SECTION 3.0

CONCEPTUAL FRAMEWORK FOR THE YUCCA MOUNTAIN GROUNDWATER SYSTEM

3.1 REGIONAL TECTONIC SETTING OF THE YUCCA MOUNTAIN GROUNDWATER SYSTEM

Conceptual considerations undertaken in this report are concerned, mainly, with the west-central portion of the Death Valley groundwater system. This portion, thereafter referred to as the Yucca Mountain groundwater system, is situated within the southern Great Basin (Plate 3.1-1). It is commonly recognized that the tectonic evolution of the Great Basin during the Cenozoic Era has produced a sharply-defined and highly-unique intracontinental region. A distinct structure of this region is considered to be the manifestation of an ongoing crustal extension. The origin of such an extension is still subject to study and debate. Several mechanisms have been proposed. They include a) oblique-extension following cessation of subduction to the west (Atwater, 1970; Christiensen and McKee, 1978); b) back-arc extension (Scholz et al., 1971); and c) active mantle upwelling (Eaton et al., 1978).

Conceptual understanding of the tectonics of the Great Basin is an important aspect of the conceptual considerations undertaken in this report. Results of various geophysical measurements and observations provide important inputs and constraints for the development of such an understanding. The observations most useful for this purpose are a) measurements of the gravitational field; b) measurements of the travel-times of various seismic waves; c) measurements of intensity of flux of terrestrial heat; and d) measurements of electric conductivity.

Plate 3.1-2 is a simplified isostatic anomaly map for the United States as compiled by Woolard, 1969. The map shows that the Great Basin region is associated with a pronounced mass deficiency. Free-air and isostatic anomalies are negative, ranging in magnitude from -10 mgal to less than -30 mgal.

Results of seismic refraction and reflection studies indicate, that in the Great Basin, the crustal thickness is only 25 to 30 km (Plate 3.1-3). This thickness is much smaller than that required to allow considerations of the isostatic compensation solely by the Airy-Heiskanen mechanism (Woolard, 1969). The isostatic compensation must be achieved, at least in part, via the Pratt-Hayford mechanism (Tatle and Tuve,

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1955). In this situation, the density and by inference the P-wave velocity in the upper mantle should be lower than the normal, 8.2 km/sec, upper mantle velocity for continental regions (Ringwood, 1969).

Plate 3.1-4 is a map showing distribution of the P-wave velocity in the upper mantle for the continental United States. This map has been compiled based on data from deep-seismic soundings, underground nuclear explosions, and natural earthquakes (Herrin, 1969). It shows, that in the western United States, there are strong regional differences in the P-wave velocity in the upper mantle. Beneath the Great Basin, the P-wave velocity is significantly lower, averaging about 7.8 km/sec.

Recent measurements have shown, that in the Great Basin, the intensity of flux of terrestrial heat is 50 to 300 percent higher than the worldwide continental average (Von Herzen and Lee, 1969; Thompson and Burke, 1974; Swanberg and Morgan, 1978; Sass and Lachenbruch, 1978; and Plate 3.1-5). This greater intensity suggests that the upper mantle and lower crust are exceptionally hot. Near the Moho discontinuity temperatures may be several hundred °C higher than in the surrounding regions. These higher temperatures may be responsible for a channel of high electric conductivity, as noted beneath the Great Basin, by Porath, 1971.

Plate 3.1-6 is a schematic representation of the conceptual understanding of the Great Basin as interpreted, based on geophysical data, by Scholz et al., 1971. The crust is thin, averaging about 30 km. It overlies a low-seismic velocity mantle. The coincidence of geophysical anomalies suggests that the upper mantle is partially melted. Apparently there is no mantle lithosphere; the low-velocity zone extends up as far as the Moho discontinuity.

Based on interpretations of the above geophysical observations, it is reasonable to assume, that in the Great Basin, viscosity of the upper mantle is low, within a range of 10^{20} to 10^{22} poise and locally less (Artyshkow and Mescherikov, 1969; Cathles, 1975). In this situation, it is also reasonable to expect that a convective flow of mass and heat, either thermal or gravitational, occurs in the upper mantle of this region. The following reasoning is offered in support of this statement. Cathles, 1975, indicated that the critical Rayleigh number (stability limit for the Rayleigh instabilities) for the mantle, assuming the "fixed-fixed" boundary conditions which give the largest critical Rayleigh number, is $R_{ac} = 1{,}700$. The Rayleigh number (stability parameter for the Rayleigh instabilities) is

$$R_a = \frac{\alpha \cdot \rho \cdot g \cdot d^3 \cdot \Delta T}{K \cdot \mu \cdot}$$

where:

α– coefficient of thermal expansion, $\alpha = 2\text{x}10^{-5}.^{\circ}\text{C}^{-1}$;

ρ– density, $\rho = 4$ gm/cc;

g– gravitational acceleration, $g = 10^3$ cm/sec^2;

d– depth;

ΔT– non-adiabatic temperature difference;

K– thermal diffusivity 10^{-2} cm^2/sec; and

μ– viscosity.

For a layer 100 km thick, the Rayleigh number is $R_a = 8.5 \cdot \Delta T$, if mean viscosity of such a layer is $\mu = 10^{21}$ poise; and $R_a = 85 \cdot \Delta T$ for a layer with $\mu = 10^{20}$ poise. The Rayleigh instabilities (buoyancy exceeds viscous retardation) develop if the non-adiabatic temperature difference (ΔT), between the top and bottom of the layer, exceeds 200°C for $\mu = 10^{21}$ poise and 20°C for $\mu = 10^{20}$. Estimates of reasonable geothermal gradients for the upper mantle indicate that such requirements may be satisfied with ease (Birch, 1969). Furthermore, these estimates suggest, that if viscosity of the upper mantle in the Great Basin is less than 10^{21} poise, the Rayleigh number may be vastly larger than the critical Rayleigh number (i.e. $R_a \gg R_{ac}$).

Results of experimental and theoretical considerations of the Rayleigh instabilities indicate that cells of convective motion become narrower and more concentrated as the Rayleigh number progressively exceeds the critical Rayleigh number. A convective system tends to establish a mode of convection that maximizes heat transfer. This means, that for the Rayleigh number (R_a) near the critical Rayleigh number (R_{ac}), convection cells with nearly square cross-sections are preferred. At large values of the Rayleigh number, however, the preferred cell geometry depends on R_a; with increasing R_a the aspect ratios of convection cells become smaller. Consequently, if mean viscosity of the upper mantle is around 10^{20} poise and

the non-adiabatic temperature difference (ΔT) is equal to a few hundred °C, one may expect narrow convection cells which take the form of plumes with cylindrical cross-sections, or elongated rolls with sheet-like zones of upwelling.

Consideration of upward movements of matter in the upper mantle as Rayleigh instabilities is rather gross over-simplification of the phenomenon. Nevertheless, the results of such considerations are useful in demonstrating that postulations of convective motions, in the "soft" upper mantle, are reasonable propositions and deserve a closer look and further inquiry. These results, however, are not regarded as an unequivocal demonstration of either presence or absence of the upper mantle instabilities in a particular region.

Consideration of the upper mantle flow as a Rayleigh instability implies that such flow results from uniform heating of a horizontal layer of fluid bounded on the upper and lower surfaces, but unaffected by the end conditions. Fluid is regarded as initially motionless and incompressible or, for the Jeffrey's problem, only slightly compressible. The stability calculations are made assuming that the upper mantle is composed of the Newtonian fluid, with homogenous chemical and physical composition, and with uniform in time and space viscosity. It is not difficult, however, to raise rational objections to all of these assumptions and develop some further reservations. Weertman, 1970; Rayleigh and Kirby, 1970; Green and Radcliffe, 1970; and Carter et al., 1972, have pointed out that the constitutive law that governs stress-strain behavior over most of the upper mantle, is most likely non-linear. This law may be best expressed through the familiar power creep equation in the form $\dot{\epsilon} = A \cdot \exp\left(-\frac{Q}{RT}\right) \cdot \sigma^n$. The flow mechanism differs from that of a substance in the amorphous state. Depending upon physical conditions (temperature, pressure, and strain rates), the upper mantle flow occurs as some combination of the Nabarro-Herring creep (diffusive flow), plastic deformation (dislocation glide), polygonization (dislocation climb), and syntectonic recrystallization (hot working) (Carter et al., 1972). Knopoff, 1964; 1967, has pointed out that a significant vertical heterogeneity in chemical and/or physical composition of the mantle will inhibit vertical movements of matter, provided the depthward increase in density is sharp enough. The effect of depthward increasing pressure on viscosity of the upper mantle materials may be greater than the contrasting effect of temperature. Consequently, in accordance with the Nabarro-Herring theory of diffusive creep or the Andradean concept of viscosity, one may postulate, that in the upper mantle, there is a downward increase in viscosity (Magnitsky and Zharkov, 1969). One may also claim that the effective viscosity of a material described by the logarithmic creep law increases with time $[\mu = \frac{E}{\phi(t)}]$. In a convective system, this has the effect of

slowing the convective rate, which in turn raises the temperature gradient and increases the driving force. The end result is that the effective viscosity may depend on the rate of heat transfer and may fluctuate with it (Lomnitz, 1974).

While exploring the feasibility of the upper mantle convection, whether thermal or gravitational, one may of course counterbalance all of the above and other objections. Allan et al., 1967, have pointed out that, in the presence of lateral temperature gradients, fluid is always unstable. This is in contrast to the Rayleigh instabilities, where circulations do not start until the vertical temperature gradient exceeds a certain critical value. The upper mantle convective flow may be augmented by downward segregations of heavier materials. Such segregations cause hydrostatic disequilibrium and result in local releases of gravitational energy, so-called chemical convection (Lyustikh, 1969). Fractional melting of the upper mantle material results in upward migrations of low melting components that comprise basaltic magmas. Denser materials, however, remain and may sink deeper into the Earth's interior. Order of magnitude calculations show that density variations of about 1 part in 10^5 will produce flow at a rate of about 1 cm/year (Runcorn, 1969). Furthermore, the fractional melting produced basaltic magmas may collect in independent liquid masses, so-called asthenoliths. The ascent of such asthenoliths will further augment gravitational or thermal instabilities. Finally, in consideration of upward movements of matter in the upper mantle it is critically important to know whether local mean density of the asthenospheric material (low-velocity channel) is lower than that of the overlying upper mantle material (between the Moho discontinuity and the low-velocity channel) (Beloussov, 1969). Some hypotheses presume that, locally, this may be the case indeed. The resulting "density inversion" causes the layered system to be gravitationally unstable (Biot, 1961; Ramberg, 1981). Contrasts of viscosity, density, and layer thickness determine whether and how fast the potential diapiric motions occur.

Considering our fragmentary knowledge of the physico-chemical composition of the Earth's mantle, and of conditions present there, it is not surprising that diversification of opinions, regarding feasibility of mantle convection, is great. Runcorn, 1962, accepts convection currents throughout the entire mantle. However, Bernal, 1961, restricts such currents to the lower mantle, and Tozer, 1965, to the upper mantle. The upper mantle was considered as a potential barrier to convection by Knopoff, 1967; as a stimulating agent by Vening Meinesz, 1964; and, if the process is initiated at a sufficient depth and with sufficient energy, as a conquerable barrier by Verhoogan, 1965.

In summary, even a casual review of geophysical literature indicates that there are many reasons to conclude, that in the Great Basin, the structure of the Earth is not radially symmetric. In the lower crust and upper mantle important lateral heterogeneities exist. The presence of such heterogeneities strongly suggests, that in the upper mantle of the Great Basin, convective motions involving mass and heat are taking place now and have taken place in the past. On one hand, the structural heterogeneities may be regarded as the driving mechanism for such motions. On the other hand, however, the motions may be regarded as a process whereby the structural heterogeneities are produced. In either case, however, there is a firm base to believe that there may be a fundamental cause-effect relationship between the upper mantle dynamics and tectonic activities at and near the Earth's surface; such as earthquakes, fault movement, and volcanism. How do such activities and their fundamental causes affect the long-term behavior of a near-surface groundwater system? The following pages are devoted to seeking an answer to this interesting and potentially important question.

3.2 GENERAL DESCRIPTION OF EXTENSION DOMINATED TECTONIC ENVIRONMENTS AND THEIR RELATIONSHIP TO GROUNDWATER SYSTEMS

The contemporary tectonic environment of the southern Great Basin, as of many other deformational systems that are dominated by extension and alkalic-tholeiitic volcanism, may be regarded as a manifestation of fundamental tectonic processes operating deep within the Earth's interior. It is reasonable to assume that first-order expression of such processes is a convective flow of heat and mass, from the asthenosphere into the overlying continental lithosphere. Such a flow amounts to diapiric upwelling of a high-temperature, low-density, and low-viscosity asthenospheric material. The process, by introducing large quantities of energy into the crust, controls and otherwise influences many geologic processes operating near the Earth's surfaces, possibly including the flow of groundwater. For the purposes of conceptual consideration undertaken in this report, an understanding of this process may be of considerable importance.

An informative illustration of a convective flow of heat and mass in the upper mantle was developed by Bridwell and Potzick, 1981. The essence of this illustration is shown on Plates 3.2-1 through 3.2-4.

Plate 3.2-1 presents the results of numerical simulations of the convective system that operates in the upper mantle of the Rio Grande Rift. The process was modeled symmetrically as a coupled thermomechanical continuum in two dimensions. The convective system is shown as the velocity field and as the deformed mesh. The corresponding temperature and viscosity fields are reflections of the process operating for 6 x 10^6 years and causing an increase in surface flux of terrestrial heat of about 1 HFU (1 HFU = 1 μ cal/cm^{2}s). Plate 3.2-2 shows that convective flow in the upper mantle introduces significant changes into structure of the upper mantle and crust. Such changes include a) tectonic thinning of the crust; b) magma bodies; and c) locally increased temperature and decreased density and viscosity of materials forming the upper mantle and crust.

The convective flow of heat and mass in the upper mantle may be regarded as a steady-state process. In this situation, there are two tectonic factors that need to be reckoned with a) steady-state flux of terrestrial heat, that in terms of its planar distribution, is heterogeneous ($\partial t/\partial x \neq 0$ and, therefore, $\partial t/\partial z \neq$ const.) and b) steady-state straining of the crustal base, that is responsible for the base parallel shear stress gradient ($\partial \tau_{zx}/\partial x \neq 0$). Plate 3.2-3 shows both of these factors operating in association with the

previously considered convective system. The resulting surface flux of terrestrial heat is strongly heterogenous. This is caused by a non-uniform planar distribution of temperature in the lower crust. The shear stress gradient is the result of outward flow in the upper mantle. This gradient causes the overlying crust to be subjected to continued extension.

Material forming the upper mantle, and containing the convective system, is gravitationally unstable. The hot, low density and viscosity, mass tends to move buoyantly upward. This causes a local on-going uplift of the Earth's surface. In the crust, the state of *in situ* stress varies in time and space. The continuously provided strain energy is utilized by advancing deformation of the crust through uplift, faulting, tilting, and viscous flow at deeper levels. Plate 3.2-4 shows the distribution of shear and normal stresses developed in the crust as a function of distance away from central portion of the convective system. The example shown corresponds to a system operating for 7×10^6 years, and causing surface flux of terrestrial heat of about 2.5 HFU. The stresses were computed assuming, unrealistically, that no deformation occurs in the crust. Consequently, the stresses are not reduced to account for faulting, uplift, or any other forms of deformation; tensile stresses exceed those that can be sustained by crustal rocks, also the vertical normal stress exceeds the lithostatic pressure.

A more explicit illustration of the *in situ* stress field in an extension dominated deformational system was developed by Spencer and Chase, 1989. These authors have used the Airy stress functions to calculate *in situ* stress conditions resulting from several hypothetical tectonic configurations. For the purposes of these considerations, particularly noteworthy are the results of calculations of *in situ* stress conditions that arise in elastic medium subjected to a uniform extension and a sinusoidally varying, upward directed, normal stress acting at the base of a system. Such boundary tractions simulate tectonic configurations where a) extension originates from outside of the deformation area and b) isostatically uncompensated upper mantle exerts an upward directed stress that is balanced by a much larger wavelength, lower amplitude uplift of the land surface. Effect of the sinusoidally varying vertical normal stress is to produce a sinusoidal flexure at the base of a system, that in terms of the boundary tractions, amounts to an application of the flexural stresses. Such flexural stresses are analogous to the shear stress gradient $\left[\tau_{zx_{(x)}}\right]$ resulting from outward flow in the upper mantle. Hence, the stress considerations performed by Spencer and Chase, 1989, are helpful in developing an understanding of the *in situ* stress field operating in the upper crust above a convective system in the upper mantle.

Plates 3.2-5 and 3.2-6 present the results of stress calculations for the above stated boundary tractions. Plate 3.2-5 presents failure-surface trajectories for three different ratios of the end normal stress (N) to the sinusoidally varying base normal stress (b). The failure-surface trajectories are lines showing distribution of points for which a potential of shear failure is greatest. Such lines are oriented 30° with respect to trajectories of the major principal stress (σ_1). The plate shows that the failure-surfaces are listric; they flatten downward and upward, and merge into a subhorizontal detachment zone. The inclination and depth of occurrence of the detachment surface increase with the increasing extensional stress (decreasing value of the ratio N/b). Plate 3.2-6 is a contour diagram showing relative magnitudes of shear stresses for two different ratios of the end normal stress (N) to the base normal stress (b). The stress magnitudes shown are in units of the maximum base normal stress. The zone of coalescing failure-surfaces is a region of low to moderate shear stresses. In this region, however, the normal stresses acting across the detachment zone are also low; so that potential for shear failure along low-angle surfaces is greatest. The increasing magnitude of the end normal stress (N) causes the magnitudes of shear stresses to increase.

Both of the diagrams are helpful in developing an understanding of the mechanics of low-angle normal faulting, so-called detachment faulting. Such faulting is widely regarded as an integral element of contemporary tectonics of the Great Basin, including the region of Yucca Mountain (Scott, 1985, and Section 3.2.1). Nucleation and propagation of detachment faults require the *in situ* stress field in which trajectories of the greatest principal stress (σ_1) are rotated off a vertical plane. Such trajectories are not the result of horizontal extension, that if devoid of deviatoric stresses from other sources, will produce failure-surfaces with dips of about 60°. To produce low-angle normal faults, additional stresses must be introduced to obtain the desired *in situ* stress configuration. Application of a basal shear, in combination with either horizontal compression or extension, leads to the state of *in situ* stress that is required for the occurrence of detachment faulting. However, in an extension dominated tectonic environment, opportunities for the development of laterally extensive and deep basal shear stresses are limited. A best bet is for the isostatically uncompensated upper mantle. Such conditions of the upper mantle, of course, are the result of convective flow of mass and heat, whether thermal or gravitational.

Returning to general considerations of groundwater systems in extension dominated terrains, simultaneous introduction of both forms of energy (thermal and strain energies) leads to the establishment of two fields of tectonic energy. For time scales of geologic proportions, both of these fields are interactive and continuously changing in time and space. The first field is a geothermal field. At any given time, such

a field is expressed by the flux of terrestrial heat through the Earth's surface. For given boundary conditions and for a given spatial distribution of thermal properties, distribution of this flux reflects a set of equipotential surfaces that describe spatial distribution of points of equal temperature.

The second field involves strain energy. At any given time, such a field is expressed through ongoing deformation of the Earth's surface. For given boundary conditions and for a given spatial distribution of material properties, such deformation reflects the *in situ* stress field that is contained within deforming crustal medium.

If it is assumed, that a) thickness of the system remains const.; b) boundary conditions remain unchanged; and c) material properties are temperature-independent, then time-dependence for both tectonic fields is caused by the steady-state straining of the base of the deformational system. For the *in situ* stress field, time-dependence is caused directly by the continued introduction of strain energy, and the resulting on-going deformation. Partially, such time-dependence is expressed as a change in positions of stress trajectories relative to a fixed reference, and as a change in shape of stress trajectories relative to each other and relative to the reference. During most of the time, the rate of these changes is relatively small. During short intervals of time, however, the rate may be significantly increased. This occurs in association with faulting that may entail a substantial restructuring of the *in situ* stress field.

At certain stress levels, the effective thermal conductivity structure of a deforming fractured medium is influenced by the *in situ* stress field (Walsh and Decker, 1966; Brace, 1972). Consequently, if the development of the *in situ* stress field reaches a certain stage, continued introduction of strain energy will cause time- and space-dependent alteration of the effective thermal conductivity structure. Although the thermal boundary conditions remain unchanged, time-dependence of the *in situ* stress field leads indirectly to time-dependence of the geothermal field. This is expressed as a drift of local values of geothermal gradients. The *in situ* stress induced changes in a geothermal field are expected to be occurring at exceedingly small rates. However, magnitudes of the resulting local reductions of vertical geothermal gradients are expected to be meaningful (e.g., in a range of 5 to 15°C/km) (Sections 4.3.3 and 4.3.4).

For the purposes of these conceptual considerations, it has been assumed that the *in situ* stress field and the geothermal field are the only factors that relate the Yucca Mountain groundwater system and tectonic environment of the southern Great Basin. Idealization of such a conceptual linkage is shown on

Plate 3.2-7. The shear stress gradient $\left[\tau_{zx_{(x)}}\right]$, at the base of a groundwater system, causes the flow of groundwater to occur in a cyclically deforming fractured medium. High and spatially heterogenous flux of terrestrial heat $\left[\mathrm{HFU}_{(x)}\right]$ causes the flow of groundwater to occur as a coupled heat-fluid flow.

For the purposes of all subsequent discussions, it has been assumed that both the increased flux of terrestrial heat and the shear stress gradient are related to a convective system in the upper mantle. It should be recognized, however, that this assumption is not a critical one. What is important, is the assumption that both forms of energy are being introduced into the groundwater system simultaneously. If the increased flux of terrestrial heat and the deforming nature of a fractured medium are related to some other tectonic process, or combination of such processes, their combined hydrologic effect will not be different.

A groundwater system is an energy field in itself. Such a field is expressed by fluxes of groundwater through boundaries of a system. For given boundary conditions and a given hydraulic conductivity structure, these fluxes reflect a set of equipotential surfaces that describe spatial distribution of points of equal hydraulic head. How, and to what degree may a groundwater system be influenced by the two other fields of tectonic energy? A main purpose of subsequent inquiries is to seek an answer to this important question. A real question that is being considered is: What are potential hydrologic implications of the volcano-tectonic setting of the Yucca Mountain groundwater system? What is desired is an understanding, on qualitative or conceptual basis, of long-term behavior of this system.

The remainder of this section has been subdivided into three parts. The first part, Section 3.3, is concerned with relationships between a changing *in situ* stress field and a groundwater system. The *in situ* stress field is assumed to be operating in a fractured medium, that also contains the groundwater system. The geothermal field is regarded as having no influence on the other two fields.

The second part, Section 3.4, deals with relationships between a geothermal field and a groundwater system. In this part, it is assumed that neither of these two fields is influenced by the *in situ* stress field.

The third part, Section 3.5, is concerned with partial three-way coupling between a geothermal field, an *in situ* stress field, and a groundwater system. It has been assumed that neither the geothermal field nor the groundwater system influences the *in situ* stress field. The *in situ* stress field, however, influences both the geothermal field and the groundwater system.

3.3 CONCEPTUAL CONSIDERATIONS OF GROUNDWATER SYSTEMS DEVELOPED IN A DEFORMING FRACTURED MEDIUM

3.3.1 Contemporary Tectonic Environment of the Yucca Mountain Groundwater System

Tectonic deformation of the area bordering and corresponding to the Yucca Mountain groundwater system, during the Pliocene and Quaternary times, is known based on various structural geology studies performed at or near the Nevada Test Site. Data concerning contemporary deformation of the area are also available. These data were obtained based on studies of regional and local seismicity and based on local geodetic measurements.

Long-term extension rates across the southern Great Basin, at latitude 37°N, are of the order of 1 cm/year or $+0.07 \times 10^{-6}$ m/m per year (Wernicke et al., 1982). These rates are average values calculated for the entire province, between the southern Sierra Nevada and the central Colorado Plateau, for the last 15×10^{6} years. At any given time, however, extension seems to be confined to relatively narrow belts, as appears to be the case today in the Death Valley region, rather than being uniformly distributed across the entire province. Local extension rates, therefore, may be different than the average and may differ with time.

From a tectonic perspective, the groundwater system is situated within the southeastern portion of the Walker Lane, a northwest trending belt of right-lateral shearing, and of continental proportions (Stewart, 1967; 1968). As shown on Plate 3.3.1-1, the Walker Lane belt is located in the southwestern portion of the Great Basin, along the California-Nevada border. Displacements along this belt occur, in part, as dextral slip and, in part, as dextral bending. Total displacement may be as large as 130 to 190 km (Stewart et al., 1968). Some faults, located within the Walker Lane, are covered by undeformed Upper Miocene or Pliocene volcanic rocks, while others displace Pliocene and Quaternary deposits.

Plates 3.3.1-2 through 3.3.1-4 show locations of more prominent Late Pliocene-Quaternary faults developed within the groundwater system, and in the area of the proposed high-level nuclear waste repository. Surface displacements of Quaternary age are known to be associated with many of the local faults. Movements of Late Pleistocene and Holocene age have been documented on several such faults (Knauss, 1981; Carr, 1984; Whitney et al., 1986; and Reheis, 1986). Typically, lengths of individual faults range

from several kilometers to tens of kilometers. Displacements of Quaternary materials are measured in meters and were developed during repeated ruptures.

In some areas within the Walker Lane, the structural grain is notably deflected from the regional trend. Such areas appear to have been rotated about vertical axes in response to dextral shear couple. Recent paleomagnetic studies of Miocene ash flows indicated that an area situated near the proposed repository has been internally rotated 30° clockwise since Middle Miocene (Scott and Rosenbaum, 1986). This area is 25 km long and is situated within central portions of the groundwater system.

Rotations along horizontal axes are also known in the region. Southward or southwestward tilting of, at least, western portions of the area of interest has been suggested by Carr, 1984. Volcanic units emplaced 8 to 15 x 10^6 years ago pinch out abruptly southward, which suggests that they were emplaced on a surface that sloped northward or northwestward. By contrast, present topographic surfaces slope generally southward from a regional topographic high in central Nevada. Southward or southeastward tilting of 3.0 to 7.0 m/km, during the past 3 x 10^6 years, is indicated by a southward decrease in altitudes of playa deposits (Carr, 1984). Contemporary tilting is suggested by an apparent tendency of centers of playa deposition to shift southward with time.

Seismicity of the southern Great Basin is shown on Plate 3.3.1-5. The area shown roughly corresponds to the Death Valley groundwater system. The events were recorded during a period from August 1, 1978 through December 31, 1983, by a 53 station seismic network. The pattern of local seismicity is of a widespread but diffuse background, corresponding principally to the east-west seismic belt of Greensfelder et al., 1980, punctuated by clusters of intense activity and by larger nearly aseismic areas. The events are of low magnitude $1 \leq M_L < 3$. The focal depth of hypocenters ranges from 1 to 17 km (Rogers et al., 1987). The depth distribution has two broad peaks at 0 to 2 km and 5 to 8 km and a distinct minimum at 3.5 km. Depth profiles in active areas frequently show earthquake patterns extending from the ground surface down to a depth of 10 km or more. Focal mechanism solutions, determined based on radiation patterns of some 29 events, indicate that strike-slip faulting is a dominant mode of elastic stress release. For most of the solutions, the B-axis, which is a pole to a plane containing the slip vector, plunges steeply. Only 6 or 7 solutions yielded the B-axis whose plunge is less than 45°.

In 1970, trilateration networks were established in two areas of the Nevada Test Site: Yucca Flat and Pahute Mesa. The geodolite measurements were conducted by USGS personnel in 1972, 1973, and 1983 (SAIC, 1985). The data obtained, for a block 40 km long and 20 km wide and located in Yucca Flat, can be fitted to a uniform strain field. The horizontal principal strains are almost exactly north-south and east-west. The strain rate in a north-south direction is -0.10×10^{-6} per year (shortening), and in an east-west direction the strain rate is $+0.08 \times 10^{-6}$ per year (extension).

An important record of regional tectonic straining was reported by Smith and Kind, 1972. In late 1970 and early 1971, six strainmeters were operated on and near the Nevada Test Site. Plate 3.3.1-6 shows locations of these strainmeters and presents some of the results obtained. Inspection of these results reveals that discontinuities in the strain rate were observed at multiple locations on December 9, January 2, and January 18. The time variation in the rate of strain, throughout the period of observations, is remarkable. Over the two-month period, the peak regional change in strain calculated from the combined data from all strainmeters was 10^{-5}. The authors concluded: "In the two-month interval analyzed, we detected a pulsating strain field that increased over a period of six weeks and decayed over a shorter interval of time. The principal axes show northwest extension, and their directions remain fixed as the magnitude of the field varies."

To account for the deformation observed, at and near the Nevada Test Site, there are few tectonic models that are being considered. Two such models appear to be of particular relevance. The first model requires the presence of one or more subhorizontal detachment faults located at deep crustal levels (Scott, 1985). These detachment faults may be regarded as a crustal expression of either the steady-state shear strains or the upward directed normal stresses from the isostatically uncompensated upper mantle (Section 3.2). In this model, faults observed at the ground surface are listric normal faults extending upward from detachment surfaces. A main feature of this tectonic model is a horizontal extension developed in response to shear displacements occurring along the detachment faults.

The second tectonic model has been proposed by Carr, 1984. According to this model, surface deformation is related to dextral shearing within the Walker Lane. The Nevada Test Site appears to be located where a right stepping in the Walker Lane occurs. This right stepping is associated with a zone of northeast trending faults showing a left-lateral sense of slip. Based on principles of basic wrench tectonics, such faults may be regarded as antithetic strike-slip faults (Wilcox et al., 1972; Moody and Hill, 1956; deBoer,

1967). Commonly, during advanced stages of wrench deformation, these antithetic features inherit some extension aspects of deformation and become nearly vertical, normal faults with negligible lateral displacements. Another important characteristic of wrench deformation is the development of tension fractures and normal faults. The development of such structures occurs parallel to the acute bisectrix of synthetic and antithetic strike-slip faults (Wilcox et al., 1973). It is possible that north-south trending faults, which are very common, at and near the Nevada Test Site, represent this type of structure (Plate 3.3.1-4).

A main feature of the tectonic arrangements, described by the second tectonic model, is a dextral shear couple acting along margins of the Walker Lane. Extension is related to this shear couple and is localized in areas where the right stepping occurs.

It is possible that the most appropriate tectonic model for the area is a hybrid involving the above two models. The observed deformation may be, in part, related to the Walker Lane dextral shear system. This deformation, therefore, is somehow related to lateral plate tectonic interactions occurring along the western margin of the North American continent. Partially, however, the deformation may be reflecting detachment faulting at mid-crustal levels. Such faulting, together with the resulting crustal thinning, may be related to a convective flow of heat and mass in the upper mantle.

For the purposes of these hydrologic considerations, it is not essential to know which one of the above tectonic models provides the best explanation of the observed structural arrangements and kinematic movement plans. What is important is the fact that regional deformation occurs on a continuous basis and that it involves extension. The presence of additional factors, such as compression and rotation about either vertical or horizontal axes, are important from the point of view of structural geology. Such factors may only indicate the presence of local heterogeneities in the stress field. They may, nevertheless, have some regional tectonic significance. However, for the purposes of these conceptual hydrologic considerations, these factors are not of critically essential importance.

This brief summary of deformation of the Yucca Mountain groundwater system during the Plio-Quaternary time indicates that the area is tectonically active. A substantial body of evidence points toward complicated displacement field with surface displacement velocities far exceeding those expected to be accommodated solely by steady-state viscous or visco-elastic flow. Consequently, an important question emerges: What are the hydrologic implications of this circumstance? Subsequent discussions are intended

to provide some answers to this question. The discussions were developed keeping in mind that a) assessment is performed for the purposes of evaluating performance of a high-level nuclear waste repository and b) time scales of geologic proportions are of interest.

3.3.2 The Changing In Situ Stress Field

3.3.2.1 General

Plate 3.3.2.1-1 presents a summary of the contemporary strain rates and the accumulated shear strains for surface deformation of the Yucca Mountain groundwater system and of the regions surrounding it. The strain rates shown are far too large to be accommodated, without an episodic rupture along some local faults, by viscous or visco-elastic flow. Occurrences of earthquakes and the presence of surface ruptures along a large number of local faults fully support this inference. In this situation, it is reasonable to assume that a changing, in time and space, *in situ* stress field is involved. For such a stress field, strain energy (w) contained within it changes with time $(\partial w/\partial t \neq 0)$. The rate of this change, although at most times exceedingly small, is time-dependent $(\partial w/\partial t \neq \text{const.})$, and is variable for different parts of the field. During short intervals of time the rate may become large, as is the case for a seismogenic fault rupture, a fault creep, or for a vibratory ground motion induced fault slip.

A changing, in time and space, *in situ* stress field is a cornerstone of the conceptual considerations undertaken in this report. It will be shown, that such an *in situ* stress field is of critical importance for considerations of suitability of the Yucca Mountain site to accommodate a high-level nuclear waste repository. In this context, it is rather difficult to identify another single technical topic of equal importance. It is essential, therefore, to gain some understanding of a changing *in situ* stress field. Such an understanding may then be used for evaluations of hydrologic and geothermal properties of a deforming fractured medium.

3.3.2.2 The In Situ Stress Field During a Single Cycle of Tectonic Deformation

Before general hydrologic considerations of a deforming fractured medium may be performed, it is necessary to develop a conceptualization of processes involved in a simple tectonic deformation. To this end and for the sake of simplicity, the area of interest is considered as composed of a number of arbitrary strain domains (Plate 3.3.2.2-1). The size of an individual strain domain is measured in tens or hundreds of km^2. It is assumed that boundaries of all domains are, respectively, parallel to each other. It is further assumed that initial boundary tractions for each domain are equal to each other, and equal to the boundary tractions present at margins of the entire deformation area.

Each of the strain domains is assumed to contain a separate fault, which with time, will become a site of a characteristic, for a given domain, earthquake. For each of the domains, the characteristic earthquake occurs at different times, and with a different frequency. This different response time is caused by a different length and orientation assumed for each of the faults. It is further assumed that the dilatancy/fluid diffusion model, as proposed by Scholz et al., 1973, provides a representative description of earthquake nucleation process.

The above idealization of the deformation process is not an accurate description of an actual deformation. For such a deformation, some of the faults are interconnected and the boundary tractions of an individual domain are not only related to the boundary tractions of the entire deformation area, but also they are influenced by developments in the neighboring domains. For the purposes of these conceptual considerations, however, this inaccuracy does not entail a serious conceptual consequence.

It is convenient and realistic to think about tectonic deformations as occurring in repeated cycles. Both the duration and frequency of such cycles are, or may be, different for each individual domain. The characteristic earthquake, with a repeat time $-t_c$, separates one deformation cycle from another.

Plates 3.3.2.2-2 and 3.3.2.2-3 present an idealization of the *in situ* stress field changes occurring in association with a single cycle of deformation in a given strain domain. The cycle can be subdivided into two phases. The first phase begins with the onset of build-up of tectonic strain in response to regional tectonic deformation. The rate of such build-up is very low, but locally it will gradually increase. During this phase, both normal stresses $(\sigma_{xx}$ and $\sigma_{yy})$ and shear stresses $(\tau_{xy}$ and $\tau_{yx})$ show small time-dependence. In the case where shear couple is involved, time-dependence pertains not only to the magnitude of the principal stresses $(\sigma_1$ and $\sigma_3)$ but also involves orientations of these stresses. There is no frictional sliding, therefore, more or less homogeneous strain conditions prevail, and the time-dependent, at a scale of hundreds of meters or so, stress gradients either do not exist or are developed only locally, and represent "remnant" stress concentrations from previous cycles of deformation.

The second phase of deformation begins at a time when, somewhere within the strain domain, a local failure is initiated. Around such a failure, the *in situ* stress field is in so-called conditions of "limit equilibrium." Throughout this report, the phrase "limit equilibrium" is used to denote the *in situ* stress conditions where a) shear stress along some fractures is equal to the frictional resistance of these fractures, this

means that locally the Coulomb-Navier criterion for shear failure is satisfied and b) local shear displacement along some fractures exceeds the peak-shear displacement.

During the second phase of deformation, the strain domain consists of two parts, each with distinct *in situ* stress conditions. The first part extends from the ground surface down to a surface $Z_{(x,y,t)}$, and contains the *in situ* stress field in the conditions of "limit equilibrium." In this part, gravitational loads are relatively small and, consequently, the values of peak-shear displacement are also low; so that the "limit equilibrium" conditions may emerge early in the deformation process. Around yielding fractures the *in situ* stress trajectories (orthogonal lines representing directions of the principal stresses) are curved. Relative to a fixed reference, the positions and curvatures of such trajectories change with time, indicating that the *in situ* stress gradients are time-dependent $\left[\partial^2_{(\sigma_{xx},\tau_{yx})}/\partial t \partial x \neq 0\right]$. Furthermore, at a time when deformation is sufficiently advanced, the *in situ* stress field contains points where the principal stresses have equal values, so-called "isotropic" points, and points where the mean stress has a very small value, so-called "singular" points (Ramsey, 1967; and Section 3.3.3.2).

The second part of the strain domain is situated below the surface $Z_{(x,y,t)}$. Within this part, gravitational loads and the resulting values of peak-shear displacement are greater and, consequently, longer deformation times are required for the onset of "limit equilibrium" *in situ* stress conditions. Generally, the stress gradients are independent of time, although both normal and shear stresses change with time. The time-dependent stress gradients exist only in local areas and represent "remnant" stress concentrations from previous cycles of deformation.

During the second phase, progress of deformation is expressed in two-fold manner. First, there is a time-dependent enlargement of the upper part of the strain domain. Such an enlargement amounts to a downward migration of the surface $Z_{(x,y,t)}$. Second, above the surface $Z_{(x,y,t)}$ the *in situ* stress field undergoes spontaneous and induced, by some external factors, restructurings occurring on either continuous or episodic basis. These restructurings involve a) spatial migration of "singular" and "isotropic" points and b) increasing density of occurrence of such points.

Toward the end of the second phase of deformation, another important factor comes into play. This factor is a dilatant zone of the characteristic earthquake. Plate 3.3.2.2-4 shows diagrammatically a sequence of events associated with the dilatant zone. The rising tectonic stress causes the region around the

focus of the subsequent characteristic earthquake to dilate by opening cracks and fractures normal to the least principal stress (σ_3). The focus is situated at a depth substantially greater than depth of the surface $Z_{(x,y,t)}$, say at a depth ranging from 10 to 15 km. The increase in fracture porosity causes the fluid pressure to decrease inducing inward migration of fluids from the surrounding areas into the dilatant zone. The drop in fluid pressure causes a temporary rise in frictional resistance of the fault. As the migrating fluids fill the cracks, however, the fluid pressure rises again and, consequently, frictional resistance of the fault decreases. Seismic failure eventually occurs and is accompanied by a collapse of the dilatant zone and a partial relief of the shear stress. This allows the cracks and fractures to relax, and the fluids they contain must be expelled rapidly upwards in the direction of easiest pressure relief. Upflow of fluids, from the collapsing dilatant zone, takes place through the fault and adjacent fractures. As the source region of the characteristic earthquake returns to its pre-dilatant state, the rates of fluid upflow decrease to zero.

Intermittent outflows of fluids, occurring in association with seismogenic fault ruptures, is known as the process of "seismic pumping" (Sibson et al., 1975; and McCaig, 1988). Quantities of fluids involved in such outflows may be considerable, in a range of 10^8 to 10^{10} liters. Significant mineralization may, therefore, be produced by such seismically induced fluid circulations.

Based on the above description of the characteristic earthquake nucleation process, it can be inferred, that at the end of the second phase, the lower part of the strain domain below the surface $Z_{(x,y,t)}$, consists of two areas. The first area corresponds to the dilatant zone, and contains in its interior the focus of the subsequent characteristic earthquake. Within this area, the time-dependent stress gradients are developed. The second area is situated outside the dilatant zone and below the surface $Z_{(x,y,t)}$. In this area, at a time when the volume representing the dilatant zone is very small, the stress gradients are essentially independent of time, although both normal and shear stresses change with time.

It can be expected that volume of the dilatant zone changes continuously. Initially, it increases until a certain characteristic for a strain domain, value is reached. This time-dependent expansion of the dilatant zone causes straining of the outside area, and may cause emergence of the "limit equilibrium" *in situ* stress conditions in this area. During terminal stages, interaction between the outside area and the dilatant zone is reversed. Stretching of the outside area reached the characteristic value. The outside area begins to expand, thus causing contraction of the dilatant zone.

Collapse of the dilatant zone is associated with releases of large quantities of strain energy stored within the strain domain, and with outflows of fluids from the dilatant zone. At this time, the *in situ* stress field begins to return to its initial state, and the strain domain enters another cycle of tectonic deformation. The process is repeated over and over again, as long as regional tectonic forces motivating these developments remain active.

Association of the dilatant zones with large seismic failures is a well known geophysical phenomenon (Nur, 1972; Frank, 1965; Scholz and Kranz, 1974; Scholz et al., 1973; Whitcomb et al., 1974; Aggarwal et al., 1973; Sadovsky et al., 1973; Semenov, 1969). The size of such zones, in terms of the dilatant volume (ΔV) and the dilatant strain $(\Delta V/V)$, may be very substantial. Sibson et al., 1975, for example, have shown that the dilatant zone for the Matsushiro earthquake swarm, which energetically was equivalent to a single event with magnitude M=6.3, had the dilatant volume of about $5x10^6 m^3$ and the dilatant strain of about 10^{-5}.

It is important to develop an understanding of a sequence of events that is triggered by the characteristic earthquake in the upper part of the strain domain, above the surface $Z_{(x,y,t)}$. To this end, it is assumed that this part consists of a stack of horizontal layers with small thickness (ΔZ) (Plate 3.3.2.2-5). Each of the layers consists of a number of blocks with small unit length (L). The blocks are separated from each other by a fracture. The mechanical properties of the system are such that the stress-strain relationship involves a constant E_1, for the blocks, and another constant E_s for fractures. Initial confining stress for the entire stack consists of a) horizontal component of the lithostatic pressure and b) small compressive stress, representing a "remnant" stress from the previous cycles of deformation. The stack is subjected to uniform extension operating with the rate $\dot{\epsilon}_{xx}$. As extension continues, the stack is transformed from the initial undeformed state into the deformed or dilated state (Plate 3.3.2.2-5). Above the surface $Z_{(x,y,t)}$, the "limit equilibrium" stress conditions are locally present and, consequently, some fractures are yielding. Below the surface $Z_{(x,y,t)}$, however, the cumulative effect of extension, on the lateral confining stress, is smaller; fractures do not yield, and the "limit equilibrium" *in situ* stress conditions are not present.

It is assumed that the characteristic earthquake, or collapse of the dilatant zone, is accompanied by a large-scale transfer of fluids via the mechanism of "seismic pumping." Such a transfer occurs along the ruptured fault and along a few adjacent fractures. A substantial volume of fluid is involved and, therefore,

the fluid outflow reaches the upper part of the strain domain, above the surface $Z_{(x,y,t)}$, where the "limit equilibrium" *in situ* stress conditions are developed.

Plate 3.3.2.2-6 presents an idealization of the response of the upper part of the strain domain to a sudden build-up of pore pressure. Two cases are considered. The first case is a uniform rise of fluid pressure. Response of the system to such a rise is a more or less uniform increase in aperture of fractures. This occurs because increased fluid pressure facilitates "contraction" of the stretched individual blocks. The deformational system, however, remains in the dilated state. Consequently, the *in situ* stress field, although somewhat restructured, retains its "limit equilibrium" status.

The second case involves a local build-up of fluid pressure occurring along a few widely spaced fractures and faults. It is assumed that the amount of fluid involved is large, and that the build-up occurs very fast. The deformational system responds to such a build-up by developing large-scale separations of wallrock. Such separations are localized at fractures along which the build-up of fluid pressure took place and occur at the expense of other areas that were not involved in the fluid pressure build-up. In such areas, spacing between individual blocks is reduced, and consequently, the degree of dilation is also reduced.

There are two factors that control size of the separation. The first factor is the amount of "recoverable" extension contained between any two fractures that were involved in the fluid pressure build-up. For example, if such fractures are spaced 10^3 m apart, and the extension was accumulating at a rate of 10^{-7} per year during 10^4 years, then the potential separation is 1 m for each of these fractures.

The second factor is the volume of fluid involved in the fluid pressure build-up for each individual fracture. As the wallrock separation increases, the volume representing the space between fracture walls increases rapidly. For the pressure build-up to be sustained, this rapidly increasing space must be always filled up with fluid. Soon after the separation begins, the volume representing space between the separated walls will match the volume of fluid involved in the transfer. At this time, further build-up of fluid pressure is arrested, and growth of the separation may be terminated.

It can also be expected that a) the transfer of fluids from depth will involve fluids that are mineralized and b) the transfer will be followed by a convective flow of heat and fluid. In this situation, precipitation of minerals along the separated fractures is to be expected. This process will cause a gradual reduc-

tion in apertures of separations, eventually leading to their complete closure and "fusing" back together with the fractured medium. At this time, the "limit equilibrium" *in situ* stress conditions are terminated, the upper part of the strain domain is "fused" together, and the deformational system is ready to enter another cycle of tectonic deformation.

At this point in the analyses, it is anticipated that some readers may have considerable difficulty in intuitively accepting both the existence and the postulated mechanism of formation of the large-scale separations of wallrocks. Prior to rejecting off-hand the phenomenon, however, the dubious reader is urged to consult geology of the Yucca Mountain area, rather than his or her anthropomorphic concepts of near-surface deformational processes. Some intuitive appreciation of the wallrock separation phenomenon may be acquired by examining a few natural outcrops located in the vicinity of the Yucca Mountain site. It is expected that, to this end, the most beneficial areas may be the following three locations:

- South shoulder of Highway 95; approximately five miles east of Lathrop Wells, just east of Skeleton Hills;

- Devil's Hole in Ash Meadows, Nevada. At this location, the following three lines of reasoning strongly suggest that the Devil's Hole is not an ordinary karst feature: a) karst is not known in the region; b) the planar opening extends deep below the base of the local groundwater system; and c) the opening occurs in discharge areas of the carbonate aquifer where there are abundant indications of persistent deposition, but not dissolution, of calcium carbonate.

- Trace of the Furnace Creek fault, between Death Valley Junction and Ubehebe Craters, California. Specific recommended stops are a) in the vicinity of Navel Spring; b) the west side of Titus Canyon; and c) the northwestern sector of the Grapevine Springs area.

The large-scale wallrock separations, specifically ones that a) are formed in response to a tectonic event; b) involve high strain rate; c) involve, in a passive or active manner, subsurface fluids; and d) exhibit, individual or combined, sizes that are measured in terms of at least tens of centimeters, are essential with respect to the conceptual considerations undertaken in this report. Without such separations, these considerations are null and void. Short of either reserving judgments or bridging the gap between our intuitive visualization of the near-surface deformational processes and interpretative necessitites, that are

dictated by the direct field observations, it is rather difficult, if not impossible, to accept the results of all subsequent analyses. It should be understood, however, that the fact that wallrock separations, with the specified above characteristics are present is, by far, much more important than the actual circumstances of their formation.

3.3.2.3 The In Situ Stress Field in a Cyclically Deforming Fractured Medium

Based on the above analyses, it is possible to develop a conceptual understanding of the *in situ* stress field changes that occur during repeated cycles of tectonic deformation. Such understanding is of critical importance for subsequent hydrologic considerations of a deforming fractured medium. For the purposes of this report, of main interest are the *in situ* stress changes taking place in the upper part of the strain domain, at a depth smaller than 2 or 3 km. In this part, which is above the surface $Z_{(x,y,t)}$, the characteristic earthquake acts merely as a catalyst responsible for resetting the *in situ* stress conditions. To be effective in such a role, the characteristic earthquake must produce only large-scale separations of wallrock. It is not of particular importance, therefore, to know a lot about this earthquake, including a) exact magnitude and focal depth; b) slip mechanism; c) stress drop and seismic moment; and d) details of physics of the nucleation process.

For the purposes of this report, it is assumed that wallrock separations are accomplished via the mechanism of "seismic pumping." This assumption, however, is not of critical importance. If wallrock separations are produced through some other means, their effect on the *in situ* stress field above the surface $Z_{(x,y,t)}$ will not be different.

Plate 3.3.2.3-1 is a schematic representation of individual strain domain at a time of nucleation of the characteristic earthquake. The strain domain consists of three parts: a) the upper part, above the surface $Z_{(x,y,t)}$; b) the lower part; and c) the dilatant zone. For the sake of simplicity, it is assumed that the strain domain is subjected only to uniform regional extension operating with a constant rate $\dot{\epsilon}_{xx}$. The deformation process is advanced and, consequently, the surface $Z_{(x,y,t)}$ is situated at a considerable depth (e.g., a depth ranging from 2 to 3 km).

The total width of strain domain $\left[L_{d_{(t)}}\right]$ was acquired progressively during n cycles of tectonic deformation $\left[L_{d_{(t)}} = L_d^{in} + \dot{\epsilon}_{xx} \cdot t, \text{ and } t = n \cdot t_c\right]$. At any given time, the width of the domain is the sum of fault width $\left[L_{f_{(t)}}\right]$ and width of the wallrock $\left[L_{w_{(t)}}\right]$; so that, $L_{d_{(t)}} = L_{f_{(t)}} + 2L_{w_{(t)}}$.

Plate 3.3.2.3-2 presents an idealization of the time-dependent "growth" of the width of the strain domain. In-between the characteristic earthquakes, the accumulation of tectonic strain is uniform throughout the entire domain. At a time of occurrence of the characteristic earthquake, however, the accumulated extensional strain $\left(\dot{\epsilon}_{xx} \cdot t_c\right)$ is concentrated along the separated fault and adjacent fractures. The combined thickness of wallrock fluctuates between a minimum L_w^{in} at the onset of a tectonic cycle and a maximum $L_w^{max} = L_w^{in} + \dot{\epsilon}_{xx} \cdot t_c$ at the end of a cycle. The combined thickness of the fault, however, increases in a step-like fashion. The individual increments have a maximum value of $\Delta L_f = \Delta L_w^{max} = \dot{\epsilon}_{xx} \cdot t_c$. The sum of such increments is equal to the total increase of the width of the strain domain, that was acquired during time t and through n cycles of deformation.

Plate 3.3.2.3-3 presents an idealization of the *in situ* stress changes occurring in a cyclically deforming fractured medium. This history is shown in the form of a stress vs. time plot that would result from monitoring the *in situ* stress changes, at a single reference point $P_{(x,y,z)}$, throughout duration of three sequential cycles of tectonic deformation. It is assumed, that at the end of each cycle, the reference point is situated above the surface $Z_{(x,y,t)}$. The *in situ* stress is represented by the Mohr circle, and the Coulomb-Navier envelope of failure represents fractured medium. Relative to the envelope of failure, the Mohr circle fluctuates between two positions. Initially, when the reference point is situated above the surface $Z_{(x,y,t)}$, the Mohr circle is positioned away from the failure envelope and, consequently, the shear displacement $\left(u_s\right)$ is less than the peak-shear displacement $\left(u_{sp}\right)$ (Plate 3.3.2.3-3). As the surface $Z_{(x,y,t)}$ migrates past the reference point, the stress circle touches the failure envelope and the "limit equilibrium" *in situ* stress conditions are established. The shear stress (τ) is equal to the frictional resistance of some fractures $\left[\tau_{\max} = C + \sigma_n \cdot \tan\phi\right]$, and such fractures begin to yield. From this time until the end of a cycle, the value of shear stress $\left(\tau\right)$ is controlled by the frictional resistance $\left(\tau_{\max}\right)$.

With the occurrence of a characteristic earthquake, the "limit equilibrium" *in situ* stress conditions are terminated, and the stress circle moves away from the failure envelope. Reduction in the amount of accumulated extensional strain in the wallrock (resulting from wallrock separation) causes an increase in the mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3)$, and the resulting change in position of the stress circle. With the onset of

subsequent cycle of deformation, the migratory movement of the stress circle toward the failure envelope is resumed. The process repeats itself as long as deformation continues.

At this point, an important question emerges: **How realistic is the above understanding of cyclic deformation and the resulting temporal fluctuations of the *in situ* stress, specifically for the Yucca Mountain area**? Some answers to this question may be provided by direct field observations made during a traverse across the area. In making such a traverse, an observer would encounter five sizeable and youthful faults (Plates 3.3.2.3-4 and 3.3.2.3-5). These faults are a) the Fran Ridge fault; b) the Paintbrush fault; c) the Bow Ridge fault; d) the Solitario Canyon fault; and f) the Windy Wash fault. The observer would also notice the presence of two types of thought provoking materials. These materials occur along traces of each of the local faults and consist of a) "mosaic" breccias and b) veins and aprons composed of opaline silica, calcite, with minor sepiolite. The combined thickness of these materials, relative to the total length of the traverse, is considerable (Plate 3.3.2.3-5).

The "mosaic" breccias consist of angular fragments of the country rock embedded in a matrix composed mainly of opaline silica and calcite. Remarkable characteristics of these breccias are a) "matrix" support texture (individual clasts are not in direct contact with each other); b) very large volumetric strain, ranging from several percent to tens of percent; and c) very isotropic character of the volumetric strain (the amount of dilation is similar in all three mutually perpendicular directions). All of these characteristics indicate that the "mosaic" breccias are either "fragmentation" breccias or "explosive" breccias. The term "explosive" breccia is employed here to denote a breccia for which formation of the clasts and the matrix is the result of a common process and, in the case of clasts, requires large and fast build-ups of fluid pressure (e.g., hydrothermal eruptions, wallrock fragmentation of an active geothermal system by vibratory ground motion, and fragmentation of host rock caused by very rapid increases of the mean effective stress followed by the geothermal circulation of fluids). In contrast, the term "fragmentation" breccia is used here to denote a breccia for which formation of the clasts and the matrix is the result of two independent processes occurring at the same time and place (e.g., mass wasting and filling-up of a bedrock opening within which minerals are precipitating from upwelling deep-seated fluids; and rapid cementation of a mass wasting fill of a topographic depression, whether structural or erosional, by mineral rich fluids discharging from nearby faults and fissures). Hanson et al., 1987, on the basis of field inspection, diagnosed the Yucca Mountain breccias as "**hydrothermal eruption breccias.**"

Another important characteristic of the "mosaic" breccias is their age. Although, little is known about this topic there are, however, some bases for constraining the problem. First, the "mosaic" breccias contain fragments of mineral products from the vapor phase mineralization and from devitrification of the country rock (Hanson et al., 1987). This suggests that the formation of "mosaic" breccias occurred after deposition of the country rock. This inference is supported by two other observations: a) "mosaic" breccias do not appear to be "stratum (ash sheet) bound" and b) occurrences of these breccias are not restricted solely to pyroclastic materials, they also occur in carbonates of Paleozoic age. Second, the "mosaic" breccias commonly form wallrocks of the calcite-silica veins. Some of these veins extend from the bedrock into alluvium of Late Quaternary age.

Taken together, the above two cross-cutting relationships constrain the age of formation of the "mosaic" breccias as between Late Miocene to Late Quaternary time. Recognizing the considerable thickness of these breccias and their age, relative to the time of solidification of the country rock, an additional inference may be drawn. This inference is that the "mosaic" breccias are polygenetic, and that they represent a considerable time span.

The calcite-silica-sepiolite veins are emplaced along fault planes in both the Quaternary alluvium and the Late Miocene bedrock. Thickness of individual veins ranges from fractions of a centimeter to a maximum of about 1 m. Combined thickness of the veins is in a range of approximately 10 m. Typically, the veins exhibit banded textures, indicating the involvement of fluids with alternating geochemical characteristics, either during multiple events or during a single event. The veins emplaced along the Bow Ridge fault and the Windy Wash fault contain thin stringers of uncemented basaltic ash that displays petrographic and chemical affinity with one of two volcanic centers situated nearby (Whitney et al., 1986). The age of these centers has been determined, based on radiometric dating of the associated lavas, as ranging from 1.2×10^6 year B.P. to possibly less than 10^5 year B.P. (Vaniman et al., 1982; and Crowe and Vaniman, 1985).

Based on the above direct field observations, it seems reasonable to conclude that a considerable "widening" of the Yucca Mountain area has occurred during the Late Miocene - Late Quaternary time span. Assuming that a) combined thickness of the "mosaic" breccias and the calcite-silica veins is 5 x 500 m = 2500 m (Plate 3.3.2.3-5) and b) average volumetric strain represented by the "mosaic" breccias is $\Delta V/V = 10\%$; total "widening" of the Yucca Mountain area is equal to about $\Delta L_d = 250$ m. The "widen-

ing" seems to be concentrated along traces of the local faults, suggesting that such "widening" may have occurred in a step-like fashion, as represented schematically on Plate 3.3.2.3-2. It is reasonable, therefore, to assume that deformation of the Yucca Mountain area occurs as a cyclic process, and that such a process includes episodic separations of wallrock. It is also reasonable to further assume that such separations, together with the ongoing extension, causes the *in situ* stress to undergo temporal fluctuations (Plate 3.3.2.3-3).

3.3.2.4 Summary and Conclusions

The analyses presented above allow for the following four conclusions to be drawn. First, the *in situ* stress field developed in a deforming fractured medium is characteristically variable in time and space. In surficial parts of a deformation area, the "limit equilibrium" *in situ* stress conditions are occasionally present. The *in situ* stress field is characterized by four features: a) stress gradients are time-dependent; b) "singular" and "isotropic" points are present; c) density of occurrence of "singular" and "isotropic" points varies in time and space; and d) depth extent of the "limit equilibrium" *in situ* stress conditions varies in time and space.

Second, in a continuously deforming fractured medium, the *in situ* stress field undergoes cyclic changes from "no limit equilibrium" at the onset of a tectonic cycle to "limit equilibrium" at more advanced stages of such a cycle. The characteristic earthquake acts as a "catalyst" resetting the *in situ* stress conditions. To be effective, the characteristic earthquake must either produce or be accompanied by large-scale separations of wallrock. This separation, but not the characteristic earthquake, is the essential factor in the cyclic deformation of a fractured medium.

Third, for the purposes of hydrologic considerations of a deforming fractured medium, it is not of critical importance to know a lot about a seismo-tectonic setting of a groundwater system. It is sufficient to know that a) active faults are present; b) deformation occurs on more or less continuous basis and that it involves tectonic extension; and c) separations of wallrock are occasionally produced by the characteristic earthquake or by some other factor. Seismo-tectonic details, such as a) exact magnitude and focal depth of the characteristic earthquake; b) seismo-tectonic zonation; c) slip mechanism; d) stress drops and seismic moments; e) physics of the nucleation process; f) overall plate-tectonic setting; however, are not of critical importance. Much more important is recognizing that a continuous tectonic extension will ulti-

mately cause a) emergence of the "limit equilibrium" *in situ* stress conditions and b) cyclic changes in the *in situ* stress field. Such recognition, however, is a matter of common sense considerations, that require neither extensive instrumental seismology studies nor extensive seismo-tectonic efforts.

Fourth, in the Yucca Mountain groundwater system, flow occurs in a deforming fractured medium. The results of various measurements and observations indicate that this conclusion is secure beyond a reasonable doubt. Furthermore, the geologic record as expressed in surficial rocks of the Yucca Mountain area may be interpreted as indicating involvement of a cyclically deforming fractured medium. This means that the *in situ* stress field in this area may be undergoing cyclic changes from "no limit equilibrium" conditions at the onset of a tectonic cycle to "limit equilibrium" at the advanced stages of such cycle.

3.3.3 Hydrologic Importance of a Changing In Situ Stress Field

3.3.3.1 Introduction

The main factor that relates hydraulic properties of a deforming fractured medium and a changing *in situ* stress field, as considered in this report, is conducting aperture of fractures. In a situation where a) fracture shear stress (τ_f) reaches a time-dependent limit set by the fracture frictional resistance [$\tau_{\max_{(t)}} = C + \sigma_{nfeff_{(t)}} \cdot \tan\phi$] and b) introduction of extensional strain continues, such a relationship is of fundamental importance. The main purpose of subsequent analyses is to provide justification for this statement. The analyses are presented in three parts. The first part, Section 3.3.3.2, is concerned with the state of *in situ* stress around a deforming (yielding) fracture. The second part, Section 3.3.3.3, deals with the hydraulic conductivity structure in a deforming fractured medium; mainly, with its time-dependent and cyclic character. The third part, Section 3.3.3.4, presents summary and conclusions.

3.3.3.2 The In Situ Stress Field Around a Deforming Fracture

It is assumed here that a fractured medium fails in a brittle or semi-brittle manner according to the combined Griffith-Navier-Coulomb envelope of failure. Such envelope is represented by curve ABC on Plate 3.3.3.2-1. If the stress circle touches the failure envelope between B and C a shear failure results. If the stress circle touches the failure envelope between A and B a hybrid extension-shear failure develops. Only if the stress circle touches the envelope at point A will pure tensile failure result. For tensile failure, the Griffith criterion requires that the stress difference ($\sigma_1 - \sigma_3$) is smaller than 4T; and for shear and hybrid extension-shear failure the stress difference ($\sigma_1 - \sigma_3$) must be greater than 4T. The failure criterion for weakest parts of a fractured medium is given by line OX, which represents the failure envelope for material with zero cohesion (pre-existing fractures).

The principle of effective stress states, that at failure, the stress circle represents two components: a) bedrock stress ($\sigma_1 - \sigma_3$) and b) fluid pressure (p). The effective mean stress is equal to $\frac{1}{2}(\sigma_1 + \sigma_3) - p$. Failure of a fractured medium may be introduced by increasing either the stress difference ($\sigma_1 - \sigma_3$) or fluid pressure (p).

Keeping in mind these elementary remarks, it is possible to develop an understanding of the state of *in situ* stress around a deforming fracture. Such an understanding is of critical importance for subsequent hydrologic considerations of a deforming fractured medium.

Plate 3.3.3.2-2 presents the assumed initial distribution of the *in situ* stress along a fracture, prior to local failure. The normal stress (σ_{nf}) increases linearly with depth in accordance with the increasing lithostatic load. This causes the fracture frictional resistance [$\tau_{\max} = C + \sigma_{nf} \cdot \tan\phi$] to also increase with depth. Location of the fracture, relative to the σ_1 *in situ* stress trajectory, is such that gradients of both the normal stress (σ_{nf}) and the shear stress (τ_f) are present along the fracture strike. The curvature of stress trajectory is caused either by a pocket of "remnant" stress from previous cycles of deformation or by local deformation of an adjacent "weaker" fracture.

It is further assumed that the overall *in situ* stress field changes in such a manner that the stress difference ($\tau_{\max} - \tau_f$) decays with increasing time. The increasing with time shear stress (τ_f), the decreasing normal stress (σ_{nf}), or both, are responsible for that. The rate of decay is assumed to be constant in time and space.

Plate 3.3.3.2-3 presents an idealization of changes in stresses and displacements, occurring along the depth of a deforming fracture, as a function of time. With increasing time, the shear stress (τ_f), relative to the frictional resistance ($\tau_{\max}$), increases. Relative to the peak-shear displacement (u_{sp}), the shear displacement (u_s) also increases. If, at a given depth, the shear stress (τ_f) reaches the frictional resistance ($\tau_{\max}$) then the shear displacement (u_s) becomes equal to the peak-shear displacement (u_{sp}). Locally, the "limit equilibrium" stress conditions are established; the fracture begins to yield and further increases in the shear stress are not possible. The depth to which the shear stress is at the limit $\left(\text{depth above which } \tau_{\max} - \tau_f = 0\right)$ increases with increasing time. The post-peak shear displacement $\left(u_s - u_{sp}\right)$ and the vertical extent of "an-elastic" deformation also increase with time.

Plate 3.3.3.2-4 presents an idealization of changes in stresses and displacements, occurring along the strike of a deforming fracture, as a function of time. As the deformation progresses, the stress difference $\left(\tau_{\max} - \tau_f\right)$ decreases. This decrease is caused by the increasing τ_f, by the decreasing σ_{nf}, or both. Eventually, the stress difference $\left(\tau_{\max} - \tau_f\right)$ becomes equal to zero, and "an-elastic" deformation of the fracture

is initiated. From this time on, the post-peak shear displacement $\left(u_s - u_{sp}\right)$ and the lateral extent of "an-elastic" deformation increase with increasing time.

Reacting to the variable magnitude of shear displacement $\left[u_{s_{(y)}}\right]$, fracture wallrock is compressed in the upper half and stretched in the lower half, and vice versa across the fracture plane. Such a differential straining causes further changes in the *in situ* stress in wallrocks of a deforming fracture. In the upper half, there is an increase in the mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3)$ and a decrease in the stress difference $\left(\sigma_1 - \sigma_3\right)$. In the lower half, however, the response is different; both the mean stress and the stress difference decrease with time. Eventually, in both walls of a deforming fracture, "isotropic" and "singular" points will emerge.

The analyses of the *in situ* stress field changes occurring around a deforming fracture indicate, that above the surface $Z_{(x,y,t)}$, four different stress states are present (Plate 3.3.3.2-5). Outside the zone of "an-elastic" deformation, the mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3)$ and the stress difference $\left(\sigma_1 - \sigma_3\right)$ are such that the stress circle is situated away from the failure envelope. Within the zone of "an-elastic" deformation, however, the stress circle touches the failure envelope, indicating that the "limit equilibrium" stress conditions are present. The remaining two stress states occur near edges of the zone of "an-elastic" deformation. For both of these states, the stress difference $\left(\sigma_1 - \sigma_3\right)$ is small and allows for satisfying the Griffith criterion for tensile failure, provided there is sufficient reduction in the mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3)$. Around "singular" points, either in existence or developing, the required reduction in the mean stress is very small. Around "isotropic points," however, the required reduction is substantial.

The progress of deformation is marked by downward migration of the surface $Z_{(x,y,t)}$. Fractures at a progressively greater depth are subjected to the "limit equilibrium" stress conditions. At a constant depth, the post-peak shear displacements $\left(u_s - u_{sp}\right)$ attain progressively greater values, and an increasingly greater number of fractures becomes involved in yielding. This means, of course, that the downward migration of the surface $Z_{(x,y,t)}$ is accompanied by a) an increase in the density of occurrence of "singular" and "isotropic" points and b) an increase in the maximum depth of development of such points.

3.3.3.3 The Hydraulic Conductivity Structure in a Deforming Fractured Medium

Measurements have shown that the hydraulic conductivity of fractures is sensitive to confining pressure. This sensitivity has been attributed to changes of conducting aperture caused by changes in the nor-

mal effective stress (Walsh, 1981). To account for such effects, various relationships have been proposed (Barton et al., 1985). For the purposes of these conceptual considerations, a simplified but realistic model has been adopted (Plate 3.3.3.3-1). The conducting aperture, $a_{con} = a_{res} + a_n$, depends linearly on confining stress (through compliance K_a) until the critical closure pressure (σ_{cc}) is reached and then only residual aperture remains. If shear dilation occurs, however, the basic relationship is assumed to be modified as shown by the dashed line. The conducting aperture, $a_{con} = a_{res} + a_d$, depends on the post-peak shear displacement ($u_s - u_{sp}$) through compliance K_s.

Both hydraulic properties of a fractured medium: hydraulic conductivity and storativity, are directly related to conducting aperture. Adopting the analogy of planar parallel plates to represent walls of a fracture, and assuming the Newtonian viscous fluid with dynamic viscosity μ, and density ρ, the hydraulic conductivity of a fracture is

$$K_f = a_{con}^2 \cdot \frac{\rho \cdot g}{12\mu}$$

or,

$$K_f = a_{con}^2 \cdot \text{const.}$$

The storativity of a fracture per unit area is

$$S_f \cong a_{con}$$

Plate 3.3.3.3-2 presents a relationship between the conducting fracture aperture and the effective normal stress, assuming that the shear stress along a fracture is equal to zero. This plate also shows the resulting time-dependence for a) fracture conducting aperture; b) fracture storativity; and c) depth above which both of these properties are time-dependent [depth to the surface $Z_{(x,y,t)}$]. The relationships shown were developed assuming that a) there is a limit for normal fracture dilation and b) the reduction in initial confining stress is continuous. At a time when σ_{nfeff} is reduced to a point where it is equal to the critical closure pressure (σ_{cc}), both the hydraulic conductivity and storativity begin to show time-dependence. The time vs. hydraulic conductivity relationship is non-linear; it is linear for storativity. Time-dependence of depth to the surface $Z_{(x,y,t)}$ is caused by the increasing with depth value of the initial confining stress.

Assuming that the rate of decay of σ_{nfeff} is const., more deformation time is needed to accumulate the required stress reduction at a greater depth.

Plate 3.3.3.3-3 shows an idealized joint response during shear dislocation for selected values of constant normal effective stress. This plate also shows the relationship between shear displacement (u_s) and the shear dilation component of the conducting aperture (a_d). The relationships shown were developed assuming that a) there is a limit to the magnitude of a_d and b) compliance K_s is greater than zero. If, at a given depth, the shear stress reaches the frictional resistance, then the component a_d becomes greater than zero and related to the value of post-peak shear displacement.

Keeping in mind the assumed time-dependence of the shear stress (τ_f), the diagram shown on Plate 3.3.3.3-3 may be used to develop a relationship between time and the shear component of conducting aperture (a_d). Such a relationship is shown on Plate 3.3.3.3-4. The resulting time-dependence of both hydraulic properties is also shown. Increasing τ_f causes a) storativity; b) hydraulic conductivity; and c) depth to the surface $Z_{(x,y,t)}$, all to increase with time. The relationship between shear dispacement (u_s) and depth is important. It shows that the shear displacement, and with it the shear component a_d, attain greatest magnitudes at the ground surface. With time, both u_s and a_d increase until limit U_{ad} is reached.

Both the limit U_{ad} and compliance K_s express an "ability" of fracture walls to move away from the fracture plane in response to shear dislocation. Without getting involved in complex analyses, it can be inferred that such an "ability" is related to two time-dependent factors: a) the effective normal stress (σ_{nfeff}) and b) the lateral extent of the zone of "an-elastic" deformation. The significance of both of these factors is analogous to factors (body weight and slenderness ratio) that control the critical stress for buckling deflections, as given by the modified Euler's equation. In a deforming fractured medium, both the effective normal stress and the lateral extent of "an-elastic" deformation are related to a degree of advancement of deformation. Consequently, it is reasonable to assume that the "ability" of fracture walls to bend and, locally, move away from the fracture plane increases with the progress of deformation. Initially, when the post-peak shear displacements are very small, values of the limit U_{ad} and compliance K_s are also very small. As more and more extensional strain is introduced into a deforming fractured medium, however, both of these values increase and, eventually, may become very large.

Plates 3.3.3.3-5 and 3.3.3.3-6 are schematic representations of time-dependent enhancement of the hydraulic properties for a segment of deforming fracture. Initially, when the shear stress (τ_f) is smaller than the frictional resistance $\left(\tau_{\max}\right)$, conducting aperture is const. and equal to residual aperture $\left(a_{con} = a_{res} = const.\right)$. When the stress circle touches the failure envelope, and later, conducting aperture becomes time-dependent, its value is $a_{con_{(t)}} = a_{res} + a_{d_{(t)}}$. The values of hydraulic conductivity and storativity increase accordingly. As deformation continues, there is a local decrease in the mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3)$ and in the stress difference $(\sigma_1 - \sigma_3)$. When the effective value of the mean stress becomes very small, the Griffith criterion for tensile failure is satisfied. At this time, conducting aperture is equal to the sum of three components, and its value is $a_{con_{(t)}} = a_{res} + a_{d_{(t)}} + a_{n_{(t)}}$. The values of hydraulic conductivity and storativity are dramatically increased.

Plate 3.3.3.3-6 presents a relationship between time and vertical extent of a segment of deforming fracture that undergoes enhancement of the conducting aperture. The relationship is caused by the increasing with depth frictional resistance of fractures. Assuming that the rate of decrease of the stress difference $\left(\tau_{\max} - \tau_f\right)$ is const., more deformation time is required for initiation of shear failure at a greater depth. Plate 3.3.3.3-6 also presents relationships between time, space, and total enhancement of conducting aperture $\left[a_{d_{(t)}} + a_{n_{(t)}}\right]$. The total enhancement increases with time. The greatest magnitudes occur at the ground surface where frictional resistance of fractures $\left(\tau_{\max}\right)$ is least. With depth, the magnitudes of total enhancement diminish, and at the surface $Z_{(x,y,t)}$ become equal to zero.

Plate 3.3.3.3-7 presents an idealized history of changes of the hydraulic conductivity structure for a cyclically deforming fractured medium. This history represents a summary of the above discussions and is shown in the form of a plot which would result from a simultaneous monitoring of the *in situ* stress and the hydraulic conductivity, at a single reference point $P_{(x,y,z)}$, during three sequential cycles of tectonic deformation. For each of the cycles, starting at the time of onset of the "limit equilibrium" stress conditions, there is a progressive enhancement of the hydraulic conductivity structure. At more advanced stages of deformation, "isotropic" and "singular" points develop around the reference point. This allows for the occurrence of Griffith failure around this point. From this time until the end of a cycle, there is an ongoing interaction between the fluid pressure $\left(p\right)$ and the mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3)$. Such an interaction is represented schematically as back and forth reversals in the value of hydraulic conductivity. Conducting aperture is controlled by the value of effective mean stress $\frac{1}{2}\ (\sigma_1 + \sigma_3) - p$, which in turn, is controlled by the combined Griffith-Coulomb-Navier failure envelope. Either tensile or hybrid shear-tensile

opening of fracture causes a drop in fluid pressure (p), and produces a value of the mean effective stress $\frac{1}{2}$ $(\sigma_1 + \sigma_3) - p$ that can be tolerated by the failure envelope. Occasionally, the "singular" point changes its position. Around the former location of the "singular point", the mean stress $\frac{1}{2}$ $(\sigma_1 + \sigma_3)$ increases; the fracture closes and the fluid pressure rises. The overall system is delicately balanced and coupled via the principle of effective stress. This means that the fluid pressure controls the *in situ* bedrock stress and vice versa. A change in one of the factors requires an adjustment in the other factor. Losses of energy from the system are compensated by a) continuous tectonic straining and b) inflow of fluids from surrounding regions. The system remains coupled until the end of a tectonic cycle. At this time, wallrock separation causes a substantial increase in the mean stress $\frac{1}{2}$ $(\sigma_1 + \sigma_3)$, and abrupt termination of the "limit equilibrium" stress conditions. Conducting apertures of fractures diminish to their residual values and, consequently, the hydraulic conductivity structure of a deforming fractured medium "returns" to its initial state, at the beginning of a cycle. Cyclic enhancements of the hydraulic conductivity structure occur over and over again providing, of course, that deformation continues.

3.3.3.4 Summary and Conclusions

The analyses presented above indicate that a changing *in situ* stress field is clearly of fundamental hydrologic importance. Such a stress field causes the hydraulic conductivity structure of a deforming fractured medium to be a) time-dependent and b) at times, substantially enhanced. For a cyclically deforming fractured medium, the hydraulic conductivity structure undergoes cyclic changes. At the onset of a tectonic cycle, the values of hydraulic properties (conductivity and storativity) are relatively low and independent of time. As deformation progresses, however, the "limit equilibrium" *in situ* stress conditions are established. From this time until the end of a tectonic cycle, time- and space-dependent enhancements of the hydraulic conductivity structure occur. Both the degree and spatial extent of such enhancements increase with time. They attain a maximum value at the ground surface (degree) and at the end of a tectonic cycle (spatial extent). With the occurrence of characteristic earthquake, and the resulting wallrock separation, the enhancement is either substantially reduced or eliminated, and "normal" conditions of the hydraulic conductivity structure are restored.

3.3.4 Conceptual Model of a Groundwater System in a Deforming Fractured Medium

Plate 3.3.4-1 presents a conceptual model of a groundwater system developed in a deforming fractured medium. In this model, both vertical boundaries are considered as "no fluid-flux" or "const. head" boundaries. Hydraulic pressures acting at both vertical boundaries differ by a constant amount, which is represented by the term cx. The horizontal boundary is considered as a "no fluid-flux" boundary.

It is assumed that a groundwater system contains a thick vadose zone. Prior to initiation of tectonic straining, the groundwater flow occurs as a steady-state process. The medium is strongly fractured along various directions. Below the water table, fracture flow is a dominant mechanism of groundwater movement. Furthermore, a groundwater system is assumed to be situated in climatic conditions such that recharge, through the vadose zone, is very small and initially, therefore, this recharge occurs as interstitial flow.

The fractured medium is a deforming one and, consequently, the *in situ* stress field is changing in time and space. The deformation is advanced; the *in situ* stress field is evolved, and the surface $Z_{(x,y,t)}$ is situated at a depth ranging from zero to a few kilometers. Above the surface $Z_{(x,y,t)}$, the hydraulic properties of the medium (conductivity and storativity) are both time- and space-dependent. Below this surface, however, the hydraulic properties are independent of time, although they may be space-dependent.

Across the surface $Z_{(x,y,t)}$, there is a time- and space-dependent transfer of groundwater, which is represented by the term $\pm R_{(x,z,t)}$. This term expresses the volume of groundwater added, around point $P_{(x,z)}$ to unit volume of the system. Such additions are caused by differences in the value of hydraulic potentials above and below the surface $Z_{(x,y,t)}$. Time-dependence of the term $\pm R_{(x,y,t)}$ is caused by the downward migration of the surface $Z_{(x,y,t)}$ which, in turn, is the result of ongoing tectonic straining of a groundwater system.

The above conceptual model is useful in developing an understanding of long-term behavior of groundwater systems that operate in a deforming fractured medium. Such an understanding is essential for informed judgments concerning magnitudes of hydrologic disturbances that are possible in these systems. Such understanding is also essential for any meaningful interpretations of the origin and conceptual

importance of some contemporary characteristics of a groundwater system (e.g., temporal behavior of hydraulic potentials; spatial distribution of hydraulic potentials, including configuration of the water table; spatial distribution of hydraulic properties; response to vibratory ground motion; response to hydraulic stressing; and bodies of perched and semi-perched water).

Plates 3.3.4-2 through 3.3.4-5 present an idealized history of developments in a groundwater system that is responding to a changing *in situ* stress field in association with a cycle of tectonic deformation. The history is presented in the form of a depth vs. hydraulic potential relationship that would be observed, at four different time intervals, in a single reference well drilled at station $P_{(x,y)}$. Time t_o represents the onset of a tectonic cycle, and time t_3 represents the end of this cycle. For the purposes of simplicity, it has been assumed, that initially, the value of hydraulic potentials is constant with depth.

Plate 3.3.4-2 presents the assumed distribution of hydraulic potentials with depth at the onset of a tectonic cycle. At time t_o, the surface $Z_{(x,y,t_o)}$ is situated at or near the ground surface and, consequently, along the entire depth of the observation well the effective normal stress $\left[\sigma_{nfeff_{(t_o)}}\right]$ is greater than the critical closure pressure $\left(\sigma_{cc}\right)$, and the shear stress $\left[\tau_{f_{(t_o)}}\right]$ is smaller than the frictional resistance of fractures $\left(\tau_{\max}\right)$. Conducting apertures of fractures are, therefore, relatively small and consist of only residual components. The water table is located at the depth $d_{w_{(t_o)}}$.

Plate 3.3.4-3 presents a depthward distribution of hydraulic potentials during early stages of deformation. At time t_1, only along the surface $Z_{(x,y,t_1)}$ and below it, the effective normal stress $\left[\sigma_{nfeff_{(t_1)}}\right]$ is always greater than the critical closure pressure $\left(\sigma_{cc}\right)$, and the shear stress $\tau_{f_{(t_1)}}$ is smaller than $\tau_{\max}$. Below the surface $Z_{(x,y,t_1)}$, therefore, conducting apertures are equal to residual apertures. Above the surface $Z_{(x,y,t)}$, or within the vadose zone, for some fractures $\sigma_{nfeff_{(t_1)}} = \sigma_{cc}$ and/or $\tau_{f_{(t_1)}} = \tau_{\max}$ and, therefore, conducting apertures of such fractures have been increased, and with it local values of the hydraulic properties were also increased. In the vadose zone a tendency toward fracture flow has emerged. Because depth of the surface $Z_{(x,y,t_1)}$ is smaller than the depth $d_{w_{(t_o)}}$, the water table remains where it was during time t_o.

Plate 3.3.4-4 presents a depthward distribution of hydraulic potentials for advanced stages of deformation. At time t_2, the surface $Z_{(x,y,t_2)}$ is located far below the depth $d_{w_{(t_o)}}$. Above this surface, deforming

fractured medium is severely dilated and its hydraulic conductivity structure is enhanced accordingly. Between time t_1 and time t_2, thickness of the vadose zone increased from $d_{w_{(t_1)}}$ to $d_{w_{(t_2)}}$. Mechanism of flow through the vadose zone is fracture flow resulting from the locally enhanced values of conducting aperture. Volume of water released from storage in association with lowering of the water table was accommodated, partially, through increased out-flow facilitated by the increased values of hydraulic conductivity, and partially through increased storage below the water table caused by the increased values of storativity. The depth distribution of hydraulic potentials shows an important zonation. In Zone I the value of hydraulic potential is const.; the value of pore pressure increases with depth in accordance with the increasing hydrostatic pressure. Locally, in this zone $\sigma_{nfeff_{(t_2)}} = \sigma_{cc}$ and/or $\tau_{f_{(t_2)}} = \tau_{\max}$ for the values of pore pressure $p = p_{(z)}$. In Zone II, however, the value of hydraulic potential may be depth-dependent. Potentially, the amount of "overpressure" is equal to $\Delta p_{(z)}$. Locally, in this zone $\sigma_{nfeff_{(t_2)}} = \sigma_{cc}$ and/or $\tau_{f_{(t_2)}} = \tau_{\max}$, but only for the values of pore pressure $p = p_{(z)} + \Delta p_{(z)}$. Zone III begins at the surface $Z_{(x,y,t_2)}$. Within this zone, the value of "overpressure" attains a maximum $\Delta p_{(z)} = \Delta p_{\max}$. Below a certain depth, the value of hydraulic potential is const. and equal to the value of overall hydraulic potential at time t_o.

Plate 3.3.4-5 presents a depthward distribution of hydraulic potential at time t_3, which represents the end of a tectonic cycle. At this time, depth of the surface $Z_{(x,y,t)}$ has been reduced, from its former value at time t_2, to a very small value. The *in situ* stress enhancement of the hydraulic conductivity structure has been eliminated, and the values of hydraulic properties returned to their initial values at time t_o. Thickness of the vadose zone decreased from $d_{w_{(t_2)}}$ to $d_{w_{(t_3)}}$. Such decrease was caused by two factors: a) "overpressure" $\Delta p_{\max}$ in Zone III and b) water released from storage in the depth interval from $Z_{(x,y,t_2)}$ to $d_{w_{(t_2)}}$. The water table is at the depth $d_{w_{(t_3)}}$, which is smaller than the depth $d_{w_{(t_0)}}$. Such a position is not in accordance with the value of hydraulic potentials at a greater depth. The water table, therefore, is unstable and with time it will decay into the equilibrium position as represented by $d_{w_{(t_0)}}$. The elimination of *in situ* stress enhancement of the hydraulic conductivity structure causes the interstitial flow to be re-established in the vadose zone.

Plate 3.3.4-6 presents an idealized history of position of the water table during the entire cycle of tectonic deformation at a single point $P_{(x,y)}$. The history represents a summary of the above discussions and is presented in the form of a hydrograph that would result from a continuous monitoring of the water table in a single well between time t_0 and time t_3. A main feature of the hydrograph is the long-term progressive lowering of the water table. During advanced stages of deformation, however, the water ta-

ble lowering is not uniform in time and space. Occasionally, excursions from the long-term trend occur. Such excursions are represented on Plate 3.3.4-6 as, back and forth, fast fluctuations of the water table. The excursions are of relatively short duration and involve both positive (hydraulic mounds) and negative (hydraulic sinks) departures from the long-term trend. The excursions reflect local developments in the *in situ* stress field. Locally, rates of deformation and magnitudes of the in situ stresses are changed in response to a) spatial migrations of "singular" and "isotropic" points; b) changing density of occurrence of such points; and c) coupled interactions between fluid pressure and the *in situ* stress. The *in situ* stress changes may, or may not, be associated with or accompanied by vibratory ground motion. Such changes may be induced by some external factors, or they may develop spontaneously. The resulting perturbations of the water table are of a relatively small spatial extent and, therefore, are unstable; they cannot be justified by the overall distribution of hydraulic potentials. With time, they will decay into a more tolerable equilibrium position.

Plate 3.3.4-7 presents an inferred relationship between configuration (shape) of the water table and the *in situ* stress field. The stress field is represented by the surface $Z_{(x,y,t)}$. Three simple configurations were selected at this surface: a) step-like; b) step-like with a depression; and c) tilted surface. There are three important effects resulting from the assumed *in situ* stress field configurations. These are a) progressive lowering; b) development of areas where lateral hydraulic gradients, and the resulting local slopes of the water table, are very steep; and c) flattening or reduction of local lateral gradients of hydraulic potential, and the resulting flattening of the water table.

3.3.5 Why is it Important to Know that a Groundwater System is Developed in a Deforming Fractured Medium?

The analyses presented above indicate, that in a groundwater system developed in a deforming fractured medium, the water table is a very special one indeed. Specifically, the position and configuration of such a water table are not permanent features. Both of these features are related to a changing *in situ* stress field. Such a stress field controls conducting apertures of fractures and, through that, it also controls the hydraulic conductivity structure of a deforming fractured medium. As the *in situ* stress field undergoes cyclic changes, position and configuration of the water table also undergo such changes. Consequently, the water table constitutes a very poor representation of spatial distribution of hydraulic potentials, that actually motivate the groundwater system. To obtain meaningful representations of such potentials, it is necessary to perform measurements of the hydraulic head below the zone of *in situ* stress influence, that is below the surface $Z_{(x,y,t)}$.

In the case of hydrologic considerations performed for the purposes of developing a high-level nuclear waste repository in the vadose zone, two issues of fundamental importance are associated with the possibility that a groundwater system is developed in a deforming fractured medium. These issues are a) mechanism of flow in the vadose zone and the resulting groundwater travel-time and b) temporal stability of the water table, including its short- and long-term aspects.

Flow mechanism in the vadose zone was considered while presenting the history of developments in a groundwater system responding to a changing *in situ* stress field (Section 3.3.4). In such a system, the surface $Z_{(x,y,t)}$ migrates downward, in response to ongoing tectonic deformation, and reflects the *in situ* stress enhancements of conducting apertures of fractures. Magnitudes of such enhancements are depth-dependent. The greatest enhancements occur near the ground surface, and they gradually diminish downward. With increasing conducting aperture, likelihood of fracture flow in the vadose zone increases. Relative measure of such likelihood, therefore, is depth of occurrence of the surface $Z_{(x,y,t)}$. If this surface is situated at a depth of about 1.5 km, as is the case for central and eastern sectors of the Yucca Mountain area (Section 4.2.5), conducting apertures in the vadose zone may be sufficiently large to support local fracture flow, regardless of the mean annual rate of infiltration of rainfall. Furthermore, in the vadose zone, the *in situ* stress conditions are such that numerous "singular" points are present. Around such points the mean stress is very small $\left[\frac{1}{2}\ (\sigma_1 + \sigma_3) \rightarrow 0\right]$. Any build-up of fluid pressure, resulting from infiltration of surface rainfall, causes the Griffith type of failure and further enlargement of the local values of

conducting apertures. It is unlikely, therefore, that stressed in such a manner, fractures may retain fluids for a length of time that is required for introduction of these fluids into the rock matrix.

Expectation of a very small groundwater travel-time, through the Yucca Mountain vadose zone, is supported by preliminary results of recently performed studies of chlorine content of samples from the Yucca Mountain tuff "pile" (Norris, 1989). These studies have detected "a large pulse of ^{36}CL" in tuff samples that were obtained from a depth of approximately 1,300 feet below the ground surface. Chlorine is deposited globally both as rainfall and as dry fallout. It originates from sea salts, that are introduced into the stratosphere by atmospheric winds. A very small fraction of the chlorine atoms, less than 5 parts in 10^{13}, in the fallout consists of ^{36}CL. The ^{36}CL atoms are the result of cosmic ray reactions in the atmosphere (Norris et al., 1985). The sharply increased ^{36}CL content of the deep Yucca Mountain tuff sample, relative to the expected background, indicates that the source of ^{36}CL is not the cosmogenic fallout. Instead, it seems reasonable to attribute the ^{36}CL increased content to the ^{36}CL bomb pulse that was deposited globally after test detonations of high-yield nuclear weapons. These detonations were conducted, at sea level, in the Pacific Ocean between 1952 and 1959. It is likely that, during the last 20 or 30 years, the ^{36}CL bomb pulse was introduced deep into the Yucca Mountain vadose zone by infiltrating meteoric precipitation.

Rapid infiltration of meteoric precipitation, locally occurring in the form of fracture flow, is also suggested by the results of repeated downhole temperature profiles obtained from the Yucca Mountain vadose zone. One particularly noteworthy instance of fracture flow is evident in temperature profiles from Well UE-25a7 (Plate 3.3.5-1). Within the first 400 feet of the borehole, a progressive with time increase of the *in situ* temperature occurred during 1981 (profiles March 1981, April 1981, and December 1981). It is likely, that this increase represents the long-term recovery of thermal equilibrium that was disturbed by the drilling process. A dramatic change in the thermal behavior is indicated by the March 1983 temperature profile. In the uppermost 400 feet of the borehole, an erratic "cooling" of the borehole is evident. It is reasonable to attribute such "cooling" to infiltration of fluids from a major precipitation event that occurred in early March of 1983 (Sass et al., 1983).

Temporal stability of the water table was also considered in Section 3.3.4. Such stability is related to stability of the hydraulic conductivity structure, and through that, it is also related to temporal stability of the *in situ* stress field. In a cyclically deforming fractured medium, the *in situ* stress field under-

goes continuous changes that are expressed in a three-fold manner: a) downward migration of the surface $Z_{(x,y,t)}$; b) spontaneous and/or induced minor restructurings occurring above the surface $Z_{(x,y,t)}$; and c) large-scale restructurings taking place at the end of a tectonic cycle. It is logical to expect that the on-going *in situ* stress field changes will be reflected in three types of water table instabilities: a) long-term progressive lowering; b) short-term instabilities; and c) long-term changes in position and configuration. In the context of performance of a high-level nuclear waste repository, all of these instabilities entail profound implications. Short-term instabilities of the water table, such as excursions discussed in Section 3.3.4 and shown on Plate 3.3.4-6, if occurring with a meaningful frequency and amplitude, may constitute an important "pumping" mechanism for movement of gases and fluids in the vadose zone. Long-term changes in position and configuration of the water table may influence radionuclide migration path and time. In extreme cases, such changes may result in flooding of the repository, and expulsions of large quantities of fluids at the ground surface.

At this point an important question emerges: **How large are the *in situ* stress related changes in the position of the water table**? Some answers to this question may be provided by relationships shown on Plate 3.3.5-2. The maximum potential rise of the water table at point $P_{(x,y)}$, at the end of a tectonic cycle, is the sum of two components. The first component, $\Delta h^{ov}_{(x,y)}$, is related to "overpressure" $\Delta p_{\max}$. The maximum potential magnitude of this component is equal to the amount of "overpressure" present at point $P_{(x,y)}$ $\left[\Delta h^{ov}_{(x,y)} = \Delta p_{\max_{(x,y)}}\right]$.

The second component, $\Delta h^{sr}_{(x,y)}$, arises from reduction of storativity of a deforming fractured medium. Such a reduction occurs at the end of a tectonic cycle; it is caused by large-scale restructuring of the *in situ* stress field, and the resulting change in conducting aperture ($a_{con} = a_{res} + a_d + a_n \rightarrow a_{res}$). The maximum potential magnitude of the "storativity" component $\left[\Delta h^{sr}_{(x,y)}\right]$ is related to two factors: a) the ratio of storativity in "strained" state to storativity in "unstrained" state $\left[S^{\sigma}/S^{in} = (K^{\sigma}_f/K^{in}_f)^{\frac{1}{2}}\right]$ and b) the difference in the value of hydraulic potential in Zone III between point $P_{(x,y)}$ and a reference point located downstream at the edge of the zone of the restructured *in situ* stress field, $\left[\Delta h^{in}_{(x,y)} = \Delta p_{\max_{(x,y)}} - \Delta p^{R}_{\max}\right]$.

If, at point $P_{(x,y)}$, the ratio K^{σ}_f/K^{in}_f is equal to 100 and the difference $\Delta h^{in}_{(x,y)}$ is equal to 10 m, then the maximum potential value of the "storativity" component is $\Delta h^{sr}_{(x,y)} = 100$ m. Assuming that the

value of "overpressure" $\Delta p_{\max_{(x,y)}}$ is equal to 70 m, then the total potential rise of the water table at point $P_{(x,y)}$ is $\Delta h^{\sigma}_{(x,y)} = \Delta h^{ov}_{(x,y)} + \Delta h^{sr}_{(x,y)} = 70$ m + 100 m = 170 m.

An interesting situation arises when a) the ratio $K^{\sigma}_{f}/K^{in}_{f}$ is very large, in a range of 10^4 and more and b) the value of "unenhanced" hydraulic conductivity K^{in}_{f} is very small, so that a sizable hydraulic gradient exists in Zone III. Under such conditions, the values of fluid pressure build-up, that are required to "bleed-off" the volume of fluid released from storage, are very large, in a range of tens of bars. The country rock cannot withstand fast fluid pressure build-up of such magnitudes. It will tend to break down and form local pockets of "explosive" breccia. The resulting hydraulic conductivity of the country rock is locally substantially enhanced and, therefore, fluids released from storage may be discharged with relatively small increases of fluid pressure.

It is possible that some of the Yucca Mountain "mosaic" breccias were initially formed as described above. To account for the observed high degree of fragmentation, one may conclude that the process requires some lateral confinement. Such confinement, however, is not present at the ground surface. It is not surprising, therefore, that the "explosive" breccias currently occurring at the ground surface are relatively old. Most likely, these breccias were formed at a time when the contemporary bedrock surface was situated at some depth, and under cover of younger rocks. These younger rocks were since removed by erosion. Younger, less cemented and developed "explosive" breccias, if such exist, should be developed at depths where the required lateral confinement is present. At and near the ground surface, the fluid pressure build-ups should be expressed as various forms of wallrock separation. It is possible that bedrock openings which, in the Yucca Mountain area, contain the calcite-opaline silica veins and the "fragmentation" breccias are, in fact, surface counterparts of the hypothetical subsurface "explosive" breccias.

In summary, failure to recognize that a deforming fractured medium is present, at a potential site of a high-level nuclear waste repository, may entail serious errors in judgment regarding intrinsic radionuclide isolation capabilities of this site. This is true, in particular, in the case of Yucca Mountain where a common denominator for all long-term performance objectives is a small flux of fluids through the vadose zone. In this situation, it is essential to make sure that a transfer of large quantities of fluids from below the water table is not a likely possibility. Most certainly this is not the case, however, if the local *in situ* stress is the controlling factor for the local groundwater system.

3.3.6 How Can Recognition of the Involvement of a Deforming Fractured Medium be Achieved?

The most important characteristic of a groundwater system developed in a deforming fractured medium is its transient behavior. Most of the time, such a transient behavior involves small, in time and space, oscillations of the values of hydraulic potentials. Such oscillations occur in conjunction with a slow downward drift of these potentials. Large-scale changes in position and configuration of the water table occur in association with seismogenic fault ruptures. Such ruptures are accompanied by large-scale restructurings of the *in situ* stress field, which in turn, cause drastic reductions of conducting apertures. Both of these aspects of transient behavior may be used to achieve recognition that a groundwater system is developed in a deforming fractured medium. Conceivably, there are two approaches that could be utilized in such an undertaking.

The first approach involves examination of the geologic record as expressed through compositions and textures of various geologic materials of Plio-Quaternary age. The main purpose of such an examination is to establish whether or not this record contains expressions of a) progressive "widening" of the area and b) large-scale fluctuations of the water table. Both of these processes should be expressed as occurrences of geologic materials whose volume, textures, and compositions are indicative of a) large, post-depositional, volumetric strains of the country rock and b) involvement of groundwater in their formation.

The second approach involves evaluation of contemporary characteristics of a groundwater system. Such characteristics, however, must be defined based on the results of extensive studies of a groundwater system. They include a) temporal behavior of the water table; b) distribution of hydraulic potentials as a function of depth; c) temporal behavior of hydraulic potentials as a function of depth; d) responses of hydraulic potentials to vibratory ground motion; e) spatial distribution of *in situ* stress; f) spatial distribution of hydraulic properties; g) responses of a fractured medium to hydraulic stressing; and h) responses of a groundwater system to large-scale restructurings of the *in situ* stress field caused, for example, by detonations of nuclear devices.

Temporal behavior of the water table is, perhaps, the most readily accessible characteristic of a groundwater system. This characteristic may be used for the purposes of establishing whether or not a deforming fractured medium may be present. To this end, the results of measurements of depth to the water table, made as a function of time and at several locations, are required. For a groundwater system con-

taining a thick vadose zone and situated such that arid climate conditions prevail, presence of a) meaningful oscillations of the water table and b) long-term progressive downward drift of the water table; signals that a deforming fractured medium may be involved.

Another easily accessible characteristic of a groundwater system is the **distribution of hydraulic potentials as a function of depth**. With reference to Plate 3.3.4-4, such a distribution may be used to evaluate whether or not "overpressure" $\Delta p_{(z)}$ is, or may be, present at a greater depth. Care should be taken, however, while interpreting the origin of "deep" hydraulic potentials. The downward increase of the value of hydraulic potentials may be caused by a number of factors; only one such factor is a deforming fractured medium. Without the proper conceptual understanding and without additional data, it is easy to confuse a plausible explanation with a correct interpretation.

Pattern of **temporal behavior of hydraulic potentials as a function of depth** is another characteristic that may be used to determine that a groundwater system is developed in a deforming fractured medium. For such a system, the temporal behavior of hydraulic potential at a shallow depth is different than that at a greater depth, where "overpressure" $\Delta p_{\max}$ is present. The presence of spontaneous transient behavior, involving either oscillations or downward drift, at a shallow depth and the absence of such behavior in the "overpressure" zone indicates, that most likely, a deforming fractured medium is present. Such distinct patterns of behavior are the result of depth-dependent influence of the *in situ* stress on conducting apertures. Above the surface $Z_{(x,y,t)}$, conducting apertures are locally enhanced and time-dependent. Even a very small change in the *in situ* stress, resulting from ongoing extension, causes continuous increases of conducting apertures. Such increases are expressed as a downward drift of the value of hydraulic potentials. Furthermore, if the surface $Z_{(x,y,t)}$ is situated at a considerable depth, the *in situ* stress field is evolved and undergoes spontaneous and externally induced restructurings. Such restructurings are related to a) spatial migration of "isotropic" and "singular" points; b) increasing density of occurrence of such points; and c) *in situ* stress – fluid pressure coupled interactions. The restructurings cause local values of conducting apertures to change episodically. This, of course, is expressed as local oscillations of the values of hydraulic potentials. Below the surface $Z_{(x,y,t)}$, however, conducting apertures consist solely of residual components. Such apertures are not sensitive to small changes of the *in situ* stress resulting from the ongoing deformation. Furthermore, such apertures experience neither spontaneous nor externally induced changes. Consequently, both aspects of transient behavior of hydraulic potentials (downward drift and oscillations) are absent.

Recognition of a deforming fractured medium may also be achieved by observing **responses of hydraulic potentials to vibratory ground motion** produced either by a natural earthquake or by an underground nuclear detonation. The observations should be made above and below the surface $Z_{(x,y,t)}$, and using fast-recording pressure transducers. Above the surface $Z_{(x,y,t)}$, the *in situ* stress field is coupled with the field of hydraulic potentials. It can be expected, therefore, that transient perturbations of the *in situ* stress field, introduced by vibratory ground motion and consisting of alternating compression and extension, will be amplified in time and space. Such amplifications should be expressed as a) relatively large oscillations of hydraulic potentials and b) oscillations occurring long after passage of seismic waves. In contrast, below the surface $Z_{(x,y,t)}$, the *in situ* stress field and the field of hydraulic potentials are not related to each other. In this portion, therefore, the vibratory ground motion induced oscillations of hydraulic potentials should be different. The magnitude and duration of such oscillations may be readily justified in terms of the observed dynamic strain and duration of vibratory ground motion.

Recognition of a deforming fractured medium may further be achieved based on the results of *in situ* **stress determinations**. Such determinations should be performed as a function of depth and at several locations. Because the depth distribution of in situ stress is required for considerable aquifer thicknesses, the *in situ* stress determinations must be made utilizing the technique of hydraulic fracturing. In a deforming fractured medium there is a characteristic zonation of the *in situ* stress conditions as a function of depth. Such a zonation is shown on Plate 3.3.4-4. The aquifer is divided into three zones each with distinct *in situ* stress conditions. Zone I extends from the ground surface down to a depth Z_1. In local areas within this zone, the normal effective stress (σ_{nfeff}) is equal to the critical closure pressure (σ_{cc}) and/or the shear stress (τ_f) is equal to the frictional resistance of some fractures $(\tau_{\max})$. Such conditions occur for values of fluid pressure $p = z \cdot \varrho$ (ϱ is density of water and z is depth below the water table). Zone II extends from the depth Z, down to the surface $Z_{(x,y,t)}$. Locally, within this zone $\sigma_{nfeff} = \sigma_{cc}$ and/or $\tau_f = \tau_{\max}$, for values of fluid pressure $p = z \cdot \varrho + \Delta p_{(z)}$. Zone III extends from the surface $Z_{(x,y,t)}$ downward. Within this zone, σ_{nfeff} is smaller than σ_{cc}, and τ_f is smaller than $\tau_{\max}$. This holds for values of fluid pressure $p = z \cdot \varrho + \Delta p_{\max}$.

In a deforming fractured medium, the characteristic zonation of the *in situ* stress is reflected in **spatial distribution of the *in situ* values of hydraulic properties** (conductivity and storativity). Such a relationship is caused by the depth-dependent *in situ* stress enhancement of conducting apertures. Within the *in situ* stress Zone III, below the surface $Z_{(x,y,t)}$, conducting apertures are relatively small and consist

of only residual components. For a medium with small secondary porosity, the *in situ* values of hydraulic conductivity are comparable with the laboratory derived values. Above the surface $Z_{(x,y,t)}$, however, there is an upward increase in the *in situ* values of hydraulic conductivity. The increase is caused by upward increase in the values of post-peak shear displacement $(u_s - u_{sp})$. The largest values of *in situ* hydraulic conductivity occur in the *in situ* stress Zone I. If the values of residual aperture are very small , then differences in the *in situ* values of hydraulic conductivity between Zone I and Zone III may be very substantial, say a factor of 10^3 or 10^4 (Section 4.2).

Observations made during hydraulic stressing of a groundwater system may also be used in evaluations of whether or not a deforming fractured medium is present. For a system in which the surface $Z_{(x,y,t)}$ is located at a considerable depth, one or two kilometers, the *in situ* stress field is evolved and contains numerous "isotropic" and "singular" points. Around such points the Griffith tensile failure, and the resulting dramatic increase in the *in situ* values of hydraulic conductivity, may be induced by a very small increase in fluid pressure. Consequently, on a hydrofracture record the presence of a "singular" point should be expressed as a virtual absence of the "break-down" pressure. Similarly, on a Cooper-Bredehoeft injection record, the presence of a "singular" point should be expressed as a very large velocity of slug movement, say 10^2 m per minute.

As far as the involvement of a deforming fractured medium is concerned, perhaps, the most informative are observations of **responses of a groundwater system to large-scale restructurings of the *in situ* stress field**. A detonation of a sizable nuclear device, with a yield of about one megaton, entails such a large-scale restructuring. The resulting changes in a groundwater system may be monitored in time and space using a system of properly instrumented observation wells. The involvement of a deforming fractured medium should be expressed by a) large oscillations of the values of hydraulic potentials occurring long after passage of the detonation induced shock waves and b) sustained, but localized, rises of the water table resulting in formation of hydraulic mounds.

It is important to recognize that data required to achieve a more or less reliable definition of a character of fractured medium in the Yucca Mountain area are already available. Surficial geology of this area is known based on detailed geologic mapping and trenching performed during the last 10 years. Subsurface conditions are known based on examinations of nearly 10 km of rock cores extracted from Yucca Mountain. Spatial distribution of the *in situ* stress is known based on hydraulic fracturing measurements

performed in four deep wells, down to a depth of approximately 1,500 m. Observations of responses of the local groundwater system to hydraulic stressing were made during the Cooper-Brederhoeft injection tests. Such tests were carried out in eight wells, down to a depth of about 2.0 km. Spatial distribution of the *in situ* values of hydraulic conductivity is known based on numerous pumping and injection tests. Such tests were made in seven deep wells, down to a depth of about 2.0 km. Since 1981, periodic measurements of hydraulic potentials were carried out in 28 deep wells. In a few wells, the measurements were made at a few discreet depth intervals. In late 1983 continuous monitoring was initiated, and by 1986 the present network, consisting of about 20 deep wells, for continuous monitoring became operational. Several natural tele-seismic events (e.g., the Mexico earthquake of 1986 and the Los Angeles earthquake of 1987) and numerous underground nuclear detonations have occurred during the monitoring period. Finally, at the Nevada Test Site, observations of responses of the local groundwater system to large-scale restructurings of the *in situ* stress field were made in association with detonations of a few nuclear devices, with a yield of up to one megaton.

It seems reasonable to expect that **reliable and informed judgments concerning the character of fractured medium in the Yucca Mountain area be made on the basis of actual data and known facts**. Development of such judgments, however, solely on the basis of a) numerical simulations of some other hydrologic circumstances; b) experience and expertise limited to groundwater supply and groundwater management problems; and c) weakly supported hydrologic beliefs and conventions, cannot be regarded as a wise and responsible course of action.

3.4 CONCEPTUAL CONSIDERATIONS OF A GROUNDWATER SYSTEM INFLUENCED BY TERRESTRIAL HEAT

3.4.1 Geothermal Setting of the Yucca Mountain Groundwater System

Regional tectonic setting of the Yucca Mountain groundwater system, as summarized in Section 3.1, suggests, that in addition to the *in situ* stress, there is another tectonic factor, that potentially, may influence groundwater flow. Such factor is terrestrial heat that is transferred from the mantle, through the crust, and into the ground surface. An understanding of the geothermal setting of the Yucca Mountain groundwater system is, therefore, an important aspect of the conceptual considerations undertaken in this report. Such understanding has been developed based on a) the results of studies of composition of lavas associated with the Plio-Quaternary volcanism in the area; b) the results of studies of structure of the upper mantle and crust, as expressed in spatial distribution of the P-wave velocity; and c) the results of measurements of intensity of flux of terrestrial heat.

The Yucca Mountain groundwater system is situated within central portions of the Death Valley - Pancake Range Volcanic Zone (Plate 3.4.1-1). This zone contains notable expressions of small scale basaltic volcanism of Plio-Quaternary age (Crowe et al., 1986). Isotopic and chemical compositions of lavas related to this volcanism are noteworthy. Detailed petrologic studies of these lavas were performed by Vaniman and Crowe, 1981; Vaniman et al., 1982; Semken, 1984; and Crowe et al., 1986. The results of these studies revealed that the lavas display hawaiite affinity. Significant features of these rocks are a) an "alkaline classification; b) the presence of groundmass olivine and calcic clinopyroxene; c) scarcity of groundmass calcium-poor pyroxene; and d) andesine - normative feldspar compositions" (Crowe et al., 1986). Both the evolved and primary basalts are present. The rocks are highly enriched in light rare-elements, even when compared to typical alkalic basalts (Vaniman et al., 1982). They exhibit a broad range of isotopic compositions. The chondrite-normalized strontium and neodymium isotopic ratios for these rocks range from $\epsilon_{SR} = 42.2$ to $\epsilon_{SR} = -9.0$ and from $\epsilon_{ND} = -10.4$ to $\epsilon_{ND} = 3.8$ (Semken, 1984). Isotopic compositions of the evolved and primary basalts appear to be similar.

The observed enrichment in light rare-earth elements and in strontium, coupled with a low rubidium/strontium ratios of the southern Nevada Test Site basalts, was attributed to the upper mantle origin of these basalts (Vaniman et al., 1982). The depth of magma generation was estimated to be within a

range of 35 to 75 km. Semken, 1984, concluded that the unique geochemical and isotopic characteristics of the basalts are best accounted for by assuming that the mixing of lower crust and upper mantle materials was involved in the generation of these special magmas. The presence of evolved and primary basalts, both with similar isotopic compositions, was interpreted to indicate that some mixing and the resulting contamination of the upper mantle materials occurred prior to magma generation. Semken, 1984, further concluded that convective motions were involved in transporting the contaminated upper mantle materials from near the depth of Moho discontinuity down to the source depth of alkalic basalts.

The occurrences of very young alkalic basalts, together with their unique geochemical and isotopic characteristics, allow for the following two inferences to be drawn. First, in the region of the Yucca Mountain groundwater system, the upper mantle is partially melted. It seems reasonable to assume that viscosity of the mantle is low, less than 10^{22} poise (Cathles, 1975). Second, in the upper mantle, convective motions appear to be present. These motions may be responsible for the transport of contaminated mantle materials from near the Moho discontinuity into the source regions of alkalic basalts.

The seismic velocity structure of the area occupied and surrounding the Yucca Mountain groundwater system is known based on studies of the P-wave travel time residuals performed by Monfort and Evans, 1982. The data for this study were obtained during 1980 by recording 98 teleseismic events at as many as 53 sites. The study revealed a local seismic velocity structure down to a depth of approximately 280 km.

The results of detailed analyses of data obtained from seismic monitoring stations located at and near the Nevada Test Site are of particular relevance. The locations of these stations are shown on Plate 3.4.1-2. The analyses were performed as linear inversions, using the Aki-Christofferson-Husebye damped-least-squares technique. This technique allows for definition of the relative P-wave velocity structure of an area in three dimensions.

The portion of the Earth covered by the monitoring array was subdivided into five layers. These layers are located at the following depth intervals: 0 to 15 km, 15 to 31 km, 31 to 81 km, 81 to 131 km, and 131 to 231 km. Each of the layers was then subdivided into rectangular blocks, each 35 km long. For each of the blocks, the relative P-wave velocities were computed using the damped-least-squares inversions, and assuming that the average velocity of each layer increases with depth. The relative velocities were computed as ratios of the velocity in each block to the average velocity in a layer hosting this block. The rel-

ative P-wave travel-time residuals were used, therefore, the average layer velocity does not correspond to the absolute velocity in this layer. In other words, the technique allows for detection of velocity contrasts but does not produce absolute velocity information.

Plates 3.4.1-3 through 3.4.1-7 present the relative P-wave velocity structure for the region of Yucca Mountain groundwater system. There are two noteworthy features of this structure. The first feature is a low-velocity trough. This, more or less continuous trough, extends from Death Valley, California to about 50 km north of the Nevada Test Site (Plates 3.4.1-3 and 3.4.1-4). The trough underlies the Yucca Mountain groundwater system, and appears to be present in both the upper and lower crusts. Within the trough, there are two localized low-velocity anomalies. Both of these anomalies are located in the upper crust (Plate 3.4.1-3). One anomaly occurs in the area of Pahute Mesa, near or within recharge areas of the groundwater system. The other anomaly is situated in the eastern portion of the groundwater system, near the southwestern corner of the Nevada Test Site, in the area of Skull Mountain-Wahmonie.

The second feature is a pronounced variability of the P-wave velocity in the upper mantle. Lower than average velocities occur beneath central and southern portions of the groundwater system. Such velocities occur in all three layers considered (Plates 3.4.1-5, 3.4.1-6, and 3.4.1-7). Higher than average velocities occur in the northern portion of the groundwater system. Across the Nevada Test Site, a distance of about 100 km, the difference in the P-wave velocity is a respectable 6 percent. The Yucca Mountain area appears to be situated near the margin of a low-seismic velocity zone. This zone is present in both the lower crust and upper mantle. South of Yucca Mountain, the P-wave velocities are lower than the average values. North of Yucca Mountain, however, the upper mantle shows higher than average values of the P-wave velocity.

There seems to be a rather noticeable correlation between the location of the low-velocity trough on Plates 3.4.1-3 and 3.4.1-4, and the location of the Death Valley-Pancake Range Volcanic Zone on Plate 3.4.1-1. This correlation suggests that a) the P-wave residuals map reflects real heterogeneities in the seismic structure of the crust and b) the seismic structure reflects thermal alterations of rigidity of the crust. The two localized anomalies in the upper crust (Plate 3.4.1-3) may be interpreted as indicative of the presence of relatively large magma chambers. Some support for this interpretation may be inferred from the results of geothermal measurements performed within and adjacent to the Nevada Test Site (Sass et al., 1976, and Sass and Lachenbruch, 1982). The low-velocity anomaly located in the area of Skull Mountain-

Wahmonie, is associated with increased intensities of flux of terrestrial heat. Such intensities range from 3.1 HFU in Well UE-25-a3, through 2.2 HFU in Wells TW-3 and TW-4 to 2.0 HFU in Well TW-5 and 1.8 HFU in Well TWF (Sass and Lachenbruch, 1982). Also, the low-velocity anomaly in Pahute Mesa seems to be associated with perturbations of the local geothermal field. The value of the vertical geothermal gradient measured in Well PM-1 ranges from 24 to 29°C/km (Sass et al., 1976). In another well, however, the geothermal gradient is noticeably increased. Measurements performed in Well PM-2 revealed that the value of this gradient ranges from 57 to 49°C/km (Sass et al., 1976; Plate 4.3.2-9).

The above interpretations of seismic velocity structure of the crust are not unequivocal, and alternative interpretations are possible. One such interpretation is that the heterogeneities in the P-wave velocity structure are not related to the distribution of *in situ* temperature. Instead, the heterogeneities are related to accumulations of low-seismic velocity materials in structural troughs, and in a sequence of volcanic calderas of Tertiary age. This alternative interpretation, however, does not appear as an attractive one for the following reasons. First, the low-velocity trough appears to be present in the lower crust, where accumulations of less dense materials are unlikely. Second, the location of the trough coincides with the location of Plio-Quaternary volcanic centers. Isotopic and geochemical characteristics of lavas from these centers are consistent with the assumption, that within the trough, *in situ* temperatures in the lower crust are relatively high. Third, the locations of two localized low-velocity anomalies appear to coincide with increases of the local values of geothermal gradients.

Clearly, presence of the substantial (6.2 percent in a lateral distance of approximately 100 km) (Plate 3.4.1-6) heterogeneities in the seismic velocity structure of the upper mantle at the Nevada Test Site is important. Such heterogeneities indicate, that in the upper mantle of this region, there are significant gradients of either density (phase or chemical transitions) or, most likely, large-scale fabric of materials forming the upper mantle (spatial distribution of perferred crystal orientations that results from high temperature and pressure creep). It is interesting to note, that at the Nevada Test Site, the degree of seismic anisotropy is comparable to that noted during careful seismic refraction studies of the upper mantle performed in the Pacific Ocean and in the Mediterranean Sea (Berry and Knopoff, 1967; Raitt et al., 1969; Morris et al., 1969; Keen and Barrett, 1971). Since the latter anisotropies were initially discovered, considerable attention was devoted to refining understanding of physico-chemical constitution and rheological state of the upper mantle (Ringwood, 1969; Green and Radcliffe, 1972; Baker and Carter, 1972). Particularly important are studies concerning the constitutive equations governing the high temperature and pres-

sure, low strain rate creep, and of atomic processes responsible for it (Carter et al., 1972). The results of such studies were used to attribute the observed lateral seismic anistropy in the upper mantle to kinematic fabric of materials comprising the upper mantle. Such fabric is the result of syntectonic recrystallizations, that accompany creep of ultramafic materials under conditions of high temperature and pressure. The lateral variability of preferred crystal inclinations is the result of non-homogenous flow, during which the flow lines are variably inclined with respect to a horizontal plane. The overall large-scale fabric of the upper mantle materials, however, is not a static (frozen) feature. Rather, it is a kinematic feature reflecting configurations of the local strain (stress) field in the upper mantle.

Considering the relatively low viscosity (10^{20} to 10^{22} poise) of the upper mantle in the Great Basin, it is rather difficult to comprehend the presence of lateral gradients, involving either density or kinematic fabric, without involving convective motions, be it of thermal or gravitational origin. The upper mantle anisotropy, as expressed in the Nevada Test Site seismic structure, may be regarded either as a cause (density) or as an effect (kinematic fabric) of such motions. In either case, however, it seems reasonable to conclude that the lateral gradients of the P-wave velocity correspond to the lateral temperature gradients. It appears, that apart from doubting the validity of definition of the local seismic velocity structure as performed by Monfort and Evans, there can no longer be any serious reservations concerning this statement.

The analyses of the seismic velocity structure of the Yucca Mountain groundwater system region allow for two inferences to be drawn. First, the convective motions seem to be occurring in the upper mantle of this region. This inference is similar to the one drawn earlier that was based on geochemical and isotopic compositions of lavas related to the Plio-Quaternary volcanism. Second, in terms of their size, convective systems seem to be narrow and concentrated suggesting, that in the upper mantle of this region, the actual Rayleigh number substantially exceeds the critical Rayleigh number.

In the Yucca Mountain groundwater system, and in the region surrounding it, geothermal conditions are known based on downhole measurements of *in situ* temperatures, and based on studies of silica contents of groundwater. The results of such studies were published by Sass et al., 1976; Swanberg and Morgan, 1978; Sass et al., 1980; Sass and Lachenbruch, 1982; Sass et al., 1983; and Sass et al., 1987.

The distribution of intensities of flux of terrestrial heat at the Nevada Test Site are presented on Plates 3.4.1-8 and 3.4.1-9. These intensities were computed by combining the results of downhole measure-

ments of *in situ* temperature with the results of laboratory measurements of thermal conductivity (Sass et al., 1976).

Based on computed intensities of flux of terrestrial heat, a large area was distinguished within the southern Great Basin. In this area anomalously low, by regional standards, intensities of flux of terrestrial heat occur. The area includes the Nevada Test Site and was named the Eureka Low by Sass et al., 1976. The anomaly was attributed to a complex hydrologic disturbance.

The location of the Eureka Low is shown on Plate 3.4.1-10. This plate also presents the 2.5 HFU contour of the intensity of flux of terrestrial heat. This contour was established, based on silica contents of groundwater, by Swanberg and Morgan, 1978. The silica contents suggest, that for most of the Eureka Low including the Nevada Test Site, intensities of "deep" flux of terrestrial heat are equal to or greater than 2.5 HFU. Such intensities are substantially greater than average intensities computed on the basis of downhole measurements of *in situ* temperature. Recognizing this inconsistency, Sass and Lachenbruch, 1982, concluded that a) groundwater "is carrying off much of the Earth's heat in the upper 3 km and is delivering it elsewhere" and b) "the regional heat flow from beneath the zone of hydrologic disturbance in the Eureka Low may be the same as that characteristic of the Great Basin, in general, (80 m Wm^2 or 2 HFU) or it could be as high as 100 mWm^{-2} or 2.5 HFU."

The conductive flux of terrestrial heat with intensities ranging from 2.0 to 2.5 HFU (1 HFU = 1 $\mu cal/cm^2sec$) is not trivial by terrestrial standards. The following two lines of reasoning are offered in support of this statement. First, as shown on Plate 3.4.1-11, an increase in the conductive surface flux of terrestrial heat from 1.4 to 2.5 HFU amounts to a temperature change along the Moho discontinuity, at a depth of approximately 30 km, from 500 to about 1,000°C. At a pressure of about 25 Kbars, incipient melting of pyrolite (a hypothetical primitive mantle material) containing 0.1 percent H_2O begins at a temperature of about 1200°C (Ringwood, 1969). Second, apart from areas containing shallow magma bodies and excluding areas of locally convective increases of heat flux, the range of 2.0 to 2.5 HFU is almost as high as intensity of the conductive flux of terrestrial heat gets on the planet Earth (Plate 3.4.1-12). A compilation of over 400 measurements of the intensity of flux of terrestrial heat from the Eurasian continent yielded a total range from 0.4 to a maximum of 3.3 HFU; the mean areal intensities range from 1.0 to no more than 2.1 HFU (Lubimova and Polyak, 1969). A compilation of results of world-wide measurments of intensities of oceanic heat flux was prepared by Lee and Uyeda, 1965. The average intensities of

oceanic heat flux are a) 0.99 ± 0.61 μcal/cm^{2}sec for oceanic trenches; b) 1.28 ± 53 μcal/cm^{2}sec for back-arc basins; and c) 1.82 ± 1.56 μcal/cm^{2}sec for oceanic ridges. Somewhat similar values were found in a relatively recent update of heat flow in the East and Southeast Asian Seas (Anderson, 1980). Across the Japan and Bonin Trenches, the intensities of flux of terrestrial heat range from 1.5 to 3.0 HFU in the back arc basins (Sea of Japan and Southern Shikoku Basin) to less than 1.5 HFU over both of the trenches.

In summary, the review of geothermal setting of the Yucca Mountain groundwater system indicates that there are many reasons to believe that the input of thermal energy into this system is substantial. The average intensity of flux of terrestrial heat may be as high as 2.0 to 2.5 HFU. Such conditions appear to be related to convective motions occurring in the upper mantle of this region. In this situation, it is reasonable to expect, that at deep crustal levels, there are substantial heterogeneities in the planar distribution of *in situ* temperature. The presence of such heterogeneities seems to be reflected in the local seismic velocity structure. Lower than average seismic velocities may indicate areas where *in situ* temperatures, at deep crustal levels, are high and the resulting intensities of flux of terrestrial heat are also high. Positive seismic velocity perturbations may indicate areas where the intensities of heat flux are lower.

The high and variably distributed flux of terrestrial heat may play an important role in the Yucca Mountain groundwater system, including its transient behavior during tectonic disruptions. It seems worthwhile, therefore, to seek some understanding of the thermally influenced flow of groundwater, its hydrologic importance, and in the case of a high-level nuclear waste repository, its potential performance implications.

3.4.2 Simultaneous Flow of Fluid and Heat - T/H Coupled Flow

With increased temperature, the density of water decreases. In a column of water containing a temperature gradient, the density contrast between warm fluid, deeper within the column, and colder fluid, at a shallower depth, introduces a buoyancy gradient acting in a vertical direction. The magnitude of such a gradient depends on a change in water density caused by a given rise in temperature:

$$i_b = \frac{\alpha_w}{Z} \int_{Z_2}^{Z_1} dTdz$$

where:

i_b - buoyancy gradient;

α_w - coefficient of thermal expansion of water, $\alpha_w = 3.85 \cdot 10^{-4}$ per °C; and

$\int_{Z_2}^{Z_1} dTdz$ - change in temperature between depth z_1, and z_2.

The total hydraulic gradient combines the hydraulic head gradient and the buoyancy gradient in a manner shown on Plate 3.4.2.-1. If a small and uniform buoyancy gradient is introduced there is no resulting perturbation in the flow field. However, if a high and/or heterogeneously distributed buoyancy gradient is introduced, perturbations in the flow field occur. Such perturbations are expressed as a change in the position and configuration of the flow lines, and as a change in the specific discharge of fluid across a reference unit area. The specific discharge of fluid, moving through a permeable medium, in response to both the pressure gradient and the buoyancy gradient, is given by

$$q_f = -L_1 \frac{\partial h}{\partial L} - L_2 \frac{\partial t}{\partial L}$$

where:

q_f - specific discharge of fluid;

L_1 and L_2 - constants of proportionality;

$\frac{\partial h}{\partial L}$ - pressure gradient; and,

$\frac{\partial t}{\partial L}$ - temperature gradient.

In most hydrologic situations, $L_1 \frac{\partial h}{\partial L} \geq L_2 \frac{\partial t}{\partial L}$. Under these conditions the assumptions that a) the groundwater system is adequately represented by one of the mathematical models for "simple" flow; b) the hydraulic head distribution at vertical boundaries of the system is a suitable representation of the total potential energy contained in this system; and c) the hydraulic conductivity is the only important phenomenological coefficient, are justified.

In a groundwater system that is developed in an area which is volcanically active and, which consequently contains a strongly heterogeneous temperature field, the value of $L_2 \frac{\partial t}{\partial L}$ must be considerable relative to $L_1 \frac{\partial t}{\partial L}$. In this situation, one is facing two, more or less, interactive energy fields (i.e., the geothermal field and the groundwater field). Interaction between these two fields is expressed as a simultaneous flow of heat and fluid under the influence of two interdependent energy gradients. Such a simultaneous flow is a reflection of the thermodynamic concept of coupled flow. For such a flow:

$$q_f = -L_{11} i_1 - L_{12} i_2; \text{and}$$

$$q_h = -L_{21} i_1 - L_{22} i_2.$$

where:

$q_{f,h}$ - specific discharge of fluid or heat;

i_1 - hydraulic pressure gradient;

i_2 - temperature gradient; and,

$L_{11}, L_{12}, L_{21}, L_{22}$ - phenomenological coefficients.

A mathematical model for the simultaneous flow of fluid and heat was described by Thunvik and Braester, 1980. The model consists of a system of coupled non-linear partial differential equations. As shown on Plate 3.4.2-2, governing equations for the coupled flow of heat and fluid are a) the equation for conservation of fluid mass, derived from basic conservation laws and Darcy's Law and b) the equation for conservation of thermal energy. There are also two equations of state (Plate 3.4.2-2). These equations express pressure- and temperature-dependence of water density and viscosity.

3.4.3 Thermal Instabilities of Fluids in a Porous and Fractured Medium

3.4.3.1 General

Thermal stability of fluids has been examined both theoretically and experimentally for two geometries: a) a horizontal layer of fluid bounded on upper and lower surfaces, but unaffected by end considerations and b) fluid in a vertical channel or tube. The initial analyses were performed by Rayleigh, 1916. These analyses considered a horizontal layer of incompressible fluid heated uniformly from below, and explained observations that fluid motion does not begin until temperature at the lower surface exceeds temperature at the upper surface by a critical amount. After fluid motions begin, a mosaic of convection cells develops. The individual cells are usually hexagonal with a width approximately equal to the depth of fluid layer. Fluid moves up the center of each cell and down near the cell boundaries.

The Rayleigh number, a dimensionless group, contains factors that govern thermal stability of a fluid, and is expressed as a ratio of buoyant forces to dissipating forces (viscous retardation). For a horizontal layer of incompressible fluid, heated uniformly from below, the Rayleigh number is

$$R_{ah} = \frac{\alpha \cdot g \cdot L^4 \cdot G}{\mu \cdot \lambda_f}$$

An explanation of all mathematical notations used in Section 3.4.3 is presented on Plate 3.4.3.1-1.

Thermal instability of a fluid begins if the Rayleigh number (R_{ah}) exceeds a stability limit, which in this case is the critical Rayleigh number (R_{ac}). For a horizontal layer of fluid, the critical Rayleigh number is equal $R_{ac} = 4\pi^2$.

Jeffreys, 1930, considered thermal stability of a slightly compressible fluid, and concluded that the critical temperature gradient required for the onset of thermal convection is given by the critical Rayleigh number provided the excess of the actual temperature gradient over the adiabatic temperature gradient, G-G_0, is used (the adiabatic temperature gradient is $G_0 = \alpha \cdot g \cdot T \cdot c_f^{-1}$).

For an incompressible fluid contained in a vertical channel, and heated from below, the Rayleigh number is

$$R_{at} = \frac{\alpha \cdot g \cdot L_c \cdot G}{\mu \cdot \lambda_f}$$

where:

L_c - is the characteristic radius of a channel.

Hales, 1937, has calculated values of the critical Rayleigh number (R_{ct}) for an infinitely long channel. This author has shown that the values of R_{ct} depend on the ratio of thermal conductivities of the fluid and of the channel walls, and whether flow is symmetric (flow up the center and down the walls or vice versa) or antisymmetric (flow up one wall and down the opposite walls). The published results vary from $R_{ct} = 68$ to $R_{ct} = 452$.

Verhoeven, 1968, has calculated variations of the critical Rayleigh number with the aspect ratio for a channel with finite length. The calculations were made assuming that the flow is antisymmetric, and that there is a negligible conduction of heat. The results indicated that the critical Rayleigh number increases from $R_{ct} = 68$ to $R_{ct} = 110$ as the aspect ratio is reduced from ∞ to 3.

3.4.3.2 Thermal Convection of Fluids in a Porous Medium

Conditions of stability of a thermally stratified fluid in a saturated, homogeneous, porous medium was given by Turcotte and Schubert, 1982. Such conditions were established assuming, that a) prior to the onset of convection, fluid is motionless; b) thickness of the porous medium is finite but the lateral extent of such a medium is large, so that boundary conditions along vertical planes do not need to be considered; and c) the porous medium is contained between two impermeable and isothermal boundaries. Under these conditions, the Rayleigh number is

$$R_{pa} = \frac{\alpha \cdot g \cdot \varrho_f^2 \cdot C_f \cdot K \cdot L \cdot \Delta T}{\mu \cdot \lambda m}$$

The critical Rayleigh number R_{pc} is related to geometry of the convection cell. A minimum value for this number is $R_{pc} = 4\pi^2$, and occurs for rectangular-hexagonal cells for which the aspect ratio is equal to 1.0.

Plate 3.4.3.2-1 presents conditions required for the onset of thermal convection $(R_{pc} = R_{pa})$ in a horizontal layer of porous material with permeability (K) and thickness (b). It is obvious that rather modest values of permeability are required for the onset of thermal convection, even in regions where the values of geothermal gradient are in a range of 20 to 30°C/km. The required values of permeability range from 1 millidarcy, for a layer with thickness b = 5 km, to 1 Darcy, for a layer with thickness b = 200 m.

Effects of vertical mass discharges on the stability of a thermally stratified fluid, in a saturated porous material and of large lateral extent, was considered by Homsy and Sherwood, 1976. The stability parameter for such a situation is the flow modified D'Arcy-Rayleigh number. This dimensionless number consists of two terms. The first term is the ratio of buoyant forces to dissipating forces, as expressed through the Rayleigh number (R). The second term, $\gamma \cdot \beta \cdot \Delta T$, expresses temperature variations of the fluid viscosity, coupled with the superimposed axial velocity. The flow modified D'Arcy-Rayleigh number is

$$R_1 = R + \gamma \cdot \beta \cdot \Delta T$$

Plate 3.4.3.2-2 presents a relationship between the critical D'Arcy-Rayleigh number (R^2) and the dimensionless through-flow strength parameter (γ). Stability limits obtained from both the linear stability considerations and the energy stability considerations are shown. The relationships indicate that the increased through-flow strength, either upflow or downflow [increasing $|\ \gamma\ |$], results in increasing the value of the critical D'Arcy-Rayleigh number (R^2). Physical interpretation of this observation is, that the mass discharge "compresses" temperature gradients in the fluid toward one or the other boundary. Consequently, the maximum temperature difference is felt over an increasingly smaller layer thickness. In other words, the effective thermal depth becomes an increasingly smaller fraction of the total layer thickness.

It is important to note, that even though the direction of through-flow does not reduce the value of stability limits, the direction of flow does substantially affect the overall stability of fluid. This occurs because the flow modified D'Arcy-Rayleigh number is related to the direction of mass discharge. Mass

discharge in the positive direction (upward flow) results in $\gamma > 0$ and, therefore, increases the D'Arcy-Rayleigh number, which is a destabilizing effect. The opposite is true for the downward flow. This differing effect is caused by the motion of thermally stratified fluid, and does not involve buoyant force. Such effects seem to be analogous to the Saffman-Taylor instabilities for displacement of immiscible fluids of differing viscosity. For the positive mass discharge, the fluid flow is from regions of low viscosity into regions of higher viscosity. In the case of immiscible fluids, this situation is known to be destabilizing relative to the opposite flow direction.

Stability limits for a thermally stratified fluid, with a small constant upward velocity, were also considered by Sutton, 1970. The analyses were made using the linear stability theory and assuming that flow is occurring in a planar vertical channel of width $(2a)$ and height (h). The channel walls were assumed to be adiabatic, while the top and bottom of a channel were assumed to be isothermal. Variation of the stability limit (R_c) with the initial velocity of flow, for different values of the aspect ratio $(s = a/h)$, was found. Such variation is shown on Plate 3.4.3.2-2. The critical Rayleigh number (R_c) is shown to be strongly dependent on the aspect ratio(s). For the through-flow velocity equal to zero, the value of R_c ranges from $4\pi^2$, for $s = 1.0$, to more than 250, for $s = 0.1$. As the velocity of flow increases, the stability limit (R_c) at first decreases, and then increases rapidly. For larger values of the aspect ratio, however, the initial decrease is small.

An important feature of thermal convection of fluids in porous materials of large lateral extends is, that for all practical purposes, it is a steady-state process. Several authors have shown, that when the Rayleigh number exceeds the critical Rayleigh number, the steady-state solutions always exist (Wooding, 1957; Donaldson, 1962; Elder, 1967; Palm et al., 1972; Strauss, 1974). In some special situations, however, quasi-static fluctuations occur (Caltagirone, 1975). In such situations, the actual Rayleigh number substantially exceeds, say one order of magnitude, the critical Rayleigh number.

3.4.3.3 Thermal Convection of Fluids in Faults and Fractures

For the purposes of conceptual considerations undertaken in this report, of particular relevance are analyses of thermal convection of fluids in faults and fractures. Such analyses were made by Beck, 1972; Zebib and Kassoy, 1977; Lowell, 1977; and Murphy, 1979.

Lowell, 1977, considered thermal convection of fluids in channels of finite vertical dimension, large lateral extent, and of a narrow aperture. The analyses were made assuming that walls of such channels are perfectly conducting; so that temperature distribution along these walls remains undisturbed by convection. Beck, 1972, and Zebib and Kassoy, 1977, assumed the vertical walls to be adiabatic. For the purposes of evaluating long-term behavior of geothermal systems, however, neither of such extremes in the boundary conditions is an acceptable simplification. This occurs because a) thermal conductivity of most rocks is roughly 5 times that of groundwater and b) wallrock conducts heat to and from convecting fluid at a variable rate (dependent upon amplitudes of the convective temperature perturbations, and upon duration that convection has been in effect).

The analyses of thermal convection of fluids in faults and fractures with consideration of heat transport, not only within convecting fluids but also through wallrock, were performed by Murphy, 1979. The main purpose of such analyses was to examine conditions required for an initially motionless fluid to either be set or set itself spontaneously into self-sustained circulation. The linear stability theory was used throughout the analyses.

For a given geothermal system, the Rayleigh number (R_a) is dependent upon fault/fracture permeability (K) and upon value of the vertical geothermal gradient (G). The value of R_a is

$$R_a = \frac{\alpha \cdot \varrho \cdot g \cdot h^2 \cdot K \cdot G}{\mu \cdot \lambda m}$$

The critical Rayleigh number (R_c) is a function of a) the aspect ratio (convection cell height-to-width ratio) and b) the "strength" of thermal interaction between convecting fluid and wallrock hosting the convection cell. Such relationships are shown on Plate 3.4.3.3-1. If wallrock is adiabatic (the heat transfer grouping is equal to 1.0), the critical Rayleigh number (R_c) is strongly dependent upon the aspect ratio. The minimum value for this number is $R_c = 4\pi^2$, and occurs for the aspect ratio equal 1.0. As the "strength" of fluid-wallrock thermal interaction increases (increasing value of the heat transfer grouping), the critical Rayleigh number also increases. The value of this limit, however, is influenced less and less by the geometry of convection cell.

The relationships presented on Plate 3.4.3.3-1 are useful in developing an understanding of the onset and subsequent behavior of thermal convection of fluids in faults and fractures. The process may be envisaged as occurring in three different stages: the late subcritical stage, the neutral stability stage, and the supercritical stage.

The late **subcritical stage** begins when the difference between the critical and the actual Rayleigh numbers $(R_c - R_a)$ attains a certain small value. Prior and during this stage, the critical Rayleigh number (R_c) is constant and independent of time. The actual Rayleigh number (R_a), however, increases with time. Such an increase is caused by some external factors which may involve a) temperature difference between top and bottom of a fault; b) permeability of a fault; c) upward through-flow of fluid or gas; and d) upward or downward propagation of a fault. In the late subcritical state, fluid is unstable with respect to perturbations of finite amplitude and duration, and involving fluid temperature, pressure, and velocity. Both the amplitude and duration of the perturbations required for the onset of thermal instabilities (i.e., exponential growth of the perturbation with time) are related to the difference $R_c - R_a$; with time, fluid becomes sensitive to smaller and smaller perturbations. Early in the process, when the difference $R_c - R_a$ is appreciable, the required perturbations must be produced by some external factors (e.g., an upward through-flow of gas or fluid, fault displacement, increase in flux of terrestrial heat, and lateral inflow of cold water at the top). With time, as the difference $R_c - R_a$ diminishes, amplitudes and durations of the required perturbations become smaller and, eventually, thermal instabilities may be produced spontaneously.

The **neutral stability** stage begins when the differnce $R_c - R_a$ approaches zero. During this stage, the critical Rayleigh number (R_c) becomes time-dependent. Murphy, 1979, has shown that the value of $R_{c_{(t)}}$ diminishes as a function of time$^{-\frac{1}{2}}$. This time-dependent decrease is caused by thermal interactions between fluid contained in "artificially" or spontaneously produced instability and wallrock of the incubating convection cell. Wallrock conducts heat from circulating fluid at a rate dependent upon a) duration that circulation has been in effect and b) amplitude of the resulting temperature perturbation. Thermal skin with a time-dependent thickness penetrates wallrock; so that heat must be conducted through increasing skin thickness. This, of course, causes a continuous reduction in the amount of heat lost from the incubating convection cell, and results in a continuous decay of the critical Rayleigh number. Eventually, R_a exceeds R_c and unstable growth of thermal instability commences.

The **supercritical stage** begins when the difference R_c - R_a is smaller than zero. At this stage, any perturbations in fluid temperature, pressure, and velocity will grow exponentially with time. The rate of such growth may be determined using the first-order small perturbation theory. As shown on Plate 3.4.3.3-2, dimensionless perturbation growth rate is dependent upon the Rayleigh number (R_a). At low values of R_a, the growth rate varies as the square of R_a, but at large values of R_a the relationship becomes more linear. Plate 3.4.3.3-2 also shows, that for a given value of R_a, the perturbation growth rate is dependent upon the aspect ratio of convection cell. The maximum value of the growth rate occurs for values of the aspect ratio ranging from 1 to 4. During the supercritical stage, thermal interactions between fluid of exponentially growing instability and wallrock have a significant stabilizing effect. As the size of the instability increases, the amount of heat lost into wallrock also increases. Consequently, the average temperature gradient within the system decreases, causing an overall reduction in the thermal energy that fuels convection. Reacting to such developments, the critical Rayleigh number (R_c) increases and the Rayleigh number (R_a) decreases. This provides a limit for the maximum value of the growth rate and, therefore, for a size of the mature convection cell. When $R_c = R_a$, the growth rate attains, at first, a constant value. From this point on, however, convection dampens and the value of growth rate diminishes with time.

Two outcomes are possible for the subsequent state of affairs. The first outcome is that the growth rate is equal to zero, and the convection continues at steady-state. In this situation, heat losses from the system are being compensated by heat conduction from the far-field. A delicate balance exists between the forces fostering and resisting convection. Any meaningful change in such balance either encourages (positive growth rate) or discourages (negative growth rate) this process. Reacting to various stimulations, the geothermal system continuously searches for new equilibrium positions. It is not difficult to imagine, that in an active tectonic environment, there are numerous opportunities to continuously stimulate the convection cell which tries to operate at steady-state conditions. Change in permeability and/or in effective thermal conductivity, through mineral deposition or fault creep; change in upward flow of gas or fluid, resulting from volcanic or seismic activity; change in flux of terrestrial heat, resulting from volcanic or seismic activity; and change of inflow of cold water at the top of a convection cell - are examples of such opportunities. Consequently, it is rather difficult to think about convection of fluids in faults and fractures as a steady-state process. Instead, the transient nature of such a process is apparent.

The second possible outcome is that the growth rate becomes negative and, eventually, convective circulation of fluids ceases. In this case, depletion of thermal energy was not compensated by heat conduction from the far-field. The cessation of convection, however, is only temporary. It is followed by a period of thermal recovery, during which heat conduction from the far-field restores the original state. Eventually, the whole process repeats itself; so that the characteristically cyclic nature of thermal convection of fluids in faults and fractures becomes apparent.

3.4.3.4 Summary and Conclusions

Based on discussions presented above, the following three conclusions can be drawn. First, rather modest values of the geothermal gradient are required for the onset of thermal convection in a porous medium whose permeability exceeds few millidarcies. Requirements for the onset of thermal convection in a fractured medium are strongly dependent on a) aspect ratios of convection cells that are possible in this medium; b) "strength" of thermal interaction between fluid and wallrock; c) permeability of a fractured medium; and d) value of the vertical geothermal gradient. For a wide fault zone with high permeability, however, such requirements are not extraordinarily stringent. If such fault zones are situated in regions where a) values of the vertical geothermal gradient range from 30 to 60°C/km and b) wallrock consists of a material whose thermal conductivity is low; then, it is reasonable to expect that thermal convection of fluids in faults and fractures is possible.

Second, thermal convection of fluids in faults and fractures is characteristically a cyclic and transient process. Thermal interactions between convecting fluid and wallrock have a significant effect on this process. In the case of a marginally stable system (in the state of neutral stability), such interactions cause the critical Rayleigh number to be time-dependent. Its value diminishes as time increases. With time, therefore, even an initially subcritical system becomes unstable and will experience exponential growth. As thermal energy fueling the system is locally depleted and isothermal state is approached, the instability dampens and, eventually, ceases until a period of thermal recovery allows its resumption.

Third, a geothermal system operating in a fault zone, regardless of its actual stability state, is quite sensitive to various stimulations resulting from volcano-tectonic activity. Both the critical Rayleigh number and the actual Rayleigh number are sensitive to such stimulations. Consequently, it may be expected that tectonic stimulations, occurring in one form or another, may cause substantial restructurings of a

geothermal system. Such restructurings may involve a) geometry of a geothermal system; b) growth rate of a thermal instability; and c) transformation of a geothermal system from one stability state into another.

3.4.4 Conceptual Model of a Groundwater System Developed in a Fractured Medium and Influenced by Terrestrial Heat

Plate 3.4.4-1 presents a conceptual model of a coupled heat-fluid groundwater system developed in a fractured medium. In this model, both vertical boundaries are assumed to be adiabatic with respect to heat flow, and "no flux" or "const. head" boundaries with respect to fluid flow. The hydraulic pressures acting at both vertical boundaries differ by a constant amount, which is represented by the term cx. Both horizontal boundaries are considered as "heat-flux" boundaries. The upper boundary is also a "fluid-flux" boundary. The lower boundary, however, is considered as a "no fluid-flux" boundary.

The upper horizontal boundary is situated at a shallow depth (Z_o); so that temperature along it does not differ significantly from a mean annual temperature at the ground surface. The lower horizontal boundary is located at a considerable depth (Z_1), say at a depth of approximately 15 km. This boundary contains a distributed heat source, which is represented by the term $R_{hZ_{1(x)}}$. The intensity of this source varies spatially from 1.5 HFU to about 2.5 HFU $(\text{HFU} = 1.0\ \mu\text{cal/cm}^2\text{s})$, but is independent of time. Along the lower boundary, the difference in temperature is substantial, in a range of 150 to 250°C. The average temperature difference, between the lower and upper boundary, is also substantial, in excess of 300°C.

Coupled heat-fluid flow occurs in a strongly fractured medium, whose permeability ranges from a few darcies to a few millidarcies, or less. Wide zones of high permeability are associated with important faults. With depth, however, there is a tendency for the overall permeability to decrease, and for the zones of high permeability to become narrow and less frequent. A strongly heterogeneous temperature field causes the gradients of temperature not to be colinear with the gradients of hydraulic potential; so that fluid buoyancy gradients are considerable relative to hydraulic head gradients. This causes the dominant mode of transport of terrestrial heat to be convection.

An important feature of the model is a space- and time-dependent depth $Z_{c_{(x,t)}}$. This depth defines a surface $Z_{c_{(x,y,t)}}$ that divides the flow system into two parts, each with different character of coupled heat-fluid flow. In the lower part, overall permeability is low and zones of higher permeability are narrow. Consequently, the aspect ratios of potential convection cells in faults and fractures are small and, therefore, the values of critical Rayleigh number are large (Plate 3.4.3.3-1). Coupled heat-fluid flow is assumed to be

occurring as a "porous medium equivalent" flow and, consequently, convection of fluids occurs as a steady-state process (Section 3.4.3.2). Thermal interactions between convecting fluids and rock hosting convection cells do not affect such a process. The horizontal and vertical temperature gradients, therefore, are independent of time $[\partial^2 T_{m_{(x,z,t)}}/\partial_x(\partial_z)\partial t = 0]$.

In the upper zone, however, overall permeability and permeability contrasts are greater, and zones of high permeability are becoming increasingly wider with a shallower depth. Requirements for the onset of thermal convection of fluids in faults are less stringent (Section 3.4.3.3). Consequently, it is assumed, that coupled heat-fluid flow occurs in the form of convection in faults and, therefore, is a transient process. Thermal interactions between convecting fluids and wallrocks control such a process, and cause the lateral and vertical gradients of temperature to be time-dependent $[\partial^2 T_{m_{(x,z,t)}}/\partial_x(\partial_z)\partial t \neq 0]$.

Space-dependence of the surface $Z_{c_{(x,y,t)}}$ is the result of two characteristics of the flow system. The first characteristic is a spatial variability of the hydraulic conductivity structure, which is represented by the term $K_{f_{(x,z)}}$. The second characteristic is a maximum value of the vertical geothermal gradient, that is possible at a given point and at a time when a geothermal system is in the neutral stability state. Such a maximum value is represented by the temperature difference $T_{z_{c(x)}} - T_{z_{o(x)}}$. This difference is related to a) local intensity of the heat source $R_{hZ1_{(x)}}$ and b) local values of the effective thermal conductivity $K_{h_{(x,z)}}$.

Time-dependence of the surface $Z_{c_{(x,y,t)}}$ is caused by the transient and cyclic character of convection of fluids in faults. During the subcritical and neutral stability states, when the temperature difference $T_{z_{c(x)}} - T_{z_{o(x)}}$ attains its maximum value, the depth $Z_{c_{(x,t)}} = 0$. The maximum value for the depth $Z_{c_{(x,t)}}$ occurs during the supercritical state. Such maximum value may be attained either immediately upon initiation of the exponential growth or, with time, in response to various tectonic stimulations (Section 3.4.3.3).

The above conceptual model is useful in developing a general understanding of long-term behavior of groundwater systems that involve coupled heat-fluid flow in a fractured medium. Such an understanding is essential for informed judgments concerning potential magnitudes of hydrologic disturbances that are possible in these systems.

The analyses presented below have been developed assuming that during a period of interest: a) there is no change in climatic conditions and, therefore, volume of fluid contained within the groundwater system remains constant; b) initial boundary conditions for the overall system remain unchanged; c) apart from minor fracturings caused by local faulting, the hydraulic conductivity structure of the system remains unchanged; and d) apart from fluid-wallrock thermal interactions, the effective thermal conductivity structure of the system also remains unchanged. A relatively small portion of the system was selected for the analyses. The location of such a portion is shown on Plate 3.4.4-2. The size of it, relative to the overall groundwater system, is small. The depth extent is no more than a few kilometers, and the lateral extent is measured in several kilometers. It is assumed, that initially, the vertical and horizontal boundaries are "constant flux" boundaries, in time and space, for both heat and fluid. Consequently, the initial volume of fluid in transit through the selected portion is Q_{tr}^{in} = const. The assumption of "constant flux" boundary conditions is not an important one, and does not affect the end result; it was made solely to avoid cumbersome notations and drafting.

Plates 3.4.4-3 and 3.4.4-4 present a transformation of the boundary conditions resulting from the supercritical growth of thermal instability in the selected portion of a groundwater system. Both vertical boundaries (x_1 and x_2) and the lower horizontal boundary are changed from the "const. flux" boundaries to the "time- and space-variable flux" boundaries. Furthermore, the rates of flux change are different for all three boundaries and are also time-dependent. This is caused by three factors: a) the time-variable rate of growth of thermal instability ; b) the space-variable distribution of thermal and hydraulic properties; and c) the space-variable distribution of the initial gradients of hydraulic potential and temperature.

Time-dependent changes in the boundary conditions, for all three boundaries, affect a local groundwater system. The impact is expressed in three-fold manner: a) change in the initial gradients of temperature; b) change in the initial volume of fluid in transit; and c) change in the initial gradients of hydraulic potential. Plate 3.4.4-5 presents an idealized history of changes in all of these parameters occurring at a single point $P_{(x,z)}$. The onset of supercritical growth of thermal instability is associated with increases in the values of $T^c_{m_{(x,z,t)}}$, $Q^c_{tr_{(x,z,t)}}$, and $h^c_{(x,z,t)}$. When the growth rate is positive, the increases are progressive and non-linear with time. As the fluid-wallrock thermal interaction exerts its stabilizing influence, the growth rate becomes constant, at first, and eventually equal to zero. At this time, the values of $\Delta T^c_{m_{(x,z,t)}}$, $\Delta Q^c_{tr_{(x,z,t)}}$, and $\Delta h^c_{(x,z,t)}$ attain their maximum. As long as the geothermal system operates at steady-state, the values of these parameters remain fixed. During the period of decay of thermal insta-

bility, the growth rate is negative, and the values of all three parameters begin to return to their initial values.

Results of the above analyses indicate that thermal instabilities of fluids in faults, being a transient and cyclic process, introduce important time- and space-variable heterogeneities into local portions of a groundwater system. It is reasonable to expect that such heterogeneities will be reflected in position and in configuration of the water table overlying them. In areas underlaid by hydraulic and heat sources, the water table tends to bend upward to form a hydraulic mound. In contrast, in areas occurring over hydraulic and heat sinks the water table bends downward. Such departures introduce further complications into shallow parts of the groundwater system by damming and otherwise interfering with the initial flow. The resulting configuration of the water table may be very complicated, and may include local mounds, sinks, and sharp steps.

At this point, an important question emerges: **How large are thermal rises of the water table**? Some answers to this question may be provided by relationships shown on Plate 3.4.4-6.

The maximum potential rise of the water table, at a point $P_{(x,y)}$, is the sum of two components. The first component, $\Delta h^{cb}_{(x,y)}$, arises from thermal expansion of convecting fluid. The magnitude of this component is related to three factors: a) coefficient of thermal expansion of water $(\alpha_w = 3.85 \cdot 10^{-4}$ per °C); b) maximum average change in water temperature at point $P_{(x,y)}$; and c) depth extent of convection cell at point $P_{(x,y)}$. For example, if the depth extent of the convection cell is 5 x 10^3 m, and the average change in water temperature is 30°C, then the "buoyant" rise of the water table at point $P_{(x,y)}$ is $\Delta h^{cb}_{(x,y)} \cong 60$ m.

The second component, $\Delta h^{ch}_{(x,y)}$, arises from the increased volume of fluid in transit through an area occupied by the convection cell. The magnitude of this component is again related to three factors: a) the ratio of maximum volume of fluid in transit during convection to the initial volume of fluid in transit - max Q^c_{tr}/Q^{in}_{tr}; b) the initial value of hydraulic gradient (i_{in}), between point $P_{(x,y)}$ and a closest reference point downstream for which $Q^c_{tr} = Q^{in}_{tr}$; and c) lateral distance (L), between point $P_{(x,y)}$ and the downstream reference point. For example, if the initial hydraulic gradient is $i_{in} = 10$ m $\cdot$ km^{-1}, the lateral distance between point $P_{(x,y)}$ and the reference point is $L = 5$ km, and the ratio max $Q^c_{tr}/Q^{in}_{tr} = 2$, then the

potential "hydraulic" rise of the water table at point $P_{(x,y)}$ is $\Delta h^{ch}_{(x,y)} = 100$ m. The total potential thermal rise of the water table at point $P_{(x,y)}$ is $\Delta h^{c}_{(x,y)} = 60$ m + 100 m = 160 m.

Although it is rather difficult to judge the values of parameters controlling the magnitude of "thermal" rise of the water table, for a particular field situation, it is rather obvious, that potentially, such magnitudes may be very substantial. This means, that in a coupled heat-fluid flow system, there is a potential for a special relationship between the vadose zone and the zone of saturation. **These two zones may supply water to each other.** The introduction of water from the saturated zone into the vadose zone occurs in association with a thermally induced hydraulic mound. Such a mound grows until the volume of fluid flowing into it equals the volume of fluid flowing out of it. For a mound that has developed sufficient dimensions to reach the ground surface, the removal of water involves surface flow and evaporation (seep, spring, etc.). In the case of less developed mounds, however, the removal of water must involve subsurface flow through the vadose zone, and leads to the establishment of perched and semi-perched bodies of water. The formation of such bodies is shown schematically on Plate 3.4.4-7. It can be noted, that the development and subsequent preservation of the perched waters is related to a subhorizontal stratum composed of a material whose interstitial porosity is high but stiffness is low; so that fracture permeability within such a stratum is low. The gravitational flow of fluid from the mound into the subhorizontal stratum causes this stratum to become saturated. From this time on, the stratum retards a downward infiltration of meteoric water from recent precipitation.

It is important to recognize, that the bodies of perched water owing their existence to thermal upwellings, always contain a mixture of "old" meteoric water from below the water table and "young" meteoric water directly from infiltration of atmospheric precipitation. Relative proportions of these two constituents are variable with time, and are related to two factors: a) intensity of local precipitation and resulting infiltration and b) length of time a given body of perched water was in existence. In this situation, it should not be difficult to understand why for some old thermally induced bodies of perched water their origin may be subtly expressed, and, at first, imperceptible to an unaware observer. Such bodies may appear as "ordinary" perched waters, the origin of which has nothing to do with thermal upwellings of the water table. Chemical and isotopic compositions of the water contained within such bodies, together with their location within and over a stratum with low hydraulic conductivity, will favor a superficial and incorrect interpretation that their origin is not related to the past upwellings of the water table.

3.4.5 "Mixed" vs. "Forced" Convection Systems: Why is it Important to Tell Them Apart?

A **"forced" convection system** is a system for which spatial variability of the hydraulic conductivity structure is the main factor accounting for the convective aspects of groundwater flow. The system is thermally "passive." Heterogeneities in the geothermal field are only expressions, but not the cause, of the convective nature of flow. In such a system, configuration and position of the water table may constitute a meaningful and durable expression of characteristics of this system. Such characteristics are a) spatial distribution of hydraulic properties; b) regional hydraulic gradients (i.e., altitudes of the recharge areas relative to the discharge areas); and c) total volume of water contained within the flow system.

A **"mixed" convection system** is a system where thermal energy plays a dominant role in establishing the convective character of groundwater flow. Such a system is thermally "active." Heterogeneities in the geothermal field are not only the expressions, but are also the cause, of the convective flow of heat and fluid. Indirectly, such heterogeneities reflect fundamental tectonic processes operating deep within the Earth's interior. For the "mixed" convection system, the water table is not necessarily a permanent feature. As discussed in Sections 3.4.3 and 3.4.4, in the "mixed" convection system, local boundaries are either the "space-variable flux" boundaries, for porous or "porous equivalent" flow, or the "space- and time-variable flux" boundaries, for fracture flow. As the flux conditions along these boundaries change, either spontaneously or in response to tectonic stimulations, position and configuration of the water table change accordingly.

It is important to reckon that differences in the appearances of "mixed" and "forced" convection systems may be subtle and hardly discernible to a casual or non-inquisitive observer. In most situations, it is possible to explain a given data set obtained from a "mixed" convection system by assuming that a "forced" convection system is involved, and vice versa. This troublesome flexibility exists because the typical hydrologic data base is a) fragmentary and limited in scope and b) covering a relatively short period of time.

What sets the "mixed" convection system apart from the "forced" convection system is its long-term stability. If the end result of hydrologic inquiries is the assessment of such stability, reliable distinction of one system from the other is of great conceptual importance. In this context, dismissal of a given data set as being explainable by assuming the involvement of "forced" convection does not appear as a sound and

prudent practice. Instead, such practice appears as a good example of circular logic, whereby the statement "it is, because we have assumed it to be" is possible.

In performing numerical simulations of groundwater systems for various purposes, a traditional approach has evolved that is not particularly sound but which, nevertheless, is firmly entrenched in usage. This approach amounts to adoption of a somewhat arbitrary plane as the base of a groundwater system and assuming "no-flux" boundary conditions for such a plane. This assumption is reasonable, but only for the "forced" convection system. In the case of "mixed" convection system, however, this assumption is not valid. "Space-variable flux" boundary conditions must be used for the base of such a system. Furthermore, these boundary conditions pertain to both heat and fluid, and in the case of thermal convection in faults, may be time-dependent. Mathematical models must account for such circumstances, otherwise a groundwater system is misrepresented. Calculations and interpretations based on wrong boundary conditions and based on incorrect mathematical representations may be grossly misleading and irrelevant.

The difference in the boundary conditions causes the response of a "mixed" convection system to tectonic disruptions to be far more dramatic than that of a "forced" convection system. In "forced" convection systems, the influence of tectonic disruptions is limited to a) alteration of litho-stratigraphic relationships; b) additional local fracturing; and c) change in altitude of the discharge areas relative to the recharge areas. A degree of change in all of these factors is relatively small and, therefore, the resulting alterations in the hydraulic conductivity structure and in the distribution of hydraulic potentials are likewise small. Such alterations may readily be accommodated by minor changes in position and configuration of the water table.

Tectonic disruptions occurring within "mixed" convection systems, however, involve an additional factor. Such a factor is the increased flux of heat and fluid through the lower horizontal boundary. If significant quantities of mass and energy are involved, considerable and long-lasting alterations of the water table are to be expected. Changes in position of the water table may be introduced by changing climate, regardless of whether "forced" or "mixed" convection systems are involved. Climatic changes will cause gradual increases or decreases of the volume of water passing through vertical boundaries of the system. Reaction of the water table to such changes is somewhat uniform. Its configuration is not altered in any appreciable degree, although its position may change considerably.

It is common for climatically induced changes in the water table to be recorded by occurrences, at or near the ground surface, of characteristic geologic materials. Compositions, textures, and morphology of such materials indicate that they were formed in association with ponds, seeps, and paleo-springs. A map pattern of occurrence of these deposits defines a surface whose configuration resembles configuration of the contemporary water table.

Surface expulsions of groundwater, caused by tectonic disruptions in a "mixed" convection system, will likewise be recorded by similar deposits. The map pattern of occurrence of such deposits, however, does not display any fixed relationship to the contemporary water table. Furthermore, the ages of such deposits do not bear a clear relationship to the history of global climatic changes.

In summary, reliable and unequivocal determination that a "mixed" convection system is either present or absent is of great conceptual importance. Confusing this system with a "forced" convection system may entail serious errors in judgment regarding past and future patterns of behavior of a groundwater system. It is essential, therefore, that hydrologic considerations performed for the purposes of developing a high-level nuclear waste repository are free of such errors.

3.4.6 "Forced" vs. "Mixed" Convection Systems: How to Tell the Two Apart?

Short of having at one's disposal information indicative of dramatic manifestations of geothermal conditions, such as geysers and superheated water or vapor dominated systems, it is rather difficult to distinguish between "forced" and "mixed" convection systems. At or near steady-state conditions, hydrologic characteristics of both systems are similar. These characteristics include a) complex spatial distribution of hydraulic potentials; b) complicated configuration of the water table; c) considerable degree of thermal heterogeneity; and d) considerable degree of variability of groundwater chemistry.

Apart from a degree of transient behavior, either spontaneous or during tectonic disruptions, there does not appear to be a unique and readily available hydrologic characteristic that could be used to distinguish one system from the other. The problem, however, may be solved by relying on "circumstantial" evidence as provided by two other disciplines of geoscience: geology and geophysics. In this context, the following two questions are of particular relevance: a) Does the geologic record contain expressions of transient behavior of a groundwater system during Plio-Quaternary time, that are compatible with a "mixed" convection system? and b) Are geologic and geophysical characteristics of a region compatible with the assumption of a "strong" and/or heterogeneously distributed thermal source at deep crustal levels? The information required to answer these questions includes the following: a) volcanic history of the area; b) crustal structure; c) contemporary- and paleo-thermal regimes; d) character (textures, composition, morphology) and location of Plio-Quaternary deposits that could be related to springs and seeps; and e) character of Plio-Quaternary materials in the subsurface (fragmentation breccias, veins, metasomatites, and alteration products).

The question of possible involvement of a "mixed" convection system may be further examined based on the results of involved hydrologic studies directed toward some contemporary features of a groundwater system. Such an approach, however, may not yield a clear-cut answer unless the studies are targeted toward youthful features. For a system containing a thick vadose zone, three features of the contemporary water table are noteworthy. As shown on Plate 3.3.4-7, such features are a) sharp, linear "steps"; b) hydraulic mounds; and c) bodies of perched and semi-perched waters.

The data base required for serious interpretations of the origin of each of the above features and, thereupon, for the assessments of their conceptual importance, is very substantial indeed. Ideally, such

a data base should include the following categories of information: a) gradients of hydraulic potentials; b) gradients of temperature; c) gradients of chemical and isotopic compositions of groundwater; d) spatial distribution of hydraulic and thermal properties; and e) spatial variability of mineral content of the host rock. Furthermore, considerable care should be devoted to interpretations of such a data base. A sharp "step" in the water table may indicate the presence of a) vertical lithologic hydraulic barrier; b) lateral heterogeneity in the *in situ* stress field; or c) thermal upwelling of groundwater. The hydraulic mound aspects of a local configuration of the water table may be the result of a) locally increased precipitation and resulting infiltration; b) injection of fluid, from below the water table, in association with a faulting event; or c) thermal upwelling of the water table. A body of perched water may owe its existence to a) local retardation of infiltrating "young" meteoric water; b) local retardation of fluid injected, from below the water table, in association with a faulting event; or c) local retardation of fluid from a thermal rise of the water table.

In dealing with all of the above possibilities, it is easy to confuse a merely plausible interpretation with a correct one. This is true, in particular, if the interpretations are made insisting that the following factors do not need to be considered: a) full range of conceptual possibilities; b) geothermal character of an area as indicated by geologic and geophysical data and observations; c) possible history of performance of a groundwater system as suggested by geologic record; and d) significance of such interpretations in the context of long-term performance of the proposed high-level nuclear waste repository.

3.5 CONCEPTUAL CONSIDERATIONS OF A COUPLED HEAT-FLUID GROUNDWATER SYSTEM DEVELOPED IN A DEFORMING FRACTURED MEDIUM

3.5.1 General

Examination of Plates 3.3.1-2, 3.4.1-1, and 3.4.1-3, one against the other, reveals an interesting possibility: at and near the Nevada Test Site, volcanism and faulting are related to a common cause. Such a possibility is suggested by contemporaneity of both of these processes, and was noted previously by several authors. This possibility is also suggested by an apparent spatial relationship between the location of major faults, the location of Plio-Quaternary volcanic centers, and the location of a low-seismic velocity trough in the crust.

In light of such possibility, the main tectono-physical characteristics of the Yucca Mountain groundwater system (crustal stresses, resulting dislocations, and flow of terrestrial heat) may be attributed to a deep thermal anomaly. Such anomaly may be responsible for the abnormally low values of the P-wave velocity in the lower crust and upper mantle. This, together with isotopic and geochemical characteristics of the local basalts suggests, in turn, that the anomaly is related to a convective flow of heat and mass in the upper mantle (Sections 3.1 and 3.4.1). From a tectonic perspective, therefore, bulk structure of the area may be regarded as **an immature rift structure**.

The term "rift" is used here to denote a large-scale geologic structure whose origin is related, either directly or indirectly, to processes operating in the upper mantle underlying it (Beloussov, 1969). No regional or continental importance of any kind is implied by the usage of this term. It is possible that the structure of the area is a rift *sensu stricto*, that is formed in direct response to the underlying convective system in the upper mantle. It is also possible, however, that involvement of the upper mantle is somewhat "passive," and that overall structure of the area is predominantely related to plate-tectonic interactions occurring along the western margin of the North American continent. If this is the case indeed then bulk structure of the area may be regarded as a "divergent rift" or a "pull-apart" basin. In either case, however, involvement of the upper mantle amounts to simultaneous introduction of both thermal energy and strain energy into upper parts of the overall system. Consequently, further resolution of the issue of structural character of the area is a moot point of very limited conceptual importance.

For the purposes of understanding a full range of potential hydrologic implications resulting from the tectono-physical setting of the Yucca Mountain groundwater system, it is necessary to explore mutual relationships between the groundwater system, the *in situ* stress field, and the geothermal field. This is achieved through considerations of a coupled heat-fluid groundwater system in a deforming fractured medium.

3.5.2 Geothermal Field Developed in a Deforming Fractured Medium

Before considerations of a coupled heat-fluid groundwater system developed in a deforming fractured medium may be performed, it is necessary to understand the relationship between a geothermal field and a changing *in situ* stress field. To this end, and for the sake of simplicity, it is assumed that such a relationship is one-sided. The geothermal field is influenced by the *in situ* stress field, but not the former by the latter. It is further assumed that a deforming fractured medium contains in pores and fractures only thermally "non-convecting", but flowing and heat transporting fluids; so that the flow of terrestrial heat involves conduction and "advection", but does not include Rayleigh instabilities.

Such idealization of the geothermal field is not a realistic description of actual circumstances. Changing *in situ* temperature will have a significant influence on the *in situ* stress field by a) causing volumetric strain of a medium and b) altering long-term stress-strain relationship for this medium. Also, it is rather difficult to imagine how the assumption of "non-convecting" fluid may be valid for real circumstances. For the purposes of these conceptual considerations, however, both of these assumptions do not entail serious consequences.

The law of heat conduction, in one dimension and at steady-state, may be expressed through the following equation:

$$q_h = -K_h \cdot \frac{\Delta T}{\Delta L}$$

where:

q_h - heat flux;

K_h - thermal conductivity of a medium; and

$\frac{\Delta T}{\Delta L}$ - thermal gradient.

In a fractured medium, the flow of heat consists of three components: a) heat flow through solid; b) heat flow through pore space; and c) heat flow through fracture space. The total heat flow may be expressed as the sum of these three components, so that

$$q_h = q_1 + q_2 + q_3$$

Thermal conductivity of a fractured medium (K_h) may be regarded as again related to three factors. Such factors represent thermal conductivity of solid (K_{hs}), of pore space (K_{hp}), and of fracture space (K_{hf}). It is reasonable to assume that two of these factors, K_{hs} and K_{hp}, are independent of the *in situ* stress and, therefore, are also independent of time, even for the changing *in situ* stress field. Such assumption, however, cannot be justified for the factor that represents fracture space (K_{hf}). Adopting the analogy of fluid flow, this factor may be regarded as related to conducting aperture $\left[K_{hf_{(t)}} = f(a_{con_{(t)}})\right]$.

By performing analyses similar to those presented in Section 3.3.3, it can be shown that K_{hf}, and with it the effective thermal conductivity and diffusivity of a deforming fractured medium, is stress- and time-dependent. Such relationships are caused by the *in situ* stress enhanced "strength" of through-flow of groundwater. This causes the apparent or effective thermal conductivity (K_{heff}) of a deforming fractured medium to increase (Prats, 1982). Additionally, the increasing value of conducting aperture causes the increase in fluid storativity. This, in turn, causes the bulk volumetric heat capacity to increase. Consequently, for a deforming fractured medium, the effective thermal diffusivity (D_{eff}) may be regarded as related to the conducting aperture, and increasing with it. The effective thermal diffusivity is defined as the ratio of the effective thermal conductivity to the volumetric heat capacity.

Plates 3.5.2-1 and 3.5.2-2 are schematic representations of time-dependence of both thermal properties for a deforming fractured medium. The relationships shown were developed recognizing that both thermal properties are related to the conducting aperture, which in turn, is controlled by the changing *in situ* stress field.

Plate 3.5.2-1 shows, that in a deforming fractured medium, the *in situ* values of both thermal properties undergo time-dependent enhancement in response to the changing *in situ* stress field. Initially, when the shear stress (τ_f) is smaller than frictional resistance of fractures $\left[\tau_{\max} = C + \sigma_{nf} \cdot \tan\phi\right]$, conducting apertures are relatively small and consist solely of the residual component $\left[a_{con} = a_{res}\right]$. At this time, therefore, the values of both thermal properties are independent of time. When the "limit equilibrium" stress conditions are established, however, conducting apertures become time-dependent, and their values

are equal to $a_{con_{(t)}} = a_{res} + a_{d_{(t)}}$ or $a_{con_{(t)}} = a_{res} + a_{d_{(t)}} + a_{n_{(t)}}$. This causes time-dependent enhancement of the *in situ* values of thermal properties.

Plate 3.5.2-2 presents a relationship between time and depth of occurrence of the surface $Z_{(x,y,t)}$. This surface denotes the bottom of a portion of the deforming fractured medium that experiences time-dependent enhancement of the thermal conductivity structure. With time, the depth extent of such portion increases, reflecting occurrence of the "limit equilibrium" stress conditions in progressively deeper parts of the geothermal field. Plate 3.5.2-2 also presents a relationship between time, space, and magnitude of the *in situ* stress induced enhancement of the thermal conductivity structure. In a deforming fractured medium, the magnitudes of such enhancement increase with time and are space-dependent. The greatest magnitudes occur at the ground surface, where frictional resistance of fractures $(\tau_{\max})$ is least. With depth, magnitudes of the enhancement diminish, becoming equal to zero along the surface $Z_{(x,y,t)}$.

Plate 3.5.2-3 presents an idealized history of changes in the *in situ* values of thermal properties for a cyclically deforming fractured medium. The history is shown in the form of a plot that would result from a simultaneous monitoring of the *in situ* stress and the *in situ* values of thermal properties, at a single reference point, and throughout three sequential cycles of tectonic deformation. For each of the cycles, beginning at a time when the "limit equilibrium" stress conditions are established, there is an initially progressive increase in the *in situ* values of thermal properties. At more advanced stages of deformation, the *in situ* values of thermal properties are assumed to attain some quasi-const. values. The geothermal system remains in the enhanced state until the end of a tectonic cycle. At this time, there is an abrupt termination of the "limit equilibrium" stress conditions. Conducting apertures diminish to their residual values and, consequently, the thermal conductivity structure returns to the initial state. The periodic enhancements of the structure occur over and over again, provided cyclic deformation of a deforming fractured medium continues.

Plate 3.5.2-4 presents a conceptual model of a geothermal field developed in a deforming fractured medium. In this model, both vertical boundaries are considered as adiabatic boundaries. Both horizontal boundaries, however, are considered as heat conducting boundaries. The upper boundary (Z_0) is isothermal, whereas the lower boundary (Z_1) contains a distributed heat source. The intensity of such a source is given by the term $R_{hZ_{1(x)}}$, and is regarded as independent of time. Only "non-convecting" fluids are

present and, consequently, the flow mechanism for terrestrial heat does not involve thermal convection (i.e., Rayleigh instabilities).

An important feature of the model is the time- and space-dependent surface $Z_{(x,y,t)}$. Time-dependence of this surface is the result of ongoing tectonic deformation of the geothermal field. Below the surface $Z_{(x,y,t)}$, both of the thermal properties are independent of *in situ* stress and, therefore, are independent of time. Flow of terrestrial heat is at steady-state, and as such may be represented by Poisson's equation. Above the surface $Z_{(x,y,t)}$, however, the thermal properties are related to the *in situ* stress and, therefore, are time-dependent. Flow of terrestrial heat occurs as a transient process and should be represented by the equation for non-equilibrium flow.

The law of conservation of energy requires that the sum of thermal energy contained in the geothermal system and thermal energy lost from this system is constant in time and equal to the thermal energy provided by the distributed heat source $R_{hZ_{1(x)}}$. Thermal energy lost from the geothermal system is represented by the flux of terrestrial heat through the upper horizontal boundary.

The above conceptual model may be used to deduce main characteristics of a geothermal field developed in a deforming fractured medium. Such characteristics are shown on Plate 3.5.2-5. The first characteristic is that the geothermal field contains spatially variable lateral and vertical gradients of temperature $[\partial temp/\partial x(\partial y)(\partial z) \neq \text{const.}]$. Such gradients are caused by heterogeneities in the *in situ* stress field which cause the development of corresponding heterogeneities in the spatial distribution of thermal properties.

The second characteristic is time-dependence of the geothermal gradients. Above the surface $Z_{(x,y,t)}$, the effective thermal conductivity of a deforming fractured medium increases progressively. This causes the heat flux through the upper boundary of the geothermal system to increase. The balance of thermal energy, however, is maintained by reducing the overall amount of thermal energy stored within the system. Such a reduction is accompanied by a) time-dependent decrease in the values of vertical geothermal gradients $[\partial^2\text{temp}/\partial z\partial t \neq 0]$ and b) time- and space-dependent downward migration of isothermal surfaces $[\partial^2\text{temp}/\partial x(\partial y)(\partial z)\partial t \neq 0 \neq \text{const.}]$. Both of these effects are shown on Plate 3.5.2-5.

It is important to note that it is rather difficult to obtain accurate measurements of the intensity of flux of terrestrial heat based on geothermal measurements performed in boreholes that are too shallow

to penetrate the surface $Z_{(x,y,t)}$. Putting aside hydrologic problems and heat transfers caused by lateral geothermal gradients, translations of the measured values of vertical geothermal gradients into intensities of flux of terrestrial heat require knowledge of the **effective thermal conductivity**.

It is standard practice to perform measurements of thermal conductivity on core samples or drill cuttings, using either the Birch-type divided-bar apparatus or the needle probe technique (Sass et al., 1971). Because of the small size of specimens involved in such measurements, however, it is unlikely, that if a deforming fractured medium is present, a true *in situ* value of the effective thermal conductivity is obtained. The actual value of the effective thermal conductivity may be greater than the value of thermal conductivity obtained on the basis of laboratory measurements. Consequently, the intensity of flux of terrestrial heat may be underestimated by a considerable amount, in a range of 50 to 100 percent (Section 4.3.2).

3.5.3 Coupled Heat-Fluid Groundwater System Developed in a Deforming Fractured Medium

Plate 3.5.3-1 presents a conceptual model of a coupled heat-fluid groundwater system developed in a deforming fractured medium. In this model, both vertical boundaries are considered as adiabatic with respect to heat flow, and "no flux" or "const. head" boundaries with respect to fluid flow. Hydraulic pressures acting at these boundaries differ by a constant amount, which is represented by the term cx.

Both horizontal boundaries are considered as "fluid-flux" and " heat-flux" boundaries. The upper boundary (Z_0) is isothermal and isobaric. The lower boundary (Z_1) contains a distributed heat source, represented by the term $R_{hZ_{1(x)}}$, and a distributed fluid source, represented by the term $R_{fZ_{1(x)}}$. The intensities of both of these sources are space-dependent, but they are independent of time.

Mass and energy transfer occurs as a simultaneous flow of heat and fluid under the influence of two interdependent energy gradients. Heat flow involves both convection and conduction, while fluid flow occurs in fractures. The groundwater system is developed in a deforming fractured medium. Deformation of such a medium occurs in the form of repeated cycles of tectonic deformation (Section 3.3.2). The deformational system contains the time- and space-dependent surface $Z_{(x,y,t)}$. Locally above this surface, either the shear stress (τ_f) is equal to frictional resistance of fractures $(\tau_{\max})$ or the effective normal stress (σ_{nfeff}) is equal to the critical closure pressure (σ_{cc}). Below this surface, however, both σ_{nfeff} and τ_f are, respectively, either greater than σ_{cc} or smaller than $\tau_{\max}$. Above the surface $Z_{(x,y,t)}$, the assumed *in situ* stress conditions cause the thermal and hydraulic properties to be time-dependent. Below the surface $Z_{(x,y,t)}$, however, both of these sets of properties are independent of *in situ* stress field and, therefore, are independent of time.

Plate 3.5.3-2 presents a conceptual model of flow in the vadose zone, for a coupled heat and fluid groundwater system operating in a deforming fractured medium. In this model, both vertical boundaries are considered as adiabatic and "no fluid-flux" boundaries. Such assumption causes the model to be an untrue representation of the actual flow conditions. However, if the distance x_2 is very large relative to the depth of water table Z_{wt}, this simplifying assumption does not entail profound implications. This is not so, however, if the conceptual model is intended to represent a limited portion of the vadose zone. In such a situation, both vertical boundaries shall be considered as "flux" boundaries with regard to heat, fluid, and gas. Furthermore, fluxes of heat, fluid, and gas may vary in time and space.

The system contains the time- and space-dependent surface $Z_{(x,y,t)}$. Below points x_0 and x_1, the surface is located at a depth greater than depth of the water table (Z_{wt}). Between points x_1 and x_2, however, the surface $Z_{(x,y,t)}$ is situated above the water table.

Both horizontal boundaries are considered as "fluid- and heat-flux" boundaries. The upper boundary (Z_0) is isothermal and isobaric. It contains a distributed fluid source, $R_{sf_{(x)}}$, which represents flux of rainwater through the vadose zone. The lower boundary is located at depth $Z_{wt_{(x)}}$; it represents the contemporary water table. This boundary contains a distributed heat source, $R_{hZ_{wt_{(x,t)}}}$, and a distributed fluid source (sink), which is represented by the term $R_{fZ_{wt_{(x,t)}}}$. Below the water table, between points x_1 and x_2, thermal and hydraulic properties are independent of time and the intensities of fluid and heat sources (sinks), therefore, vary from one point to another, but are independent of time. Between points x_0 and x_1, however, thermal and hydraulic properties are *in situ* stress- and time-dependent. Consequently, the intensities of both the heat source and the fluid source (sink) are variable in time and space. Such conditions represent short-term excursions and long-term "rises" of the water table. Both of these instabilities are the result of either restructurings of the *in situ* stress field (Section 3.3.4), or transient thermal upwellings of groundwater (Section 3.4.4).

In the vadose zone, the flow of heat involves convection and conduction, and the flow of fluids and gases occurs as fracture flow. Movements of fluids and gases are very complex, and are the result of interaction between three factors: a) gravity; b) heat flow; and c) transient gradients of pore pressure. Transient gradients of pore pressure are the result of random, upward and downward, fluxes of fluid through the lower horizontal boundary. Fluids residing in the vadose zone are the result of mixing of infiltrating rain water with water injected from below the water table.

During early stages of each tectonic cycle, the surface $Z_{(x,y,t)}$ is relatively shallow and its configuration is simple. Progress of deformation is expressed in a two-fold manner: a) downward migration of the surface $Z_{(x,y,t)}$ and b) increasingly complex configuration of this surface. At the end of a tectonic cycle, the characteristic earthquake and the resulting separation of wallrock causes termination of the "limit equilibrium" stress conditions and, consequently, the surface $Z_{(x,y,t)}$ returns to its former position. The process is repeated, over and over again, as long as regional tectonic straining remains active.

A coupled heat and fluid flow system developed in a deforming fractured medium is a remarkable thermodynamic substance. The most noteworthy feature of such a substance is, perhaps, its evolving character. The evolving character of geologic systems, and potential importance of such systems for performance considerations of a high-level nuclear waste repository, was emphasized by some investigators (Harper and Szymanski, 1980; Szymanski and Harper, 1980; Harper and Szymanski, 1982; Donath, 1982).

Evolution of a geologic substance, that corresponds to a coupled heat and fluid flow system developed in a deforming fractured medium, may be envisaged as occurring via a series of similar evolutionary loops (Plate 3.5.3-3). During the course of evolution, the form of the substance becomes increasingly more advanced. The phrase "form of the substance" is used here to denote an integrated appearance of a system as expressed at any given time by a) intensity of fracturing; b) deformed state; c) energy balance; and d) alteration of initial litho-stratigraphic framework (veins, explosive and fragmentation breccias, hydrothermal and metasomatic mineralization and alteration, seep deposits, and tuffa mounds, etc.). At each stage of evolution, the energy balance of the substance may be simplified by assuming that such a balance consists of three time- and space-dependent components: a) strain energy, as represented by the *in situ* stress field and the displacement field; b) thermal energy, as represented by the geothermal field and the overall flux of terrestrial heat; and c) potential and kinetic "hydraulic" energy, as represented by the field of hydraulic potentials and the overall flux of groundwater.

A single evolutionary loop, in terms of its duration and periodicity, corresponds to a tectonic cycle (Section 3.3.2). Throughout the duration of each cycle, the energy balance of the substance undergoes a systematic transformation. At the onset of a cycle, the energy balance is dominated by the thermal and hydraulic components. The contribution of the strain energy component is relatively small. Consequently, the *in situ* values of thermal and hydraulic properties are relatively small, and the resulting intensities of geothermal and hydraulic fields are relatively high. At some stage of deformation, the "limit equilibrium" stress conditions are established, and progressive dilation of fractured medium is initiated. This causes overall time- and space-dependent enhancements of the thermal and hydraulic conductivity structures. Gradually, as the *in situ* stress gradients become "stronger," intensities of the geothermal and hydraulic fields become "weaker". The flow system begins to lose its geothermal identity. The water table deepens and becomes flatter. If the loss of geothermal identity is substantial, above the surface $Z_{(x,y,t)}$, the flow system will appear as a "simple" groundwater system developed in a deforming fractured medium (Section 3.3.4). While dilation takes place, storativity of a deforming fractured medium increases. The volume of

water released from storage in association with the "weakening" of lateral hydraulic gradients is accommodated, not only through the increased out-flow but also, through the increased storage deeper within the system. As the surface $Z_{(x,y,t)}$ migrates downward, the volume of "extra" water in temporary storage increases. Such increase is associated with a progressive transfer of water from shallower and cooler levels to deeper and warmer levels. The transferred water must come into thermal equilibrium with the host medium. Some of the terrestrial heat will be used for that purpose. Clearly, a deforming fractured medium, while it is being strained, stores not only fluid but also thermal energy.

It is important to understand a sequence of events that occurs at the end of a cycle of tectonic deformation. Such a sequence is shown diagrammatically on Plate 3.5.3-3. At this time, the intensity of *in situ* stress field is "strong." This means, that a) the surface $Z_{(x,y,t)}$ is situated at a substantial depth; b) the *in situ* stress trajectories are strongly curved (in situ stress gradients are "strong" or "steep"); and c) the *in situ* stress field contains a large number of "singular" points. The *in situ* values of hydraulic properties are high and, consequently, the water table is low and flat. Between the water table and the surface $Z_{(x,y,t)}$, the dilated system contains a large volume of fluid in temporary storage. Furthermore, below the surface $Z_{(x,y,t)}$, the hydraulic "overpressure" $(\Delta p_{\max})$ is present. Likewise "weakened" is the geothermal field. Above the surface $Z_{(x,y,t)}$, the in situ values of thermal properties are relatively high and, consequently, the values of vertical geothermal gradients are reduced. Between the water table and the surface $Z_{(x,y,t)}$, the dilated system contains a large amount of thermal energy in temporary storage. Convection of fluids in faults is "dampened", and there is a tendency for "strong" fluid-wallrock thermal interaction (Section 3.4.3.3). Below the surface $Z_{(x,y,t)}$, however, there is a sharp increase in the value of vertical geothermal gradient (Section 3.5.2; Plate 3.5.2-5). The *in situ* values of thermal properties are lower and, therefore, strength of fluid-wallrock interaction is reduced. Both, the increased value of vertical geothermal gradients and the reduced "strength" of fluid-wallrock interaction cause the potential for convection of fluids in faults and fractures to be high.

With the occurrence of characteristic earthquake and the resulting development of wallrock separations, the circumstances under which a coupled heat and fluid flow system operates are substantially altered. The intensity of *in situ* stress field is "weakened." This means, that a) the surface $Z_{(x,y,t)}$ is situated in close proximity to the ground surface; b) the *in situ* stress gradients are small (*in situ* stress trajectories are "flat"); and c) the "singular" points either do not exist, or their existence is restricted to pockets of "remnant" stress. In wallrock, the *in situ* values of thermal and hydraulic properties returned to their

initial values, and are relatively low. Around and within the separated faults and fractures, however, the *in situ* values of hydraulic conductivity are very high. High hydraulic conductivity, together with the reduced "strength" of fluid-wallrock thermal interaction, encourages convection of fluids in the separated faults and fractures (Section 3.4.3). In the post-characteristic earthquake period, there are other factors that further encourage convection of fluids in faults (Section 3.4.3). Examples of such factors are a) upflow of fluids resulting from seismic and volcanic activities and causing an increase in the value of D'Arcy-Rayleigh number; b) upflow of gas resulting from seismic or volcanic activity and causing a decrease in density of fluids; c) local increase in flux of terrestrial heat, etc. Both, the decreased hydraulic properties of wallrock and the increased thermal instability of fluids in separated faults provide an opportunity for the occurrence of thermal and hydraulic rises of the water table (Sections 3.3.4 and 3.4.4). It is likely, that some of the fluids involved in these rises are mobilized from a considerable depth; it is reasonable, therefore, to expect that such fluids are mineralized. Precipitation of minerals along the separated fractures and within pockets of "explosive" or "fragmentation" breccia may be expected. Such precipitation will cause a gradual decrease of apertures of separations and, for pockets of "explosive" breccia, reduction of pore space. Eventually, the process will cause complete closure of separations, filling-up of pore space, and fusing back together the fractured medium. With the occurrence of subsequent cycles of tectonic deformation, the above sequence of events is repeated; so, that the cumulative volumetric strain of the system increases progressively in a manner shown on Plate 3.3.2.3-2.

Accretionary aspects of evolution of a geologic system, corresponding to a coupled heat and fluid flow system in a deforming fractured medium, are of major diagnostic importance. Such aspects may be used to unravel the true identity of a system, at a time when such identity is masked by the advanced ongoing deformation.

It is highly probable that, the "mosaic" breccias and calcite-silica veins described in Section 3.3.2-3 record the accretionary evolution of the Yucca Mountain geologic system. These breccias and veins are, therefore, the key to a **satisfactory** understanding of the long-term behavior of this system. Such an understanding, in turn, is the key to **responsible** assessments of the suitability of the Yucca Mountain area to accommodate a high-level nuclear waste repository.

It is important to recognize, that in the case of the Yucca Mountain geologic system, all uncertain pieces of the puzzle may be refined by application of simple and standard geologic procedures and tech-

niques. Trenching may provide a firm base for determining magnitude of the total volumetric strain $(\Delta V/V)$ for the area, as represented by the "mosaic" breccias and veins. Petrographic and mineralogic studies of petrogenesis involved in the formation of "mosaic" breccias may provide a solid base for unequivocal determinations of whether or not the "mosaic" breccias are a) younger than solidification of the country rock; b) polygenetic; and c) representing time spans measured in terms of 10^6 years.

In view of the potential importance of the issues involved, **it is essential that uncertainties regarding all aspects of the "mosaic" breccias and calcite-silica veins, should such exist, be addressed prior to substantial commitments of resources.** Always, it is a wise and prudent practice to follow carpenter's rule "measure twice and cut once." In the case of a high-level nuclear waste repository this seems to be a sound advice indeed.

SECTION 4.0

CHARACTERISTICS OF THE DEATH VALLEY GROUNDWATER SYSTEM IN LIGHT OF THE EXISTING DATA

4.1 GENERAL DESCRIPTION OF THE DEATH VALLEY GROUNDWATER SYSTEM

The Death Valley groundwater system, including the Yucca Mountain groundwater system, is situated in southern Nevada adjacent to parts of southeastern California (Plate 4.1-1). The system was distinguished somewhat arbitrarily from a much larger hydrologic entity which is the Alluvial Basins Groundwater Region (Heath, 1984). It is centered around the Nevada Test Site, areal extent of which takes up a significant portion of the groundwater system.

It became customary to consider the Death Valley groundwater system as composed of three subbasins (Waddell et al., 1984). These subbasins are: a) the Oasis Valley subbasin; b) the Alkali Flat - Furnace Creek Ranch subbasin; and, c) the Ash Meadows subbasin (Plate 4.1-2).

Climate of the region ranges from arid on the valley floors to subhumid on the crest of the highest mountains. Altitudes range from greater than 3,600 m in the Spring Mountains to below sea level in Death Valley. Precipitation ranges from less than 50 mm/year in Death Valley to greater than 700 mm/year in the Spring Mountain area. The mean daily maximum temperature, at an altitude of approximately 700 m, ranges from 13°C in January to 40.5°C in July; the mean daily minimum temperature for the same months ranges from 0.5°C to 24.5°C; temperatures at greater altitudes are as much as 3°C to 10°C lower. In Death Valley temperatures greater than 49°C are common during the summer months.

Plate 4.1-3 shows distribution of the average annual precipitation and the location of recharge and discharge areas. The recharge areas were identified assuming, that the empirical relationship between precipitation and recharge, as developed by Eakin, 1963 and Eakin et al., 1963, is valid for this region. This relationship suggests that no recharge occurs where the mean annual precipitation is less than 200 mm. The groundwater discharge occurs as spring flow, seeps, evapotranspiration, groundwater withdrawal, and outflow to other groundwater basins.

In the Death Valley groundwater system, the water conducting medium is composed of two units that involve sharply different hydraulic properties and water flow mechanisms. The first unit is a "valley fill aquifer" that is composed of various detrital sediments of Late Tertiary and Quaternary age. These alluvial-fan, fluvial, fanglomerate, lake-bed, and mud-flow sediments are only locally cemented, and consequently, exhibit large interstitial porosity. The groundwater flow mechanism, therefore, involves predominantly porous flow through the material matrix. Thicknesses and stratigraphic interrelationships of these deposits are highly variable. Thicknesses of more than 500 m occur in some areas, nevertheless, areas where alluvium is saturated are limited and, consequently, this hydrogeologic unit is of minor importance. The exception is beneath most of the discharge areas, where the valley fill is present in saturated thicknesses sufficient to constitute an important aquifer.

The second unit comprises the bulk of the water conducting medium of the Death Valley groundwater system. This unit consists of a great variety of sedimentary rocks that are well indurated, slightly metamorphosed, and include limestones or marbles, quartzites, and various argillites; and of volcanic rocks, that include various pyroclastic deposits, intrusives, and extrusives. The age of these rocks ranges from Precambrian-Early Cambrian, through Devonian, Carboniferous and Permian, to Tertiary. The lithologic character of these rocks is such, that by and large, interstitial porosity is quite small, therefore, groundwater movement involves fracture flow. Locally, however, interstitial porosity may be considerable as, for example, is the case for non-welded tuffs, fluvially reworked tuffs, and tuffaceous sandstones. For these rocks, the flow mechanism may involve porous flow through the medium matrix. The thickness of the unit is variable, but involves many thousands of meters. Small portions of the total thickness, 10 or 20 percent, are made up of rocks whose interstitial porosity is sufficient to support significant matrix flow.

In the Death Valley groundwater system, spatial relationships between various lithic types are very complex. Such relationships are the result of a very complex evolution of this system, which during its various stages, included sedimentary, metamorphic, volcanic, structural, and hydrothermal alteration processes. The resulting degree of litho-stratigraphic complexity is such that a three-dimensional distribution of hydraulic properties can only be known on a general and conceptual basis, regardless of the degree of subsurface exploration effort.

Notwithstanding the large degree of complexity involved, there is a very substantial data base pertaining to the Death Valley groundwater system. This data base has been developed as the result of a

very substantial commitment of resources made by the U.S. Government in association with activities involving detonations of nuclear devices as well as involving the nuclear waste repository siting and evaluation process. **The dollar value of this data base may be expressed in terms of billions of dollars.** It should not be a surprise, therefore, that such a data base contains a lot of information that may be used in testing the validity of hydrologic concepts developed in Section 3.0. This testing is an essential prerequisite for developing a validated conceptual model for the Yucca Mountain groundwater system. In turn, the validated conceptual model, is a cornerstone for assessments of suitability of the Yucca Mountain site to accommodate a high-level nuclear waste repository.

The following pages are devoted to a review of various geologic and hydrologic data from the Death Valley and Yucca Mountain groundwater systems. The review was performed keeping in mind the conceptual framework as established in Section 3.0. The discussions have been subdivided into four parts. The first part, Section 4.2, deals with the hydraulic conductivity structure of the groundwater systems. The second part, 4.3, is concerned with geothermal conditions in these groundwater systems. Section 4.4 is devoted to analyses of responses of the groundwater system to detonations of nuclear devices. Finally, Section 4.5 deals with hydrologic and geologic features that may be indicative of habits of long-term behavior of the Death Valley and Yucca Mountain groundwater systems.

For the purposes of the conceptual considerations undertaken in this report, the measured values of hydraulic potentials as a function of time are most important. These data were collected over a 10-year period during which the Yucca Mountain site was evaluated to determine its suitability to accommodate a high-level nuclear waste repository. Continuous and periodic monitorings were performed in a number of deep wells covering an area of approximately 25 km^2. At the present time, only some data from periodic monitoring are available. **The data from continuous monitoring, however, are not yet available for inspection. Such data are of critical importance for final evaluations of validity of the proposed conceptual understanding of the Yucca Mountain groundwater system. Without relying on these data, the conceptual considerations undertaken herein cannot yet be carried out to their logical and completely reliable conclusion.**

4.2 THE HYDRAULIC CONDUCTIVITY STRUCTURE, ITS CHARACTERISTICS AND ORIGIN

4.2.1 Introduction

The hydraulic conductivity structure of a groundwater system is, perhaps, the most important characteristic of this system. Such structure controls spatial distribution of hydraulic potentials, and through that, it also controls: a) position and configuration of the water table; b) vertical gradients of hydraulic potentials; c) flow lines; and d) corresponding groundwater travel-times. Moreover, there is a direct linkage between long-term stability of a groundwater system and long-term stability of its hydraulic conductivity structure.

An understanding of the hydraulic conductivity structure, including its spatial and temporal aspects, is essential for any meaningful consideration of a groundwater system. For the purposes of hydrologic considerations of a potential site for a high-level nuclear waste repository, particularly important are assessments of long-term stability of the hydraulic conductivity structure. Such assessments may be performed by first identifying factors that control the hydraulic conductivity structure, and then by evaluating durability of these factors.

It is a common hydrologic practice to assume that the hydraulic conductivity structure, for all practical purposes, is independent of time. Such assumption is build on a premise that the hydraulic conductivity structure is related to durable factors: a) litho-stratigrahic framework that controls the distribution of primary and secondary porosity; b) structural framework that controls the distribution of density of fracturing; c) overburden stress that controls aperture of fractures; and d) local alterations of secondary porosity, through mineral precipitation or dissolution. The assumption of time-independence of the hydraulic conductivity structure forms a foundation for a widely held belief that a) groundwater flow is, essentially, a steady-state process and b) responses of various groundwater systems to ordinary tectonic stimulations are similar, and are of relatively minor spatial and temporal extents. Any transient behavior observed is commonly attributed either to precipitation induced changes in the volume of groundwater or to some external factors (i.e., vibratory ground motion, fluid injection or withdrawal, drilling disturbances, earth tides and barometric disturbances, and malfunction of recording devices).

The assumption of durability of the hydraulic conductivity structure is seldom questioned and subjected to any validation efforts. This occurs because, in ordinary applications, hydrologists do not have a need for addressing dynamic and long-term aspects of behavior of groundwater systems. Furthermore, a typical hydrologic data base does not allow for unequivocal addressing of that issue. This occurs because the data base is usually fragmentary, limited in scope, and only covers a relatively short period of time.

Circumstances are different, however, if one is concerned with the Death Valley groundwater system and, in particular, with the Yucca Mountain area. There are three aspects of such circumstances. First, the inquiry is performed for the purposes of evaluating effectiveness of a high-level nuclear waste disposal system. Hence, both dynamic behavior and long-term stability of the groundwater system are of critical importance. Second, the currently existing and available data base is one of truly extraordinary proportions. Such data base contains most elements that are required to identify, on more or less conclusive basis, factors that control the hydraulic conductivity structure. Furthermore, this data base is also sufficient to make judgments regarding durability of such factors. Third, in the Death Valley system, the groundwater flow occurs in a continuously strained and deforming fractured medium. For such a medium, a reasonably well supported case can be made that a changing, in time and space, *in situ* stress field is involved. Under certain conditions, such stress field controls conducting apertures and, through that, it also controls long-term stability of the hydraulic conductivity structure.

A main purpose of the analyses presented below is an assessment of possible involvement of the changing *in situ* stress field in, locally, controlling the hydraulic conductivity structure of the Death Valley groundwater system. The analyses are presented in five parts. The first part, Section 4.2.2, describes the results of laboratory and *in situ* measurements of hydraulic conductivity. The second part, Section 4.2.3, deals with configuration and position of the water table. The third part, Section 4.2.4, is concerned with distribution of vertical gradients of hydraulic potentials. The fourth part, Section 4.2.5, presents a review of *in situ* stress data. In this part, an attempt is made to spatially associate hydrologic characteristics of a given area with the state of *in situ* stress in this area. Finally, Section 4.2.6 presents a summary and conclusions.

4.2.2 General Description of the Hydraulic Conductivity Structure

In the Death Valley groundwater system, spatial distribution of hydraulic conductivity is known based on numerous measurements performed *in situ* and in the laboratory. Pumping tests, injection tests, and swabbing tests, were all used to determine the *in situ* values of hydraulic conductivity. In addition, constant-head testing of core samples was employed to obtain values of the saturated matrix hydraulic conductivity.

The pumping tests were made by pumping water out of an entire borehole or out of a major segment of a borehole. While the test borehole was stressed by pumping, borehole-flow surveys were made to determine locations of productive zones and local rates of flow. The flow surveys were performed using a slug of iodine-131.

The Cooper-Brederhoeft injection tests were made for multiple intervals of each test borehole. During an individual test, a borehole segment was isolated between two inflatable packers. A slug of water was then injected into the isolated interval. Pressure transducers with ground-surface readouts were used to monitor a decay of the slug.

The laboratory measurements of the saturated matrix hydraulic conductivity were made using vertical core samples. Most of the tested cores were extracted from deep wells drilled in the area of Yucca Mountain, and in the vicinity of Yucca Flat. The measurements were made using distilled water and the constant-head technique. It appears, that for most of the measurements, there was no compensation or adjustment made for the overburden load.

With the above introduction, it seems appropriate to perform a review of the hydraulic conductivity data from the Death Valley groundwater system. A main purpose of such a review is to identify a pattern in spatial distribution of the *in situ* hydraulic conductivity. Once identified, this pattern may be used as a starting point in searching for factors that may control the hydraulic conductivity structure. The problem may be simplified by recognizing that there are three broad conceptual possibilities. Such possibilities are a) litho-stratigraphic framework, that may control spatial distribution of both primary (interstitial) and secondary porosity (fracture residual aperture, including dissolution); b) structural framework, that may

control spatial distribution of both primary and secondary porosities; and c) *in situ* stress field, that may control spatial distribution of conducting apertures.

The analyses presented below are made with reference to two main litho-stratigraphic complexes (i.e., Paleozoic carbonates and the Tertiary tuff "pile") that comprise the bulk of the fractured medium in the Death Valley groundwater system. Both complexes are present at the Yucca Mountain site and are of major importance for considerations of the local groundwater system.

Plate 4.2.2-1 presents the results of about 100 laboratory measurements of the saturated matrix hydraulic conductivity. The measurements were performed using core samples extracted from five deep wells drilled in the Yucca Mountain area. They represent an approximately 2 km thick section of the Tertiary tuff "pile" in this area. Plate 4.2.2-2 presents the values of saturated matrix hydraulic conductivity plotted against stratigraphic depth. The values of matrix hydraulic conductivity are reasonably well constrained. They range from 10^{-3} to about 10^{-6} m/d, averaging 5×10^{-5} m/d. There does not appear to be any discernable relationship between the values of matrix hydraulic conductivity, the stratigraphic depth, and the lithologic character of rock units involved. In terms of the laboratory values of hydraulic conductivity, the entire section appears to be reasonably homogenous.

Winograd and Thordarson, 1975, reported values of the saturated matrix hydraulic conductivity for the Paleozoic carbonates. These values were obtained for 13 core samples taken in 8 boreholes drilled around Yucca Flat and Frenchman Flat. The values of saturated matrix hydraulic conductivity for carbonates are similar to those from the tuff "pile". They range from 4×10^{-3} to about 8×10^{-7} m/d, averaging 4×10^{-4} m/d.

Plates 4.2.2-3 and 4.2.2-4 present the results of selected pumping tests performed in boreholes penetrating the Paleozoic carbonates and the tuff "pile". The average hydraulic conductivity ranges from 2×10^{2} to 3×10^{-2} m/d for the carbonates, and from 3×10^{1} to 4×10^{-2} m/d for the tuff "pile". These surprising similarities, for these two so different litho-stratigraphic complexes, are shown on Plate 4.2.2-5. Clearly, a straightforward and intuitively appealing notion that the carbonate rocks constitute the most transmissive parts of the groundwater system is not readily supported by the actual data. This, of course, does not mean that the carbonate rocks are insignificant in controlling the groundwater system. There

are, however clear indications, that locally, rocks comprising the tuff "pile", despite their lithologic character and the resulting resistance to dissolution, may be of equal importance.

The large differences in the laboratory and the *in situ* values of hydraulic conductivity, for both of the litho-stratigraphic complexes, indicate that interstitial porosity is not a decisive factor. In the Death Valley groundwater system, the hydraulic conductivity structure seems to be related, mainly, to secondary porosity. The locally comparably high values of the *in situ* hydraulic conductivity, for both the tuff "pile" and the carbonates, suggest that factors responsible for the high secondary porosity may exert their influence without paying much attention to lithologic character.

Also, it is unlikely, that in the Death Valley groundwater system, the Paleozoic carbonates owe their high secondary porosity to dissolution processes. Extensive examinations of the carbonate rocks, as exposed in outcrops and represented by core samples from deep wells, were performed by Winograd and Thordarson, 1975. These authors concluded, that "Neither stratigraphically controlled regional solution of the carbonate rocks nor significant solution below the major Tertiary- pre-Tertiary unconformity is evident. The absence of stratigraphically controlled solution is not unexpected because of the absence of disconformities marked by significant erosion in the Paleozoic rocks older than Late Devonian. No field evidence suggests that sinkholes or karst topography exist in, or were developed on, the carbonate rocks beneath the Tertiary- pre-Tertiary unconformity. Field and subsurface evidence from a few holes indicates that the fractures within the uppermost part of the carbonate rocks are commonly filled by tuffaceous or lacustrine detritus derived from the Tertiary rocks above the unconformity."

Based on the above observations, it seems reasonable to conclude, that in the Death Valley groundwater system, secondary porosity is predominately related to fracturing. Any heterogeneity in spatial distribution of such porosity is related either to variability of density of fracturing or to variability of fracture conducting aperture. Understanding the contribution that each of these two factors makes is of great conceptual importance. Furthermore, such understanding is of critical importance for any considerations of long-term stability of the groundwater system. Involvement of the enhanced conducting apertures together with the changing *in situ* stress field, leads to the conclusion that the hydraulic conductivity structure is time-dependent and subject to, more or less, dramatic changes. This is not so, however, for the involvement of density of fracturing and only residual apertures. It is unlikely, that in time scales of 10^4 years, both of these factors undergo significant changes.

An unique opportunity for examining the above two possibilities is provided by the results of *in situ* measurements of hydraulic conductivity performed at Yucca Mountain. In this area, the density of measurements is sufficiently high and permits definition of a reasonably well pronounced pattern in spatial distribution of the *in situ* hydraulic conductivity. Most of the measurements were made in rocks of the tuff "pile". Only one borehole penetrated the Paleozoic carbonates. Consequently, it is not possible to develop any generalizations regarding spatial distribution of the *in situ* hydraulic conductivity in these rocks.

Plates 4.2.2-6 through 4.2.2-11 present the *in situ* hydraulic conductivity data obtained from six deep wells drilled in the area of Yucca Mountain. In terms of location and stratigraphic depth, these wells are equivalent to those that provided core samples for laboratory measurements of the saturated matrix hydraulic conductivity (Plate 4.2.2-2).

Plates 4.2.2-13 and 4.2.2-14 are generalized cross-sections presenting spatial distribution of the *in situ* hydraulic conductivity for Yucca Mountain, relative to the stratigraphic framework of this area. Locations of both cross-sections are shown on Plate 4.2.2-12.

Examination of the cross-sections reveals, that at Yucca Mountain, a very heterogenous hydraulic conductivity structure is present. The *in situ* values of hydraulic conductivity range from 1.8×10^{1} to about 10^{-6} m/d. There is a well expressed tendency for these values to decrease depthward, northward, and westward away from borehole UE-25p#1.

With some stretch of imagination, it is possible to envisage the local hydraulic conductivity structure as composed of three layers. In the uppermost layer, the *in situ* values of hydraulic conductivity range from 1.8×10^{1} to about 10^{-2} m/d. The layer extends to a depth ranging from about 800 m in Well USW H-1 to more than 1,800 m in Well UE-25p#1, and from about 750 m in Well USW H-6 to more than 1,200 m in Well USW H-4 (Plates 4.2.2-13 and 4.2.2-14).

In the second layer, the *in situ* values of hydraulic conductivity range from 10^{-2} to about 10^{-4} or 10^{-5} m/d. Spatially, the layer is ill constrained and appears to be relatively thin.

In the third and lowermost layer, the *in situ* values of hydraulic conductivity are less than 10^{-4} m/d. Within the explored section, the layer is present only in western and northern sectors of Yucca Mountain. This layer was encountered in two boreholes: USW H-1 and USW H-6 (Plates 4.2.2-13 and 4.2.2-14).

At Yucca Mountain, there are three noteworthy features of the hydraulic conductivity structure. First, in the lowermost layer the *in situ* values of hydraulic conductivity are comparable with laboratory values of the saturated matrix hydraulic conductivity. This indicates, that under certain conditions, fracturing related enhancement of the hydraulic conductivity structure may be very small and hardly noticeable. Evidently, within deeper parts of the tuff "pile", residual apertures are exceedingly small.

Second, boundaries separating individual layers of the hydraulic conductivity structure are colinear, neither with stratigraphic boundaries nor with lines of equal overburden stress. This suggests that neither the litho-stratigraphic framework controlled density of fracturing nor the overburden stress alone controlled aperture of fractures is responsible for the observed distribution of *in situ* hydraulic conductivity.

Third, the texture of hydraulic conductivity structure appears to be somewhat systematic. There is a reasonably well expressed tendency for the *in situ* values of hydraulic conductivity to progressively decrease depthward, and with northward and westward increasing distance away from borehole UE-25p#1.

Clearly, the systematic depthward decrease of the *in situ* values of hydraulic conductivity, occurring at a spatially variable rate, is an important feature. Such a feature may be interpreted as indicating involvement of a changing *in situ* stress field. As explained in Section 3.3.3.3 and shown on Plates 3.3.3.3-6 and 4.2.2-15, the total *in situ* stress enhancement of the conducting aperture $\left[a_{d_{(t)}} + a_{n_{(t)}}\right]$ is depth-dependent. Below the surface $Z_{(x,y,t)}$, both of the components of such enhancement are equal to zero. Consequently, for very small values of residual apertures, the *in situ* values of hydraulic conductivity are comparable with laboratory values of the saturated matrix hydraulic conductivity. Clearly, this seems to be the case for the lowermost layer of the hydraulic conductivity structure at Yucca Mountain.

Immediately above the surface $Z_{(x,y,t)}$, however, the values of post-peak shear displacement $\left(u_s - u_{sp}\right)$ are small and, consequently, the values of compliance, K_s and limit U_{ad} are also small (Section 3.3.3.3). The resulting enhancement of conducting aperture is small, and consists of only the $a_{d_{(t)}}$ component. The

in situ values of hydraulic conductivity are increased, but only modestly so. Clearly, this seems to be the case for the second layer of the hydraulic conductivity structure at Yucca Mountain.

Some distance above the surface $Z_{(x,y,t)}$, local values of the post-peak shear displacement are larger and, consequently, local values of the mean stress $\frac{1}{2}(\sigma_1 + \sigma_3)$ and the resulting values of closure pressure (σ_c) are very small. The enhancements of conducting apertures are large and consist of both components $[a_{d_{(t)}} + a_{n_{(t)}}]$. Relative to the lowermost layer, the *in situ* values of hydraulic conductivity are very large. Clearly, this seems to be the case for the uppermost layer of the Yucca Mountain hydraulic conductivity structure.

In summary, examination of the hydraulic conductivity data reveals few interesting characteristics of the hydraulic conductivity structure in the Death Valley groundwater system. This structure appears to be dominated by fracture related secondary porosity. Neither interstitial porosity nor solution openings appear to be important factors. In terms of laboratory values of the saturated matrix hydraulic conductivity, the two main litho-stratigraphic complexes of the groundwater system, the Paleozoic carbonates and the Tertiary tuff "pile", appear to be similar. Within each complex, however, there is a considerable variability of the *in situ* values of hydraulic conductivity. For the Yucca Mountain tuff "pile", these values range from about 2×10^1 to 10^{-6} m/d. At this location, there appears to be a systematic pattern in spatial distribution of the *in situ* values of hydraulic conductivity. These values diminish depthward, northward, and westward away from borehole UE-25p#1. Such distribution may be accounted for by assuming that the local *in situ* stress field is the controlling factor. This assumption, unlike any other conceptual possibility, may readily be justified in terms of the available field data. This assumption, therefore, deserves further testing along lines proposed in Section 3.3.6. The following pages are devoted to such testing.

4.2.3 Configuration of the Water Table

Configuration of the water table is the most readily discernible characteristic of a groundwater system. Such configuration constitutes an integrated expression of two factors that control a groundwater system: a) the boundary conditions and b) the hydraulic conductivity structure. In Section 3.0, it was demonstrated, that in the tectonic environment of the Yucca Mountain groundwater system, both of these factors may be influenced by tectonic processes. It is important, therefore, to evaluate whether or not configuration of the water table in this system contains any indications that this may be the case indeed.

In a tectonically controlled groundwater system, configuration of the water table is the result of complex interactions involving various parts of this system. Such interactions are caused by time- and space-dependent changes in the hydraulic conductivity structure, and involve spatially and temporally variable gradients of hydraulic potentials and temperature. The resulting configuration of the water table is complex and includes a) hydraulic mounds; b) hydraulic sinks; c) areas where slopes of the water table are very steep; and d) areas where inclinations of the water table are very low (Plates 3.3.4-7 and 3.4.4-7).

Before proceeding with subsequent discussions, it is useful to define the meaning of the term "water table" as it applies to the Yucca Mountain groundwater system. To this end, it is worthwhile to examine Plate 4.2.3-1. This plate presents data concerning a degree of matrix saturation as encountered in two boreholes: USW H-1 and UE-25b#1. It shows that the degree of matrix saturation, above and below the water table, does not differ significantly. Below the water table, the degree of matrix saturation ranges from 70 to 95 percent. Similar ranges were obtained for rock samples collected 100 and 300 m above the water table. Based on these data, it seems reasonable to consider the water table as a plane below which most fractures contain water, but the rock matrix is not always fully saturated. Above this plane, however, either none or only some fractures contain water. Nevertheless, the degree of matrix saturation is relatively high, particularly in lower parts of the vadose zone.

Plate 4.2.3-2 presents the position and configuration of the water table in the Death Valley groundwater system. Both are known based on numerous boreholes drilled for either water supply or subsurface exploration. Altitudes of the water table range from over 1,900 m above mean sea level, in northern parts of the system, to below sea level in Death Valley. Configuration of the water table is remarkable; the basic pattern consists of groundwater "plateaus" separated by sharp steps. Within an individual "plateau",

altitudes of the water table vary little; horizontal gradients of hydraulic potentials are small, measured in tens of meters per kilometers or less. Between the "plateaus", however, slopes of the water table as large as 10 and 30 percent are present. Changes in altitude of the water table measured in hundreds of meters occur within relatively short distances. Uncommonly large hydraulic gradients occur in both the recharge and discharge areas of the groundwater system (Plates 4.2.3-3 and 4.2.3-4). They are also present in central portions of the groundwater system. One such feature extends from Beatty, Nevada, through northern Yucca Mountain, to Yucca Flat and beyond. This feature is developed across various geohydrologic and physiographic units, and across various structures. Plate 4.2.3-5 shows this feature as it is developed in the northern sector of Yucca Mountain.

Comparison of Plate 4.2.3-5 with Plates 4.2.2-13 and 4.2.2-14 reveals, that at Yucca Mountain, the high lateral gradient of hydraulic potentials may be justified in terms of the local hydraulic conductivity structure. This gradient occurs at locations where the most transmissive layer of the hydraulic conductivity structure is either absent or relatively thin. There does not appear to be, however, any indications that a narrow, vertical hydraulic barrier is involved.

It is interesting to note, that both very low and very high lateral gradients of hydraulic potentials are present in both the Tertiary tuff "pile" and in the Paleozoic carbonates. Examples of the low gradient in both of the lithologic complexes are found in the area of Yucca Flat and Frenchman Flat (carbonates) and between Yucca Mountain and Amargosa Desert (tuff "pile") (Plate 4.2.3-2). The areas of southern Indian Springs Valley and northern Yucca Mountain are examples of the high gradients developed in the carbonates and tuff "pile", respectively (Plates 4.2.3-5 and 4.2.3-7).

The indiscriminate occurrences of the highly variable lateral gradients of hydraulic potentials indicate, that in the Death Valley groundwater system, the hydraulic conductivity structure lacks continuity and litho-stratigraphic consistency. This conclusion is similar to the one drawn earlier based on the results of *in situ* measurements of hydraulic conductivity (Section 4.2.2).

It is interesting to examine the water table in an area where boreholes are closely spaced. Such spacing allows for a greater resolution of configuration of the water table. In the area of Yucca Flat, numerous exploratory borings were drilled in association with underground testing of nuclear devices. Plate 4.2.3-8 presents the configuration of the water table beneath Yucca Flat. Along the trace of Yucca Fault, two no-

table features are present. One is a hydraulic mound some 125 feet tall. The other feature is a hydraulic sink some 25 to 50 feet deep.

Configuration of the water table is not a characteristic, that in and by itself, allows for conclusive resolution of the issue of possible involvement of tectonic factors in controlling the hydraulic conductivity structure of a groundwater system. In the absence of other data, one may only speculate about the origin and possible conceptual meaning of observed configurations. A basic pattern of plateau-step-plateau sequence, for example, may be accounted for by assuming that narrow groundwater barriers are locally present in a medium whose hydraulic transmissivity is large otherwise. A sink-mound sequence, may be explained by assuming that local withdrawals (pumping) and additions of fluids (infiltration) are responsible.

It should be recognized, however, that those are merely plausible explanations that may or may not be correct. In the absence of other data, there are two other conceptual possibilities that may equally explain the observed configuration of the water table. As discussed in Section 3.4.6, such possibilities are a) an *in situ* stress field and b) a geothermal field. In both of these cases, the observed configurations of the water table are not permanent features. Such configurations constitute indirect expressions of tectonic processes that control either the local hydraulic conductivity structure (*in situ* stress) or the local boundary conditions (terrestrial heat). Should the controlling factors change, either spontaneously or in association with a tectonic event, position and configuration of the water table will be adjusted accordingly.

In summary, the review of data pertaining to configurations of the water table in the Death Valley groundwater system reveals that features expected to be present in a tectonically controlled groundwater system are also present in this system. Spatial variability of the water table altitudes is uncommonly large. Magnitudes of the resulting lateral gradients of hydraulic potentials are also surprisingly large. In most cases, the origin of these spectacular features is not understood. An exception to this rule, however, is present in the Yucca Mountain area. In Section 4.2.5, it will be demonstrated, that in this area, local *in situ* stress is the most likely cause.

4.2.4 Vertical Gradients of Hydraulic Potentials

4.2.4.1 General

An important characteristic of a groundwater system, one that is ignored far too often, is the distribution of vertical gradients of hydraulic potentials. The presence of such gradients determines how well hydraulic equipotential surfaces are represented by the water table. If there are significant vertical gradients, the water table constitutes a very poor representation of these surfaces. Calculations and various interpretations made on the basis of such a representation may be grossly misleading.

In terms of their origin, vertical gradients of hydraulic potentials may be related to many factors that either control or influence a groundwater system. For the purposes of these conceptual considerations, however, it is useful to recognize two broad conceptual possibilities. The first possibility is that vertical gradients of hydraulic potentials are independent of tectonic factors and, therefore, are independent of time. Such vertical gradients are the result of a) local, lithology or structure related, heterogeneity in the hydraulic conductivity structure or b) position of an area in a groundwater system – for a "simple" system, the vertical gradients occur in both recharge (downward) and discharge (upward) areas.

The second possibility is that vertical gradients of hydraulic potential are related to tectonic processes and may, therefore, be time-dependent. For "mixed" convection systems, vertical gradients are related to the "heat-fluid-flux" boundary conditions present at the base of a system (Section 3.4.4). In areas underlaid by hydraulic and heat sources, the upward directed components of flow must be present. The downward components of flow, however, must be present in areas that are situated over hydraulic and heat sinks.

For a groundwater system developed in a deforming fractured medium, vertical gradients of hydraulic potentials are related to the depth-dependent progress of deformation (Section 3.3.4). In such a system, the characteristic depthward distribution of hydraulic potentials involves "overpressure" $\Delta p_{(z)}$ (Plate 3.3.4-4). The value of "overpressure" attains a maximum Δp_{max}, at some depth below the surface $Z_{(x,y,t)}$. At the onset of a new cycle of tectonic deformation, however, the depthward distribution of hydraulic potentials is different. Such a distribution is shown schematically on Plate 3.3.4-5. It involves a gradually decaying "underpressure". With increasing depth, the value of "underpressure" becomes smaller.

With the above introduction, it seems appropriate to examine the data base with an objective of establishing whether or not, in the Death Valley groundwater system, vertical gradients of hydraulic potentials show any signs of being related to tectonic processes. To this end, three areas, located within the Nevada Test Site, were selected for closer examination: a) Rainier Mesa and Yucca Flat; b) Pahute Mesa; and c) Yucca Mountain.

4.2.4.2 The Area of Rainier Mesa and Yucca Flat

Vertical gradients of hydraulic potentials were encountered in the area of Rainier Mesa and Yucca Flat. Such gradients were detected based on measurements of hydraulic potentials as a function of depth, and based on observations of position of the water table made in association with drilling a number of exploratory boreholes.

Plate 4.2.4.2-1 presents a general hydrologic situation in the vicinity of Rainier Mesa and Yucca Flat. A characteristic feature of this area is a difference in the values of hydraulic potentials between shallow rocks of the Tertiary tuff "pile" and deeper, mainly, carbonate rock of Paleozoic age. From one location to another, the difference ranges from small (tens of meters) to very large (hundreds of meters). There is, however, a consistent pattern; with increasing depth, the values of hydraulic potentials diminish.

In the vicinity of Rainier Mesa, the downhole measurements of hydraulic potentials were performed in four wells: U12e.03-1, U12e.M-1, Hagestad #1, and Test Well #1. Locations of these wells are shown on Plate 4.2.4.2-2. The results of these measurements were reported by Thordarson, 1965. They are also shown on Plate 4.2.4.2-3. Vertical gradients are developed, mainly, within about a 1 km thick sequence of the Tertiary tuffs. Both the stratigraphic position and lithologic character of these tuffs are similar to those of the Yucca Mountain tuffs. Around Rainier Mesa, the values of vertical hydraulic gradients are surprisingly large. They range from 0.3 to as much as 2.5 m/m. There appears to be a tendency for these values to decrease with increasing depth. The resulting values of "underpressure" are also very large, ranging from 160 to as much as 547 m (Winograd and Thordarson, 1975).

Observations made during drilling of exploratory boreholes in the area of Yucca Flat provided an opportunity for examining vertical gradients of hydraulic potential in this area. Winograd and Thordarson, 1975, reported that "when Well #88-66 was 2,045 feet deep, the static water level was 1,915 feet beneath

the surface. After the well had been deepened to 2,535 feet and cased (but not cemented) to 2,121 feet, the water level was 1,959 feet beneath the surface. No further changes were made in the well for 5 months thereafter, but the water level gradually declined another 20 feet. After penetration of carbonate rocks, at a depth of 2,550 feet, the water level in the well fell to 2,055 ± 2 feet. The water level within the well remained at this depth during subsequent deepening of the well to a depth of 3,411 feet." A similar decline of the value of hydraulic potentials was observed in Well #83-69a. Winograd and Thordarson, 1975, further reported that "at well depths of 1,875 and 1,970 feet, the measured static water level was 1,716 feet beneath the land surface. After the hole was deepened to 2,430 feet, the water level dropped to 1,732 feet below the land surface. Later the hole was deepened to 2,620 feet, and the static water level was 1,780 feet below the land surface."

Plate 4.2.4.2-4 presents an idealization of changes of hydraulic potentials as a function of depth for the area of Rainier Mesa and Yucca Flat. The characteristic hydrologic feature of this area is the depthward progressive decline of hydraulic potentials. In most wells, the decline occurs in the Tertiary tuff "pile". In the sequence of Paleozoic carbonates, the values of hydraulic potentials appear to be constant. In Well #83-69a, however, the decline occurs in both the tuff "pile" and carbonates.

The depthward decline of hydraulic potentials, together with the results of *in situ* measurements of hydraulic conductivity, indicate that the hydraulic transmissivity of shallow parts of the tuff "pile" is substantially less than transmissivity of the underlying rocks, mainly, Paleozoic carbonates. This conclusion is pretty straightforward and there is little room for alternate interpretations. Nevertheless, the fact that high transmissivity contrasts, between the tuffs and carbonates, are present in only select areas seems to be curious and invites further inquiry and speculation.

As discussed in Sections 4.2.2 and 4.2.3, neither the results of *in situ* measurements of hydraulic conductivity nor the distribution of lateral gradients of hydraulic potentials support a broad generalization that the carbonate rocks constitute the most transmissive parts of the Death Valley groundwater system. Equally transmissive parts of this system are locally composed of the rocks of tuff "pile". In these parts, the transmissivity contrasts are virtually non-existent (e.g., borehole UE-25 p#1 on Plate 4.2.2-13).

A possible explanation for the selective occurrence of the high tuff-carbonates transmissivity contrasts may be constructed by involving two reasonable assumptions. The first assumption is, that regardless

of the *in situ* stress influences, the *in situ* hydraulic conductivity of the Paleozoic carbonates is generally high. In these rocks, residual apertures appear to be large. In the tuff "pile", however, residual apertures are very small and, consequently, in a "non-limit equilibrium" state, these rocks show low *in situ* values of hydraulic conductivity. Such values are comparable with laboratory values of the saturated matrix hydraulic conductivity.

The second assumption is, that at any given time, the groundwater system consists of various strain domains (Section 3.3.2.2). For each of the domains, the degree of advancement of the contemporary cycle of tectonic deformation is different and, consequently, contemporary accumulation of tectonic strain within each domain is different. In some domains, tectonic straining is advanced and depth extent of the "limit equilibrium" stress conditions is substantial and exceeds the depth of the water table. In such domains, conducting apertures in the tuff "pile" are enlarged and, consequently, the *in situ* values of hydraulic conductivity are high. In other domains, however, tectonic straining is at early stages of development. Within such domains, therefore, depth extent of the "limit equilibrium" stress condition is small and restricted to the vadose zone. Below the water table, conducting apertures in the tuff "pile" are small and the resulting *in situ* values of hydraulic conductivity are likewise small.

Taken together, the above two assumptions form the base of a testable hypothesis. According to this hypothesis, the high tuff-carbonates hydraulic transmissivity contrasts should occur only in areas where the "limit equilibrium" stress conditions either do not exist or the development of such conditions is at early stages. In areas where the "limit equilibrium" stress conditions are deep, however, the transmissivity contrasts should not be present. In Section 4.2.5, it is shown, that for two portions of the Death Valley groundwater system, this seems to be the case indeed.

Returning to the pattern of vertical hydraulic gradients in Rainier Mesa and Yucca Flat, it may be noted that the observed pattern is similar to the one shown on Plate 3.3.4-5. The latter pattern was postulated as characteristic of an area that had experienced large-scale restructuring of the *in situ* stress field, and just entered a new cycle of tectonic deformation.

Based on this similarity, it is tempting to speculate, that in Rainier Mesa and Yucca Flat, vertical gradients of hydraulic potentials may be related to tectonic processes (Section 4.5.2.4). Some support for such a possibility may be inferred from the following three sets of data. First, locally within Yucca Flat

waters that normally occur in the Paleozioc carbonates may be encountered in the overlying tuff "pile". One such example was encountered in Well #81-67, and was reported by Schoff and Moore, 1964. The results of chemical analyses of water samples from this well indicated that "mixed water", or sodium-potassium-calcium-magnesium type, occurs in the sequence of Tertiary tuffs.

Second, in Rainier Mesa there are major differences in chemical compositions of interstitial and fracture waters (Sections 4.4.5 and 4.5.2.4). Such differences were initially noted by Keller, 1960 and 1962. Utilizing interpretations of downhole electric logs and the results of laboratory measurements of electric conductivity of cores saturated with water of differing salinity, this author concluded that specific conductance of interstitial water is 25 to 35 times as great as that of fracture waters. The estimated total dissolved-solid content for interstitial water ranged from 3,700 to 5,100 ppm. In contrast, the total dissolved-solid content for fracture water in Rainier Mesa is typically less than 1,000 ppm, (White et al., 1980; Russell, 1987).

Third, the results of *in situ* stress measurements performed in Rainier Mesa indicate that the "limit equilibrium" stress conditions are shallow and restricted to the vadose zone (Section 4.2.5.3). Below the water table, conducting apertures are small and consist only of residual components.

4.2.4.3 The Area of Pahute Mesa

In Pahute Mesa, vertical gradients of hydraulic potentials are evident in the results of downhole measurements of hydraulic potentials. Such measurements were made in 16 deep wells. Discreet measurement intervals were isolated from the remainder of a borehole with inflatable packers. The results of these measurements were reported by Blankenagel and Weir, 1973 (Plates 4.2.4.3-1 and 4.2.4.3-2).

Plate 4.2.4.3-3 presents an idealization of the depthward distribution of hydraulic potentials in the area of Pahute Mesa. Within the first 2,500 feet below the water table, hydraulic potentials are reasonably constant. The value of pore pressure increases linearly with depth, in accord with the increasing hydrostatic pressure. At a greater depth, however, the depthward distribution of hydraulic potentials changes. Boreholes situated in the western sector of Pahute Mesa revealed that "overpressure" conditions are present. Boreholes situated in the eastern sector of the area, however, revealed that the values of hydraulic potentials decrease with depth, indicating that "underpressure" conditions are present.

The above spatial distribution of hydraulic potentials may be accounted for by assuming, that in Pahute Mesa, the "heat-fluid-flux" boundary conditions are locally developed along the horizontal boundary of the groundwater system. The entire area appears to be underlain by a hydrothermal convection system that is expressed as a fluid-source in line with a fluid-sink. The following three sets of data support this interpretation. First, Pahute Mesa contains a pronounced low-seismic velocity anomaly in the upper crust (Plate 3.4.1-3). Assuming that the anomaly is related to a local magma chamber, the presence of a hydrothermal convection system in this setting is not surprising. Second, downhole measurements of *in situ* temperature, performed in Pahute Mesa, revealed the presence of substantial horizontal and vertical geothermal gradients (Plate 4.2.4.3-6). At a depth of about 700 m, between Well PM-2 and Well UE-20f, the average value of the horizontal geothermal gradient is equal to about 6°C per mile. Third, the results of chemical analyses of water samples revealed, that in Pahute Mesa, there are noticeable lateral gradients involving chemical composition of groundwater (Plates 4.2.4.3-4, 4.2.4.3-5). Over the heat-fluid source, the content of sulfate and chloride anions is noticeably increased.

An alternate interpretation of the depthward distribution of hydraulic potentials in Pahute Mesa is that the "forced" convection system is present. According to this interpretation, heterogeneities in the local hydraulic conductivity structure are responsible for local upwellings of deeper fluids. As discussed in Section 3.4.6, the appearance of such a system, in terms of its contemporary characteristics, would not be significantly different from that of the thermally driven system. One feature that may lean in favor of the thermal system interpretation, however, is the fact that Pahute Mesa constititutes the recharge area of the local groundwater system. In upstream portions of the recharge area, the presence of a sizeable hydraulic source is hardly expected.

The absence of vertical gradients of hydraulic potentials in the first 2,500 feet below the water table, is noteworthy. Such an absence may indicate that the hydrothermal system is "dampened". Without additional data, one may only speculate about the nature of this damping. One such speculation is that the *in situ* stress enhancement of the local hydraulic conductivity structure is responsible. Not much is known about the state of *in situ* stress in Pahute Mesa and, therefore, it is not possible to either confirm or reject this speculation.

4.2.4.4 The Area of Yucca Mountain

Vertical gradients of hydraulic potentials are also present in the area of Yucca Mountain. Such gradients are known based on downhole measurements of hydraulic potentials performed in 7 deep wells: UE-25b#1, UE-25p#1, USW H-1, USW H-3, USW H-4, USW H-5, and USW H-6. The locations of these wells are shown on Plate 4.2.4.4-1. The measurements were made at discreet depth intervals that were isolated from the remainder of a borehole with inflatable packers. The results of these measurements were reported by Robison, 1984 (Plate 4.2.4.4-2).

Plate 4.2.4.4-3 presents an idealization of the depthward distribution of hydraulic potentials for the Yucca Mountain area. Between the water table and a depth of approximately 1,200 m, the value of hydraulic potential is more or less constant and independent of depth. Below this depth, however, higher values of hydraulic potentials were observed in three wells: UE-25p#1, USW H-1, and USW H-3. The highest increase was observed in Well USW H-1, where it was equal to about 53 m. In boreholes UE-25p#1 and USW H-3, the increase was 20 and 23 m, respectively.

Examination of Plate 4.2.4.4-2 reveals an interesting and potentially important characteristic of the local groundwater system. In boreholes where the value of hydraulic potentials increases with depth, the water level, as measured in an open hole, does not at all reflect higher pressures occurring at a greater depth (Plate 4.2.4.4-3). Evidently, groundwater enters the borehole at a depth where higher pressures exist and rises upward toward the interval of lower pressure, where it must enter the formation without noticeably raising the overall fluid level. This seems to be a curious circumstance indeed. There is, however, a logical explanation for it. In the Yucca Mountain area, the height of a water column in a borehole may be controlled by the local *in situ* stress. Locally, in upper parts of the system, either the effective normal stress (σ_{nfeff}) is equal to the critical closure pressure (σ_{cc}), or the shear stress (τ_f) is equal to the frictional resistance of fractures $(\tau_{\max})$ (Section 3.3.3). The system seems to be delicately balanced. A small rise in the value of fluid pressure causes local dilation of some fractures and, through that, causes increases in the *in situ* value of hydraulic conductivity. Such fractures may now accommodate additional volumes of water without noticeably raising the overall fluid level in a borehole.

The above explanation is similar to that proposed by Brederhoeft, 1987. Based on the review of *in situ* stress data from Yucca Mountain, this author stated that:

> " **the height of water table may be controlled by the state of stress. A rise in the water table produces hydraulic fractures which in turn increases the permeability which allows higher rates of flow which tends to lower the height of the water table below Yucca Mountain. This leads to our conclusion that the system is quite delicately balanced.**"

Clearly, the above interpretation requires, that in the Yucca Mountain area, deforming fractured medium be present. This inference is identical to the one drawn earlier, based on the texture of local hydraulic conductivity structure (Section 4.2.2). The increased values of "deep" hydraulic potentials do not appear to be the result of some imaginary lateral hydraulic barriers. Instead, the increased values of such potentials appear to be the logically expected result of the transient hydraulic transmissivity contrasts, that are developed in response to advanced tectonic straining of the Yucca Mountain area.

As discussed in Section 3.3.5, involvement of a deforming fractured medium, at a potential site of a high-level nuclear repository, is a serious matter. As such, this matter deserves open-minded considerations as well as early recognition. This may be achieved by examining the existing data along lines proposed in Section 3.3.6. As shown on Plate 4.2.4.4-4, in a deforming fractured medium the characteristic depthward distribution of hydraulic potentials should be accompanied by, at least, three other characteristics: a) different temporal behavior of hydraulic potentials above and within the "overpressure" zone [above and below the surface $Z_{(x,y,t)}$]; b) presence of the "limit equilibrium" stress conditions and the resulting low *in situ* values of the closure pressure (σ_c); and c) spatial correspondence between the *in situ* stress gradients and the gradients of hydraulic potentials.

The following discussions address all of the above characteristics, specifically, for the Yucca Mountain groundwater system. The remainder of this section deals with the temporal behavior of hydraulic potentials. The *in situ* stress data are considered in Section 4.2.5.

In the Yucca Mountain area, downhole measurements of hydraulic potentials were carried out in 28 deep wells. In some of these wells, packers and piezometers were installed so that the measurements could be made at discreet depth intervals. Since late 1982 and early 1983, most of the wells were monitored periodically. The measurements were made, approximately once-a-month, using reference and reeled steel tapes, single and multiconductor cables, and pressure transducers. The results of these measurements were reported by Robison et al., 1988.

In later 1983, continuous monitoring was intitiated in Well UE-25p#1. In 1983 and 1984, continuous records were obtained intermittently from wells USW H-1 and USW H-4. By 1986, the present network of about 20 deep wells for continuous monitoring become operational. The results of these measurements, however, are not yet available.

Plates 4.2.4.4-6 through 4.2.4.4-21 present the results of periodic measurements of hydraulic potentials from 14 wells. The locations of these wells are shown on Plate 4.2.4.4-5. The data are presented in three categories. The first category consists of hydrographs that were obtained at and near the water table, and above the "overpressure" zone. With reference to Plate 3.3.4-4, this category represents the temporal behavior of hydraulic potentials in Zone I. The data for this category are shown on Plates 4.2.4.4-6 through 4.2.4.4-16. A consistent pattern observed on all of the hydrographs is a small progressive decay of the value of hydraulic potentials during 1984.

The second category of hydrographs consists of data obtained within the "overpressure" zones. These hydrographs, therefore, represent the temporal behavior of hydraulic potentials in either Zone II or Zone III (Plate 3.3.4-4). The data for this category are presented on Plates 4.2.4.4-17 through 4.2.4.4-19.

The third category consists of hydrographs from two wells: Well UE-25 WT#6 and Well USW H-6 at a depth of 1,127 to 1,220 m. These hydrographs are shown on Plates 4.2.4.4-20 and 4.2.4.4-21. On the basis of information presented in Section 4.2.5.4, it may be concluded that both of these hydrographs were obtained from below the surface $Z_{(x,y,t)}$, and that they represent the temporal behavior of hydraulic potentials in Zone III (Plates 3.3.4-4, 4.2.5.4-98, and 4.2.5.4-99).

A consistent feature observed on all of the hydrographs from the second and third categories is the absence of the progressive decay of hydraulic potentials during 1984. In some of the wells, the values of hydraulic potentials remained constant (Plates 4.2.4.4-17 and 4.2.4.4-19). During 1984, however, in most of the wells there was a time-dependent increase of the value of hydraulic potentials (Plates 4.2.4.4-17a, 4.2.4.4-20, and 4.2.4.4-21). It is likely that such an increase represents a post-drilling recovery of well segments with very low hydraulic transmissivity.

The above observations indicate, that in the area of Yucca Mountain, the temporal behavior of hydraulic potentials differs depending upon positions of measurement intervals relative to the vertical hy-

draulic gradients. If the measurements are made at discreet depth intervals, such different behavior may be observed in a single well [e.g., Well USW H-1 (Plates 4.2.4.4-12, 4.2.4.4-12a, 4.2.4.4-17, and 4.2.4.4-17a); Well USW H-3 (Plates 4.2.4.4-13 and 4.2.4.4-18); and Well USW H-6 (Plates 4.2.4.4-15 and 4.2.4.4-21)].

To account for the above observations, there are only three possibilities: a) to discount the observations as unrepresentative or unreliable; b) to attribute the observations to some unknown, but mysterious nevertheless, causes; and c) to accept the explanation shown on Plate 4.2.4.4-4, that the difference in temporal behavior of hydraulic potentials is the expected and natural characteristic of a groundwater system indicating that this system operates in a deforming fractured medium.

At Yucca Mountain, the depthward distribution of hydraulic potentials exhibits another interesting feature. Such feature may be identified by comparing depth of occurrence of the "overpressure" zones from Plate 4.2.4.4-2 with spatial distribution of the *in situ* values of hydraulic conductivity from Plates 4.2.2-13 and 4.2.2-14. In Wells USW H-1 and USW H-3, the "overpressure" zones occur below the upper layer of the local hydraulic conductivity structure. In these two cases, therefore, the "overpressure" conditions appear to be related to the hydraulic transmissivity contrasts, that in turn, seem to be caused by advanced tectonic straining of the Yucca Mountain area. The differences in depthward distribution of the values of hydraulic potentials may be attributed to the differences in resistance against fluid flow in the two layers with contrasting hydraulic transmissivities.

In Well UE-25p#1, however, the *in situ* values of hydraulic conductivity, within and above the "overpressure" zone, are equally high (Plate 4.2.2-13). The hydraulic transmissivity contrasts, that for two other cases correlate with the "overpressure" zones, are clearly absent. The origin of "overpressure" conditions in Well UE-25p#1, therefore, must be related to some other cause.

Depthward distribution of the values of hydraulic potentials in Well UE-25p#1 indicates that across a high-angle fault contact between the tuff "pile" and carbonates, a very steep hydraulic gradient is present (Plate 4.2.4.4-2). This steep gradient, in turn, indicates the presence of a narrow hydraulic barrier that prevents equilibration of hydraulic pressures across the fault plane. This barrier is situated in the hanging wall of the fault, and may correspond to the fault conformable layer of "calcified ash", the presence of which was reported by Carr et. al., 1986. Along the fault plane, however, hydraulic transmissivity appears

to be relatively high. Such conditions are indicated by downhole temperature logs obtained during and shortly after completion of Well UE-25p#1. These temperature logs are presented on Plate 4.2.4.4-22.

Taken together, high hydraulic transmissivity along the fault contact and steep hydraulic gradient across the fault contact indicate that the tuff "pile" and the Paleozoic carbonates are hydraulically connected and, despite the fault conformable barrier, there is an upward migration of fluids. Such migration seems to be reflected in chemistry of groundwater samples collected from above and below the fault contact. The results of chemical analyses of such samples are shown on Plate 4.2.4.4-23. In terms of chemical and isotopic $(\partial^{13}C)$ composition, fluids from the tuff "pile" appear as diluted fluids from the Paleozoic carbonates. Upward migration of fluids from the carbonates into the tuff "pile" is further indicated by downhole temperature profile obtained after completion of Well UE-25p#1. This profile is shown on Plate 4.2.4.4-24. Around the Tertiary-Paleozoic contact, the temperature profile is concaved, suggesting upwelling fluids.

Based on the above considerations, it seems reasonable to attribute the "overpressure" conditions in Well UE-25p#1 to a local fluid source, that is to locally developed "upward-fluid-flux" boundary conditions. Based on interpretations of geothermal data performed in Section 4.3.3, it is reasonable to conclude that the "upward-flux" boundary conditions also pertain to local flow of terrestrial heat. Considering the high hydraulic transmissivity encountered in Well UE-25p#1, it is reasonable to postulate, that the upwelling of fluids, and the resulting depthward increase of hydraulic potentials in this well, may be caused by thermally driven convection of fluids. The local hydrothermal system is developed within and around the Paleozoic-Tertiary fault contact (Section 4.3.3).

4.2.4.5 Summary and Conclusions

Examination of the hydrologic data in this section suggests that, in the Death Valley groundwater system, vertical gradients of hydraulic potentials may be related to tectonic processes. The data are consistent with the assumption, that locally, tectonic processes either control or influence both the boundary conditions and the hydraulic conductivity structure. Such relationships, in turn, appear to be expressed through the development of spatially variable vertical gradients of hydraulic potentials.

In the vicinity of Rainier Mesa and Yucca Flat, depthward distribution of hydraulic potentials indicates that the "underpressure" conditions are present. Such conditions, in turn, indicate that hydraulic transmissivity of the Tertiary tuff "pile" is substantially less than that of the Paleozoic carbonates. Within the tuff "pile", low hydraulic transmissivity is widespread and, if compared with transmissivity of the similar tuffs at Yucca Mountain, is quite surprising. Evidently, in the vicinity of Rainier Mesa and Yucca Flat, the *in situ* stress enhancement of the local hydraulic conductivity structure is either very minimal or not present at all. Depth extent of the "limit equilibrium" stress conditions must be small and, at most, restricted to the vadose zone. Such a state of *in situ* stress requires that contemporary tectonic straining of the area is either at early stages of development or it does not involve extension.

In Pahute Mesa, both "overpressure" and "underpressure" conditions are present. This may be attributed to differences in the local boundary conditions. Lateral gradients involving groundwater temperature and chemistry, together with seismic velocity structure of the area, suggest that such boundary conditions may be the result of thermal convection of fluids.

At Yucca Mountain, the values of hydraulic gradients increase depthward, indicating presence of the "overpressure" conditions. In two boreholes, USW H-1 and USW H-3, the "overpressure" conditions appear to be related to the hydraulic transmissivity contrasts present within the Tertiary tuff "pile". It is almost certain that such contrasts are transient and related to advanced tectonic straining of the Yucca Mountain area. In Well UE-25p#1, however, presence of the "overpressure" conditions cannot be likewise justified. At this location, the "overpressure" conditions appear to be the result of thermal convection. The local hydrothermal system seems to be developed within and around the high-angle fault contact between the Tertiary tuff "pile" and the Paleozoic carbonates.

4.2.5 The In Situ Stress Field

4.2.5.1 General

In their considerations of natural groundwater systems hydrologists seldom, if ever, involve the *in situ* stress field. They prefer to regard fractured medium as a "static" medium whose hydraulic properties are not related to *in situ* stress field or to other time-dependent factors. It is in the petroleum industry, stimulation in particular, that the state of *in situ* stress receives some of its greatly deserved attention. It is recognized that, in a fractured medium, the *in situ* stress controls the conducting aperture and the *in situ* values of closure pressure (σ_c). Often, however, such recognition is limited to an immediate vicinity of a wellbore and to a uniform *in situ* stress field.

In performing reservoir analyses, petroleum engineers recognize the influence state of *in situ* stress may have on flow of gas, fluid, and heat. Conventional analyses for the stress-fluid flow coupled problem are based on an assumption that interaction is two-way. The volumetric strain influences fluid density while the resulting fluid pressure affects both the compressibility and the state of effective stress of the coupled system. Three-dimensional analytical treatment for this problem was developed by Biot, 1941. The analyses are based on an assumption of linearly elastic and infinitesimal strain behavior of a porous material.

For a fractured medium, the coupling between a changing *in situ* stress field and a field of hydraulic potentials is extremely complex and is a poorly understood problem. It appears that coherent analytical treatment for this problem has not been established. A description of a changing *in situ* stress field, in time and space, is the most important and difficult aspect of this problem.

Throughout this report a thought is advanced, that in the Death Valley groundwater system, the *in situ* stress field varies in time and space. In Sections 3.3.3 and 3.3.4, it has been shown, that eventually, such a stress field will control local hydraulic conductivity structure and, through that, it will also control local distribution of hydraulic potentials, including position and configuration of the water table. In view of potential conceptual importance of such magnitude, it seems essential to perform a review of what is known about the state of *in situ* stress in the Death Valley groundwater system.

As far as the state of *in situ* stress is concerned, the Nevada Test Site is the most thoroughly investigated location anywhere in the world. Numerous *in situ* stress determinations were made in association with underground testing of nuclear devices. Additionally, comprehensive studies of the state of *in situ* stress were made in association with various activities related to evaluations of feasibility of underground disposal of high-level nuclear wastes. The results of all of these studies were widely reported in literature (Haimson et al., 1974; Ellis and Edge, 1976; Ellis and Magner, 1980; Smith et al., 1980; Haimson, 1981; Ellis and Magner, 1982; Creveling et al., 1984; Stock et al., 1985; and Stock et al., 1986).

At the Nevada Test Site, the *in situ* stress determinations were performed in three broad areas: a) the Climax Stock; b) the area of Aqueduct Mesa and Rainier Mesa; and c) the area of Yucca Mountain. As a result, an extensive data base was established. This data base offers a unique opportunity to further test the validity of some hydrologic concepts developed in Section 3.0. There are two concepts that are of particular importance. The first concept pertains to origin of the "overpressure" and "underpressure" conditions as encountered at Yucca Mountain and around Yucca Flat, respectively. The second concept involves character of the local hydraulic conductivity structure at Yucca Mountain. The subsequent analyses explore both of these issues.

4.2.5.2 The Climax Stock

The Climax Stock is situated in the northern sector of the Nevada Test Site (Plate 4.2.5.2-1). This intrusive body consists of quartz monzonites and granodiorites of Cretaceous age. It intrudes sedimentary rocks of Pre-Cambrian and Paleozoic age. The Climax Stock and its host rocks are overlain by the Tertiary tuff "pile".

General hydrologic setting of the Climax Stock was summarized by Murray, 1981, and is shown on Plate 4.2.5.2-1. In the northeastern sector of the stock, the water table is at an altitude of above 1,100 to 1,200 m above sea level; in the southwestern sector, the water table is at an altitude of about 800 to 900 m above sea level. Beneath the Spent Fuel Test Facility, the water table occurs at an altitude of 975 m above sea level, or at a depth of about 563 m below the ground surface. A borehole drilled south of the Climax Stock revealed that the stock is situated within the Rainier Mesa-Yucca Flat "underpressure" region (Plate 4.2.5.2-1). As discussed in Section 4.2.4.2, within this region the values of hydraulic potentials in the Paleozoic carbonates are lower than those in the overlying tuff "pile". In the vicinity of Climax

Stock, the value of "underpressure" is equal to 43 m. Immediately below the water table, permeabilities are low in a range of 10^{-5} darcies (Murray, 1981).

Within the Climax Stock, comprehensive investigations of the state of *in situ* stress were undertaken. The investigations were made in two tunnel complexes: the Pile Driver Tunnel (418 m below the ground surface) and the Hard Hat Tunnel (260 m below the ground surface). Both of these tunnel complexes were excavated, initially, for the purposes of underground testing of nuclear devices. The Pile Driver Tunnel was later extended to accommodate the Spent Fuel Test Facility. The *in situ* stress determinations were made in rocks with very high stiffness. The modulus of deformation for these rocks ranges from 70 to 80 GPa ($\approx 10^7$ psi).

Prior to the emplacement of spent nuclear fuel in the Test Facility, the *in situ* stress determinations were performed at two sites. At each site, the measurements were made in three non-parallel boreholes, using the U.S. Bureau of Mines (USBM) three-component borehole deformation gauge, and employing the overcoring technique. The results of these measurements were used to calculate the magnitudes and orientations of the principal stresses, following the computational procedure described by Paneck, 1966.

The results of early attempts to define the state of *in situ* stress in the Climax Stock were reported by Ellis and Magner, 1982. These results are also shown on Plates 4.2.5.2-2 through 4.2.5.2-4. The Mohr's representations of the measured *in situ* stresses are presented on Plates 4.2.5.2-5 and 4.2.5.2-6. The Coulomb-Navier envelope of failure is also shown. For those and all of the subsequent Mohr's representations of a state of *in situ* stress, it has been assumed, that cohesion of pre-existing fractures is equal to zero and, that coefficient of friction for fractures ranges from 0.60 ($\phi = 30°$) to 1.0 ($\phi = 45°$).

At both of the sites, the overcore measurements were made at a shallow borehole depth, ranging from 4.8 to 6.5 m from the tunnel rib. It is unlikely, therefore, that the results of such measurements are representative of the far-field *in situ* stress conditions. Instead, these results appear to represent stress concentrations induced by underground openings. They are considered as suspect and, therefore, will not be used in the subsequent analyses.

After spent fuel was removed from the Test Facility, comprehensive investigations of the state of *in situ* stress were undertaken. The overcore measurements were made in two distinct regions surrounding

the Spent Fuel Test Facility. One such region includes two fairly long pillars, that are situated between the central canister room and the north and south heater drifts. Four short test boreholes were drilled into the pillars. The main purpose of these boreholes was to investigate stress concentrations introduced by underground openings. The results of these investigations are not considered here.

The second measurement region is located in a virgin bedrock, that was reached by four test boreholes drilled outward from the south heater drift and from the tail drift extension. Two of the boreholes (ISS-8 and ISS-9) were horizontal, and the other two (ISS-10 and ISS-11) were inclined 45° upward and downward, respectively. The main purpose of these boreholes was to investigate the state of *in situ* stress, that is unaffected by underground openings and by heating of the country rock. The study was conducted employing two types of strain-relief measurement instruments: a) the USBM borehole deformation gauge and b) the Commonwealth Scientific and Industrial Research Organization gauge, so called CSIRO gauge. Both of the gauges were modified to include a thermistor heat-sensing element. Such modification allowed for direct measurements of ambient temperature at the actual test locations. Twenty-eight USBM gauge tests and 22 CSIRO gauge tests were successfully conducted in the four far-field boreholes.

To obtain data concerning the magnitude and orientation of the principal stresses, both the USBM and CSIRO gauges rely on strain relief produced by overcoring. During a test, one of the gauges is inserted into a 38-mm-diameter borehole, and then it is coaxilly overcored with a 150-mm-diameter core barrel. The gauges monitor changes in size and shape of the central borehole. The USBM and CSIRO gauges, however, are fundamentally different in the manner in which they measure borehole deformation. The USBM gauge uses six strain cantilevers, mounted so that they measure three equally spaced borehole diameters in a single transverse plane. The instrument cannot measure axial deformation of the central borehole and, therefore, cannot measure all the strain components that are required to determine the complete *in situ* stress tensor. The information obtained, however, is sufficient to derive two normal stresses that exist in a plane perpendicular to the borehole axis. These stresses are combined with the results of strain-relief measurements from, at least, two other non-parallel boreholes to obtain the magnitudes and orientations of the principal stresses. Such procedure yields a reliable result but only if the *in situ* stress field does not contain steep stress gradients. The CSIRO gauge consists of nine variously oriented strain gauges encased in a thin plastic cylinder. The instrument is irretrievably bonded to a borehole wall using a special epoxy-based adhesive. The nine strain gauges are arranged such that they measure all three

principal strain components. This allows for direct calculation of the principal stresses using the thin-walled cylinder theory.

A total of 22 three-dimensional determinations of the far-field state of *in situ* stress are available for the Climax Stock. The data base consists of 3 USBM gauge sets and 18 CSIRO gauge solutions. The data were reported by Creveling et al., 1984. They are also shown on Plates 4.2.5.2-7 through 4.2.5.2-11. Each of the plates presents the upper-hemisphere polar equal-area projections of the principal stress axes, along with tabulated magnitudes of the principal stresses.

Plates 4.2.5.2-13 through 4.2.5.2-32 present the Mohr's representations of the *in situ* stresses from the Climax Stock. Such representations are useful in evaluating whether or not the "limit equilibrium" *in situ* stress conditions are present at a particular measurement station. They are also useful in estimating *in situ* values of the closure pressure (σ_c). This pressure corresponds to a value of pore pressure build-up that is required for initiation of the Griffith failure along a suitably oriented pre-existing fracture. Such value, of course, is equal to a value of the least principal effective stress.

A summary of the *in situ* stress data for the Climax Stock is presented on Plate 4.2.5.2-33. This summary is shown in the form of closure pressure (σ_c) vs. depth plot. Additionally, each "limit equilibrium" data set is identified. The plot was prepared without considering the USBM data. The CSIRO gauge data obtained in borehole ISS-9 at a depth of 46.9 feet were also excluded.

Examination of the data reveals four characteristics of the state of *in situ* stress in the Climax Stock, around the Spent Fuel Test Facility. First, the free-field *in situ* stresses are characterized by a significant degree of spatial heterogeneity. The heterogeneity involves both the orientation and magnitude of the principal stresses. The data are of sufficient quality and quantity to conclude that the heterogeneity reflects real state of *in situ* stress (Creveling et al., 1984). Clearly, significant shear and normal stress gradients are present.

Second, the state of *in situ* stress deviates significantly from a gravitational stress field, both in the relative stress magnitudes and orientation. An unusual aspect of the *in situ* stresses is significant deviation of the measured vertical stress from the expected overburden stress. Given a representative density of 2.65 g/cm^3 for the Climax Stock granite, the 418 m of overburden is expected to produce a vertical

normal stress of 10.9 MPa ($\approx 1.6 \times 10^3$ psi). The determined magnitudes of the normal vertical stress, however, are commonly different. A comparison of the measured vertical normal stresses with those expected was reported by Creveling et. al., 1984. This comparison revealed that the average vertical stress measured only in borehole ISS-11 agrees with the expected overburden value of 1.6×10^3 psi. The average vertical stress in boreholes ISS-8 and ISS-9 together, and in borehole ISS-10, are about 60 and 120 percent of the expected value, respectively. Considering high quality of the borehole deformation data and essentially linear-elastic and isotropic behavior of the granite overcores, such discrepancy is far too large to be attributed entirely to measurement errors. A real *in situ* stress "anomaly" appears to be present.

Third, the "limit equilibrium" stress conditions were encountered in only 4 out of 17 data sets. The *in situ* values of closure pressure (σ_c) are generally high, ranging from 24 to as much as 91 bars. The average value of closure pressure is equal to approximately 46 bars. With the exception of borehole ISS-10, at a borehole depth of 64.4 feet, the "limit equilibrium" stress conditions are accompanied by relatively high values of the closure pressure, ranging from 36 to 49 bars. These high values indicate that isolated occurrences of the "limit equilibrium" stress conditions are not caused by advanced extensional straining of the Climax Stock. It is possible, that either "remnant" tectonic stress concentrations or stress concentrations resulting from nearby underground nuclear detonations are responsible for isolated occurrences of the "limit equilibrium" stress conditions. It is equally possible, however, that such conditions are the result of contemporary tectonic straining of the Climax Stock. This straining seems to be dominated by compression and/or shear couple; involvement of extensional straining seems to be limited.

Fourth, in the vicinity of Climax Stock, depth extent of the "limit equilibrium" stress conditions appears to be small. It is unlikely that such conditions are present below the water table. It is expected, therefore, that the hydraulic conductivity structure of the tuff "pile' in this area is not influenced by the local *in situ* stress field. The *in situ* values of hydraulic conductivity are low. Such low values, together with higher *in situ* values of hydraulic conductivity in the underlying Paleozoic carbonates, allow for the existence of characteristic for this area hydraulic "underpressure" conditions.

4.2.5.3 The Area of Rainier Mesa and Aqueduct Mesa

Rainier Mesa and Aqueduct Mesa are located in the northern sector of the Nevada Test Site, east from Yucca Flat (Plate 4.2.5.3-1). Both of these mesas are flat-topped mountains bounded by steep cliffs

on all four sides. They are composed of gently dipping rocks of the Tertiary tuff "pile". The combined thickness of these rocks ranges from 700 to 1,400 m (Thordarson, 1965). Stratigraphic position and lithologic character of these rocks are comparable to the tuffs present in the Yucca Mountain area. Similarly as at Yucca Mountain, the tuff "pile" overlies the Paleozoic carbonates.

Four tunnel complexes have been excavated on western slopes of the mesas. Some of the tunnels penetrate nearly 3 km into the mesas and contain up to 13 km of underground workings. It is within these workings that underground nuclear detonations are carried out.

The tuffs hosting the four tunnel complexes are typically low strength and low modulus materials. The unconfined compressive strength of these rocks averages at about 20 MPa. Elastic moduli range from 1.4 to 10 GPa, averaging less than 6.9 GPa.

Hydrologic settings of Rainier Mesa and Aqueduct Mesa have been described by Thordarson, 1965; Winograd and Thordarson, 1975; and Russell, 1987. The top of the zone of saturation has been established based on numerous test boreholes. In the tuff "pile", altitudes of the water table vary up to 100 m; the mean altitude is about 1,820 m above sea level (Thordarson, 1965). In the Paleozoic carbonates, however, potentiometric surface is substantially lower. In Well #87-22, for example, it was encountered at an altitude of about 1,273 m above sea level (Winograd and Thordarson, 1975). The value of "underpressure" ranges from about 160 to 547 m. Laboratory testing of core samples yielded an interstitial porosity ranging from 25 to 38 percent. The representative range of the saturated matrix hydraulic conductivity, for the tuffs of Rainier Mesa, is from 0.2×10^{-9} to 9×10^{-9} m/sec (Russell, 1987).

Comprehensive investigations of the state of *in situ* stress were also undertaken in the areas of Rainier Mesa and Aqueduct Mesa. The investigation consisted of numerous USBM overcoring tests and a sequence of hydrofracturing tests. The overcoring tests were made in a number of horizontal and vertical boreholes drilled from various tunnels. In the vicinity of tunnel complex U12n, 7 hydrofracturing tests were made in a vertical borehole drilled from the slope of Rainier Mesa. The combined available data base consists of 9 three-dimensional *in situ* stress determinations, made with the USBM gauge, and 13 hydrofracturing tests.

The hydraulic fracturing technique offers an opportunity for direct measurement of the magnitude of the least principal stress and, under certain conditions, may indirectly yield the magnitude of the greatest principal stress (Hubbert and Willis, 1957; Haimson and Fairhurst, 1967). The technique involves creation of a fracture in an isolated interval of a borehole, and observations of the behavior of such a fracture during repeated borehole pressurization cycles. A one- to two-meter long segment of a borehole is sealed off between two inflatable rubber packers, and is connected to the ground surface by a tubing. The technique is based on analytical solution for the state of stress around a cylindrical hole in an elastic and isotropic medium. If a test borehole is drilled parallel to one of the principal stresses, hydraulic fracture formed during a test will be oriented perpendicular and parallel to the least and major principal stresses, respectively. The pressure at which the fracture closes is referred to as the "instantaneous shut-in pressure." It corresponds to the magnitude of the least principal effective stress. The borehole pressure required to initiate the hydraulic fracture is known as the "breakdown pressure". This pressure is equal to

$$\sigma_b = 3\sigma_3 - \sigma_1 + T - p$$

where:

σ_b - breakdown pressure;

σ_1 and σ_3 - major and least principal stresses;

T - tensile strength of rock; and

p - pore pressure.

Once the values of σ_1 and σ_3 are obtained, the value of the intermediate principal stress (σ_2) is still required to characterize the complete *in situ* stress tensor. Commonly, it is assumed that the intermediate principal stress is oriented vertically, and its magnitude is equal to the value of overburden stress.

The results of *in situ* stress determinations, from Rainier Mesa and Aqueduct Mesa, were reported by Ellis and Magner, 1982 (overcoring), and Haimson, 1983 (hydraulic fracturing). Both of the data sets are also shown on Plates 4.2.5.3-2 through 4.2.5.3-16. The Mohr's representations of all of these data are presented on Plates 4.2.5.3-17 through 4.2.5.3-38. Plate 4.2.5.3-39 is the summary diagram for the area of Rainier Mesa and Aqueduct Mesa. The diagram shows depthward distribution of the *in situ* values of

closure pressure (σ_c), each "limit equilibrium" data set is identified. In preparing the diagram, all of the available 22 data sets were used.

The results of *in situ* stress determinations made in Rainier Mesa and Aqueduct Mesa indicate the following four characteristics of the state of *in situ* stress in this area. First, the local *in situ* stress field is characterized by a pronounced heterogeneity. Such heterogeneity involves both the orientation and magnitude of the principal stresses. The degree of this heterogeneity, however, is not as great as that in the Climax Stock. In proceeding from north to south, along a 6 km long western slope of the mesas, the orientation of σ_1 changes gradually from north-northeast to northeast and then back to north-northeast. Such apparently systematic change may account for greater uniformity of the *in situ* stresses from hydraulic fracturing measurements. These measurements were made at a single location, whereas the overcoring measurements were made in four tunnel complexes, separated one from another by a distance of 1 to 2 km. In both of the data sets, there is a tendency for the least principal stress (σ_3) to be oriented northwest-southeast. The plunge of this stress is variable, ranging from 2° to a maximum of about 50°, and averaging 19°. The greatest and intermediate principal stresses tend to be oriented from north-south to northeast-southwest. There does not appear to be any consistency in the plunge of these stresses. Low plunges of the greatest principal stress, however, seem to be more frequent.

Second, in Rainier Mesa and Aqueduct Mesa, the *in situ* stresses appear to be mostly tectonic in origin, of course, with some modifications caused by topography of the area. Tectonic origin of the *in situ* stresses is suggested by a) the strongly deviatoric character of the principal stresses; b) the presence of significant shear stresses acting in horizontal and vertical planes, as indicated by strong variability of the orientations and plunges of the principal stresses; and c) a tendency for the major principal stresses to be inclined at a relatively low angle. Magnitudes of the normal vertical stress, determined based on the results of overcoring measurements, noticeably deviate from the expected overburden values. At a depth of 338 m below the ground surface, magnitude of the measured vertical normal stress is 5 percent lower than expected. The discrepancy increases depthward; at a depth of 400 m it averages 10 percent, increasing to 16 percent at a depth of 442 m. It is likely, that such depthward increase indicates a depthward increasing prominence of shear stresses.

Third, in the areas of Rainier Mesa and Aqueduct Mesa, the "limit equilibrium" *in situ* stress conditions are not widespread. These conditions were encountered in 7 out of the 22 data sets. The *in situ*

values of closure pressure (σ_c) are lower than those from the Climax Stock. These values range from 10 to 57 bars. The average value of closure pressure, however, is still considerable, in a range of 32 bars. It is interesting to note that, in contrast to the Climax Stock, the "limit equlibrium" stress conditions are accompanied by lower than average *in situ* values of the closure pressure. This suggests that in the areas of Rainier Mesa and Aqueduct Mesa extensional straining may be responsible for local occurrences of the "limit equilibrium" stress conditions.

Fourth, in the areas of Rainier Mesa and Aqueduct Mesa, depth extent of the "limit equilibrium" stress conditions is variable, but relatively small. Locally, such conditions are developed to a depth of about 442 m. For the majority of test locations, however, depth extent of the "limit equilibrium" conditions is less than 400 m. Below the water table, the *in situ* hydraulic conductivity of the tuff "pile" is, largely, unaffected by the local *in situ* stress field, and, therefore, remains relatively low. Similarly as around the Climax Stock, this low hydraulic conductivity, together with the higher hydraulic conductivity in the underlying carbonates, allows for the presence of large "underpressures."

4.2.5.4 The Area of Yucca Mountain

The area of Yucca Mountain is located along the western boundary of the Nevada Test Site (Plate 4.2.5.4-1). This north-south elongated mountain consists of eastward dipping rocks of the Tertiary tuff "pile". Local stratigraphic section is bounded by the Tiva Canyon Member of the Paintbrush Tuff at the top and by the Lithic Ridge Tuff at the bottom. In the central sector of the area, thickness of the tuff "pile" is greater than 3 km. In the eastern sector of the area, however, a structural "high" is present, and thickness of the tuff "pile" is substantially lower. In exploratory borehole UE-25p#1, for example, the tuff "pile" was found to be about 1,200 m thick. Similarly, as in the area of Rainier Mesa, the tuff "pile" overlies the Paleozoic carbonates, specifically the Lone Mountain Dolomite and Roberts Mountain Formation, both of the Silurian age.

Local hydrologic setting of Yucca Mountain has been considered previously (Sections 4.2.2, 4.2.3, and 4.2.4). There are two noteworthy features of this setting. First, the local hydraulic conductivity structure is characterized by an extreme degree of spatial heterogeneity. Such heterogeneity is clearly expressed in both the *in situ* values of hydraulic conductivity and the horizontal and vertical gradients of hydraulic potentials. Second, depthward distribution of hydraulic potentials indicates presence of the "overpressure"

conditions. Such conditions are in contrast to the two previously considered areas, and offer an opportunity for further considerations of hydrologic importance of the state of *in situ* stress.

In considerations of the suitability of the Yucca Mountain site to accommodate a high-level nuclear waste repository, one of the main technical issues involves the origin and long-term stability of the local hydraulic conductivity structure. Previously it has been demonstrated, that a) the characteristic three-layer texture of the local hydraulic conductivity structure (Section 4.2.2); b) presence of the "overpressure" conditions (Section 4.2.4.4); and c) the characteristic temporal behavior of hydraulic potentials (Section 4.2.4.4), are all consistent with the assumption that the local *in situ* stress field is the main controlling factor. Further testing of this assumption may be performed in light of the *in situ* stress data. As was explained in Section 3.3.6, two questions are of particular relevance: a) Does the contemporary *in situ* stress field at Yucca Mountain include the four stress states shown on Plate 3.3.3.2-5? and b) Are the gradients of hydraulic conductivity and/or of hydraulic potentials conformable with the gradients of *in situ* stress? Another related question is: Is the contrasting hydrologic setting of Yucca Mountain, relative to the Climax Stock and Rainier Mesa and Aqueduct Mesa, justifiable in terms of the *in situ* stress data?

It is important to recognize that the currently available data are of sufficient quality and quantity to provide somewhat reliable answers to all of the above questions. Furthermore, a modest investment of time and money is required to achieve substantial improvement of these data, should such improvement be deemed necessary.

The data regarding the state of *in situ* stress at Yucca Mountain are provided by two sources: the hydraulic fracturing measurements and the Cooper-Brederhoeft injection tests. The hydraulic fracturing measurements were carried out in four deep wells, whereas the Cooper-Brederhoeft injection tests were made in another eight deep wells. The combined data base consists of over 120 data sets, covering an area of about 30 km^2 down to a depth of about 1,800 m.

The hydraulic fracturing measurements were made in the following wells: USW G-1, USW G-2, USW G-3, and UE-25 p#1. The locations of these wells are shown on Plate 4.2.5.4-1. A total of 20 measurements were made in both the vadose zone and below the water table. The results of these measurements were reported by Stock et al., 1985, and Stock et al., 1986. They are also shown on Plates

4.2.5.4-2 through 4.2.5.4-6. The Mohr's representations of the hydraulic fracturing data are shown on Plates 4.2.5.4-7 through 4.2.5.4-26.

With reference to conceptual considerations undertaken in this report, there are two obvious limitations of the hydraulic fracturing measurements as performed at Yucca Mountain. First, the majority of measurements were made in unfractured, narrow, and widely spaced segments of test boreholes. Consequently, two main topics of interest, in studies of the state of *in situ* stress in a deforming fractured medium, received little attention. Such topics are a) gradients of *in situ* stress developed at a scale of a fracture and b) minimum values of the mean stress $\frac{1}{2}(\sigma_1 + \sigma_3)$. Second, during execution of a hydraulic fracturing test, an operator seeks segments of a borehole where build-up of hydraulic pressure may be achieved with pumps of relatively modest discharge capacities. In other words, the operator avoids segments of a borehole that appear as highly transmissive and where both the "instantaneous shut-in pressure" and "breakdown pressure" are not discernible. In studies of the state of *in situ* stress in a deforming fractured medium, however, such segments are of major diagnostic importance. As was explained in Sections 3.3.3.2 and 3.3.3.3, in a deforming fractured medium there are regions where the *in situ* values of the mean effective stress $\frac{1}{2}(\sigma_1 + \sigma_3)_{\text{eff}}$ are very small and, consequently, the *in situ* values of closure pressure (σ_c) are likewise very small. Such a medium may also contain "singular" points, around which $\sigma_c \rightarrow 0$. For these points, under conditions of slightly increased fluid pressure, the *in situ* value of hydraulic conductivity is very large. A positive and reliable detection of both "isotropic" and "singular" points is of major importance in studies of the state of *in situ* stress in a deforming fractured medium.

Despite their limitations, the hydraulic fracturing measurements performed at Yucca Mountain yielded very important and revealing data. This is particularly apparent if such data are considered together with the results of Cooper-Brederhoeft injection tests.

During a Cooper-Brederhoeft injection test, a segment of a test borehole is sealed off between two inflatable rubber packers, and is then connected to the ground surface by a tubing. A surface controlled valve and a pressure transducer, with the ground surface readout, are attached to the tubing bottom. A slug of water is introduced into the tubing. Pressure recorders are used to monitor leakage above and below the packers. A virtually instantaneous pulse of hydraulic pressure is introduced, into the isolated interval, by opening the bottom value. A decay of the pressure pulse is monitored using the bottom pressure transducer and the ground surface readout. Typically, the results of monitoring are presented in the form

of a plot of the log of elapsed time against either the remaining hydraulic pressure (H) or the ratio of remainig hydraulic pressure (H) to applied hydraulic pressure (H_o).

The Cooper-Brederhoeft injection tests are normally employed to determine the *in situ* value of hydraulic conductivity for a specific segment of a formation. Under certain conditions, however, the injection tests are similar to the hydraulic fracturing experiments and, likewise, may produce data regarding the state of *in situ* stress. This occurs if a) *in situ* values of the closure pressure (σ_c), or the "instantaneous shut-in pressure", of a formation are relatively small and b) values of the applied hydraulic pressure (H_o) exceed the values of σ_c. In the case of injection tests made at Yucca Mountain, both requirements were satisfied. The results of hydraulic fracturing measurements indicated that the values of "instantaneous shut-in pressure" are quite small. For example, the minimum values of this pressure in boreholes USW G-1, USW G-3, and UE-25p#1 were 35, 37, and 39 bars, respectively. The values of applied hydraulic pressure (H_o), however, were very substantial. These values ranged from a minimum of 19.6 to a maximum of 54.6 bars. The majority of injection tests was made using the value of H_o ranging from 30 to 50 bars.

When compared with the hydraulic fracturing measurements, the Cooper-Brederhoeft injection tests offer two distinct advantages. First, in contrast to the hydrofracture measurements, the injection tests emphasize areas where the *in situ* values of closure pressure are lowest. Second, large segments of a borehole are involved in testing.

Interpretation of the Cooper-Brederhoeft injection tests, as performed here, is based on analyses that were presented in Section 3.3.3.2. A main point in performing such interpretation is identification of a value of fluid pressure build-up that is required to initiate the Griffith failure along pre-existing fractures. Such value, of course, is equal to the *in situ* value of closure pressure (σ_c) or, in terminology employed in hydraulic fracturing, to the value of "instantaneous shut-in pressure".

Plate 4.2.5.4-27 presents the interpretation key that was used in performing interpretations of injection curves from Yucca Mountain. The key was developed, recognizing that conceivably, there are three relationships between the closure pressure (σ_c) and the applied hydraulic pressure (H_o). Each of these relationships has a distinct expression on the H/H_o vs. $\log(t)$ plot. Such expressions are shown on Plate 4.2.5.4-27 as curves A, B, and C.

Curve A is characteristic of a situation where the ratio σ_c/H_o exceeds 1. In this situation, the velocity of slug movement is controlled by residual aperture and/or interstitial permeability, as present within the isolated interval, and by the remaining hydraulic pressure (H). The resulting H/H_o vs. $\log(t)$ curve is characteristically S-shaped. The average velocity of slug movement is measured in tens of centimeters per minute or less.

Curve B is characteristic of a situation where the ratio σ_c/H_o is smaller than 1, but greater than some small number, say 0.2. Early during the injection test, when the value of H relative to σ_c is large, the velocity of slug movement is controlled by the aperture of an opened pre-existing fracture. The average velocity of slug movement is measured in tens of meters per minute. However, during later stages of the test, when the value of H is smaller that σ_c, the velocity of slug movement is controlled again by residual apertures. It is measured in centimeters per minute or less. On the H/H_o vs. $\log(t)$ curve, the *in situ* value of closure pressure (σ_c) is indicated by the characteristic inflection.

Curve C is characteristic of a situation where the value of σ_c relative to H_o is very small, such as around "singular" points. During the entire injection test, the velocity of slug movement is controlled either by the aperture of an opened pre-existing fracture or by the orifice of injection tubing. The velocity of slug movement may exceed 100 m per minute.

In the case of Yucca Mountain, the *in situ* values of closure pressure (σ_c), derived from the B-type of injection curves, are of major practical and conceptual importance. The following three points underline such importance. First, the interpretation that these curves indicate a minimum value of the local closure pressure appears to be secure beyond a reasonable doubt. Apart from doubting the validity of measurements and security of packers, there does not appear to be an alternate explanation for the unique shape of these curves. Furthermore, in few instances the values of closure pressure from the injection curves are directly comparable to the values of "instantaneous shut-in pressure," as produced during the hydraulic fracturing measurements. Examples of such comparability are a) boreholes USW G-3 and USW H-3, at a depth of about 1,000 m; b) boreholes USW H-1 and USW G-1, at a depth of about 700 m; and c) boreholes UE-25p#1 and UE-25b#1, at a depth of about 1,100 m (Plates 4.2.5.4-98 and 4.2.5.4-99).

Second, presence of the *in situ* values of closure pressure substantially lower than 30 percent of the effective overburden stress is a diagnostic feature for a changing and reversible *in situ* stress field, that is in

advanced "limit equilibrium" conditions. The following reasoning is offered in support of this statement. The measured *in situ* value of closure pressure fixes a position of the stress circle relative to the failure envelope. Slope of this envelope, which corresponds to the value of friction coefficient for natural fractures, controls the size of the stress circle [maximum possible value of the stress difference $(\sigma_1 - \sigma_3)_{max}$]. Assuming that the friction coefficient is equal to 0.60, the maximum possible value of the stress difference is $(\sigma_1 - \sigma_3)_{max} = 2\sigma_c$. This yields the maximum possible value of the major principal stress $\sigma_1 = 3\sigma_c$. For example, presence of a borehole segment with $\sigma_c = 20$ bars, at a depth of 1,000 m, indicates that the maximum possible value of the major principal stress is $\sigma_1 = 60$ bars. Value of the effective normal stress produced by 1,000 m of overburden and 500 m of water, however, is equal to $\sigma_{ov.eff} \cong 220$ bars - 50 bars = 170 bars. Clearly, in such a situation, the value of σ_1 and the resulting values of $(\sigma_1 - \sigma_3)_{max}$ and $\frac{1}{2}$ $(\sigma_1 + \sigma_3)_{max}$ are substantially lower than those from the overburden load. As was explained in Section 3.3.3.2, such discrepancy indicates that local values of the post-peak shear displacement $(u_s - u_{sp})$ are greater than zero, and that wallrocks of some fractures are stretched. Of course, such circumstances may arise only in a deforming fractured medium and may not be a permanent feature of the *in situ* stress field. Sooner or later the *in situ* values of $(\sigma_1 - \sigma_3)$ and $\frac{1}{2}$ $(\sigma_1 + \sigma_3)$ will be adjusted to be in accord with, at least, the overburden load.

Third, a reasonably secure recognition of the abnormally low *in situ* values of closure pressure from the B-type of injection curves paves a way toward a reasonable interpretation of the C-type of injection curves. As discussed in Section 3.3.3.2, the presence of "singular" points in a deforming fractured medium, that otherwise displays substantial reductions in the values of $(\sigma_1 - \sigma_3)$ and $\frac{1}{2}$ $(\sigma_1 + \sigma_3)$, is only a natural and logical expectation. Consequently, **interpretation that the C-type of injection curves indicates the presence of "singular"points is a viable and reasonable alternative to a true but hollow statement that a highly transmissive bedrock was encountered.**

Returning to the Yucca Mountain data base, the Cooper-Brederhoeft injection tests were carried out in eight deep wells: USW H-1, USW H-3, USW H-4, USW H-5, USW H-6, UE-25b#1, UE-25p#1, and USW G-4. The locations of these wells are shown on Plate 4.2.5.4-28. The injection tests were made below the water table, to a depth of approximately 1,800 m. About 120 injection curves were produced. These curves are shown on Plates 4.2.5.4-29 through 4.2.5.4-94. The data are presented in four categories. Plates 4.2.5.4-29 through 4.2.5.4-34 present all the A-type of injection curves and, for each tested segment, indicate a lower limit for the value of σ_c. Such limit is equal to the value of applied hydraulic pressure

(H_o). Plates 4.2.5.4-35 through 4.2.5.4-48 present the B-type of injection curves. The plates also show the values of σ_c as identified from each of the injection curves. These values correspond to the values of hydraulic pressure at which the injection curves exhibits the characteristic inflection. Plates 4.2.5.4-49 through 4.2.5.4-70 present all of the C-type of injection curves. The fourth category of injection curves contains all curves for which no interpretation was made. Such curves are shown on Plates 4.2.5.4-71 through 4.2.5.4-94.

A summary of the *in situ* stress data from Yucca Mountain is presented on Plates 4.2.5.4-95 and 4.2.5.4-96. This summary is presented in the form of a plot showing depthward distribution of the closure pressure (σ_c) and identifying each "limit equilibrium" data set. The plot was prepared using the data from all hydraulic fracturing tests and from the B-type of injection curves. Additionally, the Yucca Mountain data base is sufficiently broad to allow construction of two *in situ* stress cross-sections. Such cross-sections are shown on Plates 4.2.5.4-98 and 4.2.5.4-99. Locations of the cross-sections are shown on Plate 4.2.5.4-97. These locations are identical to the locations of *in situ* hydraulic conductivity cross-sections, that are shown on Plates 4.2.2-13 and 4.2.2-14. The *in situ* stress cross-sections are based on the hydraulic fracturing data as well as on the data from A-, B-, and C-type injection curves. The cross-sections also show locations of the "overpressure" zones relative to the surface $Z_{(x,y,t)}$. Location and configuration of this surface was interpreted based on the results of hydraulic fracturing measurements, supplemented by the values of σ_c from the B-type injection curves.

The combined data base allows for recognition of four important characteristics of the contemporary *in situ* stress field in the area of Yucca Mountain. First, **the local *in situ* stress field is characterized by a pronounced heterogeneity involving the magnitude of the least principal stress (σ_3).** It is not known whether or not the orientation of this stress is also involved. Clearly, important lateral and vertical *in situ* stress gradients are present. The most noticeable is a lateral *in situ* stress gradient between the northern and central sectors of Yucca Mountain. In the northern sector, the "limit equilibrium" stress conditions are absent and, similarly as in the Climax Stock and around Rainier Mesa and Aqueduct Mesa, the *in situ* values of closure pressure are high. In the central and eastern sectors, however, the *in situ* values of σ_c are significantly lower. The "limit equilibrium" stress conditions extend down to a considerable depth, and far below the water table. Furthermore, in some boreholes vertical gradients of the least principal stress were encountered. Such gradients are expressed as a sudden increase in the *in situ* value of closure pressure, that is far too large to be accounted for by the increasing overburden load. Examples of

such increases are a) boreholes USW H-1 and USW G-1, at a depth of about 700 m; b) borehole USW H-6, at a depth of about 1,000 m; and c) borehole UE-25p#1, at a depth of 1,300 m. Both the lateral and vertical gradients of *in situ* stress indicate, that at Yucca Mountain, the surface $Z_{(x,y,t)}$ is colinear neither with topography nor stratigraphy of this area. This surface, although poorly constrained by the available data, seems to be striking northeast and dipping southeastward (Plates 4.2.5.4-98 and 4.2.5.4-99).

Second, **the Yucca Mountain *in situ* stress field is mostly tectonic in origin**. Such origin is indicated by two lines of evidence: a) presence of the non-conformable with respect to topography *in situ* stress gradients and b) presence of the abnormally low *in situ* values of $(\sigma_1 - \sigma_3)$ and $\frac{1}{2}(\sigma_1 + \sigma_3)$. The relatively steep and systematic inclination of the surface $Z_{(x,y,t)}$ suggests that the Yucca Mountain *in situ* stress heterogeneity is produced locally, most likely in response to creep along a local fault. Such a fault seems to be located somewhere in the eastern sector of the area. The possible northeast strike of the surface $Z_{(x,y,t)}$ suggests that the causative fault may likewise be oriented northeast.

Third, **the Yucca Mountain *in situ* stress field exhibits all diagnostic characteristics of an evolved *in situ* stress field developed in a deforming fractured medium** (Sections 3.3.3.2 and 3.3.6). The "limit equilibrium" stress conditions are clearly evident in the results of both the hydraulic fracturing measurements and the Cooper-Brederhoeft injection tests. In the central and eastern sectors of the area, the depth extent of such conditions is substantial, and exceeds 1,300 m. In the northern sector, however, this depth extent is much smaller; in borehole USW G-2, for example, the surface $Z_{(x,y,t)}$ occurs at a depth smaller than 300 m. The *in situ* values of closure pressure range from very small to a maximum of about 240 bars. Typically, the *in situ* values of σ_c are less than 30 percent of the effective overburden stress. There is a well expressed tendency for such values to diminish southeastward. As expected, the *in situ* values of σ_c from the B-type of injection curves are consistently lower than those from the hydraulic fracturing measurements. This discrepancy is attributed to the limitations of the hydraulic fracturing procedure employed, rather than to uncertainties associated with interpretations of the injection curves. The presence of numerous C-type injection curves suggests that the local *in situ* stress field contains "singular" points. The frequency and maximum depth of occurrence of such points increases southeastward, in accord with dip direction of the surface $Z_{(x,y,t)}$. At Yucca Mountain, the presence of "singular" points is not surprising. On the contrary, the presence of such points only confirms the previous conclusion, that in the Yucca Mountain area, a deforming fractured medium is present.

Fourth, in contrast to the Climax Stock and Rainier Mesa and Aqueduct Mesa, **the Yucca Mountain *in situ* hydraulic conductivity structure is strongly influenced by the local *in situ* stress field.** Such influence may be best demonstrated by a comparison of the hydraulic conductivity cross-sections on Plates 4.2.2-13 and 4.2.2-14 with the *in situ* stress cross-sections on Plates 4.2.5.4-98 and 4.2.5.4-99. The spatial correspondence of the gradients of hydraulic conductivity with the gradients of *in situ* stress is evident. The *in situ* stress enhancement of the hydraulic conductivity structure is clearly depth-dependent, and seems to be responsible for development of the "overpressure" conditions in accord with the conceptual model as developed in Section 3.3.4.

4.2.5.5 Summary and Conclusions

The analyses of *in situ* stress data from the Death Valley groundwater system allow for two broad conclusions to be drawn. First, in this groundwater system, the *in situ* stress field is characteristically variable in space. Two different *in situ* stress regimes were identified. The first regime is present in two neighboring areas (i.e., the Climax Stock and Rainier Mesa and Aqueduct Mesa). In both of these areas, depth extent of the "limit equilibrium" stress conditions is small, and appears to be less than the thickness of the vadose zone. The *in situ* values of closure pressure are high, generally in excess of 30 percent of the overburden stress. At a depth of about 400 m, the values of σ_c range from 25 to 90 bars. In both of these areas, the *in situ* stress field exhibits a pronounced heterogeneity, that involves both the orientation and magnitude of the principal stresses. An unusual aspect of the local stress field is a discrepancy between the measured values of vertical normal stress and the calculated values of overburden stress. The discrepancy is too large and too persistent to be attributed entirely to measurement errors. In the Climax Stock, the measured vertical stress is about 60 and 120 percent of the calculated overburden stress. In the area of Rainier Mesa - Aqueduct Mesa, the measured vertical stress is consistently lower than the calculated overburden stress. Magnitude of the discrepancy increases depthward, from 5 percent at a depth of 325 m to about 18 percent at a depth of 442 m. The second *in situ* stress regime was identified at Yucca Mountain. In this area, depth extent of the "limit equilibrium" stress conditions ranges from less than 300 m to more than 1,300 m. With the exception of the northern sector of Yucca Mountain, the "limit equilibrium" conditions are present below the water table. The surface $Z_{(x,y,t)}$ appears to be striking northeast, dipping moderately toward southeast. Such configuration of this surface is the result of lateral and vertical gradients involving, at least, the magnitude of the least principal stress (σ_3). The *in situ* values of closure pressure are abnormally low, particularly in the central and eastern sectors of Yucca Mountain. At a depth

of 1,000 m, the *in situ* values of closure pressure range from 15 to 50 bars. Furthermore, the presence of numerous C-type injection curves strongly suggests that the *in situ* stress field contains "isotropic" and "singular" points, at which $(\sigma_1 - \sigma_3) \rightarrow 0$ and $\frac{1}{2}(\sigma_1 + \sigma_3)_{eff} \rightarrow 0$, respectively. **Presence of the closure pressures with *in situ* values substantially lower than 30 percent of the effective overburden stress is, perhaps, the most significant feature of the Yucca Mountain stress field**. Magnitude of the discrepancy is surprisingly large; locally such magnitude may exceed 60 percent. Differences in the two *in situ* stress regimes indicate that the character of tectonic straining involved in the two respective areas is different. With reference to discussions presented in Section 3.3.2.2, it is tempting to speculate that the two stress regimes represent two separate strain domains. In each domain, the degree of advancement of the contemporary cycle of tectonic deformation is different. At Yucca Mountain, the contemporary cycle is mature and, consequently, the amount of accumulated tectonic strain is significant. Around Yucca Flat, however, the tectonic cycle appears to be at early stages of development. Relative to Yucca Mountain, the amount of accumulated extensional strain is much smaller.

Second, in the Death Valley groundwater system, the *in situ* stress field is largely tectonic in origin. This conclusion is not surprising; it is consistent with the active tectonic character of this region, and with the estimates of rates of contemporary tectonic straining (Section 3.3.1). This conclusion is also in accord with conclusions drawn earlier by other workers (Ellis and Magner, 1980; Ellis and Magner, 1982; and Stock et al., 1986). In the Climax Stock and in Rainier Mesa and Aqueduct Mesa, tectonic origin of the *in situ* stress field is suggested by a) strongly deviatoric character of the measured stresses; b) tendency for the major principal stress to be inclined at a relatively small angle; and c) presence of significant shear stresses acting in vertical and horizontal planes. At Yucca Mountain, tectonic origin of the *in situ* stress field is indicated by two other lines of evidence: a) the presence of lateral and vertical *in situ* stress gradients, that are not conformable with local topography and b) the presence of abnormally low *in situ* values of the mean effective stress $\frac{1}{2}(\sigma_1 + \sigma_3)$, and the stress difference $(\sigma_1 - \sigma_3)_{eff}$. The Yucca Mountain *in situ* stress heterogeneity appears to be produced locally, most likely in response to creep along a nearby fault. The resulting abnormally low *in situ* values of the closure pressure indicate that a) an evolved deforming fractured medium is present; and b) the *in situ* stress field is transient and subject to substantial and sudden changes. Under such conditions, it is reasonable to anticipate, that in the not too distant future, both the mean stress $\frac{1}{2}(\sigma_1 + \sigma_3)$ and the stress difference $(\sigma_1 - \sigma_3)$ will acquire values that are compatible, at least, with the effective overburden stress. This means, that **at Yucca Mountain, the expected *in situ* stress changes are large, measured in terms of tens of bars.**

4.2.6 The Hydraulic Conductivity Structure - Overall Conclusions and Summary

The review of hydrologic and *in situ* stress data from the Death Valley groundwater system reveals that this system has a very interesting and unique hydraulic conductivity structure. The following three points emphasize the most important features of this structure.

First, in the Death Valley groundwater system, the hydraulic conductivity structure is characterized by extreme spatial heterogeneity. Such heterogeneity is clearly expressed in the results of a) *in situ* measurements of hydraulic conductivity; b) *in situ* measurements of lateral gradients of hydraulic potentials; and c) *in situ* measurements of vertical gradients of hydraulic potentials. The *in situ* values of hydraulic conductivity range from 2×10^2 to 3×10^{-2} m/d for the Paleozoic carbonates, and from 2.8×10^1 to 10^{-6} m/d for the Tertiary tuff "pile". Secondary porosity, specifically fracture porosity, is a main factor that dominates the hydraulic conductivity structure. The spatial distribution of the *in situ* values of hydraulic conductivity involves two distinct patterns. In the vicinity of Yucca Flat, the *in situ* hydraulic conductivity of the tuff "pile" is uniformly low, in a range of 10^{-5} m/d. The depthward decreasing values of hydraulic potentials, however, indicate that the *in situ* hydraulic conductivity increases with depth. This increase appears to be related to the occurrence of Paleozoic carbonates, that underlie the tuff "pile". Several pumping tests, performed in the carbonate rocks, indicated that the *in situ* values of hydraulic conductivity are high, ranging from 2×10^2 to 3×10^{-2} m/d. In this area, the depthward increasing hydraulic transmissivity may be regarded as related to the litho-stratigraphic framework. It appears that this framework, through some means, controls residual apertures. It is possible that, in the "non-limit equilibrium" state, such apertures are very small for the tuff "pile", but large for the carbonate rocks. The result is that the hydraulic transmissivity of the tuff "pile" is substantially lower than that of the carbonates. At Yucca Mountain, the *in situ* values of hydraulic conductivity differ significantly depending upon depth and location. They range from 1.8×10^1 to about 10^{-6} m/d. There is a well pronounced tendency for the hydraulic transmissivity to diminish depthward, northward, and westward across the area. In contrast to Yucca Flat, transmissivities of the Paleozoic carbonates and, locally, the tuff "pile" are equally high. The local hydraulic conductivity structure appears to be composed of three south-eastward dipping layers. Such layers are discordant with respect to both topography and stratigraphy of the area. **This indicates that the overall texture of the local hydraulic conductivity structure in the Tertiary tuff "pile" is related neither to the litho-stratigraphic framework nor to the spatial distribution of overburden load.**

Second, in the Death Valley groundwater system, **the spatially variable** *in situ* **stress field has a clear hydrologic expression.** Such expression is consistent with the conceptual model of a groundwater system in a deforming fractured medium, as developed in Section 3.3.4. The hydrologic influence of the *in situ* stress field is expressed in a two-fold manner. On one hand, the *in situ* stress field seems to be the factor that controls the areal distribution of vertical gradients of hydraulic potentials. In areas where "underpressure" conditions exist, the *in situ* values of hydraulic conductivity for the tuff "pile", relative to the underlying carbonates, are low. As expected in these areas, depth extent of the "limit equilibrium" stress conditions is small, and such conditions are restricted to the vadose zone. The *in situ* values of closure pressure are high, and comparable or exceeding those expected from the overburden load. In areas where "overpressure" conditions exist, however, the local hydraulic transmissivity of the tuff "pile" is comparable to that of the carbonates. Again as expected, the *in situ* stress conditions in such areas differ from those in the "underpressure" areas. Depth extent of the "limit equilibrium" stress conditions is very substantial, and exceeds depth of the water table. Not surprisingly, the *in situ* values of closure pressure are abnormally low. On the other hand, the *in situ* stress field is also a factor that controls texture of the local hydraulic conductivity structure, and the resulting distribution of lateral and vertical gradients of hydraulic potentials. Such control is evident at Yucca Mountain. It is best expressed through spatial conformance of lateral and vertical gradients of a) *in situ* stress; b) *in situ* hydraulic conductivity; and c) *in situ* hydraulic potentials. The cause-effect relationship involving the *in situ* stress field and the hydraulic conductivity structure is further indicated by three lines of evidence:, a) the three-layer texture of the hydraulic conductivity structure; b) the temporal behavior of hydraulic potentials in various parts of the local groundwater system; and c) the abnormally low *in situ* values of the mean stress $\frac{1}{2}(\sigma_1 + \sigma_3)$ and the stress difference $(\sigma_1 - \sigma_3)$.

Third, **considering potential utilization of the Yucca Mountain area as a site for a high-level nuclear waste repository, there are two particularly disturbing characteristics of this area.** The first characteristic is the abnormally low *in situ* values of the closure pressure as measured in the area. Such values indicate that both the mean stress $\frac{1}{2}(\sigma_1 + \sigma_3)$ and the stress difference $(\sigma_1 - \sigma_3)$ are substantially lower than those produced by the overburden stress. At a depth of 1,000 m, magnitude of the discrepancy for the mean stress is as large as 20 and 40 bars. This indicates that a) the observed *in situ* stress conditions are temporary and b) anticipated increases in the *in situ* mean stress are large, in a range of tens of bars. Both of these deductions are similar to those inferred in Section 3.3.2.3, based on the analyses of surficial geology of the Yucca Mountain area. The second characteristic is the very high

degree of the *in situ* stress enhancement of the local hydraulic conductivity structure. Such high degree of enhancement is best expressed by the ratio of hydraulic conductivity in the upper layer of hydraulic conductivity structure to hydraulic conductivity in the lowermost layer. Locally, the value of such ratio exceeds 10^4, and may be as large as 10^6. Assuming that the *in situ* stress is the controlling factor indeed, then it follows that the ratios of K_f^{σ}/K_f^{in} and S^{σ}/S^{in} exceed 10^4 and 10^2, respectively (Plate 3.3.5-1). In such a situation, it may be expected that the volume of fluid released from storage, in association with termination of the "limit equilibrium" stress conditions, is very large. This, together with the expected magnitudes of likely increases of the mean stress, indicates that potential rises of the water table are also large. It is possible that the resulting build-ups of fluid pressure will cause breakdown of the host rock, leading to the development of "explosive" breccias. This deduction is also in accord with the results of analyses of surficial geology of the Yucca Mountain area (Section 3.3.2.3).

4.3 GEOTHERMAL SETTING OF THE YUCCA MOUNTAIN GROUNDWATER SYSTEM

4.3.1 Introduction

Studies of a geothermal field may provide important insights into tectonic and hydrologic processes operating in a given region. In Section 3.0, it has been demonstrated that such processes may influence two factors that control a geothermal field: a) the effective thermal conductivity structure and b) the thermal boundary conditions. As was explained in Section 3.5.2, the effective thermal conductivity structure, in addition to being related to the litho-stratigraphic framework, may also be related to the *in situ* stress field. It is reasonable to expect, that in a deforming fractured medium, the hydraulic conductivity structure and the effective thermal conductivity are closely related. Both of these structures may reflect spatial and temporal characteristics of the *in situ* stress field, and may be undergoing space- and time-dependent enhancements in response to changes in this field.

Thermal boundary conditions describe spatial and temporal aspects of the input of thermal energy into a groundwater system. At the base of a system, say at a depth of about 15 km, the thermal boundary conditions reflect a distribution of *in situ* temperature in the lower crust and upper mantle. Such distrubution, in turn, is related to fundamental tectonic processes that operate deep within the Earth's interior, and may be reflected in geophysical characteristics of the system (Sections 3.1 and 3.4.1). Shallower within the crust, "deep" thermal boundary conditions become increasingly more distorted by groundwater circulations and, consequently, thermal and hydrologic boundary conditions are closely related. For "forced" convection systems, the "upward-flux" hydrologic boundary conditions are the cause of corresponding thermal boundary conditions (Section 3.4.5). For "mixed" convection systems, however, the cause-effect relationship is reversed.

Interdependence of both the boundary conditions and the conductivity structures suggests that there should be close correspondence between potentials of geothermal and hydraulic fields. In a deforming fractured medium, such correspondence involves also the *in situ* stress field, and is expressed in both temporal and spatial sense (Sections 3.5.2 and 3.5.3).

In the Death Valley groundwater system, geothermal conditions are known based, mainly, on downhole measurements of *in situ* temperature. Such measurements were made in a large number of boreholes.

Depth of these boreholes ranged from a few hundred meters to a maximum of about 4 km. The majority of geothermal data, however, was obtained in boreholes less than 2 km deep.

The results of geothermal measurements, if considered together with the data and interpretations presented in Section 4.2, offer an opportunity for further testing the validity of hydrologic concepts as developed in Section 3.0. Furthermore, the geothermal data, in and by themselves, are important for considerations of long-term stability of the Yucca Mountain groundwater system, and for considerations of hydrology of the vadose zone (Sections 3.4 and 3.5).

With reference to the Yucca Mountain site, three questions are of particular relevance: a) Is thermal convection of fluids in faults feasible at this site?; b) Is the local geothermal field developed in a deforming fractured medium?; c) Is the temporal behavior of the local geothermal field compatible with the conceptual model of a heat-fluid coupled flow in a deforming fractured medium?

As discussed in Section 3.4.3, requirements for the onset of thermal convection are not extraordinarily stringent. This is so, in particular, for convection cells with aspect ratios ranging from 0.5 to 2, and for a fractured medium with permeabilities in excess of tens of millidarcies. In regions where the average value of vertical geothermal gradient exceeds 40°C/km, thermal convection of fluids in wide fault zones is to be expected. Such a phenomenon should be more or less clearly expressed in characteristics of the local geothermal field. Based on geothermal characteristics alone, however, it is easy to confuse a thermally driven convection system with a "forced" convection system. To distinguish two such systems apart, it is necessary to involve other characteristics of the flow system, mainly habits of long-term behavior (Section 3.4.6).

In Section 3.5.2, it has been demonstrated that involvement of a deforming fractured medium will also be expressed through characteristics of a local geothermal field. Such characteristics are a) high degree of thermal heterogeneity; b) spatial conformance of lateral and vertical gradients of *in situ* stress, hydraulic potential, and *in situ* temperature; and c) time-dependence of geothermal gradients.

A coupled heat-fluid flow system in a deforming fractured medium combines characteristics of the above two systems. For such a system, the values of vertical geothermal gradients and the intensities of flux of terrestrial heat are sufficiently high to permit and sustain thermal circulations of groundwater. In

addition, the hydraulic and thermal conductivity structures of such system undergo cyclic changes (Section 3.5.3). During advanced stages of a tectonic cycle, "true" geothermal characteristics of the system are expressed with corresponding intensities, but only below the surface $Z_{(x,y,t)}$. Above this surface, the geothermal field is attenuated or dampened by the "limit equilibrium" stress conditions. The flow system is characterized by transient behavior and the resulting variability of its appearance with time (Plate 3.5.3-3).

A main purpose of the subsequent analyses is an assessment of two topics: a) involvement of terrestrial heat in influencing and/or controlling local boundary conditions and b) involvement of the local *in situ* stress field in controlling the effective thermal conductivity structure. The analyses are presented in four parts. The first part, Section 4.3.2, describes the results of downhole measurements of *in situ* temperatures from the Nevada Test Site. The second part, 4.3.3, is concerned with the Yucca Mountain geothermal field. In this part, an attempt is made to spatially associate the local geothermal and hydraulic fields and, through that, to associate both of these fields with the Yucca Mountain *in situ* stress field. The third part, Section 4.3.4, deals with possible time-dependence of the Yucca Mountain geothermal field. Finally, the fourth part, Section 4.3.5, presents a summary and conclusions.

4.3.2 Geothermal Conditions at the Nevada Test Site

At the Nevada Test Site, downhole measurements of *in situ* temperature were made in nearly 100 boreholes. The locations of some of these boreholes are shown on Plate 4.3.2-1. The data base established was summarized and reported by Sass et al., 1980; Sass and Lachenbruch, 1982; Sass et al., 1983; and Sass et al., 1987. Plates 4.3.2-2 through 4.3.2-7 present some of the geothermal data. The presentation is in the form of "worm" diagrams showing depthward distribution of *in situ* temperature for each test borehole. Six different areas from the Nevada Test Site are represented. A summary of the geothermal data is presented on Plates 4.3.2-8 and 4.3.2-9.

Examination of the geothermal data reveals, that at the Nevada Test Site, the geothermal field is characterized by a substantial degree of spatial heterogeneity. Sass et al., 1980; Sass and Lachenbruch, 1982; and Sass et al., 1983, attributed such heterogeneity to local "groundwater circulations". Additionally, these authors recognized two areas, and possibly three areas, where thermally driven groundwater circulations are evident. The first area is located in the southwestern sector of the Nevada Test Site, in the vicinity of Well UE-25a#3 (Plate 4.3.2-1). The second area is Yucca Mountain; the local hydrothermal system is developed in the Paleozoic carbonates, and was encountered in Well UE-25p#1 (Sass et al., 1983). The third area is located in Pahute Mesa; the local hydrothermal system is centered around Well PM-2 (Plate 4.3.2-9).

At the Nevada Test Site, at a depth of 500 m, the *in situ* temperatures range from about 22°C, in the vicinity of Yucca Flat and Rainier Mesa, to as much as 47 to 50°C, in Pahute Mesa and around Skull Mountain (Plate 4.3.2-9). Near the ground surface, the values of vertical geothermal gradients range from 10 to 72°C/km (Plate 4.3.2-8).

A pattern outlined by the planar distribution of *in situ* temperature, if considered together with the seismic velocity structure from Plate 3.4.1-3 and hydrologic data from Section 4.2.4-2, is quite revealing. Two observations are of particular importance. First, the lowest *in situ* temperatures occur in the area of Rainier Mesa and Yucca Flat. As discussed in Section 4.2.4, in this area hydraulic "underpressures" are present. This coincidence suggests that the thermal deficit, indicated by the anomalously low values of local geothermal gradients, may be accounted for by assuming that downward percolations of groundwater are responsible. The groundwater seems to be carrying off some of terrestrial heat from the tuff "pile" into

the underlying Paleozoic carbonates. It is also possible that lateral "through" flow of groundwater, in the transmissive carbonates, prevents efficient upward transfer of terrestrial heat. Such an interpretation is consistent with conclusions drawn earlier by Sass and Lachenbruch, 1982. Based on regional studies of the distribution of intensity of flux of terrestrial heat at the Nevada Test Site, these authors concluded that a) "groundwater is carrying off much of the Earth's heat in the upper 3 km and delivering it elsewhere" and b) "the regional heat flow from beneath the zone of hydrologic disturbance may be the same as that characteristic of the Great Basin in general (80 mWm^{-2} or 2 HFU) or it could be as high as 100 mWm^{-2} or 2.5 HFU."

Second, the two local seismic velocity anomalies from Plate 3.4.1-3, one in the area of Pahute Mesa and another in the southwestern corner of the Nevada Test Site, seem to be reflected in the planar distribution of *in situ* temperature. As shown on Plate 4.3.2-9, the highest *in situ* temperatures occur near and over both of the anomalies. It is possible, that at both locations, the intensities of flux of terrestrial heat are enhanced by the presence of local heat sources, most likely magma bodies. The locally measured high values of vertical geothermal gradients may not be representative for the entire region. Such values, however, seem to be sufficiently high to locally support thermal circulations of groundwater (Section 3.4.3). Not surprisingly, systems of hydrothermal circulation seem to be present in the vicinity of both anomalies. Particularly well expressed is the Pahute Mesa system. As discussed in Section 4.2.4.3, the presence of this system is suggested by spatial conformance of the lateral gradients of a) *in situ* temperature; b) groundwater chemistry; and c) hydraulic potentials. In the southwestern sector of the Nevada Test Site, presence of the hydrothermal circulation system was interpreted by Sass et al., 1980. The interpretation is based on analyses of a thermal profile from Well UE-25a#3. The intensity of flux of terrestrial heat, around this well, was estimated to be as great as 3 HFU ($\approx$ 120 mWm^{-2}).

The above interpretations of geothermal and seismic velocity data are not unequivocal; alternative interpretations are possible. One such interpretation is that the seismic velocity anomalies are not indicative of the local heat sources. Instead, the anomalies reflect accumulations of low-seismic velocity materials in volcanic calderas. If one accepts such alternative interpretation then it follows, that the local values of geothermal gradients measured around both anomalies may indeed be representative for the Nevada Test Site. It is also possible, however, that these values reflect local upwellings of groundwater. Such upwellings, instead of being the result of thermal convection, are caused by local hydraulic conductivity contrasts that facilitate local occurrences of "forced" upward circulations of groundwater.

In the area of Yucca Mountain, at a depth of 500 m, the *in situ* temperatures range from 29 to 37°C (Plate 4.3.2-9). These temperatures yield the average value of vertical geothermal gradients ranging from 24 to 40°C/km (Plate 4.3.2-8). Across Yucca Mountain, there appears to be a westward increase in the *in situ* temperature. In the vicinity of Quaternary volcanic centers in Crater Flat, the *in situ* temperatures are 5 to 8°C higher than the corresponding temperatures in the western and eastern sectors of Yucca Mountain (Plate 4.3.2-9).

At Yucca Mountain, a network of more than 30 wells allows for a relatively fine definition of the local geothermal field. Main characteristics of this field are presented in the subsequent section.

4.3.3 The Yucca Mountain Geothermal Field

At Yucca Mountain, a set of precise downhole temperature profiles was obtained from more than 30 boreholes covering an area of approximately 100 km^2. These boreholes were drilled to a depth ranging from 300 to about 1800 m. Repeated temperature profiles were obtained from 18 deep boreholes. The measurements were made employing standard USGS logging procedures, and using the Fenwall K 212 E thermal probe (Sass et al., 1973). Single temperature profiles were obtained in air columns of 17 wells of the WT-series. These wells were drilled to monitor position of the water table. The air column logs were made using a specially designed thermistor probe having a very small thermal mass. The probe equilibrates with a column of still air to within 1 percent of a step temperature change in a period of about four minutes. A total of 204 laboratory measurements of thermal conductivity were made using core samples extracted from Yucca Mountain. The measurements were made at an ambient temperature of about 25°C, and employing the needle-probe technique (Sass et al., 1971). All of the Yucca Mountain geothermal data were developed and reported by Sass et al., 1983 and Sass et al., 1987.

The Yucca Mountain geothermal data are also presented on Plates 4.3.3-1 through 4.3.3-43. The location of all measurement boreholes are shown on Plates 4.3.3-1 and 4.3.3-2. Plates 4.3.3-3 through 4.3.3-21 present the repeated temperature profiles from deep boreholes. Such profiles are shown as time-series. The leftmost profile is the earliest. Later profiles are identified by month and year and are stepped to the right by a specified amount relative to the first curve. Plates 4.3.3-22 through 4.3.3-38 present the temperature profiles made in air columns in test Well UZ-1 and the WT-series. All measurement points are identified by discrete symbols joined by a straight line. The measurement procedure employed consisted of lowering the thermistor probe in 30 and 60 m increments, then reading thermistor resistance at 1 minute intervals until the resistance change in 1-minute was less than 1 percent of the accumulated change, or for 20 minutes, whichever was less. Data reduction involved converting resistances to temperatures, and then interpreting a most likely equilibrium temperature. Sass et al., 1987 considered, that even though detailed features on individual thermal profiles may be questionable, the overall temperature profiles constitute a reasonably accurate representation of the actual conditions. Plates 4.3.3-39 through 4.3.3-43 present the results of laboratory measurements of thermal conductivity. Such measurements were made using core samples from five deep wells: USW G-1, USW G-2, USW G-3, USW G-4, and UE-25p#1. Upon collection, each sample was wrapped in aluminum foil and dipped in hot wax to preserve *in situ* moisture conditions. The measurements were made at an ambient temperature of 25°C, using the needle-probe technique. The

data base consists of 57 thermal conductivities representing tuffs in the vadose zone, 134 conductivities representing tuffs below the water table, and 13 conductivites representing the Paleozoic carbonates.

A summary of the Yucca Mountain geothermal data is presented on Plates 4.3.3-44 through 4.3.3-52. Plates 4.3.3-46 through 4.3.3-48 present three geothermal cross-sections for the area. The location of these cross-sections are shown on Plates 4.3.3-44 and 4.3.3-45. In constructing the cross-sections, measurement wells were projected onto topographic cross-sections, on which the water table is also plotted. The depth to a given *in situ* temperature is indicated at 5°C intervals and temperatures are jointed to show inferred isothermal lines. Plates 4.3.3-49 and 4.3.3-50 present a summary of geothermal data for the Yucca Mountain vadose zone. Plate 4.3.3-49 displays all the air column temperature profiles plotted with a common origin. Plate 4.3.3-50 shows the planar distribution of *in situ* temperatures at a depth of 350 m. The temperatures shown were not corrected to account for local, topography related, differences in the mean annual surface temperature. Plate 4.3.3-50 also shows the location of two local faults: the Paintbrush fault and the Solitario Canyon fault. Plates 4.3.3-51 and 4.3.3-52 present the results of laboratory measurements of thermal conductivity. Plate 4.3.3-51 presents two histograms of thermal conductivities for the tuff "pile". The data representing the vadose zone are shown separately from those representing tuffs below the water table. Plate 4.3.3-53 presents an estimate of the intensities of flux of terrestrial heat for the Yucca Mountain area. The estimate is based on temperature profiles from deepest segments of six boreholes: USW G-1, USW G-2, USW H-5, USW H-6, USW VH-1, and USW VH-2. All of these boreholes are situated in the northern and western sectors of the area, where the uppermost layer of the local hydraulic conductivity structure is either relatively thin or absent (Plates 4.2.2-13 and 4.2.2-14). In both of these sectors, the surface $Z_{(x,y,t)}$ is situated at a relatively shallow depth (Plates 4.2.5.4-98 and 4.2.5.4-99). Consequently, it is reasonable to assume, that in deeper portions of both sectors, the *in situ* stress enhancement of the local thermal conductivity structure is either very minimal or non-existent, It is likely, therefore, that in these portions the laboratory values of thermal conductivity are representative of the bulk effective thermal conductivity.

The Yucca Mountain geothermal data, if considered together with the hydrologic and *in situ* stress data, allow for recognition of four main characteristics of the local geothermal field. First, **the geothermal field is characterized by a surprisingly large degree of spatial heterogeneity**. Important lateral and vertical geothermal gradients are clearly evident. Deep below the water table, the highest *in situ* temperatures occur in northern and western sectors of Yucca Mountain. Toward the southeast, *in situ*

temperatures diminish progressively. In the vadose zone, however, the highest *in situ* temperatures occur in the southern sector (Plate 4.3.3-50). Toward the north, *in situ* temperatures diminish progressively. Superimposed on this overall pattern, two thermal highs are present. One of such highs occurs in the vicinity of Well UE-25p#1, around the fault contact between the Tertiary tuff "pile" and the Paleozoic carbonates (Plates 4.3.3-46a and 4.3.3-47a). As discussed in Section 4.2.4-4, depthward distribution of hydraulic potentials indicate, that along this fault contact, upwelling of fluids is taking place. Another thermal high occurs in the vicinity of Well USW WT-10 (Plate 4.3.3-48a). The location of this high coincides with the Solitario Canyon fault. In the vadose zone, at a depth of 350 m, the *in situ* temperatures range from 23 to about 43°C (Plate 4.3.3-49). The local values of vertical geothermal gradients range from 20 to as much as 60°C/km. The highest *in situ* temperatures occur along and around traces of the Paintbrush fault and the Solitario Canyon fault (Plate 4.3.3-50). In most of the deep wells, the values of vertical geothermal gradients in deep borehole segments are substantially higher than those in shallow segments. The increase is quite sharp and may readily be identified on the depth vs. temperature plots. Such an increase is apparent in the following wells: UE-25b (Plate 4.3.3-5, depth 900 m), USW H-1 (Plate 4.3.3-7, depth 625 m), USW VH-2 (Plate 4.3.3-8, depth 1,050 m), USW G-1 (Plate 4.3.3-9, depth 1,000 m), USW G-2 (Plate 4.3.3-10, depth 800 m), USW H-1 (Plate 4.3.3-13, depth 1,000 m), USW H-4 (Plate 4.3.3-15, depth 950 m), USW H-5 (Plate 4.3.3-16, depth 1,050 m), and USW H-6 (Plate 4.3.3-16, depth 1,050 m).

Second, **the laboratory values of thermal conductivity indicate, that at Yucca Mountain, the tuff "pile" is reasonably homogenous.** In the vadose zone, the laboratory values of thermal conductivity are slightly higher than those below the water table (Plates 4.3.3-51 and 4.3.3-52). The mean values of thermal conductivity, however, are not significantly different. For the vadose zone, distribution of thermal conductivities is bi-model, with peaks at about 1 and 2 $Wm^{-1}K^{-1}$. Tuff samples from below the water table, however, yielded nearly normal distribution. The mean value is equal to about 1.72 $Wm^{-1}K^{-1}$; deviations from the mean are relatively small. Samples of the Paleozoic carbonates yielded considerably higher values of thermal conductivity. Such values range from 4.67 to 5.36 $Wm^{-1}K^{-1}$, averaging at about 4.95 $Wm^{-1}K^{-1}$ (Plates 4.3.3-43 and 4.3.3-52).

Third, **spatial distribution of *in situ* temperature indicates, that at Yucca Mountain, the geothermal and hydraulic fields are clearly interrelated.** Such a relationship is indicated by two lines of reasoning. The first line involves origin of the two thermal highs, one around the Solitario Canyon fault and other around the Paintbrush fault. It is almost certain that both of these highs are caused by lo-

cal upwellings of groundwater. The presence of such upwellings is indicated by depthward distribution of hydraulic potentials (Section 4.2.4.4). Their presence is also indicated by isotopic and geochemical characteristics of groundwater samples that were collected in the vicinity of the Solitario Canyon and Paintbrush faults. As shown on Plate 4.3.3-54, water samples from the tuff "pile" in Wells USW H-3 and UE-25p#1 have the $\partial^{13}C$ ratios higher than about -7 o/oo. Such ratios represent an increased amount of ^{13}C relative to both atmospheric carbon dioxide ($\partial^{13}C$ ratios ranging from -7 to -9 o/oo) and carbon dioxide derived from plants in soil horizon ($\partial^{13}C$ ratios ranging from -12 to -25 o/oo) (Kerrisk, 1987). Furthermore, such ratios are consistently higher than typical $\partial^{13}C$ ratios for groundwaters residing in the tuff "pile" ($\partial^{13}C$ ratios ranging from -7 to -13 o/oo) (Plate 4.5.2.5-9). The ratios, however, are comparable with the $\partial^{13}C$ ratios for the Paleozoic carbonates ($\partial^{13}C \cong -2$ o/oo), and for groundwater residing in these rocks ($\partial^{13}C$ ranging from -2.3 to -5.0 o/oo) (Plate 4.5.2.5-10). As shown on Plate 4.3.3-54, water samples collected from the tuff "pile" around the two faults (Wells USW H-3, USW H-6, and UE-25p#1) are supersaturated with respect to calcite. The saturation indexes, calculated for these samples using the EQ-3 chemical equilibrium computer program, are positive. This is in sharp contrast to all remaining groundwater samples from Yucca Mountain; such samples usually yield negative values of the calcite saturation indexes (Plate 4.3.3-54). The high *in situ* values of hydraulic conductivity, measured around both of the thermal highs, indicate that rather modest values of vertical geothermal gradients are required for the onset of thermal convection. In this situation, it is reasonable to assume that both upwellings of groundwater are the result of thermal convection. Such a deduction is similar to the conclusion drawn earlier by Sass et al., 1983. Based on the analyses of Yucca Mountain geothermal data, these authors stated that

> **"...heat flow is likely to vary between about 30 and 70 mWm^{-2} (.75 to 1.7 HFU) within the very small area. This in turn suggests very shallow (in the range of 2.5 to 5 km) heat sources and sinks as the cause of the variation. The most likely sources and sinks would be hydrologic.... From the present series of measurements, it seems clear that various fluids are moving about in the unsaturated zone, that water is moving in a very complicated manner within the saturated zone to depths on the order of 1 km, and that in the Paleozoic rocks beneath the tuffs, there is also a complex hydrothermal circulation system."**

Although not likely, it is possible, nevertheless, that both groundwater upwellings are caused by hypothetical hydraulic transmissivity contrasts that may exist somewhere in the southern sector of Yucca Mountain. The resulting "forced" convection would be practically indistinguishable in appearance from thermal convection (Section 3.4.6). The second line of reasoning, that indicates interdependence of the geothermal and hydraulic fields, involves apparent similarities in the texture of the Yucca Mountain hydraulic conductivity structure and the spatial distribution of *in situ* temperature. Apart from the two thermal highs, deeper isothermal lines seem to be conformable with layers comprising the local hyraulic conductivity structure (Plates 4.2.2-13, 4.2.2-14, 4.3.3-46a, and 4.2.2-47a). In Wells UE-25b, USW H-1, USW H-5, and USW H-6, the sharp increase in the values of vertical geothermal gradients occurs at or near the bottom of uppermost layer of the local hydraulic conductivity structure. The thermal deficit, apparent within a few hundred meters below the water table, may be attributed to the depth-dependent enhancement of the local thermal conductivity structure. As was explained in Section 3.5.2, such enhancement, along with the enhancement of the hydraulic conductivity structure, is caused by the local *in situ* stress field that controls the distribution of conducting apertures. The very low values of closure pressure (σ_c), encountered below the water table, indicate that the Yucca Mountain tuff "pile" is severly dilated (Section 4.2.5.4). The resulting retention of terrestrial heat, similarily as retention of subsurface fluids, in the upper portions of tuff "pile" is not as efficient as that below the surface $Z_{(x,y,t)}$. At Yucca Mountain, the hydraulic and the effective thermal conductivity structures appear to be closely related. Being controlled by the common factor, both of these structures exhibit similar spatial characteristics. It is likely that long-term behavior of both structures is likewise similar.

Fourth, beneath the zone of hydrologic and thermal disturbance, the Yucca Mountain geothermal field appears to be the same as that characteristic of other parts of the Great Basin. In places where the *in situ* stress influences are minimal, the values of vertical geothermal gradients range from 40 to as much as 60°C/km (Plate 4.3.3-53). Such values are sufficiently high to allow for the onset of thermal convection of fluids, even in a fractured medium with modest hydraulic transmissivity (Sections 3.4.3 and 3.4.4). The relatively **high values of deep vertical geothermal gradients indicate, that at Yucca Mountain, the intensity of flux of terrestrial heat is considerable.** It is possible that such intensity ranges from 70 to as much as 100 mWm^{-2} (1.75 to 2.5 HFU) (Plate 4.3.3-53). There appears to be a tendency for the intensity of heat flux to increase southwestward, in a direction of Crater Flat where several volcanic centers of Quaternary age are located. Furthermore, the increase seems to be occurring across the interface

between the higher and lower than average seismic velocities in the lower crust and upper mantle (Plates 3.4.1-4 and 3.4.1-5).

4.3.4 Possible Time-Dependence of the Yucca Mountain Geothermal Field

In Section 3.5.2, it was shown that a geothermal field developed in a cyclically deforming fractured medium varies not only in space but also in time. Such variability is related to the *in situ* stress that controls the effective thermal conductivity structure. Throughout an individual cycle of tectonic deformation, the *in situ* temperatures fluctuate in a manner shown on Plate 3.5.2-5. This characteristic offers another opportunity for testing the validity of the conceptual models developed in Section 3.5. To proceed with such testing, however, an estimate of values of vertical paleo-geothermal gradients is required. Such an estimate may be made based on the results of isotopic and radiometric studies of vein mineralization from the Yucca Mountain vadose zone.

Samples of calcite veins, emplaced along various fractures in the vadose zone, were collected from cores extracted from three deep wells: UE-25p#1, USW G-2, and USW G3. The locations of these boreholes are shown on Plate 4.3.4-1. The samples were analyzed for uranium and stable isotope contents and were dated using the uranium-series method. The results of such analyses were reported by Szabo and Kyser, 1965. They are also shown on Plates 4.3.4-2 and 4.3.4-3. Two calcite samples yielded an average radiometric date of 2.8×10^4 years B.P., four sampes yielded an average date of about 1.7×10^5 years B.P., and four samples have an average age of about 2.8×10^5 years B.P. In addition, four calcite samples yielded minimum dates of precipitation greater than 4×10^5 years B.P.

Plate 4.3.4-4 presents the stable isotope data for the Yucca Mountain calcite samples. The presentation is in the form of a $\partial^{18}O$ vs. $\partial^{13}C$ plot. Temperatures shown represent the calculated isotopic composition of calcite in equilibrium with meteoric water having the $\partial^{18}O$ ratio of -9 and the $\partial^{13}C$ ratios of either -11.5 or -8 (Szabo and Kyser, 1985). Fractionation factors of oxygen are those suggested by O'Neil et al., 1969, and for carbon are those derived by Friedman, 1970.

Plate 4.3.4-5 presents an estimate of paleo-temperatures based on the stable isotope content of calcite samples. This estimate is also shown on Plate 4.3.4-6 which, in addition, shows the contemporary *in situ* temperatures at Yucca Mountain. Differences between the contemporary and paleo-temperatures are substantial, ranging from few °C, near the ground surface, to as much as 15°C, at a depth of about 600 m.

Depthward decrease in ^{18}O and increase in ^{13}C contents, shown on Plate 4.3.4-4, were interpreted as being caused by the depthward increase of *in situ* temperature (Szabo and Kyser, 1985). These authors estimated that the increase corresponds to an average value of vertical geothermal gradient of about 43°C/km. Such a value is about 15°C/km higher than the average value of contemporary geothermal gradient as measured in the vadose zone in boreholes UE-25p#1, USW G-2, and USW G-3. The estimated value of paleo-geothermal gradient, however, is almost identical to the values measured beneath the zone of hydrologic distubrance in Wells USW G-2, USW H-5, and USW H-6 (Plate 4.3.3-53).

The above comparison suggests, that at Yucca Mountain, the potentials of geothermal field may be time-dependent. This characteristic, if combined with other characteristics of the Yucca Mountain geothermal field, indicates that **the conceptual model developed in Section 3.5.2 constitutes a viable representation of the local geothermal conditions**. Short of questioning validity of various measurements and interpretations performed by other workers, it is rather difficult to offer an alternate explanation that equally well accounts for the entire data base.

4.3.5 Summary and Conclusions

The review of geothermal data from the Death Valley groundwater system allows for the following three main conclusions to be drawn. First, the overall geothermal field is characterized by a pronounced spatial heterogeneity. Such a heterogeneity appears to be the result of interactive relationship between the geothermal field and the groundwater system. The interaction is expressed in a two-fold manner. In areas where "underpressure" conditions are present, the groundwater system dominates and controls shallow parts of the geothermal field. Downward percolations of groundwater, together with lateral flow in the Paleozoic carbonates, seem to be responsible for the near-surface thermal deficit. The groundwater system effectively masks the geothermal field. Downhole temperature profiles appear as primarily conductive and yield anomalously low values of vertical geothermal gradients. Such values range from 12 to about 28°C/km. In Pahute Mesa and in the southwestern sector of the Nevada Test Site, a reasonably well supported case may be made that the local geothermal field plays a more active role. In both of these areas, hydrothermal convection systems appear to be present. Such systems are developed above and near the local seismic velocity anomalies identified in the middle crust. It is possible that both of these anomalies reflect the presence of magma bodies. Such bodies may be responsible for the locally increased intensities of flux of terrestrial heat, and the resulting upward migrations of fluids. The downhole temperature logs indicate that local values of vertical geothermal gradients may be as high as 40 to 70°C/km.

Second, a well expressed and defined thermal heterogeneity is present at Yucca Mountain. The spatial aspects of this heterogeneity, if considered together with the hydrologic and *in situ* stress data, indicate that the anomaly is caused by two factors. On one hand, upwellings of groundwater along the Paintbrush fault and the Solitario Canyon fault are responsible for the increased temperatures along and around traces of both faults. The increased temperatures are evident in both the vadose zone and below the water table. On the other hand, the *in situ* stress enhancement of the local thermal conductivity structure is responsible for the overall thermal deficit in upper parts of the geothermal field. Between the Solitario Canyon fault and the Paintbrush fault, the thermal deficit occurs within the uppermost layer of the local hydraulic conductivity structure, where anomalously low values of the closure pressure were measured. Both, the geothermal and hydrologic charcteristics of the area are remarkably consistent with the conceptual models of heat and fluid flows in a deforming fractured medium (Sections 3.3.4 and 3.5.2). The results of radiometric and isotopic analyses of vein filings from the vadose zone suggest that, within the Yucca Mountain geothermal field, potentials vary in time. This, together with upwellings of groundwater

along the two local faults, indicates, that at Yucca Mountain, flows of terrestrial heat and groundwater are coupled. The entire system displays a strong resemblance to a system represented by the conceptual model of a coupled heat-fluid flow in a deforming fractured medium (Section 3.5.3).

Third, within the Nevada Test Site, and probably within the entire Eureka Low, the intensity of flux of terrestrial heat appears to be the same as that characteristic of the Great Basin in general. The intensity of this flux ranges from 70 to about 100 mWm^{-2} (1.75 to 2.5 HFU). Beneath the zone of near-surface hydrologic disturbance, the values of vertical geothermal gradients range from 40 to 60°C/km. A comparison of these values with requirements for the onset of thermal convection indicates, that considerations of possible involvement of terrestrial heat in influencing the Yucca Mountain groundwater system, are fully justified. A mystery posed by the Eureka Low may be solved by recognizing, that within this low, it is rather difficult to obtain representative measurements of the intensity of flux of terrestrial heat. There are two factors that complicate such undertakings. In areas where "underpressure" conditions exist, downhole temperature profiles from shallow boreholes, are not representative of the actual geothermal conditions. In such areas, the true geothermal field is effectively masked, or substantially altered, by vertical and lateral movements of groundwater. In areas where "overpressure" conditions exist, laboratory values of thermal conductivity underestimate the values of effective thermal conductivity. The laboratory values do not account for the *in situ* stress enhancements of the local thermal conductivity structure.

4.4 HYDROLOGIC EFFECTS OF UNDERGROUND NUCLEAR DETONATIONS

4.4.1 Introduction

Hydrologic observations made in association with stressing of a fractured medium may reveal important characteristics of a groundwater system. Specifically, the results of such observations may be used to evaluate whether or not a deforming fractured medium is involved (Section 3.3.6). Furthermore, these results may also be useful for evaluations of potential magnitudes of the *in situ* stress related changes in hydraulic conductivity and storativity. This information, in turn, may be used to estimate magnitudes of potential hydrologic disturbances resulting from large-scale restructurings of the *in situ* stress field. Such estimates, of course, must be made with reference to a specific groundwater system and accounting for local hydrologic and tectonic circumstances.

In a deforming fractured medium, the *in situ* stress and hydraulic fields are coupled via the principle of effective stress (Section 3.3.3). Even a small change in one of the two fields causes the development of local instabilities, and leads to transient perturbations in both of these fields. In hydrologic terms, the perturbations are expressesd as oscillations of hydraulic potentials, that take place long after the provocation. After some time, the coupled system finds a new equilibrium configuration. Such a new configuration may require the presence of sustained, but local, changes in position of the water table.

Prior to performing a review of hydrologic effects of underground nuclear detonations it may be useful to develop some understanding of phenomena involved in such detonations. When a nuclear detonation occurs, large amounts of energy are released in a very short time. The material in the nuclear device is instantly vaporized and raised to a temperature of several million degrees Kelvin. A shock wave emerges from this hot, high-pressure region vaporizing some of the surrounding host rock. At that time, a cavity has been formed having a radius of a few meters and containing rock vapors at a pressure of about 1 megabar. The cavity continues to expand until the pressure within it drops to a value of the confining overburden stress. The cavity then has a radius of tens of meters, depending upon the yield of the detonated device. While the cavity is growing, the shock wave from the explosion is moving away from the cavity. It arrives at the ground surface, in few tens of msec after the explosion, and causes the ground surface to rise momentarily; then the shock wave is reflected back. The passage of shock wave causes failure in much of the host rock between the detonation point and the ground surface.

The above discussion pertains to so-called early-time phenomenology. Late-time phenomenology commences after the cavity has been stabilized and the shock waves have been dissipated. At that time, the rock vapors condense, leaving water vapor as the major component of the cavity gases. The cavity gases cool by Newtonian heat transfer into the cavity walls. The transfer of heat from the cavity into the surrounding bedrock established a high thermal gradient in the host rock. Eventually, after a period of minutes or hours, the cavity can no longer support itself and collapses. The collapse propagates upward through the material that has been fragmented by the shock wave and forms a rubble chimney.

Away from the detonation site, passage of vibratory ground motion causes stressing of a fractured medium. Such transient stressing is similar to that resulting from natural earthquakes. The fractured medium is alternately compressed and dilated through many cycles of strain. During a single cycle, part of the medium, perhaps one-quarter of the shock wave wavelength wide, is under either compressional or extensional strain, for a time equal to about one-quarter of the shock wave period.

Values of dynamic strain caused by an earthquake with magnitudes $M_b = 7.0$ and $M_b = 7.5$ were given by Wood, 1985. The dynamic strain produced by such an earthquake ranges from 10^{-3}, near the earthquake epicenter, to about 10^{-5}, some 200 to 250 km away from the epicenter (Plate 4.4.1-1). Wood, 1985, reported that the Handley underground nuclear detonation produced a peak dynamic strain of about 1.5×10^{-5} at a distance of about 200 km. The detonated device had a reported yield of about 1 megaton, which is equivalent to an earthquake with $M_b = 6.5$. The peak dynamic strain resulting from a 1 megaton detonation is on the order of 10^{-4}, at a distance of 20 to 30 km.

An underground nuclear detonation also causes a long-term straining of the host bedrock. An attempt to quantify such a straining was made, at the Nevada Test Site, in association with two underground detonations: the Benham and Handley events. Both of these events occurred in Pahute Mesa and had a reported yield of about 1 megaton.

Weart, 1969, reported the results of strainmeter measurements made in association with the Benham event. The measurements were made using 3 strainmeters installed at distances of 28, 30, and 71 km away from the detonation site. The instruments were capable of detecting long-period or permanent strain accumulations, in addition to measuring the dynamic strains resulting from vibratory ground motion. Following the Benham detonation, strainmeters at all three locations detected a step-like compressional strain

that decayed to the pre-detonation level within a period of about 1 hour. The amplitude of induced strain step was 3×10^{-8}, at the installation 30 km south of the detonation site. The remaining two installations, 28 and 71 km west from the detonation site, showed the amplitudes of strain step equal to 1.8×10^{-7} and 0.35×10^{-7}, respectively.

Extensive geodetic surveys of post-detonation deformation of the ground surface were undertaken in association with the Handley event. The results of this survey were reported by Bucknam and Dickey, 1971. Three kinds of measurement techniques were employed: a) high-accuracy leveling; b) horizontal distance measurements, within triangulation networks; and c) horizontal distance and angle measurements, within quadrilateral networks.

Location of the leveling lines, relative to the Handley detonation site, is shown on Plates 4.4.1-2 and 4.4.1-3. Plate 4.4.1-4 presents vertical displacements caused by the Handley event, as measured along three level lines. Around ground zero and in the northeastern sector, the detonation produced a large subsidence. A comparison of pre- and post-detonation topographic maps revealed that the magnitude of this subsidence ranged from 14 m at ground zero, to about 1.5 m at the 460 m perimeter (Bucknam and Dickey, 1971). Level lines A and C showed that the subsidence was elongated eastward and northward. Displacements recorded along line C diminished from about -1.2 m at the 460 m perimeter, to about -0.04 m at a distance of about 4 km. Along the eastern segment of line A, the subsidence decreased from about 0.14 m at a distance of 2 km, to less than 0.02 m at a distance of 6 km. In the western sector of the area, an uplift was recorded along parts of line A and along line B. The magnitude of this uplift ranged from about 0.40 m, 1.5 km away from the detonation site, to less than 0.02 m at a distance of about 3 km.

Plate 4.4.1-5 presents the results of measurements of horizontal deformation of the ground surface within triangulation network B. Directions and magnitudes of the principal horizontal strains are shown. These were calculated based on the measured distance change between individual reference points of the network. Clearly, the strain field induced by the Handley detonation lacks consistency and radial symmetry. The explosion produced strains were accommodated by rotational movements of individual blocks, in addition to radial contraction and expansion of these blocks. A least-square regression analysis of the magnitude of the radial strain vs. distance from ground zero showed that the radial strain diminished from 2.1×10^{-4}, at a distance of 2 km to 1.9×10^{-5} at a distance of 6 km. Locally, however, the induced strain was as large as $\pm 4 \times 10^{-4}$ at a distance of 4 km away from the detonation site.

Plate 4.4.1-6 presents the results of measurements of horizontal deformation of the ground surface within quadrilateral networks P and G. These networks were located 4 and 12 km away from the detonation site, respectively. The maximum horizontal strain detected within P network was equal to about $+1.3 \times 10^{-4}$ (extension). Within G-network, the maximum horizontal strain was -1.2×10^{-5} (compression).

A review of hydrologic observations made at the Nevada Test Site, in association with five underground nuclear detonations, is presented in the following sections. Sections 4.4.2 and 4.4.3 deal with the Aardvark and Bilby explosions. Both of these explosions were detonated in Yucca Flat. Section 4.4.4 describes hydrologic responses observed in Pahute Mesa in association with the Handley event. Section 4.4.5 describes effects of underground nuclear detonations on aqueous chemistry as observed in the tunnels of Rainier Mesa.

4.4.2 The Aardvark Event

The Aardvark event was one of the first underground nuclear detonations for which an attempt was made to develop an understanding of hydrologic disturbances resulting from such detonations. The nuclear device had a reported yield of about 35 kilotons. It was detonated beneath Yucca Flat, 71 m above the local water table. The emplacement chamber was situated within rocks of the tuff "pile". The location of the Aardvark detonation site, relative to the Bilby event, is shown on Plate 4.4.2-1.

The results of limited hydrologic observations made in association with the Aardvark event were reported by Garber, 1963, and later by Borg et al., 1976. Two different conceptual understandings of hydrologic disturbances have been proposed. A relatively simple understanding was put forth by Garber, 1963, and later by Knox et al., 1965. An alternate, more complex understanding was proposed by Reed, 1970.

A sustained increase of the local water table following the Aardvark event, was documented by Garber, 1963. This author reported the results of a two-year period of monitoring the water table in Well #7. The well is situated 307 m away from the Aardvark detonation site. Below the water table, the well is open over a 165 m vertical interval. Plate 4.4.2-2 presents the observed history of the water level in Well #7. Following the Aardvark event, the water level rose at least 17 m. Over a two-year period the water level slowly declined to the original level.

Garber, 1963, proposed several mechanisms by which the Aardvark detonation may have caused the observed hydrologic effect. Among those is a suggestion that straining of the host rock resulting from the growth of detonation cavity may have been responsible. This author speculated that a spherical shell of compressed rock, surrounding the 96-m-diameter cavity may have extended below the water table. Such straining caused a reduction of hydraulic storativity of the host rock, and produced an expulsion of groundwater upward and outward into the vadose zone. The slow rate of dissipation of the induced hydraulic anomaly was considered to have been the result of low hydraulic conductivity of the tuff "pile".

A similar understanding of hydrologic disturbance produced by the Aardvark detonation was proposed by Knox et al., 1965. According to this understanding, the detonation induced straining of the nearly saturated host rock caused a reduction in hydraulic storativity of this rock. The reduction occurred in a spherical shell centered on the detonation cavity, and brought the strained rock to full saturation. Ex-

pulsion of excess water from the shell saturated more rock beyond. Furthermore, collapse of the detonation cavity produced a cylindrical chimney within which hydraulic storativity was substantially enhanced. The overall hydrologic effect of the Aardvark explosion was to create a large annular hydraulic mound centered on the detonation site. The mound contained a central depression established by drainage of groundwater into the rubble chimney.

The above conceptual understanding has been tested by reproducing, after the fact, the history of water level observed in Well #7. This testing was performed by Knox et al., 1965, using an axially symmetric model of transient groundwater flow in an unconfined aquifer. A groundwater mound, about 200 m tall and 250 m in radius, constituted the initial boundary conditions. The model yielded a history of the mound dissipation shown on Plate 4.4.2-3. Adjusting parameters used in the model, the authors were able to approximately reproduce the observed water level rise and subsequent decline in test Well #7 (Plate 4.4.2-2).

A different conceptual understanding of the hydrologic disturbance produced by the Aardvark event was proposed by Reed, 1970. This author concluded that a complex thermo-hydraulic coupling was responsible for the observed changes in water level. The detonation has created a dynamic situation whereby groundwater flowed away from the thermal anomaly produced around the detonation cavity. Because the explosion induced enhancement of hydraulic conductivity markedly decreases to background, presumably a short distance from the detonation site, the outward radial flow of groundwater would be retarded until a threshold hydraulic gradient is attained. The magnitude of the water table rise, at the zone of lessening hydraulic conductivity, was considered to be related to the hydraulic conductivity contrast produced.

4.4.3 The Bilby Event

The Bilby event occurred in central Yucca Flat on September 13, 1963. The detonated nuclear device had a reported yield of approximately 200 kilotons, and was buried 2,339 feet below the ground surface. The Bilby event was the first nuclear explosion detonated in Yucca Flat below the water table. The device was emplaced within rocks of the tuff "pile".

Post-detonation geologic reconnaissance revealed that the Bilby explosion triggered displacements along two local faults, the Yucca fault and the Area #3 fault (Germain and Kahn, 1969). The length of rupture was 15 x 10^3 ft for the Yucca fault, and about 9 x 10^3 ft for the Area #3 fault. The maximum vertical fault displacement was 4 in. for the Yucca fault, and 6 in. for the Area #3 Fault. In the vicinity of detonation site, ground heaving was observed. The magnitude of such heaving was equal to about 140 in.

To evaluate the nature and magnitude of hydrologic disturbances produced by the Bilby explosion, measurements of water levels were made in three observation wells. All of the observation wells were situated near the detonation site. The results of these measurements were published by Hale et al., 1963; Garber, 1971; Corchery and Dinwiddie, 1976; and Borg et al., 1976. Around the detonation site, repeated electric resistivity and self-potential surveys were made. The results of such surveys were interpreted and reported by Scott et al., 1964.

The number of observation wells, employed in studies of hydrologic disturbances produced by the Bilby event, was far too small to permit a reliable and complete interpretation of the resulting hydrologic anomaly. Both the initial configuration and dissipation history of the anomaly are poorly known. Consequently, phenomenology and areal pattern of the induced groundwater flow are largely a matter of conjecture and speculation.

The location of the Bilby detonation site is shown on Plate 4.4.3-1. The rise of fluid pressure produced by the Bilby shock wave is shown on Plate 4.4.3-2. Plates 4.4.3-3 and 4.4.3-4 are hydrographs obtained from two of the three observation wells shortly after the explosion. After the event, borehole U3-cn PS#2 was drilled into the Bilby rubble chimney. The water level rise, so-called chimney in-fill, was

observed during a period from November, 1963 to May, 1968. The results of these observations are presented on Plate 4.4.3-5.

Immediately after the detonation, fluid pressure rose about 76 m in observation Well U3-4, which is located about 600 m from ground zero (Plate 4.4.3-2). Observation Well #7, which is located about 1,270 m away from ground zero, recorded a fluid pressure rise of about 20 m. Observation Well E, which is completed in the Paleozoic carbonates and is located about 2,400 m from ground zero, recorded a fluid pressure rise of about 3 m.

A day after the Bilby event, the water level in Well U3 cn-4 had risen at least 146 m (Plate 4.4.3-3). The rise was then followed by dissipation of about 40 m of hydraulic head during a period of about 1 day. Similar initial drop of fluid pressure was observed in observation Well #7. Within a day of the event, fluid pressure in this well declined a few meters (Plate 4.4.3-4). This decline was then followed by a gradual rise of fluid pressure. Over a period of about 10 days, the additional rise was equal to about 10 m.

The results of electric resistivity and self-potential measurements were interpreted as indicating that the Bilby explosion has produced the water table rise, not merely pore pressure increases (Scott et al., 1964). Near the detonation site, the water table rise was interpreted to be equal to several tens of meters. The rise diminished progressively with distance away from the detonation site. Several hundred meters away, the rise was no longer detectable. Post-detonation electric resistivity measurements were interpreted to indicate that the groundwater mound decayed to its original configuration within a period of about 3 months.

The in-fill of rubble-chimney, produced by the Bilby explosion, was monitored in a vertical borehole (Garber, 1971). This borehole was drilled into the chimney from a position very close to ground zero. The Bilby chimney extended from the ground surface to a depth of about 700 m. The groundwater was first detected at a depth of about 590 m, two months after the event. The history of subsequent rise of fluid pressure in the rubble chimney is shown on Plate 4.4.3-5. A period of slightly more than five years was required for the water table to return to the pre-detonation position.

The observed in-fill history of the Bilby rubble chimney was reproduced, after the fact, by Fenske and Charnell, 1971. A one-dimensional model of transient groundwater flow in an unconfined aquifer was used

for this purpose. A simple depression, centered about the chimney's axis, in a flat original potentiometric surface was assumed as the initial boundary conditions.

Based on preliminary analyses of the hydraulic anomaly produced by the Bilby explosion, Hale et al., 1963, concluded that the growth of detonation cavity caused a reduction of hydraulic storativity of the host bedrock. Such a reduction was expressed as the observed rise of fluid pressure. The rise may have been augmented by a thermally induced outward flow of groundwater from the detonation site. Following the fluid pressure rise, compensating groundwater flow commenced. The hydraulic anomaly produced was presumed to be in the form of a groundwater mound containing a depression centered around the rubble chimney. Dissipation of the anomaly occurred through radial flow into the central depression and outward away from the hydraulic mound.

Local hydrologic conditions around Yucca Flat, and presumably around the Bilby and Aardvark detonation sites, were discussed in Sections 4.2.4.2 and 4.2.5. The characteristic feature of this area is the depthward decrease of the value of hydraulic potentials. The value of local "underpressure" ranges from 43 m in Well #88-66 to 160 m in Well #87-62 (Winograd and Thordarson, 1975). Values of the closure pressure are high and comparable or exceeding those expected from the overburden load. Below the water table, the *in situ* stress enhancement of the local hydraulic conductivity structure is either very minimal or non-existent. The hydraulic conductivity structure is dominated by interstitial porosity and/or residual apertures of fractures. Such conditions may be responsible for the observed pattern of temporal behavior of the detonation induced fluid pressure build-ups. As shown on Plates 4.4.3-2, 4.4.3-3, and 4.4.3-4, the characteristic feature of this pattern is the absence of abrupt, oscillatory fluctuations of fluid pressure. Evidently, in the vicinity of Yucca Flat, the detonation induced build-ups and subsequent dissipations of fluid pressure are controlled by residual apertures of fractures and/or interstitial porosity. The magnitudes of induced fluid pressure build-ups, relative to the *in situ* values of closure pressure, are not sufficient to cause dilation of fractures. This is in sharp contrast to Pahute Mesa, where the detonation induced fluid pressure build-ups are accompanied by abrupt reversals and long-lasting oscillatory fluctuations (Section 4.4.4).

4.4.4 The Handley Event

The Handley event occurred on March 26, 1970, in Pahute Mesa. The nuclear device had a reported yield of about 1 megaton. It was detonated in emplacement hole U-20 m. The location of this hole is shown on Plate 4.4.4-1. Many measurements and observations pertaining to deformational and hydrologic effects of the Handley event were made. The results of these measurements were described by Dudley et al., 1971; Morris, 1971; and Bucknam and Dickey, 1971.

Within a few kilometers of the Handley detonation site, measurements of the induced deformation of the ground surface were made using high-precision leveling and Geodimeter distance measurements (Section 4.4.1). The data showed subsidence caused by collapse of the detonation cavity and significant bedrock straining out to at least 12 km from the detonation site. Displacements along local faults were observed as much as 8 km away. A maximum of 60 cm of vertical displacement occurred on a fault east of the detonation site. Prior to the Handley event, long-term creep was observed along the Boxcar fault.

In anticipation of the Handley event, pressure transducers were installed in 13 pre-existing wells. The most distant monitoring site was located 112 km northwest of the detonation site. The location of some of the monitoring sites is shown on Plate 4.4.4-2. Two wells at the Amargosa Tracer Site were monitored with ± 2.5 psid transducers. Recording of the fluid pressure response was made with an oscillograph operating with a chart speed of 0.64 cm/sec. One of the two wells, Sample Hole #1, was monitored as an open well. The other well, Tracer Well #3, was monitored at two different depth intervals that were isolated with pneumatic packers. The packers were set in a 150 m section of unperforated casing. Transducers were installed in the isolated casing interval and beneath the lower packer, in a transmissive section of the Paleozoic carbonates.

The dynamic fluid overpressure produced by the Handley shock wave is presented on Plate 4.4.4-3. At a distance of about 5 km from the detonation site, the value of dynamic overpressure was equal to about 100 m. The overpressure attenuated with distance from the detonation site; in Well UE-18r, located 20 km away, it was equal to about 7 m. Beyond a distance of 50 km, the value of overpressure was less than 2 m. Plate 4.4.4-4 presents portions of hydrographs obtained from the two Amargosa Tracer Site wells. There is a marked difference in the records generated in the isolated casing interval and in the well interval open to the aquifer. During high-frequency vibratory ground motion, the upper sensor registered

about 10 to 20 percent of the effect observed in the aquifer. Subsequent fluid pressure fluctuations were observed only in the aquifer; within the isolated casing interval, fluid pressure remained unaffected.

Hydrologic observations made in close proximity to the Handley detonation site are of particular importance. Four wells situated in Pahute Mesa (PM-1, PM-2, UE-20, and UE-20p) and one well situated between Buckboard Mesa and Pahute Mesa (UE-18r) were observed at the detonation time. Water levels and fluid pressures were monitored in these wells for 3 months after the event. The locations of these wells are shown on Plate 4.4.4-5. The results of monitoring were reported by Dudley et al., 1971, and Corchary and Dinwiddie, 1976. They are also shown on Plates 4.4.4-6 through 4.4.4-8.

The monitoring of dynamic fluid overpressure, produced by the Handley shock wave near the detonation site, was only partially successful. Apparently, the Rustrak recorders used in monitoring the Pahute Mesa wells were too slow to reveal cyclic responses to the shock wave. Nevertheless, the dynamic fluid overpressure of more than 92 m was recorded in Well UE-20f, located about 5 km away from the detonation site (Plate 4.4.4-6).

Sustained and oscillatory fluid pressure changes were observed for more than a month after the Handley event. Such changes were observed in two wells: UE-20f and UE-20p. In Well UE-20f, significant changes in water level were recorded for 214 minutes after the detonation time. These changes are summarized on Plate 4.4.4-7. The magnitude of greatest positive excursions was 24.5 m. There were many positive excursions with magnitudes ranging from 5 to 15 m. The magnitude of greatest negative excursion was -28.7 m; there were numerous excursions in a range of -10 m. During the first 214 minutes after the event, the maximum change in fluid pressure was equal to about 53.2 m. The activity ended abruptly, but a residual increase of about 1.1 m remained. Well UE-20f had shown significant sustained fluid pressure increases after some previous detonations, most notably the Jorum event (Dudley et al., 1971). This well, approximately 4,172 m deep, penetrates into the aftershock hypocentral zone of a seismic event triggered by the Benham detonation (Hamilton and Healy, 1969).

In Well UE-20p, a drop of fluid pressure of about 3 m was recorded at the detonation time. The initial drop was followed by a 45-minute recovery to 0.5 m above the pre-detonation water level. This peak decayed in about 15 minutes. A sustained rise in fluid pressure of about 23 m occurred on March 29, four days after the event. The rise was followed by further erratic fluctuations of fluid pressure. The maximum

value of fluid pressure rise was about 50 m (Plate 4.4.4-8). The monitoring of fluid pressure was interrupted on March 30, when the cable was pulled from the recorder shelter by an unidentified person. On May 6, 1970, another fluid pressure rise occurred. The recorder was off scale (> + 8 m) for 10 of the 14 days of the fluid pressure rise (Plate 4.4.4-6).

In Well UE-20p, the well casing is perforated in a depth interval from 348 to 783 m. Within this depth interval, high values of hydraulic conductivity were measured. The water table was at a depth of about 280 m below the ground surface. In this situation, fluid pressure increase in the well could dissipate through upward seepage with only a slight increase in position of the water table. Obviously, very large volumes of groundwater must have been involved to produce the recorded fluctuations of fluid pressure in this well.

Local hydrologic conditions in Pahute Mesa and around the Handley detonation site, were discussed in Section 4.2.4.3. Two noteworthy features were observed and interpreted in this area. The first feature is the possible presence of the "limit equilibrium" stress conditions within the first 2,500 feet below the water table. It is likely, that below the water table, the in situ values of closure pressure are low and conducting apertures of fractures are enhanced by the *in situ* stress. Such interpretation , however, is a matter of speculation because the results of direct measurements of the closure pressure are not yet available. The second feature is spatial distribution of the vertical gradients of hydraulic potentials (Plates 4.2.4.3-1 and 4.2.4.3-3). In the western sector of Pahute Mesa, the "overpressure" conditions are developed. The value of "overpressure" may be considerable. In Well UE-20f, for example, the value of "overpressure" was found to be equal to at least 142 feet (Plate 4.2.4.3-2).

Considering the above conditions, the following interpretation of the observed behavior of hydraulic potentials around the Handley detonation site seems to be reasonable. The detonation shock wave produced the dynamic fluid overpressure value of which was in a range of 100 m, at a distance of about 5 km from the detonation site. The dynamic overpressure together with the shock wave triggered a large-scale restructuring of the *in situ* stress field (Section 3.3.6). Such restructuring was expressed through seismic shocks, fault displacements, and deformation of the ground surface (Morris 1971; Bucknam and Dickey, 1971). The restructuring resulted in reduction of magnitude of the shear stresses along some fractures and/or increase of magnitude of the effective normal stress. Such changes, in turn, caused increases in the *in situ* values of closure pressure, and resulted in local reductions of hydraulic conductivity and storativ-

ity. The reductions were expressed as the fluid pressure rises observed in the monitoring wells. Groundwater involved in these rises is either water released from storage or it came from the deeper "overpressure" zone (Section 3.3.4). Evidently, the resulting fluid pressure build-ups were too high relative to the magnitudes of shock wave induced increases of the closure pressure. Consequently, dilation of previously closed fractures occurred resulting in increases of the local values of hydraulic conductivity and storativity. The fluid pressure dropped and fractures closed again. The process repeated itself over and over again until the fluid-bedrock coupled system found a new equilibrium configuration. Such configuration required the presence of sustained, but local, changes in position of the water table.

4.4.5 Effects of Underground Nuclear Detonations on Aqueous Chemistry

Effects of underground nuclear detonations on rates of discharge from local tunnel seeps and associated changes in aqueous chemistry, including content of stable isotopes, were investigated by Russell, 1987. The study was conducted in underground workings beneath Rainier Mesa, before and after two announced nuclear detonations on April 6, 1985, and April 10, 1986.

A tunnel seep discharge was monitored before and after the detonations. The discharge record obtained in association with the April 10, 1986 detonation is shown on Plate 4.4.5-1. A two-fold increase in the rate of groundwater discharge is evident. The apparent "bomb" pulse lasted for approximately 18 days.

Every five days, samples of groundwater discharging from two different seeps were collected and analyzed in the laboratory. The results of these analyses revealed that, corresponding with the increased rate of discharge, there was also an increase in the total dissolved solids of seep water. Changes in the content of specific ions are shown on Plates 4.4.5-2 through 4.4.5-5. The data represent a response of two different seeps to two different detonations. At the times of detonations, an increase in the concentration of most dissolved species is indicated by the data. The April 6, 1986 explosion was accompanied by a particularly large increase in the concentrations of sodium, sulfate, and bicarbonate. The large increase in sulfate is important because the presence of "even small quantities of sulfate minerals or their weathering products have never been reported within the rock formations of Rainier Mesa" (Russell, 1987). Changes in water chemistry, before and after the April 6 explosion, in the form of a Stiff diagram, are shown on Plate 4.4.5-6. Plate 4.4.5-7 is the Stiff representation of the changes in groundwater chemistry associated with the April 10, 1986 detonation.

The groundwater samples were also analyzed to determine their stable isotope content. The results obtained are shown on Plates 4.4.5-8 and 4.4.5-9. There are recognizable changes in isotopic contents as represented by oxygen-18 and deuterium.

Based on the above data, it is reasonable to conclude that straining of the host rock resulting from an underground nuclear detonation mobilizes water from some sources. The mobilized waters seem to be chemically and possibly isotopically different from waters residing in fractures. Russell, 1987, has proposed

that shock waves are forcing interstitial fluids out of the rock matrix. Such interpretation is similar to that proposed earlier by other authors to account for the observed hydrologic effects of underground nuclear detonations in Yucca Flat (Sections 4.4.2 and 4.4.3). Also, the interpretation that chemical composition of the Rainier Mesa interstitial fluids is different from that of fracture waters is in accordance with conclusions drawn earlier by two other authors. As discussed in Section 4.2.4.2, differences in chemical compositions of interstitial and fracture waters were noted by Keller, 1960 and 1962. A similar conclusion was also drawn by White et al., 1980 (Section 4.5.2).

Based on the above considerations, it is rather obvious that, within some members of the Rainier Mesa tuff "pile", interstitial fluids are strongly held by capillary forces. Notwithstanding the very substantial hydraulic gradients, such fluids either remain stagnant or infiltrate downward with exceedingly small velocities. This is in sharp contrast to fracture waters, for which the residence time is measured in terms of only a few years (Clebsch, 1960; Henne, 1982; and Russell, 1987). Large differences in the residence times indicate, that in the vadose zone of Rainier Mesa, the dominant mode of infiltration of "young" meteoric water is fracture flow. This infiltration mechanism operates without paying much attention to the fact that the *in situ* values of closure pressure are relatively high and, consequently, conducting apertures are small. In this situation, it is difficult not to question the expectation, that in other parts of the Death Valley groundwater system, infiltration of meteoric water occurs via the mechanism of interstitial flow. This is particularly true for areas, such as Yucca Mountain, where the *in situ* values of closure pressure are very low and, consequently, conducting apertures are very large.

4.5 LARGE-SCALE FLUCTUATIONS OF THE WATER TABLE AS POSSIBLY EXPRESSED IN THE GEOLOGIC RECORD OF THE DEATH VALLEY GROUNDWATER SYSTEM

4.5.1 Introduction

A careful and **open-minded** examination of the geologic record is an important part of the conceptual considerations of the Yucca Mountain groundwater system. Ultimately, this record will be a decisive factor in validating any conceptual understanding of this system.

If interpretations of various hydrologic, geologic, and geophysical data, as advanced in this report, are, at least, partially correct then the geologic record must contain some expressions of the cyclic and evolving behavior of the groundwater system during Plio-Quaternary time. As discussed in Section 3.5.3, termination of an individual evolutionary loop of a given portion of the groundwater system, is expected to be accompanied by injections of large volumes of fluid from below the water table into the vadose zone. It is expected, that in a geothermal setting similar to that of the Death Valley groundwater system, the injected fluids will contain large amounts of dissolved solids and may have an exotic chemistry. Both of these characteristics virtually assure that the process will leave behind an imprint in the geologic record, kind of "geologic foot prints." Examinations of the geologic record should be directed toward identifying and characterizing such "foot prints". Their presence may directly confirm that the flow system operates in a cyclically deforming fractured medium, and that it involves a heat-fluid coupled flow. Furthermore, the "foot prints" may yield data regarding a duration of single evolutionary loop of the system, and a magnitude of the associated hydrologic disturbances.

Based on the conceptual models presented in Section 3.0, it may be expected that past large-scale fluctuations of the water table are expressed in two-fold manner. First, the fluctuations may be expressed through occurrences of perched and semi-perched bodies of groundwater. As was explained in Sections 3.4.4 and 3.5.3, the formation of such bodies is expected to be occurring in association with hydraulic and thermal rises of the water table. The high values of ensuing hydraulic gradients cause the injected fluids to be introduced into the host rock matrix and thus preserved long after eventual cessation of injection. Chemical composition of the injected fluids, and initially their temperature, should be different from that of "ordinary" meteoric waters infiltrating through the vadose zone. For a hydraulic mound, that is either active or very young, the differences may be very pronounced and readily indicative of the origin of flu-

ids involved. For an older hydraulic mound, however, interactions between injected fluids, host rock, and infiltrating meteoric waters may cause the injected fluids to loose their thermal and geochemical identity. It is reasonable to expect, that for fracture water, loss of identity is much greater than for interstitial water, which may be strongly held by capillary forces. Hence, studies of chemical and isotopic compositions of interstitial waters offer the best opportunity for conclusive resolution of the question concerning phenomenology involved in the formation of a given body of perched or semi-perched water.

Second, large-scale fluctuations of the water table should also be expressed through occurrences of characteristic geologic materials. Textures, compositions, and forms of occurrence of such materials are indicative of their affiliation with deep-seated groundwaters. At the ground surface, the materials are expected to be occurring in the form of tuffa mounds, fragmentation breccias, paleo-seeps, and travertine aprons with associated feeder dykes. In the subsurface, the expected forms of occurrence of the materials are veins, explosive and fragmentation breccias, metasomatic replacements and alterations of wallrock. Furthermore, the materials are expected to be occurring at altitudes that cannot be justified by a) position and configuration of the contemporary water table; b) changes in position of the water table caused by past climatic fluctuations; and c) tectonic deformation of the land surface.

A review of geologic and hydrologic literature concerned with the Death Valley groundwater system was performed. A main purpose of such a review was to evaluate whether or not the expected expressions of past large-scale fluctuations of the water table are present. The results of this review are presented below. Section 4.5.2 deals with hydraulic features such as mounds, perched, and semi-perched waters. Section 4.5.3 is concerned with various geologic materials that may be indicative of habits of long-term behavior of the groundwater system.

It is important to recognize, that none of the examples described below, is considered as a conclusive and unequivocal demonstration of the validity of the proposed conceptual understanding of the Yucca Mountain groundwater system. In all cases considered, the currently available data base does not allow for a conclusive definition of origin of fluids responsible for the formation of a given geologic or hydrologic feature. By the same token, however, casual dismissal of a given example, as being explainable in terms of the traditional understanding, is not regarded as a sound and prudent practice. Both the geologic and hydrologic examples considered are important for validating any conceptual understanding of the Yucca Mountain groundwater system. As such they deserve careful and **open-minded** considerations. Such con-

siderations, however, must be made based on "surgically" obtained data, and keeping in mind both the range of conceptual possibilities that are permissable and **the significance of such possibilities in the context of long-term performance of the proposed high-level nuclear waste repository**.

4.5.2 The Vadose Zone

4.5.2.1 General

The vadose zone constitutes the upper portion of a groundwater system. It is located between the land surface and the uppermost continuous potentiometric surface, which is the water table. In the Death Valley groundwater system, thickness of the vadose zone ranges from a few meters, at and near the discharge areas, to as much as 600 m, in the central and recharge areas. In the vadose zone, water content, expressed as a percentage of interstitial space, is highly variable. It ranges from 60 to 75 percent, ±15 percent. In local areas, however, the degree of saturation is higher and fractures contain "free" waters. Such waters may be removed from the host rock by pumping and gravitational drainage.

Spatial distribution of water content in the vadose zone may be viewed as resulting from mixing of two patterns of distribution. The first pattern is a systematic change in moisture content as a function of depth and time (Ross, 1984). Near the ground surface, moisture content and suction potential fluctuate with precipitation events and with season. At depth, however, the fluctuations damp out, and there is a constant downward flux of water under the unit head gradient. Both, the suction potential and the effective hydraulic conductivity, are constant. Near the water table, moisture content is high and suction potential has a small negative value.

The second pattern is a random occurrence of bodies of "free" waters. Such bodies owe their existance either to local retentions or to local additions of fluids. Their origin may be related to many factors, some of which are a) local heterogeneity in the hydraulic conductivity structure, that may be caused either by lithologic composition or by an *in situ* stress field; b) locally increased infiltration, that may be the result of either atmospheric precipitation or man action; and c) injection of fluids from below the water table, that may be caused by either thermal or *in situ* stress rises of the water table.

For the purposes of these conceptual considerations, it is useful to recognize three forms of occurrence of the bodies of "free" water in the vadose zone. These are a) perched waters; b) semi-perched waters; and c) hydraulic mounds. The perched waters are separated from the underlying continuous body of groundwater by a sequence of "undersaturated" rocks. The semi-perched waters have higher hydraulic head than the underlying body of groundwater. In contrast to the perched waters, however, the semi-perched wa-

ters are not separated from the underlying body of groundwater by an "undersaturated" sequence. Such waters, similarly as the perched waters, owe their existence to local hydraulic conductivity contrasts, that restrict downward movement of fluids and thus prevent equilibration of hydraulic pressures across them. Within a hydraulic mound, however, the values of hydraulic potentials are higher than those in the upstream areas; locally, equipotential surfaces are bent upward. Such a bending is caused by local upwelling of fluids, that may be related either to "forced" or to "mixed" convection. Unlike both the perched and semi-perched bodies of water, hydraulic mounds owe their existence to local "upward-fluid-flux" boundary conditions, rather than to the local hydraulic conductivity contrasts.

Once encountered, bodies of "free" water pose an interesting and very important conceptual dilemma. Conceivably, there are two conceptual possibilities that need to be considered. The first possibility is, that a body of "free" water represents meteoric water directly from atmospheric precipitation that has been caught during downward infiltration through the vadose zone, by rocks whose hydraulic transmissivity is very low. Most, if not all, of the researchers concerned with the Death Valley groundwater system regarded this possibility as an adequate explanation of all "free" water occurring in the vadose zone of that system. The second possibility is, that at least, some bodies of "free" waters are the result of tectonic injection from below the water table. These bodies, therefore, represent tectonically induced hydraulic mounds, either active or relict.

It is rather obvious, that selection of a correct conceptual possibility regarding the origin of bodies of "free" water in the vadose zone, entails very profound implications as far as conceptual understanding of the Yucca Mountain groundwater system is concerned. Consequently, such a selection shall be made with due caution and based on an adequate data base. Such a data base should include the following categories of information: a) chemical and isotopic composition of fluids involved; b) local tectonic and hydrologic setting; c) contemporary and paleo-temperatures; d) volume of fluid; and f) values of hydraulic potentials, including their time- and space-dependence.

In the Death Valley groundwater system, bodies of "free" water are known to be present at several locations. Such bodies were encountered in many exploratory boreholes (Walker, 1962; Winograd and Thordarson, 1975; and Czarnecki, 1987). They were also encountered in underground workings in Rainier Mesa and Climax Stock (Thordarson, 1965; and Isherwood et al., 1982). At the Nevada Test Site, bodies of perched water feed several springs, that yield small quantities of water. Examples of such springs are

Cane Spring, Topopah Spring, and Pavits Spring (Thordarson and Robison, 1971; and Schoff and Moore, 1968).

For purposes of subsequent considerations, four examples of bodies of "free" water were selected for closer examinations. These examples are from the following locations: a) Greenwater Range; b) Skull Mountain; c) Rainier Mesa; and d) Climax Stock. The locations of these areas are shown on Plate 4.5.2.1-1.

4.5.2.2 The Greenwater Range Hydrologic Anomaly

Recent hydrologic studies, performed in the vicinity of Furnace Creek Ranch - Franklin Lake Playa, revealed that a groundwater "divide" may exist underneath Greenwater Range (Czarnecki, 1987). Greenwater Range separates Amargosa Desert from the Death Valley tectonic depression (Plates 4.5.2.1-1 and 4.5.3-8). Under Greenwater Range, the hydraulic head is as great as 875 m above mean sea level. Upstream, in Amargosa Desert, the hydraulic head is at an altitude of about 615 m above mean sea level. Downstream, in Death Valley, the water table is situated at about mean sea level. Clearly, a contemporary potentiometric high of about 260 m is present. A question that emerges is as follows: Is this potentiometric high caused by locally increased recharge from meteoric precipitation, or are tectonic factors involved in the formation of this feature? Unfortunately, the current data base does not allow for a satisfactory resolution of that question.

As shown on Plate 4.1-3, the average annual precipitation over the Funeral-Greenwater Range ranges from 2 in. to a maximum of 6 in. The empirical relationship between precipitation and recharge established by Eakin et al., 1963, and Eakin, 1963, suggests that no recharge occurs in areas where the annual mean precipitation is less than 200 mm, that is less than about 8 in. In light of this information, a straightforward and intuitively appealing notion that a locally increased infiltration of rainwater is responsible for the Greenwater Range potentiometric high is open to questions, and seems to be doubtful at best. An alternative explanation, one with much greater conceptual implications, is that the Greenwater Range potentiometric high is caused by the local "upward-fluid-flux" boundary conditions. Fluids that are contained within the mound originated from below the water table, and may represent either the *in situ* stress or thermal rise of the local water table. The feature appears to be youthful. A relatively modest subsurface exploration effort is required for unequivocal resolution of this controversy.

4.5.2.3 The Skull Mountain Hydrologic Anomaly

Plates 4.5.2.3-1 through 4.5.2.3-3 present the hydrologic setting of the Skull Mountain area. This area is located in the southern sector of the Nevada Test Site, between Frenchman and Jackass flats. Hydrology of this area is known based on subsurface explorations with boreholes, and was described by Winograd and Thordarson, 1975. Around Skull Mountain, altitudes of the "regional" water table range from 2,420 to 2,380 feet above mean sea level (Plates 4.5.2.3-2 and 4.5.2.3-3). A sharp increase in position of the water table occurs west of the Mercury Highway (Plate 4.5.2.3-3). This increase was encountered in two wells. The first well, Well #73-68, was drilled to a depth of 1,504 feet. Winograd and Thordarson, 1975, reported that "drillers first reported water at a depth of approximately 660 feet. Several days after completion of drilling, the water level stood 518 feet below land surface," which corresponds to an altitude of about 2,982 feet above mean sea level. This altitude is more than 500 feet higher than the regional water table downstream in the neighboring Frenchmen Flat (Plate 4.5.2.3-3). The second well, Well #77-68, is located in the northwestern sector of the area (Plate 4.5.2.3-3). Winograd and Thordarson, 1975, reported that "Well #77-68 penetrated 1,093 feet of the valley fill aquifer and 150 feet of the welded tuff aquifer. The water level in Well #77-68 stands at a depth of 796 ± 5 feet, or at an altitude of about 2,784 feet above mean sea level; this altitude is approximately 370 feet higher than water level in either the valley-fill or the welded-tuff aquifers east of the Mercury Highway" (Plate 4.5.2.3-3). The water level in Well #77-68 is also 364 feet higher than the water levels upstream, north and northeast from Skull Mountain, (Plate 4.5.2.3-2). Winograd and Thordarson, 1975, concluded that "because of the moderate to high transmissivity of the valley-fill and the welded-tuff aquifers, the anomalously high water level in Well #77-68 cannot be ascribed to perching or semi-perching of water within strata of low gross transmissivity or to the existence of steep hydraulic gradients within the aquifers."

Farther west of the Mercury Highway, west and south of Wells #73-69 and #77-68, perched waters were encountered in three wells: #73-66, #75-66a, and #75-66 (Plate 4.5.2.3-3). In addition, two small springs are found in the area: Cane Spring and Pavits Spring. The springs occur at altitudes of 4,060 and 3,940 feet above mean sea level, respectively. The perched waters, feeding the springs and encountered in the three wells, are situated as much as 1,200 to 1,600 feet higher than the "regional" water table downstream in Frenchman and upstream in Yucca Flats.

Based on groundwater chemistry data compiled by Winograd and Thordarson, 1975, (Plate 3), it appears that around Skull Mountain there is a noticeable increase in the sulfate content of groundwater. North, east, and west from Skull Mountain, the sulfate content is low, say less than 2 milliequivalents per liter. Immediately around Skull Mountain, however, the sulfate content is several times greater. The sulfate rich waters were encountered in three wells: #68-69, #73-66, and #74-61. Plate 4.5.2.3-4 presents the results of chemical analyses of water samples obtained from Well #68-69, and from three depth intervals in test Well #73-66.

Well #68-69 is located in central Mercury Valley, a few kilometers southeast from Skull Mountain. A sample of water bailed from this well yielded highly anomalous results (Plate 4.5.2.3-4). The total dissolved solids content was found to be equal to 5,420 milligrams per liter. Sodium constituted about 75 percent of the total cations, and sulfate about 95 percent of the anions.

The results of chemical analyses of water samples swabbed during the drill-stem tests in Well #73-66 (depth intervals 77 to 693 feet and 1,565 to 1,695 feet), and pumped from the Paleozoic carbonates (depth 3,140 to 3,400 feet) are shown on Plate 4.5.2.3-4. The location of Well #73-60 is shown on Plate 4.5.2.3-3. The sulfate content increases depthward from 34 to 181 mg/L. The sample of perched water, from depth interval 1,565 to 1,695 feet, was high in total dissolved solids (981 mg/L), high in sodium (424 mg/L), but very low in calcium (4 mg/L).

Well #74-61 is located in central Jackass Flat, approximately 8 km west of Skull Mountain. Winograd and Thordarson, 1975, reported that the sulfate content of water from this well is high, about 9 meg/L. In contrast, the sulfate content of water from two nearby wells, Wells #74-57 and #73-58, is less than 0.5 meg/L. These differences suggest that "dissolution of pyrite" in upstream areas, hydrothermally altered Calico Hills for example, should not be held responsible for the locally increased content of sulfate anions (Winograd and Thordarson, 1975).

Geothermal conditions around Skull Mountain were considered in Section 4.3.2. It can be recalled that a) the area contains the low seismic velocity anomaly (Plate 3.4.1-3); b) *in situ* temperatures of groundwater are higher than those in the surrounding regions; and c) around well UE-25a3, the hydrothermal convection system is present. The *in situ* temperature of perched water, encountered in Well #73-66 at a depth of 1,565 to 1,695 feet, is equal to 33.5°C (Plate 4.5.2.3-4). This temperature is about 12°C

higher than the corresponding temperature of semi-perched waters at Rainier Mesa (Plate 4.3.2-3 and 4.3.2-4). In the Paleozoic carbonates, the *in situ* temperature of groundwater in Well #73-66, at a depth interval 3,140 to 3,400 feet, was found to be 64.5°C (Plate 4.5.2.3-4). This temperature is 10 to 30°C higher than the corresponding temperatures at Yucca Mountain and around Yucca Flat (Plates 4.3.2-2 and 4.3.2-7).

The possible depthward increase in the value of hydraulic potentials within the body of perched water, as noted during drilling Well #73-68; the progressive increase in altitudes of the water table in Wells #77-68 and #73-68; and an anomalously high water level in the valley-fill aquifer as encountered in Well #77=68, relative to the upstream areas; all suggest that groundwaters involved in the Skull Mountain hydrologic anomaly may not be "ordinary" meteoric water infiltrating through the vadose zone. Such a suspicion is somewhat strengthened by both the groundwater chemistry and geothermal data. It is possible, that the anomalous waters may be related to the underlying aquifer and are, in fact, expressing a deep rooted hydraulic mound. Spatial correspondence between the location of the inferred hydraulic mound and the low-seismic velocity anomaly on Plate 3.4.1-3 suggests that the proposed explanation has merit and should be explored further. In any event, it is rather obvious that the current data base is not sufficient to either substantiate or refute that or any other explanation.

4.5.2.4 The Rainier Mesa Hydrologic Anomaly

The hydrologic setting of Rainier Mesa is shown on Plates 4.5.2.4-1 and 4.5.2.4-2. This setting is known based on extensive exploration with boreholes, and based on direct observations made in the complex of underground workings excavated for the purposes of underground testing of nuclear devices. The hydrologic setting of Rainier Mesa was described previously in Sections 4.2.4-2, 4.2.5-3, and 4.4.5.

The position of the water table has been determined based on numerous test borings drilled into Rainier Mesa. Altitudes of the water table vary up to 100 m. The mean altitude of the water table is about 1,820 m above mean sea level (Thordarson, 1965). This altitude is several hundred meters higher than east, west, and north of Rainier Mesa (Plates 4.5.2.3-1 and 4.5.2.4-1). Clearly, the water table is bent upward forming a hydraulic mound-like feature. The downhole measurements of hydraulic potentials have shown that the values of such potentials diminish depthward (Plate 4.2.4.2-3). This indicates, of course, that the feature is not caused by currently active upwelling of groundwater. Instead, the feature seems

to be related to local hydraulic transmissivity contrasts that prevent equilibration of hydraulic potentials across them. The feature, therefore, may be regarded as a body of semi-perched water. This conclusion is pretty straightforward and there is little room for alternate interpretations. A principal question, however, that needs to be addressed is: What is the origin of fluids forming the Rainier Mesa hydrologic anomaly?

Several hydrologic studies have demonstrated, that at Rainier Mesa, the overall movement of groundwater is downward, from the recharge area at the top of the mesa, through fractures in the Rainier Mesa Member, and then through the underlying Paintbrush Tuff (Thordarson, 1965; White et al., 1980; and Russell, 1987). Groundwater travel-time, between the top of the mesa and the tunnel complex, is no more than a few years (Clebsh, 1960; Henne, 1982; Russell, 1987). Upon reaching more saturated rocks or less permeable zeolitic tuffs, infiltrating meteoric waters are retarded. Thus formed perched waters slowly drain through the tuff "pile", either into the underlying Paleozoic carbonates or into the tunnel complex. Perched waters are primarily found in fractures. The majority of such water-bearing fractures are normal faults with small displacements that seldom exceed several centimeters of net slip. An examination of fractures in tunnels revealed that 50 to 60 percent of faults were water bearing (Thordarson, 1965). Some fractures drain and dry out within a few weeks after mining. Others, that may be relatively close by, have acted as continuous seeps since the initial excavation, albeit at much lower discharge rates than initially recorded.

Geochemistry of groundwater from Rainier Mesa was investigated by White et al., 1980. Samples were obtained from both rock matrix and free-flowing fractures. Samples of fracture water were collected from seeps encountered in various drifts and adits. Samples of interstitial water were obtained from selected sections of core, that were sealed in wax at a time of recovery. Interstitial fluids were extracted either by squeezing in multiple-cycle triaxle compression apparatus or by centrifugation at 2,000 revolutions per minute.

The results of chemical analyses of the Rainier Mesa interstitial waters are presented on Plate 4.5.2.4-3. Plate 4.5.2.4-4 presents the results of chemical analyses of fracture waters. The relative mole fractions of cations and anions, for both interstitial and fracture waters, are shown on Plates 4.5.2.4-5 and 4.5.2.4-6.

General concentrations of individual cations for fracture water are similar to those of interstitial water (Plates 4.5.2.4-3 and 4.5.2.4-5). The total cation concentrations are also very similar. White et al., 1980,

reported that the total average cation concentration in interstitial water is equal to 2.7 meg/L, which is only slightly higher than the 2.19 meg/L value for fracture water. A comparison of the concentrations of anions in fracture water with those in interstitial water, however, reveals that there are differences in the compositions of both types of water. In all samples representing fracture water, biocarbonate is the dominant anion (Plate 4.5.2.4-6). The chloride and sulfate anions are present in very low concentrations. This is in sharp contrast to interstitial waters, for which concentrations of chloride and sulfate anions are significantly higher.

At Rainier Mesa, the differences in chemical compositions of fracture and interstitial waters were also postulated by Keller, 1962, and later by Russell, 1987. The later author specifically noted that "the large increases in sulfate for these waters are anomalous because the presence of even small quantities of sulfate minerals and their weathering products have never been reported within the formations of Rainier Mesa." White et al., 1980, stated that "the mechanism responsible for differences in anion composition between interstitial and fracture waters is not obvious." This author examined a few possibilities that may account for the elevated sulfate and chloride concentrations in interstital water. The first possibility was that drilling fluids may have contaminated intersitial waters. Considering low interstitial permeability of rocks involved, such a possibility was not regarded as a likely one. The low permeability would inhibit widespread dissemination of contaminants through the rock matrix. Also, dye tests conducted by the Defense Nuclear Agency during drilling operations showed, that for zeolitic tuffs, drilling contamination is negligible (White et al., 1980).

The second possibility examined was, that the composition of interstitial water reflects a composition of relict water high in chloride and sulfate that was present at a time of deposition of the host rock. The preservation of such waters, over a time span of about 10^7 years, would require that a large percentage of pore space be insulated from the overall hydrologic system. This may be a reasonable proposition, but only for zeolitic tuffs, that exhibit relatively high porosity but low interstitial permeability. A relatively high interstitial permeability of the overlying Paintbrush Tuff, however, suggests that this proposition is not a plausible one.

The third possibility is that the type and extent of chemical reactions between the host rock and the interstitial and fracture waters are dissimilar. No prior reason is apparent, however, why chloride and sulfate anions are concentrated selectively in the interstitial water alone. Also, the fact that the total dis-

solved solid concentrations and the cation ratios for both types of water, are similar, tends to discount widely varying dissolution mechanisms.

A possibility which is proposed here is that the elevated sulfate and chloride concentrations in interstitial waters are expressing a relict hydraulic mound. The formation of such a mound resulted from a tectonic event that mobilized deep-seated groundwaters, in accordance with the conceptual model developed in Section 3.5.3. When this hypothetical mound was active, the mobilized fluids were introduced into the matrix of the host rock. At the present time, the mound is no longer active and the injected fluids are being gradually replaced by the infiltrating waters from meteoric precipitation. The differences in chemical compositions of interstitial water relative to fracture water reflect a difference in the replacement time for both types of water.

4.5.2.5 The Climax Stock Hydrologic Anomaly

The hydrologic setting of the Climax Stock is very complex and poorly understood; it was described by Schoff and Winograd, 1961; Walker, 1962; Murray, 1981; Isherwood et al., 1982; and previously in Sections 4.2.4-2 and 4.2.5-2. Geologic and potentiometric conditions around the Climax Stock are shown on Plates 4.5.2.5-1 and 4.5.2.5-2. In northern and eastern sectors of the area, the water table is situated at an altitude ranging from 1,200 to 1,300 m above sea level. South of the Climax Stock, however, altitudes of the water table range from 755 to 993 m above mean sea level. Murray, 1981, concluded that immediately below the Spent Fuel Test Facility the water table is at an altitude of about 975 m above mean sea level. Along the trace of Yucca Flat there is a well expressed potentiometric trough (Plate 4.5.2.5-2). Within and around the trough, the values of hydraulic potentials in the Paleozoic carbonates are less than those in the overlying tuff "pile". West and northwest of the Climax Stock, altitudes of the water table increase rapidly, and the sense of vertical hydraulic gradient changes. In the Paleozoic carbonates, the potentiometric surface is at an altitude of 1,382 m above the mean sea level, which is about 50 m higher than the potentiometric surface in the tuff "pile" (Plate 4.5.2.5-2).

The potentiometric data indicate that the Climax Stock is situated between a body of semi-perched water, in southwestern sector of the area, and a hydraulic mound, in the northeastern sector (Plates 4.5.2.4-1 and 4.5.2.4-2). In addition, bodies of perched water were encountered in several exploratory boreholes (Walker, 1962). Such bodies were also expressed as minor seeps into the complex of underground

workings (Walker, 1962; and Isherwood et al., 1982). Plate 4.5.2.5-3 presents the location of exploratory boreholes drilled into the Climax Stock. This plate also shows the potentiometric levels measured in the local bodies of perched water. Although the boreholes are located within an area of less than 4 km^2, the altitudes of fluid level range from 1,298 to 1,490 m above mean sea level; a total range of about 200 m. Such erratic levels indicate that bodies of perched water exist only locally, and that movement of fluid between adjacent bodies is severely restricted. Walker, 1962 noted that in some exploratory boreholes, fluid levels after bailing, recover to their approximate initial position. This observation indicates that, at least, some of the bodies of perched water are being replenished, and that such bodies are either of considerable spatial extent or represent "overflows" from some external sources. Similar conclusions may be drawn based on two other observations: a) seep discharges into the complex of underground openings, although occurring with rates ranging from 0.01 to 0.1 gpm, are continuous and b) fracture hydraulic transmissivity, as determined based on bailing and injection tests, is moderate and ranges from 2 to 140 gallons per day per foot (Walker, 1962).

Two sets of geochemical data, representing the Climax Stock perched waters, are available. Walker, 1962, reported the results of chemical and radiochemical analyses of three water samples: a) from borehole #31, at a depth of 122 m; b) from a seep in the shaft, at a depth of about 77 m; and c) from a seep in the cross-cut tunnel, at a depth of about 233 m below the ground surface. The results of these analyses are shown on Plate 4.5.2.5-4. For comparison purposes, this plate also presents the results of chemical analysis of water sample from Well #3. This well is located in Yucca Flat, and provided water used in drilling of all exploratory boreholes considered by Walker, 1962.

Isherwood et al., 1982, reported the results of chemical, radiochemical, and isotopic analyses of groundwater samples collected at five sites in the Climax underground workings. The locations of these sites are shown on Plate 4.5.2.5-5. The results of geochemical analyses of the five samples are presented on Plate 4.5.2.5-6. Plate 4.5.2.5-7 presents the results of reported isotopic analyses. The saturation indexes for seven common minerals, calculated for the five groundwater samples using the EQ-3 chemical equilibrium computer code, are shown on Plate 4.5.2.5-8. This plate also shows the total number of supersaturated minerals for each of the five samples.

Even a casual examination of Plates 4.5.2.5-6 through 4.5.2.5-8 indicates that the Climax Stock perched waters are intriguing, to say the least. These waters "fail to match the commonly held views re-

garding the nature of deep granitic waters" (Isherwood et al., 1982). The following three points underlie the highly unusual character of the Climax Stock perched waters. First, analyses of the chemical compositions of the groundwater samples, using the EQ-3 chemical equilibrium computer code, revealed that perched waters are not in equilibrium with many mineral phases present in either the host rock or fracture filling. The waters are supersaturated with respect to more than 38 mineral phases incorporated into the EQ-3 data base, including quartz, calcite, K-feldspar, muscovite, and kaolinite. The computed composition of groundwater in equilibrium with the host rock is one reduced in iron, aluminum, calcite, silica, and strontium.

Second, the most striking characteristic of the Climax Stock perched waters is the heterogeneity of chemical composition of individual samples. The total dissolved solids content ranges from 736 to as much as 1,910 mg/L. All of the samples contain uncommonly large amounts of sulfate anion. The concentration of this anion, however, is variable and ranges from 325 to 1,060 mg/L. There are also major differences in the concentrations of remaining constituents, including a) chloride, from 11 to 160 mg/L; b) biocarbonate, from 65 to 316 mg/L; c) sodium, from 50 to 250 mg.L; d) calcium from 0.9 to 118 mg/L; and e) magnesium, from 0.9 to 118 mg/L. Five out of the total of eight samples contained relatively high amounts of natural uranium (fission produced uranium isotopes were not detected), ranging in concentration from 1.8 to 26 ppm. The remaining three samples, however, contained uranium in concentrations below the employed detection limit (less than 100 ppb). The $^{234}U/^{238}U$ activity ratio, for two samples, ranged from 1.05 ± 0.02 to 1.115 ± 0.012 (Plate 4.5.2.5-7).

Third, the Climax Stock perched waters display chemical and isotopic affiliations with "deep-regional" groundwaters residing in the Paleozoic carbonates. Such affiliations are indicated by two lines of evidence. First, the $\partial^{13}C$ ratios for water samples from station CGW-1 are -3.54 and -4.57 o/oo (Plate 4.5.2.5-7). Such ratios indicate the increased ^{13}C content in, at least, some of the Climax Stock perched waters, relative to both atmospheric CO_2 ($\partial^{13}C \cong = -7$ o/oo) and shallow waters residing in rocks of Cenozoic age (Plate 4.5.2.5-9). The $\partial^{13}C$ ratios for the Climax Stock perched waters, however, are very similar to those for waters residing in the Paleozoic carbonates (Plate 4.5.2.5-10). Second, the relative proportions of major constituents in the Climax Stock perched waters are similar to those from "deep-regional" groundwaters discharging from the Paleozoic carbonates. Such similarities are illustrated on Plate 4.5.2.5-11.

Considering together the potentiometric and geochemical data, the following speculation regarding origin of the Climax Stock perched waters seems to be justified. The potentiometric data on Plate 4.5.2.4-1 indicate that, between the Climax Stock and Emigrant Valley, local potentiometric "high" is present. Within this high, the potentiometric levels are, at least, 50 m higher than altitudes of the water table upstream. As shown on Plate 4.5.2.5-2, the values of hydraulic potentials in the Paleozoic carbonates are higher than those in the overlying tuff "pile". This relationship indicates that local upwelling of deep fluids, rather than semi-perching of shallow waters, is responsible for formation of the potentiometric "high". The geochemical characteristics of the local perched waters may be accounted for, assuming that these waters represent "overflows" from the active hydraulic mound. Mixing of such "overflows" with the infiltrating waters directly from atmospheric precipitation may be responsible for the observed variability of chemical composition of individual samples.

4.5.2.6 Summary

In summary, the body of evidence regarding "free" waters in the vadose zone of the Death Valley groundwater system permits, at least, a **suspicion** that some such waters may have been injected from below the water table. This suspicion is consistent with the conceptual model for a heat-fluid coupled groundwater system developed in a deforming fractured medium, as established in Section 3.0. Should this suspicion be confirmed by further studies, the resulting conceptual implications are profound indeed. In this situation, the following two conclusions would be warranted:

a) In the Death Valley groundwater system, position and configuration of the water table are not durable characteristics. Both of these characteristics are related to tectonic factors, and are subject to possibly random changes; and

b) Magnitudes of potential rises of the water table are very substantial. Such rises may have long lasting hydrologic effects.

4.5.3 The Geologic Record

In the Death Valley groundwater system, surficial deposits composed of carbonate, opaline silica, and gypsum are known from many locations. According to Szabo et al.,1981, carbonates occur in a variety of forms. "They commonly fill fractures along faults, form dense, strongly cemented deposits in alluvium and are **precipitated by spring water**. The carbonates were broadly classified into five groups. The hard, dense, and finely crystallized carbonates **precipitating from groundwater are referred to as travertines**. The softer and more porous forms of the precipitated carbonates are defined as tuffaceous travertines. The surficial conglomerates or rock fragments and minerals, strongly cemented by authigenic carbonates are called calcretes. The secondary accumulations of cementing carbonate in the host material of a soil environment are identified as soil caliches. Finally the dense calcium carbonate in fractures in drill holes is referred to as calcite veins", Similar description of the carbonate accumulations at the Nevada Test Site was provided by Knauss, 1981. This author stated that "the carbonates with which we are concerned in this work are primarily shallow subsurface accumulations, in most cases within the zone of weathering. The terminology used to define or describe these accumulations is confusing and unnecessarily complex. For our purposes, we may consider the secondary carbonates sampled to be of three main types: travertine, calcrete, and caliche. **Travertine is a hard, dense, finely crystalline carbonate that precipitates from groundwater**. The modifier tuffaceous denotes a softer, more porous variety of travertine. Calcrete is a surficial sand and gravel cemented into a hard mass by calcium carbonate. Caliche is a secondary accumulation of carbonate usually associated with soil (soil caliche) but the term is used also to describe the cementing material itself (e.g., along a fault)."

Samples of all types of the carbonates have been collected from various locations throughout the Nevada Test Site and dated, using the uranium-series disequilibrium method. The results of these investigations were reported by Knauss, 1981, and Szabo et al., 1981. Locations of sample sites are shown on Plates 4.5.3-1 and 4.5.3-2. The descriptions of these sites are provided on Plates 4.5.3-3 and 4.5.3-4. Analytical data and the resulting uranium-series ages for various carbonates are presented on Plates 4.5.3-5 through 4.5.3-7.

The travertine and tuffaceous travertine deposits, that were studied, occur in three broad areas of the Nevada Test Site. Such areas are a) south of the Climax Stock along the trace of Boundary Fault (samples #BF LLL 1-1 and #BF USGS 5-5a, Plate 4.5.3-6b); b) around Skull Mountain (samples #30, #32,

#45, #46, #47, #CSF-1, and #CSF-2, Plates 4.5.3-5 and 4.5.3-6); and c) around Yucca Mountain (samples #106 and #199, Plate 4.5.3-5b). The uranium-series ages of the travertines range from more than 7 x 10^5 years B.P. to as little as 2 x 10^3 years B.P. (Plates 4.5.3-5 and 4.5.3-7). The travertines occuring along the Boundary fault yielded ages greater than 4 x 10^5 years B.P. Around Skull Mountain, however, three generations of travertines were identified. The uranium-series ages of these travertines are a) more than 7 x 10^5 years B.P.; b) between 1.1 x 10^5 and 7 x 10^4 years B.P.; and c) less than 3 x 10^4 years B.P. The tuffaceous travertines of Yucca Mountain yielded uranium series ages of about 7.8 x 10^4 and 3 x 10^4 years B.P. **All of the above travertines occur at locations where the contemporary water table is at a depth ranging from 120 m to several hundred meters below the ground surface.** In the Climax Stock and Skull Mountain, the travertines occur in proximity to the contemporary hydrologic anomalies that were considered in Section 4.5.2. It seems reasonable to suspect that factors responsible for the development of these anomalies may also be responsible for the recurring deposition of travertines.

At the Nevada Test Site, a deposit of gypsum, a few meters thick, was found near Whamonie. The outcrop of this deposit is located about 6 km northwest of Skull Mountain (Vaniman et al., 1988). The deposit occurs at the ground surface, in a topographic setting that precludes playa deposition, and in an area where the vadose zone is more than 500 m thick. It consists of almost pure gypsum with minor calcite, **and leaves little doubt that it represents a warm-water paleo-spring discharge** (Vaniman et al., 1988). Not much is known about the age of activity of this spring; field appearance of the outcrop, however, is one of a young deposit. The Whamonie paleo-spring mound occurs in a hydrologic setting where a) groundwaters are enriched in sulfate anion (Section 4.5.2); b) the intensity of flux of terrestrial heat is locally increased, (Section 4.3.2); c) the localized low-seismic velocity anomaly is present (Section 3.4.1); and d) hydrothermal convection system is locally developed (Sass et al., 1980). In this situation, it is reasonable to **speculate** that factors which are responsible for the Whamonie gypsum mound, may also be involved in the contemporary hydrology of Skull Mountain. In order to pursue such a hypothesis any further, however, it is necessary to establish the age of activity of the Whamonie mound. This age is a single most important data need that is required for a complete understanding of contemporary hydrology of the Skull Mountain area. A very modest field effort, consisting of mapping and trenching, is required to resolve this important issue.

A spectacular swarm of calcitic veins occurs along the eastern margin of the Death Valley tectonic depression. The swarm occurs along the trace of Furnace Creek fault, parallel to the western edge of the

Funeral-Greenwater Mountain Range. It was described by Winograd and Doty, 1980; and Winograd et al., 1985.

The calcitic veins are emplaced in nearly vertical fractures that are developed in well cemented fanglomerates of the Funeral Formation of Pliocene age. The veins vary from a millimeter to a meter in thickness, locally occur in swarms, and commonly are finely laminated. Individual veins may be traced vertically for tens of meters, and horizontally for over hundreds of meters. In places, **visible transitions of vein to tuffa mound may be observed. It is clear, therefore, that the veins mark sites of episodic fossil discharges of groundwater**. The veins originated as low-temperature, less than 50°C, precipitates from calcium carbonate saturated waters rising along extension, and most likely hydraulic, fractures. Three samples of the veins have been dated using the uranium-series method. The radiometric ages obtained range from 1.7×10^6 to 7×10^5 years B.P. (Winograd et al., 1985; and Winograd, 1988, personal communication). A minimum age of formation of the veins is not known. It is possible, however, that some of the veins may be considerably younger than 7×10^5 years B.P.

Plate 4.5.3-8 is a diagrammatic northeast-southwest cross-section that illustrates topography of the Funeral-Greenwater Mountain Range, from Death Valley, California, to Ash Meadows and Amargosa Desert, Nevada. The direction of regional groundwater flow is from northeast toward southwest (Plate 4.1-2). In Amargosa Desert, the ground surface is at an altitude of about 2,400 feet above mean sea level; in Death Valley the ground surface drops to below sea level. The Furnace Creek fault swarm of calcitic veins occurs at altitudes ranging from approximately 2,900 to 3,100 feet above mean sea level. Such altitudes are more than 500 feet greater than altitudes of the ground surface downstream and upstream along the path of regional groundwater flow.

There are two possibilities that may account for the above paleo-hydrologic anomaly. Either the Funeral-Greenwater Mountain Range has been uplifted some 500 feet, relative to the Amargosa Desert, during the last 7×10^5 years, or the calcite veins reflect repeated occurrences of tectonically generated hydraulic mounds, at least, during a time span from 1.7×10^6 to 7×10^5 years B.P. The presently available data do not allow for a conclusive resolution of this question.

Suspicion that the Funeral-Greenwater Range paleo-hydrologic anomaly is of tectonic origin, however, is strengthened by contemporary configuration of the water table. As discussed in Section 4.5.2.2, under

Greenwater Range, the water table is some 260 m higher than upstream in Amargosa Desert (Plate 4.5.3-8). This contemporary hydrologic anomaly may be representing a tectonically generated hydraulic mound. Further studies are required to either confirm or reject this proposition.

At and around Yucca Mountain, various vein deposits of calcite, opaline silica that includes opal CT (opal with cristobalite- and tridymite-type structure), with lesser quartz, opal A, sepiolite, and smectite are associated with Late Quaternary faults and surficial deposits (Swadley et al., 1984; Vaniman et al., 1988). The faults that contain the calcite-silica veins are a) the Windy Wash fault; b) the Solitario Canyon fault; c) the Bow Ridge fault; d) the Paintbrush fault; and e) the Fran Ridge fault. The locations of these faults are shown on Plates 4.5.3-9 and 4.5.3-10.

The calcite-silica deposits occur as veins emplaced along fault planes in both the unconsolidated alluvium of Late Quaternary age and in the Late Miocene bedrock. They also occur as aprons developed over unconsolidated alluvial and eolian sediments. Examples of both forms of occurrence of the calcite-silica deposits are shown on Plates 4.5.3-11 and 4.5.3-12. Thickness of the deposits ranges from tens of centimeters to a maximum of about two meters. Typically, the deposits exhibit banded textures, indicating the involvement of fluids with alternating geochemical characteristics, either during multiple events or during a single event. The veins that are emplaced within the Bow Ridge fault and the Windy Wash fault contain thin stringers of uncemented basaltic ash that displays petrographic and chemical affinity with one of two volcanic centers located nearby (Whitney et al., 1986). Ages of these centers have been determined, based on radiometric dating of associated lavas, to be approximately 1.2×10^6 years B.P. and possibly less than 10^5 years B.P. (Vaniman et al., 1982; Crowe and Vaniman, 1985).

Traces of the Windy Wash fault, the Solitario Canyon fault, the Ghost Dance fault, the Bow Ridge fault, the Paintbrush fault, and the Fran Ridge fault are marked by occurrences of the "mosaic" breccias (Section 3.3.2.3 and Plate 4.5.3-10). Commonly, these breccias form wallrocks of the calcite-silica veins. The breccias are older than the veins emplaced within them. Thickness of the breccia zones is highly variable, ranging from several meters to tens of meters and more.

Various origins for the calcite-silica veins have been proposed including precipitation from descending meteoric waters and from slightly elevated temperature ascending water. It was also considered, that the

deposits may have formed by multiple processes involving **deep-seated groundwaters**, perched groundwater, surficial runoff water, or **water flow induced by seismic activity** (Hanson et al., 1987).

Stable isotope analyses were performed on three samples representing the calcite-silica deposits. Two such samples represent deposits from the Bow Ridge fault, and one sample represents the calcite-silica deposits from eolian sediments overlying the Fran Ridge fault. The samples have the following $\partial^{18}O$ and $\partial^{13}C$ per mil isotopic composition referenced to the Standard Mean Ocean Water (SMOW) and the Pee Dee Belemnite (PBD) respectively (O'Neil, 1984 and 1985):

The Bow Ridge Fault	**$\partial^{18}O$ (SMOW)**	**$\partial^{13}C$ (PBD)**
Sample: T-14 FB (calcite)	19.6	-5.3
Sample: T-14-3a-w (calcite)	20.3	-7.5
The Fran Ridge Fault		
Sample: FR 6 (calcite)	19.8	-7.0

Plate 4.5.3-13 presents a comparison of the stable isotope compositions of the three samples with the stable isotope compositions of samples of the subsurface veins that were considered in Section 4.3.4. The comparison suggests that temperatures of formation of the deposits, represented by samples T-14 FB, T-14-3a-w, and FR-6, range from approximately 24 to about 27°C. Plate 4.5.3-14 presents a comparison of these temperatures with the contemporary temperatures as measured in Wells USW G-2, USW G-1, and USW H-1. Temperatures of formation of the calcite-silica deposits appear to be 5 to 8°C higher than the contemporary *in situ* temperature. This may suggest, but does not prove, that warmer subsurface fluids are responsible for the formation of the calcite-silica deposits.

It is important to note, that both the mineralogic and the stable isotope compositions of the Yucca Mountain veins are similar to those of Late Pliocene deposits occurring in the Amargosa basin. This contemporary and paleo-spring fed structural basin contains an extensive sequence of surficial deposits that were precipitated in and around shallow lakes or ponds, swamps, and springs. The sequence consists of a variety of mineral phases, including the entire mineralogic assemblage that comprises the Yucca Mountain veins and aprons. A thorough investigation of the Amargosa basin deposits was performed by Hay et al., 1986. The investigation revealed that the deposits consist of carbonate rocks intercalated with opal CT, sepiolites formed in low salinity waters, and chemically precipitated Mg smectites. The presence of

opal CT and sepiolite at both locations is important. This presence indicates that both of these mineral phases, occurring separately or in combination, are not uniquely characteristic of deposits resulting from "pedogenic" processes that involve fluids directly from meteoric precipitation, either in run-off or in infiltration modes. Instead, the Amargosa basin example demonstrates that "pedogenic" processes resulting from ground surface dispositions of fluids discharging from deep-seated seeps and springs may yield the opal CT-sepiolite sequence.

The stable isotope analyses of a large number of samples revealed, that the Pliocene deposits of the Amargosa basin have the $\partial^{18}O$ (SMOW) ratios averaging about + 19 o/oo for calcites, and about + 24 o/oo for dolomites; the $\partial^{13}C$ (PDB) ratios for the Amargosa basin deposits range from about 0 to -3.5 o/oo (Hay et al., 1986). Similarities of the $\partial^{18}O$ ratios suggest, that at Yucca Mountain and in the Amargosa basin, crystallization of calcites may have occurred at similar but relatively low temperatures. Hay et al., 1986, postulated, that in the case of Amargosa basin, the temperature of fluids in which the carbonates were precipitated, may have been in a range of 27 to 35°C. This estimate, together with one shown on Plate 4.5.3-13, indicates that the 15°C estimate of temperature of formation of the Yucca Mountain veins, as performed by O'Neil, 1984 and 1985, is doubtful.

The differences in the $\partial^{13}C$ ratios, between the Yucca Mountain calcites and the Amargosa basin calcites, ranging from 1.8 to 7 o/oo suggest that fluids responsible for the respective precipitations of calcites contained slightly different amounts of ^{13}C. Such differences in the ^{13}C contents, however, are not surprising and are to be expected. The Amargosa basin calcites were precipitated in fluids that discharged directly from the Paleozoic carbonates. Samples of such fluids yield the $\partial^{13}C$ ratios ranging from -2.2 to -5 o/oo (Plates 4.2.4.4-23, 4.3.3-54, and 4.5.2.5-10). Typically, the $\partial^{13}C$ ratios for fluids that could have been responsible for deposition of the Yucca Mountain veins, however, are lower. Such fluids, even if initially from the Paleozoic carbonates, must have passed through the Tertiary tuff "pile". As shown on Plates 4.3.3-54, 4.5.2.5-9, and 4.5.2.5-10, the $\partial^{13}C$ ratios for fluids from the tuff "pile" range from -5.6 to -14 o/oo, and from -3.4 to -6.8 o/oo for fluid samples representing the "mixed" carbonate-tuff systems.

In summary, both the mineralogic and the stable isotope data permit speculation that the Yucca Mountain veins and aprons may have been precipitated from either perched or deep-seated subsurface fluids. Locally, "pedogenic" appearances of the Yucca Mountain deposits may be appropriately attributed to "pedogenic" processes that were stimulated by ground surface dispositions of fluids discharging from

nearby faults and fissures. As far as the available data base is concerned, however, there are no indications whatever that the "pedogenic" processes were motivated exclusively by fluids directly from meteoric precipitation, whether in run-off or in infiltration modes. The very small areal extent of catchment areas, that possibly could have been involved, together with the very small rates of $CaCO_3$ and SiO_2 precipitations, as indicated by the presence of loose and uncemented Late Quaternary-Holocene deposits, suggest that attributing the formation of Yucca Mountain veins and aprons to fluids directly from meteoric precipitation is doubtful at best.

In the vicinity of Yucca Mountain, **lake beds, or most likely pond or seep deposits**, were described by Hoover et al., 1981. Such deposits occur near the southern end of Crater Flat at altitudes 120 and 150 m higher than altitudes of the contemporary water table. They consist of white to buff, laminated to thick-bedded, poorly indurated, sandy and tuffaceous marls, siltstones, and diatomites. Bedding characteristics and the presence of diatomaceous beds both indicate an environment of aqueous deposition. The diatoms identified in three samples indicate: "...alkaline, at times slightly saline, shallow water. Many of the species identified are found in alkaline, somewhat saline spring systems, but they are not restricted to this environment" (Hoover et al., 1981).

Near the southern end of Crater Flat, occurrences of **spring deposits** were also described by Hoover et al., 1981. Such deposits consist of pale brown, vuggy tuffa to white relatively dense siliceous calcium carbonates containing scattered large pores. They form mounds, diameter of which ranges from 5 to about 20 m. It appears that some of these deposits were dated by Szabo et al., 1981, using the uranium-thorium method. Samples #106 and #199 are of particular relevance. The location of both of these samples is shown on Plate 4.5.3-1. Sample #199 represents a "**nodular tuffa spring deposit**". It gave the uranium-thorium age of about 3×10^4 years B.P., indicating that "**suggested spring activity was at altitude of about 838 m as late as 30,000 years ago**. Present water table is about 120 m lower (Winograd and Thordarson, 1975)." Sample #106 represents "**seep deposited tuffa or calcrete intercalated in Q2 alluvium - shows some evidence of spring-water deposition.**" The sample yielded the uranium-thorium age of $7.8 \times 10^4 \pm 5 \times 10^3$ years B.P. It was collected at an altitude of about 940 m above mean sea level. This altitude is about 200 m higher than the altitude of the contemporary water table.

At Yucca Mountain, the veins composed of calcite and opaline silica are known to occur in the subsurface. Szabo and Kyser, 1985, reported the results of radiometric and stable isotope analyses performed

on samples of the calcite-silica veins from the vadose zone. The results of such analyses were considered in Section 4.3.4. The elevated temperature of formation of these veins suggests that their origin may be related to deep seated fluids, rather than to infiltrating meteoric water as assumed by Szabo and Kyser, 1985. Radiometric ages of the veins range from more than 4×10^5 years B.P. to 3×10^4 and 2.8×10^4 years B.P.

Veins that are most certainly related to deep-seated fluids occur in other cores extracted from the vadose zone at Yucca Mountain. Examples of such veins are a) borehole RF-3, at a depth ranging from 238 to 248 feet; b) borehole UE-25 A#4, at a depth ranging from 318 to 336 feet; and c) borehole UE-25 A#5, at a depth from 281 to 290 feet. The veins, however, were not investigated and, therefore, neither their age nor isotopic composition are known.

Carr et al., 1986, reported the occurrence of possibly metasomatic alteration of hanging wall of the large fault forming the contact between the tuff "pile" and the Paleozoic carbonates. Borehole UE-25p#1 intersected, at a depth of about 1,200 m and immediately above the fault, at least 33 m thick sequence of calcified ash-flow tuff. Locally, secondary calcite comprises more than 50 percent of the total volume of rock. Clearly, a long history of migration of carbonate-rich fluids from the nearby fault into the country rock is evident. Potentiometric data, together with the isotopic and geochemical characteristics of the local fluids, indicate that such migration is occurring at the present time (Section 4.3.3).

In summary, the geologic record representing Plio-Quaternary time seems to contain possible expressions of past large-scale fluctuations of the water table. This conclusion, if correct, together with the data and interpretations presented in Sections 4.2 and 4.3 form a harmonious picture of the Yucca Mountain groundwater system. Although it is rather difficult to judge the reliability of some information from geologic literature, the picture seems to be remarkably consistent. It seems reasonable to conclude, that the conceptual model of a heat-fluid coupled groundwater system in a deforming fractured medium constitutes a viable representation of the Death Valley and Yucca Mountain groundwater systems. Magnitudes of hydrologic disturbances, that seem to be possible in these systems, are very large, and measured in terms of hundreds of meters. The duration of a single evolutionary loop of the system appears to be less than 10^5 years.

SECTION 5.0

SUMMARY, CONCLUSIONS, AND RECOMMENDATIONS

5.1 SUMMARY AND CONCLUSIONS

Having performed the analyses presented in Sections 3.0 and 4.0, the following three main conclusions seem to be reasonable and warranted at this time:

i) **Examination of the extensive data base pertaining to the Yucca Mountain groundwater system, in light of the conceptual framework as established in Section 3.0, reveals that this system is considerably different from the flow system currently envisaged by the Yucca Mountain Project.** The conceptual understanding of the groundwater system, as used in performing suitability assessments for the purposes of developing the Final Environmental Assessment (DOE, 1986) and in establishing an approach to the forthcoming site characterization, is far too simple and much too far removed from reality. Simply stated, this conceptual understanding completely ignores the volcano-tectonic setting of the Yucca Mountain site.

According to the currently accepted conceptual understanding, the Yucca Mountain groundwater system is a "simple" system that may be represented by one of the commonly used mathematical models of groundwater flow (Section 2.2). The water table is regarded as a meaningful and durable expression of the hydraulic conductivity structure and boundary conditions. Position of the water table is assumed to be related solely to the volume of fluid passing through the system's vertical boundaries. Such a volume of fluid is a function of climatic conditions and remains fixed as long as these conditions remain unchanged. It is assumed that all moisture in the vadose zone is the result of infiltration of a small portion of local rainfall. Furthermore, it is assumed that movement of water in the vadose zone involves matrix flow, and that such movement occurs under the influence of gravity alone.

Within the context of currently accepted conceptual understanding, the influence of tectonic disruptions is limited to a) local alteration of litho-stratigraphic relationships; b) additional local fracturing; and c) minor changes in altitudes of discharge areas relative to altitudes of recharge areas. Magnitudes of tectonically induced changes in these factors are relatively small; consequently, the resulting alterations of the hydraulic conductivity structure and the boundary conditions are likewise small. Such alterations may

readily be accommodated by minor changes in position and configuration of the water table. It follows then, that the impact of such alterations on the long-term performance of a high-level nuclear waste repository is negligible.

A completely different picture emerges if one considers the volcano-tectonic setting of the Yucca Mountain Site and how it affects the local groundwater system. As indicated in Section 3.4.1, the area of Yucca Mountain contains continental alkalic basalts of Plio-Quaternary age. These basalts, as a matter of fact, are trace-element enriched hawaiites with isotopic compositions indicative of upper mantle origin. The isotopic and geochemical characteristics of the basalts are best accounted for by assuming that a convective system is developed in the upper mantle of this region. The presence of such a system may be inferred independently based on interpretations of various geophysical data (Sections 3.1 and 3.4.1). From the perspective of local and regional hydrology, the upper mantle convective system is responsible for two important tectono-physical factors. The first factor is high, and most importantly, spatially heterogenous flux of terrestrial heat. Terrestrial heat and gravitational hydraulic pressures acting at vertical boundaries of the groundwater system are the energy sources, that together, are responsible for the movement of groundwater. In the thermodynamic sense, groundwater flow is a heat-fluid coupled flow. For a groundwater system involving such flow, any plane adopted as the base must be considered as a "variable-upward-flux" boundary with respect to both heat and fluid. Mathematical models used in numerical simulations of such a flow system must account for this circumstance. Otherwise, calculations and interpretations may be grossly misleading, and with reference to the Yucca Mountain groundwater system, are quite irrelevant.

The second factor is strain energy, that because of the upper mantle convective system, is being supplied into the groundwater system on a continuous basis. The resulting changing *in situ* stress field plays an important role in controlling the resistance of a fractured medium against the coupled flow of heat and fluid. Near the ground surface, the *in situ* stress field controls the conducting apertures of fractures, and through that, it also controls the local effective thermal and hydraulic conductivity structures. For time periods of geologic proportions, the entire groundwater system behaves as a three way, H-F-M, coupled thermodynamic substance. For purposes of these conceptual considerations, this substance has been simplified and referred to as a heat-fluid coupled flow system in a deforming fractured medium. The substance evolves through a series of similar evolutionary loops, that for a given portion of the groundwater system, are closely tied to cycles of tectonic deformation (Section 3.3.2). During a single evolutionary

loop, energy content of the substance undergoes systematic transformations (Section 3.5.3). At the onset of a deformation cycle, the *in situ* stress gradients are fairly low with few aberrations. Such a state of *in situ* stress allows conducting apertures of fractures to be low. The resulting low hydraulic transmissivity and low effective thermal conductivity allow for the existence of high gradients of hydraulic potentials and *in situ* temperature. At that time, the water table is high and the flow system exhibits strong "mixed" convection aspects. As deformation progresses, "steeper" *in situ* stress gradients begin to emerge. In some areas around yielding fractures, the stress difference diminishes progressively and eventually produces "isotropic" and "singular" points. Conducting apertures of fractures increase in a time-dependent manner (Section 3.3.3.3). Such time-dependent increases cause the gradients of temperature and hydraulic potentials to "weaken." The water table deepens, and near the ground surface, the groundwater system looses its strong "mixed" convection or geothermal aspects. The termination of a deformational cycle is associated with a significant rearrangement of the local *in situ* stress field. The rearranged and "weakened" *in situ* stress field allows for reappearance of the normal "strong" gradients of *in situ* temperature and hydraulic potentials. The water table returns to its former position and the convective flow of heat and fluids returns to its normal intensity. At that time, the bedrock separations that accompanied the *in situ* stress field rearrangements are being filled with minerals precipitating from convecting fluids.

As demonstrated in Section 4.0, the assumption of a heat-fluid coupled system in a deforming fractured medium is in full accord with all known characteristics of the Yucca Mountain groundwater system. This is in sharp contrast to the conceptual understanding of this system as used to a) perform assessments of the suitability of Yucca Mountain for site characterization and b) develop strategies for demonstrating compliance of this site with all applicable Federal requirements. Not accounting for the true nature of the local groundwater system resulted in serious errors in judgment regarding anticipated long-term performance of the proposed high-level nuclear waste repository.

ii) **Conceptualization of hydrologic processes operating in the Yucca Mountain vadose zone yields a completely different picture than the one currently envisaged by the Yucca Mountain Project.** Two issues of fundamental importance are associated with the possibility that the Yucca Mountain groundwater system is developed in a deforming fractured medium (i.e., the mechanism of flow in the vadose zone and temporal stability of the water table, including its short- and long-term aspects, and involving both climatic and tectonic factors). Two other issues of nearly equal importance are associated with the possibility that the Yucca Mountain groundwater system involves a heat-fluid coupled

flow (i.e., chemistry of interstitial water in the vadose zone and spatial and temporal variability of flux of terrestrial heat through the vadose zone).

A relative measure of likelihood of **fracture flow in the vadose zone** is depth extent of the "limit equilibrium" stress conditions (Section 3.3.5). The results of *in situ* stress determinations, made at Yucca Mountain, revealed that the depth extent of such conditions is considerable, in a range of 1,200 to 1,500 m (Section 4.2.5.4). This indicates, that in the Yucca Mountain vadose zone, conducting apertures of fractures are very large. It is likely that such apertures may support local fracture flow, despite a low annual average rate of infiltration of rainfall. Furthermore, in the Yucca Mountain vadose zone, the state of *in situ* stress is such that numerous "singular" points are present. Around such points the mean stress and the stress difference are very small and, consequently, any build-up of fluid pressure causes further enlargement of local conducting apertures. It is unlikely that stressed in such a manner fractured medium may retain fluids for a period of time that is required for introduction of these fluids into rock matrix. Both of the above lines of reasoning suggest that the expectation of solely interstitial flow in the Yucca Mountain vadose zone lacks proper foundation. This is true, in particular, if one is concerned with movements of water through rocks with either high saturation or low interstitial porosity, as is the case for the tuffaceous beds of Calico Hills and for the densely welded Topopah Spring Member of the Paintbrush Tuff, respectively.

In a deforming fractured medium, temporal stability of the water table is not related solely to global or regional climatic fluctuations. Such stability is also related to tectonic factors, specifically the local *in situ* stress field. As discussed in Section 3.3.5, both short-term and long-term instabilities of the water table may be expected at the Yucca Mountain site. **The short-term instabilities** involve several meters, perhaps tens of meters, of displacement of the water table. They are short-lived, say days or weeks at most, and occur with a frequency related to a degree of development of the local *in situ* stress field. Such short-lived instabilities are caused by continuous but minor restructurings of the local *in situ* stress field (Sections 3.3.3.3, 3.3.4, 3.3.5, and 3.3.6). They may occur spontaneously, or they may require an external triggering mechanism. Fault creep, vibratory ground motion induced slip, local and temporary volumetric strain, are examples of minor restructurings of the local stress field. As discussed in Sections 4.1 and 4.2.4.4, the results of continuous monitoring of the water table, performed at Yucca Mountain in a number of deep wells during the last six years, are not yet available in an accessible and reliable form. Consequently, it is not known with certainty whether or not the short-lived instabilities of the water table are

currently occurring at the Yucca Mountain site. Such instabilities, however, are of major diagnostic importance. They may confirm directly that the Yucca Mountain groundwater system, indeed, operates in a deforming fractured medium. Also, the short-lived instabilities, if occurring with a meaningful frequency and magnitude, constitute a "pumping" mechanism for gaseous transport through the vadose zone.

The long-term displacements of the water table are caused by large-scale restructurings of the local *in situ* stress field. Such restructurings occur at the end of tectonic cycles and, at Yucca Mountain, with a periodicity measured in terms of tens of thousands of years. A magnitude of the resulting potential displacements of the water table is a sum of three components: a) rise of the water table caused by "overpressure" Δp_{max} (Section 3.3.4 and Plate 3.3.5-1); b) rise of the water table resulting from reduction of storativity (Section 3.3.5 and Plate 3.3.5-1); and c) rise of the water table resulting from tectonically induced restructuring of the local hydrothermal system (Sections 3.3.4 and 4.3.3; Plate 3.4.4-6). The existing hydrologic and geothermal data indicate

1) The greatest magnitude of "overpressure," that is known to exist at Yucca Mountain, is $\Delta p =$ 55 m (Plates 4.2.4.4-12 and 4.2.4.4-17a). The total magnitude of Δp_{max}, however, is unknown. It is possible, nevertheless, that this magnitude is approximately equal to the difference in altitudes of the water table in northern and southern sectors of Yucca Mountain. As shown on Plate 4.2.3-5, this difference is equal to at least 300 m;

2) At Yucca Mountain, potential build-ups of fluid pressure that are caused by fluids released from storage may be very large, and measured in terms of tens of bars. Such a state of affairs is suggested by two lines of evidence. First, the ratio S^{σ}/S^{in} appears to be surprisingly large, within a range of 10^2 to 10^4 (Section 4.2.6 and Plate 3.3.5-1). Second, below the zone of *in situ* stress enhancement of the local hydraulic conductivity structure, the average value of the lateral hydraulic gradient may be within a range of 10 to 20 m $\cdot$ km^{-1} (Plate 4.2.3-2). In this situation, even for a relatively small zone of the restructured *in situ* stress field, the value of fluid pressure build-up that is required to "bleed-off" the volume of fluid released from storage is very large indeed; and

3) Potential rise of the water table resulting from a restructuring of the deep Yucca Mountain hydrothermal system may be significant and measured in terms of, at least, tens of meters. The following two lines of reasoning are offered in support of this statement. First, the geochemical and

isotopic characteristics of fluids encountered around the Solitario Canyon fault, specifically the $\partial^{13}C$ ratios and the calcite saturation index, may be interpreted as indicative of local upwelling of fluids from the Paleozoic carbonates (Section 4.3.3). Assuming that such upwelling is the result of thermally driven convection, it follows that the depth extent of the local hydrothermal system is equal to at least a local depth of occurrence of the Paleozoic carbonates. Based on interpretation of the results of magnetic and gravity surveys, depth of occurrence of the Paleozoic carbonates was estimated to be ranging from 3.4 km, at the northern end of Yucca Mountain, to 4.0 km in Crater Flat (Carr et al., 1986). Second, the results of isotopic analyses of vein mineralization from the Yucca Mountain vadose zone suggest that contemporary values of the vertical geothermal gradients are some 15°C/km lower than those in the not too distant past (Section 4.3.4). Assuming that the discrepancy is, at least partially, related to past restructurings of the Yucca Mountain hydrothermal system then it follows, that the expected average change in water temperature may be in a range of few tens of degrees Celsius, say 20 to 30°C. In this situation, the potential "bouyant" rise of the water table alone is in a range of 32 to 48 m (Section 3.4.4).

The above crude estimates indicate, that at Yucca Mountain, magnitudes of the long-term displacements of the water table involve several tens and perhaps a few hundred meters. Displacements of the water table of such magnitudes are likely to profoundly impact the long-term performance of the proposed high-level nuclear waste repository. Also, in extreme cases such displacements may result in flooding of the repository and expulsion of groundwater at the land surface. A critical point in the overall conceptual considerations of the Yucca Mountain groundwater system is to establish unequivocally whether or not the geologic record representing Plio-Quaternary time contains expressions of the past large-scale displacements of the water table. As discussed in Section 4.5.3, there are indications that this may be the case indeed. Thus far, however, only superficial studies of the calcite-opaline silica vein deposits were performed during the last 10 years. Consequently, the current data base does not allow for an unequivocal resolution of this critical issue. Nevertheless, the possibility that the veins constitute a record of large-scale displacements of the water table in the past is real and, by all means, not remote.

The assumption that the Yucca Mountain groundwater system is developed in a deforming fractured medium and, that it involves a heat-fluid coupled flow, leads to another important issue: **the origin and chemistry of interstitial waters in the Yucca Mountain vadose zone.** If interpretations performed in this report are correct then it follows that interstitial waters represent, at least in part, relict hydraulic

mounds developed in association with high groundwater stands at the end of tectonic cycles. Chemistry of such waters should be different from that of "ordinary" meteoric waters infiltrating through the vadose zone. At Yucca Mountain, downward movement of infiltrating meteoric waters may involve interstitial flow, but only during early stages of deformation, when intensity of the local *in situ* stress field is low and conducting apertures of fractures consist of only residual components. Because of the high water level stand, however, it is unlikely that significant replacements of interstitial water, by infiltrating meteoric water, could occur at that time. Such replacements may be significant, but only during more advanced stages of deformation, when the water table is deeper. At that time, however, the intensity of *in situ* stress field is substantially increased and, consequently, conducting apertures are enlarged, so that infiltration of meteoric water occurs as fracture flow. As far as it could be determined, there are no published data concerning chemistry of interstitial water from the Yucca Mountain vadose zone. Consequently, there is no basis to either confirm or reject the hypothesis that interstitial waters are relict waters that originated from below the water table. Nevertheless, there are some unpublished data that seem to indicate that chemistry of interstitial waters is different from that of waters residing in fractures. Such data were obtained by Dr. A. Yang (USGS, Denver, CO) through chemical analyses of samples of interstitial waters that were mechanically squeezed out of cores extracted from the Yucca Mountain vadose zone.

The conceptual understanding of the Yucca Mountain groundwater system, as developed in this report, indicates that **flux of terrestrial heat through the vadose zone** lacks spatial uniformity and temporal consistency. Similar conclusions may be drawn based on the Yucca Mountain geothermal data. The downhole temperature profiles indicate that the local geothermal field contains very substantial lateral geothermal gradients (Section 4.3.3). The results of isotopic analyses of vein mineralization from the vadose zone suggests that potentials of the local geothermal field may be time-dependent (Section 4.3.4). Both of the above variabilities are caused by tectonic factors that control the local effective thermal conductivity structure and the local thermal boundary conditions. Clearly, the assumption of "const. flux" thermal boundary conditions, for the base of the vadose zone, cannot be justified. Moreover, the local intensity of flux of terrestrial heat is substantial, and may be within a range of 70 to 100 mWm^{-1}. In this situation, terrestrial heat is a factor that needs to be reckoned with in considering movements of fluids, vapors, and gases in the vadose zone.

iii) **The conceptual understanding of the Yucca Mountain groundwater system, as developed in this report, points out serious limitations of the Yucca Mountain site to effectively**

isolate radionuclides from the biosphere. Such limitations are greater by far than those currently envisaged by the Yucca Mountain Project. Without recognizing these limitations, the issue resolution strategies and performance allocations, as expressed in the Statutory Site Characterization Plan, are interesting propositions, but in terms of the actual site conditions are of limited practical value. The existing data indicate that it is very probable that full implementation of the proposed study plans will not yield data that are required for a successful completion of the licensing process. On the contrary, implementation of the proposed study plans will undoubtedly yield data that will bring limitations of the Yucca Mountain site into a full light. At that time, substantial changes in the site characterization "philosophy" will be required. Furthermore, not relying on timely and reasonable interpretations of the current data base, the Yucca Mountain Project failed to develop a realistic comprehension of the local hydrologic conditions. The resulting misunderstanding of hydrology of the Yucca Mountain vadose zone resulted in overly optimistic assessments of the suitability and licenseability of the Yucca Mountain site.

The most important licensing concern is **a potential rise of the water table.** If such a tectonically induced rise is possible at Yucca Mountain, then the geochronologic data indicate that the annual probability of occurrence of this rise is more than 10^{-5}, perhaps within a range 3 to 5 x 10^{-4} (Sections 4.3.4 and 4.5.3). It is likely, that in the context of long-term performance requirements as set forth in 10 CFR 60 and 40 CFR 191, the water table rise constitutes an "anticipated process and event." As such, the water table rise must be acounted for in demonstrating compliance of the Yucca Mountain disposal system with three long-term performance objectives of 10 CFR 60. Such performance objectives are a) the life of the waste package; b) the release rates of radionuclides from the "engineered barrier system"; and c) the overall releases of radionuclides into the "accessible environment." The water table rise has a potential of significantly altering radionuclide migration path, rate, and time. An occurrence of the water table rise during early stages in the life of the repository, when temperature of waste packages and the surrounding host rock is above the boiling point of water, entails a particularly profound impact on the overall repository performance. Vaporization of fluids entering the repository would cause substantial increases in the vapor pressure and, therefore, would accelerate gaseous transport from the repository to the ground surface. Subsequent cooling of the repository, to below the boiling point of water, is likely to be accompanied by a long-term convective flow of fluids from the repository to the ground surface.

It is expected, that chemistry of fluids entering the repository in association with the tectonic rise of the water table, is different from that of fluids currently residing in fractures. The impact of tectonic rise

on the life of waste packages, therefore, is two-fold. Corrosion of waste packages will be accelerated due to a) increased amounts of water contacting the packages and b) more adverse chemistry of this water.

A tectonically induced rise of the water table, with magnitudes sufficient to flood the repository, is not the only licensing concern that results from the proposed understanding of the Yucca Mountain groundwater system. This understanding brings into sharp focus serious reservations with regard to compliance of the Yucca Mountain site with **the groundwater travel-time** requirement as set forth in 10 CFR 60. Compliance of the Yucca Mountain site with that requirement is assured, but only if meteoric water infiltrates through the vadose zone via mechanism of interstitial flow. In the case of fracture flow, if infiltration occurs at all, the travel-time through the vadose zone is, most certainly, substantially less than the required 10^3 years.

A gaseous transport through the vadose zone is another licensing concern resulting from the proposed understanding of the Yucca Mountain groundwater system. Such a transport is caused by the short-lived instabilities of the water table and by the heterogeneous flux of terrestrial heat into the vadose zone. The gaseous transport may have some impact on the compliance of the Yucca Mountain site with two performance objectives of 10 CFR 60. Such performance objectives are a) the release rate of radionuclides from the "engineered barrier system" and b) the overall releases of radionuclides into the "accessible environment."

Finally, the proposed conceptual understanding of the Yucca Mountain groundwater system suggests that **chemistry of interstitial waters in the vadose zone** may be substantially different from that of fracture waters. Such differences may be important with regard to considerations of regulatory compliance of the Yucca Mountain site. Three long-term performance objectives of 10 CFR 60 may be impacted. Such performance objectives are a) the life of waste packages; b) the release rates of radionuclides from the "engineered barrier system"; and c) the overall radionuclide releases into the "accessible environment."

Most of the above licensing concerns are not new concerns. They were raised previously, in one form or another by various parties, most notably by the U.S. Nuclear Regulatory Commission and the State of Nevada. The proposed conceptual understanding of the Yucca Mountain groundwater system reinforces those concerns and provides a uniform theoretical background for them.

5.2 RECOMMENDATIONS

The proposed conceptual understanding of the Yucca Mountain groundwater system poses a number of serious questions regarding the suitability of the Yucca Mountain site to safely accommodate a high-level nuclear waste repository. Recognizing the importance of such questions, and keeping in mind the long-established policy of the DOE to always address the critical suitability issues first, the following eight recommendations are offered:

1) A thorough examination and reasonable interpretation **of the data collected during the last six years of continuous monitoring** of the Yucca Mountain water table should be performed without any further delays. A main purpose of this task is to establish unequivocally whether or not a) short-lived instabilities of the water table are present and b) response of the local water table to externally generated vibratory ground motions is abnormal. Furthermore, it is important to determine a frequency of occurrence of the short-lived instabilities as well as their magnitude;

2) A comprehensive investigation of **the state of Yucca Mountain** *in situ* **stress field** should be undertaken. Such an investigation may be performed utilizing the presently existing network of boreholes, and employing both the hydraulic fracturing technique and the Cooper-Brederhoeft injection tests. A main purpose of this task is three-fold: a) reliable and unequivocal detection of "singular" points; b) confirmation that the *in situ* values of closure pressure are abnormally low; and c) fine definition of spatial distribution of the *in situ* values of closure pressure. The measurements should be made using small values of the applied hydraulic pressure and employing various diameters of injection tubing;

3) **A deep exploratory borehole** should be drilled in the western sector of Yucca Mountain. The main purpose of this borehole is to gain further understanding of the depthward distribution of hydraulic potentials in this area. Specifically, it is of critical importance to determine a magnitude of the Yucca Mountain "overpressure." The borehole should be drilled to a depth that is sufficient to penetrate the Paleozoic carbonates for a distance of a few hundred meters. Below a depth of about 1,500 m, a series of drill stem tests should be made. After completion of the borehole, a few sets of packers and piezometers should be installed, so that the downhole measurements of hydraulic potentials could be made at discreet depth intervals;

4) A complete and conclusive investigation of the **calcite-opaline silica-sepiolite vein deposits** from Yucca Mountain should be undertaken without any further delays. A purpose of this task is to establish unequivocally whether or not the geologic record contains "foot prints" of past large-scale fluctuations of the water table. Specifically, a firm position must be established whether or not flooding of the repository, within the context of 10 CFR 60 and 40 CFR 191, constitutes an "anticipated process and event." The investigation should be carried out in both the surface and subsurface, and should include radiometric age determinations;

5) A thorough study of **the Yucca Mountain "mosaic" breccias** should be undertaken without further delays. A main purpose of this task is to establish whether or not a) the "mosaic" breccias record a substantial volumetric strain of the Yucca Mountain area, in a range of several percent and b) the "mosaic" breccias are younger than solidification of the country rock, polygenetic, and representing time spans measured in terms of 10^6 years. Specifically, a firm position must be established whether "widening" of Yucca Mountain is an ongoing and cyclic process. The studies should include limited trenching, high resolution mapping in selected areas, and petrographic analyses of selected samples to determine petrogenesis involved in formation of the "mosaic" breccias;

6) An investigation of **chemistry of fluids residing in the Yucca Mountain vadose zone** should be performed. A main purpose of this task is to establish whether local interstitial waters are chemically and isotopically different from fracture waters. The investigation should consist of chemical and isotopic analyses of samples of interstitial waters that were mechanically squeezed out from cores following the procedure used by Dr. A. Yang of the USGS. The core samples will be extracted from a few boreholes that should be drilled for that purpose;

7) An investigation **of the known bodies of perched and semi-perched water**, that occur within the vadose zone of the Death Valley groundwater system, should be performed. A main purpose of this task is to evaluate the possibility that some of these bodies represent tectonically induced hydraulic mounds, either remnant or active. There are three outstanding candidates for this study: a) the apparent hydraulic mound underneath Greenwater Range; b) the hydrologic anomaly in the vicinity of

Skull Mountain; and c) the apparent hydraulic mound between Climax Stock and Emigrant Valley. The investigation should be sufficiently broad in scope to permit a definition of lateral and vertical gradients of a) hydraulic potentials; b) *in situ* temperature; and c) isotopic and chemical composition of groundwater; and

8) An investigation of the **Whamonie gypsum paleo-spring mound** should be performed. A main purpose of this task is to establish age of activity of the Whamonie paleo-spring discharge. Specifically, a firm position must be established whether or not the apparent upwelling of deep, sulfate rich, waters is of Late-Quaternary age. The investigation should include high resolution mapping, limited trenching, petrographic analyses of selected samples, and radiometric age determinations.

The recommended investigations were designed to obtain data that are judged to be required for validating the proposed herein conceptual understanding of the Yucca Mountain groundwater system. The presence of a) short-lived instabilities of the water table and abnormal responses of the water table to externally generated vibratory ground motions; b) "singular" points and abnormally low values of the closure pressure; c) sizable "overpressure"; d) "foot prints" of past large-scale fluctuations of the water table during Quaternary time; e) post-depositional and polygenetic "mosaic" breccias; f) meaningful differences in chemical compositions of interstitial and fracture waters; g) perched and semi-perched waters resulting from injection of fluids from below the water table; and h) ground surface discharges of warm, sulfate rich, fluids during Late-Quaternary time, would confirm that the proposed understanding of the Yucca Mountain groundwater system is correct. **In this situation, and within the context of current Federal regulations, serious considerations should be given to abandoning the Yucca Mountain site and declaring it as unlicenseable for the purposes of permanent disposal of high-level nuclear waste. At a minimum, a fundamental revision of the strategies for demonstrating regulatory compliance of the Yucca Mountain site, as developed in the Statutory Site Characterization Plan, will be required. In all sincerity, the DOE would be well advised to perform the recommended investigations in a truly expeditious manner, preferably prior to the commitment of substantial resources such as those associated with *in situ* testing in the exploratory shaft and with the implementation of currently proposed study plans. After all, it is an American proverb "if in doubt, don't."**

closure and post-closure. Directly, tectonic processes govern compliance of a site with tectonic guidelines. Indirectly, other guidelines (i.e., rock characteristics, geohydrology, and erosion) are also involved.

A main purpose of this report is to offer a proposal regarding a relationship between the tectonics of the Yucca Mountain site region, the groundwater system operating at this site, and in the region surrounding it. The intention is to provoke initiation of an understanding, on a qualitative or conceptual basis, of this groundwater system. Such an understanding could then be used to a) develop a focused and effective exploration program; b) perform, if necessary, a re-evaluation of the existing data; c) develop a description of the natural system; and d) develop a rational approach to the performance assessment and the performance allocation process.

This report has been divided into four parts: Section 2.0, Mathematical and Conceptual Models of Groundwater Systems - Statement of the Problem; Section 3.0, Conceptual Framework for the Yucca Mountain Groundwater System; Section 4.0, Characteristics of the Death Valley Groundwater System in Light of the Existing Data; and Section 5.0, Conclusions and Recommendations.