Electrical Conductivity under Western North America in Relation to Heat Flow, Seismology, and Structure

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Western North America between the coastal ranges and the Great Plains, and from the Mexican-United States border to the Trans-Canada Highway has been studied by means of three large two-dimensional arrays of recording magnetometers. The paper discusses the conductive structures found by these arrays in the upper mantle of the region. Highly conductive mantle material at least 100 km thick for conductivity 0.2 S/m underlies the Basin and Range Province and ridges of still higher conductivity (or greater thickness for a given conductivity) underlie the Wasatch Fault Belt and Southern Rockies. Under the Great Plains the upper mantle is more resistive and is of sub-shield type. North of the Basin and Range the Cordillera have a thin layer (10-20 km at 0.2 S/m) of conductive material in the upper mantle and the conductivity indicates a lower level of tectonic activity than in the midlatitudes of the U.S.A. Under the Colorado Plateau the conductivity is of Great Plains type or intermediate between this and that beneath the Basin and Range. Alternative conductive models satisfying variation-field anomalies are discussed with special reference to the poor depth and excellent lateral resolution of magnetometer arrays. The distribution of heat-flow is compared with that of electrical conductivity and it is shown that there is generally close correspondence and full support for association of high conductivity with high temperature. Seismological velocity-depth profiles, station terms in P and S teleseismic times and other parameters are considered and are again consistent with the conductivity and geothermal information. The case for partial melting is considered. It is strongly supported by seismological parameters in the low-velocity layer and gives the most plausible hypothesis for the contrasts of order 100 in conductivity required by the magnetometer array results. Three tentative suggestions are made concerning tectonics of the region.

1. Introduction

In recent years work with magnetometer arrays has revealed considerable structure in the electrical conductivity distribution under western North America. The conductive structures are closely related to heat flow and to seismological parameters, as this review will attempt to show. Further, all of these geophysical properties show correspondences to tectonic provinces de-
lineated from surface geology and indicate that these provinces must be understood in relation to structures between the surface and about 400 km depth.

The major tectonic provinces of the region are outlined in Fig. 3. The east front of the Rocky Mountains divides two superprovinces, those of central and western North America. Within the western superprovince this paper will make frequent reference to the Basin and Range Province, characterized by young volcanic rocks; to the uplifted Colorado Plateau and Southern Rockies; and to the Wasatch Front with its faults striking north-south through Utah at the boundary between the Basin and Range and the Colorado Plateau. Further south the boundary between the Basin and Range Province and the Great Plains is a feature of importance. In southwestern Canada and northwestern United States the most significant distinction is that between the great Plains and the Cordillera as a whole.

2. Electrical Conductivity Structure

Conductive structures have been discovered and modelled through observation of local anomalies in time-varying magnetic fields recorded by three large arrays of magnetometers. The source fields were mainly polar magnetic substorms, giving sufficient field amplitudes in the period range $20 \text{ min} < T < 120$.

![Fig. 1. Locations of three large two-dimensional arrays of magnetometers used in western North America in the years 1967-1969.](image-url)
min, but daily variation harmonics at periods 8, 12 and 24 hours were also used. The locations of the three arrays are shown in Fig. 1, with the year in which each was operated. These array studies were jointly undertaken by the University of Alberta and the University of Texas at Dallas, and each used forty or more three-component magnetometers.

Fig. 2. Two-dimensional conductive models which fit observed magnetic variation fields in western United States. (a) refers to an east-west section at latitude 38°N and its fit to the data is shown in Porath and Gough (1971). (b) is an alternative model for 38°N due to Porath (1971) to fit the same data as (a). (c) refers to an east-west section at 32°N fitted by Porath and Gough (1971) to data of Schmucker (1964). Depths are exaggerated twofold relative to the horizontal scale. Conductivities are shown in Siemens/metre (1 Siemens = 1 ohm⁻¹).
The array of 1967 revealed the presence, during substorms, of two concentrations of induced current flowing along north-south conductors, one under the Wasatch Front in Utah and the other under the Southern Rockies of Colorado (Reitzel et al., 1970). Separation of the components of substorm fields showed that the local currents were internal and allowed estimation of normalized anomalous fields in three components and at three periods (Porath et al., 1970). It was found possible to fit the observed normalized anomalous fields within the precision of the data, by induction in either of two model structures (Porath and Gough, 1971; Porath, 1971). These conductive structures are two-dimensional and induction is by unit east-west horizontal magnetic field. Reference to the papers last quoted will show that the two models, which are reproduced in Figs. 2(a) and 2(b), fit the observations equally well. They refer to an east-west profile at 38°N. Each model is made up of ridges and steps on the surface of a highly-conducting medium. The "conductive topography" is at very different depths in the two cases, and these models exemplify the non-uniqueness of these models and the poor depth discrimination of a magnetometer array. The recorded fields can in fact be modelled by models like those shown but at any depth less than that in Fig. 2(a), which is near the maximum depth which fits the anomalous fields. By contrast an array locates a conductor on the map with accuracy limited only by the spacing between the magnetometers.

The models of Fig. 2(a) and (b) were in fact chosen to express two different hypotheses with regard to the physical cause of the high conductivities in the structures. On the planetary scale the principal variation of conductivity is radial and a steep rise to values above 1 S/m is required in the depth range 400–700 km, to fit long-period induction data (Banks, 1969, 1972). It is usually assumed that the conductivity rise is mainly controlled by temperature (Tozer, 1959) though pressure may also be significant (Mao and Bell, 1972). In electrical terms, Fig. 2(a) shows a model of perturbation of the boundary of a conductive half-space which approximates the rise in conductivity below 400 km. If the rise is assumed to be temperature-controlled then any local elongated source of heat will raise the surface of the good conductor to form a ridge. In particular if vertical transfer of material is occurring the conductive ridges may lie in upcurrents.

The model of Fig. 2(b) starts by assuming a conductive layer in the depth range which contains the seismic low velocity layer and elaborates topography on this. The causal hypothesis here is that the high conductivities are associated with partial melting.

Provided a conductive layer is at least a skin-depth thick where it is thinnest, conductors in the half-space below it affect the fields at the surface negligibly. In the highly-conductive parts of the model of Fig. 2(a) the conductivity is 0.2 S/m and fields of period 2 hr penetrate it with skin depth about
100 km. In this model, as in 2(b) and 2(c), the conductive layer could be chosen to have any thickness \( \geq 100 \text{ km} \). The assumption of infinite thickness in model 2(a) is chosen to approximate the radial conductivity of the earth; the finite thickness in model 2(b) is chosen to fit the hypothetical association of high conductivity with partial melting in the seismic low-velocity layer.

At the boundary between the Basin and Range and Great Plains Provinces just north of the U.S.-Mexican border Schmucker (1964) discovered his Rio Grande anomaly and found that his data were fitted by a ridge-plus-step structure with a highly conductive mantle high under the Basin and Range, low under the Great Plains and highest just west of the boundary. The 1968 array confirmed this general picture and showed that the southern end of the Wasatch Front anomaly becomes merged in the general high conductivity of the mantle beneath the Basin and Range (Porath and Gough, 1971). The eastern half of the 1968 array worked at low efficiency because of an unsuccessful modification to the Dallas instruments. Finite-difference numerical modelling was used to fit

Fig. 3. Electrical conductivity of the upper mantle and tectonic provinces of western North America. The density of stippling gives a qualitative indication of variations of conductivity. NACP: North American Central Plains crustal conductor.
a model of a step plus ridge to Schmucker's data (Porath and Gough, 1971) and this is reproduced as Fig. 2(c). The conductive ridge near the superprovince boundary can be regarded as continuous with that under the Southern Rockies (Fig. 3). The structure is however more like that under the Wasatch Front. The model of Fig. 2(c) is of the same “deep” variety as that in Fig. 2(a) and the data could undoubtedly be fitted by a model more like the western half of Fig. 2(b) at seismic low-velocity layer depth.

The 1969 array extended the investigation of western North America north of the Basin and Range Province and into the Canadian Cordillera. Here Hyndman (1963), Caner (1970), and Cochrane and Hyndman (1970) had shown that vertical component fields of periods less than 2 hr were attenuated in the Cordillera. This was confirmed by the array (Camfield et al., 1971). Two small, approximately north-south trending anomalies in Y and Z were found near the east front of the Northern Rockies (Camfield et al., 1971). Two-dimensional modelling was used to fit a steps-plus-ridges type of conductive model to these anomalies and to the normal fields on each side (Porath et al., 1971). Caner (1970) and Cochrane and Hyndman (1970) had fitted the normal fields with a model having a conductive layer in the lower crust of the western region. Camfield and Gough have added daily-variation long-period data to the substorm data, and find the mantle step-and-ridge model of Porath et al. (1971) to be in conflict with the daily variation data. Specifically, this model would produce an attenuation in the Cordillera of the vertical components of long-period fields which is not observed. To produce attenuation at periods less than 2 hr only, a conductive layer of limited thickness is required. The analysis of the broad-band array data for $30 \text{ min} \leq T \leq 1440 \text{ min}$ will be published shortly by Camfield and Gough. Without anticipating that paper the present review notes only that the principal feature of the conductive structure is a thin conductive layer, some 10 km thick with $\sigma = 0.2 \text{ S/m}$, in the upper mantle west of the Northern Rockies along a section at 48°N. If this is compared with either of the models, Fig. 2(a) or 2(b), of the structure in Utah and Colorado, it is clear that the conductive structures are thicker and, by inference, tectonic activity has been more intense at 38°N than at 48°N.

The conductive structures of the upper mantle of western North America are schematically represented by the density of stippling in Fig. 3. The highest conductivities are found beneath the Wasatch Front and Southern Rockies. The whole Basin and Range Province has highly conductive material under it. The northern limit of the conductive mantle is well located by anomalies due to induction in it, and is close to the boundary between the Basin and Range Province and the Columbia Plateaux, near 43°N (Camfield et al., 1971). The broad-band data from the 1969 array show that the Wasatch Front conductive ridge also ends at about this latitude (Camfield and Gough, 1974).

A structure is shown in Fig. 3 which produces one of the largest known
geomagnetic variation anomalies. It runs nearly north-south under the Black Hills of South Dakota and close to the eastern border of Montana. This is the North America Central Plains conductivity anomaly reported by Camfield et al. (1971). Anomaly half-widths show that this body is certainly in the crust and there is evidence that its conductivity is a consequence of its composition rather than of high temperatures (Gough and Camfield, 1972). Attempts to model the anomalous fields on two-dimensional assumptions were unsuccessful and showed that the currents are concentrated in a narrow conductor in the crust, but that this conductor joins large rock volumes of unknown geometry in which part of the induction occurs (Porath et al., 1971). In many continental regions a highly resistive layer in the crystalline rocks of the lower crust separates conductive rocks near the surface, in which saline ground-water in pores and fractures provides the high conductivity, from the thermally activated highly-conductive mantle. Since the local conductor under the Black Hills is in the crust it is probable that the regions of induction are also in the crust. There is, however, an alternative possibility. The anomalies of the Southern Rockies upper mantle conductor and the Central Plains crustal conductor merge. This is consistent with overlap of the conductors at different levels, or with a link between them through the resistive layer. The question is open. A detailed array study of the North American Central Plains anomaly was made in 1972 by Alabi, Camfield and the writer and when its results are available they may throw light on the question. Heat flow results are discussed in Section 3 but do not resolve the problem.

A serious limitation of present knowledge of the conductivity under western North America is set by the magnetometer spacing in the three arrays, which was typically 100 km (Fig. 1). Small structures can be missed by such an array or can be mistaken for large or deep structures as a consequence of spatial aliasing. Further studies with magnetometers closely spaced along east-west profiles would be highly desirable. Such profiles near 38°N and 32°N would be of interest and might set reduced maximum depths for the currents if steep gradients in anomalous fields were found. Such studies might discriminate in favour of the low-velocity-layer depth (Fig. 2(b)) and against the deep model of Fig. 2(a), but only if large gradients were found.

Extension of array studies to the north becomes difficult because of the proximity of the auroral zone and the rapid increase in complexity of the substorm source fields. No such problem inhibits southward extension into Mexico, and this is greatly to be desired.

3. Heat Flow

In recent years several hundreds of measurements of the heat flow through the crust have been made in the western United States. An important factor
in this development has been the demonstration that short boreholes to 200 m or even less could be made to yield reliable heat flow data by attention to, and avoidance of, near-surfaces sources of error (Roy et al., 1972). Two major groups contributing heat-flow data have recently given reviews (Sass et al., 1971; Roy et al., 1972), and the reader is referred to those. For the present purpose Fig. 4 is reproduced from Sass et al. (1971) and Fig. 5 from Roy et al. (1972). Both maps show compilations of observed heat flow. Some observations are shown on both maps.

Observations in plutonic rocks have been shown by Birch et al. (1968) to bear a linear relation to the radioactive heat generation per unit volume in the rock at the surface. This means that such observations can be corrected to yield the contribution from the mantle. Roy et al. (1972) outline the procedure and the implications of this important development. They also give a map of corrected heat flows attributable to the mantle (Roy et al., 1972, Fig. 12). Such corrected heat flows are much more relevant to tectonics than the observed

Fig. 4. Heat flow in western United States after Sass et al. (1974, Fig. 4). The authors are thanked for permission to reproduce this figure.
values. Unfortunately only a minority of the observations, those in plutons, can be corrected and these are biased toward uplifted basement areas. In particular, values in the Great Plains cannot be corrected. In consequence the more numerous observed values are here discussed. The map of corrected heat flows given by Roy et al. (1972) shows the same features as Fig. 5 in the area covered by the magnetometer arrays.

Figures 3, 4 and 5 show general agreement in several areas. High conductivity and high heat flow characterize the Basin and Range Province. An exception is an area of low heat flow in Nevada (Fig. 4) attributed by Sass et al. (1971) to anomalous ground-water circulation there. The Wasatch Front is near a boundary, in Figs. 4 and 5, between high heat flows in the Basin and Range and low values in the Colorado Plateau. The Southern Rockies give high heat flow and high conductivity. The heat flow maps place the western limit of the thermal structure west of the conductivity anomaly in Fig. 3, but the
difference is not significant in terms of the magnetometer spacing in the array.

The map of Sass et al. (1971), Fig. 4, shows a 1.5 HFU contour turning east to the southeast corner of Colorado. The magnetometer array data disagree strongly with this but are consistent with the contours drawn by Roy et al. (1972) (Fig. 5).

Roy et al. (1972) support a hypothesis due to Blackwell (1969), based on heat flow in the northwestern United States. This hypothesis combines the Basin and Range, Northern Rockies (including the Idaho Batholith) and Columbia Plateaux in a single geophysical province called the Cordilleran Thermal Anomaly Zone. In Section 2 it was pointed out that the electrical conductivity structures are much thicker (for a given conductivity) at latitude 38°N than at 48°N, and that the thick conductive layer under the Basin and Range has a northern limit near the northern limit of the physiographic province. It is suggested that further observations will show significantly larger heat flows in the Basin and Range than in its neighbour provinces to the north, especially if flow from the mantle is isolated by the method of Birch et al. (1968).

In Fig. 5 Roy, Blackwell and Decker have a ridge of high heat-flow running north-northeast from the Southern Rockies to include a single high heat-flow observation in the Black Hills of South Dakota. In Fig. 4 Sass and others show four observations in the Black Hills, two high (≥1.6 HFU) and two low (≥1.5). It has been remarked in Section 2 that the anomaly in magnetic variations is ambiguous, allowing either a link through the lower crust from the upper mantle conductor under the Southern Rockies to the crustal conductor which strikes through the Black Hills, or overlap of these conductors without connexion. A link could occur by way of rising molten material and high heat flows in the Black Hills support this. But there are two low values to three high ones. There are hot springs in the Black Hills but these, and the high heat flows, could be related to small-scale local tectonic movement rather than a rising magma column. The question remains open, and with it the question whether a low heat-flow area lies between the Southern Rockies and Black Hills or not. Sass et al. (1971) do not encounter the issue because of the eastward turn of their contour which leaves all of Colorado and an indeterminate large part of the Great Plains in the high heat flow province. The inconsistency of this with the electrical conductivity has been pointed out.

The differences between the maps of electrical conductivity and of heat flow have been considered because each can assist interpretation and correction of the other. Discussion of the differences should not, however, be allowed to obscure the impressive resemblance of Figs. 3, 4 and 5. There can be no serious doubt that the highly-conductive regions of western North America are conductive because they are hot. Whether lateral temperature variations in the solid state produce the conductive structure, or partial melting, or both, is a further question which is taken up in Section 5.
4. Structure from Seismology

A wealth of seismological information has appeared in the last two decades related to western North America. Active observational programmes, natural earthquakes covering a wide range of epicentral distances and the nuclear explosions at the Nevada Test Site in the Basin and Range Province have combined to place the region among the seismologically best-observed on the planet. Refraction profiles have revealed regional variations of the $P_n$ velocity in the upper mantle just under the $M$-discontinuity (Herrin and Taggart, 1962; Pekiser 1963) and in crustal thickness and composition (Pekiser and Robinson, 1966). $P_n$ velocities are lower in the western superprovince than under the Great Plains, and lowest under the Basin and Range; and a thin silicic crust is characteristic of the western region. Healy and Warren (1969) give a useful review of seismic parameters for the crust and topmost mantle.

The use of an array of seismometers for inversion of the apparent phase velocity to a velocity-depth profile was initiated by Niazi and Anderson (1965) and developed by Johnson (1967). The array used was in Arizona and the velocity profiles refer to the western superprovince. Johnson’s CIT 204 P-velocity profile is shown in Fig. 6 together with profiles by Green and Hales (1968), Archambeau et al. (1969), and HelMBERGER and Wiggins (1971). All show major discontinuities near depths of 400 km and 650 km, and these have become sharper in the most recent profiles as better fits to amplitudes and times have been achieved. P-velocity profiles for the central continent by Barr (1967), Lewis and Meyer (1968), and Green and Hales (1968) are shown in Fig. 6 for comparison with those for the western superprovince. Some of these

Fig. 6. Seismic velocity-depth profiles in western U.S.A. and central North America. References to the source papers are in the text.
eastern profiles were derived mainly from times with little use of amplitudes. However, it is clear that the 400 km and 650 km discontinuities are common to the whole continent. At depths less than 200 km there is a marked low-velocity layer in the western region but not in the central region, bearing out the suggestion originally made by Lehmann (1964) that shallow, low-velocity layers might be confined to the western mountain regions.

The work of Ibrahim and Nuttli (1967) led to the two S-velocity depth profiles shown in Fig. 6 for the western superprovince. That due to Anderson and Julian (1969) gives an alternative inversion of the data of Ibrahim and Nuttli. These authors proposed their S-velocity profile as representative of the United States as a whole. This is doubtless correct for the deeper half of the profile. But almost half of the seismological stations used by Ibrahim and Nuttli (1967, Fig. 3) were in the southwest quarter of the U.S.A., and all but one of the earthquakes they used were along the western margins of the Americas or in the West Indies. Most of the stations in the eastern half of the continent were more than 15° from the sources. So far as the upper mantle is concerned their study is therefore representative of the western regions. For this reason in Fig. 6 the models of Ibrahim and Nuttli (1967) and of Anderson and Julian (1969) are assigned to the western region.

The S-profiles for the central region are subject to more uncertainty than those for the west. The profile due to Brune and Dorman (1963) is derived from surface-wave studies and cannot give the resolution attainable from body waves. The S-profile of Lewis and Meyer (1968) has a pronounced upper mantle low-velocity layer but this profile was offered by its authors only with reservations.

![Fig. 7. Velocity-depth profiles for compressional seismic waves in four parts of the western superprovince. Data are from Archambeau et al. (1969). The provinces to which the profiles apply are: CIT 109P, Eastern Basin and Range–Northern Rockies; CIT 110P, Colorado Plateau–Southern Rockies; CIT 111P, Basin and Range Province; CIT 112P, Snake River Plains–Northern Rocky Mountains.](image-url)
Fine structure of the upper mantle in the western superprovince has been studied by Archambeau et al. (1969), who made inversions from $P$ times and amplitudes to velocity-depth profiles for four refracting regions (Fig. 7). These authors used two sources in Nevada. $P_n$ velocities from Herrin and Taggart (1962) and $P$ station delays due to Cleary and Hales (1966) were used as constraints on the inversions. Both geometric spreading of the elastic waves and anelastic damping were considered, with a single model for the variation of $Q$ with depth used in all cases. The profiles sampled several provinces within the western superprovince. All have major, rapid increases in compressional wave velocity near depths of 400 and 650 km and all have low-velocity zones in the top 150 km of the mantle. The differences between the regional models are in the depth range 0–150 km and affect the thicknesses of the low-velocity layer and its high-velocity lid, and the range of velocities in both. The model CIT 111P is most representative of the Basin and Range Province west of the Colorado Plateau. This region has been shown by the magnetometer arrays to have a highly conductive upper mantle. In this model, shown in Figs. 6 and 7, Archambeau et al. (1969) show no lid at all, the low-velocity layer extending up to the $M$-discontinuity. The regions related to the profiles of Fig. 7 are indicated in the figure legend.

Seismic refraction profiles and array phase-velocity data necessarily provide average velocities over the profile length or arc distance. They therefore give excellent depth resolution but for a structure averaged over 1000 km or so. A province the size of the Basin and Range can be studied without confusion with adjoining provinces but the Colorado Plateau is too small to be isolated from neighbouring regions, and velocity structures corresponding to the conductive ridges under the Wasatch Front and Southern Rockies cannot be seen at all in refraction. Seismic refraction resolves well in depth but poorly on the map. As has been seen, magnetometer arrays resolve well on the map but poorly in depth.

Arrival times of steeply incident body waves from distant sources can be analysed to isolate station residuals. While such station terms do not locate the velocity anomaly in depth they do resolve low- and high-velocity regions on the map with precision limited only by the station spacing. Seismograph arrays used in such a time-term mode therefore provide information comparable to that from magnetometer arrays and complementary to that given by seismic refraction from near sources. Cleary and Hales (1966) used $P$ times from 25 earthquakes, in the distance range 32° to 100° and well spread in azimuth, to 126 stations in North America and found station residuals up to 1 sec early in the central region and up to 1 sec late in the Basin and Range. Carder et al. (1966) found a similar pattern of $P$ residuals from three nuclear explosions in the Pacific and one in each of three continents, North America, Africa and Asia. Azimuthally dependent station residuals from $P$ times have
been estimated by Herrin and Taggart (1968) at 321 North America seismograph stations using 400 earthquakes and 30 large explosions in the distance range 20° to 105°. All of these teleseismic studies show maximum P delays in the Basin and Range Province near the Nevada-Arizona border, and earliest arrivals in inland eastern states.

Travel times of S waves from 20 earthquakes to 46 stations in the United States have been analysed by Doyle and Hales (1967). All the earthquakes and 44 of the stations were subsets of those used by Cleary and Hales (1966) in studying P residuals. The arc distances were in the range 28° to 82°. Station residuals for S had a range of 8 sec compared with 3 sec for P; the residuals in pairs for common stations were correlated with coefficient 0.75 and the regression slope of S residual on P residual was 3.72. As for P, the S station residuals are positive in the western superprovince and negative in the central and eastern United States.

A mobile array of 15 or 20 seismograph stations would produce extremely interesting fine-structure information in terms of delay absorption of teleseismic waves, in any tectonically active region and in particular in western North America. The essential data are P and S times and amplitudes. Numerous simple three-component stations would be much preferable to a few sophisticated ones.

5. Geophysical Evidence of Partial Melting

The seismic refraction evidence reviewed in Section 4 shows that there is a pronounced low-velocity layer in the upper mantle under the western superprovince of the United States, and that this layer has a boundary near the east front of the Rocky Mountains. Twelve typical models, which include those shown in Figs. 6 and 7, place the layer in the depth range 28 to 216 km and assign to it thicknesses 52 to 118 km and velocity ranges 0.2 to 1.3 km/sec.

The question of the most probable depth of the low-velocity layer has been considered by Hales et al. (1968). Variations of station P time anomaly with source distance were computed for layers of various extents and at various depths. These variations were compared with the local Pn arrival time difference (10 sec) and the teleseismic P difference (1–2 sec) between the Basin and Range Province and the central United States. The comparison indicated an upper mantle position for the low-velocity layer; Hales et al. suggest that it is probably in the depth range 100 to 200 km.

The P and S station residuals, and the relation between them are particularly useful in relation to the explanation of the low velocities within the layer. Hales and Doyle (1967) have shown that the observed relations between P and S, discussed in the last section, indicate that the change in S velocity is about 1.25 (change in P velocity), that there is a rise in Poisson's ratio in the low-velocity layer, and that the data are satisfied if the shear modulus alone varies,
with the bulk modulus and density sensibly constant. Using data of Soga et al. (1966) on the temperature dependence of shear velocity, Hales and Doyle (1967) show that even with a temperature 600°C above that in a normal region, over a depth range of 1000 km, the observed ratio of S to P delays cannot be accounted for without partial melting. It is therefore probable that partial melting occurs at least in part of the low-velocity layer. Archambeau et al. (1969) considered the possibility of partial melting and found support for it in the P velocities, but did not discuss the more diagnostic ratio of S to P velocities. In a pilot study of amplitude ratios of S to P at five observatories in the United States, McGinley and Anderson (1969) found large variations, with P amplitude, and still more the S/P amplitude ratio, heavily attenuated at Basin and Range Province stations. These authors pointed out that this supported the probability of partial melting in the upper mantle. They also remarked that the very high S/P amplitude ratio at Barker, Oregon, together with P time residuals like those in central North America, might be explained if underthrust oceanic lithosphere there provided a seismic "window" into the high-velocity mantle. Anderson and Sammis (1970) used ultrasonic data from Anderson et al. (1968) on the P and S velocities and their temperature and pressure partial derivatives to estimate a critical gradient, (dT/dz)c, which would give constant velocities with depth. This temperature gradient would have to be exceeded for a low-velocity layer to form in the absence of partial melting. Assuming either constant composition and phase or the Green and Ringwood (1967) transition from aluminous pyroxene peridotite to garnet peridotite, the known parameters of the low-velocity layer with the ultrasonic data required dT/dz of order 20°/km and heat-flow over 1.0 μcal cm⁻² sec⁻¹ from the mantle. Hence Anderson and Sammis concluded that partial melting, probably with some free water, was consistent with the known properties of the low-velocity layer and with known heat flow values.

In Section 2 the magnetic variation fields were shown to be consistent with conductive structures anywhere in the depth range 50–200 km, which is the depth range in which the seismic low-velocity layer is placed by published models. It is therefore relevant to ask whether a conductivity increase of order 100 (required to fit the magnetic data) could result from 1 or 2% melting of olivine. Presnall et al. (1972) have shown that the conductivity of an artificial basaltic mixture rises by about two orders of magnitude on melting at low pressure. However Chan et al. (1973) have pointed out that the melt conductivity must rise by 10⁴ to produce an increase of 10⁴ in bulk conductivity as a result of 1% melting. They consider ways in which an increase of 10⁴ might occur in the melt conductivity, and suggest that one or more of several mechanisms could well cause such an increase. Increased concentration of iron in the melt fraction is probably important and polaron hopping might provide a mechanism with a suitably small energy gap. Further work in the physics of
partially molten olivine should prove rewarding. Provisionally it is reasonable to associate the high conductivities with partial melting, especially as intrinsic semiconduction by olivine in the solid state seems unlikely in view of its large energy gap (Chan et al., 1973).

It may be worthy of note in passing that Gough and Porath (1970) have pointed out that if the conductive structures are deep between (120 and 350 km) as in Fig. 2(a), and are associated with heat-flow transmitted conductively, the source structures must be of order 200 my old. Since geological evidence indicates much smaller ages for volcanic rocks in the Basin and Range and for uplift of the Southern Rockies it may be that this argument should be turned round and used to argue that the conductive and thermal structures are shallower, as in Porath's model (Fig. 2(b)).

The evidence is still inconclusive but partial melting between 50 and 150 km depths would produce the seismic low-velocity layer, and lateral changes in the thickness of the partial melting and of the percentage of melting could produce the electrical conductivity structures. Closely spaced arrays of magnetometers might substantially reduce the maximum depth at which the conductors could be located.

6. Tectonics

Many papers have contributed to the discussion of the tectonics of western North America in terms of underthrusting of lithosphere plates and of overriding of the East Pacific Rise by the continent. Much of the mass of evidence is geological and this is no place to try to integrate this into a grand scheme of tectonics. Only three points will be noted as contributions from the magnetometer array studies to the tectonic picture.

1. North of the Basin and Range Province the evidence is that the conductive layer is thin and at a depth of 50 to 100 km (Section 2). Such a layer could arise through friction and consequent partial melting or through hydrated minerals in ocean-floor rocks if the continental lithosphere had overridden oceanic lithosphere between the Northern Rockies and the west coast, without subduction.

2. The main conductive structure of western North America is in the upper mantle under the Basin and Range Province. Here the conductive layer reaches a development exceeded only along the western edges of the Basin and Range where it abuts against the Colorado Plateau in Utah and against the Great Plains in the southern half of New Mexico. The conductivity is consistent with the hypothesis that the East Pacific Rise underlies the Basin and Range.

3. The structures in conductivity and heat flow under the Wasatch Front, Colorado Plateau and Southern Rockies (Figs. 2–5) could be produced by a recent subduction of a lithosphere plate moving eastward under the continent.
from the Basin and Range and dipping eastward from the Wasatch Front. The conductive ridge and high heat flow under the Southern Rockies, as well as the uplift of these mountains, could then be a result of ascent of andesitic or granitic magma from the subducted plate as has been suggested for the Andes (Gough, 1973).

These suggestions are tentative and at the present stage it is no doubt best to add the conductivity structure to the other geophysical and geological information which will eventually receive a coherent explanation in terms of tectonic processes.

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