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EXECUTIVE SUMMARY

PHYSICAL PROCESSES OF MAGMATISM AND EFFECTS ON THE POTENTIAL REPOSITORY: SYNTHESIS OF TECHNICAL WORK THROUGH FISCAL YEAR 95

Principal Investigator: G.A. Valentine

This chapter summarizes data collection and model calculations through FY 95 under Study Plan 8.3.1.8.1.2 Physical Processes of Magmatism and Effects on the Potential Repository. The focus of this study plan is to gather information that ultimately constrains the consequences of small-volume, basaltic magmatic activity at or near a potential repository. This is then combined with event probability estimates, described elsewhere in this synthesis report, to yield a magmatic risk assessment.

There are two basic classes of effects of magmatism that are considered here: (1) *Eruptive effects*, whereby rising magma intersects a potential repository, entrains radioactive waste, and erupts it onto the earth's surface. (2) *Subsurface effects*, which includes a wide range of processes such as hydrothermal flow, alteration of mineral assemblages in the potential repository system, and alteration of hydrologic flow properties of the rocks surrounding a potential repository.

- Eruptive (direct) effects - We present data on the possible entrainment of waste into eruptive magma. These data were collected from analog volcanoes and represent a range of eruption mechanisms including Strombolian, Hawaiian, and hydrovolcanic. Strombolian and Hawaiian eruptions generally entrain little wall rock material, and are unlikely to have a strong influence on total system performance assessments. Hydrovolcanic eruptions entrain 10^3 - 10^4 times more wall rock debris and could influence performance assessments, depending mainly on the probability of such eruptions.
- Subsurface (indirect) effects - We present field data from analog sites where the plumbing of small basaltic volcanoes is exposed, and where the volcanoes penetrated silicic tuffs similar to those at Yucca Mountain. Geometry of shallow intrusions is determined by a complicated interplay between pre-existing structures, shallow stress field variations, and rock mechanical properties. The intrusion geometry in turn determines the alteration of the host rocks. Chemical and mineralogical studies of the host tuffs indicated that for shallow, small-volume basaltic intrusions alteration is limited to within a few tens of meters of the intrusion itself. There is no evidence for extensive, long-lived hydrothermal systems. Numerical modeling studies suggest that the farthest reaching "hydrothermal" process in the vadose zone, which is convection of pore air, is limited to within 2.5 km of intrusions of relevant sizes.
- Magma system dynamics - examination of processes of magmatic activity from melt generation through storage, transport, and eruption are important for providing some "deterministic" constraints on probability calculations. The low magma flux dynamics of the Yucca Mountain region are not well understood -- we review what is known and recommend future directions.

PHYSICAL PROCESSES OF MAGMATISM AND EFFECTS ON THE POTENTIAL REPOSITORY: SYNTHESIS OF TECHNICAL WORK THROUGH FISCAL YEAR 95

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5.1. Introduction

This chapter is a revision of Chapter 5 of Crowe et al. (1995). Here we synthesize work carried out as part of Study Plan 8.3.1.8.1.2 Physical Processes of Magmatism and Effects on the Potential Repository. The plan was not carried to completion, therefore this chapter should not be viewed as a final report on the study plan. Rather it is a report on results obtained primarily in FY 93-95, about one half of the originally planned work.

The chapter consists of three main parts, corresponding to the three activities of the original study plan: "Eruptive Effects," "Subsurface Effects," and "Magma System Dynamics." The material reported for Eruptive Effects is the result of field studies at analog volcanoes where, due to their geologic setting, the quantity of debris entrained by rising magma could be constrained as a function of depth below the volcanoes. This provides critical information for determining how much radioactive waste could be entrained into magma and erupted onto the surface were a volcano to form directly through the potential Yucca Mountain repository. The data provide bounds on the quantity of debris that can be entrained into rising magma per meter along the volcanic plumbing, for a range of eruption mechanisms including Strombolian, hydrovolcanic, and effusive. These data can be used directly in performance assessment calculations of waste erupted onto the earth's surface, given a volcanic event that intersects a potential repository. However, the implications of these data on performance of the potential repository cannot be fully stated until they have been used in rigorous, total performance assessments.

A more difficult topic is the effects of magma intrusions (whether accompanied by surface eruption or not) on potential repository performance, either via mechanical or chemical/mineralogical alteration of the potential repository system, or by hydrothermal flow. This topic comprises the Subsurface Effects portion of the study plan. Here we report results of field research at two sites where basaltic magmas intruded silicic tuffs at shallow depths. These results constrain the compositional and mineralogical effects of the intrusions on tuff wall rocks, and factors governing the geometry of the intrusions. The implications of these data for performance of the potential repository cannot be quantitatively assessed until they have been incorporated into rigorous total system performance assessments. However, it appears that for the size and shape of intrusions that are relevant for the Yucca Mountain setting, alteration of the host rock is limited to at most a few tens of meters distance from the intrusions themselves. We also report some of the first stages of a planned extensive modeling study aimed at predicting such processes for a variety of intrusion sizes and shapes. Most of the theory and modeling results are of such a preliminary nature that they cannot be used in total system performance assessments. However, preliminary calculations suggest that unless an intrusion is within about 2.5 km of the potential repository, it will not perturb the unsaturated zone of the repository system.

The Magma System Dynamics part of the study plan dealt with the physics of magmatism and the potentially unique processes driving volcanism in the Yucca Mountain region. An ultimate aim of this part was to provide some deterministic predictions to supplement the probabilistic risk approach. In this chapter we present a review of the current state of related literature.

5.2. Eruptive Effects

5.2.1. Summary of Eruptive Effects Approach and Results

As magma rises through the lithosphere it may entrain wall rock debris, and, if the magma penetrates a potential radioactive waste repository, some waste material could be entrained as well. Here we describe a suite of field studies aimed at quantifying this entrainment process in a manner that can eventually be used in performance assessment. The entrainment process depends on the local hydrodynamic regime of the magma (e.g., velocity, temperature, bulk density), the extent of interaction of magma with groundwater, and the mechanical properties of the wall rocks. Wall rock entrainment results in local flaring of dikes and conduits, which in turn affects the hydrodynamics of magma ascent and eruption. We studied upper-crustal xenoliths erupted from small-volume basaltic volcanoes of the Lucero volcanic field (west-central New Mexico; Valentine and Groves, 1996) and of the San Francisco volcanic field (north-central Arizona) in order to assess the relative importance of various entrainment mechanisms during a range of eruptive styles including strongly hydrovolcanic, Strombolian, and effusive processes. Total xenolith volume fractions range from 0.3-0.9 in hydrovolcanic facies to $<10^{-4}$ - 10^{-2} in most Strombolian facies. The volcanoes erupted through thick, well-characterized sequences of Paleozoic and lower Mesozoic sedimentary rocks, so that in some cases erupted xenoliths can be correlated with sedimentary units and hence depth ranges. The abundance of xenoliths from a given subvolcanic unit is divided by that unit's thickness to obtain an average entrainment rate (xenolith volume fraction derived per unit depth in the conduit). Shallow (less than a few hundred m) entrainment rates are very sensitive to the degree of hydrovolcanic activity. Deeper entrainment is more sensitive to the mechanical properties of the wall rocks, and in the cases studied here is thought to depend mainly on brittle failure related to offshoot dikes, pore pressure buildup, and thermal stresses. The entrainment rate can be used as a source term in multiphase numerical models of conduit flow. Based on the data presented here, theoretical models of conduit flow and erosion are justified in neglecting the contribution of mass and momentum from entrained material during basaltic eruptions driven by magmatic volatiles, but not in eruptions driven by hydrovolcanic processes.

5.2.2. Introduction and Previous Work

The ascent and eruption of basaltic magmas through the upper lithosphere are governed by the initial volatile content and rheological properties of the magma, the pressure at the magma source, interaction between magma and external water, the solid mechanics of dike or conduit opening, and the geometry of the dike or conduit through which the magma travels. Numerous authors have studied the source pressure and the behavior of volatiles in ascending basaltic magmas and their role in accelerating the mixture (e.g., Szekely and Reitan 1971; McGetchin and Ullrich 1973; Wilson and Head 1981; Spera 1984; Vergnolle and Jaupart 1986). Similarly, much research has been done on the physical processes associated with magma-water interaction (e.g., Wohletz and McQueen 1984; Wohletz 1986; Zimanowski et al. 1991; Dobran and Papale 1993). The coupled solid and fluid mechanics of dike propagation, which is the dominant form of basaltic magma transport through the lithosphere, has been addressed by numerous workers, as reviewed by Rubin (1993) and Lister and Kerr (1991). Delaney and Pollard (1981) and Bruce and Huppert (1990) addressed the transition from flow through dikes to flow through conduits that occurs in many basaltic eruptions (e.g., Richter et al. 1970) and is also evidenced in exposed neck and dike complexes. Many workers have addressed the heat transfer between rising basaltic magmas and their wall rocks (e.g., Fedotov 1978, 1981; Delaney and Pollard 1982; Spence and Turcotte 1985; Fabre et al. 1989; Bruce and Huppert 1990; Carrigan et al. 1992).

An outstanding problem in our understanding of basaltic eruption dynamics has been the coupling between magma hydrodynamics and dike/conduit geometry at shallow depths. Conduit geometry partly determines the pressure and velocity distribution of the magma and hence the

eruption dynamics and resulting facies, but the pressure and velocity also affect conduit geometry through various entrainment or erosion processes (for the remainder of the paper, unless otherwise specified, the term *conduit* is used to refer to the plumbing for a volcanic eruption in general, whether it is a dike or circular pipe). To date there have been two end member approaches to the ascent of magma at shallow depths. One assumes that the conduit has a constant width (or diameter), in which case for a given mass flux the pressure and velocity distribution can be found (e.g., Dobran 1992) and implies that the wall rocks are perfectly rigid. The second assumes that the conduit walls are very weak so that any local change in flow pressure relative to the lithostatic pressure in the country rock is immediately accommodated by wall failure. In this case the pressure in the rising magma is always equal to the lithostatic pressure, and for a given mass flux the velocity and conduit geometry can be calculated. This second approach has been the most commonly used one for determining eruption conditions of both basaltic and silicic eruptions (Wilson et al. 1980; Wilson and Head 1981).

In reality wall rocks are neither perfectly rigid nor perfectly weak, but a more realistic theoretical treatment is very difficult and poorly constrained by a lack of knowledge of the various mechanisms by which wall rock can be eroded or entrained off conduit walls. Macedonio et al. (1994) formulated a theory that accounts for three erosion mechanisms: abrasion from pyroclasts impacting conduit walls, erosion due to shear stress at the wall, and conduit wall collapse where the magma pressure is sufficiently different from the lithostatic pressure. They used the two-phase hydrodynamic model of Dobran and Papale (1992) to compute conditions along the conduit and then used these conditions to calculate the contributions of the three erosion mechanisms. The next step in developing volcanic conduit models is to fully couple the hydrodynamic calculation with the resulting erosion so that the full feedback mechanism can be understood. However, as Macedonio et al. (1994) point out: "A much more comprehensive study (than existing field studies) of the origin of lithics within the conduit and their relationship with the eruption dynamics is badly needed for future progress in quantifying erosion processes" (parenthetical statement added).

The following entrainment mechanisms have also been identified: (1) Spalling of rock into the low-pressure cavity at the tip of a propagating dike (e.g., Rubin 1993; Lister and Kerr 1991). Variations in proportions of wall-rock types would be expected in the erupted xenolith assemblage due to the variations in mechanical properties. However, this mechanism is highly transitory and would only be reflected in early erupted materials. (2) Entrainment of xenoliths at levels in a conduit where explosive magma-water interaction fragments the wall rocks (e.g., Self et al. 1980; Wohletz and McQueen 1984; Wohletz 1986; Lorenz 1986; Houghton and Schmincke 1986; Barberi et al. 1988; Heiken et al. 1988; White 1991; Zimanowski et al. 1991; Godchaux et al. 1992). (3) Entrainment due to shear stresses exerted on the wall by the flowing magma. Macedonio et al. (1994) treated the case where roughness elements extend from the conduit wall into the flowing magma stream. Another type of shear erosion, which is most applicable where the wall rocks are unconsolidated sediments or pyroclastic material, is grain-by-grain entrainment. (4) Isolation of country rock blocks by small offshoot dikes that extend from main dikes. Such offshoot dikes can propagate some distance parallel to the main dike, and at some point may rejoin the main dike. In this case, the wall rock between the main and offshoot dikes is then included into the flowing magma. (5) Cracking by thermal stresses (McBirney 1959) and pore pressure buildup (Delaney 1982) and entrainment of fragments into the magma. (6) Interaction between rising magma and wet sediments, which can be a special case of (2) and (3) above (e.g., Kokelaar 1982; Busby-Spera and White 1987; Leat and Thompson 1988; White 1991). In such interactions the entrainment process is essentially that of two viscous fluids, variably complicated by vaporization of water in the wet sediments and quenching of parcels of magma.

5.2.3. Analog Studies in the Lucero Volcanic Field, New Mexico

5.2.3.1. Introduction

In this section we present data pertaining to wall-rock entrainment from two small basaltic volcanoes in the Lucero volcanic field, New Mexico, that exhibit a range of eruptive mechanisms from hydromagmatic to Strombolian and effusive. Because of the well-constrained stratigraphy of sedimentary rocks below the volcanoes, we are able to quantitatively estimate the fraction of debris eroded per meter depth down the conduits. From this information we can assess to some degree the relative importance of various wall rock entrainment mechanisms for different rock types and for different eruptive mechanisms.

5.2.3.2. Geologic Setting

The Lucero volcanic field is a region of isolated basaltic cones, lava flows, flow-capped mesas, buttes (representing highly eroded volcanic centers), and associated shallow intrusive bodies that straddles the eastern edge of the tectonically stable Colorado Plateau and its transition to the Rio Grande rift (Figure 1). The volcanic field covers about 2000 km² and is named after the adjacent Sierra Lucero, a monoclinal uplift. The field was defined by Baldridge et al. (1987) after being previously mapped and described in fragments for mineral resource evaluations that focused mainly on the subvolcanic sedimentary rocks (Jicha 1958; Wengert 1958; Moench 1964; Zilinski 1976; Machette 1978; Osburn 1982, 1984). Baldridge et al. (1987) established the chronological and geochemical trends of the field, which includes a range of basaltic compositions from basanites to alkali-olivine basalts, tholeiites, and evolved alkali basalts. Eruptions took place during three phases of activity: 8.3-6.2, 4.3-3.3, and 1.1-0 m.y.a.

We chose two volcanic centers, Alkali Buttes and Volcano Hill (Figure 1), because of their wide range of eruptive facies and because a combination of field and petroleum well-log data constrains the country rock stratigraphy below them. Baldridge et al. (1987) report a K-Ar age of 0.8 ± 0.1 m.y. for Volcano Hill, and interpret Alkali Buttes to be roughly the same age based on regional geomorphic relations. The erupted magmas were nepheline-normative, alkali-olivine basalts (Baldridge et al. 1987).

The Colorado Plateau is an excellent setting for the detailed study of xenolith derivation as a function of depth because of its well established, "layer cake" stratigraphy. Although the eruptive centers we describe here are only about 30 km from the eastern edge of the Plateau, the rocks beneath them are relatively unaffected by faulting. In addition to well-known stratigraphy, the various sedimentary units below the volcanoes are, for the most part, easily distinguished in hand sample. Stratigraphic descriptions and thicknesses were obtained mainly from Jicha's (1958) and Zilinski's (1976) reports; most of their data were obtained directly from exposures in the Sierra Lucero uplift. These thicknesses were confirmed in the vicinity of the Alkali Buttes and Volcano Hill from logs of two petroleum wells (Figure 1). The logs were obtained from the New Mexico Bureau of Mines and Mineral Resources: (1) Spanel-Heinze oil-well log, #1-M Santa Fe Pacific 9612, Log No. F278, 5-5N-7W, total depth 4992 feet, drilled 1959; (2) Sun Oil Company oil well log, #1 Pueblo of Acoma, 2-7N-7W, total depth 4794 feet, drilled 1960, logged by E.R. Hill. The stratigraphic units are, in descending order with simplified descriptions:

Chinle Formation (Upper Triassic) -- c. 510 m thick below volcanoes. Mainly red to reddish brown and purple silty mudstone and clay shale with thin sandstone lenses. Lower 200 m has zones of feldspathic sandstone, and chert and limestone pebble conglomerates.

San Andres Formation (Lower Permian) -- c. 130 m thick. Mainly gray fine to medium grained limestones, silty limestones, gypsum, and gypsiferous shale. Distinctive petroleum odor when broken.

Glorieta Sandstone (Lower Permian) -- c. 60 m thick. Massive, pale yellow, fine-grained, well-sorted quartzose sandstone with abundant cross bedding. Erosionally resistant formation.

Yeso Formation (Lower Permian) -- c. 400 m thick. Alternating beds of pale red, yellow, and buff gypsiferous shale with grayish white to pink sandstone. Highly gypsiferous.

Abo Formation (Lower Permian) -- c. 280 m thick. Mainly dark red, silty sandstones and shales with lenses of limestone.

Madera Formation (Pennsylvanian) -- c. 230 m thick. Various combinations of tan, gray, white, and red dense limestone interbedded with white, coarse-grained sandstone and light gray to red-brown shale.

Sandia Formation (Pennsylvanian) -- c. 75 m thick. Tan-white and tan-brown, dense, crystalline limestone interbedded with white coarse-grained sandstone and conglomerate.

Granite Wash Formation (Precambrian) -- Coarsely crystalline granite.

We estimate the thicknesses to be accurate to within 10% of the thickness of each formation, based on combined outcrop and well data. Most observed xenolith fragments were from the Chinle, San Andres, or Glorieta formations (depths less than about 700 m).

5.2.3.3. Alkali Buttes Volcanic Center

Alkali Buttes (Figures 2) are erosional remnants of an approximately 2 km long, north-northeast trending chain of four to five vents that is subparallel to major structures in the region. This chain probably formed a single continuous landform prior to erosion. The absence of paleosols or reworked deposits indicates that the chain erupted over a relative short time, although the erosional removal of deposits between South and North Alkali Buttes precludes any determination of relative timing of the two.

Description of South Alkali Butte. South Alkali Butte is the remnant of a tuff ring or tuff cone (only the inward-dipping, crater-filling deposits are preserved). The earliest preserved deposits (Figure 3) are massive tuffs containing abundant lapilli and block-sized xenolith clasts; these deposits are only exposed along the northeastern quadrant. Juvenile clasts in the tuffs range from ash to as large as ~40 cm, and occur both as scattered clasts throughout the tuff and as lapilli-rich, clast-supported lenses. The lenses are 0.5-2 m long and up to 0.3 m thick, and some are reverse-graded. Dense juvenile clasts are typically angular and blocky and in some cases have cauliflower bomb textures. Vesicular clasts in the tuff have ragged margins where basalt may extend as fingers into the tuff and vice versa, indicating that the clasts were still partly molten upon deposition. The ash matrix of the tuffs contains large proportions of disaggregated xenolithic material (discussed in detail below). The abundance of xenoliths and blocky nonvesicular juvenile clasts indicates that these tuff deposits probably represent hydrovolcanic activity at the early stages of eruption. We infer that these deposits are relatively massive due to constant churning and redeposition within or near the vent, as opposed to fallout and pyroclastic surge deposits (not preserved here) outside the crater, which would typically be better bedded and show a variety of sedimentary structures.

The tuffs are overlain by beds of poorly vesicular basalt lapilli with much lower xenolith content than the tuffs (Figure 3). At South Alkali Butte some of these lapilli beds contain an ash matrix that comprises less than 30% by volume of the deposits, while other beds have little or no ash and are clast supported. This facies is interpreted to represent explosive ejection of quenched magma in the upper part of the conduit, driven mainly by hydrovolcanic explosions, and records the transition from dominantly hydrovolcanic to Strombolian eruptions.

Overlying the lapilli beds are beds of coarse, vesicular scoria and small bombs, which are in turn overlain by variably agglutinated coarse spatter beds interlayered with thin lava flow units (Figures 2, 3). At South Alkali Butte this part of the sequence tends to exhibit upward coarsening from coarse lapilli to large bombs. Spindle and ribbon bombs are common in these deposits. Bedding is crudely defined by coarse bombs. This facies, referred to as vesicular scoria and spatter, is interpreted to represent Strombolian eruptions driven by vesiculation.

Near the original crater center, the spatter beds are overlain by deposits consisting of a mixture of mud derived from the Chinle formation and large, highly vesicular, fluidal basalt clasts. The basalt clasts are highly altered so that they are blue-gray in color and are very soft (chalky), but vesicles and bomb textures are preserved. The lowest exposures of this unit are composed of approximately equal amounts of reddish mud matrix and basalt bombs, and the proportion of basalt clasts tends to increase upward in the deposit. Some of the larger bombs (20-30 cm) have yellow-green alteration zones (palagonite?). The degree of alteration of basalt clasts decreases upward and the matrix gradually changes to a greenish-orange color. The deposit is thickest (~3-4 m, base not exposed) and basalt clasts tend to be subhorizontal near the inferred main vent of South Alkali Butte (Figure 2), the deepest part of the crater, and drapes up onto and pinches out over the vesicular scoria and spatter beds that form the rim of what remains of the crater (the original rim has been eroded away). This mixed Chinle mud-basalt clast deposit is capped by lavas. Near the original center of the crater these lavas have pillow-like structures and may be pillow lavas; they locally protrude downward into the Chinle-basalt mixture.

Interpretation of South Alkali Butte Eruptive Sequence. We interpret this sequence of deposits as recording decreasing hydrovolcanic activity and increasing Strombolian activity through the main phases of eruption (lapilli and block-rich tuffs, poorly vesicular lapilli, scoria, and spatter/lava flow units). The initial hydrovolcanic eruptions were driven by interaction between rising magma, wet sediment, and/or free pore water. At Alkali Buttes it is likely that most of the magma-water interaction took place within the Chinle formation, as indicated by the abundance of Chinle-derived material in the ash matrix of the tuffs (see below). According to Jicha (1958) most wells in the region are drilled into and produce water from the Chinle formation, indicating that much of the formation is saturated. These wells produce water at very low rates (c. 4-8 liters per minute; Jicha 1958), indicating low permeabilities. The generally low permeabilities indicate that it is possible that mixing between magma and wet sediment was an important mechanism for initial hydrovolcanic activity at Alkali Buttes, in a situation very similar to that described by White (1991) for the Hopi Buttes hydrovolcanic eruptions which penetrated a surface unit (Bidahochi formation) that is similar in its characteristics to the Chinle formation. Magma-wet sediment interaction is evidenced by interfingering contacts between basalt clasts and surrounding xenolithic matrix material (White 1991; Leat and Thompson 1988).

This main phase was followed by an extended period of weak lava effusion and mud boiling in the lowest parts of the crater (to produce the mixed Chinle mud-basalt clast unit) and eventually production of a small volume of lava that flowed onto and into the mixed Chinle mud-basalt clast unit. This mixed deposit can be thought of as having been emplaced as a weakly erupted peperite, and was probably a water-saturated slurry at the time -- an extremely muddy crater lake. This deposit was still very wet when subsequent lava flows were emplaced over and intruded down into it, hence the possible formation of pillows.

A small area, about 50 m in diameter, on the interior of the center is littered with large (decimeter to meter) xenolith and dense basalt blocks. The xenoliths are mainly Chinle, San Andres, and Glorieta rocks. These blocks are interpreted to record a late-stage phreatic explosion. Although many of the blocks originated at 510-700 m depth, we think that the locus of this phreatic explosion was probably high within the throat of the volcano. This is supported by the presence of large, dense basalt blocks, and implies that the deeper-seated xenoliths were probably recycled

from earlier stages of eruption. Possible spring deposits (gray to yellow carbonates with massive, vuggy, and laminated textures) cap the lava in the center of the original crater.

Description of North Alkali Butte. North Alkali Butte consists of the remnants of three scoria/spatter cones and associated craters (Figure 2). The southern two cones filled in a previously formed tuff ring or tuff cone, of which only the interior, inward dipping deposits are preserved. As at South Alkali Butte, the lowest preserved deposits of this tuff ring/cone phase are lapilli tuffs (only exposed at station N4; Figures 2, 3), overlain by poorly vesicular basalt lapilli beds. The latter deposits are dominantly clast supported with local thin layers and lenses that are richer in ash and are matrix supported. Local coarse lenses may represent small grain avalanche deposits. This facies is overlain by vesicular scoria and spatter beds that are variably agglutinated and are associated with the three cones (Figure 3). Small lava flows were also erupted from the cones.

The northern part of North Alkali Butte is poorly exposed, and contains only one outcrop indicative of hydrovolcanic activity (Station N1, Figure 2) -- a deposit of coarse juvenile lapilli and blocks set in a shaley or mud matrix. Juvenile clasts in this deposit range from broken bombs and angular clasts to clasts with textures indicative of fluidal mixing with the wet-sediment matrix (White, 1991; Leat and Thompson, 1988).

Possible spring deposits, similar to those described at South Alkali Butte, occur on the northeast flank of North Alkali Butte. An active spring is present at the southern end of North Alkali Butte; it is collinear with the possible spring deposits on North and South Alkali Buttes, along a north-northeasterly trend that, like the trend of the vents, is subparallel to regional structures. Spring deposits also occur about 0.5 km northeast of the northern end of North Alkali Butte along the same trend.

Interpretation of North Alkali Butte Eruptive Sequence. North Alkali Butte went through similar eruptive phases to South Alkali Butte, but with a more dominant component of Strombolian eruption to produce the scoria and spatter cones. Treating Alkali Buttes as a single chain of vents, we suggest that hydrovolcanic activity was strongest near the southern end of the chain, possibly reflecting a local concentration of groundwater paths (e.g., fractures), and decreased northward.

Structural Control on Hydrovolcanic Activity at Alkali Buttes. As will be seen below, Volcano Hill, which erupted through the same sedimentary sequence as Alkali Buttes, shows little or no evidence of magma-water (or wet sediment) interaction. It is likely that the ascent path of magma beneath Alkali Buttes was controlled by a crustal structure that also affected groundwater, whereas the magma path beneath Volcano Hill was not. This is supported at Alkali Buttes by, in addition to the presence of hydrovolcanic deposits: (1) evidence such as possible pillow lavas that the original crater at South Alkali Butte was very wet (either with standing water or a mud slurry); (2) the persistence of springs both shortly after eruption and into present times along a trend that approximately coincides with the vent locations and extends beyond them; and (3) the fact that structural control plays a role in determining quality of groundwater from the Chinle formation in modern wells and springs of the region, indicating its importance to groundwater flow (Jicha 1958); (4) the north-northeast trend of Alkali Buttes and other vents in the same valley indicates structural control - Baldridge et al. (1987) suggested that this is the location of a hinge line for the Sierra Lucero uplift. We think that there may be fractures at depth related to this hinge line, although there is not obvious evidence for significant faulting.

5.2.3.4. Volcano Hill

Volcano Hill (Figure 4) is a large scoria cone, rising about 180 m above the surrounding terrain; we estimate that the original cone extended about 50 m higher but has since been eroded. The cone is open to the west; where a lava flow extends from the central crater west and north to a

distance of about 5 km. It appears that the cone and lava flow formed simultaneously by lava fountain and Strombolian activity (similar to the 1984 Puu Oo cone, Hawaii; Wolfe et al. 1987a). The volcano shows no evidence of major breaks in eruptive activity and is therefore considered to be monogenetic. It is composed almost entirely of variably welded, coarse spatter and scoria; there is no evidence of any substantial hydrovolcanic activity. Individual clasts are moderately to highly vesicular, and fluidal shapes and textures are very common in coarser clasts. Individual beds are moderately to poorly sorted, based on visual inspection, but contain virtually no pyroclasts smaller than lapilli size. Except in the innermost part of the crater, the crudely defined bedding dips radially outward from the center of the cone. Inward dipping beds, typically 0.5-1 m thick, of alternating welded spatter and lava, occur in the deepest part of the crater, and represent the near-vent facies of the volcano (Figure 4). These lava beds probably formed by complete coalescence of hot spatter clasts, and the alternation with non-coalesced spatter reflects variations in the local clast temperature and accumulation rate and hence magma discharge at the vent (Head and Wilson 1987). A chain of three minor vents extends to the S-SW from this main vent. A highly altered, basaltic dike is exposed in a canyon to the north of the main cone. This dike lies on the projected trend of the other vents. The trend is subparallel to regional structures, as is the trend of vents at Alkali Buttes.

5.2.3.5. Xenolith Data

Maximum xenolith and juvenile clast size. Maximum clast sizes were obtained by measuring the long axis of the three largest clasts (MJ - juvenile clasts, or ML - xenolith clasts), within a few m² of outcrop area and taking the average at each station (Figure 5); they are discussed here in terms of the four main eruptive facies represented at Alkali Buttes and Volcano Hill (lapilli tuff, poorly vesicular lapilli, vesicular spatter and scoria, and lava). Overall, MJ is larger (typically 80-100 cm) in the vesicular scoria and spatter facies than in the other two pyroclastic facies. The opposite is true, with some exceptions, for ML; lapilli and block-rich tuffs tend to have the largest xenolith clasts of the pyroclastic facies. The largest xenolith clasts measured in the tuffs are 2.2 m and 1.5 m in long dimension, and are from the San Andres and Glorieta formations, respectively. Other xenolith clasts of comparable size were found at one station (S4) in the vesicular spatter and scoria units at South Alkali Butte. These clasts are 2.7 and 3.0 m in long dimension, and are both from gypsum beds that are characteristic of the topmost Yeso Formation. Thus large meter-size blocks were erupted from depths of 500-700 m during both hydrovolcanic and magmatic eruptions.

Xenolith volume fractions - measurement technique and sources of uncertainty. Many basaltic volcanic rocks are either lavas or very coarse-grained or consolidated (except for fallout tephra deposits), so that standard sieve techniques and component analyses are not very useful for determining volume fractions of xenoliths. Our method is as follows. An area (typically 1 m²) is marked off on the face of an exposure. All xenoliths (including those within juvenile clasts) with long dimensions greater than 1 cm are measured (long and short dimensions) and identified with their original formation, and thus depth range, in the subvolcanic sequence. Hand samples are collected from each measured area for analysis of smaller grain sizes with hand lens and microscope. The long and short dimensions of each clast are used to estimate its area in the exposure face, approximating the shape of the clasts as rectangles. Dividing the total clast areas for each subvolcanic unit by the exposure area yields area fractions represented by each unit. To convert these area fractions to volume fractions (F) we use the following formula:

$$F = (\text{area fraction})^{3/2} \cdot (1.18)^3 \quad (5.1)$$

The quantity 1.18 is applied to correct for the fact that individual clasts are not necessarily exposed such that their true long and short dimensions can be measured. This factor has been shown to account for random slices through spheres (Cashman and Marsh 1988; Mangan et al. 1993). Strictly speaking, this factor does not apply to angular fragments such as most of the measured

xenoliths described here. The error introduced by this simplified correction is small compared to other sources of error in the xenolith concentration estimates (see below). Samples from poorly vesicular lapilli facies and vesicular scoria and spatter facies contain virtually no material finer than lapilli size. Hand lens examination of samples from these facies confirmed that the field measurements described above account for virtually all of the xenoliths in them.

The block and lapilli rich tuff facies, on the other hand, contain appreciable quantities of ash (up to c. 90% volume), which was analyzed by standard point counting techniques on a petrographic microscope. 500 points were counted on each thin section, with spacing of about 1 mm between traverses and 1 mm between points along traverses. There are four main types of matrix grains in the tuffs. The first type is juvenile basalt grains, which occur as black, poorly vesicular, relatively phenocryst-poor varieties and as amber brown, vesicular, phenocryst rich clasts. The second type of matrix grain ranges in color from milky or purplish white to reddish brown, is massive, and appears the same under parallel and crossed polarizers. These features are typical of mudstones and claystones (e.g., Williams et al. 1954), and the general characteristics of the particles indicate that they are fragments of Chinle Formation. Some of these fragments may also contain zeolites. Quartz, the third important matrix grain, occurs mainly as individual grains, typically less than 1/3 mm in diameter, that are subangular to rounded and occasionally broken. Rare fragments of sandstone, about 1 mm in size, were found in a sample from site S9; the quartz grains in these clasts are rounded. Because most of the matrix quartz consists of individual grains, we interpret it to be derived mainly from the Chinle Formation, originating either as dispersed sand grains in the mudstones and shales, or from poorly cemented sand lenses within the formation. Some of the quartz grains and rare sandstone fragments were possibly derived from the Glorieta Formation, but this is less likely because of the strongly cemented nature of that unit. The final important type of matrix material is calcite (and some zeolite?) which lines void spaces and fractures in the matrix. This material formed after deposition of the tuffs. Detailed component data for matrix material from these tuffs are provided in the footnotes for Table 5-1.

Sources of error include, in addition to those mentioned above, the measured exposure area (less than ~10%), departure of the exposure surface from true two-dimensionality (probably <10%), irregular shapes of xenolith clasts (errors perhaps as high as 50%), uncertainty in correlation of clasts with correct subvolcanic units, and uncertainty in unit thickness directly beneath the volcanoes. The unconsolidated, fine-grained nature of much of the Chinle Formation may result in its volume fraction being underestimated because: (1) it may have mixed with magma and disaggregated into fine particles that are not recognizable; and (2) disaggregated Chinle material may have been deposited as fine ash in more distal pyroclastic deposits that are no longer preserved. Some bombs, especially at Volcano Hill, contain swirly patches of Chinle material, indicative of initial stages of the mixing into the magma. Because most sedimentary sections on the Colorado Plateau contain at least one formation that is composed of mudstone or fine shales, this problem is largely unavoidable. Similar studies at other areas where such units reside at different depths below the volcanoes are in progress; this will allow us to filter out such complications. Although these errors seem somewhat high, we think that most of the xenolith volume fraction estimates are accurate to within a factor of about two, and almost certainly within an order of magnitude.

Xenolith volume fractions. Calculated volume fractions of each stratigraphic unit in the country rock are listed in Table 5-1, where they are grouped by facies (and thus by eruption mechanism). Total xenolith volume fractions vary somewhat systematically with facies (Figure 5), the hydrovolcanic tuffs having the highest xenolith contents (by volume), ranging from 0.32 - 0.91. Poorly vesicular lapilli deposits range from 0 - 6×10^{-3} . Vesicular scoria and spatter facies have xenolith contents ranging from 2×10^{-5} - 7×10^{-3} , with most sites ranging between 10^{-4} - 10^{-3} . These are consistent with data reported by Crowe et al. (1983a) from other scoria cones in the

southwestern U.S. (Table 5-1). Only two sites in lava were analyzed for xenolith contents, and these are within the same range as the vesicular spatter and scoria deposits.

Entrainment rates. Entrainment rate for each subvolcanic unit is obtained by dividing the volume fraction of xenoliths from a given unit by the unit's thickness. The value obtained is the *average* xenolith volume fraction derived per unit depth within each unit (see also Mastin 1991). As average values, these entrainment rates do not account for local flaring or narrowing of conduits within individual subvolcanic units. We assume that feeder dikes/conduits retain near-vertical trajectories as they propagate through the various subvolcanic units, although it is possible that in some units dikes may bend and therefore effectively have a longer path through those units (e.g., Baer and Reches 1991; Baer 1991). Entrainment rates for each of the measured sites are listed in Table 5-1. We focus our discussion here on five stratigraphic sections, two at South Alkali Butte (sections S1, and S7 including site S8), two at North Alkali Butte (sections N4 and N6) and one at Volcano Hill (Figures 2-4).

Sections S1 and N6, which include poorly vesicular lapilli facies and vesicular spatter and scoria facies (Figures 6, 7), do not show simple systematic variations in entrainment rates as a function of stratigraphic height in the eruptive sequence. Note that sites S1-1 (not plotted because it contained no xenoliths), S1-2, and S1-3, all of which were measured at the same level in the same lapilli layer, show as much variation as the entire vertical sequence at section S1, which includes variations in facies (see Figure 3 for stratigraphic locations of sites). Entrainment rates within the San Andres (510-640 m depth) and Glorieta (640-700 m depth) tend to be larger than the average entrainment rate in the overlying Chinle Formation (note that high entrainment rates at very shallow depths, associated with a flaring vent structure, would not be recorded on these plots because xenolith derivation is averaged over the entire thickness of a unit). The low apparent Chinle entrainment may also be in part due to disaggregation of unconsolidated Chinle material as it was entrained and subsequent dispersal as ash that is not preserved in these proximal deposits. However, entrainment rates for San Andres and, more consistently, Glorieta materials are relatively constant in all eruptive facies at Alkali Buttes and at Volcano Hill.

The section measured at Volcano Hill consists only of vesicular scoria and spatter facies (Figure 8). All but one of the six measured sites contain xenoliths from the Glorieta Formation, which was typically entrained at a rate of between 10^{-6} and 10^{-5} volume fraction/m. Half of the sites also contain Chinle material, but at relatively low entrainment rates. There is no obvious systematic change in the pattern of wall-rock entrainment as a function of stratigraphic height in the cone.

Sections that include block and lapilli rich tuff facies show some systematic variation in entrainment rates (Figures 9, 10). The two stratigraphically-lowest sites at section S7 (including site S8) and the lowest site at section N4, all in the tuff facies, have large entrainment rates in the uppermost 510 m (Chinle Formation); this reflects the large quantity of xenolithic debris in the ash matrix of the tuffs. However, entrainment rates of the deeper San Andres and Glorieta formations are similar to estimates at other sites for the other three facies. Higher up in the sections, the vesicular scoria and spatter deposits are very poor in xenoliths and have much lower entrainment rates, especially of Chinle material.

In summary, the entrainment rate data indicate there is a relatively consistent degree of wall rock erosion at depths of 510-700 m, and especially between 640-700 m (Glorieta Formation). Pyroclastic deposits interpreted as recording early hydrovolcanic phases at Alkali Buttes are fundamentally different only in that the shallow (0-510 m, Chinle Formation) entrainment rates are much larger than in other facies. The deeper entrainment rates are essentially the same in the tuffs as in other facies.

5.2.3.6. Interpretations and Conclusions from Lucero Volcanic Field

The width of dikes or conduits beneath volcanoes has two components: that produced by elastic deformation and that produced by erosion or entrainment of the dike walls. The xenolith data presented here allow us to assess the relative importance of the latter component. For example, the total eruptive volume of the Alkali Buttes volcanic center is estimated to be on the order of 10^8 m^3 (about half of which is preserved). We estimate that about 6000 m^3 of Glorieta xenoliths, which were entrained at a relatively consistent rate of about 10^{-6} m during all eruptive phases, were ejected at Alkali Buttes. Assuming that at depth the plumbing for Alkali Buttes was characterized by a dike of about 1.9 km length (the approximate length of the chain of vents), the average dike widening from entrainment was about 5 cm. This estimate is consistent with maximum xenolith sizes in the vesicular scoria and spatter facies, most of which are between 5-10 cm in long dimension. The local effects of entrainment could be much more dramatic, for example where a meter-sized block of wall rock is entrained, but this would probably not have a large effect on eruption dynamics.

During strongly hydrovolcanic phases of eruption, on the other hand, wall rock entrainment can have a much larger effect on dike width. Although we cannot resolve entrainment variations within the uppermost 510 m below Alkali Buttes, it is likely that at very shallow depths, say less than 100 m, the entrainment component of dike widening could have been even more significant. At depths below the Chinle Formation, though, the entrainment component during hydrovolcanic phases was similar to that during magmatic phases.

The relationships between xenolith derivation data and facies (eruption mechanisms) allow us to assess the relative importance of the various entrainment mechanisms. For example, the xenolith data show that entrainment of Chinle material was much larger during hydrovolcanic phases than during magmatic phases. During the hydrovolcanic phases explosive magma-water (or magma-wet sediment) interaction was the dominant entrainment mechanism. The relatively consistent entrainment of deeper Glorieta, and, to a lesser extent, San Andres formations is more difficult to explain. One possibility is that the Glorieta Formation is an aquifer and that during magma ascent steam explosions occurred there, causing wall rock entrainment. The Glorieta Formation, however, is typically strongly cemented and probably has a very low matrix permeability. It is thin and tends to crop out as steep slopes or cliffs, and therefore does not have a large recharge area in the region (Jicha 1958). It is highly jointed and may have a high fracture permeability, but the volume of water that could flow through these fractures would be relatively small. While the possibility that entrainment of Glorieta xenoliths is caused by steam explosions cannot be ruled out, we do not think that it is the best explanation. Another mechanism for entrainment is abrasion of the conduit walls above the fragmentation level (where the volume fraction of gas in the vesiculating magma becomes sufficiently high that the magma fragments and the flow is characterized by magma clots entrained in a rapidly accelerating gas stream). However, it is unlikely that the fragmentation level of these basalts was deeper than about 400 m, which is the approximate fragmentation depth for a basalt containing a large mass fraction of water (2%). Also, this deep entrainment is recorded in lava flows which may have formed from purely effusive activity in which the magma did not fragment. Finally, the fact that the Glorieta Formation is the most competent of the subvolcanic stratigraphic units rules out the possibility that it was preferentially entrained by shear erosion.

The most likely mechanisms for entrainment of Glorieta xenoliths are: (1) inclusion of wall rocks between the main dike(s) and small offshoot dikes, and (2) cracking of wall rocks due to thermal stresses (McBirney 1959) and pore-pressure buildup (Delaney and Pollard 1981; Delaney 1982). Both of these mechanisms involve brittle failure of wall rocks, which addresses the apparent paradox that the Glorieta Formation is the most competent of the subvolcanic units and yet was preferentially entrained by the rising magmas. Ongoing studies of the type reported here are

focusing on areas where competent and incompetent stratigraphic units reside at different depths below volcanoes in order to see if the pattern described here is similar at other sites.

A final point relates the xenolith data here to the conduit modeling approach of Macedonio et al. (1994). They modeled conduit flow in a rigid conduit, then used the flow field to estimate erosion of the conduit walls as a function of depth in the conduit. In their governing equations the contribution of mass and momentum by entrained wall rock was neglected. Our data indicate that for basaltic eruptions driven by expansion of magmatic volatiles this approximation is probably quite adequate. The volume fraction of xenoliths (relative to the total volume of solids) entrained per meter is so small that even under the most aggressive conditions of magmatic eruption that occurred at the Lucero volcanic centers, neglecting the xenolith mass and momentum contributions would result in very small errors compared to the other uncertainties inherent in modeling geologic processes. During hydrovolcanic phases, however, it would be critical to account for the effects of the added xenolith mass, momentum, and energy. The entrainment rate values in Table 5-1 can be used as a range of source terms for the governing equations of Macedonio et al. (1994; $\hat{C}_w$ in their equations 3 and 5).

5.2.4. Analog Studies in San Francisco Volcanic Field, Arizona

5.2.3.1. Introduction

In this section we present data pertaining to wall-rock entrainment from three small basaltic volcanoes in the San Francisco volcanic field. These data are from Strombolian facies deposits. Up through the end of FY 95, detailed field studies were being conducted in this volcanic field on four volcanoes and a wider variety of eruptive facies including strongly hydrovolcanic. These field-based studies are not yet complete and therefore are not reported here in detail. In general, the preliminary data from San Francisco volcanic field is quite consistent with that obtained in the Lucero volcanic field, indicating that similar conclusions would have been reached if the research had been carried to completion.

5.2.3.2. Geologic Setting

The San Francisco Volcanic Field (SFVF; Figure 11) is located just north of the physiographic boundary between the tectonically-stable Colorado Plateau Province and the Basin and Range Province, which is characterized by extensional tectonics. About 600 vents are present in the field, and a wide range of lava and pyroclastic compositions occur there including tholeiitic basalts to dacites and rhyolites, as well as more alkaline compositions such as basanites, hawaiites, benmorites, and trachytes (Holm, 1986; Arculus and Gust, 1995). Most of the SFVF is covered by mafic lavas and cones, with a few focal points for silicic volcanism (Figure 11). The most prominent silicic center is the composite cone known as San Francisco Peaks, which reaches an altitude of c. 4000 m above sea level (2000 m above the surrounding plateau). The youngest volcano in the field is the basaltic cone and lava flows of Sunset Crater, which erupted at c. 1064 A.D. (Smiley, 1958; Holm, 1986, 1987; Holm and Moore, 1987; Self, 1990). Volcanism began in the area about 9 million years ago (Damon et al., 1974; Ulrich et al., 1984).

A wide variety of volcanic studies have been conducted in the SFVF, especially over the past two decades. A United States Geological Survey team has published a comprehensive base of geologic maps, geochemical analyses, radiometric ages, and paleomagnetic data (Moore et al., 1976; Ulrich et al., 1984; Tanaka et al., 1986; Ulrich and Bailey, 1987; Wolfe et al., 1987b,c; Newhall et al., 1987; Moore and Wolfe, 1987). Numerous studies have focused on xenoliths erupted in the volcanic field, but these have only been concerned with fragments of deep crust or mantle materials (Cummings, 1972; Stoeser, 1974; Pushkar and Stoeser, 1975; Chen and Arculus, 1995).

We chose four basaltic volcanoes for detailed study (Dry Lake, South Sheep Hill, Rattlesnake Crater, and South Merriam; Figure 12) because they represent a range of eruptive mechanisms and because they sit either directly on sedimentary rocks or are separated from the sedimentary rocks by no more than 50 m of volcanic rocks. However, because the field studies are incomplete, data are only reported for part of the Dry Lake sequence and South Merriam). Based on Ulrich et al. (1984), we estimate the following stratigraphy beneath the selected volcanoes, listed in descending order from the surface:

Earlier volcanics - < 50 m thick, mainly basaltic lava flows. Less than 9 Ma.

Kaibab Formation - ~150 m thick, yellowish-gray to light gray silty dolomite, dolomitic sandstone, minor sandstone, and dolomitic limestone. Lower Permian.

Coconino Sandstone - ~300 m thick, also includes Toroweap Formation. Light colored, orange-tan, fine-grained, cross stratified sandstone. Lower Permian.

"Red Beds" - ~760 m thick, includes Hermit Shale, Supai Formation, Redwall Limestone, and Temple Butte Limestone. Dominantly red-colored shales, sandstones, and limestones. Middle Devonian through Lower Permian.

We do not differentiate the formations within the "Red Bed" category because it is difficult to distinguish them in hand-sized xenoliths studied here.

5.2.3.2. East Merriam Cone

East Merriam Cone is a partly quarried cone just east of the larger Merriam Cone (Figures 12, 13). According to Moore and Wolfe (1987) the eruptive products of this cone (their vent 3036B) are microporphyritic olivine basalts, with an age of 0.15 ± 0.03 Ma. The quarry, where our data were collected, exposes massive, parallel bedded, well-sorted scoria fall layers. The scoria clasts are highly vesicular and glassy, nearly to the point that they could be termed pumice. We interpret these deposits to have been laid down by very energetic, probably violent Strombolian eruptions. Due to inaccessibility of steep quarry walls, only two xenolith abundance measurements were taken near the base of the quarry. The results for these are presented in Table 5-2.

5.2.3.3. Dry Lake

Dry Lake volcano (Figure 12, 14) is a complicated center that apparently began as a cluster of Strombolian scoria cones that were then partially truncated by maar-crater collapse during final hydrovolcanic phases (Mayr, 1994). The eruptive products are plagioclase phyric hawaiites, with an age between 0.69 - 1.41 Ma (Wolfe et al., 1987c). Extensive quarrying has exposed a thick section through one of the scoria cones and its mantle of late-stage hydrovolcanic deposits. Xenolith contents in the scoria are provided in Table 5-2. The deposits in which these measurements were taken consist of massive, parallel, clast-supported beds of red scoria lapilli and small blocks. These beds record fallout from Strombolian eruptions.

5.2.3.4. Discussion

The limited data that have been analyzed at San Francisco volcanic field are consistent with those reported above for vesicular scoria and spatter facies at the Lucero volcanic field, and with data reported by Crowe et al. (1983a). Further field and laboratory work would be necessary to complete the suite of studies identified in the study plan. These would mainly consist of completion of field work and laboratory analysis of xenoliths from Rattlesnake Crater, South Sheep Hill, and the hydrovolcanic portion of Dry Lake. Preliminary data suggest that data from hydrovolcanic facies would be consistent with those from Alkali Buttes in the Lucero volcanic

field, both in terms of overall xenolith content and derivation as a function of depth below the volcanoes.

5.3 Subsurface Effects

5.3.1. Summary of Subsurface Effects Approach and Results

In order to constrain the effects of magmatic intrusions on a potential repository we took a combined approach applying field analog studies and numerical modeling. Field analog studies, emphasized during the first phases of the study plan implementation, focused on determining the factors that control shallow basaltic intrusion geometry, and the geochemical and mineralogical effects of basaltic intrusions on silicic pyroclastic host rocks. These studies are reported here. Numerical modeling studies were due to be phased in during the final half of the project, and as a result only the very first stages of such modeling are reported here.

5.3.2. Controls on Geometry of Shallow Basaltic Intrusions - Studies from Paiute Ridge Analog Site, Southern Nevada.

5.3.2.1 Summary

Late Miocene basaltic sills and dikes in the Paiute Ridge area of southern Nevada show evidence that their emplacement was structurally controlled. Basaltic dikes in this area formed by dilating pre-existing vertical to steeply E-dipping normal faults. Magma propagation along these faults must have required less energy than the creation of a self-propagated fracture at dike tips and the magma pressure must have been greater than the compressive stress perpendicular to the fault surface. N- to NE-trending *en echelon* dikes formed locally and are not obviously attached to the three main dikes in the area. The *en echelon* segments are probably pieces of deeper dikes, which are segmented perhaps as a result of a documented rotation of the regional stresses. Alternatively, changes in orientation of principal stresses in the vicinity of each *en echelon* dike could have resulted from local loads associated with paleotopographic highs or nearby structures. Sills locally branched off some dikes within 300 m of the paleosurface. These subhorizontal bodies occur consistently in the hangingwall block of the dike-injected faults, and intrude Tertiary tuffs near the Paleozoic-Tertiary contact. We suggest that the change in stresses near the earth's surface, the material strength of the tuff and Paleozoic rocks, and the Paleozoic bedding dip direction probably controlled the location of sill formation and direction of sill propagation. The two largest sills deflected the overlying tuffs to form lopoliths, indicating that the magma pressure exceeded vertical stresses at that location and that the shallow level and large size of the sills allowed interaction with the free (earth's) surface.

5.3.2.2. Introduction

Many studies report that dikes intrude normal to the regional direction of least principal stress, or to the local direction of least principal stress in the vicinity of large structures such as volcanic centers (e.g., Delaney and Pollard, 1981). Recently, some attention has been given to the mechanics by which those dikes inject into the crust. Specifically, it has been suggested that the propagation of dikes occurs by dilation of pre-existing fracture surfaces (e.g., Delaney et al., 1986) and by generation of self-propagating tip fractures in advance of the magma (e.g., Rubin, 1993; Lister and Kerr, 1991). By analyzing the distribution and geometry of dikes and their structural setting, it should be possible to distinguish the dominant emplacement mechanisms and possible structural controls on those mechanisms.

In many cases, subvertical dikes feed subhorizontal sheet intrusions (loosely called sills in this paper, although the strict definition of a sill requires that it be concordant with bedding or foliation of the host rock). Currently, three main factors have been identified that determine whether a dike will feed a sill. (1) If dike propagation is being driven by its buoyancy relative to

local wall rocks, then when a dike reaches a height in the lithosphere where its magma is no longer buoyant (the level of neutral buoyancy, or LNB) further upward propagation will cease. Further influx of magma is accommodated by lateral flow at the LNB and depending on whether σ_3 (least principal stress) is horizontal or vertical at the LNB, the dike will become a bladed dike or a sill, respectively (Lister and Kerr, 1991). (2) The emplacement of a dike changes the stress field because a dike exerts pressure on its walls. Under some conditions (Parsons and Thompson, 1991) the vertical stress could become the least principal stress and sills could form by continued magma influx. This process is enhanced where dikes cross zones of ductile rocks where ambient differential stresses are low, as discussed by Parsons et al. (1992). (3) If the host rock can behave as a fluid, then fracture propagation ceases and the dike magma simply spreads out laterally at its LNB as a density current. An example of this is where a propagating dike encounters another intrusive body which is still partially molten (Parsons et al., 1992; Valentine, 1993).

The first two factors for sill formation given above, which apply when the host rock behaves as an elastic solid, should have the following trademarks in the field. First, a dike which experiences these processes will likely experience them along most or all of its length. This is because the orientation of the stress field associated with a self-propagating dike can be assumed to be relatively constant at a given level along its plane, except in the immediate vicinity of its edges. Ideally, such a dike would form a sill along its entire length, and the sill would extend outward from both sides of the dike. Second, where a sill forms, its feeder dike should not extend above the level of the sill (except for a minor overshoot distance; Lister and Kerr (1991)). In essence, the entire dike would be deflected horizontally to form a sill. It is worth noting that both of these models for sill formation are most commonly applied to large-scale magma transport problems, for example to explain horizontal reflectors (inferred to be sheet intrusions) at lower-crustal depths that are recorded by seismic surveys in extending terrains (see Parsons et al. (1992) for a brief review). Sill formation at smaller scales and shallower depths is not as well understood although it is commonly assumed that these large-scale models apply at all scales. For example, small-scale (tens of meters thick, hundreds of meters in plan dimension) sill formation at the Paiute Ridge intrusive complex, described in this paper, was attributed to an LNB mechanism by Crowe et al. (1983a,b). By completing detailed analysis of the distribution and geometries of dikes and sills and their structural setting, it is possible to distinguish the dominant emplacement mechanism(s), thereby providing information about the local and regional stress state and possible structural controls on magma propagation during intrusion.

Here we describe the basaltic intrusive complex at Paiute Ridge, Nevada, which contains both dikes and sills emplaced at shallow depths. However, unlike sills produced by the mechanisms given above, the sills in this complex feed off of short segments of much longer dikes, the sills extend only from one side of their feeder dikes, and the dikes extend significantly higher than the level of sill formation. The role of pre-existing structures on dike and sill formation is emphasized. This area provides a favorable setting in which to document and explain dike and sill emplacement mechanisms because the contact relations are generally well exposed.

5.3.2.3. Geologic Setting

The Paiute Ridge intrusive complex is composed of dikes, sills and small remnants of scoria cones and lava flows of alkali basalt (hawaiite) composition (Table 5-3, Figure 15). The complex was originally mapped by Byers and Barnes (1967) and has briefly been described by Crowe et al. (1983a) and Valentine et al. (1992, 1993). Paleomagnetic evidence (Ratcliff, 1993; Ratcliff et al., 1994), radiometric age determinations (our data and Byers and Barnes, 1967), and petrologic studies (F.V Perry, unpublished data) indicate that the basalts were emplaced during a single, brief magmatic pulse that had a total volume of a few tenths of a cubic kilometer. No cross-cutting relationships have been found in the intrusions. The $^{40}\text{Ar}/^{39}\text{Ar}$ age of this event is 8.59 ± 0.07 (2 σ) Ma; it is part of a group of alkali basalts that erupted within the south-central

Great Basin in the past ~9 Ma, coinciding with the waning of silicic volcanism in the region (Vaniman et al., 1982).

The structural setting at the time of intrusion was that of a small horst and graben system that consisted dominantly of N- to NNW-trending normal faults. These faults, which formed from roughly E-directed extension during the early Tertiary (Minor, 1995), displace a sequence of middle Miocene silicic tuffs (from the Southwestern Nevada Volcanic Field located 25 km to the west, see Vogel and Byers (1989), Sawyer et al. (1994)), that had been deposited on rolling hills of Paleozoic carbonates. Eruptive lavas and scoria deposits occur only in the graben regions, whereas dikes and sills occur beneath the original floor. Most of the sills formed within the silicic tuffs (Figure 16) at original depths estimated between 150-250 m (based on the original topography, as reconstructed from erosional remnants of rocks that overlie the sills).

5.3.2.4. Descriptions of Dikes and Sills at Paiute Ridge

Dikes. Basaltic dikes in the study area have intruded dominantly within a 2 km-wide graben between the Carbonate and Paiute Ridges (Figure 15). The longest dikes occur along pre-existing, NNW-trending, steeply E-dipping normal fault surfaces that cut Paleozoic sedimentary rocks and overlying Tertiary tuffs (Figure 15). There are many other faults in the area, including: (1) those with similar (NNW) strikes that either dip more shallowly east or dip toward the west and; (2) those with E to NE strikes. At the current depth of exposure, however, none of these faults is occupied by dikes. Additionally, there are a few small dikes that radiate out from the volcanic neck (Figure 15) and do not appear to follow pre-existing structures.

The three longest fault-entrenched dikes (PR, M and E on Figure 15) range in length and width from 400 m to nearly 5 km and from 1.2 m to 9 m, respectively. The western-most, or Paiute Ridge (PR), dike varies in strike from N to NNW and intrudes Cambrian carbonates (Halfpint member of the Nopah Formation) in the south and nonwelded, bedded Tertiary tuffs (Tp in Figure 15) toward the north. The dike visibly occupies the steeply E-dipping Paiute Ridge fault at Slanted Buttes (Figure 15) and along the trend of the fault at its southern tip. Along the central segment, the dike does not reside along a mappable a fault surface, however, it connects the fault-entrenched tips and most likely occupies a fault surface as well. The dike nowhere appears displaced by the Paiute Ridge fault, which has a throw of less than 15 m across the Rainier Mesa member at Slanted Buttes. Dike widths are highly variable as they pinch and swell along strike. PR dike is thickest where it feeds the sill at its southern end (Figure 15) and along its longest (100 m) central segment.

The PR dike is composed of a main continuous dike and at least 24 dike-parallel segments that vary between 4 and 100 m in length. These segments are isolated and occur within tens of meters of the parent dike. In several places along strike, two segments diverge from the parent dike and enclose a lens of host tuff. In the northern half of the PR dike, where its shallowest intrusion level is preserved at Slanted Buttes, it splits into two parallel dikes and encloses fault breccia. The breccia is in smooth contact with the dike and contains vesiculated basalt and fused tuff. Also at this tip, elongated vesicles in the dike are folded into an open fold with a nearly horizontal axis.

PR dike contains ubiquitous near-vertical joints that result in a pervasive platy texture with plates parallel to the dike-host contact. Conversely, with the exception of local cooling joints in fused wall rock (extending 10-20 cm into the wall rock, perpendicular to the dike margin), joints are never visible in the host rock along the length of the dike. The contact between basalt and the tuff host rock is consistently smooth and shows no brecciation. Along strike at this contact, the tuff host rock is fused or welded to varying degrees: in places the tuff is completely welded and forms black vitrophyre that grades rapidly away from the contact, over a distance of ca. 20 - 100 cm, into nonwelded tuff that is apparently unaffected by the dike. In other places the tuff is only partly welded at the contact and black fiamme are elongated parallel to the contact. We infer that

this "contact welding" is the combined result of heat from the dike and compressive stress exerted by the flowing magma on the wall rocks. Welded host rock commonly contains vesicles that are elongated vertically and parallel to the margin. In some places, welded tuff coats the basalt and displays rills or elongate smooth ridges (flutes). Most rills plunge nearly vertically, however, a sub-horizontal rill is present in the central part of the dike. At the dike tip, exposed at Slanted Buttes, scoria patches crop out near the dike-host contact.

The two eastern dikes, M and E (Figure 15), show geometries and textures similar to those of PR, however, M dike is much shorter and does not feed a sill, and E dike was emplaced closest to the paleosurface and feeds two sills. M dike radiates out from a volcanic neck to the southeast (Figure 15) and, although it is poorly exposed, it visibly occupies a normal fault, oriented N40°W, 61°E, that displaces the Rainier Mesa member and older tuffs (see Byers and Barnes, 1967). M dike varies in width between 2 and 4 m, and has an exposed length of 400 m. Its host rocks are only Tertiary tuffs, which show no brecciation or jointing near the dike contact. Near the neck, M dike is a single continuous dike, and farther from the neck it is composed of several segments. The upper termination of the dike veers west, narrows, and terminates at an elevation of 1579 m. Texture within the dike is characterized by a vertical platy fabric that parallels the dike margins.

E dike is the eastern-most dike studied within the graben. Like M dike, it extends from the volcanic neck to the south, and was emplaced into Tertiary tuffs in a N to NNW direction. Near the neck, the dike visibly occupies a NNW-trending, steeply E-dipping normal fault that displaces bedded tuffs 3.5 m and does not cut the dike. The exposed dike length is 2020 m and the width varies from 3 m near the northern tip, to 6 m near the southern sill. E dike is less segmented than PR dike, and is composed of fewer than a dozen segments. In at least three places near the contact with the Rainier Mesa tuff, (Figure 15), and one location near the volcanic neck, the dike locally diverges to form two segments that engulf an 8 meter wide lens of the tuff host rock.

The texture of E dike is characterized by the pervasive vertical platy fabric common to M and PR dikes. Adjacent host tuffs are not jointed nor brecciated, except for local vertical jointing of the Rainier Mesa tuff which is intruded by the dike at its shallowest level. The contact of the dike and host tuff is preserved in places and varies from partly to completely welded in the same manner as described above for the PR dike. Where complete welding has occurred, vesicles are vertical and parallel the dike margin. Contact welding of the host tuffs formed oblate fiamme that parallel the dike-host contact. Visible thermal effects on the wall rocks disappear within one meter of the dike margin.

The southern termination of E dike, near the volcanic neck, shows sub-horizontal rills on the near vertical dike-host contact, while the same contact at the northern termination shows vertical rills. At its northern termination, the dike ends in three left-stepping *en echelon* segments. The margins of these segments contain completely welded tuff host rock (black vitrophyre) and occasional patches of scoria.

Sills. Subhorizontal basaltic sheets (here loosely termed sills), that are roughly planar and locally conformable with bedding, have intruded nonwelded bedded tuffs (Tp, Figure 15) in the graben east of Paiute Ridge. Sills occur consistently as extensions off of fault-conforming dikes into the hangingwall block of the NNW-trending normal faults. The sills appear to inject exclusively into tuffs at positions near the contact between the Paleozoic sedimentary rocks and overlying tuffs: the sills do not inject into the Paleozoic rocks themselves.

The western-most sill, W, forms east of PR dike in the hangingwall of the Paiute Ridge fault. It is fed by PR dike, which widens near the sill and can be traced to within about three meters of the sill's western margin. Based on reconstructions (Figure 16), W sill was emplaced in nonwelded bedded tuffs within tens of meters above the presently-buried Paleozoic-Tertiary

contact. The minimum thickness of W sill is difficult to determine because of erosion of the uppermost part of the sill and overlying tuffs, but is estimated at about three meters. Its maximum thickness appears to be less than 20 m and its maximum exposed plan dimension is ca. 600 m. The depth of sill intrusion was at most 250 m below the earth's surface (Figure 16), and 160 m below the highest level reached by the parent PR dike.

The lower contact of W sill with nonwelded bedded tuffs is generally conformable with bedding as is the upper sill-tuff contact where it is preserved. Textures within the sill are highly variable. At the western margin, the base of the sill has a pervasive platy fabric in which the plates (a few cm thick) are parallel to the subhorizontal, S-dipping tuff-sill contact. Where this contact is irregular the basalt is either not jointed or has joints that follow the shape of the contact. Above the basal contact, the platy fabric alternates with thicker (>10 cm) subhorizontal seams of massive basalt. Just north or south of the point where the dike and sill connect, the subhorizontal platy fabric is absent: in two places, 1.5 m and 10 m above the basal contact, the basalt is either irregularly vertically jointed, or has a chaotic, jumbled appearance. On the north side of the sill, where the sill and tuff dip on average 44° toward the south, rills plunging toward 235° are preserved on the under side of the sill. The margins of the sill dip moderately inward on all sides. Thin, discontinuous sills, < 1 m thick and < 10 m long, are locally present beneath the main W sill; these parallel the main sill and are separated from it by 2-3 m of country rock.

Two sills, ES and EN, inject into the hangingwall of the fault utilized by E-dike. The sills are fed by E dike which widens near the sill contacts and can be traced directly into the sills. Based on reconstructions (Figure 16), the southern-most sill, ES, injects bedded tuffs (Figure 15) within 15 m of the presently-buried Paleozoic-Tertiary contact. ES is approximately 450 m in the long dimension and is up to 24 m thick. The depth of intrusion was probably c. 175 m below the earth's surface, only 60 m below the shallowest level attained by the parent dike.

Along its lower contact with the nonwelded bedded tuffs, ES sill is parallel with the trend of bedding in some places, and is disconformable elsewhere. Tuff host rock underlying the ES sill contains fractures normal to the sill-tuff contact. The fractures typically have sharp lower tips and extend into the host tuff less than 2.5 meters. Through the center of the sill, a set of steeply E-dipping faults, comprising a N-trending normal fault zone, displaces ES sill approximately 4 meters down to the east.

EN, the northern-most sill attached to E dike (Figure 15), is emplaced in nonwelded bedded tuffs just above the Paleozoic-Tertiary contact and continues laterally toward the east to intersect a horst of fractured Ordovician limestones. EN is approximately 380 m across the long dimension and as much as 46 m thick. The depth of intrusion was probably less than 100 m below the earth's surface, at approximately the same height as that attained by the parent dike at its northern tip. EN sill appears generally discordant with bedding of the underlying tuff. Rills exposed on the underside of the basalt in contact with tuff plunge toward 227° on an EW-striking, 68° S dipping surface (Figure 17).

En Echelon Segments. *En echelon* segmented dikes occur in the area between Paiute Ridge and Carbonate Ridge, however, they were not examined in detail for this study, and so only their gross geometries are mentioned here. The *en echelon* dikes crop out west of PR dike and north and northeast of E dike. The two western dikes trend north and northeast and do not follow recognizable preexisting surfaces such as faults or lithologic contacts. The eastern-most *en echelon* dike trends north, and although it does not inject any pre-existing structures, it does follow the trace of an inferred N-trending fault that terminates south of the dike's southern tip (Byers and Barnes, 1967). These dikes do not follow the NNW trend of the dikes discussed above, and do not appear to be associated with any sills.

5.3.2.5. Interpretation

The three major dikes in the Paiute Ridge region occupy fault surfaces and are not cut by those faults. Thus, the dikes appear to have utilized pre-existing fault surfaces with NNW strikes. This implies that the structurally-controlled magma emplacement was energetically more favorable than if the magma had created its own path by fracturing intact rock at the upper dike tip. Sills fed by these dikes intrude into tuffs (except for the distal end of EN sill) and occur consistently in the hangingwall of dike-injected faults, close to and above the Paleozoic-Tertiary contact. The very shallow emplacement of sills (between 250 m and 100 m below the paleosurface), and deflection of overlying tuffs, suggests that sills interacted with the earth's surface during intrusion. Below, we explore conditions under which such shallow emplacement could occur.

Dikes. Decades of observations of dike geometries have documented that, in the absence of local stress perturbations, dikes typically form normal to the direction of the least compressive stress, σ_3 . More specifically, to propagate a vertical dike, σ_3 must be horizontal and the magma pressure within a dike, P_m , must exceed σ_3 (Pollard, 1987). If magma injects a pre-existing plane of weakness, then for a vertical plane, P_m must exceed σ_3 and the friction across the crack surface; and for a plane dipping less than 90° , magma pressure in the crack must exceed σ_3 plus a component of the vertical stress, σ_v . Furthermore, if the strike of the pre-existing planes is not normal to the direction of σ_3 , magma pressures must be much greater than σ_3 in order to dilate them: Delaney et al. (1986) determined that if P_m is only slightly greater than σ_3 or if differential stresses are high (Pollard, 1987), only cracks nearly perpendicular to σ_3 can be dilated. If, on the other hand, P_m is greater than the greatest horizontal compressive stress, σ_H , a crack of any orientation can be dilated.

The normal and normal-oblique faults, dilated by dikes, formed between 11.45 Ma and 8.5 Ma under a stress field characterized by a SW- to WSW-directed σ_3 (Minor, 1995). Subsequent to this faulting event and into present times, the regional stress field underwent a gradual clockwise rotation of 65° , resulting in a NW-directed σ_3 (Haimson et al., 1974; McGarr and Gay, 1978; Minor, 1995). Basaltic dikes were probably emplaced near the end of the period characterized by the WSW-directed σ_3 , before the 65° rotation, based on following information: 1) The age of the Paiute Ridge basalts is 8.59 ± 0.07 Ma, which corresponds to the end of the period of WSW-directed σ_3 ; and 2) Dikes were emplaced exclusively along NNW-striking faults. If NE-striking faults, and the associated NW-directed σ_3 , had existed during the time of intrusion, dikes would have preferentially occupied the NE- rather than NNW-striking faults.

The dikes occupy only steeply dipping ($>60^\circ$) planes even though other, more shallowly-dipping planes of weakness existed during intrusion: 1) to the south, where the dip of the Paiute Ridge fault shallows to 50° , magma did not inject the fault; and 2) other preexisting fault surfaces, notably those with shallower or west dips (Byers and Barnes, 1967; Goff et al., 1996), were also not intruded. This suggests that although P_m exceeded σ_3 , plus a small component of vertical stress, along steeply-dipping faults, it was not great enough to overcome the additional component of σ_v on shallowly-dipping faults.

Some estimates of the physical conditions, such as driving pressure and energy requirements, that accompanied dike intrusion can be made based on the above qualitative analysis. For uniform loading stresses, and neglecting gravitational forces on the solid host rock, magma in a dike does not sense the density of the surrounding rock, but only the ambient stresses on the dike

plane (Rubin, 1995). Therefore, the width to length ratio (w/l) of a dike is related to the pressure available to deform the dike walls and the resistance of the host rock to deformation, (Pollard, 1987; Rubin, 1995) as expressed by:

$$\frac{w}{l} = \frac{P_d}{\mu / (1-\nu)} \quad (5-2)$$

where $P_d = P_m - \sigma_3$ is the driving pressure, μ is the elastic shear modulus, and ν is Poisson's ratio. For the dike intruding the Paiute Ridge fault, the average width is 5 m and the length is 4.6 km. Estimated average shear modulus and Poisson's ratio for Paleozoic host rocks are 1 GPa and 0.25, respectively, and for the Tertiary tuffs are 800 MPa and 0.31 (Van Burkirk et al., 1978), respectively. Therefore, the driving pressure for the basaltic magma in these dikes was probably on the order of 1.4 to 2.6 MPa, which is within the range of 1 to 4 MPa reported by others (e.g., Pollard, 1987). These low driving pressures might be the reason some of the dikes apparently did not intersect the earth's surface. Such insufficient pressures could have occurred for many reasons, including a decreasing magma supply with time, inhibited flow due to solidification of the magma, or lateral redirection of the magma into adjacent sills.

From these driving pressures, we can obtain a measure of the energy consumed in the fracture process associated with dike intrusion. G is the mechanical energy release rate per unit increase in crack length. The propagation path of a dike will be that which minimizes G (Delaney et al., 1986; Rubin, 1995). The amount of energy required to propagate the dike can be determined by assessing the critical energy release rate¹, G_c , using Equation 21 of Pollard (1987):

$$G_c = \frac{\pi l P_d^2 (1-\nu)}{4\mu} \quad (5-3)$$

Using the parameters specified above in (5-2), G_c is approximately 6 MPa*m for dikes that intruded intact rock. This value must be greater than the critical energy release rate for dilating a pre-existing plane of weakness because the latter does not require the additional energy needed to fracture intact rock at the dike tip. In other words, for dikes at Paiute Ridge, it was energetically favorable for magma to utilize pre-existing fault surfaces than to create new fractures in intact rock, and so G_c can be used as an estimate of the minimum energy required for dikes to form by fracturing of intact rock and a maximum of that required for injecting pre-existing planes. At present, few other values of G_c , for dike formation, have been published. It will be interesting to compare energy requirements for different intrusion scenarios as they become available.

Flutes and rills on the dike-host rock contact can be interpreted as flow direction indicators. Such features observed in PR dike are dominantly vertical and therefore indicate vertical flow. No such features were observed in M dike. Indicators in E dike suggest that flow in the northern part of the dike was dominantly vertical, while toward its southern end flow was lateral. We suggest that the dike was emplaced first by flowing up along a fault plane. Flow gradually became focused

¹ Note that $K=K_c$ is a commonly used propagation criterion, however, its use in cases such as this is questionable. Rubin (1993) has shown that the stress intensity factor, K , which is a measure of the stresses at the crack tip, cannot be considered solely a material property, but in fact, varies with confining pressure and thus is not a physically meaningful quantity. Additionally, the use of $K=K_c$ as a propagation criterion is valid only if the region of inelastic deformation is small compared to crack length (see Rubin, 1995). For many dikes, zones of inelastic deformation at the dike tip are long, and a significant amount of energy could be dissipated by mechanisms other than simple extension of a crack surface (see Shlyapobersky and Chudnovsky, 1994). As such, if $K=K_c$ is inadequate to determine conditions necessary for propagation, it is preferable to analyze the energy requirements of the system through the criterion G_c because of its ability to incorporate the effects of the host rock as in addition to the crack configuration and loading.

at the southern end of the dike to form the volcanic neck or plug. As this conduit became established, magma flowing up the southern end of the dike was diverted toward the conduit, resulting in lateral flow indicators. As the conduit was established it exerted pressure on its walls. Because of its roughly cylindrical shape, this pressure resulted in tension on the conduit walls and the formation of radial dikes (Figure 15). One of these radial dikes was "captured" by a preexisting fault as it propagated away from the plug to form M dike. This sequence of events, where magma rises in a dike but gradually becomes focused into a central conduit is very common for basalts (e.g., Richter et al., 1970; Delaney and Pollard, 1981; Bruce and Huppert, 1990).

Sills. Sill geometries and spatial distributions in the Paiute Ridge area suggest that the recently proposed mechanisms for sill formation, (i.e., along the level of neutral buoyancy or LNB (Lister and Kerr, 1991) or at the point of dike induced stress reorientation (Parsons et al., 1992; Parsons and Thompson, 1991)), do not adequately explain their existence. The buoyancy force in an LNB mechanism depends on the integrated density of vertical columns through the dike and through adjacent wall rock to the surface, not on the local density contrast between magma and immediately adjacent wall rock. With this in mind, it seems unlikely that an LNB mechanism could have acted along such localized segments of the main dikes. In addition, along most of the dikes' lengths, there is evidence that magma continued to rise to a level significantly higher than the sills, and, the sills propagated from only one side of the dikes. Similar arguments can be made against the mechanism of dike-induced stress reorientation (Parsons et al., 1992; Parsons and Thompson, 1991) for sills at Paiute Ridge. This mechanism would most likely have caused sills to form along most of the length of a dike, barring any externally (i.e., non-dike related) induced major stress heterogeneities along the length of the dike.

Considering that the magma senses and responds to the ambient stresses, the sills must have intruded at levels where the maximum vertical stress was lower than the maximum compressive horizontal stress. Additionally, at the point of injection, the magma pressure must have exceeded either the strength of the rock or the cohesive strength of a subhorizontal pre-existing plane of weakness, such as a bedding plane or lithologic contact, and also must have exceeded the vertical stresses at the level of sill injection (<300 m). Vertical stresses at this depth can be estimated, (using $\sigma_v = \rho gh$), at roughly 5.2 MPa in the Tertiary tuffs, and ca. 6.5 MPa under the estimated volume of Cambrian rocks. Therefore, P_m for the tuff-intruding sills must have been greater than 5.2 MPa. In the absence of any tectonic compression, vertical stresses exceed horizontal stresses at all depths below the earth's surface (see Figure 18 and compare $\sigma_v = \rho gh$ with $\sigma_h = (\nu/1-\nu)\rho gh$). For conditions to have favored sill formation, then, some additional horizontal compressive stress must have existed along segments of PR and E dikes at points where the sills formed: if a small (3 to 4 MPa) horizontal stress had existed, the horizontal stresses would have exceeded the vertical stresses at depths shallower than 300 m (Figure 18). We suggest, therefore, that a horizontal compressive stress existed during sill formation along at least short segments of the dike-hosting faults. Such a stress regime would be similar to that reported for present-day horizontal stresses at a depth of 380 m near Paiute Ridge: Haimson et al. (1974) report *in situ* stresses that range from 8.8 (NE-directed σ_1) to 2.4 MPa (NW-directed σ_3), so the magnitude of compressive stress that we infer is reasonable. We emphasize, though, that because the sills formed only along very short segments of the dikes and their host faults, this compression was probably localized. This local phenomenon might be due to contrasting rock properties on different sides of the faults and/or local stress variations due to topographic effects.

If, indeed, ambient stresses are the dominant factor controlling the direction and geometry of magma injection, the consistency with which sills inject into a particular structural setting (into Tertiary tuffs within the hangingwalls of faults, just above the contact with Cambrian carbonates), must be related to stresses in the vicinity of parent dikes. Magma migrating upwards along dikes would approach such a contact (T-Cambrian) and sense a drastic change in ambient stresses and a

difference in those stresses on either side of the fault: vertical stresses on the western side of PR dike would have been roughly 6 to 7.6 MPa in Cambrian carbonates (at about 250 m depth) and those on the eastern side, in Tertiary tuffs, would have been about 5.2 MPa (Figure 16). Furthermore, adjacent to PR dike, bedding planes are W-dipping in both the Cambrian and Tertiary rocks. The dip direction of these potential planes of weakness would favor the injection of sills upwards toward the east, rather than downwards towards the west (Figure 16) similar to the fluid-filled cracks that seek the earth's surface analyzed by Weertman (1980). Based on an analogous structural setting, it is likely that the same mechanical configuration controlled the injection of the two sills fed by E dike. Thus, the consistent occurrence of sills in particular geometric situations that are highly spatially localized, suggests that their emplacement is controlled by the ambient stresses and rock type. These two factors were highly variable during intrusion because of the spatial distribution and orientation of tuff, Paleozoic rocks (that also formed paleo-hills), and major faults.

W sill and the largest sill near Carbonate Ridge (Figure 15) have deflected the overlying tuffs and dip inward around their margins to form lopoliths. The collapsed top of these bodies suggests that the magma interacted with the overlying free (earth's) surface. This geometry is not shared by the two smaller sills (ES and EN) fed by E dike. We suggest, following results of the laccolith study by Pollard and Holzhausen (1979), that the size of the sills is an important factor in the formation of the lopolitic geometry at Paiute Ridge. Pollard and Holzhausen (1979) suggest that, because of tensile stresses over the central part of a sill, a large body of magma will interact with the earth's surface if it is close to the surface. We suggest that the same physical constraints control the formation of lopoliths: larger bodies that are closer to the earth's surface experience more interaction (i.e., bending of overlying strata) with that surface. Pollard and Holzhausen (1979) calculate that the transition from a sill to a laccolith occurs at a ratio of half-length to depth (l/d) of approximately 1.6. We support this theory with the ratios for lopoliths in the Paiute Ridge area: the l/d of W sill is between 1.2 and 1.88 (depending on the estimated depth); the l/d of the largest sill is approximately 8.5. The smaller sills are not lopoliths and have l/d ratios between 1 and 2. Notably, these sills have a much smaller area than the lopoliths, and therefore could not exert as much force on the overlying rocks to cause bending. Therefore, the *aerial extent* of the sill, in addition to simply the l/d ratio, must also be an important factor in whether it becomes a laccolith or lopolith. There are two reasons why the roofs of Paiute Ridge intrusions sag inward (forming lopoliths) rather than bow upward: (1) The magma may have been vesicular at these shallow depths during intrusion, so that while magma initially pushed the roof up, as the volatiles leaked out and the intrusion cooled the roof gradually sagged downward. (2) The host rocks were originally nonwelded silicic tuffs, but the heat and pressure of the intrusions caused welding, pore collapse, and attendant decrease in rock volume adjacent to the intrusions. This accounts for inward sagging of both the roofs and floors of the lopoliths.

En Echelon Dikes. The mechanisms responsible for the formation of the *en echelon* dikes at Paiute Ridge are not known, but based on their geometries and the stress history, we offer two possible models. The trend of the line connecting the *en echelon* segments varies from N to NE, in contrast to the NW-trending dikes discussed previously. It is possible that these segments represent intrusion during the transition from the SSW- to NW-directed σ_3 around 8.5 Ma (see above). If these dikes intruded during a rotating stress field, their orientation would reflect a plane normal to the direction of the incremental σ_3 as it changed from SSW to NW. Additionally, the rotation of principal stresses about an axis parallel to the propagation direction of the dike, as described by Delaney and Pollard (1981) and Pollard (1987), could have resulted in segmentation of those dikes. In the Paiute Ridge area, rotation would have been clockwise about a vertical axis near the final extent of dike ascension (at depths <250 m), resulting in right-stepping N-trending segments of NE-trending dikes (Figure 15). Alternatively, if the *en echelon* dikes formed in the same (non-rotating) stress field as did the other non-segmented dikes, it is possible that surficial loads could be responsible for the near-surface segmentation. Structures, topographic relief, or

intrusive bodies near the intruding dikes, could have induced a rotation of local principal stresses, resulting in dike segmentation. It is possible that further age constraints could reveal which of the two models is most likely.

5.3.2.6. Conclusions

The location and geometry of Miocene basaltic intrusive bodies in the Paiute Ridge area are strongly controlled by pre-existing structures and regional stresses. The longest dikes intruded normal to σ_3 , along pre-existing faults with critical dips and strikes: only NNW steeply dipping planes were injected. To achieve this geometry, magma pressures must have been greater than σ_3 , and also have exceeded friction across the fault surfaces. Driving pressures were apparently not sufficient to intrude along faults of other orientations nor along self-propagated paths. *En echelon* dikes trend at an angle to the fault-conforming dikes, and could have formed during the documented 65° stress-field rotation after ca. 8.5 Ma. Sills, which occur adjacent to dikes within 300 m of the paleosurface, consistently occur in the hangingwall block of the dike-injected faults. We interpret their location in Tertiary tuffs, just above the Paleozoic-Tertiary contact, to indicate that intruding magma sensed a significant local change in ambient stresses along the fault planes. Additionally, we suggest that the dip of bedding surfaces helped localize the direction of sill injection: magma preferred to flow upwards towards the east rather than downward toward the west. The two largest sills formed lopoliths indicating that the magma pressure exceeded the vertical stresses at that location and that the shallow level and large aerial extent of the sills allowed interaction with the free (Earth's) surface. We infer that sills at depths greater than those now exposed at the surface, would be less likely to form lopoliths, and that they would preferentially form in the Tertiary tuffs rather than in the Paleozoic sedimentary rocks. Finally, we speculate that the processes that governed dike propagation and local shallow sill formation at Paiute Ridge may operate in other shallow rift settings (e.g. continental rifts and rift zones of shield volcanoes, as has recently been addressed by Johnson (1995) who notes the importance of structure-related stress variations in stalling dikes in the east rift zone at Kilauea volcano).

5.3.3. Alteration of Silicic Tuffs Around Shallow Basaltic Intrusions

5.3.3.1. Introduction

This section reports studies of alteration of silicic pyroclastic rocks by shallow basaltic intrusions. Two sites were chosen for study because combined they are good analogs for the various rocks in the vadose zone at Yucca Mountain and allow us to constrain the type and spatial scale of alteration that would accompany a basaltic intrusion into the potential repository. The two sites are: (1) The Slanted Buttes area of Paiute Ridge, eastern Nevada Test Site. At Paiute Ridge variably vitric and zeolitized silicic tuffs (some of which correlate to those at Yucca Mountain, some 40 km away) were intruded by basaltic dikes and sills; (2) Grants Ridge, New Mexico. At Grants Ridge a thick sequence of unconsolidated rhyolite ignimbrite, fallout, and reworked volcanoclastic deposits were intruded by a basaltic plug that fed a scoria cone eruption. Erosion of the site has since produced a natural cross section through the scoria cone, its feeding system, and the pyroclastic host rocks. We gathered data to constrain geochemical and mineralogical alteration effects at both sites. The basic results are that alteration is quite limited, typically only found within 5-10 meters of the intrusions. The next phase of these studies is detailed numerical simulation of the thermal evolution of the sites, which would serve two purposes. The first would be a better understanding of the field-based observations and therefore more confident prediction of magmatic effects on potential repository performance. The second purpose would be for benchmarking numerical models of hydrothermal processes using these field examples that are at the same scale as the potential repository. This would lend confidence to engineering designs of thermal loading of the potential repository. Unfortunately, the studies have been terminated so that the modeling phase will not proceed.

5.3.3.2. Paiute Ridge/Slanted Buttes Analog Site

5.3.3.2.1. Summary

Vitric and zeolitized tuffs of middle Miocene age intruded by shallow high-alumina alkali basalt magmas developed localized contact metamorphic zones at Slanted Buttes in the northeastern part of the Nevada Test Site, Nevada. This study was part of a natural analog investigation to evaluate the effects of volcanism on the performance of a potential high-level radioactive waste repository environment at Yucca Mountain. Despite the similarity in size between the two dikes (~9-m wide), the contact metamorphic zones ranged from 0.5 to about 6 m. The dike responsible for the localized thermal zone (0.5 m) is about 2 km from a possible source plug, whereas the other one is within 100 m. The contact aureole beneath the 45-m thick sill is also localized and extends to about 5 m. The sill is an offshoot from a dike that is about 1 km away from a plug. Shallow emplacement into an unsaturated environment and loss of heat from the dikes and sill may be responsible for the localized nature of the contact metamorphic zones. Devitrification is enhanced in the presence of aqueous solutions.

Silicic and zeolitized tuffs within the contact aureoles are either devitrified and/or fused and quenched to form vitrophyres. Within the contact zone, the vitric content of the Ammonia Tanks Tuff of the Timber Mountain Group was totally replaced by feldspar, quartz, and cristobalite. The Tiva Canyon Tuff of the Paintbrush Group beneath the sill contains a vitrophyre and a devitrified zone side by side and a clinoptilolite-rich tuff occurs about 3 m away. Variations in major, trace, and rare earth elements are subtle except for water content.

Based on the natural analog studies, there is no indication for extensive hydrothermal circulation and alteration, brecciation and deformation related to magmatic intrusion, and vapor phase recrystallization during the magmatic intrusions into the vitric and zeolitized tuffs.

5.3.3.2.2. Introduction

We studied contact metamorphism of vitric and zeolitized tuffs intruded by basaltic dikes, sills, and plugs, in order to understand thermal, degassing, and volatilization effects on the host rocks. Contact aureoles form due to changes in a) volatile contents, b) mineralogic and isotopic compositions, c) non-volatile major and trace element compositions, and d) textural modifications (Barton et al., 1991). Contact aureoles around shallow magmatic bodies provide information about the alteration processes related to thermal, hydrothermal (metasomatic), and structural modifications that are generally manifested by coarsening, recrystallization, mineralogical and chemical changes, anatexis, and deformation and brecciation (Barton et al., 1991; Kerrick, 1991). Although loss of volatiles can be reversed if there is an appropriate source, other processes such as mobilization of certain major, trace and rare earth elements, loss of pore fluids, radiogenic argon, organic-derived components, and halogens can be irreversible. Such chemical and physical changes provide information on the thermal history of the host rocks and the spatial extent of the contact metamorphism. Here, we present results of field and laboratory studies that address the thermal effects of shallow basaltic dikes and a sill on vitric and zeolitized tuffs. We collected vitric and zeolitized silicic tuff samples within and outside contact aureoles to determine the physical and chemical modifications induced by the basaltic intrusions.

5.3.3.2.3. Geological Background and Field Study

The study site is located about 40 km northeast of Yucca Mountain along the northeastern part of the Nevada Test Site, Nevada, USA (Fig. 19). The Slanted Buttes ridge north and northwest of Paiute Ridge is intruded by dikes, sills, and conduit plugs, representing a shallow (100-200 m depth) intrusive plumbing system for a small volume volcanic center (Crowe et al., 1986). Due to erosion and faulting, only a small part of the eruptive lavas and scoria remain perched on top of vitric and altered silicic tuffs of the southwest Nevada volcanic field. The alkali basalt intrusions are late Miocene in age (8.7 Ma) and intruded into host rocks consisting of massive and bedded middle Miocene Calico Hills Formation, Paintbrush and Timber Mountain Group tuffs (Byers and Barnes, 1968; Crowe et al., 1986). According to Crowe et al. (1983a),

the basaltic intrusions are confined to a series of gently tilted fault blocks that are within a north-northwest-trending graben system of 15- to 20-km long and 4- to 8-km wide. Most of the faulting predates the basaltic intrusion in this area.

The shallow basaltic plugs, dikes, and sills were intruded along pre-existing structural weaknesses except for few that terminate at fault boundaries (Crowe et al., 1983a). The sill-like bodies are mostly local offshoots of the elongate dikes (Fig. 19). The scoria patches represent zones where pressure in the magma was reduced and volatiles exsolved, indicating that present exposures are very near to the original paleosurface. Some of the intrusion margins (e.g., plugs) also display intrusion breccias in which basaltic fragments are incorporated in a chaotic matrix of sintered to glassy hornfelsed tuff. The dikes cut across stratigraphic units, whereas most of the sill-like intrusions were emplaced along stratigraphic contacts. In most cases, the dikes occur radially around conduit plugs composed of jointed basalt, scoria patches, and partially fused and stoped silicic tuff blocks of the Timber Mountain tuff. Thicker sills have coarser (diabase) texture. Phenocrysts of the high-alumina alkali basalts consist of plagioclase, clinopyroxene, and iddingsitized olivine. Local zones of reddish, vesicular scoria occur within and near plug boundaries and intrusion margins.

Samples were collected from thermally altered zones of vitric and zeolitized tuffs. A number of samples (PR95-1 to PR95-15) were collected transverse to a 9-m thick dike from about 30 and 6 m on the north and south sides, respectively, where the dike was intruded into the middle Miocene Ammonia Tanks Tuff (11.6 Ma; Sawyer et al., 1994) of the Timber Mountain Group located along the southeastern part of the Slanted Buttes (Fig. 19, Location A). The ash flow is vitric and is matrix-supported, moderately welded, massive, pinkish brown, and pumice rich with cavernous weathering. Quartz, sanidine, and minor amounts of biotite are present as phenocrysts. The effect of the dike on the pumice-rich ash flow tuff is limited to about 4 m from the contact zone and it is represented by change of color (i.e., from pinkish brown to dark pinkish orange and light to dark gray), hardness, degree of fusion and recrystallization, and foliation and jointing parallel to the strike of the dike. The dike along the contact zone is strongly jointed parallel to the axis of the dike and forms thin plates that are easily broken and eroded creating a channel between resistant ledges of fused ash flow tuff. The dike is aphyric along the contact with the tuff while it is partially altered to smectite in the middle part.

Another suite of samples (PR95-18 to PR95-23) were collected beneath a 45-m thick sill that branches from a dike (Location B, Fig. 19). The base of the sill has a sharp contact with the underlying tuff. No breccias, stoped country rock or lithic fragments were noted along the contact. Similar to the dike, the base of the sill is jointed and forms thin plates parallel to the contact. The tuffs at this locale belong the Paintbrush Tuff and the Calico Hills Formation, based on petrographic and major and trace element chemical data.

The contact aureole below the sill is about 5-m thick and is represented by a 0.5-m thick vitrophyre (PR95-18) that laterally changes to an orange brownish baked and hardened tuff (PR95-21). This unit was identified as the Tiva Canyon Member of the Paintbrush Group based on petrographic and chemical data. A hornfelsic zone that contains a 10-cm thick light green opaline layer of concretions (PR95-19) crops out about 1 m below the vitrophyre. A baked tuff also occurs between the vitrophyre and the sill. About 3 m beneath the sill, a tuff (PR95-20) that is massive, pinkish, moderately welded contains silicified clasts (pumice?), abundant obsidian fragments, welded pumice, and stretched vesicles. This tuff also belongs to the Paintbrush Group. The pinkish tuff grades to a poorly welded zone that also contains a greenish opaline layer like an overlying outcrop closer to the base of the sill. A light gray bedded tuff (PR95-23) with abundant lithic fragments crops out about 35 m below the base of the sill. The tuff contains abundant quartz with minor amounts of feldspar and biotite and was correlated to the Calico Hills Formation on the basis of mineralogy and major and trace element chemistry. Fractures and joints within the contact

aureole are filled with opal and minor calcite, whereas the joints at the base of the sill are devoid of secondary minerals.

Four samples were collected from a 0.7-m thick poorly consolidated, coarse and fairly matrix-free pumice fall that was intruded by a 9-m wide dike at the south side of the northern part of the Slanted Buttes (Fig. 19, Location C). The dike forms a resistant ledge along the contact with the host rock, whereas the middle section is altered and eroded and forms a deep erosional channel (>1.5-m deep) parallel to the strike of the joints and the dike. A sample from the dike (PR95-28) is partially altered and contains abundant amygdules of silica. The bedded pumice fall occurs below the Rainier Mesa Tuff and probably belongs to the Paintbrush Group (12.8-12.7 Ma, Sawyer et al., 1994). The contact aureole is about 0.5-m wide and gradational. The tuff along the contact zone is fused and glassy. It is about 20 to 30-cm wide and a sample (PR95-24) from this zone is hard, glassy, dark gray, and generally foliated parallel to the strike of the dike. It also contains opaline nodules of altered pumice clasts. It grades to partially recrystallized pumice (PR95-25) where the incipient bedding is still preserved. Two more pumice fall samples collected about 1.3 (PR95-26) and 1.8 m (PR95-27) from the dike show no apparent thermal effect from the intrusion. They are friable and bedded with millimeter-thick laminations.

5.3.3.2.4. Analytical methods

Bulk samples were pulverized and analyzed using X-ray diffraction (XRD) on an automated Siemens D-500 diffractometer using Cu Ka radiation, 0.02° 2θ steps, and counting times of 1 s per step for all mounts ($2-36^\circ$ 2θ) to determine changes in mineralogy in the host rock during contact metamorphism. Quantitative mineralogic analyses were also conducted on select samples (Chipera and Bish, 1995). The major and trace element compositions of bulk powders were analyzed using a Rigaku 3064 wavelength dispersive X-ray fluorescence (XRF) and the intensities were processed to concentrations using XRF-11 to determine changes in bulk-rock compositions (metasomatism) due to fluid-rock interaction in a contact aureole. Analyses and statistics of the unknown samples are based on a model that uses intensities for 21 standard rocks.

Selected samples were analyzed for F, Cl, B, Br, S, and P_2O_5 using an Ion Chromatograph Dionex 4000i with conductivity and UV/Vis detectors and other accessories to estimate volatile components gained and/or lost during contact metamorphism (Shevenell and Goff, 1993). Bulk samples were analyzed from within and outside the contact metamorphic zone. Standards that were lithologically similar to the unknown samples were also analyzed to determine the quality of the analytical work.

The effect of contact metamorphism on the rare earth element compositions of vitric and zeolitized tuffs was also determined by instrumental neutron activation analysis (INAA) on selected samples at Washington University, St. Louis (Korotev, 1978). Two NBS standards were included with the unknown samples.

5.3.3.2.5. Analytical Results

X-ray diffraction analysis. Changes in mineralogical compositions were noted in the Ammonia Tank Tuff samples collected at different intervals on both sides of the 9-m wide basaltic dike (Fig. 19, Location A). These results are presented in Table 5-4 and Figure 20. The silicic samples generally exhibit a trend of decreasing volcanic glass contents and increasing amounts of feldspar, quartz, and, cristobalite toward the dike (Fig. 20). However, a sample closer to the dike (~ 3 m) on the north side of the dike contains about 40 % clinoptilolite, 4 % of cristobalite, and a higher content of feldspar (Fig. 20b). On the south side, samples collected within about 6 m of the dike do not contain volcanic glass (Fig. 20f). Trace amounts of clinoptilolite are also present in one of the samples collected on the south side of the dike. Recrystallization and alteration within the dike is also noticeable because of the increased content of cristobalite and smectite in the basaltic samples collected along the contact zone with the tuff (Fig. 20e). Based on the

mineralogical composition, the effect of the dike is much more apparent in samples collected within about 2.0 m from the contact (Table 5-4). For example, the volcanic glass content systematically decreases from about 60% at about 15 m to none about 2 m away from the contact. Conversely, feldspar, quartz, and cristobalite contents increased from 45 to 69 %, 4 to 9%, and 0.4 to 13%, respectively. Physical changes in color and hardness, absence of volcanic glass, and the increase in quartz, feldspar, and cristobalite contents toward the dike are important indicators of the lateral extent of the contact metamorphic aureole.

There is no systematic mineralogic and volcanic glass variation in the altered Tiva Canyon Tuff within the contact aureole created beneath the 45-m thick sill about 0.6 km northwest of the previous dike (Fig. 19, Location B). Unlike the Ammonia Tank Tuff intruded by a dike, the older tuff beneath the sill indicates variable contact metamorphic effects (Fig. 21, Table 5-4). The tuff (PR95-18) along the contact zone is fused and quenched and contains about 80% volcanic glass and 6 % clinoptilolite (Fig. 21a), whereas another sample (PR95-21) of the same unit collected along the contact about 10 m west of the previous sample is baked and hardened and contains no glass but the same amount of clinoptilolite and higher contents of feldspar, quartz and cristobalite (Table 5-4, Fig. 21d). About 1.5 m beneath the sill, a greenish layer (PR95-19) parallel to the bedding of the tuff and the contact with the sill is enriched in quartz, feldspar, and opal (Fig. 21b). However, about 3 m below the sill, another sample (PR95-20, Fig. 21c) contains abundant volcanic glass (65 wt %), whereas two other samples collected at about 8 (PR95-22) and 35 m (PR95-23) have no volcanic glass (Fig. 21e). Instead, these two samples belonging to the Paintbrush Group and the Calico Hills Formation, respectively, are dominated by clinoptilolite (Table 5-4).

More samples were collected from a pumice fall that was intruded by a 9-m thick dike along the northwestern extension of the dike system that disappears into the south side of the Slanted Buttes (Fig. 19, Location C). All samples collected along the contact zone and about 1.8 m west of the dike contain about 80% volcanic glass (Table 5-4, Fig. 22a,b). The sample along the contact is totally fused and hard with abundant spherical glass blebs most of which are still intact. It is characterized by higher contents of quartz and cristobalite compared with samples collected further from the dike (Table 5-4).

Geochemical Results. Variations are noted in major and trace element compositions of the various vitric and zeolitized tuff samples intruded by dikes and a sill. The Ammonia Tanks Tuff samples (PR95-1 - PR95-10 and PR95-12 - PR95-15) exhibit slight increase in SiO_2 , K_2O , Na_2O , TiO_2 , and Fe_2O_3 contents toward the dike, whereas the loss-on-ignition (LOI) shows the opposite trend (Table 5-5, Figures 23, 24). Al_2O_3 and MgO are minimally affected (Figs. 23c and 23f). Although within the error range of the analysis, the high-alumina basalt sample from the contact zone shows subtle depletions in most of the major elements compared to another sample collected from the middle of the dike. Depletion and enrichment of Na_2O and LOI, respectively (Figs. 23b and 23e) are noted in the basalt specimen from the contact zone.

The Ammonia Tanks Tuff shows insignificant variations in its chondrite-normalized trace element and rare earth element (REE) concentrations except for Ba, K, Rb, Sr, Tb, and Yb (Figs. 24a,b, Table 5-6). Except for Eu, Nb, and Tb enrichment, the basalt sample from the middle part of the dike contains lower trace element and REE contents compared with the Ammonia Tank Tuff (Figs. 24a,b).

The major and trace element compositions of the vitric and zeolitic tuffs beneath the sill are variable (Fig 23). The layer directly below the sill is the Tiva Canyon Tuff, the uppermost unit of the Paintbrush Group and is partially fused and transformed to a 0.5-m thick vitrophyre along the contact zone. It contains higher contents of major and trace elements and REE compared with the underlying opaline and zeolitized tuffs below it (Figs. 23,24). The vitrophyre has different major and trace elements compositions compared with the underlying samples (Tables 5-5, 5-6). For

example, an opalized Tiva Canyon Tuff sample collected about a meter below the vitrophyre is enriched in silica (79.4 % wt), depleted in Al_2O_3 (8.8% wt) and TiO_2 (0.79% wt), and contains lower content of LOI compared with the other samples collected above and below it (Table 5-5). This is also true for the trace elements and the REE contents except for Co and Ba (Table 5-6). Unlike the major, trace, and REE contents, the LOI generally appears to increase away from the sill (Fig. 23e).

Samples from the virophyre and a tuff about 3 m below it were also analyzed for B, Br, Cl, F., Li, P_2O_5 , and S because these elements along with water form the bulk of volatile contents escaping during volcanic eruptions (Table 5-6). The vitrophyre sample contains slightly higher B, Cl, and S, whereas the sample furthest from the sill is enriched in F, Li, and P_2O_5 (Table 5-6).

Samples from the pumice fall beneath the Rainier Mesa Tuff intruded by a 9-m thick dike exhibit insignificant modifications in their major and trace element compositions like the other tuffs intruded by a sill and a dike along the eastern part of the Slanted Buttes. Except for Na_2O and Zr contents that decrease away from the contact zone, there is no systematic geochemical variation with reference to the dike (Fig. 23, Table 5-5).

5.3.3.2.6. Discussion of Physical, Mineralogical, and Chemical Changes within the Contact Aureoles. The tuffs within the contact aureoles of the dikes and sill at Slanted Buttes, exhibit variable physical and chemical changes although the intruding magmatic bodies have similar basaltic compositions (Table 5-4). For example, the volcanic glass in the dike-intruded Ammonia Tank Tuff was totally recrystallized to quartz, feldspar, and cristobalite within an asymmetrical contact aureole (Fig. 19, Location A). Despite the thickness (≥ 45 m) and size of the sill, the contact aureole below the sill is not wider compared with the same kind of zone developed adjacent to the 9-m thick dike in the southeastern part of the Slanted Buttes. The Tiva Canyon Tuff beneath the sill is partially recrystallized into a quartz and feldspar-rich zone in one location and into a 0.5-m thick dark brown and densely welded vitrophyre a few meters away (c. 10 m). The contact aureole below the sill is about 5-m wide or less (Fig. 19, Location B).

The poorly-compacted but well-sorted pumice fall along the southern part of the Slanted Buttes was also fused to a light gray vitrophyre characterized by abundant and intact spherical glass blebs and recrystallized coarse quartz phenocrysts. Here, the contact aureole is less than a meter wide although the dike is as wide as the other one from the eastern part of the Slanted Buttes.

Processes related to hydrothermal circulation and alteration appear to be minimally developed in the intruded country rocks. Brecciation or stoping beneath the sill and adjacent to the dikes where the samples were collected is minimal except around the plugs. The mafic intrusions along the contact are characterized by narrowly-spaced joints that are parallel to the strike of the intrusive body.

The dike along the south eastern part of the Slanted Buttes (Location A, Fig. 19) is within 100 m from a plug, whereas the sill (~1 km) and the other dike (>2 km) are farthest from the plug area and this may explain for the localized nature of the contact metamorphic aureoles. According to Delaney and Pollard (1982), insignificant contact metamorphic zones around dikes and sill may be attributed to shallow emplacement and higher cooling rates of these thin magmatic bodies. The absence of compositional variation in the basaltic rocks probably suggests minimal residence time, rapid ascent, and emplacement. Dikes generally exhibit compositional and textural zonations because of multiple injections of magma during dike intrusions (Delaney and Pollard, 1982). However, this is not the case in the dikes and sill of the Slanted Buttes area because they generally exhibit similar chemical compositions within a dike (i.e., edge versus center of dike) and between dikes and sills, and flows in the plug that are widely separated from each other (Crowe et al., 1986). Moreover, the dikes along the contact with the host tuffs are vesicular indicating the escape of volatiles, a possible evidence for shallow emplacement.

Although the tuffs were intruded by the same high-alumina alkaline basalt magma, the effects of contact metamorphism were variable. For example, contact metamorphism of the Ammonia Tank Tuff is represented by a localized devitrified zone. However, the Tiva Canyon Tuff within the contact aureole is characterized by a devitrified zone adjacent to a fused and quenched glass. Experimental studies indicated that the rate of devitrification of natural rhyolitic glass significantly increased in the presence of water and alkali-rich solutions (Lofgren, 1970). According to Lofgren (1970, 1971), the devitrification process involves hydration followed by devitrification whereby water diffuses through silicate glass forming hydroxide ions (OH^-), breaking the bridging of Si-O bonds by OH^- , and making available the newly freed SiO_4 tetrahedra for the formation of quartz and feldspar crystals. The hydration and devitrification rates increased one to five orders of magnitude in the presence of Na and K compounds compared with pure water. Lofgren (1971) also recognized in his experimental runs the existence of a devitrification zone clearly marked behind hydration fronts. He attributed this incomplete devitrification to lack of sufficient water and/or alkalis or low temperature ($\sim 240^\circ\text{C}$) and pressure (0.5 kb) where nucleation ceased as the supply of water diminished. The limited devitrification processes noted in the host rocks in our study may be related to the intrusion of the basaltic magmas into an unsaturated environment.

The mineralogical and geochemical data indicate the extent of physical and chemical interactions between the intruding basaltic magmas and the vitric and zeolitic tuffs. The dikes and sill show variable physical modification represented by minor deformation in the form of narrowly-spaced jointing parallel to the strike of the dikes or sill, contact welding, vesiculation along the contact with the host rock, and alteration. The dikes along the contact zone are generally altered and break into thin plates of the same thickness and dimensions. Although the dikes intrude tuffs of different ages and compositions, the high-alumina alkali basalts exhibit minor chemical interaction with the host tuffs. Comparison of basaltic specimens from the edge and center of a dike and a plug about 100 m to the south (Fig. 19, Location A) exhibits minor variations in major and trace element compositions. For example, the sample from the plug is less altered with lower SiO_2 , K_2O , LOI, Cr, Ni, Rb, and Ba and higher Fe_2O_3 , MgO, Na_2O , and Zn compared to the dike samples (Table 5-5). A basalt sample from the contact with the Ammonia Tank Tuff is partially altered and contains smectite and somewhat higher LOI, SiO_2 , V, and Zn and lower CaO, Na_2O , K_2O , Ni, Sr, Y, and Zr compared with another sample from the center of the same dike (Table 5-5). Basalt specimens from the contact with the tuffs have higher LOI contents (3.43-4.4 % wt) compared with samples collected more than 4 m away (1.62-1.89% wt) from the basalt-tuff contact. The geochemical results suggest that there was some transfer of elements to and from the dike and the host tuff during the magmatic intrusion. For example, the fused tuff was dehydrated, whereas the basalt was hydrated. The basalt contains a slightly higher content of silica consistent with a devitrification process, whereas the adjacent tuff inherited minor amounts of Ni along with the recrystallization of volcanic glass to feldspar and silica.

The host tuffs adjacent to the dikes are fused and massive with none of the platy jointing found in the adjacent basaltic intrusions. Moreover, there is no breccia zone developed at the sampling locations except for a narrow baked and fused zone. According to Delaney (1980), pore fluid pressure increases due to heating in impermeable rocks like shales and siltstone during dike intrusion results in higher effective stresses that possibly lead to brecciation. Hydraulic rupture related to hydrothermal brecciation triggered by boiling and rapid release of confining pressure of fluids in fractures and intergranular pores was documented in subsurface core samples obtained during the experimental Continental Scientific Drilling Project in the Valles caldera, Jemez Volcanic field, New Mexico (Hulen and Nielson, 1988). Thus, the absence of breccias around the dikes and sills at the sampling points may be related to the permeable nature of the host tuffs, absence of pore fluids, heat loss from the dikes and sill and short duration of heat transfer to the host rocks,

and possibly due to shallow emplacement of the dikes along preexisting fault zones and fracture systems.

Hydrothermal activity related to magmatic intrusion into vitric and zeolitized tuffs in the Slanted Buttes area of the NTS appears to be insignificant except in the immediate contact zone which is generally less than five meters wide around the 9-m thick dike and beneath the 45-m thick sill. This is clearly indicated by the mineralogy and the chemical compositions of the intruded tuff samples. The devitrification process of the Ammonia Tank Tuff appears to have taken place in unsaturated environment because of the localized nature of the contact aureole and the similarity of the alkali contents of the samples from within and outside the contact aureole. According to Lofgren (1970), the devitrification rate is increased one to five orders of magnitude in the presence of alkali solutions compared with pure water. Generally, this is reflected in the increased potassium or sodium contents of the feldspar minerals. Thus, the insignificant difference in chemical compositions between the tuffs within and outside the contact aureole indicates minimal hydrothermal activity. The B, Br, Cl, F, Li, S, and P_2O_5 contents of samples from the Ammonia Tank Tuff outside the contact aureole and a recrystallized basalt from the contact with the tuff shows variable enrichments in certain elements characteristic of silicic and basaltic rocks except for S and F contents (Table 5-6). The results cannot be used to assess the effect of contact metamorphism because the two samples are more than 30 m apart. However, two samples from the contact aureole beneath the sill show no apparent trend in their B, Br, Cl, F, Li, S, and P_2O_5 contents although these elements are generally sensitive to elevated temperatures. The vitrophyre (PR95-18) is directly beneath the sill and contains higher contents of Cl, B, and S, whereas the other sample (PR95-20) which is about 3 m below it has higher F (Table 5-6).

Different processes operated simultaneously during the intrusion of the sill into the Tiva Canyon Tuff compared with the changes observed in the Ammonia Tanks Tuff within the contact aureole. For example, mineralogic analysis of the Tiva Canyon Tuff beneath the sill suggests that the tuff along the contact was totally devitrified while a few meters away (~10 m), the same unit was altered into a 0.5-m thick densely welded vitrophyre with about 80% volcanic glass (Table 5-4, Fig. 21). The presence of minor amounts of clinoptilolite in the vitrophyre and devitrified tuff suggests that the Tiva Canyon Tuff was zeolitized before the intrusion. The clinoptilolite grains are remnants of a pre-intrusion alteration process within the contact aureole rather than post-intrusion diagenetic phases. Because the volcanic glass was destroyed by devitrification there was no raw material for clinoptilolite diagenesis after the basaltic intrusion. About a meter below the vitrophyre, the tuff (PR95-19) is banded with greenish layer that is replaced by opal. The cristobalite probably formed by devitrification after original deposition of the tuff rather than from the effect of the sill. No volcanic glass is present in the lowermost two samples that are about 8 m and 35 m beneath the sill. These altered tuffs are dominated by secondary minerals such as clinoptilolite (69-89%), opal-CT 28%), and cristobalite (2%). From the mineralogic and geochemical data of the Tiva Canyon Tuff below the sill, it appears that the volcanic glass was altered to clinoptilolite and opal-CT prior to the intrusion of the sill. The occurrence of a 10-cm thick opaline zone about a meter below the vitrophyre corroborates this suggestion. It is long established that hydrolysis of glass raises the silica activity (a_{SiO_2}), pH, and activity ratios of $(Na^+ + K^+)/H^+$ creating a conducive environment for the crystallizations of smectite, zeolites, opal-CT (Hay, 1995). Although the upper stability limit of clinoptilolite and opal-CT is about 100°C (Levy and O'Neil, 1989; Iijima and Ogihara, 1995), these secondary minerals exhibit no apparent effect from the sill. During burial diagenesis, clinoptilolite alters to analcime and to albite in the temperature ranges of 76°C - 91°C and 120°C - 124°C, respectively (Iijima and Ogihara, 1995). However, clinoptilolite alters to analcime in the presence of Na^+ cation as has been well documented in saline, alkaline-lake deposits (Surdam and Sheppard, 1978; Moncure et al., 1981). X-ray diffraction analysis of samples from the contact aureole indicates no analcime. Lack of Na^+ cation may be a possible explanation for the absence of analcime and albite in the altered tuffs within the contact aureole. In fact, devitrification rate is higher in the presence of alkali solutions

compared with pure water (Lofgren 1970) and this may explain the localized nature of the contact aureole below the 45-m thick sill. The alkali contents (Na_2O and K_2O) of the Tiva Canyon samples from the contact aureole are depleted except for a sample from the vitrophyre zone and this may be related to loss of alkalis during zeolitization (Table 5-5). Generally the alteration of volcanic glass is accompanied by the release of alkali cations and silica (Hay, 1995). The abundance of clinoptilolite in the lowermost samples from the Paintbrush Group and the Calico Hills Formation tuffs suggests that thermal effect from the sill was minimal.

The dike-intruded Ammonia Tank Tuff within the contact aureole lost all its volcanic glass due to recrystallization; however, a pumice fall at the east side of Slanted Butte does not show the same kind of mineralogical modification although it was intruded by a basalt of the same composition and thickness. The occurrence of insignificant alteration around the dike may be related to low-temperature and insufficient fluids or volatiles associated with the intruding basaltic magma. The dike at this location is more than 2 km from the possible source plug (Fig. 19, Location C).

5.3.3.2.7. Implications for Magmatic Hazards and Potential Repository Heat at Yucca Mountain

The natural analog studies of magmatic intrusions are aimed at understanding the range of subsurface processes that may occur in a potential repository system. This includes the potential repository itself and the surrounding natural barriers. Subsurface processes can be divided into a) magmatic intrusions and eruptions and related thermal perturbation associated with a magmatic intrusion (short term) and b) those that occur over a long term as the hydrologic system is re-established with altered hydrologic and mineralogical properties (Valentine et al., 1993). The long term effects include the possibility of perched water near low-permeability intrusive bodies (e.g., dikes and sills), possible fast paths along intrusion-induced fractures, and reduced chemical retardation properties of the country rock resulting from hydrothermal alteration. Results of field studies indicate the extent of alteration processes related to small and shallow basaltic intrusions into vitric and zeolitized tuffs. Such results are important for performance assessment of basaltic magma intrusions into or close to a potential repository environment. The field studies also provide data that can be used to evaluate the effects of heat from decay of radioactive waste in a potential repository environment. The field and laboratory data from the natural analog studies could be used to build confidence and intuition and to calibrate modeling of such processes with similar physical and chemical environments if the need arises (e.g., contact metamorphism from heat generated from a high-level radioactive waste in a potential repository environment).

The field and analytical data from the Slanted Buttes, located about 40 km northeast of Yucca Mountain, provide information about the contents of lithic fragments in the intrusive bodies, volatile contents and thermal and degassing effects of the dikes and sill, and degree of alteration of host rocks adjacent to the basaltic intrusions. The sill and dikes were possibly intruded into a shallow unsaturated environment a few hundred meters from the paleosurface as indicated by reconstruction of stratigraphy above the intrusions (Valentine et al., 1993; Ratcliff et al., 1994). This is consistent with the limited nature of the contact metamorphic aureole and devitrification and the presence of vesicles along the edges of the basaltic intrusions. The depth of intrusion may be comparable to the potential repository environment at Yucca mountain. The Slanted Buttes sites can be used as natural analogs to delineate the different geologic processes that could be potentially initiated around a high-level radioactive waste if a basaltic magma intrudes into or close to a potential repository.

Based on these studies, there is no indication for extensive hydrothermal transfer from the basaltic intrusions into the host silicic and zeolitized tuffs. Field and analytical evidence from the localized contact metamorphic aureoles and devitrification of the silicic tuffs adjacent to the intrusion, minimal hydrothermal alteration, and presence of low-temperature minerals at close proximity to the intrusions suggest insignificant effects of the shallow basaltic intrusions on the

vitric and zeolitized tuffs. In fluid-deficient systems the rate and extent of water-rock interactions are significantly diminished (Esposito and Whitney, 1995). Although there are remnants of scoria deposits that suggest explosive eruptions, the host rocks adjacent to the sills and dikes do not record the effects of volatiles or water. If the field data are consistent with the emplacement of the intrusive bodies into unsaturated vitric and zeolitized tuffs, the volatile contents of the basaltic intrusions were insignificant. In fact, the absence of widespread devitrification of the tuffs along the contact zone appears to reflect lack of aqueous solutions (Lofgren, 1971). Fracture zones in the host rocks are filled with opal and calcite. These secondary minerals are possibly related to zeolitization processes that occurred prior to the basaltic intrusions.

Volatile contents of the basaltic intrusions (e.g., plugs, dikes, and sills) appear to have been small consistent with the localized nature of the contact metamorphic zone. Planar joints developed within the basalt intrusions along the contacts are devoid of secondary minerals. Thermal and degassing effects related to the basaltic intrusions are also insignificant as indicated by the higher of B, Cl, F, and S contents of samples within the contact aureole. This may be related to the limited nature of hydrothermal circulation and alteration developed around these basaltic bodies consistent with intrusions into unsaturated environment.

5.3.3.2.8. Conclusion

Localized contact metamorphic zones were developed in vitric and zeolitized tuffs intruded by high-alumina alkali basalt magma. The dikes are about 9-m wide, whereas the sill is about 45-m thick. The dikes were intruded into vitric tephra of the Ammonia Tanks Tuff of the Timber Mountain Group and units of the Paintbrush Group. The sill which is an offshoot of a dike that is about 1 km from a plug intruded into a zeolitized Tiva Canyon Tuff. The sizes of the contact aureoles around the intrusions are variable and range from about 0.5 to 6 m. Although the tuffs are poorly consolidated and permeable, hydrothermal circulation and alteration related to the intrusions were not widespread. The localized nature of these processes may be related to rapid loss of heat from the intrusive bodies, shallow emplacement into an unsaturated environment, and lack of fluids within the intruded tuffs. Devitrification of tuffs is enhanced in the presence of aqueous solutions and alkalis.

Mineralogical modifications within the intruded tuffs indicate the extent of contact metamorphism. Volcanic glass in the Ammonia Tank Tuff is totally recrystallized to feldspar, quartz, and cristobalite within the contact zone. Major, trace, and rare earth element compositions are subtly modified compared with the mineralogy. Water content is clearly modified in the basalt and the adjacent tuffs. Unlike the Ammonia Tanks Tuff, a pumice fall intruded by a dike of the same size formed a fused glassy tuff with about 80% glass and with no apparent variation in chemistry and mineralogy compared with a sample collected outside the contact aureole. The Tiva Canyon Tuff below the sill was devitrified and also altered to a vitrophyre within a short distance directly below the sill.

Based on the field and laboratory results, the effect of the basaltic intrusions was confined to the immediate contact zone and there were no widespread hydrothermal circulations and alterations in the poorly-welded tuffs. The occurrence of low-temperature clinoptilolite and opal also suggests that thermal transfer into the adjacent country rock was minimal. The lack of widespread brecciation around the dikes and sill may be consistent with intrusion into a shallow unsaturated environment. The natural analog studies at Slanted Buttes could be used to delineate the different geologic processes that could be potentially initiated around a high-level radioactive waste if a basaltic magma intrudes into or close to a potential repository or heat generated from a potential high-level radioactive waste repository.

5.3.3.3. Grants Ridge Analog Site

5.3.3.3.1. Introduction

We studied contact metamorphic effects on poorly consolidated late Pliocene ash flow tuffs and overlying volcanoclastic layers that were intruded by a basaltic plug and capped by thick scoria deposits. The study site is located in west-central New Mexico within the southwest confines of the Pliocene Mt. Taylor volcanic field (Fig. 25). The area is characterized by northeast-trending discontinuous mesa. The plug responsible for the widespread basaltic lavas capping the mesa is located in the northeastern part of this discontinuous mesa (termed East Grants Ridge) about 12 km southwest of Mt Taylor.

Systematic studies of contact metamorphic processes require the understanding of hydrothermal activity, structural deformation, brecciation, thermal, degassing, and volatilization effects on the host rocks during magmatic intrusion. The East Grants Ridge site provides an ideal geologic section where a thick (~150 m wide) basaltic plug intruded into vitric, unwelded, porous silicic rocks. We collected samples at different locations from the plug and the host rocks to map the aerial extent of alteration related to the intrusion. The effects of an intrusion on the host rocks are generally manifested by physical changes such as compaction and change in color. Moreover, such effects could be mapped by geochemical modifications that lead to enrichments and depletions in the host rocks due to recrystallization and hydrothermal activity. Widespread hydrothermal activity can develop in such an environment provided there is sufficient water. Deformation and brecciation of the host rocks caused by the intruding magmatic body create radial fracture systems that provide important fluid pathways.

Here, we present results of our field and laboratory studies. Mineralogical and geochemical data on bulk and pumice samples from within and outside the contact metamorphic zone provide information on the behavior of major, trace, rare earth element, and gaseous volatile contents of these rocks.

5.3.3.3.2. Geologic Background

East Grants Ridge is made up of basaltic and silicic rocks that are unconformably exposed on top of sandy shale and silty sandstone units of the Upper Cretaceous Mancos Shale (Kerr and Wilcox, 1963; Thadden et al., 1967). These rocks are well exposed along the east side of this north-east trending discontinuous mesa (Fig. 25). Initial volcanic activity at East Grants Ridge was represented by pyroclastic eruption that was followed by dome-building felsite, obsidian, and perlite lavas. The silicic tephra is represented by crystal-poor ash flows, pumiceous falls, and surge deposits that are mostly exposed along the southwest flank of the rhyolite-obsidian-perlite dome complex. According to Kerr and Wilcox (1963), the lava dome has a crystal-poor rhyolite core surrounded by flow-banded obsidian and perlite flows that hydrated under magmatic conditions. The transition from a rhyolite core to an obsidian zone is gradational with no interruption in the flow banding, suggesting contemporaneous emplacement (Kerr and Wilcox, 1963). Vapor-phase crystallization is represented by the deposition of tridymite, topaz, and garnet in lithophysae within the rhyolite and obsidian flows. The silicic rocks of East Grants Ridge were classified as topaz rhyolites because of high concentrations of F (0.5 %), Li, and Rb and the occurrences of vapor phase garnet, topaz, and tridymite in lithophysae in the rhyolite lavas (Christiansen et al., 1983). K/Ar dating of rhyolite from the dome complex yielded apparent ages of 3.3 Ma (Lipman and Mehnert, 1979).

The silicic tephra sequence of East Grants Ridge is covered by 0.5-2 m thick lithic-rich reworked and partially baked volcanoclastic sedimentary unit that contains lenses of coarse pebbles. This unit is poorly consolidated and sorted and exhibits graded bedding. However, the upper 20 to 30 cm of this orange sedimentary layer is moderately consolidated because of baking effects of the overlying scoria deposit. Although the volcanoclastic layer crops out on both sides of the plug, its lateral exposure is discontinuous because of talus and landslides accumulations from the overlying scoraceous deposit. A localized outcrop of this reworked tuff was identified and

sampled about 300 m on the northeast side of the plug. The volcanoclastic unit is totally fused and brecciated adjacent to the plug (Fig. 26, samples 5a and 5b inset map).

The silicic pyroclastic deposit, rhyolite and perlite lavas, and the volcanoclastic unit at the southwest side of East Grants Ridge are overlain by basaltic lavas and thick scoria deposits (Fig. 25b). The silicic pyroclastic and volcanic rocks are intruded by the 150 m thick basaltic plug that supplied lava and scoria to the surface. Essentially the study area represents a scoria cone (with attendant lava flows) that has been dissected by erosion and possibly faulting. A cross section of the cone, its feeder plug, and the wall rocks (silicic pyroclastic and volcanoclastic deposits) is exposed in the steep flank of Grants Ridge. The plug is obscured at the base of the slope by talus (Fig. 26). A localized contact metamorphic aureole formed around this vertically-jointed basaltic plug. The contact aureole is partly obscured because of talus and landslide blocks from the overlying scoria deposit and from the spalling of the jointed plug wall.

The olivine basalt lava flows and scoria beds contain fused rhyolite and pumice clasts. These basaltic lavas cover a wide area and form flat mesa tops of the Grants Ridge, West Grants Ridge, and Black Mesa (Fig. 25a). According to Kerr and Wilcox (1963), the basaltic eruption culminated with the formation of several cones of basaltic scoria. These scoria cones are exposed above the plug and are intruded by sills and dikes (Fig. 25b). Thick flows of bedded scoria deposits occur close to the vent area. Most of these scoria beds thin out and disappear within a short distance from the vents. The plug is surrounded by agglutinated scoria deposits and the basaltic lavas exposed above the silicic tuffs occur about 100 m west of the plug.

Although the basaltic flows from East Grants Ridge were not dated, K/Ar ranges of 2.9 to 1.6 Ma were reported from lavas of the Mt. Taylor volcanic field (Lipman and Mehnert, 1979) (Fig. 25a). A K/Ar age of 2.57 Ma was reported on a basalt flow from Black Mesa, located along the southwest part of Grants Ridge (Laughlin et al., 1994; Fig. 25a). Intermediate to felsic rocks with similar apparent ages (3.73-2.01 Ma) were also reported from the Mt Taylor volcanic field to the northeast of Grants Ridge (Perry et al., 1990).

5.3.3.3.3. Field Studies

Samples were systematically collected from the basaltic plug, silicic ignimbrite, and volcanoclastic unit on both sides and at different distances from the intrusive body located at the southwestern part of East Grants Ridge (Fig. 26). The ash flow tuff samples consist of matrix and pumice clasts from each location. Sampling was dictated by the availability of outcrops and accessibility because of steep cliffs and slopes created by pumice quarries and the accumulation of talus and landslides around the plug, respectively. As indicated in Figure 26, sample points are distributed from the contact zone to about 300 m away on both sides of the plug.

Four basaltic samples were collected from different parts of the plug (Fig. 26). Two of the samples located on the west (GR95-4) and east (GR95-9) sides of the plug were within a few meters of the contacts with the silicic tuffs and the volcanoclastic layer. Here, the plug is sparsely porphyritic with plagioclase phenocrysts and coarse pyroxene crystals. Close to the contact zone, the plug is fractured, poorly jointed, and vesiculated. The other two samples (GR95-13 and GR95-16) were collected from the central part of the plug away from the contact metamorphic zone. This part of the plug is moderately porphyritic with an aphyric matrix, massive, and vertically jointed.

Most of the samples for this study were collected from the ash flow tuff because the focus of this study was to evaluate the thermal effects of the basaltic intrusion on the poorly-welded pumice-rich pyroclastic rocks. The contact metamorphic zone around the approximately 150-m wide plug is not extensive and may be less than 5 m wide. The ash flow and pumice clasts along the contact zone were fused and vesiculated, but only welded within about 1 m of the contact. The fused pumice clasts are characterized by large spherical vesicles that are held together by reddish

fused matrix. The fused pumice and the matrix of the ash flow tuff formed a honey-comb-like mesh. A total of six samples were collected from the contact zone along the west (GR95-1a, GR95-1b, GR95-2, and GR95-3) and east (GR95-8a and GR95-8b) sides of the plug. However, the pumice clasts (GR95-8a) from the east side of the contact zone are not fused like those from the western part of the aureole (GR95-1a). More ash flow tuff and pumice clasts were taken from the west side at about 25 m (GR95-6a and GR95-6b), 45 m (GR95-7), and about 300 m (GR95-15a and GR95-15b) (Fig. 26). A large pumice quarry exposed a section of the ash flow tuff 40 to 50 m west of the plug. No alteration effect was noted in the exposed walls of the quarry except for localized fusion of the tuff around a dike that intruded into the tuff. The quarry wall closest to the plug is pumice-rich, soft, and unconsolidated. Two of the samples came from the eastern (GR95-7) and western (GR95-15a and GR95-15b) walls of this abandoned open pit mine to determine for signs of devitrification. Two more samples (GR95-5a and GR95-5b) were obtained from the volcanoclastic layer directly below the thick scoria deposit about 30 m west of the plug (inset map, Fig. 26).

More samples were collected from the east side of the plug. The pumice-rich ash flow tuff (GR95-8a and GR95-8b) and the overlying volcanoclastic sediment (GR95-21) samples from the contact zone are fused, brecciated, and intruded by vesiculated basaltic lava. However, as described above, the pumice clasts exhibit minimal thermal effects compared to the ash flow matrix and the volcanoclastic unit. Two more ash flow tuff and pumice samples were collected at about 10 m (GR95-10) and 40 m (GR95-11) east of the plug. The baked volcanoclastic unit directly below the scoria deposits was traced to about 300 m east of the plug and more samples were obtained along strike at about 10 m (GR95-17), 40 m (GR95-18 and GR95-19), and about 300 m (GR95-20) away.

About 500 m east of the basaltic plug, a 240-m high fault section exposes silicic tephra and rhyolite, perlite, and vitrophyre lavas of the dome complex that are capped by basaltic lavas and scoria deposits. Samples of a rhyolite (GR95-12) and vitrophyre (GR95-14) lavas were collected from this section to compare the mineralogical and chemical composition of the silicic tuffs and lavas from the contact aureole, the poorly-consolidated pumice-rich ash flow tuffs, and the rhyolite-obsidian-perlite lava dome.

5.3.3.3.4. Analytical methods

Bulk ash flow tuff matrix, pumice clasts, and volcanoclastic samples were pulverized and analyzed using X-ray diffraction (XRD) on an automated Siemens D-500 diffractometer using Cu K α radiation, 0.02° 2 θ steps, and counting times of 1 s per step for all mounts (2-36° 2 θ) to determine mineralogic variations among the different samples collected around the plug and beyond.

The same powders were used to analyze for major and trace element compositions using a Rigaku 3064 wavelength dispersive X-ray fluorescence (XRF). The intensities from each sample were processed to concentrations using XRF-11 to determine changes in bulk-rock compositions due to fluid-rock interaction during the basaltic intrusion. Analysis and statistics of the unknown samples are based on a model that uses intensities for 21 standard rocks. Volcanic glass chemistry was obtained on five samples using an automated SX50 Cameca electron microprobe. The probe was operated using an accelerating potential of 15 kV, 15 nA beam current and a fixed beam size of 10 mm. Standards used for calibration include both natural silicic and basaltic glasses, a set of feldspars, iron oxide, pyroxene, and amphiboles.

Selected samples were analyzed for F, Cl, B, Br, Li, S, and P₂O₅ using an Ion Chromatograph Dionex 4000i with conductivity and UV/Vis detectors and other accessories to estimate volatile components gained and/or lost during the contact metamorphism (Shevenell and Goff, 1993). Bulk ash flow tuffs and pumice samples from within and outside the contact

metamorphic zone were analyzed. Standards that were lithologically similar to the samples were also analyzed to determine the quality of the analytical work.

Finally, the effect of contact metamorphism on the rare earth element compositions of the basalt, ash flow tuffs, pumice clasts, and volcanoclastic unit were determined by instrumental neutron activation analysis (INAA) on selected samples at Washington University, St. Louis (Korotev, 1978). Two NBS standards were analyzed with the samples for controlling the quality of the analytical work.

5.3.3.3.5. Mineralogical and Geochemical Results

X-ray Diffraction Analysis. The ash flow tuff, pumice, and volcanoclastic samples collected within the contact metamorphic aureole were modified in their mineralogic compositions compared with those samples collected outside this zone (Figs. 27a). As indicated in Figure 27a, a rhyolite lava (Fig. 27a-a) collected about 500 m east of the contact metamorphic aureole exhibits the same X-ray diffraction patterns as a pumice (Fig. 27a-b), a bulk ash flow tuff (Figs. 27a-d), and a volcanoclastic sediment (Fig. 27a-f) collected from the contact metamorphic zone. The volcanic glass in the tuffaceous samples was totally recrystallized and the mineralogy is dominated by quartz, cristobalite, tridymite, and sanidine except for minor amounts of micaceous minerals in the rhyolite lava. Although the rhyolite and vitrophyre lavas are enriched in vapor phase topaz and garnet (Christiansen et al., 1983), none of these minerals were indicated by the X-ray diffraction patterns. Two other ash flow samples (Figs. 27a-c and 27a-e) collected outside the contact aureole are dominated by volcanic glass except for minor amounts of phenocrystic quartz and feldspar and vapor phase tridymite and cristobalite crystallized during deposition. These poorly-consolidated and volcanic glass-dominated ash flow tuff samples exhibit minimal physical and/or mineralogical effects from the basaltic intrusion represented by the plug.

In Figure 27b, a reworked tuff (Fig. 27b-a), a vitrophyre lava (Fig. 27b-b), and an ash flow tuff (Fig. 27b-c) are plotted with three pumice samples (Figs. 27b-d, e, and f) collected at different distances from the plug (Fig. 26). An ash flow tuff and a volcanoclastic sediment from the contact aureole and from directly beneath the thick scoria deposit, respectively, are devoid of volcanic glass and are dominated by quartz, sanidine, cristobalite, and tridymite. Trace amounts of gypsum, micaceous, and clay minerals are present in these samples. The pumice samples are glassy except for small amounts quartz and feldspar phenocrysts, cristobalite from devitrification during cooling, and secondary calcite in one of the samples (Fig. 27b-f). The vitrophyre has identical X-ray diffraction pattern as the pumice samples. Although three of the pumice samples were collected along the contact with the plug, two of them (Figs. 27a-b and 27b-c) were totally recrystallized, whereas the other one (Fig. 27b-e) is glassy like the pumice clasts collected outside the contact aureole and the vitrophyre lava. Although dominated by pumice clasts and fine ash matrix, the volcanoclastic sedimentary layer was recrystallized and contains no or minimal amount of volcanic glass. These samples were collected either from the contact aureole of the plug or directly beneath the scoria deposit and were significantly altered by heat from the basaltic flows.

Outcrops around the plug are poor because of talus cover. However, the pinkish to orange color of the patchy outcrops of the ash flow tuff and the volcanoclastic sediment is one indicator of the zone of contact metamorphism. Although gradational, the abundance of volcanic glass in the ash flow tuff samples appears to generally indicate the aerial extent of the contact metamorphic aureole. For example, at about 10 m from the plug, the ash flow samples contain about 60 to 70 percent volcanic glass and this increases to more than 90 percent about 20 m away. The contact metamorphic zone is also moderately defined by the degree of compaction of the poorly-consolidated ash flow and reworked tuffs. The compaction is gradational and is consistent with the degree of devitrification of the volcanic glass.

Geochemical Analysis. Major, trace, and rare earth element compositions of the basaltic, silicic, and volcanoclastic rocks indicate variations related to the thermal effects of the plug

and the scoria deposit (Table 5-7). Although most of the samples show a linear trend, major oxides plotted against silica indicate notable scattering in their concentrations (Fig. 28). Samples from the edge (contact zone) and the center of the plug indicate a range in major element concentrations. The two basalt samples from the center of the plug (Fig. 26) are low in silica and P_2O_5 and higher in CaO, Fe_2O_3 , MgO, TiO_2 compared with the other two basalt samples from the contact zone with the silicic rocks (Table 5-7). The enrichment in silica, alkalis, and P_2O_5 is possibly related to alteration within the contact metamorphic aureole.

Similar major element variations are also noted within the volcanoclastic samples. Those from outside the contact aureole are enriched in MgO, TiO_2 , Fe_2O_3 , CaO, P_2O_5 and LOI (loss on ignition) and depleted in SiO_2 , and the alkalis (Fig. 28). Like the basalt specimen from the contact zone, the reworked tuffs appear to have gained silica and alkalis because of devitrification consistent with the recrystallization of the volcanic glass in these rocks.

The silicic ash flow tuff, lavas, and pumice clasts plot close to each other and show minor variation in their major element compositions (Fig. 28). Except for higher alkalis and lower LOI contents in samples from the contact aureole, most of the ash flow tuff samples contain similar concentration of the major oxides (Table 5-7). The pumice clasts are more enriched in their silica and alkali contents but contain no or minor amounts Fe_2O_3 , CaO, TiO_2 , MgO, and P_2O_5 compared with the ash flow tuffs. Moreover, the LOI content is higher for the ash flow tuff samples regardless of their location. The rhyolite and vitrophyre samples generally plot within the range of the silicic tuff samples. For example, the vitrophyre is chemically similar to the pumice clasts than the ash flow tuffs. The rhyolite lava has the highest silica content (75.32 wt %) and lowest LOI (0.4 wt %) than all the silicic rocks. For the rest of the oxides, the difference was insignificant.

The major element compositions from the microprobe analysis of volcanic glasses from five samples collected at different intervals away from the plug are generally consistent with the results of the bulk samples obtained using XRF (Table 5-8, Fig. 29). The scattering in alkalis and water contents is similar to the major oxide plots shown in Figures 28, 29. However, a distinct grouping is indicated by the FeO and F plots. A sample (GR95-15b) collected about 300 m west of the plug contains twice as much F compared to other samples collected with about 25 m from the intrusion (Table 5-8). Conversely, samples (GR95-6, 7, and 8) outside the contact aureole (Fig. 26) are enriched in their FeO contents (Fig. 29).

The effect of the basaltic intrusion on the volatile contents of the silicic ash flow tuffs were also checked on four samples (GR95-7, 8a, 8b, and 11) within about 50 m from the plug (Fig. 26). These samples were checked for their B, Cl, F, Li, P_2O_5 , and S concentrations (Table 5-9). The F content increases away from the plug consistent with the electron microprobe data (Fig. 30). Cl also appears to show the same trend except for higher concentration in one of the samples from the contact aureole (Table 5-9). Although not by much, S content is higher in samples within the aureole, whereas B, Li, and P_2O_5 show no particular trend (Fig. 30).

Unlike the silica and major element plots, no linear trends are indicated by the trace elements except for Sr (Fig. 31). Moreover, the silica and trace element plots indicate more scattering compared to the major elements. The basalt samples from the center of the plug contain higher Ni, Cr, and V consistent with their elevated concentrations of MgO, TiO_2 , Fe_2O_3 , CaO, P_2O_5 (Fig. 28). Those basalt samples close to the contact zone have either similar or slightly higher Ba, Nb, Rb, Sr, Zn, and Zr contents. Similar to the major element plots, the volcanoclastic sediments are also separated into two groups on the basis of their trace element compositions. Those from outside the contact aureole are enriched in Sr and Zr, whereas higher Rb, Nb, Y, Zn, and Zr concentrations are present in the thermally altered rocks (Fig. 31).

The silicic rocks contain no Cr and Ni and this is also true for the pumice clasts and lavas with regard to Ba and V. The ash flow tuffs and pumice clasts contain variable concentrations of

trace elements. For example, the pumice clasts have slightly higher contents of Rb, Nb, and Y, whereas the tuffs contain more Ba, Sr, Zn, and Zr contents (Table 5-7, Fig. 31). The trace element compositions of the rhyolite and vitrophyre lavas are similar to the rest of the silicic tuff samples except for slightly higher Rb and insignificant amount of Sr. In fact, the vitrophyre contains no Sr.

Finally, selected samples from different locations with reference to the basaltic plug were analyzed for their rare earth element (REE) concentrations to check for signs of enrichments or depletions because of the contact metamorphic effect (Fig. 26). Chondrite-normalized REE and multi-element plots are given for the selected samples in Figure 32 (Nakamura 1974; Thompson et al., 1984). Basalt samples from the contact zone (GR95-4) and the center of the plug (GR95-16) show no significant variations. However, the overlap is not perfect and the sample from the contact zone appears to be slightly enriched (Figs 32a,b). The basalt samples are enriched in light REE, Ba, and Sr contents compared with the silicic rocks, whereas Yb, Lu, Rb, Th, K, and Nb are higher in the silicic rocks. Variations in the REE contents are more apparent in the silicic rocks. For example, the ash flow tuffs and pumice clasts from the contact aureole are depleted in light REE, whereas the opposite is indicated by the rhyolite (GR95-12). In the multi-element plot (Fig. 32b), the silicic rocks exhibit minimal variations in their Rb, Th, Nb, Ce, Nd, Sm, Hf, and Yb. Of all the elements, Ba, Eu, and Sr show the greatest variations.

5.3.3.3.6. Discussion

Despite the size of the basaltic plug (~150 m wide) hydrothermal alteration in the silicic rocks of East Grants Ridge is minimal. Field and laboratory studies provide no evidence for hydrothermal activity in the poorly consolidated pyroclastic deposits intruded by the olivine-bearing basalt. In fact, because of the unwelded nature of the ash flow tuff near the basaltic plug, it was extensively mined for pumice (Fig. 26). However, the effect of the basaltic intrusion and the thick scoria deposit is distinctly apparent in a narrow zone along the contact with the silicic tuffs. This is indicated by physical changes in brecciation, deformation, injection of basalt along bedding planes, compaction, and coloration. The pyroclastic rocks and volcanoclastic sediment below the scoria are brownish-orange because of baking. These same rocks adjacent to the plug show the effect of more heat than those below the scoria by being dark orange, fused, deformed, and brecciated. Although the silicic rocks are not fully exposed around the plug, the extent of the contact metamorphic aureole is about 5 m wide or less based on the color transition from orange to light gray and degree of compaction.

An attempt was also made to define the aerial extent of the contact metamorphic aureole using mineralogical compositions of the different samples collected from both sides of the plug (Figs. 27a,b). Generally, the volcanic glass of silicic rocks are recrystallized or devitrified at elevated temperatures in the presence of fluids (Lofgren, 1970). According to Lofgren (1970), the rate of devitrification is minimal under dry conditions and pure water. Unlike samples within the contact aureole, ash flow tuff and pumice clasts within 10 m of the plug are dominated by elevated background of amorphous silica scattering indicating the presence of abundant volcanic glass. Exceptions were noted to this observation. Pumice clasts on the west side of the contact zone were fused and totally devitrified, whereas on the east side, similar samples from the contact consist of volcanic glass. According to Lofgren (1970, 1971), the devitrification process involves hydration followed by devitrification whereby water diffuses through silicate glass forming hydroxide ions (OH^-), breaking the bridging of Si-O bonds by OH^- , and making available the newly freed SiO_4 tetrahedra for the formation of quartz and feldspar crystals. The hydration and devitrification rates increased one to five orders of magnitude in the presence of Na and K compounds compared with pure water. Thus, variation in mineralogical composition of the silicic rocks was a useful diagnostic tool to trace the extent of the thermal effect because of devitrification process. However, this was not the case with all the samples. The role of aqueous solutions appears to be an important factor in the recrystallization of volcanic glass intruded by basaltic magmas.

Variation in major element compositions among the basaltic and silicic rocks are not extensive and are consistent with a minimal amount of water-rock interaction during the basaltic intrusion. However as indicated in Figure 28, the basalt samples from the same plug show minor variations because of localized interaction along the contact with the tuff. Increase silica and alkalis in the basalt samples from the contact zone suggest localized elemental exchanges between the mafic and silicic rocks. The volcanoclastic samples from the contact zone also show the same kind of enrichment. Despite higher porosity due to poor compaction in the volcanoclastic rocks, a sample 10 m away from the plug has the same major and trace element chemistry as a sample collected about 300 m away reflecting the absence of hydrothermal alteration due to insufficient aqueous solutions during the intrusion.

Comparison of major element chemistry of volcanic glass obtained by electron microprobe analysis clearly indicates the effect of the intrusion on the silicic rocks (Fig. 29). Localized water-rock interaction involving the silicic and basaltic rocks within the contact aureole is indicated by minor enrichments in alkalis of the silicic rocks while they were depleted in their FeO, H₂O, and F contents. A similar approach using the volatile contents (B, Br, Cl, F, Li, P₂O₅, and S) of bulk ash flow tuffs to map the aerial extent of the effect of the intrusion was also informative (Fig. 30). The F data from this and the probe analyses are consistent with each other. With the exception of one sample, the Cl data shows the same pattern as the F variation. Although not by much, the S content is higher in the silicic samples close to the basalt intrusion. This is because S is the major volatile component of basaltic intrusion. Thus, microprobe data and wet analysis of volatile contents of pure glass separate appears to be a useful tool to delineate the aerial extent of contact metamorphic zone created by an intrusion.

The trace element and REE compositions of the basaltic, silicic, and volcanoclastic rocks corroborate the information provided by the major element data. This geochemical pattern is attributed to minimal water-rock interaction within the contact aureole and adjacent rocks. No mineralogical and/or chemical evidence was apparent that would suggest migration of sufficient aqueous solutions within the contact zone of the silicic and basaltic rocks and the adjacent poorly-consolidated tuffs. Evidence is provided by the similarity in trace element and REE contents of samples collected within and outside the contact zone.

5.3.3.3.7. Summary and Conclusions

Physical features (e.g., color variation and degree of compaction), mineralogical, and chemical data from the silicic tuffs around the basaltic plug indicate variable thermal effects. Despite the size of the basaltic intrusion (~150 m wide), the contact metamorphic zone developed around the basaltic plug is confined to a narrow zone of about 5 m wide. This is similar in size to the contact aureoles developed in vitric and zeolitized tuffs intruded by dikes and a sill at Paiute Ridge along the north eastern boundary of the Nevada Test Site (Section 5.3.3.2.). The poorly-cemented ash flow tuff along the contact aureole is baked, fused, deformed, and injected with basalt lava. This thermal effect gradationally decreases away from the intrusion. Variation in color and the degree of compaction of the silicic ash flow tuff provide evidence for the localized nature of the thermal effect.

Variations in mineralogical and geochemical compositions are consistent with the changes in color and compaction of the silicic rocks. The volcanic glass, water composition, fluorine, chlorine, and iron contents increase away from the intrusion. Conversely, the amount of recrystallized minerals and the alkali contents are higher along the contact.

The field and laboratory data suggest that the effect of the intrusion on the unwelded pumice-rich ash flow tuff was minimal. Pumice was mined at close proximity on both side of the plug indicating insignificant contact welding and hydrothermal alteration. Intrusion into an unsaturated environment appears to be the most probable cause for the limited water-rock interaction in the porous ash flow tuffs. Because devitrification is generally enhanced by aqueous solution, the abundance of volcanic glass within

a short distance (~ 5 m) from the intrusion suggests a dry environment. Thus, in the absence of aqueous solutions, the width of the contact metamorphic aureole is generally confined to a narrow zone regardless of the size of the intrusion.

5.3.4. Theoretical and Modeling Studies of Hydrothermal Processes

5.3.4.1. Introduction

This section (5.3.4) describes results from the first stage of theoretical and modeling studies that were conducted for the magmatic effects study plan. The results shown here comprise some of the first steps toward a more comprehensive set of numerical simulation studies that were planned for the latter half of the effects project. First, we describe analysis of the onset of convection of a compressible gas in porous media. This was a necessary first step because hydrothermal effects of interest for the potential Yucca Mountain repository would occur in the vadose zone. This differs from the classical hydrothermal system in saturated media. The theoretical analysis below showed how the different fluid properties of air (vadose zone) and liquid water affect convection, and was also used to verify the computer code FEHM. The analysis is being published as a separate paper (Stauffer et al., in press, *International Journal of Heat and Mass Transfer*). Section 5.3.4.3 describes preliminary numerical simulations aimed at determining the maximum distance that effects (in this case perturbed flow fields) extend from a shallow basaltic dike or sill. This problem is important because it defines the area surrounding the potential repository within which an intrusion might affect potential repository performance even if the intrusion does not directly intersect the potential repository.

5.3.4.2. Finite Amplitude Analysis of Natural Convection of Compressible Gas in Porous Media

5.3.4.2.1. Summary

The theory of thermally driven convection of dry air in a porous medium is reviewed. The critical Rayleigh number for air is the same as for liquid, $4\pi^2$, but the thermal gradient used is decreased by the adiabatic gradient of air. Because of the differences in the physical properties of air and water, initiation of convection requires the product of gradient and permeability to be thousands of times greater for air than for water. Finite amplitude analysis of the problem for $Ra < 300$ show that (1) the code predicts the onset of convection in an air filled porous medium, (2) at low thermal gradient, Ra vs. Nu curves are nearly the same for air and water, (3) the slope of the Ra vs. Nu curve matches well with experimental data reported by others for water, and (4) time to steady state decreases approximately as the square root of Nusselt number.

5.3.4.2.2. Nomenclature for Section 5.3.4.2

		<u>Greek symbols</u>	
c_p	heat capacity of fluid at constant pressure	α	fluid thermal expansivity
g	gravitational constant	ϕ	porosity
Γ	$\Delta T/H$, temperature gradient	κ	equivalent thermal conductivity
H	thickness of system	μ	viscosity of fluid
K	permeability	ρ	fluid density
Nu	Nusselt Number	<u>Subscripts</u>	
P	pressure of fluid	c	critical
Q	heat content		

Ra	Rayleigh Number	eff	effective
T	temperature of gas and matrix (Kelvin)	m	matrix
ΔT	change in temperature across system	s	steady state
t	time		
U	internal energy		<u>Superscript</u>
v	Darcy fluid flux		Total derivative with respect to time
W	work		
z	downward unit vector		

5.3.4.2.3. Introduction

The onset of convection in incompressible liquid in a porous media has been considered many times, and a comprehensive treatment of this subject can be found in Nield and Bejan (1992). Much less work has been done, however, on the problem of ideal gas convection in porous media. Strauss and Schubert (1977) include the compressibility term in their linear stability analysis (LSA), but restrict their analysis to steam saturated water (i.e. no free vapor phase). Saadajain (1980) neglects the pressure/volume contribution to the energy equation in his treatment of ideal gas porous convection. Nield (1982) corrects Saadajain and provides many insights into the differences between air and water, but provides only a brief LSA. Zhang et al. (1994) presents a complete LSA for moist gas convection in porous media that reduces to the solution of Nield when moisture content is set to zero. The reader is referred to this work for further details.

Nield (1982) and Zhang et al. (1994) both show that the onset of convection in an ideal gas can be characterized by a critical Rayleigh number (Ra). The exact form of the definition of the Rayleigh numbers, however, is altered because of the effects of compressibility. To first order, the only difference is that for an incompressible fluid like water, we have the Rayleigh number:

$$Ra_{liquid} = \frac{\alpha H^2 K g c_p \Gamma \rho^2}{\mu \kappa} \quad (5-4a)$$

whereas for air, we have a slightly different definition of the Rayleigh number:

$$Ra_{air} = \frac{H^2 K g (c_p \Gamma - g) \rho^2}{T \mu \kappa} \quad (5-4b)$$

The two definitions of the Rayleigh numbers are almost identical. For an ideal gas, $\alpha = 1/T$. The $c_p \Gamma$ term in Ra_{liquid} is replaced by $(c_p \Gamma - g)$ in the Ra_{air} definition. This result is analogous to Jeffrey's (1930) conclusion that the Ra for compressible fluid convection (no porous medium) must be modified to account for the adiabatic gradient. The change in the effective temperature gradient arises from the pressure dependence of the heat content of a fluid parcel. In an incompressible fluid, the internal energy depends only on the temperature. In a compressible gas, if a packet moves into a lower pressure region, it expands and cools adiabatically reducing the effective temperature gradient. This means for air, Γ must be larger than an adiabatic gradient, $g/c_p \approx 10^{-2} \text{ }^\circ \text{K/m}$, before the Rayleigh number even has a positive value. Below the adiabatic gradient, air convection is not possible for any values of system height (H) or permeability (K). Thus if we use Ra_{liquid} but substitute an effective temperature gradient

$$\Gamma_{eff} = \Gamma - \frac{g}{c_p}$$

for Γ and $1/T$ for α , the two Rayleigh numbers are identical (Nield, 1982).

It is the values of the physical quantities appearing in the two Rayleigh number definitions that dominate the difference between the convective stability of subsurface water and air. Simple algebra shows that for a given geothermal gradient the critical permeability for onset of convection in a typical porous medium is three orders of magnitude greater for dry air than liquid water (Table 5-10).

The dramatic increase in the permeability needed to initiate convection in air is explained by the factors noted previously. While the lower viscosity and higher thermal expansion coefficient favor convection in air, these are more than off-set by the p^2 factor present in Eqs 5-4a and 5-4b. Convection occurs when it is more efficient to move energy by the physical exchange of matter than by conduction of heat. While it is easy to convect large volumes of air, those volumes of air carry very little energy, and that heat is rapidly dissipated by the conduction through the matrix, damping the convective motion.

The purpose of this paper is to examine convective flow in a porous medium filled with dry air heated from below. We present the theory, report finite amplitude results for Rayleigh numbers Ra up to 300, discuss the relationship between time to steady-state, Nusselt number (Nu) and Ra for these systems, and compare the results to those for similar systems saturated with liquid.

The motivation for this work comes from our study of the effects of volcanic intrusions into partially saturated rock. We seek to understand the spatial and temporal scale over which an intrusive event can have a significant effect on the transport of gases, both from the intrusion itself (e.g., SO_2) or at some distance from the intrusion.

First, we review the governing equations and verify the ability of the numerical code we use for our finite amplitude calculations to reproduce the critical Rayleigh number.

5.3.4.2.4. Governing Equations

The basic equations which govern low velocity (Darcy flow regime) convective flow in porous media are well developed in the literature (e.g. Nield and Bejan, 1992; Strauss and Schubert, 1977; and Nield, 1982). In the following form, rewritten from Nield and Bejan (1992), Strauss and Schubert (1977), and Nield (1982) for clarity, the equations apply to single phase compressible as well as Bousinesq approximate incompressible fluids.

$$\phi \frac{\partial \rho}{\partial t} + \nabla \cdot \rho \mathbf{v} = 0 \quad (5-5a)$$

$$\mathbf{v} = \frac{K}{\mu} (\nabla P - \rho g \mathbf{z}) \quad (5-5b)$$

$$(\rho c_p)_m \frac{\partial T}{\partial t} + (\rho Q')_f = \kappa \nabla^2 T \quad (5-5c)$$

where the matrix density, ρ_m , includes a correction for porosity. Equation 5-5c can be reformatted in the more traditional form of Nield (1982). By substitution of $Q' = U' - W'$, and bearing in mind the total derivative of a quantity (Q) includes an advective transport contribution as:

$$Q' = \frac{\partial Q}{\partial t} + \mathbf{v} \cdot \nabla Q \quad (5-5d)$$

We arrive, after some algebraic manipulation and thermodynamic substitutions, at:

$$(\rho c_p)_m \frac{\partial T}{\partial t} + (\rho c_p)_f T' - \alpha T P' = \kappa_{eff} \nabla^2 T \quad (5-5e)$$

The difference between the convection of an incompressible liquid and the convection of a compressible gas arises from three basic causes. The first is the significantly smaller amount of energy contained in a given volume of gas compared to the same volume of liquid. The advective heat transport term, $\rho Q'$, of Eq 5-5c is directly proportional to the density, ρ . For example, even though the rate of change per unit mass, Q' , is roughly the same for both water and gas, the fact that the density of air is three orders of magnitude lower makes convection in air much less efficient. The second cause is the much smaller buoyancy force (ρg , Eq 5-5b) for air, again due to the very large difference in the densities. For air,

$$\delta \rho g = \rho g \left(\frac{\delta P}{P} - \frac{\delta T}{T} \right) \quad (5-6a)$$

where the first term on the right hand side corresponds to the isothermal compressibility, β . For the incompressible case, only the thermal expansion term is retained as:

$$\delta \rho g = \rho g \alpha \delta T \quad (5-6b)$$

The third is that the energy content of a given volume of air is pressure dependent whereas the energy content of a given volume of liquid is not. For air,

$$\rho Q' = c_p \rho T' - P' \quad (5-7a)$$

In contrast, for a Bousinesq liquid,

$$\rho Q' = c_p \rho T' \quad (5-7b)$$

Secondary factors that lead to real gases being more stable than an ideal gas include variations in α , c_p and μ with temperature (Nield, 1982). Comparison between air and water material properties and their derivatives with temperature can be located in Nield and Bejan (1992) and also Bejan (1993).

5.3.4.2.5. Finite Amplitude Analysis

In order to consider finite amplitude convection in air, one must go beyond the assumptions used in the linear stability analysis. This is especially important with air, since the strong temperature dependence of the thermodynamic properties of air leads to large differences in the Rayleigh number calculated at the hot versus cold ends of the convective cell. Note that in this paper, we will report all Ra values calculated at the mean temperature of the system after Davidson (1986).

For the numerical analysis of the convection problem we have used the finite element heat and mass transfer code FEHM (Zyvoloski et al., 1995a,b). Originally developed for use in geothermal reservoir modeling, FEHM is capable of simulating multi-phase transport of air, water, water vapor, and dilute solutes in fractured and porous media. FEHM solves these equations for the highly nonlinear equations for air convection using the Galerkin technique to discretize the spatial derivative, and a standard first order approach to the time discretization. Lobatto integration is used to evaluate integrals in the finite element equations and a modified Newton Raphson iterative procedure is used to solve the resulting system of equations. Dependence of material properties on pressure and temperature is represented as a cubic polynomial. The polynomial fits are based on National Bureau of Standards steam tables (Haar et al., 1984).

In our calculations, we assume that (1) flow is not affected by quadratic drag or boundary phenomenon (i.e., the creeping Darcy regime), and that (2) thermal equilibrium exists between the air and rock. The violation of either of these would require extra terms to modify the treatment of the heat transfer. We address (1) first. The transition from creeping Darcy flow to inertial flow is determined by the value of the porous Reynolds number, Re_p . If $Re_p \ll 1$, all inertial and boundary effects are negligible. Even under the liberal assumption that our pore diameter is as large as 10^{-3} m, Re_p approaches unity only if the air velocity is above 0.015 m/s. Maximum air velocities reached during this study were $2e-4$ m/s at $Ra = 300$. With respect to issue (2), thermal equilibrium is reached if a temperature perturbation in the convective cell moves slowly compared to the time required to exchange energy over pore-scale lengths. Computations of energy transfer rates for FEHM (Zyvoloski et al., 1995a) show that thermal equilibrium is valid for water velocities below 10^{-5} m/s. Scaling the water results to air by the product of relative heat capacity, thermal conductivity, and density (Table 5-10), we come to the conclusion that a typical porous medium should be in approximate thermal equilibrium at air velocities below $1.7e-3$ m/s.

A simple two dimensional regular grid representing a 200 m square, broken into 49×49 rectangular elements, has been used for the simulations. Top and bottom boundaries have fixed temperatures while the side boundaries permit no flow with respect to either mass or energy (Figure 33). Note that this choice of no flow lateral boundaries restricts the horizontal motion, and limits the possible planforms. For all the simulations presented here we use an initial mid-domain temperature of 60°C . Top and bottom boundaries are set to $60^\circ\text{C} \pm 1/2\Delta T$. For each case, a non-flow calculation was used to generate the initial temperature-pressure distributions as input for the stability test. Convection was set in motion by finite temperature perturbations ($0.1 - 1^\circ\text{C}$) at selected bottom nodes.

Our major measure of the dynamics of convection in these systems is the computed Nusselt number, Nu , which is the ratio of the total flux to the purely conductive flux implied by the temperature gradient. We calculate Nu at the top and the bottom of the system numerically after the system has reached steady state. The criterion for determining a steady-state is that a small rate of change in the temperature has been achieved. As one can see in Figure 34, this time is well-determined. Error in the calculation of Nu is estimated to be between 2 and 3%.

5.3.4.2.6. FEHM Results

We begin by validating FEHM's ability to predict the onset of convection in air. As we show below, FEHM reproduces the critical Rayleigh number remarkably well (Figure 35). We determined Ra_c by a straight line fit through the results for four calculations at low Rayleigh extrapolated to $Nu = 1$. Nonlinear effects decrease with temperature gradient, so we used a low thermal gradient of $20^\circ\text{C}/\text{km}$ for this analysis. By definition at Ra_c the time to set up convection is infinitely long so we must extrapolate the finite amplitude results to estimate the threshold value.

The Ra_c determined from these FEHM results is 39, remarkably close to the linear stability prediction of $4\pi^2$.

Figure 36 shows all of our finite amplitude results for air and for water on a Ra vs. Nu plot. As this figure shows, the results for air and water are nearly identical. See Table 5-10 for property values used in these simulations. The thermal gradient for these simulations was $20^\circ\text{C}/\text{km}$ for air and $300^\circ\text{C}/\text{km}$ for water. These results are in good agreement with a scaling analysis reported by Nield and Bejan (1992) for systems saturated with water and experimental data reported by Kaneko et al. (1974) and Elder (1967).

The relationship between Nu and time to steady state (t_s) is shown in Figure 37. As this figure shows, the time to establish the convective flow is on the order of 1000 years for the system simulated here, and decreases as the strength of the convection increases. t_s scales in a simple way, decreasing in approximately direct proportion to the square root of Nu . This is true for a variety of thermal gradients and different system heights, scaled in Figure 37 such that the product, $(K H \Gamma)$, remains constant. Scaling to determine times needed to reach steady state has proved useful in our current work. Another point of interest is that as Nusselt number approaches 1 from above, time to steady state quickly exceeds the time for a conductive profile to equilibrate, in agreement with observations reported by Elder (1967).

We use Nu rather than Ra to bypass the problems of Ra_{air} varying so greatly in value from the top to the bottom of the system. Instead of using a mean value for the system temperature and plotting Ra versus t_s after Davidson (1986), we have chosen to examine time to steady state as a function of the Nusselt number. Solution of the LSA equations for stability of integer wavelength cells predicts that multiple planforms are stable for a given Ra , each with a unique Nu (Nield and Bejan, 1992). Final planform pattern depends solely on initial conditions, thus a plot of Ra versus t_s would have several time points for a given Ra . Nusselt number provides a single value for energy transport across a system at steady state and is therefore a more consistent parameter to plot versus time to steady state.

5.3.4.2.7. Conclusions

The equations governing convection of air in a porous medium are very similar to those of water. The compressibility of the air, however, does affect the apparent temperature gradient. Because the air cools adiabatically when it rises, the effective temperature gradient is correspondingly reduced. By including this term it is possible to define a Rayleigh number appropriate for the prediction of the onset of convection in air. The resulting critical Rayleigh number for air is the same as for water, $4\pi^2$, when the gradient in the incompressible Rayleigh number is lowered by the adiabatic gradient of the gas. Even though the critical numbers are identical, it is much more difficult to induce convection in air than in water. In fact, the product of gradient and permeability must be thousands of times larger because of the very much lower density of the gas, which makes convection in air much less efficient in transporting energy than in water.

Finite amplitude analysis of the problem using the FEHM computer code show that (1) the numerical code predicts the onset of convection in air very well, (2) the Ra vs. Nu curves for air (at low thermal gradients) and water are nearly identical, (3) the slope of this Ra vs. Nu curve matches well with theory and experimental data reported by others for water and (4) time to steady state decreases approximately as the square root of Nu .

5.3.4.3. Spatial Scales of Effects on the Potential Yucca Mountain Repository Due to Convective Air Flow Induced by Igneous Intrusions

5.3.4.3.1. Summary

The purpose of this study is to provide information that will influence the calculations of $E2$ by defining the spatial scale over which an intrusive event can have a significant effect on the potential Yucca Mountain high-level waste repository. Our approach is based on numerical simulations of thermally induced convective flow of air where the thermal driving force is the heat from magmatic intrusions. Based on a series of simulations of highly simplified systems that represent Yucca Mountain, we conclude that the spatial scale over which an intruded dike or sill can have a significant effect on a potential repository is on the order of kilometers. "Intrusion" gas travels less than about 500 m and the horizontal distance over which an intrusion affects convective air flow is less than 2.5 km. Future calculations will address the role of more realistic stratigraphy, dual permeability fracture models, and topography; consider 3D domains and the flow of water (liquid and vapor) in addition to air flow; and include more complete parameter and sensitivity analyses.

5.3.4.3.2. Introduction

The following probabilistic framework is being used to estimate the risk of magmatic activity to the potential Yucca Mountain repository:

$$R_{mag} = [Pr(E2 \text{ given } E1)Pr(E1)] * C_{mag}$$

where R_{mag} is the risk posed by magmatic activity. $E1$ is the recurrence rate of magmatic events in the region surrounding Yucca Mountain (based on the estimated recurrence rate during the Quaternary and Pliocene), $E2$ is the probability that a future magmatic event intersects the potential repository or is close enough to it to have a significant effect on its performance, and C_{mag} is the sum of all consequences associated with a magmatic event that is sufficiently close to a potential repository to have an adverse effect on it.

The purpose of the present study is to provide information that will influence the calculations of $E2$ by defining the spatial scale over which an intrusive event can have a significant effect on a potential repository. Our approach is based on numerical simulations of thermally induced convective flow of air at Yucca Mountain where the thermal driving force is the heat from magmatic intrusions. At this point, the definition of "a significant effect" is unclear. For the purposes of this study, we are examining the spatial scales associated with two effects. The first is arrival of "intrusion" gas, which might contain corrosive gases such as SO_2 , at the potential repository. The second is a change in air flow at the potential repository.

Previous modeling studies have examined the effects of thermally induced convective flow at Yucca Mountain where the thermal driving force is provided by radioactive waste buried in the potential repository (Tsang and Pruess, 1987; Lu and Ross, 1993; Buscheck and Nitao, 1993). While these studies are relevant in terms of the flow and transport processes involved, they address a very different problem. One difference is the magnitude and duration of the thermal driving force. In the hot potential repository case, the thermal source lasts thousands of years and maximum temperatures are about 100°C. A dike or sill intrudes at magmatic temperatures, about 1200°C, but heat input lasts only a few years after which the intrusion gradually cools. More important is the difference in the questions being asked in these two cases. The concern in the hot potential repository problem is the transport of vapor-phase contaminants from the potential repository to the surface or water table, and therefore the focus is on flow and transport in the vertical direction. The concern in the magmatic intrusion problem is primarily one of flow and transport in the horizontal direction.

5.3.4.3.3. Modeling Approach

The FEHM computer code is used to perform the series of simulations of air convection reported here. FEHM is a nonisothermal, multiphase flow and transport code that was developed by George Zyvoloski and others at Los Alamos National Laboratory with the support of a variety of sponsors, primarily the Yucca Mountain Site Characterization Project (YMP). For details on FEHM, see Zyvoloski et al. (1991; 1995a,b).

We have made several simplifying assumptions. Our domain is a 2D box of homogeneous materials. An equivalent continuum model is used for the permeability. The role of more realistic stratigraphy, dual permeability fracture models, and topography will be considered in the future. Our simulations account for the flow of air only. Simulations which include liquid water and water vapor are also left for future work. Comparisons of simulations using linear expressions for air properties to those using real air properties show no significant differences. Since "linear" air simulations run much faster, we used this approach in this study.

Our model domain is a 2-D box of homogeneous material with the properties of Topopah Springs Tuff (unit TSW2), the rock in which the potential repository at Yucca Mountain would be located. A few simulations were performed to investigate the model sensitivity to values chosen for porosity and permeability. The domain is 700 m in height, which represents the nominal distance from the surface to the water table at Yucca Mountain, and 6 km in length, depending on the problem. The grid is in Cartesian coordinates and contains 3600 nodes. A typical grid is shown in Figure 38. Due to the symmetry of the problem, simulations are performed in a box only 3 km in length with the dike or sill located at the left wall.

Properties of TSW2

permeability ^a	10^{-12} m^2
porosity ^a	0.14
rock density	2200 kg/m^3
rock heat capacity	1000
rock thermal conductivity	$2.7 \text{ W/m}^\circ\text{C}$

^a Wilson, M. L. et al., (1994)

The following boundary conditions and initial conditions were used. The left wall represents a symmetry line in this problem, and is therefore impermeable to air flow and heat transport. The bottom boundary represents the water table. It is an impermeable boundary with a constant temperature of 38°C. The top boundary is set to 1 atm pressure, allowing for air to flow in to or out of the top. Air that leaves the system through the top wall is allowed to carry its heat with it, but the boundary is otherwise set to 20°C and air that enters the system is constrained to be 20°C. Equilibrium values of pressure and temperature obtained from 1-D background simulations before an intrusive event are maintained at the right boundary. These 1-D equilibrium temperature and pressure values were used as initial values for the 2-D domain as well.

The dike or sill is modeled by setting "dike" or "sill" nodes to 300°C for 4 years beginning at time equal to 0. Calculations based on equations given in Turcotte and Schubert [1982] for total heat into a semi-infinite half-space and the heat flux from a cooling dike or sill show that the total heat input in this way is approximately equivalent to that from a 1 m thick 1200°C intrusion. We believe this simplification is adequate for the far-field calculations considered here. Simulations are

run out to 1,000 years by which time the perturbations in the temperature field introduced by the magmatic intrusions have (or have very nearly) disappeared. A dike is modeled as a series of nodes along the left wall beginning at the water table and extending upwards. Dikes of various heights above the water table (180 m, 360 m, 560 m, and 680 m) are considered. A sill is modeled as a one-node thick zone of material extending 100 m from the left wall at various distances above the water table (0 m, 360 m and 520 m). Both permeable and impermeable intrusions are modeled for intrusions not along a side boundary.

A particle tracking routine is used to display the advective transport. To estimate the spatial scale for the arrival of air that has been in contact with the intrusion, particles are introduced at the "dike" or "sill" nodes (adjacent to the "sill" nodes for the impermeable sill cases). The initial particle field contained 10,000 particles. The paths of these particles are evaluated throughout the simulation and the maximum horizontal distance traveled is noted. To assess the spatial scale for a change in air flow at the potential repository location as a result of convection induced by the magmatic intrusion, particles are introduced along a line at the potential repository horizon (halfway between the ground surface and the water table) beginning ~600 m from the left boundary on day one of the simulation. We use the spread of the particles as a function of distance after 1,000 years as a measure of the magnitude and direction of the air flow at that distance.

5.3.4.3.4. Results

The results of the dike and sill simulations using the parameters given above are shown in the form of temperature and particle fields at 1 day, 4 years, 100 years, 500 years and 1000 years (Figs. 39-45). These plots show that the horizontal distance over which an intrusion affects convective air flow is always less than 2.5 km, and that the "dike" or "sill" particles, representing magmatic volatiles, never travel more than ~500 m horizontally. The sensitivity of these results to values of porosity and permeability values was investigated for one case, the 560 m dike, by running simulations with the porosity changed by +/- 50% and the permeability changed by +/- a factor of 2 from the values given above (Figs. 46,47). These simulations show that the distance the "potential repository horizon" is affected by the dike increases with increasing porosity and increasing permeability, but the zone of influence for effects on air flow and "intrusion" air are still less than 2.5 km and 500 m, respectively. Maximum flow velocities in our simulations are on the order of millimeters per day.

5.3.4.3.5. Discussion and Conclusions

Our basic result is that magmatic intrusions at Yucca Mountain would not cause significant large scale convection at horizontal distances greater than a few kilometers. "Intrusion" gas travels less than about 500 m laterally and the horizontal distance over which an intrusion affects convective air flow is less than 2.5 km. This makes sense considering that while it is easy to move large volumes of air, because of the very low density of air, those volumes carry very little energy. Further, the small amount of heat being physically transported by the air is dissipated by conduction through the matrix, which rapidly damps the convective motion. We stress, however, that these calculations are limited to highly simplified systems. Future work should address the role of more realistic stratigraphy, dual permeability fracture models, and topography; consider 3D domains and the flow of water (liquid and vapor) in addition to air flow; and include more complete parameter and sensitivity analyses.

5.4. Magma System Dynamics: Segregation, Transport, and Local Storage of Basaltic Magma

5.4.1. Summary

Basaltic magma of the Yucca Mountain region (YMR) originates in the upper mantle probably by processes of decompression melting. Once a melt is formed, it segregates from its solid mantle residue through a process of two-phase flow involving buoyant rise of the melt accompanied by deformation and compaction of the matrix. The mechanics of melt segregation are controlled by the equations of conservation of mass and momentum. The form of melt-matrix equations has been determined but not for all assumptions of boundary conditions. A controversial aspect of computer modeling of the melt segregation equations is the existence of solitons. Some workers argue that the formation of solitons may lead to spatial and temporal episodicity of melt segregation. Alternatively, the features may be short-lived and incapable of affecting surface volcanic patterns. Small-volume basaltic magmatism in the YMR is probably a product of relict mantle instability from athenospheric upwelling in the Miocene and/or continuing low rates of extensional deformation. There is nearly complete agreement that basalt magma ascends via fluid-assisted fracture propagation as dikes. The formation and form of magma transport in dikes have been described by many authors using different assumptions dependent on whether the shapes of magma-formed fractures are controlled by the rock properties or the dynamics of flowing magma. The latter view requires assessment of elastic deformation of the country rock and fracture, flow, and gas-properties at the dike tip. Field evidence shows that dikes can be modeled as planar cracks formed in brittle solids by pressurization and dilation associated with magma injection. Magma in dikes will rise until it erupts, solidifies, or reaches the level of neutral buoyancy (LNB). The position of the LNB in the Yucca Mountain area may be influenced by the density interface between low-density basin fill, and the Paleozoic rocks. Other processes such as crack-rate propagation, wall-rock permeability, magma-gas content, or mechanisms of volatile concentration in a dike tip could negate the controls of the LNB. One additional issue is the control of dike orientations by preexisting fractures. Field observations show that dikes can both fill fractures or fault planes or propagate by magma-induced fracture in directions controlled by the stress field. The structural controls of basaltic centers in the YMR suggest that feeder dikes may follow northwest-trending structure at depth but divert at shallow levels to north-northeast strikes following the maximum compressive stress direction.

5.4.2. Introduction

This section provides a brief overview of the processes of accumulation and ascent of basalt magma in a continental extensional setting and emphasizes the mechanisms of melt segregation and transport of magma via magma-driven fracture through the mantle and crust. Consideration is given as well to storage of basalt in the mantle and crust and at its LNB, where magma may propagate laterally. There has been substantial progress, particularly in the 1980's, in mathematically describing the processes of segregation, migration, and eruption of magma. We review the relevant aspects of these processes, focusing primarily on the constraints that can be developed for the physical dimensions of magma systems, the mechanical interactions of magma and country rock, and the rates of operation of magmatic processes. These constraints are applied where possible in attempting to understand the time-space patterns of basaltic volcanism in the YMR.

5.4.3. Melt Generation And Segregation

We assume, first, that partial melt of basaltic composition originates in the upper mantle at depths of 45 to 60 km. There is a wealth of petrologic information on the geochemical processes of melt formation constrained mostly by the major, trace-element and isotopic composition of erupted basalt (see Chapter 4). The processes controlling the formation of the magma are perhaps best known for the generation of mid-ocean ridge basalt (MORB). Melting of oceanic mantle almost certainly occurs because of adiabatic upwelling beneath mid-ocean ridge spreading centers. Here, oceanic lithosphere is upheld on a reservoir of convecting asthenospheric mantle of probable uniform composition. Recent reviews of processes controlling melting in the upper mantle are provided by Kinzler and Grove (1992), Eggins (1992), and Cordery and Morgan (1993). They note that decompression melting results when mantle rises faster than it can exchange heat with the surrounding rocks (adiabatically). Excess heat is

lost through melt production, and that loss is controlled by the latent heat of fusion of mantle peridotite, the heat capacity, the ambient mantle temperature, and the pressure and compositional fields of the mantle solidus (Kinzler and Grove 1992). The degree of melting generally increases with continuing decompression. Melting continues until either conductive cooling from above lowers the temperature below the solidus or a basaltic component is extracted. The extracted melt may migrate through the matrix or move preferentially in channels but most probably moves by porous flow (D'Arcy flow). Melt velocities are assumed to be much higher than the velocity of upwelling mantle (Cordery and Morgan 1993). Percolation allows the melt to maintain equilibrium with the matrix; channel segregation isolates the melt from the matrix (Eggs 1992). The two processes have different effects on the resulting melt geochemistry.

Melt production in the mantle below the YMR is constrained by the unique upper mantle of the southern Great Basin. First, the area is located on the north edge of the amagmatic gap (see Chapter 3; Farmer et al. 1989; Jones et al. 1992). The region south of Yucca Mountain exhibited no magmatism during the Mesozoic and Cenozoic despite major plutonic episodes in the late Mesozoic (Farmer and DePaolo 1983), and profound mid-Cenozoic extension (Jones et al. 1992). Second, the isotopic composition of strontium and neodymium in basalt in the YMR shows that it is underlain by preserved lithospheric mantle (Farmer et al. 1989). The isotopic characteristics of basalt generated in the lithospheric mantle have remained uniform for basalt melts from about 10 Ma to about 100 ka, the age of the youngest basaltic volcanic center in the Yucca Mountain region (YMR) (Farmer et al. 1989). The presence of thick and cold lithospheric mantle reduces significantly the ability of athenospheric upwelling to penetrate the crust and generate magma that feeds surface volcanism (Nicolas 1990). Finally, while initiation of a melting anomaly probably occurs in the athenospheric mantle, the isotopic composition of the basalt of the YMR requires that the erupted melt component was derived from or equilibrated with the lithospheric mantle (Farmer et al. 1989).

Once a melt is formed by decompression melting, it must be segregated from its solid residue and ascend before solidification to erupt eventually at the surface. Sleep (1974) noted melt can move as a two-phase flow involving the solid rock and the melt that exists in the matrix at grain intersections. The melt-filled grain intersections form a porous three-dimensional network in the rock. The basic dynamic process is that when melt is formed, it will tend to rise and segregate driven by buoyancy. A relatively new aspect of understanding of buoyant magma transport is that movement of melt in a porous rock matrix needs to account for the effects of compaction of the solid matrix. The equations for a two-phase flow system rising through the buoyancy of the melt phase accompanied by deformation and compaction of the matrix have been described by many workers (for example, Scott and Stevenson 1984; McKenzie 1984; Fowler 1984; Scott and Stevenson 1986, 1989).

Ribe (1987), Scott and Stevenson (1989), Fowler (1990a), and Cordery and Morgan (1993) provided the most recent reviews of the basic equations of melt segregation and compaction. They noted that the mechanics of melt segregation are controlled by conservation of mass and momentum. Summarizing and using the form of the compaction-migration equations of Fowler (1990a), conservation of mass for the melt requires:

$$\chi_t + \nabla \cdot [\chi u^l] = S$$

where χ is the liquid mass fraction, u^l is the liquid velocity, and S is the melting rate. Darcy's law for the melt velocity is:

$$u^l - u^s = -k\chi\Delta[p_l + \rho_l gy]$$

where u^s is the solid velocity, k is the permeability coefficient, p_l is the liquid pressure, ρ_l is the liquid density, g is gravity, and y is the vertical coordinate. The viscous compaction relation requires:

$$p_s - p_l = -\left(\frac{\eta_m}{\chi}\right) \left[\frac{\tau_m}{p_s - p_l} \right]^{n-1} \Delta \cdot u^s$$

where p_s is the solid pressure, η_m is the mantle viscosity scale, and τ_m is the mantle deviatoric stress scale. Conservation of energy requires:

$$\rho_s L S + \rho_s c_p \frac{dT}{dt} - \beta T \left[\frac{\rho_s}{\rho_l} \chi \left(\frac{\partial p_l}{\partial t} + u^l \cdot \Delta p_l \right) + \right.$$

$$\left. \left(\frac{1-\chi}{1+r_\chi} \right) \left(\frac{\partial p_s}{\partial t} + u^s \cdot \Delta p_s \right) \right] = k T$$

where L is the latent heat, c_p is the specific heat at constant pressure, T is the temperature, β is the thermal expansion coefficient, d/dt is the material derivative, r is the shrinkage ratio, and k is the thermal conductivity. Finally, the Clapeyron relation requires:

$$T = T_0 + \Gamma p_l$$

where T_0 is a reference temperature and Γ is the slope of the Clapeyron curve.

The equations for matrix flow include conservation of mass:

$$\Delta \cdot u^s = -\Delta \cdot [\chi(u^l - u^s)]$$

conservation of momentum:

$$-\Delta p_s + \Delta \cdot \tau + \Delta [(\rho_s / \rho_l) \chi (p_s - p_l)] + \rho g = 0$$

and the stress/strain rate relation:

$$\tau_{ij} = \eta \left(\frac{\partial u_s^i}{\partial x_j} + \frac{\partial u_s^j}{\partial x_i} \right)$$

Fowler (1990a) notes that there is general agreement on the form of the matrix and melt flow equations but there are widely varying assumptions concerning the boundary conditions of the modeled processes. There are two areas of concern. First, a common assumption is that all the segregated magma

ascends through the lithosphere via dike propagation. An alternative view is that some component of the partial melt solidifies during segregation and ascent. Second, the dynamics of melting are often neglected by assuming the melting rate S to be zero and ignoring the energy equation. A proposed alternative (Fowler 1990a) is to establish the boundary conditions by applying mass conservation, force balance, surface energy balance, and the continuity of temperature and free energy. Cordery and Morgan (1993) assume that the percentage of melt present is small (less than a few percent), and melt is extracted as a steady state process.

A number of workers (Scott and Stevenson 1984, 1986, 1989; Richter and McKenzie 1984; Scott et al. 1986; Scott 1988) have examined, through computer modeling and laboratory experiments, flow through compacting media using modified forms of these equations. Their results reveal the somewhat surprising existence of solitary wave solutions called solitons or magmons (Scott and Stevenson 1984, 1986). Solitons are waves that propagate through the compacting media as a porosity pulse responding to pressure gradients. Assignment of reasonable permeability and rheological properties to these features gives wavelengths of kilometers and velocities of centimeters per year. The volume of liquid within a magmon of this size is about 1 km^3 (Scott and Stevenson 1986). Most authors have been cautious about applying concepts of solitary waves to surface volcanic patterns but this phenomenon could lead to spatial and temporal episodicity of melt segregation if the patterns are not removed or obscured by processes of crustal transport. Cordery and Morgan (1993), for example, note that an assumption of steady state extraction eliminates the existence of magmons.

Fowler (1990b) examined applications of a compaction model for melt transport. He argued that solitary waves can propagate primarily under conditions of zero melting ($S = 0$). Under conditions when $S \neq 0$, solitary waves move at a nonconstant speed of $2z^{1/2}/\zeta$, where z is the vertical coordinate ($z = 0$ at the base of the melt zone) and ζ is the compaction height. The amplitude of the solitary waves decays according to $1/z^{1/2}$ and may exhibit nonlinearity and diffusion, leading to degrading. Fowler (1990b, his Appendix B) also obtained nonlinear equations for solitary waves by adding other melting variables and argued that the waves are only short-term features. He verified the concept of compaction length but argued that other complications such as bulk viscosity, melt refreezing, and fracturing need to be considered. Scott and Stevenson (1989) agree that solitary waves are probably not important under conditions of high melt production (fast-spreading). They suggest, however, that solitary waves could provide a mechanism for explaining oscillating magmatic rates at slow spreading ridges. This suggests the wave phenomenon could be important for areas of low melt production associated with small-volume continental basaltic volcanism typical of the YMR. The exceptionally small volume of Pliocene and Quaternary basaltic volcanism of the YMR may be especially appropriate. However, this requires that the episodic nature of solitary wave phenomena be preserved through the processes of ascent and eruption of magma. An interesting but highly speculative concept is that the episodic nature of polygenetic eruptive models may originate or be influenced by oscillatory solitary waves. Scott and Stevenson (1989) point out that the existence of solitary waves has only been indicated from one- and two-dimensional modeling; it has not been explored yet for three-dimensional modeling. The issue of three-dimensional modeling of melt migration has been addressed by Spiegelman and Wiggins (1993). They suggest the form of melt migration in three dimensions is in spherical waves of melt fraction. These waves form from perturbations to steady-state flow, and can link to form a channel network aiding melt propagation. Spiegelman and Wiggins (1993) note that solitary waves may contribute to episodicity of magmatism derived from the mantle.

Nicolas (1986, 1990) considered constraints on the depth of melting and melt extraction of basaltic magma. While most of the considerations are for MORB, they can be applied by inference to continental settings. First, the melting process is thought to occur primarily in the asthenosphere, which corresponds to the convecting mantle where heat is transferred by convective flow and the system is adiabatic (Nicolas 1990). Second, in general, a diapir initiated in the asthenosphere will rise accompanied by melting until it reaches the lithosphere. Here it may keep rising but melting ceases (McKenzie 1984; Nicolas 1990). If the asthenosphere diapir has a slow ascent rate, it will barely penetrate the lithosphere. Again, we note the

presence of the thick, largely unmodified and presumably cool lithospheric mantle beneath the YMR (Farmer et al. 1989). This cold upper mantle should inhibit the development of large zones of upwelling in the region. The diapirs under these conditions are probably less energetic with low upward velocities, low degree of extracted melt (<7%), and limited penetration of the lithosphere (Nicolas et al. 1987; Nicolas 1990).

A pertinent question to consider is the mechanisms responsible for the generation of the volcanic rocks in the YMR. The setting of the Yucca Mountain site in the Basin and Range province, the tectonic association with intense episodes of Miocene extensional faulting, and the exceptionally large volume of Miocene volcanism in the Southwest Nevada Volcanic Field almost certainly requires a genetic link between volcanism and large-scale athenospheric upwelling (for example, Eaton et al. 1978; Christiansen and McKee 1978; Lachenbruch and Sass 1978; Eaton 1982; Jones et al. 1992). The peak of upwelling must have coincided with the peak of tectonism and volcanism (12 Ma) (see Chapter 3), and may have significantly eroded and induced melting in the lower lithospheric mantle. The dramatic decrease in the volume of volcanism in the late Miocene in the YMR must have been in response to a decrease in the rate of athenospheric upwelling. The small-volume episode of postcaldera basaltic volcanism that has continued from the Miocene to the Quaternary clearly must not be associated with earlier large-scale athenospheric upwelling. Instead the basalts may be formed by two alternative processes (or a combination of processes): (1) discontinuous melting of hydrous heterogeneities in what is presently the lower lithospheric mantle, and (2) continuing, but at diminished rates, extensional deformation in the Basin and Range province. Tectonism, as recorded in rates of extensional faulting, has waned dramatically in the YMR since the Miocene but continues to the active (see Chapter 3). The modern low rates of extensional faulting are compatible with the generation of only small volumes of basaltic magma (<0.5 km³ in the Quaternary). Generation and eruption of larger volumes of basaltic magma in the YMR is probably inhibited by the presence of cool lithospheric mantle (Farmer et al. 1989) and waning tectonism.

5.4.4. Magma Ascent

Once basaltic melt is segregated, it must move to the surface rapidly to avoid solidification with decreasing temperature. There is nearly complete agreement that transport occurs through fluid-assisted fracture propagation. This conclusion is based on the global observation of basalt dikes in a range of tectonic settings in both oceanic and continental crust. Fissure eruptions at many volcanoes prove that magma is fed from linear dikes. Dikes are exposed commonly in the vent facies of dissected volcanoes. Dikes are exposed in country rock beneath and adjacent to deeply dissected volcanoes. Dike swarms are commonly observed crosscutting cratonic and ultramafic rocks. Flow of basaltic magma through fractures provides the most compelling mechanism for transmitting magma through cool country rock without temperature loss resulting in solidification (Lister and Kerr 1991).

Extraction of magma into conduits opened by melt-fracture requires that the melt pressure exceed the yield pressure of mantle peridotite. The depth of this process can be constrained best by the depth of earthquakes and deep harmonic tremor; this information is known only for oceanic island basalt. Aki and Koyanagi (1981) examined the depth distribution of harmonic tremor below Kilauea Volcano, Hawaii. They distinguished tremor from earthquakes by the duration and period of the former. The amplitude of tremor is sustained longer than earthquakes, and distant stations show the same period as close stations. Deep tremor is separated from shallow tremor by a uniform distribution of amplitude over distance for the latter. Aki and Koyanagi (1981) found seismic evidence of deep tremor to a depth of about 55 km that may mark the depth of onset of magma transport by magma-fracture. They also noted that the reduced displacement of tremor (directly proportional to the rate of magma flow) is an order of magnitude smaller than estimates of magma supply for Kilauea (Swanson 1972; Shaw 1980, 1987). Aki and Koyanagi (1981) suggested, based on this discrepancy, that a significant component of magma is transported through dikes that may propagate aseismically. Ryan (1988) showed that the seismicity patterns below Kilauea Volcano reveal the presence of a primary magma conduit that is concentrically zoned to 34 km. He interprets the zonal structure to be a central region of higher permeability surrounded by a region of numerous dikes.

The formation and form of magma transport in dikes have been described by many authors. Pollard (1976, 1987), Pollard and Muller (1976), and Delaney and Pollard (1981) described theoretical studies of dikes and calculated solutions for the equilibrium shapes of magma-formed fractures largely from a prospective of rock mechanics. Lister and Kerr (1991) note that these analyses are valid only if the intrusion shapes are determined after the magma stops flowing but has not solidified. Moreover, they suggest the dynamics of magma fracture must incorporate properties of the flowing magma, including elastic and inelastic deformation (Rubin 1993) of the country rock and fracture at the dike tip.

Spence et al. (1987) derived equations governing the steady, vertical propagation of a two-dimensional dike driven by buoyancy from a linear source. Lister (1990) extended their analysis and showed that elastic deformation and country rock strength are important primarily at the dike tip. The most recent review of fracture propagation and magma transport in dikes is by Lister and Kerr (1991). We draw heavily from their work in the following summary.

Field observations of dikes show that country rock around intrusions moved apart; there has been little sliding or offset parallel to the dike margins (Pollard 1987, p. 14). This suggests that dike intrusions can be modeled as planar cracks formed in brittle solids by pressurization and dilation associated with injection of magma (Lister and Kerr 1991). This treatment requires consideration of elastic pressures, stresses at the fracture tip, density-driven buoyancy, viscous pressure drop associated with flowage of magma, deviatoric tectonic stress normal to the dike, and the overpressure of the magma (Lister and Kerr 1991).

The response of mantle and crustal rocks to fracture is dependent on the stresses applied and the time scale of rock response. Observed and inferred propagation velocities of magma in dikes (tens of cm to m sec⁻¹) suggest a short time scale of rock response. The strain imparted to the rock by dike emplacement is measured by the ratio of dike thickness to length (Pollard 1987); it is typically < 10⁻³ to 10⁻⁴. These conditions suggest rock response can be modeled as an elastic response (Lister and Kerr 1991). Rubin (1993) noted the zones of inelastic deformation are produced at the tips of propagating dikes.

Elastic response in rocks is measured by the shear modulus μ and the Poisson's ratio ν ; typical values are 25-35 GPa and 0.22-0.28 for basalt, respectively (Lister and Kerr 1991). Dike geometry can rarely be established from field relations but can be approximated as a two-dimensional, blade-shaped crack or a three-dimensional penny-shaped crack (Maaloe 1987; Pollard 1987). Representative lengths of dikes are about 2 km long and a meter wide (Maaloe 1987).

Fractures propagate primarily by extension of preexisting microcracks governed by the stress field and bond strength near the crack tip (Lister and Kerr 1991). The stress field near the tip of a crack has the form (Lister and Kerr 1991):

$$\sigma_{ij} = K f_{ij}(\theta) / (2r)^{1/2}$$

where r and θ are plane polar coordinates centered on the crack tip, f_{ij} are a function of θ , and K is a coefficient known as the stress intensity factor. Crack propagation is governed by a critical stress intensity factor called the fracture toughness K_c (Pollard 1987).

Magma will tend to rise by buoyancy if its density is less than the density of the surrounding rock. The LNB is the level of neutral buoyancy where magma is neither positively nor negatively buoyant. The position of this zone in the crust will determine whether, and at what depth, magma may form lateral intrusions or magma chambers (Walker 1989). The difference in gravitation force on a magma body is equivalent to a hydrostatic pressure gradient on the magma given by (Lister and Kerr 1991):

$$\frac{dP_h}{dz} = (\rho_r - \rho_m)g$$

where $\rho_m(z)$ and $\rho_r(z)$ are the densities of magma and rock at a depth z below the surface. If the density difference is assigned a general value of $\Delta\rho$, during ascent of magma, then the total hydrostatic pressure is $\Delta P_h \approx \Delta\rho gh$ where h is the height of rise. Generally, magma is less dense than the mantle and the lower crust. The LNB in Hawaii is on the order of 3 to 6 km below the surface (Walker 1989).

The mean velocity of magma averaged across a dike width (w) depends on ∇p , the spatial gradient in the fluid pressure. For laminar flow the velocity u is:

$$u = -\frac{w^2}{3\eta} \Delta p$$

where η is the magma viscosity. Conservation of fluid volume shows that the thickness of a dike-induced fracture varies with (Lister and Kerr 1991):

$$\frac{\partial w}{\partial t} = \Delta \bullet (uw)$$

An approximation of the flow velocity u in a fracture is

$$u = \frac{l}{t}$$

where l is the length of the fracture and t is the time since initiation of the fracture (see Lister and Kerr [1991], Fig. 1 for the definition of h , l , and w for fractures of different geometry). A typical pressure drop in laminar flow along a fracture of length l may be estimated from combining the preceding two equations:

$$\Delta P_v = \frac{\eta l^2}{w^2 t}$$

The crust of the earth is in a complex state of stress and the regional lithospheric stress deviates from the lithostatic value dependent on location. If the tectonic stress perpendicular to a dike is σ then the shape of a dike depends on σ by an effective normal stress $\sigma - \Delta P_i$. For most cases the effective overpressure $\Delta P_o = \Delta P_i - \sigma$ (where ΔP_i is the internal pressure). Lister and Kerr (1991) argue that σ does not vary significantly with depth; ΔP_o can be assumed to be constant.

Lister and Kerr (1991) evaluated the balance between ΔP_e , ΔP_f , ΔP_h , ΔP_v , and ΔP_o . The ΔP_e is approximately equal to mw/l where m is the shear modulus divided by $1 - \text{Poisson's ratio}$, w is fracture

thickness, and l is shorter than the other two dimensions. The ΔP_f is the internal pressure required to propagate a magma fracture, ΔP_h is hydrostatic pressure, and ΔP_v is the pressure drop by laminar flow along the length of a fracture. They used the following equations from Weertman (1971) and Pollard and Muller (1976) assuming a constant Δp :

$$p(z) = P_o + \rho g z$$

where $p(z)$ is the magmatic pressure and

$$w(z) = \frac{h}{m} (\Delta P_o + \frac{1}{2} \Delta \rho g z) (1 - \frac{z^2}{h^2})^{1/2}$$

where $w(z)$ is the dike width and K the stress intensity at the dike tips is

$$K(\pm h) = h^{1/2} (\Delta P_o + \frac{1}{2} \Delta \rho g h)$$

Lister and Kerr (1991) examined the relations among the last three equations to evaluate the closure of ends of dikes and the conditions of dike propagation. They note that a dike must exceed a vertical extent of 100 m and a width of 2 mm to propagate upwards by buoyancy (p. 10,055). Further, as h increases ΔP_h increases and ΔP_f decreases. This requires that as h gets large $\Delta P_f \ll \Delta P_h$ and demonstrates that rock toughness or resistance to fracture is much less important than the hydrostatic pressures (however, see Rubin 1993 for alternative views based on field observations). A dike will propagate until it reaches the LNB or solidifies. The relation between ΔP_f and ΔP_h suggests there is limited advantage gained by magma ascending along preexisting fractures (Lister and Kerr 1991), a conclusion that is not always consistent with field observations (Delaney et al. 1986; Pollard 1987; Rubin 1993). An exception to this conclusion would be if the dike tip is deflected by existing fractures or joints and therefore preferentially follows fractures or if inelastic deformation depends upon the near-tip stress field. Lister and Kerr (1991) argue that the rate of propagation of magma is controlled primarily by viscous resistance of the flow of magma into the dike tip and that a dike is unable to close from the bottom upwards because it is difficult to expel viscous magma from a closing crack.

Lister and Kerr (1991) note that $\Delta P_h \approx \Delta P_e$ when:

$$\frac{h^2}{w} = \frac{m}{\Delta \rho g}$$

Since dikes are generally only one or two meters wide, $h^2/w \gg m/\Delta \rho g$, and therefore $\Delta P_h \gg \Delta P_e$. They conclude that the transport of magma in feeder dikes is dominated by the balance between buoyancy forces and viscous drag. They also argue that the local hydrostatic pressure gradient does not

need to be positive throughout ascent. It needs to be positive only when averaged over the length of the conduit. Other considerations are that a dike is held open against elastic stresses ΔP_e by a small fluid overpressure ΔP_o . The latter is determined by the supply rate of magma from below, a variable that has not been well defined in magma transport models. The elastic pressures are large at the dike tip. Overall, propagation and transport of a feeder dike are controlled by the balance between $dP_h/dz \approx dP_v/dz$ (Lister and Kerr 1991). This allows the derivation (Lister and Kerr 1991; see also Spence and Turcotte 1990, their equation [1]):

$$\frac{\partial w}{\partial t} - \frac{\partial}{\partial z} \left(\frac{\rho_r - \rho_m g w^3}{3\eta} \right) = 0$$

5.4.5. Eruption or Intrusion of Basaltic Magma

The preceding equation shows that (neglecting temperature effects) magma will rise through the crust to be erupted at the surface or will stagnate and can spread laterally at the LNB. Ryan and Blevins (1987) and Walker (1989) discussed the importance of the LNB, particularly for the Hawaiian volcanoes. Important problems related to these concepts are the degree of overshoot of ascending magma beyond the LNB and the lateral spread of magma at the LNB. Both could lead to eruption, the former if the LNB is located at a shallow level in the crust, the latter if the combination of lateral propagation of dikes and topographic irregularity lead to breaching of the surface. The controls of lateral movement of magma at the LNB forming blade-like dikes have been discussed extensively for Kilauea Volcano, Hawaii and Krafla Volcano, Iceland by Rubin and Pollard (1987).

Buoyancy-driven magma ascent following the concepts of Lister and Kerr (1991) has several important applications to the YMR. The shallow structure of the region is characterized by low-density (basin-fill) deposits in Crater Flat and possibly beneath Yucca Mountain. The density interface between these deposits and underlying higher density deposits (Paleozoic rocks) may be about 2 to 4 km below the Yucca Mountain area based on interpretations of gravity and seismic refraction data (Snyder and Carr 1984). This density interface could control the depth of the LNB. In this case, the depth of the density interface suggests the LNB may be deep beneath the region, perhaps considerably below the depth of the potential repository. However, an unknown variable is the degree of magma overshoot through and above the LNB. Additionally, if the LNB was a completely effective barrier to ascent of magma, the basaltic rocks of the Crater Flat area would not be present. Either the LNB is not an effective barrier or other processes such as crack-propagation rate, wall-rock permeability, or initial gas content of magma may locally or temporally be more important. If the LNB is an important barrier to magma ascent, the depth of the density interface suggests that magma should not propagate to the surface but instead spread laterally at the LNB. Consideration of the significance of this phenomenon may be important mostly for evaluating the effect of intrusions on the waste isolation system of the YMR.

The concept of an LNB could provide a means of explaining the length of propagation of the longest cluster of basalt centers in the YMR, the Quaternary basalt of central Crater Flat. The length of this aligned chain is 12 km, significantly longer than typical dike lengths (Crowe et al. 1983b; Maaloe 1987; Pollard 1987; Rubin and Pollard 1987). An alternative interpretation is that the chain represents magma that upwelled beneath the Red Cone and Black Cone centers and propagated laterally as a bladed dike, at a shallow level. Note however, that topography in the Crater Flat basin would not favor the formation of bladed dikes. However, if dike propagation did occur, it would follow the direction of the maximum compressive stress direction. Makani Cone and the Little Cone centers may represent eruptive centers formed at the ends of a propagated dike system. This would require the length of dike propagation to be only 6 km (1/2 the cone cluster length). Additionally, it could explain the small volume of the end

centers (Little and Makani Cones) compared to the Red Cone and the Black Cone centers. This interpretation may be tested by examination of petrology data and modeling of aeromagnetic data for Crater Flat. The basalt magma of the centers is distinctly more magnetic than the alluvial fill of Crater Flat (Kane and Bracken 1983) and it may be possible to model the aeromagnetic data, enhanced with ground magnetic data, to test for the presence and geometry of lateral feeder systems beneath the centers.

The presence of sills and lopolithic centers at the Paiute Ridge area is possibly consistent with trapping of magma at the LNB. These intrusions formed just above the interface between the Paleozoic carbonates and the primary and reworked pyroclastics of the Paintbrush Tuff (distal facies of the units exposed at Yucca Mountain), a potential density contrast and possible zone of weakness (Crowe et al. 1983b). Support for this interpretation is provided by the lopolithic structure of some intrusions (Byers and Barnes 1967; Crowe et al. 1983b). Sagging of the floor of some intrusions may have been caused by emplacement of magma with a density greater than the density of the underlying country rock. Alternative interpretations other than the LNB control of the intrusions of the Paiute Ridge area are possible. Perhaps the most convincing evidence against LNB control of intrusion depth is that many intrusions were fed by dikes that propagated to the surface. If the LNB had been an effective barrier, the magma would not have reached the surface. In some cases, the intrusions formed locally above the carbonate-pyroclastic interface and not at the density interface (Valentine et al. 1992). The local formation of intrusions may be controlled by asperities along fault planes that the dikes occupy (Valentine et al. 1993). The sagging of the floor of intrusions to form lopoliths may be associated with locally intense welding of the country rock, and the attendant porosity reduction, not density sagging.

An additional mechanism that may be important for the transport of magma in dikes is the possibility of exsolution of volatiles at the tips of the dikes. Lister and Kerr (1991) note that the width w of a dike approaches 0 at the crack tip. This requires the mean velocity to be very small or the pressure gradient very large. Extension of a fluid-filled crack requires low fluid pressure in the tip promoting exsolution of volatiles. Lister (1990) has, using assumptions of thermodynamic equilibrium, evaluated numerically the effects of volatiles on the dike tip. He uses the equation for the solubility of water in a basalt melt from Wilson and Head (1981), and notes that magmas are saturated in volatiles at pressures corresponding to a lithostatic overburden of at most a few km and probably not more than a few hundred meters. When p_{vol} is large under a lithostatic overburden exceeding the volatile saturation pressure, volatiles should occupy only a small region of the tip. The length of this volatile-filled region extends as the dike propagates toward the level of $p_{vol} = 0$ and the magma is saturated in volatiles (Lister and Kerr 1991). At this volatile saturation level, the rate of bubble nucleation from the magma will be too large to maintain thermodynamic equilibrium between the volatiles in the dike tip and the magma. Volatile exsolution will extend throughout the melt and the system would have to be modeled as a two-phase compressible flow. This mechanism would decrease the density and propagate dikes above the LNB of a volatile undersaturated melt (Lister and Kerr 1991; p. 10,070). Lister (1990) approximated the length of the volatile-filled region and noted that it can become surprising large below the depth of volatile saturation (Lister and Kerr 1991; their fig. 18). This again, may be a critically important mechanism to promote overshooting of the LNB and eruption of magma at the surface.

The viscosity of basalt magma is strongly temperature dependent (Ryan and Blevins 1987). A small drop in temperature caused by heat loss to the country rock promotes a viscosity increase, and an increase in viscous pressure. Delaney and Pollard (1982) note that a magma traveling more than a few kilometers from an upper crustal source at typical propagation velocities will solidify in a few hours (assuming a 2 meter dike thickness). The time for complete solidification of a dike can be approximated by (Lister and Kerr 1991):

$$t_s = \frac{w^2}{\lambda k}$$

where λ is a numerical coefficient dependent on the solidification temperature and thermal properties of the magma and solidification layer and k is the thermal diffusivity of the magma. This calculation neglects the advective supply of heat caused by the flow of the magma and overestimates the likelihood of dike solidification. Bruce and Huppert (1990) emphasized the importance of latent heat of solidification and the effect of thermal advection of magma on the temperature profile of a dike. They used two-dimensional modeling to show that for dike widths of greater than about 1 m, flow blockage from solidification will be reversed by continued supply of heat through magma flow. This reversal can exceed heat loss into the country rock, and result in melting and expansion of dike widths. Bruce and Huppert (1990) also described preliminary three-dimensional modeling of a range of other effects, the most important being the well-described localization of fissure eruptions in distinct conduits through the duration of a fissure eruption. An unconsidered implication of the Bruce-Huppert model, however, is the effect of melting on magma composition. Geochemical and isotopic studies of basalt magmas in a range of settings show that it is rare for basalt to be contaminated with country rock, particularly shallow country rock. If the advective supply of heat is significant over a depth range, country-rock contaminated basalt should be observed more commonly.

Lister and Kerr (1991) conclude that a dike cannot propagate further than a critical length approximated by:

$$l_s = P_e w$$

where Pe is the Peclet number. They suggest that while models are not yet capable of incorporating the mechanics of dike propagation and effects of solidification, several important conclusions can be drawn. First, solidification rates are significant for narrow dikes (tens of cms) over distances of a few kilometers. Narrow dike regions freeze and become narrower, and wide regions tend to melt and become wider. Second, loss of heat from the narrow dike tip will cool the magma and increase viscosity increasing the need for source pressure to maintain flow into the dike tip. This can lead to inflation of a dike behind a blocked dike tip. It may provide a mechanism for observed large dike widths in frozen magma conduits beneath eruptive centers. Third, the earlier discussions of the unimportance of elastic stresses associated with dike propagation suggest the growth of a chilled layer may be easily overcome by dike expansion.

A final problem for this section is the issue of whether a dike path follows a preexisting fracture or fracture system or the dike propagation direction is determined by the stress field. To examine this question, basic data have to be gathered on the geometry of dike systems, their relation to local fractures, and the relationship to other features such as magma reservoirs. Pollard (1987) has noted that many dikes display minor irregularities called cusps, steps, buds and segments. He notes that these irregularities have a length scale that is much less than the dike length, and are not important unless they are near the dike tip. Pollard (1987; his Fig. 9) showed that the principal direction of dike propagation is parallel to the long dimension of these features.

The geometry of a dike is determined at least in part by the path established by the dike tip. A long recognized feature of dikes is that they are emplaced commonly perpendicular to the least compressive stress direction (Pollard 1987). Delaney et al. (1986) described mafic dikes that intruded sedimentary rocks on the Colorado Plateau. These dikes are associated with joint sets in country rock that are closely spaced near the dikes but increase in spacing away from the dikes. They suggest that the joints are formed by fracturing of the host rock by tensile stress generated by magmatic pressure beyond the tips of propagating dikes. The joints become juxtaposed with the dike body with continued propagation. Delaney et al. (1986) contrast dikes with self-generated fractures propagated perpendicular to the least compressive stress direction with dikes that parallel regional joint sets. They suggest that magma can invade older or existing joints if the magmatic pressure exceeds the horizontal stress acting across the joint plane. This may be optimized by two situations. First, it may occur if the horizontal principal stress

difference is small compared to the magmatic driving pressure. Second, it may also occur if joints are nearly perpendicular to the direction of least compressive stress.

Delaney et al. (1986; their appendix A) developed criteria for identifying dikes intruded along older joints. These include field evidence for slip that is substantial in comparison to dilation and formation of crack splays if the dike propagates beyond the ends of the joint. In the latter case, magma must create its own fracture, and would turn or splay toward the direction of maximum principal stress.

Pollard (1987) recognized three fracture modes for the orientation of dikes. The first is a planar dike that follows a fracture produced by the least compressive stress acting perpendicular to the dike plan. The second is a curved dike. This is produced by a spatial rotation of the least compressive stress about an axis parallel to the dike periphery. The third is a segmented dike produced by a spatial rotation of the least compressive stress about an axis parallel to the propagation direction. Pollard (1987) presented criteria needed to verify control of dike paths by preexisting fractures that are not parallel to the least compressive stress direction. These are that the fractures must be older than the dikes, and have comparable planar dimensions. Second, shear displacements must be found along the dike indicating that dilations were accompanied by slip induced by the resolved shear stress across the dike.

Baer and Reches (1991) examined dike propagation mechanisms for a dike swarm crossing different rock units. They noted that intrusion mechanisms were different for each rock unit indicating host rock properties played a role in controlling dike propagation mechanisms, directions, and detailed geometry. Dikes in a stratified sequence formed segments contained within distinct stratified layers. Dikes in massive sandstones formed smaller segments, and had associated dike fingers with intermittent smooth portions and patches with slickensides. The dike propagation through the massive sandstones was inferred to form through alternating stages of fluidization, viscous flow and brittle deformation. These dike systems were emplaced horizontally, along dike-generated fractures. Baer (1991) suggest the dikes propagated at the LNB and were driven by the density difference between the magma and host rock.

We have limited information about the geometry and relationship of feeder dikes for Pliocene and Quaternary basalt centers in the YMR because of the minor degree of erosional modification of the centers. The primary exposure of dikes is in cone scoria where the dikes exhibit highly irregular geometry that is unrelated to the regional stress field (Crowe et al. 1983b).

Several inferences can be offered on the probable form of subsurface dikes in the area. First, the regional alignment of major vents for all Pliocene and Quaternary basalt centers of Crater Flat are northwest (see Chapter 3). This alignment is inferred to be controlled by structural features at depth, and may not be controlled by the shallow stress field. This inference is based on the repeated appearance of basalt centers ranging in age from 4.8 to < 0.1 Ma along a preferential northwest trend. The recurrence of temporally distinct events (contrasted with coeval cone clusters) on the same directional trends, suggests the presence of a long-lived subsurface structural control of the ascent of magma. Additionally, as shown in Chapter 3, the direction of alignment of Pliocene and Quaternary volcanic centers coincides with the surface of maximum magma eruption volumes. The strong correlation between center alignment and eruption volumes provides significant support for the rise of magma at depth along northwest-trending structures. This conclusion is in strong contrast with Lister and Kerr (1991) who, as noted earlier, conclude that rock is weak, and preexisting fractures are not important in dike propagation. Rubin (1993) presented arguments that the mechanical properties of country rock are important. The most compelling argument is the apparent discrepancy between laboratory measurements of fracture energy compared with field observations of country rock effects from dike propagation. Field evidence demonstrates that inelastic deformation at dike tips can be much larger than those produced at the tips of tensile cracks in laboratory experiments. Moreover, deformation of country rock is observable adjacent to dikes for a range of rock types.

An alignment of clusters of scoria cones of probable similar age provide probably the best indicator of the local trend of shallow feeder dikes (for example, Nakamura 1977). These alignments, in

the YMR, include the north-south cluster of the 3.7 Ma centers, the north-northeast cluster of the Quaternary basalt of Crater Flat, and the north-northeast cluster of the basalt of Sleeping Butte. These directions are perpendicular or near-perpendicular to the least compressive stress direction (Stock et al. 1985). The Lathrop Wells volcanic center consists of only one scoria cone, and no structural trend can be assigned to a single cone. However, the predominant direction of fissure systems in the center are northwest. The two distinct structural settings of basalt centers in Crater Flat (northwest-trending localization of vents of different ages; northeast-trending, coeval cone clusters) probably requires reorientation of dikes at a shallow depth.

A preferred model for dike emplacement in the Yucca Mountain area is ascent of pulses of basalt magma at depth along northwest-trending structures followed by a shallow north-northeast reorientation of dikes (90 degrees) parallel to the maximum compressive stress direction. This is consistent with the lengths of the different structures. The northwest-trending Crater Flat volcanic field (Crowe and Perry 1989), extends for over 50 kilometers. This exceeds the maximum length of known dikes except bladed dikes, propagating laterally from a shallow magma reservoir. Northeast trending cone alignments range from 2.6 to 12 km, consistent with formation from individual feeder dikes. The dike model may be tested with geophysical data, particularly aeromagnetic data. It may be possible to use geophysical methods to determine if there is a change in dike orientation from north-northwest to northwest with depth. If this reorientation is recognized, the depth of the reorientation compared with the depth of the potential repository at Yucca Mountain may provide key information on the likely depth of occurrence of basalt intrusions. This dike reorientation could coincide with a reduction in confining pressure for two regimes of dike propagation (Rubin 1993): northwest following structure at higher confining pressure, and northeast associated with magma-induced fractures at lower confining pressures.

Finally, it is difficult to establish a relation between vent alignments, and local structure for individual basalt centers in the YMR. Alignments of fissures, vents, scoria mounds and scoria cones define conjugate northwest and north-northeast trends (Crowe and Carr 1980; Vaniman et al. 1982; Crowe et al. 1983; Crowe 1990, Smith et al. 1990; Ho et al. 1991). These directions parallel the expected directions of dike and structural trends. But the data sets are inconsistent (see Chapter 3). Some basalt centers are located along structures or intersections of multiple structures; others appear independent of structure. Exposure of dikes in country rock associated with the Pliocene and Quaternary basalt centers is insufficient to identify structurally controlled or stress-controlled dike features using the criteria of Delaney et al. (1986) and Pollard (1987). If these features could be identified, northwest-trending dikes would be expected to be structurally controlled (fracture controlled), and north-northeast trending dikes would be stress-controlled.

5.4.6. Future Work

Future expanded work on processes of magma dynamics was planned as part of Study Plan 8.3.1.8.1.2, *Physical Processes of Magmatism and Effects on the Potential Repository*. Our intent is to use the foundation of surveyed literature from Section 5.4 as a starting point for further examination of the physical controls of magma generation, evolution, ascent, and eruption in the YMR. The purpose of this work is to provide an independent cross-check on probabilistic studies. Rates of volcanic activity for probabilistic volcanic risk assessment are established primarily from the volcanic record of the YMR (see Chapter 7). The magma dynamics studies will examine physical processes of magmatism in the YMR using geophysical and petrologic data to constrain rates of magma production. The purpose of the work was to ensure that magmatic processes as represented in probabilistic estimates are compatible with physical and geochemical data for the YMR. A schematic overview of this work is provided by Fig. 5.1, a cross-section of the crust-mantle in the YMR. Generation of basalt melt in the region may have occurred in either the convecting asthenosphere or in the lithospheric mantle. Isotopic studies of basalts of the YMR show that they must be derived from or have equilibrated with lithospheric mantle to obtain their distinctive Sr and Nd isotopic compositions. This provides two important constraints on models of basalt genesis. First, if melt initiation is assumed to occur in the asthenosphere, then extracted melt must move through the lithospheric mantle through matrix percolation and not channel segregation. The latter process would isolate the melt from the matrix. Second, there must be substantial vertical continuity of

preserved lithospheric mantle in the YMR. The Younger postcaldera basalt has decreased systematically in volume and exhibited geochemical trends from the Pliocene to the Quaternary that are consistent with decreasing degrees of partial melting or deepening of the depth of melting. Yet the isotopic composition of Sr and Nd have remained uniform.

The issue of formation of solitary waves associated with melt migration remains controversial. Such waves if preserved through magma ascent, are most likely to occur for areas of low magma flux when melt extraction is non-steady state. Both conditions may be appropriate for the waning tectonic setting of the YMR. The velocity and volume of melt (1 km^3 extracted; $< 0.1 \text{ km}^3$ erupted) associated with solitary waves could contribute to the episodic nature of polygenetic volcanism at some basalt centers. The dynamics of these processes will be examined to determine if their time constraints and volumes are consistent with observed patterns of volcanism in the YMR.

The evolved compositions and scarcity of phenocryst phases in the lavas of the Younger postcaldera basalt (particularly the Quaternary lavas) requires that they fractionated extensively prior to ascent and eruption, and the fractionating phases were removed (see Chapter 4). The most logical place for that fractionation and crystal removal to take place is at the mantle-crust density face, below the plagioclase stability field.

Finally, the ascent of basalt magma below the YMR almost certainly occurs rapidly in the form of basalt dikes. We will examine the volume constraints on the minimum volumes required to initiate an ascent pulse of magma, and the ratio between the volume of magma resident in ascent versus eruption. Unique aspects of Quaternary basaltic volcanism in the YMR are their small volumes and episodicity. Average event recurrence times for the formation of new volcanic centers are about 200 to 300 ka; event recurrence times for polygenetic events are on the order of 10 to 100 ka. It would be useful to compare these temporal constraints with dynamic processes of magma ascent. Magma ascent in the shallow crust may be controlled at a depth of 2-10 km by the presence of deep-seated structural features; shallow propagation of dikes may occur as swarms following the local stress field (maximum compressive stress direction). Assessment of the importance of the LNB, crack-tip fracture interaction, and the local stress field may be important in identifying high probability sites for future eruption of basalt magma.

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"Section 5.2.3. Analog Studies in the Lucero Volcanic Field, New Mexico" is "Q." This material has been published (Valentine and Groves, 1996), and the data tracking number for this record package segment is LA000000000129.001. Data in this section are derived from field notebook TWS-EES-5-6-93-01. Point counting was done following Los Alamos YMP procedure TWS-EES-DP-102, R2.

"Section 5.2.4. Analog Studies in San Francisco Volcanic Field, Arizona" is "Q." Data in this section are derived from field notebook TWS-EES-5-6-93-01.

"Section 5.3.2. Controls on Geometry of Shallow Basaltic Intrusions - Studies from the Paiute Ridge Analog Site, Southern Nevada" is "Q". Data in this section are partly derived from field notebook TWS-EES-5-01-93-01.

"Section 5.3.3.2. Paiute Ridge/Slanted Buttes Analog Site" is "Q." Data in this section are derived from field notebooks TWS-EES-5-01-93-01 and LA-EES-1-NBK-95-006.

"Section 5.3.3.3. Grants Ridge Analog Site" is "Q." Data in this section are derived from field notebooks TWS-EES-5-01-93-01 and LA-EES-1-NBK-95-006.

Subsections 5.3.4.2.1 through 5.3.4.2.4 of "Section 5.3.4.2 Finite Amplitude Analysis of Natural Convection of Compressible Gas in Porous Media" is "Q." The remaining subsections report results of numerical modeling carried out in a prototyping mode and are not "Q".

"Section 5.3.4.3. Spatial Scales of Effects on the Potential Yucca Mountain Repository Due to Convective Air Flow Induced by Igneous Intrusions" reports results of numerical modeling carried out in a prototyping mode and are not "Q."

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Table Headings for Chapter 5

Table 5-1. Xenolith abundance data for Lucero Volcanic Field, New Mexico.

Table 5-2. Volume fractions of sedimentary xenoliths from South Merriam and Dry Lake volcanoes, Strombolian facies.

Table 5-3. Representative major element oxide composition of Paiute Ridge alkali basalt, determined by x-ray fluorescence.

Table 5-4. Estimated and calculated(with errors) mineralogical compositions of selected tuffs from the Ammonia Tanks Tuff (a), the Tiva Canyon Tuff and other Paintbrush Group tuffs (b), the Calico Hills Formation (d), and pumice fall of the Paintbrush Group (e).

Table 5-5. Major and trace element compositions of the Ammonia Tanks Tuff (PR95-1 to 10 and 12 to 15) and a dike within the tuff (PR95-11), the Tiva Canyon Tuff and other Paintbrush Group tuffs (PR95-18 to 22), the Calico Hills Formation (PR95-23), a pumice fall of the Paintbrush Group (PR95-24-27) and a basalt within the pumice fall (PR95-28).

Table 5-6. Rare earth element and other major element compositions of selected samples analyzed by instrumental neutron activation analysis (INAA). Concentrations are given in parts per million. Selected samples were also analyzed for B, Br, Cl, F, Li, S, and P_2O_5 using Ion Chromatograph.

Table 5-7. Major and trace element compositions of silicic tuffs (GR95-1, 2, 3, 6, 7, 8, 10, 11, 15, and 18) and lavas (GR95-12 and 14), volcanoclastic sediments (GR95-5, 17, 19, 20, and 21, and basalt flows (GR95-4, 9, 13, and 16) from East Grants Ridge, west-Central New Mexico. Samples were analyzed using an X-ray fluorescence. Rare earth element and other major element compositions of selected samples of silicic tuff (GR95-6, 8, 7, and 15) and lava (GR95-12), and basalts (GR95-4 and 16), analyzed using instrumental neutron activation analysis (INAA). Concentrations are given in parts per million.

Table 5-8. Major element compositions of volcanic glass of selected tuffs (GR95-1, 6, 7, 8, and 15) from East Grants Ridge, west-central New Mexico analyzed using an electron microprobe. Compositions are given in weight percent. Water content calculated by difference.

Table 5-9. Halogens and other volatile concentrations of selected tuff samples from East Grants Ridge, west-central New Mexico. Concentrations are given in part per million.

Table 5-10. Physical properties of air and water at 293K and 373K.

Figure Captions for Chapter 5

Figure 1. Simplified geologic map of the northern part of the Lucero volcanic field, showing late Cenozoic basalts, Alkali Buttes, and Volcano Hill. Also shown are the locations of the Sierra Lucero uplift and two petroleum wells, all of which provided stratigraphic control on subvolcanic stratigraphy. Inset shows location of map area within the New Mexico and relative to the Colorado Plateau and Rio Grande Rift. Adapted from Figures 1 and 2 of Baldrige et al. (1987).

Figure 2. Geologic map of Alkali Buttes, showing distribution of volcanic facies, bed orientations, and station locations.

Figure 3. Four stratigraphic columns from Alkali Buttes (measured sections S1, S7, N4, N6 - see Figure 2 for map locations). Thicknesses have been corrected for the average dip at each measured section. The columns for stations S7 and N4 record the entire succession of pyroclastic facies at Alkali Buttes, with lapilli and block-rich tuff at the base. Stratigraphic locations of lithic abundance measurements (Table 5-1, Figures 6, 7, 9, 10) are shown.

Figure 4. Geologic map of Volcano Hill, showing distribution of volcanic facies, bed orientations, and station locations.

Figure 5. Maximum sizes of juvenile (MJ) and xenolith (ML) clasts and total volume fraction of xenoliths. The observations are arranged according to facies types. Note that the vesicular spatter and scoria facies has the largest values of MJ, and that the highest xenolith volume fractions occur in the lapilli and block-rich tuff facies. One example of the poorly vesicular lapilli facies from South Alkali Butte, indicated with an arrow, lacks xenoliths and is arbitrarily plotted at a volume fraction of 10^{-5} .

Figure 6. Entrainment rate plots for measured section S1 at southern Alkali Butte. Site numbers identified in lower right-hand corners of plots (see Figure 3 for stratigraphic location of each site). S1-2 and S1-3 were measured in the same layer of poorly vesicular lapilli. S1-4 and S1-5 are from vesicular scoria and spatter facies.

Figure 7. Entrainment rate plots for measured section N6 at northern Alkali Butte. Site numbers identified in lower right-hand corners of plots (see Figure 3 for stratigraphic location of each site). N6(low) is from poorly vesicular lapilli facies, N6(high) is from vesicular scoria and spatter facies.

Figure 8. Entrainment rate plots for Volcano Hill. The entire section is vesicular scoria and spatter. Site numbers identified in lower right-hand corners of plots (see Figure 4 for map location of each site). Note the relatively consistent entrainment of Glorieta material.

Figure 9. Entrainment rate plots for measured section S7 (including site S8, which is stratigraphically lower than S7-1) at southern Alkali Butte. Site numbers identified in lower right-hand corners of plots (see Figure 3 for stratigraphic location of each site). S8 and S7-1 are from lapilli and block rich tuff, S7-2 and S7-3 are from vesicular spatter and scoria facies. Note that the entrainment rates for Chinle material in the lapilli and block rich tuff are much higher than in other facies, entrainment of San Andres and Glorieta material is similar to other facies shown in previous figures.

Figure 10. Entrainment rate plots for measured section N4 at northern Alkali Butte. Site numbers identified in lower right-hand corners of plots (see Figure 5 for stratigraphic location of

each site). N4(low) is from lapilli and block rich tuff, while N4(high) is from poorly vesicular lapilli facies.

Figure 11. Outline of San Francisco Volcanic Field, north central Arizona. With the exception of the major silicic centers shown, most of the enclosed area is covered by basaltic rocks. Inset shows location within Arizona, and the approximate boundary between the Colorado Plateau and Basin and Range Provinces. Adapted from Arculus and Gust (1995).

Figure 12. Geologic map of eastern SFVF, showing distribution of basaltic vents exposed at the surface. Locations of the four centers studied (Dry Lake, Sheep Hill South, Rattlesnake Crater, and Merriam East) are shown, as well as that of Sunset Crater (the youngest center in the volcanic field). Major highways indicated. Map based on Ulrich et al. (1984).

Figure 13. Simple geologic map of Merriam and East Merriam cones. Quarry is where lithic abundances were measured.

Figure 14. Simple geologic map of Dry Lake volcano, showing station locations.

Figure 15. Simplified geologic map of the Paiute Ridge area, emphasizing basaltic rocks. Modified from Goff et al. (1995), Crowe et al. (1983) and Byers and Barnes (1967). Nomenclature of volcanic units follows Sawyer et al. [1994]. Cross section lines in Figure 2 are indicated. Pz - Paleozoic basement rocks (mainly carbonates) undivided; Tp - undivided bedded tuffs including units belonging to Tunnel Formation, Indian Trail Formation, Paintbrush group, comprised of fallout, pyroclastic flow deposits, and reworked nonwelded tuffs with varying degrees of zeolitic alteration; Tm - Timber Mountain tuffs, undivided (includes Rainier Mesa member); Qac - Quaternary alluvium and colluvium. PR - Paiute Ridge fault and dike; W - W sill; M - M dike and fault; E - E dike and fault; ES - ES sill along E dike near volcanic neck N; EN - EN sill along E dike.

Figure 16. Cross sections through Paiute Ridge area (A-A' and B-B' in Figure 1) showing present day topography and geology, with inferred topography (dashed) at the time of basaltic activity. Letters are the same as in Figure 1. Note that the sills (W sill in B-B' and ES sill in A-A') intrude Tertiary tuffs just above the Tertiary-Cambrian unconformity. X and Y (of section B-B') represent points at which vertical stresses are unequal (6-7.6 MPa and 5.2 MPa respectively based on estimated overburden; see *Interpretation* section) within the hangingwall and footwall of the Paiute Ridge fault.

Figure 17. Lower hemisphere stereogram of structures at Paiute Ridge. Great circles show surfaces and kinematic indicators: fm = fault surface confining m dike and slickenside; de = E dike with flow indicator; dp = PR dike with flow indicator, Sen = layering in EN sill, Sw = layering in W sill. Isolated dots represent poles to dike orientations: p = PR dike, e = E dike, m = M dike; and + indicate layering: w = layering in W dike, b = layering in tuff adjacent to intrusive bodies, e = layering in E dike. px is an anomalous fabric within PR dike that is perpendicular to its strike. Fabric elsewhere along dikes is coplanar with the dike margin.

Figure 18. Possible stress gradients for conditions during intrusive activity in the Paiute Ridge area. Depth is from paleosurface. Diamond = σ_h , X = σ_v , and Circle = σ_h + tectonic compression. Intrusion of sills would be favored at the depth where $\sigma_v < \sigma_h$. With a small tectonic compression (4 MPa) σ_h can exceed σ_v at depths shallower than approximately 300m (depth = 250 m is approximate Tertiary-Paleozoic boundary). Calculations for σ_v and σ_h are

standard ($\sigma_v = \rho gh$ and $\sigma_h = (\nu/1-\nu)\rho gh$: see e.g., Engelder [1993]). Rock properties are described in the text.

Figure 19. Geologic map of the Slanted Buttes, Half Pint Range, Nevada. Inset map shows the location of Slanted Buttes. The geologic map indicates the sample locations of tuffs intruded by dikes and a sill.

Figure 20. X-ray diffraction patterns of representative bulk samples from the Ammonia Tanks Tuff at the southeastern part of the Slanted Buttes. Samples collected at different intervals from the dike. For example at 34 m (A=PR95-1), 2.8 m (B=PR95-8), 2.4 m (C=PR95-9), 1.6 m (D=PR95-10), recrystallized basalt from the contact (E=PR95-11), and 6 m (F=PR95-15). Patterns represent smectite (Sm), mica (M), clinoptilolite (Cp), feldspar (F), analcime (An), quartz (Q), cristobalite (Cr), and pyroxene (P).

Figure 21. X-ray diffraction patterns of representative bulk samples from the Tiva Canyon Tuff at the southeastern part of the Slanted Buttes. The patterns represent a vitrophyre directly below the sill (A=PR95-18), a meter below the vitrophyre (B=PR95-19), 3 m (C=PR95-20), devitrified tuff from the contact (D=PR95-21), and 8 m below the sill (E=PR95-22). Patterns represent smectite (Sm), mica (M), clinoptilolite (Cp), feldspar (F), quartz (Q), cristobalite (Cr), and opal (Op).

Figure 22. X-ray diffraction patterns of representative bulk samples from a pumice fall in the northern part of Slanted Buttes. Samples collected at the contact with the dike (A=PR95-24) and 1.8 m away (B=PR95-27). Patterns represent feldspar (F), quartz (Q), and cristobalite (Cr).

Figure 23. Variation diagrams of major elements (wt %) versus SiO_2 . Analysis calculated to 100 % volatile free. Symbols represent samples from the Ammonia Tanks Tuff (open squares=PR95-1 to 7; solid squares=PR95-8 to 10 and 12 to 13), dike and plug samples in Ammonia Tank Tuff (open square filled with cross), Tiva Canyon Tuff (solid circle=PR95-18 and 19, open circle=PR95-20 to 23), pumice fall (solid triangle=PR95-24 to 27), and dike in pumice fall (open triangle=PR95-28).

Figure 24. Conventional chondrite-normalized rare earth element plots of the Ammonia Tanks Tuff (A=PR95-1, 8, and 10), a dike within the same tuff (A=PR95-16), and the Tiva Canyon Tuff (C=PR95-18 to 22). Plots B and D represent multi-element chondrite-normalized spidergrams of the same samples plotted in A and C (Nakamura, 1974; Thompson et al., 1984).

Figure 25. Distribution of volcanic rocks of the Pliocene Mt. Taylor and the Pleistocene Zuni-Bandera volcanic fields (A) of west-central New Mexico (Laughlin et al., 1993). A detailed geologic map of the East Grants Ridge located along the southeastern part of the Mt. Taylor volcanic field (A) is also given (B). The basaltic plug is located in the middle of the map (Thaden et al., 1967).

Figure 26. Panoramic view of the basaltic plug and the silicic ash flow tuff (White) of East Grants Ridge taken from the main highway looking northwest. Numbers represent sample locations of silicic tuffs and lavas, volcaniclastic sediments, and basalt flows from the plug.

Figure 27. (a) X-ray diffraction patterns of representative silicic tuffs and lavas and volcaniclastic sediments collected at different distances from the plug. The patterns represent a rhyolite lava (A=GR95-12), pumice from the contact (B=GR95-1a), bulk tuff (C=GR95-6b, D=GR95-8a, and E=GR95-15a), and a volcaniclastic sediment (F=GR95-21). Patterns represent mica (M), feldspar (F), quartz (Q), and cristobalite (Cr). Samples C and E were collected about 25 and 300 m, respectively, away from the plug (Fig. 26). (b) X-ray diffraction patterns of representative silicic tuffs and lavas and volcaniclastic sediments collected at different distances from the plug. The patterns represent a volcaniclastic sediment (A=GR95-20), a vitrophyre (B=GR95-14), bulk tuff (C=GR95-1b), pumice (D=GR95-6, E=GR95-8b, and F=GR95-15b). Patterns represent gypsum (G), mica (M), clinoptilolite (Cp), feldspar (F), quartz (Q), cristobalite (Cr), and calcite (Ca).

Figure 28. Variation diagrams of major elements versus SiO_2 (wt %). Analysis calculated to 100 % volatile free. Symbols represent samples of basalts (solid squares), pumice clasts (open squares), rhyolite lava (open square filled with cross), volcanoclastic sediments (solid circle), and ash flow tuff (open circle).

Figure 29. Variation diagrams of major elements compositions of volcanic glass from microprobe analyses versus SiO_2 (wt %). Analysis calculated to 100 % volatile free. Symbols represent selected samples of silicic tuffs: GR95-1 (solid triangle), GR95-6 (open squares), GR95-7 (solid circle), GR95-8 (inverted open triangle), and GR95-15 (open circle).

Figure 30. Plots of elemental concentrations of B, Cl, F, Li, P_2O_5 , and S versus distance from the basaltic plug. Legends of symbols are given in boxes.

Figure 31. Variation diagrams of trace element compositions in parts per million versus SiO_2 (wt %). Symbols represent samples of basalts (solid squares), pumice clasts (open squares), rhyolite lava (open square filled with cross), volcanoclastic sediments (solid circle), and ash flow tuff (open circle).

Figure 32. Chondrite-normalized rare earth element plots of silicic tuffs and basaltic lavas. Symbols and the corresponding samples are given in the legend. Plot B represents multi-element chondrite-normalized spidergrams of the same samples plotted in A (Nakamura, 1974; Thompson et al., 1984).

Figure 33. Domain, boundary and initial conditions for finite amplitude numerical simulations reported in this study.

Figure 34. Rate of change of temperature at a point vs. time. We defined steady-state approximately the 95% percentile of these curves.

Figure 35. Nu vs. Ra for air from low Ra FEHM simulations.

Figure 36. Nu vs. Ra for air and water from FEHM simulations. Filled circles represent results for air convection. Open circles represent results for water convection.

Figure 37. Time to steady-state vs. Nu for air from FEHM simulations for systems with several different thermal gradients and different heights.

Figure 38. Domain and boundary conditions.

Figure 39. Particle and temperature fields for 180 m dike at various times. The particle contour separates the region with particles present from the region with no particles present.

Figure 40. Particle and temperature fields for 360 m dike at various times. The particle contour separates the region with particles present from the region with no particles present.

Figure 41. Particle and temperature fields for 560 m dike at various times. The particle contour separates the region with particles present from the region with no particles present.

Figure 42. Particle and temperature fields for 680 m dike at various times. The particle contour separates the region with particles present from the region with no particles present.

Figure 43. Particle and temperature fields for a sill at the water table at various times. The particle contour separates the region with particles present from the region with no particles present.

Figure 44. Particle and temperature fields for a permeable and an impermeable sill at 360 m at various times. The particle contour separates the region with particles present from the region with no particles present. The temperature fields for these two simulations appear identical.

Figure 45. Particle and temperature fields for a permeable and an impermeable sill at 520 m at various times. The particle contour separates the region with particles present from the region with no particles present. The temperature fields for these two simulations appear identical.

Figure 46. Particle fields for a 560 m dike at various times for different porosity values. The particle contour separates the region with particles present from the region with no particles present.

Figure 47. Particle fields for a 560 m dike at various times for different permeability values. The particle contour separates the region with particles present from the region with no particles present.

Figure 48. Schematic cross section of the southern Great Basin showing the distribution of the crust, lithospheric mantle, and asthenospheric mantle. Two end member processes for generation of basaltic melts are possible. The first is decompressive melting of the asthenosphere. Basalt derived from this source is inferred to migrate by channel segregation through the thinned lithospheric mantle without equilibrating, therefore maintaining its asthenospheric compositions for Sr and Nd. The second end member, which is more appropriate to the YMR, is generation of basaltic melt by hydrous melting of the heterogeneous lithosphere. An alternative hybrid source for basaltic magmas of the YMR is decompressive melting of the asthenosphere followed by matrix percolation of melt through the lithospheric mantle allowing equilibration with the lithospheric mantle. Both the hydrous melting end member and the hybrid asthenospheric-lithospheric equilibrated melts must be trapped and fractionate at the crust-mantle density interface, followed by rapid ascent through the crust.

Table 5-1. Xenolith Abundance Data for Lucero Volcanic Field

Facies Type	Site ^a	Parameter ^b	Chinle	San Andres	Glorieta	Abo	Madera	Total Volume Fraction
LAV	N5	AF	1.6×10^{-3}	1.2×10^{-3}				1.8×10^{-4}
		VF	1.1×10^{-4}	6.8×10^{-5}	...	...	...	
		VF/m	2.2×10^{-7}	5.2×10^{-7}				
VA	VH7	AF			3.4×10^{-3}			3.3×10^{-4}
		VF	...	...	3.3×10^{-4}	...	...	
		VF/m			5.5×10^{-6}			
VESICULAR	S1-4 ^c	AF	2.7×10^{-3}	2.0×10^{-3}	2.1×10^{-3}		1.1×10^{-3}	6.0×10^{-4}
		VF	2.3×10^{-4}	1.5×10^{-4}	1.6×10^{-4}	...	5.8×10^{-5}	
		VF/m	4.5×10^{-7}	1.2×10^{-6}	2.6×10^{-6}		2.5×10^{-7}	
CULAR	S1-5 ^c	AF	2.7×10^{-3}		1.1×10^{-3}			2.9×10^{-4}
		VF	2.3×10^{-4}	...	5.9×10^{-5}	...	...	
		VF/m	4.5×10^{-7}		9.8×10^{-7}			
AR	S7-2	AF		2.1×10^{-3}				1.6×10^{-4}
		VF	...	1.6×10^{-4}	...	...	...	
		VF/m		1.2×10^{-6}				
SCOR	S7-3	AF	2.8×10^{-3}					2.4×10^{-4}
		VF	2.4×10^{-4}	...	...	...	...	
		VF/m	4.8×10^{-7}					
ORIA	N2	AF	6.9×10^{-4}	4.2×10^{-4}	1.6×10^{-3}			1.5×10^{-4}
		VF	3.0×10^{-5}	1.4×10^{-5}	1.1×10^{-4}	...	...	
		VF/m	5.9×10^{-8}	1.1×10^{-7}	1.8×10^{-6}			
AAND	N3	AF	6.4×10^{-3}					8.4×10^{-4}
		VF	8.4×10^{-4}	...	...	...	...	
		VF/m	1.6×10^{-6}					
SPATTER	N6 (high)	AF	1.8×10^{-3}	5.0×10^{-5}	2.6×10^{-2}			7.0×10^{-3}
		VF	1.3×10^{-4}	5.8×10^{-7}	6.9×10^{-3}	...	...	
		VF/m	2.5×10^{-7}	4.5×10^{-9}	1.2×10^{-4}			
R	VH1	AF			1.1×10^{-3}			6.1×10^{-5}
		VF	...	...	6.1×10^{-5}	...	...	
		VF/m			1.0×10^{-6}			
TTER	VH2	AF			5.3×10^{-4}			2.0×10^{-5}
		VF	...	...	2.0×10^{-5}	...	...	
		VF/m			3.4×10^{-7}			
R	VH3	AF	1.1×10^{-3}					5.6×10^{-5}
		VF	5.6×10^{-5}	...	...	...	...	
		VF/m	1.1×10^{-7}					

Table 5-1 (Continued)

Facies Type		Site ^a	Parameter ^b	Chinle	San Andres	Glorieta	Abo	Madera	Total Volume Fraction
P O R L Y I V E S I C U L A R	L	VH4	AF	6.7×10^{-4}		2.2×10^{-3}			
			VF	2.8×10^{-5}	...	1.7×10^{-4}	...	...	2.0×10^{-4}
			VF/m	5.5×10^{-8}		2.8×10^{-6}			
	A	VH5	AF	5.4×10^{-3}		2.0×10^{-4}			
			VF	6.5×10^{-4}	...	4.6×10^{-6}	...	...	6.5×10^{-4}
			VF/m	1.3×10^{-6}		7.7×10^{-8}			
	I	VH6	AF			1.8×10^{-3}			
			VF	...	...	1.3×10^{-4}	...	...	1.3×10^{-4}
			VF/m			2.1×10^{-6}			
	P	S1-1	AF						
			VF	...	...	...	...	...	0
			VF/m						
	L	S1-2 ^c	AF	2.7×10^{-2}		2.9×10^{-4}	1.4×10^{-3}		
			VF	6.1×10^{-3}	...	8.0×10^{-6}	8.5×10^{-5}	...	6.2×10^{-3}
			VF/m	1.2×10^{-5}		1.3×10^{-7}	3.0×10^{-7}		
	I	S1-3 ^c	AF	1.3×10^{-3}	3.6×10^{-3}	1.6×10^{-3}		4.7×10^{-3}	
			VF	8.1×10^{-5}	3.6×10^{-4}	1.1×10^{-4}	...	5.4×10^{-4}	1.1×10^{-3}
			VF/m	1.6×10^{-7}	2.7×10^{-6}	1.8×10^{-6}		2.3×10^{-6}	
	C	N4 (high) ^d	AF	3.7×10^{-2}	9.0×10^{-3}	1.4×10^{-2}			
			VF	1.2×10^{-2}	1.4×10^{-3}	2.8×10^{-3}	...	...	1.6×10^{-2}
			VF/m	2.3×10^{-5}	1.1×10^{-5}	4.7×10^{-5}			
	U	N6 (low)	AF	1.0×10^{-3}	6.5×10^{-4}	6.6×10^{-4}			
			VF	5.3×10^{-5}	2.7×10^{-5}	2.8×10^{-5}	...	...	1.1×10^{-4}
			VF/m	1.0×10^{-7}	2.1×10^{-7}	4.6×10^{-7}			

Table 5-1 (Continued)

Facies Type		Site ^a	Parameter ^b	Chinle	San Andres	Glorieta	Abo	Madera	Total Volume Fraction
L A P I L L I & B L O C K	R I C H	S7-1 ^c	AF	5.4×10^{-3}	4.4×10^{-3}	2.9×10^{-3}	...	...	.41
			VF	6.5×10^{-4} (0.41)	4.8×10^{-4}	2.6×10^{-4}			
			VF/m	8.1×10^{-4}	3.7×10^{-6}	4.3×10^{-6}			
	T U F F	S8 ^c	AF	7.3×10^{-2}	9.3×10^{-4}	5.1×10^{-3}	...	...	.32
			VF	3.2×10^{-2} (0.30)	4.6×10^{-5}	6.0×10^{-4}			
			VF/m	6.5×10^{-4}	3.6×10^{-7}	1.0×10^{-5}			
	N4 (low)	S9 ^c	AF	5.7×10^{-3}	3.0×10^{-3}	4.9×10^{-3}	...	...	.46
			VF	7.1×10^{-4} (0.46)	2.7×10^{-4}	5.6×10^{-4}			
			VF/m	9.0×10^{-4}	2.1×10^{-6}	9.3×10^{-6}			
O T H E R	C O N S	Lathrop Wells ^f	AF	0.13	2.6×10^{-4}	8.8×10^{-4}	...	...	.91
			VF	2.5×10^{-3} (0.91)	6.9×10^{-6}	4.3×10^{-5}			
			VF/m	1.8×10^{-3}	5.3×10^{-8}	7.2×10^{-7}			
O T H E R	C O N S	Lathrop Wells ^f					...	...	9×10^{-5}
O T H E R	C O N S	Lathrop Wells ^f					...	...	2.2×10^{-4}
O T H E R	C O N S	Lathrop Wells ^f					...	...	3.2×10^{-4}
O T H E R	C O N S	Lathrop Wells ^f					...	...	3.8×10^{-4}
O T H E R	C O N S	Lathrop Wells ^f					...	...	6.0×10^{-4}

^a Site nomenclature - S, South Alkali Butte; N, North Alkali Butte; VH, Volcano Hill. First numeral is site number. Numeral after dash is measured area number where several areas were studied at a single site.

^b AF is the calculated area fraction of xenoliths; VF is calculated volume fraction; VF/m is entrainment rate of xenoliths (volume fraction derived per meter in dike/conduit).

^c Varying proportions of xenoliths at these measured areas were difficult to correlate with a subvolcanic stratigraphic unit, mainly because these measurements were made early in the study before we were familiar with detailed characteristics of subvolcanic units. Uncorrelated xenoliths make up the following portions of these sites: S1-4, 63%; S1-5, 91%, S1-2, 6%; S1-3, 21%. Total volume fraction estimates for these sites are accurate, but entrainment rates for individual formations should be viewed with caution. Shales were assumed to be derived from Chinle Fm., limestones from San Andres Fm., and sandstones from Glorieta Fm., for entrainment rate calculation.

^d Highly friable sandstone clasts were not correlated with a subvolcanic unit when data were collected. We assume here that these were derived from similar sandstone horizons within the Chinle Fm.

^e AF values were determined for xenoliths >1cm (long dimension) at the outcrop, as described in text. VF values not in parentheses correspond to the field data, while values inside parentheses are the volume fractions within the matrix (<1 cm size grains) determined by point counting of thin sections. The two main matrix constituents not listed in the table are calcite, which lines and/or fills pore spaces, and juvenile basalt fragments. These comprise the following volume fractions: site S7-1 - 0.55 basalt, 0.04 calcite; site S8 - 0.59 basalt, 0.11 calcite; site S9 - 0.49 basalt, 0.05 calcite; site N4(low) - 0.05 basalt, 0.04 calcite. See text for discussion of stratigraphic correlation of matrix xenolith grains.

^f Data from the other scoria cones in Arizona and Nevada, reported in Crowe et. al. (1983). These would fall under vesicular scoria and spatter facies in the present report. SFF - San Francisco Field, Arizona.

Table 5-2. Volume fractions of sedimentary xenoliths from South Merriam and Dry Lake volcanoes, Strombolian facies.*

South Merriam Cone**

	<u>Station MS-1</u>	<u>Station MS-2</u>
Area fraction	3.8×10^{-4}	1.6×10^{-3}
Volume fraction	1.2×10^{-5}	1.1×10^{-4}

Dry Lake Volcano (Strombolian facies)***

	<u>Station 3</u>	<u>Station 4</u>	<u>Station 5</u>
Area fraction	9.3×10^{-4}	0	1.5×10^{-3}
Volume fraction	4.8×10^{-5}	0	9.6×10^{-5}

*Area fraction and volume fractions determined for xenoliths >1cm (long dimension) at outcrop, as described in text.

**All xenoliths observed at this site were identified as Kaibab Limestone.

***All xenoliths observed at these stations were basaltic lava fragments, Kaibab Limestone, or Coconino Sandstone.

Table 5-3. Representative major element oxide composition of Paiute Ridge alkali basalt*, determined by x-ray fluorescence analysis.

Oxide	Weight %
SiO ₂	48.92
TiO ₂	2.84
Al ₂ O ₃	17.35
FeO	10.95
MgO	5.28
CaO	8.12
Na ₂ O	4.17
K ₂ O	1.70
P ₂ O ₅	0.58
TOTAL	99.92

* Sample B-13, reported in Appendix A of *Crowe et al.* [1986]

Sample Number	Distance to dike/sill (m)	Volcanic glass	Mica	Feldspar	Quartz	Cristobalite	Tridymite	Opal-CT	Clinoptilolite	Calcite	Smectite
PR95-1P1 ^a	34.0	60	1.2	31.6	4.8	0.8					
PR95-2P1 ^a	28.9	60	1.2	36.8	1.6	0.4					
PR95-3P1 ^a	22.4	60	0.8	35.6	3.2						
PR95-4P1 ^a	14.6	60	1.2	32.4	5.6	0.4					
PR95-5P1 ^a	12.0	50	1.5	44.5	4.0						
PR95-6P1 ^a	7.8	50	2.0	41.0	7.5						
PR95-7P1 ^a	4.4	55	1.8	37.9	4.9	0.4					
PR95-8P1 ^a	2.8		2.0	50.0	5.0	4.0			40.0		
PR95-9P1 ^a	2.4	31	2±1	55±8	7±1	4±1	1±1				
PR95-10P1 ^a	1.6		1±1	69±10	9±1	13±1	9±1				
PR95-11P1	contact zone			83.0		1.0					16.0
PR95-12P1 ^a	contact zone			69±10	9±1	9±4					
PR95-13P1 ^a	1.25		1±1	64±9	13±1	11±1	12±1				
PR95-14P1 ^a	2.85		1±1	62±9	16±1	10±1	10±1		1±1		
PR95-15P1 ^a	6.0		1±1	57±8	24±2			6±3	3±1	9±2	
PR95-18P1 ^b	contact zone	80.0		13.4	0.4	0.8			5.6		
PR95-19P1 ^b	1.5			34.0	9.0			57.0			
PR95-20P1 ^c	3.0	65.0	0.4	25.5	1.8	0.4			7.0		
PR95-21P1 ^d	contact zone		1.0	49.0	3.0	9.0			5.0		31.0
PR95-22P1 ^c	8.0			8.0	2.0	2.0			89.0		
PR95-23P1 ^d	35.0			2.0	1.0			28.0	69.0		
PR95-24P1 ^c	contact zone	80.0		15.2	4.2	0.6					
PR95-27P1 ^c	1.8	80.0		17.8	2.2						

Table 5-4

Sample #	PR95-1	PR95-2	PR95-3	PR95-4	PR95-5	PR95-11	PR95-12	PR95-13
SiO2	68.84	70.65	69.14	70.44	67.38	47.28	71.98	72.84
TiO2	0.256	0.281	0.259	0.254	0.254	2.508	0.268	0.277
Al2O3	12.72	13.75	13.13	13.09	12.96	16.84	13.38	13.51
Fe2O3	1.36	1.46	1.33	1.35	1.33	11.09	1.47	1.47
MnO	0.077	0.089	0.082	0.081	0.077	0.167	0.09	0.082
MgO	0.35	0.46	0.56	0.48	0.42	4.92	0.27	0.24
CaO	2.9	0.93	2.19	1.97	3.47	7.77	1.15	0.72
Na2O	3.17	3.38	3.11	3.1	2.94	3.57	3.48	3.83
K2O	5.17	5.41	5.33	5.52	5.66	1.65	6.24	5.61
P2O5	0.048	0.044	0.042	0.057	0.06	0.567	0.046	0.049
LOI %	5.25	3.26	4.42	3.85	4.93	3.44	1.18	0.65
Total	94.89	96.45	95.18	96.34	94.54	96.35	98.37	98.63
V	18.6	14.8	15.1	14.5	14.3	266.1	23	11.1
Cr						47		
Ni		5.9	6.1		6.6	48.1	11.1	
Zn	47.5	52.9	46.3	44	42.2	55.8	35.6	42.6
Pb	147.7	169	165	173.1	171.3	20	183	172.7
Sr	258.6	116.3	108.9	117.9	135.4	563.6	79.6	100.7
Y	32.5	28.4	27	31.1	35.8	34.1	32.8	37.5
Zr	265	316.8	262	271.1	261.3	249.6	284.5	297.6
Nb	26.2	27.6	28.1	24.5	27.3	36.1	31.2	30
Ba	712.4	564.2	487.4	556.2	511.5	371.5	500.1	585.3

Sample #	PR95-6	PR95-7	PR95-8	PR95-9	PR95-10	PR95-17	PR95-18	PR95-19
SiO2	70	68.38	67.93	71.06	73.15	46.78	72.6	79.38
TiO2	0.269	0.25	0.273	0.268	0.265	2.564	0.176	0.109
Al2O3	13.56	12.98	13.83	13.61	13.41	17.08	12.4	8.81
Fe2O3	1.43	1.34	1.49	1.42	1.44	11.59	1.14	0.79
MnO	0.086	0.083	0.087	0.089	0.087	0.171	0.116	0.075
MgO	0.61	0.84	0.54	0.47	0.25	5.19	0.29	0.31
CaO	0.92	2.46	1.67	1.02	0.89	7.93	0.56	1.1
Na2O	3.02	2.76	2.77	3.63	3.83	4.06	3.3	2.11
K2O	5.58	5.38	5.86	5.54	5.54	1.55	5.02	4.28
P2O5	0.046	0.063	0.057	0.043	0.043	0.581		
LOI %	3.35	4.75	4.53	2.57	0.77	1.62	3.43	2.1
Total	95.52	94.55	94.5	97.16	98.91	97.5	95.61	96.96
V	16.4	24.7	34.6	10.4	13.7	248.1		
Cr						32.4		
Ni					7.6	45.5	8.7	14.6
Zn	49.5	47	58.2	46	49.4	69.7	72	49.2
Pb	171.6	169.2	86.1	155.9	171.9	15.8	187.9	99.7
Sr	113.4	146	718.8	119	93.6	593.9	80.8	63.2
Y	35.6	32.1	31	32.5	30	34.3	44.8	28.4
Zr	282.6	272.2	315.9	288.1	278.3	250.6	235.3	162
Nb	30	28.8	25.7	26.9	28.6	36.8	26.1	21.8
Ba	483	444.7	689.3	527.2	532.3	330	85.5	77.5

Table 5-5

Sample #	PR95-14	PR95-15	PR95-22	PR95-23	PR95-24	PR95-25
SiO2	68.75	64.58	67.73	71.99	74.41	74.31
TiO2	0.259	0.224	0.212	0.118	0.089	0.091
Al2O3	12.71	11.29	12.53	10.47	12.18	12.23
Fe2O3	1.42	1.19	1.52	1.08	0.89	0.88
MnO	0.072	0.067	0.098	0.071	0.054	0.051
MgO	0.72	0.55	1.59	0.64	0.19	0.2
CaO	3.28	6.89	2.43	1.39	0.8	0.86
Na2O	3.1	2.57	1.56	1.62	3.25	2.93
K2O	6.02	5.96	1.98	3.88	4.73	4.59
P2O5	0.051	0.098	0.013	0.045	0.024	0.023
LOI %	2.94	6.28	9.34	8.07	2.97	2.94
Total	96.37	93.42	89.67	91.3	96.62	96.18
V	18.5	14	12.7			
Cr						
Ni		9.6			7.4	8.1
Zn	42.3	23	76.8	47.7	36.4	39.2
Rb	132.3	144	94.5	141.2	182.9	180.3
Sr	146.3	166.7	397	188.1	86.6	77.6
Y	29.1	30.2	27.3	29.2	24.8	25.6
Zr	265.5	236.9	232	135.9	82	78.2
Nb	23.8	23.7	24.2	21.2	19.1	14.3
Ba	538.3	501.7	551	548.8	282.1	291.3

Sample #	PR95-20	PR95-21	PR95-26	PR95-27	PR95-28
SiO2	70.4	67.04	74.22	74.28	45.35
TiO2	0.189	0.254	0.092	0.081	2.511
Al2O3	12.97	14.33	12.39	12.27	16.65
Fe2O3	1.33	2.13	0.89	0.82	11.11
MnO	0.108	0.067	0.052	0.05	0.154
MgO	0.72	2.38	0.34	0.16	4.23
CaO	1.11	3.05	0.76	0.76	9.25
Na2O	2.49	2.39	2.75	2.79	3.04
K2O	4.66	2.24	4.67	4.79	1.89
P2O5	0.035	0.063	0.018	0.025	0.562
LOI %	4.88	5.12	3.27	3.04	4.4
Total	94.02	93.95	96.18	96.03	94.75
V		19.5			246
Cr		10.8			38.9
Ni	6.9	7.9	7.8	7.9	49.1
Zn	60.7	38.8	28.3	24.2	71.5
Rb	179.7	38.6	164.3	181.2	30.3
Sr	187.8	497.3	86.1	85.9	566.5
Y	41.4	18.2	21.6	13.9	28.5
Zr	212.3	155.1	78.9	68.5	237.7
Nb	25.9	16.1	15.2	20.4	37.2
Ba	179.7	653.2			

Table 5-5 contd.

Sample	PR95-1	PR95-8	PR95-10	PR95-11	PR95-16	PR95-18	PR95-19	PR95-20	PR95-21	PR95-22
Na2O	3.15	2.73	3.92		3.84	3.39	2.16	2.55	2.40	1.51
K2O	5.20	5.50	5.90		1.70	5.40	4.50	4.90	2.40	2.00
CaO	2.70	1.57	0.67		8.70	0.55	1.14	0.96	3.07	2.17
Sc	2.44	2.49	2.42		22.80	1.80	1.21	2.47	4.45	2.53
Cr	0.00	0.00	0.00		50.10	0.00	0.00	2.40	11.00	3.60
FeO	1.19	1.26	1.27		10.17	0.99	0.70	1.16	1.84	1.31
Co	1.05	0.80	0.67		33.30	0.14	0.53	1.01	2.96	1.52
Ni	0.00	2.00	1.00		31.00	0.00	1.00	5.00	0.00	6.00
Zn	57.70	59.20	59.50		104.00	91.70	68.70	92.00	54.70	94.50
As	2.60	1.00	2.70		0.80	2.90	1.11	2.58	1.23	0.76
Br	1.29	0.00	0.42		0.10	0.80	0.55	0.36	0.03	0.13
Pb	143.50	85.50	174.90		23.00	184.10	102.70	179.00	38.70	90.80
Sr	270.00	728.00	99.00		590.00	88.00	67.00	192.00	488.00	380.00
Zr	295.00	302.00	289.00		240.00	238.00	162.00	215.00	148.00	225.00
Sb	0.38	0.18	0.38		0.07	0.27	0.15	0.28	0.21	0.19
Cs	3.30	1.13	2.61		0.40	4.71	2.18	6.21	1.08	4.26
Ba	701.00	717.00	551.00		368.00	77.00	104.00	182.00	655.00	534.00
La	68.70	67.90	66.70		35.40	41.20	25.80	37.60	41.20	30.20
Ce	125.60	127.80	128.40		71.40	90.90	55.60	76.30	82.90	66.30
Nd	41.00	42.30	42.00		34.00	32.40	21.10	28.80	28.30	26.10
Sm	7.18	7.36	7.48		6.90	8.57	5.53	6.72	5.18	6.22
Eu	0.83	0.83	0.79		2.09	0.42	0.28	0.50	0.86	0.39
Tb	0.81	0.84	0.94		0.98	1.25	0.85	0.99	0.58	0.88
Yb	3.09	2.99	3.39		2.93	3.89	3.20	3.35	2.03	3.13
Lu	0.47	0.45	0.49		0.42	0.56	0.45	0.48	0.30	0.46
Hf	8.21	7.95	8.14		5.49	8.55	5.99	7.47	4.41	7.84
Ta	1.64	1.72	1.85		2.19	1.65	1.13	1.55	1.17	1.48
W	1.00	0.60	0.60		0.30	1.80	0.50	1.30	0.60	0.50
Au	0.50	1.00	0.00		1.00	0.00	1.50	2.00	0.30	0.80
Th	22.40	24.10	25.40		3.92	25.30	18.02	23.70	18.01	21.50
U	4.01	2.62	5.16		0.85	5.03	2.85	4.25	1.58	2.57
B	21.00			14.00		30.00		20.00		
Br	1.00			1.00		1.00		1.00		
Cl	371.00			38.00		134.00		78.00		
F	457.00			831.00		641.00		758.00		
Li	16.00			8.00		8.00		13.00		
S	93.00			65.00		29.00		21.00		
P2O5	422.00			4921.00		93.00		348.00		

Table 5-6

Sample	GR95-1A	GR95-1B	GR95-2	GR95-3	GR95-4	GR95-5A	GR95-5B	GR95-6A
SiO ₂	72.44	71.23	75.53	74.67	48.34	73.01	70.78	72.83
TiO ₂	0.02	0.11	0.05	0.08	2.72	0.12	0.07	0.02
Al ₂ O ₃	13.08	13.86	13.19	12.88	16.33	13.33	12.43	13.00
Fe ₂ O ₃	0.82	1.32	0.95	1.19	12.21	1.45	1.10	0.77
MnO	0.12	0.11	0.12	0.12	0.17	0.09	0.08	0.12
MgO		0.63	0.16	0.42	5.05	0.28	0.33	
CaO	0.57	0.99	0.70	0.99	7.94	0.67	2.80	0.47
Na ₂ O	3.12	3.09	4.33	3.85	3.91	3.19	3.33	3.79
K ₂ O	6.02	4.36	4.28	4.20	1.44	4.94	4.81	5.08
P ₂ O ₅	0.03	0.11	0.02	0.03	0.85	0.04	0.04	
LOI	3.02	3.74	0.59	0.95	0.77	2.35	3.49	3.24
Total	96.23	95.82	99.33	98.44	98.96	97.11	95.77	96.08
V		16.43		7.74	176.40	14.41	8.40	
Cr					53.50			
Ni		8.93			51.23			
Zn	108.30	136.65	143.52	152.07	71.83	129.84	117.78	132.15
Rb	504.59	411.86	465.06	443.63	26.13	435.57	435.69	521.14
Sr	7.14	55.84	24.13	48.08	744.19	36.83	48.15	8.59
Y	73.62	53.36	71.24	67.86	33.52	79.20	80.25	67.20
Zr	102.59	128.56	115.54	120.70	232.57	151.88	112.96	99.64
Nb	175.17	154.89	162.32	154.46	38.93	169.69	176.76	171.46
Ba		113.38	72.48	95.10	430.17	142.04	88.00	
Total	96.34	95.95	99.46	98.57	99.19	97.25	95.90	96.20

Sample	GR95-6B	GR95-7	GR95-8A	GR95-8B	GR95-9	GR95-10	GR95-11	GR95-12
SiO ₂	71.72	70.47	72.40	73.56	48.64	71.53	71.90	74.50
TiO ₂	0.07	0.11	0.02	0.05	2.62	0.08	0.02	0.02
Al ₂ O ₃	13.33	12.87	13.07	12.95	15.77	12.50	12.84	13.42
Fe ₂ O ₃	1.07	1.22	0.77	0.93	12.51	1.09	0.79	0.84
MnO	0.12	0.11	0.13	0.14	0.18	0.12	0.12	0.13
MgO	0.29	0.35		0.16	5.16	0.44		
CaO	0.62	0.94	0.52	0.67	7.91	1.15	0.60	0.47
Na ₂ O	3.21	3.43	3.25	3.72	3.92	3.63	4.21	4.82
K ₂ O	4.73	3.81	5.99	4.88	1.60	3.73	4.36	4.36
P ₂ O ₅		0.02			0.86	0.02		
LOI	3.75	5.41	2.69	2.15	0.11	4.52	4.28	0.36
Total	95.17	93.31	96.15	97.05	99.17	94.28	94.84	98.56
V	8.11	9.90			196.11	8.89		
Cr					74.91			
Ni		6.47	6.42		56.36			6.03
Zn	142.98	125.69	28.22	99.05	108.66	130.71	109.97	119.59
Rb	463.92	442.51	509.51	470.91	31.42	432.44	500.05	524.81
Sr	39.56	55.83	7.74	39.47	697.36	75.25	5.99	9.27
Y	69.35	68.97	81.38	76.28	31.62	68.42	70.25	80.11
Zr	112.58	125.84	99.41	106.41	245.07	110.13	100.41	107.03
Nb	163.29	150.82	169.85	150.61	41.91	150.90	161.59	176.15
Ba	64.14	91.16			428.20	56.96		

Table 5-7

Sample	GR95-13	GR95-14	GR95-15A	GR95-15B	GR95-16
SiO2	46.14	72.17	59.62	70.48	47.00
TiO2	2.88	0.02	0.02	0.11	2.86
Al2O3	15.53	12.93	10.51	12.76	15.99
Fe2O3	12.89	0.78	0.64	1.25	12.92
MnO	0.17	0.12	0.09	0.11	0.17
MgO	5.96		0.10	0.35	5.68
CaO	8.24	0.47	10.05	0.95	8.17
Na2O	3.57	4.41	3.28	3.44	3.80
K2O	1.37	4.50	3.66	3.76	1.41
P2O5	0.65			0.02	0.72
LOI	1.37	3.16	11.05	5.59	0.14
Total	97.40	95.39	87.97	93.21	98.74
V	225.08			10.72	219.47
Cr	74.91				72.22
Ni	69.52			10.76	64.74
Zn	108.48	132.65	99.05	96.25	104.29
Rb	26.16	508.20	406.61	444.69	24.29
Sr	666.88		20.23	66.82	693.27
Y	31.53	76.56	73.13	70.40	32.01
Zr	221.68	99.44	72.70	122.31	238.19
Nb	38.28	171.13	128.08	146.30	40.43
Ba	425.18			79.83	405.62
Total	97.63	95.51	88.07	93.33	98.97

Sample	GR95-17	GR95-18	GR95-19	GR95-20	GR95-21
SiO2	68.58	72.55	69.60	67.55	73.79
TiO2	0.57	0.02	0.55	0.83	0.12
Al2O3	13.91	13.00	13.19	12.36	12.46
Fe2O3	3.66	0.79	3.34	4.32	1.34
MnO	0.09	0.12	0.09	0.10	0.08
MgO	1.28	0.08	1.32	2.14	0.34
CaO	1.77	0.63	1.60	2.48	0.95
Na2O	2.54	4.02	2.27	2.17	3.76
K2O	3.35	4.94	3.20	2.56	4.36
P2O5	0.17		0.16	0.16	0.02
LOI	3.39	3.18	3.95	4.18	1.71
Total	95.94	96.16	95.31	94.68	97.23
V	52.41		65.56	68.00	8.42
Cr	28.69		26.82	37.63	
Ni	8.81	6.09	14.64	21.91	7.88
Zn	83.83	113.95	81.33	87.84	114.58
Rb	204.07	509.51	235.74	186.10	390.43
Sr	285.85	8.84	198.99	239.35	64.75
Y	39.94	73.57	51.36	41.65	64.86
Zr	313.33	99.13	278.59	259.13	142.11
Nb	54.93	167.51	85.98	64.66	155.34
Ba	692.07		399.00	386.38	131.09

Table 5-7 contd.

Samples	GR95-6A	GR95-8a	GR95-6B	GR95-8B	GR95-7	GR95-15B	GR95-12	GR95-4	GR95-16
Na ₂ O	3.82	3.35	3.35	3.81	3.54	3.44	4.88	3.97	3.98
K ₂ O	5.20	6.00	4.80	5.00	3.80	3.80	4.30	1.10	1.10
CaO	0.46	0.58	0.66	0.68	0.83	1.08	0.53	7.90	8.60
Sc	2.13	2.17	2.85	2.39	3.16	3.13	2.25	16.39	18.66
Cr	0.00	0.00	3.20	2.10	5.40	5.00	0.00	53.40	76.40
FeO	0.67	0.68	0.95	0.82	1.08	1.06	0.70	10.90	11.81
Co	0.07	0.14	0.95	0.72	1.37	1.40	0.13	33.70	39.50
Ni	3.00	0.00	0.00	8.00	0.00	0.00	1.00	34.00	57.00
Zn	164.70	33.60	172.10	132.70	163.00	147.40	173.20	121.00	134.00
As	10.00	11.20	9.20	8.50	8.70	8.50	5.80	0.50	0.80
Br	0.26	0.24	0.45	0.38	0.36	0.23	0.12	0.00	0.40
Rb	510.00	508.00	466.00	469.00	440.00	434.00	522.00	27.00	26.00
Sr	7.00	9.00	41.00	54.00	58.00	82.00	6.00	750.00	710.00
Zr	70.00	103.00	102.00	78.00	109.00	106.00	109.00	199.00	210.00
Sb	1.28	1.42	1.30	1.47	1.31	1.31	0.92	0.05	0.07
Cs	13.92	14.29	14.71	14.64	15.09	14.79	11.85	0.22	0.35
Ba	3.00	0.00	63.00	48.00	91.00	77.00	15.00	439.00	444.00
La	6.49	6.74	9.31	7.93	10.64	10.46	7.00	33.30	32.70
Ce	19.70	19.90	25.70	22.10	27.50	26.90	21.40	70.70	69.90
Nd	9.90	10.60	12.60	11.60	12.60	14.00	11.70	39.00	38.00
Sm	3.89	3.99	4.19	4.01	4.27	4.21	4.13	8.85	8.53
Eu	0.07	0.08	0.17	0.14	0.24	0.23	0.08	2.83	2.77
Tb	0.93	0.89	0.88	0.90	0.88	0.92	0.97	1.17	1.14
Yb	7.72	7.86	7.32	7.34	6.97	6.89	7.99	2.30	2.27
Lu	1.19	1.22	1.14	1.13	1.08	1.06	1.24	0.32	0.31
Hf	8.30	8.57	8.48	8.27	7.91	7.82	8.89	5.18	5.49
Ta	26.60	27.00	25.20	24.30	22.20	22.40	27.70	2.53	2.49
W	3.10	2.90	2.70	2.30	3.30	3.10	3.20	0.50	0.50
Au	0.60	1.20	0.50	0.80	0.30	0.60	0.00	0.00	0.00
Th	22.60	23.00	22.70	21.60	21.20	20.90	25.20	3.11	3.11
U	7.93	7.98	7.14	7.43	6.86	6.78	7.93	1.16	0.92

Table 5-7 contd.

SAMPLE	SIO2	TIO2	AL2O3	FEO	CAO	MGO	MNO	K2O	NA2O	SRO	BAO	P2O5	F	TOTAL
gr95_1	74.00	0.04	13.00	0.13	0.24	0.00	0.06	6.44	2.73	0.02	0.00	0.01	0.26	96.81
gr95_1	74.23	0.00	12.97	0.11	0.30	0.00	0.10	6.02	2.93	0.01	0.03	0.06	0.23	96.88
gr95_1	73.87	0.01	12.97	0.20	0.27	0.00	0.06	6.11	2.97	0.06	0.01	0.00	0.19	96.63
gr95_1	74.76	0.00	12.81	0.20	0.36	0.00	0.07	5.38	3.06	0.04	0.00	0.00	0.23	96.79
gr95_1	73.95	0.05	13.03	0.15	0.16	0.00	0.07	6.12	2.96	0.01	0.00	0.00	0.21	96.62
gr95_1	72.98	0.03	12.87	1.02	0.22	0.00	0.18	6.30	2.91	0.01	0.00	0.00	0.22	96.64
gr95_1	74.09	0.03	13.02	0.18	0.25	0.00	0.08	6.05	2.91	0.06	0.00	0.02	0.22	96.81
gr95_1	74.35	0.04	13.03	0.21	0.36	0.02	0.11	5.91	2.93	0.05	0.03	0.00	0.20	97.15
gr95_1	73.55	0.02	13.23	0.17	0.37	0.00	0.09	5.88	3.21	0.07	0.00	0.00	0.20	96.69
gr95_1	73.03	0.00	12.91	0.18	0.20	0.00	0.03	6.22	2.97	0.09	0.04	0.03	0.24	95.85
gr95_1	73.54	0.08	12.99	0.16	0.15	0.00	0.07	6.12	2.92	0.05	0.00	0.00	0.23	96.21
gr95_1	73.70	0.07	13.04	0.24	0.25	0.00	0.06	6.68	2.68	0.02	0.04	0.02	0.19	96.90
gr95_1	72.92	0.01	12.96	0.20	0.21	0.00	0.15	5.66	2.90	0.11	0.00	0.01	0.22	95.25
gr95_1	74.55	0.00	12.99	0.14	0.21	0.00	0.05	5.79	2.85	0.04	0.03	0.00	0.21	96.76
gr95_1	73.52	0.02	12.92	0.27	0.19	0.00	0.10	5.95	2.80	0.02	0.05	0.03	0.19	95.98
gr95_1	73.48	0.00	13.02	0.37	0.23	0.00	0.06	6.48	2.49	0.03	0.00	0.01	0.22	96.30
gr95_1	74.70	0.01	12.82	0.14	0.18	0.01	0.14	6.29	2.62	0.06	0.02	0.02	0.21	97.11
gr95_1	73.85	0.00	12.92	0.13	0.13	0.00	0.09	6.34	2.65	0.03	0.00	0.01	0.22	96.29
gr95_1	74.65	0.00	12.84	0.15	0.13	0.00	0.02	6.11	2.69	0.00	0.00	0.01	0.17	96.69
gr95_1	77.37	0.00	11.99	0.32	0.34	0.00	0.14	4.33	2.79	0.06	0.01	0.01	0.21	97.47
gr95-6b	71.88	0.03	12.82	0.74	0.44	0.00	0.14	4.58	3.97	0.01	0.00	0.00	0.24	94.73
gr95-6b	72.42	0.01	12.97	0.65	0.44	0.00	0.14	4.48	3.99	0.04	0.07	0.04	0.23	95.39
gr95-6b	72.26	0.08	12.68	0.64	0.40	0.00	0.11	4.67	3.95	0.00	0.06	0.00	0.24	94.98
gr95-6b	72.60	0.10	12.88	0.74	0.49	0.00	0.12	4.32	4.17	0.04	0.01	0.00	0.21	95.57
gr95-6b	72.63	0.02	12.88	0.64	0.45	0.01	0.13	4.40	4.05	0.00	0.00	0.02	0.24	95.35
gr95-6b	71.61	0.06	12.48	0.60	0.41	0.00	0.10	4.65	3.77	0.00	0.00	0.00	0.21	93.81
gr95-6b	69.54	0.09	12.27	0.65	0.37	0.00	0.10	4.97	3.36	0.05	0.00	0.01	0.22	91.54
gr95-6b	72.68	0.00	12.74	0.67	0.44	0.00	0.09	4.37	4.04	0.06	0.00	0.00	0.24	95.23
gr95-6b	72.83	0.12	12.66	0.77	0.44	0.00	0.11	4.79	3.76	0.00	0.00	0.00	0.20	95.58
gr95-6b	72.09	0.01	12.81	0.72	0.45	0.00	0.12	4.47	4.03	0.03	0.00	0.00	0.24	94.85
gr95-6b	72.93	0.08	12.72	0.60	0.44	0.00	0.16	4.60	3.81	0.00	0.05	0.01	0.22	95.53

Table 5-8

SAMPLE	SIO2	TIO2	AL2O3	FEO	CAO	MGO	MNO	K2O	NA2O	SRO	BAO	P2O5	F	TOTAL
gr95-6b	72.17	0.05	12.74	0.74	0.45	0.00	0.11	4.37	3.99	0.05	0.01	0.02	0.27	94.87
gr95-6b	72.45	0.04	12.84	0.68	0.46	0.01	0.09	4.37	4.06	0.00	0.04	0.01	0.26	95.20
gr95-6b	72.10	0.00	12.62	0.64	0.48	0.01	0.09	4.41	3.92	0.05	0.02	0.00	0.22	94.46
gr95-6b	72.20	0.09	12.85	0.66	0.44	0.00	0.05	4.43	4.09	0.03	0.04	0.04	0.24	95.05
gr95-6b	71.73	0.01	12.70	0.68	0.46	0.00	0.06	4.36	4.10	0.00	0.03	0.00	0.27	94.28
gr95-6b	73.00	0.03	12.69	0.71	0.47	0.00	0.13	4.47	3.99	0.03	0.00	0.00	0.22	95.64
gr95-6b	72.42	0.00	12.67	0.69	0.46	0.00	0.09	4.53	3.94	0.04	0.09	0.00	0.25	95.05
gr95-6b	69.31	0.00	12.38	0.63	0.44	0.00	0.07	4.22	4.01	0.01	0.01	0.00	0.24	91.22
gr95-6b	72.37	0.04	12.92	0.75	0.46	0.00	0.15	4.32	4.05	0.00	0.02	0.03	0.25	95.26
gr95-6b	72.53	0.00	12.75	0.76	0.45	0.01	0.06	4.34	3.92	0.00	0.00	0.00	0.26	94.97
gr95-6b	71.28	0.01	12.65	0.71	0.43	0.00	0.10	4.54	4.00	0.02	0.00	0.00	0.25	93.89
GR95-7	71.58	0.01	12.61	0.63	0.45	0.00	0.10	4.41	3.86	0.01	0.04	0.01		93.70
GR95-7	71.14	0.00	12.68	0.62	0.43	0.00	0.18	4.52	3.90	0.01	0.07	0.00		93.54
GR95-7	71.69	0.02	12.59	0.64	0.44	0.00	0.11	4.31	3.91	0.01	0.00	0.00		93.73
GR95-7	71.23	0.03	12.71	0.75	0.47	0.00	0.11	4.41	3.86	0.00	0.00	0.05		93.61
GR95-7	70.76	0.00	12.69	0.62	0.46	0.00	0.14	4.31	3.81	0.01	0.02	0.00		92.82
GR95-7	72.24	0.00	12.72	0.62	0.44	0.00	0.15	4.31	3.84	0.01	0.00	0.01		94.34
GR95-7	70.85	0.02	12.71	0.59	0.44	0.00	0.06	4.27	4.11	0.05	0.00	0.01		93.11
GR95-7	70.36	0.00	12.66	0.58	0.47	0.00	0.19	4.25	3.94	0.00	0.00	0.00		92.44
GR95-7	72.70	0.00	12.78	0.84	0.40	0.01	0.09	4.45	3.98	0.00	0.00	0.00		95.25
GR95-7	70.41	0.01	12.60	0.67	0.47	0.00	0.04	4.18	3.82	0.01	0.00	0.00		92.21
GR95-7	72.09	0.06	12.63	0.68	0.44	0.00	0.11	4.11	3.92	0.01	0.04	0.02		94.10
GR95-7	70.50	0.02	12.67	0.70	0.45	0.00	0.09	4.24	3.80	0.03	0.02	0.00		92.50
GR95-7	71.60	0.07	12.54	0.66	0.45	0.00	0.11	4.18	3.59	0.04	0.00	0.01		93.24
GR95-7	69.47	0.08	12.83	0.69	0.92	0.00	0.04	3.64	3.70	0.03	0.06	0.00		91.47
GR95-7	71.84	0.02	12.87	0.77	0.21	0.00	0.08	4.29	4.28	0.02	0.05	0.03		94.45
GR95-7	71.55	0.10	12.71	0.81	0.40	0.00	0.07	4.47	4.18	0.04	0.00	0.00		94.32
GR95-7	71.07	0.05	12.73	0.57	0.45	0.00	0.12	4.21	3.62	0.01	0.00	0.02		92.85
GR95-7	71.03	0.08	12.54	0.72	0.45	0.00	0.12	4.11	3.46	0.02	0.00	0.00		92.53
GR95-7	70.90	0.02	12.68	0.68	0.42	0.00	0.09	4.11	3.59	0.01	0.01	0.02		92.52
GR95-7	71.29	0.00	12.61	0.65	0.42	0.00	0.11	3.86	3.47	0.03	0.05	0.03		92.51

Table 5-8 coontd.

SAMPLE	SIO2	TIO2	AL2O3	FEO	CAO	MGO	MNO	K2O	NA2O	SRO	BAO	P2O5	F	TOTAL
GR95-7	71.69	0.00	12.75	0.70	0.42	0.00	0.07	4.31	4.16	0.05	0.04	0.01		94.20
GR95-7	69.63	0.03	12.62	0.62	0.43	0.01	0.06	3.13	3.46	0.00	0.00	0.01		89.98
GR95_8	73.21	0.04	13.02	0.17	0.47	0.00	0.10	4.98	3.68	0.02	0.01	0.01	0.15	95.80
GR95_8	72.29	0.04	12.83	0.14	0.46	0.00	0.05	4.98	3.62	0.00	0.02	0.00	0.15	94.51
GR95_8	71.97	0.04	12.73	0.85	0.49	0.00	0.16	5.49	3.34	0.05	0.00	0.05	0.15	95.24
GR95_8	72.46	0.01	12.99	0.10	0.41	0.00	0.10	4.98	3.74	0.05	0.05	0.01	0.14	94.97
GR95_8	73.11	0.07	12.93	0.11	0.47	0.00	0.13	4.95	3.64	0.00	0.01	0.00	0.14	95.51
GR95_8	68.40	0.00	12.09	0.12	0.46	0.00	0.06	5.25	3.27	0.00	0.01	0.00	0.15	89.74
GR95_8	74.39	0.00	13.00	0.12	0.45	0.00	0.07	4.87	3.80	0.02	0.00	0.01	0.15	96.82
GR95_8	73.70	0.10	13.02	0.21	0.48	0.00	0.11	4.71	3.80	0.00	0.08	0.00	0.15	96.30
GR95_8	73.59	0.04	13.10	0.15	0.31	0.00	0.14	5.29	3.75	0.05	0.00	0.04	0.14	96.54
GR95_8	73.49	0.06	13.05	0.12	0.45	0.00	0.11	4.75	3.72	0.05	0.00	0.00	0.15	95.89
GR95_8	74.19	0.10	13.02	0.09	0.47	0.01	0.08	5.42	3.54	0.05	0.06	0.01	0.19	97.13
GR95_8	72.51	0.00	12.71	0.16	0.50	0.00	0.10	5.05	3.51	0.07	0.02	0.02	0.19	94.76
GR95_8	72.91	0.04	12.93	0.09	0.48	0.00	0.08	4.97	3.69	0.02	0.00	0.00	0.12	95.28
GR95_8	71.89	0.03	12.84	0.18	0.47	0.00	0.08	6.32	2.77	0.01	0.03	0.02	0.16	94.73
GR95_8	73.10	0.00	12.83	0.11	0.47	0.00	0.10	5.20	3.50	0.00	0.00	0.00	0.16	95.40
GR95_8	73.05	0.03	12.84	0.10	0.47	0.00	0.11	4.73	3.87	0.03	0.02	0.00	0.16	95.34
GR95_8	72.68	0.02	12.97	0.14	0.45	0.00	0.07	6.26	3.09	0.01	0.04	0.03	0.18	95.85
GR95_8	73.39	0.05	12.91	0.24	0.45	0.00	0.16	5.98	3.18	0.05	0.04	0.05	0.17	96.61
GR95_8	72.34	0.00	12.89	0.09	0.43	0.00	0.09	4.88	3.72	0.00	0.02	0.01	0.20	94.59
GR95-15	72.05	0.01	12.86	0.69	0.45	0.00	0.05	4.38	3.96	0.01	0.06	0.00	0.45	94.78
GR95-15	72.38	0.04	12.63	0.62	0.44	0.00	0.13	4.36	3.97	0.07	0.02	0.00	0.42	94.91
GR95-15	72.07	0.01	12.76	0.70	0.42	0.00	0.13	4.58	3.65	0.01	0.09	0.03	0.43	94.70
GR95-15	72.23	0.00	12.73	0.72	0.45	0.00	0.13	4.20	3.93	0.00	0.00	0.00	0.47	94.67
GR95-15	72.51	0.06	12.70	0.61	0.43	0.01	0.10	4.24	3.97	0.00	0.00	0.01	0.47	94.91
GR95-15	72.01	0.06	12.64	0.66	0.44	0.00	0.12	4.62	3.77	0.04	0.02	0.00	0.47	94.65
GR95-15	70.99	0.00	12.48	0.71	0.42	0.00	0.13	4.49	3.55	0.01	0.01	0.01	0.43	93.05
GR95-15	72.07	0.05	12.67	0.69	0.45	0.00	0.08	4.60	3.81	0.01	0.03	0.00	0.48	94.74
GR95-15	70.95	0.00	12.47	0.70	0.46	0.00	0.11	4.37	3.87	0.00	0.00	0.00	0.44	93.17
GR95-15	72.06	0.06	12.78	0.67	0.46	0.00	0.09	4.50	3.87	0.00	0.00	0.02	0.45	94.76

Table 5--8 contd.

SAMPLE	SIO2	TIO2	AL2O3	FEO	CAO	MGO	MNO	K2O	NA2O	SRO	BAO	P2O5	F	TOTAL
GR95-15	72.36	0.07	12.78	0.71	0.42	0.00	0.13	4.39	4.08	0.06	0.02	0.03	0.49	95.34
GR95-15	72.46	0.00	12.76	0.71	0.47	0.00	0.08	4.45	3.87	0.01	0.03	0.01	0.49	95.13
GR95-15	72.12	0.05	12.65	0.68	0.44	0.00	0.12	4.30	4.02	0.02	0.01	0.03	0.46	94.69
GR95-15	71.72	0.00	12.66	0.65	0.43	0.00	0.13	4.35	3.83	0.00	0.05	0.00	0.46	94.07
GR95-15	71.72	0.05	12.71	0.70	0.46	0.00	0.08	4.35	3.95	0.06	0.00	0.05	0.47	94.40
GR95-15	71.20	0.04	12.38	0.65	0.41	0.00	0.11	4.12	3.71	0.00	0.03	0.01	0.43	92.92
GR95-15	72.62	0.09	12.82	0.66	0.45	0.00	0.05	4.55	3.96	0.03	0.03	0.00	0.47	95.52
GR95-15	72.31	0.04	12.76	0.59	0.44	0.01	0.16	4.24	4.01	0.04	0.00	0.04	0.46	94.90
GR95-15	71.47	0.01	12.70	0.60	0.46	0.00	0.07	4.48	3.99	0.03	0.00	0.00	0.48	94.09
GR95-15	71.78	0.07	12.59	0.69	0.45	0.00	0.13	4.31	3.86	0.02	0.04	0.00	0.46	94.21
GR95-15	71.70	0.02	12.79	0.69	0.48	0.00	0.17	4.59	3.88	0.05	0.00	0.01	0.45	94.64

Table 5--8 contd.

Samples	GR95-7	GR95-8a	GR95-8B	GR95-11
B	45.50	54.50	42.00	53.00
Br	0.75	0.75	1.30	0.75
Cl	307.50	125.00	619.50	239.50
F	3244.50	1736.00	1236.00	3616.00
Li	76.00	48.00	51.00	56.00
P2O5	174.50	130.00	146.00	48.50
S	101.00	78.00	183.50	59.00

Table 5-9

Table 5-10. Physical Properties of Air and Water at 293K and 373K

	<u>Air</u>		<u>Water</u>		
	293K	373K	293K	373K	
ρ	1.205	0.946	998.2	958.4	kg/m ³
α	3e-4	2.7e-3	2.1e-4	7.1e-4	K ⁻¹
κ	2.5e-2	3.2e-2	0.6	0.68	W/mK
c_p	1006	1011	4182	4216	J/kgK
μ	1.81e-5	2.18e-5	1.0e-3	2.83e-4	Pa s

Values from appendix C, Bejan (1993).

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Figure 12

Figure 13

Figure 14

Figure 16a

Figure 16b

Figure 17

Figure 18

Geologic Map of Paiute Ridge - Slanted Buttes, Half Pint Range (Nevada)

Fraser Goff, Giday WoldeGabriel, Richard Warren, and Karen Carter Krogh

Projection - UTM (Universal Transverse Mercator)
Units of Measure - Meters

Figure 19

Figure 20

Figure 21

Figure 22

Figure 23

Figure 24

Figure 25

Figure 26

Figure 27a

Figure 27b

Figure 28

Figure 29

Figure 30

Figure 31

Figure 32

Figure 33

Figure 34

Figure 35

Figure 36

Figure 37

Figure 38

Figure 39

Figure 40

Figure 41

Figure 42

Figure 43

Figure 44

Figure 45

Figure 46

Figure 47

Figure 48