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Conductivity anomalies, with special reference to the Andes

Internal and External Parts of Geomagnetic Variations

Transient fluctuations of the Earth's magnetic field are commonly known as geomagnetic variations. They arise from shifting current systems in the ionosphere or beyond and diffuse through the conductive layers of the Earth's interior with amplitude reduction and phase rotation whilst inducing currents. The depth of these currents, and thereby the depth of penetration of the incident variation field, increases from a few kilometres for fast pulsations (60 cycles/hour) to hundreds of kilometres for the slow diurnal variations ($\frac{1}{24}$ cycle/hour).

The decisive frequency–conductivity parameter is the skin-depth value

$$p = \frac{1}{(2\pi\omega\sigma\mu)^{1/2}} \tag{1}$$

where $\omega = 2\pi f$ is the angular frequency of the incident field, μ the magnetic permeability (usually set to unity), and σ the electrical conductivity of the subterranean matter. All quantities are to be measured in electromagnetic c.g.s. units (e.m.u.). In the case of a uniform conductor, p is the depth beneath its surface where the amplitude of an incident electromagnetic field is attenuated to $1/e$ of its surface value, assuming that this depth is small compared with the lateral non-uniformity of the incident field. Rewriting (1) in convenient units for geomagnetic induction problems, namely in cycles/hour for f and $\Omega^{-1}\,m^{-1}$ for σ, gives

$$p = \frac{30 \cdot 2}{(f\sigma)^{1/2}}\ \text{km} \tag{1a}$$

The surface field of the internal eddy currents is superimposed upon the primary source field from above, and we distinguish accordingly between the *internal* (induced) and the *external* (inducing) part of geomagnetic variations as observed at the Earth's surface. Both parts can be separated when the spatial distribution of the variation field is known.

We are concerned here with local anomalies of the internal part, which are caused by an unequal distribution of subterranean conductivity, involving large gradients of σ in the horizontal direction. They produce characteristic differences of simultaneously

recorded variations at adjacent sites, say less than 100 km apart, which usually cannot be attributed to the smoothly varying primary field from above. Hence, a closely spaced network of temporary magnetic recording stations is needed to detect such internal conductivity anomalies.

Two inherent limitations of this method of *geomagnetic depth sounding* should be mentioned. Since the observations are made within small areas (small in comparison with the spatial extent of the primary field) a complete separation of internal and external parts is not possible. As a consequence, the average change of conductivity with depth in the surveyed area remains unknown and any local anomalies of geomagnetic variations have to be interpreted by lateral conductivity changes within a preconceived normal, i.e. stratified conductivity distribution which must be inferred from other sources of information (cf. p. 127).

Secondly, oceans and continental surface layers form a thin conducting cover of great complexity. The flow of superficial eddy currents is therefore highly distorted. This may lead to local anomalies of the internal part, in particular near coast lines because of the outstanding conductivity contrast of sea water and rock formations on land. Such surface effects have to be taken into account before conclusions are drawn about possible anomalies at greater depth.

Conductivity and Temperature

At first sight the electrical conductivity σ of subterranean matter may not seem to be a very noteworthy parameter. There is, however, its close relation to temperature, following the general theory of semiconduction, and even small changes of temperature can cause drastic changes of conductivity. Olivine, for instance, doubles its conductivity when the temperature is raised by just 50 degc in the range[1] from 1000 to 1250°c.

We have to bear in mind, however, that semiconduction in non-metallic solids is primarily an impurity effect. Thus, minute changes of composition, in particular of the iron content, can have an equally strong effect upon σ, not counting the largely unknown influence of pressure. This limits the effective use of the conductivity as *absolute thermometer* for the Earth's interior.

There remains, however, the important aspect to use σ as *relative thermometer*, namely to infer deep-seated lateral gradients of conductivity and possibly temperature from their distorting effect upon the internal part of geomagnetic variations. Such thermal imbalances in the upper mantle could be connected with ascending and descending branches of convection cells or with local concentrations of radioactive heat sources, which may be the underlying cause for the diversified tectonic and magmatic history of the Earth's outermost layers.

Conductivity Distribution in the Upper Mantle

The electric conduction in surface rocks is mainly electrolytic through salty solutions filling pores and cracks. Their conductivity varies accordingly between $1 \ \Omega^{-1} \ m^{-1}$ for unconsolidated clastic sediments and 0·001 for dense igneous rocks. Sea water in comparison has an average conductivity of 4, copper a conductivity of $10^8 \ \Omega^{-1} \ m^{-1}$.

Rocks become insulating under pressure when their pores and cracks are closed, and the Earth's crust and uppermost mantle must be indeed very poor conductors, There is clear evidence, however, that the conductivity rises again in the upper mantle. and it is not unreasonable to relate this rise to the downward increase of ambient temperature (see the previous section).

Two methods have been in use to infer the change of conductivity with depth by means of natural electromagnetic fields. The first and classical method is based on magnetic observations alone and uses the surface ratio of internal to external parts of geomagnetic variations, averaged on a global scale. Disregarding in this way regional differences Lahiri and Price[2] gave two possible distributions, representing limiting cases, which are compatible with the internal parts of semidiurnal S_q variations and smoothed storm-time D_{st} variations. In the first model 'e', an insulating intermediate layer extends downwards from the surface to 600 km depth, where the conductivity rises abruptly to infinity. The whole model is surrounded by a thin outermost shell which has an integrated total conductivity of $5 \cdot 1 \times 10^{-6}$ e.m.u. cm, equivalent to 1280 m sea water (cf. equation (5)). In the alternative model 'd', the conductivity rises smoothly with depth beneath a surface shell of 500 m sea water. Starting with $0 \cdot 004 \ \Omega^{-1} \ m^{-1}$ beneath this shell the conductivity reaches $0 \cdot 1$ at 500 km depth and unity at 900 km depth.

The second method, introduced by Tichonov and Cagniard, uses the surface impedance, i.e. the ratio of tangential electric to orthogonal magnetic field fluctuations, as observed at one site over a wide frequency range. Lateral conductivity variations are excluded and it is assumed that the primary field is of great lateral uniformity in comparison with its depth of penetration. The magnetotelluric method yields in this way estimates for the mean layered conductivity distribution on a regional scale. The analysis of pertinent observations at various places proved the existence of a high-resistivity zone between surface layers of great complexity and highly conducting matter in the upper mantle.

Returning to the first mentioned magnetic method we observe that the slow diurnal variations propagate with negligible attenuation through the upper mantle above 500 km and cannot yield more than an upper limit for the conductivity existing here. Detailed information about this depth range comes therefore mainly from fast variations around 1 cycle/hour with a reduced depth of penetration. Rikitake[3] was the first to attempt a World-wide analysis of bays and other short-period events. It became soon evident, however, that their internal part is subject to numerous local anomalies and that the upper mantle must be extremely non-uniform as far as its conductivity is concerned. In particular, standard magnetic observatories seemed to have the tendency to lie close to anomalous zones, which facilitated their detection[4].

Since then interest has focused on these *induction anomalies* of fast variations. They have been found at many places around the World, even though the depth of their origin and their significance for the upper-mantle structure are not always clear. The prominent coastal anomalies near large and deep oceans[5], for instance, coincide with an outstanding superficial conductivity contrast and can be interpreted—at least partially—as surface effect (see p. 129). The Rio Grande anomaly in the southwestern United States, on the other hand, is presumably of deep origin, since the overall surface conductivity is rather low here and without marked changes within the zone of anomalous variations[6].

At times it appeared as if such induction anomalies could be found everywhere, and clearly, when the anomalous becomes the norm, its significance for the unusual diminishes. But this is not so and there are large areas where the normal behaviour of the variations indicates a stratified internal conductivity structure. This applies for southern Arizona and New Mexico between the Colorado river and the Rio Grande anomaly, where we observe exceedingly small but uniform Z variations[6]. Another example is Bavaria, where a north–south profile from Upper Palatinate across the Bavarian 'Molasse' into the Alps failed to give indications for internal conductivity anomalies[7]. Hence, the well-known north German anomaly does not seem to have a counterpart in southern Germany.

Outline of the Data Reduction

Considering the internal conductivity σ at the level z beneath the Earth's surface we distinguish between its constant normal part $\bar{\sigma}$ and its variable anomalous part σ_a:

$$\sigma = \bar{\sigma} + \sigma_a \tag{2}$$

The transient magnetic field vector $F(t, P)$ is accordingly the sum of a normal plus anomalous part:

$$F(t, P) = \bar{F}(t, P) + F_a(t, P) \tag{3}$$

$\bar{F}(t, P)$ represents the smoothly varying external plus internal surface field above the averaged conductivity distribution $\bar{\sigma}(z)$, while the induction anomaly $F_a(t, P)$ is, by definition, of internal origin alone. We have to find the perturbation σ_a as a function of depth from an observed induction anomaly F_a on the basis of a presumed normal distribution $\bar{\sigma}(z)$.

The flow of eddy currents in a stratified substratum is parallel to its surface. Hence, the internal and external parts of $\bar{F}$ have matching distributions at the Earth's surface, provided that the depth of penetration of the normal variation field is small in comparison with its lateral non-uniformity. This is a justified assumption in the case of fast variations and in the absence of overhead current concentrations (jets). We obtain then the normal part of the tangential H (northward) and D (eastward) variations by smoothing the observed variation field within the surveyed area. Local deviations from this smoothed level are considered as anomalous parts of H and D.

The normal part of the observed vertical Z variations is given by the overall depth of the eddy currents in relation to the lateral non-uniformity of the primary field. Let their mean depth be represented by a perfect conductor at the frequency-dependent depth h. (This substitution accounts of course only for the in-phase component of the induced surface field.) Then

$$\bar{Z} = h(\bar{H}_x + \bar{D}_y) \tag{4}$$

where $\bar{H}_x$ denotes the northward gradient of $\bar{H}$ and $\bar{D}_y$ the eastward gradient of $\bar{D}$ ($\bar{H}, \bar{D}, \bar{Z}$ are the components of the normal variation vector $\bar{F}$).

The midlatitude bay field, for instance, has in H a relative northward gradient of 5% per hundred kilometres and a negligible eastward gradient in D. Thus, by setting

$h = 200$ km, we obtain $\bar{Z}/\bar{H} = 0 \cdot 1$ as normal ratio of vertical to horizontal variations. This reflects the well-known fact that the inducing and induced fields above a conductive substratum supplement each other in the tangential components, but oppose each other in the vertical component, yielding a nearly tangential transient surface field under normal conditions. Consequently, internal conductivity anomalies which disturb this sensitive balance between external and internal Z variations are more obvious in Z than in H and D, where the anomalous parts are superimposed upon substantial normal parts (cf. figure 2).

The second step of the data reduction is a statistical correlation analysis between the thus separated anomalous and normal parts, involving numerous magnetic disturbances of the same general type (e.g. bays) but of different form and intensity. This postulated correlation is necessarily linear, since the governing equations, Maxwell's field equations, establish linear relations between the electromagnetic field components and their time and space derivatives. We obtain as result for each survey station a 3×3 matrix of transfer functions connecting the components of $\bar{F}$ and F_a in the frequency domain. They describe the induction anomaly as a function of frequency and location in a statistically condensed form and provide the proper basis for the subsequent interpretation.

It remains to verify the truly internal origin of the anomalous surface field, normalized in this way. This can be done by applying appropriate separation methods to its spatial distribution for each resolved frequency component, thereby eliminating unwanted contaminations of external origin. Siebert and Kertz[8] proposed a convenient method for two-dimensional fields which Hartmann[9] and Weaver[10] extended to three-dimensional distributions. Price and Wilkins[11] used in their treatise on the S_q field a somewhat different, but also very suitable, separation technique. The separation involves in either case elaborate numerical calculations and it is carried out preferably as final step of the data reduction. We may presume that the statistical treatment of numerous events help to minimize random contributions of external origin to F_a.

Interpretation of Induction Anomalies

At the outset we have to estimate the possible effect of near-surface conductivity variations upon the internal part. Following Price[12] it is convenient and, for the frequencies around 1 cycle/hour which are considered here, also permissible to treat the outermost layers (oceans and geological strata on land) as a thin surface sheet of variable *total conductivity*:

$$\tau = \int_0^d \sigma(z)\,\mathrm{d}z$$

separated by an insulating intermediate zone from highly conductive matter further down. The integration is carried out from the outer to the inner face of the sheet, d denoting its thickness.

Let $\bar{\tau}$ be, in analogy to $\bar{\sigma}$, the averaged total conductivity of the surface layers in the surveyed area and h the depth of a substitute perfect conductor in reference to the mean depth of the deeply induced currents. Theoretical considerations show that

$$\eta_s = 4\pi\omega\bar{\tau}h$$

E   NATO II

controls as dimensionless *induction parameter* the relative strength of those eddy currents which are induced in the surface layers. Their contribution to the internal part is greater than the contribution of deeply induced currents when $\eta_s > 1$, and vice versa.

The inclusion of h accounts for the dampening effect of the inductive couple between superficial and deep eddy currents, assuming again that their depth is small when compared with the spatial wave length of the primary field.

With regard to bays we may insert $\omega = 2\pi$ cycles/hour and $h = 200$ km. This gives $\eta_s = 7$ for $\bar{\tau} = 16 \times 10^{-6}$ e.m.u. cm ($= 4$ km of sea water, $\sigma = 4\ \Omega^{-1}\ m^{-1}$) and $\eta_s = 0\cdot18$ for $\bar{\tau} = 4 \times 10^{-7}$ e.m.u. cm ($= 4$ km of rock formations, $\sigma = 0\cdot1\ \Omega^{-1}\ m^{-1}$). We see that bay disturbances penetrate with little attenuation by eddy currents through continental surface layers of the indicated conductivity but not through large and deep oceans, at least not for the value of h postulated here. This discrepancy would explain the anomalous behaviour of bays near coast lines, but it becomes obvious at the same time that prominent inland anomalies of bays could hardly arise from superficial conductivity contrasts alone. It may be added that the proper mean value $\bar{\tau}$ is not the arithmetic but the harmonic mean over a variable total conductivity.

Observations with a self-contained D variometer, lowered to the bottom of the Pacific Ocean offshore from California, revealed that the D amplitude of bays is reduced indeed to one-quarter of its surface value beneath 4 km of sea water[13]. This implies that about three-quarters of the internal part above the ocean comes from eddy currents induced in the ocean.

Corresponding observations beneath continental surface layers, say in deep bore holes, have not been made yet, but a preliminary estimate for their shielding effect upon geomagnetic variations can be derived from magnetotelluric measurements. Let E/H be the surface impedance and $\bar{\tau}$ the total conductivity of surface layers at a given site. The difference between the tangential variations above (H) and below (H^-) these layers is equal to the integrated sheet-current density $E\tau$, multiplied by 4π. Hence,

$$\frac{H^-}{H} = 1 - 4\pi\tau\frac{E}{H} \tag{7}$$

Wiese[14] reported, for instance, as impedance of bays $E_{EW}/H = 0\cdot14$ mv/km $\gamma = 1\cdot4 \times 10^4$ e.m.u. for the observatory Niemegk near Potsdam, situated above the highly conducting sediments of northern Germany. Inserting $\tau = 1\cdot6 \times 10^{-6}$ e.m.u. cm ($= 4$ km of sediments, $\sigma = 0\cdot4\ \Omega^{-1}\ m^{-1}$) yields $H^-/H = 0\cdot72$. Thus, only a small part of the internal bay field would be due to eddy currents in the sediments, indicating a deep-seated cause for the anomaly of bays in northern Germany.

Induction anomalies which are of truly deep origin can be interpreted on the basis of two basic models.

(*a*) We approximate the internal conductivity distribution by a stratified substratum with undulating interfaces between various layers of uniform but different conductivity.

(*b*) We use a stratified substratum with plane interfaces but assume that one or more layers are non-uniform as indicated in equation (2).

Both models merge at some distance from the anomaly into a preconceived normal distribution $\bar{\sigma}(z)$ and to be determined are either the undulations or the perturbations σ_a from an induction anomaly at the surface of the substratum. The greatest possible
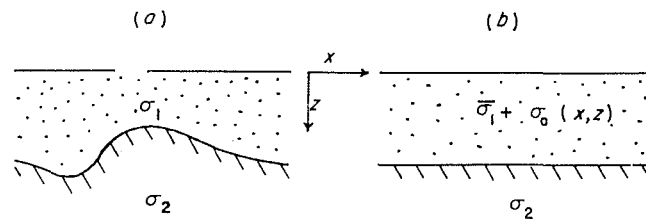
(a)   (b)

FIGURE 1   Two-layer models to illustrate the proposed types of internal conductivity anomalies

depth of these non-uniformities is given by the depth of penetration of the normal variation field which of course must reach the undulating interfaces or non-uniform layers, respectively.

Figure 1 shows these basic concepts of interpretation applied to a simple two-layer model for the Earth's interior. A straightforward treatment of the model 1(a) is possible in the limiting case that $\sigma_1 = 0$ and $\sigma_2 = \infty$. The boundary condition for a transient field in the upper non-conducting half-space requires that its magnetic vector is tangential to the surface of the underlying perfect conductor. Hence, this surface can be found by deriving the internal field-line pattern for an anomalous plus normal variation field which is given at the surface of the upper non-conducting layer and extended downwards in one way or the other. It is of course presumed that the anomalous and normal parts of the observed variations are roughly in phase.
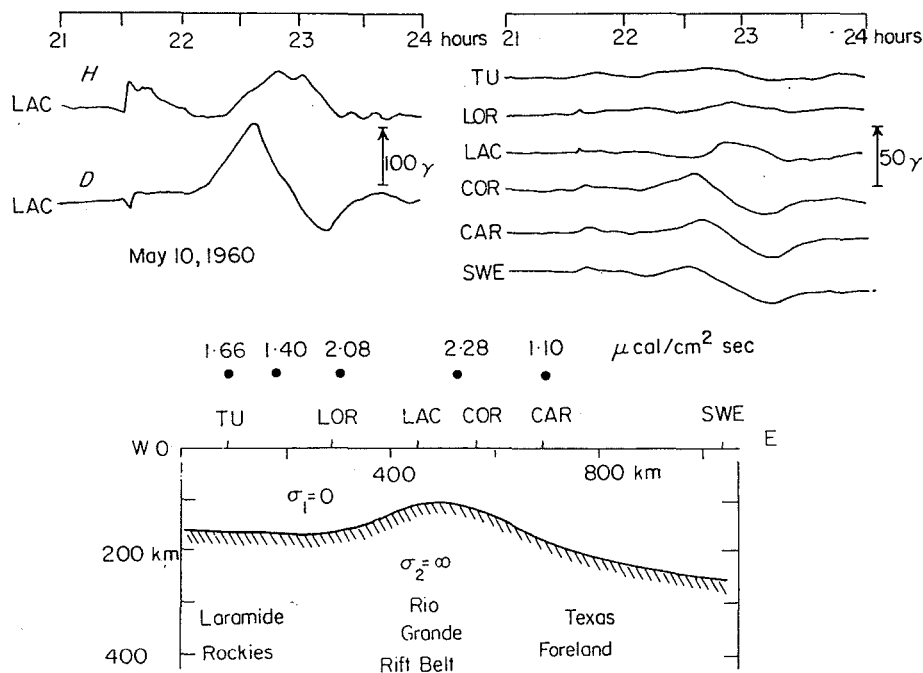

FIGURE 2   Rio Grande anomaly of fast variations in southern Arizona, southern New Mexico, and west Texas, shown for a typical bay. Subdued Z amplitudes west of the Rio Grande (Tucson, Lordsburg) suggest high mantle conductivities at shallow depth beneath the Laramide Rockies. Their conspicuous increase east of the Rio Grande (Cornudas, Carlsbad, Sweetwater) reflects in comparison, the low conductivity of the upper mantle beneath the Texas foreland. The Z reversal between Las Cruces and Cornudas can be explained by an additional rise of highly conductive matter under the Rio Grande rift belt. The horizontal variations of D and H are shown only for Las Cruces. Notice that the Z variations east of the Rio Grande have the same form as the D variations at Las Cruces, indicating a north–south trend of the conductivity structure shown below (see text). [Heat flow values from Warren[15] and Herrin and Clark[16]]

Each field line which does not intersect the Earth's surface is one possible interface between non-conducting and perfectly conducting matter. From the family of curves we choose that which merges at some distance from the anomaly into a postulated normal depth h of a substitute perfect conductor as introduced above. Each particular frequency component of the induction anomaly yields undulations around another normal depth h, indicating the varying depth of penetration of the incident field as a function of location and frequency.

The application of this field-line method to the Rio Grande anomaly of bays is shown in figure 2. In accordance with the width of the coastal anomaly in southern California we chose $h = 160$ km as the normal depth of the substitute perfect conductor under southern Arizona (Tucson) for 1 cycle/hour. Its surface then rises to $h = 100$ km under the Rio Grande rift belt and sinks to more than 200 km under the Texas foreland.

This contrast of low mantle conductivities under west Texas to high conductivities under southern Arizona and New Mexico affects even the internal part of the deeply penetrating diurnal variations as seen in figure 3. We observe that the centre of the northern S_q vortex passes during the equinoxes more or less overhead the east–west line of survey stations, which lie therefore in the range of maximum external Z variations. Hence, the observed diurnal Z amplitude is, as the sum of the external plus internal parts, a sensitive measure for the depth to the highly conducting part of the upper mantle.

It may be suggestive that the Rio Grande anomaly lies at the border of two structural provinces, the Laramide Rockies to the west and the Texas foreland to the east, the
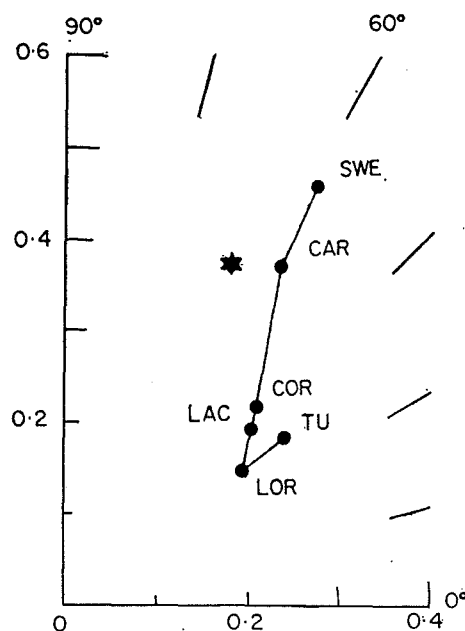

FIGURE 3   Rio Grande anomaly of the slow diurnal variations (cf. figure 2). The dots show the Z/Y ratio of the third time harmonic in polar coordinates, derived from hourly means in Z and Y ($=$ true east component) of four quiet days (April 19–22, 1960). The angle indicates a phase lead of Z relative to Y. The star shows the global average of the Z/Y ratio for the magnetic latitude of the survey stations (40° N), calculated from Chapman's equinoctial ratio of internal to external parts for the P_2^1 term (cf. Lahiri and Price[2], table 1). The reduced Z/Y ratio in southern Arizona and its gradual increase toward Texas conform with the anomalous behaviour of bays along the same profile, indicating a deep-seated cause of the anomaly

latter being a region of great tectonic stability and magmatic inactivity since Pre-Cambrian times. Furthermore, the postulated rise of highly conductive and probably hot mantle material under the Rio Grande valley coincides with a belt of intense vulcanism in recent times, high terrestrial heat flow[15-17], and unusual attenuation of seismic waves[18].

Conductivity Anomaly in the Andes of Peru and Bolivia

In 1957 the Carnegie Institution of Washington sent a seismic expedition to the Andes in South America[19]. One of its conspicuous results was the discovery of an extremely high attenuation of seismic waves which travel across the mountain range. It was suggested at that time that 'this attenuation may have some connection with the volcanic structure of the Andes', involving hot and perhaps even molten material at shallow depth.

Would this zone of high seismic attenuation appear as a zone of high conductivity in the internal part of geomagnetic variations? A field programme to test this hypothesis began in 1963 with a net of nine magnetic recording stations (Askania variographs). It has been in progress since that time as a joint venture of the Instituto Geofisico del Peru, the Instituto Geofisico Boliviano, and the Department of Terrestrial Magnetism (Carnegie Institution of Washington).

The first reconnaissance survey of 1963 revealed that the expected coastal anomaly of bays is not only missing in southern Peru but even reversed in its sign[20]. In other words, the superficial conductivity contrast between ocean and continent is more than compensated by an internal conductivity gradient in the opposite direction, bringing deep induction currents close to the surface beneath the Andes. After the second survey 1965-6 it became clear that this postulated zone of high internal conductivity ends near the eastern slope of the mountain range[21].

Let a few introductory remarks precede the detailed discussion of selected magnetograms. Peru and Bolivia are unusual countries in various aspects, but their geomagnetic distinction is the presence of the dip equator of the main field as shown in figure 4.
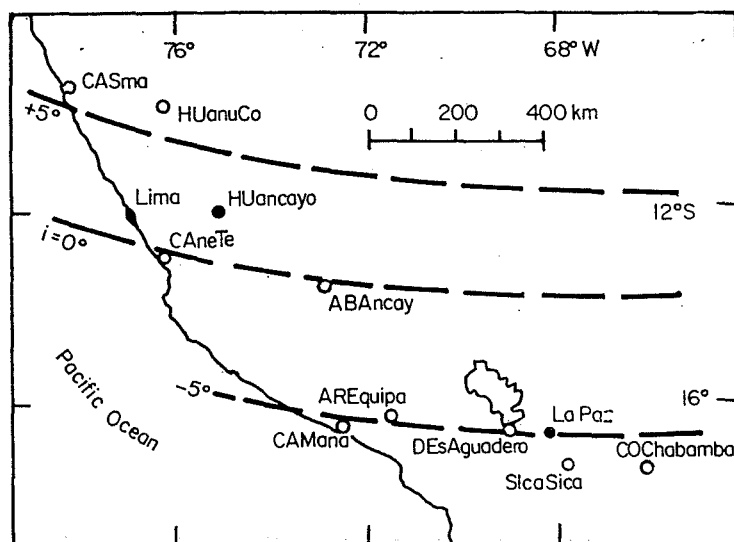

FIGURE 4 Magnetic stations during the 1965-6 survey in Peru and Bolivia, shown in relation to lines of equal dip *i*

This line of zero Z component exerts a pinching effect upon ionospheric currents on the day-lit side of the Earth, leading to an overhead current concentration which is known as *equatorial electrojet*. The surface field of this jet has a half-width of about 250 km. It may be visualized as the field of a line current, flowing 2–300 km above the line of zero dip.
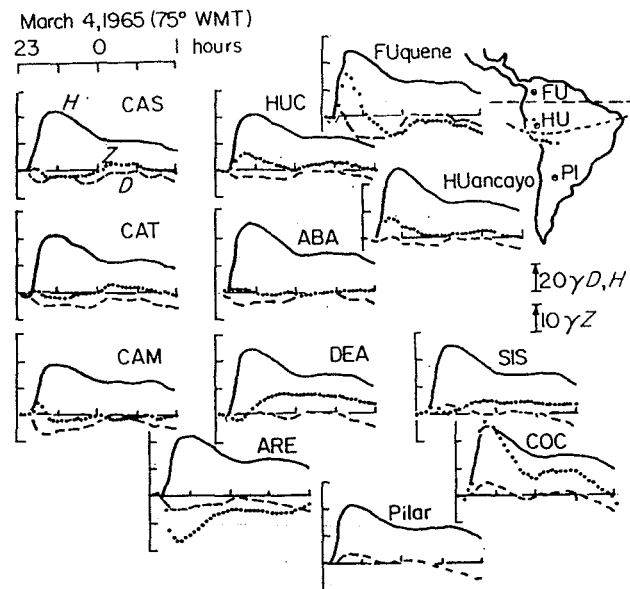

FIGURE 5   Equatorial night event (bay) as recorded at the stations of the 1965–6 survey (see figure 4) and the permanent observatories Fuquene (Columbia), Pilar (Argentina). Slight enhancement of H amplitudes at mountain stations relative to those at coastal stations and concurrent reversal of Z amplitudes (CAS–HUC, CAT–HU, ARE–COC). Both observations suggest that deep induction currents are brought close to the surface in a high conductivity channel beneath the Andes. [From Schmucker and others[21]]

The electrojet is absent during the night hours and the equatorial bay field, for example, is indeed of remarkable uniformity (figure 5). Thus, we have two distinct source fields at our disposal: the spatially smooth field of night events and the highly non-uniform jet field of day events.

Local differences of night events are undoubtedly of internal origin alone and due to subterranean conductivity anomalies. Local differences of day events, on the other hand, reflect not only the distorting effect of these internal anomalies σ_a but also the non-uniformity of the external plus internal jet field above a normal stratified distribution $\bar{\sigma}(z)$. In short, the combined anlaysis of day and night events gives us in equatorial regions the unusual opportunity to investigate concurrently the anomalous and normal conductivity distribution with observations in a limited area.

Beginning with a typical night event we infer from the traces of figure 5 that the H amplitude of the equatorial bay field hardly changes over 37° in latitude. Hence, we may expect minute Z amplitudes under normal conditions (cf. equation (4)). There is a slight anomalous increase of the maximum H deflection at mountain stations relative to those at the coast. We notice also some irregular differences in D, even though substantial changes of the compass deviation from station to station obscure their significance. Nevertheless, the overall D amplitude is small and the horizontal disturbance vector points northwards to the high-latitude centre of the ionospheric bay vortex.

A preliminary evaluation of numerous night events showed that their normal parts in X (= true north component) and Y (= true east component) are well represented by the horizontal variations at Arequipa, the capital of southern Peru half-way between the coast and the high Andes. We subtract the thus-defined normal part from the observed X and Y amplitudes and obtain for each survey station the respective components of the anomalous horizontal vector $\mathbf{B_a}$ as shown in figure 6.
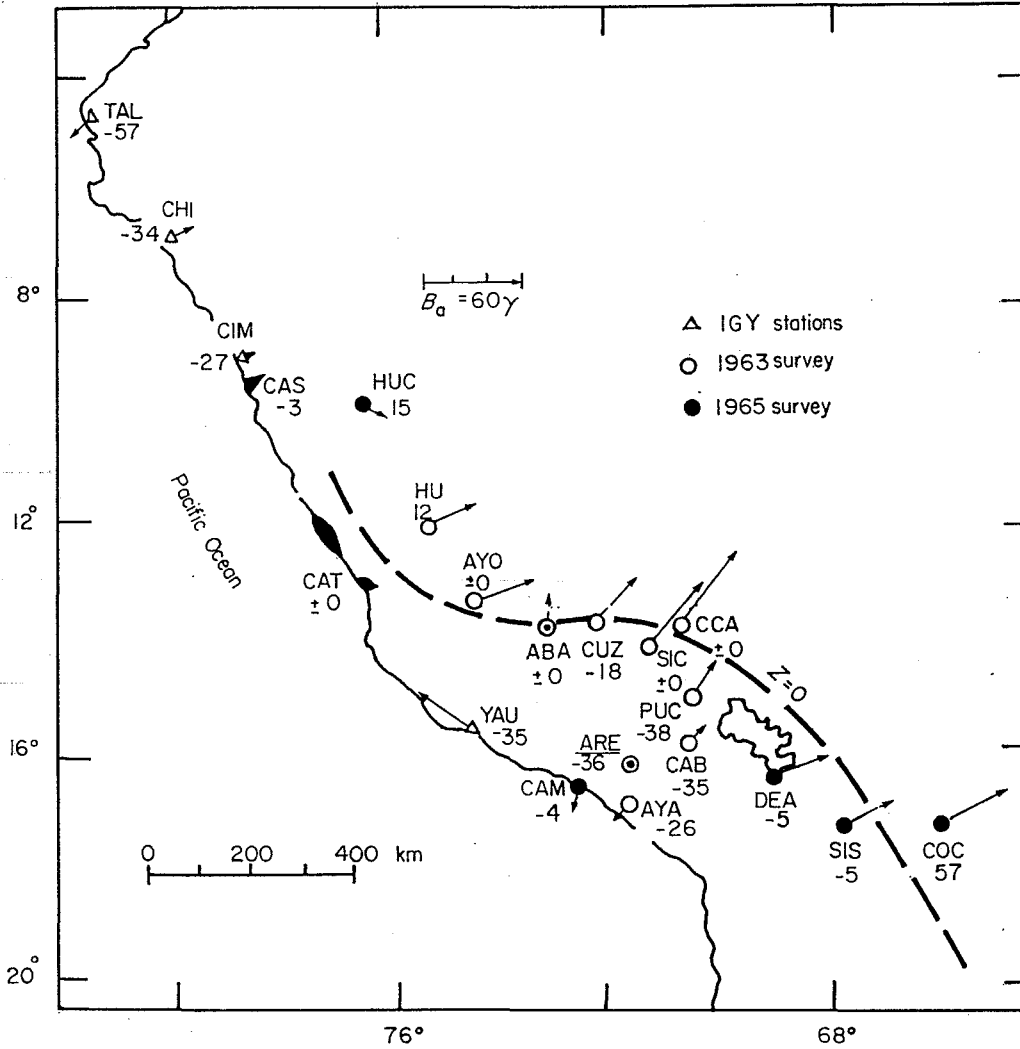

FIGURE 6 Andean anomaly of bays, normalized with the true north amplitude X of Arequipa and averaged over two events for each station. The arrows show the anomalous part B_a of the horizontal disturbance vector:

$$B_a = (X - X_{ARE})i + (Y - Y_{ARE})j$$
$$X_{ARE} = 100 \, \gamma$$

The numbers give the Z amplitude. Arrows of maximum length are found in the high mountains where they are more or less perpendicular to a line of zero Z amplitude. This line indicates the trend of the postulated high-conductivity zone in the Andes (see text). [From Schmucker and others[21]]

These vectors indicate, when rotated anticlockwise by 90°, strength and direction of the anomalous internal current field which is superimposed upon the normal east–west flow of subterranean eddy currents. We see that these currents are channelled into a high-conductivity zone which follows the general trend of the mountain range. The resultant current concentration beneath the Andes explains at the same time the

reversed anomalous Z amplitudes along the eastern and western slope which is so strikingly demonstrated by the opposite Z deflections at Arequipa and Cochabamba.

Turning now to day-time fluctuations of comparable frequency we observe that their jet field should be uniform along lines of equal dip and without variations in D when internal anomalies are absent. (The horizontal force of the main field is nearly perpendicular to the dip equator in Peru.) Hence, induction anomalies of the jet field are characterized by different day-time variations at stations of the same dip and by D variations in general, provided of course that the trend of the anomaly is not parallel to the dip equator.

The equatorial day-time fluctuations in Peru and Bolivia show indeed these criterions for internal conductivity anomalies (figure 7). The rugged Z trace of Cochabamba stands in sharp contrast with the small Z amplitudes of Sicasica, Desaguadero, and Arequipa (not shown), even though these stations lie more or less on the same isocline, namely on the southern isocline of maximum Z amplitude of the normal jet field. We conclude that Cochabamba is located above the edge of an extremely shallow concentration of internal eddy currents to the south, while the other stations are on top of it. No explanation can be offered yet for the anomalous behaviour of Z at the coastal station Camanà. Clearly visible are also anomalous D variations along the southern isocline, indicating a northward deflection of the internal jet current by the high-conductivity zone under the Andes.

The depth of this zone has to be small in comparison to the half-width of the jet field, i.e. of the order of 50 km or less. Otherwise, this narrow-spaced source field would not reach the anomalous zone at all. The skin-depth value for 1 cycle/hour of
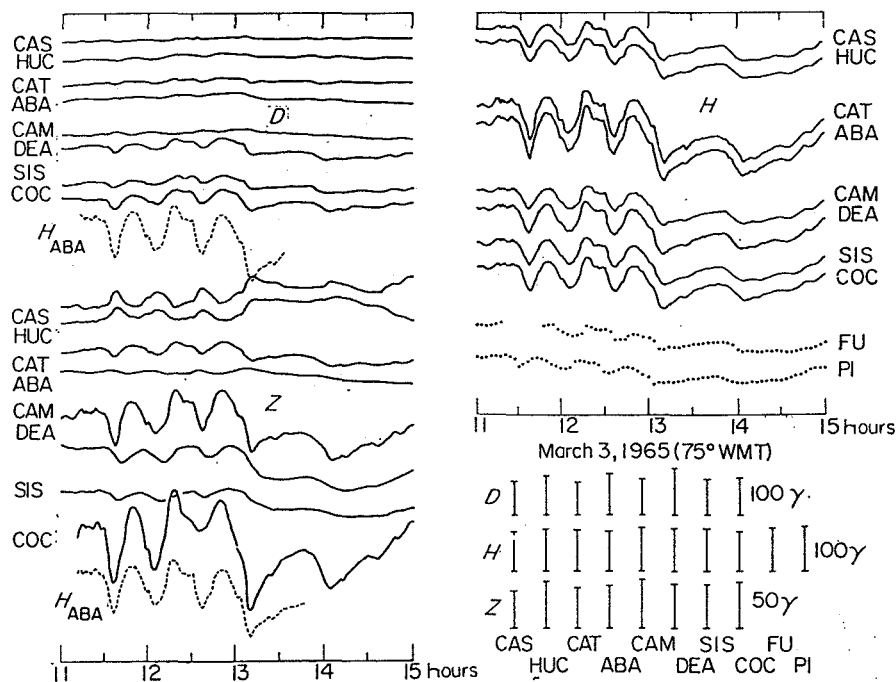

FIGURE 7 Equatorial day-time fluctuations as recorded during the 1965–6 survey (see figure 4). Maximum H amplitudes near the dip equator (CAT, ABA) reflect the electrojet effect upon the ionospheric source field. Subdued Z amplitudes at the northern stations CAS and HUC indicate that the Z component of the external jet field is nearly compensated by the field of subterranean eddy currents near 100 km depth. The large Z amplitude at COC, CAM and the D fluctuations at DEA, SIS, COC are anomalous and can be related to high internal conductivities under the Andes. [From Schmucker and others[21]]

the material existing here must be also small as against the half-width in order to permit a significant attentuation of incident day-time fluctuations by eddy currents. Hence, in virtue of equation (1a) we have to postulate a conductivity of at least $0.1\,\Omega^{-1}\,m^{-1}$ in contrast with much lower values under the adjacent Brazilian shield and the offshore Peruvian trench.

A careful intercomparison of the slow diurnal variations at the same survey stations revealed that their relatively small internal parts are not affected by the Andean anomaly. Hence, the jet field penetrates for $f = 1/12$ cycle/hour through the high-conductivity zone with negligible attenuation, which establishes $0.1\,\Omega^{-1}\,m^{-1}$ also as an upper permissible conductivity in this zone.

We recall from p. 126 that conductivities of this magnitude are expected only deep within the mantle under normal conditions. Their presence at shallow depth beneath the Andes (0–50 km) is clear evidence for the unusual thermal state or composition of the existing mantle material, reflecting perhaps the remarkably intense tectonic and magmatic history of this mountain range.

A complete review of this type of work can be found in the book by Rikitake[22].

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