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GEOPHYSICAL SETTING OF THE PAHUTE MESA–OASIS VALLEY REGION SOUTHERN NEVADA

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ABSTRACT

The Pahute Mesa–Oasis Valley region includes the northwestern part of the Nevada Test Site and a contiguous area, of roughly equal size, west of the Test Site. This region is located approximately 150 km northwest of Las Vegas, Nevada, and extends from the southernmost of the springs of Oasis Valley, directly south of Beatty, Nevada, northeastward to Pahute Mesa, an up-gradient (up the groundwater flow path) recharge area that was the site of 85 underground nuclear tests between 1966 and 1991. The latest generation of geophysical studies of this region, reported herein, was part of a multidisciplinary effort to develop better hydrogeologic models for forecasting potential impacts of radionuclide leakage from Pahute Mesa toward the major groundwater discharge areas in Oasis Valley.

Gravity, aeromagnetic, and magnetotelluric surveys of the Pahute Mesa–Oasis Valley region show that much of this region is underlain by a complex of deep basins. Based on surface and drill-hole geologic data, it has long been known that the Pahute Mesa–Oasis Valley region is partly coincident with the central caldera complex of the southwestern Nevada volcanic field. Our geophysical modeling, in combination with surface geologic control, show that most of the basin-bounding structures in this region have subsurface geometries that resemble those of other well-documented caldera-collapse ring-fracture zones. Moreover, these data form the basis for delineation of the detailed geometry of the western and northern parts of the caldera complex, areas that have frequently been a topic of geologic controversy because caldera structures in these areas are almost totally concealed under younger volcanic rock and sediment. A small minority of the geophysical features in the study area represent post-caldera tectonic structures including, most prominently, the north-trending Oasis Valley (half-graben) basin, which partly overlaps the westernmost part of the central caldera complex, and has a coincident southern boundary.

Cenozoic basin-fill deposits of the Pahute Mesa–Oasis Valley region are significantly less dense than the underlying pre-Tertiary rocks. Based on this relation, a three-dimensional inversion of the gravity data was developed that reveals a topographically complex surface at the top of the basement rocks. The major feature of this surface in the study area is the >35-km long, northeast trending, down-to-the-southeast Thirsty Canyon fault zone, the probable western boundary of the central caldera complex. West of this boundary, the top of the basement is shallow and relatively flat, and the estimated density of the basement suggests it consists mostly of siliciclastic sedimentary rocks (quartzite and argillite), which is supported by the one small exposure of basement rocks that is present in this area. These siliciclastic rocks typically are much less permeable than the higher-specific-gravity carbonate basement rocks that are exposed to the southeast of the study area. East of the Thirsty Canyon fault zone, the top of the basement is a deep, high-relief surface that is locally overlain by as much as 5 km of low-density Cenozoic strata, mostly volcanic caldera-fill deposits. Some of the high-relief basement features within

the caldera complex represent boundaries between the different caldera-collapse structures that comprise the caldera complex; others defy easy explanation and may reflect intracaldera intrusions, deeply buried pre-caldera tectonic structures, or other features of unknown origin. Detailed geophysical surveys across the Thirsty Canyon fault zone show that it separates more dense volcanic rocks, to the west, from less dense intracaldera deposits within the caldera complex, to the east. Parts of this fault zone consist of a shallow, ~2-km-wide zone of coalescing ring fracture systems exhibiting step-like differential displacements. The length and linearity of the Thirsty Canyon fault zone is atypical of caldera margin faults, indicating that it may have been localized along a preexisting tectonic fault zone.

The Thirsty Canyon fault zone juxtaposes rock types of differing permeabilities and, thus, may act as a barrier to groundwater and deflect flow along it from Pahute Mesa toward Oasis Valley. Within the Thirsty Canyon fault zone, highly fractured rocks may serve as conduits, depending upon the degree of alteration and its effect on permeability. Other major structures in the Pahute Mesa–Oasis Valley region that may influence groundwater flow include the master fault of the Oasis Valley half-graben and the southern bounding fault of the central caldera complex.

INTRODUCTION

Four areas on the Nevada Test Site (NTS), in southern Nevada (fig. 1), comprised the primary site for testing nuclear devices within the United States from the early 1950s until testing ended in 1992. One of these four testing areas is Pahute Mesa, a high plateau in the northwestern part of the NTS that was used primarily for deep tests of relatively large-yield devices. In total, 85 underground tests were conducted on Pahute Mesa between 1966 and 1991 (U.S. Department of Energy, 1994), 67 of which were detonated at or below the water table (within ~50 m), which lies at depths between ~400 and ~700 m under the mesa. Of the remaining 18 tests, 13 were close enough to the water table that the explosion cavities probably penetrated into it (Laczniak and others, 1996). Concerns about, and evidence of, leakage of radionuclides from the Pahute Mesa tests by the movement of groundwater (Laczniak and others, 1996; Russell and Locke, 1996; Kersting and others, 1999) led to the funding of a multidisciplinary effort to address the hydrogeologic framework of the region. The area covered by this effort extends from Pahute Mesa along the presumed groundwater flow path toward the closest major discharge areas in Oasis Valley, 25 and 45 km to the southwest. This report documents the results of the geophysical studies of the Pahute Mesa–Oasis Valley region, and discusses the implications of these results for the hydrogeology of the study area. Other studies in this multidisciplinary effort include hydrology (Reiner and others, 2002), geology (Fridrich and others, 1999a, 1999b) and geologic modeling (McKee and others, 1999, 2001; Sweetkind and others, 2001).

The Pahute Mesa–Oasis Valley region is part of the Death Valley groundwater flow system (Winograd and Thordarson, 1975; Laczniak and others, 1996). Whereas most of the groundwater from Pahute Mesa is thought to discharge in Oasis Valley, some component of the flow presumably continues on to more distant discharge sites in the Amargosa Desert and, ultimately, Death Valley (fig. 1).

Geophysical and subsurface geologic studies are critical to understanding the hydrogeology of this region because the major structural and stratigraphic features that control the geometries of aquifers and confining units of the Pahute Mesa–Oasis Valley region are largely concealed at the surface by younger volcanic and sedimentary units. Moreover, the water table is locally deeper than 700 m, and major water-producing units have been documented to depths of at least 2 km in parts of the region.

This study builds on the results of six previous geophysical studies of this region: (1) Healey (1968) conducted a pioneering study of the subsurface structure of the NTS using gravity data. (2 and 3) McCafferty and Grauch (1997) and Ponce and others (1999) published an aeromagnetic map and updated gravity maps of the NTS region, respectively. (4) Ferguson and others (1994) used drill-hole lithologic logs, along with gravity and seismic data, to construct interpretive cross sections across the eastern part of Pahute Mesa. (5) Grauch and others (1997, 1999) integrated regional geophysical and geological data throughout the NTS region. (6) McKee and others (1999) produced a three-dimensional interpretation of that part of the central caldera complex underlying Pahute Mesa, using gravity inversion data and drill-hole stratigraphy.

Our study expands on the above studies by more closely defining the nature of the basin boundaries and other prominent regional features (anomalies) that lie between the testing area on Pahute Mesa and the Oasis Valley discharge areas. A compilation of hydrogeologic models used to evaluate the geologic framework controlling groundwater flow in the western NTS was prepared by Bechtel-Nevada (2002).

Our investigation consisted of several parts: (1) 487 new gravity stations (Mankinen and others, 1998) were established in the Pahute Mesa–Oasis Valley area to improve regional coverage and to cross normal to the boundaries of major structures in both areas. (2) An analysis of the gravity data was conducted to provide a three-dimensional view of the basement surface, defined as the contact between pre-Tertiary rocks and younger formations. (3) Potential-field modeling of gravity and aeromagnetic data, constrained in part by basin thickness (Hildenbrand and others, 1999), was performed to develop approximations of subsurface structure. (4) A magnetotelluric (MT) survey was conducted along some of the detailed gravity traverses of Mankinen and others (1998, 1999) to provide insights on structures within the top 2 km of the crust. (5) Synthesis of the geophysical information and existing and new geologic data (Fridrich and others, 1999a, 1999b) lead to tectonic models of the Pahute Mesa–Oasis Valley region. Geologic cross sections generated by Byers and Cummings (1966), O'Conner and others,

(1966), Christiansen and Noble (1968), Orkild and others (1969), Byers and others (1976a, b), Scott and Bonk (1984), Frizzell and Shulters (1990), and Minor and others (1997) were helpful in this effort. The geologic and hydrologic cross sections of Laczniak and others (1996) were also valuable. Results of this investigation provide constraints and enhance development of conceptual geologic and groundwater flow models needed in concerted monitoring efforts at the NTS.

GEOLOGIC SUMMARY

What is now southern Nevada (fig. 1) was near the western edge of the North American craton during the Late Proterozoic and the early and middle Paleozoic. (The area shown in figure 1 will be herein referred to as the study area, whereas the term "region" will be used refer to the whole southern Great Basin). Major rock types from this time consist of Late Proterozoic to Lower Cambrian terrigenous siliciclastic deposits, overlain by lower and middle Paleozoic marine carbonate and quartzose rocks. Deposition rates in southwestern Nevada were very high in this interval because this region occupied the continental shelf. In the latter part of the Paleozoic, a new phase of tectonic activity shifted the focus of deposition to the northwest of the study area, resulting in generally reduced sedimentation rates in southwestern Nevada in the late Paleozoic. Sedimentary rocks deposited in this period are a mixture of quartzose and carbonate rocks. The Late Proterozoic and Paleozoic sequence of the NTS region, as a whole, is divisible into two major carbonate aquifers, which are separated from each other, and underlain, by two silicious confining units (Winograd and Thordarson, 1975; Laczniak and others, 1996).

Throughout the NTS region, the Mesozoic and early Cenozoic were times of mountain-building and compressional tectonism, followed by erosion. In late Eocene time, the tectonic regime changed to one of slow protracted extension, roughly synchronous with the onset of volcanism in the northernmost part of the Great Basin (at ~43 Ma; McKee and others, 1970). This earliest phase of extensional tectonism was "pre-basin-range," in that it formed broad shallow basins in the NTS region, rather than basin-range topography (Fridrich and others, 2000). From the late Eocene to the present, the foci of basin-range tectonism and of volcanism have migrated across the area of the Great Basin, generally from north to south, and from the center out toward the east and west margins, as this province-scale physiographic feature formed (McKee and others, 1970). Within this province-scale pattern, local volcanism migrated into the NTS region from the north at ~19 Ma, and the local peak period for both basin-range tectonism and volcanism was between ~14 and ~11 Ma. The focus of tectonism migrated out of the NTS region, to the south and west, between ~11 and ~7 Ma. The rate of volcanic activity greatly diminished with the local termination of significant tectonism; however, local volcanism continued, at increasing feeble rates, into the late Quaternary.

Most of the volcanic units exposed in the study area are part of the southwestern Nevada volcanic field (SNVF; Byers and others, 1976a, b; Christiansen and others, 1977; Sawyer and others, 1994), which was active dominantly between ~15 and ~7 Ma. A generalized stratigraphic section of the volcanic sequence in the study area is shown in figure 2. Whereas the specific source areas for the oldest large-volume eruptive units of this volcanic field are concealed under younger deposits, many of the largest volume pyroclastic units erupted after 14 Ma resulted in the formation of several large calderas, most of which are part of a large complex in the central part of the volcanic field. This central caldera complex has been divided into three major features, each of which is the product of multiple caldera-forming voluminous pyroclastic eruptions. These are: (1) the Silent Canyon caldera complex, source of the 13.7-Ma Grouse Canyon Tuff, the 13.25-Ma Bullfrog Tuff, and perhaps other units, on the north side; (2) the Claim Canyon caldera complex, source of the major tuffs of the 12.8-to-12.7-Ma Paintbrush Group, on the south side; and (3) the Timber Mountain caldera complex, source of the major units of the 11.7-to-11.3-Ma Timber Mountain Group, in the center. In addition, two younger calderas are located to the northwest of the central caldera complex: (1) the Black Mountain caldera, source of the tuffs of the ~9.4-to-~9.2-Ma Thirsty Canyon Group, and (2) the Stonewall caldera, source of the members of the ~7.5-Ma Spearhead Tuff (Weiss and Noble, 1989; Hausback and others, 1990; Sawyer and others, 1994; Conners and others, 1998; R.J. Fleck, M.A. Lanphere, D.A. Sawyer, and R.G. Warren, written commun., 2000). The Black Mountain caldera lies at the northwestern limit of the study area and the Stonewall caldera lies ~20 km to the northwest. Of all of these calderas, the Silent Canyon caldera complex alone is totally concealed under younger volcanic rocks, and is known only from geophysical and drill-hole data (Healey, 1968; Noble and others, 1968; Orkild and others, 1968, 1969; Simpson and others, 1986; Sawyer and Sargent, 1989; Minor and others, 1993; Sawyer and others, 1994, 1995; McKee and others, 1999, 2001).

FIELD METHODS AND DATA REDUCTION

Gravity Data

Gravity data were obtained using a LaCoste and Romberg meter, and observed gravity values (Mankinen and others, 1998) were referenced to the International Gravity Standardization Net 1971 (ISGN 71) gravity datum (Morelli, 1974). Bouguer anomaly values determined are accurate to within 0.1 mGal. An isostatic residual anomaly was calculated using a reduction density of 2400 kg/m^3, a crustal thickness of 25 km, and a mantle/crust density contrast of 400 kg/m^3 (see Simpson and others, 1986). A reduction density of 2400 kg/m^3 was chosen because it provides a reasonable estimate of the density average between 2670 kg/m^3 basement and the much lower densities of volcanic sequences, especially caldera fill.

All data were gridded at a spacing of 1 km using a minimum curvature algorithm of Webring (1981). Data from Ponce (1997) and Mankinen and others (1998) were used to produce the revised isostatic residual gravity map shown in figure 3a. A regional isostatic residual gravity map using these data (at a reduction density of 2670 kg/m^3) is provided by Ponce and others (1999).

Aeromagnetic Data

Aeromagnetic maps of the study area are available from McCafferty and Grauch (1997) and Ponce (1999). Because the aeromagnetic data were obtained from many sources of variable quality and type, McCafferty and Grauch (1997) produced a merged aeromagnetic-anomaly map that is a representation of the data as if all surveys were flown at a constant terrain clearance of 122 m. Some of the older aeromagnetic surveys were flown at a flight-line spacing of 1600 m, whereas more recent, detailed surveys were flown at 800 m or less (McCafferty and Grauch, 1997). These data were upward continued 100 m and then subtracted from the original data to produce a residual field. The magnetic anomaly data then were reduced to the magnetic pole, a technique used to center anomalies over their sources. Prominent magnetic lineaments are shown on the resulting residual, reduced-to-pole magnetic field map in figure 3b. Because a body with uniform magnetization and density has a magnetic field proportional to the derivative of the gravity field, a gravitational attraction can be derived from the magnetic data (the pseudogravity transform). This magnetic potential field, or pseudogravity, (fig. 4) removes the effects of near-surface magnetic sources to enhance regional sources. The reader is referred to Blakely (1995) for additional discussions on the utility of reduction-to-pole, pseudogravity, and continuation filters.

Magnetotelluric Data

The magnetotelluric method is an electromagnetic method that uses measurements of the natural electric (E) and magnetic (H) time varying fields on the Earth's surface to infer the subsurface electrical resistivity structure. MT sites were established at 95 locations on Pahute Mesa and in Oasis Valley (Schenkel, 1998; Schenkel and others, 1999). Of these sites, 35 were continuous profiling stations and 60 were individual soundings. The sounding sites were collected in pairs (500-m separation) in order to employ remote referencing for noise reduction (Gamble and others, 1979; Furgerson, 1982). Orthogonal components of the electric and magnetic fields were collected in the frequency band of 0.01 to 175 Hz. Continuous profiling, which combines several contiguous *Ex* fields for each setup of *Ey*, *Hx*, and *Hy* along a profile (Torres-Verdin, 1991; Torres-Verdin and Bostick, 1992), used a fixed remote reference and spanned 700 m. Electrode spacing was at either 50 or 100 m.

REGIONAL GEOPHYSICAL FEATURES

Gravity and aeromagnetic data, along with their derivative maps, enable us to define geophysical anomalies in the study area. These anomalies include areas of exceptionally high or low potential, which reflect bodies of rock that differ in their properties from the surrounding rock, as well as lineaments (figs. 3a and 3b), which typically reflect discrete structures such as faults. Often the gravity and aeromagnetic lineaments are coincident, or nearly so. Moreover, the major lineaments commonly form parts of the boundaries of the major areas of high or low potential.

One of the most prominent anomalies in the study area is the Thirsty Canyon lineament (fig. 3a) expressed by a northeast-trending linear gravity anomaly that is at least 35-km long, and which also appears, in part, on the aeromagnetic map (fig. 3b). The geologic context of this feature, in combination with data derived from detailed geophysical surveys across it, indicate that this lineament represents the western structural boundaries of both the Silent Canyon and Timber Mountain caldera complexes. The cross sectional geometry of the Thirsty Canyon lineament (TCL), as imaged on geophysical profiles, strongly supports geologic evidence indicating that this feature is a composite caldera-collapse ring-fracture zone, as discussed below. Nevertheless, the pronounced linearity and length of this feature is atypical of caldera-collapse faults, suggesting it may be localized along a preexisting tectonic fault zone that influenced caldera development.

A very large gravity low is centered over the central and eastern part of Pahute Mesa (fig. 3a). The magnitude of this anomaly has few counterparts in the western United States (Simpson and others, 1986). Healey (1968) interpreted the source of this prominent low as a thick sequence of low-density volcanic rocks within one or more volcanic depressions (calderas) (also see Noble and others, 1968; Orkild and others, 1968, 1969; Sawyer and Sargent, 1989; Minor and others, 1993; Sawyer and others, 1994, 1995; McKee and others, 1999). All of the other calderas of the SNVF that are present in the study area are also well expressed in the gravity data; however, the nature of that expression varies greatly between the different calderas. This variation reflects differences in the development of these calderas, for example, whether or not the calderas are resurgent, and whether the largely volcanic caldera-fill is exceptionally densely welded or only poorly welded.

Many of the most prominent gravity and magnetic lineaments in the study area reflect either the outer boundaries of the central caldera complex, or internal structures that represent the caldera-collapse faults that bound the individual calderas that comprise the complex. Other gravity and magnetic lineaments in the study area may reflect geologic structures such as detachment and extensional faults, and accommodation zones. For example, the high-frequency aeromagnetic data (fig. 3b) correlate well with mapped surficial extensional faults, especially on Pahute Mesa. Previously unrecognized faults and

the extensions of known faults also are apparent from such lineaments. The northeast-trending gravity and magnetic anomalies defining the Thirsty Canyon lineament on Pahute Mesa apparently terminate to the south at about 37°05'N and are replaced by gravity structures that trend east-west, and correspond to large, abrupt, lateral thickness changes in some parts of the volcanic section (the Colson Pond, Fleur-de-Lis, and Hot Springs lineaments, fig. 3a; Fridrich and others, 1999a). Moreover, several prominent, north-south-trending magnetic anomalies also are present both in the Oasis Valley basin and in the vicinity of the Transvaal Hills on the west side of the Timber Mountain caldera complex (fig. 3b). Changes in trend of anomalies south of latitude 37°05'N may represent an accommodation zone separating terranes with differing structural grains. The remaining part of this paper defines and characterizes these geophysically defined structures.

PHYSICAL PROPERTIES

The validity of our geophysical models, described in the following sections, hinges on reasonable assumptions of the physical properties of the rocks in the study area. Density of volcanic and sedimentary rocks generally increases with the degree of consolidation, and thus usually correlating with depth of burial, as well as with other factors such as increasing water content (e.g., Snyder and Carr, 1984). Although individual volcanic units can be relatively dense, the average density of basin-fill volcanic sections from ground level to depths of tens or hundreds of meters can be lowered considerably because of the presence of interbedded low-density sediments and poorly consolidated pyroclastic rocks. Large, abrupt lateral changes in average density of Cenozoic sections are particularly pronounced across the bounding faults of some calderas because intracaldera sequences frequently differ from adjacent extracaldera sequences either by including a much larger fraction of poorly consolidated material or, alternatively, by consisting dominantly of exceptionally thick and densely welded intracaldera tuffs.

Total magnetizations are difficult to characterize because the intensity of magnetization may vary widely within volcanic units, both laterally and vertically (see, for example, Rosenbaum, 1993), as well as between units because of variations in magnetic grain size and concentration of magnetic minerals. Near-surface sources will have the greatest effect on aeromagnetic anomalies and, thus, intensities typically decrease with increasing depth of burial. Chemical changes to the magnetic mineralogy commonly occur as depth increases, and these can have a significant effect on the total magnetization. Intensity of magnetization usually decreases with chemical alteration, but occasionally can be increased when the alteration produces new magnetic phases.

Resistivities of geologic units, in general, are dependent on the temperature, volume fraction, connectivity, and ionic composition of fluid in the pore spaces, as well as the conductive mineral content.

Ionic exchange between minerals and pore fluids at the pore interface is an important factor determining the resistivity of a rock (Keller and Frischknecht, 1966; Olhoeft, 1980, 1981; Keller, 1987; Palacky, 1987). Pore space is initially created during emplacement or deposition, but can be modified radically through diagenesis, alteration, and secondary mineralization. For example, the porosity of an ash-flow tuff decreases as welding of the unit increases, resulting in higher resistivities. Any geologic processes that fill or reduce pore volume or change the ionic content of a pore fluid will change the resistivity of a unit.

Although rock physical properties can vary widely for a given rock unit, we will show that sufficient data exist to estimate representative values that can be used to derive reasonable geologic models based on the inversion of geophysical data.

Density-Depth Functions

In order to model structures that may be responsible for gravity features in the study area, representative densities of the Tertiary deposits filling the basins that underlie the Pahute Mesa–Oasis Valley area must be estimated. We adopted depth intervals similar to those of Jachens and Moring (1990) in making our estimates. The accuracy of thickness estimates derived by the gravity inversion technique is dependent on the assumed density-depth relation of the Tertiary rocks, and on the initial density assigned to the basement rocks. Mankinen and others (1999) calculated a representative density-depth distribution for basin fill in the study area using numerous density data available from the NTS and vicinity ("Initial Model," table 1). Particularly useful in these calculations were the data reported by Healey (1968), Snyder and Carr (1984), Carroll (1989), Langenheim and others (1993), Ferguson and others (1994), and Warren and others (1999). Density information also was obtained by examining gamma-gamma density logs for four drill cores (PM-2, PM-3, UE20j, and UE20f) on Pahute Mesa, and from borehole gravity studies for numerous wells throughout the area (Hildenbrand and others, 1999; Phelps and others, 1999).

Data from Pahute Mesa itself, however, are anomalous. Borehole gravimeter studies in the Silent Canyon caldera complex (Ferguson and others, 1994; Warren and others, 1999) show low-density material (2100 kg/m^3 and less) extending to depths greater than 1 km, whereas densities elsewhere are usually 2450 kg/m^3 below 1 km. This finding indicates that density distributions, at times, may differ significantly between extracaldera and intracaldera sequences. A low density-depth distribution (Hildenbrand and others, 1999; McKee and others, 1999) is shown in table 1 ("Silent Canyon Model"). In other cases, a density-depth distribution higher than our preferred model is predicted. In areas of resurgence, such as Timber Mountain, multiple intrusions into the overlying volcanic sequence and concomitant hydrothermal alteration can make the basin fill much denser than elsewhere. For instance, Byers and others (1976b) reported numerous intrusions comprising an outer tuff-dike zone, inner ring

dikes, and intrusive rhyolite near the center of the resurgent dome. New drill holes near Timber Mountain (L. Prothro, Bechtel-Nevada, oral commun., 2000) reveal densities of about 2400 kg/m^3 near the top of the volcanic sequence. A hypothetical high-density distribution ("Timber Mountain Model") is thus also shown in table 1.

Because density measurements were made on samples from various drill holes and surface outcrops within the study area and surrounding region, our initial density distribution model (table 1) should be widely applicable and perhaps representative of many volcanic terrains. This density-depth function is comparable to those determined by Jachens and Moring (1990), and Blakely and others (1998, 2000) for other volcanic basin-fill deposits throughout Nevada. The Silent Canyon and Timber Mountain models (table 1) are considered to be end members of a continuum of geologically reasonable density-depth functions. Densities for specific groups of rock units in the study area are given in table 2.

Magnetic Properties

Paleomagnetic directions and total magnetizations of the volcanic rocks in the study area are needed for the interpretation of aeromagnetic anomalies. We need to consider, mainly, a representative magnetic polarity for composite rock sequences (defined below) underlying the aeromagnetic survey rather than the magnetization direction of individual units. When the magnetization direction of rocks spanning an appreciable length of time (on the order of 10^4 years) are averaged, their mean direction will correspond to one produced by a geocentric axial dipole field. In rocks as young as these, the expected direction is that of the ambient field. Hudson and others (1994), however, have documented a small (7.1° ± 6.6°), but significant, amount of post-middle Miocene, counter-clockwise, vertical-axis rotation of the central SNVF with respect to the North American craton. The angular distance between this rotated remanent magnetization direction and the ambient field assumed here (I= 60°, D=010°) is roughly 12° and thus will have generally little effect on the results.

Total magnetizations are much more difficult to characterize than are remanence directions. Intensity of magnetization, as mentioned above, varies with position within a volcanic cooling unit, as well as with rock composition, magnetic grain size, and concentration of magnetic minerals. For volcanic rocks in the study area, magnetizations commonly range between 0.1 to 1 A/m, although it is not unusual for specific units to have magnetization intensities well outside this range. Representative magnetizations can be found in Bath (1968), Rosenbaum and Snyder (1985), Ponce and Langenheim (1995), and Grauch and others (1997, 1999). In our modeling approach, we will assume initially that all the volcanic rocks have approximately the same intensity of magnetization with the exception of the Rainier Mesa Tuff, which is generally more magnetic (M.R. Hudson, U.S. Geological Survey, oral commun., 1998). Surficial sediment is weakly magnetized and has no effect on the aeromagnetic anomalies. Pre-Tertiary basement

in the study area consists mainly of weakly magnetized carbonate and siliciclastic sedimentary rocks. One exception is in an area where highly magnetic argillite is present in the Eleana Formation (Bath and Jahren, 1984) as a consequence of local contact metamorphism. Abundant Tertiary granitic dikes, sills, and other subvolcanic intrusions associated with the voluminous volcanism also are expected. Although such intrusive rocks have stronger magnetizations than sediment, they often have little effect on the aeromagnetic anomaly because of their depth of burial, particularly if beneath strongly magnetized surficial volcanic rock.

Electrical Properties

The capability of the MT method to define geologic structures depends on being able to relate electrical resistivity to the composition of rocks and geologic processes affecting those rocks. Anticipated representative resistivity ranges for rock types found in the study area, using the relations given in Schenkel and others (1999), are given in table 3. The selected values are based largely on data from Snyder (1968), Blankennagel and Weir (1973), and on information from electrical logs and lithologic descriptions in Kilroy and Savard (1995).

Modeling Subdivisions

We use density and magnetic information described in this section and in table 1 and table 2 to subdivide the rocks of the study area into six discrete packages that have distinctly identifiable physical properties (fig. 2). These six units are used as a starting point for developing the geophysical models described in the following sections. The units are (listed from the surface to basement):

Surficial sediment. This unit consists of the Oasis Valley gravels and other poorly consolidated surficial sedimentary deposits that are late Miocene to Quaternary in age. Densities are largely comparable with those of the underlying volcanic sequence (1900 kg/m^3). Most of these surficial sediments consist of clasts that were magnetized before they were eroded and redeposited; hence, their magnetizations are now randomly oriented, resulting in little or no net magnetic effect. The thickness of this unit is appreciable only in the deeper parts of some basins and caldera moat areas, where it may locally exceed 300 m.

Normal-polarity volcanic rocks (11.5 to 9.4 Ma). Major constituents of this unit are the tuffs and lavas of the Thirsty Canyon Group, the Volcanics of Fortymile Canyon (mostly lavas), and the uppermost member of the Timber Mountain Group, the Ammonia Tanks Tuff. The density of this unit generally is 1900 kg/m^3, and magnetic intensities range from moderate to strong. Normal magnetic polarity rocks predominate, especially over most of the central caldera complex. Where reversed magnetic polarity

volcanic rocks are present, they tend to be volumetrically small, or to have limited areal extent, and thus to have little effect on the aeromagnetic anomalies (Grauch and others, 1997, 1999) except on western Pahute Mesa as discussed below. Some younger small-volume volcanic units, mainly of Pliocene and Pleistocene age (Fleck and others, 1996), are present locally in the study area but were not encountered in the vicinity of our profiles and thus are not considered a separate (and younger) modeling subdivision.

Reversed-polarity volcanic rocks (12.7 to 11.6 Ma). This unit consists mainly of the Rainier Mesa Tuff member of the Timber Mountain Group, and the underlying Tiva Canyon Tuff member of the Paintbrush Group. These members have a moderately to strongly magnetized, reversed magnetic polarity, and are the oldest units commonly above the water table on Pahute Mesa, where they are relatively thin outflow tuffs. The Rainier Mesa and Tiva Canyon Tuffs are thick (>1 km) intracaldera tuffs within the Timber Mountain and Claim Canyon calderas, respectively. Their density averages 2100 kg/m^3, except when the rocks become saturated (at ~600m) and their density increases to 2300 kg/m^3 (Snyder and Carr, 1984; Mankinen and others, 1999).

Early "caldera-fill" volcanic rocks (13.7 to 12.8 Ma). Included in this unit are all of the members of the Paintbrush Group that predate the Tiva Canyon Tuff, along with the younger units of the Belted Range Group, especially the Grouse Canyon Tuff of the Silent Canyon caldera complex. Here, again, some reversed polarity volcanic rocks are present (Grauch and others, 1997, 1999), but the volumetrically important rocks, including the Topopah Spring Tuff, the Calico Hills Formation, the Bullfrog Tuff, and the Deadhorse Flat Formation, possess normal magnetic polarities. Magnetic intensities generally range from weak to moderate. These units are commonly present below the water table in most areas (between 600 and 1200 m deep under Pahute Mesa), and have an average density of 2300 kg/m^3.

Older volcanic rocks (15 to 14 Ma). This unit includes all of the Cenozoic section predating the Belted Range Group, and consists of both normal- and reversed-polarity members. Magnetizations are mostly weak to moderate, and burial depths are typically greater than 1.2 km. Given the typical deep burial and the mostly much weaker magnetization intensity of this unit relative to most of the overlying volcanic rocks, the older volcanic and sedimentary rocks generally have little effect on the aeromagnetic anomalies. One major exception to this generalization is in the northwestern part of the study area, where this unit is commonly exposed or buried at shallow depth, and has locally very strong magnetic potential, as discussed below. For modeling purposes, we assume that the total assemblage of this unit averages to a normal magnetic polarity. Densities are higher (2450 kg/m^3) than for other volcanic sequences in the

study area—mainly because of deeper burial, although exceptionally dense welding may also have occurred in the northwestern part of the study area.

Basement rocks. The major, exposed basement rocks in and near the study area consist of late Precambrian rocks of varying lithologies through late Paleozoic marine carbonate and siliciclastic sedimentary rocks comprising a section with an initial thickness estimated at ~11 km (Burchfiel, 1964). Typical densities range from ~ 2720 kg/m^3 (carbonate) to ~2650 kg/m (siliciclastics). This thick sedimentary section underwent two phases of radical structural modification. First, late Paleozoic through Mesozoic tectonic contraction imbricated sections of these sedimentary rocks to thicknesses locally as great as double or even triple the original thickness, and locally resulted in strong metamorphism of the sedimentary rocks near the bases of the imbricated sections. Some of these metasedimentary rocks, such as those at the northwest tip of Bare Mountain, are very dense (~ 2750 kg/m^3) because they formed at mid-crustal depths and include dense metamorphic minerals such as garnet (Hamilton, 1988). Subsequent late Cenozoic extension then drastically thinned the crust, such that these sections now vary laterally from thousands of meters to less than 30 m thick. In a few places, they have been removed altogether by tectonic denudation, exposing older high-grade metamorphic rocks, mostly ~1.7-Ga schists and gneisses, which have densities similar to that of the metasedimentary rocks of northwest Bare Mountain. In addition, the basement includes two generations of intrusive rocks, with typical densities of ~ 2670 kg/m^3. The first of these consists of two or three widely scattered, small Mesozoic granitic plutons. The second generation consists of Tertiary intrusive rocks that represent the subvolcanic underpinnings of the SNVF. As a practical matter, these subvolcanic Tertiary intrusions are included here as part of the basement because their physical properties are similar to most of the pre-Tertiary rocks, and differ strongly from that of the Cenozoic eruptive and basin-fill rocks.

Highs in the basement gravity field shown by Hildenbrand and others (1999) are a reflection of the density variations among the basement rocks. Saturated densities of the basement rocks measured by Snyder and Carr (1984) averaged 2660 kg/m^3, whereas Ferguson and others (1994) reported a value of 2650 kg/m^3. We assign a density of 2670 kg/m^3 to the basement rocks, a commonly used reduction density in geophysical studies (see, for example, Simpson and Jachens, 1989).

Both normal and reversed polarities may be present in the basement rocks but, again, there is a diminished effect on the aeromagnetic signal because they are deeply buried over much of the study area. This is especially true where weakly magnetized sedimentary rocks predominate, but even the initially strong magnetization of igneous rocks can be subdued by overlying strata or become weaker when burial has altered their magnetic mineralogy. One exception to a weakly magnetized basement is the argillite member of the Paleozoic Eleana Formation, which locally has a magnetization as strong as any of the

overlying volcanic rocks (Bath and Jahren, 1984). Where the magnetized Eleana Formation is present in areas of shallow basement, it has a significant effect on the aeromagnetic anomalies, as discussed below.

These fundamental units are used in developing our geophysical models and are summarized in table 2. Some deviation from this generalized sequence can be expected because of depth of burial and other reasons detailed in the discussions above. Density contrasts between these units will significantly influence interpretation of the gravity anomalies. The strongly magnetic "normal" and "reversed" volcanics will produce major aeromagnetic anomalies, whereas deeper and weakly magnetized units will have a minimal effect on the magnetic field. The subdivisions in table 2 represent a simplified view of the geology and serves as a starting point for model development. Density, magnetization, thickness and depth of burial often vary within each unit. For example, large lateral variations might be expected for some units depending on whether they are found inside or outside the caldera complexes. The density of individual volcanic units also increases by a few percent (Mankinen and others, 1999) where the rocks occur below the water table (~600 m depth on Pahute Mesa).

ANALYTICAL METHODS

Gravity Inversion

To first order, the isostatic residual gravity field (fig. 3a) reflects a pronounced contrast between dense (~2670 kg/m^3) basement and the significantly less dense (generally < 2500 kg/m^3) overlying volcanic and sedimentary rocks. Once a density-depth distribution appropriate for the study area was determined, the gravity inversion method derived by Jachens and Moring (1990), modified to include drill-hole data, was applied to separate the isostatic residual anomaly into basement and basin fields. The basin gravity field reflects variations in the thickness of low-density Tertiary volcanic and sedimentary rocks. The basement gravity field reflects changes in density related to lateral lithologic variations within the basement. To reduce edge effects, the area over which the inversion was made was extended beyond the study area (from latitude 36°30' to 37°30' and from longitude 115°52.2' to 117°).

In the inversion process, the density of basement is allowed to vary horizontally but the density of basin-filling deposits is fixed using the representative density versus depth relationships (table 1). In this iterative approach, a first approximation of the basement gravity field is derived from those gravity measurements made on exposed pre-Tertiary rocks, augmented by appropriate basement gravity values calculated at sites where depth to basement is known. This approximation (which ignores the gravity effects of nearby basins) is subtracted from the observed gravity, which provides the first approximation of the basin gravity field. Again using the selected density-depth relation, the thickness of the basin-filling deposits is calculated. The gravitational effect of this first approximation of the basin-filling layer

is computed at each known basement station. This effect is, in turn, subtracted from the first approximation of the basement gravity field, and the process is repeated until successive iterations produce no substantial changes in the basement gravity field. Adding the computed basin and basement anomalies will produce the original gravity field shown in figure 3a.

Potential-Field Modeling

Observed gravity and magnetic anomalies were used to construct initial geologic models of the subsurface using a commercially available 2½-dimensional modeling program based on Webring (1985). The program requires an initial estimate of model parameters (table 1), and varies them to reduce the weighted root-mean-square error between observed and calculated gravity and magnetic values along the profile. However, the solutions obtained are not unique because an infinite number of geometric models will have an associated field that closely matches the measured field. To constrain the models, we use the simultaneous inversion of gravity and magnetic data together with available drill-hole information (IT Corporation, 1996), published geologic reports, maps and interpretive cross sections, density-depth variations (discussed previously), and our estimate of basin thickness. Where possible, our profiles crossed several deep drill holes (PM-2, PM-3, UE20j, UE20f, and MYJO Coffer#1) or were extrapolated to gravity stations established on basement rock outcrops to provide additional constraints. If assumed densities and magnetizations are representative, the resulting geophysical models should reasonably approximate actual geologic structure.

Magnetotelluric Methodology

A MT data analysis program (M. Hoversten, Lawrence Berkeley Laboratory, written commun., 1997) provides numerous data processing and interpretative capabilities used in deriving geologic models. The forward modeling and inversion algorithms used for this study assume a two-dimensional cross section divided into cells with uniform resistivity. For regions along a profile and near the Earth's surface, the height and width of the cells used in the inversion were about 50–75 m and 50–100 m, respectively. The cells outside the area of interest increased in size based on the frequency (wavelength) of the data. The inversion codes perform a nonlinear least-squares regression on the data to calculate the resistivity of each cell. The iterative process minimizes the objective function, composed of the squared misfit and a smoothness operator. Iteration continues until the objective function reaches some tolerance value based on the noise level of the data.

Four-layer and three-layer models were selected from our initial estimates of model resistivities for the profiles on Pahute Mesa and in the Oasis Valley basin, respectively. The resolution of resulting modeled structures decreases with depth and lateral position away from the center of the profile. Without

a priori information, the maximum depth of investigation at the center of the profile is approximately equal to the length of the profile. With additional information such as independent geophysical data, however, one can make a reasonable estimate of the resistivity structure beneath much of a profile. For example, MT line 1 was further constrained by geologic and geophysical data from drill hole UE20f.

RESULTS

The 3D inversion of the basement gravity field estimates the thickness of low-density Cenozoic sedimentary and volcanic basin-fill material of the study area. Because the estimated thickness of basin fill is critically dependent on the density-depth function chosen, the general shapes and relative depths of basins should be considered accurate only on a regional scale. On a smaller scale, departures from our preferred density-depth distribution are as described previously for the Silent Canyon caldera complex and the Ammonia Tanks resurgent dome beneath Timber Mountain. In order to demonstrate effects of the differing density-depth functions, we calculated basin thickness for a test area comprising most of the Silent Canyon and Timber Mountain caldera complexes (table 1). The initial model indicated the depth of the basin beneath the Silent Canyon caldera complex to be much deeper than could be supported by geologic considerations. The Silent Canyon model, however, produced a more realistic depth for this basin and satisfied the density data from Pahute Mesa drill holes, but violated constraints from drill-hole data elsewhere in the region. The Timber Mountain model applies only to very limited areas because both average and maximum depths indicated (table 1) clearly are untenable. We note, however, that the basement surface immediately below the Ammonia Tanks resurgent dome occurs at a depth of about 3 km using this model, compared with about 1 km using our general model.

We conclude that the thickness calculated with our initial model, merged with that calculated using the Silent Canyon model for the area described by Hildenbrand and others (1999) and McKee and others (1999), provides the best estimate of depth to basement for the study area (fig. 5). However, calculated basement depths under Fortymile Canyon and Beatty Wash (fig. 5) appear anomalously shallow. Both are major drainages that exhibit gravity highs, probably resulting from use of a reduction density (2400 kg/m^3) that is too high. Kane and others (1981) demonstrated that a reduction density of 2200 kg/m^3 removes the high over the depression associated with the Fortymile Canyon area. The same reduction density eliminates the correlation of Beatty Wash with topography.

Subtracting thickness from topographic elevation yields a three-dimensional view of the elevation of the pre-Tertiary basement surface (fig. 6). In general, basement is relatively shallow (<1 km) outside the caldera systems along the western and southern parts of the study area. Basement descends to relatively great depths (~6 km) beneath the Silent Canyon caldera complex. The resurgent domes of the Black

Mountain caldera and the Timber Mountain caldera complex are clearly expressed as elliptically shaped basement highs, indicating that they may be cored by high-level Tertiary intrusions. Two nested calderas were postulated for the Silent Canyon caldera complex on the basis of drill-hole thickness and lithology of several tuffs (Sawyer and Sargent, 1989; Minor and others, 1993; Sawyer and others, 1994). Because individual calderas cannot be distinguished on the basin thickness map, we refer only to the Silent Canyon caldera complex recognizing, however, that it probably consists of one main caldera that has been internally modified by a group of younger, partially collapsed calderas (McKee and others, 1999).

The magnetic potential field highlights a number of deep and thick magnetic sources in the study area. The most prominent positive magnetic anomaly is present in a zone extending from the Colson Pond lineament to an area northeast of Black Mountain (fig. 4). An integration of geologic map data with the geophysical data indicates that this anomalous zone is the product of a number of different, contiguous features. In the north-central part of this zone, the bounding ring fracture zone of the Black Mountain caldera is clearly expressed as a partial truncation of the magnetic highs on the north and south sides of the caldera. Within the perimeter of this caldera-margin truncation, a small area of high magnetic potential is present in the south-central part of the caldera that is roughly coincident with the resurgent dome. The resurgent dome of this caldera is predominantly exogenous, consisting of a thick stack of basaltic to trachytic lavas. The central magnetic high within this caldera probably reflects the strongly magnetic character of the dominantly mafic lavas of this exogenous lava dome, along with its intrusive underpinnings.

The large magnetic high on the south side of the Black Mountain caldera is truncated on the southeast side by the Thirsty Canyon lineament—i.e., by the western caldera margin of the 11.6-Ma Rainier Mesa caldera. Some post-11.6-Ma basalts lie within the area of the high, especially the Thirsty Mountain basaltic shield volcano (fig. 4). These relatively young, small-volume basalt flows may play a role in creating the magnetic high, but the major cause must be a geologic feature that is more extensive and older than the Rainier Mesa Tuff (11.6 Ma). Part of the northwestern boundary of this large magnetic high is coincident with, and has the same trend as a small exposure of the northwestern topographic wall of the proposed (C.J. Fridrich and S.A. Minor, U.S. Geological Survey, unpublished data, 1995) ~14.2-Ma Tolicha Peak caldera (see also figure 1 in Fridrich and others, 1999a; figure 7 and table 4 in Grauch and others, 1999). On the southeastern (inboard) side of the exposed caldera wall, the tuff of Tolicha Peak is exceptionally thick and densely welded, and has high magnetic potential. Hence, the part of the large magnetic anomaly that lies to the south of the Black Mountain caldera is explained by exposed parts of the Tolicha Peak caldera fill. The gravity inversion model over this area indicates that the Cenozoic rocks are very thin. Because of the exceptionally dense welding of the intracaldera tuff of Tolicha Peak in this area, the basin-fill thickness is probably somewhat thicker than is shown in figure 5. However, the

Cenozoic basin-fill thickness must be less in this general area than it is east of the Thirsty Canyon lineament as evidenced by a small exposure of Paleozoic carbonate rocks near the western margin of the Black Mountain caldera.

That part of the large magnetic anomaly that lies to the north of the Black Mountain caldera (fig. 4) must be caused by a rock unit that predates the ~9.4 Ma Black Mountain caldera because its southern edge is truncated at the margin of the caldera. The causative rock is nowhere exposed because this part of the area is mantled by outflow tuffs of the Black Mountain caldera, and interbedded thin basalt flows. The anomaly could possibly be a continuation of the Tolicha Peak caldera.

The strongly magnetized argillite of the Eleana Formation produces, at least in part, the magnetic high near Shoshone Mountain (fig. 4). Depths to the magnetic source were estimated to range between 3.0 and 1.5 km (Oliver and others, 1995; Ponce and others, 1995), leading Grauch and others (1999) to conclude that the primary source of the high is probably a granitic intrusion or batholith. This conclusion is supported by observations made by Bath and others (1983) and Bath and Jahren (1984), suggesting that the high magnetite content in the argillite in the Yucca Mountain–Calico Hills area resulted from thermal metamorphism by an underlying pluton. If true, the argillite should not be strongly magnetized outside the area of proposed metamorphism. In any event, there is no evidence to indicate that the argillite of the Eleana Formation is present in the northwestern part of the study area.

A comparatively small, discrete magnetic anomaly occurs at Gold Meadows (fig. 4) and a similar anomaly is present in the vicinity of the Climax stock approximately 10 km east of Gold Meadows. Both anomalies are caused by composite granitic (granodiorite to quartz monzonite) plutons (Snyder, 1977; Bath and others, 1983; Orkild and others, 1983), emplaced during late Cretaceous time. Three-dimensional modeling of magnetic data (R.C. Jachens, U.S. Geological Survey, written commun., 1999) indicates that the two stocks represent cupolas of a larger intrusion at depth. Magnetizations of 1.55 A/m for the Climax stock (Bath and others, 1983) and 0.8 A/m for the Gold Meadows stock (R.C. Jachens, written commun., 1999) show that even silicic intrusive rocks can often produce significant magnetic anomalies.

Based on this evaluation of deep magnetic sources in the study area, but realizing that limited information on rock magnetic properties is available, we speculate that plutons of intermediate to mafic composition may be the source of the magnetic basement in the Thirsty Mountain–Black Mountain area. Such plutons could provide a large volume of magnetic material, and they also are likely to have been cut by a series of basaltic dikes (also strongly magnetic) that served as feeders for the lava flows of Thirsty Mountain and surrounding area.

The 3D view of the basement surface (fig. 6) highlights a number of the prominent lineaments seen on the regional gravity and aeromagnetic maps. Fifteen detailed gravity profiles (Mankinen and others,

1999) and six MT lines (Schenkel and others, 1999) were occupied to provide new details as to the nature of these features. Among these profiles, the eight shown in figure 5 are considered representative and will be described in detail below. Some of the profiles from Mankinen and others (1999) have been modified to incorporate new data from holes drilled subsequent to that study (IT Corporation, unpublished drilling reports, 1999; 2000). Major gravity features examined below include the Thirsty Canyon lineament, a gravity high separating the Silent Canyon and Timber Mountain caldera complexes, the gravity gradients along the western and southern boundaries of the Oasis Valley basin, and the gravity gradients along the southern boundaries of the Claim Canyon and Rainier Mesa calderas. Also potentially significant to groundwater flow are the east-west-trending Colson Pond, Fleur-de-Lis, and Hot Springs lineaments (fig. 3).

Thirsty Canyon Lineament

Prominent among the major anomalies seen on both regional gravity and aeromagnetic maps are those occurring along Thirsty Canyon (fig. 1) and referred to as the "NNE-trending structures" by Grauch and others (1997, 1999). We refer to the major gravity anomaly as the Thirsty Canyon lineament (TCL) (fig. 3a). This definition of the TCL encompasses gravity features 14 and 14/SCC of Grauch and others (1997, 1999). The lineament clearly extends at least 35 km from northern Pahute Mesa southward toward Colson Pond near the mouth of Thirsty Canyon. Whether the lineament extends southwest of Colson Pond is not clear from the data available. Basin thickness analysis highlights the feature and shows that there is an abrupt increase (of approximately 3 km) in depth of the basement surface southeast of the TCL (fig. 5 and fig. 6b).

Gravity and aeromagnetic anomalies (fig. 3) are approximately colinear along the southwest section of Thirsty Canyon but become more divergent toward Pahute Mesa (also see magnetic feature 13 of Grauch and others, 1997, 1999). The aeromagnetic anomalies are west of the gravity anomaly, have a more northerly strike, and we believe that they most likely express the edges of some strongly magnetic volcanic units rather than a single, deep-seated geologic structure. Note the near correspondence of these anomalies with the eastern limits of the areas of strong magnetic potential that extend northeastward from the vicinity of Springdale (fig. 4). These areas of strong magnetic potential correspond to what Grauch and others (1997, 1999) called regional magnetic highs 10 and 11, northwest of their magnetic feature 13.

Potential-field models

Four detailed gravity profiles (1, 2, 2a, and 3, fig. 5) and two MT lines (1 and 2) cross the TCL on Pahute Mesa. MT line 1 was sited along the southeastern end of gravity profile 1 near drill hole UE20f, and MT line 2 coincides with gravity profile 2a. Gravity profile 5a and MT Line 4 were located along the same

traverse at the north end of the Oasis Valley discharge area, where the continuation of the TCL is unclear. The lineament source structure(s) might simply terminate along structures expressed by EW-trending magnetic and gravity gradients at about latitude 37°05'N (fig. 3). Alternatively, the source of the TCL may continue on trend to the southwest but be masked by the upper plate of a detachment fault (Fridrich and others, 1999a,1999b). A third possibility is that the TCL is offset about 2 km left-laterally approximately 1 km northeast of profile 5a (fig. 5) and continues southwestward from that point toward the north-trending Hogback lineament (Hildenbrand and others, 1999; Mankinen and others, 1999).

Profile 1 (fig. 5) is well constrained with respect to rock properties and lithologies in drill holes PM-2 (northwest end of the profile), UE20j, and UE20f (southeast end of profile). Figure 7 shows a geophysical model (fig. 7c) of profile 1 that satisfies the constraints imposed by both the drill-hole data and estimated depths to pre-Tertiary basement, and adequately accounts for the observed magnetic (fig. 7a) and gravity data (fig. 7b). Our interpretation of geologic structure along profile 1, based on the geophysical constraints and lithologic logs (IT Corporation, 1996) for drill holes PM-2, PM-3, UE20j, and UE20f, is shown in fig. 7d. There is a distinct near-vertical offset (downthrow ~1.5 km to the southeast) at the TCL as defined on the anomaly maps. There also are similar features, but with apparently less offset, approximately 2.5 and 5 km to the northwest. The strong magnetic anomaly between holes PM-2 and UE20j (fig. 7a) corresponds closely to the northwesternmost extent of the reversed magnetic polarity Rainier Mesa and Tiva Canyon Tuffs; reversed polarity units are not present in drill hole PM-2.

In addition to the absence of reversed polarity units, the normal polarity rocks along the northwest part of profile 1 also must be more strongly magnetized than they are to the southeast in order to account for the magnitude of this magnetic anomaly. Grauch and others (1997, 1999) considered the sources of this northwestern magnetic high to be the commendite of Ribbon Cliff (part of unit #2 in table 1), and the tuff of Tolicha Peak and the dacite of Mount Helen (parts of unit #5, table 1). Although only the dacite of Mount Helen was reported from drill hole PM-2, all three units probably underlie most of the area of the magnetic high. In the models shown in figure 7, the magnetization of unit #1 has been increased from 0.8 A/m near the southeast part of the profile to 1.2 A/m near the northwestern magnetic high.

MT models

Average resistivity, lithology, and alteration (table 4) of rocks from drill hole UE20f (Warren and others, 1999) and depth to the pre-Tertiary surface (Hildenbrand and others, 1999) were used to constrain MT models along profile 1. Changes in electrical resistivity correlate well with variations in alteration and lithology. Lava flows and flow breccias are highly resistive, while the resistivity of welded tuff varies depending on the degree of welding and type and degree of alteration. In general, zeolitic rocks tend to

have lower resistivities than those identified as vitric or devitrified. Even though there may be many heterogeneous units in a sequence, they are generalized by the MT inversion and modeling process. The MT model along profile 1 (fig. 8), therefore, shows only four distinct layers (three in the upper 5 km) that can be clearly identified. These layers are a resistive surface layer (~100-250 Ωm), a conductive layer (~30-60 Ωm) extending to about 2 km, a second resistive layer (~300-500 Ωm) whose thickness decreases from east to west, and a generally conductive zone between 3.8 and 3.2 km. The lack of fine-scale resolution is due to inherent smoothness in data generated by the MT method (Schenkel and others, 1999).

Although the MT method is poorly suited for identifying individual geologic units, it is well suited for delineating gross structures such as fault zones with significant vertical offsets. The MT model along profile 1 indicates a number of near vertical changes in resistivity in the uppermost conductive layer (fig. 8). The bottom of this conductive layer is down-dropped (perhaps as much as 500 m) toward the east. This MT model closely resembles that generated by potential-field modeling (figs. 7c, 7d). Two or more vertical, seemingly subparallel breaks in resistivity, density, and magnetic properties indicate that the TCL does not express a single structure but perhaps a broad fault zone. The deeper basement and thickening of the overlying volcanic layers along the eastern part of profile 1 is consistent with this being the northwestern boundary of the Silent Canyon caldera complex.

Geologic interpretation

Our geologic interpretations of geophysical models for all profiles measured perpendicular to the Thirsty Canyon lineament are shown in figure 9. This figure shows only the portion of each of those profiles (fig. 5) in proximity to the TCL. Models for all profiles are shown in their entirety in Mankinen and others (1999). The model for profile 2 (fig. 9b) shows two subparallel structures similar to those seen along profile 1 (fig. 7d and fig. 9a). Both structures, again, indicate moderate- to high-angle offsets. The resistivity model for this area (MT line 2) is oblique to gravity profile 2 (along 2a of fig. 5) and is shown in figure 10a. Several near-vertical breaks are apparent, as is deepening of the basement toward the east near the trace of the TCL. Models generated by both methods, again, are compatible with the presence of a broad fault zone. There is a much more abrupt change in observed gravity at profile 3 (Mankinen and others, 1999) than at the other profiles crossing the TCL indicating that the main fault here must be essentially vertical (fig. 9c).

The model for profile 5a (fig. 9d) indicates that the Thirsty Canyon lineament was not crossed or, alternatively, that its expression is too subdued to be detected. The TCL also may have been offset toward the east where the model shows a deepening of the basement surface. The MT model for line 4 along this traverse, however, does indicate the presence of a 2-km wide conductive zone and vertical discontinuities

near the expected position of the TCL (fig. 10b). The MT model shows that the resistivity discontinuities at this locality are confined to relatively shallow depths, and that perhaps any vertical displacements related to the TCL are insufficient to produce a clear gravity anomaly. The observed gravity and magnetic data along profile 5a are best fit by assuming that the Rainier Mesa/Tiva Canyon Tuffs directly overlie the basement immediately southeast of the extension of the TCL. The density of these reversed volcanic rocks increases from 2100 to 2450 kg/m^3 with increasing depth. The northwestern end of profile 5a also crosses the Hogback lineament, but at an oblique angle (fig. 5). Here the Hogback lineament seems to be very near the westernmost extent of the reversed polarity volcanic rocks (fig. 9d).

Thirsty Canyon fault zone

Many workers over the years have noted that multiple calderas of the SNVF seem to have a common boundary in the vicinity of Thirsty Canyon. As discussed above, both potential-field and MT modeling indicates that a zone of en-echelon faults exists along its length and that the lineament is best described as a fault zone. This zone, which we refer to as the Thirsty Canyon fault zone (TCFZ), appears to be on the order of 2 to 3 km wide. The subparallel faults forming the zone have produced step-like incremental displacements (fig. 9) typical of a caldera ring-fracture system (e.g., Smith and Bailey, 1968).

Although most ring fracture systems are arcuate in shape, Smith and Bailey (1968) reported that others form linear faults for short distances. Nevertheless, the atypical length of the TCFZ probably indicates that it results from a coalescing of ring-fracture systems related to collapse of multiple calderas. The linearity of this inferred composite ring-fracture system was most likely caused by a zone of weakness developed initially during the early stages of basin-and-range faulting prior to eruption of the oldest tuffs in the SNVF. Grauch and others (1997) raised the possibility that such a preexisting feature could have controlled development of the calderas. If true, this ancestral TCFZ apparently accommodated caldera growth, and its step-like fault scarps formed a topographic barrier. Rose and others (1999) report mantle-derived ^{3}He in Oasis Valley groundwater and suggest that major structural features in the region, such as the TCFZ, may extend downward to the deep crust or upper mantle. We thus view the TCFZ as both a deep ancestral fault zone and a younger composite ring-fracture system related to calderas of the SNVF.

Zone between the Silent Canyon and Timber Mountain Caldera Complexes

Basin thickness variations between the Silent Canyon and Timber Mountain caldera complexes (fig. 5) reveal an apparent basement high that extends to within 3 km of the surface. Gravity profile 4 (fig. 5 and fig. 11) crosses this feature and the steep gravity gradients detected along it seem indicative of high-angle basement structures. Alternatively, the anomaly could express a large volume of dense intrusions

penetrating a ring fracture system within the structural margin of either the Timber Mountain or Silent Canyon caldera complexes. It may also reflect volcanic rocks within the caldera sequence(s) if their density was increased sufficiently by extensive hydrothermal alteration. A new drill hole located above this area did encounter alteration occurring at less than 1 km below the surface (R.G. Warren, Los Alamos National Laboratory, oral commun., 1999). The continuity of the feature, however, gives the impression that this area represents a block or septum that avoided subsidence as separate caldera systems developed to the north and south.

Gravity profile 4 also crosses the Buckboard lineament, as expressed by aeromagnetic data (fig. 3b), near the northwestern edge of the basement high. The Buckboard lineament is part of the "WNW-trending structure" (magnetic feature 18) of Grauch and others (1997, 1999). This lineament is not detected by the gravity data, which may indicate that the aeromagnetic anomaly is an expression of near-surface magnetic sources extending to a topographic wall of the Timber Mountain caldera complex. However, the MT model for line 3 (Schenkel and others, 1999), which crosses the Buckboard lineament approximately 5 km northwest of gravity profile 4, indicates the presence of a highly conductive zone southeast of the Buckboard lineament that extends from near the surface down into the second conductive horizon.

Oasis Valley Basin

The small Oasis Valley basin has long been recognized, but little agreement has existed as to its origin. Byers and others (1976b) suggested that the basin resulted from collapse following eruption of the Ammonia Tanks Tuff, whereas Christiansen and others (1977) considered it to be related to the earlier eruption of the Tiva Canyon Tuff. Later mapping failed to support either of these early interpretations, leading Sawyer and others (1994) to suggest that the concept of a caldera origin for this basin be abandoned.

More recent and detailed geologic mapping, when integrated with the geophysical data, indicates that the Oasis Valley basin is a half-graben (Fridrich and others; 1999a; 1999b). It is a tectonic basin that partially overlaps the Rainier Mesa caldera, and formed at the eastern limit of a low-angle detachment fault that extends westward of Oasis Valley, under the Bullfrog Hills. Field relations indicate that the Bullfrog Hills detachment fault itself partly predates the 11.6-Ma Rainier Mesa caldera, but the Oasis Valley basin postdates this caldera (i.e., it formed at a late stage in the evolution of the detachment fault; Fridrich and others, 1999a). Fridrich and others (1999a) consider the boundaries of the basin to coincide with the southern margin of the Black Mountain caldera on the north (fig. 1 and fig. 5), the Hot Springs and the Hogback lineaments (faults) on the south and west, respectively (fig. 3), and the western dip-slope margin of the Transvaal Hills on the east (fig. 1). Our data show that the northwestern part of this geologically defined basin (north of Colson Pond, at ~37º05'N, and northwest of the Thirsty Canyon

lineament; fig. 1) is too shallow to have significant geophysical expression. In our analysis below, we will therefore consider only the deeper parts of this basin that lie east of the Hogback lineament and southeast of the Thirsty Canyon fault zone (fig. 5 and fig. 6).

Internal structures of the Oasis Valley basin include the Thirsty Canyon, Colson Pond, and Fleur-de-Lis lineaments, all of which reflect structures that formed prior to the formation of this half-graben. Additionally, the southern boundary of the basin is coincident with the eastern part of the Hot Springs lineament (fault), which also formed prior to this basin, as discussed above. Whereas the Thirsty Canyon lineament and the eastern part of the Hot Springs fault are structures along which caldera collapse occurred (although both may have originated as tectonic structures), the Colson Pond and Fleur-de-Lis lineaments and the western part of the Hot Springs lineament are enigmatic features of uncertain origin and significance; they may be structures related to the Rainier Mesa caldera or they may be tear faults related to the detachment fault, as discussed below.

The western part of the Hot Springs lineament is simply that—it is both an aerial photo and a geophysical lineament along which there is a cluster of hot springs, but no provable geologic structure coincides with it. In contrast, large abrupt changes in the thickness of the Ammonia Tanks Tuff and of several related, slightly younger tuffs and lavas are present across both the Colson Pond and Fleur-de-Lis lineaments (Fridrich and others, 1999a). In the case of the Fleur-de-Lis lineament, the stratigraphic change coincides with a well-exposed paleo-fault scarp, of which only the 0.5 km is exposed (Fridrich and others, 1999a). The minimum magnitudes of vertical offset across the Colson Pond and Fleur-de-Lis lineaments, as indicated by a combination of structural offset and stratigraphic thickening, are 3 km and 1.5 km respectively (Fridrich and others, 1999a), which both far exceed the magnitudes of vertical offset as estimated from the gravity gradients across these features, discussed below. The probable explanation for this disagreement is that these features are tear faults separating parts of the detachment fault that have broken at different structural levels (Fridrich and others, 1999a). Thus the large change in stratigraphic thickness of Cenozoic basin fill across these faults is partly truncated at depth by the detachment fault system.

Colson Pond lineament

The Colson Pond lineament is an east-trending gravity feature normal to the Hogback lineament near the northern end of Oasis Valley (fig. 3). Detailed gravity profiles 6 and 7 (Mankinen and others, 1999) cross the feature, and the geophysical model and geologic interpretation for profile 7 are shown in figure 12. Aeromagnetic data (fig. 3b) are suggestive of an eastward continuation of the lineament as defined by gravity (fig. 3a) toward the Transvaal Hills. Gravity profiles 6 and 7 were extended to include surface outcrops of pre-Tertiary basement and provide constraints for the potential-field modeling. Although it

can be recognized in both gravity profiles, the magnitude of the associated anomaly is small (< 5 mGal). The Colson Pond lineament seems to be a feature (fault) with moderate dip and a small amount of downthrow (~0.5 km northward) of low-density material at profile 7 (fig. 12), and of basement rock at profile 6 (Mankinen and others, 1999). MT line 5 (Schenkel and others, 1999) near the eastern terminus of the lineament indicates a thickening of the surface conductive layer that could be due to alteration within a fault zone. However, the Colson Pond lineament where it intersects MT line 5 (and gravity profile 5b of Mankinen and others, 1999) is poorly expressed by gravity. The basement surface beneath gravity profiles 6 (Mankinen and others, 1999) and 7 (fig. 12) is very shallow, and magnetic anomalies observed near both profiles are modeled by assuming a strong, normal-polarity magnetization (2.0 A/m) for some of the basement rocks.

Fleur-de-Lis lineament

The Fleur-de-Lis lineament parallels, and is approximately 2 km south of, the Colson Pond lineament (fig. 3a). Only the extension of gravity profile 7 (fig. 12) crosses the feature. Although regional gravity observations are sparse in this area, the data indicate that the lineament represents a moderately dipping structure. Rocks on the north side of the Fleur-de-Lis lineament are displaced downward (~0.5 km) relative to those on the south. Only a thin layer of Tertiary volcanic or sedimentary rock overlies basement southwest of the lineament. Reversed-polarity volcanic rocks are absent along profile 7. Field relations indicate that both the Fleur-de-Lis and Colson Pond lineaments were growth faults during the Timber Mountain caldera cycle (Fridrich and others; 1999a).

Hot Springs lineament

The eastern part of the Hot Springs lineament coincides at least roughly with both the southern boundary of the Oasis Valley basin and the southwestern boundary of the Rainier Mesa caldera (fig. 3). The western part of this lineament marks the boundary between the northeastern and southeastern Bullfrog Hills structural domains (Fridrich and others, 1999a). Although it is the physiographic boundary of the present-day Oasis Valley basin, the regional geophysical data (fig. 3) suggest that this “lineament” does not, in fact, reflect a single, continuous feature; the eastern and western parts may be separate features.

No detailed gravity profile was established in the area of this lineament although some information was obtained from the extension of profile 11 (Mankinen and others, 1999), which crosses the lineament near its eastern extent. Here the geophysical model supports the interpretation that the major discontinuity is the southwestern part of the caldera-collapse ring-fracture zone of the Rainier Mesa caldera, located approximately 3 km northeast of where we show the Hot Springs lineament. Potential-field modeling reveals sharp discontinuities (faults?) near the lineament, with rocks on the south side of the lineament

denser than those to the north. Because the Hot Springs lineament coincides with a small cluster of hot springs (fig. 3 and fig. 16), a better characterization of this feature, and of the area between it and the caldera margin to the north is important to gaining a better understanding of groundwater flow in the area.

Hogback lineament

The Hogback lineament is a prominent north-south-trending gravity feature (fig. 3a) also described by Grauch and others (1997, 1999) and informally called the Hogback fault (their gravity feature 24). The lineament coincides in part with the scarp of a major concealed normal fault along the eastern side of the Oasis Mountain hogback (Fridrich and others, 1999a, b) that forms one of the major structural features in the study area and defines the eastern boundary of a distinct structural domain (northeastern Bullfrog Hills domain). Gravity profiles 5a, 9, and 10 of Mankinen and others (1999) cross the Hogback lineament. Profile 10 (fig. 5 and fig. 13), near the southern end of the lineament, has been extended to the west to include a surface outcrop of pre-Tertiary basement to help constrain the geophysical models. The Hogback lineament at profile 10 is consistent with a moderate, east-dipping fault, with the Oasis Valley basin being 2 km deeper on its eastern side. The basin progressively shallows away from this area toward the north. Reversed polarity lavas are very thin or absent west of the Hogback fault at all three gravity profiles (Mankinen and others, 1999).

Yucca Mountain–Beatty Wash

Gravity profile 13 (fig. 5 and fig. 14) begins near drill hole WT-24 on Yucca Mountain and continues northward across Yucca Wash, the Claim Canyon caldera, and ends at drill hole ER-EC-7 within the Timber Mountain caldera complex. The geophysical model for profile 13 is constrained by lithologies encountered in drill hole WT-24 (D.C. Buesch, U.S. Geological Survey, written commun., 1999) and in ER-EC-7 (IT Corp., unpublished drilling report, 1999). Density contrasts (fig. 14c) and our interpretation of geologic structure (fig. 14d) show many moderately to steeply dipping surfaces (faults) along the length of profile 13. The most northerly of the major faults indicated by our modeling is almost directly beneath drill hole ER-EC-7. It is interesting to note that drilling of this hole was stopped at 425 m because incoherent rock was encountered.

A northwest-trending fault has been thought by many workers to underlie Yucca Wash. For example, Bath and Jahren (1984) and Grauch and others (1999) have shown that a series of discontinuous aeromagnetic highs terminate in this area. Fridrich (1998) and Fridrich and others (1998) inferred a concealed strike-slip fault or accommodation zone under this wash, based on an abrupt change in the magnitude and style of structures between Yucca Mountain and the more strongly extended domain to the northeast. D. O'Leary (U.S. Geological Survey, oral commun., 2001) inferred that Yucca Wash coincides

with a strike-slip fault based on strong local strike-slip strain in bedrock exposures on the periphery of the wash. Most geologic maps of the area (Scott and Bonk, 1984; Scott, 1990; Frizzell and Shulters, 1990; Slate and others, 1999) show a fault concealed under this wash; however, other workers have disagreed and show no fault there (Christiansen and Lipman, 1965; Day and others, 1998).

In an attempt to resolve this controversy, Langenheim and others (1993) conducted six detailed gravity and ground magnetic traverses across the wash. They found no evidence of significant vertical offsets; however, their surveys show that some shallow stratigraphic units have a significant abrupt thickness change across the central area of the wash. They interpreted this result as conflicting with the concept that a concealed fault underlies Yucca Wash. Our modeling (fig. 14) does show a major structural boundary on the southwest side of Yucca Wash but directly beneath the northern edge of Yucca Mountain and not accessible to the ground traverses. The scarp on the northern part of the mountain may thus reflect reactivation of an older, deeper fault. Moreover, the geophysical surveys conducted by Langenheim and others (1993) would not necessarily reveal the presence of a small strike-slip fault having negligible vertical offset, and their finding of significant abrupt thickness changes actually supports the strike-slip fault interpretation advocated by most structural geologists who have mapped in this area.

To adequately model the strong, negative magnetic anomaly at about 5 km north of WT-24 (fig. 14) requires a thick, reversely magnetized Tertiary section. This interpretation is consistent with this part of the volcanic sequence being a thick intracaldera accumulation and is in excellent agreement with Christiansen and others (1977) who estimate that about 900 m of welded Tiva Canyon Tuff accumulated within the Claim Canyon caldera. The magnetization, in part, also is stronger than the typical exposures of Tiva Canyon Tuff, indicating that a significant thickness of the strongly magnetized tuff of Chocolate Mountain (Rosenbaum and Snyder, 1985), a late phase of the Tiva Canyon Tuff, must be present. Byers and others (1976b) reported that the tuff of Chocolate Mountain may be about 1 km thick in the Claim Canyon caldera. A block of Eleana Formation also must underlie profile 13 in order to fully account for the large, positive magnetic anomaly at about 8 km (fig. 14a). Geophysical models by Bath and Jahren (1984) and Brocher and others (1998) both include this formation as underlying parts of Yucca Mountain and vicinity.

Calderas

The synthesis of gravity, magnetic, magnetotelluric, geologic, and physical property data lead to new insights as to the structural underpinnings of the SNVC. One of our general conclusions is that many basins are sharply defined by moderately to steeply dipping structures that probably reflect caldera boundaries. These caldera faults are indicated, to a first order, by undulations in the basement surface in the geophysical models and the basin thickness map (fig. 5). We are thus able to more clearly delineate

several of the calderas based mainly on geophysics, with some additional input from geology. Our proposed caldera margins are shown in figure 1 and subsequent figures. Some of these caldera boundaries represent significant revisions to those proposed in earlier publications, and previously undetected calderas or caldera segments are suggested.

Black Mountain caldera

The resurgent dome of the Black Mountain caldera is clearly delineated by gravity data (fig. 3a) and in the map of basin thickness (fig. 5) except on its northwest flank. Here there is little density contrast between the eruptive rocks associated with the caldera and the immediately adjacent basement rocks. A small outcrop of Paleozoic carbonate rock (Ordovician Pogonip Group) west of Black Mountain (Noble and Christiansen, 1968; Minor and others, 1993; Wahl and others, 1997; Slate and others, 1999) indicates that the relatively high gravity in this area (fig. 3a) may be an expression of extensive, dense carbonate in the near surface. Here, also, lava flows filling and surrounding the caldera are more mafic and, hence, denser than typical volcanic rocks elsewhere in the study area. This combination of factors probably helps to mask the caldera signature in the gravity data. The caldera is more clearly expressed by aeromagnetic data (fig. 3b, and Grauch and others, 1999) that indicate strongly magnetized, normal polarity rocks surrounded by more weakly magnetized and/or reversed polarity rocks. The boundaries shown here (fig. 1, fig. 3, and fig. 5) are the topographic walls of the caldera (Minor and others, 1993, 1998); the ring fractures zones of the caldera lie at depth and inboard of the topographic walls and are not exposed in virtually uneroded calderas such as this.

Silent Canyon caldera complex

The general outlines of the Silent Canyon caldera complex are clearly seen on the basin thickness map (fig. 5). The western and northwestern boundaries are represented by ring-fractures along the Thirsty Canyon fault zone (fig. 9a, b). The southern boundary abuts the basement high between the Silent Canyon and Timber Mountain caldera complexes. The geophysical model (fig. 11) indicates a steep, north-dipping caldera fault at this locality. The eastern and southeastern boundaries are shown where the Grouse Canyon and Bullfrog tuffs abruptly thicken across the caldera-collapse faults that formed during the eruption of these voluminous units (McKee and others, 1999). A series of abrupt changes in the basement surface occurs in the area between the Silent Canyon caldera complex and the Eleana Range (fig. 1 and fig. 5), and some of the basin deposits are very deep although restricted in area. It is difficult to delineate any caldera boundaries in this area with any degree of certainty but if the unusual basement topography is somehow related to caldera formation, an early stage in the development of the Silent Canyon complex must be represented. Knowledge of the subsurface here is speculative due to a lack of

drill holes and, if caldera segments are present, identification of the ash-flow tuffs responsible for their formation is precluded.

Timber Mountain caldera complex

Ammonia Tanks caldera. This is the youngest caldera of the Timber Mountain Group and is expressed both by doming of the basement surface (fig. 5) that reflects resurgence (Byers and others, 1976b), and by a series of discrete, high-amplitude aeromagnetic anomalies (fig. 3b). Deep moats surround much of the resurgent dome (fig. 5) and provide an indication of the height of resurgence. Many of the high-amplitude magnetic anomalies occur in a quasi-radial pattern (see, also, plate 1 in Kane and others, 1981, and fig. 4 in Grauch and others, 1999) and may be an expression of the intrusive rocks thought to core the resurgent dome (e.g., Byers and others, 1976b). Late-stage alteration of the intracaldera tuff may also have increased the density of the caldera fill on the resurgent dome, as discussed previously. Pre-Tertiary basement is probably as much as 2 km deeper than shown in figure 5. The Transvaal Hills (fig. 1) lie on the western margin of the Ammonia Tanks caldera. The Transvaal Hills are considered by several workers (Byers and others, 1976b; Christiansen and others, 1977; Sawyer and others, 1994) to be a remnant of the Rainier Mesa caldera resurgent dome that was truncated by collapse of the Ammonia Tanks caldera. More recent mapping has shown that the structure of the Transvaal Hills is consistent with this interpretation (Fridrich and others, 1999b).

Rainier Mesa caldera. The northwestern boundary of this caldera is the Thirsty Canyon fault zone, and the northern boundary is the basement high between the Silent Canyon and Timber Mountain caldera complexes (fig. 5). There is a steep, south-dipping caldera fault at the northern locality (fig. 11). The southern boundary of the caldera abuts the Claim Canyon caldera, and the eastern margin encompasses the largest basins east of the Ammonia Tanks caldera resurgent dome (fig. 5).

Although the eastern boundary of the present-day Oasis Valley basin is the western edge of the Transvaal Hills (fig. 1), this margin is not a caldera boundary. Neither geologic mapping (Fridrich and others, 1999a, b) nor geophysical studies have revealed any deep-seated structure that could serve as a structural domain boundary separating the hills from the basin. Fridrich and others (1999a) also point out that westward tilting of the hills occurred after deposition of the Ammonia Tanks Tuff and was probably related to extensional tectonism rather than to caldera evolution. We believe that the southwestern part of the Rainier Mesa caldera underlies a large part of the Oasis Valley basin (also see Noble and others, 1991; Fridrich and others, 1999a, b).

Gravity profiles 10 (fig. 13) and 12 (fig. 15) cross the western and southern boundaries, respectively, of the Oasis Valley basin and indicate steeply dipping structures that appear to be caldera faults similar to

others described above. The western boundary also nearly coincides with the Hogback fault south of the apparent termination of the Thirsty Canyon lineament (fig. 13). A caldera wall on the west side of Oasis Mountain (fig. 1) was mapped by Noble and others (1991) and Byers and others (1976b), and Fridrich and others (1999a) suggest that this represents the upper structural wall of the Rainier Mesa caldera that was offset from the main, or lower, caldera fault (near the Hogback fault) by a shallow, subparallel detachment surface.

Claim Canyon caldera

The southwestern boundary of the Claim Canyon caldera is expressed geophysically by the farthest extent of high-amplitude aeromagnetic anomalies in that direction (fig. 3b). Basement highs form its eastern limit. It is delineated geologically by the occurrence of pre-caldera tuffs and lavas toward the southeast and southwest (Byers and others, 1976b; Sawyer and others, 1994). Collapse of the Timber Mountain caldera complex to the north destroyed much of the preexisting caldera. Intermediate basement depths beneath the caldera (fig. 5) are, in part, a reflection of structural resurgence (Byers and others, 1976b; Kane and others, 1981; Sawyer and others, 1994; and Fridrich and others, 1999b). Density may be increased over our general model by the presence of dikes intruding the volcanic section above the resurgent dome (Fridrich and others, 1999b) as discussed previously for the Timber Mountain caldera complex. Pre-Tertiary basement, again, is likely to be somewhat deeper than shown in figure 5.

Crater Flat domain

The basin thickness map (fig. 5) shows the presence of two deep basins south of the Claim Canyon and Rainier Mesa calderas. We refer to this area as the Crater Flat domain as proposed by Fridrich (1998), which includes both the physiographic Crater Flat basin (fig. 1) and Yucca Mountain. Evolution of this domain is complex and various hypotheses concerning its origin have been proposed (see Fridrich, 1998; Fridrich and others, 1998, and references therein) and a re-evaluation is beyond the scope of this report. We simply note, however, that the striking geophysical similarity of the Crater Flat sub-basins to deep basins elsewhere in the study area (fig. 5) provides permissive evidence that the former could also have resulted from caldera-forming volcanism. If true, the Crater Flat sub-basins predate eruption of the Crater Flat Group (~13.3 Ma) because the sources of these and younger units have been identified (Sawyer and others, 1994).

Potential Hydrogeologic Implications

The results of this geophysical investigation, when integrated with available geologic data, have several implications for the hydrogeology of the Pahute Mesa–Oasis Valley area. First, the basement gravity field

(Hildenbrand and others, 1999) indicates that the basement rocks in the study area have densities that are lower than those of carbonate rocks. The modeled densities are typical of siliciclastic sedimentary rocks and of granites, neither of which usually act as aquifers in the NTS region. Some Paleozoic carbonate rocks are present in the study area, as indicated by the one small exposure of Ordovician dolomite near the western margin of the Black Mountain caldera, and by the Paleozoic carbonate rocks exposed on Bare Mountain. However, deep burial and hydrothermal alteration of basement rocks has, in all probability, destroyed the permeability of any carbonate rocks that may have been present within the area of the central caldera complex. Moreover, detachment faulting and associated extreme extensional deformation, in the area to the west of the caldera complex, has probably disrupted the continuity of any carbonate rocks in that area to a degree that it is improbable that they would act as an effective aquifer. We, therefore, consider basement in the study area to be an effective barrier to flow of groundwater and that the important aquifers within the study area probably are all within the Cenozoic basin-fill.

The gravity model for the study area shows that there are great thickness variations within the Cenozoic basin-fill. The largest of these variations is between the sections that are inside the caldera complex, versus those that lie outside. Within the caldera complex, the structural and stratigraphic complexity of the Cenozoic rocks is great because this complex is an aggregate of several different overlapping and nested calderas. Whereas the hydrologic behavior of these caldera-fill rocks probably is similarly complex, there are, nonetheless, several generalizations we can make.

For example, hydrologic testing in the holes drilled for the tests under Pahute Mesa has shown that the Cenozoic section there contains several stratigraphic units that are excellent aquifers (Laczniak and others, 1996). In contrast, the water table is apparently very shallow on the top of the closed topographic high that is the Timber Mountain resurgent dome. As Timber Mountain is not high enough to be a major recharge area, the closed water table high under this feature suggests that the caldera fill under this mountain is mostly impermeable. Thus the caldera-fill strata include both aquifers and confining units. Because the stratigraphic aquifers are specific to the individual calderas in which they formed, none has regional extent or continuity. To the extent that there is large-scale continuity of the Cenozoic aquifer system within the caldera complex, it must therefore be a consequence of the major structures of the caldera complex, rather than individual stratigraphic units.

The major structures of the caldera complex are the ring-fracture zones along which caldera subsidence occurred. As discussed above, the geophysical data show that these features are broad fault zones. Surface mapping of these features (e.g., Fridrich and others, 1999b) has shown that fault breccias are typically well developed along these features, and that thick wedges of landslide breccias are commonly banked up against them, along the outer caldera walls. Thus the brecciated rocks within, and adjacent to, these caldera-collapse ring fractures zones may be excellent aquifers. On the other hand, the

water-conducting properties of these caldera-collapse fault zones frequently causes them to be the focus of hydrothermal activity at a late stage in evolution of the calderas, and protracted hydrothermal activity can decrease permeability. The geophysical data give us important information about these caldera structures, allowing us to trace these features through areas where they are concealed at the surface by younger rocks. The three major caldera-collapse structures of note in the study area are: (1) the Thirsty Canyon fault zone, (2) the Hot Springs fault, and (3) the WNW structure at the boundary between the Rainier Mesa caldera and the Silent Canyon caldera complex.

A limited barrier to groundwater flow beneath Pahute Mesa that creates a water-level discontinuity was described by Blankennagel and Weir (1973). They attributed this barrier to a permeability contrast with low permeability on the west and high permeability on the east. Distinct differences in major ion chemistry of water on either side of the feature (Blankennagel and Weir, 1973; Kilroy and Savard, 1995; Laczniak and others, 1996) support the concept of a limited-flow barrier. Part of the barrier is roughly coincident with the western ring-fracture system of the Silent Canyon caldera complex (Blankennagel and Weir, 1973). These authors also noted that aftershock epicenters from some nuclear explosions indicated possible extension of the barrier into the moat of the Timber Mountain caldera. Laczniak and others (1996) cautioned, however, that the feature's existence and hydrologic significance were uncertain beyond the immediate area. The barrier was depicted, instead, as a broad zone by O'Hagan and Laczniak (1996) in order to represent its uncertain position and areal extent (fig. 16).

The coincidence of the O'Hagan and Laczniak (1996) limited-flow barrier with part of the Thirsty Canyon fault zone (fig. 16) strongly indicates that the fault zone is a major control on groundwater flow under the central part of Pahute Mesa. The permeability contrast probably is caused by volcanic rocks on the northwest side of the TCFZ (and outside the Silent Canyon caldera system) being denser than their intracaldera counterparts. The step-like incremental displacement of the fault blocks in this area juxtapose rocks of differing permeability and apparently act as barriers to across-fault groundwater flow westward from the NTS. The permeability barrier was thought to deflect groundwater flow southwestward into Oasis Valley (Blankennagel and Weir, 1973).

Although porosity generally decreases with increasing density, dense rocks fracture easily and, thus, can provide pathways for the movement of groundwater. Substantial fracturing is to be expected in a major fault zone such as the TCFZ, and these fractures should generally strike subparallel to the feature. Fractured rocks of the TCFZ also may have created a channel for hydrothermal fluids of the long-lived volcanic systems of the Silent Canyon caldera complex. Thus, hydrothermal alteration and the resulting formation of conductive clays (Schenkel and others, 1999) could be a major feature of shallow fractures in the TCFZ. Nevertheless, groundwater within the TCFZ may be channeled southwestward, unless the fractures are sealed due to extensive alteration. If the TCFZ is an effective conduit that diverts water-flow

into Oasis Valley, the effectiveness of the structure as a barrier to westward, across-fault flow is enhanced. Thus, the Thirsty Canyon fault zone may simultaneously act as a barrier to, and a conduit for groundwater flow as suggested by Rowley (1998).

Our interpretation that the Thirsty Canyon fault zone may be a significant feature influencing groundwater flow is supported by water isotopic data (J.M. Thomas, U.S. Geological Survey, written commun., 1998). Groundwater discharging from wells in the eastern and central parts of the Oasis Valley discharge area has nearly identical deuterium and ^{18}O values to those from well U-20 WW on Pahute Mesa (fig. 16). These stable isotope values differ significantly from other wells in the study area, specifically in the Springdale and Fortymile Canyon areas (fig. 16). These isotopic differences are corroborated in a general way by previously reported water chemistry data (Blankennagel and Weir, 1973; Kilroy and Savard, 1995; Laczniak and others, 1996).

Other major geophysically expressed structural features in the study area probably also exert significant influence on groundwater flow. Colson Pond, for example, exists near the intersection of the Thirsty Canyon fault zone and the Colson Pond and Hogback lineaments (fig. 16). The alignment of major springs in Oasis Valley with the Hogback lineament is also noteworthy. Groundwater probably is deflected upward as it encounters shallow basement in this vicinity. If dense volcanic rocks along the southern margins of the Oasis Valley basin and Timber Mountain caldera complex are fractured, these fractures should parallel the fault zones and may act as conduits (aquifers) in the same manner as the TCFZ. Even if subparallel fractures are not present, flow may still be concentrated along the caldera margins because the rocks are highly disrupted or damaged. Water in this flow system may thus be deflected southwestward toward Bailey Hot Springs and Beatty. Southeastward flow across the southern moat of the Rainier Mesa caldera also is a possibility, but water chemistry data (Rose and others, 1999) indicate any such flow does not extend as far as Fortymile Canyon.

CONCLUSIONS

The Pahute Mesa–Oasis Valley area of southern Nevada is geologically and structurally complex because of the processes related to formation of the Basin and Range province, the development of the nested calderas of the southwest Nevada volcanic field, and to later tectonic activity characterized by extension, strike-slip deformation, and detachment faulting. Much of the area is now blanketed with alluvium and thick, widespread volcanic rocks that totally conceal many of the structures that potentially have a major impact on groundwater flow in the area. Despite these complexities, our geophysical investigations, integrated with recent geologic mapping (Fridrich and others, 1999a, 1999b; Slate and others, 1999),

provide new insights on the structural underpinnings of the Pahute Mesa–Oasis Valley area. For example, our study leads to a better definition of known caldera boundaries and suggests the presence of previously undetected calderas. Some geophysically defined structures are likely related to basin-and-range extensional tectonism. The Thirsty Canyon fault zone, for example, seems to represent a series of coalesced ring-fracture systems along an older Basin and Range fault. Among the many, potentially important features characterized, the Thirsty Canyon fault zone provides one of the most direct routes for groundwater flowing from the northwestern part of the Nevada Test Site to reach inhabited areas to the southwest and warrants special attention for monitoring efforts. Continued definition of major structural features will help refine sub-basin boundaries and contribute to developing a better conceptual understanding of groundwater flow in the study area.

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