Mined Geologic Disposal System (MGDS) Annotated Outline Skeleton Text for the Preparation of a License Application

November 30, 1993

Volume I of II

Revision 3

U.S. Department of Energy
Office of Civilian Radioactive Waste Management
Washington, DC

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Figure 3.1.2.2.3-52 Water-Level Recovery, Pumping Test 2: Depth Interval from 526 to 1,220 Meters.

Figure 3.1.2.2.3-53 Analysis of Water-Level Drawdown, Pumping Test 3: Depth Interval from 753 to 834 Meters; Straight-Line Solution with Dual-Porosity Model.

Figure 3.1.2.2.3-54 Residual Drawdown, Recovery Test 3: Depth Interval 753 to 834 Meters.

Figure 3.1.2.2.3-55 Water-Level Drawdown, Pumping Test 4: Depth Interval from 608 to 645 Meters.

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Figure 3.1.3.2.1.1-1 Index Map Showing the Location of Drill Holes in Relation to the CPDB

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Figure 3.1.3.2.1.1.1-1 Total Alkali Silica Diagram (Le Bas et al., 1986) for Representative Fresh and Altered Samples of Latite and Rhyolite Tuff of the Topopah Spring Member of the Paintbrush tuff.

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Figure 3.1.3.2.1.1.2-1 Illustration of the cumulative percent of phenocrysts and lithic fragments with depth in the Topopah Spring Member for a) drill hole USW G-4 and b) UE-25 a#1.

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Figure 3.1.4.1.1-1 Locations of the Climatological Data Stations in the Yucca Mountain Region [INN 3.1.4.1.1-1]

Figure 3.1.4.1.1-2 Locations of Site Specific Monitoring Sites at Yucca Mountain. [INN 3.1.4.1.1-1]

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Figure 3.1.4.1.1.1.2-2 Winter Weather Type A. Successive Time-Lapse Positions of the Low-Pressure Cyclone are Shown as the System Develops and Matures. [INN 3.1.4.1.1.1.2-2]

Figure 3.1.4.1.1.1.2-3 Winter Weather Type B. Shown are the Successive Time-Lapse Positions of the Low-Pressure System as it Develops. The System Remains Well to the North of the U.S., as High Pressure Dominates the Southwest. [INN 3.1.4.1.1.1.2-2]

Figure 3.1.4.1.1.1.2-4 Winter Weather Type C. Shown is the Belt of High Pressure which is Displaced Northwest from its Normal Position. The Low-Pressure Centers Develop in the Gulf of Alaska and off the Coast of San Diego. Then They Move Inland as Depicted in this Time-Lapse Sequence. [INN 3.1.4.1.1.1.2-2]

Figure 3.1.4.1.1.1.2-5 Winter Weather Type D. Depicted are Time-Lapse Positions of the Low-Pressure Center as it Tracks Across the Northern U.S. [INN 3.1.4.1.1.1.2-2]

Figure 3.1.4.1.1.1.2-6 Winter Weather Type E. The Strong Canadian High-Pressure Ridge is Depicted. This Ridge forces Developing Low-Pressure Systems Southward into the Great Basin. These Pacific Storms cross the Sierra-Nevada, Weaken, then Redevelop on Lee side of the Mountain Range. (From Elliott, 1943) [INN 3.1.4.1.1.1.2-2]

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Figure 3.1.4.1.1.1.2-7 Summer Southwest Monsoon. Shown is a lobe of the Bermuda High over the four-corners region. The Resulting Pressure Gradient causes a Gentle Flow of Moisture to begin from the Tropical Eastern Pacific Ocean. The Gulf of Mexico Contributes only an Insignificant Amount of Moisture to the Southwestern U.S. [INN 3.1.4.1.1.1.2-3]

Figure 3.1.4.1.1.1.3-1 Dominant Summer and Winter Moisture Sources for the Southern Nevada Area [INN 3.1.4.1.1.1.3-1]

Figure 3.1.4.1.1.1.3-2 Regression Curves Relating Annual Average Precipitation (mm) with Precipitation Gage Elevation (ft). [INN 3.1.4.1.1.1.3-3]

Figure 3.1.4.1.1.2.2-1 Precipitation Amounts [INN 3.1.4.1.1.2.2-2]

Figure 3.1.4.1.1.2.4-1 to n Wind Rose Plot (1 - n) [INN 3.1.4.1.1.2.4-1]

Figure 3.1.4.1.1.2.5.1-1 Seasonal and Annual Wind Distribution at 5,000 ft (1.524 m) Above Mean Sea Level (328 m Above Ground Level) for Yucca Flat (1957 to 1964). Note: Scale is not the same for all distributions. [INN 3.1.4.1.1.2.5.1-1]

Figure 3.1.4.1.1.2.5.1-2 Seasonal and Annual Wind Distributions at 6,000 ft (1.829 m) Above Mean Sea Level (633 m Above Ground Level) for Yucca Flat (1957 to 1964). Note: Scale is not the same for all distributions. [INN 3.1.4.1.1.2.5.1-2]

Figure 3.1.4.1.1.2.5.1-3 Upper Air Data [INN 3.1.4.1.1.2.5.1-3]

Figure 3.1.4.1.1.2.6-1 Monthly Mean Atmospheric Pressure [INN 3.1.4.1.1.2.6-1]

Figure 3.1.4.1.1.2.6-2 Hourly Mean Atmospheric Pressure [INN 3.1.4.1.1.2.6-2]

Figure 3..4.1.1.2.8-1 Pattern of Detected Lightening [INN 3.1.4.1.1.2.8-3]

Figure 3.1.4.1.1.2.8-2 Spatial Relationships Between Lightening-Strike Data and Rainfall-Runoff Data [INN 3.1.4.1.1.2.8-5]

Figure 3.1.4.1.3-1 Wind Rose Plot [INN 3.1.4.1.3-1]

Figure 3.1.4.1.3-2 Wind Rose Plot [INN 3.1.4.1.3-1]

Figure 3.1.4.1.3-3 Wind Rose Plot [INN 3.1.4.1.3-1]

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Figure 3.1.4.1.3-4 Wind Rose Plot [INN 3.1.4.1.3-1]

Figure 3.1.4.1.3-5 Wind Rose Plot [INN 3.1.4.1.3-1]

Figure 3.1.4.1.3-6 Wind Rose Plot [INN 3.1.4.1.3-1]

Figure 3.1.4.1.3-7 Wind Rose Plot [INN 3.1.4.1.3-1]

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3.1.1.1.3.1.2-1	Sketch map of the Walker Lane Belt and associated selected faults and earthquakes
3.1.1.1.4.1-1	Seismic Hazard Methodology
3.1.1.1.4.1-2	A regional map showing the major seismotectonic elements in the western great basin with an overlay of regional seismicity
3.1.1.1.4.2.1-1	Southern Great Basin Seismic Network Catalog. Figure showing the distribution of seismicity on a regional scale (322 km)
3.1.1.2.2.2-1	Additional information on the stratigraphy and distribution of Pre-Cenozoic aged rocks beneath Yucca Mountain.
3.1.1.2.2.3-1	Additional data on the volcanic stratigraphy at Yucca Mountain.
3.1.1.2.2.3-2	Additional data on the Calico Hills stratigraphic section.
3.1.1.2.4.4-1	Table of fault parameters for faults considered to be potential seismic sources for the Yucca Mountain Site
3.1.1.2.2.5-1	Additional data on the Quaternary stratigraphy at Yucca Mountain.
3.1.1.2.3.2-1	Maps of Quaternary faults (site and regional scale). Figure -1: A preliminary map of known and suspected quaternary faults within 100 km of Yucca Mountain Figure -2: A preliminary map of known and suspected quaternary faults in the vicinity of Yucca Mountain
3.1.1.2.4.2.1-1	Detailed Map of the Ghost Dance fault
3.1.1.2.4.1-1	Data from the UNR Northern Nevada and Southern Great Basin networks; Map of quaternary faults (unpublished).
3.1.1.2.4.2.2-1	Location of seismic recording stations of the Southern Great Basin Seismic Network.
3.1.1.2.4.2.2-2	Stations recording the Little Skull Mountain earthquake of 29 June, 1992 in southern Nevada.

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3.1.1.2.4.2.2-3	Attenuation of peak acceleration and peak velocity of the Little Skull Mountain, Nevada earthquake of 29 June, 1992 in southern Nevada compared with the average of the estimates derived from Joyner and Boore(1988) and Campbell (1990).
3.1.1.2.4.2.2-4	Locations of the Little Skull Mountain earthquake, the Rock Valley earthquake, and the recording station at Midway Valley.
3.1.1.2.4.2.2-5	Velocity seismograms of the Eureka Valley aftershock of June 3, 1993 (magnitude 3.9, depth 5.9 km).
3.1.1.2.4.2.2-6	Velocity seismograms of the Eureka Valley aftershock of June 8, 1993 (magnitude 4.0, depth 1.7 km).
3.1.1.2.4.2.2-7	Location map showing Nevada Test Site, Yucca Mountain seismic stations(open squares), explosion locations (solid symbols), and velocity profiles PM1, PM2 and YF1. Also denoted are the Timber Mountain Caldera and the Silent Canyon Caldera (SCC). Areas 19 and 20 comprise Pahute Mesa.
3.1.1.2.4.2.2-8	Ratios of measured PSRV at Station W-14 to average values for NTS sites.
3.1.1.2.4.2.2-9	Ratios of measured PSRV at Station W-23 to average values for NTS sites.
3.1.1.2.4.2.2-10	Ratios of measured PSRV at Station W-22 to average values for NTS sites.
3.1.1.2.4.2.2-11	Ratios of measured PSRV at Station W-21 to average values for NTS sites.
3.1.1.2.4.2.2-12	Location map of strong motion stations that recorded the Pipkin nuclear explosion at Pahute Mesa.
3.1.1.2.4.2.2-13	Profile of radial velocity time histories of the Pipkin nuclear explosion at Pahute Mesa recorded to the south at stations shown in Figure 3.1.1.2.4.2.2-15.
3.1.1.2.4.3.4-1	Seismograms recorded in a downhole array at Jackass Flats (Station 10, Well J-11) from a nuclear explosion at Pahute Mesa . Top row: acceleration; middle row: velocity; bottom row: displacement. Left column: surface station on alluvium; middle column: -61 meters near the base of alluvium; right column: -356 meters in tuff.

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3.1.1.2.4.3.4-2 Location map of downhole ground motion recording stations at Yucca Mountain and regional topography.

3.1.1.2.7.1.1-1 Information about the spatial location of the various units of the thermomechanical stratigraphy in and in the vicinity of the repository block.

Elevations of the contacts between the various units of the thermomechanical stratigraphy. The spacing between the holes should preferably be in the range of one-half to three-quarters of a mile. Additional drilling may be necessary depending on the variability of the data encountered.

3.1.1.2.7.3.1-1 Data to verify whether measurements of elastic properties of the thermomechanical units, especially the emplacement unit, made under conditions of zero confining pressure, room temperature(23° C) and full saturation (baseline conditions) are applicable to the values of these properties for in situ conditions.

Results of tests in which the elastic properties are measured on comparable samples of the thermomechanical units under the baseline conditions as well as under the expected state of these conditions in situ. If the results show that measurements made under baseline conditions are not applicable to values of these properties under expected conditions, additional testing under the expected conditions may have to be performed."

3.1.1.2.7.3.1-2 Sufficient test results on the measurement of Young's Modulus and Poisson's Ratio of the various thermomechanical units to adequately characterize these properties of the repository block.

3.1.1.2.7.3.1-3 Sufficient test results to determine the anistropy of elastic properties of the various thermomechanical units including values of Young's Modulus and Poisson's Ratio measured perpendicular and parallel to the rock fabric.

3.1.1.2.7.3.2-1 Unconfined compressive strength of the various units of the thermomechanical stratigraphy.

Sufficient test results on the unconfined compressive strength of the TSw2 unit of the thermomechanical stratigraphy so as to adequately characterize the spatial variability, if any, of the strength of this unit within the zone of interest in the repository block."

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3.1.1.2.7.3.2-2 Test results on the unconfined compressive tests of various units of the thermomechanical stratigraphy under as close to the in situ moisture content as possible. Values of unconfined compressive strength.

3.1.1.2.7.3.2-3 Effect of elevated temperatures on the strength of various units of the thermomechanical stratigraphy. Values of unconfined compressive strengths at elevated temperatures.

3.1.1.2.7.3.2-4 Test results to evaluate the strength anistropy of TSw2. Values of compressive strength of TSw2 measured both perpendicular and parallel to the rock fabric.

3.1.1.2.7.3.2-5 Geological maps of the ESF exposures showing the locations, if any, of zones of high lithophysae content in the TSw2. Information is needed to verify whether there are zones of high lithophysae content within the proposed emplacement zone of TSw2.

3.1.1.2.7.3.2-6 Test results that show the effect of lithophysae content on the mechanical properties of the Topopah Spring tuff.

3.1.1.2.7.3.3-1 Information on the tensile strength of the various units of the thermomechanical stratigraphy.

3.1.1.2.7.3.4-1 Triaxial tests on various units of the thermomechanical stratigraphy, especially TSw2 and CHn. Specifically, values of cohesion and angle of internal friction at various confining pressures and temperatures.

3.1.1.2.7.3.5-1 Long-term time-dependent deformational behavior of TSw2 at elevated temperatures expected in emplacement areas during the preclosure period. Specifically, values for steady state creep strain rate at various temperatures.

3.1.1.2.7.4.1-1 Test results to evaluate the effect of normal stress level on the coefficient of friction of natural joints. Specifically, measured values of coefficient of friction at various levels of normal stress.

3.1.1.2.7.4.2-1 Information on the effect of infilling materials on the mechanical properties of joints. Specifically, results of tests in which natural joints with various types of infilling material are tested for their mechanical properties.

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3.1.1.2.7.4.3-1 a) Test results to show the differences, if any, of the mechanical properties of healed and unhealed natural joints

b) Test results to show the effects, if any, of various test conditions such saturation level, temperature, etc., on the mechanical properties of natural joints.

The measured values of joint properties under various test conditions from these tests are needed."

3.1.1.2.7.4.5-1 Test results on the effect of short-term time-dependent factors on the mechanical properties of joints. Specifically, values of mechanical properties parameters under various displacement rates etc.

3.1.1.2.7.4.6-1 Test results showing the long-term time-dependent properties of joints at elevated temperatures expected during the preclosure period. Specifically, values of long-term displacement rates at constant normal and shear stresses.

3.1.2.1.1-1 Surface water monitoring data such as flow duration data that indicate maximum, minimum and average historical observations of surface water bodies, water quality data, snow pillow and precipitation data that is currently being recorded at all recording stations.

3.1.2.1.1-2 Provide information on the characteristics of playas including location, discharge and chemical quality of water.

3.1.2.1.1-3 Figure showing the location of all surface water bodies in the Yucca Mountain region including springs and major playas.

3.1.2.1.1-4 Update of phreatophyte data in Table 3.1.2.1.1-3

3.1.2.1.4-1 More geologic evidence of Quaternary flooding in the region of Yucca Mountain with emphasis on the Holocene may be developed during site characterization studies.

3.1.2.1.5-1 Additional flooding study results.

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3.1.2.1.5-2	Design basis water levels and elevations of all surface facilities and underground openings may be available, but could not be collected for this revision. These design basis water levels and elevations and locations of all surface facilities and underground openings may change during advanced conceptual design.
3.1.2.1.5-3	A review and reference of two T&MSS floodplain assessments, "Floodplain Assessment of Surface-Base Investigations at the Yucca Mountain Site, Nye County, Nevada" dated 8/91 and "Floodplain Assessment of Site Characterization Activities at the Yucca Mountain Site, Nye County, Nevada" dated 8/92.
3.1.2.2.2-1	Summary of selected wells monitored for water levels at Yucca Mountain
3.1.2.2.3-1	Hydraulic properties for the Lower Carbonate Aquifer
3.1.2.2.3-2	Pumping response data for the welded tuff aquifer
3.1.2.2.3-3	Results of interference tests at the C-well complex in 1983-1984
3.1.2.2.3-4	Analyses using the finite-conductivity, vertical fracture model
3.1.2.2.8-1	Update of water appropriations data used in tables.
3.1.3.2.1-1	Analyses of Samples from additional bore holes and the ESF, Mineralogy, Petrology, and Chemistry of the Topopah Spring member
3.1.3.2.1.1-1	Hydrothermal alteration of the Topopah Spring member
3.1.4.1.1-1	Locations of the regional and site-specific meteorological monitoring stations
3.1.4.1.1-2	List of parameters measured at the regional and site- specific meteorological monitoring stations.
3.1.4.1.1.1.2-1	Average position of the polar front in January.
3.1.4.1.1.1.2-2	Time-lapse movements of winter storms systems.
3.1.4.1.1.1.2-3	Summer southwest monsoon showing a lobe of the Bermuda high over the four corners region.

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3.1.4.1.1.1.3-1	Dominant summer and winter moisture sources for the southern Nevada area.
3.1.4.1.1.1.3-2	Effect of varying air flow regime on the Yucca Mountain site.
3.1.4.1.1.1.3-3	Plot of average annual precipitation as a function of station elevation.
3.1.4.1.1.2.1-1	Summary data on temperature from site and regional monitoring sites.
3.1.4.1.1.2.2-1	Precipitation data.
3.1.4.1.1.2.2-2	Precipitation amounts for each month, based on average data from Table
3.1.4.1.1.2.3-1	Relative humidity.
3.1.4.1.1.2.3-2	Wet bulb depression data.
3.1.4.1.1.2.4-1	Surface wind data.
3.1.4.1.1.2.4-2	Wind roses with data split for daytime versus nighttime wind frequencies (data plotted on terrain maps)."
3.1.4.1.1.2.5.1-1	Wind speed and directions at 1524 m above mean sea level.
3.1.4.1.1.2.5.1-2	Wind speed and directions at 1829 m above mean sea level.
3.1.4.1.1.2.5.1-3	Data from upper air winds.
3.1.4.1.1.2.5.1-4	Comparison of Desert Rock data with historical Yucca Flat upper air data.
3.1.4.1.1.2.5.2-1	Average morning and afternoon mixing height.
3.1.4.1.1.2.6-1	Standard atmospheric pressure.
3.1.4.1.1.2.7-1	Monthly and average insolation.
3.1.4.1.1.2.8-1	Evidence of tornadic weather.
3.1.4.1.1.2.8-2	Evidence of dust devil activity.
3.1.4.1.1.2.8-3	Pattern of detected lightning.

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3.1.4.1.1.2.8-4	Lightning data.
3.1.4.1.1.2.8-5	Spatial relationships between lightning-strike data and rainfall-runoff data.
3.1.4.1.1.2.8-6	Frequency of occurrence of hail.
3.1.4.1.1.2.8-7	Frequency of occurrence of fog and sandstorms.
3.1.4.1.1.2.9-1	Extreme wind speeds and probability of occurrence for the NTS and for Yucca Mountain.
3.1.4.1.1.2.9-2	Probability of occurrence of extreme temperatures.
3.1.4.1.1.2.9-3	1 and 24 hour maximum precipitation and probability of occurrence.
3.1.4.1.2-1	Composite USGS/SAIC maps of the Yucca Mountain site showing monitoring sites.
3.1.4.1.3-1	Wind roses for each stability class.
3.1.4.1.3-2	Stability Class Distributions

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3.1 DESCRIPTION OF INDIVIDUAL SYSTEMS AND CHARACTERISTICS OF THE SITE

3.1.0 Introduction

The purpose of this section is to describe the natural systems of the geologic setting in the region of the site, specifically for the "controlled area." [However, if conditions outside the controlled area affect waste isolation within the controlled area, the relevant and material information on conditions outside the controlled area will be included.] The controlled area is defined in 10 CFR Part 60 as "a surface location, to be marked by suitable monuments, extending horizontally no more than 10 kilometers 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.

3.1.1 Geologic System

The purpose of this section is to demonstrate whether the areas studied provide a representative description of the regional and site conditions.

Sources of information and data collection methods used to obtain measurements and observations are described and documented. Geological and geophysical descriptions are accompanied by maps, cross sections, photographs, and fence diagrams. The variability and uncertainty of data and information are discussed.

The representativeness of the data, the effects of varying geologic conditions, the reliability of geological and geophysical interpretations made, and uncertainties associated with the extrapolation of data and information to repository conditions are addressed in the discussions and descriptions. Conceptual models are discussed with respect to uncertainties in the data bases, the applicability and appropriateness of the geologic assumptions, the sensitivity of the model results to the uncertainty of the geologic input data, and model validation.

3.1.1.1 Regional Geology

The geologic characteristics in the region of the site are described, accompanied with sufficient data and technical analyses to demonstrate a clear understanding of the regional geology.

3.1.1.1.1 Geomorphology and Topographic Features in the Region of the Site

[This section was not revised in Revision 3 of the Annotated Outline; the skeleton text for this section will be provided in future revisions of the Annotated Outline.]

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3.1.1.1.2 Stratigraphy and Lithology of the Region

This section summarizes the regional stratigraphic and lithologic setting of Yucca Mountain. The regional setting of Yucca Mountain consists of the southern Great Basin, which was defined by DOE (1988) as primarily southern Nevada plus an adjacent area in California. This region includes Esmeralda, southern Nye, Lincoln, and Clark Counties in Nevada, and Inyo and Mono Counties in California.

To the east and southeast, the border of the southern Great Basin is defined geologically by the Colorado Plateau. To the west and south the region is bounded geologically by the Sierra Nevada and the southern basin and Range, respectively. Most of the southern Great Basin topography consists of northerly-trending, elongate mountain ranges separating narrow structural valleys with internal drainages. This pattern is less pronounced within the Walker Lane, which is an elongate, northwest-trending zone located west of Yucca Mountain that is characterized by lower, more randomly oriented mountain ranges and valleys. The only external drainage within the southern Great Basin is the Colorado River, which lies to the southeast, along the border with the Colorado Plateau.

The stratigraphy of the southern Great Basin comprises highly variable lithologies, ranging in age from Precambrian to recent. Typical stratigraphic sequence and formation names are illustrated in Figure 3.1.1.1.2-1. These rocks fall into eight general groups, based on similarities in age, lithology, and history.

Older Precambrian Crystalline Rocks. These consist of granitic intrusives, and metamorphic rocks, that are dominantly exposed in the Colorado River area and in Esmeralda County and southern Nye County, west of Yucca Mountain.

Younger Precambrian and Younger Cambrian Sedimentary Rocks. These dominantly consist of siltstones, quartzites, and carbonates believed to have been deposited in a shallow marine environment during rifting of the continental margin. Exposures of these rocks are widespread throughout the southern Great Basin.

Cambrian through Late Devonian Sedimentary Rocks. These dominantly consist of siliciclastics and carbonates believed to have been deposited in a shallow marine environment under passive continental margin conditions. Exposures of these rocks also are widespread throughout the southern Great Basin.

Mississippian through Permian Sedimentary Rocks. These rocks generally consist of shallow marine deposits, and flysch deposits (mudstone, siltstone, sandstone and conglomerate) shed off the Antler Highland during the Antler Orogeny. Exposures of these rocks also are widespread throughout the southern Great Basin.

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Mesozoic Sedimentary and Igneous Rocks. Mesozoic-aged sedimentary rocks generally consist of sandstones and shales that were deposited under fluvial or desert environments. These rocks are exposed primarily in northeastern and southern Clark County, between the Virgin Mountains on the northeast and the Spring Mountains on the southwest. Mesozoic-aged igneous rocks generally consist of granitic intrusives, and are exposed primarily in Esmeralda County west of Yucca Mountain.

Tertiary Sedimentary Rocks. These generally consist of poorly to moderately-consolidated alluvial fan and basin fill and lake deposits that were formed following the onset of early Basin-and-Range extension. Exposures are dominantly found in the Colorado River Trough and the Walker Lane, but are known to exist throughout the southern Great Basin.

Tertiary and Quaternary Igneous Rocks. The volumetrically dominant igneous rocks in the southern Great Basin consist of middle-to late-Tertiary aged rhyolitic to andesitic pyroclastic deposits, which were erupted from numerous calderas that traverse the region north of approximately 37° north latitude.

Younger igneous rocks generally consist of andesitic or basaltic lava flows, which are exposed in scattered deposits throughout the region.

Tertiary and Quaternary Surficial Deposits. The youngest deposits in the southern Great Basin region are unconsolidated alluvial and colluvial sands and gravels which are exposed in scattered deposits throughout the region.

The regional stratigraphic setting is important to the Yucca Mountain Site Characterization Project (YMP) for numerous reasons. For example, an understanding of the regional stratigraphic setting advances the discernment of the regional ground water flow configuration. Further, the assessment of seismic and volcanic hazard depends to a large extent on the record of past seismic and volcanic activity preserved in the stratigraphic section.

3.1.1.1.2.1 Precambrian Crystalline Rocks

Older Precambrian-aged metamorphic and younger Precambrian-aged igneous rocks are exposed extensively in eastern Clark County and southeastern Lincoln County (Stewart, 1980). 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 described as 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, 1980; Maldonado, 1990). The exposures in the Bullfrog Hills are mylonitized, and are located in the lower plate of a detachment fault system (Maldonado, 1990).

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These 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 the Tertiary-aged 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. In general, they consist of a westward-thickening prism of shallow marine quartzite, siltstone, shale, and carbonate which is interpreted as a rifted continental margin miogeoclinal belt (Figure 3.1.1.1.2.2-1). According to Stewart (1980), that miogeoclinal prism is divided into two depositional provinces in Nevada; from east to west, these consist of: (1) a quartzite and siltstone province; and (2) a siltstone, carbonate, and quartzite province. Stewart (1980) placed the boundary between the two provinces at approximately the eastern border of southern Nye County, trending north-south.

As described by Stewart (1980), rocks of the quartzite and siltstone province 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. Conglomerate, which contains pebbles of quartz and quartzite, is sparse in the quartzite. Limestone and dolomite occur in layers less than 1 to 100 m thick. Individual units of quartzite, siltstone, or carbonate rocks persist for hundreds of miles along the miogeoclinal belt, but are less persistent and change facies across the trend of the miogeocline. These rocks have been interpreted as deposited in a predominantly shallow water environment.

Rocks of the siltstone, carbonate, and quartzite province are finer grained, richer in carbonate, and thicker than those of the quartzite and siltstone province. 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 province; they include trilobites, archeocyathids, pelecypods, echinoderms, pelmatozoan debris, and algae. In correlative rocks of the quartzite and siltstone province to the east, fossils generally are extremely rare (Stewart, 1980). According to Winograd and Thordarson (1975), all clastic sedimentary rocks of this age (Late Precambrian through Early Cambrian) are considered aquitards. (Figure 3.1.1.1.2-1)

Typical stratigraphic formation names include the Wood Canyon Formation, of Precambrian age, and the Prospect Mountain Quartzite and Zabriskie Quartzite, both of Cambrian age. Albers (1967) argued that isopachs of the early Paleozoic section can be used as a piercing point to measure right lateral displacement across the Walker Lane, because they show a measurable right lateral displacement across the Walker Lane zone.

3.1.1.1.2.3 Middle Cambrian through Devonian Sedimentary Rocks

Middle Cambrian through Devonian-aged rocks are exposed throughout the southern Great Basin, although the Silurian System is somewhat restricted to Lincoln County and Clark County. Rocks of this early Paleozoic age group consist predominantly of carbonate and shale, with interbedded

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quartzite and sandstone, and are interpreted as the product of ongoing deposition in a passive continental marginal miogeocline. Total thickness of this section ranges from up to 500 m in western Utah, to at least 6100 m in central Nevada (Stewart 1980). 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

A marked change in the depositional environment in southern Nevada was created by the Antler Orogeny, of late Devonian and early Mississippian time. The orogeny created a roughly north-northeast trending highland, adjacent to the "Roberts Mountains Thrust," which extended from west of the site through northern Nevada and shed thick volumes of sediment into a foreland basin located roughly in the eastern half of the Great Basin. These rocks generally consist of thick flysch deposits adjacent to the Antler Highland, followed by shallow-water shelf carbonates farther east (Figure 3.1.1.1.2.4-1). The same general pattern of sedimentation continued through Permian time, although the volume of clastic sediment shed off the Antler Highland declined as the Antler Orogeny waned. According to Winograd and Thordarson (1975), Mississippian and Pennsylvanian-Permian aged rocks in the southern Great Basin are known 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-aged sedimentary rocks are present only in Clark County, where they consist of continental and marine sandstone, siltstone, and limestone of the Aztec Sandstone, Chinle Formation, and Moenkopi Formation (Triassic and Jurassic-aged). 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-aged, coarse-grained granitic plutons are exposed in Esmeralda County, west of Yucca Mountain. These plutons range in size from a few acres to the Inyo Batholith, of approximately 1000 square km area (Albers and Stewart, 1972). This absence of Mesozoic-aged sedimentary rocks suggests that throughout much of the Mesozoic era, the southern Great Basin was an emergent area, in the hinterland of the Sevier Fold and Thrust Belt, which was probably active by Middle Jurassic time. Thus, Mesozoic rocks are believed to have never been deposited over much of the southern Great Basin. Instead the Mesozoic era was characterized in the 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). The geometry of these thrust faults is an important constraint on the regional ground-water flow configuration. Wernicke et al. (1988b) have used these thrusts as offset markers, and from that offset attempted to reconstruct the Tertiary tectonic history of the Southern Great Basin. They inferred that the southern Great Basin has been extended by a factor of as much as 300 to 400 percent during the past 15 million years.

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3.1.1.1.2.6 Tertiary Sedimentary Rocks

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

3.1.1.1.2.7 Tertiary and Quaternary Igneous Rocks

The largest volume of Tertiary igneous rocks in the southern Great Basin are pyroclastic deposits of rhyolitic to trachytic composition, which were predominantly deposited before approximately 6 ma from a number of calderas located north of roughly latitude 37° north. Yucca Mountain is located on the southern flank of a cluster of four calderas known as the Southwest Nevada Volcanic Field, which erupted pyroclastic deposits between approximately 7 and 16 ma (Figure 3.1.1.1.2.7-1). These deposits are a complex mixture of pyroclastic flow and fall deposits, epiclastic deposits, and subsidiary lavas, and are believed to be approximately 3,050 m thick at the Yucca Mountain site (Stewart, 1980; Byers et al., 1976a, 1989; Best et al., 1989; Robinson, 1985).

Following the cessation of large-scale pyroclastic eruption 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 possibly as recently as 16 ka, both north and south of latitude 37° north, scattered through the region as well as within the area of the Southwest Nevada Volcanic Field (Best et al., 1989; Wells et al., 1990; Byers et al., 1976a). Carr (1990) has suggested that a roughly north-northeast trending lineation of younger basaltic volcanic centers that passes through the south end of the Southwest Nevada Volcanic Field, at Yucca Mountain, may be controlled by a structural rift. He calls this feature the Kawich-Greenwater rift.

Mining in the Yucca Mountain region is discussed in detail in Section 3.1.1.1.5. According to Cornwall (1972), most of the metal mining districts in southern Nye County that produce gold and silver, do so from quartz veins in Tertiary-aged volcanic rocks. The remaining gold and silver producing districts in that area are in Paleozoic or Precambrian sedimentary or Metomorphic rocks.

3.1.1.1.2.8 Tertiary and Quaternary Surficial Deposits

Late Tertiary to Quaternary surficial deposits exist 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. Carr (1984) presented gravity observations suggesting that these deposits may record a progressive southeastward tilt of the regional land surface in recent time.

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3.1.1.1.3 Structural Geology and Tectonic Information

An understanding of the tectonics of the site region is important to the potential repository site at Yucca Mountain for several reasons. Knowledge of the active stress field of the region, in conjunction with information on the distribution, orientation, and movement history of faults will aid in the evaluation of future earthquakes and fault displacements that could affect the site. Volcanism is a tectonic process of significance to the Yucca Mountain site for two reasons: (1) The potential repository horizon is in Miocene silicic tuff; and (2) There are Pliocene and Quaternary small volume basaltic volcanoes in the vicinity of the site. The tectonic history of the site region also exerts major control on the development and configuration of the regional hydrologic flow system.

While the size of the region and the time period of interest differ somewhat for each of these aspects of regional tectonics, in general an understanding of the tectonics of the southern Great Basin since the Miocene is needed.

3.1.1.1.3.1 Structural Features of the Yucca Mountain Region and Current State of Stress

Yucca Mountain is located in the southern part of the Great Basin, which has long been recognized as a large area where current landforms were created by extensional tectonic processes. A number of extensional mechanisms and amounts of extension have been proposed. Relatively high angle planar normal faults that cut the brittle (seismogenic) crust are able to 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 exact amount and timing of extension in the Yucca Mountain region are not fully understood. What is known is that the present Basin and Range structure, typically tilted fault blocks 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 Timber Mountain Tuff was erupted and deposited around the high standing fault blocks. It has been suggested that rates of fault movement were highest between 13 and 11.5 Ma and have decreased since that time (Scott, 1990). What is most significant to a potential repository at Yucca Mountain, however, is that faults near Yucca Mountain and in the surrounding region are favorably oriented for movement in the modern stress field. The following section discusses the structural features in the Yucca Mountain region and how they relate to the current plate tectonic setting in western North America.

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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 Yucca Mountain region, particularly the portion of the region 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 tectonics of the Basin and Range.

Estimates of current plate motion and locations of differential movement can be derived from geologic and geodetic measurements. Making the assumption that the North American Plate is "fixed," the Pacific plate is moving toward the northwest at approximately 50 mm/year. The majority of this motion, about 35 mm/year, is taken up along the San Andreas Fault. This leaves approximately 15 mm/year of northwestward (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 stable 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.

It is recognized that this model is complicated somewhat by the fact 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/contraction/shortening normal to the San Andreas Fault. There is also some uncertainty in the amount of relative motion between the Pacific and North American Plates. In the 1970s, relative motion was estimated at approximately 60 mm/year (Atwater, 1970). More recent estimates (De Mets et al., 1990) are 48 ± 1 mm/year. There is also a possibility that some motion is accommodated within the Sierra Nevadan microplate. These detailed relationships, while certainly significant on a local scale, do not have a major impact on the regional model for the Yucca Mountain site. All available geologic and geodetic evidence indicates that the Pacific Plate is moving to the northwest relative to the North American Plate. The rate of movement appears to be about 50 mm/year (Figure 3.1.1.1.3.1.1-1). The most likely distribution of that movement (west to east) is as follows: ~5 mm/year on coastal faults parallel to the San Andreas, ~35 mm/year on the San Andreas System, and ~10 mm/year between the eastern edge of the Sierra Nevada Mountains and stable North America (here assumed to be the western edge of the Colorado Plateau).

In terms of significance to the proposed repository at Yucca Mountain, 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. This is where the site is located and where the right lateral plate motions just described overlap with the extensional tectonics of the Basin and Range Province.

As was noted briefly in the introduction to this section, normal faulting (extensional tectonics) in the Great Basin was recognized by some of the earliest studies in the area (Gilbert, 1928), but

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the amount of extension has been a matter of some debate. The timing of extension and the amounts of extension in a given time period are also not fully understood. In some parts of the Basin and Range, extensional faulting may have occurred as early as late Eocene (40 Ma), though its orientation and style may differ from later extensional episodes. Evidence also exists for Oligocene normal faulting in the Yucca Mountain region (Reynolds, 1969). There is general agreement, however, that most of the extension in the Yucca Mountain region occurred in the last 20 million years (Miocene to Recent). 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 rapid, of large magnitude, and accompanied by extensive detachment faulting, 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 but cut the detachments at depth. This model is consistent with the 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 Section 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 Section 3.1.1.2.3.1.

[Note: Use of the term regional may conflict with other definitions of the term which consider regional structures to be those outside the immediate site perimeter. The term, as used here, is meant to encompass an area in which tectonic deformation is similar to that found in the geologic environment around and affecting the site. Regionally significant faults provide examples of faults in a range of deformation styles and processes which may affect the site. The style of deformation includes strike-slip, oblique-slip, and dip-slip faulting. It also includes rupture histories characteristic of specific seismic source zones (the Stillwater-Fairview Peak-Dixie Valley sequence) and interactions between tectonic and volcanic processes (Mammoth Lakes sequence). Reviewer's comments on the terms region and regionally significant are solicited.]

Before discussing regionally significant faults it is necessary to describe the Walker Lane structural trend or, as is generally referred to, the Walker Lane belt. Faults within the Walker Lane belt accommodate part of the differential motion between the Pacific and North American

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plates. Faults east of the Walker Lane belt accommodate 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 thus 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 widest at its southeastern terminus (Figure 3.1.1.1.3.1.2-1). [INN 3.1.1.1.3.1.2-1] Part of the southeastern terminus is sharply defined 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 although the southern terminus is not clearly defined. Yucca Mountain lies on the eastern side of the Walker Lane belt, north of the Las Vegas Valley shear zone.

The Walker Lane belt has had 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. It appears probable that some Cenozoic deformation took place on preexisting structures. Cenozoic deformation in the Walker Lane belt has been characterized by strike-slip faulting, oroflexural folds, and large scale extension with associated detachment faults and metamorphic core complexes. The current style of deformation in the Walker Lane belt probably began approximately 5 Ma and is perhaps best thought of as an overlap of the right lateral shear caused by Pacific/North American Plate relative motions and gravity driven extension caused by Cordillera-wide uplift (Unruh, 1993). In this modern stress field, northwest striking faults exhibit left lateral strike-slip or oblique-slip movement.

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, summarized, or estimated 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

These authors go on to speculate that there are probably 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). Their estimate of a minimum rate of 8 mm/yr of total

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right-slip across the southern Great Basin agrees well with the ~10 mm/yr estimated from very long baseline interferometry (Argus and Gordon, 1991) as discussed previously.

[Note: This review of deformation rates in the southern Great Basin remains incomplete at this time. It is anticipated that this will be completed in a future revision.]

The characteristics of regionally significant faults in the tectonic provinces are summarized below based on published paleoseismic information. These faults have a history 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 which may produce significant ground motion at the site. The faults are discussed in order of increasing distance from the site.

Garlock Fault Zone. The Garlock fault zone 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 zone 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). From well over one hundred 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: (1) 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 1200 years; (2) 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; (3) 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 1300 years; and, (4) 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 3000 years. Vertical displacements vary between 0 and 30 percent of lateral displacements and the sense of vertical displacement varies along strike.

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,

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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 1800-1360 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 Zone. The Owens Valley fault zone is a north-northwest-striking, steeply east-dipping, 100 km long, right-lateral normal-slip fault system 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 is one of the largest historical events to have occurred in the Basin and Range Province (dePolo et al., 1991; Beanland, 1993). This earthquake is discussed in more detail in Section 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 zone 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 zone, 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 zone. 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 3500 to 8300 years. In a study of the entire fault system, Beanland (1993) found evidence for three Holocene events, resulting in an average Holocene slip rate of 2 ±1 mm/yr and an average recurrence of 3,300 to 5,000 years for elsewhere on the Owens Valley fault zone 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 Zone. 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 over 200 km northwest of the proposed site. The Stewart-Monte Cristo Valley fault was the longest identified fault that ruptured during the surface wave magnitude (M_s) 7.2 1932 Cedar Mountain earthquake (Section 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., 1991). The distributed nature of the 1932 surface ruptures may be the result of several fault segments failing in sequence (dePolo et al., 1991). The complex nature of rupture in the 1932 event has led some to suggest it as a historic analog for faulting behavior in Yucca Mountain.

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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 Zone. 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 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 (Section 3.1.1.1.4.2.2). The Dixie Valley fault is significant because it illustrates some of the key issues regarding faulting behavior that are fundamental to assessing seismic hazards in the Yucca Mountain region. For example it 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 Central Nevada Seismic Belt (Pearthree et al., 1993; see Section 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).

[Note: This marks the end of the discussion on regionally significant faults. Additional faults will be added in a future revision.]

The faults discussed above are significant as analogs of fault behavior or as potential seismic sources. Faults, whether currently active or not, can be regionally important from other perspectives. Most notably, the Miocene episodes of rapid major extension accompanied by detachment faulting and the rise of metamorphic core complexes have had a strong influence on the regional ground-water flow system. The Paleozoic carbonate formations, often referred to as a regional aquifer, may have been extensively disrupted by Miocene extension. The current state of knowledge relative to the distribution of these carbonate rocks is very limited.

[Note: This last section is incomplete but raises an important issue relating to the hydrogeologic environment around the potential site. This discussion will be expanded in a future revision.]

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3.1.1.1.3.1.3 Structural Features and Seismicity

The relationship between structural features, particularly faults, and seismicity in the Basin and Range is not clear. Some zones of seismic activity, the north-south trending Central Nevada Seismic Belt 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 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. There are other complicating factors. Earthquakes may occur on subsurface faults which have no mapped surface expression. Surface faulting is generally associated only with relatively large magnitude earthquakes (6.0 to 6.5) (Smith and Arabasz, 1991) which are rare compared to the total number of earthquakes.

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 Yucca Mountain region 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 Volcanic Features of the Yucca Mountain Region

Volcanism is a tectonic process very significant to a potential repository at Yucca Mountain. 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. The importance of both types of volcanism is discussed in the following sections.

3.1.1.1.3.2.1 Silicic Volcanism

Silicic volcanism commenced in the Northern Basin and Range during the early to middle Tertiary. By the mid-Miocene (15 to 16 Ma), volcanism had migrated southward to the vicinity of Yucca Mountain and began forming the Southwestern Nevada Volcanic field (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 middle to late Miocene. Noble et al. (1991) have divided this volcanism into three stages as follows:

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1. 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 (proposed repository host rock) was the last major eruptive unit in the main stage.

2. The Timber Mountain stage (11.5-10 Ma) had as its first major magmatic event the eruption of the Rainier Mesa member of the Timber Mountain tuff (11.4 Ma). As noted previously, the Timber Mountain tuff was erupted onto a faulted topography. The Timber Mountain tuff was 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 within or near the Timber Mountain caldera.

3. The late magmatic stage (9-7 Ma) occurred mostly around the Black Mountain and Stonewall Mountain volcanic centers further from Yucca Mountain in the northwestern part of the Southwestern Nevada Volcanic field (Noble et al., 1991).

These stages represent times of voluminous eruptions. Small volume eruptions of ash and lava of varying compositions occurred sporadically throughout this time period. [See Section 3.1.1.1.2.7 - Tertiary and Quaternary-Aged Igneous Rocks and Section 3.1.1.2.2.3 - Mid-Tertiary Aged Pyroclastic Rocks for more discussion of volcanic rocks.]

Since the end of voluminous volcanic activity in the Southwestern Nevada Volcanic field, silicic volcanism appears to have migrated away from the Yucca Mountain 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 105 km northwest of Yucca Mountain. The 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. Magmatic activity of this belt includes the Long Valley caldera, the Coso Volcanic field, and the Saline Range Volcanic field in northern Death Valley (Noble et al., 1991).

Silicic volcanism in the Southwestern Nevada Volcanic field ended about 7 million years ago. In the vicinity of Yucca Mountain it ended even earlier, at the end of the Timber Mountain magmatic stage (approximately 10 Ma).

3.1.1.1.3.2.2 Basaltic Volcanism

Basaltic volcanism of the Yucca Mountain region has been divided by Crowe et al. (1993) into two episodes: basalts of the silicic episode, and post-caldera basalts. The basalts of the silicic episode, as the name implies, were erupted during the same time as the voluminous ash-flow tuffs of the Southwestern Nevada Volcanic field, particularly during the waning stages of the field. There is also a close spatial association of some of these basalts with the calderas that erupted the ash-flow tuffs. In fact, the largest volume centers of these basalts are located in the ring-fracture zone of the Timber Mountain Caldera (Crowe et al., 1993). While volumes of all

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basalts of the region are minor compared to the silicic eruptions, the basalts of the silicic episode tend to be of significant volume with most centers having more than 3 cubic kilometers dense rock equivalent. While generally termed "basalts," these dark lavas have a wide range of geochemical compositions from basalt to basaltic andesite or latite.

Geographically, Yucca Mountain is surrounded by outcrops of basalts of the silicic episode (Figure 3.1.1.1.3.2.2-1) though most of the basalts are toward the north.

In contrast to the basalts just described, the post-caldera basalts (Figure 3.1.1.1.3.2.2-2) in general tend to be younger, of small volume (less than 1 cubic kilometer) and removed from eruptive centers of the Southwestern Nevada Volcanic field. This is particularly true for the younger (Pliocene and Quaternary) basalts in Crater Flat just west of Yucca Mountain. These basalts are discussed in more detail in Section 3.1.1.2.3.5, Volcanism at Yucca Mountain.

3.1.1.1.3.3 Alternative Models

[Note: The following discussion is preliminary and not complete. The approach taken here to presenting alternative regional tectonic models is highly conceptual. No rigorous treatment of model versus available data has been attempted. While this approach would clearly not be acceptable for the site tectonic discussion, there are advantages to keeping the regional discussion conceptual. Reviewer comments are requested on this topic.]

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 coevel with the voluminous silicic volcanism of the Southwestern Nevada Volcanic field. 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 Southwestern Nevada Volcanic field. Small volume basaltic volcanism is characteristic of the current magmatic setting.

The implications of this model are that the Paleozoic carbonate rocks which underlie the Miocene tuffs, and are considered by many workers to be a regional aquifer, have been extensively disrupted by Miocene extension. This suggests that the regional flow system is complex, and

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contains several to numerous 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. These would be 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 10,000 year life of the repository. Fault planes are expected to be high angle (cutting the Miocene detachment surfaces) 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 processes. 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 considered assumes less Miocene extension. The arguments in favor of and the 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

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extensively disrupted, and that the regional flow system is interconnected and can be modeled with relative confidence.

[Note: This is the end of Section 3.1.1.1.3.3 Alternative Models. Obviously, 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.]

3.1.1.1.4 Seismology

Section 3.1.1.1.4 describes the seismotectonic setting of the region within a 322 km radius around the potential repository site (Section 3.1.1.1.4.1), summarizes the characteristics of seismicity within this region (Section 3.1.1.1.4.2), and describes the state of the regional tectonic stress field (Section 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 Section 3.1.1.1.3 on Structural Geology and Tectonics. A review of the detailed characteristics of seismicity within 100km of the potential site is contained in section 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 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 Central Nevada Seismic Belt (CNSB; Wallace, 1984), the Southern Nevada Transverse Zone (SNTZ; Slemmons et al., 1965), the Garlock Fault Zone (GFZ; 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 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].

<u>Walker Lane Belt</u>

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 is a broad northwest trending zone of diverse topography and predominately strike-slip faulting extending 700 km from the Honey Lake basin in northeastern California to Las Vegas, Nevada (Stewart, 1988). The western boundary is the Sierra Nevada range and the eastern boundary is usually defined as the western limit of northerly trending basin and range structures in the western Great Basin. In the north the Walker Lane is about 100 km wide and the eastern boundary is defined by northwest trending faults in the seismicity of Pyramid Lake and Walker Lake. The eastern

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boundary in the central zone is less distinct. To the south the belt is 300 km wide and bound by the Las Vegas shear zone and Lake Mead. The Walker Lane belt consists of nine structural blocks which 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 which delineate this region from surrounding provinces there has been considerable movement on north to north-west 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 the 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 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 Section 3.1.1.1.4.2.3.

Late Quaternary faulting in the Walker Lane Belt and in the western Great Basin in general, occurs as pulses of activity. 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 here 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

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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 is 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 degrees 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 Section 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 portions of the province with a relatively quiescent strip along the 115 meridian (Wallace, 1984). A large part of the CNSB is located in the northern Great Basin and is the locus of several significant historic earthquakes (see section 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 Zone 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 which 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 which parallel the eastern segment of the Garlock Fault Zone. 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

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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 which produced about 100 km of surface rupture along several of the northwest trending faults in the region (USGS, 1992) see section 3.1.1.1.4.2.2.

3.1.1.1.4.2 Regional Seismicity within 322 km of the Site

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 Zone. 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. A seismic source zone is a region in which earthquake potential has a uniform spatial and temporal distribution. Source zones are defined on the basis of localization of observed seismicity, orientation of regional geologic structures, and tectonic environment. Seismic source zone is a broad term and it may encompass discrete seismic sources which are linked by a commonality in geologic structure or stress field orientation or both. Seismic sources are more confined regions of the crust which may contain Quaternary faults, historic earthquakes, Quaternary volcanic features such as cinder cones, dikes, or calderas, testing sites for underground nuclear explosions (NTS), reservoir impoundments (Lake Mead), or a combination of these. This section describes the seismic sources zones in the region surrounding the potential repository site. Figure 3.1.1.1.4.1-2 shows the location of these zones. Section 3.1.1.1.4.2.2 describes significant historical earthquakes which 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 zone and continues south into the central Mojave Desert region to the southern San Andreas fault zone. Wallace considers the Garlock fault zone to segment the Eastern California Seismic Belt; to the north the zone is characterized by both dip-slip and strike-slip Quaternary faults and to the south it is dominated by right-lateral strike-slip Quaternary faults. To facilitate the characterization of seismic source zones the SNGBZ is here defined on the basis of consistency in seismological source parameters and style of faulting. For the present purpose, this feature is defined on the west as the Sierra Nevada frontal fault from Lake Almanor in northeastern California to the Garlock fault zone, 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

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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/or 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 west southwest-east 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 shows 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. The morphology of the northern Great Basin province suggests that, 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, Pearthree et al., 1993) that may have also occurred in the mid- to late-Holocene. 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

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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° and as far south has 35° 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 Section 3.1.1.2.4.1.

The SNTZ as defined above includes a region of northeast and east-northeast trending left lateral faults which 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 activity in the Great Basin (Eaton, 1982). Historic earthquakes which have been located in the SNTZ are of moderate magnitude and there is no documented surface rupture. These events are the 1902 Pine Valley, Utah, earthquake (M 6.3), the 1966 Caliente-Clover Mountain, Nevada, earthquake (M_L 6.0; see Section 3.1.1.1.4.2.2), and the 1992 Little Skull Mountain, Nevada, earthquake (M_L 5.6, see Section 3.1.1.2.4.1) near the potential repository site.

Garlock Fault Zone

The Garlock Fault Zone (GFZ) is a major seismotectonic feature of southern California (see Section 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 GFZ 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 GFZ is 265 km long, the longest contiguous left-lateral structure in the Great Basin, and extends to within 120 km of Yucca Mountain. The GFZ 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

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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 GFZ 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 GFZ, 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 (Section 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 orientation from these mechanisms is 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 GFZ (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 320 km of the potential site. The purpose of this discussion is to provide a basis for the development of a regional seismotectonic model and to describe the variation in earthquake activity which 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) which are within 320 km of the potential repository site. This map covers the time period from 1850 to 1992. Only significant aftershocks ($M > 6$) are included on this map to give a clear picture of where seismicity is clustered. Blasts, underground nuclear explosions, and seismic events related to nuclear detonations are also 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 (UNRSL),

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the University of California, Berkeley (UCB), the University of Utah (UU), and the California Institute of Technology (CIT). 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 DNAG (Decade of North American Geology), CDMG (California Division of Mines and Geology), and NEIC (National Earthquake Information Center, which publishes the Preliminary Determination of Epicenters or the PDE).

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 only for the largest events. This is a particular problem for 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 UNRSL. 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 the installation of a two component Weichert seismograph (Slemmons et al., 1965). With the use of data from a seismograph installed in Salt Lake City, Utah, in 1914 the uncertainty of location was significantly reduced. 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 1920's, in northern California starting in 1910, and in southern 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 portions of the San Andreas fault zone. The GFZ 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. In this region there are also areas which have little or no seismic activity, such as the Death Valley area, the extreme southern portion of Nevada, and the Yucca Mountain area (Gomberg, 1991a and b). The characteristics of seismicity in the vicinity of Yucca Mountain are discussed in Section 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. 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

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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 the adjacent provinces.

Seismic Gaps

In Section 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 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, 1980). 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).

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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 has begun, 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 Sou 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_L7.2 1932 Cedar Mountain rupture zone and the M_s8.0 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 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 Section 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 GFZ (Wallace, 1984). The Sierra Nevada frontal fault is considered to be the capable structure (Wallace, 1984). This zone is seismically active at the 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 Zone. The GFZ has not experienced any significant historic events along its length except for a possible association with the 1916 southern Death Valley event (see Sections 3.1.1.1.4.2 and 3.1.1.1.3.1.2 for discussions of the GFZ). Comparing energy released by historical seismicity to inferred Holocene geologic offset has led Astiz and Allen (1983) to

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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 GFZ. 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 Section 3.1.1.1.4.2.2 and focal mechanisms for smaller earthquakes in the vicinity of Yucca Mountain are discussed in Section 3.1.1.2.3.1.

All large Great Basin earthquakes which 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 which 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

{Note: Appropriate figures for this section are not available for every earthquake discussed. Preliminary figures are provided to facilitate review of the document. As the relevant information becomes available it will be incorporated into this section.]

This section summarizes what is known about the seismological and geological characteristics of significant historical earthquakes. These events represent a range of faulting behavior

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which might 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 GFZ 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 zone (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 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).

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Wonder, Nevada, 1903. In the fall of 1903 an earthquake ruptured 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) estimates 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 Section 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., 1988, 1987, 1991).

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 which 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

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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 portion 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 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 which may have an influence on the distribution of stress (Doser, 1988). These factors have also been identified in the Yucca Mountain region 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

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late Cenozoic left-lateral displacement that truncate northwest-trending right-lateral faults (Stewart, 1988). The mainshock was preceded by 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 the 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 mechanisms 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 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,

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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 which 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 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

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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 which 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 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 Southern Nevada Transverse Zone. 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).

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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 which have 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 0.7Ma. 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 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 the 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

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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.

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 6000 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

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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 1250 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 which 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). The triggering mechanism for these events is under investigation but it has been proposed by Anderson et al. (1992) that dynamic low-frequency stress associated with high amplitude long period surface waves may be the cause. Long period surface waves have relatively high strains that extend to depths of 10 km or more, the source region for many Great Basin earthquakes.

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 is made available.]

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 portion 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 Seismological Laboratory at the University of Nevada, Reno, 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 that appears to be the common element

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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 Section 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 orientation and relative magnitudes of these axes can be determined from geologic data such as the trend 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 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 portion 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, 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 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

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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 Section 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 be reasonably postulated to be present economic or subeconomic quantities. Ground water is an important resource in this area for both agriculture and human consumption. This resource is discussed in detail in Section 3.1.2.2.

The size of the region to be discussed depends, in part, on the sources of information available and the nature of the resource. In some cases the discussion will include the entire State of Nevada; in others, it will be restricted to a smaller political region or geologic unit.

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 because, eventually, YMPO will need 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.

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Yucca Mountain includes Cenozoic and underlying Paleozoic rocks. Mesozoic rocks are not known to occur below 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 USGS and the U.S. Bureau of Mines (USBM) 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 USGS and the USBM (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:

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).

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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 ground-water resource assessment are discussed in Section 3.1.2.2.

The methods of resource assessment that will be used in the evaluation of Yucca Mountain are varied and depend upon a number of factors, including: (1) the abundance of the resource at Yucca Mountain, in the vicinity of Yucca Mountain, and in the surrounding region; (2) the present technological assessment procedures for a particular resource; (3) the type of resource involved (the resource assessment for gold has many characteristics different from an assessment of geothermal energy); and (4) the 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, the 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).

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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. Thus, only those commodities will be assessed that are known to occur either in significant abundance (e.g., zeolites) or in comparable areas or settings; 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, p. 9). Nevada leads the nation in the production of brucite, magnesite, mercury, silver, and gold (USBM, 1990, p. 6-7). 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.

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.

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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 entirely 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, location, and history 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 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 to be Mesozoic, by Carr (1984).

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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 3500 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 dominantly middle to late Miocene. The sedimentary rocks contain fragments of the altered volcanics, indicating that 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, K-mica, and limonite.

Bullfrog Hills. The Bullfrog Hills 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 locally exhibit a well developed mylonitic fabric. Rocks between the two detachment faults are highly attenuated and consist of

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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, 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.

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 actual 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

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over dolomites of the Bonanza King formation. Mineralization includes fracture-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.

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 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 the 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 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, a similar high-degree of alteration or mineral assemblage, 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.

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

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al., 1981; Hoover et al., 1982). These factors, in conjunction with the lithologic variation observed and the geophysical interpretation of vertical electrical soundings (VES), induced polarization (IP) measurements, and magnetotelluric (MT) 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 (VES, IP, and MT surveys) are consistent with a complexity of structure and pervasive alteration and a probable intrusive body at depth (Hoover et al., 1982). The IP data have outlined an area of increased polarizability that may indicate mineralization (Hoover et al., 1982).

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, pp. 743-747). 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: (1) the McDermitt; (2) the Carlin; (3) the Pinson; (4) the Borealis; (5) Hog Road; (6) Jerritt Canyon; and (7) Paradise Peak. The McDermitt Mine is the principal producer of domestic mercury. The remaining mines produce mercury as a coproduct of their 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 portion of the belt along the Oregon border, with the richest deposits located in southern Oregon (Bailey and Phoenix, 1944; Fisk, 1968). Within the central portion 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.

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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: (1) sulfurous hot-spring deposits; (2) opalite blankets; (3) altered volcanic rock; (4) altered diabase dike; (5) altered interbedded sediment; (6) altered limestone; (7) high-grade metamorphic rock; and (8) 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. The 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 Tertiary 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).

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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).

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 is 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 portion of the Calico Hills at the Nevada Test Site, 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 the 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, in Nevada, has been mined continuously 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 and 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

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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 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, pp. 451-462) 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 in Nevada of fluorspar (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, pp. 1239-1242), 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 uses 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 the 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).

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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 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 area were formed in two different types of geologic environments. Those in the Amargosa District were probably formed by the alteration of glassy, air-fall volcanic ash that was deposited in the alkaline playa lakes, and 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 deposits 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 to be 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 which it is 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). Thus, no future

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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 the 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 another 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 ground-water 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 to Yucca Mountain are located at Bare Mountain in Paleozoic carbonate rocks associated with fluorite. These occurrences at Bare Mountain are described by Garside (1973), Tingley (1984), and 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 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 ground-water 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 ground water (Garside, 1973).

3.1.1.1.5.3 Geothermal Resources

Various schemes have been developed for classifying thermal 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 the steam used in the generation of electrical power; (2) intermediate-temperature systems provide the heat used in various industrial processes (Brook et al., 1979); and (3) low-temperature systems are potentially available for use in space heating and agriculture (Samuel, 1979). This

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classification is for evaluation of geothermal reservoirs and not necessarily applicable to spring or near-surface drillhole temperatures. 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 (Samuel, 1979); (2) that the volume of available fluid be large enough to sustain its use or greater than 5 km (White, 1973); (3) that the rock has an 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 following: (1) areas of deep-seated regional high heat flow that form a conductive system which may drive hydrothermal convection (Renner et al., 1975; Brook et al., 1979); and (2) igneous-related hydrothermal systems where cooling high-level silicic or mafic magma bodies provide the energy for the 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 positions 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), Trexler et al. (1979) and a recent review of geothermal energy in Nevada has been 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 which 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 a low-temperature geothermal energy resource (Bell and Larson, 1982). For example, to the west are Goss Springs (22° C), Burrell Springs (36 to 39° C), Hicks Springs (32 to 43° C),

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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 21 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. Continuous downhole temperature profiles of these drillholes reported by Sass and Lachenbruch (1982) show a range up to 70° C within one drillhole (UE-20f, located at Pahute Mesa) and having a maximum temperature of 125° C at a depth of nearly 3,700 m.

Assessments of the geothermal gradient and heat flow of an area or region are considered more valuable, when available, in the 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.

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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 (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 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 Eureka low 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) ground water 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 Eureka low, 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 Eureka low 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 the heat flow under the Eureka low 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 Eureka low and Yucca Mountain, would fall in a zone of greater than 2.5 HFU. Caution must be employed in the interpretation of this silica geothermometry data because of the following: (1) 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 a much higher solubility than quartz); (2) 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 (3) 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 resources are more difficult to evaluate because of their ability to migrate into a permeable reservoir rock. The

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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 the Yucca Mountain area provide additional documentation.

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, 1964b). 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, 1964b; 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) does report 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 (1964c), 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: (1) 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; (2) reservoirs in which hydrocarbons accumulate within fractures; (3) impermeable confining layers above many petroleum traps; and (4) 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: (1) Devonian Woodruff Formation in Eureka and Elko counties; (2) the Cretaceous Newark Canyon Formation in Eureka and White Pine counties; and (3) 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 south western-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). 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.

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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 the 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.
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 the deposition of ore minerals.

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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 between 22 to 6 million years ago and in two other major suites of rocks which consist of: (1) both intermediate and silicic ash-flow tuffs formed from 43 to 17 million years ago; and (2) 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.

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).

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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 with 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 a 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).

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.

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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 the geometry of the ore body or the age of mineralization, are excluded from Buchanan's analysis. Einaudi (1992) presents a more inclusive list of features as essential to the 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 themselves.

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.

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 the 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.

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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 a low angle normal fault 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.

3.1.1.1.6 Geophysics

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

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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: (1) 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; (2) 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 (3) 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 the characteristics of the site and the 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 *in situ* boulder deposits are found on several Yucca Mountain ridges that are well varnished and 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, 1983).

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 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: (1) it is highly unlikely that the wash incised to bedrock during each downcutting episode; (2) headcutting from the main wash did not migrate upstream into Yucca Mountain tributaries; and (3) the overall behavior of Fortymile Wash during the Quaternary has been aggradation, not degradation.

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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 the 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 which could affect the suitability of the site, such as ground-water 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 site was defined by DOE (1988) as consisting of the potential repository block, and the surrounding area from approximately Solitario Canyon on the west to Fortymile Wash on the east, and from Yucca Wash to Busted Butte on the north and south, respectively. This area is located in southern Nye County, Nevada, on the southern margin of the Middle Tertiary-aged Southwest Nevada Volcanic Field. The main emphasis of this section will be on the area as defined by DOE (1988), but adjacent or nearby areas in southern Nye County may be described as necessary.

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 Section 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 aged pyroclastic rocks erupted from the Southwest Nevada Volcanic Field. In general, the stratigraphic section can be divided into four categories, based on similarities in age, lithology, and history:

1. Pre-Cenozoic rocks are believed to underlie the pyroclastic rocks erupted from the Southwest Nevada Volcanic Field, 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-aged 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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2. 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 Southwest Nevada Volcanic Field between approximately 10 and 14 Ma.

3. 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, although the exact age of some Quaternary eruptions is controversial.

4. 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 recent-aged surficial deposits (Swadley, Hoover, and Rosholt, 1984), and analysis of several of the 36 exploratory boreholes in Yucca Mountain (DOE, 1988). These field-mapping and borehole data also have contributed to the characterization of Yucca Mountain structures such as the Imbricate Fault Zone, by the projection of stratigraphic dips into areas where the structures are poorly exposed [DOE, 1988; 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, rather than classical geologic properties. 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, the drill holes and surface outcrop data that were currently available. In addition, a zone containing prevalent zeolites (TZZ) was 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 Section 3.1.1.2.7.

3.1.1.2.2.2 Pre-Cenozoic Rocks

Very little is known on the stratigraphy of Pre-Cenozoic aged 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

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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 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. Based on exposures at Bare Mountain and Yucca Flat, to the west and east of Yucca Mountain respectively, Carr (1974) and Snyder and Carr (1982) interpreted these Pre-Cenozoic rocks as generally consisting of 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 beneath Yucca Mountain by Carr and Snyder and Carr actually can be projected there, given that they may have been offset by Cenozoic-aged strike slip faulting of the Walker Lane Belt. Based on this observation, Robinson (1985) proposed a generalized 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. The choice between these viewpoints is still unresolved, although some work on the geometry of Mesozoic-aged thrusts in the Yucca Mountain area has been done; e.g., Caskey and Schweickert (1992).

According to the Site Characterization Plan (DOE, 1988), the viewpoint of the Yucca Mountain Project on this issue is that available data are insufficient to accurately characterize the stratigraphy and distribution of Pre-Cenozoic rocks beneath Yucca Mountain.

The stratigraphy and structure of Pre-Cenozoic rocks beneath Yucca Mountain is important to repository design and waste isolation because these Upper Paleozoic carbonate rocks are a principal regional aquifer for the southern Great Basin (Winograd and Thordarson, 1975). An accurate picture of the stratigraphic geometry and distribution of this aquifer, and especially its fracture and fault geometry, is important to understanding the regional flow of ground water away from the proposed repository.

[Proposed borehole USW G-5 is intended to penetrate the Pre-Cenozoic rocks in Yucca Wash, north of Yucca Mountain. This borehole will contribute to filling the need for more information on Pre-Cenozoic rocks below Yucca Mountain; although more information than that would be useful. The borehole is part of SCP 8.3.1.4.2.1; Characterization of the Vertical and Lateral Distribution of Stratigraphic Units in the Site Area.]

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

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breccias derived from nearby calderas and volcanoes, and ranges from approximately 1,000 m to 3,000 m thickness.

The stratigraphic nomenclature currently in use at Yucca Mountain 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. The current 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). [INN 3.1.1.2.2.3-1] The stratigraphic sequence consists of seven formations, some with several members:

A. Pre-Lithic Ridge volcanic and volcanogenic rocks
B. Lithic Ridge Tuff
C. Dacitic lava and flow breccia
D. Crater Flat Tuff
    1. Tram Member
    2. Bullfrog Member
    3. Prow Pass Member
E. Tuffaceous Beds of Calico Hills
F. Paintbrush Tuff
    1. Topopah Spring
    2. Pah Canyon
    3. Yucca Mountain
    4. Tiva Canyon
G. Post-Paintbrush ash-flow tuffs at Yucca Mountain
H. 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. The stratigraphic contacts between these volcanic rock units will be discussed in terms of their geohydrologic properties, in Section 3.1.2.3.4. [This section will be added in a later review.]

A. <u>Pre-Lithic Ridge volcanic and volcanogenic rocks.</u> 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 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).

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In drillholes UW 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 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.

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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 groundmass. Lithic fragments commonly constitute less than 5 percent of the rock. Zeolite and clay minerals are scarce or absent. However, in drillhole USW G-4 the uppermost (20 m) 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).

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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, at 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.

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: (1) mudstone fragments in the Bullfrog are fewer and smaller than in the Prow Pass and occur in equal amounts with rhyolite lava fragments; and (2) 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.

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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 comprises 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 at 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 portions of one chemically zoned magma chamber.

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 at 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 the orthopyroxene found in the upper part of drillhole USW G-3 are not found in other cores.

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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 at drillhole USW G-2 to 95 m at drillhole USW G-1 over a distance of only 2.5 km. At 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.

[A more accurate characterization of the Calico Hills section will be available once borehole data from USW G-5, USW G-6, UE-25 G#7, and UE-25 G#8/WT#19 become available. These data are to be collected under SCP 8.3.1.4.2.1 - Characterization of the vertical and lateral distribution of stratigraphic units in the site area.]

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, 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

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groundmass 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 portion 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.

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 million years bp 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 portion underlain by a less densely welded vitric zone. Compositional zonation in the Tiva Canyon ranges from rhyolite in the lower

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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: (1) columnar; (2) hackly; (3) lower lithophysal; (4) red clinkstone; (5) gray clinkstone; (6) rounded step; (7) lower cliff; (8) upper lithophysal; (9) upper cliff; and (10) 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 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).

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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 on the stratigraphy of surficial deposits in the Yucca Mountain area are primarily based 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 late Pliocene for some of the lacustrine sediments to modern for the youngest alluvial deposits (Swadley et al., 1984). The deposits are grouped into five major units: (1) late Pliocene and early Pleistocene lacustrine deposits; (2) late Pliocene and early Pleistocene alluvial deposits; (3) middle to late Pleistocene alluvial and eolian deposits; (4) Holocene alluvial and eolian deposits; and (5) 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 Sections (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

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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 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.

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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 calciumcarbonate 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 yr. 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 the 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.

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

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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 is 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

3.1.1.2.3.1 Faulting at Yucca Mountain

[Section 3.1.1.2.3.1 was not revised in Revision 3 of the Annotated Outline: 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 3.1.1.2.43.2-1 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, amount and type of displacements, number, timing, and size of prehistoric earthquakes, displacement rates, and recurrence intervals. This information forms the basis for the identification of potential seismic sources of large surface-falting earthquakes in the seismic hazard assessment (Section 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 Yucca Mountain region, it is also necessary to consider 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

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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 Shaft 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.

{The selection of faults here was based on expert opinion from the USGS with some additional sources. Reviewers comments regarding which faults were included will help direct future revisions.}

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 by the U.S.G.S, the Center for Neotectonic Studies at the UNR, and the Bureau of Reclamation, 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 if these faults have ruptured independently or concurrently in the past.

Ghost Dance Fault Zone - The Ghost Dance fault zone 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 3.1.1.2.3.2-2) did include it on their Quaternary fault 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,

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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 if the Ghost Dance fault displaces Quaternary deposits or not.

Solitario Canyon Fault Zone - The Solitario Canyon fault zone 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 Quaternary alluvium 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 1.2Ma. 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 different 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 Zone - The Bow Ridge fault zone 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 278ka and involved fracturing but no 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

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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 1000 to 10,000 years and slip rates on the order of 0.001 mm/yr.

Fatigue Wash Fault Zone - The Fatigue Wash fault zone 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 Zone - The Paintbrush fault zone 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 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 multiple surface-faulting events (three to five) occurred during the middle to late Pleistocene with a suggested average recurrence interval of 1000 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).

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Windy Wash Fault Zone - The Windy Wash fault zone 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 seven faulting events occurred during the Quaternary, with four events in the last 300ka 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 thermoluminescence age of 3 to 6.5ka. Geophysical data suggests 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 and runs 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 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 Zone - The Stagecoach Road fault zone 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 runs 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 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, Reheis (1986; 1988) and Carr (1984) both 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

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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: (1) Faults that strike northeast and have dominantly left-lateral slip (e.g., the Rock Valley, Cane Spring, and Mine Mountain fault zones); (2) 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 Springs Mountain fault zones); and (3) 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 fault zones).

Oasis Valley Faults - The Oasis Valley fault includes two sets of faults 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).

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 zones (Frizzell and Shulters, 1990; Donovan, 1991).

Rock Valley Fault Zone - The Rock Valley fault zone 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 displacements (Carr, 1984). Donovan (1991) grouped the Rock Valley fault zone with the Cane Spring and Mine Mountain faults, two subparallel faults

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to the north, to form the Spotted Range-Mine Mountain structural zone, a left-lateral northeast-trending zone that cuts across Walker Lane (Carr, 1984). Quaternary fault scarps are preserved along the entire Rock Valley fault zone, but are most prominent along the central portion where scarp heights are 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 Quaternary alluvium, 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 portions 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: (1) a southern section, roughly 48 km long, that strikes north, bounds the east side of the Resting Spring Range; (2) a central section, 5 km long, which has no associated range front and has north-northwest trending fault traces; and (3) 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 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 indicate 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 Zone - The Yucca fault zone 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 the most

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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 zone is probably one of many faults, including the Carpetbag fault system and the Area Three fault, that is partially responsible for the 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 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 230ka and they calculate an average recurrence interval, for fracturing during the last 130ka, of about 25ka.

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). 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 Zone - 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 system 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 zones, the Furnace Creek fault zone (including the Fish Lake Valley fault zone), and the Emigrant Peak fault zone (Stewart et al., 1968). Scarps and offset geomorphic features are prominent along the Furnace Creek fault zone. 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 20ka, 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 3400 to 4100 yr B.P. Two of those

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events are younger than a unit dated at 1200 yr B.P. 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 Zone - The Death Valley fault zone 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 fault zone-Furnace Creek fault zone fault system. Late Pleistocene and Holocene fault scarps are prominent along the Death Valley fault zone (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 2000-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 zone. 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 2000 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.

West Springs Mountain Fault Zone - The West Springs Mountain fault zone 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 Springs Mountain fault zone, 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 Springs Mountain fault zone (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).

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Pahrump-Stewart Valley Fault Zone - The Pahrump-Stewart Valley fault zone is a northwest-striking, steeply-west-dipping, right-lateral, oblique-slip fault 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 for 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.

West Pintwater Fault Zone - The West Pintwater fault zone 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). It 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 Zone - The East Pintwater fault zone is a north-striking, east-dipping normal fault that bounds the east flank of the Pintwater Range, about 81 km east of the proposed site (Dohrenwend et al., 1991; Reheis, 1992). It is roughly 47 to 54 km long. Although the East Pintwater fault zone 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 Zone - The East Nopah fault zone 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

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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 zone 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 fault zones.

Panamint Valley Fault Zone - The Panamint Valley fault zone is an 80 km long generally north-striking, west-dipping, normal oblique fault zone 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 portion and faults close to the range front on the southern portion (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 portion 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 310° 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 2500 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 17ka, 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 2360 years. There are Pliocene and early Quaternary faults in the Panamint Range that parallel the Panamint Valley fault zone, 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 Zone - The Hunter Mountain fault zone 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 zone, but Quaternary deposits are also displaced down to the northeast in Saline Valley and down to the southwest in

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Panamint Valley (Reheis, 1992; McAllister, 1956), which is why McAllister (1956) suggested that the Hunter Mountain fault zone may be a scissors fault with the pivot point near Grapevine Pass. Although Hunter Mountain fault zone displaces alluvium that may be as young as Holocene (McAllister, 1956; Streitz and Stinson, 1974) and sections of the Hunter Mountain fault zone 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

Sections 3.1.1.2.3.3 through 3.1.1.2.3.5 were not revised in Revision 3 of the Annotated Outline; these sections will be provided in future revisions of the Annotated Outline.

3.1.1.2.3.4 Fractures at Yucca Mountain

3.1.1.2.3.5 Volcanism at Yucca Mountain

3.1.1.2.4 Seismology

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 Section 3.1.1.2.3.2. (Quaternary Faulting History) and Section 3.1.1.2.4.4 (Identification of Seismic Sources), the basis for assessment of seismic hazard at the proposed site.]

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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.

1) 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).

2) 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.

3) 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.

4) 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.

5) 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.

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6) 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 underground nuclear detonations (UNE), 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 UNE's, 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 UNE's 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 numbers of small earthquakes occur in this area which appear to be related to the UNE. Vortman (1991) has suggested that these 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).

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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).

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 Section 3.1.1.1.4.2.3.

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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).

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

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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 Section 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 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

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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.

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

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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 (EPRI) (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 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.

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[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 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

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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 (SGBSN) of the University of Nevada at Reno (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 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.

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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 larger 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. 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 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 [INN 3.1.1.2.4.2.2-5] 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.

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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 [INN 3.1.1.2.4.2.2-7] (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 [INN 3.1.1.2.4.2.2-8] (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 3.1.1.2.4.2.2-8 through INN 3.1.1.2.4.2.2-11]. 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 (NRDS) 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 3.1.1.2.4.2.2-12], and the profile of velocity recordings of the radial component is shown in Figure 3.1.1.2.4.2.2-16 [INN 3.1.1.2.4.2.2-13]. 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).

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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.]

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

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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.

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.

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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.

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.

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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 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).

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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.

[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.

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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 Section 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 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.

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3.1.1.2.5 Natural Resources

[This section was not revised in Revision 3 of the Annotated Outline; the text for this section will be provided in future revisions of the Annotated Outline.]

3.1.1.2.5.1 Mineral Resources at the Site

3.1.1.2.5.2 Energy Resources at the Site

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

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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 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 Section 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 thermomechanical stratigraphy (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 thermomechanical stratigraphy 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.

The thermomechanical stratigraphy is providing one component of the three-dimensional framework necessary to develop the three-dimensional thermomechanical model of the site. The

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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 thermomechanical stratigraphy 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 thermomechanical stratigraphy 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 ESF to construct a thermomechanical stratigraphy 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 thermomechanical stratigraphy, 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 section 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 between Young's modulus and either strain rate or confining pressure for the ranges tested (0 to 10MPA

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confining pressure and 10^{-3} to 10^{-7} per second of strain rate). Additional testing will have to be conducted to investigate the effect, if any, of these two parameters as well as temperature and sample saturation. (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 Section 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 whitish fine grained minerals.

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 TSw1. As a matter of fact the distinction between TSw1 and TSw2 is merely one of lithophysae content, aside from differences in 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.

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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 3.1.1.2.7.3.1-2 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 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.

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

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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 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:

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

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where 6_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)]

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.

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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). The results from 15 completed tests are inconclusive because other test conditions such as confining pressure, strain rate etc., were not the same for all the tests. [Nonetheless, there are some indications that increase of temperature may cause significant lowering of the strength of tuffs. 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 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

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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 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 3.1.1.2.7.3.2-4)]

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 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.

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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

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.

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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):

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

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

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.

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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 Geologic Repository Operation Area (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.

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 Section 3.1.1.2.4.

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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: (1) healed joints with infilling minerals; (2) unhealed joints with no infilling; and (3) 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 at a strain rate of 10^{-4}/second (equal to a displacement rate on the fracture surface of approximately 7×10^{-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).]

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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 section 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 section 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 Section 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.

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,

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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 degree 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)]

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.]

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[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 Section 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 Section 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 Section 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.

Fracture Attitude

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.

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).

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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.

Table 3.1.1.2.7.5-2. Volumetric Fracture Frequency in a Unit Volume (m^3) of Rock

Drill Hole	TCw	PTn	TSw1	TSw2	TSw3	CHn1
USW G-1	NA[1]	NA	3.04	5.41	15.36	0.81
USW G-4	30.87	3.95	22.35	21.56	17.39	1.53
USW GU-3	20.79	NA	9.48	40.61	12.16	2.46
UE-25a#1	8.36	NA	2.87	10.96	12.45	1.59
Mean	20.01	NA	9.44	19.64	14.34	1.60

[1]NA *Data not available.*

Using the fracture data from the four coreholes, Lin et al. (1993) estimated the fracture frequency per unit volume of rock. Their results are given in the table above. 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

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results given above 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 above.

3.1.1.2.7.6 Mechanical Properties of Rock Mass

[Sections 3.1.1.2.7.6 through 3.1.1.2.7.15 were not provided in Revision 3 of the Annotated Outline; these sections will be added to future revisions of the Annotated Outline.

3.1.1.2.7.6.1 Rock Mass Rating (RMR) and Q Rating

3.1.1.2.7.6.2 Rock Mass Strength and Lithophysae Content

3.1.1.2.7.6.3 Rock Mass Strength (Compressive) and Confining Pressure

3.1.1.2.7.6.4 Rock Mass Strength and Joint Properties

3.1.1.2.7.6.5 Rock Mass Tensile Strength

3.1.1.2.7.6.6 Rock Mass Elastic Strength

3.1.1.2.7.6.7 Rock Mass Poisson's Ratio

3.1.1.2.7.6.8 Rock Mass Properties Measured In Situ

3.1.1.2.7.6.9 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

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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 section. 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 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 section. The surface water, ground-water (both saturated and unsaturated flow), the interrelationship between the surface and ground-water flow, flooding potential, site drainage, and other hydrologic phenomena are described. The purpose of this section is to discuss whether the areas studied provide a representative description of the conditions throughout the region, around the site, and at the site.

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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: (1) surface-water bodies and physical characteristics of drainage areas; (2) surface-water monitoring network; (3) water control structures and diversions; (4) flood history; (5) flood potential; (6) chemical composition of identified bodies of surface water; (7) location, quantity, and quality of surface water extracted; and (8) 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: (1) evaluate the potential for erosion in the site area and any hazards that this may cause; (2) evaluate the potential for floods and the potential hazards to the surface facilities due to floods; and (3) help evaluate the amount of infiltration and establish the relationship between surface runoff and ground-water 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, p. 39, 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 Yucca Mountain region has a generally arid to semiarid climate, characterized by high potential evaporation (0 to 6 mm/day, USGS, pp. 3.2-49, 1993), low average annual precipitation (175 mm/yr, USGS, pp. 3.2-35, 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. p. B-2, 1966). One small lake, locally known as Crystal Reservoir, with a storage capacity of 1.84×10^6 m^3 (Giampaoli, p. 5, 1986)

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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, p. 5, 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, p. 14, 1963) located: in Oasis Valley between Springdale and Beatty (Malmberg and Eakin, p. 8, 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 ground-water 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, ground water discharges into the river south of Alkali Flat, near Eagle Mountain about 10 km south of Death Valley Junction (Walker and Eakin, p. 15, 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, 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 ground-water 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 ground-water 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 ground-water discharge north and west of Yucca Flat.

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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 ground-water 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 ground-water discharge. According to Motts (p. 9, 1970), playas exhibit four distinguishing characteristics: (1) an area that occupies a basin or topographic valley of interior drainage; (2) a smooth, barren surface that is extremely flat and has a small topographic gradient; (3) an area that infrequently contains water, which occurs in a region of low rainfall where evaporation exceeds precipitation; and (4) 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 ground-water flow system. The playa is located between Eagle Mountain and Death Valley Junction. Estimates by Czarnecki (p. 1, 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 Yucca Mountain region 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.

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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 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:

$Q = (1.486/n)*A*R^{2/3}*5^{1/2}$
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: (1) physical characteristics of the channel (channel dimensions and geometry); (2) water-surface elevations at the time of the peak stage (usually defined by high-water marks left by debris in the flow__; and (3) 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.

[Eight more continuously-recording stream gauges are currently planned to be installed during FY 1994. 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).]

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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 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 ground-water 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, flouride, 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.

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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.

[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. p. 5, 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. p. 5, 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 west of Yucca Mountain on the border and outside of the Conceptual Perimeter Drift Boundary. 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.

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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.

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. A network of 12 crest-stage gauges has operated for almost two decades. Crest-stage gauges are simple devices that record evidence of the peak-flow stage of a stream at a specific site. Individual gauges were visited monthly during the data collection program to ascertain recent occurrences of streamflow at the gauge sites. If evidence of recent streamflow was noted during a visit, an indirect measurement of the peak-flow discharge was made and recorded along with the stage of the peak flow. 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

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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. Section 3.1.4.1.1.2.2 characterizes precipitation in the Yucca Mountain region.

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, p. 15, 1984).

According to Squires and Young (pg. 12, 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 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, p. 12, 1984). They continue, "From these marks and from data on the cross-sectional area and channel slope, a peak flow of about 566.4 m^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

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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 (p. 16, 1984) used data from 12 USGS-operated, peak flow gauging stations to develop the following regression equations:

$$Q_{100} = 482\ A^{0.565}$$

$$Q_{500} = 2200\ A^{0.571}$$

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 (p. 2, 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.

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Squires and Young (1984) conclude the following:

1. 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 2.78 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.

2. 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.

3. 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 0.98 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

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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, p. 6, 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: (1) limited streamflow data reduce accuracy of model; (2) an ephemeral channel is hydrologically and hydraulically different than a perennial channel; (3) model is deterministic and does not account for channel migration as the flood passes over the alluvial fan; and (4) model does not utilize available precipitation data.

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:

1. 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.

2. 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.

3. 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).

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4. 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. 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 T&MSS 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 section. 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 Yucca Mountain region because of the arid to semiarid climate.

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Because of the ephemeral character of most of the surface water in the Yucca Mountain region, 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 ground-water interaction with playa deposits. The chemical composition of both water samples is similar to water from the tuffaceous valley fill aquifer (DOE, 1988, p. 3-25).

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 ground-water samples of the area (DOE, pp. 3-28, 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 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 Yucca Mountain region. 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., p. 15, 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., p. 22, 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., p. 23, 1984). Spring locations and spring water chemistries are discussed in detail in Sections 3.1.2.1.1 and 3.1.2.1.6, respectively. Spring water use is discussed in Section 3.1.2.2.8.

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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, p. 5, 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.

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. Playas are not considered possible sources of water supply because of the sporadic precipitation and runoff and the excessive amounts of dissolved solids (Waddell et al., p. 6, 1984). The virtual absence of perennial surface water precludes its application for beneficial use in the hydrographic study area. Section 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: (1) recharge to the valley-fill aquifer; and (2) 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, pp. 6-231, 1986)

3.1.2.2 Regional Hydrogeology

This section provides general information on the regional hydrogeology and discusses the regional ground-water conditions that may be affected by the MGDS. In addition, the existing and projected users and uses of ground water will be described.

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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 ground-water 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 ground water, 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, 1979, p. 233). 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 Yucca Mountain region 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.

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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 ground water 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 southwestern Nevada volcanic field (north of Yucca 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 ground-water 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 ground-water fluxes, flow directions, and velocities are largely unresolved.

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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, p. 227) a *regional* ground-water flow system is defined as a large ground-water flow system which encompasses one or more topographic basins; it may include within its boundaries several ground-water 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* ground-water flow system is generally confined to one topographic or ground-water 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 ground-water basin; and flow paths are relatively short when compared to regional systems. The distinction between topographic basins (hydrographic areas) and ground-water basins is that a ground-water basin is defined as the area contributing to a significant surface discharge, and may therefore be local or regional in extent. The term *ground-water flow system* is preferred over the term *ground-water basin*, although the two terms are considered synonymous. The term *interbasin flow* is retained, and is used to designate widespread lateral transfer of ground water within rocks (usually carbonates) located at depth beneath topographic basins (Naff et al., 1974, p. 12).

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 ground-water 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 ground-water 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 Ground-Water Basin" that would include Yucca Mountain. Generalized

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potentiometric contours for the Pahute Mesa area (Savard, 1989; Blankennagel and Weir, 1973) suggest that some ground-water 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 ground water 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, p. 14) refined the terminology of Winograd and Thordarson (1975), defining the Alkali Flat-Furnace Creek Ranch ground-water basin to distinguish the flow system of interest from that which discharges in Oasis Valley. Waddell retained the designation "Ash Meadows ground-water 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, p. 3695). In Waddell's terminology, the Ash Meadows and Oasis Valley ground-water 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 ground-water 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, p. 15).

The northern boundary of the Alkali Flat-Furnace Creek Ranch ground-water basin is along a somewhat arbitrary line that crosses the Cactus, Kawich, and Reveille Ranges (Waddell, 1982, p. 15). In a regional, two-dimensional flow model with no-flow lateral boundaries Waddell (1982, Table 2) 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, p. 10 and Pl. 2) 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, p. 5 and Table 1) 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, p. 44), 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 ground water in the west-central Amargosa Desert was recharged primarily by overland flow of snowmelt in or near present-day stream

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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, p. 78), and to an unknown extent from underflow across the spring line into the central Amargosa Desert. According to Winograd and Thordarson (1975, p. 85), 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, p. 85). 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, p. 78) 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, p. 85). The amount of ground water 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 ground water in Pahrump Valley to discharge at Ash Meadows. Winograd and Thordarson (1975, pp. 90-92) 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 ground water 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 ground water 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, pp. 108-109; Winograd and Friedman, 1972, p. 3701). Third, shallow water

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levels beneath the playas in Stewart and northwestern Pahrump Valleys suggest that the ground water 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 ground-water system is the possibility of underflow from Pahranagat Valley. Based on a mass-balance calculation using the deuterium content of ground water, 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 $0.23 m^3/s$ (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 Yucca Mountain region 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 Environmental-Monitoring Program state that previously unpublished data given in the reports should be considered provisional. The periods of record for localities included in the Environmental-Monitoring Program 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 FY 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, p. 1554). 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, [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:

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1. Large hydraulic gradient (LHG) 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 LHG 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.

2. 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.

3. 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.

4. Steepening of the hydraulic gradient at the latitude of Ash Meadows. This steepening may be due to structural control, but evidence is lacking.

5. 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 Yucca Mountain region 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 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 Section 3.1.2.3.10).

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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, p. C71-C73), 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, pp. C20-C30). 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 (ATCS) 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, p. 89), these data exist in unpublished form in the files of the U.S.G.S., Water Resources Division, Denver, Colorado. The ATCS is shown as locality AD-7 on Figure 3.1.2.2.2-1. The aquifer at the ATCS is believed to be a solution-modified, low-angle reverse fault zone within the Cambrian Bonanza King dolomite (Leap and Belmonte, 1992, p. 89) 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 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, p. 378).

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, p. C25) 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, p. C27).

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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, p. C28)

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, p. 28). 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 ground-water basin (Winograd and Thordarson, 1975, p. C30). 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) 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, p. 49). 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.,

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1984). Third and last, Thordarson (1983, p. 50) 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, p. 17) 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 919m 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, p. 19). 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 (1) caving and sloughing of the hole; (2) turbulent flow losses in the well and possibly the aquifer; (3) an increase in water temperature during pumping which caused the density of the water to decrease; and (4) possibly other unknown factors in the well-aquifer system (Thordarson and Howells, 1987, p. 14). 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, p. 21). 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, p. 28) 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

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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 $4x10^{-5}$ 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, p. 38). 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, p. 14) 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, 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, p. 24). 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 700m 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

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3.1.2.2.3-3 [INN 3.1.2.2.3-4]). 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, pp. 38-39).

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, p.8). 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.

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, a "large degree of uncertainty" is associated with this solution (Craig and Reed, 1991, p. 38). Maximum apparent transmissivity is probably about 480 m^2/day, based on a recharge-boundary model. The apparent hydraulic conductivity of the 777 to 788-m depth interval is approximately 7 m/day, but for the 616 to 631-m depth interval the pumping response does not appear to be analyzable (Craig and Reed, 1991, pp. 17-21). There is no slope to the straight-line portion of the drawdown curve, indicating substantial but indeterminate transmissivity, and recovery rate is more rapid than drawdown, as commonly observed, in wells penetrating fractured-rock aquifers of the NTS region.

3.1.2.2.4 Recharge and Discharge

[Text for Sections 3.1.2.2.4 through 3.1.2.2.7 are not provided; the text will be developed in future revisions of the Annotated Outline.]

3.1.2.2.5 Age of Regional Ground Water

3.1.2.2.6 Ground-Water Flow Paths

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3.1.2.2.7 Paleohydrology

3.1.2.2.8 Regional Ground-Water 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 each of the hydrographic areas, the Nevada State Engineer has determined or estimated the perennial yield of the reservoir. The maximum amount of ground water 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 ground-water management in Nevada. When these amounts are known, a quantitative statement of ground-water availability within a specified basin can be made.

Perennial yield (also referred to as safe yield; Todd, pp. 362-368, 1980) defined as the amount of water that can be withdrawn on an annual basis from a ground-water basin without depleting the reservoir (Scott et al., p. 309, 1971). When the amount of withdrawals from a basin exceeds the perennial yield, an overdraft is created that can produce undesirable effects (Todd, pp. 362-368, 1980) since the water must come from storage within the aquifer. The result is known as aquifer mining and is prohibited in Nevada (Morros, pp. 20533, 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, pp. 362-368, 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 ground water 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 ground-water pumping near Yucca Mountain that are described in this section. There are no areas of large ground water 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 ground water is used in this area because of the low population density.

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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, p. 7, 1985). The Indian Springs well is believed to be an outcropping of the regional carbonate aquifer (French, et. al., p. 21, 1984). Five commercial establishments (Bailey's Hot Springs, Fran's Star Ranch, Ray's Last Stand, 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., p. 23, 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, p. 21, 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, p. 22, 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 toxic 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.

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Most of the water pumped from the Ash Meadows subbasin is pumped from the lower carbonate aquifer (Dudley and Larson, p. 12, 1976). Virtually all irrigation withdrawals of ground water in this area were made by the Preferred Equities Corporation (Giampaoli, pg. 5, 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, p. 3, 1976). In 1971, ground-water 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, p. 1, 1976). This natural pool, formed from the collapse of the limestone bedrock, is the only habitat of the endangered species, Devil's Hole pupfish, Cyprinodon diabolis. As a consequence of court action, ground-water 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, p. 3, 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, p. 5, 1986).

The NTS receives its water from wells operated by Reynolds Electrical and Engineering Company, Inc. (REECo). 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., p. 11, 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, p. 2, 1986), which is thought to be recharged locally by the lower carbonate aquifer (Winograd and Thordarson, p. C38, 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

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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., p. 11, 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 (SWL) and a reduction in spring discharges (Giampaoli, p. 2, 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 and the fact that there is no detectable decline reported (Witherill, pp. 102-103, 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, pp. 38-110, 1973). There has not been any decline in the SWL, which suggests that the estimated perennial yield is too low (Claassen, pp. 134, 1973; Witherill, pp. 102-103, 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 ground water is withdrawn in the northern and central parts of this subbasin. In addition, very little ground water has been appropriated to the north of Yucca Mountain (upgradient) according to information filed with the Nevada State Engineer's Office.

The major ground-water 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., p. 29, 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. Ground water was also used to supply the mineral processing plant and was obtained from three wells, with reported

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capacities of 1.64 x 10^{-2} m^3/s, 6.31 x 10^{-3} m^3/s, and 7.89 x 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, pp. 1, 4, 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, p. 21, 1992).

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 ground-water discharge have been identified in the Death Valley National Monument in California (Winograd and Thordarson, p. C95, 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., p. 15, 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., p. 39, 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 (p. Y28, 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 x 10^{-2} m^3/s to 1.30 x 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 x 10^7 m^3 per year (French et al., pg. 9, 1984). An estimated 2.10 x 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., p. 39, 1984), and nearly 10,000 acre-feet per year is artificially discharged from wells in the Amargosa Valley (Coache, p. 1, 1986). Thus an overdraft of 3.70 x 10^6 m^3 per year currently exists. Water levels in wells drilled in the valley fill aquifer

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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 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 estimates (Smith and Coogan, pp. 1-14, 1984; DOE, pp. 3-32, 1986), 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. 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 ground water 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, p. 16, 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 ground water. 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 ground-water 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., p. 1555, 1993 and Czarnecki, p. 16, 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 x 10^{-4} to 2.53 x 10^{-3} m^3/s while industrial wells produce from 6.31 x 10^{-3} to 5.05 x 10^{-2} m^3/s (including irrigation and mining wells) (Giampaoli, pp. 2-7, 1986). No injection wells exist in the hydrogeologic study area.

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3.1.2.2.9 Regional Ground-Water Management Plans

The State of Nevada is responsible for protecting ground-water resources. Chapter 534 (Title 48-Water) of the Nevada Water Laws outlines and delineates the allowable uses of ground water. Morros (1982a) presents the laws and statutes relating to all aspects of ground water 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 ground water. 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 ground water belongs to the public to be put to beneficial use only, subject to existing use (Morros, p. 20545, 1982a).

The State of Nevada uses a recharge-use philosophy in its management of ground-water resources. Total annual ground-water withdrawals from any given basin may not exceed the perennial yield (Morros, p. 20533, 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, pp. 362-368, 1980).

At the first indication of an aquifer overdraft, measures are taken to carefully regulate ground-water withdrawals. A provision of the Nevada Water Laws grants authority to the State Engineer to designate ground-water basin boundaries in areas where he deems that ground-water resources are being depleted. Such designation orders have been issued in the following areas:

1. The Amargosa Desert Ground-Water Basin
2. The Pahrump Artesian Basin
3. Oasis Valley West
4. Indian Springs Valley.

Designation of a ground-water 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: (1) There is no unappropriated water in the proposed sources; or (2) The proposed use conflicts with existing rights; or (3) The proposed use threatens to prove detrimental to the public welfare (Morros, p. 18, 1982b).

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Regional agencies that have an interest in ground-water 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 ground water through a quasi-municipal permit to Las Vegas Valley; and the DOE is responsible for ground-water 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 ground-water 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 ground-water pumpage at the Yucca Mountain site on ground-water levels and springflow in the Yucca Mountain region. This plan is outlined in "Monitoring Program for Ground-Water Levels and Springflows in the Yucca Mountain Region of Southern Nevada and California," (DOE, 1991).

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 ground-water 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 ground-water 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 ground water on the NTS. The "Groundwater Protection Management Program Plan" reviews plans for protection and remediation of ground water 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 ground-water management plans and other sources, an assessment of regional ground water use projections for the foreseeable future can be made. These projections will focus on the relevant hydrographic areas of the Death Valley ground-water 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.

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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, p. 3, 1986). Recharge to the subbasin is estimated to be approximately 2.47 x 10^6 m^3 (DOE, pp. 3-136, 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., (p. 23, 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 x 10^5 m^3 per year by the end of the seventh year of construction, decrease to about 4.36 x 10^5 m^3 per year, and remain at this level for the next 25 years. The minimum average water demand for the following 23 years of operation would be approximately 9.47 x 10^3 m^3 per year (Morales, p. 19, 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., p. 39, 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., p. 39, 1984), and nearly 1.23 x 10^7 m^3 per year is artificially discharged from wells in the Amargosa Valley (Coache, p. 1, 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, pp. 1-14, 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.

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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, p. 3, 1986). Although the 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 ground-water 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 REECo. Aside from the Yucca Mountain Site Characterization Project, 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 ground water 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 ground-water overdraft problems in the past. Certified appropriations and development permits for ground water in the Pahrump Valley totaled 1.12×10^8 m^3 in 1970,

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although in recent years actual water withdrawal has averaged approximately 4.90 x 10^7 m^3 per year (DOE, pp. 3-32, 1986). The YMP is not expected to affect water resources in the Pahrump Valley, since the proposed site is within a different subbasin of the Death Valley ground-water system (Winograd and Thordarson, pp. C90-C92, 1975; Harrill, p. 3, 1982).

3.1.2.3 Site Hydrogeology

[Section 3.1.2.3 and associated subsections were not revised in Revision 3 of the Annotated Outline Process; these subsections will be provided in future revisions.]

3.1.2.3.1 Baseline Monitoring Network

3.1.2.3.2 Site Flow System Boundaries and Hydrogeologic Units

3.1.2.3.3 Potentiometric Levels, Matric Potentials, and Gradients

3.1.2.3.4 Characteristics of Hydrogeologic Units

3.1.2.3.4.1 Saturated Media

3.1.2.3.4.2 Unsaturated Zone

3.1.2.3.5 Site Ground Water Recharge

3.1.2.3.6 Site Ground-Water Discharge

3.1.2.3.7 Age of Ground Water

3.1.2.3.8 Site Pathway Analysis

3.1.2.3.9 Local Ground-Water Use

3.1.2.3.10 Paleohydrology

3.1.3 Geochemical System

[This section 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 Section 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 AO 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 project-related data.

The geochemical data collected as part of the Yucca Mountain Project 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, measurement techniques, data quality, and uncertainties used to obtain the data in this section are discussed in detail in the references from which the data were taken.

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It is important to characterize the mineralogy and petrology of the rock units in Yucca Mountain for two reasons: (1) The phases present in the rock as well as rock texture will determine how effective the geologic barrier is and (2) 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 Member, especially the lower nonlithophysal zone in which the repository will be constructed (Johnstone et al., 1984; DOE, 1988 pp. 1-63; 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 an additional low-temperature phase assemblages that included clays and zeolites (Bish, 1989).

The Topopah Spring Member of the Paintbrush Tuff will form the near-field environment for the repository, if built, making it essential that its geochemical, mineralogic, and petrographic characteristics are well understood. Regardless of the origin of the phases, their three-dimensional abundance distributions are important to measure quantitatively to permit characterization of the repository horizon. This is important to consider for waste isolation as these are the materials that will be interacting with heat from the repository and forming part of the geologic barrier between the waste package and the accessible environment.

The following section reviews the present understanding of the vertical and lateral variations in the composition and relative volumes of phases produced by the three 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 Conceptual Perimeter Drift Boundary (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

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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 Member 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 Member may occur, we may be confident that mineralogic and geochemical data from the four wells are fairly 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 Member 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 (IUGS) classification scheme (Le Bas et al., 1986), rocks of the Topopah Spring Member 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 IUGS scheme. Latite is usually used to describe rocks, such as the uppermost portion of the Topopah Spring Member, 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 Member 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 rare-earth elements (LREE) 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 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

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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 Member 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 Member 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 Section 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 Member, 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 61 m of the unit comprise the latite which is the most crystal rich part of the Topopah Spring tuff. 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 61 m below the upper contact, that the Topopah Spring Member grades into high-silica rhyolite which generally consists of less than 2 percent phenocrysts. A fair amount of scatter in phenocryst (Figures 3.1.3.2.1.1.2-1,3) abundances is observed down section, due to the difficulty in accurately measuring representative modal abundances in phenocryst poor rocks. The Topopah Spring Member also contains lithic fragments (Figure 3.1.3.2.1.1.2-1). As might be expected from a pyroclastic unit emplaced from one 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 Member, 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 Member, 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

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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 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 Member, 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 Member 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 Member, 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 Member (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 Section 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 Member 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 (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, 1988; 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.

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One of the prominent characteristics of the Topopah Spring Member 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 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 Section 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 Member 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 Member 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

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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 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 Member 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-

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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.

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 Member 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 Member 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 Member 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 Member, 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 Member in which clay and zeolite are particularly abundant, although significant variations occur from one location to another: (1) at the top of the basal vitrophyre; (2) below the basal vitrophyre; and (3) at the top of the unit above the upper vitrophyre. 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 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

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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 Member (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 may not have been detected in previous studies (Bish and Chipera, 1989).

A few percent clay are present in the Topopah Spring Member 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 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 presence 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 Member, 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 Member contain less than 20 percent disordered illite in smectite. Zeolites from the altered horizon at the top of the basal vitrophyre are rather 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 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 Member 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 suggests 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

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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 Member may have been derived from a distal, low-temperature portion of the associated hydrothermal system. In any case, available data suggest the diagenetic assemblage has remained stable for an extended period of time, perhaps since the cessation of regional silicic volcanism.

3.1.3.2.2 Geochemistry of the Site Ground Water and Gas

[Section 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

[Section 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 section. Also, an assessment of future climatic variation is provided. Information in this section 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 (CO-OP), the Federal Aviation Administration (FAA), the Department of Defense (DOD), Desert Research Institute (DRI), and the Bureau of Land Management (BLM) 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 Section 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 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 hearing 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 m. 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 compensated 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 min size, and exist on a time scale of days.

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.

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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 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.

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

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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 Yucca Mountain region. 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 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.

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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. 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

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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" (MCC) over Arizona and New Mexico. On rare occasions, an MCC 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 Yucca Mountain region 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 results in an upslope breeze which intensifies throughout the afternoon. This effect is the "valley breeze." After sunset, the ground surface quickly cools under these conditions by radiating its heat into the atmosphere. Air very near the surface is subsequently cooled. 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 ground surface is heated by the sun. These breezes are gentle because of frictional effects and the shallowness of the layer (less than 100 m).

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).

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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, meso-scale convective complex (MCC), may form covering a larger area (200 km^2). MCCs may cause heavy flash flooding. As discussed in the previous section, 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 section. 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 Project site is _____ [INN 3.1.4.1.1.1.3-2 (TBD from pcpn studies)].

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

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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 sections 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 Yucca Mountain region 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.

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.

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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).

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.

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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 ground water. 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 ground-water 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 Section 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 Section 3.1.4.2, Paleoclimatology.

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

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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 Section 3.1.4.2).

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

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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 ____ 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.

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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 (PST) and end at 1600 PST 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, 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 synaptically influenced than at lower levels.

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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 Yucca Mountain Project 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 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 (ASL).

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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 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.

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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 pyraniometer, 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 (1 langley[ly] = 1 calorie[cal]/cm^2). 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).]

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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 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 (ALDS). This system detects cloud-to-ground lightning strikes in the Yucca Mountain project area in real time. A description of the monitoring network for lightning is provided in Section 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 (WSNSO). 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 from USGS)] 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

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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 from USGS)] illustrates the spatial relationships between lightning-strike data and rainfall-runoff data.

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 {SAIC or USGS})] 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 {SAIC or USGS})] 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.

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Another measurement of extreme wind is the peak gust. This parameter has been recorded at a monitoring site near the Yucca Mountain Project 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.

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 project site (Refer to Section 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 3.4.1.1.2.9-3 (prepare when information available from monitoring data)]. The flooding potential is discussed in more detail in Section 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

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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 nine sites operated by SAIC and five sites operated by USGS. Figure 3.1.4.1.1-2 shows these locations on a map of the Yucca Mountain project 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 USGS sites have two gauges). An additional 120 precipitation gauges are operated/monitored by USGS 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 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.

- 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, and lies parallel to the Yucca Mountain ridge and is the western boundary of the Fortymile Wash in this area.

- Site 2: Yucca Mountain is located on the Yucca Mountain ridge crest, about 3.3 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 3.4 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 10 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.

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- 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. It is approximately 6 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, about 2 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 about 2 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 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 about 19 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; this includes the nine sites operated by SAIC and the five sites operated by USGS (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 to 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 (NIST) standard as is done with the meteorological stations. Also, the station is inspected for such things as cable integrity and propane supply (where installed).

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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 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 project site:

- Lightning data (Refer to description in Section 3.1.4.1.1.2.8) are collected from an existing automated lightning detection system (ALDS) operated by the WSNSO for the NTS. The network consists of seven direction finders (DF). Five are located on the NTS with additional DFs located at Blue Diamond and Warm Springs, Nevada. The YMP area is located on the western edge of the present ALDS network. The accuracy of the ALDS 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 (NOAA), Polar orbiting satellites, and geostationary (GOES) 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.

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- 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 Sections 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.

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 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.

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- The vertical temperature difference method (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:

A. Stability class A defines an extremely unstable (highly convective) atmosphere.

B. Class B is for unstable conditions.

C. Class C is slightly unstable.

D. Class D is neutral.

E. Class E is slightly stable.

F. Class F is stable

G. 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 significant features of the stability distributions are summarized in Table 3.1.4.1.3-1 [INN 3.1.4.1.3-2 (prepare table similar to table 5-11 from SCP chapter 5 for each SAIC station when data become available, example attached) and discussed in the following paragraphs.]

Class A stability was most frequently associated with winds from the ___, while winds from the ___ through _____ were most frequently associated with class B stability. For class C stability, winds from the _____ were the most common occurrence, but winds from the _____ through the _____ were also quite frequently associated with class C stability. Occurrences of class D stability were the second most commonly observed stability classification and were commonly associated with winds from the _____ and the _____ through _____. The distribution for class E stability indicates that there is a distinct shift from the generally _____ly winds associated with the unstable and neutral stability classifications (classes A through D) to predominantly _____ly winds for class E and class F. Stable atmospheric conditions (class F) were the most commonly observed stability class and the distribution clearly

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shows the predominance of winds from the _____ through the _____. Because stability classes E and F are both associated with relatively light winds, the predominance of winds from the _____ through the _____ for these classes is most likely due to drainage winds at this site that develop under synaptically calm conditions.

In summary, neutral and stable conditions (classes D, E, and F) were by far the most commonly experienced at _____ (station name) and account for ___ percent of the total observations. Stable conditions tend to be dominated by winds generally from the _____, while neutral conditions had a significant _____ to _____ly component in addition to a strong _____ly component. Unstable conditions (classes A, B, and C) occurred only ____ percent of the time, had virtually no _____ly component, and were dominated by winds from the ____ through the ____, with the most unstable classes having a stronger _____ly element than the slightly unstable category.

Determining the stability distributions is important from the standpoint of evaluating the potential impacts of particulate and gaseous emissions from the repository. The data required as input to the dispersion models used in acquiring permits for both the site characterization activities and the repository (through the environmental impact statement process) are discussed in a plan for environmental monitoring and mitigation.

3.1.4.2 Paleoclimatology

[Skeleton text not developed for Sections 3.1.4.2 through 3.1.5.3.]

3.1.4.3 Future Climatic Variation

3.1.5 Integrated Natural System Response to the Maximum Design Thermal Loading

3.1.5.1 Response of Geomechanical Subsystem

3.1.5.2 Hydrologic Response to Thermal Loading

3.1.5.3 Response of Geochemical System to Thermal Loading

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