen between USW WT-7 and USW WT-10.

- Regional earthquakes occurring on June 28-29, 1992, were observed to cause simultaneous water-level fluctuations in several wells.

5.3.5.1.2 Earthquake Effects

On June 28, 1992, two earthquakes occurred, the first near Landers, California (296 km from the Yucca Mountain area) with a magnitude of 7.5, and the second near Big Bear, California (298 km from the Yucca Mountain area) with a magnitude of 6.6 (O'Brien 1993, Figure 1, Table 1). On June 29, 1992, one earthquake occurred at Little Skull Mountain, Nevada (located in southern Jackass Flats southeast of Yucca Mountain) with a magnitude of 5.6 (O'Brien 1993, Figure 1, Table 1). Two observation wells at Yucca Mountain, USW H-5 and USW H-6, were monitored continuously at that time, producing detailed records of effects of the seismic ground motion on the water levels (upper intervals) and confluid pressures (lower intervals) in both wells. The effects of the June 28 California earthquakes are shown at the same scales in Figure 5.3-225a and b for USW H-5, and in Figure 5.3-226a and b for USW H-6.

The confined pressure responses to all three earthquakes in the lower intervals of both wells exceeded the recording scales, although only slightly for the Little Skull Mountain earthquake at USW H-5. O'Brien (1993, p. 7) explains that the apparent sustained pressure decrease at USW H-6 resulted from the packer being pushed up-hole by the pressure surge and that pressure in the monitored zone returned to normal when the packer was reset. For the Landers, Big Bear Lake, and Little Skull Mountain earthquakes, respectively, O'Brien (1993, p. 8) estimates that the double-amplitude pressure responses in USW H-5 were 1.5 m, 1.0 m, and 0.6 m and that the responses in USW H-6 were 2.2 m, 1.4 m, and 1.1 m. Water-level responses in the upper intervals to the three earthquakes were estimated or measured as 0.9 m, 0.2 m, and 0.4 m in USW H-5 and as 0.6, 0.06, and 0.22 in USW H-6. The pressure recordings from the lower intervals provide the better record of the relative effects of the three earthquakes on crustal dilitation at Yucca Mountain. Borehole construction and rock-mass hydraulic characteristics near the boreholes strongly influence the water-level responses observed in the upper intervals.

O'Brien (1993, p. 1) also reported that these three earthquakes caused short-term responses in 17 depth intervals of 14 wells being monitored hourly. Long-term effects on water-level fluctuations from the June 28 - 29, 1992, earthquakes are thought to have occurred in wells UE-25 WT#4, UE-25 WT#6, UE-25p #1, and USW WT-11 (Graves et al. 1997, pp. 59 and 61) (Figures 5.3-222 and 5.3-227). These effects are:

- In well UE-25 WT#4, the water-level rose about 0.25 m (730.92 m on June 25, 1992 to 731.17 m on July 15, 1992) and then declined back to a water level consistent with levels prior to the earthquakes (730.83 m, August 10, 1992). This complete cycle of water-level fluctuation occurred within about 42 days of the earthquakes. (Because the measurements for well UE-25 WT#4 during this time were periodic, the maximum rise and fall in water level is not known).

- In well UE-25 WT#6, there was a maximum rise in water level of about 1.07 m (1,035.02 m on June 23, 1992 to 1,036.09 m on July 17, 1992) followed by a decline in water level of about 2.66 m (1,036.09 m on July 17, 1992 to 1,034.33 m on September 22, 1992). This cycle of water-level change occurred within about 85 days of the earthquakes. The water level continued to decline in well UE-25 WT#6 (0.07 m more) through October 16, 1992. However, because the water level in this well never recovered to the levels in the well prior to the earthquake, it is believed that water-level changes after September 22, 1992, are normal fluctuations in the well after it reached its new equilibrium. (Because the measurements for well UE-25 WT#6 during this time were periodic, the maximum rise and fall in water level is not known.)

- Water-level changes in well USW WT-11 immediately following the earthquakes are believed to be due to the earthquake activity. However, it is not known if the gradual 9-month cycle of water-level change following the earthquakes was coincidental or actually due to the earthquake events. In well USW WT-11, the water level initially increased (730.72 m on June 28, 1992, to 730.75 on June 29, and June 30, 1992) and then decreased (730.72 and 730.68 on July 1, and July 2, 1992, respectively). This quick change was followed by a slow decrease in water level from July through November 1992, with a maximum change in water level of 0.24 m (730.72 m on June 28, 1992 to 730.48 m on November 26, 1992). Following this low, the water level slowly rose back to 730.72 m on April 5, 1993, which is a water level consistent with levels prior to the earthquake. This whole cycle of decline and rise in water level occurred within about 9 months of the earthquakes (July 1992 through March 1993). (Hourly data were collected for well USW WT-11 during 1992. Hydrograph data plotted in Figure 5.3-227 represent monthly mean values for all data except June 28 - July 2, 1992, which are daily mean values.)

- In well UE-25p #1, the water level declined about 0.53 m (752.44 m on June 28, 1992 to 751.91 m on July 8, 1992). Following this, the water level in well UE-25p #1 began to rise and had fully recovered from the effects of the earthquakes by December 7, 1992 (water level of 752.44 m). This complete cycle of water-level change occurred within about 191 days of the earthquakes. Hourly data were collected for well UE-25p #1 during 1992. Hydrograph

data plotted in Figure 5.3-227 represent monthly mean values for all data except June 28 - July 8, 1992, which are daily mean values.)

O'Brien (1993, p. 10 and Figure 6) states that it is difficult to determine from the hourly hydrograph record at UE-25p #1 whether the June 28 Landers earthquake or the June 29 Little Skull Mountain earthquake had the dominant effect on well UE-25p #1. Water-level excursions appear at the time of each earthquake, and the downward trend of the water level, though somewhat obscured by earth-tide effects, appears to have started on June 28.

An alternative view of the effects of the June 28-29, 1992, earthquakes on water-level change in the Yucca Mountain area is presented by Davies and Archambeau (1997). Davies and Archambeau (1997, p. 26) state that in response to the Little Skull earthquake which occurred on June 29, 1992, a long-term increase in water-level change was observed with a maximum increase of about 5 m. Wells cited as having significant water-level change were "J-11 and J-12" and "AD-11 and AD-16" (Davies and Archambeau 1997, p. 26, Figures 3 and 4). The water-level data presented by Davies and Archambeau (1997) for these wells was reviewed and compared to water-level data collected by the USGS for the same wells and for the same period of time. Comparison of the two data sets indicated that the data presented by Davies and Archambeau (1997) were flawed:

- Davies and Archambeau (1997, p. 26) reference O'Brien (1993) when discussing the locations of the four wells monitored and the fluctuations of water levels in these wells. The report by O'Brien (1993) shows well locations and water-level changes due to the June 28-29, 1992, earthquakes only in wells USW H-3, USW H-5, USW H-6, and UE-25p #1, not in wells J-11, J-12, AD-11, or AD-16.

- Davies and Archambeau (1997) attribute the water-level changes to the Little Skull Mountain earthquake without mention of the larger California earthquakes on the previous day. Figure 4 of their report shows abrupt water-level rises initiated at the same times, unlabeled but presumably June 29, for each of the four wells, but they do not acknowledge explicitly that the wells were measured at approximately monthly intervals.

- Davies and Archambeau (1997, Figures 3 and 4) located well AD-11 incorrectly and plotted a hydrograph of water-level change that was completely erroneous for well AD-11. The water-level fluctuation in well AD-11 before and after the Little Skull earthquake was only 0.04 feet (water level 225.14 feet below land surface, June 15, 1992; water level 225.10 feet below land surface, July 20, 1992) (La Camera and Westenberg 1994, p. 72, Table 2, plate 1). The location for well AD-11 shown by Davies and Archambeau (1997, Figure 3) agrees with the location for well AD-10 (La Camera and Westenburg 1994, plate 1), and the decline of water-level altitude from about 2,184 to about 2,182 feet during the second half of 1992 also agrees with the actual record for well AD-10 (La Camera and Westenburg 1994, p. 71, Table 2). However, the water-level change in well AD-10 before and after the earthquake was 2.4 feet (altitude of water level 2,181.7 feet, April 13, 1992; altitude of water level 2,184.1 feet, August 18, 1992) (La Camera and Westenburg 1994, p. 71, Table 2), not 15 feet as shown by Davies and Archambeau (1997, Figure 4). The actual dates of measurements

also do not support the abrupt rise at the end of June as depicted by Davies and Archambeau (1997, Figure 4).

- Well AD-16 reported by Davies and Archambeau (1997, Figures 3 and 4) could not be found in any other report. Consequently, the water levels reported by Davies and Archambeau could not be confirmed. However, the location of well AD-16 was found to be at the same location, and the water level at approximately the same altitude, as well AD-12 (La Camera and Westenberg 1994, plate 1). The water-level change reported for well AD-12 was compared to the water-level change reported by Davies and Archambeau for well AD-16. The water-level altitudes and change in each well were about the same, so it may be assumed that Davies and Archambeau's well AD-16 is the same as well AD-12 in La Camera and Westenberg (1994, pp. 72-73, Table 2). Again, however, the actual dates of measurements (June 15, 1992 and July 20, 1992) (La Camera and Westenburg 1994, p. 72, Table 2) do not support the abrupt rise at the end of June as depicted by Davies and Archambeau (1997, Figure 4).

- Davies and Archambeau (1997, Figure 4) report water-level increases in wells J-11 and J-12 of about 3 and 5 feet, respectively. O'Brien et al. (1995, Figures 13 and 14, Tables 15 and 16) report minimal water change in wells J-11 and J-12, with a maximum monthly fluctuation of only 0.13 feet in either well J-11 or J-12 from June through August of 1992.

The hydrographs presented by Davies and Archambeau (1997, Figure 4) all show water-level rises, which those authors represent as resulting from the strain change that is expected in a normal-faulting environment. However, in three of the four hydrographs in Figure 5.3-227, water-level declines followed the earthquakes, although O'Brien (1993, p. 10) does report a probable 28 cm rise in pressure in the lower interval of USW H-3. The National Park Service also reported a slight (<25 cm) decline in the level of the pool in Devils Hole (O'Brien 1993, p. 10), an exposed fault zone in the lower carbonate aquifer at Ash Meadows, close to the "AD" wells discussed by Davies and Archambeau (1997).

Earthquakes occurring prior to the June 28-29, 1992 events, also had some, though minimal, effects on water levels in the Yucca Mountain area. Four earthquakes occurring in California during late April 1992 produced water-level fluctuations in well USW H-5 (O'Brien 1992). On April 22, 1992, a 6.2-magnitude earthquake near Palm Springs, California (approximately 311 km from the Yucca Mountain area) caused a peak water-level rise and fall in well USW H-5 upper interval of approximately 26.1 cm. On April 25-26, 1992, three earthquakes of magnitudes 7.0, 6.5, and 6.6 occurred near Eureka, California (approximately 789 km from the Yucca Mountain area), and caused a peak water-level rise and fall in excess of 52.5 cm in well USW H-5, upper interval. Following any of the April 1992 earthquakes, the fluctuations in water levels generally lasted no more than 2 hours. No long-term effects on water-level fluctuations were seen in other wells in the Yucca Mountain area due to the April 1992 earthquakes. Short-term effects could have occurred in other wells, but only well USW H-5 was instrumented to monitor for earthquake activities during this time.

5.3.5.1.3 Potentiometric Surface

Several potentiometric maps of the Yucca Mountain area have been constructed. These include maps by Robison (1984; J.H. Robison [U.S. Geological Survey], written communication to D.L. Vieth [U.S. Department of Energy/Nevada Operations Office], September 17, 1986 [regarding revisions of Yucca Mountain water levels reported in Robison, 1984], TIC 219488), Waddell et al. (1984), Ervin et al. (1994), Tucci and Burkhardt (1995), and Czarnecki, Faunt et al. (1997).

Robison (1984, Figure 2) constructed a preliminary potentiometric-surface map of the Yucca Mountain area using 1983 water levels (Figure 5.3-228). The map, with a variable contour interval of 100, 20, and 10 m, shows contours for 1,000 to 730 m above sea level, and indicates a groundwater movement toward the south and east. The contours indicate a large hydraulic gradient in the northern part of the Yucca Mountain area, with a southward water-level decrease of 60 m per 0.5 km (gradient of 0.12), and 200 m in 1.5 km (gradient of 0.13) (Figures 5.3-222 and 5.3-228). A moderate eastward hydraulic gradient of 20 m in 0.4 km (gradient of 0.05) is shown just west of the crest of Yucca Mountain in the vicinity of Solitario Canyon. Most of the area east of Solitario Canyon and south of borehole USW H-1 has a very small hydraulic gradient. There is only a single contour (730 m) in this area. The 1983 data were not sufficiently accurate to compute a meaningful hydraulic gradient in the small-gradient area (Luckey et al. 1996, p. 21).

Two potentiometric maps were prepared by Waddell et al. (1984, pl. 3, Figure 8). One is for the area defined as the Death Valley region (discussed in detail in Subsection 3.5.2) and includes the Yucca Mountain Site, and the other defines the potentiometric surface of the immediate Yucca Mountain area. Contour lines for the Death Valley region map were drawn at 100 m intervals, and composite water levels from several hydrogeologic units were used in mapping the contours (Waddell et al. 1984, plate 3, p. 28). Water-level data sources for the regional map were from Eakin (1962, 1963, 1966), Malmberg and Eakin (1962), Walker and Eakin (1963), Malmberg (1967), Rush (1970), Thordarson and Robinson (1971), Winograd and Thordarson (1975), and Miller (1977). The direction of groundwater flow in the Yucca mountain area from the regional map is from the north, northwest, and west toward Yucca Mountain and Jackass Flats. Flow out of Jackass Flats is to the southwest toward Amargosa Valley and Amargosa Desert.

Contour lines for the Yucca Mountain Site map were variable at 100, 50, and 25 m contours; with a range of contours from 1,200 to 725 m (Waddell et al. 1984, Figure 8) (Figure 5.3-229). Water levels used for the site map represent composite water levels, which, in effect, are weighted averages of hydraulic heads in boreholes (Waddell et al. 1984, p. 59). The source of the water-level data used in the site map is assumed to be the same as for the region map by Waddell et al. (1984). The general direction of groundwater movement is the same as shown on the region map with groundwater moving from the north, northwest, and west toward Yucca Mountain and Jackass Flats. Because of the scale of the map, groundwater flow out of Jackass Flats is not shown, but is assumed to be southwest to Amargosa Valley and Amargosa Desert.

Ervin et al. (1994, plate 1) presents a revised potentiometric-surface map in the area of the small hydraulic gradient (east of Solitario Canyon and south of borehole USW H-1) (Figures 5.3-222 and 5.3-230). The large and moderate hydraulic gradients were identified and discussed in their report, but

were not contoured because no new data were available to update the map constructed by Robison (1984). The potentiometric-surfacemap prepared by Ervin et al. (1994) primarily used mean 1988 water-level data. These data are considered to be more accurate in the area of the small hydraulic gradient than the data that were available during 1983 for the report by Robison (1984). This improvement in water-level data is attributed to the improvements in the techniques used to make the 1988 water-level measurements. The potentiometric-surface map by Ervin et al. (1994) has a contour interval of 0.25 m and indicates a range of water-level altitudes from about 731.0 m beneath the crest of Yucca Mountain to about 728.5 m near Fortymile Wash. The predominant direction of the groundwater gradient is to the east-southeast (Luckey et al. 1996, p. 21).

Tucci and Burkhardt (1995, Figure 4) constructed a potentiometric-surface map of the Yucca Mountain area using 1993 average water levels (Figure 5.3-231). This map, with a variable contour interval of 20, 8, and 0.50 m, shows contours for 800 to 728 m above sea level. As in the potentiometric map by Robison (1984), this map shows areas of large, moderate, and small hydraulic gradients (Figures 5.3-222 and 5.3-231). The large hydraulic gradient is defined in an area to the north-northeast of Yucca Mountain, where water-level altitude ranges 1,034 to 738 m. The moderate hydraulic gradient is defined in an area on the west side of Yucca Mountain where the water-level altitude ranges from about 780 to 740 m, and the groundwater flow appears to be impeded by the Solitario Canyon fault and a splay of that fault. The small hydraulic gradient is defined in an area east and southeast of Yucca Mountain where water-level altitude ranges from 732 to 728 m (Tucci and Burkhardt 1995, p. 9).

Tucci and Burkhardt (1995) have defined the potentiometric surface in the Yucca Mountain area in greater detail than Robison (1984) and Ervin et al. (1994), extending contours farther east into Jackass Flats and west into Crater Flat. From the potentiometric map of Tucci and Burkhardt (1995, pp. 8-9) groundwater flows from the north and west toward Yucca Mountain, continuing east to an area just east of Fortymile Wash, where flow is to the south. Groundwater would also flow west from Jackass Flats to an area just east of Fortymile Wash, and then south toward the Amargosa Desert.

Czarnecki, Faunt et al. (1997, pp. 25 and 26, and Figure 5) presents a computer-generated potentiometric surface map, developed from water levels collected from 1952 to 1996, that was used to develop a preliminary 3-D groundwater flow model of the site saturated zone of the Yucca Mountain area (Figure 5.3-232). The map has a contour interval of 25 m and contours of 1,175 to 700 m, and extends east to Little Skull Mountain, west to Crater Flat, north to Shoshone Mountain, and south to Amargosa Valley and the Amargosa Desert. Direction of groundwater flow is from the north, northwest, and west toward Yucca Mountain and Jackass Flat. Though the potentiometric surface east of Yucca Mountain in the area of the small hydraulic gradient is not defined, the direction of groundwater movement out of Jackass Flats is south toward Amargosa Valley and Amargosa Desert. Groundwater flow from the western part of Crater Flat is also south to Amargosa Desert.

5.3.5.1.4 Potentiometric Gradients

Variations in the gradients of the potentiometric surface provide the basis for subdividing the Yucca Mountain area into three zones:

- Large hydraulic gradient, where potentiometric levels change at least 300 m in a few kilometers
- Moderate hydraulic gradient, where potentiometric levels change about 45 m in a few kilometers
- Small hydraulic gradient, where potentiometric levels change only about 2 m in several kilometers (Figure 5.3-222)

Large Hydraulic Gradient–Luckey et al. (1996, pp. 21, 24, 25) and Czarnecki, Faunt et al. (1997, pp. 27-29) present hypotheses that might explain the occurrence of the large hydraulic gradient. Luckey et al. (1996) gives the most detailed discussions of the possible causes as follows.

- The large hydraulic gradient is simply the result of flow through the upper-volcanic confining unit. The 1994 water-level altitude in borehole USW G-2 at the northern end of the large hydraulic gradient was 1,020 m and is in the upper volcanic confining unit (Table 5.3-67). The base of the upper volcanic confining unit is at an altitude of about 730 m, so the saturated thickness of the confining unit in this area is almost 300 m. A large hydraulic gradient could be expected in a thick confining unit. Because the unit that comprises the upper volcanic confining unit is not saturated to its full thickness at borehole USW G-2, the large hydraulic gradient may persist somewhat north to the area where the potentiometric surface is in the upper volcanic aquifer. The lower volcanic aquifer in this area also could have decreased hydraulic conductivity because of lithostatic pressure and hydrothermal alteration, which also would contribute to the large hydraulic gradient.

- Ervin et al. (1994, pp. 9-11) advanced the hypothesis that the large hydraulic gradient represents a perched or semiperched system (Meinzer 1923, p. 41). In a perched or semiperched system, flow in the upper and lower aquifers would be predominantly horizontal, whereas flow in the confining unit would be predominantly vertical. Winograd and Thordarson (1975, p. 50) reported that semiperched water is not uncommon at and in the vicinity of the Nevada Test Site, and perched water has been encountered in several wells at Yucca Mountain. The water levels to the north of the large hydraulic gradient represent levels in the upper volcanic aquifer, whereas water levels to the south represent levels in the lower volcanic aquifer (Table 5.3-67). The upper volcanic confining unit separates these two systems. Under this hypothesis, the large hydraulic gradient may simply be an artifact of attempting to contour points from two different and widely separated surfaces. Water levels from boreholes completed in the confining unit between these two different surfaces, where flow is essentially vertical and water levels change dramatically with depth, would be difficult to interpret. To the north of the large hydraulic gradient, flow in the upper volcanic aquifer probably is limited with most of the water being lost to downward seepage into the lower

volcanic aquifer. What little water remains in the upper volcanic aquifer at the north end of the large hydraulic gradient probably is lost in a short distance as it moves vertically through the upper volcanic confining unit and ultimately reaches the lower volcanic aquifer (Ervin et al. 1994, pp. 9-11). The permeability of the lower volcanic aquifer at the north end of Yucca Mountain may be extremely limited due to lithostatic loading and to alteration. Water in the upper volcanic aquifer also may be diverted around the large hydraulic gradient into Crater Flat or along Fortymile Wash. In this hypothesis (Ervin et al. 1994, pp. 9-11), the hydraulic gradient in the lower volcanic aquifer probably increases to the north as hydraulic conductivity decreases, but does not increase as dramatically as shown on the potentiometric-surface map (Figure 5.3-228). Borehole USW G-2 in this area produced little, if any, water from the lower volcanic aquifer (see "Flow Surveys" subsection).

- The large hydraulic gradient is due to drainage into a fault zone that moves water from the volcanic aquifers into the underlying lower carbonate aquifer (Fridrich et al. 1991, 1994). The fault is buried beneath the Calico Hills Formation and is the northern-bounding fault of a buried graben. There is no surface expression of the postulated fault, which marks the northern extent of active flow in the lower carbonate aquifer. The drain removes much of the water from the volcanic aquifers and greatly decreases flow through the volcanic aquifers to the south, which in part accounts for the small hydraulic gradient to the south and east of Yucca Mountain. Part of the water returns to the lower volcanic aquifer south of the large hydraulic gradient by upward seepage along faults, most notably the Solitario Canyon, Bow Ridge, and Paintbrush Canyon faults. Thermal evidence for the upward seepage, but which does not appear to support the drain, is discussed in Subsection 5.3.5.2.

- Another hypothesis proposed by Fridrich et al. (1994) to explain the large hydraulic gradient is the spillway hypothesis. This explanation requires the same deeply buried fault, but in this hypothesis the fault marks the effective northern limit of the lower volcanic aquifer. Fridrich et al. (1994) speculate that, as a result of the fault, the Crater Flat Group is thinner to the north of the graben, is more altered, and is much less permeable than it is to the south. This explanation envisions very little flow from the north, and what little flow there is drops abruptly at the northern-bounding fault of the buried graben.

- The large hydraulic gradient is the result of the presence, at depth, of the Eleana Formation, which is part of the Paleozoic upper clastic confining unit described by Winograd and Thordarson (1975, p. 43). This possibility may have been considered when the large hydraulic gradient was discovered in 1981, with the drilling of borehole USW G-2, but it has not been previously discussed in detail in publications. However, Fridrich et al. (1994, p. 150) noted the regional large hydraulic gradient appeared to follow the contact between the upper clastic confining unit and the carbonate aquifer. The Eleana Formation is responsible for a large hydraulic gradient on the west side of Yucca Flat in the northeastern part of the Nevada Test Site (Winograd and Thordarson 1975, p. 66). The Eleana Formation could be deeply buried beneath northern Yucca Mountain and could be affecting the flow system in that area in the same way it affects the flow system on the west side of Yucca Flat. No borehole has penetrated deeply enough to verify the existence or absence of the Eleana Formation to the north of Yucca Mountain. More than 5 km south of the large hydraulic gradient, the Eleana

Formation is not present in borehole UE-25p #1, which penetrates the pre-Cenozoic section in carbonates of Silurian age beneath a fault contact. However, the Eleana Formation is present as an altered, strongly magnetic argillite in the Calico Hills, about 15 km east of Yucca Mountain. Bath and Jahren (1984) attributed a prominent magnetic high to this argillite and suggested that the westward extension of the magnetic high beneath northern Yucca Mountain also indicated the presence of the Eleana Formation. G.D. Robinson (1985, Figure 9) projected the approximate location of the southern extent of the superposition of the confining unit on the carbonate aquifer, a location that correlates with the general locality of the large- hydraulic gradient. The absence of the Paleozoic upper clastic confining unit south of the large hydraulic gradient could provide groundwater access to the carbonate aquifer, but the overlying lower volcanic confining unit also could prevent this access.

Szymanski (1989) and, more recently, Davies and Archambeau (1997) have attributed the large hydraulic gradient to an area of decreased permeability due to relatively larger fracture-closing crustal stress (least principal stress). Davies and Archambeau (1997) report fracture-opening pressures deduced from slug-injection tests in boreholes at Yucca Mountain, including for at least two boreholes (USW G-1 and USW G-2) that were not tested by the slug-test technique. However, USW G-1 and USW G-2 were among the holes in which in situ stress and the orientation of least principal stress were determined by the production and testing of new fractures by controlled hydraulic fracturing, and observation of the induced fractures by acoustic televiewer (Stock and Healy 1988). The results of Stock and Healy (1988) and an earlier compilation of regional-stress characteristics by hydraulic fracturing and other techniques (Stock et al. 1985) do not support the assignment by Szymanski (1989) and Davies and Archambeau (1997) of different stress regimes to areas within and south of the large hydraulic gradient.

Moderate Hydraulic Gradient–The moderate hydraulic gradient is defined by boreholes with the altitudes of the potentiometric surface ranging from about 732 m to about 775 to 780 m (Figures 5.3-222 and 5.3-228) and is west of the crest of Yucca Mountain (Luckey et al. 1996, p. 25). Boreholes USW H-6, USW WT-7, and USW WT-10 (water-level altitudes of about 776 m) are located on the west side of the Solitario Canyon fault. Borehole USW H-5 (water-level altitude of about 775 m) is located to the east of the Solitario Canyon fault, but on the downthrown (west) side of a major northeast-trending splay of the Solitario Canyon fault. Boreholes USW H-1 and USW H-3 (water-level altitude about 731 m) are located on the east side of the Solitario Canyon fault. The Solitario Canyon fault is a major north-striking scissors fault, which to the south is downthrown on its western side and to the north is downthrown on the eastern side. The hinge line of the fault, where the displacement changes, is perpendicular to the fault plane and located approximately 1 km southwest of borehole USW G-2. Offset on the fault may be as much as 250 m. Toward its southern end, the Solitario Canyon fault seems to widen and have more splays. Fault gouge and secondary siliceous infillings are present along the fault. Samples of this material indicate a low-matrix porosity. Although fault gouge and siliceous infillings are surface features, they indicate that the fault, in the saturated zone, could have less permeability than the surrounding rock (Luckey et al. 1996, p. 25).

The Solitario Canyon fault seems to function as a barrier to flow from west to east. Boreholes west of the fault have higher water levels than boreholes east of the fault, except for borehole USW H-5. However, borehole USW H-5 may be connected hydraulically with the boreholes west of the Solitario

Canyon fault. According to Bentley et al. (1983, p. 20) and Robison and Craig (1991, pp. 12-13), much of the flow in borehole USW H-5 originates from the part of the Bullfrog Tuff located at an altitude of between 720 to 780 m. Thus, most of the flow to the borehole comes from an interval that is above the interval where the fault splay probably intersects the borehole. The fault splay may be less permeable as a result of fault gouge. This low-permeability interval may tend to mound water against the fault splay (Luckey et al. 1996, p. 25).

The Solitario Canyon fault could also function as a barrier because of offset of stratigraphic units, which places more permeable units against less permeable units, especially to the south where offset in the fault is greater (Ervin et al. 1994, p. 9). Some flow undoubtedly crosses the Solitario Canyon fault because there is a large difference of hydraulic potential (45 m) across it. However, most of the groundwater west of the Solitario Canyon fault probably flows south, either along the fault or through an aquifer in Crater Flat (Luckey et al. 1996, p. 25).

The moderate hydraulic gradient could be interpreted as a southward extension of the large hydraulic gradient, with or without the Solitario Canyon fault causing the moderate gradient. This extension of the large gradient could be implied by information presented by Robison (1984, Figure 2) and Fridrich et al. (1994, Figure 3). As discussed for the large hydraulic gradient, Fridrich et al. (1994) suggested that there could be substantial upwelling of water from the carbonate aquifer along the Solitario Canyon fault, as well as along other north-striking faults. If such upwelling along the Solitario Canyon fault is larger compared to along other faults, or if the hydraulic conductivity east of the Solitario Canyon fault is less than elsewhere around Yucca Mountain, the upwelling could cause the moderate hydraulic gradient. However, if upwelling is the cause of the moderate gradient, the Solitario Canyon fault seems to be unique in this respect because similar gradients are not observed across other north-striking faults in the Yucca Mountain area (Luckey et al. 1996, pp. 25-26). In addition, higher water levels occur west of the fault in Crater Flat (Figure 5.3-231), so upwelling might modify the shape of the moderate hydraulic gradient but cannot be the dominant cause.

Small Hydraulic Gradient–The small hydraulic gradient is present beneath much of eastern and southern Yucca Mountain where water levels ranged from 728 to 732 m above sea level (Figures 5.3-221 and 5.3-230); this area includes most of the boreholes at the Yucca Mountain Site. The small hydraulic gradient is bounded on the north by boreholes USW H-1 and UE-25 WT#14 and on the west by boreholes USW H-1, USW H-3, and USW WT-11. The small hydraulic gradient extends east as far as Fortymile Wash, where the water-level altitude is about 728 m, and probably extends into central Jackass Flats, where the water-level altitude at borehole UE-25 J-11 is about 732 m. The southern extent of the small hydraulic gradient includes boreholes USW WT-11, and UE-25 WT#12, but the small hydraulic gradient probably extends still farther to the south (Figure 5.3-229). The small hydraulic gradient represents a water-level change of 0.1 to 0.3 m/km (gradient of 0.0001 to 0.0003). Ervin et al. (1993, p. 1558) stated that the small hydraulic gradient could indicate highly transmissive rocks, limited groundwater flow through the system, or a combination of both phenomena. The actual case cannot be determined until further data are available (Luckey et al. 1996, p. 26, 27).

The potentiometric surface in the area of the small hydraulic gradient slopes generally eastward from Yucca Mountain toward Fortymile Wash; a smaller component slopes to the south. Limited data east of Fortymile Wash indicate that the potentiometric surface in Jackass Flats also is relatively flat and

probably slopes westward toward Fortymile Wash; again, a smaller component slopes to the south (Luckey et al. 1996, p. 27).

Vertical Gradients– Potentiometric levels were measured during hydraulic testing at multiple depth intervals at 10 sites in the vicinity of Yucca Mountain. Differences in potentiometric levels at different depth intervals in the same borehole ranged from about zero to nearly 55 m (Table 5.3-68) (Luckey et al. 1996, pp. 27-28).

Potentiometric levels generally were higher in the lower intervals of the volcanic rocks than in the upper intervals, indicating a potential for upward groundwater movement. However, at four boreholes (USW G-4, USW H-1, USW H-6, and UE-25b #1), potentiometric levels in the volcanic rocks were higher in the uppermost intervals than in the next lower intervals (Table 5.3-68). At borehole USW G-4, there was a downward hydraulic gradient with a potentiometric difference of 0.5 m between the upper and lower intervals. Both intervals are in the lower volcanic aquifer. At borehole USW H-1, the potentiometric level was about 0.2 m higher in the uppermost interval than in the next lower interval. The potentiometric level in the third interval in borehole USW H-1 was about 5 m higher than that in the uppermost interval. The upper three intervals are all in the lower volcanic aquifer. The potentiometric level in the lowermost interval of borehole USW H-1, completed in the lower volcanic confining unit, was nearly 55 m higher than in the uppermost interval. At borehole USW H-6, there was a downward gradient with a potentiometric difference of about 0.1 m between the intervals that were isolated in the lower volcanic aquifer. However, based on data from 1984 (Robison et al. 1988, p. 122, 123), there seemed to be an upward potentiometric difference of about 2 m between the lower volcanic confining unit and the lower volcanic aquifer in borehole USW H-6. At borehole UE-25 b#1, there was a downward-potentiometric difference of about 1 m between the lower volcanic aquifer and the lower volcanic confining unit (Luckey et al. 1996, p. 28)

Potentiometric levels in the Paleozoic carbonate aquifer at borehole UE-25p #1 are about 21 m higher than levels in the lower volcanic aquifer (Craig and Johnson 1984, p. 12) (Table 5.3-68). A potential for upward groundwater movement from the Paleozoic rocks to the volcanic rocks were, therefore, indicated. Because of the large difference in potentiometric levels in these two aquifers, they seem to be hydraulically separate (Luckey et al. 1996, p. 28). However, analyses of March 1984 hydraulic-test data at the C-hole Complex (boreholes UE-25c #1, UE-25c #2, and UE-25c #3, Figure 5.3-222) indicated a possible hydraulic connection between the lower volcanic aquifer and the carbonate aquifer at the C-hole Complex (Geldon 1996). During the March 1984 test of borehole UE-25c #2, the water level in borehole UE-25p #1 apparently declined; Geldon (1996) attributed this decline to a connection between the lower volcanic aquifer and the carbonate aquifer. However, water-level monitoring instruments at that time may not have been reliable, and any conclusions from data collected would be tenuous. Testing at the C-hole Complex from 1995 to 1997 did not result in any water-level decline in borehole UE-25p #1 (Geldon et al. 1997, p. 47). Further testing at the C-hole Complex (DOE 1990, Section 3.4) using more reliable water-level monitoring instruments may help to resolve the question of hydraulic connection between the lower volcanic aquifer and the carbonate aquifer at this site (Luckey et al. 1996, p. 28).

Measurement of potentiometric levels at various depth intervals in borehole UE-25p #1 shortly after the borehole was drilled indicated that:

- Potentiometric levels gradually increased from 729.9 to 734.5 m above sea level between the water table (depth of about 384 m) and deep within the lower volcanic confining unit, 1,114 m below land surface

- Potentiometric levels abruptly increased to about 752 m above sea level in the "older tuffs" between depths of 1,110 and 1,180 m below land surface

- Potentiometric levels remained at about 751 m above sea level throughout the Paleozoic carbonate aquifer to a total depth of 1,805 m (Craig and Robison 1984, pp. 6-9) (Luckey et al. 1996, p. 28).

The lowermost 70 m of the older tuffs of the lower volcanic confining unit in the borehole had potentiometric levels that were very similar to the levels in the carbonate aquifer, indicating that the lower most part of the lower volcanic confining unit and the carbonate aquifer were hydraulically connected. Such a connection could be expected in the hanging-wall rocks adjacent to a fault, and such a connection is supported by calcification of the basal tuffs in the borehole (Carr, M.D. et al. 1986, pp. 23-25). The remaining 237 m of the lower volcanic confining unit had a potentiometric level similar to that of the lower volcanic aquifer (Craig and Robison 1984, Table 2) (Luckey et al. 1996, p. 28).

No unambiguous areal patterns in the distribution of vertical hydraulic gradients around Yucca Mountain are apparent; however, some generalizations can be made as to the distribution of potentiometric levels in the lower sections of the volcanic rocks. Potentiometric levels in the lower volcanic confining units are relatively high (altitude greater than 750 m) in the western and northern parts of Yucca Mountain (Figure 5.3-233,) and relatively low (altitude about 730 m) in the eastern part of Yucca Mountain. Based on potentiometric levels that were measured in borehole UE-25p #1, the potentiometric levels in the lower volcanic confining unit in boreholes USW H-1, USW H-3, USW H-5, and USW H-6 may reflect the potentiometric level in the carbonate aquifer. Boreholes UE-25b #1 and USW H-4 do not seem to fit the pattern established by the other boreholes (Figure 5.3-233). These two boreholes penetrated only 31 and 64 m, respectively, into the lower volcanic confining unit and had potentiometric levels (about 730 m), similar to those in the lower volcanic aquifer. Penetration of the other four boreholes into the lower volcanic confining unit ranged from 123 m at USW H-3 to 726 m at USW H-1 (Table 5.3-67). Perhaps only in boreholes USW H-1, USW H-3, USW H-5, and USW H-6 are the potentiometric levels in the lower volcanic confining unit influenced by the potentiometric level in the carbonate aquifer (Luckey et al. 1996, p. 28-29).

Vertical-hydraulic gradients could have an important impact on the analysis of the effectiveness of the saturated zone as a barrier to radionuclide transport. Based on a very limited data set (five boreholes), an aerially extensive upward gradient can be inferred between the carbonate aquifer and the volcanic aquifers, which may indicate that, at least for the immediate Yucca Mountain area, radionuclide transport would be restricted to the volcanic system (Luckey et al. 1996, p. 29).

5.3.5.2 Temperature, Thermal Gradients, and Heat Flow

The thermal regime of the earth's crust is governed by several factors: the thickness of the crust; the input of heat from the underlying mantle; heat generated within the crust by radioactive decay and rock friction; heat emplaced by magma intrusions; and movement of heat in advecting fluids. In the upper few kilometers of the crust, moving groundwater may significantly perturb the conductive heat-flow field. Sass, Lachenbruch et al. (1971) and Sass, Blackwell et al. (1981) attributed the "Eureka Low," an area of regional heat-flow deficiency (<63 mW m^{-2}, or <1.5 heat-flow units) in southeastern Nevada, to downward seepage of water and subsequent lateral transport of water and its contained heat in the regional Paleozoic carbonate aquifer.

Sass and Lachenbruch (1982) reported calculated heat flows for 14 holes in the Nevada Test Site area (Figure 5.3-234), as well as measured temperature profiles in numerous additional holes. The Eureka Low extends into the Nevada Test Site from the north and, tentatively, includes the Yucca Mountain Site. Typical basin and range heat flows, 63 to 105 mW m^{-2} (1.5 to 2.5 heat-flow units), occur in a few holes in north-central NTS and near the southern and southeastern borders. Test Well 5 (TW5 in Figure 5.3-234) penetrates the Precambrian confining unit and provides the only direct measurement of crustal heat flow (84 mW m^{-2}) in the vicinity of Yucca Mountain beneath the advective flow in the lower carbonate aquifer. Anomalously high heat flow, 130 mW m^{-2}, was reported by Sass, Lachenbruch et al. (1980) for UE-25a #3, the exploratory hole in the Calico Hills, whereas anomalously low heat flow was reported by Sass and Lachenbruch (1982) and by Sass, Lachenbruch et al. (1988) for most boreholes at and near Yucca Mountain.

Sass, Lachenbruch et al. (1988) report temperatures, thermal conductivities, and calculated heat flows for both the unsaturated and saturated zones in the vicinity of Yucca Mountain. Saturated-zone heat flow at and near the potential repository site generally ranges between 30 and 55 mW m^{-2}, comparable to heat flow in the Eureka Low. Figure 5.3-235 shows the holes for which saturated-zone temperature measurements are reported in Sass, Lachenbruch et al. (1988) and heat flows calculated for linear, apparently conductive segments of the temperature profiles.

Water Table Temperature–Temperatures at the water table, which were determined from the temperature profiles in Sass, Lachenbruch et al. (1988) for all holes that reached the water table, are shown and contoured in Figure 5.3-236. As contoured, the pattern indicates a distinct high-temperature zone (34 to 38.8°C) associated with Solitario Canyon and a less prominent high-temperature zone (32 to 34°C) associated with Midway Valley. Both Solitario Canyon and Midway Valley overlie zones of north-trending faults, suggesting a genetic relationship between faulting and high-temperature anomalies at the water table. Fridrich et al. (1994) and Sass, Dudley et al. (1995) suggest that the faults provide permeable pathways for upward flow to the shallow saturated zone from a deeper aquifer, tentatively the Paleozoic carbonate aquifer.

Alternatively, the pattern of water table temperatures has been interpreted (Lehman and Brown 1996) to show southward-protruding lobes of relatively cool water (<32°C), possibly resulting from recharge or, in the case of the central lobe, from southward-deepening flow from relatively shallow, cool aquifers north of the large-hydraulic gradient (Figure 5.3-236).

Temperature Logs–Temperature logs measured in the saturated zone in boreholes east of the Yucca Mountain ridge crest are shown in Figure 5.3-237. The log for USW G-1 is composed of two approximately linear segments, indicating that fluid movement in the hole has not significantly masked the conductive thermal profile, although a small temperature reversal at a depth of about 1050 m shows that drilling effects had not completely dissipated. The log for USW H-1, which almost superimposes on that for USW G-1, is the last of four logs measured in November 1980, April 1981, November 1982, and March 1983 (Sass, Lachenbruch et al. 1988). The last three of these are nearly identical, showing that post-drilling stability was reached in USW H-1, as well as indicating indirectly that the temperature in USW G-1 was also stable except for the small residual reversal.

Below a depth of about 1,000 m in USW G-1 and USW H-1, the least-squares temperature gradients are both about 30°C/km (Sass, Lachenbruch et al. 1988). In the deepest 100 meters of USW H-1, however, there is a slight, but actual, decrease of the gradient. Though subtle, this curvature is consistent with the entry of water into the hole and then flowing upward, consistent with the large-hydraulic head (altitude 785 m) at a depth of 1800 m in that hole, as discussed above in Subsection 5.3.5.1. In the upper part of USW H-1, there is apparent upward intraborehole flow from a depth of about 775 to 625 m.

A much more distinct example of uphole flow is the log from USW G-4 (Figure 5.3-238). Below a depth of 150 m, four logs obtained between March 1983 and June 1984 are identical (Sass, Lachenbruch et al. 1988, Figure 1-12), indicating that post-drilling equilibrium had been achieved. As shown in Figure 5.3-238, the 1-D flow calculation indicates uphole flow of about 0.1 m/year. Luckey et al. (1996, Table 3) report that potentiometric levels measured during testing in January 1983 were 0.5 m higher in the upper half of the saturated zone (615 to 747 m) than in the lower half (747 to 915 m), which would produce a small gradient for downward flow. However, the stability of the convex-upward temperature profile in the borehole, indicated by four effectively identical logs from March 1983 through June 1984 (Sass, Lachenbruch et al. 1988, Figure 1-12), is rather convincing evidence that the flow is upward, suggesting that conditions in the borehole during testing may still have been perturbed by borehole construction and earlier pumping tests.

In the shallower sections of UE-25b #1 and USW H-4 (Figure 5.3-237), approximately 600 to 900 m, the logs are almost isothermal, indicating significant flow of water in the holes. Extremely small vertical-hydraulic gradients are reported by Luckey et al. (1996, Table 3), upward with a head difference of 0.1 m in USW H-4 and downward with a head difference of 1.0 m in UE-25b #1, despite the apparent similarity of the two temperature logs. Both of these hydraulic gradients are based on bottom-hole measurements in packed-off intervals during hydraulic testing. The transmissivity of the critical interval in UE-25b #1, 1006 to 1220 m, was extremely small (2×10^{-2} m^2/day), providing an apparent average hydraulic conductivity of 10^{-4} m/day. It is possible that, because of prior pumping tests, an equilibrium head was not obtained during the test of this interval, diminishing confidence in the downward gradient. Therefore, the identification of flow direction in UE-25b #1 remains inconclusive. Because of the similarity of the USW H-4 and UE-25b #1 logs and the geographic proximity of the holes, the flow direction in USW H-4 is also classed as in determinant despite the agreement between the temperature log and the very small-hydraulic gradient.

The deeper, linear segments in USW H-4 and UE-25b #1, respectively, have apparently conductive gradients of about 26 and 23°C/km (Figure 5.3-237). Flow surveys conducted while pumping these boreholes show that only 14 percent of the yield of USW H-4 was produced from depths greater than 920 m (Whitfield, Eshom et al. 1985) and that there was no detectable yield below 875 m in UE-25b #1 (Lahoud et al. 1984). These observations confirm that the temperature profile in UE-25b #1 is conductive and that in USW H-4 probably is conductive given the lack of a significant vertical hydraulic gradient. Consequently, it can be inferred that rock temperatures below depths of about 900 m are, in fact, several degrees cooler in these holes, located on the eastern boundary of the ridge, than those in USW G-1 and USW H-1, which are within the ridge in upper Drillhole Wash.

Figure 5.3-239 shows the temperature logs that were obtained from USW H-6, along the western edge of Solitario Canyon, and from three holes (USW G-3, USW H-3, and USW H-5) along the Yucca Mountain ridgecrest on the east side of Solitario Canyon. The data from these boreholes help to define the high-temperature anomaly of the water table along the Solitario Canyon fault. USW H-6 is in the hanging-wall block, that overlying the west-dipping fault, whereas the other three holes are in the foot-wall block.

The logs for USW G-3 and USW H-3 are not distinctive. That for USW G-3 shows irregularities that indicate intraborehole flow, but the direction cannot be determined with confidence. Head measurements in USW H-3 indicated an upward gradient (Luckey et al. 1996; Subsection 5.3.5.1 above). The log is permissive, but not diagnostic, for slight upward flow.

The USW H-5 log shows active intraborehole flow and, in the uppermost 100 m, probably temperature homogenization resulting from mixing of flow in the aquifer. This interval, from 700 to about 800 m depth, provided almost 90 percent of the pumped yield during a flow survey. The interval from 800 to 1,050 m displays a nearly isothermal log, but the segment is distinctly convex upward, indicating upward flow. The bottom of this interval, at the base of the Tram Tuff, yielded almost all of the remaining 10 to 12 percent of the flow-survey pumpage. Monitoring of water levels in a zone isolated below a packer set at about 1090 m shows that the bottom 110 m of the hole has a head that is slightly higher (about 0.16 m) than the composite head in the hole above the packer (Graves et al. 1997). Thus, the hydraulic data support the thermal interpretation of upward flow. The packer blocked access to the hole below 1,090 m after February 1983, but the log shown in Figure 5.3-239 (obtained in October 1983) is identical to the November 1982 log above the 1,090 m depth. From 1,090 to the total depth of 1,200 m, the earlier log was clearly conductive with a gradient of 25°C/km (Sass, Lachenbruch et al. 1988, Figure 1-16).

The thermal regime at USW H-6 is obviously different from that in the foot-wall block of the Solitario Canyon fault. The temperatures and the gradient in the head wall are significantly greater than those in the foot-wall boreholes (Figure 5.3-239), indicating the thermal effect of the fault that is projected to underlie USW H-6. The temperature profile below about 900 m is conductive with a gradient of 36°C/km.

In USW H-6, the positive temperature excursion above a depth of 625 m is difficult to interpret based on this October 1983 log alone. However, Sass, Lachenbruch et al. (1988, Figure 1-17) provide logs also for November 1982 (shortly after drilling and testing), March 1983, and March 1984.

Figure 5.3-240 shows the March 1984 log, which is very similar to that of March 1983 and which, except in the interval from 620 to 775 m, is exactly superimposed on the October 1983 log. The earlier and later March logs both show distinctly the concave-upward curvature that results from downhole flow in this interval, which is bordered above and below by the only two productive zones in the borehole (Craig and Reed 1991). It is very probable that the March logs provide direct observations of the thermal effects of winter or springtime recharge in the wash in which USW H-6 is located. The water table at this site is about 520 m below the land surface.

Borehole UE-25p #1 provided the most distinctive temperature log that have been obtained at Yucca Mountain (Figure 5.3-241). As described in Subsection 5.3.5.1, the head increased slightly with depth during testing in the volcanic rocks but rose abruptly when drilling penetrated altered tuffs and a fault zone above the lower carbonate aquifer. Although the hole was eventually cased into the Paleozoic rocks, intense uphole flow of warm water (55°C) occurred for several weeks from a depth of 1,200 m through a temporary, leaky plug. A temperature log in February 1983 showed that the upward flow exited the borehole at a depth of about 500 m into the most productive zone of the volcanic section, as shown by a borehole flow survey while pumping (Craig and Robison 1984). A persistent high-temperature excursion of the temperature log at this zone, shown in the October 1983 log (Figure 5.3-241a), indicated that the effects of penetrating into the deep-fault zone, probably the Paintbrush Canyon fault, were long lived. Subsequent logs recorded the gradual decay of the anomaly at 500 m until a stable configuration was reached. The new equilibrium profile of April 1990, shown in Figure 5.3-241b, is essentially identical to the October 1983 log except for smoothing of the temperature anomalies at depths of about 500 m and 1,300 m. Based on the geologic analysis of UE-25p #1 (Carr, M.D. et al. 1986), both of these anomalies are associated with faults penetrated by the borehole.

The basic shape of the persistent temperature profile in UE-25p #1, which is dominated by positive excursions at shallow and deep fault zones, makes it difficult to assign any part of the log to a conductive thermal gradient. However, the similarities between the logs for UE-25p #1 and USW H-6 (Figure 5.3-239) are striking. In the deeper parts of the volcanic rocks, the two holes have similar temperatures and gradients in the hanging-wall blocks above faults that are associated with high-temperature anomalies at the water table. The temperatures at depths of 1,000 m in UE-25p #1 (48°C) and USW H-6 (47°C) are distinctly greater than those for other drillholes at and near the site. However, the temperature decreases below the lower anomaly in UE-25p #1, and below a depth of 1,500 feet it is less than the temperature in USW G-1 (Figure 5.3-241a).

Figure 5.3-242 shows one interpretation of the thermal regime in the vicinity of UE-25p #1. Upwelling of the isotherms implies that fault-controlled flow parallel to the strike of the faults is accompanied by buoyant rise, and thus vertical mixing, of warm water within the fault zones. The actual flowpaths for the anomalously warm water do not necessarily rise above the confining layer at the base of the volcanic section. Rather, the mounding of the isotherms may reflect mainly conductive diffusion of the heat source at the base of the volcanics and possibly minor upward seepage along the faults. (Bredehoeft (1997) estimates by modeling that, for an assumed width of 3 m for the deeper fault, a fault-zone permeability through the saturated tuffs of about 0.1 m/day would provide sufficient mass and heat transfer to approximate the water table temperature anomaly but still provide sufficient confinement of the lower carbonate aquifer to produce the observed earth-tide response in the

borehole.) Note that this thermal cross section is also consistent with the downward reversal of the temperature anomaly in the UE-25p #1 log within the carbonate aquifer.

In this interpretation, the temperature between the two fault zones through most of the saturated Tertiary rocks, trends toward ambient as it does below the deeper fault, but it remains somewhat elevated. A conductive temperature profile, unperturbed by faulting, can be approximately reconstructed by connecting the bottom-hole and surface temperatures, varying the slope of the profile according to the thermal conductivities of the rocks. The corresponding heat flow would be somewhat less than the 42 mW m^{-2} shown in Figure 5.3-241 because the temperature at the bottom of the hole is still decreasing with depth.

Alternatively, Sass (1998) emphasizes the more regionally normal heat flow determined for the unsaturated segment (62 mW m^{-2}) of the UE-25p #1 log, together with a more traditional interpretation of the concave-upward profile to suggest downward flow between the upper and lower fault zones. However, head measurements during hydraulic testing in the borehole reveal a small but distinct upward hydraulic gradient in the tuffs (Craig and Robison 1984). Furthermore, the reversal of temperature below the lower fault zone attests to the significant warming that the fault produces. If it were merely the receptor for hydraulically driven downward flow, the lower fault should be at a lower temperature than its contiguous rocks. Given the pronounced positive temperature anomaly associated with the lower zone, the more subdued anomaly that is proposed to be associated with a fault higher in the hole appears to be reasonable. The consequent warming of the upper saturated zone would provide the heat needed to support the 62 mW m^{-2} heat flow in the unsaturated zone near the upper fault despite apparent lower heat flow into the base of the Tertiary rocks.

Borehole USW G-2 is located on the broad upland of Yucca Mountain, north of the finger-like ridges that protrude southward. As was discussed in the preceding Subsection (5.3.5.1), it is at or near the northern edge of the large southward gradient of the apparent water table. A series of eight temperature logs obtained between December 1981 and June 1984 provided effectively identical profiles (Sass, Lachenbruch et al. 1988, Figure 1-10) below the static water level which is about 525 m below land surface. The temperature profile exhibited a few steps of the gradient in the first 90 m below the water level and then a nearly isothermal step from 615 m down to about 750 m. In September 1992 and again in February and September 1995, the hole was reoccupied for studies of the large hydraulic gradient. Rather than confirming the thermal stability that had been implied by the 1981 to 1984 repetitive logging, the isothermal step had dissipated to a more moderate gradient, and by September 1995 the base of the step had risen about 30 m (Figure 5.3-243). This thermal history, a 12-m decline of the static water level between 1984 and 1995, and a possible geophysical indication of undersaturation have been interpreted to indicate the gradual downhole drainage of a perched water body that was disrupted by the borehole (Czarnecki, Nelson et al. 1995; J.H. Sass, U.S. Geological Survey, oral presentation to the YMP Saturated Zone Expert Elicitation Panel, June 6, 1997, Denver, Colorado).

The temperature log for USW G-2 below the depth of 750 m is shown in Figure 5.3-244, together with the logs for USW G-1, USW H-6, and UE-25p #1, which are repeated for reference. Although the USW G-2 log is cooler at any given depth than those for USW H-6 and UE-25p #1, the gradient is about the same as that in the lower part of USW H-6, a clearly conductive thermal regime based on both the temperature log and the hydraulic testing results. The temperatures at depth and average

gradient in USW G-1, which is the closest of the three holes to USW G-2 (Figure 5.3-235) and which also is clearly conductive, are distinctly less than those at USW G-2, indicating differences in the thermal and, thus, hydrogeologic regimes.

Conductive Heat Flow–Conductive heat-flow values are shown in Figure 5.3-235 for boreholes in which there were apparently conductive temperature-log segments, supported by hydraulic testing information that showed insufficient permeability or hydraulic gradient to produce significant intra-borehole flow. With the exceptions of UE-25p #1 and USW G-4, which did not display conductive profiles, heat-flow estimates can be made for all of the boreholes in the site area that penetrated the saturated zone significantly.

Potentially conductive segments of the temperature profiles in Figures 5.3-237,-239, and-244 were evaluated in the context of hydraulic data. The thermal gradients in the conductive segments were multiplied by the thermal conductivities of the corresponding rocks, from Sass, Lachenbruch et al. (1988, Tables 3, 3-1, and 3-2), to obtain heat flows. The results indicate that saturated-zone conductive heat flows may be geographically grouped.

The lowest heat flows, 32 and 36 mW m^{-2}, were calculated for USW H-3 and USW G-3, respectively. These boreholes are the southernmost deep holes in the site area. Both are sited on the Yucca Mountain ridge crest just east of Solitario Canyon. They penetrate similar stratigraphy and have similar thermal gradients.

UE-25b #1 and USW H-4 are both within the faulted zone that borders the potential site on the east, UE-25b #1 being at the mouth of Drillhole Wash and USW H-4 being about 1.5 km southwest in a smaller, northwest-trending wash. The heat flows calculated for these holes are 40 and 45 mW m^{-2}, respectively.

The heat flow at USW H-5, 55 mW m^{-2}, is similar to those at nearby boreholes in upper Drillhole Wash, USW G-1 and USW H-1, which have calculated values of 53 and 54 mW m^{-2}. The value for USW H-5 is based on the November 1982 log, which provided a linear, clearly conductive segment from 1100 to 1300 m depth. This interval penetrated a zeolitic lava of negligible permeability beneath the Tram Tuff (Robison and Craig 1991) and a thermal conductivity estimated to be 2.2 W m^{-1} K^{-1} (Sass, Lachenbruch et al. 1988, Table 3). Hydraulic testing of USW H-1 showed that significant permeability was restricted to zones shallower than about 750 m, so the deepest 1,000 m should be free of advective heat transport, which is consistent with its uniform temperature profile (Figure 5.3-237). The profile for USW G-1 is nearly identical to that for USW H-1 so that, despite the absence of hydraulic-test information, it is considered to be conductive.

As was discussed above, USW H-6 penetrates the hanging wall block of the Solitario Canyon fault, and its relatively high heat flow, 65 mW m^{-2}, is considered to be a consequence of its local hydrogeologic setting rather than necessarily being a marginally typical basin and range value.

USW G-2, in the northern part of Yucca Mountain and at the upper limit of the large hydraulic gradient, has the highest calculated heat flow in the site area, 65-70 mW m^{-2}, based on the deepest accessible 125 m interval (1125 to 1250 m) after the hole was bridged subsequent to the

September 1982 log. However, the September 1982 log (Sass, Lachenbruch et al. 1988, Figure 1-10) shows an irregular profile between 1,125 and 1,400 m that suggests but does not demonstrate slight downhole flow. Because downhole flow would suppress the thermal gradient, the higher end of the estimated heat flow, 70 mW m^{-2}, is considered the more likely value.

Heat flow in the site area is deficient with respect to typical basin and range values, such as the 84 mW m^{-2} determined for Test Well 5. Rather, the site-area heat flow is consistent with that of the Eureka Low (<65 mW m^{-2}), which has long been considered to be caused by regional groundwater movement (Sass, Lachenbruch et al. 1971; Sass, Dudley et al. 1995).

Hydrologic Implications of Temperature and Heat Flow–The heat-flow deficiency in the site area extends at least to near the base of the Tertiary volcanic rocks, based on the logs for deep holes such as USW G-1, USW H-3, and USW G-3. The tentatively assigned heat flow in the lower carbonate aquifer at UE-25p #1 (<42 mW m^{-2}) may indicate that the anomaly penetrates deeply into the Paleozoic rocks also. Based on these considerations, Sass, Lachenbruch et al. (1988) and Sass, Dudley et al. (1995) concluded that flow in the lower carbonate aquifer intercepts crustal heat flow, transporting it laterally toward discharge areas.

The Fridrich et al. (1994) hypothesis that a fault underlying the large hydraulic gradient provides a path for introducing downflow of cool water into the Paleozoic rocks is conceptually consistent with this general concept. However, heat flow at USW G-2, 65 to 70 mW m^{-2} is marginally deficient and may indicate some hydrologic capture of heat flux at the northern edge of the gradient. Values of 50 to 55 mW m^{-2} at USW G-1, USW H-1, and USW H-5 occur near the toe of the large hydraulic gradient, whereas lower heat flows (40 to 45 mW m^{-2}) occur farther southeast at UE-25b #1 and USW H-4. The lowest heat flows, 32 and 36 mW m^{-2}, were determined for the southernmost boreholes, USW H-3 and USW G-3, both of which were drilled on the ridge crest of Yucca Mountain on the east (footwall) side of the Solitario Canyon fault. The decrease of heat flow southward from the large gradient is not consistent with a simple model of cool recharge to the lower carbonate aquifer and southward warming of that flow as heat is collected. The density and distribution of data are not sufficient to resolve in detail the flowpaths that might cause the observed heat-flow anomalies.

The linear temperature anomalies that correlate spatially with the Solitario Canyon fault and with faults in Midway Valley clearly indicate that flow is strongly influenced by the north-south striking faults. However, as Bredehoeft (1997) concludes, only a minor upward flux from the carbonate aquifer into the Tertiary volcanics is required to produce the anomaly in Midway Valley. The relatively high temperatures and heat flows in USW H-6 and UE-25p #1, above the Solitario Canyon and Paintbrush Canyon faults, respectively, attest to the local importance of vertical mixing of dominantly lateral flow along the faults.

Alternative Interpretations–In an alternative explanation for the Eureka Low, Blakely (1988) analyzes aeromagnetic data for the State of Nevada, producing a map showing the interpreted depth to the Curie isotherm, about 580°C, the temperature at which crustal rocks lose their magnetism. Blakely suggests geographic coincidence of the Eureka Low with magnetic-basement depths on the order of 25 to 30 km. Although Yucca Mountain is outside of the area providing sufficient aeromagnetic coverage for interpretation, extrapolation over a relatively small distance would support

a predicted much shallower Curie depth, approximately 15 km. If this were demonstrated and if Curie-isotherm depth were shown to be the principal determinant of near-surface (upper few kilometers) heat flow, the low heat flow of the site area would be enigmatic. However, Sass, Dudley et al. (1995) point out that heat flow virtually doubles over a distance of a few kilometers at the site area, indicating control by groundwater movement in the upper 3 to 4 km of the crust. Such short-wavelength heat-flow variations could not be caused by differences 10 km or deeper in the crust. This does not preclude the possibility that the heat flux into the base of the hydrologically active system (i.e., into the base of the lower carbonate aquifer) is approximately normal in the vicinity of Yucca Mountain but less than normal beneath the large area of the Eureka Low.

Swanberg and Morgan (1978) used the silica geothermometer to estimate regional heat flow throughout the United States. Their results for Nevada indicated high heat flow extending from the Battle Mountain High in Northern Nevada southward through the Yucca Mountain area (including the western half of the Eureka Low) to Death Valley. The many heat-flow values determined from measured-temperature gradients and rock-thermal conductivities are incompatible with the silica-based analysis, which was not compensated for the high content of soluble silica glass in the rhyolitic rocks of the Nevada volcanic province.

In discussing the tectonic and thermal framework of Yucca Mountain, C.A. Hill et al. (1995, p. 71) cite thermal springs as characteristic of the region and heat flow "as high as 130 mW m^{-2}"as characteristic of the Yucca Mountain Site. Stuckless, Marshall et al. (1998) reply that the highest spring temperatures in the vicinity occur in Oasis Valley, west of Crater Flat, where deeply circulating groundwater from Pahute Mesa and Timber Mountain emerges at springs with a maximum temperature of 41°C. It is evident from the several temperature logs shown and discussed above that temperatures of 40 to 60°C are attainable within 2 km of the surface in an area of clearly deficient heat flow. The cited heat flow of 130 mW m^{-2} occurs in borehole UE-25a #3 (Sass, Lachenbruch et al. 1980), which is more than 10 km east of Yucca Mountain (Figure 5.3-234). The borehole penetrates argillite of the Eleana Formation before entering carbonate rocks that probably are part of the lower carbonate aquifer, which is structurally high beneath the Calico Hills. It is likely that the high heat flow in the argillite is driven by deeper warm water rising along a fault zone as is apparent in UE-25p #1 and suspected at USW H-6. Regardless of the cause in the Calico Hills, the high value of heat flow was not observed at Yucca Mountain, where there are abundant data showing regionally deficient heat flow.

5.3.5.3 Hydrochemical and Isotopic Evidence of Flowpaths and Groundwater Age

Note: Per agreement of J. Stuckless and Z. Peterman, USGS, with A. Meijer, Los Alamos National Laboratory, in the January 1998, SD/Geochem Coordination Meeting in Albuquerque, Los Alamos National Laboratory will provide a pertinent section. When it is available, the points that are important to Subsection 5.3.5 will be summarized and the section in the Geochemistry chapter will be referenced.

5.3.5.4 Conceptual Model of the Saturated Zone at Yucca Mountain

The following discussion is taken largely from Luckey et al. (1996), who provide a status of understanding of saturated-zone groundwater flow at Yucca Mountain as of 1995. Despite the passage

of time since that report was written, current (1998) concepts have not changed significantly. Updated and alternate concepts of site saturated-zone groundwater flow are presented and cited as necessary within this subsection.

5.3.5.4.1 Site Saturated-Zone Groundwater Flow

The groundwater flow system at Yucca Mountain primarily occurs in a sequence of fractured volcanic rocks and underlying carbonate rocks. Groundwater also flows within Tertiary alluvial and carbonate deposits, where the volcanic rocks pinch out, south of Yucca Mountain in the Amargosa Desert. The volcanic sequence is divided into two aquifers (the upper and lower volcanic aquifers) and two confining units (the upper and lower volcanic confining units) (Luckey et al. 1996). The division does not correlate exactly with either the stratigraphic units or the thermal/mechanical units, and is useful only as a general guideline as to where flow might occur. Czarnecki, Faunt et al. (1997) define three volcanic aquifers and confining units for purposes of simulating groundwater flow in the Yucca Mountain area; however, their lowermost aquifer does not underlie the immediate Yucca Mountain area considered here. Generally, only a small number of intervals within the aquifers, usually associated with fractures or faults, produce water. The confining units also may transmit water, but to a lesser extent than the aquifers. In both the aquifers and the confining units, fractures probably transmit more water than the rock matrix does. The less transmissive units, such as the Calico Hills Formation, probably have a much larger proportion of their flow occurring in the matrix.

Little is known about the carbonate rocks beneath Yucca Mountain, but they constitute a major regional aquifer. The carbonate aquifer probably underlies the lower volcanic confining unit beneath Yucca Mountain. Several lines of evidence indicate that the lower volcanic confining unit effectively isolates the carbonate aquifer from the volcanic aquifers. However, the altered mineralogy of the tuffs, particularly in borehole UE-25p #1, indicates past influx of deeper water, possibly localized along faults (Carr, M.D. et al. 1986, p. 24). Higher water table temperatures aligned along existing faults may indicate upwelling, probably from the carbonate aquifer (Fridrich et al. 1994). Beneath the northern part of Yucca Mountain, the carbonate aquifer may additionally be confined by the Eleana Formation. If the volcanic aquifers are essentially isolated from the carbonate aquifer, then the potentiometric surface in the volcanic aquifer would be independent of the potentiometric surface in the carbonate aquifer. If, however, the confinement is breached by faults, the potentiometric surface in one system may affect the potentiometric surface in the other system. Recent analysis of fault permeability at borehole UE-25p #1 by Bredehoeft (1997) indicates that the carbonate aquifer is well confined by a layer of low-hydraulic conductivity. However, that modeling analysis also supports the concept of upward flow from the carbonate aquifer along faults.

On a regional scale, flow is generally from the higher plateaus in the north to the Amargosa Desert in the south, probably continuing southward or westward, or both, to still lower areas of discharge. How flow moves through the Timber Mountain area is not clear because of lack of potentiometric data (Czarnecki, Steinkampf et al. 1990; Czarnecki, Faunt et al. 1997). In the immediate vicinity of Yucca Mountain, the flow direction seems to be more eastward as indicated by the potentiometric contours (Ervin et al. 1994; Tucci and Burkhardt 1995), although anisotropy may deflect part of the flow southward. Whatever part of the flow that persists eastward to reach the Fortymile Wash area would then flow south toward the Amargosa Desert.

In the area of small hydraulic gradient at Yucca Mountain, the potentiometric surface is a continuous, almost flat surface that does not change noticeably as it passes through the various faults east of the Solitario Canyon fault. The potentiometric surface slopes gently to the east at a relatively constant gradient (0.0001 to 0.0003). Potentiometric surfaces constructed for 1988 (Ervin et al. 1994) and 1993 (Tucci and Burkhardt 1995) both show this type of surface and gradient. Lehman and Brown (1996) present an alternate potentiometric surface that appears to be influenced by northwest-southeast-trending structures on the east side of Yucca Mountain; however, the data used to construct that map are not presented in their report so that the validity of their map cannot be evaluated.

The potentiometric surface in the area of small hydraulic gradient also does not seem to change as it transects aquifers and confining units. Beneath the crest of Yucca Mountain, flow is entirely in the lower volcanic aquifer and deeper units. However, at Fortymile Wash, the upper volcanic aquifer dips beneath the water table and dominates the flow system. There is no obvious break in the slope of the potentiometric surface in the transition area, which must be somewhere in Midway Valley. In the vicinity of the large hydraulic gradient in the northern Yucca Mountain area, the potentiometric surface may or may not reflect a change in hydrogeologic units, depending on the interpretation of the cause of the large hydraulic gradient.

Vertical hydraulic gradients could have an important impact on the analysis of the effectiveness of the saturated zone as a barrier to radionuclide transport. Based on a very limited data set (five boreholes), an areally extensive upward gradient can be inferred between the carbonate aquifer and the volcanic aquifers, which may indicate that, at least for the immediate Yucca Mountain area, radionuclide transport would be restricted to the volcanic system.

Potentiometric levels have been measured at multiple depth intervals at 10 sites in the vicinity of Yucca Mountain. Differences in potentiometric levels at different depth intervals in the same borehole ranged from about zero to nearly 55 m. Potentiometric levels generally are higher in the lower intervals of the volcanic rocks than in the upper intervals, indicating a potential for upward groundwater movement. However, at four boreholes (USW G-4, USW H-1, USW H-6, and UE-25b #1), potentiometric levels in the volcanic rocks are higher in the uppermost intervals than in the next lower intervals. Potentiometric levels in the Paleozoic carbonate aquifer at borehole UE-25p #1 are about 21 m higher than levels in the lower volcanic aquifer (Craig and Johnson 1984, p. 12). A potential for upward groundwater movement from the Paleozoic rocks to the volcanic rocks was, therefore, indicated.

The volume of groundwater that flows beneath Yucca Mountain is unknown. If the large hydraulic gradient to the north of Yucca Mountain and the moderate-hydraulic gradient to the west of the crest of the mountain effectively isolate the volcanic flow system from the regional flow system and, if the lower volcanic confining unit effectively isolates the volcanic and carbonate flow systems from each other, little water could be moving beneath the mountain. If these barriers are not particularly effective, then groundwater flow beneath Yucca Mountain could be much greater. The small hydraulic gradient can be interpreted as an area of either high transmissivity or small groundwater flux (Ervin et al. 1993, p. 1,558), or some combination of the two. Because there is no direct way to measure groundwater flow, resolution of the possible causes of the small hydraulic gradient would depend on both direct and indirect evidence. Sufficient testing to determine transmissivity and anisotropy on

scales of tens to hundreds of meters, combined with measured hydraulic gradients, would allow a direct estimate of flow. Detailed hydrochemical characterization, combined with analysis of rock-water interaction and water-chemistry evolution, would provide indirect evidence of the rate of flow. Water ages estimated from isotopic data also would provide indirect evidence.

5.3.5.4.2 Large Hydraulic Gradient Area

As discussed in Subsection 5.3.5.1.4, Potentiometric Gradients, several hypotheses have been proposed for the cause of the large hydraulic gradient that is believed to be present north of Yucca Mountain. Simulation of this feature in groundwater flow models has typically been through the use of low-permeability features such as large areas of low permeability (Barr, G.E. and Miller 1987; Lehman and Brown 1996; Haws 1990; Cohen et al. 1997) or a low permeability barrier (Czarnecki and Waddell 1984; Czarnecki, Faunt et al. 1997). Czarnecki, Faunt et al. (1997, p. 28) indicate that simulation of the drain conceptual model resulted in unacceptable matches between simulated and observed heads. Other conceptual models have not been simulated to date (1998).

Coppersmith and Perman (1997) discuss the results of an Expert Elicitation conducted during 1997 for saturated-zone flow and transport at Yucca Mountain. The panel of five experts reviewed data and arguments in support of the various hypothesized causes of the large hydraulic gradient during a series of three workshops. The panel chose two concepts as having significant credibility (Coppersmith and Perman 1997, pp. 3-5). In the saturated zone model, the large hydraulic gradient is interpreted to be the result of a steeply inclined water table slope within a fully saturated flow system. Factors such as topography, recharge, and geology contribute to the slope of the feature. In the perched water model, the flow system consists of a saturated interval overlying a wedge of unsaturated rocks, which in turn overlies the regional water table. The panel considered both hypotheses about equally, with a slight preference toward the perched water model (Coppersmith and Perman 1997, pp. 3-6). The panel was split in their opinion concerning the importance of the large hydraulic gradient on implications for potential radionuclide transport from the potential repository area (Coppersmith and Perman 1997, pp. 3-6).

5.3.5.4.3 Influence of Faults

Faults have been described as both barriers and conduits to groundwater flow in the region and at Yucca Mountain. Faunt (1997) characterizes the influence of faults on groundwater movement in the Death Valley region, and relates their influence to crustal stress, fracture mechanics, and structural geologic data. Also, in a regional context, D'Agnese et al. (1997, p. 90) included faults that were both conduits (northeast-southwest-trending features) and barriers (northwest-southeast-trendingfeatures) to groundwater flow. Locally at Yucca Mountain, however, the northwest-southeast-trending structures may not be barriers to flow, at least in the trend direction. Hydraulic and tracer tests in Drill Hole Wash (Lahoud et al. 1984; Ogard et al. 1983) indicate that the underlying structure, whether it is a fault or merely a fracture zone, provides continuous, permeable flow paths. Also, Geldon et al. (1997) infer from drawdown in observation wells while pumping at the C-hole Complex that northwest-southeaststructure provides a permeable connection to drillhole USW H-4.

Most researchers have concluded that the Solitario Canyon fault, on the west side of Yucca Mountain, is a barrier to west-to-east groundwater flow (Luckey et al. 1996; Czarnecki, Faunt et al. 1997) and is a probable cause for the moderate hydraulic gradient (Luckey et al. 1996; Ervin et al. 1994; Tucci and Burkhardt 1995). However, the Solitario Canyon fault may also provide a conduit for upward flow to the volcanic aquifers from the underlying carbonate aquifer (Fridrich et al. 1994). Other north-trending faults at Yucca Mountain may also provide conduits for upward flow, particularly where they correspond to linear anomalies in water temperature (Fridrich et al. 1994; Sass, Dudley et al. 1995; Lehman and Brown 1996; Bredehoeft 1997). The apparent roles of the Solitario Canyon fault as a barrier to flow across the fault but as a conduit for flow parallel to it indicates the likelihood that the fault zone itself may be poorly permeable but that the well-fractured hanging-wall rocks are highly permeable.

Several other investigators have discussed the role of faults and other structural features as possible conduits or preferential pathways for groundwater flow at Yucca Mountain (Luckey et al. 1996; Lehman and Brown 1996; Cohen et al. 1997; Geldon 1996; Geldon et al. 1997). At the C-hole Complex, the relatively large hydraulic conductivity is strongly influenced by the presence of faults and fractures that intersect the borehole (Geldon et al. 1997). The distribution of drawdown in response to extensive pumping at the C-hole Complex is believed to be related to the location of nearby structural features, and Geldon et al. postulate that the areal distribution of hydraulic conductivity in the area may be related to the presence or absence of faults. Little information is available, however, concerning hydraulic properties within the saturated zone of individual faults in the area.

5.3.5.4.4 Recharge

Recharge to the site area is assumed to be from the following sources:

- Downward and possible lateral recharge from episodic flooding of Fortymile Wash
- Throughflow from Pahute and Rainier Mesas, which is hypothesized to result in recharge along the northern border of the study area
- Throughflow from the northwestern part of the Amargosa Desert
- Minor recharge from episodic flooding of the Amargosa River channel
- Net infiltration from precipitation events (Czarnecki, Faunt et al. 1997)

Additional water may be entering the volcanic aquifers from the underlying carbonate aquifer, particularly along north-south trending structural features (see Subsection 5.3.5.4.1; Bredehoeft 1997; Czarnecki, Faunt et al. 1997).

Fortymile Wash is a major southward-draining ephemeral channel located adjacent to Yucca Mountain, and it is thought to contribute intermittent recharge to the saturated zone. Based on geomorphic/distributed-parameter simulations, Osterkamp et al. (1994) estimated recharge along the entire 95-km length of Fortymile Wash to be about 4.22×10^6 m^3/year. Based on field studies of stream

loss (Savard 1998), the total recharge in Fortymile Wash is estimated at about 27,000 m^3/year. This estimate represents a minimum value based on the inability to account for all reaches of Fortymile Wash, which may have received unobserved runoff and recharge, coupled with the minimum period of streamflow observations.

Throughflow from the northwestern part of the Amargosa Desert and Pahute and Rainier Mesas is difficult to quantify. As a result, these fluxes usually are calculated in a groundwater flow model. The regional flow model (D'Agnese et al. 1997) employs the concept that some of the recharge waters from Pahute and Rainier Mesas likely flow to Yucca Mountain. Because of uncertainties that result from the paucity of data, alternative conceptual models of flow toward Yucca Mountain include the possibility of either a groundwater divide between the Crater Flat/Yucca Mountain area and Beatty Wash, or a westward extension of a potential hydrogeologic barrier into northern Crater Flat (Czarnecki, Steinkampf et al. 1990). Both possibilities would result in diversion of more water from Pahute Mesa into the Oasis Valley subbasin, although the second case would permit some flow into Crater Flat.

The Amargosa River is an intermittent stream in the southwestern portion of the model area, where channelized flow ceases to exist. Streamflow in the Amargosa Farms area is generally very limited (Osterkamp et al. 1994). Based on channel-morphology measurements, the composite average recharge is estimated to be 0.2×10^6 m^3/year along the 15.9-km length Amargosa River from Ashton to Big Dune, an area proximate to the model area (Osterkamp et al. 1994). Recharge is, therefore, assumed to be negligible along the few kilometers of the Amargosa River in the southwestern portion of the model.

A detailed description of net infiltration to the water table in the vicinity of Yucca Mountain has been developed (Flint, A.L. et al. 1996) that shows recharge increases on the northern end of Yucca Mountain (relative to the central and southern end) and below some of the major surface water drainages. Recharge (net infiltration) averages 4.5 3.2 mm/year, but on a year-to-year basis, ranges from zero in dry years to over 20 mm/year when average precipitation exceeds 300 mm (Flint, A.L. et al. 1996).

5.3.5.4.5 Discharge

Natural discharge does not occur within the immediate vicinity of Yucca Mountain. The nearest natural discharge areas connected to the saturated-zone flow system beneath Yucca Mountain are Franklin Lake playa (also known as Alkali Flat), and possibly the major springs at Furnace Creek Ranch and the valley floor of Death Valley. Although most models of the region (D'Agnese et al. 1997; Rice 1984; Czarnecki and Waddell 1984) require a groundwater flowpath from Yucca Mountain to Death Valley, Czarnecki and Wilson (1991) postulate that a groundwater flowpath from Yucca Mountain to Death Valley (by way of the Amargosa Desert and the Funeral Mountains) was unsubstantiated with (but not inconsistent with) available data. They suggest that groundwater from Yucca Mountain ultimately discharges at Franklin Lake playa through evapotranspiration.

Discharge through groundwater withdrawals occurs within the site-model domain in the Amargosa Desert for agricultural and domestic use. This discharge, estimated in the USGS regional flow model

at about 6,300 m^3/d (D'Agnese et al. 1997), occurs mostly in the southwestern corner of the site-model domain. Pumpage from wells UE-25 J-12 and UE-25 J-13, just east of Yucca Mountain, was about 1,200,00 m^3 from each well from 1985 to 1995 (Graves et al. 1997, pp. 40-41).

5.3.5.4.6 Conceptual Uncertainties

Although the flow system at Yucca Mountain has been divided into a series of aquifers and confining units, and the general direction of flow is known, substantial conceptual uncertainties remain. Fractures probably transmit most of the water through the aquifers, but at individual boreholes. Only a small percentage of the fractures contribute flow, and faults frequently dominate the flow. Conceptually, flow in the saturated zone occurs predominately in a few fractures, but as a practical matter, numerical models would have to simulate the system either as a porous-media continuum or as a finite number of discrete features. At the scale of the Yucca Mountain Site, it is not yet known whether fractures are so pervasive that the flow system can be simulated as a classical porous medium or if discrete features need to be simulated.

How the groundwater flow system behaves in the large and moderate hydraulic gradient areas is uncertain. This uncertainty translates into uncertainty about either external or internal boundary conditions in any numerical models of the site. The uncertainties about potential groundwater divides also would translate into uncertainty about the numerical-model boundary conditions.

Although recharge is related to climate, topography, flora, soils, and unsaturated-zone flow, major uncertainties exist about the interrelation and this uncertainty translates into a general conceptual uncertainty about recharge. Perhaps equally important, the time scale at which the flow system comes into equilibrium with climate, and hence recharge, is a conceptual uncertainty.

Tests to determine flow-system parameters historically have been conducted at scales that are much smaller than the scales that ultimately need to be simulated. How to translate data from a small test scale to a larger model scale is a major conceptual uncertainty.

No matter how well the saturated-zone flow system at Yucca Mountain and vicinity is described, uncertainty will always remain. The data needed to understand, quantify, and simulate the groundwater flow system (in order to decrease uncertainty) are: recharge; storage properties, transmissive properties, discharge, and hydraulic head. Each of these hydrologic parameters have some uncertainty associated with them, and such uncertainties are discussed in detail by Luckey et al. (1996, pp. 53-55).

5.3.5.5 Site Saturated-Zone Flow Modeling

The following discussion is primarily summarized from Czarnecki, Faunt et al. (1997), who provide a much more detailed discussion of a preliminary 3-D model of the site saturated-zone groundwater flow system.

The purposes for developing a model of the saturated zone of Yucca Mountain and vicinity are to:

- Estimate groundwater flow direction and magnitude from beneath the design repository area to the accessible environment.
- Characterize the complex 3-D behavior of flow through heterogeneous porous and fractured media.
- Provide a means to account for the distribution of groundwater temperature measured in wells within the model area.
- Identify the potential role of faults as barriers or conduits to groundwater flow.
- Provide a model of the flow system for subsequent flow, heat, and radionuclide-transport modeling.

The site saturated-zone flow model simulates steady-state groundwater conditions and covers an area of approximately 1,350 km^2 over a saturated thickness of about 1.5 km, delimited by a rectangular box 45 km long and 30 km wide (Figure 5.3-245). The domain was selected to be:

- Coincident with grid cells in the regional groundwater flow model (D'Agnese et al. 1997) such that the base of the site model was equivalent to the base of layer two of the regional model
- Sufficiently large to minimize the effects of boundary conditions on estimating permeability values at Yucca Mountain
- Sufficiently large to be able to assess groundwater flow at distances as great as 30 km downgradient from the design repository area
- Small enough to minimize the number of computational nodes used in the model
- Thick enough to include part of the regional Paleozoic carbonate aquifer
- Large enough to include well control in the Amargosa Desert at the southern end of the model

5.3.5.5.1 Previous Flow Modeling

Several numerical models of groundwater flow have been developed at various scales to simulate groundwater flow in the vicinity of Yucca Mountain. Regional-scale modeling of the Nevada Test Site and vicinity was done by Waddell (1982), Rice (1984), and Sinton (1987). The areal domain of these models was essentially identical (about 18,000 km^2). Waddell and Rice examined 2-D areal flow, whereas Sinton's model incorporated the third (depth) dimension in a quasi-3-D model. D'Agnese (1994) modeled an even larger area (34,141 km^2) of the Death Valley regional groundwater flow system in three dimensions, and used geoscientific information system analyses to obtain estimates of

recharge, discharge, and hydraulic conductivity. D'Agnese et al. (1997) advanced the original model of D'Agnese (1994) in a number of ways, including the application of nonlinear regression techniques using MODFLOWP (Hill, M.C. 1992), in which observed values of hydraulic head and spring flows were used to estimate areal recharge rates, hydraulic conductivities, and other selected model parameters. A flow model of an area approximating that of Waddell (1982) has been developed for characterizing tritium migration and assessing risk subsequent to underground nuclear testing at the Nevada Test Site (DOE 1997a).

Subregional modeling included the 2-D model of Czarnecki and Waddell (1984) in which parameter estimation was used to estimate values of transmissivity through most of the model area. A smaller 2-D model of groundwater flow in the immediate vicinity of Yucca Mountain was developed by G.E. Barr and Miller (1987) to examine the sensitivity of flow and transport to uncertainties in the existing data and the response to assumed catastrophic changes in hydraulic conductivity. Buscheck and Nitao (1992) discuss the possible effects of heat from the design repository on groundwater flow in the unsaturated and saturated zones. Other models, discussed in the following subsections, simulated the potential effects of altering current climate or hydrogeologic conditions on the existing groundwater flow system.

Czarnecki (1985) used the same finite-element mesh and a slightly modified distribution of transmissivity values to estimate the effect of increased recharge on water table altitude and groundwater flow direction. In that study, an assumed 100 percent increase in precipitation over present-day conditions resulted in simulated water-level rises of as much as 130 m near the potential repository area (Czarnecki 1985, p. 21).

A cross sectional model was developed by Haws (1990) along a flowpath constructed from the flow-field vectors of Czarnecki and Waddell (1984). Haws (1990) concludes that the large hydraulic gradient resulted from a low-conductivity zone north of Yucca Mountain and that upward leakage from the underlying carbonate aquifer provided a source of water to the volcanic aquifer.

Modeling by Carrigan et al. (1991) examined potential changes of the water table altitude at Yucca Mountain resulting from seismic events. The results of that modeling indicate that the consequence of a "likely" earthquake would be a 2 to 3 m excursion of a water table that is 500 m below land surface. For an "extraordinary" seismic event, the estimated water table excursion was less than 20 m.

Dressel (1992) used a two-layer, steady-state flow model to evaluate the effects of increased areal recharge, increased recharge along Fortymile Wash, and the emplacement of hydraulic barriers on the position of the existing water table. The maximum water table rise of 20 m above the base of the potential repository was produced by simulating a combination of increased recharge and placement of a hydraulic barrier downgradient from Yucca Mountain (Dressel 1992, p. 100).

Ahola and Sagar (1992) simulated the effects of intrusive volcanic dikes on the groundwater flow field in the vicinity of Yucca Mountain. These simulations indicated either little increase in the water table elevation, or a water table drop of several meters throughout parts of the region.

Czarnecki (1992) used an existing 2-D flow model to estimate the effects of increased simulated pumpage at wells UE-25 J-12 and UE-25 J-13 on water levels. Simulated pumping rates from these wells ranged from 2.3 to 88 L/s (36 to 1,390 gal/min), and simulated values of specific yield ranged from 0.001 to 0.01. Simulated water level declines over a period of 10 years ranged from 0.09 to 13.2 m at the pumping wells and from 0.08 to 12.7 m at an observation well about 800 m south of well UE-25 J-12. The maximum simulated declines resulted from pumping each well at the maximum rate, with a specific yield of 0.001 (Czarnecki 1992, p. 19).

5.3.5.5.2 Quality Assurance Considerations

Data used in the construction of the hydrogeologic framework model and the groundwater flow model were primarily developed from published sources or obtained from publicly available sources such as the USGS National Water Information System. Nearly all of these sources originated or were published before the implementation of the accepted YMP quality assurance program in 1989. Model construction and review, however, were performed in accordance with accepted YMP quality assurance procedures and USGS policy.

The only qualified data used in the model analysis are hydraulic-test data from wells UE-25 WT#10, UE-25 WT#12, and USW SD-7 (O'Brien 1997), hydrologic data for Fortymile Wash (Savard 1995a, 1996), and permeability data (Flint, L.E. and Flint 1990). All other data used in the model analysis are unqualified. Because of the preponderance of available unqualified hydraulic-head and hydraulic-test data, no saturated zone model of Yucca Mountain can be constructed using only qualified data.

5.3.5.5.3 Preliminary 3-D Groundwater Flow Model of the Site Saturated Zone

To accomplish the objectives of the site saturated-zone flow model, a hydrogeologic framework model was first developed and used to construct the numerical groundwater flow model. The simulation code used for the site saturated-zone modeling is FEHMN (Finite Element Heat Mass Nuclear) (Zyvoloski et al. 1995). In addition, optimal hydrologic-variable values used in the numerical model were obtained through nonlinear regression techniques using PEST (Parameter Estimation Software) (Watermark Computing 1994).

5.3.5.5.3.1 Hydrogeologic Framework Model

A detailed 3-D hydrogeologic framework model was developed to characterize the complex 3-D, heterogeneous, porous, and fractured media beneath Yucca Mountain. The framework model was developed so that it could be converted into a tetrahedral mesh for use in the FEHMN groundwater flow modeling code. As a result, the framework model has many simplifications that may limit its use for other applications.

The hydrogeologic framework model used in this model (sampled at 1,500 m) is only suitable for initial calibration of a preliminary flow model using a very coarse resolution. For example, the upper volcanic confining unit is much more extensive in the coarse hydrogeologic framework model than

in reality. Because of the coarse grid increment (1,500 m), offsets across faults are much less abrupt than in reality. Hence, this coarse hydrogeologic framework model should only be used to depict the extent or the boundaries of the hydrogeologic units in a very general sense.

Additional subdivision of regional hydrogeologic units was done on an area bounded by latitude 36° N and 37° 15'N and longitude 116° W and 117° W, resulting in the identification of 18 refined hydrogeologic units (Figure 5.3-246a). These units are described in detail in Subsection 5.3.1, and they form the basis for layers in the 3-D flow model. A subarea of this refined hydrogeologic framework model used in the site model is 1,350 km^2 and extends from 533,340 to 563,340 m (30 km west to east) and 4,046,782 to 4,091,782 m (45 km south to north), UTM Zone 11 (Figure 5.3-246b). The subarea grid was chosen to be coincident with the Death Valley regional flow model (D'Agnese et al. 1997). Sixteen of the 18 refined hydrogeologic units are present within the subarea selected for the site saturated-zone flow model (Table 5.3-69).

The area of the site-scale flow model is larger than that of the 3-D site geologic framework model (CRWMS M&O 1997b) developed to support the Yucca Mountain Site unsaturated-zone model, but extends into the saturated zone as well. Due to simplifications necessary for modeling, the geologic units are lumped into hydrogeologic units. The geologic units used in the site geologic framework model can be correlated with the hydrogeologic units used in the hydrogeologic framework model (Table 5.3-70).

Construction of a 3-D framework model is based on information shown in Figure 5.3-247 and involves the following steps:

- Geologic units are classified into hydrogeologic units based on their hydraulic properties and lateral extent.
- Digital evaluation model data are combined with hydrogeologic maps to provide a series of points in 3-D space locating outcrops of individual hydrogeologic units.
- Cross-sections and lithologic logs are used to locate hydrogeologic units in the subsurface.
- Maps and cross-sections are used to locate faults.
- Structure contour maps for each hydrogeologic unit are developed by interpolating both surface and subsurface positions with gridding software which incorporates unit offsets across faults.
- A hydrogeologic framework model is developed when the structure contour maps for the individual hydrogeologic units are combined, using appropriate stratigraphic principles to control their sequence, thickness, and lateral extent.
- The potentiometric surface is used to clip the framework model at the top.

A surface hydrogeology map (Figure 5.3-246a) provided the "ground truth" for other model-building data and was the foundation upon which the rest of the hydrogeologic framework model was constructed. The published cross-sections (Figure 5.3-248) used to construct the hydrogeologic framework model were all at a scale of 1:100,000 or larger. The detailed stratigraphy was simplified into the appropriate hydrogeologic units. The simplified cross-sections were then digitized, merged, scaled, warped to fit their digitized traces, and accurately placed in 3-D space. Records for wells in the area contain lithologic data that were used to help correlate between the cross-sections.

Information on faults include fault trace maps that show where faults intersect land surface, and faults shown on cross-sections. All faults with surface traces (1:100,000 scale), regardless of length, are included in the hydrogeologic framework model. Faults in the model area can dip at almost any angle, but most are high-angle faults. Given software constraints and the flow model resolution, the faulting in the area is greatly simplified. The major simplification is that nearly all faults are treated as vertical features. Where it was thought to be hydrologically important, thrust faults were represented by repeating hydrogeologic units. Because of the relatively large grid spacing (1,500 m), these simplifications are assumed to have minimal effect on flow model results; however, no sensitivity analyses regarding these simplifications were performed.

To construct the 3-D hydrogeologic framework model, the different hydrogeologic unit tops must be interpolated between the cross-sections and wells. A gridding system and fault-handling software package was used to interpolate the hydrogeologic structure contour maps. A grid increment of 1,500 m coincident with the regional groundwater flow model of D'Agnese et al. (1997) was used, resulting in grids with 21 columns and 31 rows. The resulting structure contour maps contain a series of undulating surfaces, broken by faults.

The quality of individual structure contour maps depends on the density of the data points used to define them. Some of these surfaces, such as the upper volcanic aquifer, were relatively well defined by more than one data set (derived from surface information, lithologic logs, and cross-sections). Others, especially the units that outcrop less frequently, were less well defined and were extrapolated from sparser data and published geophysical interpretations. A relative rating of data availability for each of the hydrogeologic units appears in Table 5.3-69; the rating does not imply accuracy regarding the extent and location of each unit. Although the rating is subjective, it is partially based on the number of data points used to define each hydrogeologic unit.

These structure contour maps were stacked in stratigraphic order to build the 3-D hydrogeologic framework model. Landmark's Stratamodel SGM (Stratigraphic Geocellular Modeling) is a geologic modeling software product that uses "geologic rules" to help define the geographic extent and intersection of surfaces.

The hydrogeologic framework model has volumetric units defined by the structure contour maps of individual hydrogeologic units (such as the upper volcanic aquifer). The hydrogeologic units are numbered consecutively in stratigraphic order from bottom to top, beginning with sequence number 2 (the SGM requires the specification of an arbitrary base unit, or sequence number 1, which is not used in the actual model). Although the cells have uniform horizontal dimensions throughout the hydrogeologic framework model, the user controls the number of cell layers. In many locations, these

hydrogeologic units have large thickness. To improve the vertical resolution, the units were subdivided into "layers," each with a maximum thickness of 125 m to minimize gridding and computational problems.

The SGM software allows each cell to reflect multiple attributes, including row number, column number, sequence number, layer number, and elevation. The cells were further attributed to reflect the hydrogeologic units. For groundwater flow modeling, the hydrogeologic framework model can be used to assign appropriate hydraulic property values. The available hydraulic property data used to assign hydraulic properties to each cell are summarized in Table 5.3-71. The geology and structure represented in the hydrogeologic framework model is shown in a fence diagram through the site model (Figure 5.3-248).

The resulting hydrogeologic framework model omits many small and even intermediate-scale features within the subsurface. It does, however, represent the large-scale features as accurately as possible given the grid resolution, and, therefore provides substantial constraints for model development. For the initial simulation of groundwater flow, this resolution is probably adequate. For future flow and transport simulations, the effect of the small-scale variations will need to be considered. For example, a hydrogeologic framework model with 250 m grid spacing would much more accurately represent the offsets across faults and a change in the geometry of the units which corresponds with the large hydraulic gradient.

Gridded values from the potentiometric surface were used in the construction of the hydrogeologic framework model. The potentiometric surface was used to clip the top of the hydrogeologic framework model. The hydrogeologic framework model was then translated into a finite-element mesh through the use of automated gridding software.

After constructing the 3-D hydrogeologic framework model, an automated finite-element mesh-generation computer program, GEOMESH (Gable et al. 1995), is used to construct a computational grid of tetrahedral elements in three dimensions. The hydrogeologic framework model is converted automatically for direct input into GEOMESH. Hydraulic-head observation nodal points are also added to the mesh for spatially correct calibration points. Because the altitude at which the hydraulic head measurement applies is uncertain, the nodal points are located at the midpoint of either the water column for uncased boreholes or the midpoint of a screened or packed-off interval within the borehole. The resulting finite-element mesh is shown in Figure 5.3-249.

5.3.5.5.3.2 FEHMN

To model the saturated-zone flow system at the site scale at Yucca Mountain, several simulation capabilities were considered important, including the ability to:

- Simulate 3-D transient groundwater flow and heat transport, including 3-D representation of spatially variable permeability, porosity, and thermal conductivity
- Allow specification of constant pressure, constant hydraulic head, constant fluid and heat flux boundary conditions

- Represent discontinuous, irregularly shaped 3-D hydrogeologic units
- Permit specification of dual permeability and porosity representing both fracture and matrix flow
- Represent hydraulic-head and temperature observation points where they occur in 3-D space
- Calibrate the model with respect to observations of hydraulic-head and temperature through the use of automated parameter estimation techniques
- Directly interface the resulting flow model with radionuclide transport models used in Performance Assessment of the Yucca Mountain Site

The FEHMN simulation code was selected because it possessed these capabilities when coupled with the mesh generation software, and with the model-independent parameter estimation software.

The FEHMN computer code is capable of simulating flow and transport through both the unsaturated and saturated zones. FEHMN is a nonisothermal, multiphase flow and transport code. It can simulate the flow of water and air, and the transport of heat and contaminants, in 2- and 3-D saturated or partially saturated, heterogeneous porous media. The code includes comprehensive reactive geochemistry and transport modules and a particle-tracking capability. Fractured media can be simulated using equivalent-continuum, discrete-fracture, dual-porosity, or dual-permeability approaches. FEHMN solves three conservation equations: conservation of total fluid mass (air and water); conservation of air; and conservation of solute (contaminant). The basic conservation equations, constitutive relations, and numerical methods are described in Zyvoloski (1983, 1986), Zyvoloski and Dash (1990), CRWMS M&O (1994), and Zyvoloski et al. (1995).

FEHMN requires information about air and water properties (including density, viscosity, enthalpy, and their derivatives) as functions of temperature (T) and pressure (P). Rational function approximations are used to estimate these variables in FEHMN, where the rational functions are a ratio of polynomials. For water, polynomial coefficients were obtained by fitting data from the National Bureau of Standards/Nuclear Regulatory Commission Steam Tables (Harr et al. 1984). The density of air is assumed to obey the ideal gas law. The code also requires information about the relation between values of relative permeability, capillary pressures, and air-water saturations. Several well known functions (i.e., Brooks-Corey; van Genuchten) are available to the user.

FEHMN uses a finite-element/finite-volume method to discretize the conservation equations to be solved. Newton-Raphson iteration is applied to the fully coupled system of equations. This system of equations is solved with multidegree of freedom preconditioned conjugate gradient methods, using generalized minimum residual acceleration techniques (Zyvoloski 1986).

5.3.5.5.3.3 PEST

The parameter estimation component of the model was achieved through the use of the model-independent parameter estimation software, PEST (Watermark Computing 1994). PEST uses

nonlinear least-squares regression to estimate parameters. The benefits of using nonlinear regression include:

- Expedited determination of best-fit parameter values
- Quantification of the quality of the calibration
- Estimates of the confidence limits on parameter estimates
- Identification of the correlation among parameters (Poeter and Hill 1997, p. 250)

PEST was selected because of the ability to couple it with FEHMN without significantly changing the FEHMN software. PEST is designed to be used with virtually any model, provided that one can identify model input files, model output files, commands that invoke the model, observation data, and model parameters. PEST was used to run FEHMN and to vary user-specified model parameters prior to each run such that the weighted sum of the differences between observed and simulated values of pressure, hydraulic head, or temperature is minimized using nonlinear regression. The optimization is accomplished using the Gauss-Marquardt-Levenberg method. PEST can generally estimate parameters using fewer model runs than other estimation methods, an advantage for large models with long run times (Watermark Computing 1994, pp. 1-4).

5.3.5.5.3.4 Model Assumptions

In the preliminary site saturated-zone model, the following assumptions are applicable:

A. The hydrogeologic framework is an appropriate description of the principal hydrogeologic units and faults.

B. Permeability is invariant within each hydrogeologic unit.

C. Groundwater flow occurs in three dimensions and within the rock mass (which includes both rock matrix and fractures).

D. Groundwater flow system is isothermal at 44°C (the effect of this assumption was tested by simulating system at 20°C).

E. Hydraulic heads of the potentiometric surface along the north, south, east, and west edges of the modeled area are an appropriate data set for specifying boundary conditions along the sides of the model.

F. The system is at steady-state so that groundwater flow into and out of the flow domain is invariant with time.

G. Volumes associated with the finite-element mesh are sufficiently large so as to exceed the representative elementary volume necessary to simulate fracture flow as porous-media flow.

H. A no-flow boundary at the base of the model approximates hydrologic conditions.

I. The large hydraulic gradient is part of the saturated zone and not an artifact of perched water occurrence.

J. Recharge is assumed to occur only at the top of the model along upper Fortymile Wash; all other nodes on the top of the model are specified as a no-flow boundary.

Assumptions (E) and (H) are not and have not been supported. Areally distributed recharge likely occurs in the vicinity of Yucca Mountain, in contrast to assumption (J). All three of these assumptions represent an expedient means to assign boundary conditions, which may affect model calibration.

5.3.5.5.3.5 Boundary Conditions

The conceptual model represented within the numerical groundwater flow model assumes that recharge occurs as throughflow from the northern, eastern, and western boundary of the model, simulated with constant heads, and by minor recharge in upper Fortymile Wash. Discharge occurs as throughflow at the southern end of the model, simulated with constant heads. Model analysis shows that most of the discharge occurs along the northeastern model boundary; however, that discharge is largely an artifact of the assumed hydraulic gradient and permeability of the hydrogeologic units that occur there. Constant-head values were obtained from existing potentiometric maps of the area, and were calculated by the model for distribution vertically along the sides. No flow is assumed through the bottom of the model. The large hydraulic gradient area is assumed to be part of the saturated zone (as opposed to being an artifact of perched water occurrence), and is assumed to be caused by a buried low-permeability fault. The moderate hydraulic gradient area also is assumed to result from a low-permeability fault (Solitario Canyon fault).

The water table forms the upper boundary of the flow model. The only recharge applied to this boundary was simulated at upper Fortymile Wash. The recharge rate was estimated by PEST during model calibration. Recharge from infiltration of rainfall was not directly applied to Yucca Mountain; however, the relative amount of such infiltration is believed to be small in relation to throughflow along the sides of the model. Pumpage, known to occur south of Yucca Mountain, was not included in this simulation.

5.3.5.5.3.6 Model Calibration

The model used an automated parameter estimation routine (PEST) to minimize the difference between 94 observations of hydraulic head and those simulated by FEHMN by adjusting selected permeability and flux parameters. The greatest improvements to the model came when: vertical low-permeability barriers were added corresponding to the Solitario Canyon fault and the downgradient side of the large hydraulic gradient; and the parameter for the permeability of the upper volcanic confining unit (Calico Hills Formation) was isolated and optimized. Optimal permeability estimates for the 16 hydrogeologic units lie between high and low values for the same units reported in the literature (Table 5.3-71).

Simulated hydraulic heads generally are consistent with the observed heads. Hydraulic-head residuals (the difference between observed and simulated values) ranged from 0.1 m less than observed to 95 m

greater than observed (Figure 5.3-250). The largest class (45 percent) of hydraulic-head residuals occurred between the range of -5 to +5 m (Figure 5.3-251). In general, the model fits the observations well in small-gradient areas and not as well in larger gradient areas. The worst fitting hydraulic-head observation resulted from the inability to accurately locate the low-permeability barrier used to produce the large hydraulic gradient because of the existing node density in the finite-element mesh.

Large discrepancies between observed and simulated vertical gradients occur in USW H-1, USW H-3, USW H-4, USW-H-5, UE-25b #1, and USW p#1. Simulated flow is either largely horizontal or downward at Yucca Mountain, in contrast to the conceptual model, which indicates the potential for upward flow. These discrepancies may have important ramifications should the model be used for transport simulations, and indicates the need for additional model calibration.

Comparisons of flux from the regional model showed almost twice the amount discharging from the southern end of the site model, and substantially different amounts for the north and east sides (Table 5.3-72). The major flux differences between the two models occur in the northeast corner where a large part of the recharge from the north is diverted east and discharges in part because of the interaction of the constant-head boundaries and the imposed east-west barrier needed to represent the large-hydraulic gradient.

5.3.5.5.3.7 Sources of Error

Model discretization is coarse and, as a result, causes incomplete definition of hydrogeologic units. The flexibility of the tetrahedral elements used to construct the finite-element computational grid provides greater accuracy in representing the individual hydrogeologic units than would be available at this resolution with less flexible gridding techniques, but problems still remain. For example, many of the faults are implicit in the grid through offset of units, hence the exact location for faults will always be approximate, regardless of grid spacing. However, explicit representation of selected faults could be achieved through explicit specification of the fault as a surface within the hydrogeologic framework model, which would cause it to be defined as a set of nodes within the subsequent finite-element mesh, complete with its own set of hydraulic properties. Larger problems are identifying those faults a) for which hydraulic properties are available or could be anticipated (very little is known), and b) that are most important to represent explicitly within the flow and transport model.

Permeability is known to vary spatially within individual hydrogeologic units. The assumption of uniform permeability within each unit is a simplification of a complex system. Small-scale variations in hydraulic head likely cannot be represented without greater resolution. Even with a more refined model, permeability data to support spatial variation of permeability would be lacking. Local areas of large permeability may exist that would explain very large estimates of permeability based on hydraulic tests (e.g., within the middle volcanic aquifer) that are not represented in the final model. Specification of fault and fracture zones from independent-geologic information may be the only way to incorporate spatial variations of permeability.

An average temperature for the entire saturated zone contained within the site model has not been calculated. Different specifications of average groundwater temperature (20°C and 44°C) have an appreciable effect on viscosity (1.002 centipoise for fresh water at 20°C; 0.6067 centipoise for fresh water at 44°C). A more appropriate way to address the issue of temperature effects on flow is to simulate coupled groundwater flow and heat transport with appropriate temperature and heat-flow boundary conditions. Temperature data collected from wells throughout the model area are available for use in calibration. Using temperature and hydraulic-head data in model calibration likely would better constrain simulated results.

Hydraulic-head boundary conditions are based on a process of extrapolation and interpolation of extant data. An artifact of that process is the resultant large fluxes that occur in the north-east part of the model. Furthermore, no vertical hydraulic-head data exist at the model boundaries making it difficult to verify the resulting hydraulic-head distribution. Possible inaccuracies of assigned hydraulic-head values at the side model boundaries are an important potential source of model error.

The steady-state assumption may be invalid in areas in which groundwater withdrawals are occurring. The hydraulic-head observation data span almost 50 years of record, which results in irregularities in the potentiometric surface. The slope of the potentiometric surface toward the southwest may be indicative of groundwater withdrawal which were not specified in the model. About 73.1 kg/s of discharge by pumping wells occurs in the Amargosa Desert in the southwest part of the model, but was not represented in the model.

No flow is specified along the base of the model. This omission may explain the error in simulating lower hydraulic-head values than those observed in the deeper observation points within the model. This error results in horizontal to vertically downward flow within the model, a condition unsupported by the hydraulic-head observations. However, observed upward flow cannot be confirmed with existing data.

Finally, the representation of the large hydraulic gradient remains inconclusive. By specifying an east-west oriented barrier to flow, the observed hydraulic-head data may be better matched, but the resulting flow field is difficult to reconcile. If a buried fault does exist and is a barrier to flow, no data are available to prove or disprove its existence. Furthermore, if the large hydraulic gradient is actually an artifact of perched water occurrence, then the resulting flow field would be considerably different.

5.3.6 Site Paleohydrology

5.3.6.1 Record of Unsaturated-Zone History

Deposits of secondary minerals (mainly low-temperature calcite and opal) in fractures and cavities mark pathways of past percolation of water through the unsaturated zone. These mineral coatings preserve a history of deposition that can be deciphered by isotopic dating. Morphologic relations as well as chemical and isotopic compositions of these secondary minerals provide an interpretable record of paleohydrologic conditions within Yucca Mountain.

Detailed studies of secondary minerals within the Exploratory Studies Facility (Paces, Marshall et al. 1997) have shown multiple episodes of mineral deposition. The earliest minerals are those of the vapor-phase crystallization stage, which occurred during initial cooling of the volcanic pile. In close association with these minerals are ^{13}C-enriched calcites which may have formed during a moderate-temperature alteration of the tuffs (Broxton et al. 1987) and appears to have occurred at the time of formation of the Timber Mountain Caldera at ~10.4 Ma (Bish and Aronson 1993). Results for a few fluid inclusion from early and deep calcites support an early mineralization at somewhat elevated temperatures (Roedder et al. 1994). The discussion of secondary minerals in the remainder of this section is for paragenetically younger minerals than these early-stage ones.

Most of the secondary mineralization within the unsaturated zone occurs as coatings on the footwalls of fractures or as accumulations on the floors of lithophysal cavities, although within the bedded tuffs, calcite occurs locally as a pore-filling cement. The preferred localization of secondary minerals demonstrates precipitation under water-unsaturated conditions, and results of dating of the minerals (Neymark et al. 1998) show that the undersaturated conditions, at least at the level of the Exploratory Studies Facility, has persisted for several million years.

Textural evidence further supports deposition of secondary minerals from aqueous solutions within the unsaturated zone (Paces, Neymark et al. 1996). Figure 5.3-252 shows schematically some of the diagnostic textures. Calcite typically forms equant, blocky prisms on high-angle fracture surfaces and unusual delicate bladed crystals in lithophysal cavities and in low-angle fractures. Opal typically forms water-clear solid-hemispherical bubbles, botryoidal bubbly masses or thin sheets coating calcite substrates. Opal commonly occurs at the tops of calcite blades. Both opal and calcite are finely layered (micron to submicron), and commonly intimately intergrown. Outermost surfaces of both minerals show little evidence of dissolution, although scattered patches of basal porous zones may be related to dissolution of early-formed calcite blades. Textural evidence indicates that the deposition of low-temperature minerals occurred under unsaturated hydrologic conditions with no indication of even local zones of water ponding. Open space appears to be required for mineral deposition, and the occurrence of secondary mineral coatings on fracture foot-walls and cavity floors indicates the strong influence of gravity on water percolating through the unsaturated zone. Bladed-calcite textures, along with the occurrence of opal at the topographically highest sites, indicate mineral growth preferentially on tips and edges rather than on faces. Mineral textures require solutions to transport ions to crystal tips where conditions of mineral saturation or supersaturation are attained. A depositional scenario consistent with observed textures involves water moving down fracture surfaces as sheets or films that drip into cavities where interactions between the liquid and an independently migrating gas phase result in the slow-growing secondary mineral deposits.

Ages of the low-temperature minerals have been investigated using radiocarbon (calcite), $^{230}Th/U$ (calcite and opal), and U/Pb (opal) methods (Paces, Neymark et al. 1996; Neymark et al. 1998). Sampling for the first two techniques emphasized the outermost growth surfaces in order to determine ages of the youngest depositional events. Most ^{14}C measurements indicate at least small amounts of modern carbon resulting in a range of ages from 44 to 16 ka with the greatest number of ages distributed between 38 and 28 ka. Multiple subsamples of outermost calcite layers from the same occurrence typically show discordant ^{14}C ages well outside of analytical error. Calcite and opal dating show similarly wide ranges of ages for subsamples from the same surface; however, in the case of

^{230}Th/U, calculated ages range from 28 ka to greater than 500 ka, with most analyses between 50 and 400 ka. Like radiocarbon results, calcite and opal ages for outermost materials are not in secular equilibrium, implying that growth surfaces contain components that are less than 500 ka. No systematic age differences are observed between calcite and opal regardless of occurrence, indicating that lithophysal cavities are as likely to receive percolating water as are fracture pathways. Both radiocarbon and ^{230}Th/U ages from secondary minerals near or within the discrete zones of elevated ^{36}Cl/Cl showed similar distributions relative to those in zones with background levels of ^{36}Cl/Cl. Calculated initial ^{234}U/^{238}U ratios representing the isotopic composition of percolating waters at the time of mineral deposition range from 4 to 9.5 for samples younger than 100 ka. Like the ages, these wide ranges of initial ^{234}U/^{238}U are observed from subsamples from the same occurrence, indicating that minerals were deposited from solutions that were highly variable on a small spatial scale. The highest initial ratios, the youngest ages and the thickest Exploratory Studies Facility mineral coatings are observed within the potential repository horizon. In addition, the data show a distinct negative covariance between age and calculated initial ^{234}U/^{238}U that is not readily explained by conventional episodic/instantaneousdepositional models.

The data are consistent with a depositional model in which very thin layers are added more or less continuously at a scale much finer than sampling techniques can resolve. This hypothetical model is formulated on the assumptions that deposition proceeds at very slow but uniform rates such that all samples of finite size will integrate multiple, infinitely thin layers with continuously varying ages. Resulting isotopic measurements will represent mixtures of materials intermediate between the oldest and the youngest layers. Calculated ^{230}Th/U ages and initial ^{234}U/^{238}U for hypothetical samples follow trends that approximate the observations. In addition, the model predicts:

- Age discordance between isotopic systems with the amount and direction of discordance dependent on the respective half-lives of the isotopic systems.
- Samples will always have ages less than the theoretical maximum limit of the isotopic system if even the thinnest layers of modern materials are present on the exterior of the sample.
- The youngest ages will be from the thinnest samples of the outermost mineral coatings.

An important condition of this model is that deposition must occur very slowly and over very long periods of time. Observed rates of deposition are in accord with this condition. Rates of deposition (1 to 10 mm of mineral per million years) inferred from outermost materials are similar to those calculated assuming that the total thickness of mineral coatings accumulated since the time of tuff emplacement. Ages determined by ^{230}Th/U are more representative of true average ages than are radiocarbon ages due to the longer half-life of ^{230}Th, but young radiocarbon ages document the addition of calcite younger than about 15 to 20 ka on outermost surfaces. Interpretation of geochronological data in this scenario also implies that deposition of a finite amount of material occurred over a longer period of time than represented by the numerical age alone and as a result, lower fluxes can account for the same thickness of deposits.

Stable and radiogenic isotopes incorporated into calcite and opal at the time of formation reflect the isotopic compositions of percolating fluids, which, in turn, relate to sources and climate conditions

prevalent at the time of infiltration. Infiltrating solutions acquire carbon and strontium isotopic signatures through interactions with soils. The isotopic composition of carbon is controlled by the plant community at the surface (Cerling et al. 1989; Quade, Cerling et al. 1989) and remains largely unmodified during percolation, whereas the composition of strontium is continually modified by rock-water interaction (see Subsection 5.3.4.2.4). The isotopic composition of oxygen in secondary minerals is dependent on the composition of meteoric precipitation and the temperature and chemical composition of the mineral. Evapotranspiration of soil waters by plants has been shown to be nonfractionating (McConnaughey et al. 1995) and, therefore, infiltrating waters have the $\delta^{18}O$ of time-averaged local meteoric water. The $\delta^{18}O$ values in secondary calcite are therefore related to meteoric precipitation by known, temperature-dependent fractionation factors. Figure 5.3-253 shows that the oxygen isotopic composition for calcites from the unsaturated zone can be related to a geothermal gradient of approximately 34°C/km using a reasonable average surface temperature and initial isotopic composition. This agrees well with that reported by direct measurement (Sass, Lachenbruch et al. 1980) and that proposed on the basis of a much smaller data set by Szabo and Kyser (1990).

Values for $\delta^{13}C$ and $^{87}Sr/^{86}Sr$ for outermost calcites vary over limited ranges reflecting the compositions of modern calcite-rich soils and minor modification of the isotopic composition of strontium by rock-water interaction during infiltration. At the scale sampled, the isotopic data show no evidence of a cyclic pattern within the younger parts of the deposits that could be related to climate variation over the last 500 ka. The strontium data do show a variation with depth within individual crystals such that deeper portions are less radiogenic (Paces, Neymark et al. 1996). The increased radiogenic content as a function of time reflects the addition of ^{87}Sr through radioactive decay of ^{87}Rb in the rock with which percolation water is reacting.

Initial $^{234}U/^{238}U$ ratios show modifications from compositions typical of surface waters and pedogenic sources as a consequence of percolation. Initial $^{234}U/^{238}U$ in the Exploratory Studies Facility materials increase with depth to values as high as 8 to 9.5 at the repository horizon. Wholesale dissolution or leaching of U would not result in elevated $^{234}U/^{238}U$. Rather, the increases in excess ^{234}U fit a model of preferential dissolution of uranium isotopes during downward percolation of solutions because ^{234}U is located in more readily accessible sites than ^{238}U due to alpha-recoil decay processes. Perched and shallow saturated-zone waters beneath Yucca Mountain are consistent with the large $^{234}U/^{238}U$ ratios found in secondary mineral deposits. The high $^{234}U/^{238}U$ ratios observed in Exploratory Studies Facility minerals indicate that rates of water percolation are low enough to allow selective dissolution of ^{234}U and to allow replacement of ^{234}U to percolation-accessible sites by further decay of ^{238}U so that high ratios are maintained with time.

The mineral record can be used to estimate past water flux through the unsaturated zone if the distribution of secondary minerals is known in space and time, and if the volumes of water required to account for the mineral deposits can be estimated. Preliminary estimates of the abundance of secondary minerals have been made from ESF observations. Hydrochemical compositions of surface runoff are nearly saturated with respect to calcite and opal, and are assumed to represent shallow, infiltrating solutions. Assuming all Ca^{+2} is extracted from these waters en route to the saturated-zone, a minimum value of flux of about 2 mm/year is obtained as an average over the last 12.7 Ma.

Although similar values have recently been determined for current flux from other data sets, this value is considered to be provisional.

Surface records from the Yucca Mountain area indicate that the regional climate over the last million years was characterized by wetter and cooler conditions relative to the present for as much as 80 percent of the time (Forester et al. 1996; Winograd, Landwehr et al. 1997). However, there is no indication in the subsurface that deposition rates of calcite and opal varied greatly during this interval even though higher water tables throughout the region reflect greater recharge during pluvial periods. Evidence for nondeposition or substantial dissolution of youngest calcite or opal surfaces that might be expected if infiltrating waters were significantly more dilute is also lacking. Instead, observed ^{230}Th/U ages are more-or-less evenly distributed throughout the last 400 ka with the possibility that depositional hiatuses may have occurred during interglacials at about 100 and 200 ka. The isotopic age distributions are consistent with a model in which the rock mass remained transmissive in areas represented by secondary mineral deposits over much of the last several hundred thousand years. These observations suggest that percolation (and mineral deposition) through most of the Topopah Spring Tuff may have been buffered from variations in effective moisture except in the zones of highest flux (highly transmissive pathways associated with through-going structures).

5.3.6.2 Past Water Table Elevations at Yucca Mountain

Several lines of evidence suggest that the elevation of the water table beneath Yucca Mountain has never been more than 100 m greater than exist today.

Bish and Vaniman (1985) noted that glass within the nonwelded tuffs occurs both above and below the welded zone in the Topopah Spring Member of the Paintbrush Tuff. Bish and Vaniman (1985) state "the lower nonwelded vitric zone thins and disappears to the east where stratigraphic dip and structural displacements bring the basal Topopah Spring Member glassy zone closer to the static water level. The vitric nonwelded material may have important paleohydrologic significance because the preservation of open shards and pumice made of nonwelded glass is rare below past water levels (Hoover 1968)." The minor glass that does exist below past zones of saturation may do so because of armoring by alteration products (Bish and Chipera 1989). Where present, the base of the lower nonwelded vitric tuff occurs about 80 to 100 m above the present water table. On the basis of mineralogy, Bish and Vaniman (1985) concluded, "The only apparent change in phase assemblage near the water table in Yucca Mountain is the alteration of vitric tuff of Calico Hills and lower Topopah Spring Member." The observations by Bish and Vaniman (1985) suggest that past water levels beneath Yucca Mountain may never have been more than 100 m higher than the modern water levels.

Levy (1991) re-evaluated the base of the vitric tuff as a paleohydrologic indicator in the light of structures and uplift that occurred after devitrification. She noted that some of the devitrified tuff, which occurred farthest above the current water table, was not coplanar with the water table, and had apparently formed prior to tilting of the tuffs. Thus, some of the devitrification must have occurred more than 13 Ma. Other past water table elevations must be younger and post-date major tectonic events. The occurrence of such a stand probably cannot be dated, but might have occurred several million to several thousand years ago. Even in the absence of a date for the alteration, the base of the

nonwelded vitric tuff shown on Bish and Vaniman's cross-sections (1985) is tens of meters to more than 100 m below the base of the potential repository horizon. Thus, a possible maximum (though undated) water table stand is significantly lower than the repository horizon under consideration. If the structural interpretations proposed by Levy (1991) are taken into account, the maximum rise in the water table, for a long enough period of time to cause devitrification, is 60 m. Levy (1991), citing Dibble and Tiller (1981), notes that the time required to form zeolites from glass is on the order of 10^4 years, and thus, larger short-term fluctuations are not precluded by the elevation of devitrified tuff.

The analysis by Levy (1991) also noted that in the central and southern part of the exploration block there was almost no drill-core information. Drill cores USW SD-12 and USW SD-7 provide important data in this area. The new data generally support the previous findings, but do not provide an unequivocal limit of water table rise (see Subsection 6.1.8.4).

The studies of Trench 14 and Busted Butte identified several contrasts between calcites deposited from groundwater of the saturated zone and those deposited by waters percolating down from the soil zone. Thus calcites, which are common in drill core and the Exploratory Studies Facility, provide a potential record of vadose and saturated zone hydrology. Preliminary studies (Marshall, Whelan et al. 1992; Peterman, Stuckless et al. 1992; Marshall, Peterman, Stuckless 1993) revealed a typically more radiogenic strontium for calcites recovered more than 85 m above the current water table, although some overlap exists, particularly for the earlier calcites from the unsaturated zone. Carbon isotopic compositions (Whelan and Stuckless 1992; Marshall, Stuckless et al. 1993) generally support this same conclusion by yielding a heavier isotopic composition in the saturated zone and up to 85 m above the water table. This relative heavy isotopic composition ($\delta\ ^{13}C \approx 0$) is attributed to a contribution from the underlying carbonate aquifer (Whelan and Stuckless 1992). The lighter compositions ($\delta\ ^{13}C < -6$) in the vadose zone are attributed to interaction with soil carbon in the first few meters of infiltration (Vaniman and Whelan 1994). The carbon data also show a few isotopically light samples up to 300 m below the current water table which could indicate a lower water table (Whelan and Stuckless 1992; Whelan, Moscati et al. 1998) or a different source for carbon in the saturated zone at some time in the past.

The carbon isotopic data show a few anomalously heavy samples from USW G-4, as much as 500 m above the water table; however, these samples are not anomalous with respect to strontium isotopes or fluorescence (Whelan and Stuckless 1992). These authors note that calcite from the unsaturated zone fluoresces bright white to pale purple, whereas calcites from the saturated zone fluoresce in hues of orange to pink. In addition, saturated zone calcites exhibit a strong phosphorescence. The heavy carbon so far above the saturated zone in USW G-4 may well represent very early (Timber Mountain alteration event) calcite. The chemical anomalies that distinguish shallow from deep calcites in USW G-2 and USW GU-3/G-3, such as low abundances of transition elements and prominent negative Ce and Eu anomalies in the unsaturated zone calcites (Vaniman 1994a), are not as definitive in USW G-4 (Vaniman 1994b). The change in calcite chemistry appears to be near the present water table elevation. In her study of minerals in fractures in the same drillhole, Carlos (1987) notes that mineralogy and mineral morphology do not change at or near the current water table, but rather about 118 m above it.

Studies of fluid inclusions within the calcites from drill core support a low-temperature origin for these minerals (Roedder et al. 1994). Most inclusions are fluid only, thereby demonstrating a formation temperature less than 100°C. The few two-phase inclusions contain either CO_2 and/or CH_4. The presence of the latter suggests very low, "perhaps near-surface temperatures" (Roedder et al. 1994). Crushing of the fluid inclusions shows that they formed at low (essentially atmospheric) pressures. Had they formed under water saturated conditions with saturation existing all the way to the earth's surface, the overlying water column would have generated as much as 30 atmospheres of pressure for the deepest sample investigated.

The temperature and pressure data argue against any kind of upwelling, spring-like origin for the calcites. Upwelling waters would not be expected to equilibrate with the surrounding rock, thereby preserving a record of the geothermal gradient. This is demonstrated by the abundant warm springs in Ash Meadows (Dudley and Larson 1976) and Oasis Valley (White 1979). Furthermore, any spring-like origin would result in deep calcites forming at high pressures, which is not observed. Thus all of the data are consistent with a vadose zone origin for the calcites found above the current water table except for some which occur within the last 85 m of the unsaturated zone.

Like the mineralogic record of alteration, the calcite data do not preclude higher stands of the water table in the past because calcite may not have been precipitated. In fact, Whelan, Vaniman et al. (1994) report microscopic dissolution textures for some calcites from the vadose zone which demonstrates that infiltrating water may be either saturated or undersaturated with respect to calcite at different times. The same is likely true of the saturated zone as well.

Czarnecki (1985) addressed the issue of possible past, and by implication, future pluvial-related water table rises beneath Yucca Mountain with a two-dimensional finite element groundwater flow model of the region, together with assumptions about pluvial recharge on, and underflow into and from the region. He calculated a maximum increase in water table altitude of about 130 m beneath the site of the potential repository. This number agrees well with that calculated by Quade, Mifflin et al. (1995) on the basis of past discharge deposits. They conclude that the maximum increase in water table elevation under Yucca Mountain in response to climate change was $\leq$115 m since the last full glacial period. Thus multiple lines of evidence suggest that 130 m rise in water table elevation is a good estimate of the maximum change due to climate change. Inundation of any part of the planned repository would require a water table rise of more than 200 m. As noted earlier, the rise in water table under Yucca Mountain also suggests a past shortening of the present-day groundwater flowpaths from Yucca Mountain to discharge areas in the southern Amargosa Desert. The closest of the past discharge areas is about 15 km southwest of the site, in southern Crater Flat (see Subsection 5.2.6).

The analysis by Czarnecki (1985) may be considered conservative for several reasons. A 100 percent increase in precipitation during pluvial periods was assumed, resulting in a 15-fold increase in recharge from the modern rate (0.5 mm/year) to about 8 mm/year at Yucca Mountain. Czarnecki (1985) acknowledges that this recharge value, which was derived from the empirical Maxey-Eakin water budget method, probably is high. Half of the calculated recharge flux in the model was applied directly east of the potential repository site, along a segment of Fortymile Wash. This is consistent with the proposal of Claassen (1985) that Fortymile Wash is an area of recharge under wetter climatic conditions. The flux at Fortymile Wash causes about three-quarters of the computed water table rise

of 130 m. Czarnecki (1985) notes, however, that under a 100 percent increase in precipitation, large quantities of runoff might flow away from the area down Fortymile Wash and other drainage ways. This would have the effect of decreasing the effective groundwater recharge to less than the calculated values (Czarnecki 1985).

Other considerations were noted by Czarnecki (1985) that might have resulted in a greater simulated water table rise. Recharge into Fortymile Wash was limited to the main-stream channel near Yucca Mountain. If recharge had been included along the full length of Fortymile Wash and its distributaries, which extend beyond the town of Amargosa Valley, then a greater water table rise would have been simulated. Simulated development of discharge areas southeast of the town of Amargosa Valley helped to limit the water table rise beneath the primary repository area. If this discharge were decreased because of the possible existence of marsh deposits or eolian silts, or because of a greatly decreased ratio of vertical hydraulic conductivity to horizontal hydraulic conductivity, then the simulated water table rise might be greater. The model used by Czarnecki (1985) derived its parameters from a model presented in Czarnecki and Waddell (1984), which assumed that the groundwater system was in steady-state conditions. If the groundwater system was still equilibrating to recharge that may have occurred 10,000 to 20,000 years before present, then the values of transmissivity used in the model in Czarnecki (1985) might be too high. Large transmissivities used in recharge simulations would produce less water table rise than if smaller transmissivities were used.

Recent modeling (D'Agnese, F.A. et al., *Simulated Effects of Climate Change on the Death Valley Regional Ground-Water Flow System, Nevada and California*, Milestone Report SP23OM3, DTN GS970708312144.003, (TBV) Water Resources Investigations Report 98-4041, unpublished) using a new regional flow model (D'Agnese et al. 1997), and climates estimated to exist during the glacial maximum at 21 ka and under a climate with double the current atmospheric CO_2, produced significantly less rise in the elevation of the water table. The climate at 21 ka is estimated to have increased recharge five-fold which then leads to an increase in water table elevation beneath the potential repository site of 60 m. The doubling of atmospheric CO_2 leads to a 1.5 times increase in recharge and a water table rise beneath the potential repository site of 50 m.

The convergence of the calcite data, the mineralogic data, the implied depth of the water table from past discharge deposits and the theoretical maximum water table rise from the hydrologic models, strongly suggest that the saturated zone has not risen to the level of the potential repository and is not likely to do so in the future. The presence of calcite within the unsaturated zone demonstrates that water has percolated through the potential repository horizon and some of that infiltration has been as recent as latest Pleistocene and possibly even Holocene. The amount of water and exact timing relative to past climates is not currently known.

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