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MAGNETOTELLURIC AND GEOMAGNETIC DEEP-SOUNDING STUDIES IN RIFTS AND ADJACENT AREAS: CONSTRAINTS ON PHYSICAL PROCESSES IN THE CRUST AND UPPER MANTLE

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Abstract. Deep electrical studies are reviewed along with other supportive geophysical/geological investigations of five of the major rift zones of the world: the Baikal rift, and Rhine graben, the East African rift, Iceland and the Rio Grande rift. All of these regions exhibit anomalously low values of electrical resistivity, density and seismic velocity, either within the crust itself or at high levels in the mantle. Deep electrical studies support a model where ascending masses of material from the mantle are intimately coupled to the fractionation of a basalt melt and its accumulation at higher levels within the earth. In Iceland, an interplate rift, the accumulation and chilling of melt at the base of the crust apparently leads to a significant component of crustal underplating, whereas in intraplate rifts such as the Rio Grande rift, the emplacement of basaltic magma at high levels may lead to extensive remelting of the crust, triggering eruptive episodes of silicic magmas.

Introduction

Knowledge of the distribution of electrical conductivity in the crust and upper mantle places important constraints on models for physical properties beneath the active rift zones of the earth. Clearly the coupling of regional doming, rifting, and magma genesis in these regions involves a variety of thermal phenomena which also serve to modulate in situ values of electrical conductivity. Therefore regional electromagnetic studies, such as magnetotelluric and geomagnetic deep-sounding experiments, can establish limits not only on the high temperature experienced by materials in the solid state, but on a variety of other thermally related phenomena as well: the presence of hydrothermal fluids in pores and cracks, hydrated and dehydrated phases in metamorphosed materials, and the presence of highly conducting melt phases distributed along grain boundaries. The large number of factors influencing the electrical conductivity precludes, of course, the unequivocal interpretation of any deep electrical studies on its own merits. However, when combined with other geophysical and geological information, electromagnetic studies contribute substantially to discriminating between various hypotheses offered for processes associated with rift tectonics.

The following discussion compares deep electrical studies in five of the major rift zones of the world: the Baikal rift, the Rhine graben, the East African rift, Iceland and the Rio Grande rift. We first summarize the principal geophysical/geoelectrical characteristics of each of these regions. Then a model is proposed which embraces many of the geophysical constraints established by the studies and which also seems compatible with many of the megatectonic/magmatic elements associated with rifting environments.

Electromagnetic induction studies in rift zones using long-period natural electromagnetic fields have been reviewed previously by Garland (1975), Hutton (1977) and Jiracek et al. (1979). Although present discussion is to some degree an extension of this earlier work, our present emphasis is on the implications these studies have for constraining models of actual physical phenomena in the earth, particularly the interaction between crust and mantle processes. In the interest of keeping the present discussion of manageable size, we will not review the interesting implications of a geomagnetic deep-sounding in the Imperial Valley of California (White, 1973a,b) or of a magnetotelluric investigation in the Snake River Plains of Idaho (Stanley et al., 1977).

Overview of Rift Studies

The Baikal Rift Zone

The Baikal rift zone (Figure 1) extends for a distance of 1500 km along the southeastern edge of the Precambrian Siberian platform and, with vertical displacements of up to 6-8 km, represents the world's deepest intracontinental rift valley (Logatchev and Florensov, 1978). The morphotectonic development of this graben feature accompanied broad-scale domal uplift, apparently due to local diapiric flow in the astheno-

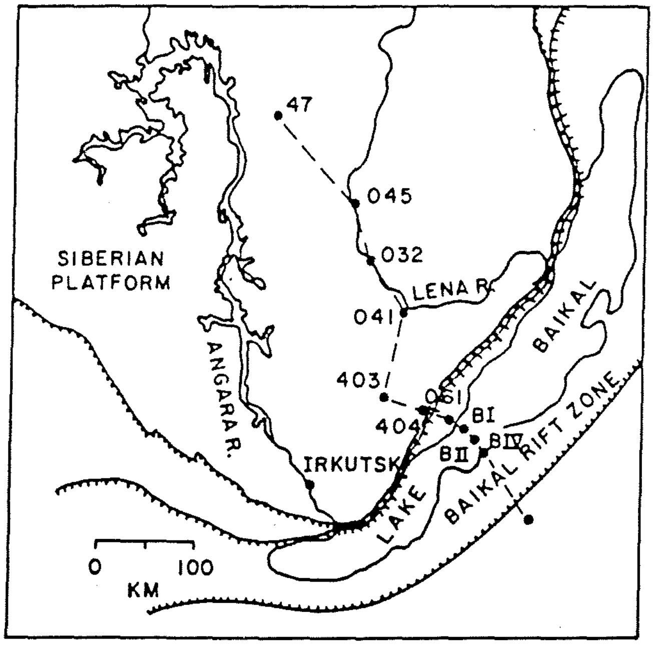

Fig. 1. Map of the Baikal region showing the location of sites for the magnetotelluric data reported by Gornostayev et al. (1970). Boundaries of the Siberian platform and the Baikal rift zone are generalized from Lubimova et al. (1972).

sphere. The Baikal rift system is isolated from the world rift system and appears to represent an autonomous complex of Cenozoic tectonism and volcanism (ibid.).

Although much of the research on the Baikal rift has not been reported in the English literature, this region is one of the most thoroughly studied of the continental rift systems (Puzirev et al., 1978). Soviet geophysicists have a long-standing tradition in the application of magnetotelluric and geomagnetic deep-sounding techniques to studying regional tectonic problems, and it is not surprising that magnetotelluric sounding experiments were among the first quantitative geophysical indicators of the anomalous nature of the crust and upper mantle beneath the Baikal rift and the adjacent regions (Gornostayev et al., 1970, Gornostayev, 1972; Pospeev and Mikhalevsky, 1976; Berdichevsky et al., 1980). Basically, the magnetotelluric interpretation indicates five general zones within the earth's interior (Figure 2): first, a conductive sedimentary layer several kilometers thick and having a resistivity on the order of 20-50 ohm-m; second, a resistive zone (> 500 ohm-m) representing crytalline crust and extending to a depth of 10-20 km; third, a conductive intracrustal zone at a depth of 12-25 km having a depth-integrated conductivity (section conductance) of S = 600 - 2500 Siemens (Mhos). The maximum thickness of this zone appears to be 20 km (if $\rho \simeq 30$ ohm-m); it could have a minimum thickness of less than 1 km if the resistivity is on the order of 1 ohm-m or less. The fourth zone appears at a depth of 20-40 km (depending on the thickness of the third zone) and seems to have a resistivity in excess of several hundred ohm-m. It is thought to be made up of peridotite containing emulsified basalt. Fifth and finally, data from several long-period sites indicate a more conductive layer (< 10 ohm-m) at depths in excess of 50-100 km beneath the rift axis and at depths exceeding 200 km to the northwest of the rift (Pospeev and Mikhalevsky, 1976). However, a recent statistical analysis of more than 300 magnetotelluric soundings in the Baikal region (Berdichevsky et al., 1980) shows that although a high-level conducting layer (depth ~12 km; conductance ~2000 s) is clearly required to satisfy the data, the presence and character of a possible deeper conducting feature (d > 50 km) is not very well resolved.

In other words, the most striking result of the magnetotelluric interpretation from the Baikal rift is the presence of the anomalous conducting zone (depth ~12-15 km; conductance ~2000 s) of limited thickness (< 30 km). The second noteworthy feature of the data is that the earth appears to be surprisingly resistive beneath the intracrustal conductor, perhaps more than several hundred ohm-m. It is likely that this deeper zone (from a depth as shallow as 20-40 km to a depth exceeding 100 km) represents an ultramafic material which, while having a relatively high temperature, may still be relatively resistive (Gornostayev, 1972).

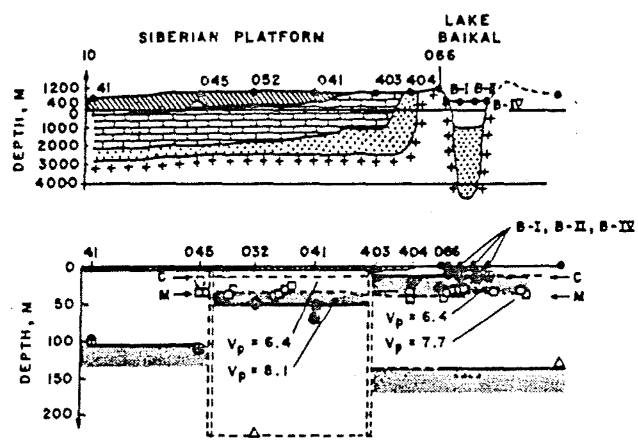

Fig. 2. Vertical section along Profile 1 (see map of Baikal region in Fig. 1) showing the detailed interpretation of magnetotelluric data by Gornostayev et al. (1970). The horizontal solid lines in the lower figure correspond to the depth to conductive interfaces. The velocities of compressional seismic waves are also shown, along with the approximate depths to the Mohorovicic discontinuity (M) and the Conrad discontinuity (C).

Heat flow work in the Baikal rift described by Lubimova et al. (1972) is compared with results from various rift areas of the world in Table 1. Clearly, the heat flow in the Baikal rift is anomalously high, although these values are restricted to a band at most 60 km wide, along the morphotectonic expression of the rift. Lubimova et al. (1972) have argued, on the basis of the narrow width of the Baikal thermal anomaly, for a relatively shallow depth of origin for this feature (< 24 km), and suggest the spatial coincidence of a long-term temperature perturbation with the high-level (10-20 km depth) electrical anomaly revealed through the geomagnetic variation and magnetotelluric deep-sounding studies described above. Moreover, Lubimova et al. point out that the static magnetic anomalies are restricted to a crustal depth less than 14 km in this region, suggesting a relatively shallow depth to the Curie isotherm (ibid.). From this evidence for high temperature at intermediate levels in the crust, they argue that the electrical anomaly may be associated with the accumulation of a basalt melt fraction derived either by partial melting in situ or by segregation and upward migration of melt from deeper zones.

Seismic investigations support many of the conclusions based on electromagnetic and deep thermal studies in the Baikal rift (Puzirev et al., 1970, 1974, 1978; Krylov et al., 1974, 1975). An intracrustal low-velocity layer (depth $\sim$ 15-17 km), with a velocity differential of -0.2 to -0.3 km/sec relative to values in normal material above and below this zone, has been detected under two sites along the margin of the rift approximately 150 km apart (Krylov et al., 1974, 1975). This low-velocity crustal layer is associated with the zone of high electrical conductivity (Puzirev et al., 1978; Berdichevsky et al., 1980). Earthquake foci beneath the rift are restricted to the volume above the anomalous intracrustal layer, and their absence at greater depth suggests a rapid transition to higher temperature, more plastic conditions (Puzirev et al., 1978). It is thought that the anomalous intracrustal layer is most likely to result from increased heating and partial melting (ibid.).

Large-scale seismic refraction experiments (line-lengths of up to 300 km) indicate the presence of anomalous low-velocity mantle material (7.6-7.8 km/sec) associated with the base of the crust beneath the entire region. This pillow-like zone (average thickness 17 km) has been mapped over an area of 200,000 km^2 and occupies a region 2-3 times wider than the actual geologic trace of the rift zone (Puzirev et al., 1978). Krylov (1976) has used estimates of the density, P-wave velocity and electrical resistivity to infer a partial melt fraction of 5-10%. Logatchev and Florensov (1978) reason that it is quite possible for mantle material to have penetrated into the lower crust, essentially in the form of semi-solid peridotite, with only an insignificant admixture of basalt magma, and feel that this interpretation is supported by the gravity, P-wave velocity, heat flow and deep electroconductivity soundings.

An alternative suggestion is that the anomalous conducting layer at intracrustal levels is caused by the release of crystalline-bound water through dehydration processes and its accumulation as an electrolytic fluid along pores and joints (Gornostayev, 1972; Pospeev and Mikhalevsky, 1976; Berdichevsky et al., 1980). Even if this is the case at high levels in the crust, there remains the distinct possibility that the electromagnetic data are indicating the accumulation of basaltic melt at depth below this, perhaps as shallow as 20 km or less (Gornostayev, 1972).

TABLE 1. Heat Flow in Major Rift Zones

Region	Heat Flow microcal/cm^2/sec	Reference
East African Rift	2.88	Von Herzen and Vacquier, 1967
Basin and Range	2.21-2.44	Warren et al., 1969
Rhine Graben	2.6	Meincke et al., 1967
Iceland NVZ	5-6	Palmason, 1973
Baikal Rift	2.0-3.4	Lubimova et al., 1972
Rio Grande Rift	2.56±0.65	Reiter et al., 1979

The Rhine Graben

The Rhine Graben (Figures 3 and 4) is a coherent structural feature, sharply bounded by master faults, for a distance of over 300 km from Frankfurt to Basel in the Federal Republic of Germany (Illies, 1970). The graben itself is superposed on a complex system of fractures and faults extending from the Netherlands into southern France, and seems to be a fundamental component in the dynamics of global tectonics (Illies, 1974).

The geophysical framework for regional studies of the Rhine graben and adjacent areas is summarized by Fuchs (1974). Several prominent features in the crust and upper mantle appear to play a profound role in the tectonic development of the region. First, a low-velocity layer (the sialic low-velocity zone) is detected at depths of only 10 km beneath the graben (Figure 5; Ansorge et al., 1970; Mueller, 1970). Second, significant crustal thinning seems to character-

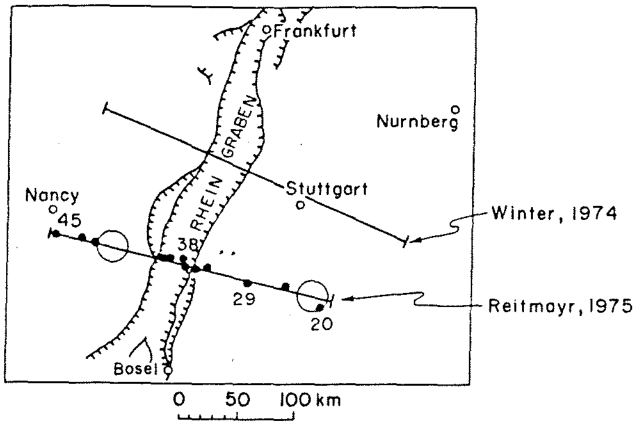

Fig. 3. Map of the Rhine graben (after Illies, 1970) showing the geomagnetic variation profile lines of Winter (1974) and Reitmayr (1975). The numbers on the southern profile refer to sites indicated in Fig. 7. The circles represent zones where, according to Reitmayr (1975), magnetotelluric measurements could be explained with one-dimensional models.

ize the graben; the crust-mantle boundary is elevated to a depth of only 25 km in the southern part of the graben (Ansorge et al., 1970), whereas in the northern graben, seismic reflection experiments indicate an uplift of the crust-mantle boundary along the flanks on either side of the graben (Meissner and Vetter, 1974). Third, there is some suggestion of a low-velocity pillow in the upper mantle (Meissner et al., 1970), and while reversed seismic refraction measurements indicate more normal mantle velocities at the base of the crust beneath the southern graben (Rhine Graben Research Group for Explosion Seismology, 1974; Edel et al., 1975), it is still necessary to invoke a low-density (hence low-velocity) pillow within the mantle itself to explain the regional gravity data (Fuchs, 1974). The pronounced thinning of the crust beneath the rift, seen seismically, would suggest a much larger positive Bouguer anomaly than actually measured; hence, an anomalous low-density (lower-velocity) compensating mass must be imbedded at some depth within more normal mantle material (Fuchs, 1974). Moreover, seismic surface wave studies using the fundamental Rayleigh wave mode (Reichenbach and Mueller, 1974) also support a low-velocity cushion in the upper mantle.

Mueller and Rybach (1974) argue that relatively high values of heat flow in the graben, along with an inferred decrease in the P-velocity of some 7 percent, are compatible with the view that the sialic low-velocity channel may be in a state of partial fusion (at a depth of 20-25 km). Local increases in the temperature gradient superposed on the regional field of increased heat flow may be explained by cooling bodies of acidic magma which have intruded into the crystalline basement from the sialic low-velocity zone (ibid.). Illies (1970) feels that abnormally high values of heat flow preclude the rupture of crustal material deeper than 10 km; rather, one would expect more plastic conditions to obtain. Fuchs (1974) proposes a model for the tectonic evolution of the rift in which the sialic low-velocity layer plays a major role in the rheological behavior of the crust.

Magnetotelluric, telluric and geomagnetic variation studies strongly support the possibility

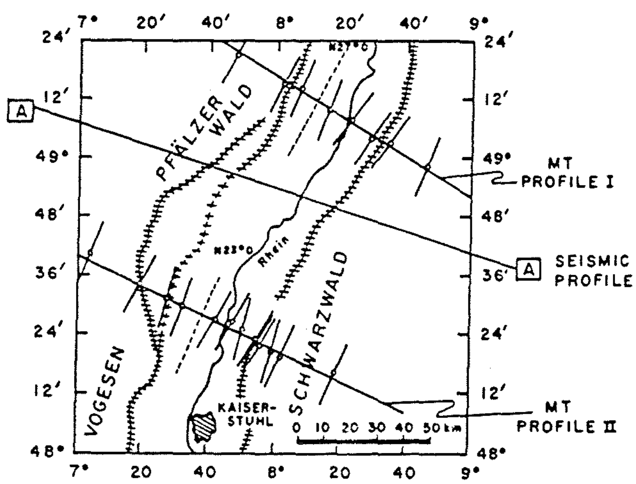

Fig. 4. The location of magnetotelluric profiles discussed by Scheelke (1972, 1974) and the seismic section described by Mueller and Rybach (1974). The seismic section shown in Fig. 5 is representative of the axial portion of the graben in this region.

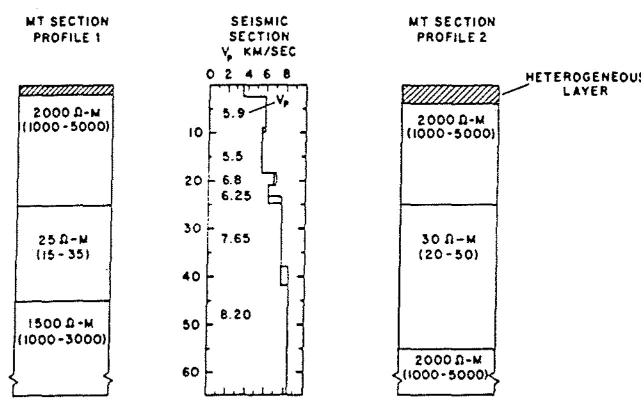

Fig. 5. A comparison of the interpretation of magnetotelluric data by Scheelke (1972, 1974) from the two profiles shown on the previous figure with the interpretation of seismic data by Mueller and Rybach (1974). The seismic velocities in the mantle are inferred from the surface wave studies of Reichenbach and Mueller (1974).

of very high temperature beneath the Rhine graben proper (Fuchs, 1974). Haak et al. (1970) used telluric measurements and Winter (1970) used geomagnetic variations to determine a depth of approximately 25 km to an anomalously conducting layer (< 30 ohm-m) beneath the axis of the rift. Because of the limited depth resolution due to the restricted period band for which data were used, there is some ambiguity intrinsic in these analyses. However, preliminary magnetotelluric measurements over a broader period range by Losecke (1970) suggest that this anomalous layer is at a depth of ∿28 km and may be associated with the pillow of low-velocity material in the upper mantle. The most reliable parameter that can be estimated for such a conducting layer, when sandwiched between resistive layers, is generally its total depth-integrated conductivity, or the so-called section-conductance. In this case, the layer conductance would be S = 1200 - 1700 Siemens (Mhos). Losecke (1970) favors an interpretation in which this is the conductance of a layer approximately 5 km thick and having a resistivity of 3-6 ohm-m. To interpret more refined magnetotelluric data with periods between 100 and 1000 seconds, Scheelke (1972, 1974) also found it necessary to assume an anomalously conducting layer (Figure 5)

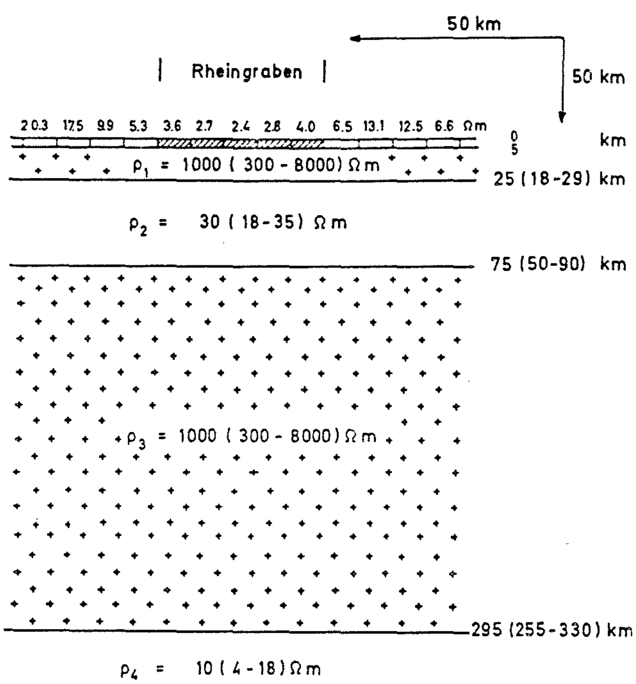

Fig. 6. The interpretation of geomagnetic variation data by Winter (1974) along the profile indicated in Fig. 3. Lateral changes in resistivity are restricted to a surface layer having a thickness of 5 km. The parameters and their estimated ranges of variability are shown for the best-fitting models.

beginning at a depth of about 25 km, although he assumed a resistivity of about 25 ohm-m and a thickness of 20 km (S = 800 Siemens).

Detailed interpretations of geomagnetic variation data along the two profiles shown in Figure 3 are presented by Winter (1974) and Reitmayr (1975). The magnetotelluric interpretation is compatible with the geomagnetic variation results, although, as shown in Figure 6, Winter (1974) favors a somewhat more resistive (∿30 ohm-m) and thicker (∿50 km) anomalous zone. The conducting layer proposed for the upper mantle by Reitmayr (Figure 7) is even more resistive (50 ohm-m), but is somewhat thinner (∿20 km) than the structure proposed by Winter. The section conductance of the layer proposed by Reitmayr (400 Siemens) is significantly smaller than that proposed by Winter (1700 Siemens). Some of this difference is explained by the fact that Reitmayr proposes the existence of a second conducting layer (∿25 ohm-m; 25 km thick) at a depth of 75 km. This layer (S = 1000 Siemens) is absent in Winter's interpretation, although he has an equivalent conductance above a depth of 75 km. These subtle differences are difficult to resolve from the geomagnetic variation data alone for the range of periods available. It seems clear, however, that the upper mantle has anomalously low resistivities (25-50 ohm-m) at depths as shallow as 25 km.

Magnetotelluric data have been analyzed by Haak and Reitmayr (1974) from two sites outside the morphotectonic expression of the rift (Figure 3): Bolstern, 120 km east of the Rhine graben, and Saint Stacl, 45 km west. The interpretation of data from both sides (Figure 7) indicates that the high-level (depth ∿25 km) conducting layer is either absent altogether or is significantly deeper (∿70-90 km) outside the rift proper. The upper mantle in these adjacent regions has a rather high resistivity (10^2 to 10^4 ohm-m) at depths on the order of 70 km (see Figure 7). It is interesting to notice the close similarity between the structure proposed for the Rhine graben by Reitmayr (Figure 7) and that proposed for the Baikal rift (Figure 2) by Gornostayev et al. (1974).

The East African Rift

The rift zone of East Africa is a system of normal faults bordering a trough-like feature 40 to 65 km wide which traverses two broad, contiguous domal uplifts in Ethiopia and Kenya (Baker et al., 1972). Domal uplift has occurred in three major pulses over the last 35-40 × 10^6 years - in late Eocene, in mid-Miocene and in Plio-Pleistocene times - and is associated with episodes of graben faulting and volcanism of intermediate to silicic type.

The Kenyan Rift. A large (350 km wide) negative-gravity anomaly (-50 milligal) is associated with the domal uplift and appears to reflect low-density, partially fused material in

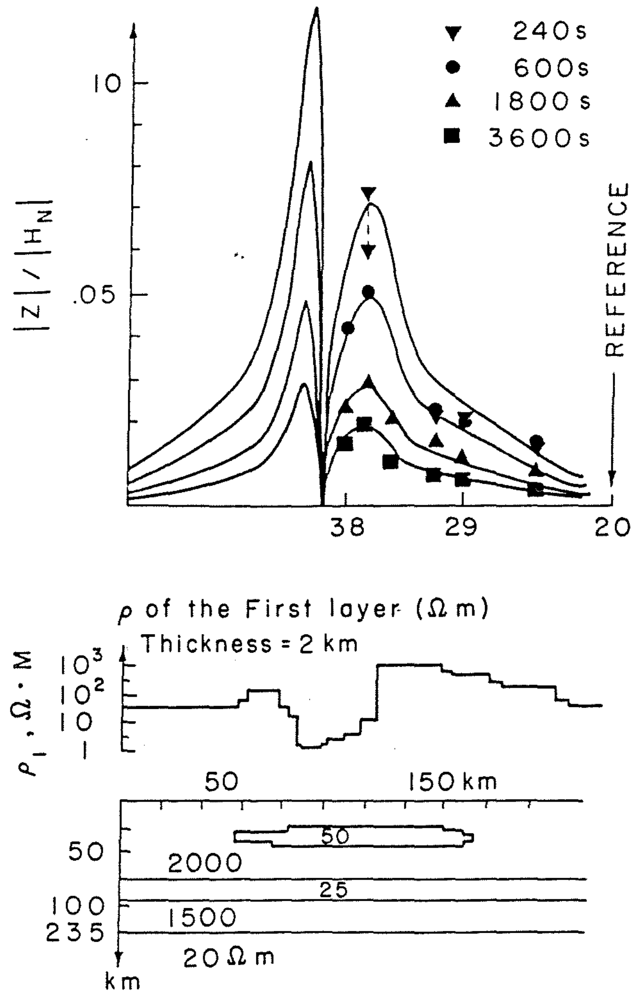

Fig. 7. The ratio of the amplitude of vertical to normal horizontal magnetic field variations from the sites occupied in the Rhine graben (Fig. 3) by Reitmayr (1975). His two-dimensional model and its theoretical $|Z|/|H_N|$ response at various periods is shown. Two low-resistivity zones are shown in the upper mantle: one laterally limited, at a depth of 25 to 45 km beneath the graben axis (ρ = 50 ohm-m); the other laterally continuous at a depth of 75 to 95 km (ρ = 25 ohm-m).

the upper mantle (Fairhead, 1976). The model proposed by Fairhead (1976) involves the upward movement of the lithosphere-asthenosphere boundary on a regional scale in which the cold, high-density lithosphere (3.34 gm/cc) is replaced by hot, lower-density asthenosphere (3.29 gm/cc) over a lateral distance of more than 250 km.

A seismic refraction experiment along the axis of the Kenyan rift south of Lake Rudolf (Griffiths, 1972) indicates an average crustal velocity of V_p = 6.4 km/sec and an upper mantle velocity of 7.5 km/sec. The crust appears to be only 20 km thick. The dispersion of seismic surface waves suggests a crust-mantle structure not unlike that of the Basin and Range province of western North America, with anomalous low-velocity mantle material (V_s = 4.25-4.45 km/sec) extending to a depth of 120-200 km (Knopoff and Schlue, 1972). Long et al. (1972) combined teleseismic observations with regional surface wave data in a study which suggested that the anomalous crust is of very limited lateral extent on either side of the morphotectonic rift feature. However, this study did confirm crustal, or lithospheric, thinning along the axis of the ridge, indicating a strong similarity between the mean compressional velocity of the mantle beneath East Africa and that beneath Iceland. The presence of such an ultra-low-velocity zone seems to indicate a mantle structure analogous to that beneath ocean ridges (ibid.).

Reconaissance geomagnetic variation studies using an array of six recording magnetic variometers (Figure 8) revealed an inductive anomaly associated with the Kenyan rift 100-150 km north of Nairobi (Banks and Ottey, 1974). These studies suggested the presence of a high-conductivity zone (< 0.1 S/m) elongated along the axis of the rift, at a depth of less than 20 km. Because of the restricted band of periods for which data were available (~25 min), Banks and Ottey (1974) were unable to determine unequivocally the depth to the top of this conducting feature. According to these investigators, a conductor at the surface, such as might be associated with conducting sediments in the valley fill, could as readily explain the observed data as a conductor somewhat deeper in the crust. However, attempting to model the effect as due simply to the conducting sediments required a conductivity of 0.2 S/m and a thickness of 5 km for the rift fill; they argued that these values were unreasonably large, and concluded that the source of the anomaly must be sought at greater depths.

A magnetotelluric traverse (Rooney and Hutton, 1977) was undertaken in the same portion of the rift (Figure 8) as the geomagnetic variation studies described above. The magnetotelluric data (Figure 9) confirmed and extended the conclusions of the geomagnetic deep-sounding experiment. In fact, the magnetotelluric investigations, because they covered a much wider frequency band, were able to resolve a more detailed model for the electrical structure beneath the rift. It appears that the entire vertical section beneath the rift is dominated by lower than normal resistivities (approximately 15 ohm-m) from the surface to depths greater than 30 km. The simplest model that fits the data is essentially a homogeneous half-space having a resistivity of approximately 10 ohm-m from the surface to a depth exceeding 35 km. It is possible, however, that a resistive layer could be present beneath the conducting sediments and not be resolved in the analysis of the data available. According to Rooney and Hutton (1977), such a resistive layer could be at a depth somewhat greater than 5 km, and might be as thick as 5 to 10 km; it would reflect the presence of a resistive, crystalline basement. Whether or not such a resistive zone is present, however, the material beneath this zone would have to be strikingly conductive, and at a depth which seems to be of the order of (or less than) the depth to the crust-mantle boundary in this region.

Although data from only a single site were analyzed in detail by Rooney and Hutton (1977), there is remarkable agreement among long-period

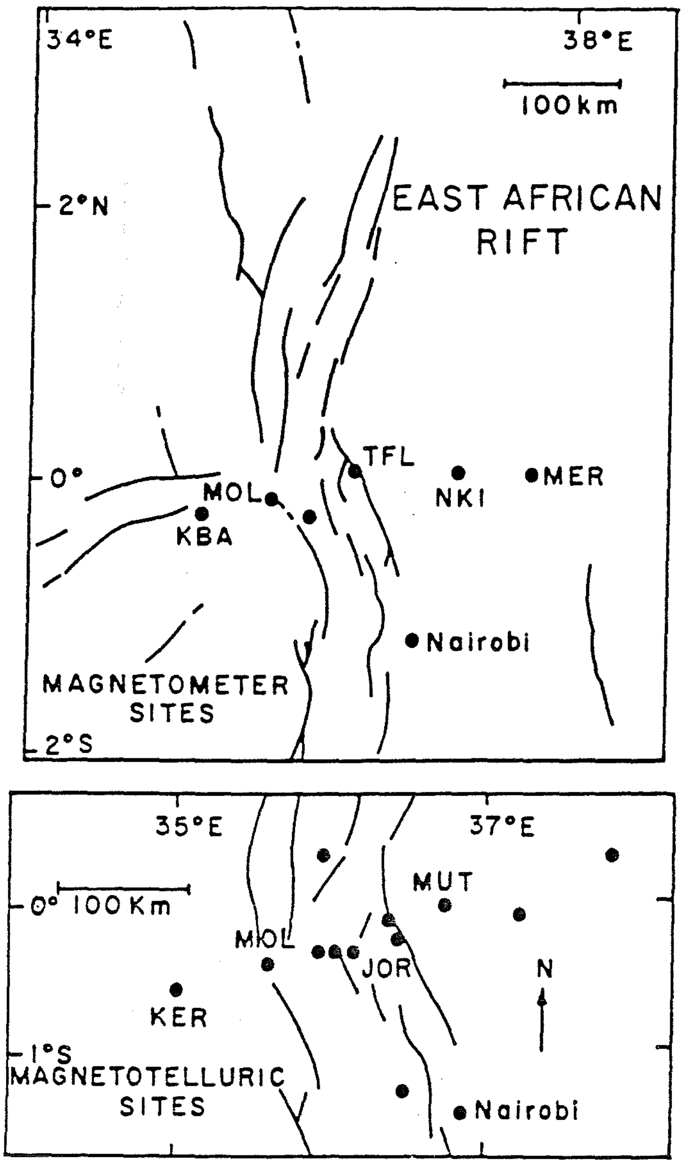

Fig. 8. The Kenyan rift in East Africa, showing the location of magnetic variometry sites (top figure, after Banks and Ottey, 1974) and magnetotelluric sites (bottom figure, after Rooney and Hutton, 1977).

data from at least three sites along a profile over 100 km long, spanning the rift. This leaves little doubt that a thick section of material having a resistivity of 15 ohm-m or less must be associated with the sub-basement of the East African rift.

Afar. The Afar triangle may be a supramarine manifestation of incipient oceanic development (Mohr, 1972). Basaltic volcanism of oceanic type is related to crustal separation, whereas large, essentially silicic central volcanoes are present along the margins of the depression and are thought to represent interactions between subcrustal magma and sialic crust (Barbieri et al., 1972). In this sense, the Afar region might be as representative of mid-ocean ridge processes as those observed in Iceland (Mohr, 1972).

Seismic sections show a strong similarity to the interpretation of Icelandic data (Lepine et al., 1972). An interpretation of gravity data suggests an attenuation of the crust from continental type in south Afar to largely oceanic type in north Afar (Makris et al., 1972); moreover, there is a strong indication of a low-density, low-velocity pod of mantle material beneath the entire region (ibid.).

Berktold et al. (1975), in discussing the interpretation of magnetotelluric measurements, note the gross similarity between the apparent resistivity data from Afar and those from Iceland described by Hermance (1973). Unusually high temperatures are inferred at depths of 15 km, and data from measuring sites outside the rift proper suggest that the conducting anomaly extends beyond the region of the rift itself, though at a somewhat greater depth. On the basis of the zone of very low resistivity which has been detected below the Afar depression and its west escarpment, a temperature of 800-1200°C is inferred at a depth of 15 km, corresponding to a mean gradient of 60°C/km. Such a temperature is compatible with partial fusion of material at this depth (Berktold et al., 1975).

The Rift in Southern Africa. It has been suggested, on the basis of recent faulting and seismic activity, that the region of southern Africa between Rhodesia and Botswana on one side and Angola and Zambia on the other may be a site for incipient rifting (Reeves, 1972; Scholz et al., 1976). Geomagnetic variation experiments by de Beer et al. (1975), using an array of 25 Gough-Reitzel recording variometers, indicate an electrical conductivity anomaly associated with this same regional trend. Even though a quantitative interpretive model of the data has not been advanced, the strength of using natural electromagnetic methods in a reconnaissance mode has been well demonstrated by these experiments.

Iceland

The position of Iceland on the Mid-Atlantic ridge has led to its extensive study as a site of

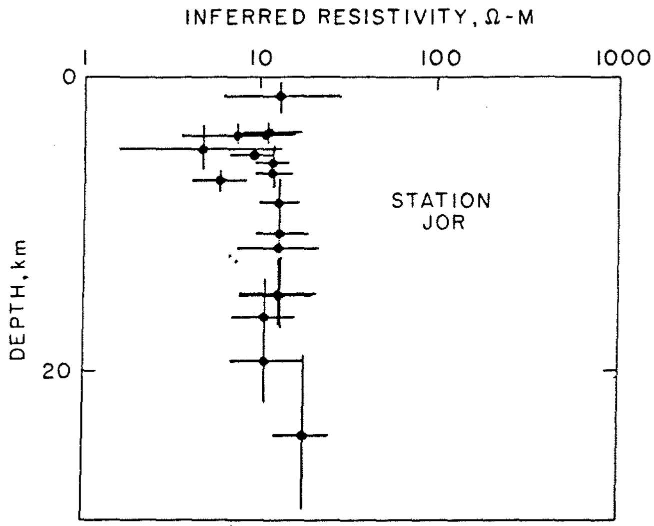

Fig. 9. Inversion of magnetotelluric parallel resistivity data from site JOR in the East African rift (see Fig. 8) (after Rooney and Hutton, 1977).

active accretion at plate margins (Bodvarsson and Walker, 1964; Ward, 1971; Palmason and Saemundsson, 1974; Beloussov and Milanovskiy, 1976). The neovolcanic zone (Figure 10), in particular, is characterized by the type of active linear volcanism associated with mid-ocean spreading centers (Palmason and Saemundsson, 1974). Figure 11 is a map of the area around Lake Myvatn. The geology between Myvatn and Jokulsa a Fjollum is characterized by a series of north-northeast striking fractures and fissures, dramatic evidence of the post-glacial tectonic activity which characterizes this region (Saemundsson, 1974). Although the neovolcanic zone has been volcanically active since the last glaciation and is apparently a center for present crustal spreading, the eastern and northwestern margins of the island, in contrast, are much more stable tectonically, having an age on the order of 10^7 years B.P. or greater.

A major episode of rifting along the boundary separating the North American and European plates began in northern Iceland on December 20, 1975 and is continuing at the present time (Bjornsson et al., 1977, 1979). During this time, crustal spreading of some 3 m has been observed in the Krafla caldera (Figure 11), and the caldera itself has been the focus of an 80-km-long fissure swarm. Needless to say, Iceland is one of the most active supramarine volcanic areas on earth (Palmason, 1976) and remains one of the best examples of the correlation of electrical conductivity anomalies and regional thermal processes in the deep crust and upper mantle (Garland, 1975).

A number of geophysical studies have been carried out in this area. Gravity measurements have been performed by Einarsson (1954), Schleusener et al. (1976) and Palmason (private communication, 1978); heat-flow work is summarized by Palmason (1973).

Seismic refraction studies in the neovolcanic zone (Palmason, 1971) suggest that velocities in the upper crust increase from 2-3 km/sec at the surface to approximately 6.5 km/sec at a depth of 4-5 km. The lower crust (V_p = 6.5 km/sec) is thought to be made up of basic intrusives and metabasalts (Palmason, 1971; Palmason and Saemundson, 1974). Material having a P-wave velocity of 7.2 to 7.4 km/sec is associated with an anomalous mantle and, although not seen on all seismic refraction lines in this region, is estimated from surface wave studies (Båth, 1960; Tryggvason, 1962) to be at an average depth of approximately 10 km. In comparison, seismic refraction studies by Pálmason (1971) indicate that the upper mantle may be as little as 8 km beneath the southwestern rift zone and may deepen to as much as 15 km beneath the southeastern portion of the island. Francis (1969), using body-wave data from earthquakes along the Mid-Atlantic ridge, and Long and Mitchell (1970), using teleseismic signals recorded at four stations over Iceland, suggest that anomalously low-velocity mantle material may extend to a depth of 200-250 km beneath the entire island.

The first indication of anomalously high electrical conductivities beneath Iceland was provided by the geomagnetic variation studies of Hermance and Garland (1968), who used a primi-

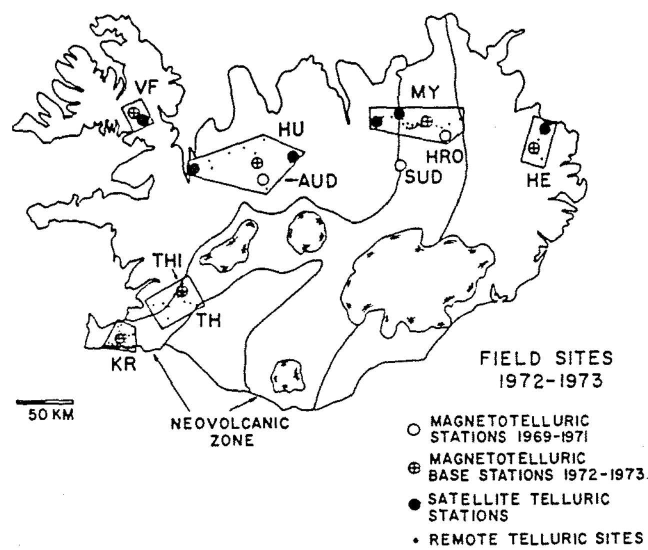

Fig. 10. Location of magnetotelluric survey areas in Iceland (after Hermance et al., 1976). Code letters refer to Krisuvik (KR), Thingvellir (TH), Westfjords (VF), Hunavatnssysla (HU), Myvatn (MY) and Herad (HE). The boundaries of the neovolcanic zone are generalized from Palmason and Saemundsson (1974).

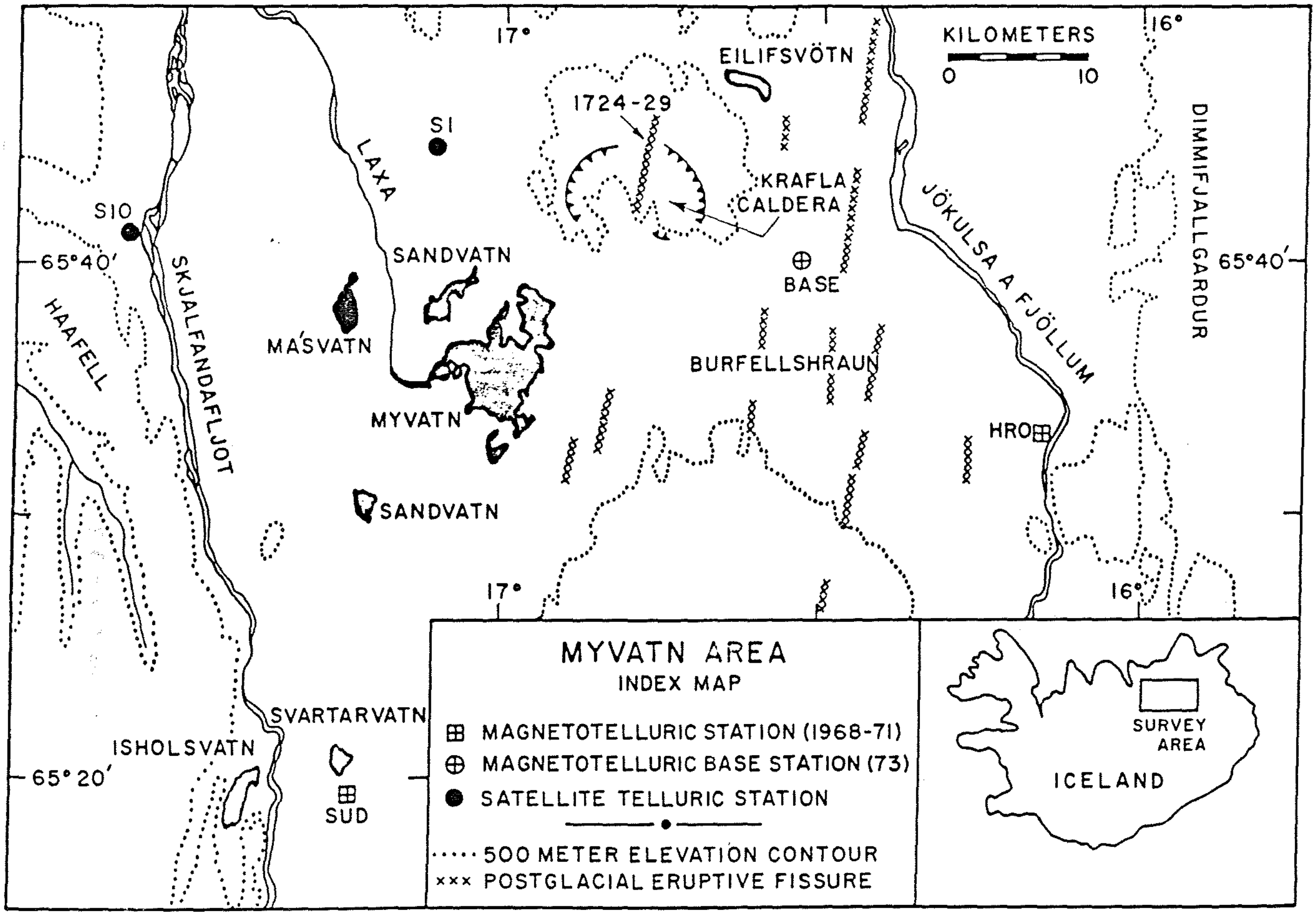

Fig. 11. The northern neovolcanic zone in Iceland showing Krafla Caldera (after Thayer, 1975; Thayer et al., 1980). Data from the site HRO are described in detail in Thayer et al. (1980), as is the interpretation of data from the base site, and the satellite sites S1 and S10. In general, features of the electrical structure beneath these sites are similar to those for the other neovolcanic zone sites shown in Figs. 13 and 14 below.

tive form of magnetic gradiometry in applying $\Delta Z/\Delta H$ ratios for magnetic bay-type events. More refined models of the deep electrical structure of the region were provided through subsequent broad-band magnetotelluric studies by Hermance and Grillot (1970, 1974), Grillot (1973), Thayer (1975), Hermance et al. (1976), Beblo and Bjornsson (1978, 1980) and Thayer et al. (1980). Electrical measurements of the upper crust ($d \leq 5$ km) using active techniques are summarized by Bjornsson (1976).

Hermance and Grillot (1974) concluded from magnetotelluric data from southwest Iceland that not only were temperatures in the upper mantle beneath this region higher than those predicted theoretically for the ocean basins and continental areas (Figure 12), but also the geothermal gradient in the mantle was less than a few °C/km, or nearly two orders of magnitude smaller than the geothermal gradient in the crust. Such low temperature gradients at shallow depth in the upper mantle, along with the implications from the teleseismic studies described above, suggest that the anomalous temperature regime beneath Iceland shown in Figure 12 extends over a minimum depth range of several hundred kilometers. Clearly, such a temperature profile calls for a significant revision of concepts regarding the petrological evolution of crustal material along the boundaries of accreting plates.

It is interesting to note that a number of sites in the neovolcanic zone exhibit a strikingly similar magnetotelluric response (Grillot, 1973; Thayer, 1975). Hermance et al. (1976) compared data from sites in the neovolcanic zone with data from one of the older (10^7 years B.P.) geologic provinces in Iceland. At the longest periods the data tend to appear quite similar, as shown in Figure 13; however, at shorter periods a significant difference can be seen.

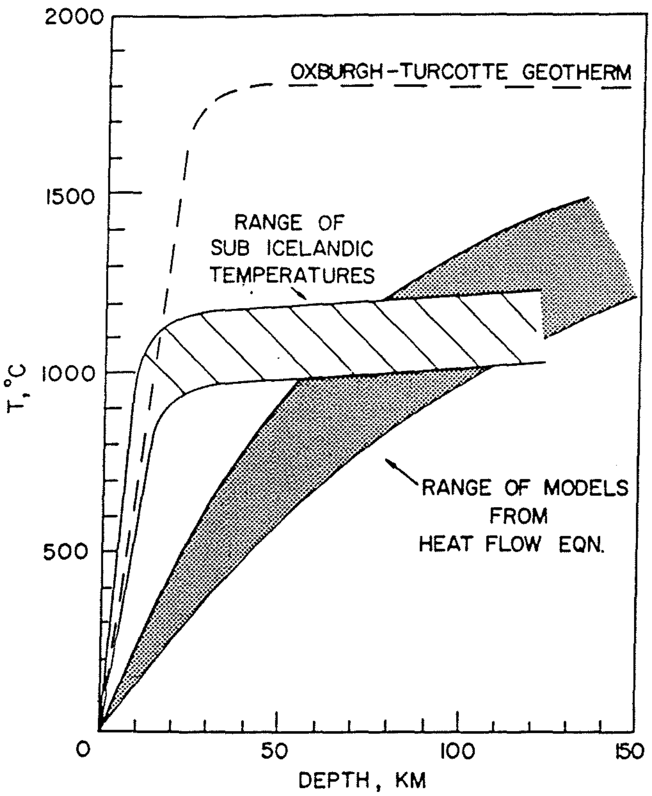

Fig. 12. The range of possible temperatures beneath Iceland as inferred from a synthesis of magnetotelluric, heat flow and seismic data (after Hermance and Grillot, 1970; 1974). Upper mantle temperatures beneath Iceland are significantly higher than those temperatures inferred from the conductive heat flow equation (e.g. Ringwood, 1975), whereas they are significantly lower than the temperatures implied by the thermal boundary layer model of Oxburgh and Turcotte (1968). The distribution of temperatures is much closer to that suggested by the kinematic modeling of Bottinga and Allegre (1976).

In order to explore possible reasons for this systematic difference, Hermance et al. (1976) applied a Monte Carlo inversion scheme to the data in Figure 13. For a given set of measurements a range of uncertainty was specified. Theoretical apparent resistivity values were then generated for a random set of layered models. Those models that generated values falling within the range of uncertainty of the observations were accepted and those that generated values outside the range of uncertainty were rejected. The results of two Monte Carlo runs are shown in Figure 14. Both runs were constrained to calculate values from models having the same range of possible resistivity. In one case, however, models were accepted that generated values falling within the envelope of data from the Tertiary basalt province; in the other case, values had to fall within the envelope of the data from the neovolcanic zone. These models were also constrained to have a constant value of 600 ohm-m for the uppermost 750 m (as required by surface active resistivity measurements). Similarly, below 100 km the value of resistivity was constrained to be 40 ohm-m. The resistivity in each of the three layers at intermediate depths was allowed to vary over the permissible ranges indicated. The only layer that shows a significant difference in the acceptable values of resistivity is that extending from 6 to 12 km in depth. Since this is approximately the base of the crust (Palmason and Saemundsson, 1974), we conclude from this study that the resistivity at the base of the crust is probably within the range of 5 to 10 ohm-m beneath the neovolcanic zone, whereas the resistivity beneath the Tertiary province is probably within the range of 15 to 30 ohm-m.

What causes a systematic lowering of the bulk resistivity of the deep crust beneath the neovolcanic zone? Three possibilities exist:

1. We could be seeing the effect of hydrothermal pore fluids. This seems improbable because the high P-wave velocity (6.5 km/sec) suggests a material having relatively low porosity. Moreover, the lowest resistivities that Hermance (1973) was able to synthesize using a highly idealized version of an electrolytic pore fluid model were above 12 ohm-m.

2. Temperatures may be high enough that electrical conduction in the solid rock lowers the resistivity to 5 ohm-m or so. Again, the results of Hermance (1973) suggest that with a geothermal gradient of 100°C/km one encounters resistivities higher than 100 ohm-m at depths less than

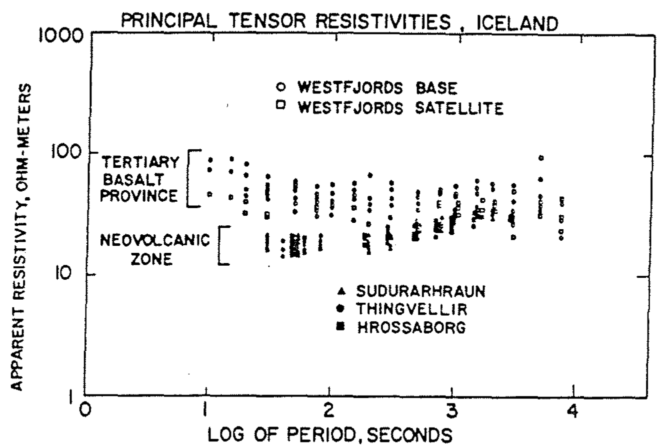

Fig. 13. A comparison between apparent resistivity data as a function of period from two sites in the Tertiary basalt province (Westfjords) and three sites in the currently active neovolcanic zone. These are principal resistivity values for the component which is felt to be least disturbed by lateral inhomogeneities (Hermance et al., 1976).

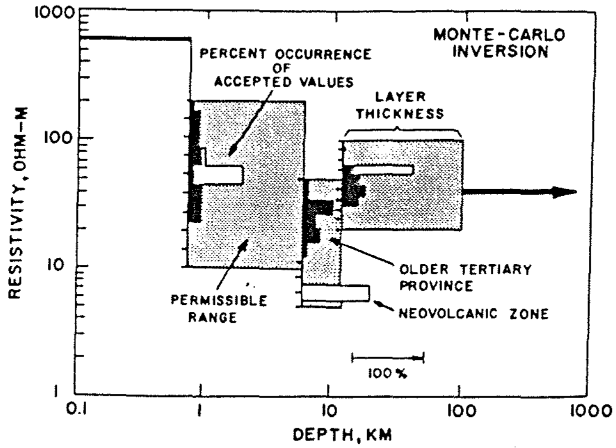

Fig. 14. Monte Carlo inversion of the two sets of data in the last figure. The surface layer and deepest layer were constrained to be 600 ohm-m and 40 ohm-m respectively. Theoretical models were calculated for which the resistivity in each layer could assume any of the values indicated by the tick-marks along the left-hand margin of the permissible range for each layer. The histograms show the percent occurrence of layer resistivities which generate theoretical values within the acceptable range of the observed data.

10 km. These values are too high to be compatible with the present results.

3. Perhaps there is a small fraction of basalt melt present in the crust on a regional scale beneath the neovolcanic zone. This would not be unlikely, considering the extensive current volcanism there. Palmason (1973), extending the earlier work of Bodvarsson and Walker (1964), concluded that a kinematic model with continuous replenishment of source material along a crustal spreading zone explained many of the geologic features in the area as well as the heat-flow observations across northern Iceland. Since crustal spreading is likely to occur over a relatively broad region, it is equally likely that the injection and accumulation of magma also occurs over a broad region. The present precision of seismic data seems able to accommodate such a possibility.

Figure 15, a schematic diagram from Thayer et al. (1980), shows the basic elements of the magnetotelluric interpretation within the framework of other investigations in Iceland. Seismic layers 1 and 2, to the left of the figure, represent the transitional zone from uncompacted pyroclastics at the surface to relatively competent basaltic material at a depth of 4 km (Palmason and Saemundsson, 1974). Layer 3 represents the lower crust (V_p = 6.5 km/sec), with the mantle ($V_p \sim$ 7.2 km/sec) at a depth of 10 km. Zones I through IV, on the right-hand side of the figure, are based on the regional electromagnetic studies described above. Zone I is a regional hydrothermal zone associated with seismic layers 1, 2 and perhaps the top of layer 3. The range in resistivity values for this layer (25-30 ohm-m) and the appropriate seismic velocities are compatible with electrolytic conduction in high-temperature hydrothermal fluids distributed in pores and cracks in relatively permeable crustal material (Hermance, 1973; Bjornsson, 1974). This leads one to the conclusion that the dominant proportion of hydrothermal activity in the Icelandic crust is regionally induced by the enormously high geothermal gradients associated with the neovolcanic zones (100-165°C; e.g. Palmason, 1973).

Zone II, beneath the zone of regional hydrothermal activity, is associated with seismic layer 3 (V_p = 6.5 km/sec). This zone seems to represent an intensely intruded region (perhaps consisting of up to 100% dikes), as proposed by Walker (1975) on geologic grounds and by Palmason (1973) on the basis of kinematic heat-flow models. Since the electrical resistivity ($\rho \geq$ 30 ohm-m) at this depth can be explained by less than a 2-5% concentration of melt (Waff, 1974; Shankland and Waff, 1977; Hermance, 1979), this implies that, although the zone could have been created completely by dike intrusives in response to crustal spreading, only a few percent of the intrustives are at any time molten.

Zone III is perhaps the most significant feature of the magnetotelluric interpretation: its manifestation is a conducting layer ($\rho <$ 10 ohm-m; $\Delta T <$ 4 km) which appears to be associated with the crust-mantle boundary in this area (depth $\sim$ 10-15 km). This layer is interpreted as a zone of magma accumulation (Hermance et al., 1976; Beblo and Bjornsson, 1978, 1980). The concentration of melt may be as low as 7.5-15% if distributed over a vertical thickness of 4 km, or it may be as high as 100% if distributed over a thickness of only 200 m (the resolution of the magnetotelluric method is not sufficient to discriminate between these alternatives). Moreover, we cannot determine from magnetotelluric measurements alone whether this zone consists of a single coherent lens or of a number of discrete tabular-like bodies of either microscale or macroscale proportions. Nevertheless, we can say with some confidence that over a depth interval of 4 km (at a depth of 10 km), there is the vertically integrated equivalent of at least 200 m of molten magma, and that, because this magma zone appears beneath a number of magnetotelluric sites over a large area, it must be a steady-state feature of the neovolcanic zone. The presence of this zone of magma has important implications for kinematic models depicting crustal spreading in Iceland (e.g. Palmason, 1973; Daignieres et al., 1975). Thayer et al. (1980) suggest that crustal underplating, as envisaged by Lachenbruch and Sass (1978) for the Basin and Range province in the western United States, may be playing an important role in crustal development in Iceland.

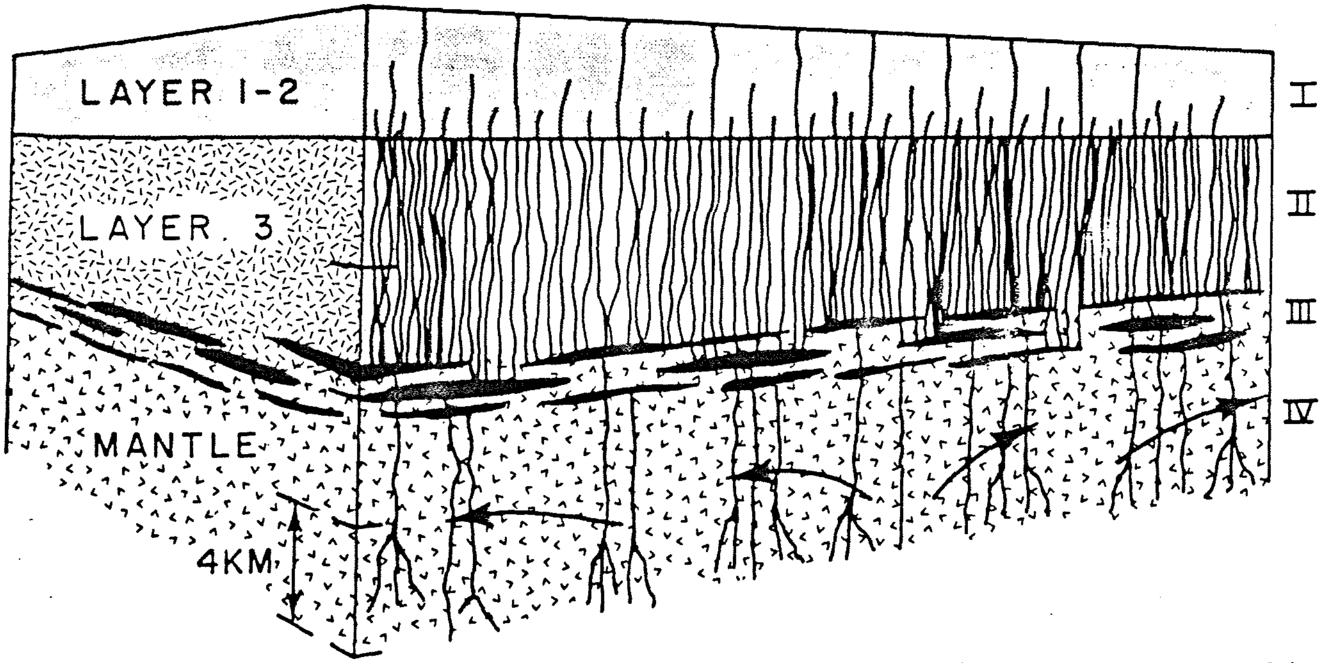

Fig. 15. A physical model for the northern neovolcanic zone in Iceland (after Thayer et al., 1980). The layers indicated on the left side of the figure are based on the seismic studies of Palmason (1971). On the right side of the figure, the Roman numerals denote zones which have distinctive electrical properties. Zone I is a permeable surface layer in which hydrothermal circulation is quite pronounced. Zone II has at most a few percent partial melt present. Zone III represents a conducting layer at the base of the crust which is thought to be a zone of magma accumulation and crustal underplating. Zone IV, the upper mantle, has only a few percent partial melt but seems to be in a state of mass or convective transport.

Underplating involves the accretion of a zone of molten magma to the base of the crust which, as it cools, leads to crustal thickening. Although deserving of a great deal more attention than the process has received up to the present time, one can nevertheless estimate the significance of such a process if it were operative. Based on the magnetotelluric interpretation described above, the crustal basal zone of magma accumulation may have a thickness on the order of 4 km. This zone is not totally molten at any one time, of course, because of transient cooling effects; in fact, the magnetotelluric interpretation suggests an effective thickness for the molten fraction of 200 m over a depth of approximately 5 km. Hermance et al. (1976) and Thayer et al. (1980) proposed, however, that melt is being more or less continuously supplied to this layer, so that an equivalent fraction of melt (.2 km/ 5 km $\sim$ 4%) is a steady-state feature of this zone. If one assumes that a total crustal spreading velocity of 1 cm/year is achieved over a lateral distance of 50 km in the northern neovolcanic zone (cf. Palmason, 1973; or Palmason and Saemundsson, 1974), one can invoke the mass-flow balance relation of Lachenbruch and Sass (1978) to argue that a two-dimensional steady-state flow of molten material representing a total flux of 4 km · 1 cm/year must be occurring over a lateral distance of 50 km. If this material simply solidifies at constant temperature, it releases a latent heat of fusion of approximately 100 cal/gm or 300 cal/cm^3. The average heat released from this underplating process, q_u, which is contributed to the regional surface heat flux, can be expressed by

$$q_u = \frac{L \cdot v_d \cdot T \cdot \ell}{W \cdot \ell} \tag{1}$$

where L is the latent heat of fusion (300 cal/ cm^3), v_d is the crustal drift velocity (1 cm/ year), T is the thickness of the underplating zone (4.10^5 cm), ℓ is a characteristic dimension along the strike of the spreading zone (a term which cancels), and W is the width of the active zone of extension ($\sim 5 \times 10^6$ cm). For the values of the parameters indicated above,

$$q_u \sim 0.8 \text{ hfu} \tag{2}$$

where 1 hfu = 10^{-6} cal/cm^2/sec. Clearly this component of the heat flux is strongly dependent on the thickness of the underplating zone, T, as well as on the total width of the region over which crustal extension is distributed, W.

If, for example, W is less than the figure assumed above, and has a value between 15 and 25 km, say, the heat flow contribution will be correspondingly greater: on the order of 2 to 3 hfu.

In short, it appears that crustal underplating, which is simply the accretion of material in the molten phase to the base of the crust along a zone of crustal extension, may play an important role in ridge processes. Although hitherto neglected as a significant contributor to the dynamics of accreting plate boundaries, the presence of such an underplating layer beneath Iceland suggests that an analogous feature may also be present beneath the submarine portions of the oceanic ridges.

In fact, Bottinga and Allegre (1976) have reflected on the relatively unsuccessful attempts by seismologists to detect and confirm the presence of high-level magma systems in the crust beneath the mid-ocean ridges. If such systems have the characteristics and dimensions typified by our studies in Iceland, they indeed may very likely represent difficult targets to resolve seismically. It is interesting in this regard to note that one of the features of the asthenospheric-upwelling model of Bottinga and Allegre (1976) is that far more liquid is produced by partial fusion in the upper mantle beneath the ridge crests than is erupted at the surface; perhaps 4 to 5 times as much liquid is formed as is actually erupted. As a consequence of these studies, along with the magnetotelluric interpretation in Iceland, one has independent lines of investigation converging on very similar models for magma genesis in the upper mantle.

The presence of the magma accumulation zone also has important implications for models explaining the emplacement of magma along fissure swarms up to 70 km away from central volcanoes (Sigurdsson and Sparks, 1978). These results suggest the possibility that, rather than flowing horizontally over these extreme distances (70 km), magma may flow vertically (< 10 km) from a pre-existing magma zone.

Zone IV, the upper mantle, is characterized by relatively small thermal gradients (Hermance and Grillot, 1974) indicative of the bulk transfer of material through diapiric-like motion. The present interpretation (Thayer et al., 1980) suggests a higher concentration of melt in the upper few tens of kilometers of the mantle than in the region directly below. Presumably the uppermost mantle is supplying magma to the accumulation zone which is underplating the crust as well as to the dike intrusives which compensate for the actual crustal spreading itself.

Within the resolution of present data, the entire area of Iceland (120,000 km^2) is underlain by mantle material at a depth greater than 10-15 km which is laterally uniform in composition and temperature - at least considerably more uniform than the crust appears to be. Such a result is similar to the results of magnetotelluric measurements in rift areas elsewhere, such as the Rhine graben (Scheelke, 1974).

The Rio Grande Rift

The Rio Grande rift (Figure 16) is characterized by a contiguous series of basin-type structures extending from southern Colorado through central New Mexico into northern Chihuahua, Mexico (Chapin, 1979). Throughout its development, the rift has undergone periods of extensive volcanic activity. Early volcanism was characterized by calc-alkalic andesites, but beginning approximately 5 million years ago and continuing to the present, basaltic lava flows have become more widespread from southern Colorado to northern Mexico (Chapin and Seager, 1975).

A number of geophysical studies suggest that major crustal and mantle processes are contributing to the evolution of the Rio Grande rift (Cordell, 1978). Some of the earliest geophysical studies in this region that dramatized the anomalous character of the rift feature itself, were the geomagnetic variation studies of Schmucker (1964, 1970) in the southern rift, and similar studies across other portions of the rift by Reitzel, Gough and Porath (see the review by Gough, 1974). This work clearly established the presence of an electrical conductivity anomaly associated with the axis of the extensional feature.

The heat flow studies of Reiter et al. (1975), Decker and Smithson (1975), Sass et al. (1976) and Reiter et al. (1978, 1979) reveal a belt of

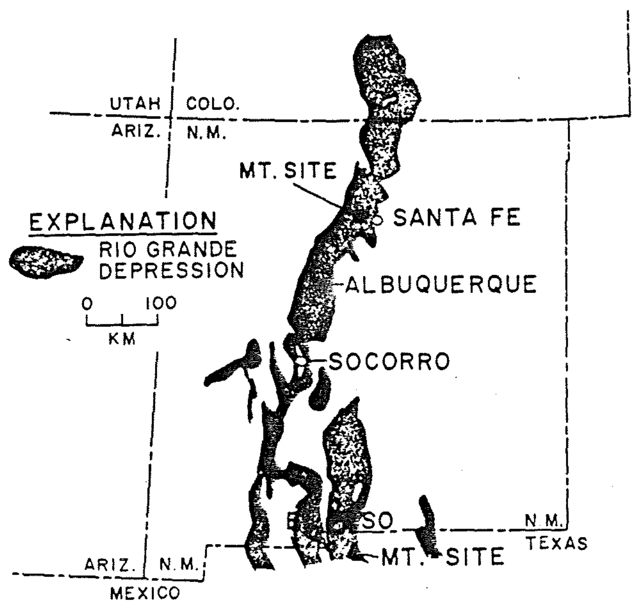

Fig. 16. The Rio Grande rift in New Mexico (after Chapin and Seager, 1975), showing the location of magnetotelluric deep-sounding experiments.

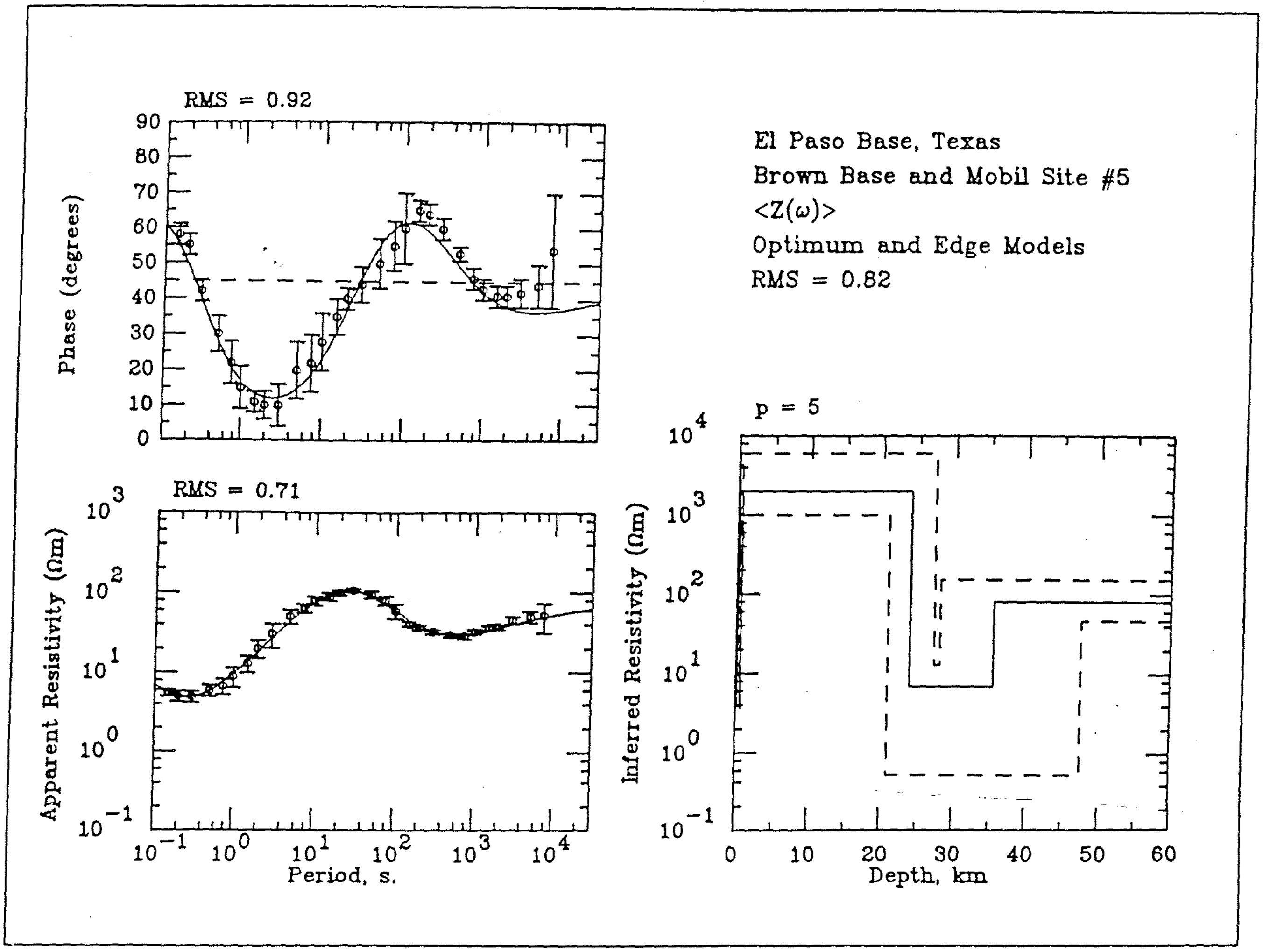

Fig. 17. Magnetotelluric data and interpretation from the southern Rio Grande rift near El Paso, Texas. The apparent resistivity, phase and associated error bars are shown as points along with the response of the optimum model determined by a nonlinear least-squares fit to both phase and apparent resistivity (continuous line). The optimum model (continuous line) and the extreme bounds on other possible models using up to the fourth eigenvalue (dashed lines) are shown in the lower right-hand graph. The dashed lines do not show allowable models, but define the allowable range for the model parameters (after Hermance and Pedersen, 1980).

anomalously high heat flow associated with the rift, particularly its western margin. Calculations by Cook (1975) using transient thermal models suggest that there have been three epochs of magma intrusion; 25 million years, 13.2 million years and 2.5 million years before present. The earliest thermal event, in this model, is associated with early rifting and calc-alkalic andesitic volcanism, while the latest intrusive episode is associated with the initiation of recent basaltic activity.

The regional seismic refraction interpretation of Toppozada and Sanford (1976) proposes a two-layer crust (V_p = 5.8 km/sec, 19 km thick; V_p = 6.5 km/sec, 21 km thick) under which lies a mantle having a lower-than-normal P-wave velocity of 7.9 km/sec. There is a suggestion that the crust thins to approximately 35 km beneath the north-central portion of the rift (Olsen et al., 1979), and that the upper mantle may have a somewhat lower velocity (7.6 km/sec); this model is in essential agreement with the surface-wave dispersion studies of Keller et al. (1979). On a smaller scale, Sanford and his colleagues (Sanford et al., 1973; Sanford et al., 1977; Shuleski et al., 1977; Rinehart et al., 1979) have used reflected P- and S-phases from local micro-earthquakes near Socorro to delineate a liquid-like layer, which they presume to be magma, having a thickness of less than several

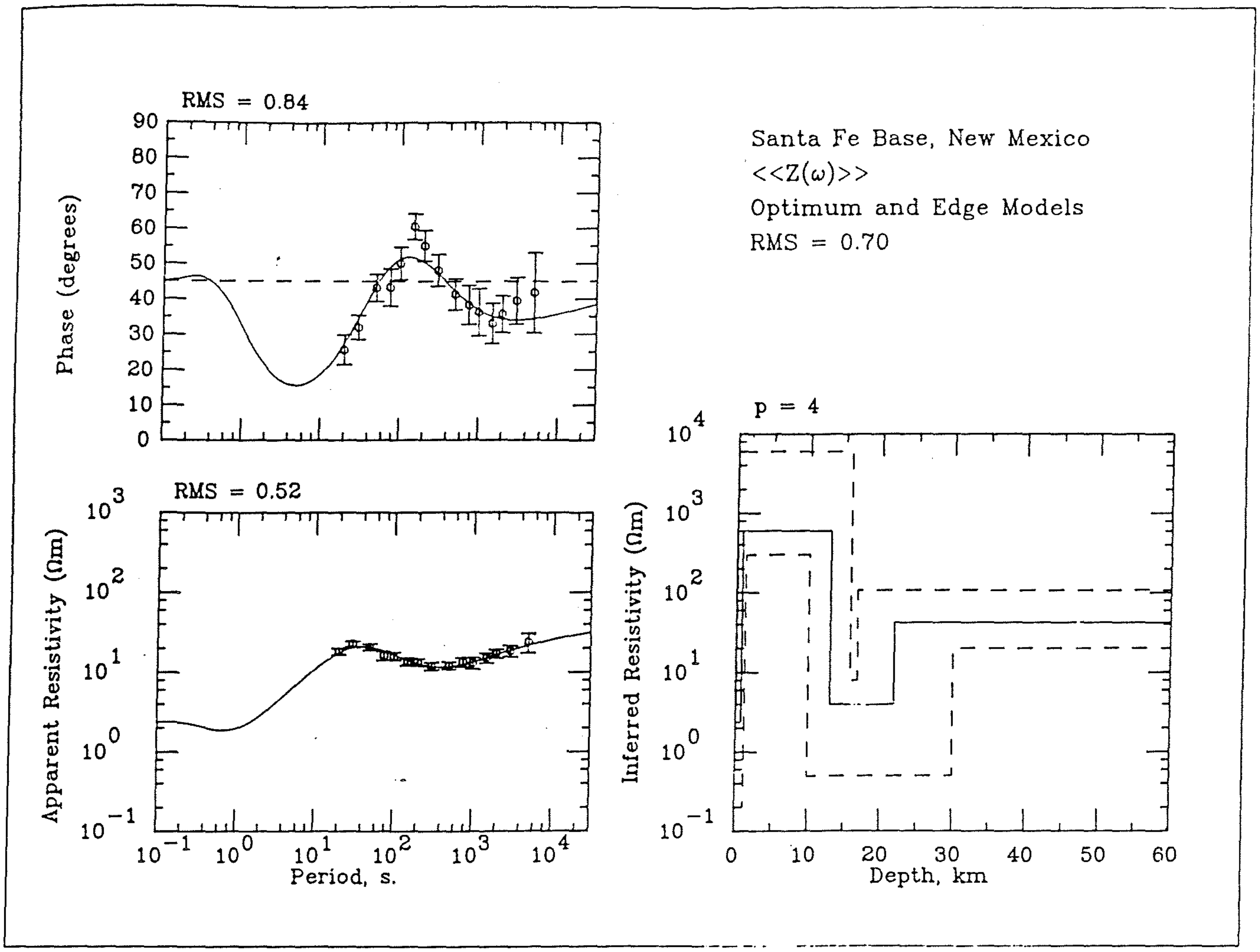

Fig. 18. Magnetotelluric data and interpretation from the north-central Rio Grande rift near Santa Fe, New Mexico (see Fig. 16). For an explanation of the presentation format, see the caption of Fig. 17.

kilometers, at a depth of 19 to 20 km beneath this portion of the rift. This zone has been mapped over a geographical extent of at least 1700 km^2. The presence of a coherent layer at this depth has recently been confirmed by the COCORP deep seismic reflection profiling in the Socorro area (Krumhansl et al., 1978; Brown et al., 1980).

Pedersen and Hermance (1976), using magnetotelluric measurements, suggest it is possible that this structure, or features related to it, may have a much wider extent. They report the presence of a conducting layer beneath El Paso at a depth of 21 to 28 km (Figure 17). Pedersen and Hermance (1978) report a similar layer beneath Santa Fe at a depth of 10 to 17 km (Figure 18). This work opened the possibility that the zone of enhanced electrical conductivity is contiguous with the tabular body previously delineated seismically by Sanford and his colleagues, and probably has a similar generic cause (Hermance and Pedersen, 1980).

Recent seismic refraction work by Olsen et al. (1979), near the site of the magnetotelluric experiment at Santa Fe described above, suggests the presence of a thin zone of anomalously low shear-wave velocity at mid-levels in the crust beneath the north-central rift. They feel this feature is a manifestation of a widespread intracrustal low-rigidity layer which is also generically related to the Socorro magma body.

Although the electrical conductivity itself can be affected by (1) electrolytic conduction in hydrothermal pore fluids, (2) conduction in the bulk silicate material itself, or (3) solid conduction through hydrated phases such as layered silicates, we find the suggestion offered by Sanford and his colleagues (e.g. Rinehart et al.,

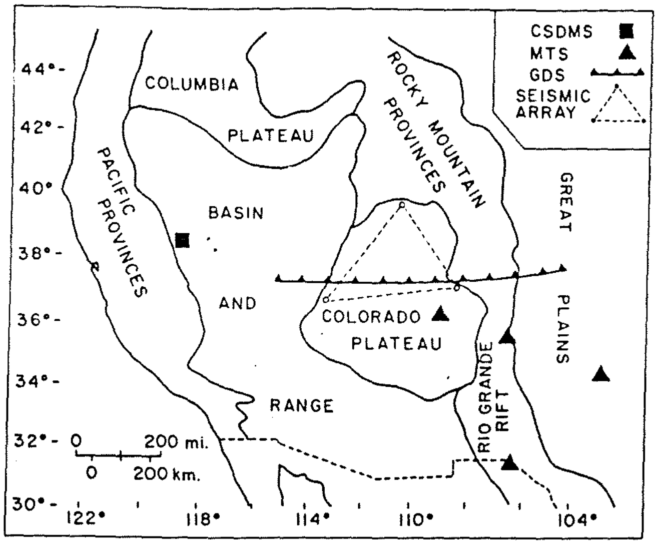

Fig. 19. A simplified tectonic map of the western United States. The sites for the following experiments are shown: the controlled-source deep magnetic sounding experiment (CSDMS) by Lienert and Bennett (1977); the geomagnetic deep-sounding (GDS) profile line of Porath (1971) and Porath and Gough (1971); the seismic array experiment of Bucher and Smith (1971) and the location of magnetotelluric sites (MTS) occupied by Brown University.

1978) for the Socorro structure the most attractive; (4) the intracrustal zone represents a generic class of phenomena associated with the accumulation of basaltic melt derived from sources at greater depth. If this is the case, we may be seeing the effects of a thin zone (1 km) comprised totally of melt, or a thicker zone containing as little as a 15-35% melt fraction.

Deep Structure of the Rio Grande Dome. As is true in many rift provinces, the Rio Grande rift is associated with regional uplift (doming) which is presently active (Cordell, 1978; Reilinger et al., 1979). The actual uplifted region bisected by the rift (Figure 19) extends laterally from the Colorado plateau to the High Plains provinces of western North America (ibid.). Long-path-length seismic surface wave studies (Biswas and Knopoff, 1974) indicate a general homogeneity of the mantle beneath this region (to a depth greater than 100 km). The seismic data presently available suggest that the mantle from the base of the crust to a depth greater than 100 km is grossly similar beneath the Colorado plateau, the Rio Grande rift and the High Plains province immediately adjacent. This region is characterized by low compressional and shear wave velocities, and apparently low densities as well.

It is instructive, therefore, to compare long-period magnetotelluric data from the Colorado plateau (Pedersen and Hermance, 1980) with similar measurements in the rift itself. The data from the Colorado plateau near Farmington, New Mexico (Figure 20) characterize the province as consisting of conductive surficial sediments a few kilometers thick, underlain by a more resistive crystalline crust down to about 28 kilometers. Beneath this zone the average resistivity is 13 ohm-m down to at least 130 km. On the other hand, seismic evidence indicates a crustal thickness of approximately 40 km for the interior of the plateau. The electrical data therefore require the transition from resistive crust to more conductive mantle material to occur in a lower crustal layer approximately 12 km thick.

Pedersen and Hermance (1980) point out that an upper limit on possible temperatures in the upper mantle is provided by comparing the range of possible upper mantle resistivities (10-15 ohm-m) to the laboratory measurements of Duba et al. (1974) on solid olivine; temperatures in excess of 1700°C would be required. Allowing for the uncertain effect due to impurities and more conductive grain boundaries, temperatures in excess of 1500°C are needed to yield resistivities less than 15 ohm-m (Shankland and Waff, 1977, Figure 2). Since reasonable temperatures in the upper mantle at the depths considered here are expected to be from 1000°C to 1400°C (Lachenbruch and Sass, 1977), it seems plausible that the bulk resistivity of the mantle may be influenced by the presence of some degree of partial melt (Waff, 1974). Shankland and Waff (1977) cite both field and experimental observations to support the hypothesis that such a melt fraction is thoroughly interconnected. Recently, petrologic and thermodynamic arguments have been advanced by Waff and Bulau (1979) to support a melt configuration which is thoroughly connected along grain edges, provided that the melt is in chemical equilibrium with the heat rock.

Using this model for the mantle beneath the Colorado plateau and the calculations of Shankland and Waff (1977), Pedersen and Hermance (1980) argue that a resistivity in the range 10 to 15 ohm-m would imply a mantle temperature of approximately 1200°C at 50 kilometers depth, assuming a water content of 0.1% for the upper mantle. At 100 kilometers depth the temperature would be about 1300°C. The partial melt fraction would vary from 8% to 3% over the same depth interval. From this analysis, a mantle geotherm for the Colorado plateau may be calculated yielding a gradient of 2.5°C/km (with an uncertainty of about the same magnitude).

An alternative estimate of the temperature gradient in the upper mantle was made by Pedersen and Hermance (1980) following a procedure originally outlined by Hermance and Grillot (1974), who argued that relative values of temperature are more precisely determined from deep electrical studies than are absolute values of temperature. Hermance and Grillot (1974) showed that the geothermal gradient, G_T, is related to the logarithmic gradient of resistivity through the expression

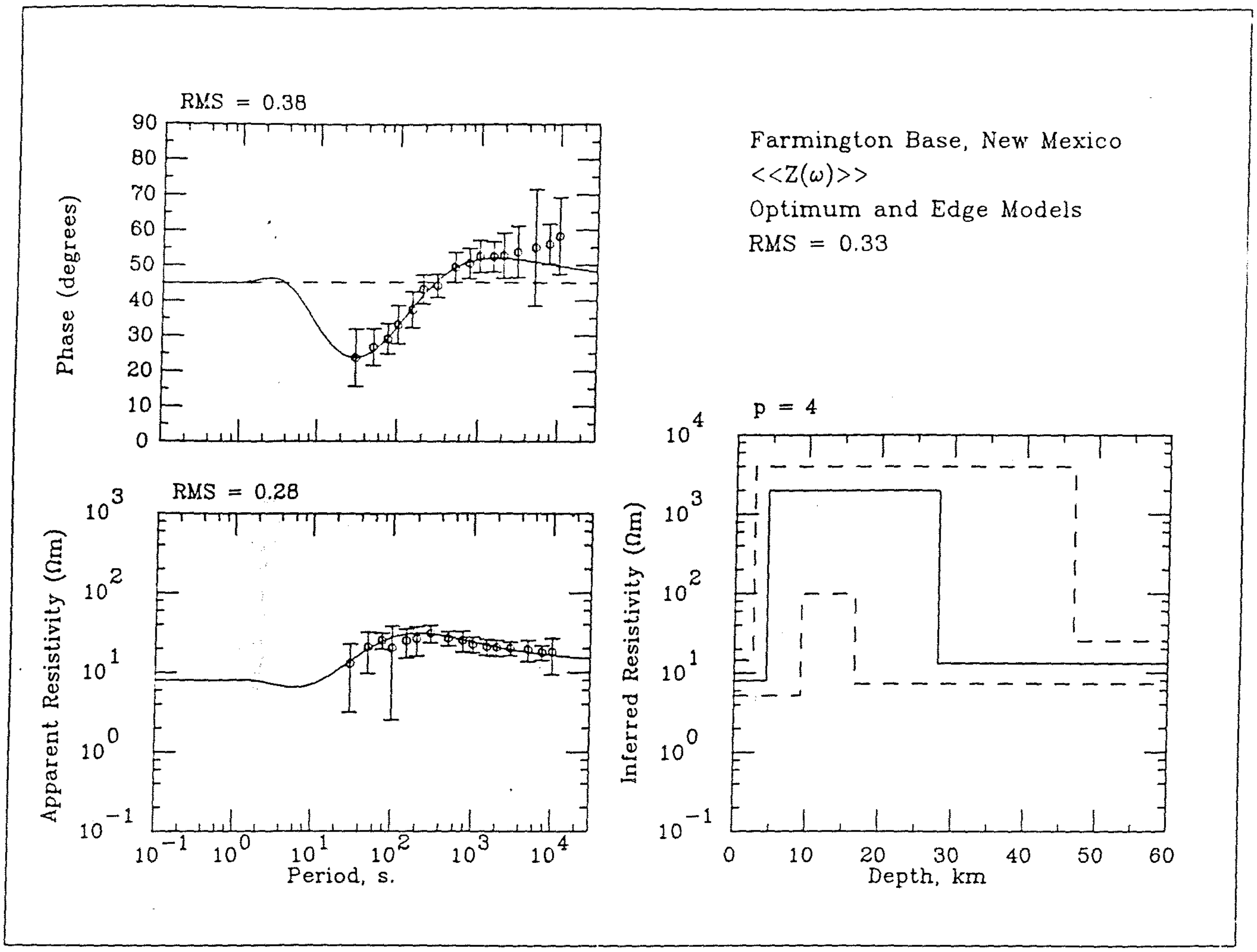

Fig. 20. Magnetotelluric data and interpretation from the Colorado plateau near Farmington, New Mexico (see Fig. 19). For an explanation of the presentation format see the caption of Fig. 17 (after Pedersen and Hermance, 1980).

$$G_T = (kT_1^2/E\Delta z)\ln(\rho_1/\rho_2) \qquad (3)$$

where k is the Maxwell-Boltzmann constant, T_1 is the temperature (°K) at the top of a vertical section, Δz is the thickness of the section, ρ_1 and ρ_2 are the resistivities at the top and bottom of the section, respectively, and E is the activation energy (on the order of 1-2 eV).

By considering the extreme range of possible resistivity gradients in the upper mantle, Pedersen and Hermance (1980) determined that the maximum variation in resistivity may be a factor of 3 over a depth range of 120 kilometers. Assuming a temperature of 1000°C at 30 kilometers depth and applying the above analysis, an estimate of 1.3°C/km was determined for the maximum geothermal gradient beneath the Colorado plateau. This value is consistent with the estimate obtained using the approach of Shankland and Waff (1977) described above. These low geothermal gradients suggest that vertical heat transfer beneath the Colorado plateau and the Rio Grande dome is dominated, as in Iceland, and presumably in many other rift areas, by a convective material transport mechanism.

Additional geophysical evidence in the form of uniform surface heat flow values of 1.5 to 1.8 hfu in the interior of the plateau (Reiter et al., 1978), the absence of long-wavelength aeromagnetic anomalies, and the need for isostatic compensation in the Colorado plateau to be largely in a low-density upper mantle (Thompson and Zoback, 1979) all supports the hypothesis of moderately high upper mantle temperature. The low (when compared to the stable continental interior) upper mantle compressional and shear velocities of 7.8 and 4.25 km/sec, respectively (Bucher and Smith, 1971; Keller et al., 1978),

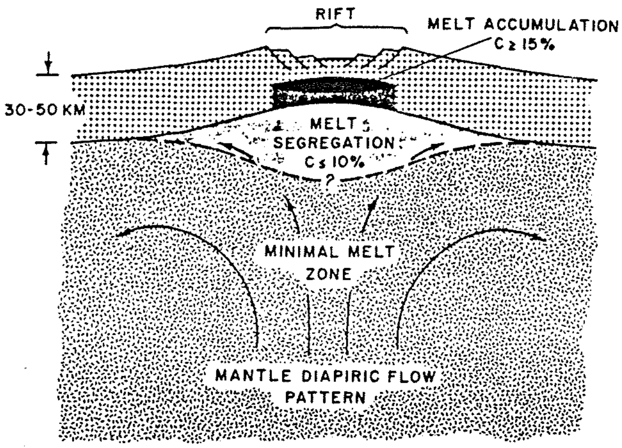

Fig. 21. A conceptual model for physical processes in the upper mantle and deep crust beneath active rift zones. The three zones indicated (the melt accumulation zone, the melt segregation zone, and the minimal melt zone exhibiting diapiric flow) each have distinctive geophysical/electromagnetic properties.

likewise indicate moderately high upper mantle temperature and are thus consistent with the magnetotelluric data. A comparison with electrical data available for the surrounding tectonic provinces indicates that the upper mantle beneath the Colorado plateau has resistivities at least as low as those for the Basin and Range, Rio Grande rift, and Great Plains, and may in fact be substantially more conducting than any of these surrounding tectonic provinces (Pedersen and Hermance, 1980).

Summary and Conclusions

A Conceptual Model for Deep Crust and Upper Mantle Processes

Many of the major rift zones of the world are very similar in morphotectonic appearance and development (Illies, 1970). Although significant differences exist in the character and degree of regional doming, rifting and magma genesis (Logatchev, 1978), in the present discussion we seek a basis for a hypothetical model that, while not satisfying all of the features of any specific rift zone, accounts for many of the features common to most of the rift zones studied.

Such a conceptual model is portrayed in Figure 21. Recent magnetotelluric investigations in Iceland and the Rio Grande rift (including the eastern Colorado plateau) support the idea of Illies (1970) that beneath rift systems (both mid-oceanic and intracontinental), ascending masses of material from the mantle are intimately coupled with regional doming and the morphotectonic development of rift features. This phenomenon is seen electromagnetically as essentially a constant value of electrical resistivity over a very large range of depths (from 30 to over 100 km), from which is inferred a small change in temperature over the same depth interval. Since crustal temperature gradients are on the order of 40 to 100°C/km in these regions, and we infer upper mantle gradients of less than a few °C/km, heat transfer mechanisms must be extremely efficient (i.e. by mass transport) in the upper mantle.

At the top of the mantle diapir, lowered values of the bulk resistivity indicate in some areas (e.g. Iceland) a slight enhancement of a partial melt fraction which may be associated with the low-velocity pillow beneath many rifts. We should keep in mind the fact that the seismic structure beneath some rifts appears to be systematically different from that beneath others. For example, the Baikal rift and the Rhine graben are associated with a low-velocity pillow at the base of the crust having a thickness of only 15-25 km. On the other hand, Iceland, the East African rift, and the Basin and Range province of western North America (of which the Rio Grande rift seems to be a part) all appear to have low-velocity material extending to very great depths (> 100 km).

The present interpretation of magnetotelluric and geomagnetic deep-sounding data from the Baikal rift and the Rhine graben seems also to favor a conductivity anomaly having a limited thickness in the crust and/or upper mantle. In contrast, magnetotelluric data from Iceland and from the vicinity of the Rio Grande rift indicate anomalous mantle conductivities over a very large depth interval (> 100 km). Electromagnetic studies in the East African rift are inconclusive on this point; although low resistivities are indicated for the full range of periods over which measurements were made, the effective depth of resolution is less than 35 km at the longest periods for which reliable data are available. Because of the close similarity between the electrical structure in the crust beneath Iceland and that beneath the most active portions of the East African rift, as well as because of the similarity in the seismic character of the upper mantle beneath these regions which extends to great depth (> 100 km), one might tentatively suggest that the character of the upper mantle beneath the East African rift is not unlike that beneath the Rio Grande rift or that beneath Iceland. Clearly this places important constraints on the possible distribution of partial melt in the upper mantle and, more importantly, on the physical character of possible diapiric motions responsible for regional tectonics.

On the other hand, areas such as the Baikal rift and the Rhine graben seem to have minimal melt in the mantle beneath the zone of melt segregation (Figure 2), whereas there seems to be clear evidence in the electrical data for a significant fraction of melt in the vicinity of the low-velocity pillow in these regions.

It does not seem plausible that melt, having segregated and formed a concentration of greater than a few percent of the total volume, can be dynamically stable (Walker et al., 1978). Therefore, melt in the pillow at the base of the crust probably migrates either laterally as part of the diapiric flow pattern, or vertically in response to a combination of buoyancy and lithostatic pressures, and accumulates at a preferred level in the crust which is presumably hydrostatically controlled.

Intracrustal low-velocity layers appear to be typical of continental rifts (Puzirev et al., 1978). Apart from the Baikal, the existence of such layers is noted under the Rhine graben (Mueller, 1970) and beneath the Rio Grande rift (Olsen et al., 1979). Although a low-velocity zone does not appear to have been detected at intracrustal depths beneath the East African rift, this may be due to a complexity of structure which masks the subtle effects of such a layer, rather than to a fundamental difference in the physical character of the rift itself (Griffiths, 1972; Long et al., 1973). Magnetotelluric interpretations place this zone of magma accumulation at the base of the crust in Iceland, whereas, from a combination of seismic and deep electrical studies, it seems to be at intracrustal levels beneath the Rio Grande rift, the Rhine graben and the Baikal rift.

The nature of the migratory zone (Figure 21) separating the region of melt segregation in the mantle from the zone of melt accumulation at the base of, or within, the crust is obscure at the present time, and its character undoubtedly varies from region to region. In some areas, such as portions of the Baikal rift and the north-central Rio Grande rift, the zone of melt accumulation seems to be a discrete layer at intermediate levels in the crust which can be clearly distinguished from its anomalous roots in the mantle. In other regions, such as the southern Rio Grande rift, Iceland and other portions of the Baikal rift, the zone of melt accumulation may be more intimately coupled to its mantle origins. The level at which magma accumulates must depend to a marked degree on its depth of origin as well as on the overall character of temperature and density gradients in the lithosphere and asthenosphere. Moreover, the frequency and mass flux with which the magma zone is replenished and sustained determines the extent to which the thermal regime of the lower crust is perturbed from a conductive environment to a convective environment (Lachenbruch and Sass, 1977).

Relationship to Extrusive Silicic Volcanism

It is well known that centers of silicic volcanism exist along some of the major rift zones such as the East Africa rift, the Rio Grande rift and Iceland. It has also been suggested by a number of workers that these silicic centers have developed through partial melting (anatexis) of crustal material through interaction with more primitive basaltic magma at depth. It is useful to speculate on the relationship of the conceptual model in Figure 21 to general aspects of silicic volcanism, as Hermance and Pedersen (1980) have done for the Rio Grande rift. Such speculation does not appear to be particularly germane to the Baikal rift (Logatchev and Florensov, 1978), but might be of interest for other of the world's rift areas.

To begin, we must recognize that many silicic centers on continents have lifetimes on the order of 10 million years (Doell et al., 1968; Smith and Bailey, 1968). Volcanism is of course not reversible, but proceeds in a specific volcanic center through a definite series of events, some of which seem to reflect similar stages among different volcanoes (Williams, 1941; Smith and Bailey, 1968). For example, the explosive eruption of a tuff phase occurs late in a volcano's history, and volcanism continues episodically (at intervals on the order of $4\text{-}6 \times 10^4$ years) within the caldera in the form of resurgent doming and rhyolite flows (Doell et al., 1968).

In our opinion, there may be a direct relationship between the high-level emplacement of basaltic magma at mid-crustal levels and surface volcanism. First, we suggest the impossibility of sustaining a single coherent silicic magma chamber for a period of 10 million years. Second, we suggest that the episodic volcanism so characteristic of the latter phases of the evolution of a volcanic center provides a clue to an important element in the evolution of these systems. In discussing Figure 21, we suggested that diapiric-type motion in the upper mantle leads

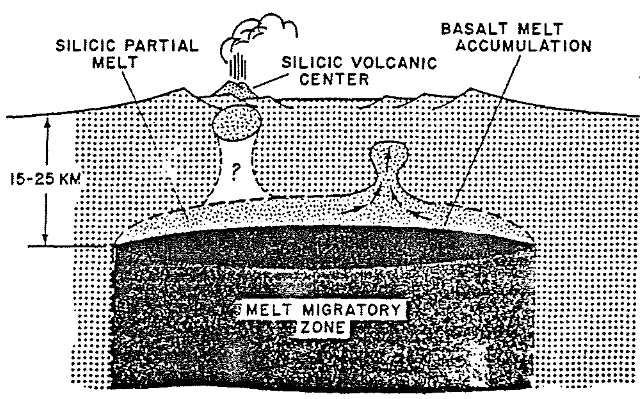

Fig. 22. A model illustrating the accumulation of basalt melt at deep to intermediate levels in the crust and its possible relationship to the remelting of the crust and the mobilization of a silicic partial melt phase. The silicic melt would be gravitationally unstable and would tend to form diapir-like ascending masses. The nature of the root connecting high-level silicic magma chambers to deeper sources in the crust remains unclear.

to partial fusion of mantle materials, resulting in the segregation and upward migration of a basaltic melt due to a combination of buoyancy and lithostatic pressures. The melt in turn reaches a preferred level in the crust (which is presumably hydraulically controlled) and a zone of magma rapidly accumulates, followed by a long cooling history. An appropriate time constant for a molten lens-like structure 2 km thick to chill below its solidus appears to be 10^4 to 5×10^4 years (Jaeger, 1964). In Figure 22, we illustrate that the emplacement of a molten (1200°C) basaltic lens at depth will perturb the surrounding country rock, which may undergo local secondary remelting (to temperatures apparently as low as 750-800°C (Bailey et al., 1976)). Assuming for the sake of illustration that the latent heat of fusion for basalt and granite are approximately equal (i.e. 100 cal/gm), then up to 400 cal/gm is available in the basalt intrusion to locally remelt the crust. In other words, in this grossly oversimplified model 1 unit of basalt can mobilize up to 4 units of silicic material.

The partial melt will segregate as a rhyolite magma (Figure 22) and, because of its lower density, will tend to rise as a silicic diapir, leading to extrusive volcanism at the surface. The episodic replenishment of the basaltic magma layer at depth is associated with the episodic silicic volcanism at the surface. Thus this model, although highly speculative, provides a basis for relating geophysically delineated structures in the deep earth, as conceptualized in Figure 21, to the genesis of major centers of silicic volcanic activity (Figure 22).

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