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mantle peridotite leads to an amplification of the thickness and degree of melting of the LVZ; III, eventually, the melt fraction segregates and drains buoyantly upwards. This upward convection of superheated basalts advances the front of melting of peridotite to 35 km depth, where we have detected it at present. The region of continuous interconnection of melt in the upper mantle is defined by stippling and the predicted deep resistivity profile beneath the Roosevelt Hot Springs at each stage is drawn at the right side of diagram.

ABSTRACT

Magnetotelluric (MT) measurements in southwestern Utah have yielded a model of resistivity structure in this area to a depth of about 100 km. The MT observations are strongly affected by Great Basin graben sedimentary fill, which constitutes conductive upper-crustal lateral inhomogeneity and requires simulation using two- and three-dimensional modeling algorithms before deeper portions of the resistivity section can be resolved. Included in our model is a layer of low resistivity ($20 \Omega\text{-m}$) residing from 35 to 65 km depth. Sensitivity tests of the data to the structure weigh strongly against the top of this layer being as shallow as 25 km and against the conductivity and thickness of the layer being highly correlated. No intra-crustal low-resistivity layer is indicated by the MT data.

The deep, low-resistivity layer in the upper mantle of southwestern Utah can be interpreted to consist of interconnected basaltic partial melt of low melt fraction (2-3%). Interpretation of conductive layers in the upper mantle of rift environments is simplified by the essentially invariant, volatile-deficient nature of most basalt production. The inferred melt fraction is probably a minimum, however; in extensional environments, some melt may reside in fissures or other discrete configurations and not contribute to reducing bulk resistivity.

The low-resistivity layer is a manifestation of the diapiric uprise and melting events concentrated in the eastern Great Basin during Late Cenozoic time. This activity occurs over an interval that extends to much greater depths than does the layer itself. In an environment like this one which has experienced pronounced extension and volcanism, we suggest that the lower crust and upper mantle may be flushed relatively clean of volatile-bearing phases. Some of these volatiles may have been introduced to the middle crust, where their long-term fate remains unclear.

INTRODUCTION

Southwestern Utah, situated on the eastern margin of the Great Basin subdivision of the Basin and Range tectonic province, contains several geothermal resource discoveries (Ward et al., 1978; Chapman et al., 1981; Ross et al., 1982). Our understanding of the near-surface characteristics of individual resource areas has progressed tremendously by the intense application of a wide spectrum of geological, geochemical and geophysical methods (Ward et al., 1981). A coherent framework relating eastern Great Basin geothermal occurrences to their tectonic environment, however, has been slow in developing for two reasons: first, geoscientific data sets sensitive to relevant deep structure are in general contaminated by upper crustal inhomogeneity; and second, any such framework must consider the evolution of neighboring geological regimes in the western United States (Eaton, 1982; and many others).

It is our purpose here to communicate recent contributions to the concept of regional controls on geothermal resources in the eastern Great Basin resulting from deep electrical resistivity investigations. These contributions stem from fundamental advances in techniques for modeling electromagnetic measurements in tectonically complex areas (Wannamaker and Hohmann, 1982; Wannamaker et al., 1982) along with progress by numerous reseachers in linking earth resistivity to physiochemical conditions. The aforesaid advances in

resistivity interpretation have been applied to an extensive collection of magnetotelluric observations in southwestern Utah, whose original purpose was thermal reservoir assessment at the Roosevelt Hot Springs known geothermal resources area (KGRA) (Wannamaker et al., 1980; Ross et al., 1982). In the course of this discussion we will derive a profile of electrical resistivity in southwestern Utah to a depth of about 100 km, propose a sequence of events describing the evolution of deep resistivity structure and physiochemical conditions in the Great Basin and the Colorado Plateau and explain the physical significance of this resistivity profile to geothermal systems in the eastern Great Basin.

TECTONIC SETTING OF THE EASTERN GREAT BASIN AND NEIGHBORING REGIONS

If dynamic interactions between the North American and Farallon (or remnants thereof) lithospheric plates unite the evolutionary histories of all tectonic regimes in the southwestern U. S. (Lipman, 1980; Eaton, 1982), then any discussion of the geological character of the eastern Great Basin must at the very least consider the Great Basin as a whole along with the adjacent Colorado Plateau. Our examination of tectonic setting focuses upon the Miocene-to-present extension that is most evident in the southwestern U. S., since this activity is postulated to dominate resistivity structure in S. W. Utah, but we will include earlier events when deemed relevant.

Lithospheric Deformation and Stress Distribution

Regional lithospheric extension has affected the Great Basin for more than 30 m.y. (Zoback et al., 1981; Eaton, 1982). Before about 20 m.y. ago, the mechanism of extension was upper crustal intrusion of calc-alkaline magmas, derived ultimately from melting events in the upper mantle associated with Farallon plate subduction (op. cit.). While possibly contributing a great deal to regional heat flow, intrusion alone does not thin the crust (Eaton, 1982).

Starting roughly 20 m.y. ago, Basin and Range structure began to develop, but initially with a NNW preferred orientation (Zoback et al., 1981). The majority of Great Basin deformation which defines today's NNE-trending horst-graben morphology took place in the last

Figure 1. Physiographic provinces of the southwestern United States, modified from Stewart (1978). Dashed line separates Great Basin from the southern Basin and Range (Eaton, 1982).

10-12 m.y. (Stewart, 1978, 1980; Zoback et al., 1981). This broadly distributed extension, with attendant basaltic volcanism, indicates stretching not just of the crust but also of the upper mantle, so that crustal attenuation will occur. Crustal thicknesses determined seismically hence can be used as a guide to average extension rates over the last 10-12 m.y.

Regional uplift of about 2 km amplitude (prior to extension) is a further characteristic of the Great Basin, and of the Colorado Plateau as well. The timing of this uplift is less clear than that of the extension, but 25 m.y. ago has been cited (Stewart, 1978; Bodell and Chapman, 1982).

Mapped crustal extensions and regional elevation suggest that the eastern Great Basin, which underlies western Utah, has experienced a rate of normal faulting greater than that of the Great Basin interior perhaps by more than 1% per m.y. (Stewart, 1978; Christiansen and McKee, 1978). While some extension in southwestern Utah may be older than 20 m.y., the bulk of the basin formation in this area seems to have occurred within the last 8-9 m.y. (Rowley et al., 1979).

Rowley et al. (1979) also discuss certain ENE-oriented topography, faults, igneous rocks, aeromagnetic anomalies and mineral deposits in eastern Nevada and southwestern Utah. The age of origin of these features, collectively named the Pioche-Marysvale belt, generally exceeds 17 m.y., and so predates the Basin and Range extension dominating today's morphology. However, Quaternary movement along some of these ENE-trending faults has been observed (ibid.). Thus, S. W. Utah contains the intersection of two major structural features defining an area of relatively enhanced tectonism.

Major normal faulting has not affected the interior of the Colorado Plateau, although some extension appears to a distance of about 60 km inwards from both western and eastern physiographic borders (Thompson and Zoback, 1979). Indeed, WNW deviatoric stresses for the Colorado Plateau seem slightly compressive, in contrast to those of the Great Basin which are tensional (op. cit.). However, older, relatively gentle, long-wavelength (~100 km) Laramide folding, which also trends NNE, is evident in this province (Hintze, 1973; Dickenson and Snyder, 1978). Laramide and Sevier deformation is also present in the eastern Great Basin, but upon it is superimposed the Late Miocene to present extension forming the horst-graben morphology (op. cit.).

Seismology

The Great Basin is a region of ongoing seismicity. The Intermountain Seismic Belt is a zone of particularly high seismicity which, in Utah, forms a roughly NNE-trending band about 150 km wide straddling the physiographic boundary between the eastern Great Basin and the Colorado Plateau (Smith, 1978). Hypocentral depths in the Great Basin appear restricted mainly to less than 15 km, and to less than about 10 km in southwestern Utah, implying that extension at greater depth is characteristic of a ductile deep crust and upper mantle. Seismicity is virtually absent today in the Colorado Plateau interior, in line with the existence of minor WNW compression.

Seismic refraction, body wave and surface wave dispersion surveys show further contrasts between the crust-mantle properties of the Great Basin, its eastern margin and the Colorado Plateau interior. Crustal thicknesses for the Great Basin interior lie between 30 and 35 km, while values for its eastern margin can be less than 25 km (Smith,

1978; Keller et al., 1979; Priestley et al., 1980; see Figure 2a).

Modest crustal thicknesses (~35 km) appear to exist up to 60 km inside the western physiographic border of the Colorado Plateau but increase to near 45 km for its interior. If this value for the Colorado Plateau can be used as a guide, and if most crustal thinning occurred in the last 10-12 m.y., then the crustal thicknesses imply an average rate of extension of nearly 2% per m.y. for the Great Basin interior, but about 5% per m.y. for much of its eastern margin.

P_n velocities (Figure 2b) for the uppermost mantle average 7.8 km/s for the Great Basin interior, fall to as low as 7.4 km/s for its eastern margin, and are about 7.8 km/s for the Colorado Plateau (Smith, 1978; Keller et al., 1979). All these values are less than the 8.1 km/s P_n velocity characteristics of the stable Great Plains to the east. The shear velocity model of Priestley and Brune (1978), which is likely more applicable to the tectonically less active Great Basin interior than to its anomalous eastern margin, shows a higher velocity ($V_s = 4.5$ km/s) mantle lid of nearly 30 km thickness, below which, beginning at a depth of about 65 km, a low velocity zone (LVZ) of $V_s = 4.0$ km/s extends down to about 160 km. Beneath this in the mesosphere, V_s exceeds 4.4 km/s and rises beyond 5.0 km/s below 400 km depth. Models compiled by Thompson and Zoback (1979) suggest that a LVZ below the Colorado Plateau (80 to 160 km depth extent, $V_p^{lid} = 7.9$ km/s, $V_p^{LVZ} = 7.6$ km/s, $V_p^{meso} = 8.2$ km/s) is closer in character to that of the Great Basin (65 to 160 km depth extent, $V_p^{lid} = 7.8$ km/s, $V_p^{LVZ} = 7.6$ km/s, $V_p^{meso} > 8.2$ km/s) than to that of the Great Plains (120 to 160 km depth extent, $V_p^{lid} = 8.1$ km/s, $V_p^{LVZ} = 7.8$ km/s, $V_p^{meso} > 8.2$ km/s).

Figure 2.- Tectonic indicators in the southwestern United States: a) crustal thicknesses, contour values in kilometers. From Smith (1978) and Keller et al. (1979); b) uppermost mantle compressional (P_n) velocities, contour values in km/s. From Smith (1978) and Keller et al. (1979); c) distribution of volcanics less than 5 m.y. old. From Christiansen and McKee (1978), Ward et al. (1978) and Luedke and Smith (1978a, 1978b, 1981); d) surface heat flow contours in heat flow units (1 HFU = 41.6 mWm^{-2}). From Sass et al. (1981) and Bodel and Chapman (1982).

2a

2b

2c

2d

Volcanism

Starting approximately 80 m.y. ago, broad areas of the western U. S. experienced shifting patterns of predominantly calc-alkaline igneous activity associated with accelerated North American-Farallon plate convergence and generally shallow subduction (Lipman, 1980). Triggered perhaps by a reduction in plate convergence rates starting about 40 m.y. ago, a southward-progressing front of this activity from 35 to 17 m.y. ago defined ENE-trending structures in eastern Nevada and western Utah, including the Pioche-Marysville igneous belt (Rowley et al., 1979; Lipman, 1980).

Mafic lavas erupted in southwestern Utah some 21 m.y. ago along with a 17 m.y. old series of basaltic dikes and flows in north-central Nevada mark the transition from calc-alkaline activity to a fundamentally basaltic mode of volcanism in the Great Basin (Lipman, 1980; Best et al., 1980; Zoback et al., 1981). By 12 m.y. ago, roughly contemporaneous with the onset of today's style of horst-graben development, basaltic and bimodal basalt-rhyolite volcanism had spread throughout the Great Basin (Christiansen and McKee, 1978). However, since 10 m.y. ago and especially since 5 m.y. ago, volcanic activity has been excluded almost completely from a growing central region of the Great Basin and concentrated near its margins (Christiansen and McKee, 1978; Eaton, 1980; see Figure 2c). Limited Pliocene basalts also have been observed near Loa, Utah, some 50 km east of the Great Basin-Colorado Plateau physiographic boundary (Best et al., 1980).

The basalts of the Great Basin as a whole are mainly alkali-olivine in composition (Leeman and Rogers, 1970), but those of its eastern margin are quite diverse. Here, basanites, alkali-olivine

basalts, hawaiites and olivine tholeiites are all commonly represented (Best and Brimhall, 1974; Best et al., 1980). Interpretation of the source region of these basalts is made difficult in some cases by evidence of complex fractionation histories (e.g., Lynch and Nash, 1980). In addition, the presence of volatiles (chiefly H_2O but with some CO_2) in small amounts (e.g., < 0.5 wt. % H_2O) is indicated in a number of the most alkalic mafic lavas in the Basin and Range (Boettcher and O'Neil, 1980; Best et al., 1980; Bergman and DuBessy, 1982; Nash et al., 1983) and in the unusual basaltic andesites of Hausel and Nash (1977).

However, the vast majority of basalts in the eastern Great Basin can be interpreted to derive from depths as great as 95 km to as shallow perhaps as 35 km by modest degrees ($< 20\%$) of melting of mantle peridotite under essentially invariant, volatile-deficient conditions (Best and Brimhall, 1974; Yoder, 1976; Wyllie, 1979a). Bailey (1978) argues that water contents of basalts in continental rift zones worldwide are quite low, in the range 0.10 to 1.0 wt.%. We suggest that ongoing extension and volcanism to a large degree flushes the upper mantle clean of volatile phases (see Irving, 1980; and Watson, 1982), and that dry solidus temperatures are being reached over a wide range of depths to produce the variety of basalts in the eastern Great Basin.

Rhyolite extrusives are observed in bimodal association with basaltic volcanism throughout the Great Basin (Smith and Shaw, 1975; Ward et al., 1978; Hildreth, 1981). These rhyolites generally are acknowledged to result through fractionation of less silicic magmas which in turn derive from lower crustal melting events in rocks near

gabbro in composition (Hildreth, 1981; Kay and Kay, 1981). While aided by high regional heat flow, these melting events are local in nature, with the ultimate source of heat being deep crustal intrusion or underplating by basaltic liquids. In light of the strong propensity of H_2O to partition into the remaining melt phase during differentiation, the normally limited water contents of the least evolved rhyolites in extensional environments ($\leq 2-3$ wt. %; Bailey, 1978; Hildreth, 1981), and especially in the eastern Great Basin (≤ 1 wt. %; Crecraft et al., 1981; Nash, 1982), suggest that temperatures well in excess of $1000^\circ C$ resulting in complete disappearance of any hydrous phases (mainly amphibole) often are achieved locally during melting. Fusion of lower crustal mafic rocks in the presence of hornblende but in the absence of water-rich vapor would occur just below $1000^\circ C$ (perhaps higher for fluorine-bearing amphibole; Nash, 1982) and yield about 0.6 vol. % per vol. % hornblende (1 vol. % hornblende contains ~ 0.02 wt. % H_2O) of a granodioritic melt possessing 3-4 wt. % water (Wyllie, 1977; Burnham, 1979a; 1979b).

Absence of water-rich vapor is favored for the lower crust at least of the eastern Great Basin and Rio Grande Rift due to the limited occurrence of hydrous minerals here implied by xenolith studies (Kay and Kay, 1981). Mafic rocks at these depths can be converted upon addition of water almost completely into compositionally equivalent amphibolite containing about 2 wt. % H_2O with no free vapor phase remaining at temperatures below those of vapor-absent melting (Burnham, 1979b). On the other hand, some CO_2 of a mantle source is introduced routinely to the deep crust (Bailey, 1978; Ehrenberg and Griffin, 1979; Roden et al., 1979; Wyllie, 1979b,

1980; Newton et al., 1980; Hildreth, 1981). At temperatures below about 700°C at 8 kb, this CO₂ will be absorbed in carbonate phases with no discrete free vapor remaining (Bailey, 1978; Wyllie, 1979a, 1979b; Jacobs and Kerrick, 1981). Alternately, if a lower crust bearing amphibole and carbonate is subjected to temperatures beyond 700°C, a CO₂-rich vapor phase (mole fraction of H₂O in the vapor, $X_{H_2O}^v \lesssim 0.3$) may evolve through joint decomposition of carbonate and amphibole and exist at temperatures below those of melting (ibid).

In any event, a free water-rich phase seems much more tenable in the rather felsic compositions of the middle to upper crust (granite to granodiorite at depths less than 15-20 km; Burnham, 1979b; Kay and Kay, 1981.) Water-bearing siliceous melts, especially the relatively hydrous granodioritic liquids defined above, may upon ascent from the mafic deep crust in sufficient amounts introduce such a free phase upon crystallization. This is because middle and upper crustal rocks of the aforesaid composition can accommodate only about 0.5 wt. % H₂O in hydrous phases (Wyllie, 1977; Burnham, 1979a, 1979b). The key to this process is the chemical contrast between the deep and the middle and upper crustal sections. A purging by magmatism of deep crustal volatiles as is especially evident in the eastern Great Basin (Nash, 1982) is interpreted by Kay and Kay (1981) to be a general feature of rift environments. It is tempting in this context to consider the seemingly more hydrated lower crust of the Colorado Plateau (op. cit.) as a standard by which to judge the water content of the lower crust at least of the eastern Great Basin prior to Late Cenozoic extension and volcanism.

Heat Flow

Values of heat flow in the Great Basin are variable, as shown in Figure 2d. A conspicuous low of $q_0 < 1.5$ heat flow units (HFU) in central Nevada, the Eureka Low, may be due to regional hydrologic recharge (Lachenbruch and Sass, 1978; J. Sass, 1982, pers. comm.). Heat flow averages about 2.2 HFU along the eastern margin of the Great Basin to perhaps 50 km inside the Colorado Plateau physiographic boundary (Bodell and Chapman, 1982). Southwestern Utah is no exception; in fact, regional heat flow here may reach 2.4 HFU (Chapman et al., 1981). Upward mass transfer by solid-state lithospheric extension and/or by crustal intrusion and underplating by basaltic magma is the root cause of the heat flow variation from the Great Basin interior through its eastern margin to the Colorado Plateau (Lachenbruch and Sass, 1978).

Bodell and Chapman (1982) show that the Colorado Plateau has a cool thermal interior with heat flux less than 1.5 HFU. This observation is not consistent with evidence that P_n velocity, the LVZ and deep resistivity structure (Porath, 1971, to be discussed shortly) below the Colorado Plateau do not differ greatly from those of the Great Basin interior (Thompson and Zoback, 1979). Since the lithosphere of the Colorado Plateau is apparently stable, surface heat flow must be by conduction only. Hence, if the regional uplift occurring 25 m.y. ago is the result of a fundamental thermal amplification below the lithosphere, then the present geotherm and surface heat flux are still largely transient (Bodell and Chapman, 1982). The ultimate causes of such amplification are the subject of much speculation; Bailey (1978) and McGetchin et al. (1980) discuss possible modes.

Gravity and Aeromagnetics

Smoothed free-air gravity data suggest that the Great Basin and the Colorado Plateau are approximately in isostatic equilibrium (Keller et al., 1979). A density model incorporating this gravity data with seismic and deep resistivity interpretations (Porath, 1971) was computed by Thompson and Zoback (1979). Their model is consistent with a crust thickening from less than 30 km to greater than 40 km eastward from the Great Basin to the Colorado Plateau over a distance of about 60 km.

Mabey et al. (1978) described the general features of Great Basin aeromagnetic observations, in particular noting ENE-trending anomalies in eastern Nevada and western Utah correlated with Mid-Cenozoic igneous activity and mineralization including the Pioche-Marysvale belt. Shuey et al. (1973) inferred a 15 km increase in the Curie depth over the Colorado Plateau relative to the Great Basin some 50 to 100 km east of the physiographic boundary, in agreement with seismicity and heat flow.

RELATION OF RESISTIVITY TO TECTONIC SETTING

The calibration of resistivity with controlling variables such as temperature, pressure and bulk composition has been approached by many investigators. However, there are some aspects of the relation of resistivity to the physical state of earth materials that have not been addressed adequately in the literature, especially for the deep crust and upper mantle. For this reason, we present our perception of the controls on resistivity structure by tectonic setting.

Upper and Middle Crustal Resistivity Mechanisms

The bulk resistivity ρ_r of upper crustal materials composed of silicates, either unconsolidated sediments or well-indurated rocks, is controlled by fluid content through Archie's Law (Grant and West, 1965, p. 390-393; Brace, 1971), which is

$$\rho_r = A \rho_e \phi^{-m}, \quad (1)$$

where ρ_e is the resistivity of the pore-filling electrolyte, including contributions from mineral surface conduction, and ϕ^{-m} is the formation factor comprised of the porosity ϕ , ranging from zero to one, and the tortuosity coefficient m . If porosity is dominated by intergranular passageways, the coefficient m is about two, while if porosity is due mainly to planar fractures or rod-like conduits, m is near unity. On the basis of laboratory experiments, Brace (1971) concluded that upper crustal materials for the greater part have

values of m near two. The factor A is also a function of pore geometry. For the special cases of randomly oriented planar fractures, A takes the value $1\frac{1}{2}$, but in general it is near one (Grant and West, 1965; Brace, 1971).

Since Brace (1971) considered pore waters that were quite fresh ($\rho_e \approx 50 \Omega\text{-m}$), he concluded that the effect of pressure on reducing porosity could increase bulk resistivities in crystalline rocks by a factor of 30 to perhaps $10^5 \Omega\text{-m}$ over the depth interval 0-10 km. However, the ability of pore fluids to carry solute increases with temperature and pressure, so that mid-crustal (in the depth range 5-20 km) resistivities for the greater part should remain below $10^4 \Omega\text{-m}$ (Olhoeft, 1981; Shankland and Ander, 1982).

Jiracek et al. (1982) have proposed that mid-crustal volatile contents, and thus bulk conductivity (inverse of resistivity), are amplified in rift environments by a trapping of volatiles beneath the brittle to ductile crustal transition (Eaton, 1982). In our earlier discussion of volcanism, we suggested a mechanism for introduction of free water to higher levels of the crust through crystallization of H_2O -laden melts. Alternately, in high heat flow environments, a flux of vapor from the upper mantle that is predominantly CO_2 may carry water up from the lower crust by lowering the activity of water beneath that required to stabilize amphibole at the ambient temperatures (Newton et al., 1980). Although CO_2 reduces the ionic strength of the fluid phase (Kay and Kay, 1981), and thus increases ρ_e , if m in Archie's Law exceeds one then bulk resistivity of a water-bearing rock is decreased if CO_2 vapors appear and increase porosity. However, a water-rich (vs. a CO_2 -rich) volatile phase appears to be the one

introduced at depth in regions like the western U.S. which have experienced subduction of hydrated oceanic lithosphere (ibid.).

Deep Crustal Resistivity Mechanisms

Earlier, we were unable to deny categorically the evolution of a free, CO_2 -rich vapor in the deep crust of regions of high heat flow (prograde metamorphism). In shield environments, on the other hand, deep crustal temperatures appear sufficiently low ($\leq 700^\circ\text{C}$ at 8 kb) that both CO_2 and H_2O are structurally bound, leaving only a remnant distribution of volatile molecules along grain boundaries with $P_{\text{fluid}} = P_{\text{H}_2\text{O}} + P_{\text{CO}_2} < P_{\text{total}}$ (Winkler, 1979; Wyllie, 1979b; Jacobs and Kerrick, 1981). This seems most consistent with the studies of Richter and Simmons (1977) and Padovani et al. (1982) which showed vanishingly small porosities in samples of deep crustal amphibolite and granulite from shield terrains. A stable volatile phase can exist at these lower temperatures only if the crust is depleted in the silicate reactants needed to produce the volatile-bearing minerals. If a highly conductive grain boundary phase becomes unimportant in lowering bulk resistivity at great depth, two mechanisms remain. For depths beyond 15 km in the Great Basin, crustal resistivity would be controlled by semi-conduction through minerals and by ionic conduction in partial melts.

Solid-State Resistivity. The electrical conductivity of minerals exhibits an Arrhenius dependence on temperature (T), which is

$$\sigma(T) = \sum_{i=1}^n \sigma_{0i} \exp(-E_i/kT) \quad , \quad (2)$$

where n different conductivity mechanisms may individually dominate (T) through a specific temperature range, the proportionality

constant σ_{0i} of the i th mechanism is essentially independent of temperature, E_i is the activation energy of the i th mechanism and k is Boltzmann's constant (Hamilton, 1965).

Even within a single rock or mineral type, there is great deal of variation in conductivity at a specific temperature (Olhoeft, 1979; Kariya and Shankland, 1983). The majority of this variation appears to stem from scatter in σ_{0i} , whereas E_i tends to show more uniformity (op. cit., also Schmucker, 1970). No systematic difference in resistivity at a specific temperature exists between anhydrous silicates and the hydrous mineral amphibole (Olhoeft, 1979, 1981). Furthermore, the effect on crustal mineral conductivities of uncertainties in the oxidation state (f_{O_2}) of the lower crust is likely to be small compared to the considerable scatter in present conductivity measurements, and so is disregarded for now (Olhoeft, 1981).

If one can ignore volatiles (i.e., water), then we believe that a combination of the gabbro and granite conductivity measurements of Kariya and Shankland (1983) provides the most reliable bounds with which to interpret deep crustal resistivities in the Great Basin. As an example, their data indicate that an increase in deep crustal temperature from 750 to 900°C would cause a decrease in resistivity by a factor of 3 to 5.

It is suggested, however, that the finite values of P_{H_2O} existing throughout the crust, even in the crust of shields where P_{fluid} normally does not reach P_{total} , could amplify solid-state conductivities. By a mechanism similar to that of Burnham (1979a) for silicate melts, water may enter feldspar crystals and increase the concentration of mobile alkali cations for electrical conduction (Yund

and Tullis, 1981; Stolper, 1982). The equilibrium population of hydrous defects in the crystals increases with P_{H_2O} , though the effect of temperature alone seems less clear. The presence of intracrystalline hydrous defects has been shown already by Duba and Heard (1980, and pers. com.) to magnify solid-state conductivity in olivine. Spear and Silverstone (1983) also discuss the dissolution of water in quartz. A further unknown effect on conductivity which may be important in extensional regimes is that of strain rate, which is known to speed metamorphic reactions (Winkler, 1979).

Resistivity of Partial Melts.- If semiconduction in the crystalline matrix is unimportant compared to ionic conduction in an interconnected melt phase (Tickle, 1967), then Archie's Law will govern the bulk resistivity of partially fused crustal rocks. By imposing $P_{H_2O} = P_{total}$, Lebedev and Khitarov (1964) demonstrated a strong dependence of melt fraction and resistivity on water content. However, in light of the model of dissolution of H_2O by Burnham (1975, 1979a), we suggest that mole fraction of water in the melt ($X_{H_2O}^m$), rather than P_{H_2O} , and temperature are the decisive intensive variables in melt phase conductivity. All else fixed, P_{total} should have a relatively minor effect on melt conductivity in the deep crust, although total pressure affects bulk resistivity through its control on the solidus temperature and hence melt fraction for a specific composition.

Furthermore, we have pointed out already that silicic melts at least in the eastern Great Basin form in a water-deficient environment. A melt of 3-4 wt. % H_2O at 950°C, such as would appear at a depth of 25 km in mafic rocks with amphibole present (Burnham,

1979a, 1979b), would have a resistivity of about $1 \Omega\text{-m}$ (Burnham, 1975), but bulk rock resistivity depends on the volume of melt (related to volume of amphibole) and the tortuosity coefficient (unknown). Products of silicic volcanism at the surface, as pointed out earlier, also indicate melting events often occur with local temperatures well in excess of those needed to stabilize amphibole in the presence of melt, making $X_{\text{H}_2\text{O}}^{\text{m}}$ uncertain.

Upper Mantle Resistivity Mechanisms

Semiconduction in minerals and ionic conduction in partial melts control the bulk resistivity of the upper mantle as well as that of the deep crust. While uncertainties in upper mantle solid-state mechanisms remain, the interpretation of partial melting seems fairly clear.

Solid-State Resistivity.- Olivine is the dominant mineral in the upper mantle (Wyllie, 1979a) and thus has been proposed to control bulk resistivity in the solid state (Duba et al., 1974). Careful measurements by Duba and Heard (1980) with f_{O_2} controlled to lie between the magnetite-wustite and wustite-iron stability boundaries (Arculus and Delano, 1981; Eggler, 1982) yielded resistivities higher by about a factor of 3 at a specific temperature than those obtained by Duba et al. (1974) at the quartz-fayalite-magnetite buffer. At mantle temperatures, pure olivine resistivities are in general rather large; for example, approximately $1000 \Omega\text{-m}$ at 1350°C .

However, upper mantle temperatures derived by relating laboratory data on single crystals of olivine to model resistivity structure based on surface electromagnetic measurements seem consistently higher than those provided by other methods such as surface heat flow

observations or petrological geothermometers (Shankland, 1981). We believe this problem, and indeed a similar one for the deep crust (Shankland and Ander, 1982), arises largely from biases in the interpreted resistivity structure, which in turn arise from an often inadequate regard for the effects of upper crustal lateral inhomogeneities (Wannamaker et al., 1982). This matter will be pursued shortly.

Nevertheless, the difficulty in estimating upper mantle temperatures using single-crystal data cannot be dismissed completely in this manner; measurements of the resistivity of bulk peridotite at subsolidus temperatures show variable but uniformly lower values than those of pure olivine (Shankland, 1981). Proposed explanations of this discrepancy include enhanced charge-carrier activities within olivine in contact with other mantle minerals (ibid.). Also, Duba and Shankland (1982) have introduced graphite as a possible means of reducing upper mantle resistivity. Arculus and Delano (1981) propose that graphite would be the stable carbonaceous phase in the upper mantle of S.E. Arizona, but the studies of Egger (1982) in Wyoming and Bergman and DuBessy (1982) in south-central Nevada suggest that graphite is not widespread in the upper mantle of the western U.S. Even if the existence of this mineral was likely, it must form a thoroughly interconnected phase to reduce bulk resistivity. The model of deep resistivity that we develop is best explained without invoking graphite.

Resistivity of Partial Melts.— As was the case for crustal rocks, the correlation of the bulk resistivity of partially molten peridotite with upper mantle temperatures requires knowledge of the

degree of melting of peridotite as a function of T , P_{total} and composition, of the dependence of the melt phase conductivity on these same three variables, and of the tortuosity of the melt in its crystalline matrix. In contrast to crustal materials, there are firmer constraints on these three subjects for mantle peridotite.

The degree of melting of peridotite as a function of temperature and pressure has been studied by Mysen and Kushiro (1977) and Scarfe et al. (1978). Their samples 66SAL-1, 66PAL-3 and PHN1611 in Figure 3 all show somewhat different solidus temperatures at 20 kbar pressure in the absence of water. However, 66PAL-3 with its solidus temperature near 1400°C seems the most likely representative of pristine upper mantle material (Mysen and Kushiro, 1977; Wyllie, 1979a; Harrison, 1981). Nevertheless, Figure 3 shows the melting behavior of dry upper mantle peridotite to be nearly invariant (Yoder, 1976), with all samples generating 25 wt. % or more melt within a few tens of degrees of their solidi through the eutectic-like melting interval A.

Also given in Figure 3 is the melting curve of PHN1611 in the presence of 1.9 wt. % H_2O (Mysen and Kushiro, 1977). Note that only about 6 wt. % melt is generated at temperatures significantly below the dry solidus ($\sim 1475^{\circ}\text{C}$ for PHN1611), just enough to dissolve completely the available water (op. cit.). At higher melt fractions, the fusion curve resembles that of the dry specimen, being displaced only to slightly lower temperatures and exhibiting essentially invariant behavior when all crystalline phases are present. Given the nature of this melting, and Burnham's (1979a, 1979b) work, we suggest that fusion at subsolidus temperatures in the presence of water, for

Figure 3. Degree of melting of three peridotite specimens as a function of temperature at 20 kbar pressure. Successive melting intervals A, B and C are identified corresponding to successive depletion of mineralogical phases: A, olivine + orthopyroxene + clinopyroxene + spinel + liquid; B, olivine + orthopyroxene + liquid; C, olivine + liquid. Hachured boxes indicate experimental uncertainties in temperature and melt fraction measurements. From Mysen and Kushiro (1977) and Scarfe et al. (1978).

both basaltic and granitic liquids, proceeds until all water is dissolved in the melt phase, beyond which further melt production requires temperatures approaching the dry solidus (Wyllie, 1971).

Amounts of water smaller than that considered in Figure 3 have correspondingly smaller effects on melting relationships in mantle peridotite (Yoder, 1976; Wyllie, 1979b). Therefore, the melt fraction determinations of Ringwood (1975, p. 156), which show up to 15 wt. % fusion at temperatures below the dry solidus in the presence of 0.1 wt. % H_2O , are believed to be in error (also see Yoder, 1976; Wyllie, 1979a; Mysen et al., 1982). The results of Ringwood (1975), however, continue to be used exclusively in the interpretation of low resistivity layers in the upper mantle (for example, Adam, 1980; Oldenburg, 1981; Shankland, 1981).

The degree of melting of peridotite as a function of temperature in the presence of carbon dioxide differs negligibly from that of volatile-free peridotite at depths less than about 75 km due to the low solubility of CO_2 in silicate melts here (Wendlandt and Mysen, 1980; Wyllie, 1980). At depths around 100 km, melt volumes exceeding 40 wt. % can be produced at temperatures below the volatile-free solidus but, to do this, several weight percent of CO_2 must be present (Wendlandt and Mysen, 1980).

As stated previously, the Late Cenozoic basalts of the eastern Great Basin are interpreted to derive from modest degrees of invariant melting of mantle peridotite in the presence of little or no volatiles (Best and Brimhall, 1974; Best et al., 1980). In fact, such conditions are evident for the great majority of basalts observed worldwide (Yoder, 1976; Bailey, 1978; Wyllie, 1979b). This

observation, coupled with the results of Figure 3, simplifies tremendously the interpretation of low-resistivity layers in the upper mantle assumed to indicate partial melting; without incurring errors of more than a few tens of degrees, dry peridotite solidus temperatures as a function of pressure can be assigned throughout such layers. This simple interpretational approach requires no knowledge of the intrinsic resistivity of the melt phase or of the tortuosity of the melt in its crystalline matrix. However, it is probably not valid for interpreting any weakly-developed, low-resistivity layering in the upper mantle beneath truly stable lithosphere, since such layering may reflect traces of melt induced by volatiles at temperatures well below the dry solidus (Wyllie, 1979a).

The interpretation of resistivity structure due to partial melting in the upper mantle can be refined by considering the intrinsic resistivity of the liquid phase and the mathematical model of the interconnection of melt in its crystalline matrix. Waff and Weill (1975), Rai and Manghnani (1978) and Waff (1980a) show that resistivities of dry basalt liquids depend strongly upon temperature but weakly upon composition, P_{total} and f_{O_2} . We anticipate the effect of dissolved water in reducing basaltic liquid resistivities will be subordinate to that in granitic magmas (Burnham, 1979b; Waff, 1980a), but it remains to be substantiated. Given the mechanism of solubility of CO_2 in basalts (Burnham, 1979b; Mysen et al., 1982), the effect of this volatile on melt resistivities should be small if any.

Until the melt phase becomes highly interconnected, its contribution to reducing bulk resistivity is negligible (Waff, 1974). Thus, low-resistivity zones corresponding to partial melts

will have abrupt upper and lower bounds, i.e., they are very much layer-like. Under conditions of uniform stress in the upper mantle, randomly oriented straight rods of conductive melt in a resistive crystalline matrix constitute an accurate solid-liquid mixing relation (Waff and Bulau, 1979; Vaughn and Kohlstedt, 1980a, 1980b; Watson, 1982). Hence, Archie's Law of equation (1) with $A = 3$ and $m = 1$ equates bulk resistivity to percent melt phase using our knowledge of melt phase resistivity (Grant and West, 1965). However, in extensional environments like the Great Basin, the preceding Archie's Law may provide only a minimum estimate of melt fraction since significant amounts of liquid might accumulate in discrete fissures and not contribute to lowering bulk peridotite resistivity (Spera, 1980; Irving, 1980).

PREVIOUS INVESTIGATIONS OF RESISTIVITY IN THE EASTERN GREAT BASIN AND NEIGHBOURING REGIONS

Techniques that map deep resistivity structure have been neither as widely applied nor as fruitful in illuminating tectonic processes in the Great Basin and the Colorado Plateau as have geoscientific methods described in a preceding section. Nevertheless, some unique indications of physiochemical conditions at depth, as well as stern warnings about the effects of upper crustal lateral inhomogeneities upon deep resistivity interpretations, have been provided by early geoelectric surveys.

Controlled Source Surveys

Experiments using galvanic resistivity methods in the Great Basin and the Colorado Plateau by the U. S. Geological Survey have been summarized by Keller (1971). As Keller stated, the techniques used at the time were largely experimental and developmental, so that only "broad-brush" conclusions about the deep resistivity structure of the western U. S. were possible.

In 1977, multifrequency dipole-dipole (3 and 6 km dipoles, separations to $n = 6$) galvanic resistivity measurements were made in southwestern Utah by the Department of Geology and Geophysics, University of Utah, in an attempt to describe any resistivity structure associated with the Pioche-Marysvale trend (Petrick et al., 1981). The data, from a profile extending 33 km northward from the

Roosevelt Hot Springs, were dominated to an unexpectedly severe degree by upper crustal, 3D lateral inhomogeneities, especially graben sedimentary fill. These inhomogeneities lead to a lack of fit of layered inverse calculations to observed data, as well as dubious 1D and 2D earth models (W. R. Petrick and A. C. Tripp, unpub.). An approximate 3D inversion of these results (Petrick et al., 1981) presents a concentration of low-resistivity structure within 2 km of the surface, evidence again of variable basin fill dominating the observations.

Lienert and Bennett (1977) and Lienert (1979) using an interstate power transmission system as an essentially infinite line current source, advanced 1-D models of resistivity in western Nevada to depths of 40 to 50 km. Included in their interpretation was a 10-15 km thick layer of low resistivity (10-30 Ω -m) whose depth to top varied from 20 to 30 km from north to south in their survey area. However, these authors did not evaluate quantitatively the effects of current-gathering in the 3D sedimentary basins common in this area of the western Great Basin. Such current-gathering can cause widespread distortions of both electric and magnetic field components to arbitrarily low frequencies even for ungrounded artificial sources (Ward, 1983).

Utilizing the same interstate power system as Lienert and Bennett, Towle (1980) has studied deep resistivity in eastern California with direct current magnetic field variations. A return current concentration along the westernmost Great Basin is permitted by the data, but so are several alternate return paths (ibid.).

Past Magnetotelluric Measurements

Early scalar MT soundings were recorded in southern Utah and northern Arizona by Plouff (1966) and summarized by Keller (1971). In general, scalar MT data are unreliable due to changing source fields and lateral inhomogeneities in the earth. In particular, the scatter in the data of Plouff (1966) was very great, resulting in deep resistivity profiles with unacceptably wide parameter variances.

Pedersen and Hermance (1981) have interpreted an MT sounding in the eastern Colorado Plateau near Farmington, New Mexico. However, this sounding is located in a high heat flow area ($q > 2$ HFU) which is probably related to the Rio Grande Rift province (Sass et al., 1981; Bodell and Chapman, 1982), and so is not relevant to consideration of the cool Colorado Plateau interior of southeastern Utah.

Thirty-five tensor MT soundings were obtained in 1977 in southwestern Utah by the Department of Geology and Geophysics, University of Utah. These observations, arranged in two N-S oriented profiles, were also aimed in conjunction with the aforesaid multifrequency dipole-dipole measurements at delineating any resistivity structure associated with the Pioche-Marysvale trend. However, just as occurred for the galvanic readings, the upper crustal horst and graben structure here in the eastern Great Basin obscured deeper structures from study by N-S profiles of MT data (J. A. Stodt and P. E. Wannamaker, unpub.).

Geomagnetic Deep Sounding

Pioneering measurements by Schmucker (1970) established geomagnetic anomalies indicating conductive upwellings beneath the Rio Grande Rift and the western Great Basin. Concerning the latter area,

Schmucker places the conductive mass in the upper mantle, while Lienert (1979) advocates a deep crustal origin.

Much of what we know about the deep electrical resistivity of the Great Basin and the Colorado Plateau is the result of geomagnetic temporal variations observed by Reitzel et al. (1970). Three components of magnetic field were recorded at periods from 20 to 200 minutes along four E-W profiles about 150 km apart, with station spacings averaging about 120 km. Porath et al. (1970) and Porath and Gough (1971) showed that anomalies in vertical and E-W horizontal fields were of internal origin (see Figure 4), with the external fields varying smoothly over the recording array.

From these estimates of normalized anomalous fields, Porath (1971) presented a preferred, two-dimensional resistivity model, also shown in Figure 4, with a 2 Ω -m layer of varying thickness from the Great Basin to the Great Plains, and with superimposed conductive ridges beneath the Wasatch Fault Belt and the Southern Rocky Mountains. This model was correlated with seismic LVZ estimates. Gough (1974, 1981) has reaffirmed the model of Porath (1971) and proposed that the conductive ridges in the upper mantle beneath western Utah and the southern Rocky Mountains reflect zones of partial melting. The secondary field amplitudes diminish to the south, and so presumably does the anomalous resistivity structure, but the features in Figure 4 appear to continue into northwesternmost Arizona.

Figure 4.- Observed normalized anomalous horizontal E-W (Y_{ia}/Y_n) and vertical (Z_{ia}/Z_n) magnetic variation fields at a period of 60 minutes in the southwestern United States (Porath et al., 1970) along with the two-dimensional resistivity cross-section preferred by Porath (1971) and Gough (1974, 1981) to explain the observations. The profile for which the cross-section is defined is drawn with dashes on the maps of observed fields.

In-Phase Y_{ia} / Y_n

In-Phase Z_{ia} / Y_n

MAGNETOTELLURIC OBSERVATIONS AT THE ROOSEVELT HOT SPRINGS KGRA

The magnetotelluric method has been widely applied in the exploration of geothermal systems, sedimentary basins and the deep crust and upper mantle (Swift, 1967; Word et al., 1971; Vozoff, 1972; Larsen, 1975; Jupp and Vozoff, 1976; Stanley et al., 1977; Ward et al., 1981). While recent advances in instrumentation and data processing (Gamble et al., 1979; Weinstock and Overton, 1981; Stodt, 1983) enable accurate measurements of tensor MT responses, models of deep resistivity derived from MT responses are commonly suspect due to an unsatisfactory treatment of upper crustal lateral inhomogeneities (Wannamaker et al., 1982).

From 1976 to 1978, ninety-three tensor magnetotelluric stations were recorded in the vicinity of the Roosevelt Hot Springs KGRA in an attempt to delineate resistivity structure associated with this geothermal system (Wannamaker et al., 1980; Ross et al., 1982; see Figure 5). The complexly 3D nature of the MT responses directly over the thermal anomaly area prevented a rigorous quantification of any resistivity structure associated with an economic hot brine reservoir or a mid-crustal magmatic heat source for the geothermal system (ibid.). However, soundings along line B-B' in Figure 5 at distance from the thermal anomaly area, carefully modeled to remove the effects of upper crustal lateral inhomogeneities, have yielded a regional resistivity profile for S.W. Utah to depths of about 100 km.

The Problem of Upper Crustal Lateral Inhomogeneities

The Milford Valley immediately to the west of the thermal anomaly area, which is enclosed by the solid contour in Figure 5, is typical of sediment-filled grabens in the Great Basin (Stewart, 1978, 1980; Ward et al., 1978; Eaton, 1982). Serious problems with interpretation of MT soundings near such electrically conductive sediments were first documented by Swift (1967) in his Arizona and New Mexico studies. A detailed examination of the MT responses associated with graben alluvial fill has been performed by Wannamaker et al. (1982), who concluded that such responses are fundamentally three-dimensional in nature and that indiscriminate use of 1D or 2D modeling algorithms will generally incur serious errors.

Wannamaker et al. (1982) demonstrated in particular that two-dimensional transverse electric (TE) modeling algorithms are inappropriate for interpreting apparent resistivities, impedance phases or tipper identified as TE in the Great Basin. The 2D TE mode involves no boundary charges and hence no current-gathering, whereas in the Great Basin, all models of MT responses due to upper crustal inhomogeneities are dominated by current-gathering (ibid.).

If MT sounding data identified as TE in the presence of conductive inhomogeneities are used in deriving 1D models of deep resistivity structure, as we perceive has been overwhelmingly the case in the literature, then such models will suffer systematically a compression of layer thicknesses and a downward bias in layer resistivities (Wannamaker et al., 1982). While concurring with Shankland and Ander (1982) that a firm relation should exist between surface heat flow and low resistivities at depth, we maintain that a relation

Figure 5.- Magnetotelluric site location map for the Roosevelt Hot Springs thermal area. Rock outcrop bounding the Milford Valley sediments is lightly stippled while the solid curve is the 400 mWm^{-2} thermal contour (Ward et al., 1978). Station numbers referred to in text are prefixed according to the year in which they were occupied.

● GEOTRONICS CORP.
(1976)

○ GEOTRONICS CORP.
(1978)

■ UNIV. OF TEXAS
AT AUSTIN (1977)

▼ UNIV. OF CALIFORNIA
AT BERKELEY (1979)

MT./AMT. STATIONS
ROOSEVELT HOT SPRINGS KGRA, UTAH

between heat flow and conductive inhomogeneities also exists at least in extensional environments (also see Eaton, 1982). It is suggested that lateral inhomogeneities can help explain the distinction drawn by Shankland and Ander regarding tectonic and shield regimes in their equating of deep resistivity to deep temperatures. We note that their interpretational electrical conductivity curve for shields approaches the dry gabbro conductivity field of Kariya and Shankland (1983) for temperatures exceeding 700°C.

Fortunately, accurate resistivity cross-sections through horst-graben morphology in the Great Basin can be obtained by selective application of a 2D transverse magnetic (TM) modeling algorithm (Wannamaker et al., 1982). Boundary charges are included in both 3D and 2D TM formulations, allowing a proper treatment of current-gathering effects upon apparent resistivity and impedance phase.

In defining data for transverse magnetic modeling at the Roosevelt Hot Springs, we have employed a coordinate system based on tipper-strike (Vozoff, 1972). Since tipper responses are band-limited in frequency (Wannamaker et al., 1982), we may choose an optimal frequency range for defining tipper-strike to minimize the contributions of secondary inhomogeneities much smaller than the Milford Valley sediments with which we are primarily concerned. Tipper responses for the Milford Valley peak at frequencies near 1. Hz with values of about 0.7. This behavior in the tipper is very similar to the computer simulation of Wannamaker et al. (1982).

Tipper-strike estimates near 1. Hz for stations along line B-B' in Figure 5 are oriented consistently around N15°E, a trend which closely parallels the Great Basin normal faulting bounding the Milford

Valley sediments (Ward et al., 1978). Hence, we impose a fixed x-axis orientation of N15°E to define the TM mode apparent resistivity ρ_{yx} and impedance phase ϕ_{yx} for all MT sites on line B-B'.

Observed Apparent Resistivity and Impedance Phase Pseudosections

The observed quantities ρ_{yx} and ϕ_{yx} of line B-B' defined in the aforesaid manner have been assembled in Figure 6 into pseudosections, whereby log frequency constitutes the ordinate and horizontal distance is assigned to the abscissa for contour plots of the apparent resistivity and impedance phase. The measured results for this line exhibit a virtually classic horst-graben response for the TM mode and bear a close resemblance to the computations of Wannamaker et al. (1982). Apparent resistivities for stations 78-5 to 78-8 of Figure 6 over the conductive graben sediments comprising the Milford Valley are monotonically decreasing functions as frequency decreases, falling from somewhat over 6 Ω -m at 100 Hz to below 0.1 Ω -m at very low frequencies for station 78-7. Values of impedance phase ϕ_{yx} remain above 45° over the basin with a maximum exceeding 70° at low frequencies.

Substantial gradients in ρ_{yx} in Figure 6 at mid to low frequencies between stations 78-4 and 78-5 and between 78-8 and 78-10 are responses due to the range front faulting bounding the graben sediments. Gradients in ϕ_{yx} at the same stations at mid to high frequencies are corresponding responses to the same faulting. High apparent resistivities in Figure 6, often exceeding 1000 Ω -m, are observed for soundings east of and including 77-13 in the Mineral Mountains, with a maximum in the contour values reached at mid frequencies ($\sim$ 0.3 Hz). Impedance phase contours over the mountains

Figure 6. Observed pseudosections of ρ_{yx} and ϕ_{yx} , identified as transverse magnetic, along line B-B' of Figure 5. Contours of ρ_{yx} in $\Omega\text{-m}$ and of ϕ_{yx} in degrees.

undergo a minimum below 20° at mid to upper frequencies (about 3 Hz) and a maximum of almost 70° at lower frequencies (~ 0.01 Hz).

Examination of the observations just over the Mineral Mountains in Figure 6 will show that ρ_{yx} at a specific frequency varies from station to station by up to an order of magnitude, with the higher values generally occurring near the center of the mountains (e.g., site 78-14). This site-to-site variation in fact changes rather little as frequency changes; specifically, the shapes of the apparent resistivity soundings on a $\log \rho_{yx}$ vs $\log$ frequency plot for stations in the mountains are all very similar but the overall levels of the soundings differ. These differences, which occur also for the other principal apparent resistivity ρ_{xy} , stem from boundary charge effects (current-gathering) in small-scale, near-surface resistivity inhomogeneities (Wannamaker et al., 1982). Such inhomogeneities, which constitute geological noise, correlate with complicated fluctuations in overburden thickness, rock porosity, and pore water content and chemistry in the upper 500 m. Because of the effects of near-surface geological noise, and of the graben sedimentary fill comprising the Milford Valley, indiscriminate application of one-dimensional inversion algorithms within the Mineral Mountains would yield confusing and erroneous models of deep crust and upper mantle resistivity structure (Wannamaker et al., 1982).

The effect of small-scale geological noise on ϕ_{yx} in Figure 6, and for that matter on ϕ_{xy} , for the soundings over the Mineral Mountains is limited to frequencies above 30 Hz. One might consider the response at station 77-13 between 1. and 10 Hz to be an exception but, as will be demonstrated in our upcoming detailed simulation, such

behavior is due to the conductive Milford Valley sediments immediately to the west. The impedance phase response in Figure 6, particularly over the Mineral Mountains at frequencies below 1. Hz, is in fact remarkably smooth.

The character of the impedance phase affirms the superiority of imposing a uniform data coordinate system based upon selective tipper-strike and known geological trends for all soundings in a specific profile (Wannamaker et al., 1982). If the principal directions of the impedance tensor, which show wide scatter from station to station even at low frequencies due to small scale geological noise, had been used to define the pseudosections of Figure 6 for TM mode modeling, substantial fluctuation in the impedance phase beyond whatever statistical scatter appears in the data would have resulted from site to site in the mountains to frequencies well below 1. Hz.

Such fluctuation in the phase would stem from a non-uniform coordinate direction in conjunction with the widespread anisotropy in impedance phase induced throughout the Mineral Mountains by the graben sedimentary fill of the Milford Valley. One of the two principal axes of this anisotropy lies close to the uniform coordinate orientation of N15°E we have imposed, enabling recovery of the impedance phase response of the Milford Valley and of the regional resistivity structure that is undistorted by near-surface geological noise (Wannamaker et al., 1982). Given the high apparent resistivities evident in Figure 6 over the Mineral Mountains, station-to-station variations in impedance phase due to small-scale resistivity inhomogeneities are forbidden to occur at frequencies below about 30 Hz by fundamental EM scaling concepts (Wannamaker et al., 1982), so

that any variations at lower frequencies introduced by an inappropriate definition of coordinate axes are not physically realizable for 2D geometries and cannot be simulated by 2D algorithms (in particular for us, 2D TM routines).

Calculated Pseudosections and Model Resistivity Cross-Section

Construction of Model Cross-Section.- While automated inversion schemes for one-dimensional resistivity models are commonplace, inverse algorithms for 2D resistivity geometries are rare (Jupp and Vozoff, 1977) and can only compute optimum media resistivities for fixed media geometries. We rely instead upon trial-and-error matching of results calculated from an assumed resistivity cross-section with the observed MT data of Figure 6 using a versatile and accurate 2D finite element forward program (Rijo, 1977; Stodt, 1978; Nutter, 1981). While this method is relatively slow and inefficient, it provides model cross-sections that are at least as reliable as those from automated inverse routines.

Any resistivity cross-section derived using our finite element routine consists of a variety of adjoining, straight-sided polygons each with a uniform interior resistivity. Clearly, there is some question of the validity of this type of model in the real world. Variations in pore water content and chemistry, hydrothermal alteration and temperature of earth materials are likely to be smooth functions of position. However, other structures such as fault-bounded masses of conductive alluvium or zones of partial melting at least in the mantle are expected to have sharp boundaries. We have presumed, in instances where the resistivity boundaries in the earth are gradational, that the flat-sided computed section whose response

fits the observations is, in some sense, close to the true structure. Larsen (1981) already has pointed out that, in one-dimensional environments, layered models provide local averages of earth resistivity when the true structure varies continuously.

Non-uniqueness in multidimensional modeling can be alleviated by a high station density in each MT profile and by a wide spectrum of data at each station. In particular concerning the latter point, data at each sounding should extend to sufficiently high frequencies that both principal apparent resistivities ρ_{xy} and ρ_{yx} , as well as impedance phases ϕ_{xy} and ϕ_{yx} , have converged to common values above some frequency. For this high frequency range, the earth will likely be effectively one-dimensional, allowing us to constrain the near-surface model resistivities close to their true values. Such isotropic behavior occurs for frequencies above 3 Hz for the valley stations and above 30 Hz for the mountain sites.

Calculated Apparent Resistivity and Impedance Phase Pseudo-sections.- The computed pseudosections of ρ_{yx} and ϕ_{yx} for line B-B' appear in Figure 7. Note that the computed and observed pseudosections are virtually identical; the observations have been fit to within data scatter almost everywhere. Examples of data scatter typical of sites over the Mineral Mountains will be forthcoming in a discussion of parameter resolution.

Model Resistivity Cross-Section.- The model of deep resistivity structure we have produced by application of our 2-D TM finite element algorithm to the observations of Figure 6 can be divided into two major parts: first, a shallow portion detailing just the extensive upper crustal lateral inhomogeneities in the upper 2 km; and second, a

deep, purely layered portion extending to about 100 km which is determined by the physiochemical conditions (temperature, pressure and bulk composition) of the middle to lower crust and upper mantle.

a.) Upper Crustal Lateral Inhomogeneities.- The upper 3 km of the finite element cross-section we have derived is shown in Figure 8. Within the Milford Valley, units within 140 m of the surface having resistivities from 3.5 to 18 Ω -m correspond to surficial clays, sands and gravels which are partially or completely water-saturated. Looking a bit deeper here, resistivities as low nearly as 1 Ω -m are caused by Pleistocene Lake Bonneville clays (Hintze, 1973, 1980), which we show residing to a maximum depth near 700 m below station 78-7. Deeper still exists a substantial thickness (> 1 km) of modest resistivity material (25 Ω -m) related to pre-Bonneville alluvium and volcanics (Rowley et al., 1979). This last unit is poorly resolved.

The resistivity media defining the Milford Valley are bounded on their eastern and western margins by steeply dipping interfaces, which we presume correspond to major normal faulting developed during the Late Cenozoic extension of this area. Dips of these interfaces are not well resolved, especially on the west side, but a steep dip for the eastern boundary faulting is favored on this profile given the abrupt lateral gradients in ρ_{yx} and ϕ_{yx} between stations 78-9 and 78-10 in Figure 6. The initial guess in our modeling process of this upper crustal resistivity section was guided by the refraction seismic model of Gertson and Smith (1979) and the gravity observations of Carter and Cook (1978), but our final section in Figure 8 shows important differences.

Over the Mineral Mountains, our model in Figure 8 shows

Figure 8. Best-fit, 2-D TM finite element resistivity cross-section fitting the observations of line B-B' in Figure 5. Resistivity values of individual media are in Ω -m and the vertical exaggeration is 6:1.

resistivities in hundreds of ohm-m above 2 km depth. These relate to crystalline rocks with pore fluid volumes of perhaps a few percent. Over the depth interval of 2 to 3 km, the resistivity of the cross-section in Figure 8 increases to 3000 Ω -m. This increase represents a decrease in porosity of the well-indurated basement rocks to well under 1%, probably due for the greater part to a decrease in fracturing below 2 to 3 km depth (Yusas and Bruhn, 1979).

b.) Deep Resistivity Profile.- The 1400 Ω -m medium in Figure 8 begins the deep layered resistivity profile introduced previously. This entire profile is shown in Figure 9 along with some alternate 1-D models considered in forthcoming parameter resolution tests. The best-fit model shows a maximum resistivity of 3000 Ω -m over the depth interval of 3 to 11 km. Subsequently, resistivity falls to 200 Ω -m by 35 km depth, whereupon an abrupt drop to 20 Ω -m is encountered. This low-resistivity material defines a deep, thick layer bottomed at 65 km depth by a sudden increase to a 200 Ω -m basal half-space.

The fact that this deeper part of our model is purely layered is not due to poor resolution on the part of the measurements of Figure 6; the data quality for all soundings in line B-B' is good to excellent and yet the data are fit within scatter by imposing a purely 1D section for depths greater than 2 km. Furthermore, through application of a 2D TM program, Wannamaker et al. (1980) showed that the data of line C-C' in Figure 5 also are fit using the preferred deep profile of Figure 9.

Establishing the physical significance of this deep resistivity profile is the major purpose of this paper. We defer addressing this purpose to a future section, where the matter is treated in detail.

Parameter Resolution Study.- The finite element cross-sections of Figures 8 and 9 are rather complicated. Strictly speaking, every vertex of every polygon in these cross-sections must be considered a parameter of the resistivity model, yet we do not pretend that every such parameter is well-resolved. Nevertheless, we maintain that a complex earth model is required to simulate the more subtle features of an ensemble of high-quality MT soundings, and that our resistivity cross-sections are substantially accurate representations of the natural resistivity structure. In any event, our tests of the sensitivity of the data to the resistivity structure will not dwell upon minute details of the finite element section but instead will involve families of polygons which we believe constitute geologically important media in the subsurface.

To begin the resolution study, we refer to Figure 10 where observed data points of both modes of apparent resistivity, ρ_{xy} and ρ_{yx} , and of impedance phase, ϕ_{xy} and ϕ_{yx} , are plotted for station 76-13 of line B-B'. Apparent resistivity anisotropy reaching an order of magnitude is observed to the lowest frequencies at which we have reliable data (~ 0.002 Hz). Impedance phase anisotropy is also evident, being as high as 25° at 3. Hz, but diminishing to near zero by about 0.01 Hz. Such anisotropy is widespread throughout the Mineral Mountains and is caused by the conductive graben sedimentary fill of the Milford Valley to the west (see also Wannamaker et al., 1982). We have already pointed out the care necessary in defining MT coordinate axes in the face of such widespread anisotropy and near-surface geological noise.

In Figure 10 are plotted also several computed sounding curves.

Figure 10.- Observed data ρ_{xy} , ρ_{yx} , ϕ_{xy} and ϕ_{yx} for sounding 76-13 of line B-B'. Solid curve represents goodness-of-fit of computed to observed results using the best-fit resistivity cross-sections of Figures 8 and 9. Curves of dots and shorter dashes relate to correspondingly drawn alternate resistivity profiles in Figure 9 used in parameter resolution tests. Curves of long dashes appearing between the two modes of data points represent the 1D forward computation using the best-fit regional profile in Figure 9.

First, consider the solid curves passing through the observed data points of ρ_{yx} and ϕ_{yx} . These curves display the calculated response of our best-fit finite element model of Figures 8 and 9 at station 76-13, a response which seems very agreeable with the trends of the observed data points. In addition, curves of long dashes lying intermediate to the principal apparent resistivity and impedance phase observations are shown in Figure 10. The response of the best-fit one-dimensional deep resistivity profile of Figure 9 in the absence of all upper crustal lateral inhomogeneities is represented by these latter curves (referred to subsequently as ρ_{1D} and ϕ_{1D}). The great difference between the solid curves and the curves of long dashes illustrates again the importance of the Milford Valley in determining the MT signatures in this area. We add that the placement of the observed principal apparent resistivities and impedance phases with respect to the one-dimensional profile response varies from station to station due to small-scale, near-surface lateral inhomogeneities, as discussed earlier. The inclusion of lateral inhomogeneities in models of deep resistivity structure is of paramount importance in tectonically disturbed regions like the Great Basin, and supercedes other considerations of the appropriateness of resistivity models vis-a-vis continuous vs. layered one-dimensional structures.

We turn now to resolution of detail in the deep resistivity profile of Figure 9. It is noted at the outset that, while lateral inhomogeneities i.e., the Milford Valley, have induced anisotropy to a high degree over the Mineral Mountains, both tensor and 1D signatures in Figure 10 bear a certain mutual resemblance. To start with, the increase in model resistivity from 2 to 3 km depth accounts for the

general rise in all apparent resistivities and for the minimum in the impedance phases between 10 and 1. Hz. Continuing below 0.3 Hz, the decrease in model resistivity at depths in excess of 11 km leads to decreasing apparent resistivities along with impedance phases greater than 45° . The factor of ten drop in model resistivity at 35 km depth is responsible for the particularly steep gradients in the apparent resistivities and the values of impedance phases about 70° around 0.03 Hz. Finally, without the abrupt rise in model resistivity at 65 km depth, the apparent resistivity data would not flatten out nor would the impedance phase data fall as strongly or rapidly as they do in Figure 10 at the lowest frequencies, especially below 0.003 Hz. We presume a maximum depth of sensitivity of the data to the structure of about 120 km, as imposition of low-resistivity material here causes computed phases at 0.002 Hz which exceed those of the best-fit model by about four degrees and which cannot be made to agree with the observations by increasing the resistivity of the 200 Ω -m material.

At frequencies less than 0.01 Hz in Figure 10, both ϕ_{xy} and ϕ_{yx} have converged to common values while ρ_{xy} and ρ_{yx} have shapes that are essentially identical except that they are offset by a factor of ten. Eventually, a convergence of this sort occurs as frequency falls for all stations in the Roosevelt Hot Springs area. This frequency dependence of the tensor sounding curves particularly illustrates that a one-dimensional, regional resistivity profile containing local lateral variations in structure, chiefly the Milford Valley sediments (see Wannamaker et al., 1982), is a viable model for the resistivity makeup here in S.W. Utah. The deep resistivity profile in Figure 10 of course must have some horizontal limits, but the aforesaid

convergence in the observations of Figure 9 indicates that such lateral bounds should be far-removed in any direction from the Roosevelt Hot Springs, possibly on the order of 100 km. The geological significance of these bounds is the subject of an upcoming section.

Continuing in our parameter resolution studies, we test a deep resistivity profile that is of different shape than the best-fit model, namely a profile whose log resistivity decreases linearly with depth (plotted with dots and labelled the "Steady Decrease" model in Figure 9). The calculated response in ρ_{yx} and ϕ_{yx} at station 76-13 using this test profile is drawn also with dots in Figure 10. With the Steady Decrease layering, computed apparent resistivities are substantially higher than observed over most of the frequency range. In addition, calculated impedance phases using this alternate profile are lower than the measured ones down to 0.01 Hz, but cross and asymptote to about 75° at the lowest frequencies. This significant disagreement between the observed and calculated values of ρ_{yx} and ϕ_{yx} leads us to reject the Steady Decrease model as a candidate for deep resistivity structure in southwestern Utah.

As a second test, let us examine the overall scale of the deep layering. By shrinking the depths and resistivities of individual layers by 30% such that layer contrasts and conductivity thickness products remain constant (Madden, 1971), we arrive at the sequence drawn with alternating dots and dashes in Figure 9 and called the "Shallow" regional model. Likewise, the responses in ρ_{yx} and ϕ_{yx} with this alternate are drawn with dots and dashes in Figure 10. A preference for the best-fit profile over the Shallow case cannot be

formed on the basis of ϕ_{yx} alone since the difference in computed impedance phases for the two 1D models is smaller than the data scatter. The fit appears significantly worse for the apparent resistivity, however, and on this basis we disqualify the Shallow test model as a possibility.

The last parameter resolution examination we make deals solely with the deep low-resistivity layer from 35 to 65 km. We maintain that this 20 Ω -m medium is not a "thin" layer (Madden, 1971), and to prove so it is replaced by a unit of equivalent conductivity-thickness product of 10 Ω -m resistivity and 15 km thickness (see Figure 9). The calculations using this "Thin Layer" model, shown with short dashes in Figure 10, differ from the measurements to a rather modest degree. The test case apparent resistivity decreases slightly from the best-fit response ρ_{yx} in the frequency interval 0.1 to 0.003 Hz, but the two calculated signatures start to converge again at the lowest frequencies. On the other hand, the computed impedance phase of the Thin Layer alternate exceeds the observations and the best-fit response in ϕ_{yx} by about four degrees near 0.02 Hz. Given that ϕ_{yx} averages near 70° at this frequency for all sites over the Mineral Mountains, the Thin Layer profile is abandoned as a replacement for the best-fit structure.

The foregoing studies exemplify systematics in the resolution of parameters of one-dimensional earths. The apparent resistivity data are primarily useful for establishing overall amplitudes and depths of resistivity structure. The impedance phase data, on the other hand, are especially good for resolving the general shape and rates of change with depth of the resistivity profile. We believe these

conclusions are largely valid for 2D and 3D earths as well.

Before we close this particular study, there is one more demonstration of the hazards of 1D interpretation of MT data in regions of extension like the Great Basin. In Figure 11 are reproduced the best-fit deep profile of Figure 9 as well as model profiles computed through direct 1D inversion of the tensor MT observations of Figure 10 (we used the routine of Petrick et al., 1977). The display of these observations with the computed 1D responses of the aforesaid model profiles in Figure 12 shows the goodness of fit possible by a 1D inversion algorithm to data taken in 3D environments. The deep resistivity models derived thereby are unacceptable, however; their departure from the preferred deep profile exceeds bounds that were already rejected in the rigorous parameter resolution studies of Figures 9 and 10.

Figure 11. Best-fit regional resistivity profile for the Roosevelt Hot Springs thermal area compared to layered models obtained through 1D inversion of ρ_{xy} and ϕ_{xy} (dotted lines) and ρ_{yx} and ϕ_{yx} (dashed lines).

Figure 12. Observed data points of ρ_{xy} , ρ_{yx} , ϕ_{xy} and ϕ_{yx} for sounding 76-13 showing the goodness of fit that can be obtained to 3D data using a 1D inverse routine. However, the resultant layered models were dismissed as unrealistic. Solid curve appearing between data points is the 1D forward response of the best-fit regional profile.

GEOTHERMAL SIGNIFICANCE OF DEEP RESISTIVITY STRUCTURE IN SOUTHWESTERN UTAH

We are not interested in resistivity for its own sake but rather for what this property can tell us about physical processes in the earth. Hence, this final section begins by relating our best-fit, deep resistivity profile in Figure 10 to the physical state of the crust and upper mantle in this area. Ultimately, comprehension of the present state will require consideration of the tectonic evolution of the eastern Great Basin during Late Cenozoic time. The statements that follow draw heavily upon our foregoing discussions of the tectonic setting of the eastern Great Basin and of the relation of resistivity to subsurface physiochemical conditions.

Deep Resistivity Profile Beneath the Roosevelt Hot Springs KGRA

Our best-fit, crust and upper mantle resistivity profile is reproduced in Figure 13 along with estimates of bulk resistivity vs. depth for relevant earth materials based upon laboratory experiments and an assumed conductive geotherm. Aqueous electrolytic conduction in rock pores and fractures, solid-state semiconduction in minerals and ionic conduction in partial melts all have domains of dominance comprising the entire deep resistivity section. The laboratory-derived resistivities must of course be considered as abstractions; given the oversimplifications upon which they were founded, especially the use of a conductive heat flow model in the actively extending

Figure 13.- Best-fit, crust and upper mantle resistivity profile for southwestern Utah compared to physical model incorporating aqueous electrolytic conduction in rock pores, solid-state semiconduction in minerals and ionic conduction in partial melts, in conjunction with conductive geotherm assuming regional heat flow of 2.4 HFU (Lachenbruch and Sass, 1978; Chapman et al., 1981). Dotted crustal bounds use the aqueous electrolytic conduction measurements of Brace (1971) and the gabbro-plus-granite, solid-state data of Kariya and Shankland (1983). The line of dashes represents buffered olivine measurements of Duba and Heard (1980, and pers. comm.). The curve of dashes and dots corresponds to bulk resistivity of partially melted peridotite using the melt fraction determinations of Mysen and Kushiro (1977), Scarfe et al. (1978) and Wyllie (1979a), the melt phase interconnection model of Waff and Bulau (1979) and the melt phase conductivity measurements on alkali olivine basalts of Rai and Manghnani (1978). Individual numbers on partial melting curve indicate percentage of liquid phase.

eastern Great Basin, departures by our best-fit resistivity profile are inevitable. Such departures in fact are wholly desirable, since in their explanation lies the means for refining and strengthening our concepts of the physical state at depth in southwestern Utah.

As explained with Figure 8, resistivities of our best-fit deep profile increase rather sharply from 600 to 3000 $\Omega\text{-m}$ over the depth interval 2 to 3 km. This high resistivity persists to depths exceeding 10 km. Bulk resistivity in these upper levels of the crust is controlled by aqueous electrolytic conduction in rock pores; for comparison with laboratory experiments, we have borrowed from Brace (1971) to construct the upper 10 to 15 km portion of the dotted curves in Figure 13. We note that Brace's resistivities have generally higher values and increase more gradually than do those of our best-fit model. However, in view of the significance of fracture porosity in the upper 2-3 km (Gertson and Smith, 1979; Yusas and Bruhn, 1979) and the uncertainties in electrolyte chemistry (Olhoeft, 1981; Shankland and Ander, 1982), we are not concerned by this disagreement. It is sufficient to conclude in light of Archie's Law that porosities at depths of several kilometers in this area are well under one percent.

From 11 to 35 km depth in Figure 13, our best-fit resistivity profile decreases in a steady manner from 3000 to 200 $\Omega\text{-m}$. We have modeled this decline with layers due to the nature of our finite element algorithm, but we believe the actual variation with depth to be smooth. The dotted curves in this diagram representing laboratory investigations also decrease with depth below about 15 km. These curves were constructed from the compilation by Kariya and Shankland

(1982) of semiconduction measurements on dry gabbroic and granitic rocks in conjunction with a conductive geotherm assuming a regional heat flow of 2.4 HFU (Lachenbruch and Sass, 1978; Chapman et al., 1981). The depth to the base of the crust in this area is about 28 km (see Figure 2), so it is here at which the dotted curves end.

Although the resistivities derived from the laboratory work and the MT observations exhibit an approximate mutual resemblance, the finite element model shows somewhat lower values and a more gradual decrease of resistivity from the middle to the lower crust. Applying the results of Kariya and Shankland (1983), the resistivity model suggests that temperatures at depth (especially around 15-20 km) exceed those produced by our conductive heat flow model. However, a reduction of mid crustal resistivities by volatiles, either by pore conduction or through a solid-state mechanism, cannot be ruled out.

The MT observations do not support the notion that temperatures at depth in the crust are high enough to create any widespread, layer-like, electrically conductive zone of interconnected silicic melt. Given the depletion of lower crustal water in the eastern Great Basin, we believe temperatures exceeding 1000°C would be necessary to do this.

At depths greater than about 28 km, below the Moho in this area, resistivity mechanisms in mantle rocks must be considered. For comparison with our best-fit profile, the inclined curve of dashes in Figure 13 displays a hypothetical resistivity due to pure olivine with f_{O_2} between the magnetite-wustite iron-wustite buffers (Duba and Heard, 1980, and pers. comm.) and with temperature defined by our conductive geotherm. Resistivities of pure olivine with this geotherm

decrease rapidly with depth, falling below 20 Ω -m by 45 km, but possess magnitudes in the depth interval 28 to 35 km that are much greater than the 200 Ω -m appearing here in our best-fit deep profile. It may well be that pure olivine is not a representative composition for determining bulk solid-state resistivities here (Shankland, 1981); on the other hand, a very resistive layer occupying this narrow depth interval may be difficult to resolve with surface MT measurements (Madden, 1971).

Much more disagreeable to us, however, is the discrepancy between shapes of the hypothetical solid-state olivine curve and our best-fit resistivity profile. In an earth with temperatures that increase steadily with depth, such as is the case with our conductive geotherm or even with any of the temperature profiles of Lachenbruch and Sass (1978) which incorporate convective components, the Arrhenius solid-state temperature dependence for any rock results in bulk resistivities which decrease monotonically with depth. Indeed, the fact that activation energies for semiconduction in earth materials also increase with temperature (Duba et al., 1974; Olhoeft, 1979) means that bulk resistivity curves determined by solid-state mechanisms and steadily increasing geotherms do not flatten out with depth. The dashed curve in Figure 11 representing pure olivine resistivity is an example of this. In our discussion of resolution of parameters, a deep resistivity profile that decreased steadily with depth was rejected as a candidate to explain the low-frequency MT observations at the Roosevelt Hot Springs; the order-of-magnitude rise of resistivity in the upper mantle in our best-fit deep model is a strict requirement. Hence, unless substantial variations in bulk chemical

composition with depth exist in the upper mantle of southwestern Utah, a solid-state resistivity mechanism alone is not sufficient to produce a deep resistivity structure resembling our best-fit profile.

We therefore look to partial melting as the cause of the 20 Ω -m deep layer. If our conductive geotherm is imposed, the curve of dots and dashes in Figure 13 arises as a conceptual variation of resistivity with depth due to an interconnected basaltic melt in upper mantle peridotite. Given the presumably unimportant effects of volatiles upon Late Cenozoic basalt production in the eastern Great Basin, we derive this curve using the dry peridotite solidus presented by Wyllie (1979a), the nearly invariant nature of basalt generation demonstrated by Mysen and Kushiro (1977) and Scarfe et al. (1978), the measurements of electrical conductivity of alkali-olivine basalt of Rai and Manghnani (1978) and the melt phase interconnection model of Waff and Bulau (1979). Individual numbers on the partial melting curve indicate percent liquid phase.

Our MT measurements as exemplified in Figure 9 are too scattered to resolve any detail within the 20 Ω -m layer, although some could exist in nature. In Figure 13, the value of 20 Ω -m on the laboratory-based partial melt resistivity curve corresponds to a melt fraction of about 3%, but this is specific to the geotherm used. If the conductive temperature profile employed had been somewhat less steep, so that it intersected the solidus at a greater depth and temperature than shown, 20 Ω -m on the partial melting curve would arise from a melt fraction less than 3%. We simply conclude that the interconnected melt fraction in our deep, low resistivity layer is rather small, on the order of 2-3%. We do not discount the

possibility that in extensional environments like the eastern Great Basin an unknown portion of the melt may reside in discrete vertical fissures to which our measurements are not sensitive.

On the basis of the MT measurements alone, we cannot exclude easily the possibility that the top of our model low-resistivity layer is as shallow as 28 km, i.e., is at the base of the crust in this area, although we expressed disfavor earlier with the top being at only 25 km. In any event, impressing dry peridotite solidus temperatures at the Moho would lead to rampant melting in the crust above. This would surely be detectable by our MT soundings, but is not.

We have not defined the laboratory-derived curve in Figure 13 beyond melt fractions of about 30% since further melting in most peridotite will not be nearly invariant. Even so, continuation of the conductive geotherm would yield yet greater melt fractions and lower bulk resistivities. Such extreme melt fractions alone condemn the purely conductive geotherm as inaccurate to arbitrarily great depth. Furthermore, the 20 Ω -m material in our best-fit deep profile is a discrete layer, and not an infinitely extending basal half-space. Supported by the approximately invariant, water-deficient character of Late Cenozoic basalt production in the eastern Great Basin, we conclude from the MT interpretation that the dry peridotite solidus is a representative P-T trajectory for the depth interval 35 to 65 km of the 20 Ω -m layer below S.W. Utah. We argue shortly that this trajectory may be valid to much greater depths in the region.

Physical State at Depth in the Eastern Great Basin and Neighbouring Regions

Our interpretation of resistivity structure at the Roosevelt Hot

Springs has provided a point sampling of physiochemical conditions at depth in southwestern Utah. It is noted at the outset that the depth interval of interconnected basaltic melt corresponding to the 20 Ω -m layer in Figure 13 is significantly shallower than the seismic low-velocity zones of either the Great Basin interior or the Colorado Plateau, which are also presumed to reflect partial melting in the upper mantle (Wyllie, 1979a). These apparently disparate geological environments can be reconciled, with help from geomagnetic deep sounding anomalies and the concepts outlined in our opening discussion of regional tectonic setting, in a model for the evolution of the eastern Great Basin during Late Cenozoic time.

Diapirism and Melting.- Of paramount importance in fitting the deep physical state we have inferred in S.W. Utah to those of neighbouring regions is the much-discussed concentration of extensional tectonism throughout the eastern Great Basin during the last 8-9 m.y. Regardless of the underlying causes of this deformation, upwelling of deep crust and upper mantle material must go hand-in-hand with lateral stretching to preserve volume (Lachenbruch and Sass, 1978). If an extension rate of 5% per m.y. as deduced earlier from crustal thicknesses is appropriate for the eastern Great Basin, then a one-dimensional estimate of upward velocity v of material at depth (related linearly to extension rate s and depth z through $v = sz$, *ibid.*) ranges from about 1 1/2 km/m.y. at 35 km through 5 km/m.y. at 100 km to over 7 km/m.y. at 160 km.

While contributing a great deal to surface heat flow and temperatures within the lithosphere (Lachenbruch and Sass, 1978), extension rates tell something about the thermal state at greater

depths as well. Consider for simplicity a spherical diapir in the upper mantle of radius 50 km, a representative E-W length scale for the region of enhanced tectonism defining the eastern Great Basin. For the thermal state of a rising diapir to be largely adiabatic (Oxburgh, 1980), its ascent time must be less than its thermal conduction time constant (Spera, 1980). If our hypothetical spherical diapir has been rising with a velocity $v = 5 \text{ km/m.y.}$ from a depth of 100 km or so during the last 8 m.y., then a distance of roughly 40 km would have been covered. In covering this distance, the diapir ascent rate merely must exceed about 0.40 km/m.y. (Spera, 1980), much less than v , for the aforesaid thermal inequality to be met. We conclude that large volumes rising here in the upper mantle tend to retain much of their sensible heat during transport.

As a mass at depth rises adiabatically, its temperature trajectory may intersect the peridotite solidus so that melting will proceed (Yoder, 1976). Upon crossing the solidus, the temperature within the ascending diapir is buffered by the melting behavior of peridotite (Oxburgh, 1980). In particular, the great size of the latent heat of fusion relative to the heat capacity of peridotite along with the nearly invariant nature of basalt production through melting interval A of Figure 3 serve to keep the temperature gradient within the diapir well within 1°C/km of the solidus during adiabatic ascent (Carmichael et al., 1974, p. 143-147; Yoder, 1976; Oxburgh, 1980; Turcotte, 1982).

However, the degree of melting is inevitably subdued through heat loss to the earth's surface from the top of the diapir aided possibly by dissipation of the internal heat of the diapir through convection

of the interconnected melt phase in its porous crystalline matrix (Spera, 1980). A restrained degree of melting is compatible with the eutectic-like compositions of Late Cenozoic basalts in the eastern Great Basin and with the apparent lack of a systematic progression with time of the major element chemistry of these basalts (Best and Brimhall, 1974; Best et al., 1980).

Evolution of Resistivity Structure in the Eastern Great Basin.- With this brief discussion of diapirism, we advance a schematic and highly simplified model for the Late Cenozoic evolution of deep resistivity in the eastern Great Basin. The model is depicted in Figure 14 in three stages: I, by about 10 m.y. ago, extensional activity responsible for the present horst-graben morphology had spread throughout the Great Basin. Attendant upwelling of mantle peridotite had resulted in widely distributed, fundamentally basaltic volcanism, presumably involving the mechanism of decompression just described, along with a layer of partial melt of low melt fraction (~5%) which did not differ greatly from that defining the present seismic low-velocity zone of central Nevada; II, by 8 m.y. ago, the well-documented concentration of extension in the eastern Great Basin was underway. Accelerated diapiric uprise of mantle material leads to a further amplification of the thickness and degree of melting of the LVZ. The extent of the amplification is difficult to gauge, as it depends on the degree to which the geotherm just above and just below the former LVZ was superadiabatic (Oxburgh, 1980); III, the melt fraction in the magnified zone of fusion can build up only to a point, after which much of it segregates and drains buoyantly upward. This convection of basalts, which in effect are superheated with respect to

Figure 14.- Schematic, highly simplified, three-stage model of evolution of resistivity structure in the eastern Great Basin: I, by 12 to 14 m.y. ago, extensional activity had spread throughout the Great Basin resulting in a layer of partial melting similar to that defining the present low velocity zone of central Nevada; II, by about 8 m.y. ago, the well-documented concentration of extension in the eastern Great Basin was underway. Diapiric uprise of mantle peridotite leads to an amplification of the thickness and degree of melting of the LVZ; III, eventually, the melt fraction segregates and drains buoyantly upwards. This upward convection of superheated basalts advances the front of melting of peridotite to 35 km depth, where we have detected it at present. The region of continuous interconnection of melt in the upper mantle is defined by stippling and the predicted deep resistivity profile beneath the Roosevelt Hot Springs at each stage is drawn at the right side of diagram.

0 100 200 300
SCALE

VERT. EXAG.
1.5:1

ρ_{1-D} at RHS

the environment into which they rise, advances the front of melting to 35 km depth and helps establish the realm of interconnected melt corresponding to the 20 Ω -m layer of Figure 13 by warming the host peridotite during passage.

The lifespan of Stage II in Figure 14 is hard to predict; it might even occur more than once. The degree of melting possible before basaltic liquid segregates and leaves a partially molten source region may depend upon the density contrast between liquid and solid, the magnitude of deviatoric stresses, the rheology of the crystalline matrix, the size of the source region and perhaps other poorly-understood means of stabilizing the melt phase in its peridotite host (e.g., Waff, 1980b; Spera, 1980; Stolper et al., 1981). The duration of Stage II is thus the interval of time during which this critical degree of melting is reached, which in turn is controlled by extension rate and non-adiabatic heat loss. One may speculate a value of 5 m.y. for this interval in the eastern Great Basin based on peaks in bimodal volcanic activity here (Best et al., 1980).

Why doesn't a low liquid fraction, say 2-3%, remain interconnected in the source peridotite at depths below 65 km in S.W. Utah, even if the bulk of the melt segregates and leaves? We should point out that the profile of Figure 13 is only resolvable to depths a bit over 100 km, so that there is no direct evidence from the MT soundings that interconnected melt is absent at deeper levels. Indeed, the preferred model of Porath (1971) in Figure 4 shows 2 Ω -m material extending uniformly to 190 km depth in southwestern Utah. Nevertheless, the vertical resolving power of geomagnetic deep sounding data is quite limited (cf. Porath and Gough, 1971) and an unpublished 2D TE

mode finite element simulation by us shows that a qualitatively good fit to the observations of Figure 4 can be achieved using the resistivity geometries of Figures 13 and 14 and allowing the Great Basin LVZ to have resistivity of 3 Ω -m while the Colorado Plateau LVZ is higher at 5 Ω -m.

We suggest that the zone of interconnected melt below the Roosevelt Hot Springs lies where it is because the very much reduced rigidity of the crystalline peridotite matrix at greater depths (Darot and Gueguen, 1981) promotes both the formation of vertical fissures under tensile stress (Spera, 1980) and the ductile collapse of the crystalline matrix as the melt leaves to enter the fissures (Stolper et al., 1981; Turcotte, 1982). Tholeiitic lavas, frequently represented in the eastern Great Basin, have been interpreted to originate within the range of our 20 Ω -m layer (Best and Brimhall, 1974; Wyllie, 1979a; Best et al., 1980). However, their occurrence at the surface may be out of proportion to their rate of production at depth, since the relatively low density of these magmas favors their ascent over that of other mafic melts (Kushiro, 1980; Stolper and Walker, 1980). It is presumed that basalt production above 65 km is sufficiently restrained to prevent destruction of the widespread interconnection of melt implied by the resistivity structure.

Thermal State at Depth in S.W. Utah.- In Figure 13, a conductive geotherm based upon a regional heat flow of 2.4 HFU for S.W. Utah predicted fusion of dry peridotite at essentially the same depth as the top of our model low-resistivity layer. However, with the high extension rates proposed for the eastern Great Basin, the foregoing coincidence should not be (Lachenbruch and Sass, 1978). A purely

solid-state stretching rate of 5% per m.y. is consistent with a depth of melting of nearly 40 km, but a surface heat flux reaching 3 HFU is to be expected (ibid.).

One would conclude on this basis that the crust and upper mantle of southwestern Utah, and probably the entire eastern Great Basin, have not reached a steady thermal state. The onset of extension is fundamentally an increase of heat flux into the lithosphere, and adjustment of surface heat flow to this form of thermal disturbance may take longer than the 8 m.y. inferred as the interval of concentrated extension in the eastern Great Basin (Lachenbruch and Sass, 1978; Rowley et al., 1979). This possibility would explain our resistivity model without invoking volatiles.

In a previous section, we assigned a temperature trajectory through our deep low-resistivity layer that was coincident with the dry peridotite solidus (nominally that of Wyllie, 1979a). Based upon the same laboratory experiments and field occurrences of basalts in S.W. Utah, such a thermal profile seems appropriate to depths near 100 km. In fact, if the abstraction of adiabatic diapiric uprise in the eastern Great Basin is favored, a dry solidus trajectory may be definitive to a depth exceeding 160 km.

Application of Evolutionary Model to Other Regions.- We believe processes similar to that depicted in Figure 14 for the eastern Great Basin are operating in other areas of the southwestern United States. For instance, the attenuated crustal thicknesses shown in northwestern Nevada in Figure 2a, the pronounced extension near Yerrington interpreted by Proffett (1977) and the concentration of rifting, heat flow and volcanism in the Long Valley and Coso areas

(Bailey et al., 1976; Duffield et al., 1980) indicate that the western margin of the Great Basin of western and southwestern Nevada and eastern California is experiencing an enhanced degree of extension relative to the Great Basin interior, although the timing of this deformation seems less uniform than that of the eastern Great Basin (op. cit.). The similarity in tectonic styles between the eastern and western margins of the Great Basin further suggests that a deep electrical structure corresponding to Stage III of Figure 14 is the likeliest for the western Great Basin and that an amplified low velocity zone with accompanying low resistivity as per Stage II is improbable in this region. Deep resistivity investigations by Schmucker (1970), Lienert (1979) and Towle (1980) permit this conclusion, but do not uniquely favor it.

Another tectonic environment which probably fits into the framework of Figure 14 is the Rio Grande Rift and related regions. Deep electrical structure of the Rio Grande Rift has been reviewed by Jiracek et al. (1979, 1982) and Hermance (1982), and from their work we conclude that this area most closely resembles Stage III of our evolutionary model. Of particular interest to us, however, is the continuation of this rift zone northward into Colorado. As described by Williams (1982), extensional deformation and volcanism diminishes in this direction although some Late Tertiary mafic volcanism persists into Wyoming. In light of the modest degree of extension, we propose that the continuation of the Rio Grande Rift into Colorado may be a good place to search for deep electrical structure representing Stage II in Figure 14. The northward increase in geomagnetic deep sounding anomalies in this area (see Figure 4) is furthermore suggestive of

this notion, but we believe that magnetotelluric profiling with a proper multidimensional interpretation is needed to assess accurately the deep resistivity section here.

Controls on Geothermal Resources in Southwestern Utah

Mapped crustal extensions, crustal thicknesses, seismic velocities, volcanism, heat flow and deep resistivity structure have all pointed to a fundamental concentration of rifting and upper mantle diapirism with a strong NNE preferred orientation that presently is affecting the eastern margin of the Great Basin and the westernmost parts of the Colorado Plateau. Coincident with this trend of enhanced tectonism is the distribution of high-temperature geothermal resources depicted by Ward et al. (1978); high-temperature ($> 150^{\circ}\text{C}$) systems are much less common to the west in the Great Basin interior while to the east in the Colorado Plateau interior geothermal resources are essentially non-existent (Brooke et al., 1978). Since the lower crustal melting events that beget upper crustal silicic magma chambers are fostered by high regional extension rates along with localization of deep crustal intrusion or underplating by basalt (Smith and Shaw, 1975; Hildreth, 1981; Crecraft et al., 1981; Nash, 1982), the gathering of geothermal resources along this trend of anomalous rifting seems inevitable.

The intersection of this NNE-oriented zone of extension with the ENE-directed Pioche-Marysvale trend contains yet a further concentration of bimodal volcanism, heat flow and geothermal resources (Rowley et al., 1979; Chapman et al., 1981; see Figure 2). Furthermore, Rowley and colleagues observe a tendency for Late Cenozoic rhyolitic centers to become progressively younger in an

easterly direction along the Pioche-Marysvale belt. This progression might mark the eastward movement of the intersection of this belt with the NNE-trending region of enhanced stretching, since the latter activity appears to be migrating gradually eastward into the Colorado Plateau (Thompson and Zoback, 1979; McGetchin et al., 1980). In any event, the intersection of these two tectonic trends defines a special zone of weakness encouraging extension, intrusion of basalts from the upper mantle, emplacement of silicic magma chambers surface igneous activity, and hydrothermal convection within the crust (Rowley et al., 1979; Hildreth, 1981).

We suggest that this focusing of magmatism has produced a corresponding anomalous resistivity structure at depth in this area which has characteristic horizontal dimensions of tens of kilometers. It is not necessary that a readily detectable reduction in bulk crustal resistivity directly reflect such a plexus of postulated intrusions; at the very least, the pronounced warming of the mid to deep crust by this localized intrusive activity could provide a resistivity target through solid-state conduction (see earlier discussion of lower crustal conductivity mechanisms). To assess this possibility, a continuation of magnetotelluric profiling from the Roosevelt Hot Springs to, say, Pine Valley in the west and to the Sevier Fault in the east is recommended.

Nevertheless, the application of magnetotellurics in delineating deep magmatic heat sources for individual geothermal systems remains highly problematical. At the Roosevelt Hot Springs, the MT response of any such source must be unraveled from the responses of adjacent graben sediments and of overlying hydrothermal alteration and

variations in overburden thickness (Wannamaker et al., 1980; Ross et al., 1982). Furthermore, if Great Basin extensional faulting is the principal control on magma emplacement, one expects a N-S elongation of the chamber. Tensor quantities ρ_{xy} and ϕ_{xy} are likely to be the MT functions most sensitive to such a geometry, but these cannot be modeled with a 2D algorithm as can ρ_{yx} and ϕ_{yx} (Wannamaker et al., 1982). To interpret these functions rigorously requires a computer routine which can simulate the MT signatures of arbitrary 3D structures in a layered earth. While the program of Wannamaker and Hohmann (1982) theoretically is capable of this, sheer expense precludes the modeling of a prospect with the degree of complexity of the Roosevelt Hot Springs at present. The existence of a magma chamber here cannot be ruled out on the basis of MT, but neither can it yet be substantiated.

CONCLUSIONS

In active extensional environments, the potential of electrical resistivity to portray the physical state at depth in the earth is great. Here, significant temperature perturbations may exist, possibly with interconnected melt phases, creating lateral and vertical contrasts in bulk resistivity of an order of magnitude or greater. Unluckily, such environments are attended to a large degree by upper crustal lateral inhomogeneities, especially graben sedimentary fill, whose resistivity contrast with the surrounding rock host is at least as high as that of the structures of interest at depth, and whose proximity to the surface obscures the interpretation of deep targets using electromagnetic measurements.

However, there is generally a strong preferred orientation of lithospheric deformation, and thus of the distribution of lateral resistivity inhomogeneities, in rift environments that is perpendicular to the direction of spreading. Three-dimensional model simulations have shown that a 2D transverse magnetic mode algorithm is most suitable for removing the distortion of MT soundings by such upper crustal structures and hence for assessing the resistivity of the deeper crust and upper mantle in extensional regimes.

Such an appropriately rigorous modeling technology has been applied to an extensive set of MT measurements at the Roosevelt Hot Springs in southwestern Utah. A resultant finite element cross-

section, whose MT response shows excellent agreement with field observations, contains resistivities in the upper 2 km as low nearly as 1 Ω -m representing brine-saturated graben sediments hosted within crustal rocks having resistivities of several hundred ohm-meters. At greater depths, a one-dimensional model resistivity structure is sufficient to fit the observations. A feature of great interest required by the data is a low-resistivity (nominally 20 Ω -m) layer residing from about 35 to 65 km in the upper mantle beneath the Roosevelt Hot Springs. In particular, sensitivity tests of the MT response to the model indicate that this low-resistivity medium is not a "thin" layer, but rather that its conductivity and thickness to some extent can be resolved independently. No low-resistivity layer in the middle to lower crust has been detected, however (cf. Eaton, 1980; Jiracek et al., 1982).

We can interpret the entire deep resistivity section to be controlled by aqueous electrolytic conduction through rock pores in the upper crust, by solid-state conduction mechanisms in dry rocks of the lower crust and most of the upper mantle, and by ionic conduction through an interconnected basaltic partial melt of low melt fraction within a solid peridotite matrix for the low-resistivity layer in the upper mantle. The interpretation of resistivity structure due to partial melting in peridotite is greatly simplified for the Great Basin, and probably for many other rift environments as well, by the essentially invariant, volatile-deficient nature of the great majority of basaltic magmas. We do note, however, that melt fraction estimates derived from currently-favored models of melt tortuosity may be underestimates in regions of extension.

The present-day deep resistivity structure of southwestern Utah derives from the evolution of the eastern Great Basin through Late Cenozoic time. The low-resistivity layer we have interpreted in the upper mantle here is but a manifestation of diapiric uprise and melting events occurring over an interval that extends to much greater depths than does the layer itself. Given the eutectic-like and volatile-deficient character of Great Basin basalts, we assign a temperature trajectory coincident with the dry peridotite solidus to depths from 35 km to perhaps in excess of 160 km in the upper mantle here in southwestern Utah. Both the resistivity structure and the timing and amplitude of extensional activity suggest that temperatures in the deep crust and uppermost mantle here exceed those which would be inferred from surface heat flow.

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