3.1 DESCRIPTION OF INDIVIDUAL SYSTEMS AND CHARACTERISTICS OF THE SITE

This section describes the natural systems of the geologic setting in the region of the site, specifically for the controlled area. Where conditions outside the controlled area affect waste isolation, the relevant and material information on these conditions will be included. The controlled area is defined in 10 CFR 60, *Disposal of High-Level Radioactive Wastes in Geologic Repositories*, Section 2, as "a surface location, to be marked by suitable monuments, extending horizontally no more than 5 km in any direction from the outer boundary of the underground facility, and the underlying subsurface, which area has been committed to use as a geologic repository and from which incompatible activities would be restricted following permanent closure." The natural systems of the geologic setting in the region consist of: geologic system, hydrologic system, geochemical system, and climatological and meteorological system.

3.1.0 Introduction

Yucca Mountain is a narrow ridge located near the southern end of the Great Basin physiographic province in the Southwest Nevada Volcanic Field (SNVF). The ridge crest trends north-south and extends approximately 20 km from the southern margin of the Timber Mountain caldera complex. The mountain is composed of Miocene volcanic rocks erupted from Timber Mountain between 7 and 16 million years (Ma) ago. The silicic volcanic tuffs that comprise Yucca Mountain and much of the structural deformation that has isolated the Yucca Mountain ridgeline are typical of mid-Tertiary basin and range extensional tectonics in southern Nevada.

3.1.1 Geologic System

The geologic system for Yucca Mountain Region (YMR) includes the geomorphology and topography, stratigraphy and lithology, structural and tectonic history and regime, seismology, natural resources, and the results of geophysical investigations.

In general terms, the geology of the region is characterized by a thick section of Precambrian and Paleozoic sedimentary rocks overlain by a sequence of Tertiary siliciclastic volcanic rocks. The older rocks have been faulted and folded in response to compressional tectonic events and the entire stratigraphic section deformed by extensional basin-and-range tectonics. Uplifted ranges, such as Yucca Mountain, are separated by basins partially filled with alluvial deposits. Silicic volcanism waned approximately 6 Ma; however, smaller basaltic eruptive centers have formed in the basins adjacent to Yucca Mountain as recently as 20,000 years ago. Geomorphic evidence suggests that, aside from the basaltic eruptions, the land surface around Yucca Mountain has been stable for at least the last million years. Neither debris transport nor earthquakes have dislodged significant amounts of material even from the steeper slopes, and erosion rates are low, even in the channels.

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There are no perennial streams or natural bodies of surface water on or adjacent to Yucca Mountain. The major drainages, 40-Mile Wash on the east, Solitario Canyon on the west, and tributary drainages, primarily along the east flank of the mountain, only flow briefly, immediately following rainstorms.

The water table under Yucca Mountain is deep, approximately 2,000 feet below the crest of the mountain. While the flow processes through the unsaturated zone in the mountain are complex, the semi-arid climate is characterized by low rainfall and high evaporation rates, and infiltration through the mountain is believed to be minimal. Groundwater recharge from precipitation on Timber Mountain flows southward past Yucca Mountain under a very low hydraulic gradient. This water discharges, in part, through springs and pools in Ash Meadows, although some flow may reach Death Valley.

3.1.1.1 Regional Geology

The YMR, as defined in the *Site Characterization Plan* (SCP), DOE/RW-0199, consists of the southern Great Basin in southern Nevada and an adjacent area in California. To the east and southeast, the Great Basin is marked geologically by the margins of the Colorado Plateau in Utah and Arizona. To the west and south, the region is bounded by the Sierra Nevada and the Transverse Ranges in California. The northern edge of the Great Basin is covered by the Snake River Plain and the flood basalts of the Columbia Plateau in Oregon and Idaho.

Typical Great Basin topography consists of north-south trending mountain ranges separating narrow structural valleys with internal drainages. The only external drainage from the southern Great Basin is the Colorado River, which flows along the margin of the Colorado Plateau, topographically and geographically isolated from Yucca Mountain. The basin and range topography is less distinct within the Walker Lane, a northwest-trending zone to the west of Yucca Mountain where the ranges are lower and less distinctly aligned.

3.1.1.1.1 Geomorphology and Topographic Features in the Region of the Site

[The skeleton text for this section will be provided in future revisions of the Annotated Outline.]

3.1.1.1.2 Stratigraphy and Lithology of the Region

The stratigraphy of the southern Great Basin comprises highly variable lithologies, ranging in age from Precambrian to Holocene. The complete stratigraphic sequence with formation names is illustrated in Figure 3.1.1.1.2-1. These rocks fall into eight general groups (based on similarities in age, lithology, and history) that are described individually in the following sections.

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3.1.1.1.2.1 Older Precambrian Crystalline Rocks

Older Precambrian metamorphic and younger Precambrian igneous rocks are exposed extensively in eastern Clark County and southeastern Lincoln County (Stewart, 1980a). According to Longwell et al. (1965), the Precambrian crystalline rocks of Clark County, Nevada are exposed in and south of the Colorado River Trough, and are a "complex of schist, gneiss, and coarse-grained igneous rocks." Tschanz and Pampeyan (1970) report exposures of Precambrian granite, pegmatite, amphibolite, and gneiss in three places in southern Lincoln County.

Schist, gneiss, and gneissic quartz monzonite are exposed in the Bullfrog Hills and the Trappman Hills of southern Nye County (Figure 3.1.1.1.2.1-1). These rocks may be as young as Late Proterozoic (Stewart, 1980a; Maldonado, 1990). The exposures in the Bullfrog Hills are mylonitized (Maldonado, 1990). The Nye County exposures also are the northernmost outcrops of Precambrian crystalline rocks in the southern Great Basin. Gold deposits of the Bullfrog Mining district are primarily located in Tertiary pyroclastic rocks overlaying these crystalline rocks (Cornwall, 1972).

3.1.1.1.2.2 Precambrian and Lower Cambrian Sedimentary Rocks

Sedimentary rocks of late Precambrian and early Cambrian age are present throughout the southern Great Basin. They consist of a westward-thickening prism of shallow marine quartzite, siltstone, shale, and carbonate which is interpreted as a rifted continental margin miogeosyncline (Figure 3.1.1.1.2.2-1). Stewart (1980a) divided the prism into two depositional systems in Nevada: an eastern quartzite and siltstone system, and a western siltstone, carbonate, and quartzite province. Stewart (1980a) placed the boundary between the two systems near the eastern border of southern Nye County, trending north-south.

Rocks of the quartzite and siltstone system typically consist of fine to medium-grained quartzite and sandstone units from 30 to 1,220 m thick, separated by units of siltstone and very fine to fine-grained quartzite from 15 to 300 m thick. Conglomerates, contain pebbles of quartz and quartzite, and are sparse in this system. Limestone and dolomite occur in layers ranging from less than 1 to 100 m thick. Individual units of quartzite, siltstone, or carbonate rock are laterally extensive along the miogeosyncline, but are less persistent and change facies of the miogeosyncline. This system has been interpreted as deposited in a shallow marine environment (Stewart, 1980a).

Rocks of the siltstone, carbonate, and quartzite system are finer-grained, richer in carbonate, and thicker than those of the quartzite and siltstone system. These rocks consist mostly of shallow-water sediments including siltstone, carbonate, and fine-grained quartzite. Fossils are relatively abundant in the Lower Cambrian strata of this system, including trilobites, archeocyathids, pelecypods, echinoderms, pelmatozoan debris, and algae. Fossils are rare in correlative rocks of the quartzite and siltstone province to the east (Stewart, 1980). Winograd and Thordarson (1975), consider all clastic sedimentary rocks of this age (Late Precambrian through Early Cambrian) in the Great Basin as aquitards. (Figure 3.1.1.1.2-1)

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Stratigraphic formations include the Wood Canyon Formation, of Precambrian age, and the Prospect Mountain Quartzite and Zabriskie Quartzite, both of Cambrian age.

3.1.1.1.2.3 Middle Cambrian through Devonian Sedimentary Rocks

Middle Cambrian through Devonian-age rocks are exposed throughout the southern Great Basin, although Silurian-age rocks are restricted to Lincoln and Clark Counties. Rocks of this age group consist of carbonate and shale with interbedded quartzite and sandstone and are interpreted as the deposits of a passive continental marginal miogeosyncline. Total thickness of this section ranges from up to 500 m in western Utah, to at least 6100 m in central Nevada (Stewart, 1980a). Strata of middle Cambrian through Devonian age are known as the Lower Carbonate Aquifer (Winograd and Thordarson, 1975) (Figure 3.1.1.1.2-1).

3.1.1.1.2.4 Mississippian through Permian Sedimentary Rocks

The Antler Orogeny created a marked change in the depositional environment in southern Nevada in late Devonian and early Mississippian time. The orogeny formed a north-northeast trending highland, adjacent to the Roberts Mountains Thrust, which extended from west of the site through northern Nevada. Large volumes of sediment eroded into a foreland basin in the eastern half of the Great Basin. These rocks consist of thick flysch deposits adjacent to the Antler Highland and shallow-water shelf carbonates farther east (Figure 3.1.1.1.2.4-1). This pattern of sedimentation continued through Permian time, although the volume of clastic sediment shed off the Antler Highland declined as the orogeny waned. Winograd and Thordarson (1975) considered Mississippian and Pennsylvanian-Permian rocks in the southern Great Basin as the Upper Clastic Aquitard and the Upper Carbonate Aquifer, respectively (Figure 3.1.1.1.2-1).

3.1.1.1.2.5 Mesozoic Rocks

Mesozoic sedimentary rocks are present in the region only in Clark County, where they consist of continental and marine sandstone, siltstone, and limestone of the Triassic and Jurassic Aztec Sandstone, Chinle Formation, and Moenkopi Formation. Tschanz and Pampeyan (1970) report the presence of cobble conglomerate deposits, which they informally call "older clastic rocks," in the Spotted Pintwater, and Northern Desert Ranges of southwestern Lincoln County. These "older clastic rocks" may range in age from Cretaceous to Oligocene.

Approximately 30 separate, Mesozoic to Tertiary, granitic plutons are exposed in Esmeralda County, west of Yucca Mountain. These plutons range in size from less than one square-km to the 1,000 square-km area of the Inyo Batholith (Albers and Stewart, 1972).

The absence of Mesozoic sedimentary rocks suggests that throughout much of the Mesozoic era, the southern Great Basin was an emergent area and no deposition occurred over much of the southern Great Basin. Instead, the Mesozoic Era was characterized in the

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region by the development of complex fold and thrust structures, such as the Keystone Thrust (Figure 3.1.1.1.2.5-1) (Stewart, 1980a).

3.1.1.1.2.6 Tertiary Sedimentary Rocks

Tertiary sedimentary rocks, such as the Esmeralda Formation and the Horse Spring Formation, crop-out throughout the southern Great Basin, and consist of poorly to moderately consolidated alluvial deposits and fresh water limestones (Jayko, 1990; Taylor, 1990). Thicknesses of these deposits are variable but can reach 1,000 m. These rocks are commonly found interbedded with volcanic deposits, and are interpreted as early, extensional, basin-fill deposits (Stewart, 1980a; Stewart and Diamond, 1990).

3.1.1.1.2.7 Tertiary and Quaternary Igneous Rocks

The most prevalent Tertiary igneous rocks in the southern Great Basin are pyroclastic deposits of rhyolitic to trachytic composition, which were deposited before approximately 6 Ma from calderas located north of 37° north latitude. Yucca Mountain is located on the southern flank of a cluster of four calderas in the SNVF, which erupted between approximately 7 and 16 Ma (Figure 3.1.1.1.2.7-1). The deposits from these eruptions are a complex mixture of pyroclastic flow and fall deposits, epiclastic deposits, and subsidiary lavas, and are approximately 3,050 m thick at the Yucca Mountain Site (Stewart, 1980a; Byers et al., 1976a, 1989; Best et al., 1989; Robinson, 1985).

Following the cessation of large-scale pyroclastic eruptions at approximately 6 Ma, the dominant form of volcanism in the southern Great Basin has been scattered, small volume, basaltic or bimodal basaltic-andesitic lava and scoria eruptions. These may have occurred as recently as 16,000 years ago (ka), both north and south of 37° north latitude, scattered throughout the region, including within the SNVF (Best et al., 1989; Wells et al., 1990; Byers et al., 1976a).

3.1.1.1.2.8 Tertiary and Quaternary Surficial Deposits

Late Tertiary to Quaternary surficial deposits survive as unconsolidated alluvial fan, pediment, and basin fill deposits throughout much of the region. Thickness or character of the deposits is highly variable, depending on the location.

3.1.1.1.3 Structural Geology and Tectonic Information

[The Technical Basis Report for Tectonics is scheduled for completion in mid FY 96 and will be the basis for updates to the tectonic sections in Revision 1].

The regional structural and tectonic setting has established the composition and configuration of the potential repository site at Yucca Mountain. Understanding past and present tectonic regimes is critical to evaluating the effectiveness of the natural barriers of the site, and includes predictive capabilities for potentially disruptive tectonic events during the lifetime of

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the repository. Information on the history of faulting and the active stress field support evaluations of the structural and seismic risks to the site. Volcanism (both in terms of the silicic eruptions that created the SNVF and Yucca Mountain, and of the younger basaltic centers that dot the southern Great Basin) must be understood to establish the regional geologic history and to understand the likelihood of future eruptions and the potential effects on a repository. The tectonic development and setting of the region exert control over the configuration and behavior of the surface and subsurface hydrologic systems.

3.1.1.1.3.1 Regional Structural Features and Current State of Stress

The Great Basin has long been recognized as an area where the major landforms have been created by extensional tectonic processes. Several extensional mechanisms and amounts of extension have been proposed. Relatively high-angle, planar, normal faults that cut the brittle (seismogenic) crust can accommodate a maximum of 10-15 percent extension. Normal faults that are at a high angle at the surface and curve to lower angles with depth (listric faults) may accommodate much greater amounts of extension. Very low-angle detachment-fault models have been developed that suggest extensive tectonic thinning of the brittle crust, accommodating extension of 200 percent or more (Figure 3.1.1.1.3.1-1). The typical Basin and Range structures, tilted fault block ranges with relatively large displacement, high-angle normal faults exposed at the surface, bounding one or both sides of each range, was well developed in this area by approximately 11 Ma. The Paintbrush Tuff (potential repository host rock) was deposited between 13.1 and 12.5 Ma. (Marvin et al., 1970). It was faulted to roughly its present configuration by 11.4 Ma when the overlying Timber Mountain Tuff erupted. Scott (1990) has suggested that rates of fault movement were highest between 13 and 11.5 Ma and have decreased since that time. Some of the faults near Yucca Mountain and in the surrounding region are oriented for movement within the modern stress field.

3.1.1.1.3.1.1 Plate Tectonic Setting

The plate tectonic setting of the southwestern United States is dominated by the interaction of the North American and Pacific Plates. In the region, particularly west of Yucca Mountain, the interaction of the North American and Pacific Plates is complicated by the overlap of right-lateral plate boundary stress and the extensional stress of the Basin and Range.

Estimates of current plate motion and locations of differential movement can be derived from geologic and geodetic measurements. The Pacific plate is moving toward the northwest at approximately 50 mm/year with respect to the North American plate (Figure 3.1.1.1.3.1.1-1). Most of this motion, about 35 mm/year, is taken up along the San Andreas Fault. This leaves approximately 15 mm/year of right-lateral movement that must be accommodated by other structures. Measurements from very long baseline interferometry indicate that the eastern edge of the Sierra Nevadan microplate (composed of the Sierra Nevada Mountains and the Great Valley of California) is moving northwest (relative to the rest of North America) at approximately 10 mm/year (Argus and Gordon, 1991). The remaining 5 mm/year is probably taken up by slip along the Hosgri, San Gregorio, and other coastal strike-slip faults parallel to and west of the San Andreas Fault.

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This model is complicated by the observation that Pacific plate motion (N36°W ± 2°) is not quite parallel to the strike of the San Andreas Fault (about N40°W). Consequently, there is some compression normal to the San Andreas Fault. There is also some uncertainty in the rate of relative motion between the Pacific and North American Plates. Relative motion was previously estimated at approximately 60 mm/year (Atwater, 1970). More recently, De Mets et al. (1990), have estimated the relative motion of 48 ± 1 mm/year. It is also possible that some motion is accommodated within the Sierra Nevadan microplate. The 10 mm/year of relative movement that occurs between the eastern edge of the Sierra Nevada Mountains and the western edge of the Colorado Plateau (the Southern Great Basin) is most important to the potential repository site.

Normal faulting (extensional tectonics) in the Great Basin was recognized by some of the earliest studies of the area (Gilbert, 1928). The amount of extension, however, has been a matter of debate. The timing of extension and rates of extension are also not fully understood. In parts of the Basin and Range, extensional faulting may have occurred in the late Eocene (40 Ma) and evidence exists for Oligocene normal faulting in the region (Reynolds, 1969). Most of the extension in the region occurred in the last 20 million years (Miocene to Holocene).

Extension was probably not uniform across the region in time or space. Several authors (Scott, 1990; Noble et al., 1991) have suggested that rapid extension migrated westward from Yucca Mountain after about 11.5 Ma. Extension may also have been nonuniform from north to south. Duebendorfer and Black (1992) present evidence that structures trending generally east-west may have accommodated different amounts of extension to the north and south.

While Miocene extension appears to have been wide and rapid, Pliocene and later extension appears to have a different style. Unruh (1991) presents convincing evidence for a Cordillera-wide uplift event starting about 5 Ma. The extension accompanying this uplift is more evenly distributed and is accommodated on high angle normal faults that are coincident with the Miocene faults at the surface. This model is consistent with current seismicity of the Basin and Range, which indicates high-angle faulting to depths of 15 km or more.

3.1.1.1.3.1.2 Regionally Significant Faults

Regionally significant faults are those structures which exhibit styles of deformation and rupture histories characteristic of tectonic subprovinces in the Great Basin. Because of the location of Yucca Mountain in close proximity to several different subprovinces (see Subsection 3.1.1.1.4.1), a review of these faults is important in developing seismotectonic models to guide hazard assessment. In addition, these faults, in particular those located in the southern Walker Lane, may be significant in terms of potential seismic activity. Faults which pose a seismic hazard to Yucca Mountain will be discussed in detail in Subsection 3.1.1.2.3.1.

Yucca Mountain stands near the southern end of the Walker Lane belt, a structural subprovince of the Great Basin. The Walker Lane belt trends northwest-southeast and is distinguished by a less distinct orientation to the basins and ranges than is seen farther east in

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the Great Basin, and by ranges that are lower with respect to the basin floors. Faults within the Walker Lane belt accommodate part of the transform motion between the Pacific and North American plates and strain due to the northwest-southeast extension of the Great Basin.

The Walker Lane belt is a northwest-trending zone of northwest-directed shear approximately 700 km long and 100 to 300 km wide (Stewart, 1990). It is characterized topographically by mountain ranges with unusual shapes or trends that lie between the Sierra Nevada Mountains on the west, and the large area of long, north-northeast trending mountains and valleys typical of Basin and Range topography on the east (Stewart, 1990). The Walker Lane belt is parallel or subparallel to the Sierra Nevada Mountains and the San Andreas Fault. It is narrow at the northwest end near Honey Lake in northern California, and widens toward its southeastern end (Figure 3.1.1.1.3.1.2-1). [INN 3.1.1.1.3.1.2-1] The southern end is truncated by the east-west trending Garlock fault. East of the Garlock fault, in the vicinity of Las Vegas, the eastern edge of the Walker Lane belt is probably bounded by the Las Vegas Valley shear zone.

The Walker Lane belt has a long and complex deformational history. Stewart (1990) summarizes and expands on the ideas of a number of workers who suggest a Mesozoic initiation of Walker Lane structures. Some of the Cenozoic deformation probably took place on preexisting structures and has been characterized by strike-slip faulting, oroflexural folds, and large-scale extension. The current style of deformation in the Walker Lane belt probably began approximately 5 Ma as an overlap between the right-lateral shear caused by Pacific and North American Plate relative motion and gravity-driven extension caused by Cordillera-wide uplift (Unruh, 1993). In the modern stress field, northwest-striking faults move with left-lateral strike-slip or oblique-slip.

In the southern Walker Lane belt there are three major faults along which right-lateral shear is distributed. From west to east, these are the Owens Valley, the Panamint Valley-Hunter Mountain, and the Death Valley-Furnace Creek fault systems (Figure 3.1.1.1.3.1.2-l). Data presented in Zhang et al. (1990), indicate that the right-slip rates in the current stress field for these fault systems are as follows:

	Slip in mm/year
Owens Valley	0 .70 - 2.20
Panamint Valley-Hunter Mountain	1.57 - 3.15
Death Valley - Furnace Creek	2 + - 2 +
Total	4.27 - 7.35

Zhang et al. (1990) speculate that there are lesser amounts of slip occurring farther east on the Pahrump Valley fault system and within the rest of southern Nevada and southwestern Utah (west of the Colorado Plateau). They estimate a minimum rate of 8 mm/yr of total right-slip across the southern Great Basin, which agrees with the ~10 mm/yr estimated from very long baseline interferometry (Argus and Gordon, 1991).

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The characteristics of specific regionally-significant faults in the tectonic provinces are summarized below, based on published paleoseismic information. These faults have histories of Quaternary movement and may be representative of fault behavior in the vicinity of the potential repository site. Some of these faults are capable of generating large earthquakes that may produce significant ground motion at the site. The faults are discussed in order of increasing distance from the site.

Garlock Fault. The Garlock fault is an arcuate, northeast-striking, near-vertical, left-lateral strike-slip fault system that forms the southern boundary of the Walker Lane belt and separates two provinces with different rates of extension within the Basin and Range, the Great Basin to the north, and the Mojave block to the south (Davis and Burchfield, 1973). The Garlock fault extends 265 km northeast from the Tehachapi Mountains in southern California, where it intersects the San Andreas Fault, to the southern Death Valley Fault Zone in the Avawatz Mountains, where it is about 125 km south of the proposed site. The Garlock fault is significant because it is potentially capable of producing a Moment Magnitude (M_w) 7.8 earthquake (McGill and Sieh, 1991).

The Garlock fault probably originated in the early Tertiary and has been active throughout the late Cenozoic (Smith, 1962; Astiz and Allen, 1983). Measurements of total left-lateral slip from offset geologic features are 48-65 km (Smith, 1962; Smith and Ketner, 1970; Smith 1975). Abundant geomorphic evidence indicates left-lateral slip in the Holocene (Clark, 1973; Smith, 1975; McGill and Sieh, 1991). Using more than 100 measurements of offset geomorphic features, McGill and Sieh (1991) determined slip per event and estimated slip rates and recurrence intervals for four areas. Their results from west to east are:

a. The El Paso Mountains area is characterized by 7 m of left-lateral slip during the most recent event, 7 m of slip in the penultimate event, 4 m of slip in the antepenultimate event, a slip rate of 4-7 mm/yr, and a recurrence interval of 600 to 1,200 years;

b. The Searles Valley area is characterized by 2-3 m of left-lateral slip for each of the past several events, a slip rate of 4-9 mm/yr, and a recurrence interval of 200 to 750 years;

c. The Pilot Knob Valley area is characterized by 2-4 m of left-lateral slip for each of the past several events, a slip rate of 3-9 mm/yr, and recurrence interval of 200 to 1,300 years; and

d. The Leach Lake-Avawatz Mountains area is characterized by 2-3 m of left-lateral slip in the most recent event, a slip rate of 1-9 mm/yr, and a recurrence interval of 200 to 3,000 years.

Vertical displacements vary between 0 and 30 percent of lateral displacements and the sense of vertical displacement varies along strike.

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McGill and Sieh (1991) also proposed many plausible rupture scenarios ranging from M_w 6.6 to 7.8; the latter assumes rupture of the entire fault. Surficial mapping and a trench exposure just west of the El Paso Mountains area, indicates 9-17 events occurred in the past 14,700 years, resulting in 80 m of left-lateral slip (Burke and Clark, 1978; Burke, 1979). They inferred 5-9 m of average slip per event and estimated a slip rate of 5.4 mm/yr. Radiocarbon ages from surfaces offset 2.7 ±0.6 m left-laterally and 0.2 m vertically, suggest the most recent event in the Avawatz Mountains occurred after 1,800-1,360 yr B.P., compared to a maximum age of 460 yr B.P. for the most recent event in Searles Valley (McGill, 1993).

Owens Valley Fault. The Owens Valley fault is a north-northwest-striking, steeply east-dipping, 100-km-long, right-lateral normal-slip fault that lies in Owens Valley, about 10 km east of the Sierra Nevada range front and 140 km west of the proposed site (Beanland, 1993; Bryant, 1993; Lubetkin and Clark, 1988; Beanland and Clark, 1987; Martel et al., 1987; Carver et al., 1969). The 1872 Owens Valley earthquake ruptured the entire fault zone and was one of the largest historical events in the Basin and Range Province (dePolo et al., 1991; Beanland, 1993). This earthquake is discussed in more detail in Subsection 3.1.1.1.4.2.2.

The Owens Valley fault is a good example of a Great Basin strike-slip fault that is associated with a range front fault (the Independence fault) that is characterized by dip-slip displacement and relatively lower slip rates (Beanland, 1993; Bryant, 1993). Detailed geomorphic and stratigraphic studies of this fault suggest that surface rupturing events since the late Pleistocene were similar in size and style of faulting to the 1872 earthquake (Carver et al. 1969; Beanland and Clark, 1987; Lubetkin and Clark, 1988; Beanland, 1993). Lubetkin and Clark (1988) studied the Lone Pine fault, a subsidiary fault of the central Owens Valley fault, and found evidence that suggests three events similar to the 1872 event occurred since 10-20 ka, yielding an average recurrence of 5,000 to 10,000 years. They calculated a Late Quaternary slip rate of 0.4 to 1.3 mm/yr for the Lone Pine fault. Martel et al. (1987) found dominantly vertical displacements along the Fish Spring fault, part of the northern Owens Valley fault. Their mean estimates of vertical displacement rates for the Late Quaternary range from 0.18 to 0.44 mm/yr and they estimate an average recurrence of 3,500 to 8,300 years. In a study of the entire fault system, Beanland (1993) found evidence for three Holocene events, resulting in an average slip rate of 2 ±1 mm/yr and an average recurrence of 3,300 to 5,000 years elsewhere on the Owens Valley fault besides the Lone Pine fault. She estimated a slip rate of 1.5 ±1 mm/yr for the last 300,000 years.

Stewart-Monte Cristo Valley Fault. This structure is a northwest-striking, right-lateral, normal-slip fault system that extends through the middle of Stewart and Monte Cristo Valleys (dePolo et al., 1991; Nakata et al., 1982). It is located in the Walker Lane belt, more than 200 km northwest of the proposed site. The Stewart-Monte Cristo Valley fault was the longest identified fault that ruptured during the 1932 Cedar Mountain earthquake (Subsection 3.1.1.1.4.2.2). These surface ruptures were not confined to a single range-front fault, but occurred over a zone 60 km long and 6-14 km wide (Gianella and Callaghan, 1934; Molinari, 1984). The longest continuous strand of surface faulting is only 17 km long. Most of the remaining surface displacement is distributed along discontinuous ruptures (dePolo et al.,

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1991). The distributed nature of the 1932 surface ruptures may be the result of several fault segments failing in sequence (dePolo et al., 1991).

Ongoing trench investigations in Monte Cristo Valley reveal flower structures, tilted beds, slickensides that plunge 6-10° north, sag pond and graben deposits, and colluvial wedges, indicating that small surface-rupturing events with right-normal slip have occurred repeatedly during the Quaternary (Yount et al., 1993a and b; dePolo et al., 1987, 1988; Bell, 1988). Molinari (1984) found evidence for two additional Holocene events and at least three, possibly five or six, surface faulting events during the latest Pleistocene and Holocene. Slip rate estimates for the last 25 to 35 ka range from 0.2-0.7 mm/yr, and estimates for the last 6 to 13 ka are 0.3-0.7 mm/yr (Bell, 1993).

Dixie Valley Fault. Dixie Valley fault is a north-northeast striking, east-dipping fault zone that includes both a range front and a piedmont fault that lie along the west side of Dixie Valley within the Central Nevada Seismic Belt (CNSB) in the northern Great Basin (Slemmons, 1957; Bell and Katzer, 1987, 1990). Complex surface rupture occurred on this fault during the 1954 Dixie Valley earthquake (Subsection 3.1.1.1.4.2.2). The Dixie Valley fault has a complex history of rupture alternating between the piedmont and range-front faults; behavior which does not fit a simple segmentation model (Bell and Katzer, 1990). Although the historic rupture was concentrated along the range-front fault, the most recent prehistoric event occurred on the piedmont fault between 1.5 ka and 6.8 ka and did not rupture the range-front fault (Bell and Katzer, 1987, 1988). The most recent prehistoric event on the range-front fault occurred in the late Pleistocene and caused 2.5-3 m of vertical displacement, similar to the 1954 rupture (Bell and Katzer, 1987, 1988). The Dixie Valley fault is also an example of variations between short-term and long-term activity, which is apparently common in the CNSB (Pearthree et al., 1993; see Subsection 3.1.1.1.4.2). The estimated slip rate for the last 6 to 13 ka is 0.5-0.8 mm/yr across the entire Dixie Valley fault, whereas the slip rate for the last 200 ka is 0.2 mm/yr (Bell, 1993).

3.1.1.1.3.1.3 Structural Features and Seismicity

The relationship between specific structural features, particularly faults, and seismicity in the Basin and Range is not clear. Some zones of seismic activity, the north-south trending CNSB for example, are clearly associated with major faults or fault systems that have experienced historic surface ruptures. There are, however, other zones of seismic activity, and areas of diffuse activity where there is no historic surface faulting. The east-west seismic belt, which includes the northern part of the Nevada Test Site (NTS), is an example of a zone of seismic activity where historic surface faulting is absent.

There are several reasons for the apparent poor correlation between earthquakes and faults. These include the short historical record relative to the long earthquake-recurrence intervals on the faults in the region, the difficulty in accurately locating epicenters in a region of low population and few seismographs, and the unknown geometry of faults at depth. Earthquakes may occur on subsurface faults that have no mapped surface expression. Surface faulting is

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generally associated only with relatively large magnitude earthquakes (6.0 to 6.5) (Smith and Arabasz, 1991), which are comparatively rare.

In terms of estimating or predicting future seismic and/or surface faulting activity, the paleo-seismic record for the Quaternary is probably more valuable than the brief historic record. Detailed studies of offset surficial deposits on faults in the YMR suggest that recurrence intervals for surface rupture are on the order of thousands to tens of thousands of years. While long by human standards, that still implies up to 200 events per fault during the Quaternary (approximately the last 2 million years).

3.1.1.1.3.2 Regional Volcanic Features

The rocks forming Yucca Mountain were formed by voluminous eruptions of silicic ash from nearby calderas during the Miocene. Later, during the Pliocene and Quaternary, small volume basaltic volcanoes erupted in the vicinity of the site.

3.1.1.1.3.2.1 Silicic Volcanism

Silicic volcanism commenced in the northern Basin and Range during the early- to mid-Tertiary. By the mid-Miocene (15 to 16 Ma), volcanism had migrated southward to the region around Yucca Mountain and began forming the SNVF (Figure 3.1.1.1.3.2.1-1). This area was a center of large-scale pyroclastic volcanism for approximately 8 Ma during the mid- to late-Miocene. Noble et al. (1991) have divided this volcanism into three stages as follows:

a. The main magmatic stage (15.2-12.8 Ma), during which a number of overlapping and nested calderas developed with the eruption of at least nine voluminous sheets of silicic ash-flow tuff. Many smaller units of lava and tuff of silicic to intermediate composition also were erupted during this stage (Noble et al., 1991). The Paintbrush Tuff represents the last major eruption in the main stage.

b. The Timber Mountain stage (11.5-10 Ma) began with eruption of the Rainier Mesa Member of the Timber Mountain Tuff (11.4 Ma). The Timber Mountain Tuff was erupted onto a faulted topography and is thin or absent on the high standing blocks of Paintbrush Tuff. The Rainier Mesa Member was followed by the Ammonia Tanks Member and two younger members, all apparently erupted from the Timber Mountain caldera.

c. The late magmatic stage (9-7 Ma) occurred mostly in the Black Mountain and Stonewall Mountain volcanic centers, farther from Yucca Mountain in the northwestern part of the SNVF (Noble et al., 1991).

Since the end of voluminous volcanic activity in the SNVF, silicic volcanism appears to have migrated outward from the region toward the margins of the Basin and Range. The nearest silicic center of Pliocene or younger age is the 2.9 Ma Mount Jackson rhyolite dome, located

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105 km northwest of Yucca Mountain. Volcanism in the vicinity of Mount Jackson appears to be a distinctly later pulse of activity, more closely related to late-Pliocene and Quaternary volcanism along the western margin of the Basin and Range. Other volcanic centers along the western margin include the Long Valley caldera, the Coso Volcanic field, and the Saline Range Volcanic field in northern Death Valley (Noble et al., 1991).

3.1.1.1.3.2.2 Basaltic Volcanism

Miocene silicic volcanism in the field was succeeded by late-Miocene to Quaternary basaltic volcanic activity. This basaltic activity can be divided into two episodes: Basalt of the Silicic Episode (BSE) that occurred during the waning stage of silicic volcanism and Post-caldera basalt (8.8 Ma to Quaternary) (Crowe et al., 1995).

The BSE occurred in three major geographic areas (Crowe et al., 1995):

- basalt exposed in the moat zone of the Timber Mountain caldera
- basalt near and flanking the Black Mountain caldera
- isolated basaltic eruption centers.

The post-caldera episode includes the Older Post-caldera basalt (OPB) that occurs north and northeast of Yucca Mountain and the Younger Post-caldera basalt (YPB) that crops out west, southwest, and south of Yucca Mountain (except the basalt of Buckboard Mesa). The OPB consists of the 8 Ma basalt of Rocket Wash, the basalt of Pahute Mesa (~ 9 Ma), the basalt of Paiute Ridge (8.0-8.7 Ma), the basalt of Scarp Canyon (8.7 Ma), the basalt of Frenchman Flat (8.6 Ma), and the basalt of Nye Canyon (6.5-7.0 Ma). Each basalt unit is a relatively small-volume eruptive center (< 1 km^3) formed by clusters of scoria cones with associated lava flows. The centers formed at or along basin-range faults and at the intersection of basin-range faults and ring-fracture zones of caldera complexes, except for the basalt of Nye Canyon. The latter unit forms a northeast-cluster of centers that does not follow local structure. Two centers of the basalt of Nye Canyon formed partly from hydrovolcanic eruptions. The basalt of Paiute Ridge consists of dissected scoria cones and flows underlain by a complex sequence of sills, dikes, and lopolithic intrusions.

The YPB consist of seven sites of Pliocene and Quaternary volcanism; six of the sites occur in a narrow northwest-trending zone called the Crater Flat Volcanic Zone (CFVZ). The oldest and largest volume basaltic volcanic center of the YPB is the basalt of Thirsty Mesa, a newly recognized Pliocene volcanic center. It consists of three coalesced, small scoria cones surmounting a lava mesa. The age of the unit is 4.8 Ma. A negative aeromagnetic anomaly was drilled by a private company and has been shown to be a buried basalt center of 3.8 Ma. The 3.7 Ma basalt of southeast Crater Flat comprises five north-trending dissected scoria cones and associated moderate-volume lava flows. It is the largest volume Pliocene or Quaternary basalt in the Crater Flat alluvial basin. The unit formed largely from Hawaiian-type fissure eruptions accompanied by outpouring of sheet-like lobes of aa lava flows. The

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basaltic andesite of Buckboard Mesa (2.9 Ma) crops out in the moat zone of the Timber Mountain caldera, northeast of Yucca Mountain. Nearly 1 km^3 of lava vented from a small scoria cone and an associated northwest- trending fissure located southeast of the scoria cone.

The Quaternary basalt of Crater Flat consists of an arcuate alignment of basalt centers extending along the axis of Crater Flat. Individual centers of the alignment include, from southwest to northeast, the Little Cones, the Red Cone, the Black Cone, and the Makani Cone centers (Crowe et al., 1995). Greater than 90% of the volume of the four centers are contained in the scoria cone and lava units of the Red Cone and Black Cone centers. The Little Cones center consists of two closely spaced, small scoria cones. Lava flows were extruded from and breached the south wall of the southwest center. The Little Cones yield K-Ar ages of 1.1 to 0.76 Ma. These ages are consistent with the degree of dissection, the degree of horizon development in soils and the magnetic polarity of the center.

The Red Cone and Black Cone centers are analogous volcanic landforms with similar eruptive histories. Each consists of a main scoria cone capped by a crater filled with agglutinated spatter, large lava blocks and scoria. The main scoria cones of both centers are flanked to the south by eroded scoria mounds that vented aa lava flows. K-Ar ages of between 0.84 and 1.55 Ma were reported for Red Cone; ages for Black Cone range between 0.8 and 1.1 Ma. Soil and geomorphic data are consistent with this range.

The Makani center is a deeply dissected remnant of a scoria cone and lava flow. K-Ar age determinations for this center range from 1 to 1.66 Ma. A continuing area of controversy for the Quaternary basalt of Crater Flat is the eruptive models for individual centers, and the age differences between each center.

K-Ar and $^{40}Ar/^{39}Ar$ ages of the basalt centers in Crater Flat suggest that the mean age of all centers is 1.0 ± 0.1 Ma. Paleomagnetic data are permissive with, but do not provide conclusive proof that the centers are all close in age and formed from a single pulse of magma. Field, geomorphic, soil, and petrologic data suggest some of the centers could be polygenetic, and may differ slightly in age. The difference in ages of the centers may be less than the detection limits of K-Ar and $^{40}Ar/^{39}Ar$ chronology methods.

The Sleeping Butte centers are located 45 km northwest of Yucca Mountain. They consist of two centers: the southwest Little Black Peak center; and the northeast, Hidden Cone center. Each consists of a small volume main scoria cone flanked by blocky aa lava flows that vented from radial dikes at the base of the cone. The Little Black Peak center appears from field, geomorphic, petrologic, and paleomagnetic data to be a simple monogenetic center with a K-Ar age of about 380 ka. The Hidden Cone center is more complex and may have formed from at least two temporally distinct eruptions. The age of the major volume of the Hidden Cone center is about 380 ka. The age of a scoria-fall eruption that mantled the main cone of the center may be as young as late Pleistocene.

The Lathrop Wells volcanic center is the youngest and most carefully studied basaltic volcanic center in the region. The center is located at the south end of Yucca Mountain, near

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the intersection of northwest and northeast trending fault systems; most of the vents are aligned along northwest-trending fissure systems. The basalt center formed from multiple eruptions that comprise four chronostratigraphic units. The oldest chronostratigraphic unit (I) consists of four groups of lava flows and local pyroclastic deposits marking the vents for the lava flows. These volcanic deposits occur along multiple, northwest-trending fissures located south, beneath, north, and northeast of the main cone. The scoria and spatter deposits of chronostratigraphic unit I were modified extensively by erosion prior to formation of chronostratigraphic unit II. Chronostratigraphic unit II consists of two sets of lava and pyroclastic deposits. The largest volume lava of the center was erupted from a northwest-trending fissure that is parallel to a northwest-trending normal fault, which displaces the Timber Mountain Tuff, but not the basaltic deposits. Small volume lobes of lava were erupted from a west-northwest trending fissure that extends from the north end of the northwest-trending fissure system. A short, northwest-trending fissure marked by scoria mounds formed at the northeast base of the main scoria cone. Widespread scoria-fall and pyroclastic-surge deposits were erupted from vents that are inferred to be largely concealed beneath the modern main cone. The eruptive events of chronostratigraphic unit III formed most of the main scoria cone and small-volume lobes of blocky aa lava flow that erupted from a vent located northeast of the main cone. Chronostratigraphic unit IV includes small volume eruptions from a cluster of inferred, small satellite vents south of the main cone that have been removed by commercial quarrying of the main cone. The eruptive events of this chronostratigraphic unit have been established through identification of local tephra beds that have distinctive major-and-trace element chemistry.

The delineation of individual chronostratigraphic units has been established on the basis of detailed field, stratigraphic, geomorphic, and soils studies. The individual chronostratigraphic units can be discriminated uniquely on the basis of major and trace-element geochemical compositions. The chemistry of the units is inconsistent with a simple, monogenetic eruption model and instead requires formation from multiple magma batches consistent with a polygenetic origin of the volcanic center (multiple, time-distinct volcanic eruptions).

Establishing the age of the multiple volcanic events at the Lathrop Wells volcanic center has been problematic. The problem has been approached by application of multiple geochronology and age-correlation methods (K-Ar, U-Th disequilibrium, cosmogonic He, thermoluminescene (TL), geomorphic, soils, paleomagnetic, and petrologic studies). Recent results show some convergence in the range of ages obtained using the different methods but the data remain permissive of multiple eruptive and chronology models. Conventional whole rock K-Ar ages of lava units of the Lathrop Wells volcanic center give a mean age of 137 ± 52 (1 sigma) ka. The age is obtained by averaging age determinations from chronostratigraphic units of probable different age. Moreover, the ages show a positive correlation between percentage radiogenic Ar and the measured age of the sample. Published $^{40}Ar/^{39}Ar$ age determinations of the same rock samples as the conventional K-Ar data set also show a positive correlation between percentage radiogenic Ar and age. The mean age of all measured $^{40}Ar/^{39}Ar$ age determinations is 162 ± 62 (1 sigma) ka, and is again obtained by averaging age determinations from separate chronostratigraphic units. At best, the K-Ar and $^{40}Ar/^{39}Ar$ ages provide maximum estimates of the ages of the two oldest chronostratigraphic

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units (I and II). The data sets cannot be used to establish the range in age of the volcanic center or to discriminate the ages of individual chronostratigraphic units. U-Th disequilibrium age determinations have been obtained for lavas of chronostratigraphic units I and III. Mass spectrometric analyses of separated phases yielded isochrons with ages of 135 ± 20-15 and 125 ± 25-30 (1 sigma) ka, in relatively close agreement with the K-Ar and $^{40}Ar/^{39}Ar$ age determinations. Difficulties with the U-Th disequilibrium measurements include a small degree of U-Th fractionation in measured phases, and a similarity in measurement ages of volcanic units that are inferred to be of probable different ages.

The ages of chronostratigraphic units at the Lathrop Wells center have been estimated additionally by measuring the accumulation of cosmogonic ^{3}He in 25 surface samples (Crowe et al., 1995). The preferred ^{3}He ages of chronostratigraphic unit I range from 80 to 90 ka, and represent minimum ages, because of coverage of the sample deposits by scoria-fall deposits of chronostratigraphic unit II. The ^{3}He ages of chronostratigraphic unit II range from 80 to 100 ka. Deposits of chronostratigraphic unit III show the greatest range in ^{3}He ages (30 to 65 ka) because many of the samples were collected from the main scoria cone, a non-resistant geomorphic feature. The age of the unit may be greater than 65 ka if the ages are interpreted as minimum ages. This interpretation does not explain however, the wide variations (40%) in replicate ages for samples collected from the geomorphically unmodified main cone. Two samples collected from the interior of volcanic deposits yielded ages of about 5 to 6 ka consistent with a zero age, and a small component of inherited ^{3}He. TL age determinations give reproducible and consistent ages of 4 to 9 ka for tephra units of chronostratigraphic unit IV that are interbedded with soils showing horizon development. TL ages of about 25 ka for baked sediments beneath lava of chronostratigraphic unit II are inconsistent with the age determinations of all other chronology methods.

Geomorphic studies of the degree of dissection of deposits of the Lathrop Wells volcanic center are consistent with an age of no older than 20 ka for the youngest events at the center. This is consistent with the TL ages of the youngest tephra units, but is somewhat younger than the measured ^{3}He ages of the main cone. The systematic differences in degree of erosion of volcanic landforms of chronostratigraphic units I, II, and III require a time-difference between each unit. Paleomagnetic data have been obtained to augment studies by Champion (1991) and Turrin et al. (1991) and now have been measured for all significant chronostratigraphic units at the volcanic center. The between-site dispersion of directions of the primary magnetization of units differs considerably for some sites, because of a combination of difficulty in sampling intact material, and lightening strikes. Paleomagnetic data obtained from two phases of sampling are not dissimilar from those reported by Turrin et al. (1991). However, the data do not support the contention that individual eruptive features have unique paleomagnetic signatures that can be confidently separated from other eruptive features. Eruptive features of the center have recorded directions of the latest Quaternary geomagnetic field that are well within the expected one-sigma range of paleosecular field variation about the spin-axis direction. There is no indication of a single volcanic event occurring during a period of unusual geomagnetic activity.

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The field relations of the volcanic units are well constrained. The chronology of the units remains uncertain and the existing data can only partly be used to test the field relations. Three alternative models are presented for the eruptive history of the Lathrop Wells center. These include respectively, a four-event eruption model (>130, 80-90, 65, and 4-9 ka), a three event eruption model (120-140, 65 and 4-9 ka), and two-event eruption model (120-140, 4-9 ka).

3.1.1.1.3.3 Alternative Tectonic Models

[The following discussion is preliminary and incomplete. The approach taken here to presenting alternative regional tectonic models is highly conceptual. No rigorous treatment of model versus available data has been attempted. There are a number of alternative models for the regional tectonic environment and this section will be expanded to include these models in future revisions. [INN 3.1.1.1.3.3-1]] The purpose for considering alternative tectonic models is to evaluate what the effect on waste isolation would be if the regional tectonics were different than the model presented in preceding sections. To highlight the differences and implications of the alternative models, the preferred model just presented will be briefly summarized and its implications for ground-water flow, seismicity and fault rupture, and the potential for future volcanism will be stated. The same format will then be followed for the alternative models.

In the preferred model, Miocene extension was large, fairly rapid, and accompanied by detachment faulting, rise of metamorphic core complexes, and oroflexural bending of ranges. Extension was not uniform in time and space and appears to have migrated westward from the Yucca Mountain area after about 11.5 Ma. Extension was at least in part coeval with the voluminous silicic volcanism of the SNVF. The current stress regime is an overlap of northwest-directed, right-lateral shear induced by Pacific/North American relative plate motions and minor extension induced by gravitational adjustments of the buoyant crust of the Basin and Range. This stress regime began approximately 5 Ma (Unruh, 1991). The recent (Pliocene and Quaternary) magmatic activity of small volume basaltic volcanism is unrelated to the SNVF. Small-volume basaltic volcanism is characteristic of the current magmatic setting.

The implications of this model are that the Paleozoic carbonate rocks that underlie the Miocene tuffs, and are considered to be a regional aquifer, have been extensively disrupted by Miocene extension. This suggests that the regional flow system is complex, and contains isolated bodies of carbonate aquifer. In terms of fault rupture and seismicity, large earthquakes and associated surface ruptures are likely on the major faults of the southwestern Walker Lane belt, including the Owens Valley, Panamint Valley-Hunter Mountain, and Death Valley-Furnace Creek fault systems. As one approaches the Yucca Mountain Site, the amount of northwest-directed shear begins to decrease and the faults and fault systems are generally shorter and/or less continuous. Earthquake magnitudes are expected to be smaller and recurrence intervals longer. However, as has been demonstrated by paleoseismic studies, the faults at and near the site are favorably oriented in the current stress field and can be expected to rupture within the next 10,000 years. Fault planes are expected to be high angle

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and earthquake focal depths to 15 km are possible. The Quaternary pattern of volcanism is expected to continue with small-volume basaltic eruptions occurring in valleys or at range margins rather than in range interiors, similar to the pattern that has developed in Crater Flat. This argues against direct disruption of the repository by volcanic processes.

The first alternate model considered here is very similar to the preferred model with the exception of the forces that induce extension in the current stress field. In this model, extension is driven not by gravity, but by shear tractions on the base of the lithosphere from mantle convection. The implications of this model are as follows.

The Paleozoic carbonate aquifer has been regionally disrupted by Miocene extension and a complex flow system has developed. In terms of fault rupture and seismicity, the detachment surfaces developed in the Miocene would still be active structures. Based on their length, geomorphic expression, and paleoseismic data, the major faults of the southwestern Walker Lane belt would be expected to generate large earthquakes and associated surface rupture. If extension has continued to move westward, the most rapid extension would also be in this area. A detachment at Yucca Mountain at a depth of approximately 3-4 km has been proposed by Scott (1990), which is in turn underlain by other deeper detachments. He suggested a tier of probably interconnected, low-angle faults that form an extensional system above a mid-crustal detachment at the brittle-ductile transition (Scott, 1990). If high-angle faults at the surface sole into a low-angle fault and the low-angle fault moves, the implication is that multiple fault movements could occur in the upper plate. The oroflexural bending of ranges and other responses to differential extension would be a continuing process. Earthquakes in the upper plate would be shallow and perhaps of limited magnitude. Deeper, larger earthquakes may or may not have direct surface expression, or may have distributed expression. Relative to the potential for future volcanism, an active convection system in the mantle suggests that volcanism of significant volume may be possible.

The second alternative model assumes less Miocene extension. The arguments in favor of and evidence supporting large amounts of extension are not universally accepted. Proponents of large-scale extension agree that extension was not uniform in space and time and large-scale extension may not have occurred everywhere. It is possible then that the region around Yucca Mountain was not subjected to large scale extension. Leaving other aspects of the preferred tectonic model the same, this implies that the Paleozoic carbonate aquifer is not extensively disrupted, and that the regional flow system is interconnected and can be modeled with relative confidence.

3.1.1.1.4 Seismology

[The Technical Basis Report for Tectonics is scheduled for completion in mid FY 96 and will be the basis for updates to the tectonic sections in Revision 1].

Subsection 3.1.1.1.4 describes the seismotectonic setting of the region within a 322 km radius around the potential repository site (Subsection 3.1.1.1.4.1), summarizes the characteristics of seismicity within this region (Subsection 3.1.1.1.4.2), and describes the state of the regional

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tectonic-stress field (Subsection 3.1.1.1.4.3). The purpose of this section is to provide a regional framework for the assessment of the seismic hazard for Yucca Mountain. It will focus mainly on the seismological aspects of the tectonic features. A review of the basic tectonic elements of the region is provided in Subsection 3.1.1.1.3 on Structural Geology and Tectonic Information. A review of the detailed characteristics of seismicity within 100 km of the potential site is contained in Subsection 3.1.1.2.4 on Site Seismology.

3.1.1.1.4.1 Regional Seismotectonic Setting

The following discussion is based on the data and the approach presented in the SCP. The text will be revised to reflect the results of planned activities and the application of the U.S. Department of Energy's (DOE's) proposed seismic hazard assessment methodology [INN 3.1.1.1.4.1-1].] In the region around the site there are several zones in which seismicity is concentrated: the Sierra Nevada-Great Basin Boundary Zone (SNGBZ) (VanWormer and Ryall, 1980), the CNSB (Wallace, 1984), the Southern Nevada Transverse Zone (SNTZ) (Slemmons et al., 1965), the Garlock fault (Astiz and Allen, 1983) and the Mojave block (Hutton et al., 1991). In addition to the seismicity concentrated in these zones, there is a broad distribution of seismic activity throughout much of the Great Basin which is not associated with any major known tectonic feature (Rogers et al., 1991). Figure 3.1.1.1.4.1-1 identifies the major tectonic elements in the western Great Basin and boundary regions (see [INN 3.1.1.1.4.1-2]).

Walker Lane belt

The main tectonic element in which the first four of these seismic zones wholly or partially lie is the Walker Lane belt (Figure 3.1.1.1.4.1-1). The Walker Lane belt consists of nine structural blocks that act more or less independently of each other (Stewart, 1988). These blocks are defined by the style of faulting which ranges from northwest-trending right-lateral slip (the Pyramid Lake, Walker Lake, and Inyo-Mono blocks), to northeast- trending left-lateral slip (the Carson, Spotted Range-Mine Mountain, and Lake Mead blocks), and to east-west trending left-lateral slip (Excelsior-Coaldale block). Two blocks, the Goldfield and the Spring Mountains, do not have significant strike-slip faults in the interior but are bounded by major northwest-trending right-lateral faults. Strike-slip faults in one block rarely extend into adjacent blocks (Stewart, 1988). Cumulative lateral offset on individual major faults ranges from a few to as much as 100 km.

Although the Walker Lane belt probably developed in the Mesozoic, most of the faults have evidence of Cenozoic movement with numerous fault zones showing Quaternary and Holocene offset (Stewart, 1988; Wallace, 1984). While it is the northwest and northeast trending strike-slip faults that delineate this region from surrounding provinces there has been considerable movement on north to northwest trending high-angle faults with dip-slip or oblique-slip motion and localized areas of large-scale extension on low-angle normal faults (Stewart, 1988). This diversity in style of Cenozoic deformation results in a disrupted pattern of topographic relief, which characterizes the Walker Lane belt and suggests significant changes in the orientation of the stress field. Repeated changes in the stress field from an

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extensional mode to a strike-slip mode accounts for the presence of basin and range structures and large-scale strike-slip faults in the same region (Stewart, 1988; Zoback, 1989). The orientation of the regional stress field and temporal and spatial changes during the Cenozoic are discussed in Subsection 3.1.1.1.4.2.3.

Late Quaternary faulting in the Walker Lane belt and in the western Great Basin in general occurs in pulses. The long-term recurrence intervals are on the order of thousands or tens of thousands of years (Wallace, 1984). The short-term recurrence intervals may be on the order of decades for certain sections of the seismic zones (e.g., the CNSB). Wallace (1984) proposes a model of strain release in which faulting occurs in a cluster of activity for a period in a subprovince of a seismic zone while adjacent areas remain relatively quiet. Subsequent to this, activity initiates in another subprovince (not necessarily an adjacent one) but does not return to the previous area of activity for thousands of years. This behavior is in contrast to that of the San Andreas fault, a first-order plate-boundary structure to which the Walker Lane is sometimes compared, in which significant displacement has occurred along almost the entire length of the fault during historic time (Jennings, 1975).

Northern Great Basin

The region to the north and east of the Walker Lane belt is characterized by a significantly different style of deformation and is designated the northern Great Basin (Figure 3.1.1.1.4.1-1). This tectonic province encompasses those regions in northwestern Nevada not covered in Stewart's (1988) definition of the Walker Lane belt and extends west to east from northern California to the Wasatch front in Utah and north to south from the Snake River Plain to the SNTZ. The boundaries of the northern Great Basin are coincident with the boundaries of the "Basin and Range deformational field I" of Wright (1976). The primary structural features in this province are the north-northeast trending horst and grabens marked by gently tilted fault blocks and steeply dipping normal faults at the range fronts. These fault blocks are grouped into regions of consistent tilt (Stewart, 1980b). The patterns of tilt domains are defined by antiformal (tilts away from) and synformal (tilts toward) boundaries at right angles to the northwest extension direction. Tilting of the ranges averages about 15° to 20° in Nevada and Utah. The regional tilt pattern may be related to stress relief extending outward from antiformal boundaries that are interpreted as initial sites of rupture during late Cenozoic extension (Stewart, 1980b). Three structural zones that separate domains of east- or west-tilted blocks trend northwesterly (parallel to the extension direction) across the northern Great Basin (Stewart, 1980b). These boundary zones accommodate contrasts in the direction, magnitude, and rate of strain due to differences in regional structural vergence across the zones. Thenhaus and Barnard (1989) define these zones as "extensional accommodation zones" and suggest that these zones play a controlling role in limiting the lengths of belts of range-bounding Quaternary faulting (See Subsection 3.1.1.1.4.2.1).

Quaternary faults are distributed throughout the northern Great Basin (Nakata et al., 1982; Wallace, 1984); Figure 3.1.1.1.4.1-2. With the exception of a region in central Nevada straddling the Utah border, late Quaternary activity is also broadly distributed throughout the province (Wallace, 1984). Holocene activity is confined to areas in the western and eastern

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parts of the province, with a relatively quiescent strip along the 115th meridian (Wallace, 1984). A large part of the CNSB is located in the northern Great Basin and is the location of several significant historic earthquakes (see Subsection 3.1.1.1.4.2.2). Small to moderate magnitude events are diffusely distributed throughout the rest of the region.

Mojave Block

The Mojave block in southeastern California is separated from the Walker Lane belt to the north by the Garlock fault and bounded by the San Andreas Fault on the west (Hutton et al., 1991); Figure 3.1.1.1.4.1-3. For the purpose of this discussion, the eastern margin of the province is the California border with Nevada and Arizona. To the south the block is bounded by the left-lateral Pinto Mountain fault (Dokka, 1986). Although this block lies within the Basin and Range physiographic province (Wallace, 1984), the tectonic processes are related to right-lateral shear along the boundary between the North American and Pacific plates. The primary tectonic structures of the Mojave block are a group of seven right-lateral, strike-slip faults that strike northwest across the central section of the province (Figure 3.1.1.1.4.1-3). The Pinto Mountain fault trends east-west across the southern edge of the block and truncates the strike slip faults (Hutton et al., 1991). In the northeast section of the province several east-west trending faults exist that parallel the eastern segment of the Garlock fault. All of these faults have been active in the Quaternary (Jennings, 1975).

The Quaternary faults of significance to this report occur in the central Mojave block. Cumulative late Cenozoic displacement on the seven northwest-trending right-lateral faults is approximately 26 to 38 km (Dokka, 1983). Recent seismicity is distributed in clusters and isolated events in the center of the block, and more densely concentrated in the southern portion of the province (Hutton et al., 1991). Significant historic earthquakes are the 1947 M_L(Local Magnitude) 6.2 Manix, which caused approximately 2 km of surface rupture along the east-west trending Manix fault (Richter, 1958); the 1975 M_L 5 and the 1979 M_L 5.7 Homestead Valley earthquakes (Hutton et al., 1991), two very shallow events which had peak depths less than 5 km; and the 1992 M 7.6 Landers earthquake, that produced about 100 km of surface rupture along several of the northwest trending faults in the region (USGS, 1992), see Subsection 3.1.1.1.4.2.2.

3.1.1.1.4.2 Regional Seismicity

Seismicity in the region surrounding the potential repository site is generally associated with one of several seismic zones in the Basin and Range province: the CNSB, the SNGBZ, the SNTZ, and the Garlock fault. Seismicity is also associated with subprovinces of the Walker Lane belt, the Mojave block, and diffusely distributed through much of northern Great Basin, with no clear alignment to any known tectonic structure. Seismic sources are more confined regions of the crust that may contain Quaternary faults, historic earthquake epicenters, Quaternary volcanic features, testing sites for Underground Nuclear Explosions (UNEs) (e.g., the Nevada Test Site), reservoir impoundments (e.g., Lake Mead), or a combination of these. This section describes the seismic-source zones in the region surrounding the potential repository site. Figure 3.1.1.1.4.1-2 shows the location of these zones. Subsection

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3.1.1.1.4.2.2 describes significant historical earthquakes that characterize seismic sources within these zones.

Sierra Nevada-Great Basin Boundary Zone

The SNGBZ, forms the western boundary of the Walker Lane belt (Stewart, 1988). This seismic zone was described by VanWormer and Ryall (1980) as extending along the Sierra Nevada front from the latitude of Susanville, California, in the north, to Bishop, California, in the south. Parts of the SNGBZ are correlative with the Eastern California Seismic Belt as described by Wallace (1984). This belt extends along the entire Sierra Nevada front south to its intersection with the Garlock fault and continues south into the central Mojave Desert region to the southern San Andreas fault zone. Wallace (1984) considers the Garlock fault to segment the Eastern California Seismic Belt. In the north, the zone is characterized by both dip-slip and strike-slip Quaternary faults and in the south, it is dominated by right-lateral strike-slip Quaternary faults. To facilitate characterization of seismic-source zones the SNGBZ is defined on the basis of consistency in seismological source parameters and style of faulting. The feature is defined on the west as the Sierra Nevada frontal fault from Lake Almanor in northeastern California to the Garlock fault, a distance of about 600 km (Figure 3.1.1.1.4.1-2). This zone is, therefore, coincident with the western limit of basin and range topography and by definition forms the western boundary of the Walker Lane belt defined by Stewart (1988). The eastern limits of the SNGBZ are not precisely defined. Based on the concentration of earthquake epicenters, the SNGBZ is about 30 km wide along its length. To the south, the eastern boundary is the White Mountains and the Inyo Mountains (Wallace, 1984).

Many of the faults in the SNGBZ show evidence of Quaternary movement and there are numerous faults with Holocene scarps and geomorphic features suggesting Holocene movement (Wallace, 1984; Bryant, 1993; Page et al., 1993). Historic surface rupture has occurred on the Owens Valley (1872), Hilton Creek (1980), Little Lake (1982), and the White Mountains and Volcanic Tablelands faults (1986) (Bryant, 1993). The sense of motion for Late Quaternary slip varies on these faults from right-lateral to dip-slip. Focal mechanisms from recent earthquakes in the SNGBZ indicate both dextral and dip-slip faulting is occurring with tension axes oriented east-west to southwest-northeast (VanWormer and Ryall, 1980; Zoback, 1989; and Vetter, 1990). This consistency in T-axis orientation is a distinguishing characteristic of the SNGBZ compared to the rest of the Walker Lane belt and the CNSB.

Central Nevada Seismic Belt

The CNSB is a zone of concentrated seismicity extending from the Mammoth Lakes area north to Winnemucca, Nevada (Wallace, 1984); Figure 3.1.1.1.4.1-2, and perhaps further north to the Oregon border (Michetti and Wesnousky, 1993). Unlike the SNGBZ, there is no defining structural element that sets this zone apart from the rest of northern Nevada; most of the faults in this region trend north to northeast. Although geologic data show repeated movement along the steeply-dipping range front faults, and some historic events also break these same faults, the association of earthquakes with known faults is not always possible.

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The morphology of the northern Great Basin province suggests that, in the late Cenozoic, the stress regime was oriented such that extension occurred in a west-northwest direction, resulting in dip-slip motion on faults oriented north-northeast (Wright, 1976). Focal mechanisms from moderate as well as large earthquakes indicate that this extension direction still predominates (Wallace, 1978; Zoback, 1989). However, normal-oblique and strike-slip motion on steeply dipping faults appear to be the main mechanism for strain relief (Zoback, 1989).

The CNSB contains many north to northeast trending faults with Quaternary and Holocene offset similar to other regions in north-central Nevada. Historic surface rupture has occurred in Pleasant Valley (1915), Wonder (1903), Rainbow Mountain-Stillwater (1954), Dixie Valley-Fairview Peak (1954), Cedar Mountain (1932), Excelsior Mountains (1934), and Chalfant Valley (1986). The clustering of activity evident along the CNSB is typical of the pulses of displacement which Wallace (1984) considers characteristic of the western Great Basin. In contrast, paleoseismic studies indicate that recurrence intervals of 10,000 years or more are typical for most Quaternary faults in the region (Pearthree et al., 1993). Faulting along the CNSB represents a temporal and spatial clustering of activity, lasting perhaps 500 years, that may have also occurred in the mid- to late-Holocene (Pearthree et al., 1993). Pearthree et al. (1993), find no evidence for other continuous north-south belts of activity in north-central Nevada. Seismic gaps, defined by the absence of significant surface rupturing events, exist along the CNSB north of the Pleasant Valley rupture zone (Sonoma Range Gap; Thenhaus and Barnard, 1989), between the Pleasant Valley and the Dixie Valley-Fairview Peak ruptures (the Stillwater seismic gap; Wallace and Whitney, 1984), and between the Excelsior Mountains and Owens Valley ruptures (the White Mountains seismic gap; Wallace et al., 1983).

Southern Nevada Transverse Zone

The SNTZ is a diffuse east-west zone of seismic activity that appears to merge with the Intermountain Seismic Belt (Smith and Sbar, 1974) in southwestern Utah to the east and extends into the Walker Lane belt in the vicinity of Mammoth Lakes in the west (Slemmons et al., 1965) Figure 3.1.1.1.4.1-2. This zone has also been referred to as the east-west seismic belt by Smith (1978). Rogers et al. (1991) suggest that this zone has considerable north-south extent and they propose including activity as far north as 40° north latitude and as far south as 35° north latitude within the SNTZ. For the purpose of this report, the SNTZ is restricted in width to an arcuate belt of seismicity approximately 150 km wide extending from the southern region of the Intermountain Seismic Belt to the Mammoth Lakes area. Yucca Mountain lies within the SNTZ. This definition excludes the seismicity in the vicinity of Lake Mead and the Mojave province of southeastern California. A discussion of seismic activity around Lake Mead, both induced and tectonic, is deferred to Subsection 3.1.1.2.4.1.

The SNTZ includes a region of northeast and east-northeast trending, left lateral faults that extend discontinuously across Nevada at this latitude. Some of these structures are inferred to have been active in the Quaternary, such as the Pahranagat Shear Zone. The southern boundary of the SNTZ roughly coincides with the southern terminus of Quaternary volcanic

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activity in the Great Basin (Eaton, 1982). Historic earthquakes in the SNTZ are of moderate magnitude and there is no documented surface rupture. These events are the 1902 Pine Valley, Utah, earthquake (M_L 6.3), the 1966 Caliente-Clover Mountain, Nevada, earthquake (M_L 6.0; see Subsection 3.1.1.1.4.2.2), and the 1992 Little Skull Mountain, Nevada, earthquake (M_L 5.6, see Subsection 3.1.1.2.4.1) near the potential repository site.

Garlock fault

The Garlock fault is a major seismotectonic feature of southern California (see Subsection 3.1.1.1.3.1.2). Figure 3.1.1.1.4.2-1 shows the dimensions of the zone as defined by Astiz and Allen (1983) and the seismicity in the area for the period 1932-1981. The Garlock fault represents a significant seismic-source zone, because of the difference in style of deformation compared to neighboring provinces. The occurrence of large prehistoric left-lateral strike-slip earthquakes and the modern left-lateral and reverse slip earthquakes distinguishes it from adjoining source zones. The Garlock fault is 265 km long, the longest contiguous left-lateral structure in the Great Basin, and extends to within 120 km of Yucca Mountain. The Garlock fault has been described by Astiz and Allen (1983) as consisting of two segments with distinct seismological and geological characteristics. The western segment is 150 km long extending from Gorman, California, to the fault's midpoint at Rand, California. This segment is characterized by relatively complex fault traces, low seismic activity (between 1932 and 1981; Astiz and Allen, 1983), and aseismic creep. The eastern segment extends 155 km from Rand to the Avawatz Mountains and has a relatively simple fault trace, a lower level of seismic activity, and no documented creep.

Abundant Holocene and Quaternary scarps and offset geomorphic features along the Garlock fault suggest that the fault has produced large earthquakes in the past (McGill and Sieh, 1991). Although no moderate or large historic earthquakes can be definitively associated with the Garlock fault, the eastern segment may have been the source for the M_L 6.1 1916 southern Death Valley earthquake which Slemmons et al. (1965) have located in the Avawatz Mountains area. A survey of seismicity during the period 1932 to 1981 (Astiz and Allen, 1983) and recent seismicity indicates (Subsection 3.1.1.1.4.2.1; Figure 3.1.1.1.4.2-1) that the Garlock fault is still active. Focal mechanisms determined for selected events from 1978 to 1981 by Astiz and Allen (1983) show left-lateral or reverse slip on planes subparallel to the trace of the Garlock fault. The P- and T-axes orientations from these mechanisms are north-northeast and west-northwest, respectively. The T-axis orientation is consistent with the extension direction from a geodetic network along the eastern end of the Garlock fault (Savage et al., 1981). The most recent significant event was a M_L 5.4 earthquake in 1988 located several kilometers north of the Garlock fault and 20 km east of the intersection with the San Andreas Fault (Hutton et al., 1991). The fault-plane solution is consistent with left-lateral reverse motion on a steeply dipping plane (McGill and Sieh, 1991).

3.1.1.1.4.2.1 Distribution of Seismicity

This section will summarize the characteristics of seismicity within the region. The purpose of this discussion is to provide a basis for the development of a regional seismotectonic

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model and to describe the variation in earthquake activity that may have an impact on seismic-hazard assessment at Yucca Mountain. The discussion focuses on seismicity in the Great Basin. Although they are encompassed by the 320 km radius, earthquakes along the San Andreas fault system are not discussed because they lack relevance to hazard issues at the potential site.

Epicentral Distribution

Figure 3.1.1.1.4.2.1-1 is an epicentral plot of earthquakes of magnitude 5 and greater (or intensity VI and greater) from 1850 to 1992 that are within 320 km of the potential repository site. Only significant aftershocks (M > 6) are included on this map to give a clear picture of where seismicity is clustered. Explosions, UNEs, and seismic events related to UNEs, are removed from this compilation. Information on the location of these induced events is from a study by Vortman (1991). Table 3.1.1.1.4.2.1-1 [INN 3.1.1.1.4.2.1-1] is a listing of magnitude 5 and greater events to accompany the map. The table lists the spatial and temporal origins of the earthquakes, the maximum reported magnitude and/or intensity, and the distance between the epicenter and Yucca Mountain.

The data for Figure 3.1.1.1.4.2.1-1 and Table 3.1.1.1.4.2.1-1 have been compiled from several sources. The main source is catalogs produced by various institutions operating seismic networks in the region. They are the University of Nevada Reno Seismological Laboratory; the University of California, Berkeley; the University of Utah; and the California Institute of Technology. Other data sources are catalogs compiled from regional sources as well as literature references, especially for pre-instrumental time periods. These catalogs are produced by the Decade of North American Geology, California Division of Mines and Geology, and the National Earthquake Information Center, which publishes the Preliminary Determination of Epicenters.

The level of completeness of these catalogs varies. For earthquakes in the 19th century newspaper accounts, anecdotal reports, and Indian legends are often the only sources of information available and exist only for the largest events. This is a particular problem in the sparsely settled Great Basin. Slemmons et al. (1965) discuss the various pre-instrumental sources used in developing their catalog of Nevada earthquakes. This catalog has been incorporated into the catalog of the University of Nevada Reno Seismological Laboratory.

The uncertainties in earthquake location for this time period are great, involving errors of as much as several hundred kilometers for non-surface rupturing events (Rogers et al., 1991). Practical instrumental recording of Nevada earthquakes started in 1916 with installation of a two-component Weichert seismograph (Slemmons et al., 1965). The use of data from a seismograph installed in Salt Lake City, Utah, in 1914 significantly reduced the uncertainty of location. Figure 3.1.1.1.4.2.1-2 summarizes the minimum magnitude of detection for the regions surrounding the site (Engdahl and Rinehart, 1991). Because the regional networks and catalogs overlap in coverage, the time-dependent detection threshold is assessed by region rather than by network. This figure shows that magnitude 5 earthquakes were locatable in Nevada starting in the 1920s, in northern California starting in 1910, and in southern

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California starting in 1932 (Engdahl and Rinehart, 1991). The arrows in Figure 3.1.1.1.4.2.1-2 indicate the time at which modern seismological analysis became possible throughout the regions. This indicates that the quality of instrumentation and station coverage were sufficient to routinely carry out detailed seismological studies.

Figure 3.1.1.1.4.2.1-1 shows that the seismicity in the western Great Basin is concentrated in the CNSB and the SNGBZ. Seismicity is also concentrated in the southern Mojave Desert and along part of the San Andreas fault zone. The Garlock fault is relatively aseismic for this time period as is much of the southern Great Basin. The SNTZ is characterized by a sparse distribution of moderate events between 38° and 36.5° N. In addition to these zones, earthquakes are diffusely distributed throughout much of Nevada, occurring in clusters of small magnitude earthquakes or as isolated events. None of the large historic earthquakes in the region have occurred outside the major seismic-source zones. There are also areas that have little or no seismic activity, such as the Death Valley area, the extreme southern part of Nevada, and the Yucca Mountain area (Gomberg, 1991a and b). The characteristics of seismicity in the vicinity of Yucca Mountain are discussed in Subsection 3.1.1.2.3.1.

Hypocentral Distribution

Figure 3.1.1.1.4.2.1-3 shows the distribution of focal depths for some areas in the Great Basin. The majority of earthquakes are located less than 20 km deep. More than 95 percent of the events occur in the upper 15 km of the crust (Rogers et al., 1991). In this upper region, hypocentral concentrations display considerable variability depending on the seismotectonic setting. Mainshock focal depths for large events, however, tend to be located at depths between 8 and 15 km (Doser and Smith, 1989). In general, seismicity in the northern Great Basin appears to occur at deeper crustal levels (Figure 3.1.1.1.4.2.1-3 top) than in the southern Great Basin (Figure 3.1.1.1.4.2.1-3 bottom). The relatively shallow depth distribution of earthquakes in the Great Basin is in contrast to two adjoining regions, the Sierra Nevada and the Colorado Plateau. Because earthquakes represent a brittle mode of deformation Smith and Bruhn (1984) have inferred that the Great Basin brittle-ductile transition occurs at a depth around 15 km, much shallower than in the adjacent provinces.

Seismic Gaps

In Subsection 3.1.1.1.4.2 the existence of seismic gaps in Great Basin seismotectonic provinces was noted. Three major earthquakes (M 7.2 to 7.8) have occurred along the CNSB, two (M 7.0, 8.0) along the SNGBZ, and one in the Mojave Desert (M 7.6). Wallace (1978, 1981), Hill et al. (1985), and Astiz and Allen (1983) have suggested that seismic gaps may exist between the rupture zones of some of these events. The concept of a seismic gap assumes stationarity in the rate of strain accumulation in space and time. For a gap to be significant, there must exist structures of sufficient length to generate a large earthquake in the zone of active tectonism. Although the existence and significance of seismic gaps seems to have been well established for plate-boundary regions (Nishenko and McCann, 1981), it is not certain how well this concept applies to a broadly deforming intraplate region such as the Great Basin (Hill et al., 1985). The localization of deformation in the seismotectonic

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provinces has led Wallace (1978, 1984) to assume continuity of strain accumulation between areas of historic rupture. The presence of contiguous structures connecting the rupture zones is not always evident at the surface, but such structures are assumed to exist at depth. Wallace (1978) has identified three seismic gaps in the western Great Basin; the Stillwater, the White Mountains, and the southern Owens Valley (Figure 3.1.1.1.4.2.1-4). Thenhaus and Barnard (1989) have proposed the existence of a fourth gap in the northern CNSB (Figure 3.1.1.1.4.2.1-5). These gaps occur between the ruptures zones of major historic earthquakes and contain structures that show evidence of prehistoric activity.

Sonoma Range Seismic Gap. The Sonoma Range seismic gap extends 40 km from the northern termination of the M_s 7.7 1915 Pleasant Valley earthquake to the Krum Hills near Winnemucca in the north (Thenhaus and Barnard, 1989; Figure 3.1.1.1.4.2.1-5). The Krum Hills are located in an area traversed by a N60°W striking structural zone which separates domains of oppositely tilted range blocks to the north and south (Stewart, 1980b). This zone is one of three in the Great Basin, termed "extensional accommodation zones" by Thenhaus and Barnard (1989) who interpret them to be barriers to rupture along north-south striking fault zones. The Sonoma Range gap terminates in the northern-most accommodation zone. The gap encompasses an area of generally west-dipping Quaternary faults with the latest scarps dated at 0.012 Ma (Wallace, 1979). Rupture of this gap would complete the pattern of strain release between the northern and central accommodation zones. Seismicity in this region is at a lower level than parts of the CNSB to the south and similar to the background rate of seismicity in the rest of Nevada (Rogers et al., 1991).

Stillwater Seismic Gap. The Stillwater seismic gap (Figure 3.1.1.1.4.2.1-4) lies between the southern termination of the Pleasant Valley earthquake and the northern termination of the M_L 6.8 1954 Dixie Valley earthquake, a distance of 40 km (Wallace, 1978, 1984). The distribution, number, and size of prehistoric scarps in the gap is similar to that in the rupture zones to the north and south; however, no scarps are younger than about 300 years (Wallace and Whitney, 1984). Because of the low level of seismic activity in the area since instrumental monitoring began, Wallace and Whitney (1984) have concluded that rupture along this gap is not imminent. The Stillwater seismic gap lies in the central accommodation zone (Thenhaus and Barnard, 1989). In their role as rupture barriers, the accommodation zones may act to distribute slip from large earthquakes along a broader area relative to the primary rupture zone. This creates a zone of diffuse scarps and disrupted structures with relatively low topographic relief, such as that found in the South Hills separating the Pleasant Valley and Dixie Valley ruptures (Thenhaus and Barnard, 1989). The implication is that slip from these two historic events may have been absorbed by the accommodation zone, diminishing the significance of the Stillwater seismic gap.

White Mountains Seismic Gap. The 130 km-long White Mountains seismic gap (Figure 3.1.1.1.4.2.1-4) lies between the M_L 7.2 1932 Cedar Mountain rupture zone and the M_s 8.0 1872 Owens Valley rupture zone (Wallace, 1978). These two historic earthquakes are discussed in more detail in the next section. The gap is in an area of complex faulting and many of these structures exhibit Quaternary activity (Wallace, 1978). Faults along the base of the White Mountains are long enough to be considered capable of generating large

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earthquakes. Hill et al. (1985) considered this region to be a temporal seismic gap based on the increase in activity of the Mammoth Lakes area to the west and the distribution of moderate-magnitude earthquakes to the north and east of the northern White Mountains. The 1986 Chalfant Valley sequence, which produced some surface rupture along this zone, only partially filled this gap (see Subsection 3.1.1.1.4.2.2).

Southern Owens Valley Seismic Gap. The southern Owens Valley seismic gap (Figure 3.1.1.1.4.2.1-4) extends from the southern termination of the 1872 rupture zone to the Garlock fault (Wallace, 1984). The Sierra Nevada frontal fault is considered to be the capable structure (Wallace, 1984). This zone is seismically active at present (Figure 3.1.1.1.4.2.1-1). Walker Pass in the southern Sierra Nevada was the location of a M_L 6.3 earthquake in 1946 (see Table 3.1.1.1.4.2.1-1) and a sequence of events in 1961. The 1946 mainshock was located at the eastern edge of the Sierra Nevada, possibly on the frontal fault, at a depth of 22 km (Chakrabarty and Richter, 1948). The largest event in the 1961 sequence was a M_L 5.3 earthquake located in the same region with a focal depth of 13 km (Dollar and Helmberger, 1985). The southern Owens Valley region is active at a microseismic level, although recent epicentral maps show these sequences being located east of the Sierra Nevada with the frontal fault north and south of Walker Pass remaining very quiet (Jones and Dollar, 1986; Hutton et al., 1991).

Garlock fault. The Garlock fault has not experienced any significant historic events along its length, except for a possible association with the 1916 southern Death Valley event (see Subsections 3.1.1.1.4.2 and 3.1.1.1.3.1.2 for discussions of the Garlock fault). Comparing energy released by historical seismicity to inferred Holocene geologic offset has led Astiz and Allen (1983) to suggest that the Garlock fault is a seismic gap and capable of generating future large earthquakes, particularly at the eastern end (Figure 3.1.1.1.4.2-1). McGill and Sieh (1991) have estimated earthquake magnitudes for several rupture scenarios along this fault, ranging from M_w 6.6 for slip on shorter segments to M_w 7.8 for rupture of the entire Garlock fault. Astiz and Allen (1983) suggest that the western segment of the fault is creeping aseismically, which means that its status as a seismic gap is doubtful (strain is not accumulating). They consider the eastern segment to be locked, therefore, it is more likely to rupture in the future. McGill and Sieh (1991) estimate that this eastern segment is capable of a M_w 7.5.

Focal Mechanisms

Figure 3.1.1.1.4.2.1-6 shows focal mechanisms for M≥ 5 Great Basin earthquakes derived from body-wave inversions for the time period from 1932 to 1986 (Rogers et al., 1991). Figure 3.1.1.1.4.2.1-7 shows focal mechanisms for earthquakes in the Great Basin recorded by local seismograph networks, compiled from various sources (Rogers et al., 1991). Focal mechanisms for individual historic earthquakes is discussed in Subsection 3.1.1.1.4.2.2 and focal mechanisms for smaller earthquakes in the vicinity of Yucca Mountain are discussed in Subsection 3.1.1.2.3.1.

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All large Great Basin earthquakes that have been studied in detail (Smith et al., 1985; Doser, 1986, 1987, 1988; Doser and Smith 1989; Rogers et al., 1991) indicate faulting on steeply-dipping (greater than 35°) fault planes that penetrate the upper 15 km of crust. The average dip for large normal faulting Great Basin earthquakes is 44° (Doser and Smith, 1989; Thatcher and Hill, 1991). In general, mainshock hypocenters for M > 7 earthquakes in this region can be located on the down-dip projection of the surface rupture, suggesting that large Great Basin events occur on steeply-dipping planar faults at depths less than about 15 km (Rogers et al., 1991). Most microearthquake activity and aftershocks also occur in steeply-dipping volumes of rock, presumably related to high-angle fault zones (Rogers et al., 1991).

The rupture mechanisms for events of all magnitudes range from strike-slip (dextral predominates over sinistral) to normal and includes solutions with varying amounts of oblique-slip (Rogers et al., 1991). Lateral slip tends to occur on faults that are oriented north (dextral-slip) or east-northeast (sinistral-slip). Dip-slip generally occurs on faults trending northeast. Oblique-slip occurs on faults with intermediate orientation. For large Great Basin earthquakes, the amounts of normal, strike-slip, and oblique-slip faulting are nearly equal (Doser and Smith, 1989).

3.1.1.1.4.2.2 Significant Historical Earthquakes

This section summarizes what is known about the seismological and geological characteristics of significant historical earthquakes. These events represent a range of faulting behavior that could be encountered in future ruptures on faults in the vicinity of the potential repository. Figure 3.1.1.1.4.2.2-1 shows the location of these earthquakes. These events are considered significant, because they are either surface rupturing earthquakes or represent the largest event in a particular seismic-source zone. Only the Garlock fault has not experienced a significant historic event.

Owens Valley, California, 1872

On March 26, 1872, one of the three largest historic earthquakes in California occurred in the vicinity of the town of Lone Pine located in the Owens Valley (Figure 3.1.1.1.4.2.2-2). This earthquake is considered by dePolo et al. (1991) to be the largest historic event in the Basin and Range province. Magnitude estimates range from M_w 7.8, based on geologic data (Hanks and Kanamori, 1979), to M_s 8.0 based on felt area (Oakeshott et al., 1972). The earthquake was felt over most of California and Nevada and probably in places in Utah and Arizona (Richter, 1958). In Lone Pine, most of the adobe dwellings were destroyed, 27 people were killed and 54 were injured (Oakeshott et al., 1972). Because seismographs were not installed in the western United States until 1887 (Richter, 1958), detailed seismological information is not available for this event.

This earthquake caused surface ruptures along 90 to 110 km of the Owens Valley fault (dePolo et al., 1991; Figure 3.1.1.1.4.2.2-2). The fault has an average strike of 340° and a dip of 80° ± 15° east-northeast (Beanland and Clark, 1993). Right-lateral slip produced by the 1872 rupture averaged 6 ± 2 m, with a maximum of roughly 10 m near Lone Pine, and

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normal slip averaged 1 ± 0.5 m (Beanland, 1993). The average and maximum net oblique-slip of 6.1 ± 2.1 m and 11 m provides a basis for an estimated seismic moment of 1.8 to 4.4 x 10^{27} dyn-cm (Beanland and Clark, 1993). A maximum vertical displacement of about 4 m occurred along the eastern base of the Alabama Hills, west of Lone Pine, where the scarps formed a graben (Beanland and Clark, 1993; LPS on Figure 3.1.1.1.4.2.2-2). Richter (1958) suggests that these structures may be a secondary feature of the rupture, because they are formed primarily in alluvium and die out in bedrock. Some left-lateral displacement was observed on related fault strands near Independence and Lone Pine (Richter, 1958).

The main surface trace of the fault consisted of a relatively straight section of strike-slip motion in the center of the Owens Valley graben (B-C in Figure 3.1.1.1.4.2.2-2) with diffuse zones of oblique to normal faulting at the northern and southern ends (A-B and C-D, respectively, in Figure 3.1.1.1.4.2.2-2; dePolo et al., 1991). At the southern end of the fault zone, strong ground motion caused extensive liquefaction of unconsolidated sediments around Owens Lake (Vittori et al., 1993). The northern and southern ends of the rupture terminate near extensional basins and both end-regions are associated with late Quaternary basaltic volcanic fields. Slip at the northern end is distributed into small closed basins (dePolo et al., 1991; A in Figure 3.1.1.1.4.2.2-2). Slip in the southern end is distributed through Owens Lake, an extensional basin formed by right-steps in the Owens Valley fault zone, and continues south to the Coso Range at point D (Vittori et al., 1993; Figure 3.1.1.1.4.2.2-2).

Wonder, Nevada, 1903. In the fall of 1903, an earthquake accompanied the rupture of the northerly trending Gold King fault on the eastern side of Dixie Valley (WON in Figure 3.1.1.1.4.2.2-1). Information about this event is incomplete; however, it represents an important event in the 60-year history of stress release along the CNSB. An investigation by Slemmons et al. (1959) reports that surface faulting extends from 5 km to 16 km. An unpublished report by F.C. Schrader in 1911 describes fissures up to 1.5 m wide and 1.5 m deep developed in alluvium. Rogers et al. (1991), estimate the magnitude at 6.5, based on the rupture length. This event occurred in the same region as the 1954 Fairview Peak - Dixie Valley earthquakes and displacement on the Gold King fault also occurred in that sequence.

Cedar Mountain, Nevada, 1932. On December 21, 1932, the M_s 7.2 Cedar Mountain earthquake occurred in Gabbs Valley, causing surface displacement there and in Stewart Valley and Monte Cristo Valley to the south (Abe, 1981; dePolo et al., 1991; Figure 3.1.1.1.4.2.2-3). The earthquake resulted in about 61 km of discontinuous faulting in a belt 6 km to 14 km wide trending south-southeast from the epicenter. This event ruptured parts of the Stewart - Monte Cristo Valley fault zone (Molinari, 1984; see Subsection 3.1.1.1.3.1.2) and several unnamed faults in the area. Displacement on the faults was principally right-lateral with tensional fractures and minor graben distributed in an en echelon pattern between fault strands. Only half of the 1932 surface ruptures broke along obvious older scarps on Quaternary or Tertiary deposits, superimposing west-facing scarps as high as 60 cm on older scarps (dePolo et al., 1987; Yount et al., 1993a). There is abundant geomorphic evidence for right-lateral normal slip for the 1932 rupture (Gianelli and Callaghan, 1934), with measurements of vertical displacements as high as 0.5 m, horizontal displacements as large as 1.8 m, and 3:1 ratios of lateral to vertical slip (dePolo et al., 1987, 1988, 1991).

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The northern termination of the 1932 rupture in Gabbs Valley is characterized by a broad zone of relatively short ruptures trending north-northeast with displacements less than a few decimeters (dePolo et al., 1991; A in Figure 3.1.1.1.4.2.2-3). One of these ruptures occurred on a fault that experienced displacement in the 1954 Fairview Peak earthquake (dePolo et al., 1991). To the south of Gabbs Valley is an area in which surface rupture did not occur in the 1932 event (B in Figure 3.1.1.1.4.2.2-3). This area contains folds in Quaternary and Tertiary sediments not observed in older basement rocks (Molinari, 1984), which may indicate that coseismic displacement at depth is accommodated by surficial folding (dePolo et al., 1991). The longest continuous strand of faulting (17 km) occurred in the northern part of Monte Cristo Valley and exhibited right-lateral oblique-slip motion (Molinari, 1984). There are additional distributed fault scarps in the southern area. The southern end of the rupture zone terminates in an extensional basin in the center of Monte Cristo Valley (C in Figure 3.1.1.1.4.2.2-3).

The source parameters derived from seismological analysis differ somewhat from those suggested by the geologic data. In a study of the 1932 earthquake based on inversion of body-waves, Doser (1988) determined the main shock to be composed of two dextral-slip sources occurring about 20 seconds apart. The first source was a M_w 6.7 event located at 14 km below the surface and the second was a M_w 6.6 event located at a depth of 12 km, about 18 km south of the first (Doser, 1988; Doser and Smith, 1989). The focal mechanisms for both events indicate nearly pure strike-slip motion with the preferred nodal plane orientations (based on the trend of the surface ruptures) striking approximately due north (Doser, 1988; Figure 3.1.1.1.4.2.2-3).

A moment magnitude of 7.1 derived from the geologic data (Doser and Smith, 1989) is larger than the moment magnitude of 6.8 calculated from the combined subevents (Wells and Coppersmith, 1993). It is possible that energy was released at frequencies longer than those used in the body-wave analysis. The assumptions implicit in determining moment from geologic data may not be valid in this case. If some of the ruptures were caused by subsequent events, the moment estimate would be lower. There is also no suggestion in the focal mechanisms of the 3:1 ratio of strike-slip to dip-slip motion observed by dePolo et al. (1987). The body-wave focal mechanisms indicate that the greatest moment released occurred along a deep strike-slip fault (Doser, 1988). Analysis of the waveforms suggest that the rupture initiated in the north and propagated south for about 10 sec, stopping in the area where the strike changes in the surface-rupture trend (Doser, 1988; B in Figure 3.1.1.1.4.2.2-3). There, a second event occurred ten seconds later, rupturing to the south (Doser, 1988). Doser (1988) does not exclude the possibility of a third subevent rupturing in a manner that would explain some of the dip-slip motion observed at the surface, but if it exists, its presence is masked in the coda of the first two subevents.

This earthquake was the first in a series of seven moderate events to occur in this part of the CNSB from 1932 to 1939. The fact that surface rupture propagated through three different valleys, crossing two drainage divides, indicates that it was not a range-front faulting event. This may be due to its location in the Walker Lake section of the Walker Lane belt. The Walker Lake section is characterized by subdued topography and several prominent

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northwest-trending right-lateral faults (Stewart, 1988). The trend of these faults is about 40° counterclockwise to the trend of faults in the northern Great Basin. The fact that the 1932 ruptures occur on more northerly trends suggests the present stress field in the Walker Lake section is similar to the northern Great Basin.

The 1932 event is an example of a complex distributive rupture that displaced multiple faults (dePolo et al., 1991). Deformation in the area is likely driven by northwest-southeast extension, accommodated on north- and northeast-trending structures. Focal mechanisms indicate right-lateral oblique-slip motion on north striking nodal planes. Also present in the area are caldera structures that may have an influence on the distribution of stress (Doser, 1988). These factors have also been identified in the YMR by Ramelli et al. (1988). The similarities in seismotectonic characteristics between Yucca Mountain and Stewart-Monte Cristo fault zone have led dePolo et al. (1989) to suggest that the 1932 Cedar Mountain earthquake could be a possible historic analog for earthquakes in the Yucca Mountain area.

Excelsior Mountains, Nevada, 1934. On January 30, 1934, a M_L 6.3 (M_w 6.1) earthquake occurred in the Excelsior Mountains (Slemmons et al., 1965; Doser, 1988). This is the second largest event in the southern section of the CNSB and it was located about 60 km west-southwest of the 1932 rupture in Monte Cristo Valley (Figure 3.1.1.1.4.2.2-1). It occurred in the Excelsior-Coaldale section of the Walker Lane belt, which is characterized by northeast-trending faults with late Cenozoic left-lateral displacement that truncate northwest-trending right-lateral faults (Stewart, 1988). The mainshock was preceded in approximately 45 minutes by a M_L 5.6 foreshock (Slemmons et al., 1965). Callaghan and Gianella (1935) observed a northeast-trending surface rupture 1.4 km long near the crest of the Excelsior Mountains. The surface rupture had predominantly down-to-the-north normal displacement of less than 13 cm and some evidence of sinistral displacement (Callaghan and Gianella, 1935). DePolo et al. (1987) suggest that this surface rupture is sympathetic in origin rather than primary. Similar sympathetic ruptures have been observed in several other earthquakes in the Mammoth Lakes and Chalfant Valley sequences (dePolo et al., 1987).

Inversion of body-waves from this event suggests a preferred nodal plane striking east-northeast, dipping 40° to the northwest (Doser, 1988; Figure 3.1.1.1.4.2.2-4, top). This focal mechanism is poorly constrained and the range in solutions permits a small amount of left-lateral motion on the fault which would be consistent with field observations. Using body-waves, the earthquake hypocenter has been located about 5 km west of the surface rupture at a depth of 14 km (Doser, 1988). Comparison of waveforms between the mainshock and the foreshock suggest the foreshock had a similar focal mechanism (Doser, 1988). A poorly constrained focal mechanism for a M_L 5.5 earthquake on August 9, 1943, approximately 40 km southeast of the 1934 mainshock, has been interpreted as being consistent with the 1934 event (Doser, 1988). Ryall and Priestly (1975) determined a composite focal mechanism from earthquakes in the Excelsior Mountains region from 1969 to 1971 (Figure 3.1.1.1.4.2.2-4, middle). This mechanism indicates right-lateral oblique-slip on a northerly-striking plane or left-lateral oblique slip on a northeast plane. The latter orientation is roughly consistent with the range of preferred nodal-plane orientations from the 1934 event (Doser, 1988). Focal mechanisms for several moderate earthquakes occurring between 1980 and 1982,

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southwest of the 1934 mainshock, indicate either right-lateral slip on northwest-striking planes or left-lateral slip on northeast-striking nodal planes (Vetter, 1990; Figure 3.1.1.1.4.2.2-4, bottom).

The spatial and temporal relation between the 1932 Cedar Mountain and the 1934 Excelsior Mountains earthquakes suggests a possible interaction between the two events. Doser (1988) has raised the possibility that the 1932 Cedar Mountain earthquake produced temporary changes in the stress field influencing in the 1934 event. This may explain the differences in the focal mechanisms between the 1932 earthquake and later events. The next two historic earthquakes are also examples of an interdependence between events in the CNSB.

Rainbow Mountain-Stillwater, Nevada, 1954. The Rainbow Mountain earthquakes (M 6.6, M 6.4; Doser, 1986) of July 6, 1954, initiated a six-year period of activity in the CNSB in which ten earthquakes greater than M 5.5 occurred, including four surface-rupturing events (Figure 3.1.1.1.4.2.2-1). The July 6 events were followed on August 24 by the M 6.8 Stillwater earthquake (Doser, 1986). All of these events produced measurable ground displacement in the region.

The first earthquake in the sequence occurred on July 6 at 1113 UT, about 10 km below the surface and 20 km east-southeast of the surface trace of the fault (Doser, 1986; Doser and Smith, 1989). This event produced 18 km of surface rupture along the east side of Rainbow Mountain (dePolo et al., 1991; Figure 3.1.1.1.4.2.2-5). East-facing scarps ranging from 3-30 cm high formed along the fault zone and extended into adjacent Quaternary basins (Tocher, 1956; Bell, 1984). At the northern and southern terminations of this fault zone, the displacement dies out in young alluvial deposits (dePolo et al., 1991; C and D in Figure 3.1.1.1.4.2.2-5). No surface lateral offset was observed (Tocher, 1956). A focal mechanism for this event from body-wave inversion shows right-lateral oblique slip along a plane striking 20°-30° west of the surface rupture and dipping 80° to the east (Doser, 1986; Figure 3.1.1.1.4.2.2-5). Doser (1986) has modeled this earthquake by assuming two subevents, 7 sec apart. The first subevent generated the greatest seismic moment (1.3×10^{25} dyn-cm) and dominates the event focal mechanism. The second subevent was located at 7 km depth, had a somewhat lower seismic moment (0.9×10^{25} dyn-cm), and a significant component of dip-slip motion. Doser (1986) suggests the second subevent, because of its shallower depth, is responsible for most of the surficial offset.

At 2207 UT on July 6, a major aftershock of the first event occurred. This earthquake was located about 10 km south-southwest of the mainshock at a depth of 8 km (Doser, 1986; Figure 3.1.1.1.4.2.2-5). Tocher (1956) and Bell (1984) observed ground displacement southeast of the Rainbow Mountain fault, but dePolo et al. (1991) apparently do not consider this to be primary rupture. Doser (1986) determined a focal mechanism for this event that indicates right-lateral oblique-slip motion on a north-northwest striking plane dipping 60° to the east (Figure 3.1.1.1.4.2.2-5).

The August 24, 1954, Stillwater earthquake occurred about 10 km north-northeast of the July 6 mainshock at a depth of 12 km (Doser, 1986; Figure 3.1.1.1.4.2.2-5). This event

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reactivated and increased the heights of fault scarps at the northern end of the July 6 ruptures (dePolo et al., 1991; hatchered lines in Figure 3.1.1.1.4.2.2-5). Ground displacement also extended 34 km to the north across the Stillwater Marsh into the Carson Sink (Tocher, 1956; Bell, 1984). Many of these scarps are east-facing, 50-75 cm high, and appear to be more continuous than those associated with the July sequence (Bell, 1984). Bell (1984) interprets the total rupture length for this event to be at least 40 km. A focal mechanism for this event based on work by Doser (1986; Figure 3.1.1.1.4.2.2-5) suggests right-lateral, oblique-slip on a north-northwest striking plane dipping 50° to the east. Teleseismic waveforms were modeled by a source function consisting of two subevents in the first 10 sec, followed by a third 12 sec later (Doser, 1986). All the subevents were located at the same depth and had the same focal mechanism. Total seismic moment for this earthquake is 6.9×10^{25} dyn-cm giving a moment magnitude of 6.5 (Doser, 1986).

Figure 3.1.1.1.4.2.2-5 shows two aftershocks of the Rainbow Mountain-Stillwater sequence, which occurred in the region prior to the Fairview Peak-Dixie Valley earthquakes (next section). The first, an M_w 5.8 event on August 31, 1954, was located at the northern end of the Stillwater rupture zone and its focal mechanism indicates right-lateral oblique slip on a north-striking plane dipping 58° to the east (Doser, 1986). The second, an M_L 5.5 event on September 1, was located in the epicentral region of the Rainbow Mountain earthquakes. A focal mechanism for this event suggests left-lateral oblique slip on a N20°E plane dipping 70° to the east. Based on an interpretation of geodetic data by Meister et al. (1968), Doser (1986) suggests that this event occurred on an unmapped fault in the area along which east-west extension may have occurred. Vetter and Ryall (1983) have also determined focal mechanisms for several small-magnitude earthquakes in this area that show a component of left-lateral slip.

Fairview Peak-Dixie Valley, Nevada, 1954. On December 16, 1954, two large earthquakes occurred to the east of the Rainbow Mountain and Stillwater earthquakes (Figure 3.1.1.1.4.2.2-6). The first was an M_L 7.3 event along the Fairview Fault east of Fairview Peak, followed four minutes later by an M_L 6.9 event which ruptured the Dixie Valley Fault at the eastern base of the Stillwater Range. Both of these earthquakes produced complex surface ruptures along a diffuse zone from Mount Anna south of Fairview Peak to the northern part of Dixie Valley covering an area 100 km by 30 km. Maximum displacements of approximately 4 m right lateral and 3 m vertical occurred along the Fairview Peak fault zone and more than 2 m vertical occurred in Dixie Valley (Slemmons, 1957). These events were felt over an area of 570,000 km^2 in five western states.

The Fairview Peak earthquake was located 15 km below the surface in the Louderback Mountains, east of Fairview Peak. Doser (1986) reports the focal mechanism to be predominantly dextral slip on a plane dipping 60° to the east. Using body-wave modeling, the source is resolved into three subevents, all showing the same style of oblique, right-lateral, strike-slip motion. Surface faulting occurred on the Fairview Peak Fault, the Gold King Fault, and the Westgate Fault. Surface displacement reaches a maximum in Bell Flat east of Fairview Peak at 3.7 m vertical. Geologic and geodetic data suggest nearly equal

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components of vertical and horizontal motion on the Fairview Peak system, whereas the body wave analysis suggests that dextral slip dominates at the expense of dip-slip motion.

The hypocenter of the Dixie Valley earthquake was located 12 km below the surface, beneath the Clan Alpine Mountains. Because the event occurred only four minutes after the onset of the Fairview Peak earthquake, the first motion information is lost in the earlier event's coda. Body wave modeling by Doser (1986) gives a best fit to a mechanism striking north-northwest, dipping 50° to the east and exhibiting pure dip-slip motion. The fault scarps produced along the Dixie Valley fault at the base of the Stillwater Range dip between 55 - 75° to the east in bedrock (Slemmons, 1957), with a maximum vertical offset of 2 m. There is some evidence of strike-slip faulting at the southern end of the rupture, where 2 m of left lateral slip was reported (Slemmons, 1957). This event ruptured a zone 43-47 km long on the west side of Dixie Valley and generally followed the prehistoric scarps on the west side of the valley as did the Fairview Peak event.

Caliente-Clover Mountains, Nevada, 1966. On August 16, 1966, an M_L 6.0 earthquake occurred near Caliente, Nevada, approximately 210 km east-northeast of Yucca Mountain (Figure 3.1.1.1.4.2.2-7). It is the largest historical earthquake to occur in the SNTZ. Doser and Smith (1989) estimate a magnitude of M_w 5.3 for this event. There was no surface faulting reported for this event (dePolo et al., 1991).

The source depth of this event is estimated at 6 km from depth phases (Wallace et al., 1983). Focal mechanism studies by Smith and Sbar (1974) and Wallace et al. (1983), indicate strike-slip motion on steeply-dipping planes oriented either north-northeast or west-northwest (Figure 3.1.1.1.4.2.2-7). Resolution of the aftershock trends is not sufficient to provide an unequivocal determination of the failure plane. Rogers et al. (1983) prefer the north-northeast interpretation because of the existence of north-trending structures in the region and an alignment of aftershocks from a study by Barcher (1967). However, additional aftershock monitoring by Beck (1970) shows alignments of aftershock clusters in an east-west direction. Smith and Sbar (1974) note the presence of east-west trending faults in the region that has both right- and left-lateral, strike-slip motion, offsetting Quaternary basalt flows and Holocene alluvial deposits.

Mammoth Lakes, California, 1978-1988. Since 1978, the area near Bishop, California, on the eastern side of the Sierra Nevada has been one of the most seismically active areas in the region. Five earthquakes of magnitude greater than or equal to 6 have occurred in the region around Long Valley, the remnant of a volcano which catastrophically exploded around 0.7 Ma. The sequence began in October, 1978, with an M_L 5.8 earthquake (event 1 in Figure 3.1.1.1.4.2.2-8) located midway between Bishop and Mammoth Lakes, followed 18 months later by the swarm-like sequence of four earthquakes (M_L 6.5, M_L 6.0, M_L 6.7, M_L 6.3; events 2-5 in Figure 3.1.1.1.4.2.2-8) within two days in May 1980. This sequence was accompanied by inflation of the resurgent dome in the west-central part of the Long Valley caldera starting in mid-1979 (Hill et al., 1985). This activity continued with moderate earthquake swarms in the southern section of the caldera that included spasmodic tremor sequences usually associated with magma injection at depth (Ryall and Ryall, 1983). In 1986, activity increased

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with the occurrence of the Chalfant sequence (see next section) to the east, near the base of the White Mountains.

Most focal mechanisms for these earthquakes show east-northeast/west-southwest extension in the southern part of the Long Valley region. This is in contrast to the usual east-west extension direction common to the rest of the SNGBZ. Focal mechanisms for the 1978-1980 sequence are shown in Figure 3.1.1.1.4.2.2-8. These solutions are based on an analysis of first motions from short-period P-waves recorded on seismic networks in California and Nevada (Cramer and Toppozada, 1980). Strike-slip mechanisms predominate for earthquakes shallower than 6-9 km and oblique or normal-slip mechanisms for deeper events. Complications in some of the patterns of radiated energy from the Mammoth Lakes sequence have led some to speculate that they are directly related to a magma injection event (Julian and Sipkin, 1985).

Source depths for the events in this sequence range from very near surface to 12 km, with depths of the four M 6 earthquakes ranging from 7-10 km. Surface rupture did occur during the May 1980 sequence-trending north-northwest in a zone 20 km long and up to 10 km wide within the Long Valley caldera and extending southeastward from the caldera along the Hilton Creek Fault. The pattern of fractures was complex involving both normal faulting (down to the east) and tensional cracking. These surface features in the caldera were generally unrelated to known faults and may be caused by ground failure during the earthquakes. The largest displacements were about 25 cm (both vertical and horizontal), but most were much smaller. Strike-slip offset was a minor component of the surface rupture, observed in only a few places, and was right lateral on north to northwest trending faults.

Chalfant Valley, California, 1986. On July 21, 1986, an M_L 6.6 earthquake occurred in Chalfant Valley in eastern California about 15 km north of Bishop, California (Figure 3.1.1.1.4.2.2-9). This region has been active since 1978, when the Mammoth Lakes sequence started, but most of that activity has been confined to the southern edge of the Long Valley caldera and the eastern Sierra Nevada front to the south. This earthquake was located southeast of Long Valley at the western base of the White Mountains. The source-depth was located 11 km below the surface and the focal mechanism (Uhrhammer, 1986) indicates right lateral slip on a plane oriented north-northwest dipping 70° to the southwest (Figure 3.1.1.1.4.2.2-9). There was a small component of dip slip motion in the event, with the east side of the fault moving up. The T-axis orientation is generally east-west for the mainshock (Smith et al., 1986). Surface rupture associated with the mainshock extended about 10 km along the White Mountains fault zone (Lienkaemper et al., 1986) with a maximum amount of right lateral slip of 4 cm.

Prior to this earthquake, seismic activity had been increasing for 18 days culminating in a M_L 5.7 foreshock 24 hours before the mainshock. This foreshock exhibited left-lateral oblique movement on a northeast trending fault plane (Smith and Priestly, 1988) and was located 7 km below the surface. Ten days after the mainshock, an M_L 6 aftershock occurred at a depth of 9 km. The earthquake generated right-lateral slip on a northwest trending plane similar to the mainshock.

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Landers, California, 1992. The largest earthquake in southern California in 40 years occurred on June 28, 1992, near Landers, California, in the south-central section of the Mojave block (Figure 3.1.1.1.4.2.2-10). The mainshock was an M_S 7.6 right-lateral, strike slip event, which ruptured sections of several mapped north- to northwest-trending faults and along several unmapped north-trending concealed faults (Hauksson et al., 1992; USGS, 1992). The Landers sequence initiated on April 23, 1992, with the M_L 6.1 Joshua Tree earthquake which was followed by a large aftershock sequence of 6,000 events. The mainshock focus was located in the 5-10 km range and the focal mechanism indicates right lateral strike slip on a nearly vertical plane trending N8°W (Hauksson et al., 1992). The mainshock was followed by an extensive aftershock sequence, which extended 85 km to the north of the mainshock and 40 km to the south, into the Joshua Tree aftershock zone. Aftershock activity was also observed off the main fault trace. The most significant sequence was associated with the M_S 6.7 Big Bear earthquake, which occurred three hours after and 30 km west of the mainshock.

Surface rupture associated with the mainshock extended 85 km from the Little San Bernardino Mountains in the south to the Rodman Mountains in the north. Rupture occurred along parts of the Eureka Peak, Johnson Valley, Homestead Valley, Emerson and Camp Rock faults, in a right-steeping en echelon pattern (USGS, 1992). A 5 km gap in this otherwise continuous zone separates the northern end of the Eureka Peak fault from the southern end of the Johnson Valley fault. No surface rupture occurred on the Pinto Mountain fault which runs through this gap. Minor surface rupture was observed on faults within 30 km on either side of the main rupture zone. The surface ruptures exhibit predominantly right-lateral offset. Measured displacement across the rupture zone averaged 2 to 3 m, with a maximum of 6.7 m observed on the Emerson fault (Rubin and McGill, 1992). Vertical displacements averaged less than 0.3 m, with a maximum vertical throw of 1.75 m on the Emerson fault.

After the Landers earthquake, a sudden increase in seismic activity in the western United States was observed up to 1,250 km from the mainshock (Reasenberg et al., 1992). The most intense cluster of triggered events occurred in the Walker Lane belt, the largest being the M_L 5.6 Little Skull Mountain, earthquake located 20 km from Yucca Mountain, on June 29, 1992 (Reasenberg et al., 1992; Anderson et al., 1992). All of the areas that experienced an increase in activity have a history of persistent seismicity and most are characterized by recent volcanism and geothermal activity, with the exception of the Little Skull Mountain event (Gomberg et al., 1992) and several smaller events in western Nevada (Anderson et al., 1992). Anderson et al. (1992), have proposed dynamic, low-frequency stress, associated with high amplitude, long-period surface waves as the cause of these events. Long-period surface waves have relatively high strains that extend to depths of 10 km or more.

Eureka Valley, California, 1993. [Note: the Eureka Valley earthquake is significant because it may be the closest surface-rupturing event to the potential repository site. Studies are still underway on the seismologic and geologic characteristics of this earthquake; therefore, this section is incomplete. This discussion is included in recognition of the importance of this event and will be updated as data are made available.]

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On May 17, 1993, an M_L 6.1 earthquake occurred 30 km southeast of Bishop, California, in Eureka Valley. The hypocenter was located 9 km below the surface in the southern part of the valley. Preliminary analysis suggests normal faulting on a northeast striking plane, perhaps paralleling a north-northeast trending inferred Quaternary fault in the area. Reconnaissance mapping has delineated a hairline surface crack extending north several kilometers along the playa floor, which may represent surface rupture or may be a secondary feature due to strong ground shaking (Ramelli, personal comm.). Aftershocks have been recorded by the University of Nevada Reno Seismological Laboratory, and field investigations are underway. This event is the largest earthquake to occur in the western Great Basin since the Chalfant Valley event.

3.1.1.1.4.2.3 State of Regional Tectonic Stress Field

In general, contemporary tectonic activity is occurring along structures developed in prior Cenozoic episodes of deformation (Stewart, 1988). To the extent that these features are oriented favorably to the present stress field, strain is relieved on these inherited structures. This implies some uniformity in the orientation of the stress fields throughout the Cenozoic. The orientation and relative magnitude of the minimum horizontal stress appears to be the common element linking past and present tectonic stress fields (Zoback, 1989). This section will briefly review some of the evidence for this. Figure 3.1.1.1.4.2.1-2 shows the focal mechanisms for moderate to large regional earthquakes and Figure 3.1.1.1.4.2.1-7 shows mechanisms from smaller events recorded by local microearthquake networks. A discussion of the relation of the stress field to the overall tectonic environment, as it relates to plate motions in western North America, is presented to Subsection 3.1.1.1.3.1.

A discussion of the tectonic stress field is facilitated by referral to the orientation of the three principal axes of the stress tensor. In this section, the minimum and maximum horizontal stress axes are denoted as S_h and S_H, respectively. The vertical-stress axis is identified as S_V. Strike-slip faulting occurs in an environment where S_H is the maximum principal stress, S_V is intermediate in magnitude, and S_h is the minimum principal stress. Normal faulting occurs in conditions where S_V is the maximum principal stress, S_H is intermediate, and S_h is the minimum stress direction. The orientations and relative magnitudes of these axes can be determined from geologic data, such as the trends of magmatic dikes (S_H is parallel to the elongated trend of the dike, S_h is perpendicular), faulting style (S_h is oriented perpendicular to the fault strike for normal and at 45° to fault strike for strike-slip faulting; S_H is perpendicular for thrust and reverse faulting), analysis of borehole elongation (S_h is parallel to the "breakout" direction), and focal mechanism studies of earthquakes (S_H, S_h, and, S_V determined from orientation of nodal planes). In addition, magnitudes of the stress tensor can be measured directly in hydraulic fracturing experiments in boreholes.

The basin and range structures in the Great Basin were developed in an episode of east-west directed extension, which began about 10 Ma and resulted in the north-south trend of the tilted range blocks (Stewart, 1988). The elevation of many of the ranges in the Great Basin and the Sierra Nevada suggest that considerable vertical movement must have taken place to position the blocks at their current heights. This indicates that in the past normal faulting

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must have predominated, although this period could have been interspersed with episodes of lateral motion. At some point in the Quaternary, a shift to the present tectonic regime occurred.

An evaluation of geologic and seismologic stress indicators for the Quaternary shows the central part of the Great Basin, including the CNSB, to be characterized by northwest-southeast extension (Zoback and Zoback 1980). A more detailed survey of the Great Basin shows some regional variations (Zoback 1989). Stress axes along the SNGBZ determined from earthquake flood mechanism are oriented approximately east-west (VanWormer and Ryall, 1980; Vetter, 1990; and Zoback, 1989). All studies indicate that the maximum and minimum principal stress axes are nearly horizontal, which favors strike-slip faulting. The earthquakes from which these data are derived are distributed throughout the brittle crust to depths of 15 km which indicate a relatively uniform state of stress in the upper crust (Zoback, 1989). In the Walker Lane belt, the orientation of S_h is west-northwest (Zoback and Beanland, 1986; Vetter, 1990). This region is characterized by strike-slip (both dextral and sinistral) and subordinate normal faulting. A hydraulic fracturing experiment at Yucca Mountain in the southern part of the Walker Lane belt gave similar results for the minimum horizontal stress direction (N60°W; Stock et al., 1985). This orientation is similar to that obtained by Rogers et al. (1987) for earthquakes in the vicinity of NTS and to that obtained by Frizzell and Zoback (1987) from fault-slip data on young faults in the area. In the CNSB, S_h is oriented northwest-southeast, based mostly on focal mechanisms of large and moderate earthquakes. Although few focal mechanisms are available from eastern Nevada, alignment of young (Quaternary) volcanic cones also suggest northwest-southeast extension (Zoback, 1989).

[Note: This ends the introduction to this topic. This discussion will be expanded in a future revision to include a detailed analysis of the spatial and temporal variations in the Late Cenozoic stress field.]

3.1.1.1.4.3 Relation of Seismicity to Geologic/Tectonic Structures and Settings

[See Subsection 3.1.1.1.3.1.3, Structural Features and Seismicity]

3.1.1.1.5 Natural Resources

This section presents a summary of our current knowledge of the natural resources present or inferred in the region surrounding Yucca Mountain. The summary includes a brief description of the resources, including metallic, nonmetallic, and energy resources, currently known in the region or could reasonably be postulated to be present in economic or subeconomic quantities. Groundwater is an important resource in this area for both agriculture and human consumption. This resource is discussed in detail in Subsection 3.1.2.2.

The size of the region for each resource to be discussed depends, in part, on the nature and occurrence of the resource.

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[In addition to the discussion of known resources, there will be discussions of the theories on how the resources form and how they relate to the geologic development of the area. This information is required to assess the resource potential of Yucca Mountain and the surrounding area. The purpose of this assessment is to identify resources, the exploitation of which could result in inadvertent intrusion into the repository. It is premature to make that assessment at this time and nothing in the text should be construed as an evaluation of the area's resource potential.]

This section concludes with a description of conceptual models that may be useful in the evaluation of resources discussed earlier in the section. These models will attempt to incorporate and explain data currently known about resources in the area. The models will also provide insight into the processes that have operated during the development of the area.

Yucca Mountain includes Cenozoic and underlying Paleozoic rocks. Mesozoic rocks are not known to occur beneath Yucca Mountain. Cenozoic rocks include tuffaceous rocks locally covered by varying thicknesses of surficial deposits. The Paleozoic rocks lie 1,200 to several thousand meters below the surface based, on drillhole data (Craig and Johnson, 1984; W. J. Carr et al., 1986) and structural projections of the Paleozoic strata beneath Yucca Mountain (Robinson, 1985).

A definition of mineral or energy resources has been established by the U.S. Bureau of Mines (USBM) and the U.S. Geological Survey (USGS) in Circular 831 (USBM/USGS, 1980). This definition includes as a mineral or energy resource, "a concentration of naturally occurring solid, liquid, or gaseous material in or on the Earth's crust in such form and amount that economic extraction of a commodity from the concentration is currently or potentially feasible" (USBM/USGS, 1980).

A distinction is drawn between identified resources and undiscovered resources. The following terms have been defined by the USBM and the USGS (USBM/USGS, 1980) (as presented below) and the classification approach is illustrated in Figure 3.1.1.1.5-1.

Identified Resources. Resources whose location, grade, quality, and quantity are known or estimated from specific geologic evidence. Identified resources include economic, marginally economic, and subeconomic components. To reflect varying degrees of geologic certainty, these economic divisions can be subdivided into measured, indicated, and inferred.

Undiscovered Resources. The existence of resources, which are only postulated, comprising deposits separate from identified resources. Undiscovered resources may be postulated in deposits of such grade and physical location as to render them economic, marginally economic, or subeconomic. To reflect varying degrees of geologic certainty, undiscovered resources may be divided into two parts:

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Hypothetical resources are undiscovered resources, similar to known mineral bodies, that may reasonably be expected to exist in the same producing district or region under analogous geologic conditions. If exploration confirms their existence and reveals enough information about their quality, grade, and quantity, they will be reclassified as identified resources.

Speculative resources are undiscovered resources that may occur either in known types of deposits in favorable geologic settings where mineral discoveries have not been made, or in types of deposits as yet unrecognized for their economic potential. If exploration confirms their existence and reveals enough information about their quantity, grade, and quality, they will be reclassified as identified resources.

USBM/USGS (1980) further states that "materials that are too low grade or for other reasons not considered to be economic may be recognized and their magnitude estimated, but they are not classified as resources." A separate category, labeled "other occurrences," is included in Figure 3.1.1.1.5-1 for materials of this nature. Further details on this classification system are given in USBM/USGS (1980).

A mineral or energy reserve is that part of the total resource that can be economically extracted or produced at the present time, or at the time the resource evaluation is conducted (USBM/ USGS, 1980). Current economic conditions define the difference between a resource and a reserve. As economic conditions change, production of a mineral reserve could become subeconomic and the mineral would be reclassified as a resource. A commodity classified as a resource could be reclassified as a reserve, as the value of the commodity increases.

3.1.1.1.5.1 Methods of Mineral-Energy-Resource Assessment

Mineral-energy-resource assessments are conducted for a variety of reasons, such as establishing a national policy, location of a company's future exploration target, or projecting future availability of a commodity. Methods used in resource assessments vary widely (Singer and Mosier, 1981) due to several factors, including varied goals, scale or size of the area considered, time available, the acceptable level of uncertainty, and the particular user of an assessment. The methods of groundwater resource assessment are discussed in Subsection 3.1.2.2.

The methods of resource assessment that will be used in evaluation of Yucca Mountain are varied and depend upon a number of factors, including:

a. Abundance of the resource at Yucca Mountain, in the vicinity of Yucca Mountain, and in the surrounding region;

b. Present technological assessment procedures for a particular resource;

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c. Type of resource involved (the resource assessment for gold has many characteristics different from an assessment of geothermal energy); and

d. Type and amount of data available to perform the assessment.

Resource assessments make wide use of two broad overlapping categories of methods: extrapolation and analogy (Singer and Mosier, 1981). A quantitative approach to resource assessment is very desirable, but the immense and varied data base required for statistical treatment or other quantitative and analytical methods is rarely available (Taylor and Stevens, 1983). Many resource-assessment methods and approaches are available (Harris and Agterberg, 1981; Singer and Mosier, 1981; Zwartendyk, 1981; Taylor and Stevens, 1983), but existing quantitative assessments often apply only in situations where a potential ore body has already been discovered and to resource assessments of large scale.

The concept of zones, belts, and knots of mineralization will serve as a starting point in the evaluation of the mineral potential of Yucca Mountain. A statistical analysis of mineral deposits of an area that covered greater than 50 percent of Nevada was attempted by Horton (1966). Horton found that silver and lead fell into statistically significant mineralized belts; whereas other elements considered, including other precious metals and other base metals, did not.

Factors such as location, grade, and tonnage have a profound influence on value and, hence, classification of a commodity as a reserve or a resource. In general, the worldwide distribution of a mineral commodity exhibits an exponential curve, when grade of the potential deposits is plotted against tonnage of the deposits (Brooks and Andrews, 1974). In other words, as the price of a mineral commodity increases or as the cost of extraction decreases, the volume of mineable material increases significantly by factors of 100 or 1,000 (Brooks and Andrews, 1974). Therefore, elements or materials below or equal to average crustal abundances may never be considered reserves. For example, copper occurs in the Topopah Spring Member at concentrations greatly below average crustal abundance (Schuryatz et al., 1986); thus, copper will never be considered a resource at Yucca Mountain. This will remain true, even with large increases in the value of copper or great improvements in extraction technology, because significant resources will always occur elsewhere with a higher grade. Therefore, only those commodities will be assessed that are known to occur either in significant abundance (e.g., zeolites) or in comparable areas or settings, or are presently, or historically have been, mined in the vicinity of Yucca Mountain.

3.1.1.1.5.2 Mineral Commodities Data

Nevada ranks approximately third in value of nonfuel-mineral production in the United States (USBM, 1990). Nevada leads the nation in the production of brucite, magnesite, mercury, silver, and gold (USBM, 1990). Gold is the State's leading commodity in terms of value (Carrillo and Schilling, 1985). Barite, clays, gemstones, gypsum, iron ore, lead, sand, gravel, and crushed stone are (were) also produced in Nevada.

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3.1.1.1.5.2.1 Precious- and Base-Metal Deposits

The uneven distribution of mineral deposits in the earth's crust has led to the idea of metallogenic provinces or zones. This concept has been applied to Nevada in an attempt to characterize the distribution of mineral deposits and to identify mineral deposits that may be characteristic of a particular area.

Roberts (1966, 1971) has identified metallogenic zones in Nevada. He suggests a general division between a western, or precious metal province, and an eastern, or base metal province. The boundary between the two falls roughly between longitudes 116° W and 117° W. The boundary generally follows the Devonian Antler Orogenic Belt and the associated Roberts Mountain Thrust (Figure 3.1.1.1.5.2.1-1).

The western province is characterized by deposits of gold, silver, antimony, and mercury. These deposits are generally hosted by clastic sedimentary rocks and felsic igneous rocks. The eastern province is characterized by lead-zinc deposits hosted by carbonate rocks and porphyry copper deposits. Yucca Mountain falls within the eastern base-metal province, but the most important mineralization in the vicinity of Yucca Mountain is precious metal and gold-silver deposits in Tertiary igneous rocks. In addition to the general mineral provinces, or zones, Roberts also identifies mineral belts that are defined by individual mineral deposits. The belt nearest to Yucca Mountain is the Beatty-Searchlight Belt. This belt is characterized by precious-metal mineralization and Roberts suggests that it may be localized by structures that are part of the Walker Lane zone.

Within the last two decades, economic trends and new geologic discoveries have expanded the precious-metal reserves in Nevada. Increases in the price of gold and silver have provided an impetus for renewal of mining operations and exploration. In turn, new exploration models have successfully proved a revealing diversity of geologic occurrences of silver and gold greater than previously perceived. For example, during this renewed exploration activity the new concept of mobilization and reprecipitation of gold along thrust faults or detachment structures was introduced. These models are revealing new, potentially important environments of precious metals. In addition, bulk low-grade, and disseminated silver-gold deposits (e.g., Carlin, Candelaria, and Goldfield) have been discovered in a variety of rock types, including hydrothermally altered tuff (e.g., Round Mountain) (Sander and Einaudi, 1987).

Many descriptions of the occurrences, locations, and histories of Nevada's gold, silver, and base-metal deposits exist in current literature. A few of these summary articles concerning southern Nevada include Ball (1907), Ransome et al. (1907), Lincoln (1923), Kral (1951), Cornwall and Kleinhampl (1964), USGS/USBM (1964), Albers and Stewart (1972), Cornwall (1972), Bell and Larson (1982), Quade and Tingley (1983), Smith et al. (1983), Kleinhampl and Ziony (1984), Tooker (1985), Mosier et al. (1986), Johnson and Abbott (1987), and Garside et al. (1987). Some of the most important mineralized districts in the vicinity of Yucca Mountain are the Tonopah, Goldfield, and Bullfrog Hills (Rhyolite) districts (Figure 3.1.1.1.5.2.1-2). These deposits were actively mined, beginning around the turn of the

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century, and now are being reassessed or exploited as bulk low-grade gold deposits. The deposits are associated with Tertiary volcanic rocks that are related to andesitic volcanic rocks (Cornwall and Kleinhampl, 1964; Cornwall, 1972; Stewart et al., 1977; Ashley, 1979; Sawkins, 1984). These deposits exhibit a range in mineralization ages from 9 to 23 Ma (Silberman, 1985; White, 1985).

Interpretations of gravity and magnetic data for the region, including Yucca Mountain, are given by Snyder and Oliver (1981), Snyder and Carr (1982, 1984), Carr (1984), USGS (1984), Bath and Jahren (1984), Robinson (1985). Generally, the region around Yucca Mountain is a gravity low (Carr, 1984) corresponding to the thick Cenozoic volcanic cover and the presence of several caldera complexes (Snyder and Carr, 1982; USGS, 1984; Robinson, 1985). However, local gravity and magnetic highs have been interpreted to include the existence of a deep pluton that may extend from northern Bare Mountain to northern Yucca Mountain, to the Calico Hills area (Snyder and Oliver, 1981; Hoover et al., 1982; Carr, 1984; USGS, 1984), and/or reflect a northeast-trending buried ridge of Paleozoic rocks extending from the gravity high of Bare Mountain to Calico Hills (Snyder and Carr, 1984). The inferred pluton could be responsible for some of the metamorphism observed at Calico Hills and Bare Mountain (Carr, 1984). Based on gravity interpretations and magnetotelluric data, the inferred pluton is greater than 2,200 m below Calico Hills (Hoover et al., 1982; Snyder and Carr, 1982), and probably much deeper below Yucca Mountain, if it exists (Robinson, 1985). The age of the inferred pluton is thought by Carr (1984) to be Mesozoic.

Goldfield. The ore deposits near Goldfield, Nevada (Figure 3.1.1.1.5.2.1-2) are classic examples of epithermal bonanza-type gold deposits (Billingsley and Locke, 1941). Important characteristics of these deposits include fracture-control of the mineralization and association with felsic volcanic rocks (Ashley, 1979). The deposits occur within an extensive area of acid-sulfate hydrothermal alteration (Ashley and Silberman, 1976; Silberman, 1985). The Goldfield deposits, which produced 119 metric tons of gold, 42.5 metric tons of silver, and 3,500 metric tons of copper, occur as linear features in volcanic rocks associated with andesite (Albers and Stewart, 1972; Smith et al., 1983).

Outcrops of pre-Tertiary rocks in the area include limited exposures of the Ordovician Palmetto formation and Jurassic intrusive rocks of quartz monzonitic composition. Tertiary rocks include Oligocene and Miocene volcanic, volcaniclastic, and sedimentary rocks.

Oligocene volcanics are dominantly silicic calc-alkaline rocks. They include flows and ash-flow tuffs, which were probably extruded from vents in the immediate vicinity. Overlying these rocks are a sequence of early Miocene volcanics. This sequence of calc-alkaline rocks is dominantly andesitic in composition. Mineralization is spatially and temporally associated with these early Miocene units (Ashley, 1979).

Postmineralization rocks include a bimodal combination of basalts and rhyolites, as well as a variety of volcaniclastic and sedimentary rocks. These rocks are most often middle to late Miocene. The sedimentary rocks contain fragments of the altered volcanics, indicating that

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the altered and mineralized rocks were exposed at the surface a short time after the mineralization event (Ashley and Silberman, 1976).

These rocks are cut by a complicated system of faults. Ashley (1979) has suggested that a ring-fracture zone may have bounded the area currently occupied by the volcanic rocks. The faults can be divided into three sets. A prominent set of normal faults trending north-northeast have accommodated 1 km or more of extension. Billingsley and Locke (1941) emphasize the relatively low dip-angles of these faults. A second set trends north from Goldfield, then turns east, following the inferred ring fracture zone. A third set trends west-northwest and follows the inferred ring-fracture zone for at least part of its exposed length.

Primary ore occurs in highly silicified zones that formed as replacement bodies along fractures and faults. Ore mineralogy includes native gold, as well as sulfide and sulfosalt minerals. The silicified zones may also contain alunite, kaolinite, pyrophyllite, and diaspore. Argillic alteration is abundant around the silicified zones and includes quartz, kaolinite, potassic mica (K-mica), and limonite.

Bullfrog Hills. The Bullfrog Hills are composed predominantly of Miocene volcanic rocks. These rocks are cut by a system of normal faults that have created a complex fracture system, that locally hosts important ore deposits. Maldonado (1990) has recognized two low-angle detachment faults within this complicated set of structures. Rocks below the lower detachment fault include metamorphosed Late Proterozoic rocks that exhibit locally a well developed mylonitic fabric. Rocks between the two detachment faults are highly attenuated and consist of fault-bounded blocks of lower and middle Paleozoic clastic and carbonate rocks. Block-faulted volcanic, volcaniclastic, and sedimentary rocks of Miocene age occur above the upper detachment fault. The Bullfrog district produced more than 2.7 million dollars worth of gold and silver between 1905 and 1921.

Au-Ag mineralization exhibits strong structural control at a variety of levels. Mineralization is localized along detachment faults and high angle normal faults. Stockwork mineralization has locally formed, filling the fractures in highly deformed and altered volcanic rocks. Mineralization commonly includes quartz, calcite, and gold. Other minerals locally associated with the mineralization include adularia, acanthite, and pyrite; and very minor chalcopyrite, tetrahedrite, sphalerite, and galena (Jorgensen et al., 1990).

Relatively close constraints can be placed on the age of hydrothermal alteration, and presumably the age of associated mineralization. Volcanic units older than 10 Ma have been affected by the hydrothermal alteration, but those units younger than 10 Ma have not been altered (Maldonado, 1990). Deformation related to movement on the detachment fault and associated higher-angle normal faults, both pre- and post-date the alteration and mineralization event.

Round Mountain. The Round Mountain gold and silver district (Figure 3.1.1.1.5.2.1-2) was discovered in 1906 (Ferguson, 1921; Shawe et al., 1986), and by 1959 had produced approximately 14 metric tons of gold from placer and lode deposits (Shawe et al., 1986). The

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lode deposits were found in fractured and altered rhyolitic welded ash-flow tuffs that are associated with quartz veining located along both steep and flat-lying faults, often containing strongly brecciated rocks (Kleinhampl and Ziony, 1984; Shawe et al., 1986). Low-grade disseminated and bulk-mineable gold deposits are the focus of present-day mining operations from which greater than one-half million ounces of gold have been obtained (Shawe et al., 1986).

Two types of alteration are recognized: Type-I alteration is characterized by the replacement of sanidine by potassic-feldspar (K-feldspar) and plagioclase by albite. Type-II alteration is characterized by strong potassium metasomatism where the tuff is densely welded. The Round Mountain deposit contains quartz-adularia veins that may contain gold (Sander and Einaudi, 1987). The district is associated with uranium, mercury, tungsten, and copper mineralization, and gangue mineralogy consisting of veins of quartz, adularia, limonite (oxidized from pyrite), manganese oxide, and subordinate alunite, fluorite, and realgar (Kleinhampl and Ziony, 1984; Shawe et al., 1986). Similar gangue mineralogy and shallow plutonism or gold-silver mineralization associated with the Round Mountain deposits (Kleinhampl and Ziony, 1984; Shawe et al., 1986) have not been found at Yucca Mountain.

Bare Mountain. Bare Mountain contains a variety of mineral occurrences not easily classified into one overall category. Gold, fluorite, and mercury have all been extracted commercially. Some of the mineralization is spatially associated with a low-angle, normal or detachment fault. Veins and replacement bodies are characteristic of other mineralized areas.

A low-angle, normal fault in Fluorspar Canyon, at the north end of Bare Mountain, places Miocene volcanic rocks in the upper plate against Paleozoic and Precambrian metasedimentary rocks in the lower plate. Regionally, this fault is correlated with the Bullfrog Hills detachment fault of Maldonado (1990) and the Boundary Canyon fault system at the north end of the Funeral Mountains. Scott (1990) has suggested that this fault may correlate with a postulated detachment fault under Yucca Mountain. Several mineral occurrences are found along and near this fault in Fluorspar Canyon.

Greybeck and Wallace (1991) have described gold occurrences in some detail in the vicinity of the Daisy Mine in Fluorspar Canyon. They identify three mineralized zones; the West zone, the South zone, and the Secret Pass zone. All three zones are near the low-angle "detachment" fault, one in the upper-plate volcanics, and two in the lower-plate metasedimentary rocks. Intense silicification and abundant fluorite mineralization accompany much of the gold mineralization. Local propylitic and argillic alteration is abundant. The mineralization exhibits strong structural control and is localized along normal faults. The best ore grades are associated with fault intersections. Within these highly faulted areas, much of the gold is finely disseminated through the host rock.

Gold mineralization is present along the eastern side of Bare Mountain in the Sterling Mine. At this location, the mineralization is concentrated near the intersection of high angle normal faults and a Mesozoic thrust fault. The thrust fault places siltstones of the Wood Canyon formation over dolomites of the Bonanza King Formation. Mineralization includes fracture-

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controlled veins of quartz, calcite, and fluorite. The gold is disseminated mainly in the siltstones of the Wood Canyon Formation. Locally, brecciation is common and the mineralization is associated with an alteration suite that includes quartz, sericite, alunite, kaolinite, mixed-layer clays, and halloysite. Geochemically, arsenic, antimony, lead, zinc and mercury are anomalous; and locally, molybdenum is very anomalous (Garside et al., 1987). The deposit has a grade of 0.04 to 4.0 ounces per ton gold and is associated with anomalous concentrations of arsenic, antimony, molybdenum, and mercury (Smith et al., 1983; Tingley, 1984). In 1980, production was estimated to be 50,000 tons with an average of 0.25 ounce per ton gold (Bonham, 1984). Production in 1984 was estimated to be 58,000 tons with 9,000 ounces of gold produced per year (Bonham, 1987). Reserves are estimated at 200,000 tons of 0.02 ounce per ton gold, with possible additional reserves of 300,000 tons of similar grade (Bonham, 1987). The ore is associated with porphyry dikes and extensive alteration (Smith et al., 1983; Tingley, 1984), including high-temperature alunite.

Wahmonie-Salyer. The Wahmonie-Salyer volcanic center or district (Figure 3.1.1.1.5.2.1-3), a location of gold and silver lode and placer interest from the 1850s until the 1930s, is predominantly a highly altered vein system believed to extend for approximately 11 km near the surface and occurs in association with volcanic andesites, dacites, and latites, along with intrusive granites (Ekren and Sargent, 1965; Quade and Tingley, 1983). Areas near Wahmonie were worked for placer deposits; production is unknown, but assumed to be low.

A shallow pluton may exist below the Wahmonie area, 20 km to the east of the site, based on the presence of: (1) a large aeromagnetic high; (2) a gravity high; (3) a zone of alteration coincident with the geophysical anomalies; and (4) two small outcrops of granodiorite (Smith et al., 1981; Hoover et al., 1982). These factors, in conjunction with the lithologic variation observed and geophysical interpretation of vertical electrical soundings, induced polarization measurements, and magnetotelluric surveys, support the inference of potential for disseminated mineralization at Wahmonie (Smith et al., 1981; Hoover et al., 1982; Quade and Tingley, 1983), particularly of precious and base metals.

The observed magnetic high at Calico Hills is, in part, believed to be associated with the magnetite-rich, Paleozoic, Eleana Formation and the gravity high due to a contrast between the density of the Paleozoic rocks and the overlying volcanics (Hoover et al., 1982). However, a pluton is postulated at depths greater than 2.5 km (Hoover et al., 1982), because of the presence of a structural dome, extensive radial and ring faulting (Maldonado et al., 1979), and alteration of the rocks (Hoover et al., 1982; Carr, 1984; USGS, 1984; Maldonado, 1985a). In addition, the electrical studies (vertical electrical soundings, induced polarization measures and magnetotelluric surveys) are consistent with a complexity of structure and pervasive alteration and a probable intrusive body at depth (Hoover et al., 1982). The induced-polarization measurements outline an area of increased polarizability that may indicate mineralization (Hoover et al., 1982).

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Mercury Resources. Mercury occurs as cinnabar (HgS) and rarely as native mercury or salt calomel (Hg_2Cl_2). Most of the worldwide reserve occurs as HgS (Bateman, 1950). The commercial unit of measure for Mercury is the flask, containing a standard 34.5 kg of mercury.

In 1990, there were eight mines in the United States, including six in Nevada, that produced mercury as a byproduct and one mine, the McDermitt Mine of Nevada, that produced mercury as a principal product (USBM, 1990). There are at least 100 mines in Nevada from which some mercury production has been recorded (Bailey, 1964). As of 1990, however, the seven mercury-producing mines in operation in Nevada are:

- the McDermitt
- the Carlin
- the Pinson
- the Borealis
- Hog Road
- Jerritt Canyon
- Paradise Peak.

The McDermitt Mine is the principal producer of domestic mercury. The remaining mines produce mercury as a coproduct of gold mining operations.

In Nevada, mercury-bearing deposits are distributed throughout the central one-third of the state in a north-south trending belt that extends into Oregon. The larger and richer deposits are located in the northernmost part of the belt, along the Oregon border, with the richest deposits located in southern Oregon (Bailey and Phoenix, 1944; Fisk, 1968). Within the central part of the belt, mercury is produced as a coproduct of gold mining operations. At the southern end of the belt, occurrences of mercury have been reported, and some old mines and prospects from the 1920s are known. Production from the southern end of the belt has been minimal (less than 100 flasks per mine), with the past maximum production estimated to be nearly 100 flasks at the Tip Top Mine, and 72 flasks at the Harvey (Telluride) Mine (Figure 3.1.1.1.5.2.1-2) (Cornwall and Kleinhampl, 1964; Tingley, 1984). The Tip Top and Harvey Mines, currently inactive, are located west of Yucca Mountain in Paleozoic rocks at northern Bare Mountain.

Occurrences and deposits of mercury are classified according to the environment or host-rock type in which they occur. Eight different types of deposits have been described:

- sulfurous hot-spring deposits
- opalite blankets
- altered volcanic rock
- altered diabase dike
- altered interbedded sediment
- altered limestone

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- high-grade metamorphic rock
- altered granite deposits.

In Nevada, mercury-bearing deposits occur primarily in igneous and sedimentary rocks. Only minor production in Nevada (less than 100 flasks per mine) has been reported from altered granitic rocks and sulfurous-type deposits (Bailey and Phoenix, 1944).

Field and laboratory studies indicate that mercury deposits form from ascending, hydrothermal alkaline solutions ranging in temperature from 100 to 200°C. Ore bodies generally occur above conduits, such as faults or breccia zones. Because mercury-bearing solutions are relatively cool, the deposits are believed to form near the surface (Bailey and Phoenix, 1944; Bailey, 1964). In Nevada, most occurrences and deposits of mercury have been found at or near the surface; the deepest known deposit is at 366 m (Bailey and Phoenix 1944).

Volcanic-type mercury deposits occur primarily in altered andesitic lava flows and agglomerates (Bailey and Phoenix, 1944). In the past, a few economic deposits, such as those at the Cordero Mine in northern Nevada, have been found in altered, silicic lava flows and tuffs (Fisk, 1968). Volcanic-type deposits are characterized by the presence of abundant pyrite or marcasite, minor amounts of silica, and the alteration of the surrounding host rock to clay. Cinnabar is the dominant mercury-bearing mineral, occurring in veins and fractures. In the larger deposits, mercury is also disseminated in the pore spaces that result from the alteration of the volcanic host-rock. Localization of the deposit is controlled by steeply dipping fault planes or fractures along which there has been little movement. No other structural controls have been recognized (Bailey and Phoenix, 1944). Approximately 75 percent of the mercury recovered in Nevada has been mined from volcanic deposits associated predominantly with altered andesitic rocks (Bailey, 1964).

Opalite-type mercury deposits are associated with a high degree of silicification of silicic lavas and tuffs. Most deposits occur in volcanic rocks of Miocene age, although some deposits have been found in younger rocks. In Nevada, the majority of deposits are horizontal, blanket-shaped bodies ranging from 150 to 230 m in diameter, and are 9 to 15 m thick (Bailey and Phoenix, 1944). Cinnabar is the dominant mineral, but native mercury, calomel, and oxychlorides are sometimes present in small amounts. Pyrite is rare or absent in the ore, but it is frequently found in the altered silicic lavas below the opalite (Bailey, 1964). The ore in opalite-type deposits is of relatively low grade (Bailey and Phoenix, 1944). Nevada has more opalite-type deposits than any other state; however, opalite deposits account for less than 20 percent of Nevada's total mercury production. Opalite deposits are considered to be economic only when the price of mercury is exceptionally high, or when the deposit is very large (Bailey and Phoenix, 1944).

3.1.1.1.5.2.2 Industrial Minerals and Rocks

This broad category includes a variety of minerals and rocks commonly equated with nonmetallic and industrial minerals and rocks (Bates, 1975).

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Nevada is the nation's leading producer of barite. Most of Nevada's barite reserve occurs as bedded deposits (Papke, 1984) within the central Nevada barite province (Shawe et al., 1969; Papke, 1984). Papke (1984) estimates that Nevada's barite reserves currently total 90 million short tons. All the barite produced is from open-pit mines in near-surface, bedded deposits.

In Nevada, the most productive areas are located in the Shoshone and Toquima Ranges, 390 and 215 km north of Yucca Mountain, respectively. Barite production in these areas is from bedded sedimentary deposits in the Devonian, Slaven Chert Formation (Brobst, 1975), a black chert and carbonaceous shale and siltstone with minor amounts of intercalated dark limestone and brown, limey sandstone. Within other districts, bedded barites occur in similar rocks of Ordovician age (Brobst, 1975).

A barite vein was identified at Mine Mountain approximately 32 km east-northeast of Yucca Mountain, in fault breccias associated with the Devonian, Devils Gate Limestone and the Mississippian, Eleana Formation (Quade and Tingley, 1983). The vein, approximately 1.5 m thick, is traceable for 100 m along strike. The ore is high-grade and associated with various sulfide minerals and minor amounts of silver (Quade and Tingley, 1983). The area was mined for mercury in the late 1920s; no barite production was ever recorded.

Barite occurrences also have been noted in the central and southeastern part of the Calico Hills at the NTS, approximately 18 km east of Yucca Mountain (Quade and Tingley, 1983). The barite occurs as a gangue mineral along contact zones between argillites of the Eleana Formation and the unconformably overlying dolomites of the Devils Gate Limestone. The minute quantity and low grade of the mineralization in this location (Quade and Tingley, 1983) precludes it from consideration as an economic deposit. Noneconomic occurrences of barite associated with calcite and fine-grained silica veins occur at Bare Mountain (Smith et al., 1983) in Paleozoic rocks.

Fluorspar has been mined continuously in Nevada since 1919, and approximately 500,000 metric tons have been produced (Papke, 1979). Fluorspar deposits, consisting essentially of the mineral flourite (CaF_2) exhibit a pronounced clustering into districts. A northwest-trending fluorspar belt crossing Nevada has been identified (Papke, 1979; Sherlock and Tingley, 1985). However, fluorspar has small commercial value in Nevada in comparison with worldwide and total United States reserves.

A large number of fluorspar deposits and occurrences located in central Nevada (Sherlock and Tingley, 1985) probably contain significant near-surface reserves. A large fluorspar district occurs in the Quinn Canyon Range located in eastern Nye County and northwestern Lincoln County (Sainsbury and Kleinhampl, 1969). Of 62 deposits in Nevada, about 30 percent are associated with intrusive igneous rocks (Papke, 1979). High-grade fluorspar (i.e., more economically viable) is commonly associated with carbonate rocks and low-grade deposits are associated with silicified or igneous rocks (Papke, 1979).

Nearly 40 percent of fluorspar deposits in Nevada are replacement-type deposits and approximately 32 percent of deposits are vein-type deposits. Three-fourths of the low-grade

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vein deposits occur in silicic to intermediate volcanic rocks or intrusive rocks (Papke, 1979). In vein-type deposits, the host rock is usually unaltered or the alteration extends only a short distance into the host rock (Papke, 1979). Thus, this type of fluorspar deposit is difficult to discover without actually intersecting the deposit through drilling or a surface exposure.

The single active area of production (Bare Mountain) in Nevada (USBM, 1990) is located near Yucca Mountain (Carrillo and Schilling, 1985; Papke, 1985). The Daisy Mine at Bare Mountain (Figure 3.1.1.1.5.2.2-1) has produced nearly half the total production and is the largest deposit of fluorspar in Nevada (Papke, 1979; Smith et al., 1983). The Bare Mountain deposits are replacement-type deposits or breccia pipes in dolomite (Papke, 1979; Smith et al., 1983). The deposits are structurally controlled by thrust and normal faults (Smith et al., 1983). The Bare Mountain fluorspar deposits include gangue minerals of montmorillonite, sericite, calcite, quartz, and kaolinite with localized minor occurrences of cinnabar, gold, stibiconite, and traces of uranium in fluorite (Cornwall and Kleinhampl, 1964; Papke, 1979; Smith et al., 1983).

Zeolite Resources. Zeolite minerals are hydrated aluminosilicates containing alkali and alkali earth cations, principally sodium, potassium, and calcium. Zeolite minerals are common authigenic phases in low-temperature alteration of silicic-volcanic rocks. Only small tonnages are mined today, approximately 275,000 metric tons per year for the eight reporting countries, including the United States (Mumpton, 1977a). The United States produced 15,569 metric tons of zeolites in 1990 (USBM, 1990), while reserves are indicated to be in excess of 90 million metric tons (Eyde, 1986).

In comparison, the large proportion of worldwide reserves known to exist is measured in trillions of tons (Sheppard, 1975). Higher-quality synthetic zeolites have been commercially produced since the 1950s (Sheppard, 1975). Zeolites have many industrial and agricultural uses, including use as an ion exchange material with radioactive materials (Mumpton, 1975, 1977b). The use of zeolites is expected to greatly increase in the future (Sheppard, 1975; Mumpton, 1975, 1977b).

Large tonnages are known to exist in the United States, particularly in Cenozoic deposits in the gulf coast and western half of the United States (Olson, 1975; Sheppard, 1975). Exploration and sampling of deposits have occurred in Arizona, Wyoming, Texas, and Nevada, and zeolites have been mined in Bowie, Arizona; Hector, California; and Jersey Valley, Nevada (Sheppard, 1975). In addition, it is estimated that 110 metric tons of zeolites, which occur as pure and readily extractable surface deposits, are available in the Basin and Range province (Sheppard, 1975). Large reserves at or near the surface are also available in Nevada (Papke, 1972).

Papke (1972) has described four major deposits in Nevada: the Eastgate deposit in Churchill County, the Jersey Valley deposit in Pershing County, the Pine Valley deposit in Eureka County, and the Reese River deposit in Lander County. These major zeolite deposits are derived from silicic ash falls that were deposited in saline or alkaline lakes Pliocene to

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Pleistocene in age (Papke, 1972). The deposits are described as being extensive, but no volume estimates are given.

Construction Materials and Other Resources. The clay deposits in the vicinity of Yucca Mountain were formed in two different types of geologic environments. Those in the Amargosa District were probably formed by alteration of glassy, air-fall volcanic ash that was deposited in alkaline playa lakes that existed during the Plio-Pleistocene time (Bell and Larson, 1982). Production of this clay is continuing (Jones, 1987) and new reserves are being defined.

The second type of clay deposit is located in the Bare Mountain district, and includes bentonite, kaolin, and halloysite; these were formed by the alteration of rhyolitic tuffs. These deposits have had minimal production (Lincoln, 1923) and are not considered to be of economic importance (Bell and Larson, 1982). Deposits located in the densely welded and nonwelded tuff that occur 2.4 km south of Beatty, Nevada, have had historical production (Cornwall and Kleinhampl, 1964). No deposits of this type are known to occur at Yucca Mountain.

Perlite, a glassy volcanic rock used as a lightweight aggregate, occurs in Quaternary to Tertiary andesite and rhyolite tuffs (Chesterman, 1975). Specifications for industrial use include that the perlite contain no more than two percent mineral grains or lithic fragments, and that the weight-percent of bound water be between two and five percent. Two occurrences of perlite have been identified in the vicinity of Yucca Mountain: one in Beatty Wash at Bare Mountain, in a glassy facies of rhyolitic ash flows, and another in the NE 1/4, Section 10, T. 12 S., R. 47 E. (Cornwall and Kleinhampl, 1964). There has been no production from these deposits (Bell and Larson, 1982) and no occurrences of perlite have been identified at Yucca Mountain. Petrographic and mineralogic studies of the tuffs present at Yucca Mountain revealed abundant lithic fragments (highly variable, from 0 to 40 percent) in most units (Warren et al., 1984), and bound water less than one percent by weight (Cornwall, 1972). Thus, if perlite is present at Yucca Mountain, it would probably not be suitable for industrial uses.

Volcanic cinders and pumice have been mined from the basaltic cones at Crater Flat, 5 to 10 km to the west and southwest of the proposed controlled area boundary. The Lathrop Wells basaltic cinder cone is currently being mined. Basaltic cones are shown in Figure 3.1.1.1.5.2.2-1.

A single deposit of silica was mined at the Silicon Mine for ceramic silica. The Silicon Mine is located in the silicified and altered rhyolite tuffs northeast of Bare Mountain, on the northwestern end of Yucca Mountain, more than 16 km from central and southern Yucca Mountain (Figure 3.1.1.1.5.2.2-1). The mine (a shallow pit) had a moderate, but unknown amount of production from 1918 to 1929 (Bell and Larson, 1982). Other sources of high-grade silica that are currently more economically and technologically feasible to mine, such as quartz sand, are readily available (Murphy, 1975). For example, the Cretaceous Baseline Sandstone is currently mined near Overton, Nevada, in Clark County (Papke, 1986).

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Thus, no future production is expected from the Silicon Mine. Unconsolidated and cemented sand and gravel deposits are ubiquitous in the area surrounding the bedrock of Yucca Mountain as well as within the Basin and Range Province.

In Nevada, 28 mines have produced 62,633 kg of uranium oxide (U_3O_8), 80 percent of which was from the Apex Mine in Lander County (Garside, 1973). No production has occurred in Nevada since 1968 (Garside, 1973). Deposits and occurrences of uranium in Nevada are concentrated in central and northern Nevada, along the State's western boundary (Sherlock and Tingley, 1985). Almost all of the production in Nevada has been from vein-type deposits. These vein deposits are irregularly distributed, small in size (Butler, 1964), and occur in rocks altered by hydrothermal activity (Butler, 1964; Schrader, 1977).

Disseminated occurrences of uranium are known to occur in silicic tuffs, such as the McDermitt Caldera Complex deposits, which are also well known as areas of mercury mineralization. The uranium mineralization and occurrences are associated with hydrothermal activity (Rytuba et al., 1979; Rytuba and Conrad, 1981; Rytuba and McKee, 1984). The Apex Mine is Tertiary age vein and disseminated uranium deposit localized in highly fractured and faulted metasediments, where the uranium mineralization is believed to have resulted from chemical interactions at an ancient groundwater table boundary (Garside, 1973; Plut, 1979).

Garside (1973) has reviewed all known occurrences of uranium in Nevada. The type of host rock and kind of uranium occurrences are highly varied. The occurrences of uranium nearest Yucca Mountain are located at Bare Mountain, in Paleozoic carbonate rocks associated with fluorite. The occurrences at Bare Mountain are described by Garside (1973) and Tingley (1984), and are discussed under fluorspar resources. Uranium deposits that occur in silicic tuffs are more appropriately compared to Yucca Mountain. Uranium occurrences in Tertiary volcanic rocks are divided into two lithologic subtypes by Garside (1973); those that occur in plugs and lava flows and those in ash-flow tuffs. Fifty-five occurrences of uranium are known in ash-flow tuffs, mostly in the counties of Elko and Washoe, Nevada (Garside, 1973). Generally, these occurrences are associated with groundwater interactions or hot thermal waters that interact with the tuff-rock and deposit uranium. The Bullfrog and Tonopah districts (Figure 3.1.1.1.5.2.1-2) have occurrences of uranium that may be associated with hot thermal waters or leaching of tuff by groundwater (Garside, 1973).

3.1.1.1.5.3 Geothermal Resources

Various schemes have been developed for classifying geothermal reservoirs. One of the most widely used classifications divides thermal reservoirs into three regimes (Muffler and Guffanty, 1979): (1) high-temperature systems, greater than 150°C; (2) intermediate-temperature systems, 90 to 150°C; and (3) low-temperature systems, less than 90°C. These thermal regimes correspond to potential uses of the thermal waters; for example: (1) high-temperature systems provide steam used to generate electrical power; (2) intermediate-temperature systems provide heat for use in various industrial processes (Brook et al., 1979); and (3) low-temperature systems are potentially available for use in space

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heating and agriculture (Sammel, 1979). Other factors that play a role in determining the potential for a geothermal resource or the value of a reserve include: (1) that the depth of thermal water be less than 3 km (White, 1973) or in low-temperature geothermal systems at depths less than 1 km (Sammel, 1979); (2) that the volume of available fluid be large enough to sustain its use or greater than 5 km^3 (White, 1973); (3) that the rock has adequate permeability to sustain its water flow (White, 1973); and (4) that the thermal fluid does not contain a high proportion of corrosive dissolved solids (White et al., 1971; Trexler et al., 1979).

The ultimate source of geothermal water is discussed in terms of two models in which thermal waters are generated from the areas of deep-seated, regional, high heat flow from a conductive system, which may drive hydrothermal convection (Renner et al., 1975; Brook et al., 1979); and igneous-related hydrothermal systems, where cooling, high-level, silicic or mafic magma bodies provide energy for hydrothermal convection (Smith and Shaw, 1975, 1979). For high-temperature systems, these two models probably overlap to some extent because a high, regional geothermal gradient alone could not sustain the high temperatures found in most regional geothermal systems, for extended periods of time (Renner et al., 1975). Low-temperature systems may occur as a result of either model, but typically are associated with conduction-dominated systems, considered to consist of either hot, dry rock or basin, aquifer systems. Low-temperature, hydrothermal systems and conduction-dominated systems are reviewed by Sorey et al. (1983).

Geothermal Systems in Nevada. The Basin and Range Geologic Province is considered to be favorable for the occurrence of geothermal systems, because of a generally high heat flow and the extensional tectonic regime that provides the conduits for thermal water flow (Mariner et al., 1983; Sorey et al., 1983). Normal faults provide near-surface conduits for the circulating waters and thus control the locations of most of the identified hydrothermal convective systems (Mariner et al., 1983). More than 300 thermal springs, spring clusters, and thermal wells occur throughout the State of Nevada (Garside and Schilling, 1979). Potential geothermal resource sites and areas have been plotted on base maps by Garside (1974), Garside and Schilling (1979), and Trexler et al. (1979). A review of geothermal energy in Nevada was published by Garside (1987). The distribution of hot springs and geothermal wells is presented in Figure 3.1.1.1.5.3-1. The majority of high-temperature springs occur in northern Nevada, along major faults that bound the State's mountain ranges (Garside and Schilling, 1979). In general, exploration in Nevada is expected to concentrate on finding reservoirs that can produce steam or hot water for electric power generation (Garside and Schilling, 1979; Bell and Larson, 1982), or supply heat for space heating (Garside, 1983b). The hottest surface and subsurface temperatures encountered are a Beowawe (25° to 212°C), Brady (70° to 214°C) Desert Peak (199° to 208°C), and Steamboat Hot Springs (21° to 187°C) (Garside and Schilling, 1979). These sites have been or are being investigated for electric-power generation.

In the vicinity of Yucca Mountain, including the Amargosa Desert and the NTS, springs have potential as low-temperature geothermal energy resources (Bell and Larson, 1982). For example, to the west are Goss Springs (22°C), Burrell Springs (36° to 39°C), Hicks Springs

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(32° to 43°C), and Beatty Municipal Springs (24°C); to the south are Scruggs Springs (30°C), Devils Hole (27° to 33°C), and Ash Meadows (24° to 34°C). These are all low temperature springs. Of the 19 wells in the vicinity of Yucca Mountain, temperatures range from 23° to 65°C with an average of 29°C (Garside and Schilling, 1979; Trexler et al., 1979). These data have been tabulated by Bell and Larson (1982). As would be expected, temperature generally increases with depth (e.g., the high temperature of 65°C was measured at a depth of 1,006 m and the low temperature of 23°C was measured at 102 m), and is generally thought to be representative of the local geothermal gradient.

Assessments of the geothermal gradient and heat flow of an area or region are considered more valuable, when available, in evaluation of hydrothermal and conductive reservoir resource potential, because they more accurately reflect the amount of energy available than do simple down-hole temperatures. Simple down-hole temperature measurements can be affected by a variety of hydraulic-flow conditions and the changing thermal conductivities of the rock along the flow path.

Heat Flow and the Thermal Gradient. Generally, heat flow and thermal gradient are considered in assessing geothermal reservoirs because heat flow, thermal gradient, and thermal conductivity of the rocks are simply related by Fourier's Law, such that:

$$r = q/K$$

where

r = the thermal gradient expressed in ^{0}C/km

q = heat flow in 10^{-6}cal/cm^{2}s

K = thermal conductivity in 10^{-6}cal/((cm)(s)(^{0}C)).

Thermal gradient is directly proportional to heat flow, but inversely proportional to thermal conductivity (White, 1973).

Average continental geothermal gradients typically range from 20° to 30°C per kilometer (Press and Siever, 1982). In Nevada, thermal gradients generally range from 30° to 60°C per kilometer (Garside and Schilling, 1979). Thus, by comparison, Nevada has a higher than average thermal gradient. The average heat flow in Nevada is 2 x 10^{-6}cal/cm^{2}s (2 heat flow units (HFU)) (Garside and Schilling, 1979). The average continental heat flow is 1.5 x 10^{-6}cal/cm^{2}s (1.5 HFU) (Sass et al., 1971). In general, Nevada would be expected to have a high potential for geothermal energy, based on the heat-flow data.

Within Nevada, several regions of abnormally high or low heat flow are present. Battle Mountain high in northern Nevada (Figure 3.1.1.1.5.3-2) has an average heat flow of 3 x 10^{-6}cal/cm^{2}s (3 HFU) (Garside and Schilling, 1979) and the Eureka Low (EL) (Figure 3.1.1.1.5.3-2) has an average heat flow of less than 0.7 to 1.5 x 10^{-6}cal/cm^{2}s, or less than 0.7

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to 1.5 HFU (Sass et al., 1971; Lachenbruch and Sass, 1977). The Battle Mountain region is interpreted as having a high heat flow due to the transient effect of recent crustal intrusion by silicic magma (Sass et al., 1971); whereas nearly the whole of the Basin and Range Province has high heat flow, because of postulated near-melting conditions in the lower crust and upper mantle (Roy et al., 1968, 1972; Lachenbruch, 1970). The EL is interpreted in two ways by Sass et al. (1971). It could be a crustal region where temperatures in the lower crust and upper mantle have been below the solidus for some time or, more likely, the result of complex, interbasin water flow.

Complex interbasin water flow could be regionally systematic with appreciable downward flow of relatively cool waters to a depth of approximately 3 km (Sass et al., 1971). Sass and Lachenbruch (1982) indicated that this latter model would require the following:
(1) groundwater is redistributing much of the earth's heat in the upper 3 km and delivering it elsewhere; (2) the heat flow at depth, beneath the EL, could be the same as that characteristic of the Great Basin (2 HFU) or may be as high as 2.3 HFU. Blackwell (1978) related the EL to a crustal region where older volcanism has occurred, in comparison to the Battle Mountain high where more recent volcanism has occurred. Finally, Lachenbruch and Sass (1977) related the heat flow in the Basin and Range to areas of rapid, active, crustal extension, such as the Battle Mountain high.

Interpretation of silica geothermometry data (used to predict heat flow) may suggest that heat flow under the EL is near the average for Nevada (2.0 to 2.5 HFU). Swanberg and Morgan (1978, 1981) have produced heat-flow contour maps of the western United States based upon regional silica geothermometry data. These maps indicate that most of Nevada, including the EL and Yucca Mountain, would fall in a zone of greater than 2.5 HFU. Caution must be employed in the interpretation of silica geothermometry data, because of the following:

a. The geothermometer is based on the temperature dependence of quartz solubility in water and has only been calibrated in a few areas in carbonate aquifers (i.e., much of Nevada is covered by silica-rich volcanics, including silicious glass, which could have much higher solubility than quartz);

b. The data base has been applied on the scale of the continental United States, so care should be taken in applying interpretation to a localized area; and

c. The silica geothermometer is based on an empirical relationship that, in turn, is related to heat flow by another empirical relationship.

Because of the provisional nature of this data it will be used for background information only.

3.1.1.1.5.4 Hydrocarbon Resources

Deposits of coal, oil shale, and tar sands can typically be identified, because they occur as large, persistent, tabular bodies. Thus, they are readily recognized on the basis of surface outcrops and stratigraphic interpretation, even in structurally complex terrains. Oil and gas

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resources are more difficult to evaluate, because of their ability to migrate into permeable reservoir rock. The generation and localization of oil and gas involve complex processes; evaluation requires knowledge of: (1) any appropriate source rocks available for oil or gas generation; (2) any appropriate reservoir rocks; (3) the thermal history of the area of consideration; (4) the stratigraphic and structural components of the area; and (5) theoretical approaches, as applied to migration models, tectonic history, and other factors.

Deposits of Coal, Tar Sands, and Oil Shales. Previous studies, such as the inventory of the mineral and water resources of Nevada (USGS/NBM, 1964), as well as Cornwall (1972) and Bell and Larson (1982), do not report the occurrence of any coal, tar sand, or oil shale deposits in either southern Nye County or the NTS. Geologic mapping and core-logging data from various drillholes near and at Yucca Mountain provide additional documentation of nonoccurrence.

Coal Resources. The sizable reserves and resource of coal in the United States have been inventoried by Averitt (1975). Although many of the important deposits are Paleozoic (Pennsylvanian), much of the coal in the western United States is Cretaceous and Tertiary. The importance of mining some of these young, near-surface, low-sulfur coal seams has steadily increased in recent years (Friedman et al., 1985). By contrast, there are no economic deposits of coal presently located in Nevada, although occurrences of thin, low-grade deposits (high ash content, minimal thermal output, or both) are known in several localities (Horton, 1964a). Averitt does not tabulate any reserve tonnages for Nevada.

The closest known coal deposits to Yucca Mountain are found near Coaldale in Esmeralda County, Nevada (Horton, 1964a; Brady, 1984a,b). The coal seams occur within a Tertiary lacustrine sequence as thin (all less than 2.1 m) bituminous rank deposits. Although the coal is bituminous, the grade is quite low because of a very high (average 50 percent) content of ash. Very small tonnages have been produced from these low-grade, structurally complex, and thin coal seams (Brady, 1984a). Kleinhampl and Ziony (1984) report that there are no known occurrences of coal in northern Nye County.

Tar-Sand Resources. No extensive tar sand or oil-impregnated sand deposits have been reported in Nevada. However, Bortz (1983, 1985) reports several surface occurrences of hydrocarbons within the northern part of the Basin and Range Province. Most notable of these, with regard to similarity to tar sands, is the well-known Bruffey Oil Seep located in Pine Valley, southeast of Elko in Eureka County, Nevada (Foster et al., 1979). Of further interest is the Blackburn oil field discovered in 1982 by Amoco Oil Company in Eureka County, located only a few kilometers west of this seep, where Quaternary conglomerate is partially impregnated by black, tar-like hydrocarbons (Bortz, 1983).

Horton (1964b), Cornwall (1972), and Bell and Larson (1982) do not report any occurrences of tar sands or similar deposits in either southern Nye County or near the NTS. In addition, no evidence of such deposits appears in any of the subsurface data from drilling at or near Yucca Mountain.

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Oil Shale Resources. Shales and related rocks can be associated with hydrocarbon resources in the following four possible relationships:

a. Oil shales, or rocks enriched in solid organic matter or kerogen, that can yield synthetic crude oil through the destructive distillation of the mined rock;

b. Reservoirs in which hydrocarbons accumulate within fractures;

c. Impermeable confining layers above many petroleum traps; and

d. A source rock whose organic matter is converted into hydrocarbons by natural thermal maturation and expelled to migrate into petroleum traps elsewhere.

It is not unusual for certain shales, as discrete stratigraphic units, to perform in more than one of these roles.

As early as 1917, synthetic crude oil was produced from Nevada oil shales at Elko, in Elko County (Knutson and Dana, 1983). A single processing plant produced approximately 12,000 barrels, over a 7 year period, from oil shales of the Eocene-Oligocene Elko Formation. However, the venture was not an economic success and no other commercial efforts in Nevada have been made. Oil shales in Nevada are described by Winchester (1923), Garside (1983a), Knutson and Dana (1983), and Palmer (1984).

In addition to the Elko Formation in Elko County, occurrences of oil shale are found within the more widespread Ordovician Vinini Formation within Elko, Eureka, Lander, and northern Nye counties in Nevada. Other formations containing oil shales are the Devonian Woodruff Formation in Eureka and Elko counties; the Cretaceous Newark Canyon Formation in Eureka and White Pine counties; and the Cretaceous(?)-Paleocene-Eocene Sheep Pass Formation in White Pine, Lincoln, and northern Nye counties (Knutson and Dana, 1983).

Known Occurrences of Oil and Gas in Nevada. For many years, Nevada totally lacked any economically productive oil and gas fields. The Eagle Springs field, discovered by Shell Oil Company in 1954, remained the only producing oil field within the State for more than two decades. From the late 1970s into the mid-1980s, six additional fields were discovered. In addition to the Eagle Springs field, four of these more recent fields (Trap Spring, Currant, Bacon Flat, and Grant Canyon) are located in the Railroad Valley area of northeastern Nye County. The Blackburn field is situated in the Pine Valley area of Eureka County; whereas the single-well Jiggs field is in southwestern-most Elko County. Both the Jiggs and Currant fields are shut-in and the former field may ultimately be considered noncommercial. Additional information and discussion can be found in Garside et al. (1977), Bortz and Murray (1979), French (1983), Jones and Papke (1984), Bortz (1985), Chamberlain (1986), and Fritz (1987).

Even though Nevada does not rank among the leading petroleum-producing states, its cumulative production has passed 16.9 million barrels (Garside and Weimer, 1986; 1987).

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Nearly 80 percent of that production has been achieved during the last decade. Sizable quantities of oil have been produced from the Grant Canyon field (Railroad Valley, Nye County), where cumulative production through August 1987 has risen to 4.7 million barrels (Veal et al., 1988a). Chamberlain (1986) has also shown a dramatic increase in the barrels of oil produced versus feet of exploratory drilling conducted since the mid-1970s.

For nearly six decades, ranches in the Fallon area of Churchill County have produced natural gas (mainly methane) from very shallow (less than 50 m) wells drilled into the organic-rich sediments of the Quaternary Wyermaha Formation (Brady, 1984b; McDaniel, 1985). In addition, shows of natural gas have been reported from several seeps (located at Bruffey, Pine Valley, Eureka County and Diana's Punch Bowl, Monitor Valley of northern Nye County) and from various exploratory wells. Except for the Fallon area, no production of gas is known in Nevada.

3.1.1.1.5.5 Conceptual Models of Resource Formation

Mineral deposits form as an integral part of the development of geologic systems. However, they commonly result from some of the most complicated aspects of those systems. Consequently, developing an understanding of these deposits requires a sophisticated combination of data collection, analysis, and interpretation. Substantial research effort over the last century has led to development of a number of conceptual models for mineral deposit formation.

The development and elucidation of conceptual models is important for a number of reasons. Conceptual models incorporate a set of characteristics common to a group of mineral deposits. The models identify physical, chemical, and geologic processes important in the development of the deposits and suggest possible consequences from these processes. Conceptual models suggest possible relations between a given mineral deposit and other features in the region. The models also identify potentially anomalous features that must be explained in another way and, consequently, may provide important clues to improve our understanding of the area. The regulatory implications of these and other conceptual models will be discussed in Section 3.3.

For the development and evaluation of conceptual models, we must begin with a classification system. This system will have an important impact on the development of models, because it will identify a set of characteristics that will serve to define and differentiate the deposit types. These characteristics will also provide the foundation for any conceptual model developed.

The types of features that have been identified in the past include the style of mineralization, the geometry of the ore body, and the associations with rock-types or structures. The style of mineralization includes the nature of the mineralization; disseminated, massive, tabular, irregular, etc. Style can also include mineral associations. It is common for particular economic metals, or minerals, to occur together and to have characteristic associated (alteration) minerals.

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The geometry of an ore body can have important implications for how it developed. Ore bodies can be tabular, cylindrical, irregular, fracture-controlled, etc. This geometry can reveal important information about processes of formation, such as controls on the flow of mineralizing solutions, etc.

Temporal and spatial associations of ore bodies are crucial to the ore forming process. A common example is lithologic control by the wallrocks that bound the ore body. The association of ore bodies with irregularities in a fracture system or with breccia zones can indicate the importance of permeability contrasts in localizing the flow of hydrothermal fluids and deposition of ore minerals.

One example of this approach is presented in Cox and Singer (1986). They identify a set of characteristics that describe and classify each deposit type. The models discussed in Cox and Singer (1986) include two parts: "Geologic Environment," that describes the environments in which the deposits are found; and "Deposit Description," that gives the identifying characteristics of the deposits. Descriptions of geologic environments include rock-types and textures, age range, depositional environment, tectonic setting(s), and associated deposit types. Identifying characteristics of the deposits include mineralogy, texture, structure, alteration, ore controls, and geochemical signature.

Further development of conceptual models can be achieved by relating mineral-deposit formation to physical, chemical, and/or geologic processes. Information about geologic setting and physical-chemical conditions of mineral-deposit formation can be incorporated into models by characterizing the processes involved in the formation.

Volcanic Systems Associated with Metallic Deposits. Approximately 60 percent of the world's known Tertiary epithermal precious- and base-metal vein districts associated with volcanic rocks have been analyzed and tabulated by Mosier et al. (1986) in terms of their geology, grade-tonnage information, and ore associations. Of these deposits, 46 percent have a clear association with rhyolite or rhyodacite tuffs. Although the ages differ, the volcanic rock associations discussed previously for the rhyolite and rhyodacite tuffs are similar and directly comparable to the associations found in volcanic rocks of the Great Basin (Figure 3.1.1.1.5.2.1-2). In the Great Basin, most of the gold production from deposits associated with volcanic rocks has been from vein deposits in andesites associated with calc-alkalic volcanic suites that formed from 22 to 6 million years ago and in two other major suites of rocks that consist of both intermediate and silicic ash-flow tuffs formed from 43 to 17 million years ago; and bimodal assemblages of both basalt and rhyolite that range in age from 17 million years old to the present (Silberman et al., 1976). The compositional range of rock types associated with most vein-type deposits generally do not occur at Yucca Mountain.

The volcanic rocks at Yucca Mountain have affinities with silicic volcanic systems (Lipman et al., 1966; W. J. Carr et al., 1986; Flood and Schuraytz, 1986). Thus, the compositional range of the erupted tuffs is different than the volcanic suites associated with vein deposits in the Great Basin.

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Tertiary, ash-flow tuff constitutes greater than half of the total surface outcrops in Nevada. However, 93 percent of all major metal-mining districts in Nevada are in lithologies other than silicic tuff (McKee, 1979). Much of Nevada, Nye County, and all Yucca Mountain at the surface consist predominantly of Tertiary ash-flow tuffs. The tuffs are the products of eruption from magma chambers that vented through numerous volcanos, now recognized as calderas, throughout the State. The age of the calderas and their associated tuffs partially overlap in space and time (Christiansen et al., 1977). The general relationship among calderas, ash-flow sheets, and mineralization has been summarized and evaluated by McKee (1979).

Areas within calderas are typically barren of economically important base and precious metals, while strongly metamorphosed and altered rocks located on the margins of calderas are more economically important. Only two of Nevada's 31 recognized calderas (6 percent) have produced 1 million dollars or more of gold, silver, copper, lead, zinc, mercury, antimony, and iron (McKee, 1979). The two mineralized districts found within calderas in Nevada are the Goldfield district (largely a gold producer) and the Opalite district (a mercury district) (McKee, 1979). Mineralization is within the margins of calderas (e.g., the Jefferson Caldera) (Shawe, 1987) and with thick piles of silicic, ash-flow tuffs within and around their calderas, elsewhere in the western United States (Bethke and Lipman, 1987).

These relations suggest that the ash-flow and caldera-forming processes are not integral parts of the ore forming system. In fact, these processes may be more important for dispersing metals than for concentrating them. However, there are good examples of areas where important ore deposits have formed in ash-flow tuff piles in the final stages of the volcanic systems or at a later time. In these cases, the mineralization shows strong evidence for fracture control and is commonly related to regional structures.

Physical and Chemical Characteristics of Precious Metal Ore Deposits. Many precious-metal and base-metal vein deposits have formed at temperatures between 50° and 300°C, and at depths less than 1,000 m from the surface (Sillitoe, 1977; Berger and Eimon, 1983; Nelson and Giles, 1985; White, 1985; Guilbert and Park, 1986). These deposits are postulated to exhibit an ore-zonation with depth (Berger and Eimon, 1983; Sawkins, 1984), with gold and silver ore mineralization nearer to the surface than sulfide mineralization (Figure 3.1.1.1.5.5-1). Both high-volume, low-grade (bulk-mineable) and low-volume, high-grade (bonanza) precious-metal deposits are known to be of the epithermal-disseminated and vein-style of origin. Gangue mineralogy also exhibits zonation with depth (Figure 3.1.1.1.5.5-1) and occurs as three-dimensional alteration halos around the epithermal deposits (Berger and Eimon, 1983; Bird et al., 1984; Cole and Ravinsky, 1984; Sawkins, 1984; Mosier et al., 1986). Alterations and ore zonations have been summarized by Berger and Eimon (1983), Bird et al. (1984), Sawkins (1984), and Mosier et al. (1986). The principal ore minerals are electrum, tellurides, argentite, native gold and silver, and the base-metal sulfide and sulfosalt mineral groups (Mosier et al., 1986). Principal gangue minerals are quartz, chalcedony, opal, adularia, alunite, calcite, chlorite, pyrite, and arsenopyrite (Mosier et al., 1986).

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Mineralization commonly shows clear evidence of fracture control of the flow of hydrothermal fluids and open-space filling during mineral deposition. The veins commonly display well developed crystals of quartz and other minerals, as well as other clear, accretionary features, such as crustification. In many cases some minor evidence for wallrock replacement accompanies this abundant evidence for open space filling. The fracture systems hosting the mineralization commonly display a well-developed, preferred orientation, commonly forming a few "sets," but in detail, the fracture geometries can be very complex. Bends, horsetails and bifurcations are common and these irregularities are important for the localization of ore.

Mineralization commonly occurs over a fairly discrete vertical zone, typically 300-400 m thick (Buchanan, 1981). In many cases the mineralized zone is terminated at depth by a decrease in ore grade. Alteration and mineralization are present below the ore zone, but precious metal contents are diminished and replaced by sulfide, carbonate, and/or silicate minerals.

These characteristics form the basis for many models of ore deposition. Different models emphasize and organize features in different ways. For instance, Buchanan (1981) considers a subset of features, including mineral-type, grade and abundance, alteration minerals and assemblages, fluid inclusion data, vein attitudes, and evidence for zonation within veins. Other types of evidence, such as geometry of the ore body or age of mineralization, are excluded from Buchanan's analysis. Einaudi (1992) presents a more inclusive list of features as essential to description and classification of ore deposits. Einaudi's list includes age; host rock lithologies and lithologic controls; structural control; alteration style, including zoning and dimensions; vein- or veinlet-type distribution and zoning; ore mineral associations; ore zone morphology, including dimensions and grade; elemental signatures; and geophysical signatures.

Models developed in this way will have important implications for many "local" problems associated with the formation of individual ore deposits.

Processes Associated with Felsic Volcanism. Felsic-volcanic rocks and their associated caldera structures host some of the world's most important precious metal deposits. These environments provide all the essential features required by most conceptual models for ore deposit formation. The formation of calderas creates an extensive fault and fracture network that can serve as an effective "plumbing system" to localize the flow of hydrothermal solutions. The igneous activity inherent in the volcanic system can serve as a heat source to drive a hydrothermal system, although this igneous activity must be subsequent to the caldera forming event. The magmas of the igneous system may serve as a source for some or all of the metals that form the deposits.

One model, developed by Sillitoe (1973), postulates that classic epithermal mineralization is a high level expression of mineralized systems that include porphyry copper-type mineralization at deeper levels.

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Smith and Bailey (1968), who suggest that mineralization is an integral part of a caldera system, have developed a seven-stage model for the development of resurgent calderas. Stage seven of their model is characterized by fumaroles and sulfataric alteration/mineralization.

Another model suggests that calderas are important from a ground-preparation point of view. Caldera collapse forms the structural framework that later localizes the flow of hydrothermal solutions and deposition of the ore minerals. The actual mineralization is associated with an igneous intrusion that is later than the caldera-forming event, but may be associated with resurgence of the caldera system.

Processes Associated with Normal Faulting. Normal-fault systems have provided important controls on the location of precious metal deposits, especially in the Bullfrog Hills. At the original Bullfrog Mine the detachment fault that forms the base of the extended volcanic rocks is the host to extensive mineralization. Normal faults above the detachment host important ore concentrations in other parts of the district.

On a regional scale, normal faulting appears to be diachronous, younger in the west and older in the east. At Yucca Mountain, the Paintbrush Tuff is involved in the main phase of deformation and the Timber Mountain Tuff fills paleotopographic lows between rotated blocks of the Paintbrush Tuff. In the Bullfrog Hills, the Timber Mountain Tuff is involved in the main part of the deformation, and the Stonewall Flat Tuff fills paleotopographic lows between rotated blocks of the Timber Mountain Tuff. Still further west, Quaternary faulting characterizes Death Valley and Panamint Valley.

In general, this system is characterized by low-angle normal faults observed in some places, such as Boundary Canyon and inferred in others, such as Yucca Mountain. At higher structural levels, the system is characterized by high-angle normal faults that bound rotated blocks, commonly composed of Tertiary volcanic rocks. Breccias are common in the fault zones that bound the rotated blocks and near the low-angle faults that form the lower bound of the system. These breccias and other areas of abundant fracturing can provide high permeability zones favorable to the flow of hydrothermal solutions. These may originate as an integral part of the deformational process or may be related to igneous activity that occurs independent of the deformation.

Processes Associated with Thrust Faulting. Regional, Mesozoic thrust-fault systems have been projected through the Yucca Mountain area by several workers (Wernicke et al., 1988b; Carr, 1984). These workers have projected thrust faults through the area in different ways based on very sparse data. The details of these projections are highly speculative. However, it is highly probable that Paleozoic rocks beneath Yucca Mountain are involved in the Mesozoic structures in some way. Models for the development of these structures may be important for the evaluation of some resources, especially oil and gas.

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3.1.1.1.6 Geophysics

The results of geophysical investigations, as needed for issue resolution and/or basic understanding of the region, will be discussed or referenced in the relevant section describing geology, seismology, and geohydrology.

3.1.1.2 Site Geology

3.1.1.2.1 Geomorphology and Topographic Features of the Site Area

The purpose of this section is to describe the geomorphology and topographic features of the site. [The data, information, investigations, and analyses will address the following:

a. Whether extreme erosion is represented at the site or an adjacent area, such that isolation of waste within the site may or may not be affected;

b. Projections of the maximum erosion rates in the vicinity of facilities important to safety (critical structures) and engineered features (i.e., exploratory shafts, waste handling facilities, etc.); and

c. Projections of the maximum erosion rates in areas (such as washes) overlying underground facilities having minimum thickness of overburden or vulnerability to denudation.

In considering projection of the maximum erosion rates, the effects of subsidence and faulting are incorporated in the analysis.]

[Appropriate geomorphic and topographic maps and cross sections will be provided to illustrate and clarify characteristics of the site and potential for erosion. Additional figures and tables will be required.]

3.1.1.2.1.1 Erosion at the Site

Long-term, average erosion rates have been calculated from dated hillslope deposits (DOE, 1993). These rates range from <0.1 cm to 0.6 cm/ka for hillslopes in the Yucca Mountain area. Erosion rates on Yucca Mountain hillslopes are very low, averaging 0.19 cm/ka. This rate is considered to be a maximum rate because well-varnished *in situ* boulder deposits that are found on several Yucca Mountain ridges show no signs of significant movement by hillslope erosion since they were formed. The degradation rate for slopes in the Yucca Mountain area is slightly lower than for hillslopes on Yucca Mountain, averaging 0.12 cm/ka, because older deposits were found on hillslopes north and east of Yucca Mountain (DOE, 1993).

Two types of stream-incision rates were calculated for the Yucca Mountain area (DOE, 1993). Minimum and maximum, hypothetical, downcutting rates were constructed for Fortymile

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Wash on the basis of stream-terrace elevations above the modern channel and the total thickness of alluvium in the valley. A minimum downcutting rate is about 42 cm/ka and the maximum is 222 cm/ka. The true, long-term downcutting rate for the present incised canyon is believed to be closer to the minimum rate, because it is highly unlikely that the wash incised to bedrock during each downcutting episode; also, headcutting from the main wash did not migrate upstream into Yucca Mountain tributaries; and the overall behavior of Fortymile Wash during the Quaternary has been aggradation, not degradation.

Several small canyons are located above the potential repository block on Yucca Mountain. Each of these canyons is cut 60-100 m into 12.7 Ma volcanic tuff and shallow eastward toward Midway Valley into which the canyons drain. The drainage system of Fortymile Wash and its tributaries was established in Miocene time and has changed little in basic plan since then. Thus, the long-term incision rate for first-order streams that cut the small canyons on Yucca Mountain is 0.8 cm/ka or less.

3.1.1.2.2 Stratigraphy and Lithology of the Site

This section summarizes the stratigraphic and lithologic setting of the Yucca Mountain Site. The stratigraphic and lithologic setting of the site is important to the YMP for two main reasons. First, the site stratigraphic context is important to understanding the characteristics of the natural system that could affect the suitability of the site, such as groundwater flow configuration; and second, an accurate description of site stratigraphy is necessary to properly assess whether the site can be constructed and operated safely, using available engineering technology.

3.1.1.2.2.1 Introduction

The stratigraphy and lithology of the site area will be summarized according to the level of detailed information available. Where little is known, some stratigraphic units will be generalized on the basis of similar age, lithology, or history. Where more is known, individual stratigraphic units will be described and discussed. The general age, distribution, lithology, and history of each unit or age group will be discussed. The geohydrologic characteristics of these units and their contacts will be discussed in Subsection 3.1.2.3.4 [which will be added in a later revision].

Stratigraphy at the Yucca Mountain Site includes rocks and unconsolidated deposits of a wide variety of types and ages, but is dominated volumetrically by mid-Tertiary pyroclastic rocks erupted from the SNVF. In general, the stratigraphic section can be divided into four categories, based on similarities in age, lithology, and history:

a. <u>Pre-Cenozoic rocks</u> are believed to underlie the pyroclastic rocks erupted from the SNVF, but little detailed information is available on the character of Pre-Cenozoic rocks at the Yucca Mountain Site. They are believed to consist primarily of Paleozoic sedimentary rocks, similar or identical to rocks exposed in mountain ranges near the Yucca Mountain Site, such as Bare Mountain or the Calico Hills.

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b. Mid-Tertiary pyroclastic rocks make-up the bulk of the site area, and form the actual host rock of the repository. They consist of as much as 3,000 m of generally rhyolitic to dacitic ash-flow, -fall, and epiclastic-flow deposits, with some interbedded lava. The contact of these rocks with the Pre-Cenozoic rocks is known as the Sub-Tertiary Unconformity. The pyroclastic rocks represent multiple-eruption events, which are believed to have erupted from the SNVF between approximately 10 and 14 Ma.

c. Quaternary igneous rocks are scattered as volumetrically minor basaltic dikes, lava flows, and scoria cones. These deposits are believed to have been erupted after cessation of pyroclastic eruptions.

d. Late Tertiary to Quaternary surficial deposits, the youngest deposits in the site area generally consist of unconsolidated collivium, alluvial sands and gravels, marsh deposits, and eolian deposits.

Stratigraphic studies of Yucca Mountain and vicinity have included geologic field mapping of Yucca Mountain (Scott and Bonk, 1984), trenching and mapping studies of Holocene surficial deposits (Swadley, Hoover, and Rosholt, 1984), and analysis of several of the 36 exploratory boreholes at and near Yucca Mountain Site. These field-mapping and borehole data also have contributed to the characterization of Yucca Mountain structures, such as the Imbricate Fault Zone, by projection of stratigraphic dips into areas where the structures are poorly exposed [Scott, 1990].

Ortiz et al. (1985) produced a reference system of stratigraphic nomenclature for Yucca Mountain based on thermal, mechanical, and hydrologic properties of the rocks. The purpose of this system was to classify the Yucca Mountain stratigraphy according to engineering properties of the rocks. This system is based on two material properties; porosity and grain density. Sixteen reference thermomechanical units were defined, based on assumptions regarding the porosity and grain-density characteristic of each of the volcanic rock-types present at Yucca Mountain, using the drill holes and surface outcrop data that were currently available. In addition, zones containing prevalent zeolites were defined. These defined units and their three-dimensional locations in drill hole and outcrop were used to generate, by computer, a three-dimensional computer thermal-mechanical model of the volcanic-stratigraphic section at Yucca Mountain. This model is discussed in more detail in Subsection 3.1.1.2.7.

3.1.1.2.2.2 Pre-Cenozoic Rocks

Very little is known about the stratigraphy of pre-Cenozoic rocks underlying Yucca Mountain. [INN 3.1.1.2.2.2-1] Based on extrapolation from outcrops at scattered locations in the vicinity, such as at Bare Mountain and Crater Flat, the pre-Cenozoic rocks underlying Yucca Mountain are believed to consist generally of Paleozoic sedimentary rocks (Robinson, 1985; Carr, 1974). The best current estimate of the thickness, stratigraphy and distribution of these rocks, and the geometry of the sub-Tertiary unconformity, is based primarily on gravity and magnetics modeling (Robinson, 1985). This modeling is supported by direct observation in

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only one borehole, UE-25p#1 which is located east of Yucca Mountain, and penetrated carbonate rocks of interpreted Silurian age from approximately 1,244 to 1,805 m depth (Muller and Kibler, 1984).

The main source of possible variation in the thickness, distribution, and stratigraphy of Pre-Cenozoic rocks is their style and history of deformation. Exposures of pre-Cenozoic rocks at Bare Mountain and Yucca Flat are interpreted by Carr (1974) and Snyder and Carr (1982) as lower and middle Paleozoic carbonates thrust over upper Paleozoic carbonates on Mesozoic thrust sheets of the Sevier Orogeny. However, Robinson (1985) questioned whether the thrusts proposed by these authors can be projected beneath Yucca Mountain, given the possibility of offset by Cenozoic strike-slip faulting of the Walker Lane belt. Robinson (1985) proposed a model in which upper-Paleozoic carbonate rocks at Yucca Mountain are distributed in an asymmetric basin, approximately 80 km wide, that is flanked by uplifted upper-Proterozoic clastic rocks.

3.1.1.2.2.3 Mid-Tertiary Pyroclastic Rocks

As noted previously, the bulk of the stratigraphic section at the Yucca Mountain Site is made up of mid-Tertiary silicic-volcanic rocks. This volcanic sequence generally consists of a series of welded and non-welded ash-flow and ash-fall tuffs and lavas, with some interbedded volcanic breccias derived from nearby calderas and volcanoes, and ranges from approximately 1,000 m to 3,000 m in thickness.

The stratigraphic nomenclature is based on the nomenclature devised from geologic mapping (e.g., Christiansen and Lipman (1965)), Lipman and McKay (1965), Orkild (1965), Byers et al. (1976a), Carr (1984), and Scott and Bonk (1984), and is shown below, from oldest to youngest. [INN 3.1.1.2.2.3-1] This nomenclature also incorporates data from geological and geophysical studies of a number of drillholes, shown in Figures 3.1.1.2.2.3-1, 3.1.1.2.2.3-2 and 3.1.1.2.2.3-3, and reported by Spengler et al. (1979, 1981), Spengler and Rosenbaum (1980), Maldonado and Koether (1983), Rush et al. (1983), and Scott and Castellanos (1984). This nomenclature will be revised to incorporate recent revisions to the site stratigraphy. The stratigraphic sequence consists of seven formations, some with several members:

- Pre-Lithic Ridge volcanic and volcanogenic rocks
- Lithic Ridge Tuff
- Dacitic lava and flow breccia
- Crater Flat Tuff
  - Tram Member
  - Bullfrog Member
  - Prow Pass Member
- Tuffaceous Beds of Calico Hills
- Paintbrush Tuff
  - Topopah Spring
  - Pah Canyon

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- Yucca Mountain
- Tiva Canyon

- Post-Paintbrush ash-flow tuffs at Yucca Mountain
- Timber Mountain Tuff

The ages, magnetic polarities, and thicknesses of mid-Tertiary volcanic rock units in the site area are illustrated in Table 3.1.1.2.2.3-1. Each formation is discussed in more detail in the following text.

a. Pre-Lithic Ridge volcanic and volcanogenic rocks. As shown in Table 3.1.1.2.2.3-1, the oldest known volcanic and volcanogenic rocks in the Yucca Mountain area were deposited approximately 14 million years before present (mybp), and underlie the Lithic Ridge Tuff. These Pre-Lithic Ridge rocks are difficult to correlate, because they are heterogeneous and show varying degrees of alteration. In addition, they are not known to be exposed at the surface; they may be exposed in the Bullfrog Hills to the west of Yucca Mountain, but this has not been confirmed (Broxton et al., 1989).

In drillholes USW G-1, G-2, and G-3, the Pre-Lithic Ridge sequence consists generally of bedded and ash-flow tuffs, quartz-latitic to rhyolitic lavas and flow breccia (Broxton et al., 1989), with thickness varying from approximately 45 m in USW G-3, to approximately 350 m in USW G-2. In drillhole UE-25p#1, pre-Lithic Ridge volcanic and volcanogenic rocks are approximately 45 m thick and consist of a rhyolitic, compositionally homogeneous, simple cooling unit, ash-flow tuff.

In drillhole USW G-1, the sequence contains reworked tuffaceous sediments, and alteration products, such as clays, calcite, and chlorite. Calcite also is present in drillhole UE-25p#1.

b. Lithic Ridge Tuff. A thick section of massive ash-flow tuff overlying the older ash-flow and bedded tuff of the Pre-Lithic Ridge sequence has been named the Lithic Ridge Tuff (Carr et al., 1984). Its thickness ranges from 185 m at drillhole USW G-2, north of the proposed repository, to 304 m in drillholes USW GU-3 and USW G-3, at the south end of the repository. The unit is nonwelded to moderately welded and has been extensively altered to smectites and zeolites.

The Lithic Ridge Tuff is distinguished by a relatively low volume-percentage of quartz phenocrysts (3 to 11 percent of the total phenocrysts) with biotite as the major mafic phenocryst, hornblende and sphene as accessory minerals, and an abundance of lithic fragments, including fragments of distinctive spherulitic rhyolite lava. These lithic fragments make up 10 to 40 percent of the rock (Broxton et al., 1989). Many slight variations in the degree of welding, phenocryst ratios, and lithic-fragment content suggest that the unit is the result of several eruptive surges.

The Lithic Ridge Tuff is separated from the overlying dacitic lava and flow breccia (where present) by as much as 14 m of bedded tuff. At the southern end of the

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repository (drillhole USW G-3), this layer consists of a moderately to well-indurated bedded tuff that is 8.5 m thick, and is composed mainly of pyroclastic fall material.

c. Dacitic lava and flow breccia. Dacitic lava and autoclastic flow breccia overlie the Lithic Ridge Tuff in deep drillholes in the northern and western parts of Yucca Mountain, but are absent elsewhere. The thickness of the unit is 22 m in drillhole USW G-2, 112 m in drillhole USW H-1, and 249 m in drillhole USW H-6. In drillhole USW G-1, most of the unit is flow-breccia made up of angular to subangular fragments of dacite, commonly from 2 to 10 cm long, which is intercalated with a few lava flows ranging from 1 to more than 17 m thick. In contrast, the unit in drillhole USW G-2 consists largely of lava with flow breccia confined to the uppermost and lowermost parts.

The phenocryst content is variable, but mafic minerals, mostly hornblende and clinopyroxene, are prominent. Much of the unit has been moderately to intensely altered to smectite clays and zeolites.

d. Crater Flat Tuff. Three rhyolitic, ash-flow-tuff sheets have been assigned to the Crater Flat Tuff on the basis of stratigraphic relations and petrologic and geochemical characteristics. They are, in ascending order, the Tram Member (Carr et al., 1984), the Bullfrog Member, and the Prow Pass Member (Byers et al., 1976b). The Crater Flat Tuff is distinguished from other units in the vicinity of Yucca Mountain by the relative abundance of quartz and biotite phenocrysts. In addition, the Prow Pass Member and, to a lesser degree the Bullfrog Member, contain distinctive lithic inclusions of reddish-brown mudstone. At Yucca Mountain, the Crater Flat Tuff overlies dacitic lavas and flow breccias in the northern part of Yucca Mountain, and the Lithic Ridge Tuff in the southern part (Broxton et al., 1989).

(1) Tram Member

Two subunits are distinguished within the Tram Member on the basis of the relative abundance of lithic fragments; the lower subunit is rich in these fragments throughout, and the upper subunit is poor (Spengler et al., 1981). Inclusions of lava fragments constitute as much as 33 percent of rock volume in the lower subunit. South and east of drillhole USW G-1, more than two-thirds of the basal part of the lower subunit characteristically contains zeolite minerals, including clinoptilolite, mordenite, and analcime. Other alteration products are illite-smectite, calcite, kaolinite, and in places, albite and pyrite. The upper 20 to 65 m of the lower subunit have less secondary alteration. In drillhole USW G-1, the upper part of the lower subunit contains 10 to 15 percent smectite and 10 to 15 percent zeolites, including clinoptilolite, mordenite, and analcime.

The upper subunit of the Tram, from 126 to 171 m thick, is partially welded and has a microcrystalline ground-mass. Lithic fragments commonly constitute less

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than 5 percent of the rock. Zeolite and clay minerals are scarce or absent. However, in drillhole USW G-4 the top 20 m of the upper subunit contain some 40 percent clinoptilolite and mordenite (Spengler and Chornack, 1984).

In drillhole USW G-2, parts of the Tram are conspicuously different (Maldonado and Koether, 1983). The upper subunit described above is absent; instead, the uppermost interval, 46 m thick, contains 50 to 95 percent lithic fragments of tuff and lava of both rhyolitic and intermediate composition. The abundance of lithic fragments nearly masks the pyroclastic texture. The lithic-rich interval rests gradationally on ash-flow tuff with 10 to 30 percent lithic fragments. The lateral stratigraphic relations between the Tram in drillhole USW G-2 and the lower two subunits of the Tram in drillholes USW GU-3 and USW G-3 suggest that it comprises at least 28 separate magmatic pulses (Scott and Castellanos, 1984).

The Tram Member typically contains between 6 percent and 13 percent phenocrysts, consisting primarily of quartz, sanidine, and plagioclase. Mafic minerals include biotite and traces of hornblende. Higher quartz contents and lack of sphene serve to distinguish the Tram member from the underlying Lithic Ridge Tuff, in addition to the differing content of lithic fragments (Broxton et al., 1989).

The Tram is separated from the overlying Bullfrog Member by 6 to 22 m of ash-fall and reworked tuff, composed primarily of zeolitic pumice clasts (Scott and Castellanos, 1984; Spengler and Chornack, 1984).

(2) Bullfrog Member

The ash-flow tuff comprising most of the Bullfrog Member is from 68 to 187 m thick. In drillholes north of USW G-4, it seems to be a simple cooling unit, in which nonwelded to partially welded zones enclose a moderately to densely welded core. However, towards the south, in drillholes USW GU-3, USW G-3, and USW G-4, it is compound; composed of two welded zones separated by a 1-m-thick bed of welded ash-fall tuff. Most of the upper part of the member is devitrified or shows evidence of vapor-phase crystallization.

Total phenocryst content in the Bullfrog Member is 12 to 17 percent, above the ash-fall layer, and 9 to 11 percent below (Spengler and Chornack, 1984). The percentage of quartz phenocrysts is lower below the ash-fall layer, whereas the percentage of alkali, feldspar phenocrysts remains constant, and that of plagioclase increases below the ash-fall layer. In drillholes USW G-1, USW G-2, and USW G-4, the upper part of the Bullfrog is partially altered to zeolites, or clays, or both; in drillhole UE-25b#1, only the lower 21 m have been slightly altered to clay; in drillholes USW GU-3, and USW G-3, little to no alteration has occurred.

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(3) Prow Pass Member

The ash-flow tuff of the Prow Pass Member is from 80 to 200 m thick, and is similar in appearance and petrologic characteristics to the Bullfrog. However, the two members are distinguished, because mudstone fragments in the Bullfrog are fewer and smaller than in the Prow Pass and occur in equal amounts with rhyolite lava fragments; and the Bullfrog contains more biotite and hornblende. The Prow Pass is also distinguished from other units of the Crater Flat Tuff by its orthopyroxene content and by the extreme resorption of quartz phenocrysts. Generally, the Prow Pass is devitrified and only slightly welded, but the upper part is vitric to partly vitric in several drillholes.

The phenocryst content is approximately 10 percent, consisting mostly of sanidine, quartz, and plagioclase (Broxton et al., 1989). The top and bottom parts are commonly altered to clay and zeolites. A zone of slight to intense zeolitization and silicification, from 25 to 55 m thick, occurs in the lower part, except in drillholes USW G-1, GU-3, and USW G-3.

A sequence of ash-fall tuff, reworked tuff, and tuffaceous sandstone as much as 20 m thick, separates the Crater Flat Tuff from the overlying tuffaceous beds of Calico Hills.

e. Tuffaceous Beds of Calico Hills. In outcrop, the Tuffaceous Beds of Calico Hills comprise a sequence of ash-flow and ash-fall tuff, volcaniclastic sediment, and rhyolitic lava (McKay and Williams, 1964). Tuffaceous units commonly occur in this unit beneath the central part of Yucca Mountain. The unit consists of massive, homogeneous, non-welded ash-flow tuffs totalling 27 to 289 m in thickness, separated in places by ash-fall and reworked tuff beds as much as 4 m thick. The non-welded tuff is mostly phenocryst-poor (less than 5 percent), but locally it is rich in lithic fragments, some as much as 6 cm long. Underlying most of the northern part of Yucca Mountain, the entire unit is zeolitized, but it remains vitric in the southern part (drillholes USW GU-3 and USW G-3). Zeolites (predominantly clinoptilolite and mordenite) constitute 60 to 80 percent of the rock volume (Spengler et al., 1979; 1981).

Tuffaceous beds of the Calico Hills in drillhole USW GU-3 are characterized by having an uppermost part that is relatively rich in mafic phenocrysts; 8 percent of the phenocrysts consist of biotite, hornblende, orthopyroxene, and opaque phases, with plagioclase as the dominant feldspar. The middle part of the unit has an intermediate phenocryst assemblage, and the lowest part is relatively rich in felsic phenocrysts; only a trace of biotite and orthopyroxene is present, and sanidine is the dominant feldspar. Also, the lowest part of these tuffs contains larger and more abundant lithic fragments than do the two upper parts. This sequence probably represents at least three separate magmatic pulses that tapped different parts of one chemically zoned magma chamber.

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Several lines of evidence suggest that the sequence of tuffs found in the core from drillhole USW GU-3, south of the proposed repository, may differ from the tuffaceous beds of Calico Hills in drillholes USW G-2 and USW G-1, north of the proposed repository [INN 3.1.1.2.2.3-2]:

(1) The hornblende and orthopyroxene found in the upper part of drillhole USW G-3 are not found in other cores.

(2) In drillhole USW GU-3, the upper part of these tuffs is characterized by a higher phenocryst content, a higher mafic-phenocryst content, and a plagioclase/sanidine ratio greater than one; the lower part is characterized by a lower phenocryst content, a very low mafic-phenocryst content, and a plagioclase/sanidine ratio less than one. This sequence of changes in phenocryst ratios is the inverse of that observed at drillhole USW G-2 (Maldonado and Koether, 1983).

(3) The tuffaceous beds of Calico Hills thin from north to south; from 290 m in drillhole USW G-2 to 95 m in drillhole USW G-1 over a distance of only 2.5 km. In drillhole USW GU-3, almost 4.5 km farther south, these tuffaceous beds are less than 30 m thick, close to the distal ends of these tuffs (Scott and Castellanos, 1984).

Ash-fall tuff and reworked-tuff beds from 1 to 17 m thick separate the tuffaceous beds of Calico Hills from the overlying Paintbrush Tuff.

f. Paintbrush Tuff. The Paintbrush Tuff is composed of four ash-flow tuff cooling units that have been named, in ascending order, the Topopah Spring, Pah Canyon, Yucca Mountain, and Tiva Canyon Members. The Topopah Spring and Tiva Canyon Members crop-out to the east and south of the Timber Mountain-Oasis Valley caldera complex. The Pah Canyon and Yucca Mountain Members are relatively small cooling units exposed south of the Timber Mountain-Oasis Valley caldera complex. At Yucca Mountain, these members thin southward, away from the caldera complex, and are absent in drillholes south of UE-25a#1. Bedded and non-welded ash-flow tuffs, and lavas, are locally present between the members of the Paintbrush Tuff at Yucca Mountain (Broxton et al., 1989).

(1) Topopah Spring

The Topopah Spring Member of the Paintbrush Tuff is a multiple-flow, compound cooling unit (Lipman et al., 1966). The description presented here is based mainly on reports by Spengler and Chornack (1984), Scott and Castellanos (1984), Scott and Bonk (1984), and Byers (1985).

A nonwelded to partially welded glassy basal zone, from 10 to 20 m thick, grades upward by increased welding into a densely welded basal vitrophyre,

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from 10 to 25 m thick. The basal vitrophyre grades abruptly upward into the lower nonlithophysal zone, which is a densely welded, devitrified, unvesiculated zone from 27 to 56 m thick. This zone grades upward into the lower lithophysal zone, from 43 to 117 m thick. Overlying the lower lithophysal zone is the middle nonlithophysal zone, from 20 to 50 m thick. This zone represents a second eruptive pulse, because the ground-mass at the base is finer grained than the groundmass elsewhere in the zone, indicating that the base has been chilled. The upper lithophysal zone, from 54 to 96 m thick, represents the vesiculated part of the second eruptive phase. Finally, a third eruptive pulse of quartz-latitic, ash-flow tuff forms the caprock zone, from 39 to 62 m thick. A thin (approximately 1 m) vitrophyre exists in most places in the uppermost part of the caprock and resulted from chilling at the upper surface of the ash flow.

The Topopah Spring Member of the Paintbrush Tuff is petrologically complex, because it is a thick, compound cooling unit with greatly varying phenocryst compositions and abundances. In the lower three-fourths of the member, within the proposed host rock, phenocryst abundances are less than 2 percent (typically 0.5 percent) and the phenocrysts include sanidine, plagioclase of andesine to oligoclase composition, and minor quantities of quartz, biotite, amphibole, iron-titanium oxides, allanite, and zircon. However, the Topopah Spring is compositionally zoned from a crystal-poor rhyolitic base to a crystal-rich quartz-latite top (Lipman et al., 1966). The upper one-fourth of the member contains more phenocrysts; up to 22 percent toward the top. Clinopyroxene occurs as minor phenocrysts in the upper part of the unit.

The major mineral variations in the host rock are due to devitrification, vapor-phase crystallization, and alteration. Small amounts of smectite (up to 15 percent) occur in the core from drillhole USW G-2. Among the devitrification products, the content of cristobalite is variable (in drillholes USW G-2 and UE-25a#1). Vapor-phase minerals consist mostly of cristobalite and alkali feldspars, although various other silicate and oxide minerals also occur (Vaniman et al., 1984; Byers, 1985).

(2) Pah Canyon, Yucca Mountain, and Tiva Canyon Members

The Pah Canyon and the Yucca Mountain members are relatively thin, non-welded ash flow tuffs, which are the distal edges of flow sheets that thicken to the northwest (Lipman and Christiansen, 1964; Orkild, 1965). The Pah Canyon attains a maximum thickness of 71 m; the Yucca Mountain, 29 m. The Pah Canyon contains up to 15 percent phenocrysts, mainly alkali feldspar, plagioclase, and biotite. In contrast, the overlying Yucca Mountain Member is a distinctively uniform, shard-rich tuff containing small amounts of pumice, phenocrysts, and lithic fragments.

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The uppermost unit of the Paintbrush Tuff at Yucca Mountain, the Tiva Canyon Member, is a multiple-flow, compound cooling unit made up of ash flows that erupted approximately 12.5 mybp from the Claim Canyon Caldera segment north of Yucca Mountain (Lipman et al., 1966; Byers et al., 1976b). The Tiva Canyon caps most of the surface of Yucca Mountain, and consists of a moderately to densely welded, devitrified central part underlain by a less densely welded, vitric zone. Compositional zonation in the Tiva Canyon ranges from rhyolite in the lower and middle parts to quartz latite near the top (Gibson et al., 1992). Phenocryst mineralogy of the Tiva Canyon and Topopah Spring are similar with the addition of phenocrysts of sphene in the Tiva Canyon.

Ten mappable units are distinguished in the Tiva Canyon (Scott et al., 1983; Scott and Bonk, 1984). In ascending order, these units are informally designated as follows:

- columnar;
- hackly;
- lower lithophysal;
- red clinkstone;
- gray clinkstone;
- rounded step;
- lower cliff;
- upper lithophysal;
- upper cliff; and
- caprock.

The rounded step subunit is laterally equivalent to the combined red and gray clinkstone subunit.

The Tiva Canyon thickens southward from approximately 90 m to nearly 140 m, then thins again to nearly 125 m in the southernmost part of Yucca Mountain. Within the Tiva Canyon, the relative proportion of the total unit represented by the thickness of the two lithophysal subunits decreases southward from approximately 30 to 10 percent. As the lithophysal zones thin, the nonlithophysal zones thicken.

g. Post-Paintbrush ash-flow tuffs at Yucca Mountain. Petrochemical work by Warren and Byers (1986) has identified a nonwelded ash-flow and ash-fall tuff sequence between the top of the Tiva Canyon Member of the Paintbrush Tuff and the base of the overlying Rainier Mesa Member of the Timber Mountain Tuff, in the vicinity of Yucca Mountain. The sequence is present in the subsurface beneath alluvial deposits on the eastern flank of Yucca Mountain, in the area proposed for the surface facilities of the repository (Carr, 1992). The sequence ranges in thickness from zero to 61 m and is intermediate in composition between that of the underlying Tiva Canyon and the overlying Rainier Mesa units. The sequence is a possible lateral stratigraphic

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equivalent of the Pinyon Pass (?) and tuff of Chocolate Mountain, located north of Yucca Mountain in the vicinity of the Claim Canyon cauldron segment (Byers et al., 1976a).

h. Timber Mountain Tuff. The 11.3 million year old Rainier Mesa Member of the Timber Mountain Tuff (Marvin et al., 1970) is locally present in valleys on the flanks of Yucca Mountain (USGS, 1984). It is a series of ash flows and ash falls that occurs only on the downthrown side of large faults (Gibson et al., 1992).

At Yucca Mountain, the Rainier Mesa Member has a maximum thickness of approximately 46 m. It is nonwelded and glassy at the base, grading upward into partly-welded devitrified tuff near the interior. Its phenocrysts (10 to 25 percent) are mainly quartz and alkali feldspar, with some plagioclase and biotite (Lipman and McKay, 1965; Gibson et al., 1992).

3.1.1.2.2.4 Younger Basalt

The youngest volcanic rocks at Yucca Mountain (10 million years old) are basalt dikes less than 1 m thick that intrude a fault and nearby fracture in the northwest part of Yucca Mountain, at the head of Solitario Canyon near Little Prow (USGS, 1984). The basalt is very fine grained and locally vesicular, with sparse microphenocrysts of olivine, plagioclase, and rare clinopyroxene, grading into an intergranular groundmass. The basalt was brecciated near its margins by later movement along the fault. Younger basaltic flows and cones are present in the area around Yucca Mountain.

3.1.1.2.2.5 Surficial Deposits

Most studies of the stratigraphy of surficial deposits in the Yucca Mountain area are based primarily on the work of Hoover and Morrison (1980), Hoover et al. (1981), and Hoover (1989), who developed an initial Quaternary/Tertiary stratigraphy and defined correlation characteristics. Wesling et al. (1992), Taylor (1986), and Lundstrom (personal communication, 1993) have used these correlation characteristics and Quaternary stratigraphic nomenclature as a basis for their work in Quaternary stratigraphy. These correlation characteristics rely on physical and morphological characteristics of landscape elements, such as landform, drainage network, soils, topographic position, desert pavement, desert varnish, depositional environment, and lithology. Other workers, such as Bull (1984), Wells et al. (1984), and McFadden et al. (1987), contributed to the understanding of dominant desert geomorphic processes and responses to climate change in the Yucca Mountain area and the southern Great Basin region (Wesling et al., 1992). The following brief description of stratigraphy is based primarily on the work of Hoover et al. (1981).

According to Hoover et al. (1981), the late Tertiary and Quaternary surficial sedimentary deposits of the Yucca Mountain area consist of colluvium; fan alluvium; eolian sand sheets; ramps and dunes; lacustrine or marsh sediments; and playa deposits. These range in age from

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late Pliocene for some of the lacustrine sediments to Holocene for the youngest alluvial deposits (Swadley et al., 1984). The deposits are grouped into five major units:

- Late Pliocene and early Pleistocene lacustrine deposits;
- Late Pliocene and early Pleistocene alluvial deposits;
- Middle to late Pleistocene alluvial and eolian deposits;
- Holocene alluvial and eolian deposits; and
- Spring and marsh deposits.

The distribution of these units is shown on a generalized surficial deposit map in Figures 3.1.1.2.2.5-1 and 3.1.1.2.2.5-2. The map units shown on these figures are described in the following Subsections (3.1.1.2.2.5 a and b).

a. Sediment Deposits. Late Pliocene and early Pleistocene deposits (shown as QTld on Figures 3.1.1.2.2.5-1 and 3.1.1.2.2.5-2) are the oldest surficial deposits in the area. They consist of lacustrine deposits of unconsolidated to moderately indurated marl and silt that locally contain beds of limestone, sand, and fine-grained volcanic ash. According to Swadley et al. (1984), the sediments were deposited in Lake Amargosa, which occupied part of the Amargosa Valley during the late Pliocene; remnants of the Lake may have persisted into the Quaternary. However, other workers, such as Hay et al. (1986) and Donovan (1991), regard these sediments as swamp or marsh deposits.

Other late Pliocene and early Pleistocene deposits (shown as QTa on Figures 3.1.1.2.2.5-1 and 3.1.1.2.2.5-2) consist predominantly of debris flows with sparse, bedded, fluvial sediments; they occur as dissected fans and fan-remnants that are adjacent to bedrock ranges and, less commonly, as isolated outcrops several kilometers from the ranges. These deposits are moderately indurated, coarse, angular, unsorted gravel with minor amounts of sand-to-clay-size material. In most exposures, they are partly cemented with calcium carbonate. Unit QTa also contains colluvium, which consists of unsorted, fine to course angular rubble on steep slopes, such as Little Skull Mountain, in the site area.

Middle-to-late Pleistocene deposits (Q2 on Figures 3.1.1.2.2.5-1 and 3.1.1.2.2.5-2) consist of fan alluvium, fluvial and eolian sands, and local lenses of volcanic ash. These deposits have been divided into five mappable units on the basis of relative age and lithology: three alluvial units, eolian dunes and sand sheets, and fluvial sand sheets. The fan deposits are typically unconsolidated, poorly to well sorted, non-bedded to well-bedded, angular-to-rounded gravel with sand and silt in the matrix. Interbeds of silty sand are locally common. These deposits generally overlie older alluvial deposits on middle-to-upper pediment slopes, and they occur as terrace deposits in larger stream valleys.

The eolian deposits occur as dunes and sand sheets in, and adjacent to, the Amargosa Valley. Ramps of fine, well-sorted sand as much as 50 m thick flank many of the hills bordering the Amargosa Valley on the north. The fluvial sand sheets occur along

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major streams and along drainages downstream from dunes. Sheets consist of water-laid, fine-grained, medium-gravelly sand or stream-reworked, windblown sand.

Holocene deposits (Q1 on Figures 3.1.1.2.2.5-1 and 3.1.1.2.2.5-2) in the Yucca Mountain area consist of fluvial sand and gravel, and eolian sand. Fluvial sand and gravel deposits typically are poorly to well bedded and moderately well sorted. Gravel deposits locally contain numerous thin beds of sand. Holocene deposits occur mainly as thin, broad fans, downstream from incised stream channels on pediment slopes. Eolian sand deposits consist of well-sorted, fine sand that occurs as small dunes and irregularly-shaped sheets over much of the Amargosa Valley. Eolian deposits commonly overlie all other units.

b. Soils. Soils formed on Quaternary deposits of Qla, Qlb, Qls, and Qle (illustrated on Figures 3.1.1.2.2.5-1 and 3.1.1.2.2.5-2) are very weakly developed and contain little or no secondary calcium carbonate or opaline silica. Soils formed on Qlc and Q2a have a weakly developed horizon of calcium-carbonate accumulation (Bk horizon) about 20 to 50 cm thick. The undersides of stones in the Bk horizon commonly have thin, discontinuous coatings of calcium-carbonate or silica or both.

Soils formed in deposits of Q2b commonly have a silica-enriched duripan (Bqm horizon) about 50 cm thick; this typically overlies a moderately developed horizon of secondary calcium-carbonate accumulation (Bk/k horizon) about 50 to 70 cm thick. Secondary silica in the duripan typically occurs as finely disseminated matrix cement and as coatings and pendants on stones. Secondary calcium-carbonate in the bk/k horizon typically occurs as coatings on stones, bridges between some of the stones, and locally as a finely disseminated matrix cement. Soils formed in deposits of Q2c, Q2e, and Q2s commonly have a silica-indurated Kqm horizon that is cemented with calcium carbonate and is about 50 cm thick. The Kqm horizon typically overlies a Bk/k horizon that is enriched in calcium-carbonate and is about 40 to 50 cm thick.

Soils formed in QTa deposits are distinguished from younger soils by their prominent Kqm horizons, which are usually more than 1 m thick. The upper part of the Kqm horizon commonly consists of thin layers enriched in calcium-carbonate and cemented by thin laminae of opaline silica.

c. Other studies. Szabo et al. (1981) report on two possible spring deposits near Yucca Mountain. The first deposit is located in Crater Flat at the south end of Yucca Mountain and is described as a seep-deposited tufa or calcrete intercalated in Q2 alluvium. The deposit shows some evidence of spring-water deposition and was dated by uranium-thorium methods at 78,000 ± 5,000 yrs. The second deposit is a nodular, tufa spring deposit at the south end of Crater Flat. The deposit is also overlain by Q2 alluvium, but relationships suggest that spring activity continued after deposition of the alluvium. This deposit has been dated at about 30,000 yr. Marsh deposits are found at several locations in the Amargosa Valley.

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Peterson (1988) evaluated geomorphic relationships and soils in the Crater Flat area, west of Yucca Mountain. His goals were to describe soils on geomorphic surfaces to define characteristics that may define a surface; and evaluate the relationship of geomorphic surfaces and soils to the surficial mapping of Swadley et al. (1984). Peterson (1988) identified five geomorphic surfaces in the Crater Flat area, and assigned ages to them based on radiometric dating, soil-development, and stratigraphic position. Based on this work, he criticized the stratigraphy of Swadley et al. (1984), for several reasons. First, he held that they had incorrectly used the Av horizon as the basis for distinguishing Pleistocene from Holocene surfaces. Second, Peterson (1988) held that Hoover et al. (1981) incorrectly treated pedogenic soil horizons, as if they date the surficial deposit in which they occur, rather than dating the onset of long-term stability of the surface developed on the deposit. Finally, Peterson (1988) held that Swadley et al. (1984) had failed to recognize several, separate, geomorphic surfaces in the Crater Flat area.

Westling et al. (1992) reviewed available literature on the Tertiary-Quaternary surficial stratigraphy of the Yucca Mountain area, particularly Midway Valley. Their evaluation of the stratigraphy of Hoover et al. (1981) and Hoover (1989) suggested that their correlation characteristics provide a useful basis for reconnaissance-level and detailed mapping, but more quantitative data are needed. In particular, they suggested that geologic mapping be extended along Yucca and Fortymile Washes, to facilitate regional correlation. They also suggested that more quantitative age dates are needed. [INN 3.1.1.2.2.5-1]

3.1.1.2.3 Site Structural Geology and Tectonics

[The Technical Basis Report for Tectonics is scheduled for completion in mid FY 96 and will be the basis for updates to the tectonic sections in Revision 1].

3.1.1.2.3.1 Faulting at Yucca Mountain

[The text for this section will be provided in future revisions of the Annotated Outline.]

3.1.1.2.3.2 Quaternary Faulting History

This section summarizes relevant published information on Quaternary faults that may pose potential seismic hazards to the repository site (Figures 3.1.1.2.3.2-1 and 3.1.1.2.3.2-2) [INN 3.1.1.2.3.2-1]. Information that is necessary for evaluating surface rupture and/or ground shaking hazards at the site is emphasized, including general fault activity and geometry, amounts and types of displacements, number, timing, and size of prehistoric earthquakes, displacement rates, and recurrence intervals. This information forms the basis for identification of potential seismic sources of large surface-faulting earthquakes in the seismic hazard assessment (Subsection 3.1.1.2.4.4) and Table 3.1.1.2.4.4-1 [INN 3.1.1.2.4.4-1]. Because the 150 year-long historical record of seismicity is too short to comprehensively characterize potential earthquake sources for the YMR, it is also necessary to consider

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paleoseismic (prehistoric earthquake) information from the geologic record. It should be noted that the resolution for detecting earthquakes in the geologic record is limited to larger earthquakes, causing surface rupture (generally greater than magnitude 6.0 to 6.5 in the Basin and Range Province).

In keeping with recommendations in the Topical Report on Methodology to Assess Seismic Hazard at Yucca Mountain, all known or suspected Quaternary faults within 100 km of the site were initially considered (Figure 3.1.1.2.3.2-1) [INN 3.1.1.2.3.2.-1]. Piety and others (1992, 1993-in review) identified 140 Quaternary faults within 100 km of Yucca Mountain in their comprehensive compilation. However, many of these faults are characterized by short traces, indistinct geomorphic expression, low rates of activity, and are distant enough from the site that they do not pose a threat from surface rupture, nor do they contribute significantly to the ground-motion hazard at the site (eg., Tolicha Peak fault zone, Bullfrog Hills faults, and Checkpoint Pass fault). Thus, included here are only the known or suspected Quaternary faults that may pose a potential threat from surface rupture (local faults within 20 km), and, faults that are significant contributors to the ground-shaking hazard as determined in the preliminary hazard assessment for the Exploratory Studies Facility (regional faults within 100 km). Also included are some faults that are not significant contributors themselves, but are thought to be somehow associated with faults that are significant contributors (e.g., the Cane Spring and Mine Mountain faults, which are thought to be associated with the Rock Valley fault zone; Donovan, 1991). Faults are listed in order of proximity to the site.

Piety et al. (1992) discuss in more detail many of the faults included here, along with less active or more distant faults not included here. Their forthcoming final report, along with results from many ongoing paleoseismic studies [INN 3.1.1.2.3.2-1] will fill many, but not all, of the significant information gaps for the following faults.

3.1.1.2.3.2.1 Quaternary Faults Within 20 km of the Site

There are nine known or suspected Quaternary faults within 20 km of the site that pose a potential threat from surface rupture and near-field, strong ground motions (Figure 3.1.1.2.3.2-2, Piety et al., 1992). Eight of these are closely-spaced (1 to 3 km), subparallel, normal faults that generally strike to the north and dip steeply to the west, including the Ghost Dance, Solitario Canyon, Bow Ridge, Fatigue Wash, Paintbrush, Windy Wash, Crater Flat, and Stagecoach Road fault zones. Ramelli et al. (1991) have suggested that recent faulting could have involved concurrent rupture on more than one of these faults. However, the paleoseismic information is presently too incomplete to conclude whether these faults have ruptured independently or concurrently in the past.

Ghost Dance Fault - The Ghost Dance fault is a north-striking, steeply west-dipping, normal fault that transects the proposed site and is roughly 3 to 9 km long (Scott and Bonk, 1984; Swadley et al., 1984; Wesling et al., 1991). Swadley et al. (1984) found no evidence for Quaternary movement during their surficial mapping and trench investigations. However, Ramelli et al. (1991) speculate that it may have had Quaternary displacement and Simonds and Whitney (Figure 3.1.1.2.3.2-2, INN 3.1.1.2.3.2-1) included it on their Quaternary fault

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map, noting the difficulty in conclusively demonstrating that the Ghost Dance fault was inactive during the Quaternary due to the general lack of Quaternary deposits overlying the fault. Detailed mapping revealed many additional fault traces in bedrock associated with the Ghost Dance fault, forming a deformation zone as wide as 213 m (Spengler et al., 1993; Figure 3.1.1.2.3.2-3). These anastomosing, near-vertical faults displace Tertiary volcanic rocks 3 to 6 m, down to the west. Ongoing and future trenching studies by the USGS [INN 3.1.1.2.3.2.1-1] may clarify whether or not the Ghost Dance fault displaces Quaternary deposits.

Solitario Canyon Fault - The Solitario Canyon fault is a north-striking, west-dipping normal fault that may be as long as 21 km and lies along the western boundary of the proposed site (Swadley et al., 1984; Scott and Bonk, 1984; Frizzell and Shulters, 1990). Swadley et al. (1984), identified offset of early Quarternary alluvium (QAL), but found no evidence for movement within the last 270ka. Fractures in a trench exposure contained reworked basaltic ash tentatively correlated with an ash from Crater Flat, with an age estimate of 1.2 Ma. From analysis of low-sun-angle aerial photographs, Ramelli et al. (1988, 1991) identified a compound Quaternary fault scarp 1 to 3 m high and 13 km long. Surfaces with various ages are offset different amounts, suggesting repeated movement during the late Quaternary. Alluvial surfaces, believed to be 8 to 11ka, appear to be vertically offset roughly 20 cm, suggesting most recent faulting could be Holocene. Older, late Pleistocene alluvial fans, inferred to be about 20 to 30ka, are vertically offset as much as 1 m. Ramelli et al. (1991) also found a 3-m high bedrock scarp that apparently formed since the middle Pleistocene. Displacement dies out at the northern end of the scarp, whereas the southern end is obliterated by Holocene erosion and deposition.

Bow Ridge Fault - The Bow Ridge fault is a north-striking, west-dipping, normal-oblique fault with dips ranging from 75° (Scott and Bonk, 1984; Menges, 1993) to near vertical (Swadley et al., 1984). It bounds the west side of Bow Ridge and the fault trace is roughly 2 km east of the site. Length estimates vary from 9 to 20 km and it may merge with the Paintbrush Canyon-Stagecoach Road fault system at its southern end (Maldonado, 1985; Frizzell and Shulters, 1990; Swadley et al., 1984). Evidence for fault scarps on Quaternary sediments is equivocal. Scott and Bonk (1984) indicated that the fault is concealed beneath alluvium along most of its length. Swadley et al. (1984) found no evidence for fault scarps in Quaternary deposits and Gibson et al. (1991) found no offsets of alluvial surfaces. However, based on interpretation of low-sun-angle aerial photographs, Ramelli et al. (1988) suggested that there are small scarps in younger deposits. Wesling et al. (1991) also identified photolineaments in alluvium and colluvium that could be fault-related.

There is also some discrepancy regarding the age of most recent faulting determined from trench investigations. Taylor and Huckins (1986) found faulted colluvium, with a Uranium-series age of 490 ± 90 ka, overlain by unfaulted eolian sediments, with Uranium-trend ages of 90 ± 50 ka and 38 ± 10 ka. They suggested that the most recent faulting was associated with a basaltic ash inferred to be as young as 270 ka, based on correlation with an ash from Crater Flat. Based on their trench investigations, Swadley et al. (1984) suggested that the most recent deformation occurred between 38 and 278 ka, and involved fracturing, but no

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displacement along the Bow Ridge fault zone. However, Gibson et al. (1991) found inconsistencies in their uranium-series and uranium-trend ages and Swan (1992) reinterpreted a trench exposure and found evidence for displacements of middle to late Pleistocene colluvium. Menges (1993) found evidence for five to six events since the middle Pleistocene, with the youngest event having occurred during the late Pleistocene. Vertical displacements ranged from 5 to 20 cm and averaged 10 cm per event. Cumulative vertical displacement was 45 cm. From striations indicating left-lateral slip, Menges estimated 11 to 28 cm of average net-slip per event and cumulative net-slip of 76 to 132 cm. Dates are forthcoming, but his preliminary interpretations suggest recurrence intervals of 1,000 to 10,000 years and slip rates on the order of 0.001 mm/yr.

Fatigue Wash Fault - The Fatigue Wash fault is a 10 to 16 km long, north-striking, west-dipping normal fault that generally bounds the west side of Jet Ridge, and is located roughly 2 km west of the proposed site (Ramelli et al. 1991; Scott and Bonk 1984; Frizzell and Shulters, 1990). Based on analysis of low-sun-angle aerial photographs, Ramelli et al. (1991) found evidence for fault scarps on Quaternary deposits and throw comparable to other faults in the Yucca Mountain vicinity. Displacement dies out to the north where the Fatigue Wash fault zone appears to merge with the Windy Wash fault (Wesling et al., 1991; Ramelli et al., 1991).

Paintbrush Fault - The Paintbrush fault is the northern part of the Paintbrush Canyon-Stagecoach Road fault system. It is roughly 3 km east of the proposed site and is one of many closely-spaced, west-dipping, generally north-striking, normal-oblique faults near Yucca Mountain (Scott and Bonk, 1984). It bounds the east side of Midway Valley, with displacement down to the west, forming a half-graben (Swan et al., 1993). A left-lateral component of displacement is also indicated by slickenlines on Tertiary rocks and offset stream channels (Whitney and Muhs, 1991). Estimates of length for the Paintbrush fault zone vary from 17 to 31 km (Carr, 1984; USGS, 1984; Maldonado, 1985; Frizzell and Shulters, 1990). Swadley et al. (1984) and Gibson et al. (1991) found no evidence for fault scarps on Quaternary deposits; however, Wesling et al. (1991) identified photolineaments in Quaternary sediments that could be fault-related.

Studies of natural and trench exposures suggest that the northern part of the Paintbrush fault zone was less active than the southern part during the Quaternary. Interpretation of exposures in trenches along the northern part of the fault near Yucca Wash suggest the most recent faulting was between 270 and 700ka with small displacements of less than a few centimeters (Swadley et al., 1984). Further south along the fault, close to where it splits into two splays, evidence for five faulting events in the last 700ka is exposed in arroyos (Whitney and Muhs, 1991). Near the base of the exposures, a soil developed on sand deposits containing the Bishop ash is inferred to be about 700ka and is vertically displaced as much as 4.1 m. The estimated net slip is 5.8 m, yielding a slip-rate of about 0.008 mm/yr (Whitney and Muhs, 1991).

A trench exposure on the western splay of the southern Paintbrush fault zone indicates that multiple surface-faulting events (three to five) occurred during the middle to late Pleistocene

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with a suggested average recurrence interval of 1,000 to 10,000 years (Swan et al., 1993). The most recent event offset late Pleistocene sediments, down to the west, with about 15 cm of dip-slip displacement. An average displacement for four earlier events was 40 to 85 cm per event. Total cumulative dip-slip displacement since the middle Pleistocene was 170 to 270 cm. Late Cenozoic slip-rates range between 0.01 to 0.006 mm/yr and appear to be decreasing during the last 13 Ma (Gibson et al., 1990).

Windy Wash Fault - The Windy Wash fault is a north-northeast-striking, west-dipping normal-oblique fault on the west side of West Ridge. It is roughly 3 km east of the proposed site and estimates of length vary from 14 to 25 km (Frizzell and Shulters, 1990; Whitney, 1992; Swadley et al., 1984; Scott, 1990). Two measurements of fault scarps on Quaternary or Tertiary deposits yield maximum slope-angles of 7° and 9°, with a scarp height of 1.5 m and a surface offset of 1 m, respectively (Swadley et al., 1984). Evidence from low-sun-angle aerial photographs is consistent with small Holocene displacements on compound Quaternary fault-scarps (Ramelli et al., 1988). Trench investigations indicate that seven faulting events occurred during the Quaternary, with four events in the last 300 ka, yielding an average recurrence interval of 75ka (Whitney et al., 1986). The maximum cumulative displacement for these events is 1.5 to 3.0 m down, to the west (Whitney, 1992). The most recent event resulted in 10 cm of apparent vertical offset of an eolian deposit with a TL age of 3 to 6.5ka. Geophysical data suggest a component of left-slip, but the amount is unknown (Whitney et al., 1986).

Crater Flat Fault - The Crater Flat fault is a north-northeast-striking, west-dipping fault that is roughly 9 to 13 km long extending along the east side of Crater Flat, about 5 to 6 km west of the proposed site (Frizzell and Shulters, 1990). The Crater Flat fault lies on trend with the "west lava fault" of Ramelli et al. (1988). From their analysis of low-sun-angle aerial photographs, they found a previously unrecognized west-dipping fault, south of Crater Flat fault and west of the southern Windy Wash fault zone, that strikes north-northeast. It is characterized by a scarp less than 1 m high, which is developed on a late Pleistocene alluvial surface with an estimated age of 17 to 30ka (Ramelli et al., 1991). If the "west lava fault" is a continuation of Crater Flat fault, the Crater Flat fault may be as long as 22 km.

Stagecoach Road Fault - The Stagecoach Road fault is the southern part of the Paintbrush Canyon-Stagecoach Road fault system. It is a northeast-striking, northwest-dipping, normal-oblique fault and its northernmost trace is roughly 10 km south of the proposed repository site (Frizzell and Shulters, 1990; O'Neil et al., 1992). Quaternary pediments are displaced 10 to 30 m down to the west and left-slip is also indicated by slickenlines on Tertiary rocks and offset stream channels (Whitney and Muhs, 1991).

Bare Mountain Fault - The Bare Mountain fault is a north-striking, east-dipping, normal-oblique fault that bounds the west side of Crater Flat and is roughly 14 km west of the proposed site (Reheis, 1988; Reheis and Noller, 1991; Carr and Monsen, 1988; Monsen et al., 1990). South of Tarantula Canyon, a 13 km long single trace extends along the base of Bare Mountain and the range-front is steep with faceted spurs, whereas north of Tarantula Canyon, the fault splits into numerous short, north-northeast trending splays, resulting in a total length

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of about 16 km (Figure 3.1.1.2.3.2-4) (Reheis, 1988). Although Hamilton (1988) found no evidence for range-front fault scarps, either in bedrock or sediments, both Reheis (1986; 1988) and Carr (1984) identified a system of range-front faults offsetting Quaternary deposits, down to the east. Fault exposures in bedrock indicate both dip-slip and right-lateral movement (Reheis, 1988; Carr and Monsen, 1988). Reheis (1988) mapped discontinuous fault scarps on alluvial fans of middle Pleistocene to Holocene age and also found stratigraphic evidence for repeated movement since the middle Pleistocene. Based on stratigraphic and geomorphic evidence, Reheis (1988) subdivided the fault into five, very short segments. Two events were identified at a site near Tarantula Canyon, one before 350ka and one younger event. Two events younger than 40 to 50ka were identified at a site to the south. At her southernmost site, Reheis (1988) found evidence for one event about 145 to 160ka and another event younger than 9ka. This later event was inferred to have a minimum displacement of 1.75 m, which yields a vertical displacement rate of about 0.2 mm/yr for the Bare Mountain fault.

Mine Mountain Fault - The Mine Mountain fault is a 20 km-long, left-lateral fault that is roughly 18 km east of the proposed site and is the northernmost fault in the Spotted Range-Mine Mountain structural zone, which also includes the Cane Spring fault and Rock Valley fault zone (Frizzell and Shulters, 1990; Donovan, 1991).

3.1.1.2.3.2.2 Significant Quaternary Faults between 20 and 100 km of the Site

There are 18 additional faults, included here as potential significant contributors to the ground shaking hazard at the potential repository site. These faults exhibit a variety of geometries and kinematic styles, reflecting complex structural interactions between the Walker Lane belt and the eastern boundary of the Basin and Range Province. Despite the complexities, the faults can generally be separated into three groups:

- Faults that strike northeast and have dominantly left-lateral slip (e.g., the Rock Valley, Cane Spring, and Mine Mountain faults);
- Faults that generally strike north and have normal or normal-oblique slip (e.g., the West Pintwater, East Pintwater, Ash Meadows, Oasis Valley, Yucca, Death Valley, Panamint Valley, Ash Hill, Carpetbag, and West Spring Mountains faults); and
- Faults that strike northwest and have dominantly right-lateral, right-lateral-normal, and normal slip (e.g., the Furnace Creek, Hunter Mountain, Keane Wonder, Grapevine, East Nopah, and Pahrump-Stewart Valley faults).

Oasis Valley Fault - The Oasis Valley fault includes two splays that generally strike north, have a total length of about 19 km, and are part of a graben in Oasis Valley, roughly 20 km northwest of the proposed site (Reheis and Noller, 1989; Reheis, 1992). These faults offset early to middle Pleistocene alluvial-fan deposits by as much as 4 m (Reheis and Noller, 1989).

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Cane Spring Fault - The Cane Spring fault is a 25 km long, left-lateral fault that is roughly 24 km east-southeast of the proposed site, is part of the Spotted-Range-Mine Mountain structural zone, and also includes the Mine Mountain fault and the Rock Valley fault (Frizzell and Shulters, 1990; Donovan, 1991).

Rock Valley Fault - The Rock Valley fault is a northeast-striking, southeast-dipping, left-oblique-slip fault with dips varying from 70° to near-vertical and estimates of fault length varying from 19 to 65 km (Yount et al., 1987; Frizzell and Shulters, 1990; Reheis and Noller, 1991; Donovan, 1991; Swadley and Huckins, 1989). It generally lies within Rock Valley, about 25 km southeast of the proposed site. Estimated total left-slip is a few kilometers (Barnes et al., 1982), with much less total vertical displacement (Carr, 1984). Donovan (1991) grouped the Rock Valley fault with the Cane Spring and Mine Mountain faults, two subparallel faults to the north, to form the Spotted Range-Mine Mountain structural zone, a left-lateral northeast-trending zone that cuts across the Walker Lane belt (Carr, 1984). Quaternary fault scarps are preserved along the entire Rock Valley fault, but are most prominent along the central part, where scarp-heights range from less than 1 m to 2.5 m and maximum slope-angles are 5 to 8° (Hinrichs, 1968; Carr, 1974; Szabo et al., 1981; Yount et al., 1987; Swadley and Huckins, 1989; Donovan, 1991; Reheis, 1992). In trench exposures, Yount et al. (1987) measured 2.5 to 3 m vertical separation, down-to-the-north, of early QAL, inferred to be older than 740ka. Vertical separation of deposits about 31 to 38ka was 10 to 32 cm, down-to-the-north. The age of the most recent prehistoric faulting could be quite young, but is not well constrained. Trench investigations indicate the youngest faulting occurred after 31 to 38ka (Yount et al., 1987) and fault scarps on deposits as young as Holocene have been identified along the southern (Donovan, 1991; Szabo et al., 1981) and central parts of the fault (Yount et al., 1987). Inconclusive stratigraphic evidence for a penultimate faulting event suggests it probably occurred between 31 to 38 and 180ka (Yount et al., 1987).

Ash Meadows Fault Zone - The Ash Meadows fault zone is a north-northeast-striking zone of short discontinuous lineaments and scarps that extends for roughly 60 km along the Amargosa River Valley and is roughly 35 km south of the proposed site (Reheis and Noller, 1991; Donovan, 1991; MicKittrick, 1988). Based on geometry, Donovan (1991) subdivided the Ash Meadows fault zone into three sections:

- A southern section, roughly 48 km long, that strikes north, bounds the east side of the Resting Spring Range;
- A central section, 5 km long, which has no associated range-front and has north-northwest trending fault traces; and
- A northern section, 7 km long, with no associated range-front and two sets of faults, one striking northwest and one striking northeast.

Fault scarps generally face west and trench exposures along the central section indicate fault dips of 50 to 58° to the west with, slickensides indicating normal dip-slip movement, down to

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the west (Donovan, 1991). Scarps along this section are 0.35 to 1.2 m high, with maximum slope angles of 10.5 to 12.5° (Donovan, 1991). Age of the most recent faulting is not well-constrained, but Donovan identified fault scarps on Holocene alluvium in the northern Ash Meadows fault zone, whereas her trench investigations on the central section suggested the most recent faulting occurred before 10ka, possibly around 38ka. Mapping along the southern Ash Meadows fault zone indicates that faults offset latest Pleistocene to early Holocene deposits, but do not offset all Holocene deposits (McKittrick, 1988). Based on a 50 m displacement of 3.2 Ma volcanic rocks along the northern section of the Ash Meadows fault zone, Donovan (1991) estimated an average vertical-displacement rate of 0.016 mm/yr.

Yucca Fault - The Yucca fault is a north-striking, east-dipping fault that is within Yucca Flat, about 40 km northeast of the proposed site (Reheis, 1992; Dohrenwend et al., 1992; Frizzell and Shulters, 1990; Swadley and Hoover, 1990; Carr, 1974). Length estimates vary from 22 km (Reheis, 1992) to 32 km (Carr, 1974). It appears listric, with dips flattening from 75° to 80° near the surface to 55° to 65° at depth (Carr, 1974). Fault scarps at the northern end displace an alluvial surface 15.3 m, down to the east (Fernald et al., 1968); whereas, an en echelon, left-stepping pattern of scarps suggests a component of right-slip as well (Carr, 1974, 1984). Geomorphic expression dies out to the south, but some displacement may be historic and induced by nuclear testing at this end (Carr, 1974; Frizzell and Shulters, 1990). The age of most recent prehistoric faulting is not well-constrained, but could be as young as late Pleistocene (Frizzell and Shulters, 1990; Reheis, 1992; Dohrenwend et al., 1992). The Yucca fault is probably one of many faults, including the Carpetbag fault system and the Area Three fault, that is partially responsible for formation of the Yucca flat structural basin (Carr, 1984). In fact, fractures on the Carpetbag fault system may actually be secondary features formed during a surface-rupturing event on the Yucca fault zone (Shroba et al., 1988).

Carpetbag Fault System - The Carpetbag fault system is a north-striking, east-dipping, normal-oblique fault system that is estimated to be 16 to 30 km long and is about 43 km northeast of the proposed site (Swadley and Hoover, 1990; Frizzell and Shulters, 1990: Shroba et al., 1988). An underground explosion at Yucca Flat produced fault scarps as high as 1.2 m, but the Carpetbag fault system generally lacks prehistoric scarps on Quaternary deposits (Shroba et al., 1988). The historic rupture also produced left-stepping en-echelon cracks and 15 cm of right-slip, indicating a minor component of lateral offset. Although middle Pleistocene deposits do not appear to be significantly offset, Shroba et al. (1988) found stratigraphic evidence for six episodes of fracturing along the Carpetbag fault system. Their uranium-series dates suggest these events occurred about 30, 45, 65, 100, and 230 ka and they calculate an average recurrence interval, for fracturing during the last 130 ka, of about 25 ka.

Keane Wonder Fault - The Keane Wonder fault is a northwest-striking, west-dipping 25 km long, normal fault that bounds the west side of the Funeral Mountains, roughly 43 km southwest of the proposed site (Reheis and Noller, 1991). The Keane Wonder fault is about 9 km northeast of the Furnace Creek fault zone and about 15 km southeast of the Grapevine fault, which has a similar orientation and sense of displacement (Reheis and Noller, 1991).

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Although the Keane Wonder fault and the Grapevine fault are spatially associated with the Furnace Creek-Death Valley fault system, the tectonic relationship between these faults has not been well-studied.

Furnace Creek Fault - The Furnace Creek fault zone is a northwest-striking, near vertical, 190-km-long, right-slip fault that is about 50 km southeast of the proposed site (Brogan et al., 1991). It is a main component of the Death Valley-Furnace Creek fault system, which extends for 350 km from the Garlock fault at the southern end to Fish Lake Valley at the northern end. This fault has accumulated as much as 80 km of total right-lateral displacement and, from south to north, it includes: the southern, central, and northern Death Valley fault, the Furnace Creek fault (including the Fish Lake Valley fault zone), and the Emigrant Peak fault (Stewart et al., 1968). Scarps and offset geomorphic features are prominent along the Furnace Creek fault. Based on scarp morphology, fault geometry, and limited timing information, Brogan et al. (1991) separated the fault into numerous segments. An alluvial fan margin is offset by 46 m of right-slip along the Redwall fan segment (Bryant, 1988). Bryant infers the fan to be about 20 ka, yielding a slip-rate of 2.3 mm/yr. Brogan et al. (1991) measured 64 m of maximum vertical separation of a late Pleistocene surface near Dyer. Hooke (1972) found evidence for four to six distinct events during the last 2ka, which yields an average recurrence of 300 to 500 years. In comparison, Sawyer (1987) found evidence for longer recurrence on the Fish Lake Valley fault zone to the north. From trench investigations, he identified three events that are younger than a unit dated at 3,400 to 4,100 ka. Two of those events are younger than a unit dated at 1,200 yr ka. Displacement was dominantly right-lateral with estimated ratios of lateral to normal slip varying from 3:1 to 10:1. Measurements of vertical displacements per event varied from 3 to 28 cm. Sawyer (1989) calculated vertical slip-rates of 0.18 to 0.32 mm/yr and inferred lateral slip-rates of 0.96 to 1.8 mm/yr.

Death Valley Fault - The Death Valley fault is a north-northwest striking, west-dipping, 75-km-long (not including scarps south of Shore Line Butte), normal-right-slip fault that lies along the west side of the Amargosa Range and the east side of Death Valley, about 55 km southwest of the proposed site (Brogan et al., 1991; Butler et al., 1988; Hunt and Mabey, 1966). It is the southernmost fault in the Death Valley-Furnace Creek fault system. Late Pleistocene and Holocene fault scarps are prominent along the Death Valley fault (Hunt and Mabey, 1966). Scarps are west-facing, heights range from 5 to 9.4 m, and maximum vertical displacement of late Pleistocene surfaces is 15 m (Brogan et al., 1991). Based on scarp morphology and offset of a 200 to 2,000-year-old surface, Brogan et al. (1991) suggest that the age of most recent faulting may be nearly historic. Butler et al. (1988) identified 35 km of right-lateral offset of an alluvial fan along the southern Death Valley fault. Total vertical relief measured on the north Ashford Mill segment is 200 m (Brogan et al., 1991). Slip rate estimates vary from 0.15 to 0.75 mm/yr for the last 2,000 years to as high as 3.5 mm/yr (Brogan et al., 1991). The latter is based on 35 km of total offset and assuming faulting began 10 Ma. Brogan et al. (1991) also found geomorphic evidence for occurrence of three Holocene events on the Artists Drive segment, which yields an average recurrence of about 3,300 years.

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West Spring Mountains Fault - The West Spring Mountains fault is a north-northwest-striking, west-dipping, normal fault that bounds the southwest side of the northern Spring Mountains and is about 55 km southeast of the proposed site (Hoffard, 1991; Donovan, 1991; Dohrenwend et al., 1991; Reheis and Noller, 1991; Reheis, 1992). It has also been referred to as the Spring Mountains, range-front fault zone (Hoffard, 1990). Hoffard (1991) grouped the West Spring Mountains fault, the East Nopah fault, and the Pahrump-Stewart Valley fault zone into the Pahrump fault system. Estimates of length vary from 30 km to as much as 60 km for the West Spring Mountains fault (Hoffard, 1991; Reheis and Noller, 1991; Reheis, 1992). Prominent fault scarps on middle Pleistocene alluvium are greater than 20 m high and a minimum surface offset of 12 m was measured across a surface inferred to be about 200ka (Hoffard, 1991). Some scarps may be as young as late Pleistocene (10 to 130ka) (Dohrenwend et al., 1991) or possibly even Holocene (Hoffard, 1991). Hoffard (1991) estimated a range of vertical displacement rates from 0.2 to 0.02 mm/yr, with a preferred estimate of 0.06 mm/yr.

Grapevine Fault - The Grapevine fault is a northwest-striking, west-dipping, 20-km-long, normal fault that bounds the southwest side of the Grapevine Mountains, roughly 58 km west of the proposed site (Reheis, 1991; Reynolds, 1969, 1976). There are prominent Quaternary fault scarps along part of the Grapevine fault, most of which face southwest, with movement down to the west, but some antithetic scarps face northeast (Reheis, 1991; Reynolds, 1969). Estimated total vertical displacement is at least 4.3 km (Mabey, 1963; Reynolds, 1969).

Pahrump-Stewart Valley Fault Zone - The Pahrump-Stewart Valley fault zone is a northwest-striking, steeply-west-dipping, right-lateral, oblique-slip fault zone that is about 60 to 70 km southeast of the proposed site (Hoffard, 1991; Malmberg, 1967; Liggett and Childs, 1973). It extends along the east side of Stewart Valley and into central Pahrump Valley. Reheis and Noller (1991) referred to the Pahrump-Stewart Valley fault zone as the Pahrump fault zone and included faults in Ash Meadows and the Amargosa Desert, where the northern Ash Meadows fault zone intersects the Rock Valley fault zone. Estimates of length vary between 50 km (Hoffard, 1991) and 70 km (Reheis and Noller, 1991; Reheis, 1992). Because of apparent differences in behavior, Hoffard (1991) separated the fault into two sections, one north of latitude 36° 05', and one to the south. Along the northern section, Hoffard (1991) found west-facing scarps with a maximum height of 5 m and no evidence of lateral displacement. Based on scarp morphology and weakly developed soils, she inferred that the age of these scarps is late Pleistocene. On the southern section, Hoffard (1991) found a 15 m high escarpment and lineaments to the west that show no evidence for vertical offsets. The youngest deposits displaced by this section are Pliocene or Pleistocene; the southern section does not displace Holocene terraces and is buried by Holocene sand dunes (Hoffard,1991). In contrast, Burchfiel et al. (1983) referred to part of the northern section as the Stewart Valley fault and from structural relations, inferred it was actually older than faults to the south; however, they did not focus on the youngest Quaternary deposits displaced by Pahrump-Stewart Valley fault zone. Malmberg (1967) estimated a total vertical displacement of 488 m and Stewart (1988) estimated greater than 16 to 19 km of total right-lateral displacement of Paleozoic rocks.

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West Pintwater Fault - The West Pintwater fault is a north-striking, west-dipping, normal fault that bounds the west flank of the Pintwater Range, about 76 km east of the proposed site (Dohrenwend et al., 1991; Reheis, 1992), and is roughly 48 to 54 km long. Prominent fault scarps on Pleistocene deposits bound a tectonically active range-front that has steep, faceted-spurs and wineglass valleys (Dohrenwend et al., 1991).

East Pintwater Fault - The East Pintwater fault is a north-striking, east-dipping normal fault zone that bounds the east flank of the Pintwater Range, about 81 km east of the proposed site (Dohrenwend et al., 1991; Reheis, 1992), and is roughly 47 to 54 km long. Although the East Pintwater fault is primarily characterized by a range-front fault placing bedrock against alluvium, at one location scarps offset deposits that may be middle Pleistocene in age (Dohrenwend et al. 1991).

East Nopah Fault - The East Nopah fault is a northwest-striking, near-vertical, right-lateral strike-slip fault that is 17 to 19 km long and lies along the east side of the Nopah Range, about 85 km southeast of the proposed site (Hoffard, 1991; Reheis and Noller, 1991; McKittrick, 1988). Quaternary fault scarps are 3 to 8 m high, face both east and west, and juxtapose older fan remnants against younger fan remnants (Hoffard, 1991). Hoffard estimated a vertical displacement rate of 0.06 to 0.6 mm/yr (which should be corrected to 0.06 to 0.006 mm/yr) from a 3 m high scarp developed on a surface inferred to be 50 to 500ka. This rate is possibly high as it is based on scarp height and not surface offset. The East Nopah fault was previously named the Nopah Range-Front fault zone (Hoffard, 1990) and it was included in the Pahrump fault system, along with the Pahrump-Stewart Valley and West Spring Mountains faults.

Panamint Valley Fault - The Panamint Valley fault is an 80 km long generally north-striking, west-dipping, normal-oblique fault that bounds the east side of the Panamint Range, about 95 km southwest of the proposed site (Zhang et al., 1990; Reheis, 1991; Smith, 1975, 1979). Its southern end is truncated by the Garlock fault and its northern end is intersected by the Hunter Mountain fault. Dips range between 60 and 90° (Murphy, 1932). Sense of displacement appears to vary along different parts of the fault. Normal-slip down to the west is dominant on the central part and faults close to the range-front on the southern part (Zhang et al., 1990; Reheis, 1991). As much as 440 m of uplift occurred during the late Quaternary (Smith, 1979). Faults several hundred meters west of the range front, along the southern part are dominated by right-slip (Smith, 1979; Zhang et al., 1990). From his detailed mapping, Smith (1979) identified Holocene mudflow levees with 20 m of right-lateral displacement; dip-slip displacement was somewhat less. He also found landslide deposits offset by 3 to 4.5 km of right slip. Smith (1979) estimated the average right-lateral displacement for the most recent event to be 2.0 ± 0.6 m and suggested that this event occurred at least several hundred years ago. Scarps with heights ranging from 0.6 to 1.8 m and maximum slope-angles of 27 to 31° are inferred to be single-event scarps, whereas scarps 2 to 6 m high were inferred to be multiple-event scarps. Smith also inferred an average Holocene recurrence of 700 to 2,500 years. From analysis of low-sun-angle aerial photographs and detailed mapping, Zhang et al. (1990) found a shoreline from the highstand of Lake Panamint, age estimated to be 17 ka,

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offset by 27 m of right-slip. They also found ridges offset 24 ± 4 m, 27 ± 4 m, and 37 ± 4 m. Based on scarp morphology, Zhang et al. (1990) believe that the most recent event occurred a few hundred years ago on a 25- to 30-km-long section of the fault south of Ballarat. From measurements at six localities, they determined an average right-lateral displacement of 3.2 ± 0.5 m for this event. Near Goller Wash Canyon, the right-slip was 4 m and the dip slip was 0.4 to 1.2 m for the most recent event. Scarps from older events accumulated 6 to 7 m, possibly during two events, and 11 ± 2 m, possibly during three to four events. From the above, Zhang et al. (1990) estimated right-lateral displacement rates ranging from 1.74 ± 0.65 to 2.36 ± 0.79 mm/yr and the average recurrence to be between 860 and 2,360 years. There are Pliocene and early Quaternary faults in the Panamint Range that parallel the Panamint Valley fault, however, these faults apparently have not moved during the late Quaternary (Reheis, 1991). There are also two clusters of Quaternary faults in Panamint Valley that lie between the Panamint Valley fault zone and the Ash Hill fault to the west (Reheis, 1991).

Hunter Mountain Fault - The Hunter Mountain fault is a northwest-striking, 78-km-long, oblique-slip fault that is about 95 km southwest of the proposed site (Reheis, 1991; Streitz and Stinson, 1974; McAllister, 1956). It extends along the southwest side of Saline Valley, over Grapevine Pass, and along the northeast side of Panamint Valley. Right-lateral displacement and near-vertical dips are indicated for much of the Hunter Mountain fault, but Quaternary deposits are also displaced down to the northeast in Saline Valley and down to the southwest in Panamint Valley (Reheis, 1992; McAllister, 1956), which is why McAllister (1956) suggested that the Hunter Mountain fault may be a scissors fault with the pivot point near Grapevine Pass. Although Hunter Mountain fault displaces alluvium that may be as young as Holocene (McAllister, 1956; Streitz and Stinson, 1974) and sections of the Hunter Mountain fault are some of the most active faults in the area (Reheis, 1991), very little is known about its prehistoric behavior. Zhang et al. (1990) calculated a long-term displacement rate of 2 to 2.7 mm/yr based on 9.3 ± 1.4 km of displacement of a 4 Ma basalt layer.

Ash Hill Fault - The Ash Hill fault is a north-northwest-striking, west-dipping, normal-right-slip fault that is roughly 100 km southwest of the proposed site (Zhang et al., 1990; Reheis, 1991; Smith, 1975). Ash Hill fault lies along the west side of Panamint Valley, bounding the east side of the Argus Range and the west side of Ash Hill. It roughly parallels the Panamint Valley fault zone, but the Ash Hill fault is only 32 to 38 km long (Reheis, 1991). Fault scarps cut young alluvial fans, leaving only modern streams not offset (Smith, 1975). Many scarps face uphill (Reheis, 1991). A shoreline from the highstand of Lake Panamint is displaced 12 m, down to the west, along Ash Hill fault and a late Quaternary fan is offset by 183 m of right slip (Smith, 1979). Hopper (1947) noted many stream channels offset by 24 to 61 m of right-slip. Streitz and Stinson (1974) showed Ash Hill fault as faulting Holocene alluvium against Pleistocene deposits and Pliocene volcanic rocks.

3.1.1.2.3.3 Folding at Yucca Mountain

Subsections 3.1.1.2.3.3 and 3.1.1.2.3.4 were not provided in this Revision of the Annotated Outline; these sections will be provided in future revisions of the Annotated Outline.

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3.1.1.2.3.4 Fractures at Yucca Mountain

3.1.1.2.3.5 Volcanism at Yucca Mountain

Volcanism is a significant aspect of the tectonics of the Yucca Mountain Site. The following discussions of both the history and nature of occurrence of silicic and basaltic volcanism are presented because tectonic processes active within the Quaternary are an important part of the evaluation of the Yucca Mountain Site as a potential high-level waste repository.

The rocks found at Yucca Mountain, to a few kilometers depth, are predominantly Miocene silcic volcanics derived from nearby nested caldera complexes and centers that were active 7-16 Ma. The geochronology of these events has recently been reviewed by Sawyer et al. (1994). In the region of Yucca Mountain, these silicic volcanic events and depositional events are intertwined with relatively minor sedimentary depositional events and minor (volumetrically) basaltic lava flows and cinder-cone development.

The volcanic rocks that formed Yucca Mountain were emplaced during eruptive cycles of the Timber Mountain caldera complex (Christiansen and others, 1977; Byers et al., 1976; Broxton et al., 1989; Byers et al., 1989; Sawyer et al., 1994). The Yucca Mountain Site, including the surface rocks and rocks extending to the depth of the potential repository horizon, comprises volcanic units of the Paintbrush Tuff. The Paintbrush Tuff is a major outflow ignimbrite of the Claim Canyon caldera segment of the Timber Mountain caldera complex (Lipman et al. 1966a; Sawyer et al., 1994).

During the late Neogene (10-2 Ma) and Quaternary (2-0 Ma) small-volume, mostly polygentic, basaltic centers (Figure 3.1.1.2.3.5-1) erupted lava flows, air falls, and cinder cones. The likely nature, rate of occurrence, eruptive style, and emplacement mechanisms of possible future basaltic volcanism are important to the evaluation of the proposed Yucca Mountain Site if it is expected that these natural processes will likely lead to loss of the waste-isolation capabilities of the site. The following discussions are intended to provide an overview of the background information available on both silicic and basaltic volcanism, which then can be used to make assessments of the likelihood of future volcanic events and, thus, their likelihood to directly or indirectly influence the performance of a potential repository at the Yucca Mountain Site.

3.1.1.2.3.5.1 Silicic Volcanism

Introduction

The record of silicic volcanism in the YMR is part of an extensive, time transgressive pulse of mid-Cenozoic volcanism that occurred throughout much of the southwest United States. The Yucca Mountain range is in the south-central part of a major Cenozoic volcanic field that covered an area exceeding 11,000 km^2. The field has been named the SNVF (Christiansen et al., 1977; Byers et al., 1989; Figure 3.1.1.2.3.5.1-1).

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The time-space distribution of volcanic activity in the Basin and Range province has been described by many authors (Armstrong et al., 1969; McKee, 1971; Lipman et al., 1971; Lipman et al., 1972; Christiansen and Lipman, 1972; Snyder et al., 1976; Stewart et al., 1977; Stewart and Carlson, 1978; Christiansen and McKee, 1978; Cross and Pilger, 1978; Smith and Luedke, 1984; Luedke and Smith, 1984; Axen et al., 1993). During the Mesozoic, magmatism in the Cordillera was distributed in linear belts parallel to the continental margin (Armstrong and Ward, 1991). In the southwestern United States, these belts became locally inactive or disrupted about 80 Ma and formed the Laramide magmatic gap (Armstrong, 1974). Renewed silicic magmatism following the Laramide hiatus initiated about 50 Ma in the northeastern part of the Great Basin. Sites of eruptive activity migrated south and southwest, progressively, in time and space across an area of Nevada and adjoining parts of Utah (Stewart et al., 1977; Armstrong and Ward, 1991). The loci of eruptive centers during this silicic volcanism were distributed along arcuate east-west trending volcanic fronts (Stewart et al., 1977). The leading edge of the migrating front during successive increments of time marked the area of most intensive volcanic eruptions. There was a dramatic waning of volcanic activity in the lee or backside of the front (Stewart et al., 1977), and virtually no volcanic activity ahead of the front. The period of most voluminous silicic volcanic activity in the YMR occurred between 15 to 11 Ma. The YMR marks the southern limit in the spread of time-transgressive volcanic activity.

Two significant changes in the regional volcanic and tectonic patterns occurred about 13 to 10 Ma, at the approximate latitude of the YMR. First, the progressive southern migration of volcanism halted. Silicic-eruptive activity continued in diminished volumes and migrated predominantly to the southwest and southeast, following less systematic patterns than the migration of preceding volcanic activity. Post-Miocene volcanic activity approached present positions along the east and west margins of the Great Basin (Christiansen and McKee, 1978; Smith and Luedke, 1984). These changed patterns in migration of volcanism left a conspicuous amagmatic gap extending from the southern edge of the NTS, south to the latitude of Las Vegas (Figure 3.1.1.2.3.5.1-2). It coincides with a major increase in the regional gravity field that forms the southern boundary of the gravity low of the Great Basin (Eaton, 1978). This latitude (35° N) also coincides with the approximate position at 10-20 Ma of the boundary between the incoherent, subducted slab or "slab gap" south of the Medocino fracture zone and persisting subducted slab to the north (Severinghaus and Atwater, 1990).

The second major change at about 10 Ma in the YMR was a transition in the composition of volcanic activity. This change is consistent with a time-transgressive switch across the southwestern United States, from predominantly silicic volcanism to bimodal basalt-rhyolite volcanism (fundamentally basaltic volcanic activity; Christiansen and Lipman, 1972). The exact timing of this transition in the YMR cannot be precisely defined. It may be marked by a transition from near homogeneous quartz latite and rhyolite ignimbrite (> 13 Ma) to eruptions of compositionally zoned ash-flows units (< 12 Ma) that range from high-silica rhyolite at their base (SiO_2 > 75%) to quartz-latitic caprocks (SiO_2 < 70 %). This transition in the NTS region is probably marked by eruption of the Paintbrush Tuff (Christiansen and Lipman, 1972; Christiansen et al., 1977). Alternatively, the transition may be recorded by the

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widespread appearance of moderately voluminous basaltic volcanism (Crowe, 1990). The age of this basaltic-volcanic activity is bracketed between eruption of the Timber Mountain Tuff (11.2 Ma) and the peralkaline ash-flow sheets of the Black Mountain caldera complex (8.5 Ma; Crowe, 1990). Yucca Mountain consists of a thick sequence of ignimbrites derived from multiple caldera-forming, eruptive cycles of the Claim Canyon and Timber Mountain-Oasis Valley caldera complexes. The existence of caldera complexes of Crater Flat, which border Yucca Mountain to the west, is considered as controversial by some workers. Numbered contour lines locate the position of arcuate, migrating fronts of silicic volcanism during increments of Cenozoic time. Yucca Mountain is located at the southern limit of the southern migration, mid-Cenozoic, silicic volcanism. The Yucca Mountain area borders the northern edge of a zone in southern Nevada distinguished by the absence of Mesozoic and Cenozoic volcanism and plutonism. This zone, referred to as the amagmatic gap, includes much of the central Basin and Range subprovince of Wernicke (1992) and Jones et al. (1992). [Section needs to be completed] [INN 3.1.1.2.3.5.1-1]

3.1.1.2.3.5.2 Basaltic Volcanism

Field and geochronology data for the basaltic-volcanic rocks of the YMR define two episodes (Crowe, 1990). The first episode of bimodal basalt-rhyolite volcanism is called BSE. This episode postdates, but is close in age, to most of the silicic eruptions of the TM-OV complex. For this report, we describe only the basaltic units present in and around the Timber Mountain highland and the YMR. Other basaltic units are present, but are not described. These basalts are associated primarily with the Black Mountain-Stonewall Mountain calderas.

The second basaltic episode includes spatially scattered, small-volume centers marked by scoria cones and lava flows of alkali basalt. These rocks range in age from 8 Ma to Quaternary. They are divided into two cycles (Crowe, 1990), the Older and Younger Post-Caldera Basalts (OPB and the YPB). The field relations and geochronology of the basalt cycles of the YMR are described in following sections.

The level of detail of studies of basaltic units of the YMR varies with age of the volcanic units. Pre-Pliocene basalt units of the region (> 6.5 Ma) were studied primarily in reconnaissance. The outcrop relations of the units, as depicted on published quadrangle maps of the USGS, were checked in the field. Samples were collected of the basalt units. They were examined petrographically, and analyzed geochemically (Crowe et al., 1986). The stratigraphic relations of the pre-6.5 Ma basaltic-volcanic units were evaluated in the field and geochronology data (whole rock K-Ar age determinations) were obtained, where required to discriminate the ages of the rocks.

By contrast, the basaltic units of the YPB have been the focus of much more detailed studies. All units of this cycle have been mapped or remapped at scales of 1:12,000 to 1:4,000. The eruptive sequences of individual centers were assessed largely through the detailed mapping and, additionally, on the basis of new geochronology data (Crowe, 1990; Crowe et al., 1992). The rocks have been analyzed petrographically and geochemically. Geochronology data,

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using conventional K-Ar methods, has been supplemented through application of age-calibrated geochronology methods in order to cross-check ages of the youngest Quaternary volcanic centers. This increase in the level of detail of geologic and geochronologic data with decreasing age of basalt centers is the result of a purposeful attempt to focus work on assessment of the Pliocene and Quaternary volcanic history of the YMR. It is this part of the geologic record that provides the most important basis for forecasting the risk of future volcanic activity.

3.1.1.2.3.5.2.1 Basalt of the Silicic Episode

The BSE crops-out throughout the YMR. These basaltic rocks are identified by a suite of characteristics; however, none of the features are individually unique. The first is the close association, in space and time, with the eruptive activity of the calderas of the TM-OV complex. The BSE postdates closely the eruption of the Timber Mountain Tuff (11.2 Ma). Many of the basalt units of this episode stratigraphically underlie the Thirsty Canyon Tuff (8.5 Ma; Noble et al., 1992). Second, the largest volume centers of the BSE are located in the ring-fracture zone of the Timber Mountain caldera complex and record a waning phase of the caldera-related volcanic activity (Figure 3.1.1.2.3.5.2.1-1) [INN 3.1.1.2.3.5.2.1-1]. Third, all centers of the BSE were large-volume eruptive units (> 3 km^3, dense-rock equivalent). Their present surface outcrops form major topographic features (eroded shield volcanoes or lava mesas, where basalt lavas cap modern topographic ridges). Third, the BSE exhibits a wide range in geochemical composition (basalt to basaltic-andesite or latite) and a much larger range of compositional variation than the YPB (Crowe et al., 1986; Crowe, 1990).

The BSE is divided into three major groups, based on geographic distribution. These are: (1) basalts exposed in the moat zone of the Timber Mountain caldera, (2) basalt near and flanking the Black Mountain caldera, and (3) basalts of the Yucca Mountain area (Crowe et al., 1995).

a. Mafic Lavas of Dome Mountain. The most important occurrence of basaltic rocks by volume is in the southeast edge of the Timber Mountain caldera (Figure 3.1.1.2.3.5.2.1-1). Here, mafic lavas are interbedded with two major successions of rhyolite lava. They overlie the rhyolite of Fortymile Wash and underlie the rhyolite of Shoshone Mountain (10.3 ± 0.3 Ma; Noble et al., 1992; Minor et al., 1993). The basalt and rhyolite define a bimodal association marking the waning postresurgence stage of the Timber Mountain caldera (Christiansen et al., 1977). The largest volume of basaltic magma is the mafic lavas of Dome Mountain. These lavas have been described by Luft (1964), and Marsh and Resmini (1992). The likely structural control for eruption of the mafic lavas of Dome Mountain is the ring-fracture system of the Timber Mountain caldera. Further information on these basalts can be found in Crowe et al. (1994).

b. Basaltic Rocks of the Black Mountain Caldera. Basaltic-volcanic rocks crop out on all but the eastern flanks of the Black Mountain caldera (Figure 3.1.1.2.3.5.2.1-1). These basalt units overlap in age with the peralkaline volcanic units of the caldera. They

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underlie, are interbedded with, and overlie, ash-flow units of the Thirsty Canyon Tuff (Minor et al., 1993). Many of the outcrops of basalt are located in the Nellis Bombing and Gunnery Range. Because of restricted access, the basalts have been sampled only at scattered localities around Black Mountain (Crowe et al., 1986). A sequence of eroded basalt vents and lava flows are exposed south of Black Mountain (Minor et al., 1993). These lavas underlie the Thirsty Canyon Tuff. Basalt lava exposed north of Black Mountain (Minor et al., 1993) underlies and locally overlies the Thirsty Canyon Tuff.

c. Basaltic Volcanic Rocks, Yucca Mountain Area. Basaltic-volcanic rocks of the BSE comprise three geologic units in the eastern Yucca Mountain area, and two additional units at and southwest of Yucca Mountain. Two of the eastern units were shown on published geologic maps (Ekren and Sargent, 1965; Sargent and Stewart, 1971; Sargent and others, 1970). These are the basaltic andesite of Skull Mountain and the basalt of Kiwi Mesa (Figure 3.1.1.2.3.5.2.1-1). A third unit, the basalt of Jackass Flats, has been separated in more recent geologic reports (Crowe et al., 1986; Frizzel and Shulters, 1990).

The south and southeast perimeters of Jackass Flat are bounded by Skull and Little Skull Mountains. These small ranges are capped by a sequence of mesa-forming lava flows and flow-breccia that locally exceed 60 m in thickness. This unit is the basaltic andesite of Skull Mountain. The lavas are distinguished in the field by the presence of abundant phenocrysts of bipyrimidal quartz (Crowe et al., 1986). Moreover, they are the only known lavas in the region that are subalkaline in chemical composition (Crowe et al., 1983a, 1986). This quartz-bearing basaltic andesite of Skull Mountain overlies the lavas of the Wahmonnie-Salyer center and the Paintbrush and Timber Mountain tuffs. It has been dated by the whole-rock, K-Ar method at 10.2 ± 0.5 Ma.

Several small outcrops of dissected lava are flanked by alluvium in the central part of Jackass Flats (Figure 3.1.1.2.3.5.2.1-2; Frizzel and Shulters, 1990). These lavas have been dated by the K-Ar method at 11.0 ± .4 and 11.2 ± .4 Ma. Correlative lavas (petrographically and geochemically) are interbedded with the upper part of the sequence of basaltic andesite of Skull Mountain at Little Skull Mountain, but not at Skull Mountain (Crowe et al., 1986). Samples of the lavas at Little Skull Mountain have been dated at 8.4 ± .4 Ma, which is inconsistent with the age of the inferred correlative lavas of Jackass Flats.

The Solitario Canyon fault forms the western edge of the potential Yucca Mountain Site (Scott and Bonk, 1984). The north trace of the fault, near its intersection with Drill Hole Wash, is intruded by a basaltic dike. The dike is exposed locally along the northwest edge of the block (Figure 3.1.1.2.3.5.2.1-2) and in trench 10. Exposures of the dike in trench 10 show that the dike both intrudes the fault plane and has been offset by subsequent episodes of movement on the Solitario Canyon fault. North of trench 10, the strike of the dike swings to the northwest. It crops-out discontinuously, forming en echelon dikes in the Tiva Canyon member of the Paintbrush Tuff. This change in strike coincides with an abrupt change in the trend of the topography of ridges and valleys, from north-south to northwest-trending, and

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occurs at the north end of the exploratory block of Yucca Mountain. The dike has been dated at about 10.0 ± .4 Ma (Carr and Parrish, 1985).

The older basalt of Crater Flat (Crowe et al., 1986) crops-out along the south slope of an arcuate ridge, forming the southern boundary of Crater Flat (Figure 3.1.1.2.3.5.2.1-2). The outcrops consist of the remnants of oxidized vent-scoria, intruded by dikes and small plugs. These units are probably eroded roots of at least one, and more likely several scoria cones. Directly east of the scoria deposits, there is a sequence of thin lava flows and flow breccia that is petrographically similar to and probably derived from the plugs or cones. These lavas are overlain by slide blocks of Paleozoic carbonate. They have been dated by the whole rock, K- Ar method at 10.5 ± .1 Ma. The lavas dip to the northeast and are inferred to underlie the south and southwest parts of Crater Flat. The combined field, geochronology, geochemical and aeromagnetic data suggest that the west and southwest parts of Crater Flat are floored by an extensive basalt unit of 10.5 ± 0.1 Ma. Further information can be found in Crowe et al. (1994).

3.1.1.2.3.5.2.2 Postcaldera Basalt of the YMR

The second recognized episode of basaltic volcanism in the YMR is the postcaldera basalt. These basalt centers occur at sites either well removed from eruptive centers of the TM-OV complex or are younger than, and cannot be related in time to, the silicic-magmatic activity. The basalts consist of small volume (<1 km^3) centers marked by clusters of scoria cones and lava flows (except for the basalt of Thirsty Mesa, which has a volume of > 3 km^3). They are divided into two groups, the OPB and the YPB. This division is based on differences in ages and geographic distribution (Figure 3.1.1.2.3.5.2.2-1).

a. OPB. The OPB occur at five localities, all north and east of Yucca Mountain (Figure 3.1.1.2.3.5.2.2-1). The OPB was erupted along either north-northwest trending Basin and Range faults or at the intersection of Basin amd Range faults with the ring-fracture zone of older calderas. Field evidence shows that some of the basalt units were erupted contemporaneously with movement on the extensional faults (Crowe et al., 1983b, 1986). The OPB ranges in age from 9 to 6.3 Ma (Crowe et. al., 1995).

(1) Basalt of Rocket Wash. A platy series of thin, basalt lava flows, informally named the basalt of Rocket Wash, overlies the Thirsty Canyon Tuff in the northwest edge of the Timber Mountain caldera (Figure 3.1.1.2.3.5.2.2-1). This basalt, which was mapped by Lipman et al. (1966b), O'Connor et al. (1966), and Minor et al. (1993), erupted along a north-trending normal fault that marks the approximate edge of the ring-fracture zone of the Timber Mountain caldera. The vent for the basalt lavas was located at the northern edge of the mapped outcrops. The basalt yielded a K-Ar age of 8.0 ± 0.2 Ma. Further details are provided in Crowe et al. (1994).

(2) Basalt of Pahute Mesa. Three spatially separate, but related, basalt units of the OPB crop-out in the northern part of the ring-fracture zone of the Silent Canyon

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caldera. These basalts, which are informally named the basalt of Pahute Mesa, are localized at the intersection of north-northeast trending Basin and Range faults with the ring-fracture zone of the Silent Canyon caldera (Figure 3.1.1.2.3.5.2.2-1). They are divided into three units, each separate geographically. These are a western basalt, a central basalt, and an eastern basalt.

A sample collected from a western basalt feeder-dike was dated at 9.1 ± 0.7 Ma. The central basalt yielded a K-Ar age of 8.8 ± 0.1 and 10.4 ± 0.4 Ma. The older age is inconsistent with the stratigraphic relations of the central basalt center (overlying the Thirsty Canyon Tuff), and is probably anomalously old. The eastern basalt is identical, petrographically, to the western basalt and has not been dated. Further details are provided in Crowe et al. (1994).

(3) Basalt of Paiute Ridge. The Half Pint Range, which forms the eastern margin of Yucca Flat, exposes a complex sequence of sills, dikes, and lopolithic intrusions of basaltic composition (Byers and Barnes, 1967). Dissected scoria cones and lava flows are associated with the intrusive bodies, indicating surface eruptions accompanied the formation of the intrusions (Crowe et al., 1983b; Valentine et al., 1992). The sills and lopolithic bodies occur in the interior of a northwest-trending graben. Subsurface magma intruded generally upward and locally eastward from dikes that are coplanar with northwest-trending Basin-Range faults. Eastward-intruded magma formed sills by lifting and locally intruding the overlying sequence of tuff. At several localities, primarily in the axis of the graben, the sills sagged to form lopolithic intrusions (Byers and Barnes, 1967; Crowe et al., 1983b; Valentine et al., 1992). Some of the sills differentiated in situ to form pods of syenite (Byers and Barnes, 1967).

The surface lava from the basalt of Paiute Ridge has been dated by the whole rock K-Ar method at 8.5 ± 0.3 Ma (Crowe et al., 1983b). The geologic relations of the surface-basalt outcrops and dike and sill complexes record a complex history of Basin-Range faulting in the Halfpint Range. Dikes of basalt intrude and are coplanar with several northwest-trending faults. The faults and main graben system must, therefore, predate the basalt (>8.5 Ma). Sills of basalt are offset, and downfaulted outcrops of cone-scoria are present in the north-center of the graben. Both relationships require continuing offset on the faults, after eruption and intrusion of the Miocene basalt. A possible correlative basalt with the basalt of Paiute Ridge has been reported in the subsurface of southern Yucca Flat (Drill hole UE1-h; 784 feet below the surface) and has been dated at 8 ± 0.3 Ma.

Recent paleomagnetic investigations by Ratcliff et al. (1993) have demonstrated that emplacement of the basalt of Paiute Ridge occurred during a geomagnetic polarity transition, probably from reverse to normal polarity. All the sills, dikes, lopolithic intrusions, and lava flows give transitional paleomagnetic data and,

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together, define a relatively smooth "path" for the field reversal. These data are of considerable interest for the emplacement dynamics of mafic magmas in such a setting. Given our current understanding of the duration of geomagnetic polarity reversals, it is conceivable that the entire complex of mafic intrusions at Paiute Ridge was emplaced in less than a few thousand years.

A separate, but petrologically related basalt, the basalt of Scarp Canyon, crops out southeast of the basalt of Paiute Ridge and west of Nye Canyon. The basalt of Scarp Canyon yielded a whole-rock, K-Ar date of 8.7 ± 0.3 Ma. Further details are provided in Crowe et al. (1994).

(4) Basalt of Nye Canyon. The last major unit of the OPB is the basalt of Nye Canyon (Figure 3.1.1.2.3.5.2.2-1). This unit consists of at least four centers; one is buried beneath alluvial deposits of northeast Frenchman Flat (Carr, 1974). There are three surface centers of basalt that define a northeast-trending alignment of basalt centers. The basalt of Nye Canyon is the oldest basalt unit in the YMR that shows a northeast structural trend. That is the predominant direction of dike orientation and alignment of clusters of centers in post-Pliocene basalt in the region (Crowe et al., 1983; Crowe and Perry, 1989; Crowe, 1990).

The northeastern center has been dated at 6.3 ± 0.2 Ma. The middle and southern basalt centers are eroded tuff rings (Crowe et al., 1986) formed by mixed Strombolian and hydrovolcanic activity. The interiors of both tuff rings are partly to completely filled by scoria deposits and lava, and record episodes of mixed Strombolian and Hawaiian eruptive activity that followed the hydrovolcanic eruptions. The middle basalt center is geochemically the most primitive basalt in the region (Crowe et al., 1986; Farmer et al., 1989). A plug from the middle center has been dated at 6.8 ± 0.2 Ma. The south center has been dated at 7.2 ± 0.2 Ma.

A fourth basalt center was intersected in a drill hole in alluvium beneath Frenchman Flat (Carr, 1984). It has been dated at 8.6 Ma. This basalt may be correlative with the basalt of Nye Canyon. Alternatively, its K-Ar age suggests it may be correlative with the basalt of Scarp Canyon (Figure 3.1.1.2.3.5.2.2-1).

b. YPB. A brief, but regionally significant, hiatus in volcanic activity followed eruption of the basalt of Nye Canyon. No volcanic rocks of the YMR, including both surface and subsurface rocks, have yielded age determinations in the range of 6.3 (basalt of Nye Canyon) and 4.8 Ma (basalt of Thirsty Mesa). The second cycle of the post caldera basalt, the YPB, includes all volcanic rocks younger than the basalt of Nye Canyon. In order of decreasing age, these basalts include the basalt of Thirsty Mesa, the basalt of Amargosa Valley, the basalt of southeast Crater Flat, the basaltic andesite of Buckboard Mesa, the Quaternary basalt of Crater Flat, the basalt of Sleeping Butte and the Lathrop Wells basalt center (Crowe, 1990). These basalt centers, except the basaltic andesite of Buckboard Mesa, all occur in a narrow northwest-trending zone,

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located west and south of the Yucca Mountain Site (Figure 3.1.1.2.3.5.2.2-1). This zone has been named the CFVZ (CFVZ; Crowe and Perry, 1989). The basalt centers of the YPB, with two exceptions, consist of clusters of probably coeval, small-volume centers aligned along predominantly northeast (one is aligned to the north) structural trends. The exceptions are the basaltic andesite of Buckboard Mesa and the Lathrop Wells center. The basaltic andesite of Buckboard Mesa consists of a main scoria cone located on a northwest-trending fissure. The Lathrop Wells cone consists of only one center.

(1) Basalt of Thirsty Mesa. The basalt of Thirsty Mesa comprises a thick accumulation of fluidal lava and local feeder vents marked by dissected scoria cones surmounting the lava mesa (Figure 3.1.1.2.3.5.2.2-1). These lavas, viewed from the west, appear to form a shield volcano, but the volcanic landform is actually more complex than a simple shield. The lavas were erupted onto a pre-existing plateau formed by ignimbrite of the Thirsty Canyon Tuff. The flows form a convex upward mesa surface above the plateau.

The scoria vents have been dissected significantly, but still maintain a rounded, mound-form and uphold the high-standing topography of the mesa. Minor et al. (1993) reported an $^{40}Ar/^{39}Ar$ age of 4.6 Ma for the basalt of Thirsty Mesa. Crowe et al. (1983a) have obtained $^{40}Ar/^{39}Ar$ ages for a sample collected from the west base of the lavas of Thirsty Mesa and a dike sample collected from one of the summit scoria cones. The ages are, respectively, 4.68 $\pm$ 0.3 and 4.88 $\pm$ 0.4 Ma (Table 3.1.1.2.3.5.2-1). These ages are in good agreement with the new $^{40}Ar/^{39}Ar$ ages, the age reported by Minor et al. (1993), and geomorphic data (Crowe et al., 1992). The basalt of Thirsty Mesa has a reversed magnetic polarity. Adequate geochronology data have been obtained to document the age of the basalt of Thirsty Mesa. Further details are provided in Crowe et al. (1994).

(2) Basalt of Amargosa Valley. A negative aeromagnetic anomaly was identified in the Amargosa Valley, several kilometers south of the town of Amargosa Valley (Kane and Bracken, 1983; Crowe et al., 1986; Figure 3.1.1.2.3.5.2.2-1). The anomaly was drilled in 1991 by a private company (Harris et al., 1992). Samples of cuttings of the basalt were obtained and submitted for K-Ar age determinations. A $^{40}Ar/^{39}Ar$ age of 3.85 $\pm$ 0.05 Ma was obtained for the cuttings under YMP controls (Table 3.1.1.2.3.5.2-1). [INN 3.1.1.2.3.5.2-1] Turrin (1992) reported an $^{40}Ar/^{39}Ar$ age of a basalt sample from the exploratory drill hole of 4.4 $\pm$ 0.07 Ma. This is significantly older than the age of 3.85 Ma, and additional age determinations will be obtained to investigate the discrepancy in ages. Further details can be found in Crowe et al. (1994).

(3) Basalt of Southeast Crater Flat. The basalt of southeast Crater Flat has been described by Crowe and Carr (1980), Vaniman and Crowe (1981), Vaniman et al. (1982), Crowe (1990), and Perry and Crowe (1992). The basalt unit consists

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of an alignment of north-trending, dissected scoria cones and associated moderate volume, lava flows (Figure 3.1.1.2.3.5.2.2-2). The scoria cones extend south in an en echelon pattern from the east central part of Crater Flat. The former presence of individual centers is indicated primarily by the presence of eroded deposits of oxidized cone scoria. The scoria contains aerodynamically shaped bombs, and spatter. The large size of the bombs (> 0.5 m) and local agglutination of the spatter indicates near-vent deposition. However scoria outcrops are deeply eroded, discontinuous, and retain no primary constructional topography.

The geometry of vent zones can be reconstructed by detailed mapping of three surface features. Lavas of the basalt of southeast Crater Flat are the most aerally extensive of the YPB. They form sheet-like flows that flank, to the south and east, the described vent zones. In no case, however, can the lavas be traced directly to their individual feeder sources. Vents distinguished in the field can be identified in the aeromagnetic data by north-south elongation and local near-circular patterns of magnetic contours. These patterns are not observed in the eastern outcrop areas of the basalt of southeast Crater Flat where they are concealed beneath alluvium or are faulted against Miocene tuff. This suggests that there are probably no major eruptive centers concealed beneath the alluvial deposits.

The basalt of southeast Crater Flat was dated initially at 3.85 $\pm$ 0.13 and 3.64 $\pm$ 0.13 Ma. Sinnock and Easterling (1983) obtained ages of 4.27 $\pm$ 0.46, 3.73 $\pm$ 0.06, and 3.89 $\pm$ 0.17 Ma for six, replicate, whole-rock K-Ar age determinations obtained from splits of the same sample collected at one site in the basalt of southeast Crater Flat. The samples were analyzed at three separate chronology laboratories (values reported as unweighted averages with 1 sigma standard deviation of the replicate measurements). A second sample site was collected from outcrops of the basalt of southeast Crater Flat and analyzed at the same three laboratories. This site yielded average ages of 4.22 $\pm$ 0.08, 3.69 $\pm$ 0.09, and 4.00 $\pm$ 0.13 Ma, respectively (Sinnock and Easterling, 1983). Unpublished K-Ar ages of about 3.91 $\pm$ 0.20 and 3.75 $\pm$ 0.12 Ma have been obtained by the USGS for a dike exposed in cone scoria and a lava, respectively, for the basalt of Crater Flat. New $^{40}Ar/^{39}Ar$ ages of 3.65 $\pm$ 0.06, 3.69 $\pm$ 0.06, and 3.75 $\pm$ 0.04 (2 sigma) Ma have been obtained for the basalt of southeast Crater Flat (Table 3.1.1.2.3.5.2-1). The consistency of age determinations for this unit at multiple analytical laboratories suggests the chronology of this unit is well established by the existing radiometric ages. All measured samples of the basalt of southeast Crater Flat have a reversed magnetic polarity (Vaniman and Crowe, 1981; Vaniman et al., 1982). The similarity in age of the basalt of Amargosa Valley, and the basalt of Crater Flat suggests the units were probably erupted about the same time, but as spatially separate units.

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(4) Basaltic Andesite of Buckboard Mesa. The basaltic andesite of Buckboard Mesa is the second largest basalt by volume of the postcaldera basalt cycles of the region (1 km^3, dense-rock equivalent). It comprises aphyric to sparsely porphyritic lavas erupted in the northeast part of the ring-fracture zone of the Timber Mountain caldera (Figure 3.1.1.2.3.5.2.2-1). The lavas were erupted mainly from a scoria cone in the northwest part of the outcrop area (Scrugham Peak). Additional lava flows vented from a fissure that extends southwest for about 5 km from the base of Scrugham Peak. This fissure is marked both by a subdued linear ridge in the present-day topography and by the presence of scoria and agglutinated spatter exposed in craters excavated during high-explosive experiments (Lutton, 1968). Further details can be found in Crowe et al. (1994)

Whole-rock K-Ar ages of 2.82 $\pm$ 0.04 and 2.79 $\pm$ 0.10 Ma were obtained for core of the basaltic andesite of Buckboard Mesa recovered from drill holes WDH-11 and WDH-12. Unpublished K-Ar ages of 2.93 $\pm$ 0.03 Ma for the kaersutite-bearing flow, and 3.07 $\pm$ 0.29 for the main flow unit have been obtained by the USGS. These ages are consistent with the positive magnetic polarity of the lavas. Additional samples of the basalt of Buckboard Mesa have been submitted for $^{40}Ar/^{39}Ar$ age determinations (Table 3.1.1.2.3.5.2.2-1). [INN 3.1.1.2.3.5.2.2-2]

(5) Quaternary Basalt of Crater Flat. A series of four Quaternary basalt centers form a north-east trending, slightly arcuate cluster of basalt centers extending along the axis of Crater Flat. From southwest to northeast, these centers consist of, respectively: Little Cones, Red Cone, Black Cone, and the Makani cone (Crowe et al., 1995).

(a) The Little Cones. The Little Cones consist of two separate cones of small dimensions. The southwest cone is breached on the south side probably from extrusion of a lava flow that is buried beneath alluvial deposits. The dimensions of this inferred flow can be approximated from the lobate boundaries of a mass of a negatively magnetized body, shown on the drape aeromagnetic data of Kane and Bracken (1983).

The northeast cone is a symmetrical scoria cone and is slightly smaller than the southwest cone. There is no evidence of extrusion of lava from this cone, based either on examination of surface exposures or the aeromagnetic data of Kane and Bracken (1983).

A K-Ar age of about 1.1 $\pm$ 0.3 Ma was obtained for the feeder dike south of the Little Cones (Crowe et al., 1982; Vaniman et al., 1982). An unpublished K-Ar age of a sample collected at the same site was dated by the USGS at 0.76 $\pm$ 0.20 Ma. A recently obtained $^{40}Ar/^{39}Ar$ age of the Little Cones feeder dike is 1.02 $\pm$ 0.10 Ma (Table 3.1.1.2.3.5.2-1). Both of the Little Cone centers have reversed magnetic polarity, consistent with the

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negative aeromagnetic anomaly associated with the centers (Kane and Bracken, 1983). The reversed magnetic polarity and the current K-Ar ages indicate the Little Cones center was probably formed in the Matayama reversed epoch. Further details can be found in Crowe et al. (1994).

(b) Red Cone and Black Cone Centers. The Red and Black cone centers are analogous volcanic landforms with similar eruptive histories. Each consists of a main scoria cone surmounted by a summit crater filled with agglutinated spatter, large lava blocks, and scoria (Vaniman and Crowe, 1981; Smith et al., 1990; Feuerbach et al., 1990; Ho et al., 1991). The main cones are flanked by scattered scoria mounds, which are erosional remnants of satellite vents. Some or perhaps all the lavas of both centers were erupted from the satellite vents (Vaniman and Crowe, 1981; Vaniman et al., 1982; Smith et al., 1990; Ho et al., 1991).

A series of satellite cones or scoria mounds is exposed south of the main cone (Vaniman and Crowe, 1981; Ho et al., 1991). These are clustered along northeast and northwest trends. Locally, the mounds were the source vents of aa lava flows that flank the southern part of the main cone and surround the scoria mounds (Smith et al., 1990). These lavas extend for as much as 1 km from the vents. The outcrop edges of the lavas have abrupt flow-fronts with steep flow foliations indicating the present edges are close to the original flow-margins. However, there has been sufficient removal of flow-top clinker and flow margin breccia to expose the massive interior lobes of some of the aa flows.

There is limited evidence of preservation of original primary flow topography of the lava flows of both the Red Cone and Black Cone centers. The lava surfaces have been smoothed from a combination of erosion, deposition of loess and pedogenic processes.

An initial K-Ar age determination of 1.5 $\pm$ 0.1 Ma was obtained for a lava flow directly east of the main cone of the Red Cone center (Crowe et al., 1982; Vaniman et al., 1982). Sinnock and Easterling (1983) obtained ages of, respectively, 1.53 $\pm$ 0.31, 1.12 $\pm$ 0.27, and 1.55 $\pm$ 0.15 Ma for splits of a sample collected from the same lava flow. Smith et al. (1990) reported ages of 0.98 $\pm$ 0.10, 1.01 $\pm$ 0.06, and 0.95 $\pm$ 0.08 Ma for samples from the Red Cone center. Unpublished ages by the USGS for two samples of the same lava site, located east of the main cone, are 0.84 $\pm$ 0.15 and 1.07 $\pm$ 0.34 Ma. There is a slightly larger spread in K-Ar ages of the Red Cone center than other centers. Additional samples have been submitted for $^{40}Ar/^{39}Ar$ age determinations from the Red Cone center to provide final documentation of the chronology of this center.

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Potassium-Ar ages of 1.09 ± 0.3 and 1.07 ± 0.4 Ma were obtained for the Black Cone center (Crowe et al., 1982; Vaniman et al., 1982). Unpublished K-Ar ages by the USGS are 0.80 ± 0.06 for a lava at the south edge of the Black Cone center, and 0.83 ± 0.09 Ma for a spatter sample from the crater-fill at the summit of the cone. New $^{40}Ar/^{39}Ar$ ages of the Black Cone center are 1.05 ± 0.14, 0.96 ± 0.15, 0.94 ± 0.05 (2 sigma) Ma (Table 3.1.1.2.3.5.2-1).

The K-Ar ages for both the Red and Black Cone centers are consistent with the degree of horizon development in the soils and the observed, reversed, magnetic polarity of the centers. The combined chronology data and the reversed magnetic polarity of both centers suggest they belong to the Matuyama reversed epoch. Further details are discussed in Crowe et al. (1994).

(c) Makani Cone. The Makani Cone is the northernmost basalt center of the Quaternary basalt of Crater Flat. It is the most deeply incised of the four Quaternary basalt centers. The Makani Cone consists of an eroded aa lava flow and small remnant of cone-scoria; the latter is probably part of an original scoria cone (Vaniman and Crowe, 1981).

Potassium-Ar ages of 1.14 ± 0.3 and 1.07 ± 0.04 were obtained for the Makani center (Crowe et al., 1982; Vaniman et al., 1982). Unpublished ages of 1.66 ± 0.5 and 1.04 ± 0.03 Ma have been obtained for the Makani center. The older age was obtained for a sample of a basalt dike that is moderately altered. The age may be anomalously old. Additional $^{40}Ar/^{39}Ar$ age determinations will be obtained for the center to attempt to resolve the chronology data.

(d) Eruptive Models for the Quaternary Basalt of Crater Flat. One area of continuing controversy is in reconstructing the eruptive history of individual centers of the Quaternary basalt of Crater Flat. There are two problems. First, it has been difficult to evaluate whether each center formed in a single eruption (monogenetic) or multiple, time-separate eruptions (polygenetic), analogous to the Lathrop Wells center (Crowe et al., 1988; Wells et al., 1990; Crowe and Perry, 1991; Crowe et al., 1990; Perry and Crowe, 1992). Small-volume basalt centers have traditionally been assumed to be monogenetic centers (Wood, 1980). Second, it is difficult to establish the age of eruptive activity at each center and to use these data to compare the ages of eruptive activity at the other centers. A total of 21 K-Ar and $^{40}Ar/^{39}Ar$ age determinations have been obtained for the Quaternary basalt of Crater Flat (Table 3.1.1.2.3.5.2.2-1). [INN 3.1.1.2.3.5.2.2-2] The mean age of the centers is 1.11 ± 0.25 Ma. Exploratory statistical analyses of the K-Ar ages, using the stem and leaf, box, and probability plots, show that the data distribution is non-normal, and four samples are identified as outliers.

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If these samples are removed, the mean age of the centers is 1.00 $\pm$ 0.11 Ma, and the data show a near-normal distribution with no outliers. The statistically refined data-set can be interpreted to indicate the centers are coeval. However, based on analog studies of the Lathrop Wells center (Crowe et al., 1992), the time between polygenetic events may be less than the standard deviation of the mean ages. Thus, the precision of the chronology methods may be insufficient to test the polygenetic model. Judgment on the ages and eruptive models of the basalt centers require additional $^{40}Ar/^{39}Ar$ age determinations. [INN 3.1.1.2.3.5.2.2-2]

Champion (1991) suggested all Quaternary basalt centers of Crater Flat record a single, reversed, polarity remanent magnetization on the basis of field and paleomagnetic analyses of samples collected from 20 sites. These paleomagnetic data were interpreted by (Champion 1991) to permit the inference that the Quaternary basalt centers of Crater Flat formed contemporaneously (single magma-pulse) with each center being of monogenetic origin (formed in one brief eruptive cycle). However, this interpretation requires several critical underlying assumptions. First, Champion (1991) argued that modern (Holocene) secular variation in the intensity of the earth's magnetic hold occurs at a rate of about 4 to 5 times field per century. He did not assess the temporal variability of the geomagnetic field during the period of eruption of the Quaternary basalt of Crater Flat. Second, the paleomagnetic data for the Quaternary centers of Crater Flat were assumed to adequately represent all volcanic events at these centers. The sample sites for collection of paleomagnetic material were not identified (Champion, 1991), so this assumption cannot be tested. Third, and related to the second argument, is the reliability of the overall paleomagnetic data. Champion (1991) presented only site mean directions of remanent magnetization for the Quaternary basalt of Crater Flat. Both lava and scoria were sampled for the paleomagnetic studies (Champion, 1991). The latter material may be a less reliable recorder of the ancient geomagnetic field. Without presentation of data, and in particular the demagnetization results, it is impossible to assess the variability of the field magnetization directions. The dispersion of sample directions at a site level is particularly important, if the secular variation is less than the assumed 4 to 5 times per century. Fourth, the four basalt centers extend in a northeast-trending arc for a distance of about 12 km. This is longer than expected for basalt feeder dikes (Crowe, et al., 1995). More information [INN 3.1.1.2.3.5.2.2-3] is required to determine if the centers were fed from multiple dike systems and if the geochemistry of the centers is consistent with single or multiple pulses of magma. Finally, the paleomagnetic data must be considered in light of the evidence of geochemical diversity in the lavas (Perry and Crowe, 1992). The assumptions associated with the paleomagnetic study must be more carefully evaluated before accepting the

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conclusion that each center is monogenetic and, especially, that all centers formed from a single magmatic event.

The recognition of time-separate events at basalt centers must be based on establishing unequivocal time-gaps between eruptive events, such as the presence of soil-bounded unconformities (Crowe et al., 1992). The variable K-Ar age determinations provide permissive, but not conclusive, evidence of polygenetic events. Alternatively, the precision of the K-Ar methods may be insufficient to test the polygenetic model. Smith et al. (1990) and Ho et al. (1991) presented evidence, based on field interpretations that Red Cone and Black Cone centers could be polygenetic.

(6) Sleeping Butte Centers. The Sleeping Butte volcanic centers are located 47 km northwest of the potential Yucca Mountain Site (Figure 3.1.1.2.3.5.2.2-1). The Sleeping Butte centers consist of two, spatially separate, small-volume (< 0.1 km^3) basaltic centers. The basalt centers comprise a main scoria cone, flanked by small, satellite scoria cones. Each center erupted multiple lobes of blocky aa lava flows that extruded from the base of their main scoria cone. The southwest center, the Little Black Peak Cone, erupted through fanglomerate deposits. The northeast center, the Hidden Cone, erupted through and draped the north-northeast-facing slope of Sleeping Butte. This prominent topographic mount is upheld by resistant outcrops of moderately welded ignimbrite of Miocene age, located about 1 km inward of the range-front of the Pahute Mesa-Black Mountain highland. The two centers are aligned in a north-northeast direction, the preferential direction of alignment of post-Pliocene volcanic centers in the YMR (Crowe et al., 1983; Crowe and Perry, 1989). The separation of the centers is 2.6 km, measured from crater center to crater center. These centers have been mapped at a scale of 1:5,000 (Crowe and Perry, 1991).

(a) The Little Black Peak Center. The oldest volcanic unit of the Little Black Peak volcanic center includes two mound-shaped accumulations of basalt scoria and volcanic bombs (Qs3; Figure 3.1.1.2.3.5.2.2-3) located at the southern margin of the main cone. These mounds are similar to the scoria mounds described at the Red Cone, Black Cone (Smith et al., 1990; Crowe, 1990), and Lathrop Wells centers (Crowe et al., 1992). There are no distinct vents or craters at the crest of the mounds. However, the abundance of large bombs (> 1 m in diameter) suggests the mounds were vents for basaltic eruptions. Lenticular dikes are exposed locally in the mounds. The dikes fed short, lobate lava flows, exposed west of the eastern scoria mound (Crowe and Perry, 1991). Further details are discussed in Crowe et al. (1994). Minor et al. (1993) reported an age of about 350 ka for the Little Black Peak volcanic center.

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(b) Hidden Cone Center. The Hidden cone center consists of a main scoria cone formed on the north-facing slope of Sleeping Butte (Figure 3.1.1.2.3.5.2.2-4). The center was constructed in two stages: (1) a main scoria cone formed from central-vent eruptions of mildly explosive Strombolian type, accompanied by extrusion of multiple lava flows at the northeast flank of the cone, and (2) a thin sequence of scoria-fall deposits was erupted during mildly explosive Strombolian eruptions. These deposits mantled the eroded slopes of the previously formed main scoria cone.

Champion (1992) and Minor et al. (1993) noted the presence of a possible second lava flow associated with the Hidden Cone center, located northwest of the base of the north cone. Champion (1992) reported a K-Ar age of about 385 ka for the flow, consistent with the age of the lava flows of the Hidden Cone center. This western lava flow was identified in preliminary mapping as being part of the Hidden Cone center. However, detailed field examination showed that the western lava flow could not be traced to the Hidden Cone. Instead, it appears to be associated with deeply dissected scoria deposits that mark one of a series of basalt centers stratigraphically beneath the Thirsty Canyon Tuff. Additionally, the western lavas are overlain by a thick soil with dramatically greater horizon development than the Hidden Cone lava flows. The new K-Ar age determination provides evidence that the western flow could be associated with the Hidden Cone center. [Additional field work will be conducted to verify the stratigraphic position of the western lava flow.] [INN 3.1.1.2.3.5.2.2-4]

Whole-rock, K-Ar age determinations have been obtained for both centers of the basalt of Sleeping Butte. They are 0.29 ± 0.11, 0.32 ± 0.15, and 0.24 ± 0.22 Ma (Crowe et al., 1982). The analytical uncertainty of individual analyses generally exceeds 100 ka (Crowe and Perry, 1991). Turrin (1992) reported $^{40}Ar/^{39}Ar$ ages of about 380 ka for lavas of both of the Sleeping Butte centers. Minor et al. (1993) reported an age of about 350 ka for the Hidden Cone center. The degree of erosional dissection of the lavas is consistent with these age determinations. Additionally, the degree of horizon development of soil on the lavas and cone-apron deposits, and the degree of dissection and development of a cone-slope are consistent with a minimum age of at least several hundred-thousand years (Crowe and Perry, 1991). A U-Th disequilibrium age measurement is being processed for the western lava lobe of the Little Black cone. Cosmogonic, surface-exposure ages, using the ^{3}He method, will be obtained for surface scoria deposits of the youngest event of the Hidden Cone. $^{40}Ar/^{39}Ar$ age determinations will be obtained of both Little Black Peak and Hidden Cone multiple-eruptive units [INN 3.1.1.2.3.5.2.2-1].

Preliminary paleomagnetic data from a limited number of sites in the lava and scoria deposits of both basalt centers were interpreted as a single or closely grouped direction of remanent magnetization (Champion, 1991). Champion (1991) inferred from these data that both centers formed in a single eruptive event and are, therefore, monogenetic volcanic centers. Such an interpretation is consistent with the geology and stratigraphy of the Little Black Peak center (Crowe and Perry, 1991). However, the monogenetic classification of the Hidden Cone center may not be verifiable using paleomagnetic data. There are no deposits of agglutinated spatter

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associated with the youngest eruption of the Hidden Cone center that would provide reliable indicators of the field direction at the time of formation of this deposit. More geochronology work [INN 3.1.1.2.3.5.2.2-2] is planned to evaluate the different models of the eruptive history of the Hidden Cone center (Crowe and Perry, 1991). Further details are discussed in Crowe et al. (1994).

3.1.1.2.3.5.2.3 The Lathrop Wells Center

[Sections 3.1.1.2.3.5.2.3 and 3.1.1.2.3.5.2.4 will be included in future revisions of Chapter 3.]

3.1.1.2.3.5.2.4 Tectonic Setting in Relation to Basaltic Volcanism

3.1.1.2.3.5.2.5 Petrologic and Geochemical Constraints on Basaltic Volcanism

Basaltic volcanism has been the predominant form of volcanism in the Great Basin over the last 5-10 Ma. Great Basin basalts are primarily alkalic in composition indicating small degrees of partial melting of relatively deep mantle sources. The compositions of basalt have become more undersaturated with time, both in the Basin and Range province as a whole, and within individual volcanic fields, suggesting that the intensity of melting in the mantle has declined with time on a regional scale. The volume of basalt flux into the crust has also declined in a broad sense throughout the middle and late Cenozoic, considering that the enormous volumes of ash-flow tuff erupted in the mid-Cenozoic were fueled by a comparable or greater volume of basaltic magma. The transition from the eruption of evolved magmas to basalt in the late Cenozoic can be attributed to changes in plate tectonic processes and crustal properties with time. Isotopic and trace-element studies of basalt have documented the role of both asthenospheric mantle and lithospheric mantle in the generation of basaltic magma in the western United States. The regional distribution of these sources correlates with tectonic setting and history, and may control the volume and compositional distribution of basalt relative to major physiographic boundaries.

Basalt eruptions within the CFVZ follow many of the same patterns seen in other basalt fields of the western United States. Eruptive volumes decrease and the depths of magma chambers possibly increase with time, suggesting an overall decrease in basalt flux from the mantle with time. An unusual aspect of several of the CFVZ basalt centers is that they erupted during multiple eruptive episodes spanning several tens of thousands of years (polygenetic volcanism). Geochemical data show that eruptive episodes at individual volcanic centers are geochemically distinct and cannot be related by the evolution of a single magma. Instead, the geochemical data indicate that several separate magma batches, each representing an independent episode of dike intrusion, were involved in the formation of individual polygenetic centers. The pattern of repeated dike intrusions at single volcanic centers within the YMR provides spatial constraints on where magma will ascend during the lifetime of a polycyclic volcanic center. Further detailed discussion of these topics and how they relate to the Yucca Mountain area can be found in Crowe et al. (1994).

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3.1.1.2.3.5.2.6 Summary of Basaltic Volcanism

[Sections will be included in future revisions of Chapter 3]

3.1.1.2.4 Seismology

[The Technical Basis Report for Tectonics is scheduled for completion in mid FY 96 and will be the basis for updates to the tectonic sections in Revision 1].

3.1.1.2.4.1 Local Site Seismicity

[Note: This section is summarized from existing information in the SCP. Recent data from the Southern Great Basin Seismic Network has not been made available. When these data are released this section will be updated and expanded. This discussion is intended as a guide to the reviewer on the organization and direction of the section. The section provides a basis for the development of ground motion models, a description of data used in determining site response, and, along with Subsection 3.1.1.2.3.2. (Quaternary Faulting History) and Subsection 3.1.1.2.4.4 (Identification of Seismic Sources), the basis for assessment of seismic hazard at the proposed site.]

In the vicinity of the potential repository site, the seismicity is distributed in a broad band trending east-west from the Utah border to California. Figure 3.1.1.2.4.1-1 [INN 3.1.1.2.4.1-1] shows the distribution of earthquakes within 100 km of the site. The earthquakes generally have strike-slip and normal faulting mechanisms with focal depths ranging from near-surface to 12 - 15 km. The number of resolved focal mechanisms indicate that approximately half of the solutions are strike-slip and half are dip-slip (Rogers, et al., 1992).

In this region, as elsewhere in the Great Basin, there is a general lack of correlation between the distribution of epicenters and Quaternary faults, especially on faults on which east-west extension has occurred. These faults include the northwest trending right-lateral faults and northeast trending left-lateral faults of the Walker Lane.

The general features of Great Basin seismicity are summarized below.

a. Seismic activity is generally expressed in clusters of earthquakes distributed in an east-west belt between latitude 36 and 38° north (the southern Nevada seismic zone).

b. The earthquake clusters are not readily associated with the surface traces of known faults. These clusters may align with local structural grain, and composite and single event focal mechanisms suggest that nodal planes correlate with regional stress directions. The clusters of events may be diffusely distributed around mapped faults, covering areas larger than the surface projections of the rupture.

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c. Right lateral slip on northerly trending faults is the predominant mode of stress release near the site. Subordinate faulting on east-northeast (left-lateral) and northeast (normal) has been observed as well as oblique slip on structures of intermediate orientation with the appropriate dip angles.

d. The majority of the earthquakes are located between 5 and 15 km in depth. There are some areas in which the seismicity is shallow (less than 5 km) and a few events have been located at depths greater than 15 km, Rogers et al. (1983), show that most of the seismic energy released in the southern Great Basin occurs at depths less than 12 km.

e. A zone of quiescence centered on Yucca Mountain is apparent in all studies describing the seismicity in the southern Great Basin. Brune et al. (1992) and Gomberg (1991a and b) have shown that this zone is a real feature of the seismicity and not an artifact of network design.

f. The most active regions within the southern Great Basin the northern and southeastern sections of the NTS, the area west of Bare Mountain fault, the region at the northern end of the Death Valley-Furnace Creek fault system, and the Pahranagat Shear Zone (Rogers, et al., 1987). The region surrounding the NTS is the site of testing of nuclear devices and the level of seismic activity may reflect tectonic release following the blasts. Blasting (chemical) has also occurred in the Bullfrog Hills west of the Bare Mountain Fault and some of this activity in Figure 3.1.1.2.4.1-1 may reflect these events. The specific characteristics of these active zones are described in the next section.

3.1.1.2.4.1.1 Northern NTS

The northern section of NTS includes the Timber Mountain caldera, Pahute Mesa, and Yucca Flat. These areas have been the focus of considerable seismic energy release. This region has also been the location of nuclear detonations since 1951. Determining which events are related to UNEs, either cavity collapse events or induced events, is problematic. It is reasonable to expect that there is the normal level of background seismicity in the area. A study to determine the number of artificial and induced seismic events in this area suggest that the natural seismicity in the area reflects the background activity in the southern Basin and Range (Vortman, 1991).

In 1979 and 1983, several swarms of microseismicity have occurred in this area unrelated to the UNEs, in the vicinity of Dome Mountain and Thirsty Canyon. Focal mechanisms suggest both normal and strike slip faulting on northeast and north trending structures (Rogers et al., 1987).

The relation between UNEs and seismicity has been discussed by Rogers et al. (1987), Vortman (1991), and Hamilton et al. (1971). The exact number of induced versus natural events is not as important as the fact that a considerable number of small earthquakes occur in this area which appear to be related to the UNE. Vortman (1991) has suggested that these

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induced events are responding to the change in stress as the UNE shock wave passes. Some events are triggered by the arrival of the UNE phase, others are responding to changes in an altered stress field caused by the explosion. A UNE may cause a stress change on the order of bars, a fraction of the lithostatic stress in the hypocentral region. This indicates that some areas of the southern Great Basin are in a state of critical stress in which a small perturbation in the load on a fault causes the release of accumulated strain.

3.1.1.2.4.1.2 Southern NTS

The southern portion of the NTS is a seismically active region relative to other areas in the southern Great Basin. Some of the events in this area may be related to underground testing; however, there seem to be considerable numbers of small to moderate earthquakes related to natural tectonic stress release (Gomberg, 1991a and b). The largest event in this region is the M_L 5.6 Little Skull Mountain which occurred in 1992 (see below).

Seismicity in the southern NTS region is generally diffusely distributed, with a relatively high degree of microseismic activity. Seismicity in the Jackass Flats area, 15 to 20 km east of the potential site, shows some weak northeast trends in epicentral alignment. Focal mechanisms from Rogers et al. (1987) indicate sinistral slip on northeast structures and (or) dextral slip on northerly trending structures. The dominate trend of mapped faults in this region is northeast. Seismicity in the Rock Valley area also shows northeast trends, parallel to the Rock Valley Fault Zone.

A conspicuous hole in the seismicity pattern exists in the vicinity of Yucca Mountain. A rigorous analysis of the earthquake detection threshold for the Southern Great Basin Seismic Network suggests that this zone of quiescence is real (Gomberg, 1991b). An experiment in high resolution monitoring of seismicity at the potential site by Brune et al. (1992) also confirms the existence of the quiescent zone. The significance of this feature is open to speculation. Modeling of the strain field in southern Nevada by Gomberg (1991a) suggests that this area is not accumulating significant amounts of strain, an interpretation consistent with the observations of seismicity.

3.1.1.2.4.1.3 Northern Amargosa Valley - Sarcobatus Flat

This region encompasses the areas west and northwest of the Bare Mountain Fault, 25 to 90 km from the site. Seismicity in the northern Amargosa Valley is diffusely distributed in the vicinity of Beatty and the Bullfrog Hills. Some of this activity may be related to mining (Vortman, 1991). In Sarcobatus Flat, earthquakes have occurred in four clusters since the advent of instrumental monitoring (Rogers et al., 1983, 1987). These clusters are spaced roughly 10 to 20 km apart in a northerly trend along the length of the valley. Focal mechanisms for the three southern clusters suggest dextral slip along north to northeast trending structures. The fourth cluster is located in the northern part of the valley and shows normal faulting on a northeast trending fault (Rogers et al., 1987).

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3.1.1.2.4.1.4 Northern Death Valley Region

There is a concentration of seismicity along the Furnace Creek Fault Zone in northern Death Valley. These earthquakes are diffusely distributed over an area much larger than the mapped surface traces of the faults. A concentration of epicenters extends northeast from this area through the Gold Mountain-Mount Dunfee region. The largest event in this area was an M_L 4 at Gold Mountain. A focal mechanism for this event suggests sinistral slip on a northeast trending nodal plane (Rogers et al., 1987). A cluster of events occurred in a northeast alignment near Mount Dunfee in 1983. A composite focal mechanism from some of these earthquakes suggests oblique normal faulting on a northeast-striking plane (Rogers et al., 1987). In 1993, west of northern Death Valley, the M_L 6 Eureka Valley earthquake occurred on a northeast trending fault. This event, 110 km northwest of Yucca Mountain, is discussed in Subsection 3.1.1.1.4.2.3.

3.1.1.2.4.1.5 1992 M_L 5.6 Little Skull Mountain Earthquake

On June 29, 1992, an M_L 5.6 earthquake occurred beneath Little Skull Mountain approximately 20 km southeast of Yucca Mountain. This event, although relatively small, is significant because it is the largest event recorded near the potential repository site. The event triggered several strong motion instruments in the region. The largest recorded acceleration was 0.21 g at Lathrop Wells, 15 km from the epicenter. It also caused some damage to the Field Operations Center on the NTS. The epicenter is located in a cluster of small events representing high seismic activity in the NTS that have been recorded since 1978 (Rogers et al., 1991).

The mainshock of the Little Skull Mountain earthquake was located at 10 km below sea level in a region of predominately northeast striking structures. The fault plane solution indicates normal faulting of a east-northeast striking plane dipping to the southeast. The fault motion is mainly down dip with some northeasterly directed slip. The nearly horizontal northwest oriented T-axis indicates that the principal stress direction is in accordance with the regional observed N50 W direction. The magnitude of the event has been determined at M_b 5.6 (Anderson et al., 1992), M_w 5.6 to 5.8 (Walter, 1993). No surface rupture has been reported for this event. The mainshock fault plane is roughly on strike with the Rock Valley fault system (Sheehan et al., 1993a).

Numerous aftershocks of this earthquake have been recorded, the largest being an M_L 4.4 on July 5, 1992. The aftershock epicenters form an "L" shaped pattern with the long arm extending northeast and the shorter one extending southeast. The focal depths are tightly clustered between 7 and 11 km. Focal mechanisms indicate both normal faulting consistent with the mainshock and strike-slip faulting in approximately equal proportions (Sheehan et al., 1993b). The trend of the aftershocks (N35E) and the surface projection of the aftershock plane aligns with an inferred southwestern extension of the Mine Mountain Fault (Sheehan et al., 1993b).

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The timing of this event has led some to suggest that the Little Skull Mountain earthquake was triggered by the M_s 7.5 Landers, California, earthquake, which occurred less than 24 hours earlier (Anderson, et al., 1992). An increase in microseismic activity in the vicinity of Yucca Mountain was observed in the coda of the Landers event as recorded by the Yucca Mountain Microearthquake Array (Brune, et al., 1992). This activity increased over the next 23 hours culminating in the Little Skull Mountain sequence. A similar pattern of increased activity in eastern California and western Nevada has been reported, though no events as large as the Little Skull Mountain event occurred (Anderson et al., 1992). Triggering of earthquakes has been observed before in the southern Great Basin from nuclear detonations (Rogers et al., 1983) and from the filling of Lake Mead (Rogers and Lee, 1976). These events can range in magnitude from very small to M 5. It is interesting to note that in the days prior to the Landers mainshock two small events occurred in the vicinity of the Little Skull Mountain mainshock. In February 1992, several small earthquakes (the largest an M_L 2.4) also occurred in this region. This suggests that the Little Skull Mountain region was near failure prior to the Landers earthquake which may have advanced the time of rupture. How much of an advance in timing is not known. Anderson et al. (1992) observed an increase in seismicity at Mammoth Lakes following the Landers mainshock. In the same region, 50 km to the southeast, the May 17, 1993, M_L 6 Eureka Valley, California, earthquake also occurred on a northeast trending structure similar to the Little Skull Mountain fault plate orientation. Presumably this region was also near failure during the Landers mainshock, thus one would have expected it to rupture if the triggering effects were significant.

3.1.1.2.4.2 Ground Motion Model

3.1.1.2.4.2.1 Introduction

This section describes the model that will be used to estimate ground motions at the Yucca Mountain Site for use in seismic design and evaluations. The model includes the explicit estimation of uncertainty in ground motion estimates, and can be used to develop ground motion estimates for use in either deterministic or probabilistic approaches. The ground motion products that can be generated using this model include peak ground motion parameters such as peak acceleration and velocity, response spectral ordinates at specified periods, power spectral density functions, and ground motion time histories representative of specific spectra. The model includes the contributions of source, path, and site effects to ground motions, and considers all the effects currently recognized in engineering models of strong ground motion as important contributors to ground motions on rock and stiff alluvium resulting from crustal earthquakes. In the following description of the model, the focus is on source and path effects. Local site effects on ground motions are considered in Subsection 3.1.1.2.4.4. The description of the model is followed by a summary of what is known about ground motion characteristics at Yucca Mountain based on available data.

A large volume of strong motion recordings is available from the Yucca Mountain Site to provide a basis for developing a site-specific characterization of ground motions at the site. The data include recordings of both earthquakes and NTS explosions, both at the surface and downhole. Surface recordings exist at both rock and alluvial sites, and have topographic

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conditions ranging from flat to steep. Also, as a result of the extensive subsurface investigations that have been undertaken and are planned at the Yucca Mountain Site, our knowledge of the physical properties of the site and the wave propagation characteristics of the site is potentially very extensive, allowing a detailed characterization of ground motion response at the Yucca Mountain Site.

Summary of Model for Estimating Ground Motions at Yucca Mountain. The ground motion model includes both empirical and numerical components. The empirical component is derived from analyses of a large set of earthquake and explosion strong motion recordings from the western United States. In the development of ground motion estimates, an approach based on regression analyses is augmented by an approach based on statistical analyses of suites of strong motion recordings selected to match the magnitude, closest distance, site conditions, and other site-specific aspects (such as faulting mechanism) of the design basis earthquakes or explosions. The primary data base of strong motion recordings includes sites having geological and seismic wave velocity characteristics compatible with the Yucca Mountain Site. Data bases including a broader range of site characteristics may be required to evaluate effects discussed below such as the magnitude dependence of dispersion, or dependence of ground motions on style of faulting. The empirical ground motion model describes the dependence of peak acceleration, peak velocity, and response spectral ordinates on earthquake or explosion magnitude and closest distance to the source. Separate ground motion attenuation models are given for deep sources (earthquakes that do not rupture close to the surface) and shallow sources (explosions and earthquakes that rupture close to the surface). Separate ground motion models are given for vertical and horizontal motions. The dependence of ground motions on the type of faulting is described. For dip-slip faulting, the difference between ground motions on the foot wall and hanging wall is described as a function of closest distance. The attenuation relations are adjusted at close distances to account for near fault directivity effects at the longer periods. The dispersion in ground motions about their median value is described as a function of magnitude and period. A comparison is made between the predictions of these models and ground motions from basin and range earthquakes and explosions, and earthquakes from other regions considered representative of earthquake source conditions at the Yucca Mountain Site. Modifications to the empirical model are made based on this comparison if warranted by the data.

The empirical ground motion model described above is based on a large set of recorded data that are applicable to the site in a general way, but do not reflect the actual geometrical orientation of the faults near the site nor the seismic wave velocity structure beneath the site. Also, the empirical data are too sparse to provide definitive information on issues such as the dependence of ground motion on style of faulting, and on the characteristics of near-fault ground motions. The purpose of the numerical model is to provide site-specific estimates of these source geometry and wave propagation effects, and to augment the empirical data in quantifying style of faulting and near fault effects on ground motions. Ground motions generated using the numerical model can be used both to guide the development of attenuation relations, and to generate suites of time histories for use in estimating the ground motions of deterministic earthquakes and explosions to augment the statistical approach based on recorded data.

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The numerical component of the ground motion model is a set of procedures that use physical models of the earthquake and explosion source and the wave propagation path to generate ground motion time histories. The characteristics of the site (the shallow part of the path beneath Yucca Mountain), described in more detail below, are also included in the model. The earthquake source model uses a discretization of the rupture surface to represent the faulting process. The radiation from the fault elements is described by either theoretical models or by the use of empirical source functions derived from strong motion recordings of small earthquakes. The explosion source is described either by theoretical models or by empirical source functions. Similarly, the Green's functions representing the propagation path are described either by theoretical models or by the use of empirical Green's functions derived from strong motion recordings at the Yucca Mountain Site from explosions and small earthquakes. The selection of theoretical or empirical approaches to modeling the source and wave propagation effects is based on considerations of the frequency range of the ground motions being estimated and on the quality and appropriateness of empirical source functions and Green's functions. Using the numerical model, the effects of faulting style, difference between foot wall and hanging wall motions, and near fault directivity are estimated using the specific geometry of the faults in relation to the site. The uncertainty in ground motion estimates derived from the numerical model is derived from estimates of modeling and random uncertainty and parametric uncertainty.

Data Analysis Methodology for Development of Ground Motion Model. Important developments have occurred in the past several years in the methodology used to derive empirical ground motion attenuation relations using regression. A summary of modern regression methods is given by Joyner and Boore (1993). In particular, the procedure used to separate inter-event and intra-event randomness using the random effects model has been rationalized (Abrahamson and Youngs, 1992).

In the numerical component of the ground motion model, recorded data are first used to check the assumptions on which the numerical model is based. For example, if surface waves dominate the ground motions from shallow sources, as indicated by the data described below, the ground motion model needs to include surface waves. Having thus established the appropriateness of the formulation of the ground motion model, the data are next used to test the efficacy of the model in explaining the data. This modeling of the data can be used to refine the selection of parameters in the model that describe the source, path and site. It can also provide an estimate of the modeling uncertainty associated with the use of the model, as described below.

A wide range of ground motion modeling approaches that can augment the information provided by empirical ground motion attenuation models are described by the Electric Power Research Institute (1990). The selection of the modeling approach used should take into consideration the kinds and quality of data available to constrain the ground motion model parameters and provide inputs into empirical source function and empirical Green's function methods, the characteristic features of ground motions recorded at the site, and the adequacy of the model in describing these ground motion characteristics over the frequency range of importance at the site. An approach based on empirical source functions and simplified

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theoretical Green's functions was used for near-fault calculations of high frequency ground motions at the Diablo Canyon site. The availability of strong motion recordings at the site from the Little Skull Mountain earthquake sequence and other regional earthquake sequences may make the use of empirical source function and empirical Green's function summation procedures particularly advantageous because they can include site response, discussed further below, in a site-specific empirical manner, and avoid the uncertainties involved in empirical measurements of site response and numerical modeling of site response. Another widely used method is the band-limited white noise-random process model, which is based on a spectral description of the source and path effects. The data required to develop estimates of the parameters required by this model are available in the seismograms described above. It may be appropriate to use different modeling methods for long period ground motions, which are characterized by relatively deterministic behavior, and high frequency ground motions, which are characterized by relatively stochastic behavior.

[To explain the large surface waves observed from nuclear explosions and shallow earthquakes, it may be necessary to use two-dimensional or three-dimensional velocity models to describe the wave propagation characteristics of shallow paths.] A finite difference method appropriate for these calculations is described by Helmberger and Vidale (1988), and examples of the applications of these models are given by Vidale and Helmberger (1988), Frankel and Vidale (1992), and Graves (1993).

Estimation of Uncertainties. An empirical ground motion model is actually a mathematical model selected to fit a set of recorded data. In using empirical ground motion models, it is therefore necessary to include the uncertainty associated with selecting this mathematical model as part of the overall uncertainty in estimating ground motions. One method of accomplishing this is to use more than one empirical ground motion model. For example, Campbell (1990) uses models that include depth of sediments and depth to crystalline rock as parameters, which yield lower standard errors than other models that do not use such parameters. The use of multiple attenuation relations also incorporates differences in ground motion models that result from different selections or categorizations of data.

The use of the random effects model has provided estimates of inter-event and intra-event contributions to variance in empirical ground motion attenuation relations, and clearly demonstrated the reduction in variance with increasing earthquake magnitude (Youngs et al., 1993). For peak acceleration, this reduction is more pronounced for horizontal motions than for vertical motions. There is also a dependence of variance on period, with variance tending to increase at the longer periods. For soil sites, this period dependence can be reduced by incorporating the depth of soil as a term in the regression, and for rock sites by incorporating depth to basement rock (Campbell, 1990). There are significant differences between "soft rock" and "hard rock" ground motions at moderate and long periods.

A rigorous procedure for estimating the uncertainty in ground motion estimates derived from numerical modeling has been developed by Abrahamson et al. (1990). The uncertainty consists of two contributions: The first, which consists of modeling and random uncertainty, represents the variance between recorded and calculated ground motions in situations where

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the basic information required for calculating the motions (such as earthquake source parameters, seismic velocity structure of the path, and site conditions) is known. It is estimated from the discrepancy between recorded ground motions of past earthquakes and ground motions calculated using the known source, path and site models that pertain to these events. The second, which consists of parametric uncertainty, represents our uncertainty in the source parameters of future earthquakes that may affect the site in question, and the uncertainty in the path and site effects at that site.

3.1.1.2.4.2.2 Characteristics of Ground Motions at Yucca Mountain

Earthquake Data. A large volume of strong motion recordings is available from the Yucca Mountain Site to provide a basis for developing a site-specific characterization of ground motions at the site. The data include recordings of both earthquakes and NTS explosions, both at the surface and downhole. Surface recordings exist at both rock and alluvial sites, and have topographic conditions ranging from flat to steep. Also, as a result of the extensive subsurface investigations that have been undertaken and are planned at the Yucca Mountain Site, our knowledge of the physical properties of the site and the wave propagation characteristics of the site is potentially very extensive, allowing a detailed characterization of ground motion response at the Yucca Mountain Site.

Seismic data are currently being recorded by several networks. Microearthquake activity is monitored by the Southern Great Basin Seismic Network of the UNR. In addition to these high gain stations which cover a wide area, UNR also operates portable strong motion instruments and broadband velocity instruments located at the Yucca Mountain Site. There are plans to deploy downhole as well as surface stations in Midway Valley. Both UNR and the USGS operate portable seismic networks to record aftershock sequences and investigate crustal structure. The US National Seismic Network has a very broadband instrument at Shoshone Peak, located within NTS.

Figure 3.1.1.2.4.2.2-1 [INN 3.1.1.2.4.1.1-1] shows the locations of the 62 current analog stations of the Southern Great Basin Seismic Network. The open triangles just north of the southern NTS border are stations installed after the 6/29/92 Little Skull Mountain earthquake; the remainder were installed in the period from 1978-1981. Except for 10 stations, these are vertical-component only. In addition, about eight portable instruments were deployed by UNR around Little Skull Mountain for a period of six months after the earthquake. These were within about 10 km of the epicenter and all recorded three components. Similarly, the USGS deployed about 10 portable instruments, but for a lesser period of about three months after that earthquake. Between April and June of 1993, the UNR deployed eight three-component instruments in Midway Valley, just east of Yucca Mountain, to gather data for ground-motion studies.

The earthquakes for which Yucca Mountain Site recordings are available include the Landers and Big Bear earthquake sequences of June 28, 1992, the Little Skull Mountain earthquake sequence of June 29, 1992 (URS Blume, 1992), the Southern Utah earthquake of September 2, 1992, the Rock Valley earthquake sequence of May 15, 1993, and the Eureka Valley

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earthquake sequence of May 17, 1993. This earthquake sequence was recorded on a profile that extends from west to east across Yucca Mountain. A partial list of some of the larger events is given in Table 3.1.1.2.4.2.2-1. The seismograms of these events are being published by the UNR (Sheehan et al., 1993a).

These earthquake recordings provide an abundance of data for the development of empirical source functions and empirical Green's functions for use in site-specific numerical modeling of ground motions. They also provide an abundance of data for the empirical characterization and numerical modeling of wave propagation effects and site response, including the reduction of ground motions with depth below the ground surface.

Characteristics of Earthquake Ground Motions at Yucca Mountain. To date, earthquake ground motions at Yucca Mountain have been estimated using attenuation relations derived principally from strong motion recordings of earthquakes in the western United States. This strong motion data set has not included any recordings from the vicinity of Yucca Mountain. However, the occurrence of the magnitude 5.6 Little Skull Mountain earthquake on June 29, 1992 provided ground motion data which can serve as a check of these empirical attenuation relations (URS/John A. Blume & Associates, 1992; Anderson et al., 1993). Locations of some of the stations that recorded the earthquake are shown in Figure 3.1.1.2.4.2.2-2 [INN 3.1.1.2.4.2.2-2].

The magnitude 5.6 mainshock produced peak velocities about 50 percent larger than those predicted by the average of the Joyner-Boore (1981) and Campbell (1990) attenuation relations at Lathrop Wells at a distance of 15 km, and about twice as large at an NTS site at a distance of 30 km (Figure 3.1.1.2.4.2.2-3). [INN 3.1.1.2.4.2.2-3] The recorded response spectra at 5 percent damping are compared with those calculated using the average of the Joyner-Boore (1981) and Campbell (1990) relation in Figure 3.1.1.2.4.2.2-4.

Velocity seismograms of four earthquakes recorded on rock in Midway Valley have been examined. The first is a magnitude 3.2 aftershock of the June 29, 1992, Little Skull Mountain earthquake, which occurred at a depth of about 9.5 km at a distance of about 16 km from the recording station. The second is a magnitude 4.0 earthquake in Rock Valley that occurred on June 15, 1993, at a very shallow depth (nominally 0 km). The third and fourth are magnitude 4 aftershocks of the May 17, 1993 Eureka Valley earthquake at depths of 6 and 2 km, respectively. The parameters of the earthquakes are listed at the bottom of Table 3.1.1.2.4.2.2-1, and the locations of the first two epicenters and the recording station are shown in Figure 3.1.1.2.4.2.2-5 [INN 3.1.1.2.4.2.2-4].

The three component velocity seismograms of the first two earthquakes are shown in Figure 3.1.1.2.4.2.2-6 [INN 3.1.1.2.4.2.2-5]. The seismograms of the deeper event have "S" waves that are much larger in amplitude than the surface waves. These seismograms may be representative of the motions of large earthquakes that control the seismic hazard at Yucca Mountain. In contrast, the seismograms of the shallow event have surface waves that are larger than the "S" waves. It appears that the shallow source depth of this earthquake caused the generation of scattered higher mode surface waves in the strongly heterogeneous upper

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crust; these arrive following the fundamental Rayleigh wave. It is shown below that this seismogram strongly resembles the seismograms of nuclear explosions. The three component velocity seismograms of the two Eureka Valley aftershocks are shown in Figures 3.1.1.2.4.2.2-7 and 3.1.1.2.4.2.2-8 [INN 3.1.1.2.4.2.2-6]. The peak velocities of both seismograms are dominated by body waves. However, surface waves are present in both seismograms, and are more pronounced for the shallower event.

Since most of the energy release occurs at depths greater than 3 km in large earthquakes, the seismograms of the shallow Rock Valley earthquake may not be representative of the ground motions of moderate magnitude earthquakes that dominate the seismic hazard at the site. However, analysis of ground motion characteristics at the site addresses the possible effects of shallow faulting, especially from large magnitude earthquakes, on the generation of surface waves at the site.

Explosion Data. There is a vast data set of strong motion recordings of about 660 NTS nuclear explosions spanning the time period 1951 through 1992. The explosions were recorded on a network of approximately one hundred stations, including stations at Yucca Mountain. Some of the publications that describe the Yucca Mountain recordings are Phillips, 1991; Subramanian et al., 1990; Walck and Phillips, 1990. NTS explosion sources and Yucca Mountain strong motion recording stations are shown in Figure 3.1.1.2.4.2.2-9 (Walck and Phillips, 1990). The Yucca Mountain stations that have monitored nuclear explosions include 11 stations inside or within a few km of the perimeter of the repository, four of which also have downhole recordings.

These explosion recordings provide an abundance of data for the development of empirical source functions and empirical Green's functions for use in site-specific numerical modeling of ground motions. They also provide an abundance of data for the empirical characterization and numerical modeling of wave propagation effects and site response, including the reduction of ground motions with depth below the ground surface.

Characteristics of Explosion Ground Motions at Yucca Mountain. The strong motions of NTS explosions at Pahute Mesa recorded at four Yucca Mountain stations were analyzed by Phillips (1991). The four sites, shown in Figure 3.1.1.2.4.2.2-10 (Sites 14, 21, 22 and 23) have significantly larger ground motions than at other stations at comparable distances within NTS, as seen in Figures 3.1.1.2.4.2.2-11 through 3.1.1.2.4.2.2-14 [INN 3.1.1.2.4.2.2-7 through INN 3.1.1.2.4.2.2-10]. Amplifications of horizontal PSRV in the frequency range of 0.5 to 5 Hz are about a factor of 2 on average. More specifically, the amplification factors for horizontal motions are approximately 2.1, 2.4, 2.0, and 1.8 for frequencies of 0.5, 1, 2, and 5 Hz; the corresponding amplification factors for vertical motions are 2.4, 2.4, 2.3, and 1.8.

Strong motion recordings along profiles extending from NTS southward to the Nuclear Rocket Development Station at Jackass Flats (Weetman et al., 1970) provide insight into the composition of the wave field at Yucca Mountain from NTS explosions. The location of stations that recorded the Pipkin explosion are shown in Figure 3.1.1.2.4.2.2-15 [INN 3.1.1.2.4.2.2-11], and the profile of velocity recordings of the radial component is shown in

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Figure 3.1.1.2.4.2.2-16 [INN 3.1.1.2.4.2.2-12]. Surface waves appear to propagate from near Pahute Mesa to Jackass Flats, and have peak amplitudes that are comparable to or larger than those of the "S" waves. The seismograms are similar to the seismogram of the very shallow Rock Valley earthquake shown at the bottom of Figure 3.1.1.2.4.2.2-6. This suggests the possibility that the amplification effects observed in the Yucca Mountain recordings of NTS explosions are due to the shallow path between NTS and Yucca Mountain, and not to site effects. The shallow structure for this path to Yucca Mountain (path PM2 in Figure 3.1.1.2.4.2.2-9) is shown in the center panel of Figure 3.1.1.2.4.2.2-17 (Walck and Phillips, 1990).

It is important to determine whether surface waves cause the amplification effects observed in the Yucca Mountain recordings of explosions. This can be done by comparing the ratio of body waves to surface waves recorded at Yucca Mountain with the ratio recorded at other stations at a comparable range that reflects the average ground motion characteristics of the NTS data set as a whole. If the amplification is due to surface waves and not body waves, it may not be present in the recordings of earthquakes except in those that have significant energy release at shallow depths. The earthquake seismograms shown in Figures 3.1.1.2.4.2.2-6 through 3.1.1.2.4.2.2-8 indicate that the large surface waves evident in the explosion data recorded near Yucca Mountain may not be present in the recordings at Yucca Mountain of earthquakes that occur at depths greater than a few km.

Summary of Ground Motion Characteristics at Yucca Mountain. A significant aspect of ground motions recorded on rock at the Yucca Mountain Site is the presence of large amplitude surface waves in the seismograms of nuclear explosions and shallow earthquakes. These surface waves are absent in the seismograms of local and regional earthquakes occurring at depths greater than a few km based on review of a small data set. Since the response of underground openings is primarily controlled by peak velocity, which occurs at a period of about 1 second (or longer for nuclear explosions), these surface waves may contribute significantly to the seismic hazard at the site. The ground motion model used in seismic hazard analysis addresses both deep earthquake sources, for which body waves dominate the ground motions on rock at close distances, and shallow sources, for which surface waves may dominate at the Yucca Mountain Site. An important issue for source characterization is the expected incidence of shallow faulting and its magnitude dependence. A corresponding issue for ground motion characterization is the depth below which surface wave excitation becomes secondary to body wave excitation.

To make full use of the abundant recorded data, it is important that detailed seismic velocity and density measurements be made in the site region. [In particular, downhole measurements of seismic velocities in wells, especially those that contain or have contained seismic instrumentation, were used for the development of wave propagation and site response models using recorded data.]

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3.1.1.2.4.3 Seismic Hazard Assessment

3.1.1.2.4.3.1 Summary of Site Response Model Required for Seismic Hazard Assessment

The site response model describes the modification of incident ground motions by the shallow geologic structure that underlies Yucca Mountain, including the effect of topography and depth of burial on ground motions at the site. In one form, it can be described by a modification to be applied to ground motion peak values and spectral ordinates derived from empirical ground motion attenuation relations, statistical analyses of recorded motions, or numerical modeling. This modification, which may influence the overall level of the ground motions and their spectral shape, can be derived empirically from analysis of ground motions recorded at the site or estimated from numerical modeling of site response. In the numerical modeling approach, the site response can be a module of the numerical ground motion model that is interfaced with the source and path modules. If the numerical ground motion model uses strong motion recordings at the site from earthquakes or explosions as empirical source functions or empirical Green's functions, the response of the site can be incorporated in a site-specific empirical manner. This would obviate the need to separately include site response in the ground motion model, and avoid the uncertainties involved in empirical measurements of site response and numerical modeling of site response.

3.1.1.2.4.3.2 Data Analysis Methodology for Development of Site Response Model

Site response can be estimated from the data in two ways. The first is to use the recorded data to empirically evaluate site effects such as the variation of surface ground motion levels within the Yucca Mountain Site and the variation of ground motion levels with depth below the surface. The second approach is to use a model of the seismic velocity structure of Yucca Mountain to numerically compute the response of a given site to strong ground motions.

In the empirical approach to site response analysis, a method is required to establish a relevant reference level against which the site ground motions are compared. This reference level may be an empirically based model or a numerical model. An example of an empirically based model is an attenuation relation for an appropriate site category.

In the numerical approach to estimating site response, a detailed model of the shallow seismic velocity structure of the Yucca Mountain Site is required to compute the response of the site to strong ground motions. The purpose of the model is to provide a means of estimating the site response for seismic sources, such as the faults in the immediate vicinity of the site, for which recorded data are not available. The seismic velocity model is derived from borehole, seismic refraction, and seismic reflection data. The model is further constrained by testing it against recordings of earthquakes and NTS explosions both at the surface and downhole. The procedure for calculating site response using this model may be interfaced with the numerical ground motion model described above which addresses the source and path effects.

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3.1.1.2.4.3.3 Estimation of Uncertainties

The variance in ground motions described by empirical ground motion attenuation models is derived from the variability in recorded data for a given site category. The variance thus reflects not only the variance at a given site from event to event, which is relevant at a specific site, but also the variability in response from site to site, which is only relevant at a specific site if there is no knowledge of its site response. To incorporate an estimate of site response, the estimate of variance in the empirical attenuation relation is replaced with that derived from site data. Generally, an individual site is expected to have a lower variance than that of empirical attenuation relations for a given site category. However, if the variance of the site response is derived from small earthquakes, it may not be applicable to larger earthquakes because of the observed tendency of the variance to decrease with increasing magnitude.

It is found that for larger magnitudes, the intra-event contribution to the variance is larger than the inter-event contribution (Youngs et al., 1993). The relatively small inter-event contribution indicates that differences in ground motions between different events of the same magnitude are relatively small, especially at large magnitudes. Most of the variability in ground motions is attributable to individual path and site effects. This indicates that if the path and site response characteristics at a given site are known, the variance in the ground motion may be significantly lower.

If the site response is estimated using numerical modeling and interfaced with a numerical model that addresses source and path effects, the methodology described above for estimating uncertainty (Abrahamson et al., 1990) can be used to describe the variance in site response and include it in the overall uncertainty in ground motions. The estimate of modeling uncertainty could be derived from differences between recorded surface motions and surface motions calculated using downhole motions together with a wave propagation model and a model of the structure between the downhole and surface recorders. The estimate of parametric uncertainty could be derived from varying the parameters of the structure model and the parameters that describe the incident wave field such as its orientation, wave composition, phasing, and frequency content.

3.1.1.2.4.3.4 Characteristics of Site Response at Yucca Mountain

Data. A large volume of strong motion recordings is available from the Yucca Mountain Site to provide a basis for developing a characterization of site effects on ground motion. The data include recordings of both earthquakes and NTS explosions, both at the surface and downhole. Surface recordings exist at both rock and alluvial sites, and have topographic conditions ranging from flat to steep. These earthquake and explosion recordings provide an abundance of data for the empirical characterization and numerical modeling of wave propagation effects and site response, including the reduction of ground motions with depth below the ground surface.

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As a result of the extensive subsurface investigations that have been undertaken and are planned at the Yucca Mountain Site, our knowledge of the physical properties of the site and the wave propagation characteristics of the site is potentially very extensive, allowing a detailed characterization of ground motion response at the Yucca Mountain Site. To make full use of the abundant recorded data, it is important that detailed seismic velocity and density measurements be made in the site region. [In particular, downhole measurements of seismic velocities in wells, especially those that contain or have contained seismic instrumentation, are needed for the development of wave propagation and site response models using recorded data.]

Amplitude Reduction Due To Depth of Burial at Yucca Mountain. Estimates of amplitude reduction due to depth below the surface have been derived from downhole data recorded at Yucca Mountain. Strong motion data from explosions have been recorded at the surface and downhole at several Yucca Mountain Sites. These recordings provide an opportunity to evaluate the variation of ground motions with depth beneath Yucca Mountain.

An example of surface and downhole recordings at a nearby site in Jackass Flats is shown in Figure 3.1.1.2.4.3.4-1 [INN 3.1.1.2.4.3.4-1]. The three rows of time histories from top to bottom are for acceleration, velocity, and displacement. The three columns of time histories from left to right are for the surface motions on alluvium; the motions at the base of alluvium at a depth of 61 meters; and the motions at a depth of 356 meters below the surface in tuff. The amplitude scale is the same for each depth. The increase in amplitude as the wave approaches the surface, and the frequency dependence of this increase, are evident in these time histories. As in the seismograms shown in Figure 3.1.1.2.4.2.2-16, these seismograms have relatively large surface waves.

Appendix A-4 of Subramanian et al. (1990) describes strong motion data from explosions recorded at the surface and downhole at Yucca Mountain Sites. The data most relevant to Yucca Mountain are from three sites, all on tuff, two of which are located on ridge crests (W12/30 and W28) and one located at the base of a ridge (W25) as shown in Figure 3.1.1.2.4.3.4-2 [INN 3.1.1.2.4.3.4-2]. While none of these sites is located within the perimeter of the repository, site W12/30 is located at the ridge crest just south of the perimeter, and the motions at 360 meters depth should be relevant to the deeper parts of the ramps. Site W29 is located east of the mountain, near the entrance of the northern ramp, and the motions at 83 meters depth at this site should be relevant to the shallow part of the ramps.

The amplification of peak velocity from a depth of about 360 meters to the surface, averaged over approximately 10 events, is 1.58, 2.07, and 2.25, respectively. The reduction factors that would adjust surface peak velocities to about 360 meters depth at sites W12/30, W28, and W25 are 0.63, 0.48, and 0.44. The reduction factors for peak horizontal and vertical velocity and acceleration are listed in Table 3.1.1.2.4.3.4-1.

The amplification at W25, which was expected to be deamplified by topography, is larger than that at W12/30 and W28, which was expected to be amplified by topography. Thus, the influence of topographic effects on the data is not clear from this set of data. However, as

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discussed below, the presence of steep topography at some sites requires that the effects of topography and depth of burial be separated.

The analyses of amplitude reduction that have been performed to date have used the whole seismograms, which at intermediate and long periods are dominated by surface waves. However, shear waves are also present in these seismograms, and by windowing and analyzing these separate arrivals, the influence of depth of burial on both shear waves and surface waves could be evaluated.

The reduction factors derived from explosion data recorded on Yucca Mountain are comparable with those that have been observed downhole from earthquake data in rock. For example, Anderson et al. (1993) found a 2 Hz horizontal response spectral reduction factor of about 0.65 between the top of Little Skull Mountain and the tunnel 100 meters beneath the surface from the September 7, 1992 aftershock of the Little Skull Mountain earthquake at an epicentral distance of 3.5 km; the corresponding reduction of the vertical motions was 0.5 (Figure 3.1.1.2.4.3.4-3). The reduction factors at 1 Hz for the southeastern Utah earthquake of September 7, 1992 were 0.65 and 1.0 for horizontal and vertical motions respectively (Figure 3.1.1.2.4.3.4-4). These reduction factors are listed in Table 3.1.1.2.4.3.4-1.

A more complete model of the reduction of amplitude with depth can be obtained by modeling the depth variation of the recorded data, and using more general source models (including earthquakes) to estimate the reduction of ground motion with depth. To test these models, downhole seismic velocity and density measurements are taken in the wells in which downhole seismograms have been recorded. Differences in seismic velocity profiles may provide an explanation of the variation from well to well that is observed in the reduction of ground motions with depth.

Topographic Effects in Ground Motion Recordings at Yucca Mountain. Strong motion data from explosions have been recorded at the surface and downhole at Yucca Mountain Sites. These recordings provide an opportunity to evaluate the influence of topographic effects on ground motions at Yucca Mountain. In particular, the availability of downhole as well as surface recordings should allow the evolution of the topographic effects to be traced. The seismograms of the explosions recorded at Yucca Mountain are dominated by surface waves. However, shear waves are also present, and the influence of topographic effects on both shear waves and surface waves can be evaluated.

Topographic conditions at Yucca Mountain have been reviewed by Phillips (1991). The Yucca Mountain Site is characterized by relatively steep topography, especially on its western flank, where there is a drop of about 180 meters over a horizontal distance of about 300 meters, giving a shape ratio of 0.6. On its much less steep eastern flank, there is a drop of about 270 meters over a horizontal distance of about 2,250 m, giving a shape ratio of 0.12. If the empirical data compiled by Geli et al. (1988) is used, the slope of the western flank is expected to produce significant topographic effects, while the slope of the eastern flank is not.

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[Using a shear wave velocity of 1.76 km/sec for the upper few hundred meters of tuff from Subramanian et al. (1990), and the estimated mountain widths of 750 m to 2,100 m from Phillips (1991), the broadband amplification is centered in the frequency range of 2.4 Hz to 0.84 Hz, following Geli et al. (1988).] Amplification is observed in this frequency range, as described above in the analysis of explosion data. However, if the topographic effect were responsible for the amplification, the amplification at sites located away from ridge crests is not expected, as is observed at station W25, as previously described. This suggests that the amplification is not related in a simple way to topographic effects.

Since the ramps reach the surface on the eastern side of the mountain, it seems unlikely that the shallow parts of the ramps will be exposed to significant topographic effects. While topographic effects may be significant at the surface on the western side of the mountain, it is unclear how they may influence the motions near the deeper parts of the ramps at depths of 200 to 300 meters.

Summary of Site Response Characteristics at the Yucca Mountain Site. Downhole recordings show the effects of amplification of ground motions as they approach the surface. Topographic effects may be present in the data. The most prominent aspect of the ground motion response at the Yucca Mountain Site is the presence of large amplitude surface waves in the seismograms of nuclear explosions and shallow earthquakes. These surface waves are absent in the seismograms of local and regional earthquakes occurring at depths greater than a few km based on review of a small data set, indicating that this ground motion response may be attributable to path effects rather than being due to effects local to the site. For this reason, it is discussed above in the section on source and path effects on strong ground motion.

To make full use of the abundant recorded data, it is important that detailed seismic velocity and density measurements be made in the site region. [In particular, downhole measurements of seismic velocities in wells, especially those that contain or have contained seismic instrumentation, are needed for the development of wave propagation and site response models using recorded data.]

3.1.1.2.4.4 Identification of Seismic Sources

Identification of faults which may be the source of significant ground motion at the potential site depends on the dimensions of the capable structures and the recurrence interval of slip on the structures. These sources are selected on the basis of both geologic and seismologic data, because the historic record is too short to encompass multiple earthquake cycles. Faults considered to have seismogenic potential which may impact Yucca Mountain are listed in Table 3.1.1.2.4.4-1. These faults generally occur within 100 km of the site. The paleoseismic data which form the basis for Table 3.1.1.2.4.4-1 are reviewed in Subsection 3.1.1.2.3.

Seismic source characterization is based on fault parameters which can be measured in the field such as fault offset, length, sense of motion, slip rates, and recurrence intervals. Relation of fault dimensions to magnitude is based on statistical correlations of fault area with

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earthquake magnitude (Wells and Coppersmith, 1993). The preferred magnitude scale for this assessment is the based on the seismic moment. This moment magnitude (M_w) scale (Hanks and Kanamori, 1979) provides a consistent estimate of earthquake size over all magnitudes. The M_w scale is readily interpretable in terms of fault dimensions because the seismic moment is proportional to the product of the average fault slip and the fault rupture area.

In Table 3.1.1.2.4.4-1, the set of faults considered to be potential sources is from a compilation of faults with known or suspected Quaternary offset by Piety, et al. (1993, in review) within 100 km of the site, a distance within which ground motions are expected to be significant.

3.1.1.2.5 Natural Resources

[The text for this section will be provided in future revisions of the Annotated Outline.]

3.1.1.2.6 Geophysics

[The results of geophysical investigations that are conducted for issue resolution and to develop a basic understanding of the site will be discussed and/or referenced in the relevant section discussing geology, seismology, and geohydrology.]

3.1.1.2.7 Geoengineering

This section describes the geoengineering properties of the site with specific focus on that part of the site where the proposed repository will be built. Geoengineering properties include the physical, mechanical, thermal, thermomechanical, and other relevant special properties of the various units of geologic material constituting the site. The state of stress at the site is also described in this section.

A knowledge of the geoengineering properties is essential for the design and construction of the various subsystems of the geologic repository system and for developing an understanding of the response of the site to the construction and existence of the repository. Such an understanding is necessary to examine and evaluate the ability of the site to meet the performance and design criteria contained in 10 CFR 60.

The primary emphasis in the description of the geoengineering properties in this section is on the repository block. Specifically, the repository block is that part of the controlled area containing the subsurface section of the Geologic Repository Operations Area (GROA) excluding the inclined accesses. Geoengineering properties of near-surface geologic material in certain areas outside the repository block are also described in this section. These are needed to design the surface facilities.

Knowledge of the geoengineering properties of the site is the outcome of the analysis and synthesis of the results of experimental observations both in the laboratory and in situ in the context of the knowledge of other properties of the site, especially geologic, geophysical, and

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geochemical. Experimental observations are, thus, the primary source of the geoengineering properties data base.

3.1.1.2.7.1 Three-Dimensional Geoengineering Model

Data on geoengineering properties are used to create a three-dimensional geoengineering model of the site. This constitutes one component of the three-dimensional geologic model. The three-dimensional model illustrates the variability, if any, of rock properties in any direction. The reference framework for collecting geoengineering/thermomechanical data is described in the two following subsections. These are the thermomechanical stratigraphy (TMS) and location map of drillholes and other data collection points. (Figure 3.1.1.2.7.1-1)

[The data base is not extensive at this time: in the early stages of the project, samples of welded and non-welded tuff from the Grouse Canyon member of the belted range tuff from G-tunnel at Rainer Mesa were used for testing for thermal and mechanical properties. A meaningful three-dimensional geoengineering model cannot be constructed, until additional samples from the repository block have been tested [INN 3.1.1.2.7-all].]

3.1.1.2.7.1.1 Stratigraphic Framework for Testing/Analysis

The stratigraphic framework of the site is described in Subsection 3.1.1.2.2, Stratigraphy and Lithology of the Site. This framework is in adherence to the current knowledge about the petrogenesis of the units of rock occurring in the region and does not consider their various properties that are of significance in the design, construction, and performance (pre- and postclosure) of the repository. These latter properties can be classified into three broad groups; hydrogeologic properties, mechanical/thermal/thermomechanical properties, and geochemical properties.

To provide a systematic basis for facilitating the characterization of the mechanical, thermal, and thermomechanical properties of the units of rock and for analyzing the response of the geologic setting to the repository, a TMS was developed. The TMS was developed by designating a lithologic unit in whole or part or a group of contiguous unit(s) and/or parts of a unit as a thermomechanical unit. The grouping is done on the basis of similarities in such characteristics of the rock as degree and presence of welding, degree and presence of vitrification; reworking, bedding, amount of lithophysae, etc. These characteristics are considered to show similarities in mechanical, thermal, and thermomechanical properties.

The TMS acknowledges that there are variations in the geoengineering properties of the rock mass at Yucca Mountain in the vertical plane and provides a systematic framework for investigating further variations, if any, of properties in the vertical and lateral directions within the thermomechanical units. The TMS is also used as a reference framework for designing the repository and for analyzing its post- and pre-closure performances. Mathematical models simulating the repository or part(s) thereof use the TMS as a reference framework for the purposes of design and performance assessment.

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The TMS is providing one component of the three-dimensional framework necessary to develop the three-dimensional thermomechanical model of the site. The TMS enables the variability, if any, of properties in any vertical plane to be modeled (or described) in a systematic manner. The other component of the three-dimensional framework is the plan map showing the areal location of drillholes or other points of data collection. Such a plan map provides the framework for modeling the variability, if any, of properties in any horizontal plane. The three dimensional thermomechanical model is one aspect of the three-dimensional geologic model.

The basic TMS is described in Table 3.1.1.2.7.1.1-1. The TMS was first formalized by Ortiz et al. (1985), who developed the concept proposed by Lappin et al. (1982). The TMS described in Table 3.1.1.2.7.1.1-1 is a minor modification of the one by Ortiz et al. (1985) in that the names and designators of some of the thermomechanical units are changed and two additional units are created so that the nomenclature and designators of the TMS are fully compatible with those of the geologic stratigraphy.

The TMS and geologic stratigraphy at the hole USW G-4 is shown in Figure 3.1.1.2.7.1.1-1. [Additional data from within and the vicinity of the repository block will be collected from boreholes and from the Exploratory Studies Facility (ESF) to construct a TMS which would be considered an acceptable representation of the repository block [INN 3.1.1.2.7.1.1-1)].]

3.1.1.2.7.1.2 Geographic Distribution of Holes

Figure 3.1.1.2.7.1-1 shows the locations of the five boreholes from which samples were collected for testing of geoengineering properties. The expected lateral boundary of the proposed repository is also shown in this figure. Locations of data collection points, other than drillholes, can also be shown in such a plan. Figure 3.1.1.2.7.1-1 describes the lateral distribution of the data points and can be the basis for illustrating the variability, if any, of the rock properties in the horizontal plane. In combination with the TMS, this figure forms the framework for creating the three-dimensional geoengineering model.

3.1.1.2.7.2 Spatial Variability and Sampling Approach

[The skeleton text for this subsection is not ready at this time.]

3.1.1.2.7.3 Mechanical Properties of Intact Rock

A knowledge of the mechanical properties of the in situ rock mass is necessary to understand and predict the response of the rock mass to the construction and existence of the repository over its required performance period(s). The in situ rock mass properties are a function of the properties of intact rock or rock matrix and the nature and extent of the discontinuities and flaws present in the rock in situ. These are the joints, fractures, faults, bedding planes, lithophysaes, vugs, inclusions, etc. Generally, these heterogeneities make the properties of the rock mass lower than that of the rock matrix. In any case, determination of the properties of intact rock is a first step toward determining the properties of the rock mass.

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Properties of rock matrix are measured in the laboratory by testing a regularly shaped test-piece, in many cases a right, circular cylinder. The process of collecting samples and preparation of test-pieces causes exclusion of gross inhomogeneities. Although micro inhomogeneities may still remain in a test piece, properties of rock measured in the laboratory on laboratory scale test pieces are considered intact rock properties. The term "intact rock property" or "property of rock matrix" is used in this sense throughout the License Application unless stated otherwise.

Subsections of this section describe the matrix mechanical properties of various units of rocks in the repository block. Both strength properties and properties relative to deformational behavior are covered. Effects of various test conditions on the measured values are described.

3.1.1.2.7.3.1 Elastic Properties

Young's modulus and Poisson's ratio are the primary mechanical deformation properties of rock and are indicators of the elastic response of the rock to stress. The elastic properties are input parameters in simulating by mathematical modeling the response of the rock to the excavation of repository rooms, emplacement holes, access openings, and other drifts.

Elastic properties have been measured in compression only and are usually assumed to be the same in tension. Since the tensile strength of rocks is a small fraction of the compressive strength, elastic properties of rocks are not usually measured in tension.

To date, measurements have been made almost exclusively on fully saturated samples at atmospheric pressure (unconfined), room temperature (23°C), and a nominal strain rate of 10^{-5}/second. [Applicability of these test results to other temperatures, pressure, saturation, and strain rate will be investigated by future testing] [INN 3.1.1.2.7.3.1-1)].

Figure 3.1.1.2.7.3.1-1 is a representative plot of the axial stress versus axial strain measured on a sample of welded devitrified tuff from the Topopah Spring member. The figure demonstrates the linearity of the deformation response up to a stress level of approximately 95 percent of the failure stress. This behavior is typical for intact samples of this material, suggesting that the rock matrix of the Topopah Spring member is an elastic material at least for the applied test conditions.

Table 3.1.1.2.7.3.1-1 gives the existing data on Young's modulus and Poisson's ratio of the various units of interest of the TMS, including the mean, standard deviation (wherever possible), and the number of tests are tabulated. [This illustrates that a limited number of tests have been completed and extensive additional testing will have to be conducted [INN 3.1.1.2.7.3.1-2].]

Effect of Test Conditions on Elastic Properties. Test conditions such as confining pressure, sample moisture content, temperature, strain rate etc., can affect the results of laboratory tests for elastic properties. A preliminary study (Nimick et al., 1985) on samples from the Topopah Spring member from hole USW G-4 could not establish any definitive relationship

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between Young's modulus and either strain rate or confining pressure for the ranges tested (0 to 10 MPa confining pressure and 10^{-3} to 10^{-7} per second of strain rate).

Price et al., 1987 performed a series of compressive tests on samples from the non-lithophysal sections of the Topopah Spring Tuff collected from the outcrop at Busted Butte. These tests were conducted at confining pressures of 0, 5 and 10 MPa and temperatures of 22°C and 150°C on both saturated and oven-dried samples at a strain rate of 10^{-5} per second. In a few of the tests strain rates of 10^{-7} and 10^{-3} per second were employed. The results of these tests shown in Figures 3.1.1.2.7.3.1-2 through 3.1.1.2.7.3.1-9, indicate that increase in either confining pressure or temperature tends to cause flattening of the stress-strain curve and hence lowering of Young's modulus. (The alpha-numeric label to each curve denotes the sample number.) Additional testing will have to be conducted to investigate the effect of temperature, confining pressure, and other conditions on the elastic properties. [INN 3.1.1.2.7.3.1-1].

Anisotropy and Elastic Properties. Rock properties often depend on the direction of measurement relative to such characteristics as bedding, schistosity and dominant orientation of layered silicate minerals. The dynamic elastic moduli for samples of the densely welded Topopah Spring member from drillhole USW GU-3 showed that anistropy of elastic properties for orientations parallel and perpendicular to the rock fabric is insignificant. (Price et al., 1984). [Additional tests will have to be performed to investigate the anistropy of elastic properties [INN 3.1.1.2.7.3.1-3].]

Static and Dynamic Elastic Properties. Young's modulus and Poisson's ratio of rock can be determined by measuring the longitudinal wave velocity and shear wave velocity. The elastic properties measured in this manner are referred to as the dynamic elastic properties. It is well known that the dynamic Young's modulus of rocks is significantly higher than if it is measured by static methods. (Jaegar and Cook, 1969). Price et al. (1984) determined the dynamic elastic properties of the Topopah Spring member by tests on 10 samples from the hole USW GU-3. The ratio of the average dynamic Young's modulus to the one measured by static methods was found to be 1.30. This is well within the range of ratios reported by Jaegar (1969) and Lama & Vutukuri (1978).

Effect of Lithophysae Content on Elastic Properties. As described in Subsection 3.1.1.2.2, the tuffs of the Topopah Spring member do contain varying amounts of lithophysae. These are generally spherical to sometimes slightly flattened cavities of up to several tens of centimeters in diameter found primarily in the Topopah Spring unit. The lithophysal cavities often have on their inner wall a thin layer of feldspars, silica minerals, and other vapor-phase mineral deposits.

Although abundance of lithophysae voids of up to 70 percent by volume in cores (hole USW G-2) has been noted, the present limited data from drillcores indicate that there are two zones of relatively high lithophysae content in Topopah Spring welded unit, lithophysae-ric (TSw1). As a matter of fact the distinction between TSw1 and Topopah Spring welded unit, lithophysae-poor (TSw2) is merely one of lithophysae content, aside from differences in

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mechanical properties. TSw2 stratigraphically starts immediately below the upper lithophysal zone. TSw2 is proposed to be the repository horizon because of its low lithophysae content.

Considering the geographical distribution of the existing boreholes (Figure 3.1.1.2.7.1-1) the possibility exists that there are zones of higher lithophysae content in TSw2. Since lithophysae are considered to affect the mechanical properties of the tuff, it is necessary to gather data about how much the properties are affected, if at all.

Price and Bauer (1985) analyzed a large number (111) of test results and established a relationship, by the method of linear least square fit, between Young's modulus(E) in GPa and functional porosity (n) in volume fraction, as given below:

$$E = 85.5\, e^{-6.96n}$$ where e is the base of natural logarithm.

The functional porosity, n, in the relationship is the total of lithophysal pore volume, non-lithophysal pore volume and volume of clay minerals, expressed as a fraction of sample volume. The measured values of n ranged from approximately 0.1 to approximately 0.4 and the above relationship can be considered valid within that range. It should be pointed out that some of the values of n in this analysis are not measured values but are estimates based on measured values of n of nearby samples. Also, while most of the tests for mechanical properties including Young's modulus were performed on samples saturated in the laboratory, some were performed on air-dried samples. The correlation coefficient for the above fit is 0.93. Figure 3.1.1.2.7.3.1-10 shows the measured and estimated data and the curve given by the above equation.

The functional porosity, especially at higher values, is a measure of lithophysae content of the Topopah Spring tuff. The above relationship, thus, indicates how Young's modulus is affected by lithophysae content. It should be pointed out that this relationship is not applicable to Topopah Spring welded unit, vitrophyre (TSw3), the vitrophyre.

Price (1983) also investigated whether any correlation exists between Poisson's ratio and functional porosity. No meaningful correlation could be established.

3.1.1.2.7.3.2 Compressive Strength

Compressive strength of rock is its ability to withstand compressive stress without failure. Compressive strength of intact rock or rock matrix is measured in the laboratory by subjecting a cylindrical test piece to a compressive load parallel to its axis until it fails. Compressive strength is the maximum stress at failure and is computed from the maximum load and the cross-sectional area of the test piece. Compressive strength measured this way is often referred to as the unconfined compressive strength as against the confined (or triaxial) compressive strength which is determined by subjecting the cylindrical test piece to a uniform lateral confining pressure in addition to the axial load. Unless specified otherwise, the term compressive strength in this document means the unconfined compressive strength.

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It should be pointed out that unless otherwise stated in specific cases, all compressive tests reported herein were performed on fully saturated samples. The effect of saturating the samples is discussed in the section on Water Saturation Level (Moisture Content) below.

Compressive strength of a wide range of tuff samples from the Yucca Mountain area have been measured and reported by a number of workers. (Olsson and Jones, 1980; Olsson, 1982; Price and Jones, 1982; Price and Nimick, 1982; Price et al., 1982a,b; Price, 1983; Price et al., 1984; Nimick et al., 1985). Compressive strengths of the various members of the Paintbrush tuff are given in Table 3.1.1.2.7.3.1-1. Mean value, standard deviation, and number of tests are given for each unit at various hole locations. [The number of tests performed so far is limited; extensive additional testing will be carried out to understand the variations, if any, both laterally and vertically.] [INN 3.1.1.2.7.3.2-1]

Effect of Test Conditions on Compressive Strength. Results of compressive strength tests in the laboratory are known to be sensitive to a number of test conditions. (Obert et al., 1946; Stagg and Zienkiewicz, 1968). These are size and shape of samples, loading rate, temperature, loading direction and sample orientation, and moisture content (saturation) of sample. Each of these conditions is considered in the following subsections.

Sample Size and Shape. It has been long known that the compressive strength of rocks measured in the laboratory decreases as the size of the test piece increases. It is thought that inherent microscopic cracks critically oriented with respect to the applied stress during testing play a significant role in initiating the failure process. Since any piece of rock contains some amount of microscopic cracks the probability of one or more cracks being critically oriented during testing increases rapidly as the size of the test piece increases. This results in the applied failure stress being lower for larger sized samples. This phenomenon is true for testing performed on large volume (several cubic meters) of rock in the field.

The other factor affecting the measured compressive strength of rocks is the length to diameter ratio of the test piece. It has been observed that compressive strength decreases as the length to diameter ratio increases (Obert et al., 1946). This ratio is sometimes referred to as the aspect ratio or the slenderness ratio. Friction between the steel loading platens and the two ends of the rock sample is thought to cause this phenomena. As the slenderness ratio decreases, the restraining forces due to friction at the two ends of the sample become more and more successful in delaying failure.

Obert et al. (1946) came up with an empirical relationship which enabled the compressive strength of samples of slenderness ratio between 2 and 1/3 to be normalized to a strength value of an equivalent 1:1 length : diameter ratio sample. This idea of normalizing test results (to enable valid comparison) was later refined to develop a standardized test procedure for a slenderness ratio 2:1 (ASTM, 1986).

Price (1986) performed a series of unconfined compressive tests at a strain rate of 10^{-5}/second on saturated samples of Topopah Spring tuff from outcrops at Busted Butte near Yucca Mountain. The sample diameters varied from 25.4mm to 228.6mm and the length : diameter

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ratios were kept constant at approximately 2:1. On the basis of these results, summarized in Table 3.1.1.2.7.3.2-1 and plotted in Figure 3.1.1.2.7.3.2-1. Price developed the following relationship between compressive strength and sample diameter:

$$\delta_u = 1944.D^{-0.846} + 69.5$$

where δ_u is the compressive strength in MPa and D is the sample diameter in millimeters. These test results, plotted in Figure 3.1.1.2.7.3.2-1, appear to indicate that compressive strength reaches an asymptotic value as the sample size increases. This asymptote can be considered analogous to the compressive strength of rock mass in situ, which is a useful rock property in the design of subsurface openings. Extensive additional testing on samples from TSw2 under as close to the in situ moisture content (saturation) as possible will have to be performed before such an asymptotic compressive strength value can be utilized in drift design.

Water Saturation Level (Moisture Content). The moisture content of the sample can have a significant effect on the measured compressive strength of rocks. (ASTM, 1986). Olsson and Jones (1980) performed a series of drained unconfined compressive tests on 18 closely spaced samples of the Grouse Canyon member of the Belted Range tuff from the G-Tunnel. Nine of the samples were oven-dried before testing and nine were tested under fully saturated conditions. Three samples each were tested under three different loading rates for either group. The results are given in Table 3.1.1.2.7.3.2-2 and are plotted in Figure 3.1.1.2.7.3.2-2.

The data in Table 3.1.1.2.7.3.2-2 show that saturating the sample causes the compressive strength to be lowered by an average of 30 percent from that of oven dry samples. The scatter in the test results also increased because of saturation by a factor of 3 to 5, as measured by the percent standard deviation. The measured value of Young's modulus was also lower by an average of 10 percent in the case of saturated samples.

Since the Grouse Canyon member is known to be free from lithophysae, the above effects of saturation on the test results for the Topopah Spring member (known to have lithophysae) can be even more severe.

Price and Jones (1982) report performing unconfined compressive tests on four closely spaced samples of the tuffaceous beds of Calico Hills from hole USW G-1 under comparable test conditions. Two of these samples were saturated and the other two were air dried at room temperature. The average strength of the saturated samples was 18 percent lower.

Although the proposed repository horizon of TSw2 is unsaturated, almost all the tests to date have been performed on fully saturated samples. Early tests were done on saturated samples because originally the plans were to have the repository in the saturated zone below the water table. Also, fully saturated samples provide one standard, definite boundary condition of testing. [Nonetheless, it is necessary to perform tests under as close to in situ saturation conditions as possible. A sufficient number of tests on room dry samples will be performed [INN 3.1.1.2.7.3.2-2].]

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Loading Rate. The measured strength of rocks depends on the loading rate. Faster loading results in higher value of the measured strength. Loading rate is typically specified by the rate at which strain is produced or the strain rate.

The data from five series of experiments on site-specific tuffs (Price and Jones, 1982; Price et al., 1982b; Nimick et al., 1985; Nimick and Schwartz, 1987) and from two series on Rainier Mesa tuffs (Olsson and Jones, 1980) are listed in Table 3.1.1.2.7.3.2-3. The test series show average strength decreases of 3 to 14 percent per order-of-magnitude decrease in the strain rate. The sequence of experiments on the Topopah Spring Member reported by Price et al. (1982b) showed virtually no strain rate effect on strength, but the three other test series on this material showed decreases of 5 to 14 percent per order-of-magnitude decrease in the strain rate (Nimick and Schwartz, 1987). The results showing no rate dependence may reflect the physical and mineralogical variability of the samples tested. The core used by Price et al. (1982b) was from an interval ranging from 371.3 to 390.0 m in depth (drillhole USW G-1) and, therefore, probably had a range of physical and mineralogical characteristics.

Martin et al. (1993) reports performing a series of constant strain rate unconfined compressive tests on saturated samples of the Topopah Spring member collected from the outcrops at Busted Butte near Yucca Mountain. Twenty samples were loaded to failure at strain rates ranging from 10^{-9}/second to 10^{-3}/second under ambient pressure and temperature. The results show the compressive strength increased as the strain rate increased from 10^{-9}/second to 10^{-5}/second. However, for the highest strain rate of 10^{-3}/second there was a decrease in strength compared to that for the 10^{-5}/second rate.

The strain rate is not of much significance in the performance of the repository during the postclosure period. It is of some significance with respect to preclosure performance, especially as it may relate to creep (time dependent) deformation at elevated temperatures. For ambient temperature and pressure conditions, a standard strain rate needs to be employed for all tests. A strain rate of 10^{-5}/second is considered appropriate.

Temperature. The strength of most engineering materials (metals, plastics, concrete, rocks) decreases with increasing temperature. Experimental data about the effect of elevated temperature on the strength of tuff are limited (Olsson and Jones, 1980; Olsson, 1982; Board et al., 1987; Price et al., 1987). As mentioned in Subsection 3.1.1.2.7.3.1, Price et al., 1987, performed compressive tests on samples of Topopah Spring Tuff from Busted Butte at 22°C and 150°C. The results shown in Figures 3.1.1.2.7.3.1-8 and 3.1.1.2.7.3.1-9 show some lowering of strength because of temperature. Sufficient tests will have to be performed to determine this effect [INN 3.1.1.2.7.3.2-3].]

Loading Direction Relative to Rock Fabric Orientation. Measured strength of rock is often affected by the direction of loading relative to such characteristics as bedding planes, schistosity, and preferred orientation of silicate minerals. Generally, the strength is higher if the direction of loading is perpendicular to such preferred planes of orientation. This characteristic, referred to as strength anisotropy, should be considered in the design of underground openings in rock. The tuffs at Yucca Mountain do have a preferred rock fabric

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orientation which is roughly horizontal. Thus the axis of core from vertical drillholes is approximately perpendicular to the rock fabric orientation. Price et al. (1984) reports testing five samples of the Topopah Spring member from the hole USW GU-3 in compression with the loading direction parallel to the rock fabric orientation. Samples adjoining four of these five samples were tested with the loading direction perpendicular to the rock fabric orientation. The average strength value of the four samples parallel to rock fabric are approximately ten percent less than the average strength of the six samples perpendicular to rock fabric. This would indicate that the Topopah Spring member may have some strength anisotropy. [Sufficient additional tests will be performed to evaluate the validity of this initial finding [INN 3.1.2.7.3.2-4].]

Effect of Lithophysae Content on Compressive Strength. The Topopah Spring tuffs at Yucca Mountain contain varying amounts of lithophysae. Based on drillcore data, two separate zones of relatively high lithophysae content have been identified in the Topopah Spring Member. The unit TSw2 is expected to be relatively lithophysae-free.

Price and Bauer (1985) reports performing compressive tests on 10 samples of lithophysae-rich Topopah Spring member collected from the outcrops at Busted Butte near Yucca Mountain. These samples were what is generally referred to as high lithophysae content, having measured porosities between 31 to 40 percent. The samples were submerged in a water bath in vacuum for a length of time needed to make them saturated. The tests were performed at a strain rate of 10^{-5}/second. The mean strength (16.2 ± 5 MPa) of these 10 samples was significantly lower (in some cases by an order of magnitude) than that of lithophysae-poor TSw1 and TSw2 reported in Table 3.1.1.2.7.3.1-1. Although the data in Table 3.1.1.2.7.3.1-1 are from tests on samples from cores and hence not quite comparable, lithophysae will have, not unexpectedly, a lowering effect on the strength of the Topopah Spring member.

[It may be necessary to establish the relationship between lithophysae content and the strength of Topopah Spring tuff. This will be needed if the ESF exposures show that there are zones of relatively high lithophysae content even in TSw2 [INN 3.1.1.2.7.3.2-5]. Sufficient tests will then have to be performed to evaluate the effect of lithophysae content on mechanical properties [INN 3.1.1.2.7.3.2-6].]

Functional Porosity and Compressive Strength. Price and Bauer (1985) analyzed a large number of compressive test results and came up with the relationship between the unconfined compressive strength (σ_u) in MPa and functional porosity (n) in volume fraction, as given below:

$$\sigma_u = 4.04\, n^{-1.85}$$

The correlation coefficient for the above fit is 0.93. Functional porosity is defined in the same way as in the Section on the Effect of Lithophysae Content on Elastic Properties. The range of functional porosity for which the above relationship is valid is approximately 0.1 to approximately 0.4. The above relationship is not valid for TSw3, the vitrophyre. As demonstrated by the test results mentioned in the subsection above on the effect of

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lithophysae on compressive strength, it may not be valid for higher values of functional porosities due largely to lithophysae. The relationship may not be valid for the densely welded Tiva Canyon member of the Paintbrush tuff, because the data on the basis of which it is derived included only one test result from the Tiva Canyon member.

For a relationship like the above to be useful, additional testing must be done so that the effects of lithophysal porosity and non-lithophysal porosity can be separated.

3.1.1.2.7.3.3 Tensile Strength

The tensile strength of Yucca Mountain tuffs were measured by the Brazilian method by Blacic et al. (1982) using 21 sets of saturated samples from five different lithologic units. The results are given in Table 3.1.1.2.7.3.3-1.

The tensile strength and porosity data in Table 3.1.1.2.7.3.3-1 were used by Price (1983) to develop the following relationship between tensile strength, T (MPa) and porosity, n (expressed as percent fraction of volume):

$$T = 27.2 - n \times 0.837$$

The correlation coefficient for the above fit is 0.89. The range of porosities for which the above relation is considered to be valid is approximately 9 percent to 30 percent. Data for the Vitrophyre (TSw3) of the Topopah Spring member were not used in this correlation.

Knowledge of the tensile strength of rock is needed in stress measurement by hydrofracturing. Tensile strength is also useful in defining the failure envelope of intact rock. [Sufficient testing will be performed by the Brazilian method to have a reliable estimate of the tensile strength of TSw2 [INN 3.1.1.2.7.3.3-1].]

3.1.1.2.7.3.4 Shear Strength of Intact Rock and Triaxial Test

Mathematical simulation of the response of rock to human activities such as mining and drilling requires the use of failure criterion for the rock. The commonly used criterion is the Coulomb criterion (sometimes referred to as the Mohr-Coulomb criterion), which defines the limiting state of stress for static equilibrium within the material at which inelastic deformation begins (Jaeger and Cook, 1979). The criterion itself is expressed as follows:

$$\tau = C_0 + \sigma \tan \phi$$

where

τ = shear stress on the failure plane at the onset of failure
σ = normal stress on the failure plane at the onset of failure
C_0 = cohesion
ϕ = angle of internal friction

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The shear stress at the onset of failure is also the shear strength of rock which is defined by its two components, cohesion (C_0) and angle of friction (ϕ). Results of unconfined compressive tests and triaxial (confined) compressive tests can be used to determine τ and ϕ.

Twenty series of triaxial compression tests on 90 tuff samples have been performed (Olsson and Jones, 1980; Olsson, 1982; Price and Jones, 1982; Price et al., 1982b; Morrow and Byerlee, 1984; Nimick et al., 1985). The experimental results for the test series in which all samples were obtained from a single location or depth were fit by using linear least squares to obtain the Coulomb parameters (cohesion and angle of internal friction) as listed in Table 3.1.1.2.7.3.4-1. The correlation coefficients for the fits also are given in Table 3.1.1.2.7.3.4-1. For nine of the test series, the coefficients are less than 0.8, suggesting that the correlation between compressive strength and confining pressure is not significant. The discussion that follows is limited to the test series for which correlation coefficients greater than 0.8 were obtained.

Four of the test series were conducted on room-dry samples. Three of these four were at ambient temperature, two were on welded tuffs, and one was on nonwelded tuff (tuffaceous beds of Calico Hills). The friction angles in the two welded tuffs are similar (68.0 and 66.7°), whereas the cohesion of tuff from the Topopah Spring Member is lower than that of the Tiva Canyon Member. Both the cohesion and the friction angle are lower for the nonwelded tuff than for the welded tuffs. The one test series conducted on room-dry samples at 200°C used moderately welded tuff from the Bullfrog Member. The test results cannot be directly compared with those for ambient-temperature tests because of differences in both porosity and temperature.

The seventh test series, for which a relatively high correlation coefficient was obtained, was conducted at 200°C. Comparison with the result from ambient temperature tests on similar material (Bullfrog Member) suggests that the friction angle decreases with increasing temperature but that cohesion is unaffected.

The observations in the preceding paragraphs are inconclusive because they are based on limited data. Mechanical tests on tuff samples have shown large variability at any single set of test conditions, especially for compressive strength. Thus, the trends inferred from the limited available data may be either spurious or real. [More definitive conclusions cannot be made until more data become available [INN 3.1.1.2.7.3.4-1].]

In general both the cohesion and the friction angle are inversely related to sample porosity (Price, 1983). Relationships between the Mohr-Coulomb parameters of angle of internal friction (ϕ) and cohesion (C_0), and functional porosity (n) have been derived (Nimick and Schwartz, 1987):

$$\phi = \text{Sin}^{-1} \left[(0.079n^{-1.856})/(2 + 0.079n^{-1.856})\right]$$

$$\text{and} \quad C_0 = 51.139 \tan \phi$$

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where n is a volume fraction, ϕ is in°, and C_0 is in MPa. The functional porosity in the above relationship is the total porosity due to both lithophysal and non-lithophysal voids as well as the volume occupied by clay minerals. These two relationships are not in themselves least-squares fits, therefore do not have associated correlation coefficients. Data from approximately 122 tests have been used in the derivations. The range of values of the functional porosity for which the above relationships is valid in approximately 0.10 to approximately 0.40.

3.1.1.2.7.3.5 Time-Dependent (Creep) Properties of Rock

If a solid is subjected to a load (stress) within its elastic limit, it instantaneously experiences an amount of deformation (strain) which disappears on the removal of the load (stress). If the load (stress) is maintained at the same level, the solid will continue to deform beyond the instantaneous deformation at a slow rate depending on the level of the applied stress. This continuing deformation with time in spite of no increase in stress is referred to as time-dependent deformation or creep deformation.

Many rocks are susceptible to creep deformation. The rate of creep depends on the type of rock, the level of stress, and other factors such as temperature and confining pressure. At low stress levels most rocks do not experience creep strain. On the other hand, rock salt which is highly prone to creep may experience measurable creep strain even at a low stress level. Elevated temperature is considered to aid creep deformation of rock.

Because of the retrievability requirement of the regulations most of the subsurface sections of the GROA will have to be accessible until the end of the preclosure period which may be as long as 80 years or longer. The areas receiving waste early will be subjected to the heat of the emplaced waste for much of the preclosure period. For these reasons the creep behavior of TSw2, the host rock, needs to be investigated.

Limited preliminary results suggest that creep deformation is unlikely to cause any insurmountable problem. [Sufficient additional testing both in the laboratory and in the ESF will have to be performed to evaluate the response of the host rock in this regard [INN 3.1.1.2.7.3.5-1].]

3.1.1.2.7.4 Mechanical Properties of Discontinuities

Discontinuities such as joints, bedding planes, faults, and fractures cause the properties of the rock mass in situ to be different from those of the intact rock or rock matrix tested in the laboratory. In general, strength and deformational properties of rock mass will be lower than those measured in the laboratory. An understanding of the mechanical properties of discontinuities is necessary to evaluate their effect on rock mass properties.

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Properties of discontinuities, their attitudes, frequency and distribution are important parameters used in computer-aided mathematical simulation of the response of the rock mass to the creation and existence of subsurface openings. They are also taken into account in the design of ground support systems.

This section and its subsections address the mechanical properties of joints, beddings, fractures, and faults. These features, their nature, origin, abundance, frequency, attitude, etc., are described broadly for the site region in Subsection 3.1.1.2.4.

3.1.1.2.7.4.1 Properties of Natural vs. Simulated Discontinuities

The natural joints present in the tuffs at Yucca Mountain can be divided into three groups: healed joints with infilling minerals; unhealed joints with no infilling; and unhealed joints with infilling minerals. Frictional properties of the joints with infilling will depend on the nature of the infilling minerals.

The condition of cores from Yucca Mountain with unhealed joints has been unsuitable for representative laboratory mechanical tests. It is expected that joints or faults containing gouge or other infilling may have a lower coefficient of friction than clean, unfilled joints (Byerlee, 1978). The coefficient of friction for typical saturated gouges generally ranges from 0.2 to 0.6 (Morrow et al., 1982), whereas the coefficient of friction for dry clay gouges has a range of 0.2 to 0.7 (Shimamoto and Logan, 1981). Lower values for the coefficient of friction will result in lower shear strengths for the joints, for the same values of cohesion and applied normal stress.

Teufel (1981) determined the shear strength in triaxial compression of a simulated joint in the welded Grouse Canyon Member as a function of normal stress, time of stationary contact, displacement rate, and saturation. Joints were simulated by using a right-circular cylinder with a precut inclined at 35° to the cylinder (load) axis. Room temperature tests were conducted at confining pressures from 5 to 40 MPa, at axial displacement rates from 10^{-2} to 10^{6} cm/second, with both dry and fully saturated samples.

The shear strength of a simulated joint in welded tuff fits the linear relation:

$$\tau = C_0 + \sigma \tan \phi$$

where τ is the shear strength, C_0 is cohesion, σ is the applied normal stress, and ϕ is friction angle of friction.

The coefficient of friction ($\tan \phi$) at the initiation of slip was found to be independent of the normal stress for air-dried samples, with a value of 0.64 at a displacement rate of 1.2×10^{-4} cm/second. Similar results ($\tan \phi = 0.59$) were obtained for air-dried samples of partially welded tuff (Prow Pass Member) at a displacement rate of 10^{-3} cm/second (Olsson and Jones, 1980). Data provided by Morrow and Byerlee (1984) can be used to derive a coefficient of friction of 0.59 for the initiation of slip in saturated samples of the Topopah Spring Member

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at a strain rate of 10^{-4}/second (equal to a displacement rate on the fracture surface of approximately 7 x 10^{-4} cm/second). The independence of the coefficient of friction with respect to the confining pressure and the corresponding normal stress across the sliding surface is consistent with rock-friction literature as reviewed by Byerlee (1978). However, at low normal stresses the coefficient of friction of rough natural joints may have some dependence on normal stress. [Adequate testing will have to be performed to investigate this [INN 3.1.1.2.7.4.1-1].]

Data presented by Morrow and Byerlee (1984) suggest that the coefficient of friction increases with progressive shear displacement along a joint. For saturated samples of the Topopah Spring Member at a strain rate of 10^{-4}/second, the increase would be from 0.59 to 0.76.

3.1.1.2.7.4.2 Effect of Infilling on Joint Properties

[This Subsection will describe the effect, if any, of infilling minerals on the mechanical properties of joints [INN 3.1.1.2.7.4.2-1].]

3.1.1.2.7.4.3 Properties of Healed and Unhealed Joints

[This Subsection will describe how the process of healing affects the properties of joints [INN 3.1.1.2.7.4.3-1].]

3.1.1.2.7.4.4 Effect of Test Conditions on Joint Properties

[This Section and subsections thereof will describe test results to evaluate how test conditions such as state of saturation, level of applied stress, etc., affect, if at all, the measured properties of joints [INN 3.1.1.2.7.4.3-1].]

3.1.1.2.7.4.5 Time-Dependent (Short-Term) Behavior of Joints

Two categories of time-dependent behavior of joints need to be considered: the long-term and the short-term. The long-term, time-dependent behavior is relevant in terms of the long-term behavior of the rock mass around the repository, especially at an elevated temperature. Such long-term behavior is addressed in Subsection 3.1.1.2.7.4.6.

This section deals with short-term time-dependent behavior as it relates to rate of shear displacement during testing, sliding velocity, the stick-slip phenomenon, etc.

Short-term, time-dependent effects on joint strength have been addressed in two ways. Constant displacement-rate tests have provided insight into the effect of changes in sliding velocity on the sliding coefficient of friction. Hold times in these tests provided a preliminary evaluation of the increase in the static coefficient of friction of the joint as a function of time. The data were taken to attempt to quantify the joint strength increases that occur with time for use in evaluating the maximum in situ joint strength.

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To determine the effect of sliding velocity on the coefficient of friction of a precut joint, a series of room-temperature tests were conducted at a confining pressure of 10 MPa and axial displacement rates from 10^{-2} to 10^{-6} cm/second (Teufel, 1981). As shown in Figure 3.1.1.2.7.4.5-1 the coefficient of friction for oven-dry samples increased from 0.62 at 10^{-2} cm/second to 0.66 at 10^{-6} cm/second, a 6 percent increase in the coefficient of friction over four orders of magnitude decrease in displacement rate. These results are consistent with the work of Dieterich (1978) and Teufel and Logan (1978) for granites and sandstones, respectively. For water-saturated joints, the displacement rate effect is slightly higher, although at a 9 percent increase in the coefficient of friction over four orders of magnitude decrease in displacement rate, the effect is still small. Over the observed range of displacement rates the coefficient of friction for water-saturated precuts is slightly greater than that for dry precuts (Teufel, 1981).

To evaluate time-dependent joint strength increases, the time dependence of the frictional shear strength of oven-dried and water-saturated Grouse Canyon Member welded tuff was investigated in triaxial compression by examining the response of 35° precuts (Teufel, 1981). A confining pressure of 10 MPa was used. The test procedure was slightly different from that used in the previous quasi-static tests. In the tests of time-dependent behavior, axial load was increased until slip occurred along the precut at a constant sliding velocity of 1.2×10^{-3} cm/s. The test was stopped for a given time under load, then was started again at the same sliding velocity. This procedure was repeated for several different durations of contact.

The results of the test resembled the stick-slip phenomenon observed in many other rock types (Paterson, 1978). A plot of the change in shear stress as a function of displacement (Figure 3.1.1.2.7.4.5-2) shows that when a test was stopped for 60 seconds then started again, there was an abrupt increase of approximately 0.4 MPa in the shear stress necessary for slip; then there was a drop back to the former stress level as slip continued at the previous sliding velocity of 1.2×10^{-3} cm/second. With an increase in the time of stationary contact (2,400 s), both the stress rise required to initiate slip and the corresponding stress drop were larger. However, the stress required for stable sliding did not change significantly. A plot of the static coefficient of friction at peak stress versus the logarithm of the time of contact for oven-dried samples (Figure 3.1.1.2.7.4.5-3) shows that as the time of contact increased, there was a consistent increase in the friction resisting the initiation of sliding. During these tests, the displacement rate was 1.2×10^{-3} cm/s. The data in Figure 3.1.1.2.7.4.5-3 are considered a lower bound on the static coefficient of friction. Extrapolation of these data to much longer times is not warranted, if only because of the variability of fracture surfaces in situ.

As also shown in Figure 3.1.1.2.7.4.5-3, the time-dependent increase in the coefficient of friction is enhanced with water saturation. The increased time dependence of the frictional resistance of the water-saturated tuff is attributed (Teufel, 1981) to hydrolytic weakening and time-dependent stress corrosion of asperity contacts on the sliding surface.

[Sufficient additional testing will be performed to develop an understanding of the short-term, time-dependent behavior of joints [INN 3.1.1.2.7.4.5-1].]

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3.1.1.2.7.4.6 Time-Dependent (Creep) Properties of Joints

[This section will describe the long-term, time-dependent behavior of joints, especially at the elevated temperatures expected to be prevailing during the preclosure period. Such knowledge will be useful in assessing the long-term stability of the repository openings and in ensuring that operating plans for retrieving emplaced waste are workable.]

[No data exists on the creep properties of joints at elevated temperatures. Sufficient data will be collected, especially in the thermomechanical tests in the ESF in this regard [INN 3.1.1.2.7.4.6-1].]

3.1.1.2.7.5 Nature and Extent of Joints at the Yucca Mountain Site

The nature and extent of the joints, fractures, faults, etc., in the site region as well as at the site (repository block) are described in Subsection 3.1.1.2.4.3. In this section, fracture data relative to the repository block is analyzed to assist in assessing the impact of the discontinuities on the response of the rock mass to the construction and long-term performance of the repository. The data presented in Subsection 3.1.1.2.4.3 are analyzed and summarized here with specific emphasis on those characteristics relevant to repository performance in the preclosure period.

As can be seen in Subsection 3.1.1.2.4.3, there is a fair amount of data on the fractures in the site region. However, with respect to fracture data within the repository block, especially the subsurface category, the source is limited to observations on the cores from the drillholes in the block shown in Figure 3.1.1.2.7.1-1. Lin et al. (1993) performed a comprehensive analysis of the fracture data from cores, USW G-1, USW G-4, USW GU-3, and UE-25a#1 and came up with some summaries as described below.

<u>Fracture Attitude</u>

The attitude of a fracture as described by its strike and dip directions and dip magnitude is an important parameter in assessing its impact. Complete and reliable data about fracture attitude cannot be obtained from cores alone unless the cores were oriented. Oriented cores provide the needed information about the orientation at which the core was in situ before being removed from the rock mass. Borehole televiewer logs enable one to estimate the strike and dip directions; however, the reliability of such estimates drops in the case of near vertical fractures and nominally vertical holes.

Limited amounts of oriented cores were collected from holes USW GU-3 and USW G-4 at selected 3-meter intervals within each geologic member, and fracture attitude information was recorded. Continuous sampling was performed using borehole television which measured the fracture strike only. Results, reported in Spengler and Chornack (1984) and Scott and Castellanos (1984), are in the form of stereonets and strike histograms for oriented core and borehole television data respectively.

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On the basis of these results from hole USW GU-3, Lin et al. (1993) suggest that in the Tiva Canyon member there may be two concentrations of joint orientations: a broad trend striking N30°W to due north with near vertical dips in both the northeast and southwest directions, and a more concentrated trend striking roughly N50° W with dips of 12° NE. These orientations are present in the strike rosette developed from borehole television observations of 133 fractures, although they are not the dominant orientation. The borehole television data indicate a dominant trend between N18°W and N36°E (dip not recorded).

With respect to the Topopah Spring Member in USW GU-3, Lin et al. (1993) found a trend similar to that of the Tiva Canyon Member: a concentration striking N10° W and dipping 75° to 90° NE and SW and a concentration with strike trending N25° E and dipping 10° SE. A concentration striking N45° E and dipping 80° to 90° NW and SE is also there. Borehole television data extended only 10 meters into the Topopah Spring Member and meaningful data were not obtained.

Data from USW G-4 showed that in the Tiva Canyon Member there is a concentration striking N22° E with dips of 65° to 90° NW and a second one striking N50° W to east-west with high angle dips. The strike data recorded by borehole television indicate a relatively uniform distribution between N45° W and N60° E with some concentration due north.

In the hole USW G-4 in the Topopah Spring Member, a concentration of north striking joints with high angle dip is observed. The strike data based on bore television observation indicate a strike distribution between N15° W and N60° E, with local concentration at N40° E and due north.

The trends in fracture orientation described above are summarized in Table 3.1.1.2.7.5-1. [The interpretations are based on very limited data and must be considered preliminary.]

Fracture Frequency

The frequency of fractures per linear meter of each hole, reported by Lin et al. (1993), is summarized in Table 3.1.1.2.7.5-2 for the different thermomechanical units.

Using the fracture data from the four boreholes, Lin et al. (1993) estimated the fracture frequency per unit volume of rock. Their results are given in Table 3.1.1.2.7.5-3. It should be pointed out that the procedure by which Lin et al. (1993) extrapolated the data from cores from near vertical holes to fracture frequency per unit volume is known to result in overestimates if the joints are steeply dipping. Lin et al. (1993) recognized this possibility and attempted to verify the procedural correction factor by a statistical analysis technique. Nonetheless, there is uncertainty about the results given in Table 3.1.1.2.7.5-3 because of the limited nature of the basic data; the possibility exists that volumetric fracture frequencies for some of the thermomechanical units are not as high as in the table.

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3.1.1.2.7.6 Mechanical Properties of Rock Mass

[Subsections 3.1.1.2.7.6 through 3.1.1.2.7.15 will be added to future revisions of the Annotated Outline.]

3.1.1.2.7.6.1 Rock Mass Rating (RMR) System and Q Rating

3.1.1.2.7.6.2 Rock Mass Time-Dependent Properties at Elevated Temperature

3.1.1.2.7.7 Thermal Properties of Intact Rock

3.1.1.2.7.7.1 Thermal Conductivity

3.1.1.2.7.7.2 Heat Capacity of Intact Rock

3.1.1.2.7.7.3 Thermal Expansion

3.1.1.2.7.8 Thermal Properties of Rock Mass

3.1.1.2.7.8.1 Rock Mass Thermal Properties Measured In Situ

3.1.1.2.7.8.2 Rock Mass Thermal Properties vs. Intact Rock Thermal Properties

3.1.1.2.7.8.3 Rock Mass Thermal Properties During Cooling Phase

3.1.1.2.7.9 Existing Stress Field

3.1.1.2.7.10 Special Engineering Properties

3.1.1.2.7.11 Engineering Properties of Surficial Material

3.1.1.2.7.12 Excavation Characteristics of Rock Mass

3.1.1.2.7.13 Estimated Water Inflow

3.1.1.2.7.14 Ground Support Requirements

3.1.1.2.7.15 Overview of Mechanical, Thermal, and Thermomechanical Models Using Geoengineering Properties

3.1.1.3 Future Variation in Geologic Process

[Future changes that might be expected to occur in the geologic system are assessed in this Subsection. This assessment includes the potential for earthquakes on faults in the geologic system; the potential for new and extended faulting and fracturing in currently unfaulted and

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unfractured areas, including location and nature of ruptures, and the potential for volcanism, including location, extent, and proximity to the site. Also described are features of interest, how analogs were used; the time frame of the comparative study; and any differences in scale or features between the analog and the site.]

Potential for Earthquake on any Faults. [Describe the potential for earthquake on the faults identified within 10 km of the repository.]

Potential for New and Extended Faulting. [The movement potential of faults in the vicinity of Yucca Mountain is being studied. The results will be provided when available.]

Potential for Volcanism. [Studies are to be performed on the potential for volcanic eruption in the vicinity of the Yucca Mountain Site and will be described here when completed.]

3.1.2 Hydrologic System

A description of the hydrology of the site and the surrounding area is presented in this Subsection. The surface water, groundwater (both saturated and unsaturated flow), the interrelationship between the surface and groundwater flow, flooding potential, site drainage, and other hydrologic phenomena are described. The purpose of this Subsection is to discuss whether the areas studied provide a representative description of the conditions throughout the region, around the site, and at the site.

3.1.2.1 Surface Water Hydrology

Knowledge of the site surface-water hydrology is important because surface-water is a major source of infiltration to the unsaturated zone that could affect radionuclide transport from the site to the accessible environment. In addition, future surface-water flooding may pose a hazard to repository facilities and could effect waste isolation.

This subsection contains a description of the surface-water hydrology of the region and the site. This description includes:

a. Surface-water bodies and physical characteristics of drainage areas;

b. Surface-water monitoring network;

c. Water control structures and diversions;

d. Flood history;

e. Flood potential;

f. Chemical composition of identified bodies of surface water;

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g. Location, quantity, and quality of surface water extracted; and

h. Projected surface-water uses.

3.1.2.1.1 Description of Surface-Water Bodies and Physical Characteristics of Drainage Areas

Characterization of surface hydrology is required to provide baseline data necessary to resolve several issues. Knowledge gained through the surface hydrology studies can be used to evaluate the potential for erosion in the site area and any hazards that this may cause; evaluate the potential for floods and the potential hazards to the surface facilities due to floods; and help evaluate the amount of infiltration and establish the relationship between surface runoff and groundwater recharge that occurs in the study area.

The regional surface-water system that encompasses Yucca Mountain includes most of the Death Valley Basin Hydrographic Region of California-Nevada and a small part of the Central Hydrographic Region of Nevada (Rush, 1968). In the Death Valley Basin, valleys drain into California's Death Valley by the Amargosa River and its tributaries. In contrast, the Central Hydrographic Area consists of closed basins that contain playas. The hydrographic area shown in Figure 3.1.2.1.1-1 consists of eight smaller units as described by Waddell, et al. (1984) (Plate 1). Table 3.1.2.1.1-1 lists the approximate areas of each of these smaller units. Yucca Mountain lies on the boundary of the Crater Flat and Fortymile Canyon-Jackass Flats hydrographic areas. The boundary lines of the hydrographic area are drawn principally along topographic divides.

The YMR has a generally arid to semiarid climate, characterized by high potential evaporation (0 to 6 mm/day, DOE, 1993), low average annual precipitation (175 mm/yr, DOE, 1993), and infrequent storms. Surface runoff results from regional storms that occur most commonly in winter, and occasionally in autumn and spring and from localized thunderstorms that occur mostly during the summer.

There are no perennial streams in or near the Yucca Mountain area. Perennial surface water in the area is found only in springs, some short sections of the Amargosa River, and marshes around the edge of the salt pan in Death Valley (Hunt, et al., 1966). One small lake, locally known as Crystal Reservoir, with a storage capacity of 1.84×10^6 m^3 (Giampaoli, 1986) occurs in the Ash Meadows part of the Upper Amargosa hydrographic area (Figure 3.1.2.1.1-1). The water for the reservoir is supplied via a concrete flume from Crystal Pool (Giampaoli, 1986).

The major ephemeral stream system in the Death Valley Basin hydrographic area is the Amargosa River. This river originates in Oasis Valley, continues southeastward through the Amargosa Desert past Death Valley Junction, then southward another 75 km where it turns northwestward and terminates in Death Valley (Figure 3.1.2.1.1-1). The river carries floodwater following cloudbursts or intense storms, but is normally dry except for a few short reaches that contain water from springs (Walker and Eakin, 1963) located: in Oasis Valley

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between Springdale and Beatty (Malmberg and Eakin, 1962), Ash Meadows northeast of Death Valley Junction, and near Shoshone, about 40 km south of Death Valley Junction. Base flow to these segments of the river is maintained by groundwater discharge during the winter, when evapotranspiration is at a minimum. During the summer, discharge from the springs is almost entirely lost by evapotranspiration. During winter, groundwater discharges into the river south of Alkali Flat, near Eagle Mountain about 10 km south of Death Valley Junction (Walker and Eakin, 1963).

Analysis of the general character of streamflow was based on scattered data from peak-flow stations located around the Yucca Mountain area. Figure 3.1.2.1.1-2 shows the locations of 12 crest-stage gaging sites that were part of a historical monitoring network located in the general area of Yucca Mountain. Table 3.1.2.1.1-2 summarizes the streamflow information obtained from each of these 12 sites. Data collection at these sites was discontinued in 1980. [Stations 2-6 have been reactivated as part of on-going studies [INN 3.1.2.1.1-1], and provide surface-water monitoring data when available.]

Springs and seeps are another source of perennial surface water in the study area. Springs can be classified on the basis of their groundwater source, (i.e., water-table springs and perched springs). Water table springs discharge where the land surface intersects the water table. Perched springs discharge where the land surface intersects a local groundwater body that is separated from the main saturated zone below by a zone of relatively lower permeability and an unsaturated zone. Seeps are springs with very low discharge rates, commonly so low as to preclude measurement. Areas of spring discharge are grouped into several distinct locations within the hydrogeologic study area. The primary areas are southern Oasis Valley, Ash Meadows, Furnace Creek Wash, and an area of minor groundwater discharge north and west of Yucca Flat.

An inventory of springs, seeps, and phreatophyte areas in the hydrogeologic study area has been compiled and is presented in Table 3.1.2.1.1-3. This table is based on data from Thordarson and Robinson (1971); Malmberg and Eakin (1962); Pistrang and Kunkel (1964); Schoff and Moore (1964); Miller (1977); Winograd and Thordarson (1975); Dudley and Larson (1976); White (1979); and Waddell et al. (1984). Phreatophyte areas are locations where heavy growths of plant species that tap the water table provide considerable amounts of groundwater discharge through evapotranspiration. [INN 3.1.2.1.1-4], Update phreatophyte areas information in Table 3.1.2.1.1-3.

Playas are another area of groundwater discharge. According to Motts (1970), playas exhibit four distinguishing characteristics:

a. An area that occupies a basin or topographic valley of interior drainage;

b. A smooth, barren surface that is extremely flat and has a small topographic gradient;

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c. An area that infrequently contains water, which occurs in a region of low rainfall where evaporation exceeds precipitation; and

d. An area that is fairly large in size (generally more than 609 to 914 m in diameter).

Franklin Lake playa is one of the principal discharge areas of the Furnace Creek Ranch-Alkali Flat groundwater flow system. The playa is located between Eagle Mountain and Death Valley Junction. Estimates by Czarnecki (1990) of maximum evapotranspiration, as a residual of the energy balance equation, ranges from .36 m/a during the winter months to 1 m/a during the summer months. [Playas have not been included in Table 3.1.2.1.1-3 because of the lack of data concerning quantities of discharge and the locations of many of these areas. [INN 3.1.2.1.1-2, Provide information on the location and characteristics of playas when available].]

[Figure 3.1.2.1.1-3 shows major perennial and ephemeral surface-water bodies in the YMR including springs [INN 3.1.2.1.1-3], Figure 3.1.2.1.1-3.]

3.1.2.1.2 Surface-Water Monitoring Network

The surface-water monitoring network collects information on precipitation, surface runoff, and streamflow that may be used in flooding and infiltration studies. A stream-gauging network has been designed to characterize surface-water hydrology. Stream-gauging sites were selected to document streamflow in major, hydrologically important, and representative drainage basins at the Yucca Mountain Site.

The stream gauging network consists of primary, secondary and tertiary sites. Primary sites utilize continuous stream-stage recorders to measure runoff events. Secondary sites monitor only the peak or crest stream-stage that occurred during a given time. Tertiary sites monitor stream discharge by observation of flow through hydraulic structures and indirect measurements. Primary and secondary sites were located based on their hydrological importance, accessibility and proximity to existing stream gaging facilities. Tertiary sites were located where field reconnaissance indicates that runoff has occurred. The three site approach increases confidence that ephemeral, unpredictable, and non-uniform runoff events are adequately characterized by utilizing alternate methods of investigation.

Fourteen primary streamflow measurement sites are currently in operation. These measurement sites are equipped with continuous stream-stage recorders, crest-stage recorders, and precipitation gauges. Three of the primary sites use stilling wells to obtain continuous stream-stage measurements. Each continuously recording stream gauge located at a primary streamflow measurement site is visited monthly (or at least six times per year) for stream-stage record collection and equipment maintenance, and as soon as possible after each runoff event. Figures 3.1.2.1.2-1 and 3.1.2.1.2-2 show the locations of these stations.

Continuous streamflow rates at primary streamflow measurement sites can be derived using stage-discharge relations. Flow measurements are made after-the-fact using indirect

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measurement techniques. Indirect (after-the-fact) determinations of peak flow for any given runoff event are necessary because the unpredictable, short-lasting, and hazardous nature of streamflow in this semi-arid environment reduces the likelihood of making many direct measurements of flow. Indirect measurements use hydraulic energy equations such as the Manning equation to compute peak discharge after the flood peak has subsided rather than during the flood. The Manning equation is:

$$Q = (1.486/n)*A*R^{2/3}*S^{1/2}$$

where Q = discharge (in cubic feet per second)
n = Manning's coefficient of roughness
A = cross sectional area of flow (in square feet)
R = hydraulic radius of flow (in feet)
and S = slope of the energy gradient.

The Manning equation uses different methods for different conditions, but all methods use physical characteristics of the channel (channel dimensions and geometry); water-surface elevations at the time of the peak stage (usually defined by high-water marks left by debris in the flow__; and hydraulic factors based on physical characteristics, water-surface elevations, and discharge, such as roughness coefficients and discharge coefficients. Scour chains are one method used to measure roughness coefficients. Scour chains are installed at several primary stream gauging sites. Scour chains are vertically suspended lengths of chain buried in individual vertical pits in the alluvial streambed. Any scour of the streambed during periods of streamflow will deform the upper segment of a chain and allow determination of the maximum cross-sectional depth and shape of channel scour for a specific runoff event.

Additional continuously recording stream gauges are currently planned to be installed. These gages may be installed at new sites or converted from existing crest-stage gauge stations. The locations of these eight sites have not been determined [INN 3.1.2.1.1-1].]

Currently 11 secondary measurement sites are equipped with crest-stage gauges that record only the peak level of flow for a specific runoff event. Five of these sites have been operational since the 1960s, providing historic streamflow records. This core has been supplemented by six additional crest-stage gauges. These crest-stage gauges do not record any data on rising or receding stages of flow, the timing of flow or the duration of flow. Precipitation gauges are also located at these sites.

The crest-stage gauge network is visited periodically, generally monthly (or at least six times per year), in concert with operational visits to the continuously recording stream-stage gauges. These sites are also visited as soon as possible during or following runoff events. At the crest-gauge stations, as at the continuously-recording, stream-gauge stations, flow measurements are made after-the-fact to relate to recorded peak stages.

Peak flows are also measured, following flooding, on numerous miscellaneous drainages where no continuous streamflow recorders or crest-stage gauges are located (tertiary

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measurement sites). Stream discharge at tertiary sites is monitored by observation of flow through appropriate hydraulic structures (culverts, open channels, etc). Miscellaneous measurements are also made at these ungauged sites within drainages that have primary or secondary measurement sites. The tertiary measurement sites were selected as targets of opportunity when a field reconnaissance of runoff suggested that these normally-unmonitored drainages, or drainage components, have experienced recent storm runoff of a character that requires additional measurement to enhance the knowledge of streamflow and flooding. The data thus collected supplements and complements other streamflow data collected systematically and periodically at primary and secondary sites throughout the network. Because of the non-uniform nature of desert flooding, data collected at tertiary sites are a valuable supplement to data collected at primary and secondary sites. Peak-flow measurements at tertiary sites are also made using indirect measurement techniques.

Chemical and isotopic samples of selected streamflows are collected and used to define the geochemical character of runoff in order to assist in groundwater recharge studies. The data collected through streamflow sampling include: onsite pH, water temperature, specific conductance and concentrations of sodium, calcium, magnesium, potassium, carbonate, bicarbonate, sulfate, chloride, strontium, silica, lithium, aluminum, iron, manganese, fluoride, bromium, iodine, carbon -12, -13 and -14, oxygen -16 and -18, tritium and deuterium. During runoff events, water samples are collected manually. [INN 3.1.2.1.1-1, Provide results of surface-water monitoring including chemical analyses, streamflow measurements, and other data when available.]

Runoff and streamflow at the Yucca Mountain Site are direct responses to precipitation. Collection of additional precipitation, temperature, and snowpack data is necessary in order to obtain intensities of precipitation and other information at specific points of interest for the runoff monitoring activity. These additional data supplement and complement meteorological data collected under different studies.

Rain gauges (recording and storage gauges) are used to measure the duration, magnitude, and distribution of rainfall. Thirty-one gauges are currently operational, 14 at the existing primary measurement sites, 11 at the existing secondary sites, and 6 at precipitation recording sites. Recording gauges are of the tipping bucket or weighing bucket type; data from these devices are recorded by analog or digital devices. Storage gauges are either standard aluminum cans or plastic containers with graduated markings. These devices are monitored by observers. Snow accumulation, storage, and dissipation are monitored by observers and snow pillows. (A snow pillow consists of an inflated elastometric bladder connected to a pressure transducer in such a way that the weight of snow falling on the bladder may be determined by the pressure changes in the bladder). Continuous recording thermometers or thermographs are used to measure temperature variations at selected stream gauge sites. Temperature data are used to relate runoff and weather conditions, support precipitation-runoff modeling and support geochemistry data.

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[Eventually all crest-stage gauge sites will be converted to full sites for a total of 51 full sites that have continuous stage recorders, crest-stage recorders, and rain gauges [INN 3.1.2.1.1-1].] Table 3.1.2.1.2-1 lists the sites, and their location, purpose, status, and equipment.

3.1.2.1.3 Water Control Structures and Diversions

Knowledge of existing, and proposed man-made, surface-water structures is important because these structures may influence the potential for flooding at the site. However, the presence of man-made water control structures near the Yucca Mountain Site is unlikely to be a problem because there is not enough surface-water in this area to be considered a resource.

The Yucca Mountain Site has an arid climate and few perennial surface water bodies. Perennial surface-water bodies include short reaches of the Amargosa River and spring-fed ponds (Waddell, et al., 1984). In addition, several small, man-made reservoirs used for agricultural, milling, and mining purposes are scattered throughout the southern portion of the hydrologic study area. The largest man-made reservoir is Crystal Reservoir, a 1.84×10^6 m^3 reservoir (Giampaoli, 1986) located in Ash Meadows Wildlife Conservation Area and used for recreational purposes only.

There are no perennial streams in the Yucca Mountain vicinity. The major stream system in the region is the Amargosa River which is ephemeral except for short, spring-fed reaches (Waddell, et al., 1984). As such, sufficient supplies for domestic, municipal, commercial, agricultural, or industrial uses have not been available.

Fortymile Wash is an extensive wash located several miles east of Yucca Mountain on the border and outside of the Conceptual Perimeter Drift Boundary (CPDB). Because of the size of this wash, a detention basin or diversion structure could be developed to control flood water in this wash. Depending on the size, design, and location of such a structure, the repository design could be impacted. Potential affects include changes in recharge in the wash and flooding of surface facilities and openings. In any case, a thorough study of the impacts of such a structure would be necessary before a water control design in Fortymile Wash was implemented.

No man-made water control structures currently exist that could influence conditions at the site. Existing man-made water control structures such as the Crystal Reservoir are either too small or too distant to influence site conditions. In addition, there are no known plans to construct dams or reservoirs in the vicinity of the Yucca Mountain Site. Because of the limited supply and poor quality of existing perennial water sources in the region, the projected use of surface-water supplies is limited to preservation of wildlife habitats and recharge to the valley-fill aquifer. Further, the absence of a perennial river within the boundaries of the hydrographic study area precludes the development of hydroelectric power and further limits the potential for application of surface-water for beneficial use.

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3.1.2.1.4 Flood History

Information on the site's flood history can be used to determine the capability of the site to withstand floods and flood waves. Moderate to extensive flooding along major drainages within the study area is usually the result of regional storm systems that occur most commonly during the winter and occasionally during the spring and fall. Small areas may also discharge intense precipitation; however, intense areally-restricted flooding is usually caused by summer thunderstorms. Flash floods constitute a major hazard throughout the Great Basin and specifically in southern Nevada. These floods usually do not cumulate to cause regional floods, but their intensive character renders them potentially destructive over limited areas. Flash floods and associated debris flows are among the most important geomorphic processes currently active in the region and local area.

Data have been collected by the USGS since the early 1960s on streamflow and flooding in ephemeral stream channels throughout Nevada (see Subsection 3.1.2.1.2). The resultant data base is a reasonably reliable record of the magnitude and frequencies of peak flows at the crest-stage gauging stations during the tenure of operation of the gauges. The location of these 12 crest-stage gauging stations is shown on Figure 3.1.2.1.1-2. Table 3.1.2.1.4-1 summarizes the date, peak discharge, maximum unit peak flood discharge (peak discharge divided by upstream drainage areas), and level for each gauging site. [The stations were discontinued in 1980. Stations 2-6 on Table 3.1.2.1.4-1 were reactivated as part of ongoing site characterization studies [INN 3.1.2.1.1-1].]

Table 3.1.2.1.4-1 shows several major floods that occurred in some of the washes during the operation of the monitoring network from 1964 through 1980. The most notable flood was associated with a large winter storm that occurred over the upper Amargosa River drainage basin (drainage area, 1,217 km^2) in February 1969. This storm caused an estimated peak flow of 453 m^3/s near Beatty with a peak discharge per unit area of 0.37 $(m^3/s)/km^2$ (based on station 6). The upper Amargosa drainage basin is immediately west of Fortymile Wash basin and has a similar physiographic terrain. Also, in August 1968, a peak flow of 97.1 m^3/s with a peak discharge per unit area of 0.34 $(m^3/s)/km^2$ occurred at station 3 in an unnamed tributary to the Amargosa River near Mercury, Nevada, on the NTS. This drainage basin has an area of 284.9 km^2 and a terrain similar to that of Fortymile Wash. Subsection 3.1.4.1.1.2.2 characterizes precipitation in the YMR.

Because no significant surface-water bodies or water-control structures are located near the site, no surges, seiches, tsunamis, dam failure, or ice jams have occurred that could affect the site. In addition, no evidence for flooding in the recent past induced by nearby landslides has been reported (USGS, 1984).

According to Squires and Young (1984), "Geomorphic studies on the NTS have indicated that some of the alluvial surfaces along Fortymile Wash are thousands of years old. Such ages might imply that the surfaces have not been flooded since they formed several thousand years ago. However, distinct high-water marks are observed along Fortymile Wash in the vicinity of cross-section FM-4," indicating that the alluvial surfaces along Fortymile Wash were

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inundated. Cross-section FM-4 is located just south of the junction of Drillhole and Fortymile Wash. Survival of these alluvial surfaces may be explained by a previous observation in Colorado that a fine-grained alluvial surface overtopped by a flash flood was virtually unaffected (Squires and Young, 1984). They continue, "From these marks and from data on the cross-sectional area and channel slope, a peak flow of about 20,000 ft^3/s is estimated. Documentation of similar flooding in nearby washes indicates that this flood peak probably occurred during February 1969." [Site characterization studies are currently seeking additional evidence of Quaternary flooding. [INN 3.1.2.1.4-1], Provide evidence of Quaternary flooding.]

[Flow duration data that indicate maximum, minimum, and average historical observations of surface-water bodies is being recorded at 14 continuous, stream-stage recording stations; however, data from these stations are not currently available. [INN 3.1.2.1.1-1, provide results of surface-water monitoring].]

3.1.2.1.5 Flood Potential

Long-range flood predictions are necessary to estimate the location, magnitude, and potential consequences of future flooding of the site. Long-range flood predictions are difficult to make, particularly in arid regions because of the limited streamflow data available. Current flood-prediction methods for this area generally involve some form of statistical evaluation of available regional streamflow data.

Two detailed studies of flood-prone areas in the vicinity of Yucca Mountain provide a basis for estimating the magnitudes of floods with various recurrence intervals. Squires and Young (1984) analyzed the flood plain of Fortymile Wash and its southwestern tributaries. An analysis was also made of flood plains of Topopah Wash, a sizable drainage running from the Shoshone Mountain area into Jackass Flats, about 10 to 15 km east of Yucca Mountain (Christensen and Spahr, 1980). Figure 3.1.2.1.5-1 shows the flood-prone areas in Fortymile Wash based on the Squires and Young study and Figure 3.1.2.1.5-2 shows the site topography and potential flood areas.

Squires and Young (1984) analyzed the flood plain of Fortymile Wash and its southwestern tributaries in which the magnitudes of the 100-year and 500-year flood peaks and "regional maximum" flood were estimated. The southwestern tributaries include Busted Butte, Drillhole and Yucca Washes. The report concluded that most floods are caused by excessive runoff associated with convective storms.

Squires and Young (1984) used data from 12 USGS-operated, peak flow gauging stations to develop the following regression equations:

$$Q_{100} = 482\ A^{0.565}$$

$$Q_{500} = 2200\ A^{0.571}$$

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where A = the tributary drainage area in square miles,
Q_{100} = the magnitude of the 100 year flood in cubic feet per second, and
Q_{500} = the magnitude of the 500 year flood in cubic feet per second.

The standard errors for the estimates of the 100- and 500-year floods are .070 and 0.77 log units, respectively. These relatively large errors result primarily from the short period of record (15 to 20 years) and the extreme areal variability of flood flow in arid climates. However, because of the proximity of the gauging stations to the site and the similarity of the physiographic characteristics of the basins, the data taken from the 12 gauging stations should be representative of Fortymile Wash. In addition, the regression approach used by Squires and Young (1984) is believed by them to be the best available method for peak-discharge determinations.

The estimates of "regional maximum" floods made by Squires and Young (1984) are based on a graphical boundary curve developed by Crippen and Bue (1977). Crippen and Bue (1977) used data for historic floods of unusually large magnitude in a five-state region that includes Fortymile Wash without reference to recurrence interval to define a boundary curve of maximum discharges that have occurred in drainages of various sizes. The region studied included the states of Arizona, California, Nevada, New Mexico, and Utah. These "maximum" flood values, however, do not represent the upper limits of physically possible floods for these drainages.

Squires and Young (1984) conclude the following:

> Fortymile Wash within the flood-study area is a well-defined incised channel, with a cross-section of 15 to 21 m depth and 300 to 450 m width. The estimated values of the 100-year, 500-year, and regional maximum floods indicate that the flow would stay within the confines of the wash. Estimated depths of flood water in the stream channel would range from 0.9 to 2.4 m for the 100-year flood, 1.8 to 3.3 m for the 500-year flood, and 6.4 to 8.8 m for the regional maximum flood. Corresponding mean velocities would be from 1.8 to 2.7 m/s for the 100-year flood, 3.3 to 4.3 m/s for the 500-year flood, and 7.0 to 8.5 m/s for the regional maximum flood.
>
> The drainage basin of Busted Butte Wash varies from a shallow valley with meandering ephemeral streams to a deeply incised canyon in the upstream reaches. Drill Hole Wash is characterized by deep canyons extending from Yucca Mountain to its mid-drainage area. Both washes would have estimated flood-water depths of 0.3 to 1.2 m in the stream channel during the 100-year flood, and the corresponding mean velocities would range from 1.2 to 2.4 m/s. The 500-year flood would exceed bank capacities at several reaches of the washes. Depths and mean velocities would range from 0.9 to 3.0 m and 1.5 to 3.3 m/s. The regional maximum flood would inundate all central flat-fan areas in these two watersheds. Flood-water depths in the stream channels would range from 1.5 to 3.7 m, with mean velocities varying from 2.1 to 4.9 m/s.

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Yucca Wash is contained within an incised channel that is about 14 m deep and 240 m wide at its confluence with Fortymile Wash. The 100-year, 500-year, and regional maximum floods would stay within the steep-side-slope stream banks that contain the flood plain. Flood-water depths in the stream channel would range from 0.9 to 1.5 m for the 100-year flood, 1.5 to 2.7 m for the 500-year flood, and 2.7 to 7 m for the regional maximum flood; corresponding mean velocities would vary from 1.5 to 2.7 m/s for the 100-year flood, 2.4 to 3.7 m/s for the 500-year flood, and 2.7 to 6.7 m/s for the regional maximum flood.

The extent of erosion and sediment movement caused by flood flow in Fortymile Wash and its tributaries that drain Yucca Mountain is not known quantitatively. Qualitatively, however, on the basis of knowledge of record major floods elsewhere in southern Nevada (Glancy and Harmsen, 1975; Katzer, et al., 1976), erosion of or deposition in channels and flood plains probably would occur during the 100-year, 500-year, and regional maximum floods. Evidence of erosion and deposition was observed in some channels during field surveys. Channel erosion or aggradiation in the existing streambeds could alter the flood-flow characteristics of cross-sectional area, width, mean velocity, and maximum depth listed in the report by Squires and Young (1984).

The study by Christensen and Spahr (1980) defined flood-prone areas for the 100- and 500-year floods and maximum potential flood for the eastern part of Jackass Flats. The 100- and 500-year floods are identified by a recurrence interval; the average interval of time within which a flood of a given magnitude will be equalled or exceeded once. The maximum potential flood is based on maximum floods known to have occurred somewhere in the region without reference to the recurrence interval (Christensen and Spahr, 1980).

Christensen and Spahr (1980) used a regional regression model developed by USGS personnel in Carson City, Nevada, to estimate flash floods. Data from 71 gauged basins were used to develop the regression equations shown in Table 3.1.2.1.5-1. Only 19 of the basins used were in southern Nevada and none was located on the NTS. The primary variables in this model are basin size, altitude, and latitude. Although a regional regression flood model analysis can yield useful results, the following problems can occur when applying this model in an arid climate:

a. Limited streamflow data reduce accuracy of model;

b. An ephemeral channel is hydrologically and hydraulically different than a perennial channel;

c. Model is deterministic and does not account for channel migration as the flood passes over the alluvial fan; and

d. Model does not utilize available precipitation data.

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The Christensen and Spahr study also analyzed hydraulic characteristics of stream channels to estimate discharge, area, width, mean velocity, and maximum depth in channels for the 100-year, 500-year, and maximum potential flood. Hydraulic characteristics were based on natural flow conditions, 47 channel cross-sections, Manning's roughness coefficients obtained in the field, and estimated discharges for the three floods. Manning's equation was used to compute flood depths at each cross-section for the three floods. Table 3.1.2.1.5-1 shows the Christensen and Spahr flash flood model.

From their investigation of Topopah Wash, Christensen and Spahr (1980) concluded the following:

> The areas prone to 100-year floods closely parallel most main channels, with few occurrences of out-of-bank flooding of the areas between the main channel and adjacent secondary channels. Out-of-bank flooding would result in a water depth of less than 0.6 m, with a mean velocity as high as 2 m/s occurring on the steeper slopes. Flood-water depth in the stream channels would range from 0.3 to 2.7 m, with mean velocities of 0.9 to 2.7 m/s.
>
> The 500-year flood would exceed the discharge capacity of all stream channels except Topopah Wash and some upstream tributaries. Out-of-bank flooding of areas between the adjacent channels would result in water depths as much as 0.9 m, with mean velocities greater than 2 m/s. Flood-water depth in the stream channels would range from 0.3 to 3.7 m, with mean velocities ranging from 0.9 to 4 m/s.
>
> The maximum potential flood would inundate most of Jackass Flats. Out-of-bank flows in the areas between adjacent channels would have a depth as much as 1.5 m, with a mean velocity as high as 4 m/s. Flood water in the stream channels would have depths of 0.6 to 7 m, with velocities of 1.2 to 7.9 m/s. (Christensen and Spahr, 1980).
>
> Channels and floodplains would be severely eroded during the 100-year flood, and this erosion would be more widespread during the 500-year flood and the maximum potential flood. Channels eroded from their present condition would alter flood flow characteristics shown in the Christensen and Spahr (1980).

The maximum potential flood for Topopah Wash was estimated using data from maximum floods that have been observed on other streams having similar flood potential. Crippen and Bue (1977) developed a graphical boundary curve that defines maximum discharge that has occurred in drainages of varying sizes and is based on quantitative measurements of flood-flow peaks in a five-state region that includes Arizona, California, Nevada, New Mexico, and Utah. The maximum potential flood has no reference to recurrence interval or to the 100- and 500-year floods.

The rugged terrain at the Yucca Mountain Site and local convective storms can cause intense flooding to occur periodically in the normally dry washes that drain Yucca Mountain ridge.

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Protection from this flooding was considered in choosing the proposed locations for the surface facilities, shafts, and ramps in the current conceptual design. USGS topographic maps with 6.1 m contour intervals on a 1:24,000 scale were used together with direct observations to choose the proposed locations. New topographic maps of the Yucca Mountain area at a scale of 1:5,000 and a contour interval of 2 m have also been compiled and will be used in some aspects of the final design.

In general, maximum flood flows as derived from Crippen and Bue (1977) contain graphs of peak discharges versus drainage area of measured historical floods, with envelope curves above the plotted floods. The envelope curves represent the maximum potential flood for a given drainage area. The graph used for estimating maximum flows at the site is based on a region covering all or parts of Nevada, California, Utah, Arizona, and New Mexico. This methodology provides estimates of the flood discharges at the site suitable for conceptual design. [Results of additional flooding studies will be provided here. [INN 3.1.2.1.5-1], Results of additional flooding studies.] [INN 3.1.2.1.5-3, Review and Reference two Technical and Management Support Services floodplain assessments.]

[Note: The design basis water levels and elevations of all surface facilities and underground openings have been developed, but could not be compiled for this revision [INN 3.1.2.1.5-2], Design basis water levels and elevations of underground openings.]

3.1.2.1.6 Chemical Composition of Identified Bodies of Surface-Water

The chemical composition of bodies of water that could be affected by releases from the proposed facility is described in this Subsection. The seasonal cycles of physical and chemical limnological parameters, bottom and shoreline configuration, sedimentation rates, sedimentation gradation analysis, and sorption rates are not applicable to the discussion of the chemical composition of surface-water in the YMR because of the arid to semiarid climate.

Because of the ephemeral character of most of the surface water in the YMR, historical and recent records are limited. Only two water samples have been taken from the Amargosa River; one from a reach near Eagle Mountain and the other from Carson Slough, a major tributary that joins the Amargosa River near Alkali Flat. Table 3.1.2.1.6-1 shows the chemical composition of this water. The sulfate and chloride are indicative of groundwater interaction with playa deposits. The chemical composition of both water samples is similar to water from the tuffaceous valley fill aquifer (DOE, 1988).

Samples of surfacewater from the Yucca Mountain area were collected in July and August 1984. Water samples were collected from the main stream channel of Fortymile Wash and from two of its principal tributaries, Drill Hole and Busted Butte Washes. Results of the analyses are also shown in Table 3.1.2.1.6-1. All the samples are significantly lower in solutes than groundwater samples of the area (DOE, 1988). Samples of surface water collected during periods of runoff and flooding were also collected during storms in 1993. [INN 3.1.2.1.1-1], Provide surface-water monitoring data when available. Table 3.1.2.1.6-2 shows the chemical composition of selected springs in the hydrographic study area based on a

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study by McKinley, Long, and Benson (1991). [There is no information available on the chemical composition of discharge from playas. [INN 3.1.2.1.1.-2], Characteristics of playas.]

3.1.2.1.7 Location, Quantity, and Quality of Surface Water Extracted

The location, quantity, and quality of surface water extracted are important to identify and to establish baseline information on the quality and quantity of water used that could be impacted by the site. Because of the arid climate and limited perennial surface-water supplies, little surface-water is extracted in the YMR. The only surface-water bodies currently used are a few springs, spring-fed pools, and man-made reservoirs scattered throughout the southern portion of the hydrologic area that extract water for domestic, agricultural, commercial, milling, and mining purposes.

The Furnace Creek Inn and Ranch Facilities at Death Valley National Monument in Death Valley, California are supplied with water from three springs: Travertine Spring, Texas Spring, and Nevares Spring. The springs are owned by the National Park Service and serve approximately 800 permanent residents and 2,000 visitors mainly during the winter months. The springs also supply irrigation needs of the park (French, et al., 1984). Table 3.1.2.1.7-1 lists the quality of water extracted from Texas and Nevares Spring.

The Beatty Water and Sanitation District is a separate public entity with 250 customers in the Beatty area. These customers are served by spring water from Indian Springs (not to be confused with the community of Indian Springs, Nevada) located northeast of the town of Beatty (French, et al., 1984) and three other wells. Table 3.1.2.1.7-1 lists the quality of water extracted from Indian Springs. One family located in Rhyolite, Nevada, is also supplied with water from Indian Springs (French, et al., 1984). Spring locations and spring water chemistries are discussed in detail in Subsections 3.1.2.1.1 and 3.1.2.1.6, respectively. Spring water use is discussed in Subsection 3.1.2.2.8.

Several small, man-made reservoirs used for agricultural, milling, and mining purposes are scattered throughout the southern portion of the hydrologic study area. The largest, a man-made reservoir, known locally as Crystal Reservoir, is located in the Ash Meadows area 52 km southeast of Yucca Mountain. This reservoir receives its water from Crystal Pool spring, located approximately 0.8 km to the north. Crystal Reservoir, which has a capacity of 1.84×10^6 m^3, is used for recreational purposes only. Another small reservoir with a capacity of 5.31×10^4 m^3 is used for milling. This and many other reservoirs are fed by wells. Other small, spring-fed reservoirs with capacities much less than 4.94×10^4 m^3 exist near Point of Rocks and Death Valley Junction (Giampaoli, 1986).

The Amargosa River is not used for water supply. The flow regime has been intermittent to ephemeral through historic times, except for the spring-fed reaches. As such, sufficient supplies for domestic, municipal, commercial, agricultural, or industrial uses have not been available.

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No other sources of surface water exist within the hydrographic study area. Frenchman Lake and Yucca Lake are playas that contain water only after rainstorms. Because of the sporadic precipitation and runoff, and the excessive amounts of dissolved solids (Waddell et al., 1984), playas are not considered possible sources of water supply. The virtual absence of perennial surface water precludes its application for beneficial use in the hydrographic study area. Subsection 3.1.2.1.6 provides information on the chemical composition of identified bodies of surface water.

3.1.2.1.8 Projected Surface-Water Uses

It is important to predict future surface-water uses in the area to identify water supplies that could be impacted by the potential repository and identify future water needs in addition to those of the site. Arid conditions of the region and the absence of perennial streams or large lakes make surface water an unlikely future water supply regardless of whether the region's population increases. Within the hydrographic study area, surface water serves two principal functions: recharge to the valley-fill aquifer; and maintenance of aquatic wildlife habitats. Agricultural, recreational, milling, or mining operations are supplied in part by several small, man-made reservoirs scattered through the southern portion of the hydrologic study. Nevertheless, because of the limited supply, use of surface water in the study area by man would probably increase only in the event of change in climate to a wetter regime. Such a change in climate is unlikely during the period of repository construction, operation, closure, and decommissioning (DOE/RW-0073).

3.1.2.2 Regional Hydrogeology

This subsection provides general information on the regional hydrogeology and discusses the regional groundwater conditions that may be affected by the Mined Geologic Disposal System. In addition, the existing and projected users and uses of groundwater will be described.

3.1.2.2.1 Regional Flow System Boundaries and Hydrogeologic Units Overview

A variety of study techniques has been applied to the delineation of groundwater flow system boundaries in the region surrounding Yucca Mountain. These include identification of major discharge and recharge areas, recognition of the distribution of low-permeability rock units, mapping of potentiometric surfaces associated with generally-accepted hydrostratigraphic units, mass-balance water-budget calculations for hydrographic basins, evaluation of geochemical reaction paths, age-dating of groundwater, and stable-isotope studies. Most of the hydrologic boundaries of the regional aquifer systems are geologically complex and existing data only allow preliminary approximations.

Regional flow-system boundaries in southern Nevada were developed by Mifflin (1968) on the basis of both conventional hydrologic data and system classification studies of the large springs. The greatest problem of delineation in this region is location of flow-system boundaries in areas where important flow occurs at depth in carbonate rock (Mifflin and Hess,

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1979). Shallow fluid potential data may be misleading as to the location of significant boundaries.

While details of regional flow-system boundaries may not be attainable everywhere, it is possible to evaluate the sensitivity of local conditions to uncertainties in the regional boundaries. A sub-regional model was constructed using boundary conditions that are partially constrained by the results of regional modeling. Using this approach, information needs associated with flow-system boundaries can be prioritized based on the sensitivity of site-scale results to conditions at distant boundaries.

In contrast to the uncertainties associated with flow-system boundaries, the hydrostratigraphy of the YMR is well-established in terms of the general hydraulic properties of major rock units. This is not to imply that the three-dimensional distribution of these units is known in detail, rather that stratigraphic groupings can be used to identify the principal aquifers and aquitards. The assignment of aquifer or aquitard status to a particular lithology is not universally applicable, however; welded tuffs, for example, may function as aquifers or aquitards depending on fracturing.

3.1.2.2.1.1 Hydrogeologic Units

The stratigraphic and hydrogeologic units at the NTS and vicinity were evaluated by Winograd and Thordarson (1975) as part of a broader synthesis of the hydrogeologic framework of the south-central Great Basin. Six major aquifers and numerous aquitards are present in the region. The aquifers include the lower and upper carbonate aquifers; bedded-tuff, welded-tuff, and lava-flow aquifers; and valley-fill aquifers.

Carbonate Aquifers. From a resource perspective, aquifers associated with Paleozoic carbonate rocks are by far the most important water-bearing units in the region. The lower carbonate aquifer is particularly important to the movement of water between Yucca Flat and Frenchman Flat and discharge areas in Ash Meadows and Death Valley. The hydraulic continuity of carbonate rock in the subsurface is the major factor governing interbasin movement of groundwater in the region.

Only a few wells penetrate the lower carbonate aquifer on the test site, and none of these wells penetrate more than 200 m of saturated carbonate rock. Test Well UE-25p#1 encountered 561 m of Paleozoic dolomite beneath the south end of Midway Valley (Craig and Robison, 1984). Until more geologic and hydrologic data are available from saturated carbonate rocks, however, flow within the lower carbonate aquifer cannot be characterized in any detail. A major uncertainty is the extent to which the carbonate section has been compartmentalized by Cenozoic extension and thereby isolated from the regional flow systems thought to exist in eastern Nevada and western Utah.

Volcanic-Rock Aquifers. While minor aquifers are known to occur in bedded tuffs and lava flows, welded tuffs constitute the most productive volcanic-rock aquifer of the region. The voluminous ash flows from mid-Miocene caldera complexes of the SNVF (north of Yucca

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Mountain) resulted in densely-welded, compound cooling units sometimes hundreds of meters thick, which due to their brittle nature contain abundant fractures.

Ervin et al. (1993) have subdivided the "welded-tuff aquifer" of Winograd and Thordarson (1975) into upper and lower flow systems that are applicable to the Yucca Mountain area. The lower system generally corresponds to the Crater Flat Tuff, and the upper system to the Paintbrush Tuff. The upper groundwater flow system north of Yucca Mountain in volcanic rocks is conceptualized as occurring above the tuffaceous beds of Calico Hills; the lower system occurs beneath the tuffaceous beds of Calico Hills. Beneath and east of Yucca Mountain the Paintbrush Tuff is generally above the level of Regional Saturation, and the uppermost saturated zone in those areas most commonly occurs in the Crater Flat Tuff. Beneath and east of Yucca Mountain the upper system is considered to be absent.

Valley-Fill Aquifers. Alluvium, fanglomerate, lake beds, and other Cenozoic detrital materials can constitute aquifers where they occur below the level of regional saturation; central Amargosa Valley is a good example. Much of the valley-fill in the region is, however, unsaturated. Hydraulic relations between valley-fill aquifers and other underlying or adjacent aquifers have received considerable attention during water-rights disputes such as those involving the Ash Meadows Wildlife Refuge, but technical issues related to groundwater fluxes, flow directions, and velocities are largely unresolved.

Paleozoic and "Eocambrian" Clastic Aquitards. Beneath the Paleozoic carbonate section is a sequence of quartzites, siltstones, shales, and minor dolomite referred to by Winograd and Thordarson (1975) as the lower clastic aquitard. Within the Paleozoic carbonate section, the similar lithologies of the Mississippian Eleana Formation have been referred to as the upper clastic aquitard.

Volcanic Aquitards. Bedded tuffs, where hydrothermally altered or zeolitized and welded tuffs, where relatively unfractured, are considered aquitards.

3.1.2.2.1.2 Basis for Defining the Physical Boundaries of the Regional Hydrogeologic Systems

The nomenclature for flow-system description is based primarily on scale and relationship to topography. Following the usage of Mifflin and Hess (1979) a *regional* groundwater flow system is defined as a large groundwater flow system which encompasses one or more topographic basins; it may include within its boundaries several groundwater basins; interbasin flow is common and important with respect to total volume of water transferred within the system boundaries; and lengths of flow paths are relatively great when compared to lengths of flow paths of "local" flow systems. A *local* groundwater flow system is generally confined to one topographic or groundwater basin; interbasin flow is not important with respect to total volume of water transferred within the system; the majority of flux of water within the system discharges within the associated groundwater basin and flow paths are relatively short when compared to regional systems. The distinction between topographic basins (hydrographic areas) and groundwater basins is that a groundwater basin is defined as

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the area contributing to a significant surface discharge, and may therefore be local or regional in extent. The term *groundwater flow system* is preferred over the term *groundwater basin*, although the two terms are considered synonymous. The term *interbasin flow* is retained, and is used to designate widespread lateral transfer of groundwater within rocks (usually carbonates) located at depth beneath topographic basins (Naff et al., 1974).

As a starting point for identifying boundaries of regional hydrogeologic systems, the hydrographic areas proposed by Rush (1968) have been utilized by Dettinger (1992). Area boundaries were developed by the Nevada State Engineer's office to provide a framework for water-balance calculations; they generally correspond to topographic divides and may differ from groundwater flow system boundaries discussed in the scientific literature. The relationship of probable recharge and known discharge areas to the hydrographic subdivisions are shown in Figure 3.1.2.2.1.2-1. An evaluation of regional, interbasin flow in the carbonate aquifers by Dettinger (1992) included consideration of the areal extent and continuity of carbonate-rock sequences, including information on stratigraphic and structural thinning. A summary of estimates of annual groundwater inflow, recharge, and outflow is given in Table 3.1.2.2.1.2-1.

Winograd and Thordarson (1975, p. 94) introduced a change in nomenclature from that in use by the State Engineer's office and largely retained by Dettinger (1992), referring to "Oasis Valley - Fortymile Canyon Groundwater Basin" that would include Yucca Mountain. Generalized potentiometric contours for the Pahute Mesa area (Savard, 1989; Blankennagel and Weir, 1973) suggest that some groundwater flows from Pahute Mesa toward Oasis Valley and that the water beneath Pahute Mesa may, in turn, be derived from Kawich Valley to the northeast (Figure 3.1.2.2.1.2-2). Interbasin movement of groundwater in volcanic rocks is therefore implied from potentiometric data from beneath Pahute Mesa, but the flow system geometry within the Timber Mountain Caldera (Figure 3.1.1.1.4.2.1-1) is unknown due to a near-absence of potentiometric data in that area.

Waddell (1982) refined the terminology of Winograd and Thordarson (1975), defining the Alkali Flat-Furnace Creek Ranch groundwater basin to distinguish the flow system of interest from that which discharges in Oasis Valley. Waddell retained the designation "Ash Meadows groundwater basin," which has been in general usage and a well-accepted approximation of flow-system boundaries for over two decades (see, for example, Winograd and Friedman, 1972). In Waddell's terminology, the Ash Meadows and Oasis Valley groundwater basins are subbasins within the Alkali Flat-Furnace Creek Ranch basin (Figure 3.1.2.2.1.2-2). In a manner analogous to the Oasis Valley and Ash Meadows groundwater basins being tributaries to the Alkali Flat-Furnace Creek Ranch basin, Alkali Flat-Furnace Creek Ranch basin is tributary to Death Valley basin. Outflow to Death Valley basin occurs via discharge and underflow at Furnace Creek and probably as underflow in Amargosa River alluvium near Eagle Mountain (Waddell, 1982).

The northern boundary of the Alkali Flat-Furnace Creek Ranch groundwater basin is along a somewhat arbitrary line that crosses the Cactus, Kawich, and Reveille Ranges (Waddell, 1982). In a regional, two-dimensional flow model with no-flow lateral boundaries Waddell

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(1982) prescribed a recharge flux of 4.0×10^{-10} m^3/s (1.0×10^{-5} acre-ft/yr) to the area representing Pahute Mesa (Figure 3.1.2.2.1.2-3) in his model. In a later effort using a sub-regional model, Czarnecki and Waddell (1984) placed a constant-head boundary based on heads derived by Waddell (1982) at the northernmost boundary of their model (Fig. 3.1.2.2.1.2-4). In his evaluation of effects of increased recharge, Czarnecki (1985) indicates that flux across the northern boundary is prescribed as 0.2959 m^3/s (7570 acre-ft/yr), the same as that resulting from prescribed heads in Czarnecki and Waddell (1984).

Inflow to northern Jackass Flats was divided into two components by Czarnecki and Waddell (1984) based on the regional model of Waddell (1982). These components (Figure 3.1.2.2.1.2-4) include a flux from the Eleana argillite (Qjf) of 0.1835×10^{-2} m^3/s (47 acre-ft/yr) and another from the carbonates underlying Rock Valley (Qrv) of 0.1249×10^{-1} m^3/s (320 acre-ft/yr). Czarnecki (1991), in testimony before the Nevada State Engineer, indicated that a no-flow boundary south of Rock Valley is associated with the lower clastic aquitard, which would block flow and divert it toward the west.

Recharge to the Alkali Flat-Furnace Creek Ranch system also occurs as infiltration from Fortymile Wash and as underflow from the Ash Meadows and Oasis Valley systems. Claassen (1985) concluded from geochemical evidence that groundwater in the west-central Amargosa Desert was recharged primarily by overland flow of snowmelt in or near present-day stream channels during late Wisconsin time. In their base case simulation, Czarnecki and Waddell (1984) and Czarnecki (1985) applied a flux (Qfm) of 0.2563 m^3/s (6557 acre-ft/yr) to Fortymile Wash to simulate recharge by overland flow (Figure 3.1.2.2.1.2-4).

Input to the Alkali Flat-Furnace Creek Ranch flow system occurs at Ash Meadows, to a minor extent from recycling part of the 0.67 m^3/s (17,100 acre-ft/yr) of spring discharge reported by Winograd and Thordarson (1975), and to an unknown extent from underflow across the spring line into the central Amargosa Desert. According to Winograd and Thordarson (1975), the amount of such underflow could not be approximated with available data. Czarnecki and Waddell (1984) assumed an underflow flux (Qam) of $0.8990 \times 10^{-3} m^3/s$ (23 acre-ft/yr) from Ash Meadows.

Inflow from the western Amargosa Desert (Oasis Valley) was assumed by Czarnecki and Waddell (1984) to be $0.2244 \times 10^{-3} m^3/s$ (6 acre-ft/yr). This flux (Qwad) represents underflow beneath the Amargosa River channel south of Beatty (Fig. 3.1.2.2.1.2-4).

Outflow from the Alkali Flat-Furnace Creek Ranch flow system could occur either by underflow and evaporation in the vicinity of Eagle Mountain, immediately south of the sub-regional model boundary, or by underflow through the lower carbonate aquifer toward Death Valley (Winograd and Thordarson, 1975). Evaporation from Franklin Lake Playa (Alkali Flat) was assumed to be 0.4120 m^3/s (10,500 acre-feet per year) and was later estimated by Czarnecki (1990) as 0.26 m^3/s (6750 acre-ft/yr) using the energy-balance eddy-correlation technique. Total discharge from the Furnace Creek area is reported to be 0.20 m^3/s (about 5100 acre-ft/yr) (Hunt et al., 1966, Table 25 cited in Winograd and Thordarson, 1975). The

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amount of groundwater that might move southward past Eagle Mountain is unknown, and was assumed to be zero by Czarnecki and Waddell (1984).

Other information that relates to flow-system boundaries is available from numerous evaluations of the possible relationship of groundwater in Pahrump Valley to discharge at Ash Meadows. Winograd and Thordarson (1975) cited three lines of evidence to refute previous estimates that as much as 0.5 or as little as 0.1 m^3/s (13,000 or 3,000 acre-ft/yr) of the yearly discharge at Ash Meadows is derived from Pahrump Valley via underflow through Paleozoic carbonate rocks. First, the Montgomery Thrust is interpreted to isolate groundwater in the lower carbonate aquifer and in the Cenozoic aquifers in northwestern Pahrump Valley from the same aquifers in the east-central Amargosa Desert, since the upper plate contains a nearly continuous band of lower clastic aquitard rocks between Pahrump and Stewart Valleys and the east-central Amargosa Desert. Second, a comparison of the chemical and isotopic composition of groundwater in Pahrump and Stewart Valleys with that at Ash Meadows precludes movement of significant quantities of water from these valleys toward Ash Meadows (Winograd and Thordarson, 1975; Winograd and Friedman, 1972). Third, shallow water levels beneath the playas in Stewart and northwestern Pahrump Valleys suggest that the groundwater in these areas is ponded by some impermeable boundary, in contrast to lower water levels in southwestern Pahrump Valley, which is bounded by lower carbonate-aquifer rocks underlying the Nopah Range. Therefore, water from Pahrump Valley may not even constitute a source of secondary importance to the water budget at Ash Meadows.

A final topic that bears on the quantity of flux in the Ash Meadows groundwater system is the possibility of underflow from Pahranagat Valley. Based on a mass-balance calculation using the deuterium content of groundwater, Winograd and Friedman (1972) concluded that the Ash Meadows springs are fed by a mixture of recharge derived from the Spring Mountains - Sheep Range uplands, 50 to 100 km east of the springs, with underflow from Pahranagat Valley and vicinity, 145 km northeast of the springs. The computed underflow amounts to about 35 percent of the spring discharge or approximately 0.23m^3/s (6000 acre-feet/year).

3.1.2.2.2 Potentiometric Levels and Hydraulic Gradients

Sources of Data and Periods of Record. Potentiometric data from the YMR are collected and compiled through two DOE-sponsored programs, the Environmental Monitoring Program (EMP) and the Site Characterization Program. Environmental-Monitoring sites in the region are given in Figure 3.1.2.2.2-1; as yet unpublished quarterly reports from the EMP state that previously unpublished data given in the reports should be considered provisional. The periods of record for localities included in the EMP are highly variable, since much of the information is compiled from "available sources". [Regional water-level data and a hydrogeologic map of the region (from the Site Characterization Program) will be available during Fiscal Year 1994] In contrast, potentiometric levels in the uppermost saturated zone in Tertiary volcanic rocks in the Yucca Mountain area have been monitored on either a periodic or continuous basis since 1983 or 1984 (Ervin et al., 1993). The site potentiometric map and supporting data from Ervin et al., (1993) are given in Figure 3.1.2.2.2-2 and Table 3.1.2.2.2-1,

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[INN 3.1.2.2.2-1] respectively. Averaged 1988 water-level data were used in constructing Figure 3.1.2.2.2-2, based on an analysis of temporal trends and corrections for improved elevation control, equipment wear, and mechanical stretch and thermal expansion of the steel tapes used for measurement.

Cenozoic Rocks. Water levels in the valley-fill aquifer of the Amargosa Desert were compiled by Claassen (1985), who field-checked localities reported by Walker and Eakin (1963), and noted only very minor water-level changes between 1962 and 1974. Minor modifications to Claassen's 1985 map were made by Czarnecki (1990) based on new field data. Czarnecki's results were combined with those of Ervin et al. (1993) in Figure 3.1.2.2.1.2-4. Features of the potentiometric surface associated with the Cenozoic aquifer of the Yucca Mountain - Amargosa Desert region are as follows:

a. Large hydraulic gradient north of Yucca Mountain. An apparent gradient of at least 0.15 separates regions where water levels differ by approximately 300 m. The large gradient is presently unexplained, but appears to be regional in extent, trending generally SW-NE. The large hydraulic gradient may be due to damming of southward flow or to the presence of a high-permeability drain from shallow aquifers to a deeper (presumably carbonate) aquifer system.

b. Moderate hydraulic gradient west of Yucca Mountain. Higher water levels west of the Solitario Fault Zone appear to be fault-related, but there is insufficient hydrologic data at the present time to make a confident interpretation.

c. Possible upwelling between Amargosa Valley (formerly Lathrop Wells) and Ash Meadows. Claassen (1985, 1991) interprets potentiometric, temperature, and geochemical data as indicating that inflow of carbonate-aquifer water is affecting the Cenozoic aquifers in this area.

d. Steepening of the hydraulic gradient at the latitude of Ash Meadows. This steepening may be due to structural control, but evidence is lacking.

e. Contours do not close beneath Alkali Flat. This suggests that flow to the south, past Eagle Mountain, is occurring.

Pre-Cenozoic Rocks. Potentiometric levels in pre-Tertiary rocks of the YMR were compiled by Dudley (1991) as supporting information for testimony before the Nevada State Engineer on a proposed appropriation of water from Well J-13. Dudley's map (modified and incorporated in Figure 3.1.2.2.2-3) is based on data from the EMP, SCP, and USGS records of NTS hydrologic investigations.

In examining Figure 3.1.2.2.2-3, the absence of any control between Yucca Mountain and Furnace Creek Wash should be noted. Discharge at Nevares Spring, in Furnace Creek Wash, occurs at an elevation of 286 m. A steepening of the hydraulic gradient beneath the Funeral Mountains seems unlikely given the predominance of carbonate rocks in that area. If, on the

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other hand, hydraulic gradients in carbonate aquifers are relatively flat beneath the Funeral Mountains and central Amargosa Desert, as they are up-gradient from Ash Meadows, the potentiometric surface would be quite different than that portrayed on Figure 3.1.2.2.2-3. A steepening of the hydraulic gradient would be implied in the vicinity of Ash Meadows, which is consistent with the 48-meter elevation difference between Devil's Hole and Crystal Pool, which are only 3 km apart. Carbonate-aquifer fluid potentials substantially less than those in overlying Cenozoic aquifers of the central and southern Amargosa Desert would imply widespread lowering of carbonate-aquifer fluid potentials over the last million years, based on the work of Winograd and Szabo (1988) (see Subsection 3.1.2.3.10).

The prominent potentiometric trough that extends northeastward from Ash Meadows to Frenchman Flat and then north-northwestward to Yucca Flat was discussed at length by Winograd and Thordarson (1975), but subsurface information is insufficient to explain its origin. Wells that appear to define the west flank of the trough west of Yucca Flat are completed in the Mississippian Eleana Formation and younger carbonates suggesting semi-perched conditions. If the potentiometric surface in the lower carbonate aquifer is lower than in overlying units, flow leaving Yucca Flat might move in a more southwesterly direction than the contours of Figure 3.1.2.2.2-3 suggest. This possibility is significant to the assessment of hydraulic boundary conditions in the Rock Valley area for sub-regional modeling applications.

3.1.2.2.3 Characteristics Of Hydrogeologic Units

Lower Carbonate Aquifer. Hydraulic properties of the lower carbonate aquifer have been measured by drill-stem and single-well pumping tests of 10 wells (Table 3.1.2.2.3-1 [INN 3.1.2.2.3-1]; Winograd and Thordarson, 1975). Coefficients of transmissibility estimated from the specific-capacity data of the 10 wells range from 600 to 900,000 gpd per ft. (7.5 to 1.1×10^4 m^2/day). With one exception, Table 3.1.2.2.3-1 contains data for only those wells for which drawdown and recovery data were available; well 79-69a is included because its estimated transmissivity represents a measured maximum for the NTS. Hydraulic conductivities, computed by dividing the apparent transmissivity by the thickness of the producing interval, range from 0.03 to 228 m/day, and average about 1 m/day if drawdown data are used in the analysis and the one extreme value (from well 79-69a) is ignored.

Drawdown and recovery data from the Amargosa Tracer Calibration Site could not be presented in Table 3.1.2.2.3-1 and companion figures since only indirect references are available. According to Leap and Belmonte (1992), these data exist in unpublished form in the files of the U.S.G.S., Water Resources Division, Denver, Colorado. The Amargosa tracer calibration site is shown as locality AD-7 on Figure 3.1.2.2.2-1. The aquifer at the Amargosa tracer calibration site is believed to be a solution-modified, low-angle reverse fault zone within the Cambrian Bonanza King dolomite (Leap and Belmonte, 1992) based on examination of drilling logs and cores. Five wells completely penetrate this aquifer, whose thickness varies from 3.1 m to 13.7 m. Transmissivity determined by a pumping test ranges from 4,800 m^2/day to 10,900 m^2/day, indicating a minimum apparent hydraulic conductivity of 350 m/day; as indicated above, pumping response curves are unpublished. A series of

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tracer tests using tritiated water resulted in estimates of average interstitial velocity that ranged from 0.14 to 3.4 m/day (Claassen and Cordes 1975).

Response of the lower carbonate aquifer to pumping is characterized by an anomalously rapid rate of recovery after pumping (Figures 3.1.2.2.3-1 through 3.1.2.2.3-4); Winograd and Thordarson (1975) indicated that the significance of the anomalous recovery rate was not understood, and they therefore elected not to use transmissivity values obtained from recovery curves. No re-analysis of these data is known to have occurred (Winograd and Thordarson, 1975).

For the six carbonate-aquifer wells with analyzable drawdown curves, transmissivity values obtained from specific capacity data are on the order of 1/6 to 1/2 of the values derived from the second limb of the drawdown curves, and on the order of 1/10 to 1/4 of values obtained from those test with analyzable (Figures 3.1.2.2.3-1 through 3.1.2.2.3-4) recovery curves. No attempt was made to determine the storage coefficient of the lower carbonate aquifer from pump-test data, but "extremely low" effective porosities of the carbonate rocks were noted. (Winograd and Thordarson, 1975)

Carbonate rocks encountered in Test Well UE-25p#1 were evaluated with single-well pumping and recovery tests (Figures 3.1.2.2.3-5 through 3.1.2.2.3-8 and Table 3.1.2.2.3-1) and packer-injection tests by Craig and Robison (1984). An interval of less than 10 meters in the upper part of the Lone Mountain Dolomite has an apparent transmissivity of 59 m^2/day. Below this permeable zone, the next 190 meters has an apparent transmissivity of 33 m^2/day. Between 1550 and 1805 meters, the apparent transmissivity is 6 m^2/day. Hydraulic conductivities for Paleozoic rocks in the depth interval 1297 to 1805 m therefore ranged from 0.02 to greater than 5.9 m/day. The transmissivity of the full 508m interval is estimated at 111 m^2/day, which translates to an average hydraulic conductivity of 0.22 m/day. Craig and Robison (1984) did not calculate values of the storage coefficient.

The design of the packer-injection tool used during testing of well UE-25p#1 restricted determinations to transmissivities less than about 5 m^2/day (Craig and Robison, 1984). This restriction applied to 11 of the 12 packer-injection tests in Paleozoic rocks of UE-25p#1, so packer-injection data are not presented in Table 3.1.2.2.3-1.

Upper Carbonate Aquifer. No hydraulic tests are known to have been made in the upper carbonate aquifer in the vicinity of the NTS or within the Ash Meadows groundwater basin (Winograd and Thordarson, 1975). In 1981, Ertec Western, Inc. tested borehole CE-DT-5 in Coyote Spring Valley (Figure 3.1.2.2.1.2-1) as part of the MX Siting Investigation for the U.S. Air Force (Ertec Western, 1981). The transmissivity of the fractured aquifer associated with the lowermost Anchor and uppermost Dawn Members of the (Mississippian) Monte Cristo Limestone is on the order of 2.32 x 10^4 to 9.29 x 10^4 m^2/day, higher than any transmissivities yet determined in any unit in the NTS region.

Welded-Tuff Aquifer. The welded-tuff aquifer near Yucca Mountain has been evaluated as a potential water supply for the former Nuclear Rocket Development Station (Young, 1972)

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and in conjunction with the Site Characterization Program (Table 3.1.2.2.3-2 [INN 3.1.2.2.3-2]). A significant body of pumping-response and recovery data is related to testing of Well J-13 (Figures 3.1.2.2.3-9 and 3.1.2.2.3-10), as summarized by Thordarson (1983). The apparent transmissivity of 120 m^2/day, based on later-time data, is considered more representative of actual aquifer conditions; 850 m^2/day probably is a reasonable maximum value for transmissivity in the opinion of Thordarson (1983). Single-well methodologies have been used almost exclusively for testing the welded-tuff aquifer(s) at Yucca Mountain and vicinity, with three exceptions. First, interference tests were conducted at the C-well complex in 1983-84 [INN 3.1.2.2.3-2]. Second, well UE-25a#1 was used as an observation well during pumping of UE-25b#1 (Lahoud et al., 1984). Third and last, Thordarson (1983) reports that Well J-13 responded to pumping of Well J-12, 4.7 km to the south, by exhibiting drawdown of 0.37 m after three days of continuous pumping J-12 at an average discharge rate of 22.7 l/s. Young (1972) indicates that in a 4-day pumping test Well J-12 produced water at a rate of 370 gpm (23.3 l/s) with a drawdown of 2.5 feet (0.76 m). The referenced observations appear to represent the same test given the long duration and similar discharge estimates, and could be used to constrain permissible combinations of transmissivity and storativity (or specific yield) in the region between J-12 and J-13.

In borehole UE-25p#1, a 919 m composite section of Tertiary tuffs including the Crater Flat, Lithic Ridge and older tuffs was pumped at 22.1 l/s for 3150 minutes (Craig and Robison, 1984); a borehole-flow survey indicated that about 58 percent (12.8 l/s) of the production was from an interval less than 30 m thick in the upper part of the Prow Pass member of the Crater Flat Tuff. The transmissivity of the composite Tertiary section in UE-25p#1 is "on the order of 25 m^2/day" (Figures 3.1.2.2.3-11 through 3.1.2.2.3-13 and Table 3.1.2.2.3-2). If the second straight-line segment in Figure 3.1.2.2.3-10 is used, the transmissivity would be 60 m^2/day, a likely maximum (Craig and Robison, 1984). One discrete zone in the Prow Pass has been shown to have a hydraulic conductivity approaching 0.5 m/day.

Six drawdown and recovery tests were performed in borehole USW VH-1, which is located about 6 km west of Yucca Mountain in Crater Flat (Figures 3.1.2.2.3-14 through 3.1.2.2.3-17 and Table 3.1.2.2.3-2). None of the tests was wholly satisfactory because of problems caused by:

a. Caving and sloughing of the hole;

b. Turbulent flow losses in the well and possibly the aquifer;

c. An increase in water temperature during pumping which caused the density of the water to decrease; and,

d. Possibly, other unknown factors in the well-aquifer system (Thordarson and Howells, 1987).

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Apparent transmissivities from the composite section in VH-1 range from 450 m^2/day to 2200 m^2/day (Table 3.1.2.2.3-2), which correspond to average hydraulic conductivities of 0.78 to 3.8 m/day, respectively.

Three pumping tests were conducted in test well UE-25b#1 after the well had been drilled to its total depth of 1220 m (Lahoud et al., 1984). Well UE-25a#1, 107 m south-southwest of UE-25b#1, was used as an observation well during tests 2 and 3. Well USW H-1, 2.0 km northwest of the pumped well in Drill Hole Wash, was monitored during the tests but no measurable water-level decline was observed. Lahoud et al. (1984) used the Thiem equation and data from test 3 in an attempt to define the hydraulic coefficients (Figure 3.1.2.2.3-18 and Table 3.1.2.2.3-2); however, they indicated that the calculation was for illustrative purposes only and may not have any physical basis. They compute an apparent transmissivity of 340 m^2/day and an apparent hydraulic conductivity of 0.45 m/day.

Three drawdown and recovery tests were performed in borehole USW H-1, which is located in Drill Hole Wash (Figures 3.1.2.2.3-19 through 3.1.2.2.3-27 and Table 3.1.2.2.3-2). The first test was conducted when the well had been drilled to a depth of 688 m, and the second and third tests when the well had been drilled to its total depth of 1829 m and cased to 687 m (Rush et al., 1984). Nearly all the permeable rock penetrated by this well is in the Prow Pass member of the Crater Flat Tuff above a depth of 688 m. The apparent transmissivity using drawdown data from the first test is about 154 m^2/day, which corresponds to an average hydraulic conductivity on the order of 1 m/day. The most permeable zone, 652 to 653 m, has an apparent average hydraulic conductivity of 18 m/day (Rush et al., 1984, p. 40). The apparent transmissivity from the second and third tests for the depth interval 687 to 1829 m, which includes the lower 20 m of Prow Pass member and older units, is on the order of 1 m^2/day. The maximum apparent hydraulic conductivity from the deeper tests was 0.1 m/day from the depth interval 736 to 741 m in the Bullfrog member (Rush et al., 1984, pp. 52-53); below a depth of 811 m, the apparent hydraulic conductivities of all zones are less than 4×10^{-5} m/day.

During drawdown and recovery tests described above, a drawdown in well USW G-1 resulted from pumping of well USW H-1, indicating hydraulic connection between these wells (Rush et al., 1984). No attempt was made to determine aquifer properties from USW G-1 data since the drawdown was thought not to be representative of aquifer conditions.

Borehole USW H-3 was initially pumped for several brief cycles of drawdown and recovery, but could not sustain a yield of 3.13 l/s and was subsequently equipped with a smaller pump (Thordarson et al., 1985) that was satisfactory for long-term testing. The straight-line method, Brown's method for cyclic pumping, and the Theis method were used to analyze the data (Figures 3.1.2.2.3-28 through 3.1.2.2.3-30 and Table 3.1.2.2.3-2). For the depth interval 754 to 1219 m, which contains mostly the Tram member of the Crater Flat Tuff, apparent transmissivities of 0.4 and 0.5 m^2/day were computed from cyclic pumping and recovery data, respectively. After installing a packer at 822 m and the smaller-capacity pump, the Theis method yielded an apparent transmissivity of 1.0 m^2/day for the 822 to 1219m interval. Pumping responses in all three tests exhibited only small departures from ideal behavior,

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except that the interval evaluated in test 3 appeared to have a slightly higher transmissivity (1.0 m^2/day) than the slightly longer interval evaluated in test 1 (0.5 m^2/day).

Borehole USW H-4 was pumped at 17.4 l/s for 12,818 minutes (Whitfield et al., 1985). The tested interval included the full section of Crater Flat Tuff, and yielded an apparent transmissivity of 200 m^2/day by the straight-line method and 790 m^2/day by the Theis-recovery method (Figures 3.1.2.2.3-31 and 3.1.2.2.3-32 and Table 3.1.2.2.3-2). Dividing by the 700 m thickness of the tested interval gives apparent hydraulic conductivities of 0.29 and 1.1 m/day, respectively. Water-producing zones extended from the water surface at 519 m, in the Prow Pass member, to the bottom of the test hole at 1219 m in the Lithic Ridge Tuff.

Two successful pumping tests were conducted in borehole USW H-5 after the test well had been completed to its total depth of 1219 m (Robison and Craig, 1991); these are designated tests 3 and 4 since tests 1 and 2 were stopped prematurely due to generator failures. Pumping test 4 was conducted after additional perforating of the casing below the water table, and comparison of results from tests 3 and 4 appear to indicate that the original perforations were inadequate to transmit the available water. Drawdown response to pumping was consistent with two alternative conceptual models, an anisotropic unconfined aquifer or an aquifer containing a large, finite-conductivity vertical fracture (Figures 3.1.2.2.3-33 through 3.1.2.2.3-48; Tables 3.1.2.2.3-2 and 3.1.2.2.3-3 [INN 3.1.2.2.3-3]). The indicated apparent transmissivity is about 35 m^2/day, with 90 percent of the total production coming from the Bullfrog member of the Crater Flat Tuff at depths between 690 and 836 meters. The pumping response of USW H-5 indicates that fracture conductivity may be large and that nonradial flow responses need to be considered. Analyses of packer-injection tests performed in test well USW H-5 indicate disadvantages of using a long water column that were not apparent during previous testing of other wells near Yucca Mountain. These include dilation of fracture apertures, flow-velocity effects on transducer readings, and inertia and friction effects (Robison and Craig, 1991).

Two pumping tests and one recovery test (Figures 3.1.2.2.3-49 through 3.1.2.2.3-52 and Table 3.1.2.2.3-2) were conducted in test well USW H-6 after the well had been drilled to its total depth of 1220 m (Craig and Reed, 1991). No recovery data were obtained from the first test due to premature shutdown of the pump while drawdown-monitoring equipment was temporarily out of the well. During these tests, the well was cased to a depth of 581 m, with casing perforated in the depth interval from 530 to 572 m. Two zones of primary production in the Crater Flat Tuff were evident from borehole-flow surveys; the depth interval from 616 to 631 m in the Bullfrog Member produced 60 percent of the total flow, and the depth interval from 777 to 788 m in the Tram Member produced 32 percent of total flow. No measurable flow (greater than 0.05 l/s) was detected below a depth of 803 m. Two additional pumping and recovery tests (Tests 3 and 4) were conducted after the producing zones had been isolated with retrievable plugs and packers. Test 3 tested the depth interval from 753 to 834 m, and test 4 tested the depth interval from 608 to 645 m. Results of tests 3 and 4 are given in Figure 3.1.2.2.3-53 through 3.1.2.2.3-56.

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From analysis of the first two pumping tests in USW H-6 based on the transitional part of a dual-porosity solution, calculated transmissivity is about 240 m^2/day. However, the accuracy of this value is uncertain (Craig and Reed, 1991). Maximum apparent transmissivity is probably about 480 m^2/day, based on a recharge-boundary model. The 777 to 788-meter depth interval was tested with packers set, isolating the borehole from 753 to 834 meters. From this test, transmissivity was calculated to be 76 m^2/day, based on a dual-porosity model without fracture skin. If this value of transmissivity is assumed to apply entirely to the 777 to 788-meter interval using a homogeneous radial flow model, the apparent hydraulic conductivity is calculated to be about 7 m/day. A larger interval encompassing the 616 to 631-meter depth interval was tested, but testing problems resulted in the data being not analyzable.

3.1.2.2.4 Recharge and Discharge

[Text for Subsections 3.1.2.2.4 through 3.1.2.2.6 are not provided; the text will be developed in future revisions of the Annotated Outline.]

3.1.2.2.5 Age of Regional Groundwater

3.1.2.2.6 Groundwater Flow Paths

3.1.2.2.7 Regional Paleohydrology

Lowering of groundwater levels during the Quaternary Period

There is geologic evidence for former high levels of the water table in the south-central Great Basin (Figure 3.1.2.2.7-1) [INN 3.1.2.2.7-1] during the Quaternary Period. The evidence consists of tufas, spring orifices, calcitic veins, and cylindrical-lined tubes that mark the routes of former groundwater flow to springs; former water levels marked by calcitic cave deposits on the walls of Devils Hole (a fault-controlled collapse feature at Ash Meadows); and widespread marsh deposits (composed of chalk and authigenic clay) precipitated from former groundwaters. Most of these features have been described in reports by Winograd and Thordarson (1975), Dudley and Larson (1976), Winograd and Doty (1980), Khoury et al. (1982), Pexton (1984), Winograd et al. (1985), and Hay et al. (1986). The paleohydrologic significance of these deposits is the subject of papers by Winograd and Doty (1980), Pexton (1984), and Winograd and Szabo (1986).

Winograd and Doty (1980) mapped tufa and vein calcite in the Ash Meadows area of the Amargosa Desert, south of Highway 95 and east of Highway 29. They concluded that the potentiometric surface in the lower carbonate aquifer was once apparently as much as 50 m higher than the highest modern water levels in the Ash Meadows discharge area namely that in Devils Hole (altitude 719 m). Uranium disequilibrium dating of several of these calcitic veins (AM-7, AM-10, and DH-1 in Figure 3.1.2.2.7-1) show that the higher stands of water levels occurred during Pleistocene time, 1 million to 500,000 yr ago (Pexton, 1984; Winograd et al., 1985; and Winograd and Szabo, 1986).

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Still higher and older groundwater levels are recorded throughout the central Amargosa Desert by widespread marsh deposits of groundwater origin. These deposits are described by Khoury et al. (1982) and Pexton (1984). The marsh deposits have been referred to as lacustrine or playa deposits (Walker and Eakin, 1963; Denny and Drewes, 1965; and Swadley, 1983). Hay et al. (1986) present evidence based on stratigraphy, mineral composition, fossil assemblages, and lithofacies relationships that these sediments were deposited in marshes or ponds fed by groundwater rather than in lake or playa environments. In this report, they will be referred to as marsh deposits. The marsh deposits occur at altitudes as high as 790 m (Winograd and Doty, 1980) in Amargosa Flat in the east-central portion of the Amargosa Desert. This is about 70 m above the highest modern water level in Devils Hole (altitude 719 m).

A sample from the south end of Crater Flat described as nodular tufa spring deposits has been dated by the uranium-series method to approximately 30,000 yr B.P. (Szabo et al., 1981). Another sample described as seep deposited tufa or calcrete from the south of Yucca Mountain was dated to 78,000 $\pm$ 5,000 yr B.P. (Szabo et al., 1981). These dates may not be correct, however, because this dating method is only considered accurate if the material is from a closed chemical system (Swadley et al., 1984; Bradley, 1985). Szabo et al. (1981) state that these types of deposits may form open systems. In the Ash Meadows area, these marsh deposits range in age from about 2 to 3 million years (Pexton, 1984) and presumably mark the altitude of regional groundwater discharge during late Pliocene and perhaps earliest Pleistocene time.

The occurrence of the various deposits, marking Quaternary water tables, at altitudes tens to over 100 m above the modern water table probably resulted from a combination of tectonism, water-table lowering due to increasing aridity since Pliocene time, and erosion. Winograd and Szabo (1986) believe that the altitude of the water table-related deposits at Furnace Creek Wash (Figure 3.1.2.2.7-1) principally reflect uplift, rather than an actual lowering of the water table. However, they infer that regionally the water table in the lower carbonate aquifer is likely to have lowered during the Quaternary Period. The following evidence is cited in support of their hypothesis:

a. The several-thousand-meter topographic relief in Death Valley dates largely from Pliocene and Pleistocene time (Hunt and Mabey, 1966; USGS, 1984), and the movement of the floor of Death Valley probably has been downward relative to sea level and to bordering areas (Hunt and Mabey, 1966).

b. Interbasin flow of groundwater through Paleozoic carbonate rock and Tertiary welded-tuff aquifers toward Ash Meadows and Death Valley occurs today (Winograd and Thordarson, 1975). Interbasin flow of groundwater toward Death Valley likely also occurred during the Quaternary Period in response to the lowering of groundwater discharge outlets; as a result, Death Valley gradually evolved during the Quaternary into a regional sump for groundwater.

c. An increase in aridity due to uplift of the Sierra Nevada and Transverse Ranges would have resulted in a reduction of groundwater recharge.

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Another climate-related control on the decline of the regional water table may have been the level of Lake Manly, a Pleistocene lake that filled Death Valley to about 120 m above sea level (Hunt and Mabey, 1966). With the drying of the lake, groundwater base level in Death Valley would have declined. The exact age of Lake Manly and its duration are not well known. Hunt and Mabey (1966) considered the lake to be no older than the Wisconsin. Hale (1985) suggests that it is much older, perhaps of early- to mid-Pleistocene age. That the lake is either very old, had a short duration, or both, is suggested by the paucity of geomorphic and sedimentologic evidence for its existence (Hunt and Mabey, 1966).

Winograd and Szabo (1986) also suggest that the regional water-table decline postulated for the lower carbonate aquifer during the Quaternary Period was accompanied by a water-level decline in the overlying Cenozoic welded-tuff and valley-fill aquifers. They note that "The suggested progressive lowering of the regional water table throughout the Quaternary does not preclude superimposed and relatively rapid cyclical fluctuations in water level in response to the glacial (i.e., pluvial) and interglacial climates of the Pleistocene."

Evidence for increased aridity in the south-central Great Basin is discussed here because of its importance to the above-cited inference of water-table lowering throughout the Quaternary Period. Huber (1981) demonstrated that the central Sierra Nevada has been rising since early Miocene time (23.7 million years ago) and that the rate of uplift has been increasing from approximately 0.03 m/1,000 yr during early Miocene to 0.35 m/1,000 yr at present. In the past 3 million years the Sierra Nevada has risen about 1,000 m. Because the major source of cool season (November to May) precipitation for the Great Basin is the Pacific Ocean, this major uplift is likely to have significantly reduced precipitation in the region. Warm-season precipitation, which is derived largely from the Gulf of California and the Gulf of Mexico, is not considered important because this moisture is unlikely to contribute much to groundwater recharge (Winograd and Riggs, 1984). G.I. Smith et al. (1983) suggest that when the Sierra Nevada was about 1,000 m lower (3 million years ago), about 50 percent more moisture might have moved into the Great Basin. Based on paleobotanical records, Raven and Axelrod (1978) and Axelrod (1979) believe that aridity increased in the Great Basin, Mojave Desert, and Sonoran Desert during the late Tertiary and the Quaternary periods. They attributed this to uplift of the Sierra Nevada, Transverse Ranges, Peninsular Ranges, and the Mexican Plateau. On the basis of sediment depositional environments, Pexton (1984) believes that the Ash Meadows-Amargosa Desert area became more arid during the Quaternary Period. Winograd et al. (1985) describe a major and progressive depletion in the deuterium content of groundwater recharge in the Spring Mountains (Figure 3.1.2.2.7-1) during the Quaternary Period, and the most likely explanation for this is a decrease in Pacific moisture due to uplift of the Sierra Nevada and Transverse Ranges.

Lowering of the water table at Ash Meadows might have occurred during the Quaternary Period, even in the absence of vertical tectonics or increased aridity. Winograd and Szabo (1986) cite Winograd and Doty (1980), noting that "the major springs at the Ash Meadows oasis (Figure 3.1.2.2.7-1), differ in altitude by as much as 35 m and are as much as 50 m lower than the water level in Devils Hole. Thus, periodic initiation of discharge from new spring orifices (or an increase in existing discharge) in the lower portions of this oasis due to

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faulting and extensional fracturing would have resulted in new and lower base-levels for groundwater discharge." Implicit in this hypothesis is the belief that the faulting would be extensional opening new (or widening old) avenues of discharge from the buried Paleozoic carbonate rock aquifer that underlies eastern Ash Meadows and that feeds all the modern springs (Winograd and Thordarson, 1975). In support of this hypothesis, Winograd and Szabo (1986) note that most of the calcitic veins in Pliocene and younger rocks at Ash Meadows strike N. 40° ± 10°E. This is nearly at right angles to Carr's (1974) estimate of the direction of active extension in the region, namely N. 50°W. to S. 50°E. This mechanism may also have periodically lowered the water table in east-central Death Valley where the difference in altitude between the highest (Nevares) and lowest (Texas) major springs discharging from the regional carbonate aquifer is about 170 m (Winograd and Thordarson, 1975).

In summary, available data from calcitic veins and marsh deposits indicate a 50 to 70 m apparent lowering of the water table in the east-central Amargosa Desert since the middle Pleistocene and an apparent lowering of perhaps as much as 130 m since the end of the Pliocene. A strong inference can be made that the water-table altitude in the south-central Great Basin actually declined regionally throughout the Quaternary due to increasing aridity coupled with an absolute lowering of groundwater base level in Death Valley. This postulated general lowering does not preclude superimposed water-table rises in response to pluvial climates of the Pleistocene.

Groundwater flow paths during the Quaternary Period

The distribution of calcitic veins, tufas, and marsh deposits also suggest a lengthening of groundwater flow paths during Quaternary time. Regardless of whether the offset in the paleo-water table marked by these deposits was caused by water-level decline or land-surface rise, the occurrence of these deposits, kilometers to tens of kilometers upgradient from areas of modern groundwater discharge, indicates that flow to points of groundwater discharge were shorter in the past. Variations in flow-path length for the Ash Meadows area in response to different water-table levels are shown diagrammatically in Figure 3.1.2.2.7-2. [INN 3.1.2.27-2] Groundwater formerly discharged from the lower carbonate aquifer about 14 km northeast of Ash Meadows, the present discharge area. Based on the age of a calcite vein (AM-10 in Figure 3.1.2.2.7-1), this discharge occurred about 750,000 yr ago (Winograd and Szabo, 1986).

The widespread marsh deposits of the Amargosa Desert also show the location as well as the altitude of former groundwater discharge. In east-central Amargosa Desert (east of Highway 29 and south of Highway 95), these deposits were precipitated from groundwater discharging from the lower carbonate aquifer, either directly or indirectly via Tertiary aquitards. Like the calcitic veins, they occur as much as 14 km upgradient from Ash Meadows. West of Highway 29 and both north and south of Highway 95 the marsh deposits may reflect not only former discharge from the lower carbonate aquifer (which crops out in the west-central portion of the valley) but also from the Cenozoic welded-tuff and valley-fill aquifers. The

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marsh deposits in north-central Amargosa Desert are as much as 70 km from Alkali Flat, the modern discharge area in the southern Amargosa Desert.

Swadley and Carr (1987) mapped the Quaternary and Tertiary deposits in the Big Dune quadrangle, which includes the southern end of Crater Flat and the western part of the Amargosa Desert. They described the marl and silt deposits and the mounds of highly calcareous marl at the southern end of Crater Flat (the Crater Flat deposits) and north of Hwy 95 north from Roses Well (often referred to as the Horse Tooth Deposit). These and other deposits throughout the basin, including extensive deposits to the south by the California-Nevada stateline, are categorized as Pleistocene (?) and Pliocene Lacustrine Deposits that were deposited in former Lake Amargosa. They note that the Horse Tooth deposits contain vertebrate remains including *Mammuthus* sp., *Equus* sp, and a camelid; as well as noting abundant different diatoms and that some of the beds are highly tuffaceous. A sample from an ash bed was also taken from the "lake beds" in southern Crater Flat and was described as having remnant magnetic polarity normal, with no other analyses as of January 1985. Swadley (1983) identifies more of these Pleistocene (?) and Pliocene lacustrine deposits in the southern part of the Lathrop Wells quadrangle.

Hoover (1989) classified these deposits as the "waterlaid sediments of Amargosa marsh" and the boundaries of the Amargosa marsh correspond roughly to the boundaries of Swadley and Carr's (1987) Lake Amargosa. Hoover describes tuffaceous sands and clays overlain by a diatomaceous marl which is overlain by tufa north of Highway 95 (apparently the "Horse Tooth deposits" based on descriptions of the mammalian fossils). He notes that based on the vertebrate fossil evidence "deposition of the waterlaid sediments of the Amargosa marsh probably ended in early Pleistocene times" (Hoover, 1989).

Pleistocene Lakes

Reheis et al. (1993) describe the evidence supporting the existence of a pluvial lake in Fish Lake Valley, Nevada and California. The methods of detailed geologic mapping, analysis of tephra layers and correlation with known tephras, and examination of drill hole logs for evidence of lacustrine sediments were employed in the study. Reheis et al. informally named the lake Pluvial Lake Rennie. Tephrachronology indicates that the lake persisted from before 2.0 Ma until after 0.77 Ma. A highstand at an elevation of about 1,460 m with an areal extent of ~400 to 500 km^2 and a depth of ~250 meters at 0.77 Ma is documented by thick nearshore deposits of Bishop ash. After ~0.5 Ma, Pluvial Lake Rennie was either shallow or ephemeral. Comparisons with other pluvial lakes led Reheis et al. (1993) to suggest that this desiccation was due to the local cause of increasing rain-shadow effect on Fish Lake Valley from increase in elevation of the White Mountains and possibly the Sierra Nevada rather than regional climatic reasons.

Pluvial Lake Tecopa occupied a basin south of Yucca Mountain at the paleo-terminus of the Amargosa River (Morrison, 1991). Morrison estimates the drainage area at 6,000 km^2 and the highest stage covered an area of about 235 km^2. The lake existed from 3.0 Ma to later than 500 ka (Hillhouse, 1987; Dohrenwend, 1985). Based on a tephra layer (found in only one

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location in the basin), which has been correlated with an ash in a core from Tulelake, California and dated on the basis of Tulelake sedimentation rates, Morrison (1991) suggests that the lake achieved its highstand at about 160 ka. However, the youngest tephra layer found throughout the basin (Tuff A, correlated with the Lava Creek B ash from Yellowstone caldera) has been dated to 620 ka and a soil on an erosional surface of overlying alluvial fan deposits has been dated to 160 ± 18 ka by the uranium-trend method (Hillhouse, 1987; Dohrenwend, 1985). It is recognized that Lake Tecopa formed because its outlet was dammed and sometime after 500 ka the damming structure was breached and the lake drained (Hillhouse, 1987; Dohrenwend, 1985; Morrison, 1991). Calculations by Hillhouse (1987) indicate that even if the dam were still in place, water inflow into the basin is insufficient to balance evaporation and sustain a perennial lake in the basin. This suggests that precipitation or spring discharge must have been greater between 3.0 Ma to 0.5 Ma than it is now (Hillhouse, 1987).

Late Wisconsin Lakes, marsh deposits, and groundwater levels

During the late Wisconsin, 21,000 to 10,000 yr ago, average annual precipitation is believed to have been as much as 100 percent greater than modern, with average annual temperature as much as 6°C cooler (Spaulding et al., 1984). Not unexpectedly, changes in climate from late Wisconsin to present conditions have produced major changes in the hydrology of the region.

Mifflin and Wheat (1979) discuss pluvial lakes of late Wisconsin age in Nevada. In southern Nevada, they examined Frenchman Flat, Yucca Flat, Gold Flat, Kawich Valley, Sarcobatus Flat, Stonewall Flat, Emigrant (Groom Lake) Valley, Papoose Lake Valley, Pahrump Valley, and Indian Springs Valley, which generally surround the Yucca Mountain area. Of these valleys, they reported shoreline development only in Gold Flat, Kawich Valley, and Emigrant Valley. They report marsh deposits in Indian Springs, Pahrump, and Sarcobatus valleys. Mifflin and Wheat (1979) conclude that "southern Nevada did not contain perennial lakes during the Wisconsinan pluvial," but that "there were likely a number of areas of marsh environments in southern Nevada due to concentrated groundwater and spring discharge considerably in excess of present discharge. In south-central and southern Nevada significant differences in the amount and location of groundwater discharge seem apparent."

Hooke (1972) believed that a lake existed in Death Valley about 11,000 yr before present that left strand lines at an elevation of 90 m above msl on the Black Mountains (the current elevation of the low point in the Death Valley basin is 87 m below msl at Badwater). Smith and Street-Perrot (1983) agree that there was a lake between 12,000 and 21,000 yr before present but conclude that any lake or lakes were small and much shallower.

Quade (1986) reports marsh deposits in southern Indian Springs Valley. These deposits are assumed to be of late Wisconsin age, based on their stratigraphic and topographic similarity to dated deposits in northwestern Las Vegas Valley. They lie at a maximum altitude of 1,037 m, 20 to 50 m higher than the present water table in the valley-fill aquifer. If these deposits are of groundwater origin, as Quade believes, they may record a higher water table during late Wisconsin time. The marsh deposits occur above an inferred major hydraulic barrier

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(Winograd and Thordarson, 1975), and are 15 to 20 km downgradient from the Spring Mountains, a major recharge area. These factors may bear on the apparent water-table rise.

Mifflin and Wheat (1979) report marsh deposits, presumably of late Wisconsin age in Sarcobatus Flat and Pahrump Valley. They do not provide information on the location of these deposits so that we cannot compare their maximum altitudes with that of the preirrigation water table in these valleys.

Preliminary uranium-disequilibrium dating of calcitic cave deposits in Devils Hole, in central Ash Meadows suggests a 10-m rise in water level in the lower carbonate aquifer about 30,000 yr ago at Ash Meadows (Winograd and Szabo, 1986).

Estimation of future groundwater levels

Winograd and Doty (1980) using a variety of boundary conditions, suggest that during future pluvials, water levels in the regional carbonate aquifer at Ash Meadows and areas upgradient might be a few tens of meters higher than, or perhaps even lower than, modern levels. See Site Paleohydrology for estimation of possible future water levels in the Cenozoic volcanic rocks beneath Yucca Mountain and vicinity.

Winograd and Szabo (1986) concluded that there was a progressive lowering of the water table in the south-central Great Basin during the Quaternary. They noted that "a continued decline of the regional water table in the next 100,000 yr (and beyond?)" was likely.

Craig (1984) makes a comprehensive statistical-climatologic analysis of climate conditions affecting the Death Valley drainage systems that existed during the last glacial maximum. In one analysis, Craig modeled full glacial conditions without assuming a major decrease in lake evaporation rates. In this instance, precipitation increased by 90 percent and pluvial lakes formed in some valleys where lakes are known to have existed in the Late Pleistocene, but no lake formed in Death Valley and the lakes that resulted were not of a size "substantial enough to directly influence the regional groundwater potentiometric surface" (Craig, 1984). In an alternate analysis, Craig (1984) assumes a major reduction in lake evaporation rates. He states that, "If lake evaporation rates decrease by 27 percent, a major lake will form in Death Valley. This lake will be of an extent comparable to the highest recorded stand of Lake Manly during the Wisconsin." Under these climatic conditions, he states "recharge rates, and so groundwater flow conditions, will change significantly during a glacial event." Craig's (1984) basis for considering a 27 percent decrease in rate of evaporation from pluvial lakes was paleoclimatic evidence that evaporation was significantly lower during the Late Pleistocene. As validation for this assumption, he notes that the model predicted "a lake in Death Valley as large as is documented in the geologic record of the late Pleistocene." Craig's (1984) model predicts that such climatic conditions would have sustained lakes in Timber Mountain caldera, in Frenchman Flat, at Ash Meadows, and elsewhere. If Lake Manly and other major lakes existed in Wisconsin time, regional groundwater levels would indeed have been different. However, Lake Manly may be of early-to-middle Pleistocene age, when the Sierra Nevada was several hundred meters lower and the Transverse Ranges were

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not a topographic barrier to inflow of Pacific storms. Thus, the age of Lake Manly has bearing on the validity of Craig's (1984) alternate analysis for reconstruction of late Pleistocene paleoclimatology of the Death Valley area, and by extension for prediction of the effects of future pluvials on groundwater levels. Hunt and Mabey (1966) also note that "The slight erosion and sedimentation record of Lake Manly may mean that the lake was of brief duration and its level may have fluctuated rapidly. Whatever the cause, this California lake left one of the least distinct and most incomplete records of any Pleistocene lake in the Great Basin." Thus, even if Lake Manly is of Wisconsin age, the climatic changes it portends for the groundwater regime may have been short lived.

The paucity of geomorphic and sedimentologic evidence for Lake Manly makes it difficult to date the age of this major ancient lake in Death Valley. Resolving whether this lake is closer to 100,000 or 1 million years old would be of value to the paleohydrologic and paleoclimatologic analyses of Czarnecki (1985) and Craig (1984). Application of new or established methods, such as dating of desert varnish (Dorn, 1983) and uranium-series disequilibrium dating of lake-water cemented gravels, may hold some promise.

[INN 3.12.2.7-3] The reconnaissance (air photo) observations of Mifflin and Wheat (1979), regarding the absence of shorelines of late Pleistocene age in Yucca and Frenchman Flats and the presence of such shorelines in Gold Flat, Kawich Valley, and Emigrant Valley may need to be confirmed by more detailed study.]

Groundwater recharge during late Wisconsin time: carbon-14 evidence

Apparent groundwater ages determined from carbon-14 data (Benson and McKinley, 1985; Claassen, 1985) indicate that much of the groundwater within the tuffaceous rocks in the vicinity of Yucca Mountain and within the valley fill aquifer beneath the central Amargosa Desert was recharged during late Wisconsin time, about 10,000 to 18,000 yr ago. Holocene-age (less than 10,000 yr) water occurs in Fortymile Canyon at well UE-29 a#2 (4,000 yr before present) and at well J-12 (9,100 yr before present).

According to Claassen (1985), "little or no radiocarbon dilution probably occurred during the evolution of water in tuff or tuffaceous valley fill; therefore, the unadjusted ages are taken as true ages." He concluded that for groundwaters that contain as much as 20 percent of carbonate-derived constituents unadjusted ages could be as much as 1,000 yr too old, and that for a higher percentage there could be a serious problem in determining true ages from carbon-14 data. Since expected groundwater flow paths of waters sampled from Cenozoic aquifers in the vicinity of Yucca Mountain involve little or no contact with carbonate rocks, carbon-14 ages probably are near true ages, provided the well sampled did not tap water of more than one age.

Carbon-14 data from well USW H-6 (Benson and McKinley, 1985) show an increase in apparent groundwater age with depth in tuffaceous rocks. The interval 526 to 1,220 m (the entire saturated thickness penetrated) yielded water with an age of 14,600 yr; the interval 608 to 646 m yielded water 16,800 yr old; and the interval 753 to 853 m yielded water 18,500 yr

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old. A flow-meter survey (Benson et al., 1983) shows significant water yield between the water table (526 m) and a depth of 790 m. Because water pumped from the entire saturated zone (526 to 1,220 m) included older water from the two deeper zones, it is probable that water in the uppermost part of the saturated zone is considerably younger than 14,600 yr.

Groundwater with apparent age dates of 10,000 to 17,000 yr B.P. occurs in the valley-fill aquifer of the central Amargosa Desert (Claassen, 1985), 25- to 40-km downgradient from Yucca Mountain. Claassen (1985) believes, as did Winograd and Thordarson (1975), that groundwater in the valley fill represents pluvial recharge resulting from infiltration of surface-water runoff along the distributaries of Fortymile Wash. Therefore, groundwaters of about the same age as those beneath the upper zone of saturation near Yucca Mountain could have been emplaced 25 to 40 km downgradient.

[INN 3.1.2.2.7.4: Evaluate recharge along Fortymile Wash.]

[INN 3.1.2.2.7-5: Evaluate the apparent groundwater ages presented by Benson and McKinley (1985)]

Most of the wells sampled by Claassen (1985) that tap tuffaceous valley fill yielded water with apparent age less than 15,000 yr. Speculating on the absence of older water in the central Amargosa Desert, he considered that any water older than 17,000 yr B.P. might have moved downgradient, beyond his study area. Claassen (1985) favors an explanation based on climatic studies, concluding that "snowfall earlier than about 20,000 years B.P. was insufficient to result in snowmelt recharge, while subsequent climatic conditions caused such recharge."

Deducing groundwater velocities from the carbon-14 data is difficult. Claassen (1985) estimates a maximum velocity of 7 m/yr for the region between the head of Fortymile Canyon and the central Amargosa Desert. Claassen (1985) believes that actual velocities are probably much slower, but "because recharge may occur anywhere but not necessarily everywhere along surface drainageways, no probable minimum velocity can be calculated." Because recharge probably occurred by widespread infiltration along distributaries that resulted in layering of waters of differing ages, the estimation of groundwater velocities within the aquifer would be difficult even if test holes were available to selectively sample for carbon-14 ages of water with depth. If recharge of the same-age water can occur simultaneously along long reaches of Fortymile Wash and its distributaries in the Amargosa Desert, map distances may be irrelevant for calculating groundwater velocity. However, under such recharge conditions, it might be possible to draw conclusions regarding paleo versus modern recharge rates and the effects of groundwater velocity on the flushing out of old recharge water.

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Conclusions

In summary, the following conclusions can be drawn:

1. Evidence from dated calcitic veins and marsh deposits indicates that the regional water table (more correctly the potentiometric surface) in the south-central Great Basin has apparently declined 50 to perhaps 130 m during the Quaternary Period and that this decline has been accompanied by a lengthening of groundwater flow paths to points of discharge by 14 to perhaps 60 to 70 km. Considerations of the neotectonics of Death Valley, regional interbasin flow through the lower carbonate aquifer, and of the probable increasing aridity in the region during the Quaternary, collectively suggest that an actual and progressive water-table decline occurred in the region during this period.

2. Superimposed on the indicated long-term decline of water table are rises reflecting pluvial climates. In Devils Hole, at Ash Meadows, preliminary uranium-series dating of calcitic cave deposits indicates that about 30,000 yr ago the water table in the lower carbonate aquifer stood about 10 m above modern level. In southern Indian Springs Valley, marsh deposits, of presumed groundwater origin, occur as much as 20 to 50 m above the modern water table in the valley-fill aquifer; these deposits probably are late [INN 3.1.2.2.7-6] Wisconsin in age.

3. Reconnaissance observations indicate that pluvial lakes of late Pleistocene age were not present in Yucca Flat or Frenchman Flat, though such lakes apparently existed in Gold Flat, Kawich Valley, and Emigrant Valley, 70 km north and northeast of Yucca Mountain.

4. Groundwater of late Wisconsin age is present in the uppermost part of saturated volcanic rocks beneath Yucca Mountain and in the valley fill aquifer in the Amargosa Desert. The valley fill aquifer was recharged by surface-water runoff along the distributaries of Fortymile Wash.

[INN 3.1.2.2.7-7: A quantitative interpretation of the carbon-14 data to measure groundwater velocity will require test wells that permit water to be sampled selectively from several depths in the zone of saturation.]

3.1.2.2.8 Regional Groundwater Use

The regional aquifers in the hydrogeologic study area that are used for human activities are the valley fill aquifer and the lower carbonate aquifer. The welded tuff aquifer is locally important; it is developed only in the southwestern areas of the NTS in support of the former Nevada Research and Development Area.

For discussing water use in the hydrogeologic study area, the hydrographic regions delineated by the Nevada State Engineer are the most useful and practical units for consideration. For

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each of the hydrographic areas, the Nevada State Engineer has determined or estimated the perennial yield of the reservoir. The maximum amount of groundwater that can be appropriated from a given basin in Nevada is limited to its perennial yield. Thus, the relationship between the perennial yield and the amount of water being used forms the basis for groundwater management in Nevada. When these amounts are known, a quantitative statement of groundwater availability within a specified basin can be made.

Perennial yield (also referred to as safe yield; Todd, 1980) defined as the amount of water that can be withdrawn on an annual basis from a groundwater basin without depleting the reservoir (Scott et al., 1971). When the amount of withdrawals from a basin exceeds the perennial yield, an overdraft is created that can produce undesirable effects (Todd, 1980) since the water must come from storage within the aquifer. The result is known as aquifer mining and is prohibited in Nevada (Morros, 1982a). When aquifer mining (or over appropriation) is identified within a basin, the State Engineer may issue a statement known as a designation order. The designation order is a means of protecting the aquifer(s) from overuse by defining the boundaries of the area of overdraft and restricting the issuance of permits within that area. Since it is difficult to determine the perennial yield of a basin in which no pumping has occurred (Todd, 1980), aquifer overdrafts tend to be a common problem.

The following subsections discuss current water use for each community located within the three hydrogeologic subbasins of the study area, within the framework of the hydrographic areas. The users, location, rate, typical well construction and specific hydrogeologic unit source for each regional groundwater use are also identified whenever possible. When available, water withdrawals within each hydrographic area are described according to use (irrigation, municipal, livestock, domestic, and energy development). Figure 3.1.2.2.8-1 shows the locations of areas of large groundwater pumping near Yucca Mountain that are described in this section. There are no areas of large groundwater injection near Yucca Mountain.

Oasis Valley Subbasin. The Oasis Valley subbasin encompasses the western parts of the Gold Flat and Oasis Valley hydrographic areas, 147 and 228, respectively (Figure 3.1.2.2.8-2). Very little groundwater is used in this area because of the low population density.

The unincorporated community of Beatty, Nevada (population 1,623; U.S. Department of Census, 1991), receives most of its water from the Beatty Water and Sanitation District. The water is supplied from the Indian Springs well (not to be confused with the community of Indian Springs, Nevada) located approximately 6.4 km north of the town, the Summit (Bullfrog) well, a well in the town of Beatty and a new well recently drilled near the Summit well. The system supplies water to approximately 500 customers. Water from the four wells is mixed to meet the current State drinking water standards. Around 1985, well production capacity from the Indian Springs well decreased from 0.025 to 0.005 m^3/s due to overpumping and subsequent dewatering of the aquifer (Gram, 1985). The Indian Springs well is believed to be an outcropping of the regional carbonate aquifer (French, et al., 1984). Five commercial establishments (Bailey's Hot Springs, Fran's Star Ranch, Ray's Last Stand,

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Scotties's Junction, and Cottontail) also have water supplies that are available to the public. These establishments are all single users and generally serve transient populations of 25 or fewer people per day. Outlying ranches in the area around Beatty have their own wells. Rhyolite, a virtual ghost town located near Beatty, Nevada, is inhabited by two families. One family gets its water in Beatty and one family is served by a pipeline from a 75.8 m^3 water tank at Indian Springs (French et al., 1984).

At least three mines operating in the Beatty area require water for production purposes. Bullfrog Mine owned by LAC Minerals (formerly Bond Gold) is a gold/silver mining and milling operation that gets its water from four wells in Amargosa Valley. This water is used for milling, washing, and drinking. Another well is used to dewater LAC's underground operations. The LAC mineral facilities employ approximately 268 people (Nevada Bureau of Mines and Geology, 1993) and are expected to be in operation for another 5-10 years. Another mine in the area is the New Discovery Mine and Mill owned by Vanderbilt Minerals Corporation which mines specialty clays. The New Discovery Mine employs approximately 10 people (Nevada Bureau of Mines and Geology, 1993) and uses water from the Beatty municipal water system to fill a 0.95 m^3 fire-suppression tank. The U. S. Ecology Waste Site, a disposal area for low-level radioactive waste, has an on-site well used for laboratory facilities, potable water, and decontamination wash stations. The site is expected to be in operation for another 6 to 7 years. Smaller active and inactive claims may exist around Beatty, and the U. S. Air Force and the U. S. DOE also have water-using operations in this area.

Table 3.1.2.2.8-1 summarizes the known water use within the portions of Gold Flat and Oasis Valley basins that are located within the Oasis Valley subbasin. Note that although water uses do not currently exceed the perennial yield in either basin, Oasis Valley (hydrographic area 228) has been designated. The designation of this area by the Nevada State Engineer was a protective measure to prevent over appropriation.

Ash Meadows Subbasin. The Ash Meadows subbasin is located 42 to 60 km southeast of Yucca Mountain (Figure 3.1.2.2.8-1). Several small, unincorporated communities and a large part of the NTS, including the community of Mercury, are located within this subbasin. Very little water is withdrawn in this subbasin.

Most of the water pumped from the Ash Meadows subbasin is pumped from the lower carbonate aquifer (Dudley and Larson, 1976). Virtually all irrigation withdrawals of groundwater in this area were made by the Preferred Equities Corporation (Giampaoli, 1986). Water from eight major wells, supplemented by three small capacity wells, was used to irrigate more than 12 km^2 of cropland. Pumping peaked at a rate of approximately 1.2 million cubic meters per month during the years 1970 and 1971. Detailed pumping records were not maintained after 1971, while the land was controlled by Spring Meadows, Inc. (Dudley and Larson, 1976). In 1971, groundwater withdrawals associated with the planned development of a large agricultural enterprise caused a decline in the water level of the pool at Devil's Hole (Dudley and Larson, 1976). This natural pool, formed from the collapse of the limestone bedrock, is the only habitat of the endangered species, Devil's Hole pupfish,

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Cyprinodon diabolis. As a consequence of court action, groundwater withdrawals in this area are now restricted to a degree that is sufficient to maintain the water level in Devil's Hole (Dudley and Larson, 1976). The land, which was previously owned by the Preferred Equities Corporation, was turned over to Nature Conservancy in 1986 then transferred to the Federal Government (Giampaoli, 1986).

The NTS receives its water from wells operated by Reynolds Electrical & Engineering Company, Inc. The NTS accommodates a worker population of approximately 5,000 individuals, most of whom reside in Las Vegas and other nearby communities; a very small percentage of this workforce resides in Mercury on an intermittent basis. There are 12 NTS wells that currently withdraw water from the Ash Meadows subbasin for construction, drilling, fire protection, and consumption uses. Some of the water requires treatment before distribution. Table 3.1.2.2.8-2 summarizes that information for these 12 wells.

The unincorporated community of Indian Springs, Nevada (population 1,164; U.S. Department of Census, 1991) is located 80 km east-southeast of Yucca Mountain. The community receives its water from the Indian Springs Sewage Company, Inc., which is a public corporation owned by stockholders. As of 1993, there were 250 water customers, including several trailer parks with multiple connections. The company has three wells, two capable of pumping approximately 0.050 m^3/s and one well capable of pumping approximately 0.006 m^3/s. The water from these wells meets State drinking water standards and requires no chemical treatment. The water supply system is metered and is a separate system from that which supplies Indian Springs U.S. Air Force Base (French, et al., 1984).

Approximately 80 other shallow, domestic wells in the Indian Springs area are in use. Information regarding water quality and quantity for these wells is unavailable. All wells in the area are probably completed in the valley fill aquifer (Giampaoli, 1986), which is thought to be recharged locally by the lower carbonate aquifer (Winograd and Thordarson, 1975).

Indian Springs U.S. Air Force Base is located adjacent to the community of Indian Springs, Nevada. The population, including military personnel and their dependents, is approximately 900 persons. The water distribution system for this installation is owned and operated by the U.S. Air Force. Three wells, probably completed at shallow depths, supply water to the base. Two of the three wells are used for potable supply; the third well no longer provides potable water due to the presence of organic materials. All water distributed through the metered system is chlorinated to control bacteria (French et al., 1984).

Table 3.1.2.2.8-3 summarizes known water use in the hydrographic areas of the Ash Meadows subbasin. Very little water is pumped outside of the areas and communities previously discussed. Indian Springs Valley (hydrographic area 161 in Figure 3.1.2.2.8-2) is currently over appropriated. Actual water use in 1985 exceeded the perennial yield to the basin. Adverse effects resulting from overdraft include decline in the static water level and a reduction in spring discharges (Giampaoli, 1986). Information provided in Table 3.1.2.2.8-3 suggests that Yucca Flat is currently overdrafted. Well data in Table 3.1.2.2.8-3 indicate that these withdrawals may be from the lower carbonate aquifer. On the basis of this information

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and the fact that there is no detectable decline reported (Witherill, 1986), it is reasonable to conclude that Yucca Flat is not overdrawn. An overdraft also appears to exist for Frenchman Flat (hydrographic area 160 in Figure 3.1.2.2.8-2). All withdrawals in this area are from the valley-fill aquifer and have been in excess of perennial yield since 1964 (Claassen, 1973). There has not been any decline in the static water level, which suggests that the estimated perennial yield is too low (Claassen, 1973; Witherill, 1986).

Alkali Flat-Furnace Creek Ranch Subbasin. The Yucca Mountain Site is located within the Alkali Flat-Furnace Creek Ranch subbasin to the north of the Amargosa Desert (Figure 3.1.2.2.8-1). Very little groundwater is withdrawn in the northern and central parts of this subbasin. In addition, very little groundwater has been appropriated to the north of Yucca Mountain (upgradient) according to information filed with the Nevada State Engineer's Office.

The major groundwater users in the area, the town of Amargosa Valley and small rural communities of the northeastern Amargosa Desert, are located in the southwestern portion of the Alkali Flat-Furnace Creek Ranch subbasin. Most of the water to these users is supplied by wells; however, there has been some spring development. Most residences rely on individual wells, while some trailer parks, public facilities, and commercial establishments are served by small, private water companies. Table 3.1.2.2.8-4 summarizes the public water suppliers in the area, the type of well used, and the population served. Table 3.1.2.2.8-4 identifies the total number of known water wells drilled in the Amargosa Desert according to their defined uses. All wells are completed in and produce from the valley-fill aquifer.

Two mineral production operations are located in the Amargosa Desert. One operation, owned by the American Borate Corporation and located between Amargosa Valley, Nevada, and Death Valley Junction, California, was decommissioned in July, 1986. The facility consisted of a large mineral processing plant and a housing development for its employees (French et al., 1984). Water for the community was pumped from a shallow well with a capacity of 7.89×10^{-3} m^3/s. The water was treated by a reverse osmosis process before distribution to reduce the total dissolved solids. A complete analysis of water chemistry is not available, but the fluoride content is known to be greater than State drinking water standards allow. Groundwater was also used to supply the mineral processing plant and was obtained from three wells, with reported capacities of 1.64×10^{-2} m^3/s, 6.31×10^{-3} m^3/s, and 7.89×10^{-3} m^3/s. Well log data indicate that all the wells were completed in and produced from the valley fill aquifer (well logs are filed with the Office of the State Engineer). The plant closed permanently in July of 1986 (Pahrump Valley Times Star, 1986) and water use has ceased. The other operation is owned by IMV Division of Floridin, Inc. and is also located between Amargosa Valley, Nevada and Death Valley Junction, California. The operation employs approximately 53 people to mine specialty clays (Nevada Bureau of Mines and Geology, 1992).

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A small portion of the NTS receives its water from wells drilled in the Alkali Flat-Furnace Creek Ranch basin. Five wells, located in the western areas, supply water for drilling, construction, fire protection, and consumption uses. Table 3.1.2.2.8-5 summarizes the relevant data.

In addition to well production, a number of springs supply water to the region. The main concentration of springs is in Death Valley in the vicinity of Furnace Creek Ranch, approximately 50 to 60 km southwest of the Yucca Mountain Site. Many points of groundwater discharge have been identified in the Death Valley National Monument in California (Winograd and Thordarson, 1975). The water supply for the National Park Service facilities is derived principally from three groups of springs: Travertine Springs, Texas Springs, and Nevares Springs (French et al., 1984). The population served by this water supply varies during the year. From October through April, approximately 800 persons live in the area on a semipermanent basis, and an additional 2,000 persons live in the area as visitors. From May through September, the number of semipermanent residents decreases, and there are few visitors (French et al., 1984).

Three resorts are located within the boundaries of the Death Valley National Monument: the Stovepipe Wells Hotel, Furnace Creek Inn, and Furnace Creek Ranch. At Stovepipe Wells Hotel, northwest of Furnace Creek Ranch, potable water was trucked in for many years from a storage tank in the alluvial fan south of Stovepipe Wells Hotel and began trucking water to this tank from Nevares Spring. No other improvements to the system have been documented. Pistrang and Kunkel (1964) report that water from an excavated sump lined with buried drainage tile in the Furnace Creek Wash is conveyed in a buried pipeline to the Furnace Creek Inn and Furnace Creek Ranch resort. The measured flow rate ranged from 1.20×10^{-2} m^3/s to 1.30×10^{-2} m^3/s.

Water use within the Alkali Flat-Furnace Creek Ranch subbasin occurs primarily in the Amargosa Desert (hydrographic area 230 in Figure 3.1.2.2.8-2). The perennial yield to this hydrogeographic area is estimated to be 2.96×10^7 m^3 per year (French et al., 1984). An estimated 2.10×10^7 m^3/s per year of this total is naturally discharged from springs and seeps in the Ash Meadows area (French et al., 1984), and nearly 10,000 acre-feet per year is artificially discharged from wells in the Amargosa Valley (Coache, 1986). Thus, an overdraft of 3.70×10^6 m^3 per year currently exists. Water levels in wells drilled in the valley fill aquifer declined an average of 3.75 m between 1963 and 1984 (Nichols and Akers, p. 5, 1985). Total appropriations in 1985 were over 8.64×10^7 m^3 per year; if these rights to appropriate water were exercised, rapid depletion of the valley fill aquifer would result. Given the current and projected population estimates (Smith and Coogan, 1984; DOE/RW-0073), this scenario is highly unlikely.

Crater Flat (hydrographic area 229 in Figure 3.1.2.2.8-2) is currently overdrawn because of an appropriation made to Saga Exploration, Inc. for the development of the Sterling Mine, located on the east side of Bare Mountain. The mine uses its own well for its heap-leach operation and relies on municipal water for its potable water. The mine employs approximately 40 individuals, and is expected to be in operation until 1997 or 1998.

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Although an overdraft exists, no protective measures will be taken because the water has been appropriated for mining, which is considered a preferred use under the Nevada Revised Statutes. Under the statutes, overdrafts for mining are allowable for periods not to exceed five years.

Future applications to appropriate groundwater in the Amargosa Desert area will be carefully regulated by the Nevada State Engineer; the majority of applications for irrigation permits submitted through 1982 have been denied (Morros, 1982b).

In summary, although the Alkali Flat-Furnace Creek Ranch subbasin as a whole is over appropriated, actual annual water use is less than 30 percent of the perennial yield. The over appropriation problem exists primarily in the Amargosa Desert (hydrographic area 230 in Figure 3.1.2.2.8-2). Outside of the Amargosa Desert, a small number of ranches and domestic users withdraw small amounts of groundwater. The amount of water withdrawn in hydrographic areas other than the Amargosa Desert represents only a fraction of the perennial yield available to those areas (see Table 3.1.2.2.8-6) [INN 3.1.2.2.8-1]. [Update water appropriations data when available.]

The local effects of groundwater withdrawals vary among the different areas previously discussed. Some recent potentiometric maps of the hydrogeologic study area are shown in Figures 3.1.2.2.2-2 and 3.1.2.2.8-3 (Ervin et al., 1993 and Czarnecki, 1990, respectively). No drawdown is shown on either map because of limited pumping. Figure 3.1.2.2.8-4 shows the potentiometric surface in the Amargosa Desert as measured from wells in 1962 and in 1984. A decline in the water level and one depression cone can be seen. The decline in water levels, and the formation of a depression cone, resulted from withdrawals made by 27 wells in the valley. Pumping rates for domestic wells range from 6.31×10^{-4} to 2.53×10^{-3} m^3/s while industrial wells produce from 6.31×10^{-3} to 5.05×10^{-2} m^3/s (including irrigation and mining wells) (Giampaoli, 1986). No injection wells exist in the hydrogeologic study area.

3.1.2.2.9 Regional Groundwater Management Plans

The State of Nevada is responsible for protecting groundwater resources. Chapter 534 (Title 48-Water) of the Nevada Water Laws outlines and delineates the allowable uses of groundwater. Morros (1982a) presents the laws and statutes relating to all aspects of groundwater and geothermal resources, as described in Chapters 534 and 534A, respectively. Water use in the State of Nevada is governed by the Office of the State Engineer in the Division of Water Resources. A permit from the State Engineer is required to appropriate groundwater. Permits may be requested for domestic, quasi-municipal, municipal, commercial, irrigation, and agricultural purposes. In larger, incorporated communities, water supply is governed through the Nevada Water Laws by municipal water supply systems. In smaller communities, the appropriate Board of County Commissioners governs water supply. According to the Nevada Water Laws, all groundwater belongs to the public to be put to beneficial use only, subject to existing use (Morros, 1982a).

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The State of Nevada uses a recharge-use philosophy in its management of groundwater resources. Total annual groundwater withdrawals from any given basin may not exceed the perennial yield (Morros, 1982a). Unfortunately, there are many difficulties inherent in determining the actual amount of recharge to an undeveloped basin where no pumping has occurred. In many instances, aquifer overdrafts occur before the perennial yield has been firmly established (Todd, 1980).

At the first indication of an aquifer overdraft, measures are taken to carefully regulate groundwater withdrawals. A provision of the Nevada Water Laws grants authority to the State Engineer to designate groundwater basin boundaries in areas where he deems that groundwater resources are being depleted. Such designation orders have been issued in the following areas:

- The Amargosa Desert Groundwater Basin
- The Pahrump Artesian Basin
- Oasis Valley West
- Indian Springs Valley.

Designation of a groundwater basin places permits on a preference basis, but does not establish which types of permits have priority. Preference is determined on a case-by-case basis by the State Engineer. Current granting of permits in restricted areas indicates that irrigation and agricultural use have the lowest preference while domestic permits have the highest. In designated basins, quasi-municipal permits, permits for water resources for two or more houses are no longer available. Revocable permits, permits that can be revoked when municipal water is available, are currently granted by the State Engineer in designated and non-designated areas; non-revokable permits are no longer an option. The State Engineer is prohibited by law from granting a permit where there is no unappropriated water in the proposed sources, the proposed use conflicts with existing rights, or the proposed use threatens to prove detrimental to the public welfare (Morros, 1982b).

Regional agencies that have an interest in groundwater management include the U.S. Fish and Wildlife Service, the National Park Service, the Las Vegas Valley Water District, and the DOE. The U.S. Fish and Wildlife Service is concerned with the Ash Meadows Wildlife Conservation Area; the National Park Service is responsible for Devil's Hole, a part of Death Valley National Monument; the Las Vegas Valley Water District was set up through the Nevada Water Laws to provide groundwater through a quasi-municipal permit to Las Vegas Valley; and the DOE is responsible for groundwater protection and restoration at the NTS.

The U. S. Fish and Wildlife Service is currently monitoring water levels in the Ash Meadows area to establish baseline springflow and groundwater levels. The National Park Service is also monitoring in the Ash Meadows and Devil's Hole area. A monitoring program was set up jointly by the National Park Service and the DOE in 1991 to monitor the impact of groundwater pumpage at the Yucca Mountain Site on groundwater levels and springflow in the YMR. This plan is outlined in "Monitoring Program for Groundwater Levels and Springflows in the YMR of Southern Nevada and California," (DOE, 1991).

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The Las Vegas Valley Water District uses a quasi-municipal permit issued to the State Engineer to appropriate water for the Las Vegas Valley Water District. The Las Vegas Valley Water District is expected to continue to emphasize development of groundwater resources and use of a recharge system to recharge appropriated water to the regional aquifer so that this water can be utilized when needed.

The DOE is also responsible for groundwater restoration and management on the NTS. The "Groundwater Protection Management Program Plan for the DOE Nevada Field Office," currently a draft document and the "Environmental Restoration and Waste Management Site Specific Plan, Fiscal Years 1994-1998," (DOE, 1992) outline the plans to manage and restore groundwater on the NTS. The "Groundwater Protection Management Program Plan" reviews plans for protection and remediation of groundwater at the Yucca Mountain Site. The "Environmental Restoration Waste Management Site Specific Plan" looks at unique plans to assess and clean up specific contaminated areas at the Yucca Mountain Site and other facilities operated by the Nevada Operations Office.

Based on information obtained from regional agencies, their groundwater management plans and other sources, an assessment of regional groundwater use projections for the foreseeable future can be made. These projections will focus on the relevant hydrographic areas of the Death Valley groundwater system and the hydrogeologic study area. The three main subbasins relevant to Yucca Mountain are: The Oasis Valley Subbasin, the Alkali Flat-Furnace Creek Ranch Subbasin, and the Ash Meadows Subbasin (see Figure 3.1.2.2.8-2). Projected water use due to the YMP and the town of Pahrump will also be discussed.

The Oasis Valley Subbasin (see Figure 3.1.2.2.8-2) was designated in 1980 by the State Engineer to prevent over appropriation. Current water use in the Oasis Valley subbasin is minor because of the low population density (Giampaoli, 1986). Recharge to the subbasin is estimated to be approximately 2.47×10^6 m^3 (DOE, 1988).

The main population center in the Oasis Valley Subbasin is Beatty, Nevada (population 1,623; U.S. Department of Census, 1991) is unincorporated and is governed by the Nye County Commissioners. According to French et al. (1984), most of Beatty is employed by mining and service industries, the U. S. Ecology Waste Site, the DOE, and the U. S. Air Force. Future growth of Beatty and corresponding water needs depend on the growth of nearby mines.

The Alkali Flat-Furnace Creek Ranch Subbasin is located to the North of the Amargosa Desert and includes the Yucca Mountain Site (see Figure 3.1.2.2.8-2). Trailer parks, public facilities, and commercial establishments are generally served by small, private water companies while most of the residences rely on individual wells.

A small portion of the NTS receives its water from wells drilled in the Alkali Flat-Furnace Creek Ranch basin. The maximum annual water demand for the repository is anticipated to rise to a peak of 5.55×10^5 m^3 per year by the end of the seventh year of construction, decrease to about 4.36×10^5 m^3 per year, and remain at this level for the next 25 years. The

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minimum average water demand for the following 23 years of operation would be approximately 9.47 x 10^3 m^3 per year (Morales, 1986). Estimates of the quantity of water needed to support site characterization activities are considerably less. Preliminary estimates indicate that maximum total water use for all phases of the exploratory shaft construction and testing will be approximately 1.25 x 10^6 m^3 (see Table 3.1.2.2.9-1) (State of Nevada, Exhibit #40, 1992).

Water use within the Alkali Flat-Furnace Creek Ranch Subbasin occurs primarily in the Amargosa Desert (hydrographic area 230 in Figure 3.1.2.2.8-2). The perennial yield to this hydrogeologic area is estimated to be 2.96 x 10^7 m^3 per year (French et al., 1984). An estimated 2.10 x 10^7 m^3 acre-feet per year of this total is naturally discharged from springs and seeps in the Ash Meadows area (French et al., 1984), and nearly 1.23 x 10^7 m^3 per year is artificially discharged from wells in the Amargosa Valley (Coache, 1986). Thus an overdraft of 3.70 x 10^6 m^3 per year currently exists. Total appropriations in 1985 were over 8.64 x 10^7 m^3 per year; if these rights to appropriate water were exercised, rapid depletion of the valley fill aquifer would result. Given the current and projected population estimated (Smith and Coogan, 1984), this scenario is highly unlikely.

Crater Flat (hydrographic area 229 in Figure 3.1.2.2.8-2) was overdrawn because of an appropriation made to Saga Exploration, Inc. for the development of the Panama-Stirling Mine, located on the east side of Bare Mountain. Although an overdraft existed, no protective measures will be taken because the water has been appropriated for mining, considered a preferred use under the Nevada Revised Statues. Under the statutes, overdrafts for mining are allowable for periods not to exceed five years.

In summary, although the Alkali Flat-Furnace Creek Ranch Subbasin as a whole is over appropriated, actual annual water use is less than 30 percent of the perennial yield. The over appropriation problem exists primarily in the Amargosa Desert hydrographic area. Outside the Amargosa Desert, a small number of ranches and domestic users withdraw small amounts of groundwater. The amount of water withdrawn in hydrographic areas other than the Amargosa Desert represents only a fraction of the perennial yield available to those areas.

The Ash Meadows Subbasin includes several small, unincorporated communities such as Indian Springs, a large part of the NTS including the community of Mercury, and the Ash Meadows Wildlife Conservation Area. The Indian Springs Valley and Amargosa Desert hydrographic areas have been designated by the State Engineer to prevent over appropriation (see Figure 3.1.2.2.8-2).

Indian Springs, Nevada (population 1,164; U.S. Department of Census, 1991) is located 80 km east-southeast of Yucca Mountain. The community receives its water from the Indian Springs Sewage Company, Inc. which is a public corporation owned by stockholders. Over 80 wells appropriate water for the community of Indian Springs, the Indian Springs Sewage Company, Inc., and the U. S. Air Force Base. Water-level declines in the Indian Springs community caused a decrease in local spring discharge (Giampaoli, 1986). Although the

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amount of water appropriated is minor, total water withdrawals and natural discharge in the form of springs and evapotranspiration are in excess of the perennial yield.

The Amargosa Desert groundwater basin was designated by the State Engineer in 1979 to prevent over appropriation of water in the Ash Meadows Wildlife Conservation Area, the Amargosa Valley, and by rural water users in the Amargosa Desert. Although no order of preference has been established by the State Engineer, permit applications on file for wells located within the Amargosa Desert groundwater basin indicate that irrigation and agricultural use have the lowest preference. Water use at the NTS and the Yucca Mountain Site is not affected by the designation order.

The NTS receives its water from wells operated by the Reynolds Electrical & Engineering Co., Inc. Aside from the Yucca Mountain Site, which is located in the Alkali Flat-Furnace Creek Ranch Subbasin, water use is not expected to increase on the NTS in the future.

Little of the groundwater in the Ash Meadows subbasin is currently privately appropriated or used. Since much of the land within this subbasin is under the jurisdiction of Federal Agencies and has been withdrawn from the public domain, future private appropriation of water is not expected.

The town of Pahrump, located approximately 100 km southeast of Yucca Mountain, has experienced groundwater overdraft problems in the past. Certified appropriations and development permits for groundwater in the Pahrump Valley totaled 1.12×10^8 m^3 in 1970, although in recent years actual water withdrawal has averaged approximately 4.90×10^7 m^3 per year (DOE/RW-0073). As the proposed site is within a different subbasin of the Death Valley groundwater system (Winograd and Thordarson, 1975; Harrill, 1982), the YMP is not expected to affect water resources in the Pahrump Valley.

3.1.2.3 Site Hydrogeology

3.1.2.3.1 Baseline Monitoring Network

The baseline monitoring network is a system of monitoring wells, vadose zone boreholes, and other installations used to gather information on the site hydrogeologic system. The function of the baseline monitoring network is three-fold: (1) continuing measurement of existing groundwater conditions to serve as an historical data base for future observational comparison, (2) long term monitoring of potentiometric levels and hydrochemistry to demonstrate the degree of stability of the geohydrologic system during the period required for waste isolation, and (3) water level observations to record any changes of the water table that would affect the overall groundwater travel time in both the unsaturated and saturated zones beneath the repository.

Eight deep monitoring boreholes have been drilled into the unsaturated zone (YMP Site Atlas, 1992) (Figure 3.1.2.3.1-1). Of those boreholes completed only one, borehole USW UZ-1, located north of the Yucca Mountain Site (36°52'08"N, 116°27'39"W, Elev. 4426.24 Ft.

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above msl), is currently being monitored. Monitoring is accomplished with a variety of instruments permanently emplaced at various levels within this borehole (see Figure 3.1.2.3.1-1). These instruments are connected to an automatic data acquisition system at the ground surface. Data collected during monitoring includes downhole temperature, pneumatic pressure, matric potential, and pore gas sampling for hydrochemical analysis. The instrumentation installed within the boreholes is designed to permit monitoring to continue for an indefinite period of time (after the site characterization program has been completed), as required for performance confirmation.

In addition to the deep UZ boreholes, 84 shallow (less than 30 m in depth) boreholes have been completed in alluvium, with the purpose of studying infiltration processes in the near-surface unsaturated zone (YMP Site Atlas, 1992) (Figure 3.1.2.3.1-2). Each hole is cased to total depth and logged for moisture content with a neutron (soil moisture) probe on a monthly basis, as well as after significant moisture events. The objective of the Unsaturated Zone Neutron holes is to estimate net-infiltration rates in surficial materials by monitoring changes in moisture content. The Unsaturated Zone Infiltration Program, combined with data from the site precipitation monitoring network, provides input on the spatial distribution of infiltration/recharge for the boundary conditions used in the three dimensional hydrologic model of the proposed repository block.

The saturated zone monitoring system is composed of a network of 25 holes from which water level measurements of the saturated zone are obtained (Table 3.1.2.3.1-1, [INN 3.1.2.3.1-1], Figure 3.1.2.3.1-3). Thirteen of the holes are monitored continuously, and the remainder are measured periodically. The monitored sites include a water supply well (J-13), a 1533 meter geologic test hole (originally constructed to obtain data on lithology and stratigraphy), eight hydrologic test holes, in which pumping and other tests were performed to determine hydraulic characteristics of the formations, one Paleozoic exploratory hole (UE-25P#1) and 14 water table holes areally distributed to enable definition of the potentiometric surface at Yucca Mountain. The original purpose of the drillholes now being observed was not for baseline monitoring, therefore, no formal selection process was used to determine their locations or depths (SCP). However, they are situated for general coverage of the upper part of the saturated zone, through which groundwater flow paths from the repository are likely.

The construction of the geologic and hydrologic series of these holes was such that early measurements of head were composites, reflecting averages of all the members of the Paintbrush and Crater Flat tuffs that were penetrated below the water table. Subsequently, a single semipermanent packer was installed between permeable zones in five of the seven hydrologic holes (drillholes UE-25b#1, USW H-3, USW H-4, USW H-5, and USW H-6), enabling comparison of head in the zones above and below the depth of the packer setting. Although they are open to different geologic zones, the potentiometric heads in most of the wells are believed to represent a laterally continuous aquifer, as the potentiometric head pattern forms a coherent surface (Ervin, et al., 1993). Potentiometric heads in the Tertiary rocks penetrated in drillhole UE-25p#1 were measured while drilling, but these rocks were sealed off before the well penetrated the underlying Paleozoic rocks, and present measurements reflect head only of the Paleozoic interval.

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3.1.2.3.2 Site Flow System Boundaries and Hydrogeologic Units

A flow system is a complex mixture of rock, soil, liquid and vapor phase water, and air. Energy to drive this system enters in the form of heat from solar radiation, gravitational potential, barometric pressure changes and geothermal gradient. The interactions between these segments are complex; however, by grouping similarly behaved elements of the system together, limiting inquiry to an area within a natural or contrived set of boundaries whose conditions are presumed to be known, and applying the fundamental principles describing flow and storage of fluids within permeable media, a conceptual model can be constructed which can be used to make predictions about the behavior of the system.

Within this conceptual framework, the flow system beneath the proposed Yucca Mountain repository can be described as a stratified block of east-dipping hydrogeologic units composed of welded and nonwelded tuffs of contrasting hydraulic properties. The block is bounded on the west by a westward-dipping high-angle normal fault (the Solitario Canyon fault), and on the east by a set (or zone) of imbricate westward-dipping, high-angle normal faults. The upper hydrologic boundary consists of the land surface, across which water enters the unsaturated zone as infiltration and leaves the system in the form of water vapor. The water table acts as the lower boundary of the unsaturated zone and the interface with the saturated zone. The Yucca Mountain block also contains one major (i.e., the Ghost Dance fault) and a number of minor west-dipping, high-angle normal faults that, depending on location and ambient hydrologic conditions, may be pathways for or barriers to moisture flow (Montazer and Wilson, 1984).

The rocks underlying Yucca Mountain through which the groundwater moves are mostly fractured tuffs of Tertiary age; because the physical properties of strata that control movement of groundwater are not completely consistent with stratigraphic boundaries, six unsaturated zone (Montazer and Wilson, 1984) and two saturated zone (Waddell et al., 1984) hydrogeologic units have been defined (Table 3.1.2.3.2-1), and are described below.

3.1.2.3.2.1 Unsaturated Zone Hydrogeologic Units

The unsaturated zone beneath Yucca Mountain consists of six hydrogeologic units of welded and nonwelded tuffs with contrasting hydraulic properties (Figure 3.1.2.3.2-1). Fracturing in these units varies from highly fractured (up to 40 fractures/m^3 for the Topopah Spring Welded Unit [TSw]) to relatively unfractured (approximately 1 fracture/m^3 for the Paintbrush nonwelded unit [PTn]). The fractures may either enhance or retard the flow of moisture. The following discussion of hydrogeologic units is adapted from work done by Montazer and Wilson (1984).

QAL. The QAL hydrofacies is composed of irregularly distributed deposits of alluvium and colluvium. QAL ranges up to 30 meters in thickness under the washes that dissect Yucca Mountain, but is thin or nonexistent over much of the site. Percolation through the alluvial deposits is slow (for example, see Kearl, 1982). Alluvial and colluvial deposits generally have small effective hydraulic conductivity, large specific retention, and large effective

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porosity as compared to the fractured rocks. Therefore, a large proportion of the water infiltrated into the alluvial and colluvial material is stored in the first few meters of the soil and is lost to evaporation during dry periods (Montazer and Wilson, 1984).

Tiva Canyon Welded Unit (TCw). Lying immediately beneath the QAL is the TCw, consisting of devitrified ash-flow tuffs ranging from 0-150 meters in thickness across the site. The TCw hydrofacies is composed of the densely to moderately welded part of the Tiva Canyon Member of the Paintbrush Tuff lithofacies; saturated matrix hydraulic conductivity for the TCw unit has been estimated at about 10^{-11} m/s (Montazer and Wilson, 1984, Peters et al., 1984, Peters et al., 1986). Hydraulic conductivity under field conditions is considerably less, as saturation is estimated to average about 60%. Although the TCw unit is highly fractured (Scott et al., 1983, estimate 10-20 fractures/m^3), under the low rates of net infiltration most water entering the TCw unit may be drawn into the matrix before it can traverse the entire unit thickness to recharge underlying units (Montazer and Wilson, 1984).

PTn. The PTn is situated below the TCw unit, and is composed of the nonwelded and partially welded base of the Tiva Canyon Member, the Yucca Mountain Member, the Pah Canyon Member, and associated bedded tuffs, all of which are part of the Paintbrush Tuff lithofacies. The PTn hydrofacies represents 20-100 meters of partially welded to nonwelded, vitric and occasionally divitrified tuffs with few fractures (approximately 1/m^3, Scott et al., 1983) and considerably higher saturated hydraulic conductivity than TCw (estimates by Montazer and Wilson, 1984, Peters et al., 1984, and Peters et al., 1986, are on the order of 10^{-7} m/s). Hydraulic conductivity in this unit demonstrates considerable anisotropy which, combined with the low matrix conductivity of the underlying Topopah Spring unit, contributes to lateral flow within the unsaturated portion of the block (Montazer and Wilson, 1984).

TSw. Underlying the PTn unit are the moderately to densely welded, devitrified ash-flow tuffs known collectively as the TSw. These tuffs, which vary from 290-360 meters in thickness, are locally lithophysae-rich in the upper part, and are moderately to highly fractured (Scott et al., 1983, estimate 8-40 fractures/m^3). Overall porosity is low (on the order of 10%) and saturated matrix hydraulic conductivity is comparable to that of the TCw unit at about 10^{-11} m/s.

Calico Hills Nonwelded Unit (CHn). Beneath the TSw unit is a series of non- to partially welded ash-flow tuffs, 100-400 meters in thickness, designated as the CHn. Locally, these may be vitric (CHnv) or zeolitized (CHnz). While fracture densities for these two sub-units are similar (2-3 fractures/m^3, Scott et al., 1983), saturated hydraulic conductivities differ by several orders of magnitude (approximately 10^{-8} m/s for CHnv verses 10^{-11} m/s for CHnz). Due to the dip of the beds and variations in the elevation of the phreatic surface, the water table intersects lower portions of the CHn unit under the northern-most portion of the site; further to the north, beyond the boundaries of the site, the unit is fully saturated.

Crater Flat Undifferentiated Unit (CFu). The CFu consists of 0-200 meters of undifferentiated welded and nonwelded, vitric, devitrified, and zeolitized ash-flow and air-fall tuffs. Montazer and Williams (1984) estimated saturated matrix hydraulic conductivity at

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$5x10^{-10}$m/s. The CFu hydrofacies is the lowest of the unsaturated units under the southern-most one-half of the proposed repository block; towards the north, the unit becomes fully saturated. Because of the limited extent of this unit beneath the central block, its effect on the unsaturated flow system is limited (Montazer and Williams, 1984).

3.1.2.3.2.2 Saturated Zone Hydrogeologic Units

Waddell et al. (1984) divided saturated groundwater flow into two hydrofacies: (1) moderately and densely welded tuffs and (2) nonwelded and bedded tuffs. The moderately and densely welded tuffs lying below the water table are characterized by relatively low porosities, abundant fractures, and low matrix permeability. In nonwelded and bedded tuffs, porosities are greater and fractures are fewer, although fractures do occur. Unless the rock contains clays or zeolites, matrix conductivities are generally greater than in the welded tuffs. Based on laboratory tests on core from borehole UE-25a#1, saturated matrix hydraulic conductivities for the Bullfrog Member of the Crater Flat Tuff are between 10^{-8} and 10^{-9} m/s (Waddell, 1994). Cores from the same borehole yielded porosities of 15 to 25 percent. These values were determined from the most productive well zone; other zones are proportionately less permeable.

Beneath the saturated portions of the CFu and other, hydrologically similar units lie the Paleozoic carbonate rocks (Craig and Robison, 1984), equivalent to the "lower carbonate aquifer" of Winograd and Thordarson (1975). In borehole UE-25p#1 these carbonate rocks are intersected approximately 1200 meters below the land surface; due to structural complications, however, the Paleozoic carbonates generally exist at 6000+ meters in depth (Robinson, 1985). While these carbonate rocks are permeable, the potentiometric head is greater than in the overlying units, indicating small vertical permeability, with possible small upward groundwater flow. The groundwater from the upper saturated zone, therefore, does not have significant local interaction with water from the carbonate rocks.

3.1.2.3.3 Potentiometric Levels, Matric Potentials, and Gradients

Measurements of metric potential, potentiometric head, and other saturated and unsaturated zone potentials at the site of the proposed Yucca Mountain repository have been made since the mid-1980's. As more data has become available, an increasingly complex picture of the site flow system has emerged; the following subsection is intended to provide an overview of important site gradients and how they relate to our understanding of the system.

3.1.2.3.3.1 Unsaturated Zone Potentials [INN 3.1.2.3.3.1-1]

In the unsaturated zone, water is present both in liquid and vapor phases within the interstitial, fracture, and lithophysal openings. Water flow and storage is complexly three dimensional, and is controlled by the structural, stratigraphic, and climatological setting. The hydrologic evaluation of the proposed repository site constitutes a problem of two-phase, multi-component, coupled heat (geothermal) and water flow within a layered sequence of tilted, faulted, and fractured, variably saturated, tuffaceous geohydrologic units.

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In order to understand the movement and occurrence of water under such complex conditions, a detailed knowledge of the unsaturated zone potentials is imperative. Data from boreholes USW UZ-1, -4, -5, -6, -6s, -7, and -13 has provided much useful information; however, detailed in situ measurements of matric potential have been taken only at the 33 instrumentation stations vertically arrayed within borehole USW UZ-1. This borehole was drilled without water using a reverse air vacuum system to minimize disturbing local hydrologic conditions near the borehole (Whitfield, 1985). It was drilled to a total depth of 387 meters, whereupon drilling was discontinued because an apparent perched-water zone was encountered. Following drilling and geophysical logging, the borehole was instrumented at 33 depth levels over a total depth of 371 meters (Montazer et al,. 1986). At 15 of these levels, 3 well screens were embedded in dry, coarse sand columns that were isolated from each other by thin layers of bentonite, columns of silica flour, and isolation plugs consisting of expansive cement. The upper and lower screens within each of these 14 levels were equipped with access tubes for gas sampling. Thermocouple psychrometers to measure matric potential were emplaced in the upper and middle screens and pressure transducers to measure pore-gas pressure in the middle and lower screens at each level. In addition to these instruments, 18 heat-dissipation probes, which also measure matric potential, were installed within the columns of dry silica flour. Six of the heat-dissipation probe stations were also equipped with thermocouple psychrometers. All instruments were connected to an automatic data-acquisition system installed at land surface. Data loggers were programmed to take hourly readings of the thermocouple psychrometers and twice daily readings of the heat-dissipation probes during the first 50 days of operation, after which the thermocouple psychrometers were read twice daily and the heat-dissipation probes once daily. The pressure transducers and thermocouples were sampled twice hourly during the first 90 days of operation and, subsequently, have been sampled every 2 hours (Montazer et al., 1986).

Much of the data collected exceeds the useful range of matric potential over which the heat-dissipation probes were calibrated (-0.01 to -0.5 MPa). Some of the matric potential data also exceeded the maximum value (-0.5 MPa) for which the thermocouple psychrometers can be calibrated reliably. Nonetheless, the data are useful to indicate trends. The measurements of matric potential at each level within the borehole exhibited a trend during which, following the disturbance of natural conditions produced by drilling and instrumenting the borehole, the immediate environment of each instrument equilibrated with the ambient hydrologic conditions within the surrounding host rock. By May 1985, however, most of the instruments appeared to have stabilized sufficiently to yield a matric-potential profile that probably is indicative of ambient conditions within the host rock (Montazer et al., 1986).

Presuming conditions within the borehole to have equilibrated with those in the host rock, the data indicate that matric potentials within the PTn hydrogeologic unit at borehole USW UZ-1 range from -10 to -1 bars (-1 to -0.1 MPa) with a mean value of about -5 bars (-0.5 MPa) over the thickness of the unit. The matric potentials within the TSw unit range from -10 to -1 bars (-1.0 to -0.1 MPa), with an approximately constant mean value of about -3 bars (-0.3 MPa) over the thickness of the unit. Montazer et al. (1986) concluded that downward moisture flux within the TSw unit occurs predominantly as liquid-water flow within the fractures. Assuming that the nearly constant matric potential over the depth interval from 122

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to 224 meters in borehole USW UZ-1 implied unit vertical hydraulic gradient, Montazer et al. (1986) estimated that the downward liquid-water flux within the TSw unit at borehole USW UZ-1 probably is within the range of 0.1 to 0.5 mm/yr.

Pneumatic potentials (pore-gas pressures) are being monitored by downhole pressure transducers emplaced in borehole USW UZ-1 and have been measured within a piezometer nest installed in borehole UE-25a#4, about 1.6 km southeast of borehole USW UZ-1. Diurnal and barometrically induced fluctuations of gas pressure of about 0.25 kPa (Montazer et al., 1986) have been observed in these boreholes to depths of about 30 meters. Such fluctuations are apparently damped out below this depth, although seasonally induced pressure variations are observed to occur at greater depths (Montazer et al., 1986).

3.1.2.3.3.2 Saturated Zone Potentials

Potentiometric surface data from the site of the proposed repository at Yucca Mountain has been gathered from selected wells, either continuously or on a periodic basis, since 1983 or 1984 (Robison, et al., 1988). Resurveying of the land-surface altitudes of the well heads in 1984, and improved measurement corrections have resulted in greater precision of data collection, and the ability to infer the potentiometric surface more accurately. Increases in the accuracy of the water level data were made possible by corrections for more accurate borehole-casing collar measurements, equipment wear, mechanical stretch and thermal expansion of the steel tapes used for measurement (Ervin, et al., 1993).

In the area of the site, the potentiometric surface occurs principally in members of the Paintbrush Tuff and underlying Crater Flat Tuff lithostratigraphic units. Although the saturated zone monitoring wells are open to different geologic zones (Table 3.1.2.3.1-1), the potentiometric heads in most of the wells are believed to represent a laterally continuous aquifer of small hydraulic gradient within, and to the east-southeast of the proposed repository block (Ervin, et al., 1993). West of Yucca Mountain, potentiometric levels are approximately 45 meters higher than in the repository block; north and east, water levels are approximately 300 meters higher. These three zones define areas known respectively as the small, moderate, and large hydraulic gradient areas (Ervin, et al., 1993, see Figure 3.1.2.3.3.2-1).

Within the proposed repository block and to the southeast, the slope of the potentiometric surface is south to southeast, ranging from 732 to 728 meters in altitude and paralleling the regional topographic slope. The hydraulic gradient in this area is nearly horizontal as a result of [flow through highly permeable rocks, INN 3.1.2.3.3.2-1]. This area of small gradient is part of the lower volcanic flow system conceptualized by Ervin, et al. (1993), and is composed of fractured tuffs beneath the tuffaceous beds of Calico Hills, primarily in the Crater Flat Tuff. The flow system contains the potentiometric surface of the uppermost saturated zone in the small gradient area, and flow is primarily horizontal and down-gradient to the southeast of the repository block. To the west, the system is bounded by the low-permeability Solitario Canyon fault; although the system may extend north of the proposed repository, increasing lithostatic pressure would tend to close fractures and reduce permeability northwards.

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The moderate gradient area is located to the west of the proposed repository site. In this area, potentiometric heads range from 775 to 780 meters in altitude, approximately 45 meters higher than those in the repository block. Water in this area is mounded against the west side of the Solitario Canyon fault, and a splay of that fault. [The faults, composed of low permeability materials such as fault gouge and secondarily deposited silica, form a barrier to groundwater flow. Although a small amount of water crosses the Solitario Canyon fault in response to the strong gradient [INN 3.1.2.3.3.2-2], most of this water flows south, parallel to the fault (Ervin et al., 1993).]

To the north of the site, the potentiometric surface is situated at elevations ranging from 738 to 1,035 meters. This area, known as the large-gradient area, represents a semi-perched, unconfined aquifer system. As conceptualized by Ervin, et al. (1993), this system consists of an unconfined groundwater body situated above a confined water body of lower potentiometric head; hence, flow potential is downward, through a confining layer separating the two groundwater bodies. Flow in the upper and lower permeable zones is assumed to be horizontal, while flow in the confining layer separating the two zones would be predominantly vertical.

Variations in Potentiometric Head. Data in Robison and others (1988) and Gemmell (1990) indicates that water levels beneath Yucca Mountain are very stable over time. Gemmell (1990) shows that, except for water-supply well J-13, the maximum variation of 0.69 meters for the periodic water level network from 1983 through 1988 occurred at well UE-25 WT#14. However, annual average water levels in well UE-25 WT#14 exhibit a range of only 0.17 meters while the average over 1986-88 where the same equipment was used is only 0.03 meters. Considering all wells in the periodic network, the annual average water-levels from 1986 through 1988 changed from -0.10 to 0.04 meters and the average absolute change was 0.04 meters (DOE, 1990a).

Ervin et al. (1993) performed trend analyses to determine the effect of using average water-level data and data of different years to construct a potentiometric surface map of the Yucca Mountain vicinity. Their findings indicate little or no trend over the period of record, and they concluded that water levels probably would not change significantly over the next few years. Of the 27 wells included in the study, however, four showed statistically significant upward changes in water levels; wells USW WT-2, UE-25 WT#6, UE-25 WT#16, and USW H-5 exhibited changes of 0.07, 0.6, 0.2, and 0.08 meters/year, respectively. The investigators attributed the rising water levels in wells UE-25 WT#6 and UE-25 WT#16 to post-drilling equilibration, while the reasons for the trends in the remaining two wells remains unclear. The researchers noted, however, that the rate of change in the water levels demonstrated a decrease with time (Ervin, et al., 1993).

Besides year-to-year changes, water levels appear to exhibit little in the way of seasonal variation (DOE, 1990a). In regions where precipitation is greater and water levels are shallower than at Yucca Mountain, there is typically a well-defined seasonal or annual correlation between precipitation and groundwater levels. The relative lack of these seasonal fluctuations at the proposed repository site is generally attributed to extremely low rates of net

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infiltration and recharge attributable to local precipitation, the great depths to the water table, and long groundwater travel times through the unsaturated zone (DOE, 1988). The apparent lack of seasonal fluctuation may also be related to the precision of potentiometric head measurements. Ervin, et al. (1993), report the accuracy of water level data as approximately +/- 0.11 meters, with average water level precision being about 0.01 meters. Variations smaller than the accuracy of measurement will tend to be masked by experimental error.

3.1.2.3.4 Characteristics of Hydrogeologic Units [INN 3.1.2.3.4-1]

In the area of the central block of the proposed Yucca Mountain repository, strata are divided into three geologic formations: Paintbrush Tuff, tuffaceous beds of Calico Hills, and Crater Flat Tuff. These formations and their component members are distinguished stratigraphically by their petrographic characteristics; the physical properties within each formation vary considerably, largely due to variations in the degree of welding of the tuffs. Because physical properties control the characteristics of water occurrence and flow, and because physical-property boundaries do not correspond to rock-stratigraphic boundaries in most cases, the strata underlying the site have been grouped into hydrogeologic units, based principally on the degree of welding. Six unsaturated zone units and two saturated zone units have been defined, primarily through the work of Montazer and Wilson (1984) and Waddell et al. (1984).

3.1.2.3.4.1 Unsaturated Zone Hydrologic Units

Rocks in the unsaturated zone beneath Yucca Mountain are grouped into six hydrogeologic units, based on their physical properties. Hydrogeologic units consisting of vitrophyres and moderately to densely welded devitrified tuffs alternate with units consisting of nonwelded to partially welded tuffs and bedded tuffs. Welded units generally have greater fracture density and fracture permeability then nonwelded units; however, the welding process reduces matrix pore size, resulting in lower matrix permeability for welded units than for nonwelded units (Montazer and Wilson, 1984).

QAL. This hydrostratigraphic unit coincides with the rock-stratigraphic unit of the same name, and covers approximately 24% of the site surface (Wittwer, et al., 1992). The QAL ranges from 0-30 meters in thickness and is composed of alluvial and colluvial deposits, the major constituents of which are sand and gravel. As such, these deposits can have very high saturated matrix permeabilities, large specific retention, and large effective porosity. Winograd and Thordarson (1975) report saturated matrix conductivities for Quaternary alluvial valley-fill aquifers near Yucca Mountain ranging from 2.4×10^{-6} to 3.3×10^{-5}m/s. Because of the large amount of pore-space, however, under unsaturated conditions percolation rates are small, as evidenced by infiltration experiments conducted in desert alluvial material (Kearl, 1982). Baring locally impermeable barriers, vapor-phase permeability is high.

Grain size distribution of QAL deposits is highly variable, ranging from clay-sized material to boulders. Character of alluvial materials is a function of distance of transport from source, type and weathering of source material, variations in transport and depositional processes, and

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other, similar factors. Interbedding of fine- and coarse-grained material resulting from spatial and temporal variations in runoff also contributes to the highly heterogeneous nature of QAL deposits.

QAL deposits may be further modified by the presence of interbedded caliche layers. Calichification varies from slight cementation of grains to thickly-bedded calcium carbonate layers similar in nature to sparite; caliche beds are often discontinuous, although they may be areally extensive. Because well developed caliche beds have small, poorly interconnected pore spaces that demonstrate low (compared to non-calichified sediments) matrix hydraulic conductivities, caliche bedding contributes significantly to the anisotropy present in QAL deposits. In addition, where layers are continuous over large areas, caliche may impede infiltration and contribute to the formation of perched-water conditions just below the surface (DOE, 1990).

TCw. The TCw hydrofacies lies completely within the boundaries of the Tiva Canyon lithostratigraphic unit. The TCw hydrostratigraphic unit consists of moderately to densely welded tuffs 0-150 meters in thickness, which dip 5° to 10° eastward within the central block of Yucca Mountain. Although the unit has been eroded away in some washes, the majority of the surface of the site consists of TCw (Wittwer et al., 1992). The unit is highly fractured (Scott et al., 1983, estimate 10 to 20 fractures/m^3) but has low matrix permeability (Weeks and Wilson, 1984, estimated saturated hydraulic conductivity at 2.3×10^{-11} m/s) .

PTn. The PTn consists of the non- and partially welded base of the Tiva Canyon Member, the Yucca Mountain Member, the Pah Canyon Member, the non- and partially welded upper part of the Topopah Spring Member, and associated bedded tuffs. All members comprising the PTn hydrostratigraphic unit are part of the Paintbrush Tuff lithostratigraphic unit.

Within the central block, the PTn unit consists of thin, nonwelded ash-flow sheets and bedded tuffs that thin to the southeast from a maximum thickness of 100 meters to a minimum thickness of about 20 meters. The unit dips to the east at 5° to 25°, although the dip within the repository block rarely exceeds 10° (Scott and Bonk, 1984). Tuffs of this unit are vitric, nonwelded, very porous, slightly indurated, and, in part, bedded. Porosities average about 46%, but may range as high as 61% (Montazer and Wilson, 1984). In keeping with these high porosity values, the moisture content of PTn samples is also high; Montazer and Wilson (1984) report mean gravimetric water content at 19% (volumetric 27%). Fracture density is estimated at 1 fracture /m^3 (Scott et. al., 1983).

Laboratory analysis of core samples of PTn demonstrate saturated hydraulic conductivities on the order of 10^{-7} m/s (Weeks and Wilson, 1984). In situ air-permeability tests were conducted by E.P. Weeks, of the USGS, in well UE-25a#4; these tests indicated a bulk hydraulic conductivity for the non-wetting phase (air) of 4×10^{-7} m/s (E.P. Weeks, as reported by Montazer and Wilson, 1984). The close agreement of air and water hydraulic conductivities are cited by Montazer and Wilson (1984) as evidence in support of the belief that fracture permeability in the PTn unit is insignificant.

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TSw. The Topopah Spring Member is the thickest and most extensive ash-flow tuff of the Paintbrush Tuff. This hydrostratigraphic unit consists of thin upper and lower vitrophyres separated by a thick central zone of densely welded, devitrified ash-flow sheets. The thickness of the TSw varies between 290-360 meters in the area of the central block; the proposed repository horizon is located within the central and lower parts of this unit.

The moderately to densely welded, devitrified central section of the TSw unit has a small saturated matrix conductivity on the order of 10^{-11} m/s (Montazer and Wilson, 1984); bulk saturated hydraulic conductivity is much higher, owing to the fractured nature of this unit (Scott et. al., 1983, indicate fracture densities ranging from 8-40 fractures/m). Tests conducted at well J-13 (about 6 km east of the proposed repository site) indicate horizontal saturated hydraulic conductivity on the order of 10^{-5}m/s (Thordarson, 1983).

Generally speaking, overall matrix porosities in the TSw unit are low. Weeks and Wilson (1984) measured average porosity in the upper portion of the TSw at 14%, 5% of which was drained, resulting in a mean saturation of 65%. The central and lower portions of this unit, however, contain abundant lithophysal gas cavities, resulting in greater porosities. Although they increase bulk porosity, capillary barrier effects probably keep these cavities filled with air, decreasing unsaturated hydraulic conductivity and effective porosity (Montazer and Wilson, 1984).

CHn. The CHn hydrostratigraphic unit is a grouping of non- to partially welded ash-flow tuffs, 100-400 meters in thickness, which underlie the TSw hydrofacies. Because the water table intersects the CHn unit under the northern one-half of the proposed repository block, the unsaturated thickness of this hydrofacies is a function not only of the thickness of the component stratigraphic units, but of the unit's spatial relationship to the water table.

As defined by Montazer and Wilson (1984), this unit is composed of the following stratigraphic units:

a. A nonwelded to partially welded vitric layer, locally zeolitic, that is the lowermost part of the Topopah Spring Member of the Paintbrush Tuff.

b. Tuffaceous beds of Calico Hills.

c. The Prow Pass Member of the Crater Flat Tuff, which is nonwelded to partially welded where it occurs in the unsaturated zone beneath the central block.

d. The nonwelded to partially welded upper part of the Bullfrog Member of the Crater Flat Tuff, where it is above the water table.

Beneath the southern two-thirds of the central block, the upper part of the CHn unit contains some non-vitric facies; however, the lower portion of the southern two-thirds, and the entire northern third of the unit is devitrified and altered. The most common alteration product is zeolite, but clays and calcite are also known to occur (Montazer and Wilson, 1984). Because

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the vitric facies, and the devitrified, zeolitic facies posses somewhat different hydrogeologic characteristics, they are considered as separate subunits of the CHn unit, and designated CHnv and CHnz.

Both CHnv and CHnz facies have similar porosities (37% and 31%, respectively, Montazer and Wilson, 1984), and high degrees of saturation (about 90%, Montazer and Wilson, 1984). In spite of the similar porosities between the two sub-units, they exhibit very different permeabilities. Tests performed in UE-25 a#4 indicate vertical saturated hydraulic conductivities of $4x10^{-8}$ m/s for CHnv, and $9x10^{-11}$ m/s for CHnz facies (E.P. Weeks, as reported by Montazer and Wilson, 1984). Montazer and Wilson (1984) attribute this difference in permeability to extensive argillization of the CHnz unit. Effective permeabilities under unsaturated conditions were estimated by Weeks and Wilson (1984) to be an order of magnitude smaller. In addition, the CHn unit exhibits moderate anisotropy; Montazer and Wilson (1985) report mean horizontal hydraulic conductivity (saturated) of the CHnz facies at $1x10^{-9}$ m/s, about an order of magnitude greater than reported vertical conductivity. Anisotropy in the CHn unit is probably related to intercalation of tuffs with greater and lesser permeabilities; the degree of anisotropy present is moderated to some degree by the presence of discontinuous fracture systems, particularly in the zeolitic facies (Montazer and Wilson, 1984). These fracture systems increase the vertical component of conductivity; however, the density of fractures in the CHn unit is low (2-3 fractures/m^3, Scott et al., 1983), and fracture flow occurs only in the uppermost layers of this hydrofacies. Below these upper layers, imbibition of moisture to the matrix removes the liquid water from the fractures, except along structural flowpaths (Montazer and Wilson, 1984).

CFu. Under the southern one-half of the central block of the proposed repository, the lowermost unit of the unsaturated zone is the CFu. The unsaturated thickness of this unit varies from about 0-160 meters; under the northern half of the block, the unit becomes fully saturated. The CFu is composed of both welded and non-welded components of the Bullfrog Member of the Crater Flat Tuff, hence the descriptor "undifferentiated."

Fracture densities for the CFu unit are in the range of 8-25 fractures/m^3 (Scott et al., 1983), and hydraulic conductivities, both saturated and unsaturated, are highly variable due to the undifferentiated nature of this unit. Because of the limited extent of CFu beneath the central block, however, it exerts little control over unsaturated zone flow system (Montazer and Wilson, 1984).

3.1.2.3.4.2 Saturated Zone Hydrogeologic Units

The saturated zone flow system at Yucca Mountain can be divided, based on stratigraphy, into two systems; a saturated tuff system (the "lower volcanic flow system" of Ervin, et al., 1993), and a lower carbonate system. The two systems may be separated in places by the upper clastic aquitard (Winograd and Thordarson, 1975). At one point (well UE-25p#1), the head in the carbonate system is higher than the head in the tuff system, and at a number of other points the potential appears to be for upward flow (DOE, 1990a). It is, therefore, unlikely that groundwater moving through the shallow part of the saturated zone in the

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vicinity of the proposed repository block has significant local interaction with water from the lower carbonate system.

The saturated tuff system, or "lower volcanic flow system" as defined by Ervin et al. (1993), is a zone of fractured tuffs beneath the tuffaceous beds of Calico Hills, and is primarily composed of the various members of the Crater Flat Tuff. This system has not been subdivided into hydrostratigraphic units, primarily due to the lack of correlation between stratigraphic units and producing zones in existing wells. Degree of welding is probably a better indication of hydraulic properties than stratigraphic units, and hydraulic conductivity is largely a function of fracturing and subvertical faulting; therefore, all stratigraphic units are treated as a single, highly variable hydrofacies.

CFu. The saturated zone flow system as defined by Ervin, et al. (1993) consists of a "lower volcanic flow system" between zones of poorly permeable volcanic sediments. This lower volcanic flow system is composed principally of the Prow Pass, Bullfrog and Tram members of the CFu. These units are non- to densely welded ash-flow and bedded tuffs with saturated matrix conductivities ranging from $7x10^{-12}$ to $1x10^{-8}$ m/s (Whitfield et al., 1985); bulk saturated hydraulic conductivities are higher due to the presence of secondary (fracture) permeability (Rush, et al., 1984, calculated horizontal average hydraulic conductivities for CFu members as ranging from $8x10^{-11}$ m/s within the Tram member, to $2x10^{-4}$ m/s in an interval of the Prow Pass member).

CHn. Evidence indicates that the most likely direction of travel for radionuclides from the proposed repository site through the saturated zone is east to southeast, perpendicular to potentiometric contours (Ervin et al., 1993). Down gradient of the Crater Flat unit along this flow path is the zeolitized facies of the CHnz. This unit is of special interest because it is considered an important natural barrier to radionuclide transport, due to its zeolitized nature and low permeability. The CHnz unit is considered part of the "poorly permeable volcanic rocks" discussed by Ervin et al., (1993); saturated hydraulic conductivity is small (Montazer and Wilson, 1984, report mean horizontal saturated hydraulic conductivity of $1x10^{-9}$ m/s for the CHnz units), possibly due to extensive argillization (Montazer and Wilson, 1984). A degree of anisotropy which is present in much of the CHn unit may be ameliorated by the presence of discontinuous fracture systems (Montazer and Wilson, 1984). Scott et al. (1983) estimate 2-3 fractures/m^3 for this unit.

Down gradient and beyond the site boundaries, the water table intersects saturated portions of the Topopah Spring welded hydrofacies. Further information with regard to the saturated flow system is presented in Subsection 3.1.2.2.3, Characteristics of Hydrogeologic Units (Regional Hydrogeology).

3.1.2.3.5 Site Groundwater Recharge

[INN 3.1.2.3.5-1]

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3.1.2.3.6 Site Groundwater Discharge

No reports of groundwater discharge have been documented at the Yucca Mountain Site or within the controlled area. The closest known spring is Topopah Spring, about 19 km northeast of Yucca Mountain.

3.1.2.3.7 Age of Site Groundwater

This subsection presents information concerning hydrochemical evaluation of the groundwater system and the age of the site groundwater. This subsection deals primarily with problems associated with interpreting groundwater velocities derived from radiocarbon data.

Environmental isotopic data are available for groundwater samples from wells in the tuff aquifers beneath Yucca Mountain (Benson et al., 1983; Benson and McKinley, 1985). Both carbon-14 activity and "uncorrected, apparent radiocarbon age" of the samples were reported. In areas where carbonate minerals may contribute significant amounts of carbon, it can be misleading to use these apparent ages in interpreting a flow system. Such apparent ages, which can be off by more than a factor of two, have been omitted in Table 3.1.2.3.7-1. Since corrected ages (determined by modeling the initial radiocarbon activity of the water; e.g., IAEA, 1983) are not available, only the sample activity measurements were used. The measured carbon-14 activities are relatively homogeneous for an area this size, particularly when accounting for sampling, sample preparation, and analytical uncertainties. Samples from the two southernmost drillholes (USW H-3 and USW H-4) have lower carbon-14 content (mean value 11.1 percent modern carbon) than samples from the other, more northerly drillholes (mean value 20.1 percent modern carbon) (Table 3.1.2.3.7-1). The significance of this difference cannot be fully assessed without carbon-14 modeling. Nevertheless, groundwater residence times of the southern samples may be on the order of 5,000 yr greater than the northern samples. Because significant amounts of carbonate minerals do not occur in the tuff or in tuffaceous valley fill with which these groundwaters have come into contact, minimal radiocarbon dilution may have occurred from such contact and apparent ages may be more realistic than at other sites.

The δ carbon-13 ($\delta^{13}C$) values may show a spatial variability similar to those of the carbon-14. The values are consistent with recharge under very poorly vegetated conditions. The $\delta^{13}C$ value reported for sample USW H-3 is above the mean atmospheric value for carbon dioxide. The explanation of this unusual phenomenon could shed light on the differences in radiocarbon content between this sample and the northern ones. The $\delta^{13}C$ values below about -10 parts per thousand in dilute waters suggest that little carbonate rock/water and soil gas/water exchange has occurred.

The three tritium values of <62 tritium units (TU) in Table 3.1.2.3.7-1 serve only to indicate that no large amounts of bomb-test tritium have reached the aquifer at those points. The other samples, which were analyzed by a more sensitive method to give a more precise result, are consistent with trace tritium contamination of samples of very old water (very old, in the

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context of tritium dating implies several hundred years or more). Such contamination is common and not unexpected at Yucca Mountain.

The deuterium and oxygen-18 data in Table 3.1.2.3.7-1 plot on and below the meteoric line, occurring between a deuterium excess of +10 (mean continental meteoric water) to +5. Such a small deviation from the world-wide average meteoric deuterium excess is within expected local continental variation. The water is extremely depleted in heavy isotopes of both oxygen and hydrogen, consistent with high altitude, cold (winter) continental recharge or snow-melt. No evaporative enrichment or thermal alteration is observed. No complete hydrochemical or isotopic analyses are available for water from the unsaturated zone at the site. Methods (triaxial compression, high-speed centrifuge and vacuum distillation) are currently under development for extracting uncontaminated samples upon which to perform these analyses. Yang (1986) has reported some preliminary calcium and sodium concentration data, showing that calcium is elevated in the pore water relative to the groundwater, to the extent that it dominates over sodium. Complete chemical and isotopic characterization of the infiltrating pore water is needed (1) to develop an understanding of the hydrochemical nature of the water that may contact a waste package and (2) to isotopically trace the movement of these waters through the unsaturated zone.

[INN 3.1.2.3.7-1: Collect these data as a function of depth]

The carbon-14 and carbon-13 composition of the unsaturated zone CO_2 (g) phase has been determined by Yang et al. (1985) on samples obtained by the new methods developed by Haas et al. (1983). Radiocarbon from bomb-test fallout was observed to a depth of 12.2 m in borehole UZ-1 on the exploration block. Radiocarbon activity decreased to below 100 percent modern carbon below 18.3 m, which may indicate that downward gas-phase transport from the surface has not occurred beyond this depth since the mid-1960.

Unsaturated zone

Complete hydrochemical or isotopic analyses are not available for pore water taken from the site's unsaturated zone. Methods to extract this water in an uncontaminated state are still under development. Yang (1986) presents preliminary calcium and sodium concentration data; his studies indicate that calcium concentrations are elevated in the unsaturated zone and dominate over sodium. This information will be used to understand potential interactions of pore water with the waste canister and to define flow paths and velocities in the unsaturated zone.

Saturated zone

The reported radiocarbon dates of groundwater at Yucca Mountain have been uncorrected and represent apparent ages (Benson and McKinley, 1985; Claassen, 1986). Given a series of assumptions examined in more detail below, such ages may represent the time since groundwater was isolated from the CO_2 gas reservoir of the atmosphere or the unsaturated zone (i.e., the time since the water entered the water table beneath Yucca Mountain).

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Uncorrected ages are realistic when no carbon-12 dilution has occurred to the initial carbon-14 by the dissolution of carbon-14-free carbonate minerals. Claassen (1985) indicates that this is the case in the tuffaceous aquifer, since these formations are virtually carbonate-free. If such dilution occurs, the apparent radiocarbon ages may be an upper limit of the groundwater age.

The other major assumption underlying the apparent ages relates to the initial carbon-14 composition of the infiltrating water. Uncorrected ages assume that this water was in isotopic equilibrium with a gaseous reservoir at 100 percent modern carbon. This is generally valid, since both the atmosphere (before nuclear weapons testing) and soil gas (gas in the unsaturated zone in a well mixed or low CO_2 gas-residence-time system) have this composition. In a poorly mixed system with long CO_2 gas residence times where the carbon-14 in the CO_2 gas has decayed, the radiocarbon reservoir is at less than 100 pmc. In such a case, the apparent age of the groundwater is greater than the time it has been in the aquifer (by an amount proportional to the difference between 100 pmc and the carbon-14 concentration in the equilibrating CO_2 gas).

Yang et al. (1985) have shown that the carbon-14 concentration of the CO_2 gas in the soil zone can be as low as 34.2 (±1.5) pmc in the unsaturated zone at 368 m below the surface. The apparent age of a groundwater sample equilibrated with such gas is really equal to the age of the gas in the unsaturated zone (without isotopic mixing) plus the residence time of the groundwater in the aquifer since equilibration (without isotopic dilution). The apparent age is thus older than the true groundwater residence time. Since isotopic reequilibration with CO_2 gas can occur throughout the unsaturated zone, apparent ages of the unsaturated zone water (time since precipitation) are not reliable.

Well documented and tested models of initial groundwater carbon-14 activity exist. Some of these models (Mook, 1976, 1980; Fontes and Garnier, 1979) are particularly well suited for modeling waters which equilibrate with CO_2 gas in the unsaturated zone and then enter the water table. Such models can be used to examine the apparent difficulties in interpreting carbon-14 data from Yucca Mountain. Such models, which are said to give corrected ages, do not actually perform corrections of an error, but rather, they simply do not ignore the hydrochemical processes which control carbon-14 behavior.

Kerrisk (1987) has pointed out that at Yucca Mountain, gaseous CO_2 from the unsaturated zone may be added to or exchange with the carbonate of the saturated zone water. This process could lead to apparent ages that are younger than the true ages of the water in the saturated zone.

An accurate knowledge of initial carbon-14 activity is less important in calculating relative groundwater ages or velocities than it is in determining absolute chronologies. Groundwater velocities can be estimated by dividing the distance between two points along a hydrologic flow line by the carbon-14 age difference of the water at the two points. Consistent errors in estimating initial activity tend to cancel, resulting in relative ages which are more accurate than are the individual absolute ages. This is not the case when calculating the velocity of

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water to a point where the carbon-14 age is known from the assumed point of present day recharge (absolute age zero). Mixing (i.e., the two points are not actually on the same hydrodynamic flow line or the line has moved during the transit of the parcel of water) is also not considered in this calculation. These two difficulties apply to the results of Claassen (1985) who states, "For example, assume recharge near the head of Fortymile Canyon occurred 17,000 years ago and flowed to the lower end of the Canyon, where it is sampled. The velocity that is calculated, 7 m/yr, must be a maximum. Because recharge may occur anywhere but not necessarily everywhere along surface drainageways, no probable minimum velocity can be calculated. The absence of water older than about (apparent age) 10,000 yr. before present even near the head of Fortymile Canyon, would favor velocities slower than 4 m/yr."

[INN The use of initial carbon-14 modeling and relative dating along flow paths to resolve apparent difficulties of interpretation.]

3.1.2.3.8 Site Pathway Analysis

The presence of certain hydrogeologic features at Yucca Mountain significantly affect flow in the unsaturated zone of the site. These features include fractured porous media, layered units with contrasting physical properties, dipping units, and a deep water table resulting in a very thick unsaturated zone. The combination of these features provides a hydrogeologic framework that probably results in certain flow phenomena, including infiltration into fractured rocks, flow through fractured rocks, retardation of flow by capillary barriers, lateral flow, and vapor flow (Montazer and Wilson, 1984).

Based on a composite-porosity continuum model developed for fractured, porous media, Peters, et al. (1986) used a one-dimensional numerical model TOSPAC to examine water movement to the water table. Assuming percolation rates of less than or equal to 1.0 mm/year, simulations indicated travel times on the order of hundreds of thousands of years. The values assumed for percolation are less than the matrix conductivity of the least conductive units; therefore, it was assumed that no water movement would occur in fractures, because capillary forces would cause water to remain in the matrix until saturation. The model was particularly sensitive to percolation rate; a factor of 10 difference in the percolation rate could result in fracture flow and substantially reduced travel times from the land surface to the saturated zone.

Nitao et al. (1992) examined the available evidence in support of fracture flow, and found that field observations support long-distance, relatively rapid flow in the vadose zone in the vicinity of Yucca Mountain. Their evidence includes:

a. Detection of chlorine-36 and tritium at depths of 135-150 meters below the land surface (Norris, et al., 1989).

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b. The apparent movement of drilling fluid a distance of about 300 meters, from borehole USW G-1 to USW UZ-1, indicating that water can move a significant distance in short time periods (in this case, 300 meters, in less than three years).

c. Unpublished results obtained from barometric pressure response tests at the C-well complex (Luckey and Patterson, cited in Nitao et al., 1992) yield permeability values greater than 40 darcys. The most probable explanation for such high values is the existence of connected pathways of large-aperture fractures.

d. Stable isotope signatures for Trench 14 indicated that calcite fracture fillings in the upper 3 to 4 meters were probably derived from meteoric water moving through fracture zones (Quade and Cerling, 1990).

Numerical simulations of vertical percolation through Yucca Mountain driven by a constant infiltration flux predict a saturation vs. depth profile which is not consistent with measured values (Buscheck et al., 1991). Nitao et al. (1992) cite as an explanation possible occurrences of episodic infiltration pulses in fractures, probably originating from washes during flash floods. Welded units, because of their low matrix permeability, imbibe only small amounts of water. The bulk of the pulses travel until reaching higher permeability, non-welded units with higher absorptive capacities.

Assuming episodic non-equilibrium fracture-matrix flow, Buscheck et al. (1991) simulated flow through a (hypothetical) 0.1 mm aperture fracture extending from the ground surface to the water table. As a test case, the investigators assumed a ponded source on the land surface throughout the simulation. Under these conditions, the simulated wetting front reached the repository horizon in about 70 years. Using comparable assumptions, water moved from the repository horizon to the water table in 290 days.

The model of Buscheck et al. (1991) is particularly sensitive to fracture aperture. Although it took 70 years for water to move from land surface to the repository horizon through a 0.1 mm fracture, a 1 mm fracture(or equivalent aperture fracture zone) conducted water to the repository horizon in approximately one hour.

Geologic structure other than fractures can effect the flow of moisture, both in the unsaturated and saturated zones. Several faults are know to bisect the site of the proposed nuclear waste repository; their effects on the distribution of moisture flux vary. The Solitario Canyon Fault to the west of Yucca Mountain, for instance, acts as a barrier to flow, resulting in mounding of groundwater against the west side of the fault and a splay of the fault (Ervin et al., 1993). The presence of low permeability material, such as fault gouge or secondarily deposited silica, cause a barrier effect which forces flow from the west to travel parallel to the fault, in a southerly direction.

Other faults behave differently, based on morphological variations [INN 3.1.2.3.8-1]. Faults which are open, either along their entire length, or at depth, can act as conduits for flow. The presence of faults that are open, or filled with high permeability materials, can provide drains

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which divert lateral flow above the repository horizon directly to the water table; [for example, the Bow Ridge fault is a high permeability fault which acts as a preferential pathway to drain moisture from the upper unsaturated zone to the water table [INN 3.1.2.3.8-1].]

Other discontinuities in geologic media can act to promote preferential movement of moisture as well. In particular, contrasting physical properties of adjacent layers, and dipping units can combine to cause lateral flow at the boundaries between low and high permeability layers. Montazer and Wilson (1984) give a detailed description of how the capillary barrier effect can retard flux from a layer of low permeability into a layer of higher permeability; [the bedded tuff zones between the major units may act as capillary barriers that cause ponding of infiltration above the bedded tuffs. Similarly, non-welded vitric units, and the bedded tuffs themselves, may act to concentrate flux and promote areas of locally high saturation. Episodic infiltration pulses in fractures pass through the overlying welded units, which, because of their low matrix permeability, imbibe very little water over the relatively short time period of the pulses. As the pulses reach the nonwelded vitric units, the higher permeability matrix absorbs most of the water, locally raising the saturation of the units.

Regardless of the mechanism promoting higher saturation, once the level of saturation reaches a critical value, lateral flow begins down-dip towards the east. At the site of the proposed repository, lateral flow continues along the boundary between layers until it contacts the Bow Ridge fault, which acts to divert lateral flow directly to the water table [INN 3.1.2.3.8-1].

3.1.2.3.9 Local Groundwater Use

This subsection identifies all production wells that are located near the Yucca Mountain Site. [It is possible that withdrawals from these wells could affect the local flow system. Available pumping information that will aid in identifying these effects is presented. These data, in conjunction with results from planned modeling and field studies are used to quantify and qualify any potential changes to the local flow system.]

There are two wells located proximally to the Yucca Mountain Site that withdraw water on a regular basis. Wells J-12 and J-13, located approximately 12 km and 8 km southeast of the Yucca Mountain Site, respectively, are utilized and controlled by the DOE to support activities in the southwestern part of the NTS (refer to Subsection 3.1.2.2.8 for a discussion of regional groundwater users).

Water from wells J-12 and J-13 is produced from the Topopah Spring Member of the Paintbrush Tuff, which constitutes the lower portion of the welded tuff aquifer. The Topopah Spring Member occurs at a depth of approximately 180 to 347.2 m at well J-12 and 207.3 to 449.6 m at well J-13 (Thordarson et al., 1967; Claassen, 1973). Pumping-test and water-level data for both wells are shown in Table 3.1.2.3.9-1. Well-construction information can be found in Thordarson et al. (1967), Young (1972), Claassen (1973), and Thordarson (1983). Table 3.1.2.3.9-2 summarizes water production for wells J-12 and J-13 from 1983 through 1985.

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Wells J-12 and J-13 have been pumped intermittently since their completion. Before 1968, well J-12 was pumped at rates in excess of 2,000 m^3/day. After the wellbore was cleaned and deepened in 1968, production capabilities exceeded 4,500 m^3/day. Well J-13 has been shown to be capable of producing in excess of 3,500 m^3/day of water (Claassen, 1973). The specific capacity of the well underwent a slight decline in 1964 shortly after completion. Claassen (1973) hypothesized that the well was not fully developed during the 1964 tests. Subsequent pumping tests performed in 1969 support this hypothesis, and the specific capacity of 540 m^3/day/m is considered the most reliable measurement.

Pumping tests performed after 1968 on well J-13 produced draw-downs of less than 7 m with pumping rates in excess of 3,500 m^3/day (Table 3.1.2.3.9-1). Figure 3.1.2.3.9-1 [INN 3.1.2.3.9-1] indicates that even with well J-13 in continuous service, declines in the water level have been minimal. This suggests that the short-term effects of pumping on the regional potentiometric surface are probably negligible. The effects of long-term pumping on the local potentiometric surface are unknown and remain to be investigated (Robison, 1984).

The water which will be needed for site characterization activities probably will be drawn from wells J-12 and J-13. It is possible that a substitute source may be located and developed (Morales, 1986).

Other drillholes exist near the site but are pumped only periodically to obtain hydrologic and hydrochemical information. These test holes have not been used to supply water.

Pumping rates of J-13 during 1992 through the first quarter of 1994 for site characterization activities, are tabulated in Table 3.1.2.3.9-3.

3.1.2.3.10 Site Paleohydrology

Discussion of the paleohydrology of the south-central Great Basin, including Yucca Mountain, is presented in Regional Paleohydrology. In this subsection, the discussion is limited to those past and probable future hydrologic conditions that may directly affect a repository at Yucca Mountain.

Winograd and Szabo (1986) concluded that there was a progressive lowering of water table in the south-central Great Basin during the Quaternary, and that "a continued decline of the regional water table in the next 10^5 to 10^6 yr (and beyond?) in response to increasing aridity and to lowering of groundwater base level" is likely (Subsection 3.7.4). They noted, however, that this progressive water-table decline "does not preclude superimposed and relatively rapid cyclical fluctuations in water level in response to glacial (i.e., pluvial) and interglacial climates of the Pleistocene." Preliminary field evidence and the results of a computer-model analysis bearing on past and possible future water levels beneath Yucca Mountain are outlined below.

Bish and Vaniman (1985) noted that nonwelded glass occurs both above and below the welded zone in the Topopah Spring Member of the Paintbrush Tuff. Bish and Vaniman

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(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)." 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.

To help assess the implication in Bish and Vaniman (1985) that the lowest position of vitric nonwelded tuff might be related to a stand of a paleowater table, a contour map of the base of the lower vitric nonwelded tuff (in both the Topopah Spring Member and in the tuffaceous beds of the Calico Hills) will be prepared and contrasted with the modern water table. If such a contour map indicates (1) a relatively smooth surface at the base of the vitric nonwelded tuff, (2) a surface that locally cross-cuts stratigraphic horizons in the tuffs, and (3) a surface that climbs sharply to the north as does the modern water table beneath Yucca Mountain, then it is possible that the porous nonwelded vitric tuffs have indeed been altered below a paleowater table, as suggested by Bish and Vaniman (1985). Such a contour map would then presumably mark the stand of water table at the time or times of alteration of the tuffs. The occurrence of such a stand probably could not be dated, but might have occurred several million 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 proposed repository horizon. Thus, if the previously discussed work demonstrates the likelihood of a relationship of the paleowater table to the lowest occurrence of vitric tuff, a possible maximum (though undated) water-table stand would have been identified that is significantly lower than the repository horizon under consideration.

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 proposed repository. Inundation of any part of the planned repository would require a water-table rise of more than 200 m. His analysis also suggests a past shortening by two-thirds of groundwater flow paths from Yucca Mountain to discharge areas in the southern Amargosa Desert. These discharge areas are still beyond the boundary of the accessible environment.

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/yr) to about 8 mm/yr at Yucca Mountain. Czarnecki (1985) acknowledges that this recharge value, which was derived from

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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 proposed repository site, along a segment of Fortymile Wash. This flux 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 drainageways. This would have the effect of decreasing the effective groundwater recharge to much 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 yr 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.

[INN 3.1.2.3.10-1: Parameters used to model the hydrologic system more effectively and any variations induced by climatic or tectonic changes.]

During pluvial periods of the Pleistocene, perched springs and seeps could have occurred along the flanks of Yucca Mountain and these might have resulted from recharge along the exposed up-dip portion of tuff units, with water moving down-dip to discharge sites along the flank of the mountain. No conclusive evidence for former springs has been observed.

[INN 3.1.2.3.10-2: Additional paleospring investigations are planned.]

Deposits along fault zones have been investigated to provide additional evidence of paleo-hydrologic conditions. Along portions of the Bow Ridge fault, 1 km east of the eastern boundary of the proposed repository, Trench 14 has exposed deposits of carbonate, opal, and minor amounts of sepiolite (Vaniman et al., 1985; Taylor and Huckins, 1986; Voegele, 1986a, b; Quade and Cerling, 1990; Stuckless, 1991; Taylor, in press). Although this trench was originally opened to examine the history of faulting along the Bow Ridge fault, questions about the origin of these deposits caused their genesis to be studied, and the following hypotheses were considered:

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a. The deposits are the result of groundwater discharge along the fault, from the deep regional flow system and reflect former higher water levels, tectonic uplift of the deposits, or a combination of both.

b. The deposits are the result of groundwater discharge along the fault, from groundwater being forced up due to tectonic processes (volcanism, faulting, interaction between volcanism and faulting, strain release, etc.).

c. The deposits are the result of discharge along the fault zone from a shallow groundwater flow system along the flank of Yucca Mountain.

d. The deposits are shallow soil features that resulted from rain water that moved downslope from the top of a nearby low hill, along the contact between the soil zone and bedrock, and accumulated at the fault zone. The water would have evaporated slowly, leaving precipitates of the minerals that were dissolved in the water.

The fault infilling features have been cited to support the first two hypotheses. Examination of these features and the data gathered regarding the features by an independent review panel led the panel to indicate that there is no evidence to support the assertion that the water table has risen periodically hundreds of meters from deep within the crust (NAS/NRC, 1992). The same review of the geologic features indicated that the evidence strongly supports a surface-process origin from meteoric water for the vein and surface-parallel calcite and calcite silica deposits throughout the Yucca Mountain area (NAS/NRC, 1992). Other findings of the panel are discussed below.

Calcite deposits along faults, fractures, and at the surface

Deposits of calcite, with subordinate amounts of opaline silica, and minor sepiolite are associated with late Cenozoic faults and late Pliocene to Holocene surficial deposits in the vicinity of Yucca Mountain (Swadley et al., 1984; Voegele, 1986a, b; Taylor, in press). The deposits described will hereafter be referred to as calcite-silica deposits. The faults with associated calcite-silica deposits include the Bare Mountain fault (Swadley et al., 1984; Reheis, 1986), the Paintbrush Canyon fault (Swadley et al., 1984), the Bow Ridge fault (Trench 14, Figure ZZZ); (Voegele, 1986a, b; Taylor and Huckins, 1986; Vaniman et al., 1988; Quade and Cerling, 1990; Taylor, in press), the Solitario Canyon fault (Swadley et al., 1984), the Windy Wash fault (Swadley and Hoover, 1983; Whitney et al., 1986), and a N-S trending fault on the west side of Busted Butte (Grant, 1986; DOE, 1993). A discussion of the available petrologic, geochemical, and age data is presented below.

The calcite-silica deposits occur along fault planes in unconsolidated alluvium and in bedrock. The calcite-silica deposits exhibit several generations of deposition. They may exhibit laminations that are subhorizontal to nearly vertical in orientation or appear in a massive form (Voegele, 1986a; Taylor, in press). In Trench 14, where the Bow Ridge fault can be observed to displace alluvium and bedrock, the calcite-silica deposit contains angular blocks of the bedrock (DOE, 1993; Taylor, in press). Stringer veins of the calcite occur within the

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soil surrounding the faults in Trench 14 (Taylor and Huckins, 1986; Taylor, in press). In soil horizons, the calcite-silica deposits occur as coatings, which may be laminated, on stones and as finely disseminated matrix material; these soil deposits may or may not be associated with faults (Vaniman et al., 1984b; Taylor and Huckins, 1986; Taylor and Shroba, 1986; Taylor, in press). Vaniman et al. (1988) used petrographic microscopy, electron microprobe, and X-ray diffraction to examine the mineralogy of the veins at Trench 14, Busted Butte, and possible analog sites at hot and cold springs. They concluded that the deposits in the faults at Trench 14 and sand ramps at Busted Butte are closely related and the process of deposition did not require the upward flow of water from depth (Vaniman et al., 1988). More recent work by Vaniman et al. (1994) emphasizes evidence of:

- acquisition of geochemical signatures for the deposits in overlying soil deposits and not from the brecciated tuff, and
- pervasive distribution of pedogenic fungal microfossils.

These lines of evidence contradict disruptive upward injection and are consistent with only pedogenic origins.

Age of the calcite-silica deposits

Four ages for the calcite-silica deposits have been determined from three samples at one location in a soil upslope from the Bow Ridge fault in Trench 14. The uranium-series disequilibrium method was used to obtain these dates (Szabo and O'Malley, 1985). The field locations of the samples are discussed in Swadley et al. (1984). The three samples (two opals and an opal residue from acid leaching) yield ages of greater than 400,000 yr ago, and the fourth is greater than 350,000 yr ago (Szabo and O'Malley, 1985).

Calcite-silica deposits within the Bow Ridge and Windy Wash fault zones contain basaltic ash that is believed to correlate with basaltic eruptions from either Crater Flat or the Lathrop Wells cone (Taylor, in press). The basaltic eruptions in Crater Flat have been dated at approximately 1.2 Ma and possibly less than 20 ka (Taylor, in press).

Stable isotope studies on the calcite-silica deposits

Carbon and oxygen isotope work on fault-associated calcite-silica and opaline silica samples from Trench 14 indicates that the deposits may have formed at or near surface temperatures (<15°C) (O'Neil, 1984, 1985). Low temperatures of formation may also be supported by the presence of similar mineral assemblages of calcite, opal-CT, and sepiolite in soil deposits (Taylor and Huckins, 1986).

Carbon and oxygen isotope work on calcite-filled fractures from drill-holes in Yucca Mountain yield temperatures that increase with depth and range from 20 to 46° C (Szabo and Kyser, 1985). A maximum geothermal gradient of 43°C/km has been estimated from these data. The gradient represents an upper limit because the calculated temperature does not take

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into account the isotopic fractionation due to other processes. These results compare well with estimated-thermal gradients of 36 to 45°C/km derived from stable isotope studies of calcite-silica from drill core (Sass et al., 1980).

Fluid inclusion studies on the calcite-silica deposits

Limited fluid inclusion data also tend to support a low-temperature origin for the calcite and opaline silica deposits in Trench 14 because the inclusions are single phase as opposed to two- and three-phase inclusions that are typical of high-temperature deposits; however, these inclusions are extremely small and optical data alone do not provide definitive conclusions (Vaniman et al., 1984b).

Paleontological studies on the calcite-silica deposits

Paleontological studies of nineteen soil and vein samples from Trench 14 are consistent with pedogenic origin of the calcite-silica deposits. Eight samples were taken for determination of presence of calcareous microfossils. No ostracodes or other aquatic animals such as mollusks were found (DOE, 1993). The absence of these fossils indicates that they were not deposited in an environment that was saturated for periods longer than a month (Forestor, 1991). Eleven samples of soils and veins were examined for chrysophyte cysts. Rare cysts were found in two vein samples (DOE, 1993). Chrysophyte cysts are more common in areas where surface waters are recharging the hydrologic system than where groundwater is discharging (DOE, 1993). Ostracodes are indigenous to all modern spring-fed pools near the NTS, and ostracodes are ubiquitous in the fossil record of springs (Stuckless, 1986). Furthermore, fungal microfossils are pervasive in these deposits and characteristic of pedogenic, not spring, origins (Vaniman et al., 1994). These types of paleontological evidence argue against a groundwater discharge environment for deposition and are consistent with a pedogenic mode of deposition for the veins (DOE, 1993).

National Academy of Sciences' National Research Council (NAS/NRC) findings

The NAS/NRC was requested by the DOE to evaluate if the water table had been raised to the level of the proposed repository and whether or not such a water-table rise likely to occur over the proposed life of the repository. A panel, established by the NAS/NRC for this purpose, reviewed the pertinent literature, consulted with scientists involved in field and laboratory studies, and spent several days in the field visiting the sites of calcite-silica deposits. Their findings are documented in a report (NAS/NRC, 1992) and the conclusions are discussed below.

The panel's overall conclusion (NAS/NRC, 1992, p. 3) was that none of the evidence cited as proof of groundwater upwelling in and around Yucca Mountain could reasonably be attributed to that process. Further, the panel stated that "The preponderance of features ascribed to ascending water clearly (1) were related to the much older (13-10 million years old (Ma)) volcanic eruptive processes that produced the rocks (ash-flow tuffs) in which the features appear, (2) contained contradictions or inconsistencies that made an upwelling groundwater

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origin geologically impossible or unreasonable, or (3) were classic examples of arid soil characteristics recognized world wide." Based on physical and textural evidence from the veins in the trenches, the panel concluded that a sedimentary, low-temperature origin from descending meteoric water was supported rather than an origin involving upwelling of thermal water from deep within the crust. The panel concluded "that the preponderance of geochemical and mineralogical evidence supports the interpretation that the static water level has been close to its present level deep below Yucca Mountain for the last 100 ka, and possibly as long as 10 Ma, and that the carbonate rich fracture fillings exposed in trenches in the region are composed of wind blown dust cemented by material the deposited from evaporating water that had infiltrated unsaturated rock along open fractures." The panel also concluded "that to date the preponderance of evidence supports the view that the calcretes and other secondary carbonates in veins of the area formed from meteoric water and surface processes." The evidence cited in NAS/NRC (1992) has convinced the panel that the groundwater level at the proposed Yucca Mountain repository site has not reached or exceeded the level of the proposed repository at any time within the last 100 ka.

The panel considered the experience of historic earthquakes and the results of different types of modeling of earthquake responses. Credible assumptions and various types of modeling indicate that the seismic pumping mechanism is inadequate to raise the water table more than tens of meters and that significant water table excursions to the design level of the repository are unlikely.

The panel observed that at Busted Butte, slope-parallel deposits are thinner downslope from the vertical fracture fillings than the slope-parallel deposits above the fracture and that delicately calcified roots of plants are found wrapped around cobbles of volcanic rocks. Both of these observations are deemed inconsistent with formation by upwelling water. Formation of these slope-parallel deposits by upwelling water from the vertical fracture would thus require that the water flow upslope in greater quantity than downslope. Upwelling water would also be expected to flow over the land surface and form deposits at the land surface rather than specifically in the root zone. Both of these observations are consistent with the soil and vein carbonates forming from normal arid climate soil-forming processes.

Occurrence of mosaic breccias at Yucca Mountain and Busted Butte has been suggested as support for upwelling of pressurized groundwater. These breccias consist of angular fragments of local bedrock encased in a matrix of calcite and opaline silica. These rocks have been interpreted as forming as either a result of large and fast buildups and releases of fluid pressure or rapid downslope sliding of bedrock with simultaneous infilling of spaces between the fragments by minerals precipitated from upwelling deep-seated fluid. The NAS panel disagreed with both of these interpretations based upon the field evidence. Three types of breccia were noted: talus breccia, fault breccia, and small pipe-like structures with breccia cemented in them. The panel concluded that the carbonate cementing the talus breccia was deposited by evaporating rain water, progressively and simultaneously with accumulation of the talus deposit. The carbonate cementing the fault breccia was concluded to be of pedogenic or surface origin, based on grain size, structural characteristics, and the presence of older detrital zircons. The most likely origin of the small irregular breccia-filled pipes in

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the Topopah Spring Member was thought to be channeling of steam and other volatiles degassing during the cooling of the Tiva Canyon Member. As a final note, the panel concluded that there was no good evidence in support of upwelling deep hot water to account for the brecciation or the silica-carbonate cementation.

It is possible to predict the isotopic composition of calcite that would be precipitated from water of a known isotopic composition. Comparisons of predicted values of calcite isotopes ($\delta^{18}O_{VSMOW}$; $^{87}Sr/^{86}Sr$; $^{234}U/^{238}U$) for Devils Hole Nevada with measured values show a close match. Performing the same test for measured values of these calcite isotopes at Trench 14, led the panel to conclude that the lack of agreement between values predicted based on groundwater chemistry and measured calcite values demonstrates that the calcites could not have been precipitated from the modern analyzed groundwaters. Available evidence on $\delta^{18}O_{VSMOW}$ and δD_{VSMOW} suggests that ancient groundwater had a similar composition to modern groundwater over the last 300,000 years.

The geologic record indicates that the during the lifetime of the repository, a low volume basaltic intrusion is the only likely style of intrusion in the Yucca Mountain area. The panel concluded that the elastic effect of dike intrusion would result in raising the water table no more than a few tens of meters.

The panel concluded that climate variability models that call for 100% increase in precipitation are probably overly conservative because the pluvial climates in this region were probably much drier than what is normally considered pluvial.

Possible analogs for the calcite-silica deposits

Laminated calcite vein material along fractures in the Amargosa Desert of Nevada and Death Valley in California has been interpreted to represent feeders from springs (Winograd et al., 1985) in a Plio-Pleistocene marsh environment (Hay et al., 1986). The veins are thought to have originated by low-temperature (<50°C) precipitation from calcite-saturated artesian groundwater rising along tensional fractures in Pliocene and Pleistocene alluvium, fanglomerate, and lake beds (Winograd et al., 1985). On the basis of this interpretation, the distribution of these veins indicates that the water table was higher during the Pliocene and early to middle Pleistocene (Winograd and Szabo, 1986). Winograd and Szabo (1986) suggest that the regional water table has declined tens to hundreds of meters in response to tectonics and increased aridity. Although no marsh deposits are known at Yucca Mountain, the calcite-silica deposits have some mineralogical similarity to the marsh deposits in the Amargosa Desert (Hay et al., 1986; Taylor and Huckins, 1986).

3.1.3 Geochemical System

[This subsection will describe the geochemical characteristics of the site, including information on anomalies, properties, and conditions affecting the stability of geochemical characteristics. The difference between unsaturated zone and saturated zone geochemical characteristics, properties, and conditions will also be included.]

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3.1.3.1 Regional Geochemistry

[For the purpose of site characterization of Yucca Mountain, the geochemical subsystem of the geologic setting is the site (see Subsection 3.1.3.2)]

3.1.3.2 Site Geochemistry

[Sufficient data and technical analyses will be provided to describe the saturated and unsaturated zone geochemical characteristics, conditions, and properties of the site. Specifically, the geochemical conditions and processes that exist and have existed during the Quaternary (with emphasis on those that differ significantly from present conditions) will be presented. Assessments of changes that might reasonably be expected to occur in the future will be included.]

3.1.3.2.1 Geochemistry of the Site Rock

[This portion of the Annotated Outline is descriptive in some detail, of the phases and textures present in the Topopah Spring Member. The reader should also note that other important near-field topics such as a discussion of the occurrence of fracture filling minerals, the paleo-hydrothermal system, mineral stabilities under thermal loading, geochemical models of the near-field environment, etc., have not been included in this revision. Temporal constraints have imposed this limitation.

This revision will not include discussion of other units within the saturated and unsaturated zones that form the geologic barrier to the proposed repository. However, what has been presented is an example of the type of analysis anticipated for future revisions.]

Yucca Mountain was one of a number of sites on and near the NTS in southern Nevada that was considered a potential location for a waste repository. Geochemical information had been collected about this area for some time as part of the nuclear testing program at the NTS. The information presently available is a combination of data obtained before Yucca Mountain was considered a potential waste repository site (pre-1977) and data that have been collected as part of the YMP (formerly called the Nevada Nuclear Waste Storage Investigations). Much of the data presented here have been collected as part of the YMP and have been supplemented from other sources to aid interpretation or for confirmation of YMP-related data.

The geochemical data collected as part of the YMP have been obtained by an examination of samples taken from the surface or at depth from Yucca Mountain and vicinity. Mineralogical and petrological samples have come from drill cores, sidewall samples, drill cuttings, and surface outcrops. The mineralogy and petrology of many of the units exposed at Yucca Mountain, based on data from the existing drill holes, have been well characterized. [Future work during construction of the ESF and new drill holes will provide additional data on the mineralogy/petrology of Yucca Mountain rocks; [INN 3.1.3.2.1-1] Experimental methods,

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measurement techniques, data quality, and uncertainties used to obtain the data in this subsection are discussed in detail in the references from which the data were taken.

It is important to characterize the mineralogy and petrology of the rock units in Yucca Mountain for two reasons: First, the phases present in the rock as well as rock texture will determine the effectiveness of the geologic barrier. Second, rocks of the Topopah Spring Member and the tuffaceous beds of the Calico Hills immediately below it will form the near-field environment as well as an outlying zone of elevated temperature, making it imperative that the phases present in these rocks are well understood. This second point is not only related to the stability of sorptive phases under the thermal load of the repository, but also to thermal-mechanical changes that may be induced in the rock by phase changes. Sorption is likely to occur on minerals such as zeolites and clays found between the repository and the water table and below the water table. This obviously requires detailed information about the mineralogy of the rocks.

3.1.3.2.1.1 Geochemistry of the Topopah Spring Member

The phases present in the Topopah Spring Tuff, especially the lower nonlithophysal zone in which the repository will be constructed (Johnstone et al., 1984; DOE, 1988; Byers and Moore, 1987), can be thought of as resulting from three processes. First, glass and phenocrysts represent primary phases which were present in the magma prior to eruption. Vertical variations in the composition and proportion of minerals in ashflow tuffs are generally thought to represent an inverted chemical "stratigraphy" of a magma chamber before eruption (Hildreth, 1979; Lipman et al., 1966; Tegtmeyer and Farmer, 1990). During eruption, lithic fragments of country rock were incorporated into the erupting ashflows. Following emplacement of the ash-flow sheet, which occurred as multiple pulses formed a compound cooling unit (Lipman et al., 1966), hydrothermal effects acting on the glassy matrix and pumice lapilli of the tuff resulted in generally pervasive devitrification as the tuff cooled from 700°C to near ambient temperatures (Byers, 1985). In addition to devitrification, vapor phase deposition of secondary minerals occurred. Other portions of the Topopah Spring Member remained somewhat glassy, such as the upper and lower vitrophyres (Byers, 1985). Third, once the tuff began to approach ambient temperatures (<200°C), further alteration of the rock produced additional low-temperature phase assemblages that included clays and zeolites (Bish, 1989).

The Topopah Spring Member Tuff will form the near-field environment for the potential repository, making it essential that the geochemical, mineralogic, and petrographic characteristics of this unit are well understood. Regardless of the origin of the phases, their three-dimensional abundance distributions are an important part of characterization of the repository horizon because these are the materials that will be interacting with heat from the repository and forming the closest part of the geologic barrier between the waste package and the accessible environment.

The following subsection reviews the present understanding of the vertical and lateral variations in the composition and relative abundances of phases produced by the three

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processes described. Data are primarily taken from analyses performed on a set of four boreholes (USW G-1, USW G/GU-3, and especially USW G-4 and UE-25 a#1) that have been extensively studied and bound the CPDB (Figure 3.1.3.2.1.1-1). Additional data from future and existing boreholes, as well as the ESF, will be incorporated in future revisions of this document [INN 3.1.3.2.1-1]. These data will more fully capture variations in the mineralogy and petrology of the host rock. Byers and Moore (1987) note that major stratigraphic, mineralogic, and textural changes occur in the Topopah Spring Tuff between these drill holes and USW G-2. However, that well is located 1.6 miles nearly due north of USW G-1 well away from the CPDB; thus, although major lateral lithologic variations in the Topopah Spring Tuff may occur, the mineralogic and geochemical data from the four closer wells are likely to be representative of the proposed repository horizon.

3.1.3.2.1.1.1 Major Element Composition

It is important to briefly consider the major element chemistry of the Topopah Spring Tuff because it either directly controls or reflects the primary, hydrothermal, and diagenetic mineral distributions in the unit. Minor and trace element abundances do not control mineralogic distributions within the Topopah Spring Member; therefore, they will be considered only as they mirror petrologic processes operational therein. In the International Union of Geologic Sciences classification scheme (Le Bas et al., 1986), rocks of the Topopah Spring Tuff range from trachyte to high-silica rhyolite (Figure 3.1.3.2.1.1-1). Because the more mafic end member is described as latite in most references, the use of that term will be retained although it is not used in the International Union of Geologic Sciences scheme. Latite is usually used to describe rocks, such as the uppermost portion of the Topopah Spring Tuff, with sub-equal normative or modal abundances of plagioclase and alkali feldspar.

As will be shown below, modal phenocryst mineralogy (Broxton et al., 1989) of the Topopah Spring Tuff indicates a fairly sharp compositional break between the latite and rhyolite in the upper portion of the section, reflecting eruption from a zoned magma chamber (e.g., Lipman et al., 1966; Tegtmeyer and Farmer, 1990) which contained a distinct, rather than a gradational, compositional break (e.g., Schuryatz et al., 1989; Peterman et al., 1991) between rhyolite and latite magma. This is reflected in the major element chemistry where most analyses cluster into two fields with few mixed intermediate compositions (Figure 3.1.3.2.1.1.1-1)

Sm-Nd data (Figure 3.1.3.2.1.1.1-2) indicate that the latite has higher rare-earth element (REE) abundances and is more enriched in light REEs than the rhyolite. Implicit in this observation is that some magmatic process produced the partitioning of the REE, probably by fractional crystallization, if the two magmas were co-genetic. The sharp compositional break between the two magma types as recorded in the tuff is illustrated by the data plotted in Figure 3.1.3.2.1.1.1-2. The two rock types have distinct apparent Sm/Nd ratios as seen in data (circles, Figure 3.1.3.2.1.1.1-2) derived by isotope dilution (Tegtmeyer and Farmer, 1990; Peterman et al., 1991), a very precise method. However, these data were derived from whole rock samples from drill core. Lipman et al. (1966) and Schuryatz et al. (1989) noted that pumice lapilli are heterogeneous in composition throughout the unit, especially within the

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quartz latite. A mixing line between average rhyolite and latite reproduces the variation in REE abundances in latite; thus, much of the spread in the latite data is probably due to the presence of rhyolite pumices in the latite (Figure 3.1.3.2.1.1.1-2). Neutron activation data from Schuryatz et al. (1989), although somewhat less precise, were determined predominantly on pumice samples rather than whole rock powders (squares, Figure 3.1.3.2.1.1.1-2). These data tend to cluster into distinct end member compositions consistent with the hypothesis of two distinct magmas with few intermediate compositions (3.1.3.2.1.1.1-2).

The major element chemistry of the Topopah Spring Tuff also records the effects of clay and zeolite diagenesis in certain horizons. The low alkali samples in Figure 3.1.3.2.1.1.1-1 are from a diagenetically altered horizon at the upper top of the basal vitrophyre from wells USW G-4 and UE-25 a#1 (zone I of Broxton et al., 1986), and contain smectite clays and clinoptilolite-family zeolites (Mg- and Ca-rich clays and zeolites respectively) of varying proportions (Bish and Chipera, 1989). Clearly, alkalis have been lost from these samples, even in instances where alteration was relatively minor.

The relative open-system mobility of major elements in the altered samples can be assessed in a normalized enrichment diagram (Figure 3.1.3.2.1.1.1-3). This figure was produced by dividing the major-element concentration of each altered sample by a representative, whole-rock, high-silica rhyolite sample unaffected by diagenesis. This permits a comparison of diagenetically altered to unaltered samples. It is readily apparent that alteration has produced manifold enrichments in divalent cations (Ca, Mg) and water, whereas monovalent cations (Na, K) have been lost. The crystallization of clinoptilolite appears to be responsible for the enrichment of Ca and Mg, whereas smectite is responsible for the enrichment of Mg alone. The small apparent depletion of Si may be due to increased hydration during diagenesis; however, the depletion of alkalis is so pronounced that alkalis must have left the system as the depletions cannot be explained by the constant sum effect.

Peterman et al. (1991) presented a similar plot to Figure 3.1.3.2.1.1.1-3 for minor and trace elements, indicating that other divalent cations such as Sr have also been enriched within the altered horizon, whereas monovalent cations such as Rb have been lost and REE have remained relatively immobile. Peterman et al. (1991) note that the hydrothermal alteration of the Topopah Spring Tuff may be an important analog for a repository in the same unit. [These studies are being pursued under Study Plan 8.3.1.3.2.2, History of Mineralogic and Geochemical Alteration at Yucca Mountain (Bish 1991); [INN 3.1.3.2.1.1.1-1].] The mobility of some elements (e.g., Sr) is directly relevant to radionuclide migration in the near-field environment. The immobility of the REE, analog elements to the actinides, suggest that their migration may be strongly inhibited within the host rock (Peterman et al., 1991). The distribution, mineral chemistry, and age of diagenetic phases will be considered in further detail in Subsection 3.1.3.2.1.1.4.

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3.1.3.2.1.1.2 Primary Phases

Modal Abundances. To characterize the mineralogy of the Topopah Spring Tuff, especially the repository horizon, data are presented from boreholes within (USW G-4) or near (UE-25 a#1) the CPDB. An extensive review of the petrography, upon which much of this discussion is based, was presented by Broxton et al. (1982a, 1989). [Additional data [INN 3.1.3.2.1-1] will be gathered from the ESF and new boreholes. Future revisions of this document will incorporate these data as well as additional existing data.]

The uppermost ~60 m of the unit comprise the latite which is the most crystal rich part of the Topopah Spring Tuff. The latite contains up to 15-20 percent phenocrysts (Figure 3.1.3.2.1.1.2-1) with sub-equal proportions of alkali feldspar and plagioclase (Figure 3.1.3.2.1.1.2-2). From about 21-27 m from the top of the unit, the number of phenocrysts steadily decreases to less than 5 percent. It is at this point, at about 60 m below the upper contact, that the Topopah Spring Tuff grades into high-silica rhyolite which generally contains less than 2 percent phenocrysts. A fair amount of scatter in phenocryst abundances (Figures 3.1.3.2.1.1.2-1 through 3.1.3.2.1.1.2-3) is observed down section, due to the difficulty in accurately measuring representative modal abundances in phenocryst poor rocks. The Topopah Spring Tuff also contains lithic fragments (Figure 3.1.3.2.1.1.2-1). As might be expected from a pyroclastic unit emplaced from one (albeit compound) eruptive cycle, the overall abundance of xenoliths increases toward the base of the unit. The first material out of the vent would be expected to preferentially incorporate available lithic materials. Although USW G-4 is located only 1.03 km WNW of UE-25 a#1, there are considerable differences in absolute lithic abundances (Figure 3.1.3.2.1.1.2-1). USW G-4 contains a maximum of 2-3 percent lithic fragments in the high silica rhyolite within the proposed repository horizon, generally showing a steady increase down section. However, UE-25 a#1 contains two lithic-rich zones (Figure 3.1.3.2.1.1.2-1). The uppermost zone roughly coincides with the upper one third of the lower lithophysal zone (Byers and Moore, 1987) from 230 to 270 m in depth and reaches a maximum lithic content of about 8 percent. The lithic-rich lowermost zone is within the proposed repository horizon, or lower nonlithophysal zone. Here the lithic content reaches a maximum of about 14 percent. It is anticipated, therefore, that there may be considerable vertical and lateral heterogeneity in xenolith content within the Topopah Spring Tuff, including the proposed repository horizon.

In terms of relative felsic phenocryst variations (Figure 3.1.3.2.1.1.2-2), the latite is composed of plagioclase and sanidine with little or no quartz. Although both feldspars are present as discrete phases throughout the Topopah Spring Tuff, sanidine mantles plagioclase in some instances in the latite. Toward the top of the latite, sanidine dominates the assemblage with 60-80 percent alkali feldspar. The relative abundance of sanidine decreases down section. Plagioclase is the most abundant phase in the rhyolite, although it may decrease slightly in relative abundance down section as quartz increases. Alkali feldspar comprises approximately 20 percent of the felsic phenocrysts in the rhyolite. Quartz first appears consistently at a depth of 122-137 m from the top of the unit (Figure 3.1.3.2.1.1.2-2) and increases more or less systematically until it comprises up to 20 percent of the felsic phenocrysts. Within the proposed repository horizon, phenocrysts in the rock are 5-20 percent quartz, 15-30 percent

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sanidine, with the remainder comprised of plagioclase. Again, the scatter in Figure 3.1.3.2.1.1.2-2 is due to the difficulty in measuring sparse phenocryst modes.

Mafic phenocryst abundances show consistent vertical and lateral trends (Figure 3.1.3.2.1.1.2-3). Biotite is the most abundant mafic phase in the latite, comprising a maximum of 12,000 ppm (1.2 percent by volume, 10,000 ppm=1 percent). Sparse clinopyroxene is also present in the latite, and is generally absent or very sparse in the rhyolite. Fe-Ti oxides are present as several hundred ppm throughout the Topopah Spring Tuff, although they are somewhat more abundant in the latite. Hornblende is also found in small quantities throughout the unit (Broxton et al., 1989). Although mafic phenocrysts comprise up to 1-2 percent of the latite (Figure 3.1.3.2.1.1.2-3), they decrease steadily with depth until the Topopah Spring Tuff transitions into rhyolite where they comprise less than 2000 ppm with relatively constant abundances. Broxton et al. (1989) report the presence of perrierite (a REE silicate) as an accessory phase in the latite and allanite as an accessory phase in the rhyolite. Apatite and zircon can be found throughout the Topopah Spring Tuff, although sphene is absent.

At the time of emplacement, the majority of the magma was quenched as glass, now represented by matrix. Some of the glass has been preserved in upper and basal vitrophyres of the Topopah Spring Tuff (Figure 3.1.3.2.1.1.2-4). However, much of the material was subjected to sub-solidus high-temperature crystallization of the glass and vapor phase precipitation. The nature and distribution of the minerals produced by this process will be discussed in Subsection 3.1.3.2.1.1.3.

Mineral Chemistry. The alkali feldspars are intermediate in composition, ranging from Or_{38-66}, and tend to be slightly more potassic in the rhyolite (mean Or_{57}) than in the latite (mean Or_{47}), despite considerable compositional overlap with the latite. Broxton et al. (1989) also note that sanidine becomes increasingly Or-rich toward the base of the unit within the rhyolite. Plagioclase rim compositions for both rhyolite and latite range chiefly from An_{10-20} with a dominant composition at An_{17}, although the total range is from An_{10} to An_{50}. Plagioclase cores range in composition from about An_{10-35} in both latite and rhyolite, although higher An contents are more common in the latite.

Unlike the feldspars, significant differences exist in the Mg# of the biotites (Mg#=Mg/(Mg+Fe)x100). Biotite in latite ranges from $Mg\#_{60-80}$, with a strong mode at about $Mg\#_{63}$. As expected, biotite in rhyolite is more Fe-rich ranging from $Mg\#_{30-65}$ with a mean composition of about $Mg\#_{42}$. Unlike the latite, there is a broader distribution of biotite compositions in the rhyolite. Reported pyroxene compositions (Broxton et al., 1982a) in latite samples from drill hole USW G-2 range narrowly from $Mg\#_{71-74}$. Schuryatz et al. (1989) analyzed Fe-Ti oxide phases in an attempt to infer temperatures and oxygen fugacities in the magma chamber at quenching. Their study showed that both ilmenite and magnetite are present as both discrete phenocryst phases and exsolution lamellae within phenocrysts.

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3.1.3.2.1.1.3 High-Temperature Secondary Phases

Mineralogy. Although most of the mass of the Topopah Spring Tuff was originally composed of hot ash and pumice (glass), high-temperature post-emplacement processes resulted in devitrification and vapor-phase deposition. Such phases are fine-grained and difficult to quantify by point counting or electron-probe microanalysis. As a result, the rocks have been analyzed by whole-rock quantitative X-ray diffraction (e.g., Bish and Chipera, 1989). This gives an analysis of all materials in the rock regardless of their origin. X-ray diffraction is especially useful in quantifying phases produced during devitrification of the glass and high-temperature vapor-phase deposition in cases where this material comprises a majority of the rock. This is especially true for the rhyolite which generally contains less than 2 percent phenocrysts; therefore, little overprinting from the phenocrysts results. As will be seen in a subsequent section, the X-ray technique is also useful in describing the distribution of diagenetic minerals such as clays, zeolites, and oxide minerals, phases considered to be of particular importance for retarding the migration of radionuclides.

Several workers have been particularly active in characterizing the rocks at Yucca Mountain by X-ray diffraction, and several reports have been produced (e.g., Bish and Vaniman, 1985; Bish and Chipera, 1989; Chipera and Bish, 1988). These reports have been superseded by additional data and corrections to older data in a report by Bish and Chipera (1989) which provides an excellent summary of mineral distributions at the site from drill hole samples. In addition, this report gives an extensive account of the analytical details, difficulties, and uncertainties in quantitative identification of phases by x-ray diffraction. [Additional data is presently being gathered from recent boreholes, and new data will be gathered from planned boreholes and the ESF [INN 3.1.3.2.1-1].]

Although it is beyond the scope of this portion of the document to recount the analytical procedures in detail, a few items regarding precision and accuracy are worth noting as they may influence apparent mineral distributions and subsequent interpretation of the data. For instance, all feldspar, including plagioclase, is reported as alkali feldspar. The analyses do not discriminate between feldspar present as phenocrysts and feldspar produced by high-temperature alteration, although the latter is clearly dominant on a volumetric basis. Solid solution and preferred mineral orientation may affect the apparent concentration of certain phases. As a result, reported analytical uncertainties for individual phases are often on the order of a few tens of percent of the analytical total for each mineral (Bish and Chipera, 1989). Although this may seem large for any given phase, the analytical procedure is rapid, allowing many samples to be analyzed to build a good statistical data base. Some of the scatter in the data as presented may result from limitations to the level of precision attainable by this technique. However, in spite of any limitations on precision imposed by this analytical technique, x-ray diffraction has the advantage of providing positive identification of phases present in the rock even in instances where they occur in low abundances.
One of the prominent characteristics of the Topopah Spring Tuff is that it consists primarily of devitrified material between glass-rich horizons located near the top and bottom of the unit (Figure 3.1.3.2.1.1.2-4). It appears however, that a smaller proportion of glass is preserved in those horizons in USW G-4 compared to UE-25 a#1. USW G-4 exhibits a prominent clay

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and zeolite-rich zone between the lower vitrophyre and the lower nonlithophysal zone (Figure 3.1.3.2.1.1.2-4). The upper and lower portions of the unit show distinct lateral heterogeneity in the proportion of glass and authigenic/diagenetic phases, the origin and mineralogy of which will be described in Subsection 3.1.3.2.1.1.4.

Figure 3.1.3.2.1.1.2-4 shows that the devitrified central portion of the Topopah Spring Tuff consists almost exclusively of 30-35 percent microcrystalline silica polymorphs with alkali feldspar comprising most of the remainder of the rock. The presence of the metastable silica polymorphs tridymite and cristobalite attest to the relatively high temperatures, probably in the presence of water vapor, that produced the alteration. In other units, Bish (1989) noted that cristobalite appears to have reacted to quartz in horizons that have subsequently experienced temperatures in excess of about 100°C. However, it appears that this has occurred in a hydrothermal environment given the presence of diagenetic phases. Thus, the inversion of cristobalite to quartz may involve dissolution and reprecipitation from an aqueous fluid. Both high and low cristobalite are metastable phases at low temperatures and pressures, and cristobalite does not invert to its high polymorph (a displacive phase transformation) unless temperatures on the order of 260-270°C are reached (Hurlbut and Klein, 1977). It seems, therefore, rather improbable that cristobalite would undergo a reconstructive phase transformation at 100°C without some other component, such as water, to mediate its transformation. Thus, the nearly ubiquitous presence of cristobalite (except within the vitrophyres, Figure 3.1.3.2.1.1.2-4) implies that the Topopah Spring Tuff at Yucca Mountain has probably not been reheated in the presence of water for an extended period to temperatures sufficiently high to result in its inversion to quartz.

Throughout much of the unit, all three silica polymorphs are present. Within the latite at the top of the unit; however, the silica polymorphs tend to consist of tridymite and cristobalite alone (Figure 3.1.3.2.1.1.2-4). In the middle of the section in both drill holes there is a tridymite-free zone 18 m thick in USW G-4 and 40 m thick in UE-25 a#1 (Figure 3.1.3.2.1.1.2-4). This zone corresponds to the middle non-lithophysal and uppermost lower lithophysal zones in USW G-4, and the uppermost lower lithophysal zone in UE-25 a#1. Tridymite is also generally absent in the lower nonlithophysal zone, or potential repository horizon (Figure 3.1.3.2.1.1.2-4), a relationship that holds true for other drill holes (USW G-1 and GU-3) within or near the CPDB (compare data from Byers and Moore, 1987, and Bish and Chipera, 1989).

Textures. Numerous workers (e.g., Bish et al., 1982; Broxton et al., 1982a; Byers, 1985; Byers and Moore, 1987) provide detailed petrographic descriptions of the textures present in the Topopah Spring Member due to devitrification and vapor phase deposition/alteration. Bish et al. (1982) describe the interior of the Topopah Spring Member by noting petrographic differences in terms of the upper and central-lower portions of the unit. The upper interior zone roughly corresponds to the latite. This zone is densely welded and devitrified, and vapor-phase alteration is pervasive. Pumice lapilli are flattened and primary voids within them are now occupied by spherulites and aggregates of alkali feldspar and cristobalite overprinted by small, numerous pockets of tridymite. Devitrification and vapor-phase crystallization have eradicated most primary igneous textures in the pumices. Devitrified

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shards comprise most of the groundmass, and smectite clays have imparted a dark-brown hue in plane-polarized light. Vugs in this portion of the interior are lined with tridymite.

The central-lower interior portions of the Topopah Spring Member, including the proposed repository horizon, are also moderately to densely welded, devitrified, and exhibit vapor-phase alteration. In general, moderately welded horizons are interstratified with densely welded tuff. Joins between welding horizons probably represent upper and lower contacts to discrete cooling units or eruptive pulses. Bish et al. (1982) describe four stages of devitrification and vapor-phase crystallization in the pumice lapilli. Small spherulites of cristobalite sheaves or other axiolitic minerals nucleated along the outer margins of the pumices and grew inward. Second, interlocking spherulites partially occupied the primary void spaces in the pumices. Third, vapor-phase tridymite and alkali feldspar were deposited throughout the pumices. Finally, granophyric crystallization infilled remaining interstices. Clay is fairly abundant in the groundmass along with granular alkali feldspar and quartz. Plumose, asymmetrical spherulites are common in some samples as are lithophysae that are partially filled with alkali feldspar and pseudomorphs of quartz after tridymite.

Byers and Moore (1987) apply statistical textural analysis, describing textural types similar to those reviewed in the previous paragraphs, to the stratigraphic designations of the Topopah Spring Member as reported by the USGS (Spengler et al., 1979; 1981; Spengler and Chornack, 1984; and Scott and Costellanos, 1984) in several drill holes. This facilitates comparisons of vertical and lateral variations in textural types within the unit, especially within the proposed repository horizon.

The distribution of textures is plotted in Figure 3.1.3.2.1.1.3-1. As expected, the upper and lower vitrophyres are easily distinguished by abundant glass (Figure 3.1.3.2.1.1.3-1). However, this is not as apparent for the upper vitrophyre as in Figure 3.1.3.2.1.1.2-4(b) because the data do not extend as far up section (compare Figures 3.1.3.2.1.1.3-1 and 3.1.3.2.1.1.2-4). Other horizons within the Topopah Spring Tuff can be recognized on the basis of textural information combined with phenocryst and lithic abundances. The crystal-rich (sanidine and plagioclase) vapor-phase zone largely corresponding to the latite (Figure 3.1.3.2.1.1.2-2) also contains an increasing proportion of spherulites and microlites at the expense of the cryptocrystalline texture (Figure 3.1.3.2.1.1.3-1). The upper lithophysal zone, although similar in many respects to the lower lithophysal zone, is generally less welded and contains tridymite-bearing amygdules (Figure 3.1.3.2.1.1.3-1). The middle non-lithophysal zone is marked by an increase in the proportion of cryptocrystalline texture relative to spherulites, and microlites relative to the zones above and below it (Figure 3.1.3.2.1.1.3-1). This distinction is even more obvious in other drill holes where the proportion of cryptocrystalline texture is even greater (Byers and Moore, 1987). The lower non-lithophysal zone is also characterized by a high proportion of cryptocrystalline groundmass which increases significantly at the contact with the lower lithophysal zone. The lower non-lithophysal zone can be distinguished from the middle non-lithophysal zone by stronger welding and the presence of quartz phenocrysts and lithic fragments. The marked increase in cryptocrystalline material at the 270 m depth within the lower nonlithophysal zone (Figure 3.1.3.2.1.1.3-1) is unique to UE-25 a#1.

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The lateral variations in textures of the proposed repository horizon are illustrated in Figure 3.1.3.2.1.1.3-2. Most of the available data are from the center of the horizon and represent descriptions at only 3-5 discrete horizons. In spite of this, there appears to a general consistency in the textures. The percentage of cryptocrystalline material in the matrix increases deeper in the lower nonlithophysal zone at the expense of spherulites/microlites. On average, both textures appear to be present in sub-equal proportions.

3.1.3.2.1.1.4 Low Temperature Phases

The subsolidus alteration and crystallization/recrystallization of the Topopah Spring Tuff occurred over a range of magmatic to ambient temperatures. For the purposes of this document, the appearance of zeolites, clay minerals, and other minerals of low-temperature origin are considered to be diagnostic of diagenetic alteration or alteration at the low-temperature stages of waning hydrothermal alteration. Bish (1989) provides convincing arguments that the presence of clinoptilolite and disordered interlayered smectite/illite combined with persistent cristobalite within the Topopah Spring Tuff indicate diagenesis at less than about 100°C.

As was observed in Figure 3.1.3.2.1.1.2-4, diagenetic minerals comprise more than 20 percent of the rock of the Topopah Spring Tuff in certain horizons. Figure 3.1.3.2.1.1.4-1 expands the scale in Figure 3.1.3.2.1.1.2-4 in order to examine the distribution of low abundance, low-temperature phases in the rest of the unit, again relying on the data of Bish and Chipera (1989). Clay and zeolite are observed throughout the Topopah Spring Tuff, but generally in quantities of just a few percent or less (Figure 3.1.3.2.1.1.4-1).

There are three important horizons within the Topopah Spring Tuff in which clay and zeolite are particularly abundant, although significant variations occur from one location to another:

- at the top of the unit above the upper vitrophyre;
- at the top of the basal vitrophyre; and
- below the basal vitrophyre.

More recent work, however, with UE-25 and UZ-16, shows that the zeolite stilbite may form up to 14 percent of some portions of the devitrified Topopah Spring Tuff in the zone of the potential repository horizon. More work on new core samples and on ESF samples is required to understand the range of mineralogy in the potential host rock.

Byers and Moore (1987) report that the clay and zeolite-rich zone at the base of the proposed repository horizon is on the order of 1-6 m thick and they place it stratigraphically in the lower nonlithophysal zone, although Levy and O'Neil (1989) consider the horizon to represent altered vitrophyre. The zone is part of the Zeolite Interval I (Bish and Vaniman 1985) or diagenetic Zone I (Broxton et al., 1991). On the whole, it contains highly variable amounts of smectite clays and clinoptilolite-family zeolites. For example, the top of the basal vitrophyre in USW G-4 consists of 45 percent smectite and 28 percent clinoptilolite, whereas the corresponding portion of the altered zone at the base of UE-25 a#1 consists of only 5

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percent clinoptilolite and 1 percent smectite (compare Figure 3.1.3.2.1.1.4-1 and Figure 3.1.3.2.1.1.2-4). Bish and Chipera (1989) report that "clinoptilolite" may include heulandite, a compositionally similar zeolite with a lower thermal stability. On a microscopic scale, veinlets of smectite lace the relict framework of flattened glass shards, and zeolites occupy void spaces in relict pumice shards, microvesicles, and veinlets. Levy and O'Neil (1989) also noted the lateral variations in alteration which were concentrated along fractures which controlled fluid transport.

Broxton et al. (1982a) interpreted this altered horizon at the top of the basal vitrophyre as diagenetic alteration associated with locally perched groundwater. Levy and O'Neil (1989) contended that the alteration resulted from hydrothermal alteration at somewhat elevated temperatures (40-100°C) on the basis of oxygen isotope geothermometry of coexisting quartz and zeolite. In either case, the low permeability of the vitrophyre may provide a mechanism to explain the alteration of the top of glassy horizons in both the upper and lower vitrophyres, and its lateral variations in intensity.

The partially welded to nonwelded horizon below the vitrophyre is pervasively altered and contains abundant zeolites and a few percent clay (Figures 3.1.3.2.1.1.2-4, 3.1.3.2.1.1.4-1) (Bish and Chipera, 1989) that comprise the majority of the rock. However, it should be noted that this horizon is largely vitric in the western portion of the site. Broxton et al. (1986) consider the zeolites below the vitrophyre to be part of Zeolite Interval II. At one horizon within UE-25 a#1 a few percent mordenite, which is rare in the Topopah Spring Tuff (except in fracture lining assemblages from all drill holes in which it commonly occurs; Carlos et al., 1991), has been identified in addition to clinoptilolite (Bish and Chipera, 1989). A few percent opal-CT have been identified in this horizon in UE-25 a#1 (Figure 3.1.3.2.1.1.4-1). Recent work suggest that opal-CT is present wherever clinoptilolite and mordenite occur, although it was not quantified separately for cristobalite in previous studies using older XRD methods (Bish and Chipera, 1989).

A few percent clay are present in the Topopah Spring Tuff in USW G-4 above the upper vitrophyre (Figure 3.1.3.2.1.1.4-1). Above the vitric cap rock in UE-25 a#1 is a horizon with about 25 percent smectite clays, and minor zeolite (Figure 3.1.3.2.1.1.4-1). USW G-2 also has an altered horizon with about 75 percent clinoptilolite and a few percent clay. Bish et al. (1982) similarly proposed that the upper vitrophyre acted as a local paleo-impermeable zone producing perched water in USW G-2 from which essentially complete alteration into clay and zeolite resulted.

The regions labeled "other" (Figure 3.1.3.2.1.1.4-1) are the sum of mica and/or hematite as determined by quantitative X-ray diffraction, and the greater abundance of biotite in the latite accounts for the abundance of "other" minerals near the top of the unit. Although modal biotite (Figure 3.1.3.2.1.1.2-3) is observed throughout the Topopah Spring Tuff, it apparently is not sufficiently abundant to be detected throughout the entire unit (Figure 3.1.3.2.1.1.4-1). The zone of "other" minerals in USW G-4 is a mica-bearing horizon correlated with the zone of clay and zeolite mineralization at the base of the lower nonlithophysal zone.

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Mineral Chemistry. Bish (1989) reports that clays in the upper portions of drill holes USW G-1, G-2, and G-3, including the Topopah Spring Tuff, contain less than 20 percent disordered illite in smectite. Zeolites from the altered horizon at the top of the basal vitrophyre are Ca and Mg-rich in terms of exchangeable cation concentrations (Figure 3.1.3.2.1.1.4-2). They cluster rather tightly and distinctly from zeolites below the vitrophyre. As mentioned above, the zeolites at the top of the vitrophyre are considered to be part of Zeolite Interval I. Broxton et al. (1991) define this interval to include partially vitric tuffs above the water table. Zeolites above the vitrophyre in J-13, however, are below the static water level and still cluster with the other analyses. Nonetheless, they are the most alkali-rich of any of the zeolites above the basal vitrophyre.

Zeolites from the partially welded to nonwelded horizon below the vitrophyre are generally more variable in composition and usually more alkali-rich (Figure 3.1.3.2.1.1.4-2). These zeolites are also preserved above and below the present day water table. This again implies that they have retained at least part of their distinctive cation distributions. Another feature of the zeolites from this horizon is a compositional intra-borehole variability in samples separated by only tens of feet in vertical distance. This is illustrated by clusters of data points from the common boreholes enclosed by the dashed line. Particularly apparent are distinct clusters of zeolite compositions from drill hole J-13 which are separated by 11 m. Thus, it is apparent that there is considerable vertical and lateral compositional anisotropy in the zeolites of this horizon, even in saturated tuffs.

The distinct cluster of zeolite compositions from all three horizons in J-13 implies that only partial cation exchange of the zeolites, if any, has occurred in the water-saturated Topopah Spring Tuff at J-13 despite a lengthy residence time in relatively Na-rich waters (DOE, 1988). Otherwise, one might expect the zeolites to undergo cation exchange toward a common composition rather than maintaining their distinct chemical identities.

Bish (1989) demonstrated a strong geochemical interdependence between hydrothermal cristobalite, diagenetic smectite/illite, and clinoptilolite/heulandite. If this is the case, a confident age determination of one phase is probably a good estimate of the age of all the phases. K-Ar age determinations of zeolites from the Topopah Spring Tuff (WoldeGabriel et al., 1992) range from 4.2-4.3 Ma. However, zeolites may not be particularly retentive of Ar (McDougall and Harrison, 1988) and are certainly subject to cation exchange of K^+ with other cations. Thus it is not surprising that zeolites may not record their crystallization age, although some other zeolites from deep within the saturated zone at Yucca Mountain seem to record the inferred age of diagenesis (WoldeGabriel et al., 1992). K-Ar ages of the clay minerals suggest that the diagenetic assemblage may have originated at about 10-11 Ma (Bish and Aronson, 1993, in press; Broxton et al., 1991; Bish, 1989), although it should be noted that these ages can only be strictly applied to a deep diageneitc zone well below the potential repository. The ages are fairly reproducible from sample to sample, indicating that the clays probably have not been profoundly affected by Ar loss or open system behavior of K. This range of ages corresponds fairly well to the time of Timber Mountain volcanism, and infers that diagenesis in the Topopah Spring Tuff may have been derived from a distal, low-temperature portion of the associated hydrothermal system. In any case, available data

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suggest the diagenetic assemblage has remained stable for an extended period of time, perhaps since the cessation of regional silicic volcanism.

3.1.3.2.1.2 Geochemistry of Calico Hills Formation

3.1.3.2.2 Geochemistry of the Site Groundwater and Gas

[Subsection 3.1.3.2.2 was not provided in this revision of the Annotated Outline Process.]

3.1.3.2.3 Site Geochemistry Governing Radionuclide Mobility

[Subsection 3.1.3.2.3 was not provided in this revision of the Annotated Outline Process]

3.1.4 Climatological and Meteorological Systems

Past and present climatological and meteorological conditions at Yucca Mountain are described in this subsection. Also, an assessment of future climatic variation is provided. Information in this subsection provides descriptions of the conditions representative at the site and surrounding region. These descriptions are provided as supplements to the Safety Analysis Report, required for the License Application.

3.1.4.1 Present Climate and Meteorology

Baseline climate and meteorology of the site are described in this subsection. These data characterize the climatological and meteorological processes to the site and are used to design and predict the performance of a geologic repository. Both site-specific and regional data are presented. Meteorological conditions, or weather forecasts, must be considered in engineering design, surface facilities placement, and radiological safety assessment. The description of present meteorological conditions also serves as input to other investigations, such as the rainfall-runoff assessments.

Although a major emphasis throughout the site characterization process focussed on assessing the ability of the selected host rock to contain stored wastes, the climate and site-specific meteorology of the Yucca Mountain area can influence some important aspects of repository development. The design and operation of the repository must consider climatic influences to ensure that surface facilities are capable of withstanding expected meteorological conditions (e.g., the design of the ventilation system). The potential for flooding must also be considered in the siting of surface facilities. In addition to these short-term design considerations, defining the existing climatic conditions establishes the basis for comparing the future and past climates with the present climate. Further, establishing the existing climatic conditions and the infiltration rates associated with these conditions is important in evaluating whether climatic variations will affect infiltration rates and subsequent rises or declines in the water table in the Yucca Mountain area.

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3.1.4.1.1 Climate

[Long-term site-specific climatological data for Yucca Mountain have been provided by 14 monitoring stations.] Nine of these site-specific stations are designed for collecting data for use in atmospheric dispersion calculations. The other five supplement the data collection of precipitation data for rainfall-runoff studies. Also, data from other regional weather stations in the Yucca Mountain vicinity operated by the National Weather Service (NWS), NWS cooperative observers, the Federal Aviation Administration, the U.S. Department of Defense, Desert Research Institute, and the Bureau of Land Management have been used to provide a general description of the climate in the area. Figure 3.1.4.1.1-1 shows the locations of the regional monitoring sites. Figure 3.1.4.1.1-2 shows locations of site specific monitoring stations at the Yucca Mountain Project area [INN 3.1.4.1.1-1].] A complete list of the site-specific and regional monitoring stations and the parameters measured is given in Table 3.1.4.1.1-1 [INN 3.1.4.1.1-2]. The meteorological monitoring program, described in Subsection 3.1.4.1.2, provides site-specific data on the meteorological conditions likely to influence site characterization activities or repository development.

The Yucca Mountain Site is situated in an area bordering two NWS climatological zones of Nevada: south central and extreme southern (Bowen and Egami, 1983b). The distinction between these two classifications is governed mostly by elevation. Lower elevations in the vicinity of Yucca Mountain experience conditions typical of southwestern desert zones within the United States that are characterized by hot summers, mild winters, and limited amounts of precipitation. Higher elevations have less severe summer temperatures and greater but still limited amounts of precipitation. The general climatological classification of mid-latitude desert, in a modified Koeppen system presented in Critchfield (1983), also can be used to describe conditions in and around the Yucca Mountain Site. Mid-latitude desert areas are far removed from windward coasts and are dominated by tropical and polar air masses. For areas classified as mid-latitude desert, summers are dominated by continental tropical air masses and winters are dominated by continental polar air masses. Large annual and diurnal fluctuations in temperature are characteristic of mid-latitude desert areas, as is significant variability in precipitation from year to year.

3.1.4.1.1.1 Scales of Atmospheric Motion and General Climatic Perspective

Discrete atmospheric disturbances or motions occur over a wide spectrum of time and space. This spectrum is divided into four scales within which most atmospheric motions or disturbances can be categorized. These scales, as originally classified by Ligda (1951) are: *planetary scale, synoptic scale, mesoscale, and microscale.* The original scales classified by Ligda (1951) are not rigid and the atmospheric motions they represent may overlap in space and time. Figure 3.1.4.1.1.1-1 depicts these scales in relationship to each other in space and time. The following discussion provides a relative perspective of how atmospheric motions on the larger scales affect the microscale environment of Yucca Mountain.

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3.1.4.1.1.1.1 Planetary Scale

The largest scale motions occur on a planetary scale because of the unequal heating of the earth's surface. Warm buoyant air rises from equatorial regions and flows poleward in both hemispheres in a meridional fashion. This rising air induces convergence at the surface, generating the easterly trade winds, and divergence aloft, leading to the high altitude westerlies. This large scale cell of convection circulation, known as the Hadley Cell, finally sinks in the subtropical latitudes (approximately 30° [north and south]) causing a belt of high-pressure at the surface in the subtropics. A weaker circulation cell transports energy toward the poles in a pattern called the Ferrel cell. A third very weak polar cell may exist in each hemisphere (Gedzelman, 1985). Therefore, on a planetary scale, air is rising at 0 and 60° [north and south] and sinking at 30 and 90° [north and south]. This basic atmospheric mechanism is responsible for all planetary atmospheric motion.

The primary planetary features of the atmosphere which control the movement of synoptic-scale systems are the polar and subtropical jet streams or jets. The jets are a combined product of differential heating mentioned above and the torque generated by the earth's rotation. The jet streams generally circumnavigate the earth in a five-wave pattern of troughs and ridges with wavelengths of 2000-4000 km. This pattern may vary from three to eight major waves (Rossby waves) and oscillates northward in northern hemisphere summer and southward in northern hemisphere winter. The intensity and motion of the atmosphere at any given time depends on many variables including sea surface/continental temperature contrasts. The time and length scales for planetary-scale motion are on the order of weeks and 10,000 km, respectively.

3.1.4.1.1.1.2 Large-Scale Atmospheric Features

At the synoptic scale, air masses of different origin and properties, such as moisture, temperature, and pressure interact as low barometric pressure caused by warm rising air is replaced by cold air from areas of high barometric pressure. The air flow from high to low pressure is the cause of winds which vary in intensity depending on the strength of the gradient between high and low pressure centers. The strongest temperature discontinuity is usually found along the polar front. Figure 3.1.4.1.1.1.2-1 [INN 3.1.4.1.1.1.2-1] depicts the average position of the polar front in January. Low-pressure systems forming along the polar front (frontogenesis) over the oceans generally follow the jet stream or "storm track" on shore and release their precipitation as they move. Wendlend and Bryson (1981) provide more detailed insight into the mean seasonal and annual frontal positions in the Northern Hemisphere. They discuss the confluences of near-surface airstreams which represent mean surface storm tracks. Terrain features greatly affect the movement and strength of low-pressure frontal systems. Synoptic-scale systems average on the order of 1000-2000 km size, and exist on a time scale of days.

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There are two distinctly different storm patterns affecting the desert climate, one in winter, the other in summer. Winter precipitation tends to be of low intensity and long duration (days), and covers large areas. In contrast, most summer rains result from convective thunderstorms of high intensity and short duration (hours), and are limited in size.

The major air masses affecting the weather of the Yucca Mountain area during winter months originate either over the Pacific Ocean or over polar continental regions. The polar and subtropical jet streams or "storm tracks" steer developing storm systems eastward and inland. Most of the moisture carried by the Pacific air masses does not reach the Yucca Mountain area because the physical (orographic) lifting effect of the Sierra Nevada forces the air masses to higher elevations as they move eastward, thus cooling the air masses and lowering their ability to hold moisture. The moisture that cannot remain in the vapor phase, due to this cooling, falls as precipitation on the western slopes of the Sierra Nevada. When an air mass has passed the ridge of the mountains, it descends along the eastern slopes and warms again, creating what is called a rain shadow in the lee of the Sierra Nevada (Wallace and Hobbs, 1977). Yucca Mountain and the surrounding areas lie within this shadow. Polar continental air masses bring cold, dry air into the area but are not as common a winter phenomenon as are the Pacific air masses.

In spring, the west-to-east storm track typical of winter storms migrates northward as cold air retreats and warm air invades from the south. A thermally induced area of low pressure is created during the summer months over most desert regions (Wallace and Hobbs, 1977) and prevails over the southwestern United States during summer. Although this thermal low is generally associated with weak cyclonic motion (Huschke, 1959), it brings south to southwesterly winds to the Yucca Mountain area. However, this circulation pattern is essentially nonfrontal. Although summer is generally the driest time of the year, this circulation pattern can bring tropical moisture originating over the Pacific Ocean off the lower coast of California to the area, which in combination with the strong solar insolation during the summer can create thunderstorm activity. Another less frequent summer circulation pattern that brings moisture to the area is a semipermanent subtropical high-pressure system called a Bermuda High (Huschke, 1959). If this system becomes well developed, it can bring moisture from the Gulf of Mexico to the Yucca Mountain area with southeasterly winds, again resulting in thunderstorm activity. By midsummer, air-mass thunderstorms become the dominant storm type in the region. These storms are fueled by the extreme heating over the "hot deserts," and the increased moisture being adverted from the Gulf of California and/or Gulf of Mexico.

Review of historical data (Elliot, 1943) reveals that despite apparent, small-scale, erratic movements of lows, fronts, and troughs, certain large-scale patterns dominate movements of the weather systems. There is a readily recognizable overall pattern to the steering currents that establish the most probable track that storms will follow. Steering currents that move surface and upper-air lows on long, over-water trajectories to southern California and the southern Great Basin may produce precipitation across southern Nevada.

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The climatology of storm systems can be classified into five winter weather types (A-E), as well as the predominant summertime storm type (the "Southwest Monsoon"). Of the five winter storm types, only three produce precipitation in southern Nevada. Of those, one produces only an insignificant amount of precipitation. However, all five types are described here for completeness because lack of precipitation is a major consideration in characterizing the climatology of Yucca Mountain. Storms associated with the southwest monsoons have greater potential for flooding than winter storms. Figures 3.1.4.1.1.1.2-2 through 3.1.4.1.1.1.2-6 [INN 3.1.4.1.1.1.2-2] depict time-lapse movements of winter storm systems, as discussed by Elliot (1943).

Weather Type A (Figure 3.1.4.1.1.1.2-2 [INN 3.1.4.1.1.1.2-2]). A strong ridge in the upper-air steering current is present at about 150° west longitude with a trough near 110° west longitude. At the surface, the eastern lobe of the Pacific High is strong and located well to the northwest of the normal winter position. Typically, a wave pattern forms along the southern coast of Alaska or near the British Columbia coast and is directed rapidly by the upper-air steering currents down the coast over coastal Washington and onward over the Great Basin. At this point, the wave has occluded and slowed as cold air advancing behind the system begins recurving, moving the low center eastward over the Rocky Mountains near Denver, Colorado. Typically, southern Nevada experiences no precipitation from this type of weather pattern.

Weather Type B (Figure 3.1.4.1.1.1.2-3 [INN 3.1.4.1.1.1.2-2]). This scenario features a belt of high pressure dominating the eastern Pacific and the southern half of the United States. Surface lows are directed from the Gulf of Alaska eastward through northern Canada with fronts trailing into southern Canada and the northern United States. As the frontal systems pass the Continental Divide, a maritime polar air mass from the Pacific Ocean follows, forming a temporary high pressure area over the Great Basin. No precipitation results from this type of weather in southern Nevada.

Weather Type C (Figure 3.1.4.1.1.1.2-4 [INN 3.1.4.1.1.1.2-2]). This type of weather system produces precipitation in southern Nevada and the YMR. Note that the belt of high pressure is displaced northward with steering currents to the south. The cyclone centers are displaced southward to about 30° latitude, migrate onshore, then move eastward along the United States-Mexico border. This scenario is often referred to as the "San Diego Low." This weather type produces the heaviest non-thunderstorm precipitation known in southern California and Nevada. Storms generally move slowly through the area, prolonging the duration of low clouds and precipitation. As the system moves onshore, and until its core moves east of 110° west longitude, rainfall generally continues throughout southern Nevada. When the trough west of the California coast persists, a succeeding wave develops and moves onshore with similar precipitation patterns. This cycle repeats as long as the high-pressure system to the north continues to feed the trough lying west of the California coast.

Weather Type D (Figure 3.1.4.1.1.1.2-5 [INN 3.1.4.1.1.1.2-2]). This weather system is occasionally of importance to southern Nevada. However, as a general rule, it is of little significance to precipitation in southern Nevada; the movements of the surface lows and

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frontal systems are too far north and east. On the occasion when weather type D is of importance to southern Nevada, the strong sub-arctic high and steering currents are displaced slightly south of the depicted position. A polar outbreak that occurs through Alberta, Montana, and Wyoming extends into the Great Basin and southward into southern Nevada. Strong north to northeast winds occur as the cold air passes. Temperatures drop to well below normal, and light precipitation, with snow near the 1200 m level, falls on the mountains.

Weather Type E (Figure 3.1.4.1.1.1.2-6 [INN 3.1.4.1.1.1.2-2]). A cold continental high persists in northern Canada and Alaska with the eastern lobe of the Pacific high centered near 20° north latitude and 140° west longitude. The steering currents for the Pacific storm systems lie in the belt of low pressure between the two highs. The Type E weather pattern can be broken up into subgroups, depending on the strength and position of the Canadian high- pressure system. If the Canadian high develops to a moderate 1030 millibars (mb), the storm tracks are displaced further to the south. When the Canadian high reaches a strong 1040 mb and extends south-eastward over most of the North American continent, the Pacific storm tracks are at their southernmost limit for the Type E systems. In each subgroup of the Type E systems, the surface low crosses the Pacific coast at 45 to 50° north latitude, following the steering currents as it progresses eastward across the western United States. Warm tropical Pacific air is drawn northward up the west coast, ahead of the approaching low, and causes heavy precipitation along the coast. A low-pressure center then develops on the lee side of the Sierra Nevada and moves across south-central Nevada. This development, often termed the "Tonopah Low," may bring precipitation to southern Nevada, depending on the amount of atmospheric moisture available. The speed of the storm is gaged by studying the air to the north of the storm trajectory (i.e., when the air to the north is exceptionally cold, the low moves rapidly).

Southwest Monsoon (Figure 3.1.4.1.1.1.2-7 [INN 3.1.4.1.1.1.2-3]). The summer (July and August) weather regime in southern Nevada exhibits a totally different circulation pattern than is characteristic of winter. It contributes to a secondary, though less discernible, maximum in annual precipitation. Thunderstorm activity over southern Nevada is the hallmark of this pattern known as the "Southwest Monsoon." These storms have a much greater potential for flooding than the frontal precipitation in the winter months. Much more precipitable water is available in summer due to the heated atmosphere's increased ability to hold moisture. Therefore, these storms can release a large amount of precipitation in a relatively short time.

Summertime atmospheric dynamics exhibit more of a tropical rather than mid-latitude character primarily because of the location of the Rocky Mountains to the east and the proximity of the Pacific Ocean to the west. A thermally induced area of low pressure is created during the summer months over most desert regions (Wallace and Hobbs, 1977) and prevails over the southwestern United States during summer. Although this thermal low is generally associated with weak cyclonic motion (Huschke, 1959), it brings south to southwesterly winds to the Yucca Mountain area. However, this circulation pattern is essentially non-frontal. Defining the location of the moisture source(s) fueling storm activity in southern Nevada during the summer monsoon season has not been completely resolved.

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Bryson (1957), Hastings and Turner (1965), Weaver (1962), Tubbs (1972), Houghton (1969), and the U.S. Weather Bureau (1966) believe the moisture source to be the Gulf of Mexico for precipitation events in southern California and the Great Basin. However, Rasmusson (1967) found that low-level moisture flow (below 800 mb) was southwesterly while high-level flow (750-500 mb) was from the south-southeast. In 1972, Pyke stated that he believed late-summer precipitation in the desert southwest was fed by moisture from the warm waters of the Gulf of California, the Pacific Ocean (south and west of Baja California), and to a certain extent, the Gulf of Mexico. He concluded that easterly waves from the Gulf of Mexico could trigger thunderstorms, and that the most likely source of moisture is the Gulf of California or even the Pacific Ocean west of Baja. Hales (1972 and 1974) held a similar view. He analyzed summer monsoon moisture in southern Arizona and concluded that the greatest percentage of tropical moisture in the southwestern United States and northwestern Mexico comes from the Pacific Ocean via the Gulf of California rather than the Gulf of Mexico. Studies prior to 1970 seem to favor the Gulf of Mexico argument, while studies since 1970 favor the Gulf of California conclusion.

In its most active state, the monsoon manifests itself by forming "mesoscale convective complexes" over Arizona and New Mexico. On rare occasions, an Mesoscale Convective Complexes will intrude into southern Nevada, offering the greatest opportunity for localized flash flooding, hail, and strong winds. Hansen (1975) studied heavy rainfall incidents at Elko, Nevada; Phoenix, Arizona; and Morgan, Utah. He concluded that the Pacific Ocean serves as the source of low-level moisture in all three cases. Figure 3.1.4.1.1.1.2-7 illustrates a simplified circulation pattern of the Southwest Monsoon. The Bermuda High intrudes westward into New Mexico with a lobe extending over Arizona. This creates the pressure gradient which causes low-level moisture flow from the Gulf of California.

3.1.4.1.1.1.3 Mesoscale Atmospheric Features

In addition to these synoptic-scale climatic influences, the rugged terrain of the YMR can create meso-scale meteorological variations of a given parameter. Meso-scale motions generally occur on a length scale of 10 to 100 km. At this scale not only is topography a significant factor, but surface frictional effects come into play. One example of meso-scale phenomena is the mountain-valley wind effect. This circulation effect results from diurnal heating and cooling of mountain sides and valley floor throughout the cycle of a 24-hour period. This heating results in an upslope breeze which intensifies throughout the afternoon. This effect is the "valley breeze." After sunset, the ground surface quickly cools by radiating its heat into the atmosphere. Air very near the surface subsequently cools. This cooler, denser air then drains down the terrain. This effect is known as the "mountain breeze" and these conditions generally dissipate quickly after sunrise as the sun heats the ground surface. These breezes are gentle because of frictional effects and the shallowness of the layer (less than 100 m).

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In closed topographic basins such as Yucca Flat, the cold drainage air essentially fills the basin and can significantly lower the temperature in the basin with respect to temperatures at the rims of the basin. The temperature inversion (temperature increasing with height) thereby created can limit effective atmospheric dispersion within the basin.

These types of terrain-dependent winds can locally affect wind speed, wind direction, and temperature (Eglinton and Dreicer, 1984), and are most pronounced under calm (synoptic-scale) conditions during cloudless nights (Huschke, 1959).

Another important meso-scale phenomenon is the thunderstorm. A typical thunderstorm results from thermal convection when the atmosphere is statically unstable. The distribution of thunderstorms is a direct consequence of the heating patterns coupled with the availability of moisture. A number of variables influence thunderstorm activity and govern precipitation amounts, intensity, and duration. Individual storms generally complete a thunderstorm cycle in about 2 hours. Storm movement is predominantly along a northerly trajectory, but will move in a northwesterly or northeasterly direction depending on the upper-air flow pattern. As a rule of thumb, a summertime dew point of 10°C indicates that enough low-level moisture is available to generate thunderstorms. There are several types of thunderstorms. Those that result purely from local heating are called "air mass thunderstorms" and tend to be less violent. Those that develop in response to a lifting mechanism such as a cold front or mountain range (orographic effect) are sometimes more severe with heavy rain, hail, and strong winds. Although the average single thunderstorm cell may cover an area of only 10 km^2, a cluster, Mesoscale Convective Complexes may form covering a larger area (200 km^2). Mesoscale Convective Complexes may cause heavy flash flooding. As discussed in the previous subsection, the southwest monsoon can produce this feature. The most severe thunderstorms develop into squall lines ahead of a cold front; however, this effect is not common in the southwestern U.S.

Another example of meso-scale meteorological variations induced by the terrain is the variability in precipitation amounts between stations at different elevations and between those with differing exposure to prevailing storm tracks or storm occurrences. French (1983) analyzed the distribution of precipitation resulting from winter and summer precipitation regimes across southern Nevada. He concluded that two zones (one winter and one summer) of precipitation can be identified and are separated by a transition zone. These zones, south of 38.5° north latitude, contain major orographic features which influence precipitation. In winter, the flow regime is from the west, resulting in a "rain shadow" east of the mountains, as discussed in the previous subsection. In the summer, the primary source of moisture is from the southwest. The air flow curves to the east as it crosses southern Nevada as shown in Figure 3.1.4.1.1.1.3-1 [INN 3.1.4.1.1.1.3-1]. The net effect of these two phenomena is to provide some areas of southern Nevada with a relative excess of precipitation while other areas receive a relative deficit. The effect of this on the Yucca Mountain Site is _____ [INN 3.1.4.1.1.1.3-2 (to be determined from pcpn studies)].

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Orographic influences on precipitation are important to consider in an analysis of precipitation trends resulting from synoptic-scale and mesoscale weather patterns and their relations to sources of moisture.

French (1986) analyzed precipitation-elevation relations and calculated a positive gradient of 28 mm annual average precipitation per 305 vertical meters using linear regression and a sample of 63 selected precipitation stations in southern Nevada. He also calculated a gradient of 38 mm per 305 m for a sample of 12 stations on the NTS with lengths of record of 10 years or more. Havesi (1990) calculated a gradient of 31 mm per 305 m for the subset of French's sample consisting of 42 stations with lengths of record of 8 years or more. In both studies, log-linear relations were found to provide slightly improved regression results compared to linear regressions. Figure 3.1.4.1.1.1.3-2 [INN 3.1.4.1.1.1.3-3] shows a plot of annual average precipitation as a function of station elevation for the subset of 42 stations selected by Havesi (1990). The best-fit curves obtained for this sample and for the two samples analyzed by French (1986) are also plotted for comparison.

3.1.4.1.1.1.4 Microscale Atmospheric Features

At the finest scale over which distinct atmospheric disturbances occur is the microscale occurring on a typical length scale of 1 to 1000 meters. Examples of micro-scale meteorological phenomena are tornadoes and dust devils. Dust devils, a fair-weather occurrence, are a product of intense heating on the desert floor. They are typically 10 m in diameter with a rotational velocity of 10 m/s and frequently occur during summer months. Tornadoes are much more violent products of severe thunderstorms and are very rare in the southwest U.S. They are on the order of several hundred meters in diameter although some have been reported as wide as 1 km or more. Tornado life-cycles are on the order of minutes to one hour. Rotational wind velocities are on the order of 100 to 150 m/s.

3.1.4.1.1.2 Climatological Parameters

Records of meteorological parameters that together describe the climate of the region include temperature, precipitation, atmospheric moisture, surface wind speed and direction, upper air wind speed and direction, atmospheric pressure, sunshine, and severe weather phenomena. Each of these variables is discussed in detail in the subsections that follow.

3.1.4.1.1.2.1 Temperature

Temperature in the site vicinity varies widely on both a diurnal and annual basis (Quiring, 1968; Eglinton and Dreicer, 1984). Temperature data from the __-yr climatological summary for the sites at and around the YMR are presented in Table 3.1.4.1.1.2.1-1 [INN 3.1.4.1.1.2.1-1], (when summary data collected from regional sites and site data are summarized). General temperature cycles and ranges of expected temperature values for the Yucca Mountain Site are discussed below.

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The lowest temperatures generally occur during the months of November through March. Although temperatures below 0° F (-18°C) do occur, they are infrequent and commonly occur on less than 1 percent of the days from November through March. Despite the apparent frequency of temperatures below 32° F (0°C), daily maxima during the winter months remain quite mild, averaging at least 50° F (10°C).

Substantial temperature differences between average daily maximum and average daily minimum are characteristic of the area due to high insolation rates and generally low relative humidities.

Based on the data presented here, general trends can be inferred. Summer temperatures in excess of 100° F (38°C) should be expected. Winter minimum daily temperatures below 32° F (0°C) will be relatively common, but temperatures below 0° F (-18°C) will be infrequent. There will probably be substantial temperature ranges, with cool summer nights and mild winter daily maximums during most years.

3.1.4.1.1.2.2 Precipitation

Precipitation in the Yucca Mountain area is associated with two distinct atmospheric circulation patterns. The first of these patterns creates winter frontal passages associated with Pacific air masses moving toward the area from the west. Approximately __ [INN 3.1.4.1.1.2.2-1 (fill in when data are determined and available)] percent of the precipitation in the vicinity of Yucca Mountain occurs as a result of these systems during the months of November through April, even though the entire area lies in the rain shadow of the Sierra Nevada. The second type of circulation pattern that occurs in the area creates a secondary peak in precipitation in the late summer (July and August) and is a result of thunderstorm activities. These storms have a much greater flood potential than the frontal precipitation that occurs during the winter months because the storms can release significant amounts of moisture in relatively short periods of time.

The low-pressure area that typically dominates the southwestern United States during the summer results in south to southwesterly winds in the Yucca Mountain area that can transport moisture-laden air into the area. The convective activity associated with this moist air flow can result in strong (intense), isolated downpours throughout the area. If the ground surface cannot absorb this moisture quickly, runoff and flooding can occur. Flooding potential is high; a large number of variables influence thunderstorm activities and govern precipitation types, amounts, and intensity, and individual storms generally complete a thunderstorm cycle in about 2 hours (Huschke, 1959).

Specific precipitation amounts for the 14 weather stations and 116 other precipitation gages located in the vicinity of Yucca Mountain are presented in Table 3.1.4.1.1.2.2-1 [INN 3.1.4.1.1.2.2-1 (fill in when data are available)]. Monthly and annual average and maximum precipitation amounts at these stations are shown in the table. The data cover periods of record ranging from __ yr at _____ to __ yr at __ (see Figures 3.1.4.1.1-1 and 3.1.4.1.1-2 for locations).

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All the stations follow the characteristic annual precipitation cycle with a winter peak, an early summer minimum, a secondary peak in late summer followed by a secondary minimum in October. This cycle is clearly illustrated in Figure 3.1.4.1.1.2.2-1 [INN 3.1.4.1.1.2.2-2], which shows precipitation amounts for each month, based on the average data presented in Table 3.1.4.1.1.2.2-1. The maximum values for each month do not exhibit the same annual cycle evident in the monthly averages. Instead, these data indicate that maximum precipitation generally occurred during winter months with maximums for the various towers having occurred in ___ (month). This variability is indicative of individual frontal systems or storms having occurred during these months.

For the operation phase of the repository and during site characterization, the precipitation resulting from short-term thunderstorms of high intensity is a consideration in design and placement of surface facilities, including those for the exploratory shaft, because these storms can lead to flash flooding (sheet flow, stream flow, and debris flow). Data from monitoring stations provide some indication of the amount and intensity of precipitation that has occurred at Yucca Mountain. The data represent conditions for approximately __ yr. The data are included only as an indicator of the potential thunderstorm-related precipitation.

The most significant storm event recorded during the __-yr period occurred on ____. On that date, the ___ site recorded a 24-hour total precipitation amount of ___ mm, ____ mm of which _____ fell in 1 hour.

The Yucca Mountain area does receive precipitation in the form of snow, but such occurrences are uncommon (Nichols, 1986). The snow that does fall persists for only a few hours. The greatest daily amount of snowfall recorded at ____ is ___ mm, which occurred in _____. Snow is not important in terms of overall precipitation amounts but should be considered in the design of the surface facilities.

Because the repository would be located in the unsaturated zone beneath Yucca Mountain, evaluation of the long-term ability of the site to contain stored waste must include a determination of how much of the precipitation falling at the surface infiltrates as potential recharge to the groundwater. While thunderstorms are significant events and potentially damaging, they occur in the summer months when soil moisture is low and potential evapotranspiration is high and thus are not likely to result in significant groundwater recharge (Nichols, 1986). The most likely events leading to infiltration that exceed soil moisture deficit and evapotranspiration and could thus lead to percolation through the repository horizon would occur during the winter months. A series of precipitation events with no intermediate drying-out period would represent high-recharge potential (Nichols, 1986). A discussion of hydrologic infiltration and flux through the unsaturated zone is found elsewhere in Subsection 3.1.2.3.

A comprehensive precipitation monitoring network is utilized both in the immediate vicinity of Yucca Mountain and in sections of the Fortymile Wash drainage to fully evaluate the recharge potential. A description of this network is provided in Subsection 3.1.4.2, Paleoclimatology.

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3.1.4.1.1.2.3 Atmospheric Moisture

The same processes that restrict precipitation in the vicinity of Yucca Mountain provide for generally low relative humidity throughout the year. The general diurnal and annual cycle of relative humidity for the area can be derived from the data contained in the climatological summary for Yucca Flat shown in Table 3.1.4.1.1.2.3-1. Data collected for others sites is presented in Table 3.1.4.1.1.2.3-2 [INN 3.1.4.1.1.2.3-1]. On the basis of these data, relative humidity is expected to reach a minimum monthly average of approximately __ percent in June or July and reach a maximum of around ____ percent in December. Diurnal trends (also shown in Table 3.1.4.1.1.2.3-1) are for the highest relative humidity to occur during the early morning and late evening hours, with the lowest relative humidity occurring in the afternoon hours.

Although relative humidity, a temperature-dependent variable, is commonly used as an accepted measure of atmospheric moisture, it can be a somewhat misleading indicator. A less temperature-dependent indicator of atmospheric moisture is the wet-bulb temperature or depression, which can either be measured directly or derived from relative humidity, and dry-bulb temperature data using a psychometric chart for the elevation. Because the wet-bulb temperature is always less than or equal to the ambient (dry-bulb) temperature (Huschke, 1959), the difference between the dry-bulb temperature and the wet-bulb temperature is referred to as the wet-bulb depression.

The monthly average relative humidity at hour 1600 and the maximum average temperature data given in the climatological summary for _____ (Table 3.1.4.1.1.2.1-1) [INN 3.1.4.1.1.2.1-1] have been used to calculate approximate wet-bulb depression values expected to occur in the vicinity of the Yucca Mountain Site. These wet-bulb depression data were calculated using a psychometric chart given in Weast (1972). The data are shown in relation to the ambient temperature and humidity values for _____ in Table 3.1.4.1.1.2.3-3 [INN 3.1.4.1.1.2.3-2]. As shown in the table, the wet-bulb depression varies inversely with the relative humidity.

3.1.4.1.1.2.4 Wind Speed and Direction

Yucca Mountain lies in a geographical region of generally linear mountain ranges that dissect alluvial piedmont valleys with rugged, complex terrain features. Wind speeds and, more importantly, wind direction will be heavily influenced by the site-specific terrain features. Therefore, extrapolation of wind data from other sites to Yucca Mountain may not be as accurate as extrapolation of other meteorological parameters.

An analysis of winds at 14 locations nearest the YMP site provides data on conditions that might be experienced at Yucca Mountain. Descriptions of the tower locations are provided in subsequent sections on the site meteorological monitoring network (Refer to Subsection 3.1.4.2).

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Data from the towers indicate that wind direction is influenced primarily by two general types of atmospheric activity. First, large-scale pressure systems govern seasonal variations in wind direction and produce predominantly northerly winter winds and predominantly southerly summer winds. Strongest wind speeds are associated with passage of frost or thunderstorms. Secondary to the overall patterns are terrain-induced wind flow patterns and the effects of ground surface heating and cooling (mountain-valley winds). The influence of these effects is evident in the diurnal wind flow reversal from upslope flows during the day to drainage flow at night. The directions associated with these flows are an artifact of the terrain in the vicinity of the tower.

Wind speeds associated with the various flow patterns in the Yucca Mountain area are highest during the mid-afternoon hours and reach minimum speeds both shortly after sunrise and shortly after sunset. This pattern is also due to the terrain influences in which minimum mountain-valley wind speeds represent flow reversals or directional changes, and maximum mountain-valley speeds correspond to periods when intense surface heating initiates upgradient air movement. The exception to this pattern is _____. At this location, wind speed minimums occur during _____ hours in _____.

Surface wind data from the _____ stations have been summarized by frequency of occurrence for the period ____ to ____ and are summarized by season and year, as shown in Figures 3.1.4.1.1.2.4-1 through 3.1.4.1.1.2.4(n) [INN 3.1.4.1.1.2.4-1] (provide for all stations when available) as wind rose plots. Figures _____ to _____ show the same wind roses with data split for daytime versus nighttime wind frequencies, [INN 3.1.4.1.1.2.4-2] (provide wind roses on terrain maps).

[The following discussion should be provided for either all or a specific subset of the site monitoring stations.]

The spring (March through May) distribution shows that winds from the ___ occur most frequently and account for ___ percent of the observations during the season. Winds from the ____ also occur quite regularly, accounting for ____ percent of the spring observations. Overall, winds from the ____ through the ____ and ____ through ____ are dominant during the season, with nearly __ percent of the recorded observations having occurred in one of these six directions.

The summer season (June through August) distribution is dominated by winds from the ____, accounting for ___ percent of the distribution. Winds from the ____ through ____ are clearly the most common during the summer months and represent ___ percent of the total observations. The ___ component of the summer wind flow pattern is not nearly as pronounced as in the spring distribution, with winds from the ____ accounting for ___ percent of the observations. This _____ component in the summer months is most likely due to the nighttime drainage winds at this site, which were discussed previously.

Winds from the ____ during the fall (September through November) were most common and were observed ___ percent of the time, with winds from the _____ through the ____

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accounting for ___ percent of the observations. There is also a strong ____ly element of the fall wind rose, representing ____ percent of the recorded observations. The fall data show a general variability in wind direction that is not as evident in the other seasonal distributions.

The winter (December through February) distribution shows a predominance of _____ly winds that alone account for ____ percent of the total observations. Winds from the _____ through _____ are the most common observations and account for ___ percent of the total distribution. The _____ly component evident in the fall wind rose is also significant during winter months, occurring ___ percent of the time.

On an annual basis, winds from the ____ occur somewhat more frequently than winds from the _____ (____ percent from the _____ versus ____ percent from the ____). Excluding these two dominant wind flow quadrants, the balance of the annual wind rose exhibits a relatively uniform distribution, somewhat skewed to the _____ and ____ due to the terrain in the vicinity of the _____ tower.

Wind speeds occurring at _____ are generally less than __ mph (___ m/s) during all seasons, with generally higher wind speeds occurring in the _____ and somewhat lower wind speeds (overall) occurring in the ____ and _____ seasons.

The general information indicates that although extreme wind phenomena are potentially damaging, they do not present obstacles to design, construction, or operation of the proposed repository.

3.1.4.1.1.2.5 Upper Air Data

3.1.4.1.1.2.5.1 Winds

Data on meteorological conditions, specifically wind speed and direction, at levels above those measured by the various towers at the YMP site are useful in assessing the possibility of long-range transport of potential repository emissions. Data collected at the Yucca Flat weather station from 1957 to 1964 and summarized in Quiring (1968) provide upper air wind data. The upper air data from the Yucca Flat weather station are for mid-season months only (January, April, July, and October). Times of collection begin at 0400 Pacific standard time and end at 1600 Pacific standard time at 3-hour intervals, and these mid-season data are assumed to represent an entire season.

Data at 1,524 m above mean sea level (328 m above ground level), which are similar to surface observations, are shown in Table 3.1.4.1.1.2.5.1-1 [INN 3.1.4.1.1.2.5.1-1] and plotted as wind roses in Figure 3.1.4.1.1.2.5.1-1 [INN 3.1.4.1.1.2.5.1-1]. Winds from the northwest through northeast during winter (January) are shown to occur 60.3 percent of the time, with average speeds of approximately 5.4 m/s. Summer patterns at the 1,524 m level are virtually opposite the winter data, with southeasterly to southwesterly winds occurring 74.9 percent of the time at average speeds of about 5.8 m/s. Spring southerly (southeast through southwest) winds occur somewhat more frequently than do northerly (northwest through northeast) winds,

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46.2 percent versus 36.1 percent. Fall patterns balance just slightly opposite the spring data, with 44.5 percent northerly winds compared with 39.4 percent southerly winds. On an annual basis, the seasonal north-to-south change in wind direction is about equal: 43.5 percent of the time from the north (northwest through northeast) and 41.2 percent of the time from the south. Data at 1,829 m above mean sea level (633 m above ground level) presented in Table 3.1.4.1.1.2.5.1-2 [INN 3.1.4.1.1.2.5.1-2] and shown in Figure 3.1.4.1.1.2.5.1-2 [INN 3.1.4.1.1.2.5.1-2] show patterns similar to the data at the 1,524 m levels, but they are considered to be more synoptically influenced than at lower levels.

Another source of upper air data is the twice daily rawinsondes taken by the NWS at Desert Rock, Nevada. Table 3.1.4.1.1.2.5.1-3 [INN 3.1.4.1.1.2.5.1-3 (summary data from upper air winds needed)] and shown in Figure 3.1.4.1.1.2.5.1-3 [INN 3.1.4.1.1.2.5.1-3 (create figure from tabular wind data as for previous two tables/figures)]. These data represent a measure of the wind flow at mid- and upper-levels of the atmosphere. [INN 3.1.4.1.1.2.5.1-4 (discuss comparison of Desert Rock data with historical Yucca Flat upper air data)]. These winds would be the steering currents for long-range transport of potential releases from Yucca Mountain.

3.1.4.1.1.2.5.2 Mixing Heights

Mixing height is the distance above the ground to which relatively unrestricted vertical mixing occurs in the atmosphere. The mixing height is an indication of the potential vertical extent of pollutant diffusion. As is typical in desert-like environments, the mixing height in the Yucca Mountain area can be quite high during the afternoon (especially in summertime) due to strong convective activity which can mix the air vertically throughout the lower levels of the atmosphere. Conversely, mixing heights can be quite low in the morning because of nighttime radiative cooling, which causes surface-based or low-level temperature inversions to form.

Inversions restrict the vertical movement of the air in the lower atmosphere, thereby preventing atmospheric pollutants from mixing with the air above the inversion. Atmospheric mixing in regions of the lower elevations can be greatly restricted when an inversion limits the vertical mixing and surrounding terrain limits horizontal mixing. Spatial variation in mixing depth are complex depending on local elevations and ground surface properties. Mixing heights normally increase during the day and decrease during the night.

The annual average morning mixing height in the YMP area is estimated to be _____. [INN 3.1.4.1.1.2.5.2-1] and the annual average afternoon mixing height is about _____ [INN 3.1.4.1.1.2.5.2-1].

3.1.4.1.1.2.6 Atmospheric Pressure

Atmospheric pressure is a measure of the force that air exerts over a unit area in any direction. Pressure is simply the product of the mass of air in an overhead column and the acceleration of gravity. Because the acceleration of gravity is nearly a constant 9.8 m/s^2

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through the lower atmosphere, station pressure is proportional to the overhead column mass. The overhead column mass is affected by temperature changes and/or convergence or divergence of air. For example, high temperatures and/or horizontal divergence in the air column lower pressure, whereas lower temperatures and/or horizontal convergence in the air column increase pressure.

The normal atmospheric pressure at sea level for a standard atmosphere is 1013.3 mb (Berry, Bollay, and Beers, 1945). However, the atmosphere quickly thins out with height with an associated decrease in pressure, especially in the first 6000 m. The standard atmospheric pressure is about _____ mbar [INN 3.1.4.1.1.2.6-1] (fill in when information available from monitoring data) at _____ (fill in station name), elevation ____ m above sea level.

Monthly mean atmospheric-pressure statistics are shown in Figure 3.1.4.1.1.2.6-1 and Table 3.1.4.1.1.2.6-1 [INN 3.1.4.1.1.2.6-1] (prepare when information available from monitoring data, examples attached). The data in Figure 3.1.4.1.1.2.6-1 show that Yucca Mountain station pressure has a definite annual cycle, with a summer maximum and a spring minimum. The spring minimum is caused by frequent storms. The summer higher pressures result when the mid- and upper-atmosphere westerlies move to the north and weaken while the subtropical high-pressure systems move north toward the southern United States. Variations from the mean are the greatest during the spring and least during the summer.

Even though the mean pressure is highest during summer, the summer maximums are lower than during winter months. The highest recorded pressure in the __ yr record at Yucca Mountain (Station ____) was recorded in _____ (month). Winter high-pressure readings result from cold, low-level high-pressure systems. Extreme high pressure in the summer results from middle- and upper-level high-pressure systems.

Pressures adjusted to sea level were also calculated from estimated temperature and humidity profiles for elevations below those at Yucca Mountain (Table 3.1.4.1.1.2.6-1) [INN 3.1.4.1.1.2.6-1] (calculate and include in table). Unlike actual station pressures, pressures adjusted to sea level are related to temperature, reaching a peak in January and a minimum in July. The sea-level pressure follows the annual temperature cycle, whereby a theoretical atmospheric-pressure profile lower than that at Yucca Mountain would show a reduced pressure (and density) because of atmospheric heating. Indeed, during the summer months, a low-level "heat low" is a common feature on weather maps in the deserts of southern California, Nevada, Arizona, New Mexico, and northern Mexico.

Hourly station pressure is plotted in Figure 3.1.4.1.1.2.6-2 for January, April, July, and October. [INN 3.1.4.1.1.2.6-1 (will prepare when information becomes available from monitoring data, example attached).]

Note a distinct diurnal cycle prevails during all the months, with a maximum during the late morning and a minimum during the late afternoon. The diurnal cycle is associated with the alternate heating and cooling of the atmosphere. Because cold air is denser, pressure increases when the lower atmosphere is colder. The maximum atmospheric cooling from the

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prior evening occurs late in the morning, well after sunrise. The maximum pressure occurs earlier in the morning during summer when the sunrise is earliest. Conversely, mean pressure reaches a minimum in the afternoon, lagging behind the strongest sunshine and surface heating.

3.1.4.1.1.2.7 Insolation (Sunshine)

Insolation (defined as incoming solar radiation) has been measured at the Yucca Mountain since ____ (yr). Measured by a pyranometer, insolation is the energy that strikes a plane parallel to the earth's surface.

Solar energy strikes the top of the earth's atmosphere at a rate of 1.4 kW/m^2 on a surface perpendicular to the sun's rays. This rate (1.4 kW/m^2) is referred to as the solar constant.

Several factors affect the amount of radiation that reaches the earth. First, the angle of the sun above the horizon has a great effect. At the summer solstice (~ June 21), the local noon sun is closest to its zenith (highest position in the sky) and therefore provides the maximum possible surface radiation. Also, the angle of the sun during the summer at high elevations results in longer days, contributing to the increased total insolation. In contrast, at the winter solstice, the low sun angles and short days give relatively little solar radiation. Second, the thickness of the atmosphere, clouds, and aerosols also affect the amount of solar radiation received at ground level.

The atmosphere is an excellent scatterer of incoming sunlight. As sunlight reaches farther down into the atmosphere, direct-beam radiation is reduced while downward-scattered (diffuse) radiation increases. In addition, an increasing amount of solar radiation is lost to upward scattering (backscattering) as sunshine approaches sea level. Similarly, direct-beam sunlight is greatly reduced shortly before sunset and after sunrise because the beam effectively must pass through several atmospheres.

Aerosols in the atmosphere can decrease the amount of solar radiation received at the ground. Days with haze, blowing dust, or high humidity have slightly reduced solar radiation when compared to days with clean, dry air. Clouds, composed of water droplets or ice crystals, are efficient sunshine reflectors, thereby reducing incident solar radiation. However, during brief periods, ground-level solar radiation on partly cloudy days can actually exceed the maximum possible radiation occurring on clear days because sunlight is reflected by surrounding clouds.

Monthly and daily average insolation as measured at _____ [station name(s)] are presented in Table 3.1.4.1.1.2.7-1 [INN 3.1.4.1.1.2.7-1] (provide when information available from monitoring data).

3.1.4.1.1.2.8 Severe Weather and Obstructions to Visibility

Occurrences of severe weather are superimposed on the average or normal climatic conditions in the Yucca Mountain area. These events generally include thunderstorms and associated

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lightning and flash flooding, hailstorms, tornadoes, straight-line extreme winds, sandstorms, temperature extremes, freezing rain, and fog. The most important of these phenomena, with respect to development of Yucca Mountain as a repository, are the thunderstorm-derived conditions such as high winds, hail, and flash flooding.

Tornadoes, a possible source of high winds, are considered rare in Nevada but have been observed within a radius of 250 km of Yucca Mountain (Eglinton and Dreicer, 1984). The most severe of these tornadoes was classified as F-0 on the Fujita tornado intensity scale. This scale was developed to classify tornado intensity and maximum wind speed based upon the extent of resultant damage. An F-0 tornado on this scale is classified as a very weak tornado; it has winds of between 18 and 32 m/s, a path length of less than 1.6 km, and a path width of less than 16 m (Ludlum, 1982) [INN 3.1.4.1.1.2.8-1] (document from video record any evidence of tornadic weather at Yucca Mountain).

Dust devils, which are small whirlwinds containing sand or dust, occur in and around the Yucca Mountain Site during the summer months. Dust devils occasionally develop wind velocities in excess of that associated with an F-0 tornado, but they dissipate rapidly (Eglinton and Dreicer, 1984) [INN 3.1.4.1.1.2.8-2] (document from video record any evidence of dust devil activity at Yucca Mountain).

Lightning is frequently associated with thunderstorm activity. Lightning data have been collected from the automated lightning detection system. This system detects cloud-to-ground lightning strikes in the YMP area in real time. A description of the monitoring network for lightning is provided in Subsection 3.1.4.2. Because cloud-to-cloud lightning occurs nearly 10 times as frequently as cloud-to-ground lightning, strikes of consequence (i.e., resulting in measurable damage) in Nevada only average 18 per year (Eglinton and Dreicer, 1984). However, the sparse observational network may reflect a somewhat lower frequency of occurrence than actually might be experienced at the Yucca Mountain Site. Detection of cloud-to-cloud lightning has been made possible by the use of recently developed optical technology. A field test of a prototype optical lightning detector was conducted by the U.S. Weather Service Nuclear Support Office (now the Atmospheric Research Laboratory for the NTS). The test showed that cloud-to-cloud lightning activity precedes cloud-to-ground strikes by 15 minutes (Scott 1989).

Lightning-strikes have been correlated with sudden precipitation downpours (Moore et al., 1962 and 1964; Szymanski et al., 1980; Piepgrass and Krider, 1982; Goodman and Buechler, 1990). Lightning-strike data collected within the YMP area indicates whether regional topography enhances thunderstorm development, movement, and speed, or modifies precipitation amounts. Figure 3.1.4.1.1.2.8-1 [INN 3.1.4.1.1.2.8-3] (prepare when data becomes available) shows the pattern of detected lightning from the storm of _____ (date) in the Yucca Mountain area as an example indicating the concentration of lightning strikes in relation to regional topography. In addition, lightning data indicates [INN 3.1.4.1.1.2.8-4] (need to confirm with data from USGS study) where the heaviest surface-water runoff will occur. Figure 3.1.4.1.1.2.8-2 [INN 3.1.4.1.1.2.8-5] (prepare when data available) illustrates the spatial relationships between lightning-strike data and rainfall-runoff data.

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Hail is a variety of thunderstorm activity that can have quite damaging effects. Based on climatological records, one occurrence is expected annually at Yucca Mountain (Eglinton and Dreicer, 1984). Table 3.1.4.1.1.2.8-1 [INN 3.1.4.1.1.2.8-6] (prepare when (if) data is available from monitoring data) provides the frequency of occurrence of hail at Yucca Mountain during the monitoring program of the site since ____ (yr).

Obstructions to visibility in the vicinity of Yucca Mountain could temporarily disrupt activities at the proposed repository. The most likely conditions that could obstruct visibility appreciably are sandstorms or fog. The conditions conducive to fog formation (strong radiative cooling with surface moisture) occur only about twice a year in this area of Nevada and sandstorms of sufficient magnitude to reduce visibility occur only a small fraction of the time (Eglinton and Dreicer, 1984). Table 3.1.4.1.1.2.8-2 [INN 3.1.4.1.1.2.8-7] (prepare when (if) data is available from monitoring data) provides the frequency of occurrence of fog and sandstorms at Yucca Mountain during the monitoring program of the site since ____ (yr). [More detailed discussions of obstructions to visibility as they relate to safe operation of the repository are included in the environmental impact statement.]

3.1.4.1.1.2.9 Extremes

Predictions of extreme weather needed in design considerations are generally derived by extrapolating past recorded data. The associated probability of occurrence of an event is then calculated based on the frequency of actual occurrence of the event during the period of record. Of most importance in designing the surface facilities of the proposed repository at Yucca Mountain are estimates of extreme winds, temperature maximums and minimums, and extreme precipitation events.

Extreme wind speeds and associated probabilities of occurrence have been calculated for the NTS and are presented in Table 3.1.4.1.1.2.9-1 (Quiring, 1968) [INN 3.1.4.1.1.2.9-1]. Table 3.1.4.1.1.2.9-2 [INN 3.1.4.1.1.2.9-1 (prepare when information is available from monitoring data)] shows similar data based on the Yucca Mountain Site meteorological monitoring. These data are for a fastest mile of wind, which represents an average highest wind velocity as 1 mile of air passes the measurement point. Eglinton and Dreicer (1984) discuss the potential for both straight-line winds and tornadic (cyclic) winds expected to occur at Yucca Mountain. The probability of a tornado strike at Yucca Mountain, given in Eglinton and Dreicer (1984), is approximately 7.5×10^{-4} in any given year. The maximum design wind speeds cited in Eglinton and Dreicer (1984) for the NTS are a straight-line wind speed of 94 m/s (210 mph) and a tornadic (cyclic) wind speed of 28 m/s (63 mph). Both phenomena are given as having a probability of 1×10^{-6} of occurring in 1 year. These extrapolations, however, do not take into account the possibility of climatic change in the future.

Another measurement of extreme wind is the peak gust. This parameter has been recorded at a monitoring site near the Yucca Mountain Site (Yucca Flat) from 1962 to 1971 (Bowen and Egami 1983). Table 3.1.4.1.1.2.9-3 shows the historical extreme peak gust for each month and annual for the period during which measurements were gathered at Yucca Flat. The peak gust represents a maximum 1-minute wind speed.

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The probability of occurrence of extreme temperatures at Yucca Flat, also given in Eglinton and Dreicer (1984), is shown in Table 3.1.4.1.1.2.9-4 [INN 3.1.4.1.1.2.9-2]. Table 3.1.4.1.1.2.9-4 shows the measured temperature extremes from the 14 monitoring stations at the Yucca Mountain Site (Refer to Subsection 3.1.4.2, Paleoclimatology for discussion on monitoring sites). These data are estimated on the basis of measured extreme temperatures but do not account for the influence of climatic change.

Another design consideration is extreme precipitation and the potential for flooding that could occur as a result. Because the flooding potential from short-duration, high-intensity storms is high, 24-hour average precipitation amounts are not a realistic indicator of potentially damaging extreme precipitation and subsequent flood events at Yucca Mountain. Both 1- and 24-hour maximum precipitation and associated probabilities of occurrence, again based on measured extremes at the 134 monitoring sites for precipitation, are given in Table 3.1.4.1.1.2.9-5 [INN 3.1.4.1.1.2.9-3 (prepare when information available from monitoring data)]. The flooding potential is discussed in more detail in Subsection 3.1.2.1

3.1.4.1.2 Site Meteorological Monitoring Network

The key meteorological data that indicate local climate are:

- Horizontal wind speed
- Horizontal wind direction
- Air temperature
- Temperature difference (vertical)
- Dew-point temperature
- Relative humidity
- Barometric pressure
- Precipitation
- Solar radiation
- Vertical wind speed

Table 3.1.4.1.2-1 [INN 3.1.4.1.2-1] provides a listing of the meteorological monitoring station locations and elevations for the fourteen sites. Figure 3.1.4.1.1-2 shows these locations on a map of the YMP area. Table 3.1.4.1.2-2 provides a listing of the parameters monitored at each location. Each of these 14 meteorological monitoring stations is equipped with a tipping bucket rain gage (three of the sites have two gauges). An additional 120 precipitation gauges are operated/monitored around the YMP area; four gauges are tipping-bucket type, and the remainder are storage type.

A description and site selection rationale of the nine sites operated by Science Applications International Corp. (SAIC) are provided below. The NTS-60 site has a 60 m tower that includes wind sensors at the 10 m and 60 m levels. All other sites have 10 m towers with wind sensors at the top.

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- Site 1: NTS-60 (Main Site) is generally representative of the source region of possible releases from the proposed repository surface facilities. Site exposure is a reasonably open area in Midway Valley approximately 4 km north-south by 3 km east-west. The station is east of the central portion of the primary ridge of Yucca Mountain. The location is about 1.2 km east of rapidly rising terrain of the Yucca Mountain ridge, and about 2 km west of Fran Ridge.

- Site 2: Yucca Mountain is located on the Yucca Mountain ridge crest, about 3.9 km west-northwest of the NTS-60 station. This location is out of immediate dominating airflow influences of Midway Valley, but is still influenced by larger topographic features, particularly toward the north and east.

- Site 3: Coyote Wash is located in a narrow valley along the east side of the Yucca Mountain ridge, about 1.6 km east of the Yucca Mountain station. Coyote Wash is important because it is representative of many small valleys along the east side of the Yucca Mountain ridge.

- Site 4: Alice Hill is located on top of a sloping hill about 3.0 km northeast of the NTS-60 station; the hill is near the fork between Fortymile Canyon and Yucca Wash, and is just north of Fran Ridge. The elevated location of Alice Hill is advantageous for observing meteorological conditions out of the immediate influence of the floor of Midway Valley.

- Site 5: Fortymile Wash is located along the floor of Fortymile Canyon southeast of Busted Butte, about 9.2 km southeast of the NTS-60 station. The site is just east of the drainage cut in Fortymile Wash. This location is a good indicator of airflow along an airborne transport pathway between the proposed repository surface facility and the Amargosa Valley populated area.

- Site 6: WT-6 is at the WT-6 well pad, in the upper end of Yucca Wash, near the boundary line between the NTS and the Nellis Air Force Range Land, 6.1 km north-northwest of the NTS-60 site. Data from this location characterize the daytime up-valley airflow from the south and southeast through Midway Valley, and also provides up-canyon measurement during down-valley airflow relevant to Midway Valley.

- Site 7: Sever Wash is in the topographic gap between Alice Hill and Fran Ridge, 2.1 km east-northeast of the NTS-60 site. This location provides data to more completely characterize the surface nocturnal drainage flow out of Midway Valley through this gap, which is downhill from the ESF north portal.

- Site 8: Knothead Gap is in a saddle east of the gap at the northern end of Bow Ridge, in the southern end of Midway Valley, and is 1.7 km south-southwest of the NTS-60 site. This location is useful in determining the preferred nocturnal drainage flow for the area east of the proposed ESF south portal. Pathways include directly to the

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southeast through the saddle between Bow Ridge and Fran Ridge, or circling northeastward toward the Sever Wash area.

- Site 9: Gate 510 is along the southern border of the NTS along the Lathrop Wells Road, at Gate 510, south of the Fortymile Wash site. It is 18.4 km south-southeast of the NTS-60 site. Data from the Gate 510 site further characterizes the airflow along the trajectory between the proposed repository area and the populated portions of Amargosa Valley.

The extensive precipitation-data-collection network operated by the USGS involves several different activities. First, each meteorological station has a precipitation gauge (three sites have two gauges, one for rain and one for snow [heated]). Second, other automated recording precipitation stations are operating. These stations do not have the full spectrum of instrumentation as do the meteorological stations, but each precipitation station has a tipping-bucket precipitation gauge monitored by a Campbell Scientific CR10 datalogger. Some of the tipping-bucket gauges will be propane-heated to collect frozen precipitation. The CR10 datalogger is vertically mounted on a steel pole in a weather-proof enclosure. Power is provided by a 12-volt, sealed, lead-acid storage battery which is charged by a solar panel. The gauges are mounted on the ground in a position such that no tall obstacles interfere with the precipitation catch (WMO, 1983). All gauges are calibrated in millimeters. The storage of precipitation data on automated platforms is event-driven. The datalogger checks the gauge every second. When a tip is detected, with a date and time stamp, that data is recorded on magnetic tape. The tapes are collected from stations approximately every two months. During these visits, any error in the datalogger clock is recorded and the clock is reset using a National Institute of Standards and Technology standard as is done with the meteorological stations. Also, the station is inspected for such things as cable integrity and propane supply (where installed).

In addition to the automated precipitation-monitoring network, manual monitoring of precipitation is made using a relatively dense network of storage gauges. The storage gauges are used to measure total-storm-event accumulations and monthly accumulations at 116 sites. Individual storm events are defined as periods of precipitation separated by periods of no precipitation longer than 6 hours (WMO, 1983). Storm-event accumulations are recorded manually by a visual reading of the fluid level collected in each gauge. The resolution and accuracy of these measurements have been determined to be ± 0.25 mm. The timing of most events is recorded by correlating the times and dates obtained from the automated precipitation gauges, as well as direct observation of storm events over the project area by visual observation and by automated time lapse photography of Yucca Mountain. This provides an accuracy on the timing of storms for the storage-gauge network at least to the nearest day.

Precautionary measures are taken to ensure against errors. Collection gauges, when properly mounted so that the orifice is located well above the top and sides of the mounting post, are subjected to minimal splashing from the outside. Installation of wire mesh, deep enough inside the orifice to eliminate back splashing, is effective in preventing most insect

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entrapment. Oil is being successfully used to prevent evaporation. With these measures, it is believed that the network of plastic storage gauges provides an accurate assessment of the actual rainfall.

In addition to the five USGS site-specific meteorological monitoring stations and 120 additional precipitation gauges, the USGS collects the following data to help characterize the meteorology of the Yucca Mountain Site:

- Lightning data (Refer to description in Subsection 3.1.4.1.1.2.8) are collected from an existing automated lightning detection system operated by the Atmospheric Research Laboratory for the NTS. The network consists of seven direction finders. Five are located on the NTS with additional direction finders located at Blue Diamond and Warm Springs, Nevada. The YMP area is located on the western edge of the present automated lightning detection system network. The accuracy of the automated lightning detection system is reported to be approximately 1 km with an 85- to 90-percent detection rate within the boundaries of the network.

- Weather satellite data is collected from both National Oceanic and Atmospheric Administration Polar orbiting satellites and Geostationary satellites. These data are especially useful in tracking synoptic-scale storm systems and determining areal coverage of snow.

- Synoptic-scale weather maps are collected via computer link with the NWS. These data are used to keep track of major frontal systems and storm tracks.

- Time-lapse photography is used to survey weather development on Yucca Mountain and vicinity. Using a video camera, cloud development and movement are documented. The camera is operated continuously to visually document weather conditions related to meteorological conditions recorded by the automated stations. During the thunderstorm season (July-August), the photography helps to determine the influence of Yucca Mountain itself on storm development.

3.1.4.1.3 Site Meteorology

The meteorological conditions that describe the site include temperature, wind speed and direction, precipitation, atmospheric moisture and pressure, sunshine, and severe weather phenomena. These parameters are governed by local influences of terrain and vegetation. Summaries of the site data have been included in the description of the local and regional climate presented in earlier sections. To avoid unnecessary duplication of information, the reader should refer back to Subsections 3.1.4.1.1.2.1 through 3.1.4.1.1.2.9. Winds will be governed to a significant extent by the local terrain with regard to direction and speed. The highest winds will be associated with winter frontal passages and thunderstorms. One meteorological parameter not previously discussed, because it is not a characteristic of climate, is atmospheric stability, a measure of the dispersion capability of the atmosphere. A discussion of this parameter follows.

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Atmospheric stability is an important parameter with respect to dispersion of emissions (e.g., particulates and exhaust gases) from the proposed repository and has been analyzed using two different methods. Atmospheric stability measurements are calculations usually based on fluctuations of either horizontal wind direction or vertical wind speed. The turbulence measurements are:

- Standard deviation of the horizontal wind direction, also known as sigma-theta (σ_θ) or sigma-A (σ_A);
- Standard deviation of the vertical wind speed, also known as sigma-w (σ_w).

Since atmospheric stability can be calculated from various turbulence indicators, depending on dispersion model input requirements, the most common and accepted calculation methods are described below:

- The <u>lateral turbulence (σ_A) and wind speed</u> method (EPA, 1986 and 1987a) is used to estimate Pasquill stability categories (A-F) using the standard deviation of the horizontal wind directions, σ_A, scalar mean wind speed, and a day/night distinction.
- The <u>lateral turbulence method</u> used by the NRC (NRC, 1981) is almost the same as the σ_A method described above, except that a seventh category ("G") is added for yet another more stable distinction. Also, no adjustments are made for wind speed or day/night.
- The <u>vertical temperature difference method</u> (NRC, 1981) reflects long-standing NRC guidance that assigns one of seven Pasquill categories based on the temperature lapse rate, typically from 10 to 60 m-agl.

A brief description of stability is as follows:

- Stability class A defines an extremely unstable (highly convective) atmosphere.
- Class B is for unstable conditions.
- Class C is slightly unstable.
- Class D is neutral.
- Class E is slightly stable.
- Class F is stable
- The NRC class G is defined as very stable.

Annual average stability distributions can be constructed for only those occurrences of wind speed and direction by Pasquill stability class ("A" through "F") and presented in a similar manner to the data presented in Figure 3.1.4.1.1.2.4-1 through 3.1.4.1.1.2.4(n).

Wind roses for each of these stability classes are shown in Figures 3.1.4.1.3-1 through 3.1.4.1.3-7, [INN 3.1.4.1.3-1 (prepare wind roses of Stab classes when information is available from monitoring data)] for station ___ [repeat for all nine SAIC stations]. The most

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