Mantle Conductive Structures in the Western United States from Magnetometer Array Studies

H. Porath and D. I. Gough

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Summary

Geomagnetic substorm fields were simultaneously recorded during the summer of 1968 at stations of a two-dimensional array of variometers covering western Texas, New Mexico, Arizona and eastern California. This work formed part of a continuing study of upper mantle conductive and thermal structure under the western United States. Magnetograms and maps of Fourier spectral components of two substorms are used to describe conductive structures in the upper mantle and crust. Conductive structures under the Southern Rockies and Wasatch Fault Belt, discovered in a similar array study in 1967, are shown to continue southward. The Basin and Range Province as a whole overlies a highly conducting (and, by inference, hot) region in the mantle. Conditions are discussed which must be satisfied by any assumed regional field used to isolate and normalize the anomalous fields. Anomalous fields satisfying these conditions are derived from the 1967 array data supported by the 1968 array. Numerical calculations are used to show that the anomalies cannot be associated with conductors of limited thickness within the crust, unless these have improbably high conductivity. Two-dimensional ridges on the surface of a semi-infinite conductor, on the other hand, satisfactorily model the observed fields. The conductive structures can therefore be assigned with some confidence to the upper mantle. The anomalies are well represented by induction in a half-space of conductivity $0.2 \text{ (ohm m)}^{-1}$ with a ridge under the Southern Rockies and a step with superposed ridge under the Wasatch Fault Belt. The depths to the surface of the conductor in our model are 190 km under the Basin and Range Province, 120 km under the Wasatch Fault Belt, 150 km under the Southern Rockies and 350 km under the Colorado Plateau and the Great Plains. Such models are naturally not unique.

1. Introduction

During the summer of 1968 The University of Texas at Dallas and the University of Alberta again operated a combined array of variometers for geomagnetic deep sounding studies in the south-western United States. It was the purpose of this work to look for a southward continuation of the magnetic variation anomalies found in a similar array study across the central Rocky Mountain States during the summer of 1967 (Reitzel, et al., 1970; Porath, Oldenburg & Gough 1970). Variometer stations of both field seasons are shown in Fig. 1 on a simplified tectonic map of the western United States.
In 1967 we observed anomalous vertical and east–west variation fields related to concentrations of electric current flowing north–south under the Southern Rockies and the Wasatch Fault Belt. This fault belt forms the boundary between the Colorado Plateau and the Basin and Range Province in Utah. We interpreted the observed anomalies in terms of lateral changes in electrical conductivity produced by temperature variations in the upper mantle. The Southern Rockies and the Basin and Range Province are regions of high electrical conductivity. An additional increase in conductivity above the Basin and Range average was observed under the Wasatch Fault Belt, whereas the Colorado Plateau shows low mantle conductivity very similar to that of the Great Plains Province east of the Rockies. Simple models based on perfectly conducting structures were used to approximate the anomalous vertical and horizontal fields (Porath et al. 1970).

The upper mantle structure deduced from geomagnetic deep sounding studies appears to agree with that derived from variations in heat flow and from seismic data. The Colorado Plateau is a region of normal heat flow in contrast to high heat flow in the Basin and Range and the Southern Rockies (Roy et al. 1968; Blackwell 1969).

Julian (1970) has shown that the low velocity zone is thicker and shallower under the Basin and Range and Southern Rockies provinces, and deeper and less developed under the Colorado Plateau and Great Plains. There is thus a general correlation between enhanced electrical conductivity and development of the seismic low-velocity zone. Both are probably controlled by temperature. It is not implied that the conductive and seismic structures are at the same depth. Other seismological studies
Mantle conductive structures (Herrin & Taggart 1962, 1968; Cleary & Hales 1966; Doyle & Hales 1967; Archambeau, Flinn & Lambert 1969) distinguish the Basin and Range Province from the Great Plains but do not resolve the Colorado Plateau from the Southern Rockies.

The 1968 array covered the area from West Texas across New Mexico and Arizona into California. It overlapped our 1967 array, and included the variometer profile operated by Schmucker (1964) across Southern Arizona, New Mexico and West Texas. Schmucker observed that the vertical variation field was very much attenuated west of the boundary between the Basin and Range and the Great Plains in Southern New Mexico and attributed this to an abrupt increase in mantle conductivity under the Basin and Range. A reversal in the vertical variation field for periods of 15–30 min at a station just west of the boundary led Schmucker to suggest an additional local increase in conductivity, above the Basin and Range average, under the Rio Grande Rift Valley.

2. Magnetograms

The array of 1968 operated with lower efficiency than those of 1967 and 1969, because problems in the film transport system had not yet been overcome. However, about 90 per cent of the stations recorded one or other of two variation events.

**FIG. 2.** Magnetograms for a substorm of 1968 August 16.
Fig. 2 shows magnetograms for a magnetic substorm on 1968 August 16. Stations are plotted from east to west; Line 1 is the northernmost line across Arizona and New Mexico. Downward Z, northward H and eastward D are positive.

The event of 1968 August 16 is far from ideal for geomagnetic deep sounding purposes, as the vertical component shows a large long period trend, which makes any shorter period anomalies less obvious. The magnetograms show the following features in the vertical component. Large Z variations resembling the eastward horizontal fields are observed east of the Southern Rockies (SPR, GLN) and a reversal appears in Z at the edge of the Southern Rockies and the Colorado Plateau (GOC, SNF). These features indicate a north–south current concentration under the Southern Rockies, as shown in our 1967 results. The large Z amplitude at LEO and the reversal in Z at DAL are attributed to sedimentary conductive structures to be discussed later.

The variation anomaly observed by Schmucker is evident in the decrease in Z between CLB and DEM. DEM suggests a slight reversal in Z which indicates a local current concentration at the edge between the Basin and Range and the Great Plains under the Rio Grande Rift Valley (Schmucker 1964).

At the boundary between the Colorado Plateau and the Basin and Range we observe increased Z amplitudes on Line 1 due to higher conductivities under the Basin and Range (PAG and KAN). This anomaly is already very much reduced on Line 2, where the boundary between the Colorado Plateau and the Basin and Range begins to swing east. There appears to be a small reversal in Z at RED, PEA, KIN and WIN, but this may be a feature of the normal Z field. The latter interpretation is preferred, as other events do not show reversals at these stations.

Stations in the Basin and Range show strongly attenuated vertical variations except at YUM at the Arizona–California border and at MOJ, a station close to the Sierra Nevada Batholith. MOJ shows a strong reversal in Z indicating higher conductivities under the Basin and Range compared with the Sierra Nevada. Other stations operated by Schmucker (1970) over the Sierra Nevada–Basin and Range boundary give similar results. This agrees with the anomalously low heat flow observed in the Sierra Nevada and the high heat flow in the Basin and Range (Lachenbruch, et al. 1966; Blackwell 1969).

The phase relation at YUM between the vertical and horizontal fields is not clear. It appears from the event of 1968 October 2 (Fig. 3) that downward Z is in phase with easterly D. Schmucker (1970) has computed a Parkinson vector for this station using numerous events. This vector points to higher conductivities south-west of this station, which may be related to a region of high temperature and associated high electrical conductivity under the Gulf of California where the East Pacific Rise terminates.

For the event of 1968 October 2, we again observe similar variation anomalies in the vertical component. Large variations characterize stations east of the Rockies (CAN, SPR, TUL, CLC) and a reversal is observed at the Colorado Plateau–Southern Rockies boundary (SNF). A sharp attenuation of the vertical component is shown by stations west of the boundary of the Great Plains and the Basin and Range between CAZ and SNM and generally very subdued variations are observed for stations within the Basin and Range Province (VEY, GLE, SHO, KIN, AGU, BLY, COA, SNM, LUN, DOU, TUS, WHY). At the boundary between the Colorado Plateau and the Basin and Range we observe large anomalous vertical variations induced by the east–west variations (PAG, KAN). As for the August 16 event this anomaly is very much reduced on Lines 2 and 3 (WIN, RED, PEA, PHO, AGU).

There appears to be a local anomaly at GER in the Basin and Range. The Z trace at this station is noisy because this variometer had unusually low sensitivity. GER is close to a magnetotelluric station (Safford) operated by Swift (1967), which showed effects of a nearby graben filled with conducting sediments. We also attribute
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1968 OCTOBER 2

LINE 1
LINE 2
LINE 3
LINE 4

FIG. 3. Magnetograms for a substorm of 1968 October 2.

this anomaly to a local sedimentary feature, as it does not persist over greater distances and has, as we shall show, no corresponding horizontal field anomaly associated with it towards the west (Section 4). The magnetic field variations at LEO, CHI and DAL are strongly affected by sedimentary features in the Great Plains Province. Current concentrations in the conducting sediments of the 38 000 ft deep Anadarko Basin of south-east Oklahoma cause reduced vertical variations towards the south-west of the basin (CHI) and enhanced $Z$ to the north-east (LEO). This sedimentary anomaly misled Caner, Cannon & Livingstone (1967) to suggest that the Rocky Mountains anomaly swung towards the east in this region. With our closer station spacing we have established the existence of two distinct anomalies, one a mantle feature associated with the Southern Rockies and the other an upper crustal anomaly due to a sedimentary basin.

At Dallas the $Z$ variations show the effect of a boundary between two sedimentary provinces of different conductivities. This boundary is formed by the Ouachita Tectonic Belt which divides the Palaeozoic sediments of west Texas from the more conductive younger sediments of the coastal plains.

3. Spectra

The substorm events are treated as transients and Fourier transforms of the magnetograms are computed by means of the Cooley Tukey algorithm (Reitzel et al. 1970). Examples of the amplitude spectra for the event of 1968 October 2 at SPR are shown in Fig. 4. The peaks in the amplitude spectra of the three variation
components do not coincide at all periods. This can be attributed to the fact that the variation event is not a single homogeneous substorm, but a more complex event. In general, magnetic substorms are elliptically polarized, which means that phase differences occur between the two horizontal components. In the Fourier transform technique one computes an average spectrum from a finite data set. For a data set consisting of two or more substorms the spectra will depend on the amplitude ratios and phases of the horizontal components within each substorm, and on the time differences between substorms. In general, a multiple event of this kind will give maxima of the two horizontal components at different periods. Minima in one component may correspond to maxima in the other. The positions of spectral peaks will also shift over the array, as the relative phase differences between the horizontal substorm components change in the ionospheric source fields with changes in the current system (Rostoker et al. 1970).

Certain events containing substorm components with phase differences between their horizontal components give substantially different amplitude spectra for $H$ and $D$. Such events may show a strong correlation between the vertical variation spectrum and that of some horizontal component. This gives an indication of the trend of the conductive structure. Thus, in Fig. 4 peaks in the amplitude spectra of $Z$ and $D$ are correlated: SPR is close to a north–south trending structure (the Southern Rockies). Further examples of this method of defining trends are given by Camfield, Gough & Porath (1970) for magnetic events recorded in the north-western United States and south-western Canada.

Fig. 4. Amplitude spectra of the substorm of 1968 October 2 at Springer, New Mexico.
4. Maps of spectral components

For the presentation of the anomalies in the form of contour maps, amplitudes and phases for the two events have been chosen for periods as close as possible to the spectral peaks at all stations. For the August 16 event the contour maps are shown for a period of 66 minutes (Fig. 5). An increase in the east–west component \( D \) is observed over the Southern Rockies. This maximum and two extrema in the vertical field on either side of it delineate a concentration of induced currents under the Rockies. The phase map shows that the two vertical field extrema are of opposite sign. There is a \( D \) field anomaly in northern Arizona just west of the boundary between the Colorado Plateau and the Basin and Range Province. The vertical field anomaly is evident in the phase maps, but no pronounced amplitude anomaly is observed. This may be caused by destructive interference between the spectra of the long period trend observed for this particular event and the anomalous \( Z \) variation.

The event of October 2 was recorded at more stations in the central part of the array. Contour maps are shown in Fig. 6 for a variation period of 54 minutes. These maps show similar features in \( D \) and \( Z \) to those of the maps of August 16, with the exceptions of a much more pronounced \( Z \) amplitude anomaly at the edge between the Colorado Plateau and the Basin and Range and a local high at GER in Southern Arizona. There is, however, no correspondingly large anomaly in the \( D \) field west of the \( Z \) anomaly and the GER anomaly is associated with a local feature in the upper crust. A phase reversal in \( Z \) is again observed over the Southern Rockies. Phase contour maps for the horizontal components show the east–west phase shift characteristic of substorm fields, which can be attributed to westward current surges in the ionospheric source fields (Rostoker et al. 1970).

![Fig. 5. Amplitudes and phases at period 66 min. for the substorm of 1968 August 16. Units arbitrary.](https://academic.oup.com/gji/article-abstract/22/3/261/566498/566498)
In both Figs. 5 and 6 the anomalies in the eastward field at the boundary between the Colorado Plateau and the Basin and Range decrease from the northern edge of the map. The only anomaly of importance in $H$ is at this boundary towards the northern limit of the array in Fig. 6. This probably reflects a departure from north-south strike of the structure.

From the vertical field anomaly it appears that the major structure at the boundary between the Colorado Plateau and the Basin and Range in Northern Arizona is a steep increase of electrical conductivity under the Basin and Range. With the available station spacing another extremum or even a reversal in $Z$ may have been missed west of this boundary. Such a reversal was observed in our 1967 data at the Utah–Nevada border for variation periods of about 30 min. It is difficult to decide from the anomaly in the east-west fields, with the present coverage, whether we have a symmetrical or an asymmetrical structure.

5. Anomalous fields

The variometer arrays of 1967 and 1968 have yielded maps of magnetic variation anomalies in the western United States between latitudes 32° and 45° N. Because of the large north-south extent of these anomalies it appears possible to approximate the observed anomalous fields at suitably placed east-west profiles by two-dimensional conductive structures. The best position for such a profile is across southern Colorado and southern Utah into Nevada (Fig. 1). This profile is well removed from the ends of the anomalous structures. This region was covered in detail with our 1967 array and a substorm of 1967 September 1 yielded anomalous fields for the three variation periods 32.5, 50 and 89 min (Reitzel et al. 1970; Porath et al. 1970).
In addition a reasonable two-dimensional approximation of the anomalies should be possible along a profile across the boundary between the Great Plains and the Basin and Range in southern New Mexico. The anomalous fields along this profile have been obtained by Schmucker (1970) by a transfer function technique from numerous events. Our data have confirmed the anomaly pattern observed by Schmucker.

In the analysis of the 1967 data it was shown by a separation analysis that the variation anomalies are due to internal current systems (Porath et al. 1970). The normal field, which consists of external and internal fields with scale lengths much larger than the dimensions of the array, cannot be separated (Rikitake 1966, p. 148). Any separation analysis, therefore, removes only the external inhomogeneous part, which is usually small in mid-latitudes. Thus, removal of a long wavelength regional field from the observed variation fields leaves substantially anomalous fields of internal origin.

In the analysis of the 1967 data the anomalous fields in the horizontal east–west variations were obtained by removing an estimated normal field from the contour maps of the sine and cosine transforms. The normal field was estimated by fitting planes to these maps by least squares. After the vertical field had been separated into parts of external and internal origin, the normal field in the vertical component was approximated by substituting the normal horizontal fields in the field separation formula (Porath et al. 1970, p. 253). With the regional horizontal normal field determined in this way we obtained a considerable phase difference of about 35° between the anomalous and normal fields. The in-phase normalized anomalous fields were then approximated by a perfect-conductor model to indicate the shape and maximum depth of each conductive structure.

From the magnetograms, however, it can be seen (Reitzel et al. 1970, Fig. 2) that the vertical fields in the strongly anomalous regions, east of the Southern Rockies and of the Wasatch Front, are closely in phase with the east–west fields, and that phase differences cannot be larger than about 15° for the variation periods considered. This indicates that removal of the regional normal horizontal field by least squares methods does not give the correct base line from which to compute the anomalous fields. Small phase angles between the vertical fields and the horizontal inducing fields imply small phase differences between the anomalous and normal horizontal fields. One has therefore to draw base lines to profiles of the east–west fields of the sine and cosine transforms in such a way as to make the phase angle small. This is not difficult for a single anomaly, which has been mapped well outside the anomalous region and for which the regional normal fields do not change appreciably along the profile. The normal vertical field is then removed separately from the sine and cosine transform profiles to match the phase differences observed between the vertical and the horizontal inducing fields. For symmetrical anomalies the range of anomalous \( Z \) should equal the range in anomalous \( H \).

For a multiple anomaly such as is observed in the western United States, it is much more difficult to define a normal field as baseline for the anomalies. The difficulty is accentuated by two factors. Over such long east–west profiles the regional normal field varies substantially, and phase changes in the east–west horizontal fields observed across the array (Rostoker et al. 1970) impose an additional gradient on the sine and cosine transform coefficients.

The procedure chosen to define the anomalous fields along the profile shown in Fig. 1 was the following. The baselines of the profiles of the sine and cosine transforms of the east–west fields were adjusted to minimize the phase angles between anomalous and normal horizontal fields. The separated external fields (Porath et al. 1970) provided an additional control on the general form of the baselines.

To estimate the normal vertical field two criteria were used. The first is that the anomalous vertical field amplitude east of the Rockies should be half the anomalous
eastward field; this assumes that the vertical field anomaly in this region arises only from a current concentration under the Rockies and is not influenced by any other conductivity anomaly. The second criterion is that the phase difference between the normal horizontal and the anomalous vertical field is not greater than 15°. It is further assumed that normal $Z$, which arises from induction by external $Z$, is constant along the east–west profile. This is certainly only an approximation, as normal $Z$ will be smaller for stations in the Basin and Range Province, where the good conductor is at shallower depth than further east. It is, however, a good approximation, as is shown by the small east–west gradient of the external vertical field over the central array (Porath et al. 1970, Fig. 10a). The normalized anomalous horizontal and vertical fields at three periods are shown in Fig. 7. The form of the anomalous fields is very similar to those obtained previously from a separation analysis with the use of least squares methods to approximate the regional field, except near the edges of the profile, especially the western end. It must be remembered, however, that a separation is inaccurate near the edges of the array, particularly for the separated vertical fields, whereas for the determination of the anomalous fields in Fig. 7 we have used the observed fields and assumed the external inhomogeneous part to be small.

6. Models of conductive structures

There will naturally be a range of models that satisfy the observed anomalous fields, as one can only set an upper bound (15°) to the phase difference between anomalous and normal fields. Computation of transfer functions from numerous events as described by Schmucker (1970) would probably have increased the accuracy of determination of the vertical anomalous fields. For the horizontal anomalous fields the transfer function method, however, meets the difficulty discussed in Section 5, i.e. the determination of the regional normal field.
The perfect-conductor model presented earlier (Porath et al. 1970) gives a good account of the shape of the conductive structures which produce the observed variation anomalies. For a more realistic representation of the structures on the basis of the revised normalized anomalous fields we have attempted to approximate the anomalies by structures of finite conductivity using the numerical methods suggested by Madden and Swift (Swift 1967) as modified and programmed by Wright (1969). Wright calculates the anomalous horizontal and vertical fields, as well as phase differences, for two-dimensional structures by solving Maxwell's equations numerically. His method uses a transmission surface analogy as described by Swift (1967). The conductive structures are approximated by blocks of uniform conductivity and electric and magnetic fields are calculated for incident fields polarized with either electric or magnetic field vector along the strike of the structures. In geomagnetic deep sounding problems one is concerned only with the case of the electric field parallel to the strike of the conductor. A field with $H$ along the structure does not result in anomalous magnetic fields at the surface of the Earth. Wright's method takes account of induction in, and current distribution among all conductors in the model.

Models fitted to the observed anomalous fields will be necessarily non-unique. However, any model must satisfy the spatial distribution of the anomalous fields, the phase differences between normal and anomalous fields and the amplitude of the normalized anomalous fields as a function of frequency. These requirements allow limits to be set to possible models.

Structures in conducting sediments in the upper crust cannot be responsible for the anomalies observed in the western United States, as the phase differences should be characteristically 45° or more for reasonable sedimentary conductivities (Wright 1969). The same argument excludes an isolated good conductor in the lower crust of the Earth, such as has been suggested by Hyndman & Hyndman (1968) and by Caner (1970) to account for variation anomalies related to tectonic provinces. We have computed the electromagnetic response of a layer of resistivity 5 Ohm m between 20 and 30 km depths in the lower crust, with the width of the Southern Rockies (300 km), within Cantwell's normal conductivity section (Cantwell 1960). Phase differences of 35° or larger were encountered at the periods considered. In addition the amplitudes of the anomalous fields for this model varied considerably in the period range 30 to 90 min, whereas the observed anomalous fields of Fig. 7 give comparable amplitudes at all three periods. By going to higher conductivities or very much thicker conductors one can reduce these phase differences between anomalous and normal fields, but it is difficult to assume much lower resistivities than 5 Ohm m for conductors in the lower crust.

We have, therefore, approximated the observed anomalous fields by those produced by models of conductive structures in the upper mantle. The model consists of ridges and steps in a half space of resistivity 5 Ohm m in a 1000 Ohm m matrix. It would be possible to compute more complex models. However, this would be unwarranted in view of the simplifying assumptions in the model studies, such as uniform horizontal fields and two-dimensional structures. Further, the phase difference between normal and anomalous fields is subject to errors of several degrees, and superficial conductors introduce noise in the data.

Fig. 8 compares the observed normalized anomalous fields with the fields of the model for the three variation periods. For the horizontal fields the fit is reasonable considering the station spacing. The fit for the vertical anomalous fields, which are much more sensitive to the actual geometry of the structure than the horizontal fields, is less good, especially at the Wasatch Front; here the model gives consistently lower $Z$ amplitudes than the observed vertical fields. West of the Wasatch Front the model matches the reversed $Z$ amplitudes fairly well for variation periods of 32.5 and 50 min, but agreement is not so satisfactory for 89 min. A model with the top of the conductor under the Wasatch Fault Belt at 140 km depth instead of 120 km
The model proposed to approximate the conductor under the Southern Rockies is symmetrical, implying that the upper mantle structure under the Colorado Plateau is similar to that under the Great Plains. Within the accuracy of our data it would, however, be possible to make the depth to the conductor under the Colorado Plateau somewhat less than that under the Great Plains. Nevertheless, the upper mantle structure under the Colorado Plateau resembles that under the Great Plains much more closely than that under the Basin and Range Province.

We have similarly approximated the anomalous horizontal and vertical fields determined by Schmucker (1970) as transfer functions for numerous events (Fig. 9). The boundary between the Great Plains and the Basin and Range in southern New Mexico showed a similar structure to that between the Colorado Plateau and the Basin and Range in Utah. Depth values were comparable to those proposed by Schmucker on the basis of perfect conductivity. Once again the phase difference was small.

There is considerable scatter in the experimental data on the electrical conductivities, at high pressures and temperatures, of silicates that presumably make up the upper mantle. Measurements by Schult & Schober (1969) of the electrical conductivity of natural olivine show that a resistivity of 5 Ohm m (assumed in the model structure) is reached at pressures of 30 kbars and 1000°C. On the other hand, extrapolation to higher temperatures of experimental results for synthetic olivine by...
Hamilton (1965) and for peridot by Hughes (1955) indicate that this resistivity corresponds to a temperature closer to 1500°C. The conductivity-temperature dependence is also sensitive to the fayalite content. Our model shows a normal depth of the conductor under the Great Plains of 350 km. This depth would correspond to about 1500°C on the Clark & Ringwood (1964) model for continental shields and would be within the range of the temperatures determined experimentally for the resistivity of the model.

We have offered an interpretation of the variation anomalies in the western United States in terms of undulations in a highly conducting half-space modelling the relief in mantle isotherms. To illustrate the non-uniqueness of such interpretations, we have calculated a model which assumes the variation anomalies to be caused by thickening of a 5 Ohm m layer under the Southern Rockies, the Wasatch Front and the Basin and Range, but of minimum thickness only 25 km under the Great Plains and the Colorado Plateau (Fig. 10). Beneath this zone the resistivity increases again tenfold to 50 Ohm m, before it decreases to 10 Ohm m at a depth of 500 km. As is shown in Fig. 10 for a period of 30 min, the vertical and horizontal anomalous fields are almost identical with those of the previous model and the phase angles differ by only a few degrees. Geomagnetic deep sounding measurements will therefore not be able to distinguish between these two types of models. No particular geological significance is claimed for the model of Fig. 10.
Fig. 10. Calculated anomalous fields for a conductive structure of finite thickness, compared with similar structures on a conducting half space as in Fig. 8.

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H. Porath:  
University of Texas at Dallas,  
(formerly Southwest Center for Advanced Studies)

D. I. Gough:  
University of Alberta,  
Edmonton,  
Canada

References


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