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Geomagnetic Deep Sounding and Upper Mantle Structure in the Western United States

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Summary

Magnetic field time variations were observed in September 1967, with a two-dimensional array of 42 three-component variometers between latitudes 36° and 43° N and longitudes 101° W and 116° W. Fourier analysis of a polar substorm and of a storm shows that the former has a smooth spectrum and the latter a complex spectrum with many maxima. Upper mantle conductivity structure can be seen qualitatively in the original variograms, but is far more sharply defined in maps of Fourier spectral component amplitudes and phases. A ridge of high conductivity runs at a depth no greater than 200 km under the Southern Rocky Mountains between the Great Plains and the Colorado Plateau, which marks a low-conductivity region within the Cordillera. A strong conductivity anomaly runs north–south along the Wasatch Front through central Utah, and indicates the presence of an upwelling of highly conductive material at depth no greater than 120 km along the edge of a step structure which brings the conductive mantle to shallower depth under the Basin and Range Province than under the Colorado Plateau. Long-period maps from the storm suggest a rise in the conductive mantle between the north–south structures, from the Colorado Plateau southward to the Basin and Range. The daily variation shows the conductivity structures and indicates their great extent in depth. The geomagnetic deep sounding anomalies are found to be in excellent agreement with existing heat flow data, and this supports correlation of electrical conductivity with temperature. There is also good correlation with the available seismic velocity information for the upper mantle.

1. Introduction

Geophysical observations of several kinds indicate that the upper mantle of the Earth under North America is laterally inhomogeneous. Upper mantle seismic velocities of 8·0 km s^{-1} or larger are characteristic of the eastern United States and the Great Plains Province, while velocities decrease to values of 7·9 km s^{-1} or lower west of the Rocky Mountains (Herrin & Taggart 1962). A similar pattern is shown by travel-time anomalies of seismic waves at vertical incidence. *P* and *S* waves arrive early at stations in the eastern United States; late arrivals are predominant in the western United States (Cleary & Hales 1966; Doyle & Hales 1967; Herrin & Taggart 1968). As the differences between the *P* travel-time residuals and the gravity anomalies in the central and western U.S. cannot be explained by the Birch (1961) relation between velocity and density, Hales & Doyle (1967) suggested that temperature anomalies in the upper mantle, possibly accompanied by partial melting, might cause lower velocities in the western United States by lowering the values of the elastic constants.

This interpretation is supported by the distribution of heat flow data which have become available in recent years (Lee & Uyeda 1965; Roy *et al.* 1968; Blackwell 1969). High heat flow in the Southern Rockies and the Basin and Range Province correlates with low upper mantle velocities; normal heat flow is found for the Great Plains, which have high upper mantle velocities.

Seismic and heat flow data are still too scarce to delineate the boundaries of different upper mantle provinces with accuracy. However, temperature anomalies in the upper mantle will cause anomalies in the electrical conductivity, as silicates at upper mantle temperatures show semi-conductor properties (Tozer 1959). Anomalies in electrical conductivity can be detected from differences between stations in the amplitudes of natural geomagnetic variations, especially the vertical component. These differences arise from currents in regions of enhanced conductivity in the upper few hundred kilometres of the Earth. Geomagnetic deep sounding is the study of such regions by simultaneously recording the three components of geomagnetic variations at a number of stations.

2. Principles of geomagnetic deep sounding

Inhomogeneities in the electrical conductivity of the crust and upper mantle affect the internal field induced by geomagnetic variations. The variations best suited for geomagnetic deep sounding are those associated with magnetic substorms and storms, which involve intense ionospheric current systems concentrated in the auroral zones. In mid-latitudes the fields of these external currents vary smoothly with position, so that sharp local changes in amplitude and phase can be attributed to non-uniform conductivity structures in the Earth which control the induced internal currents.

The depth of penetration of the variation field is a function of its period (skin effect). Assuming *normal* conductivities for the crustal rocks it is usually safe to assume that the induced current systems of variations having periods of 30 minutes or more flow in the upper mantle. The rapid rise in conductivity at depths between 600 and 800 km established by Lahiri & Price (1939) sets an upper limit in this range to the depth attainable in geomagnetic deep sounding.

When dealing with local conductivity structures we can treat the Earth as a conductor with a plane boundary. Consider first the case of a plane Earth in which the conductivity of the crust and part of the upper mantle is negligible compared with that of the mantle below some depth where semi-conduction is predominant. The vertical component Z of the induced internal field will oppose the external field, whereas the external and internal horizontal components reinforce each other (Rikitake 1966, p. 129). Thus, vertical variations of low amplitude and enhanced horizontal variations are observed in regions above high mantle conductivity. Because of the relation between electrical conductivity and temperature, this would correspond to a region of high mantle temperature. High Z consisting mainly of the vertical component present in the external field will be observed in regions above low mantle conductivity. Provided that the inducing field is non-uniform the field on the surface of the Earth can be separated into components of external and internal origins. From the amplitude and phase relationships between external and internal components as functions of period one can estimate the conductivity as a function of depth by fitting multi-layered models, or, if the electromagnetic response of the Earth is known over a sufficiently wide frequency spectrum, by directly inverting the surface data (Siebert 1964).

Induction problems which involve lateral changes of conductivity are much more difficult to treat mathematically, and for conductors having more complicated boundaries than a sphere or a cylinder, the induced field can in general only be calculated on the assumption that a body of infinite conductivity is responsible for the observed anomaly. However, in many electromagnetic induction problems in the Earth the assumption of infinite conductivity is a good approximation, as only small phase differences between inducing and induced fields are observed for most deep-seated conductivity anomalies.

The good resolution of lateral conductivity variations by the geomagnetic deep sounding method arises from the fact that an additional vertical field is induced by the horizontal component perpendicular to the strike of the conductor. This is called the *anomalous* internal Z compared with the normal internal Z induced by the external vertical component of the incident variation field in the absence of lateral conductivity changes. The anomalous internal Z will therefore show a correlation with an external horizontal variation field in a certain azimuth. This correlation has been given a mathematical and graphical form by Parkinson (1959, 1962) and Wiese (1962).

The induction arrow as defined by Parkinson points towards regions of higher conductivity and corresponds to the direction of maximum correlation between upward Z and horizontal field component. A single event is usually not sufficient to determine Parkinson's induction arrow accurately, as even in mid-latitudes a vertical component due to external sources is present. However, the direction of the external field varies for different events so that an analysis of numerous events can give the correct azimuth of Parkinson's vector for a given site.

3. Previous work and description of array

The pioneering work on geomagnetic deep sounding in the western United States was done by Schmucker (1964), who recorded geomagnetic variations in California and along a profile across southern Arizona, New Mexico and west Texas. He found an increase in the vertical variation by a factor of about 3 east of 106° longitude, close to where the Basin and Range Province borders the Great Plains. Schmucker interpreted this as due to a rise in mantle conductivity in the Basin and Range.

Reitzel & Gough in 1966 (Gough & Reitzel 1969) employed eight variometers across the eastern front of the Southern Rockies in Colorado and found a similar increase in Z associated with the boundary between the Southern Rockies and the Great Plains. Caner, Cannon & Livingstone (1967) carried out geomagnetic deep sounding along a profile across northern New Mexico into Oklahoma and found high Z only at their eastern-most station in western Oklahoma. They suggest that the anomaly related to the eastern front of the Rockies swings towards the east in northern New Mexico and does not follow the topographic front of the Rocky Mountains.

In the summer of 1967 we began a programme of study of the distribution of upper mantle conductivity in the western United States in the hope of defining boundaries of regions with different mantle structure more accurately than had previously been possible. For this purpose 42 magnetic variometers had been designed and constructed at the Southwest Center in Dallas and the University of Alberta in Edmonton. The instruments were arranged along four east–west profiles about 1200 km in length, crossing the Southern Rockies and Colorado Plateau, and extending well into the Great Plains and the Basin and Range Province (see Fig. 1). Variometers were concentrated across the Southern Rockies and also across the Wasatch Fault zone. The evidence of vertical displacements at the Wasatch zone and the abundant volcanics to the west of it had suggested the possibility of structure in the

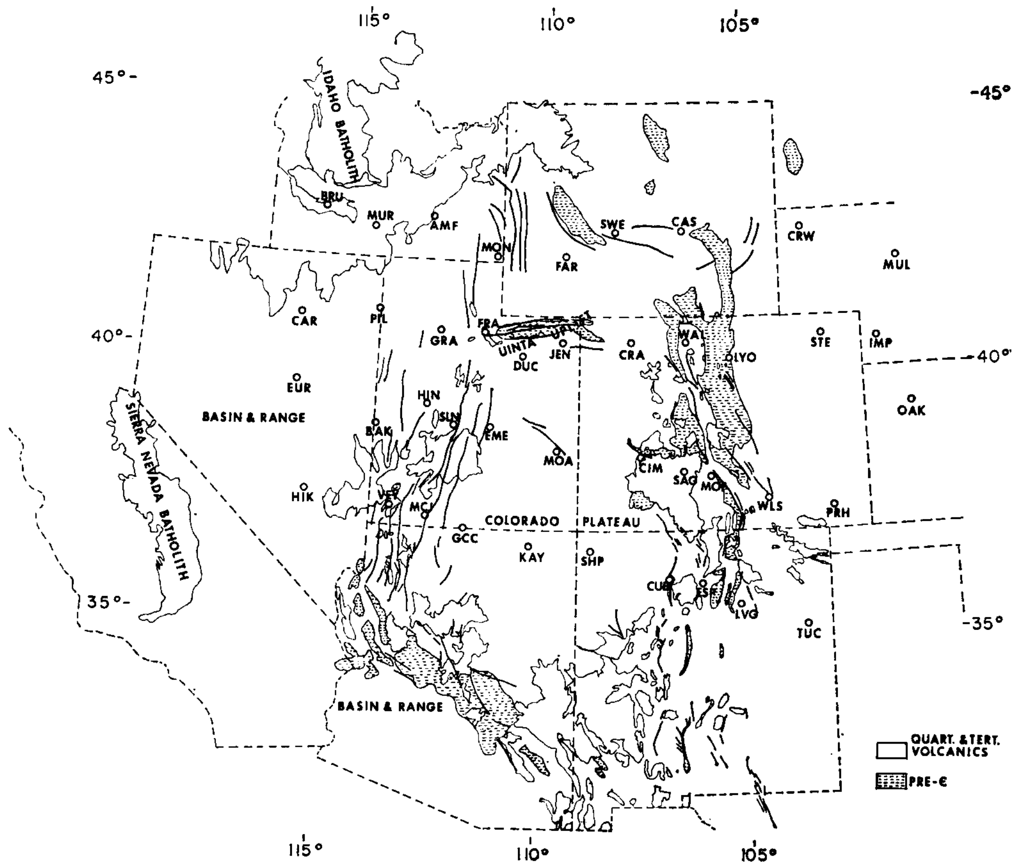

FIG. 1. Geological sketch map of the western United States showing location of the array.

isotherms there. The spacing between lines was about 150 km and that between stations in each line averaged 120 km. Small conductivity anomalies at depths less than 100 km may be missed by such an array. This risk was accepted with the aim of looking for the large features in upper mantle conductivity.

4. Instruments and procedures

The variometers (Gough & Reitzel 1967) are classical in type, with magnets suspended on torsion wires and with analogue recording on photographic film. In use they are buried in soil for thermal stability and concealment, and are visited at about 12-day intervals. At each visit standardizing fields known within 1 per cent were applied by means of permanent magnets (southern lines) or permanent magnets and coils (northern lines). With good field operation better than 2 per cent precision of calibration can be secured by these methods, though the precision attained was less in 1967 at some stations. The first deflection in a routine sequence was timed by counting second ticks from WWV or by stopwatch with WWV. The Bulova Accutron timers which controlled the variometers had reasonably constant rates of order 10 s/day or less, and time was generally reliable within one minute. An independent time check can usually be obtained from recognizable Pi-2 micropulsations near the start of an event, which are virtually simultaneous across the array (Rostoker 1968).

The data for the present paper were secured by commercial digitization at 1 minute intervals of enlarged Xerox prints of events selected by scanning the film records. The traces were smoothed by underlining them in pencil, so as to avoid aliasing from micropulsations of periods less than about 2 minutes.

5. Variation anomalies

The summer of 1967 was not magnetically active, but a few bays as well as a moderate storm have been recorded by about 80 per cent of the instruments, difficulties with the film transport being the main reason for instrument failure. The results given in this paper are derived from a substorm on 1967 September 1 and from a storm on 1967 September 20–21. Calcomp plots of these events are shown in Figs 2 and 3. The lines of variometers are numbered 1 to 4 from north to south. The variation anomaly associated with the eastern front of the Southern Rockies can be seen in the vertical component of the substorm of 1967 September 1 in the increases

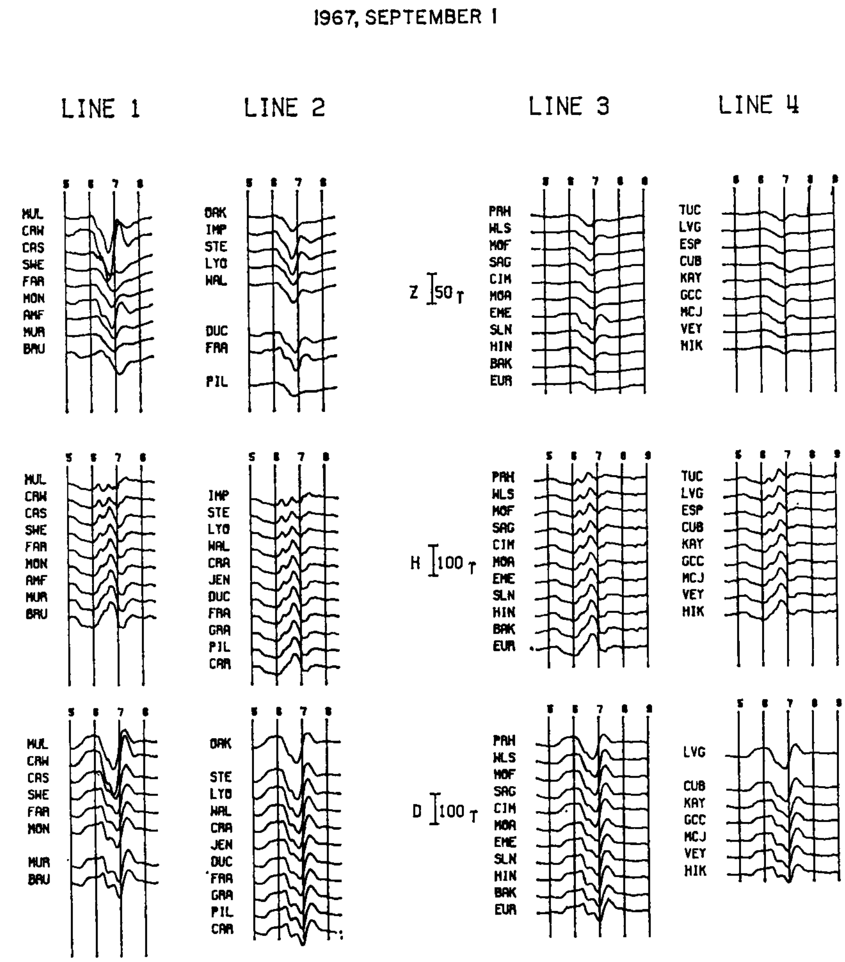

FIG. 2. Variograms of substorm 1967 September 1.

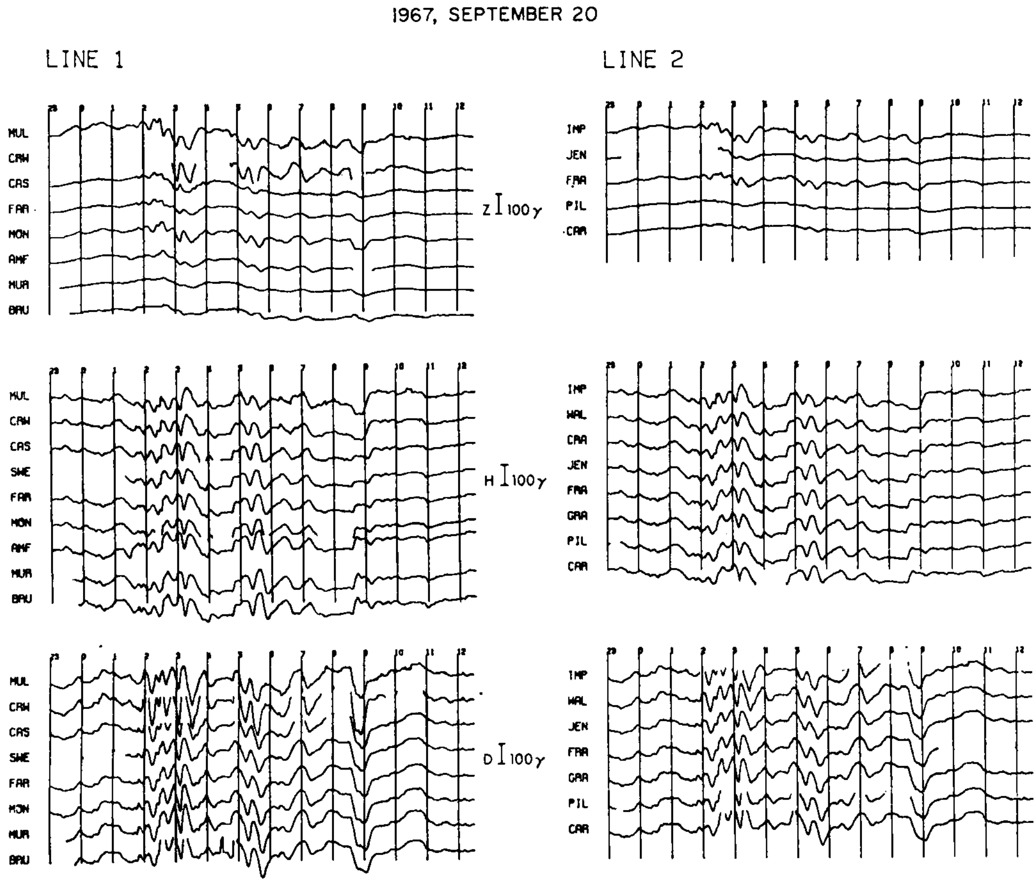

FIG. 3a. Variograms of storm, 1967 September 20–21. Northern lines.

in Z between CAS and CRW, between LYO and STE, and between SAG and WLS. Only a small increase is observed for the substorm at the easternmost station, TUC, of Line 4. The presence of the eastern front anomaly is more obvious for the storm. This difference in the response of the vertical field may be due to the different azimuth of the external field for the storm, which was mainly NE–SW, compared with NW–SE for the bay. A further bay event on 1967 September 28 also shows the presence of the eastern front anomaly in northern New Mexico by a substantial increase in Z between LVG and TUC.

Our results therefore show that the conductivity anomaly found by Schmucker in southern New Mexico and by Reitzel and Gough in Colorado does not swing east into western Oklahoma in northern New Mexico as suggested by Caner *et al.* (1967), but continues along the topographic front of the Cordillera. This interpretation is supported by subsequent observations in northern New Mexico during the summer of 1968.

Most remarkable, however, is the variation anomaly observed on all profiles over the Wasatch Front in Utah (MON, DUC, FRA, EME, SLN, MCJ, GCC), which forms the boundary between the Colorado Plateau and the Basin and Range Province. The similarity in the traces of Z and D indicates the presence of anomalous internal Z at these stations. Upward (negative) vertical variations are in phase with westerly horizontal variations. This and the fact that stations west of the Wasatch Front (PIL, EUR, HIK) have very low Z amplitudes suggest that the Basin and

Range Province is a region of high mantle conductivities. The Calcomp plot of the storm shows reversals in Z for periods of about one-half hour at BAK and PIL, stations just west of the Wasatch Front. These phase reversals suggest that a ridge in the conductivity structure lies along the edge of the Basin and Range Province at the Wasatch Front. Increased mantle conductivities at the edge of the Basin and Range Province have also been found by Schmucker (1964) in southern New Mexico. At BRU at the west end of line 1, Z resembles the H component rather than D, with upward Z following southward H. This suggests the presence of a highly conducting body south of this station. This could be the northern edge of the Basin and Range Province.

In addition, there appears to be an anomaly related to the boundary between the Colorado Plateau and the Southern Rockies. This anomaly is observed only on the two southern profiles, which have stations close to this boundary. CUB shows a reversal in Z which must be due to anomalous internal Z. Parkinson's vector points towards the Southern Rockies as a region of relatively higher conductivity compared with the Colorado Plateau. A phase change but no reversal is observed at CIM: a similar phase difference for CIM has been reported by Gough & Reitzel (1969). Even in mid-latitudes the field of a geomagnetic substorm contains a small, usually negative, external vertical component, so that a complete reversal due to higher conductivities in the Southern Rockies will be observed only when the anomalous positive internal Z is larger than the external negative Z. Stations in New Mexico of an array operated during the summer of 1968 support the interpretation

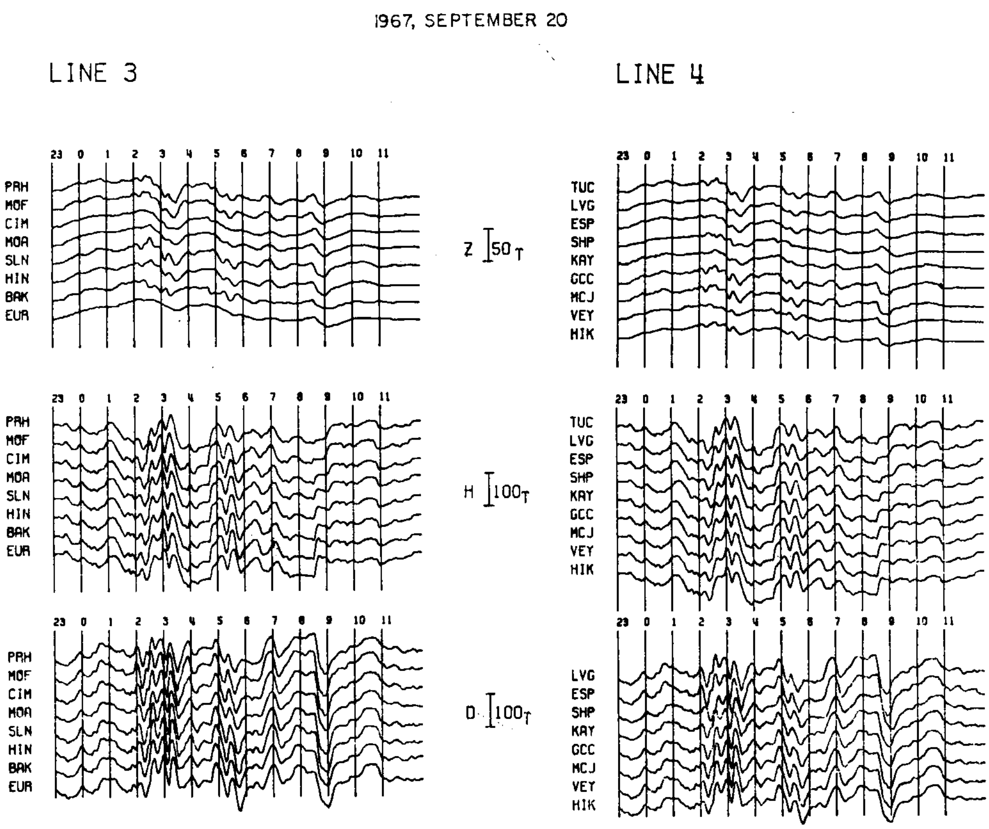

FIG. 3b. Variograms of storm, 1967 September 20–21. Southern lines.

that mantle conductivity is higher under the Southern Rockies than under the Colorado Plateau. This work will be published elsewhere.

The horizontal variations H and D of the geomagnetic substorm of September 1 vary substantially over our array. The N–S component increases whereas the E–W component decreases from east to west, in such a way that the change in horizontal fields across the array is principally a rotation of azimuth.

6. Spectral analysis

To investigate the dependence on period of the different conductivity anomalies, Fourier transforms of the substorm and storm field components were computed by the use of the Cooley–Tukey algorithm after linear removal of a trend. The beginnings and ends of the records were smoothed by a $\sin^2$-window. To secure estimates of amplitude and phase at a reasonable number of periods up to 89 minutes, for the substorm, zeros were added to the 211 data points to give a total of 2048 points. The addition of zero values for 1837 minutes after the substorm is considered justifiable because inspection of the variograms shows that the substorm fields can be regarded as a transient: while the trace after it does show small deflections, these have independent sources such as the daily variation. In the case of the storm, the records yielded 811 data points at one-minute intervals, and zero values were added to give

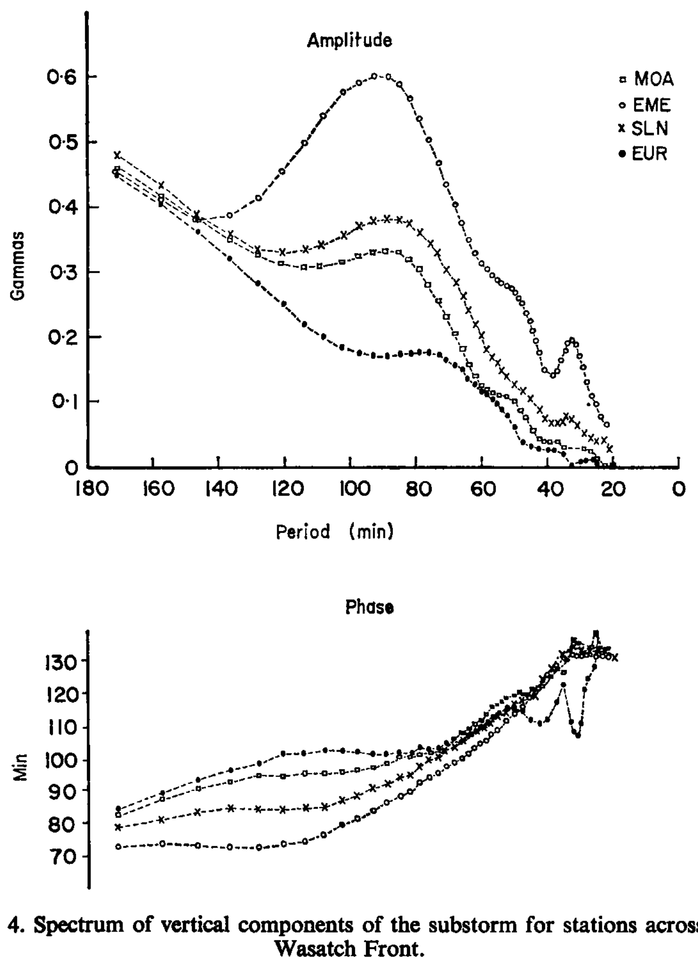

FIG. 4. Spectrum of vertical components of the substorm for stations across the Wasatch Front.

4096 in all, on the same assumption that the storm could be treated as a transient event. At some stations where the amplitudes were very large, mainly in lines 1 an 2 for the D component, fictitious values were interpolated in the gaps indicated in Fig. 3 to permit some attempt at spectral analysis. The results for those stations at which the gaps are numerous should be treated with reserve.

Figs 4 and 5 show examples of Fourier transforms of the vertical component of the substorm and the storm for stations over the Wasatch Front. The substorm has a smooth amplitude spectrum and very marked differences are observed between stations for periods from 20 minutes to 2 hours. Most spectra contain very little energy at periods less than 30 minutes except for those stations with large anomalous internal Z. Substantial phase differences are observed between stations, especially where anomalous induction occurs, but for periods exceeding 90 minutes these may be partially due to variations in the long period trend removed before Fourier analysis.

A line spectrum (Fig. 5) is obtained for the storm with peaks at periods of 155, 120, 85, 65 minutes and several peaks below 60 minutes. The short-period peaks may be distorted by side-lobes. The spectrum again contains very little energy for periods under 20 minutes. A differential shift between the energy maxima of the horizontal and vertical components is observed in the storm spectrum for periods larger than 50 minutes (Fig. 6). This is probably due to the induced internal field. The internal part becomes relatively smaller with increasing period, and this effect

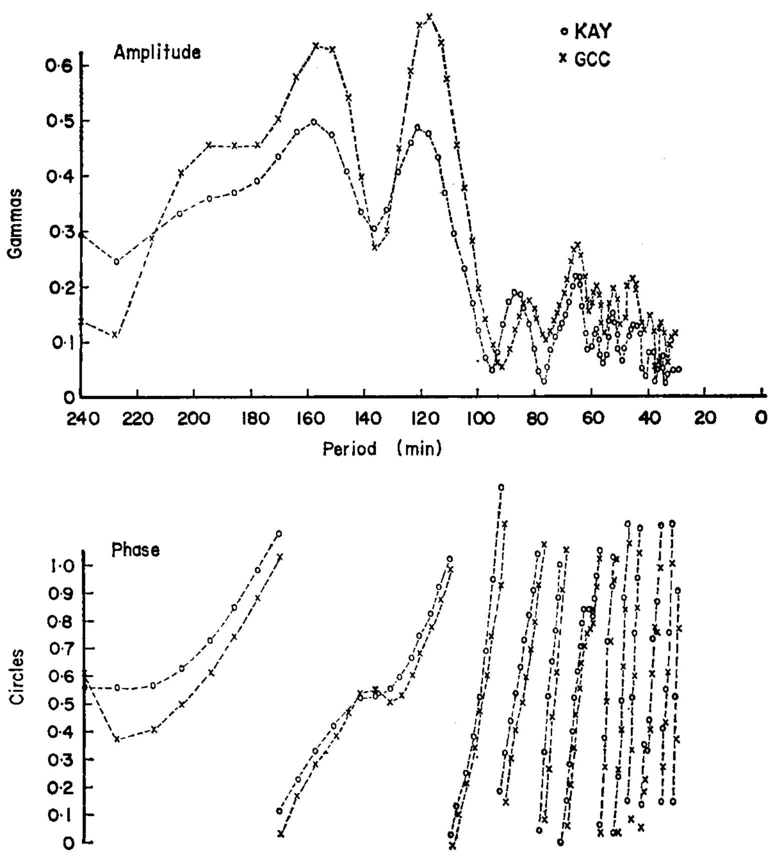

FIG. 5. Spectrum of vertical components of storm for a station over the Wasatch Front (GCC) and a station over the Colorado Plateau (KAY).

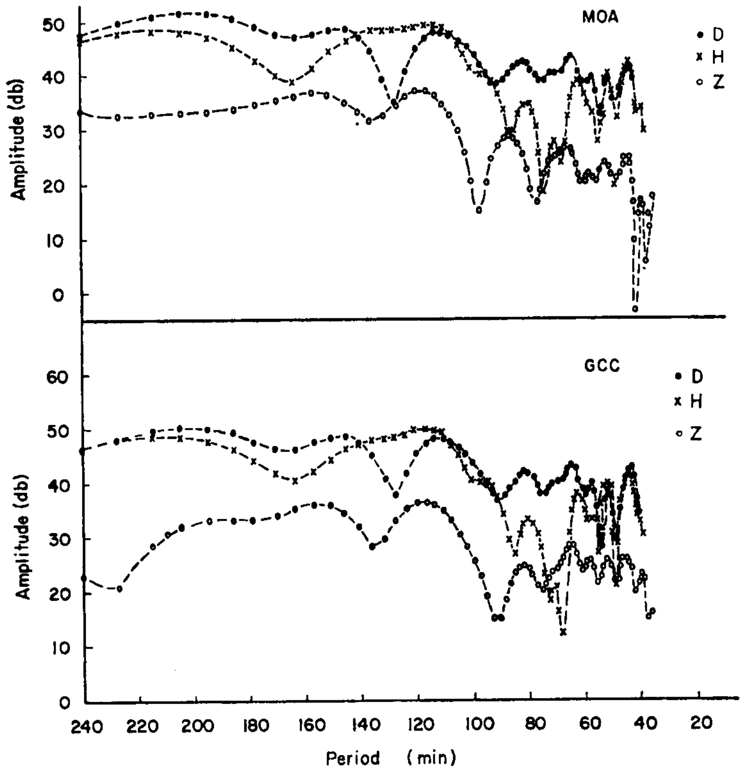

FIG. 6. Spectra of three components of storm showing differential shift of amplitude maxima. Amplitudes are relative to an arbitrary level for each station.

should shift the energy maxima of the horizontal components to shorter periods and those of the vertical component to longer periods, as is observed for the storm spectrum.

Let us look more closely at the substorm of September 1, as it will be used for the separation analysis in a subsequent paper (Porath, Oldenburg & Gough 1970). A map of the phases of the horizontal components at the stations reveals that the disturbance does not start simultaneously over our array; time differences in H and D of several minutes are observed, with later starting times going from east to west (see Figs 7 and 11). The differences in starting times can be explained by a westward surge in the source current. Phase differences in the vertical component are greatly disturbed by internal anomalies (Fig. 9).

7. Maps of spectral components

Systematic variations in amplitude and phase of spectra of Z at neighbouring stations, as exemplified in Fig. 4, suggest the mapping of these parameters. Figs 8–11 show such maps for spectral components of the fields due to the substorm of 1967 September 1. Figs 12 and 13 show similar maps derived from the input fields of the storm of September 20–21.

The vertical field amplitude maps of Fig. 8 show two prominent anomalies, one along the east front of the Southern Rockies and the other along the Wasatch Front. These anomalies show more strongly at the longer periods partly because of greater penetration of the fields but partly also because the source field contains more energy

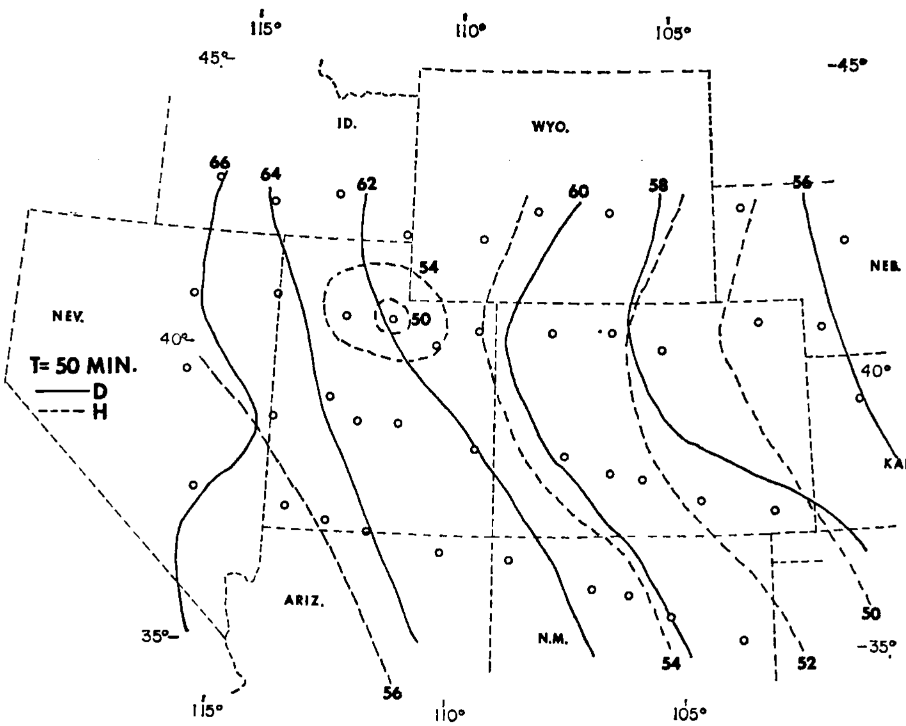

FIG. 7. Phase variations of horizontal components of substorm for 50 minutes: phases in minutes.

at these longer periods. The storm field Z amplitude maps (Fig. 12) have lower resolution than those from the substorm because fewer stations recorded the storm. Nevertheless the maps in Figs 8 and 12 for Z at $T = 45$ minutes are strongly alike.

The eastward horizontal component Y shows strong evidence of both the East Front and the Wasatch Front current systems in the amplitude maps for the substorm (Fig. 10). The maximum of Y is west of that in Z in both cases, as is to be expected for local north–south currents. In each case Y falls again to the west of the maximum, indicating a linear current. At the Wasatch Front this represents a local upwelling in the isotherms under this Front, superposed probably on a step which leaves the isotherms higher under the Basin and Range Province than under the Colorado Plateau. Schmucker (1964) reached similar conclusions with respect to the conductivity structure at the edge of the Basin and Range Province in southern New Mexico. He observed a reversal in Z at Las Cruces, a station somewhat west of the boundary, and attributed this to a local increase of mantle conductivity at this boundary above the Basin and Range average, before a steep decrease under the Great Plains. The Y amplitude maps at the East Front suggest a strip of current under the Southern Rockies between the East Front and the Colorado Plateau.

The amplitude maps of Y for the storm are strongly affected by fictitious interpolated data at some stations, especially in the north-east corner. The apparent disappearance of the East Front anomaly from the Y map at $T = 150$ minutes is probably an artefact.

An intensification of the East Front anomaly in both Z and Y is found at the northern limit of the array at Crawford, Nebraska. This may be associated with a strong local increase in conductivity related to the Black Hills thermal area of South Dakota.

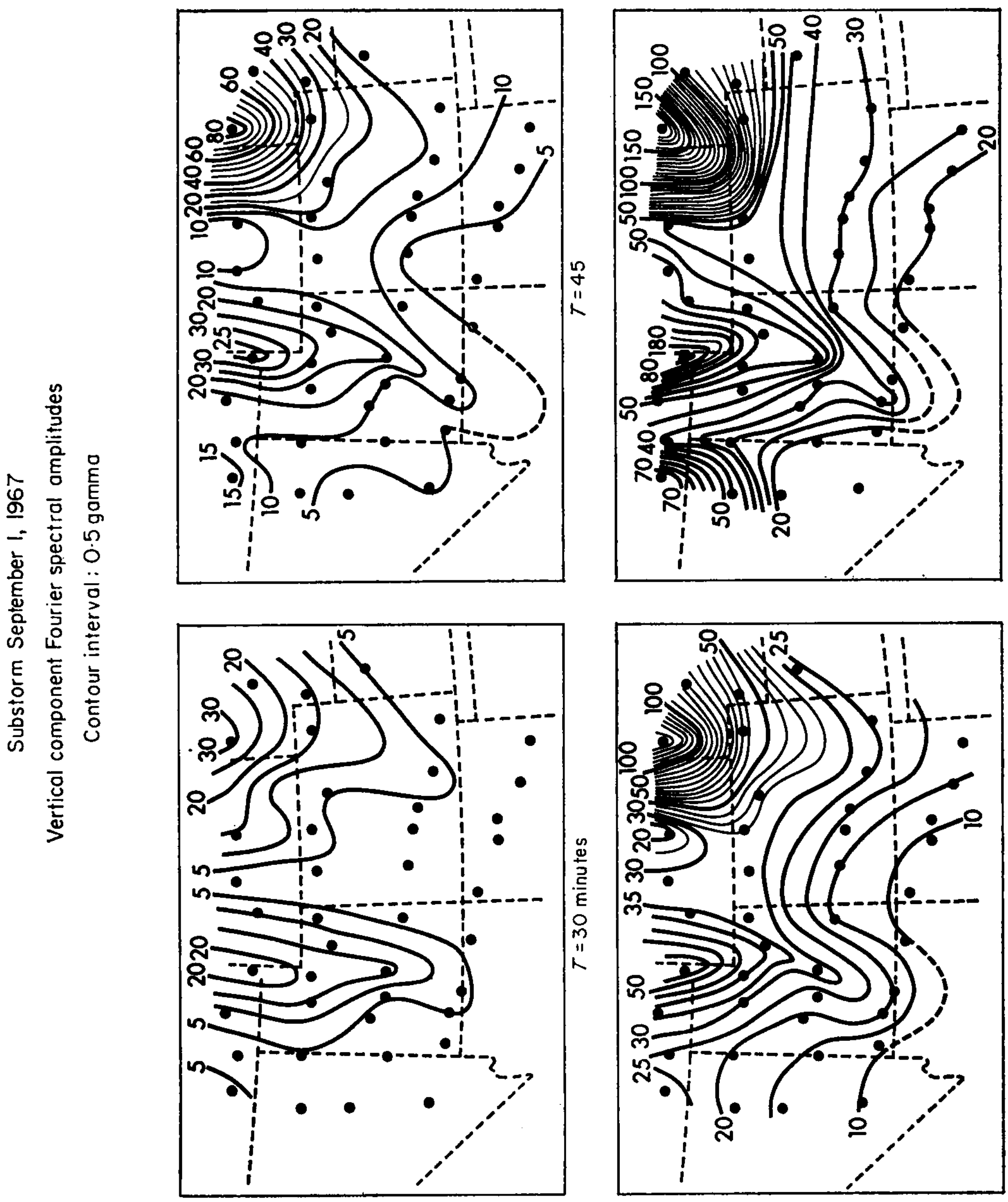

FIG. 8. Unit: 0·1 gamma.

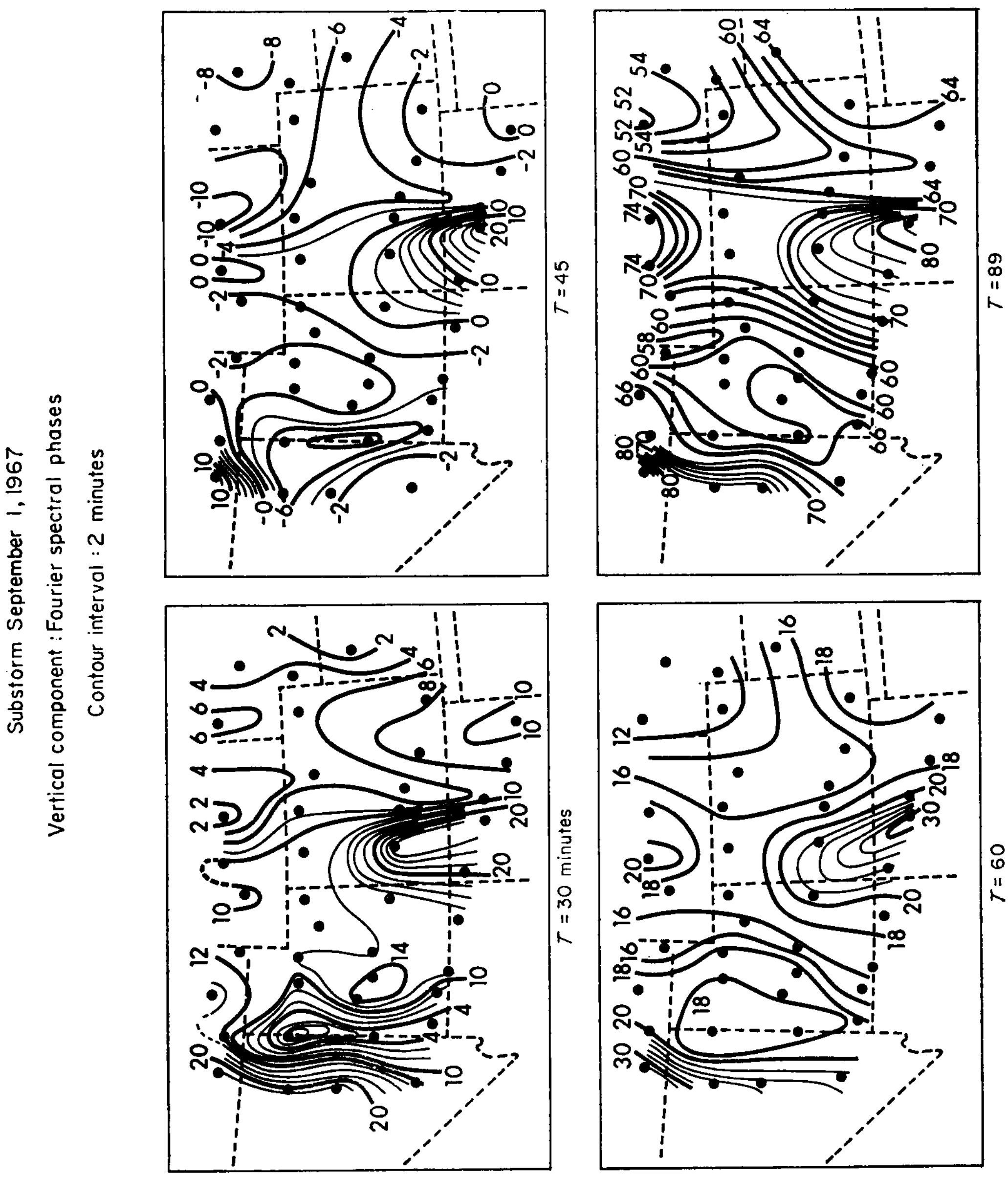
Substorm September 1, 1967
Vertical component : Fourier spectral phases
Contour interval : 2 minutes
T = 30 minutes
T = 45
T = 60
T = 89

FIG. 9.

Substorm September 1, 1967

Horizontal components: Fourier spectral amplitudes

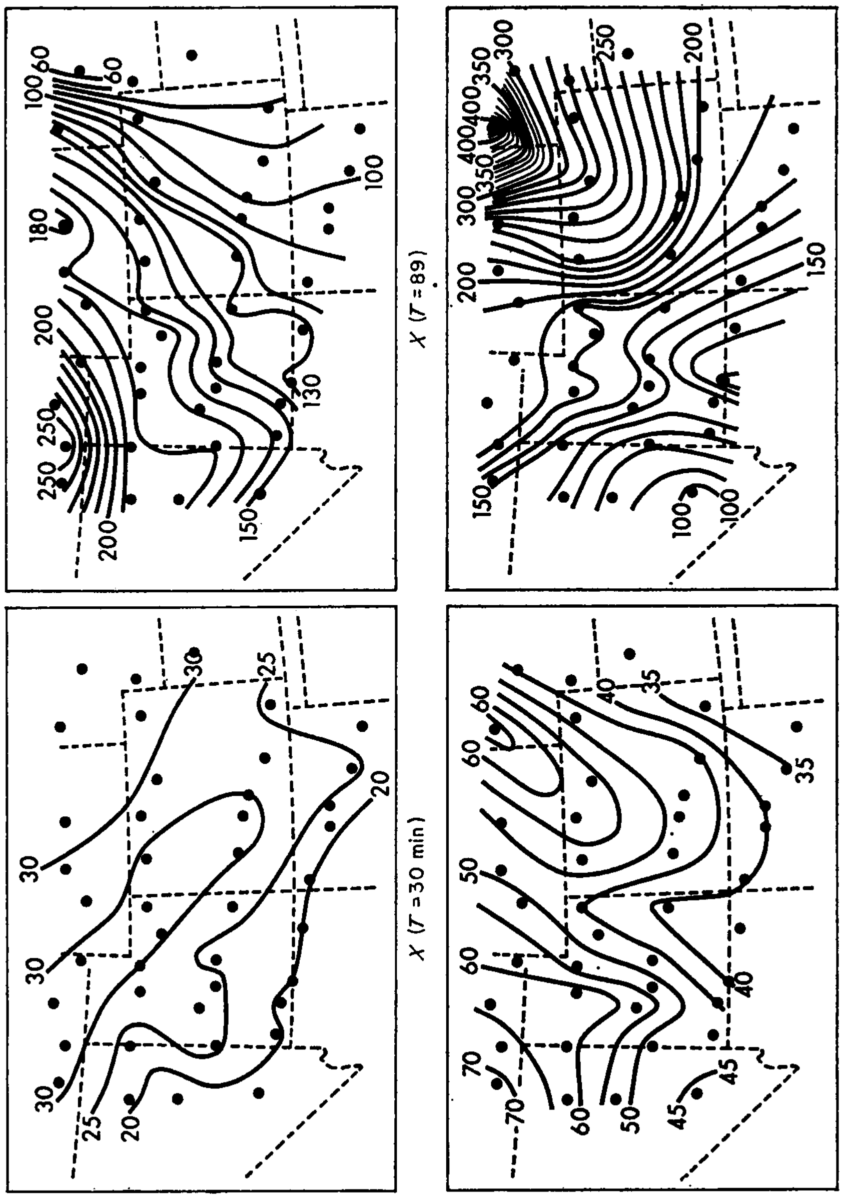

FIG. 10.

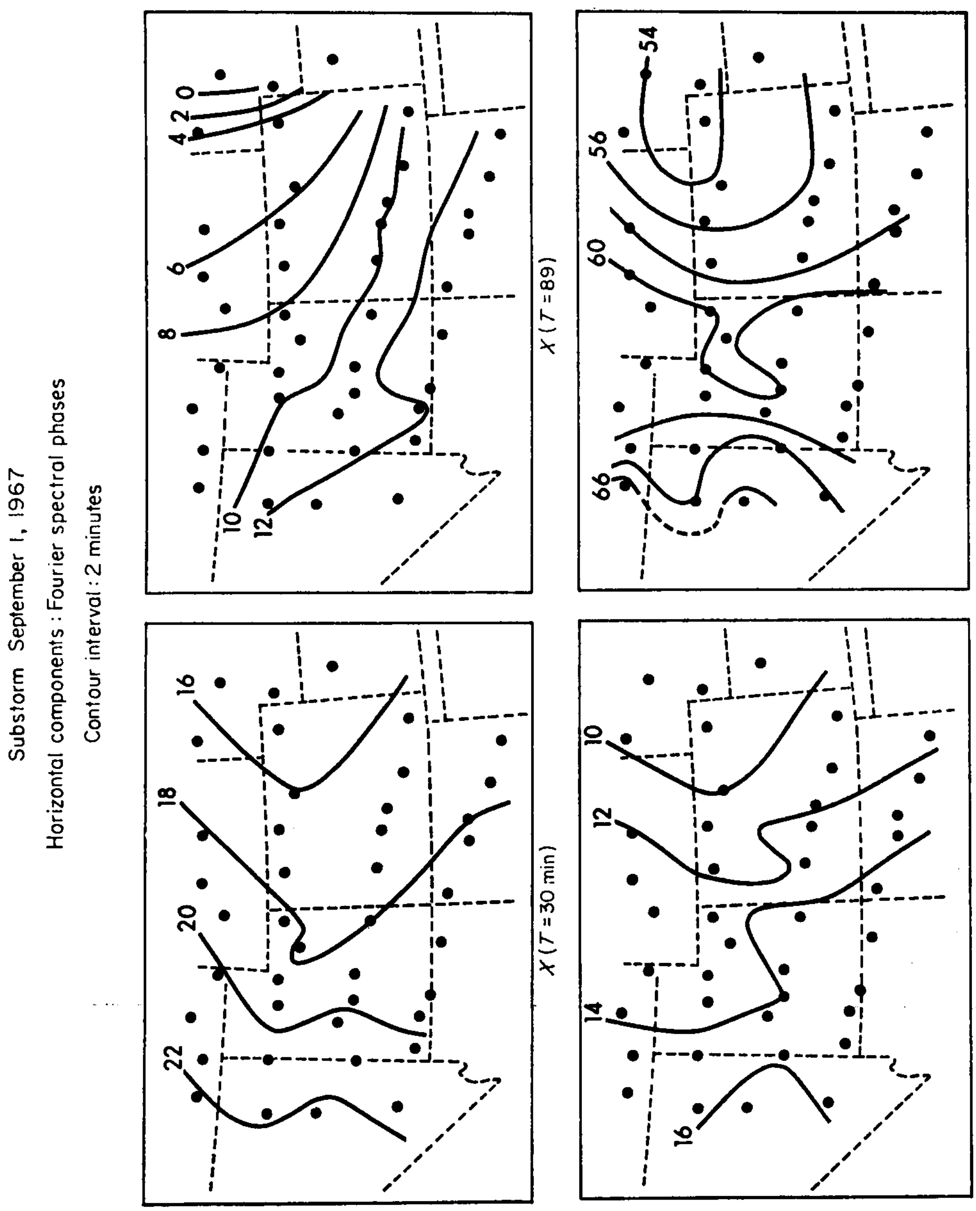
Substorm September 1, 1967
Horizontal components: Fourier spectral phases
Contour interval: 2 minutes
X (T = 89)
Y (T = 89)
X (T = 30 min)
Y (T = 30)

FIG. 11.

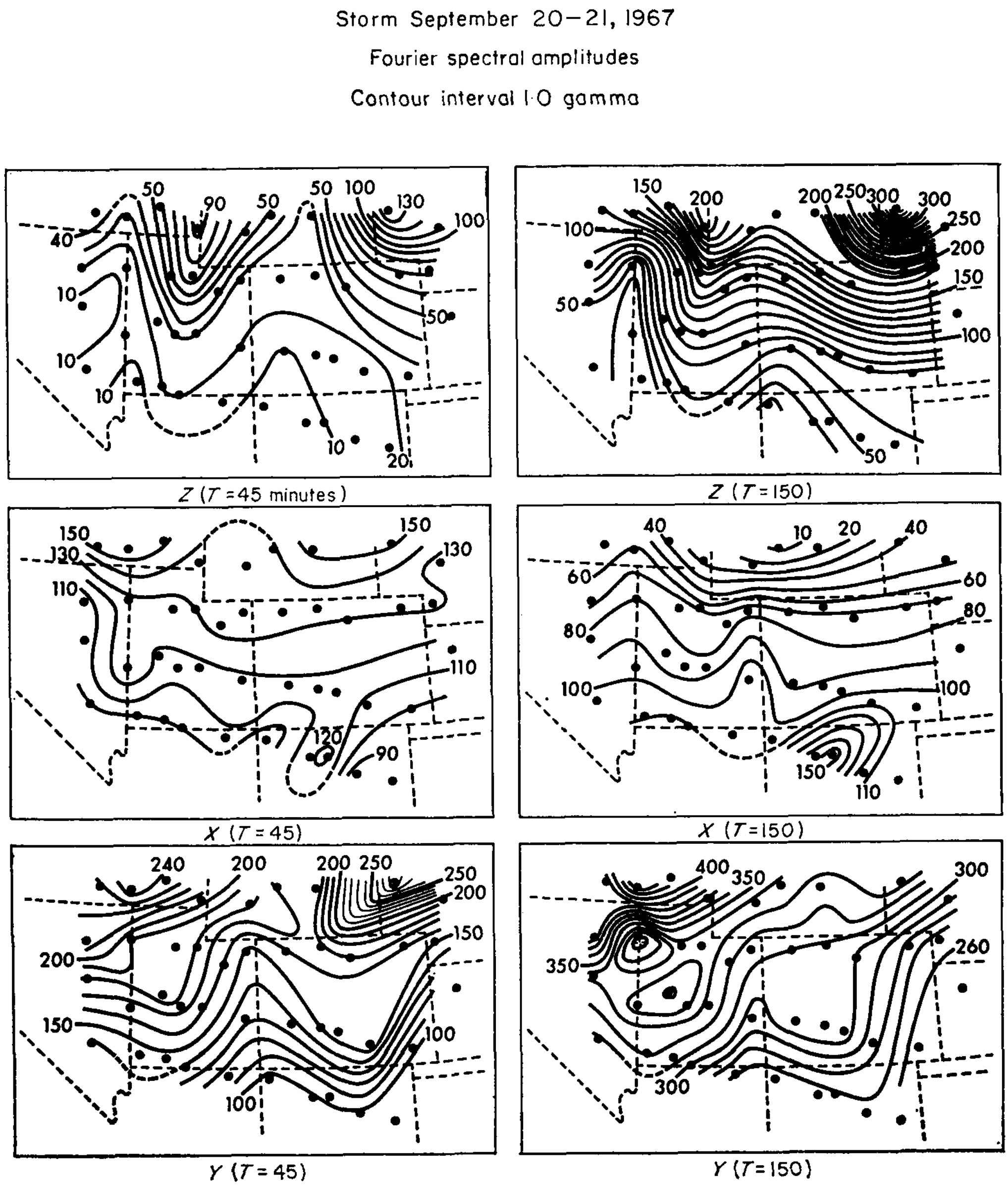

FIG. 12. Unit: 0·1 gamma.

Preliminary inspection of east–west profiles in Y and Z across the Southern Rockies and the Wasatch Front has yielded rough estimates of the depths of line currents approximating the conducting ridges, from half-widths of the Y maxima and horizontal distances between peaks for Z. The depths are of order 120 km for the Wasatch Front structure and 200 km for the Southern Rockies structure, for currents of period 89 minutes. Quantitative analysis of separated fields by means of conductivity models is discussed elsewhere (Porath *et al.* 1970).

The amplitude maps of the X component for the substorm (Fig. 10) are relatively featureless, as would be expected where the major structures strike north–south. A small anomaly in X, associated either with the Uinta uplift or with the deep sedimentary Uinta Basin (20000 ft) in northern Utah, is seen in a sine transform map at period 50 minutes (Porath *et al.* 1970, Fig. 7b). Deep basins of conducting sediments can cause variation anomalies at periods up to one hour or more (Schmucker 1964), and we interpret this anomaly as due to conducting sediments in the Uinta Basin. Superficial anomalies usually show a conspicuous phase lead of the induced to the inducing field (Schmucker 1964) and the early arrival times of X for the substorm in this region may be an expression of the Uinta Basin anomaly (Fig. 7).

The X component maps for the storm (Fig. 12) show a remarkable reversal in gradient of the amplitude between periods 45 and 150 minutes. Whereas X diminishes

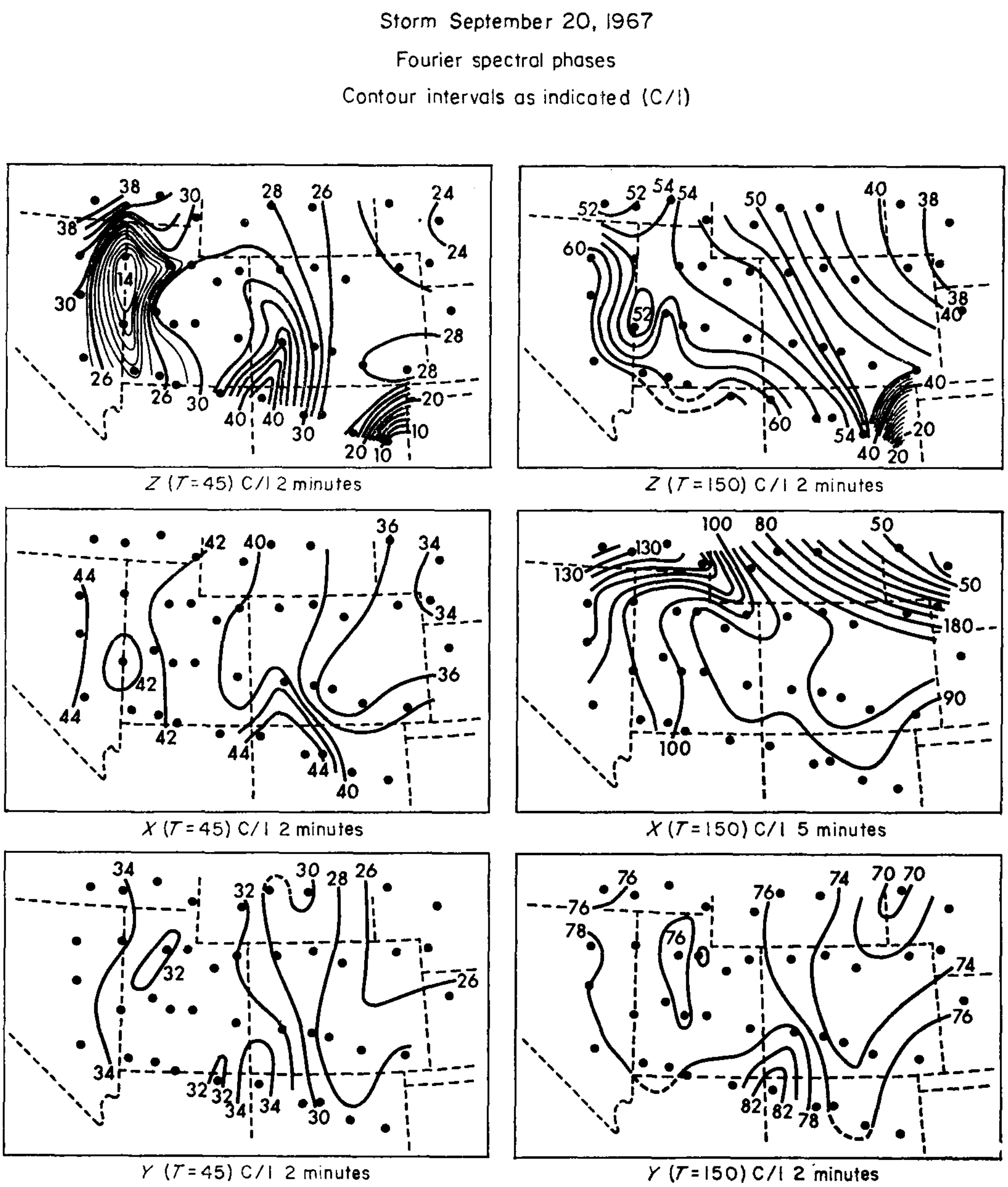

FIG. 13.

southward at period 45 minutes, it actually increases southward over the whole array at $T = 150$ minutes. Other maps, not shown, indicate that the reversal occurs close to $T = 90$ minutes. If the external field decreases southward in X amplitude, this reversal in gradient must be caused by an anomalous internal X field. The effect could be caused by a rise in the isotherms southward from the Colorado Plateau to the Basin and Range Province. The Z map shows a large gradient from north to south which is consistent with the interpretation offered for the X reversal. Failure of stations in the eastern part of line 2 to record Z for the storm makes it impossible to say whether an appropriate local maximum exists in Z: such a maximum should appear north of an east–west striking step in the conductive mantle as proposed.

The phase maps are shown not only because they contain information with regard to anomalous conductivity structures, but also because the phase must be known to enable the significance of the amplitude maps to be assessed. In general the phases vary smoothly across each map. In all cases the phase lags to the west, for a reason associated with the source current already indicated. There are anomalies in the phase superposed on the general east–west variation. A strong Z phase anomaly lies west of the Wasatch Front and is most marked at periods 30 and 45 minutes in the substorm maps (Fig. 9) and at 45 minutes for the storm (Fig. 13). At the shortest periods the phase relationships would be physically impossible for an induction process by external Z, but when it is remembered that anomalous internal Z is induced by Y the phase anomalies are consistent with local linear current along the Wasatch Front as already proposed to account for the amplitude anomaly. The

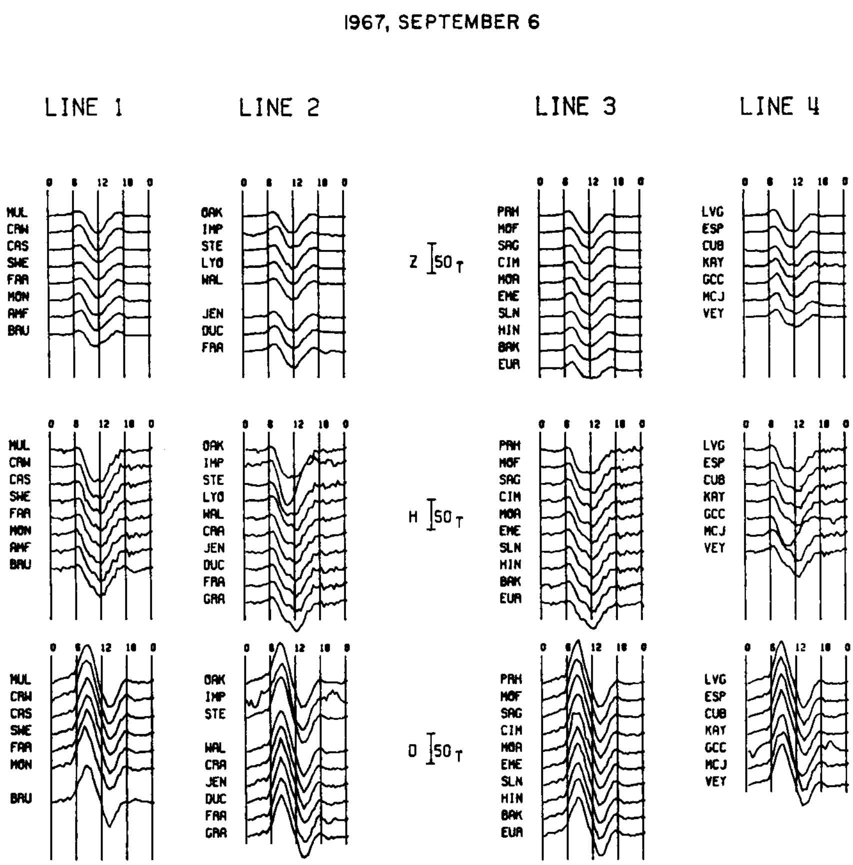

FIG. 14. Variograms of daily variation 1967 September 6.

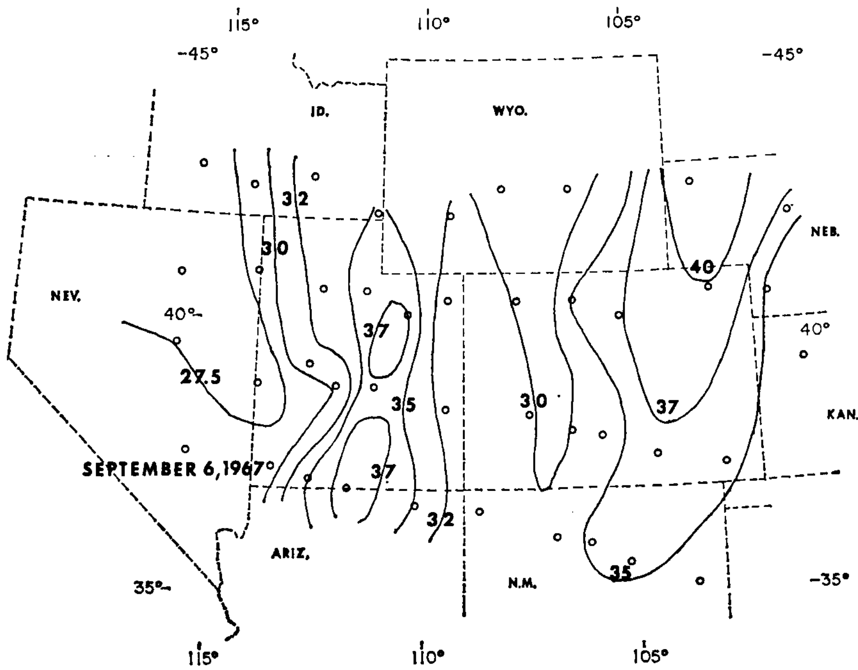

FIG. 15. Contour map of vertical component ranges in gamma for daily variation 1967 September 6.

phase anomaly in fact represents the reversed Z field west of the Wasatch Front. The decrease of the phase anomaly with increasing period indicates that the isotherms west of the Wasatch Front remain higher than under the Colorado Plateau. Another striking phase anomaly is associated with a reversal which is easily seen in the variograms at CUB in line 4. The northern limit of this anomaly is indeterminate because JEN and CRA in line 2 did not record the Z component. This anomaly is associated with the line current along the Southern Rockies already inferred from the amplitude anomaly. The persistence of this phase anomaly to $T = 89$ minutes indicates a much deeper step in the isotherms from the Southern Rockies to the Colorado Plateau than that found west of the Wasatch Front.

8. Daily variation

H, D and Z have been read every half hour for three quiet days (September 5, 6 and 11). The recording efficiency for the daily variation was somewhat less than for the events of shorter duration, as during the frequent inspections of the instruments the torsion rods were reset. This caused a considerable non-linear drift of the sensor magnets for about half a day thereafter, which could not be estimated.

A Calcomp plot of the daily variation of September 6, corrected for local time differences between stations, is shown in Fig. 14. Amplitude and phase differences between stations are small and irregular for the horizontal components. The vertical component shows a reduction for Basin and Range stations and a small enhancement at the Southern Rockies and the Wasatch Front, consistent with the behaviour of bays and storms. These features are best seen in the contour map of the ranges of the vertical component of the daily variation for September 6, shown in Fig. 15.

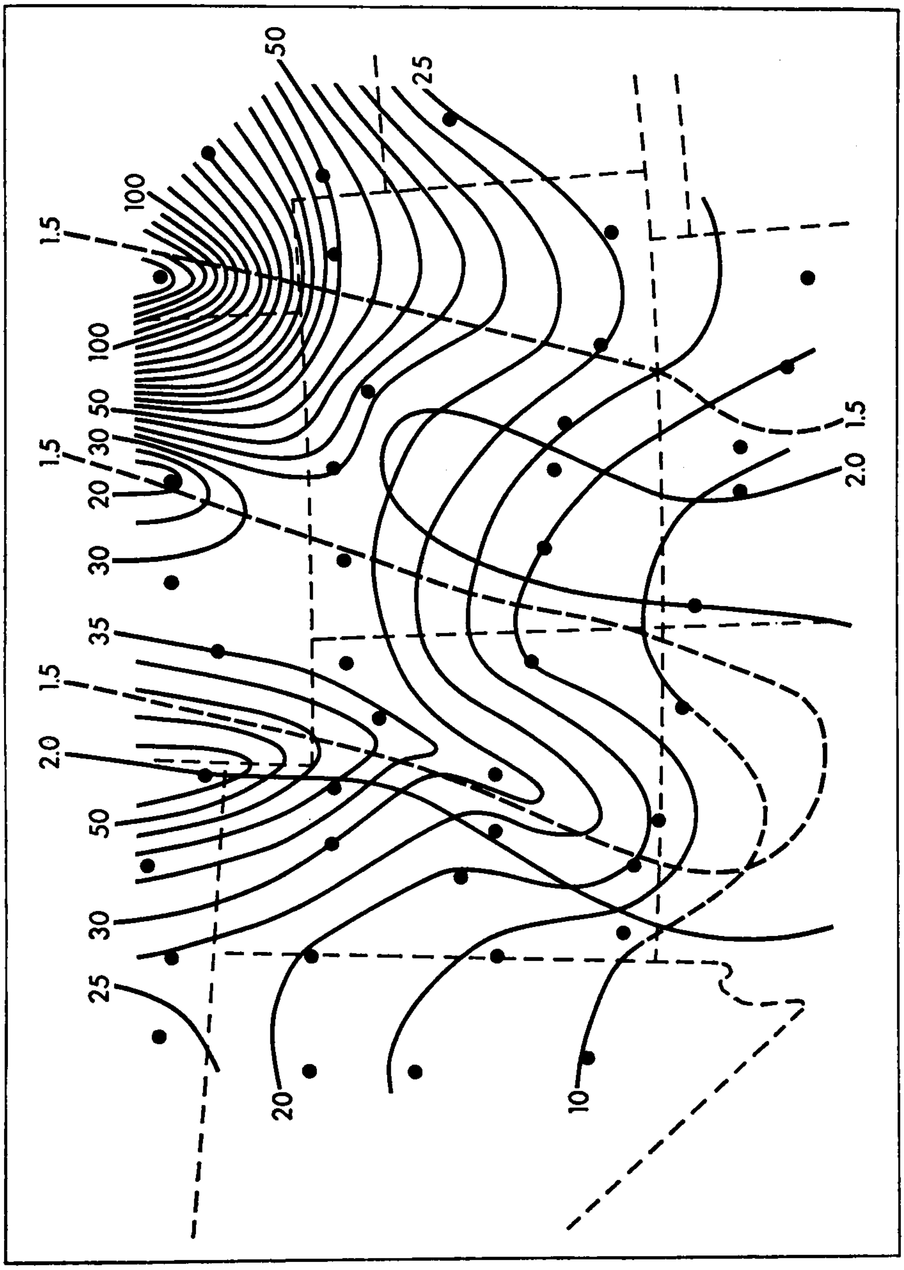

FIG. 16. Heat flow contours superimposed on the vertical component amplitude map for period 60 minutes.

$S_q(Z)$ increases at the eastern front of the Southern Rockies and the Wasatch Front. This indicates that even the deep current systems induced by the daily variation, whether at period 24, 12 or 8 hours or at two or three of these, are influenced to some degree by the conductivity anomalies of the Southern Rockies and the Wasatch Front. Induction by these long-period fields suggests that the observed conductivity anomalies extend to great depths and may be continuous with the highly-conducting mantle below 600 km depth.

9. Discussion

We have shown that the electrical conductivity anomalies follow closely the boundaries of the structural provinces. The conductivity rises from the Great Plains to the Southern Rockies and falls again under the Colorado Plateau. A second upwelling of the conductive mantle is located under the Wasatch Front and mantle conductivity decreases somewhat towards the Basin and Range. As a first interpretation one could model the observed anomalies by two north–south trending ridges in conductive mantle under the Southern Rockies and the Wasatch Front. An interpretation of maps of separated fields in terms of conductivity models of this type is the subject of another paper (Porath *et al.* 1970).

Seismic velocity data for the upper mantle are still too scarce to delineate the boundaries with the accuracy we have achieved with geomagnetic deep sounding. Our results are, however, in close agreement with the heat flow results in western North America. Fig. 16 shows the contours of the 2·0 and 1·5 μcal cm^{-2} s^{-1} isolines in the western United States superimposed on a Z amplitude contour map ($T = 60$ minutes) from the substorm (Fig. 8). The heat flow contours are taken from a map by Roy, Blackwell, Decker and Decker (in preparation) by kind permission of these authors. The heat flow results do not show the local maximum over the Wasatch Front, above the level in the Basin and Range, as indicated by the geomagnetic deep sounding results. Even with much denser distribution of heat flow observations this structure might be too narrow to be seen. Increased seismicity over the Wasatch Front (Woollard 1958) may indicate that it is tectonically different from the more stable Basin and Range.

In Fig. 16 it will be noted that the heat-flow contours dividing the Basin and Range Province from the Colorado Plateau track just west of the maximum of the anomaly in Z amplitude. They must lie virtually under the maximum of the Y anomalies and thus along the linear current. The maximum in the heat-flow which runs northward up the Southern Rockies likewise lies beside the Z maximum of our anomaly and under the Y maximum.

A recent study of the P_n velocity distribution in the United States (Tucker, Herrin & Freedman 1968) shows generally lower velocities under the Cordillera than elsewhere in the United States. Their map defines the Basin and Range Province clearly as a low-velocity region, but does not show the Southern Rocky Mountains as such, as the heat flow and geomagnetic deep sounding results would lead one to expect. The discrepancy may reflect lower resolution in the seismic information.

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References

Birch, F., 1961. The velocity of compressional waves in rocks to 10 kbars, Part 2, *J. geophys. Res.*, **66**, 2199–2224.

Blackwell, D. D., 1969. Heat flow determinations in the North-western United States, *J. geophys. Res.*, **74**, 992–1007.

Caner, B., Cannon, W. H. & Livingstone, C. E., 1967. Geomagnetic depth sounding and upper mantle structure in the Cordillera Region of Western North America, *J. geophys. Res.*, **72**, 6335–6351.

Clearly, J. & Hales, A. L., 1966. An analysis of the travel times of *P* waves to North American stations, in the distance range 32° to 100°, *Bull. seism. Soc. Am.*, **56**, 467–489.

Doyle, H. A. & Hales, A. L., 1967. An analysis of the travel times of *S* waves to North American stations, in the distance range 28° to 82°, *Bull. seism. Soc. Am.*, **57**, 761–771.

Gough, D. I. & Reitzel, J. S., 1967. A portable three-component magnetic variometer, *J. Geomagn. Geoelect., Kyoto*, **19**, 203–215.

Gough, D. I. & Reitzel, J. S., 1969. Magnetic deep sounding and local conductivity anomalies, in *The Application of Modern Physics to the Earth and Planetary Interiors*, ed. by S. K. Runcorn, Interscience Publishers, New York, in press.

Hales, A. L. & Doyle, H. A., 1967. *P* and *S* travel-time anomalies and their interpretation, *Geophys. J. R. astr. Soc.*, **13**, 403–415.

Herrin, E. & Taggart, J., 1962. Regional variations in P_n velocity and their effect on the location of epicenters, *Bull. seism. Soc. Am.*, **52**, 1037–1046.

Herrin, E. & Taggart, J., 1968. Regional variations in *P* travel times, *Bull. seism. Soc. Am.*, **58**, 1325–1337.

Lahiri, B. N. & Price, A. T., 1939. Electromagnetic induction in non-uniform conductors, and the determination of the conductivity of the Earth from terrestrial magnetic variations, *Phil. Trans. R. Soc. London*, **A237**, 509–540.

Lee, W. H. K. & Uyeda, S., 1965. *Review of heat flow data, in Terrestrial Heat Flow, Geophys. Monograph* 8, ed. by W. H. K. Lee, pp. 87–190, American Geophysical Union, Washington, D.C.

Parkinson, W. D., 1959. Directions of rapid geomagnetic fluctuations, *Geophys. J. R. astr. Soc.*, **2**, 1–14.

Parkinson, W. D., 1962. The influence of continents and oceans on geomagnetic variations, *Geophys. J. R. astr. Soc.*, **4**, 441–449.

Porath, H., Oldenburg, D. W. & Gough, D. I., 1969. Separation of magnetic variation fields and conductive structures in the western United States, *Geophys. J. R. astr. Soc.*, **19**, 237–260.

Rikitake, T., 1966. *Electromagatism and the Earth's interior*, p. 308, Elsevier Publishing Co., Amsterdam, London, and New York.

Rostoker, G., 1968. Macrostructure of geomagnetic bays, *J. geophys. Res.*, **73**, 4217–4229.

Roy, R. F., Decker, E. R., Blackwell, D. D. & Birch, F., 1968. Heat flow in the United States, *J. geophys. Res.*, **73**, 5207–5221.

Schmucker, U., 1964. Anomalies of geomagnetic variations in the South-western United States, *J. Geomagn. Geoelect.*, **15**, 193–221.

Siebert, M., 1964. Ein Verfahren zur unmittelbaren Bestimmung der vertikalen Leitfähigkeitsverteilung im Rahmen der erdmagnetischen Tiefensondierung, Nachr. d. Akad, d. Wiss. Göttingen, *Math. Phys. Klasse, Jahrgang*, 25–35.

Tozer, D. C., 1959. On electrical properties of the Earth's interior, in L. H. Ahrens, F. Press, K. Rakama and S. K. Runcorn (Editors), *Physics and Chemistry of the Earth*, Pergamon, London, **3**, 414–436.

Tucker, W., Herrin, E. & Freedman, H. W., 1968. Some statistical aspects of the estimation of seismic travel times, *Bull. seism. Soc. Am.*, **58**, 1243–1260.

Wiese, H., 1962. Geomagnetische Tiefentellurik. II. Die Streichrichtung der Untergrundstrukturen des elektrischen Widerstandes, entschlossen aus geomagnetischen Variationen, *Geofis. Pura Appl.*, **52**, 83–103.

Woollard, G. P., 1958. Areas of tectonic activity in the United States as indicated by earthquake epicenters, *Trans. Amer. geophys. Union*, **39**, 1138–1150.