The Geomagnetic Coast Effect

W. D. PARKINSON

Department of Geology, University of Tasmania
Hobart 7005, Tasmania, Australia

F. W. JONES

Department of Physics, Institute of Earth and Planetary Physics, University of Alberta
Edmonton, Alberta, Canada T6G 2J1

Near most coastlines the vertical component of the geomagnetic variation field is abnormally large and correlates positively with the inland horizontal component. This phenomenon is known as the geomagnetic coast effect. The few coastal locations at which it is absent are tectonically anomalous. The ratio of vertical to horizontal components decreases inland at a rate which depends on the geological nature of the continent. The effect varies only slightly with period, reaching a broad maximum between periods of 30 and 90 min. A number of models have been investigated by either calculation or analogue modeling, but the complete problem dealing with finitely conducting irregular oceans overlying a less conducting lithosphere has not yet been solved. Eddy currents flowing entirely in the seawater would be induced only by the vertical component of the magnetic field. Observations, combined with analogue model experiments, suggest that induction by horizontal components is also important. This requires flow of current in the lithosphere beneath the oceans.

CONTENTS

1. INTRODUCTION

Generally speaking, the geomagnetic field is unaffected by the distribution of continents and oceans. This is true of the main field, which originates almost certainly in the outer liquid core. It is also true of the time-varying fields of external origin, the sources of which are currents flowing in the ionosphere or magnetosphere. However, there are three ways in which the continent-ocean interface can influence the geomagnetic field; first, there is a systematic difference in the magnetic history of the rocks beneath oceans and continents which contribute to the local or crustal field. The result of this is the well-known difference between magnetic anomaly patterns on oceans and continents. Second, ocean currents moving in the main field generate a field by dynamo action which is not negligible. This effect has been the subject of intensive research by a small number of investigators [*Hill and Mason,* 1962; *Larsen and Cox,* 1966; *Larsen,* 1973; *Osgood et al.,* 1970] and is most clearly seen in the enhanced lunar daily variation at coastal locations. Third, the electrical conductivity of both seawater and material under the oceans differs considerably from that of the continents. This is the effect usually referred to as the 'geomagnetic coast effect' (GCE), which will be the subject of this review. In particular, we shall try to evaluate the weight of observations and opinion on the question of the relative importance of electric currents flowing in seawater against that of currents in the deeper parts of the earth. The strong correlation between the vertical and eastward variation fields at Watheroo (Western Australia) was first noticed by *Baird* [1927]. That this is an example of a more universal phenomenon associated with coastlines was first realized by *Parkinson* [1959].

The earliest investigators of electromagnetic induction in the earth [*Schuster,* 1889; *Chapman,* 1919; *Chapman and Whitehead,* 1922] used the model of a spherical conductor with uniform conductivity as studied by *Lamb* [1889] and considered periodic fields. The extension to aperiodic fields was made by *Price* [1930, 1932] and applied by *Chapman and Price* [1930] to the study of magnetic storms. The conductivity of the earth as a function of its radius was considered in a classic paper by *Lahiri and Price* [1939]. Since that time a number of workers have obtained conductivity-depth profiles for the earth. A short discussion of such global studies is given by *Price* [1973], and a review of methods and results for global geomagnetic sounding is given by *Bailey* [1973]. A feature of all of these profiles is a rapid rise in conductivity with increasing depth somewhere between 400 and 800 km. *Kertz* [1963] called the region below this the 'conductosphere', and that

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useful term will be used in this review. Although on a global scale the conductivity may be treated as a function of radius only, on a more local scale, near and on the surface of the earth, great variations in conductivity may occur over relatively short distances. Such a variation of conductivity is encountered between oceans and continents. Anomalies in conductivity not associated with coastlines may also occur. Correlations between the components of the variation field and spatial changes, particularly in the vertical component, have been known for some time from the work of *Bartels* [1954], *Wiese* [1962], *Schmucker* [1959], and others in Europe. A number of reviews [*Schmucker,* 1970*b*; *Gough,* 1973*b*; *Hutton,* 1976] of various aspects of this subject have appeared. A review of theoretical models for electromagnetic induction in the ocean has been given by *Ashour* [1973], and a review of techniques and instrumentation for the study of electromagnetic induction at sea by *Filloux* [1973]. Also, *Bullard and Parker* [1970], *Vacquier* [1972], *Cox et al.* [1970], *Hobbs* [1975], and *Larsen* [1973] have discussed the coast effect as well as other phenomena.

2. THE COAST EFFECT

The geomagnetic field, as recorded at a fixed location, is a complicated function of time. In the period spectrum between a few minutes and a few days, several phenomena occur. The most persistent is the diurnal variation, which produces a sharp line spectrum of 24-hour period, and its low-order harmonics. A more irregular phenomenon is the occurrence of magnetic storms. These are more or less isolated events that have two main features. The first is a large decrease in the magnetic north component of a few hours duration, followed by a recovery to the prestorm value in a few days. Simultaneous with the decrease of the magnetic north component is the second feature of the storm, which consists of erratic fluctuations with a wide period spectrum. Finally, there are isolated excursions from the normal value which typically last from 10 min to 2 hours. These have often been called 'bays,' but now are more often referred to as 'substorms' [*Rostoker,* 1972]. The coast effect is more noticeable in substorms and the erratic part of magnetic storms, though the other variations are influenced by coastlines as well. Even the storm field *Dst* gives rise to an appreciable coast effect [*Fainberg,* 1978]. The time variations of the magnetic field are due to electric currents flowing in the ionosphere or magnetosphere. The magnetic field produced by such currents is referred to as the 'primary field.' These varying primary fields generate electric fields, and since the earth has a nonzero conductivity, eddy currents flow in response to such electric fields. The eddy currents in turn generate a magnetic field which is called the 'secondary field.' Near its surface the earth has a complicated spatial distribution of conductivity which makes this secondary field a complicated function of position. In general, for an observation far inland, and far from concentrations of primary current in the ionosphere, the vertical component of the variation field is much smaller than the horizontal components. That is, the vectors representing ΔB, the variation field, tend to lie close to a horizontal plane. This is not principally a feature of the primary field but occurs because the upper surface of the conductosphere is nearly horizontal. On approaching a coastline the variation vectors still tend to lie close to a plane, but this plane is no longer horizontal and tilts upward toward the deep ocean. This is the coast effect. The plane in which the variation vectors lie is called the 'preferred plane'. Its equation can be written as

$$z = ax + by \tag{1}$$

where the Cartesian coordinates x, y, and z are in the northward, eastward, and downward directions. It should be emphasized that this is simply a plane of regression of ΔZ on ΔX and ΔY. A plane should not be defined unless the regression is significant. The orientation of the plane (1) can be shown on a map by the two-dimensional vector (a, b). This practice was started by *Wiese* [1962]. The vector points away from the region with higher conductivity. Independently, *Parkinson* [1962*a*] introduced a vector that is the horizontal projection of the downward normal of the preferred plane. It is of length

$$(a^2 + b^2)^{1/2}(1 + a^2 + b^2)^{-1/2}$$

and points toward the better conductor. The relation between these two vectors was pointed out by *Untiedt* [1964]. It should be emphasized that the 'vectors' are not vectors in the mathematical sense. They are lines on a plane representing directions in three dimensions. They cannot be added to give a meaningful resultant. 'Induction arrows' is a more satisfactory name. The parameters a and b are generally dependent on frequency, and it is therefore more convenient to consider Fourier transforms of the north, east, and vertical components of the variation field. If these are written X, Y, Z, then by analogy with the equation for the preferred plane we can write

$$Z = AX + BY \tag{2}$$

where, in general, all quantities are complex. It is usual to consider only a limited range of frequencies, which is equivalent to filtering the time functions. The relation between the filtered time function and the transform can be expressed by equations such as

$$Z(t) = Z_r \cos \omega t - Z_i \sin \omega t$$

$$Z(\omega) = Z_r + iZ_i$$

and A and B are transfer functions between the vertical and horizontal components of the variation field. They are normally calculated by the least squares method introduced by *Everett and Hyndman* [1967*a*]. A and B are chosen to minimize

$$\sum |Z - AX - BY|^2$$

summed over an ensemble of Fourier transforms. This differs from the method of deriving preferred planes used by *Parkinson* [1959] in that in the Everett and Hyndman method the larger-amplitude events are given greater weight.

There are some advantages in using time functions, as was pointed out by *Midha et al.* [1978].

The transfer functions are usually displayed by plotting on a map the two-dimensional vectors (A_r, B_r) and (A_i, B_i), corresponding to the real and imaginary parts of the transfer function. To maintain the custom of plotting arrows that point toward the better conductor, the sign is changed. Some investigators change the sign of the real part only; others change the sign of the imaginary part as well (compare the work of *Cochrane and Hyndman* [1974] with that of *Hyndman and Cochrane* [1971]). The parameter (A, B) is known in magnetotelluric parlance as a 'tipper.'

The concept of a preferred plane is of little use when the transfer function is complex. In fact, it can be shown that the variation field cannot be confined to a plane unless both A and B are real. In this case, $Z(t)$ is in phase with the horizontal component in the direction of the vector (A, B).

The above method infers a lack of symmetry between X, Y, and Z. If transfer functions A and B were derived by minimizing, for instance,

$$\sum |X - (1/A)Z - (B/A)Y|^2$$

quite different values would be obtained unless the residuals in both cases were small. In spite of this the temptation to consider X and Y as causes and Z as the effect should be resisted [*Bailey et al.*, 1977; *Honkura et al.*, 1977]. More generally, the relation between the primary and secondary fields can be represented by a tensor which becomes degenerate if the field is exactly confined to a plane [*Lilley and Bennett*, 1973]. *Schmucker* [1970*a*] described the relationship between the components by coefficients related to the tensor elements. In his approach he assumed that a linear relationship exists between the Fourier transforms of the anomalous internal field components and the normal field components for a single frequency, and this is expressed as

$$\begin{vmatrix} X_a \\ Y_a \\ Z_a \end{vmatrix} = \begin{vmatrix} h_x & h_y & h_z \\ d_x & d_y & d_z \\ z_x & z_y & z_z \end{vmatrix} \begin{vmatrix} X_n \\ Y_n \\ Z_n \end{vmatrix} + \begin{vmatrix} \Delta_x \\ \Delta_y \\ \Delta_z \end{vmatrix}$$

The quantities X_a, Y_a, and Z_a are the Fourier transforms of the anomalous parts of the x, y, and z components of the magnetic field, respectively; X_n, Y_n, and Z_n are the Fourier transforms of the normal parts of the x, y, and z magnetic field components, respectively; and Δ_x, Δ_y, and Δ_z are the Fourier transforms of the uncorrelated portions of the x, y, and z components. The transfer functions h, d, and z are in general complex, and their real parts are referred to as the in-phase transfer functions, while their imaginary parts are the quadrature-phase transfer functions. Using the cross-correlation and autocorrelation power spectra of the anomalous and normal field components along with the condition that the cross-correlation power spectra between the residual and normal field components be zero, Schmucker obtained a linear matrix involving the transfer functions and power spectra. For the z transfer functions,

$$z_x S_{x_n x_n} + z_y S_{y_n x_n} + z_z S_{z_n x_n} = S_{z_a x_n}$$
$$z_x S_{x_n y_n} + z_y S_{y_n y_n} + z_z S_{z_n y_n} = S_{z_a z_n}$$
$$z_x S_{x_n z_n} + z_y S_{y_n z_n} + z_z S_{z_n z_n} = S_{z_a z_n}$$

where $S_{x_n x_n}$ is the autocorrelation power spectrum of X_n and $S_{z_a x_n}$ is the cross-correlation power spectrum between Z_a and X_n, etc. Similar matrices can be obtained for the h and d transfer functions. *Schmucker* [1970*a*] represented the transfer function as a set of arrows. If we let $\mathbf{i}$ and $\mathbf{j}$ be Cartesian unit vectors in the x and y directions, respectively, the induction arrows are defined for a particular frequency as

In-phase arrow

$$\mathbf{C}_{\text{real}} = -z_{x_{\text{real}}}\mathbf{i} - z_{y_{\text{real}}}\mathbf{j}$$

Quadrature-phase arrow

$$\mathbf{C}_{\text{imaginary}} = z_{x_{\text{imaginary}}}\mathbf{i} + z_{y_{\text{imaginary}}}\mathbf{j}$$

If the vertical component is considered to be purely anomalous and the horizontal components purely normal, i.e.,

$$z_n = x_a = y_a = 0$$

then this transfer function is identical with the one defined by *Everett and Hyndman* [1967*a*]. Written in this way, the in-phase arrow is in accordance with the orientation of the *Parkinson* [1962*a*] arrow. Schmucker also defined a second set of arrows, called perturbation arrows, but we will not be concerned with these here.

3. Summary of Observations

During the last decade, geomagnetic deep sounding has become a standard method of probing the earth. Therefore there is a large amount of data available that report the ratio of vertical to horizontal components of the variation field. Summaries and discussion of such work are given by *Lilley* [1975], *Adam* [1976], and *Gough* [1974]. Much of the work is designed to search for mantle anomalies not connected with the coast effect. Indeed, many surveys seem to be designed to avoid coastal regions because of this effect (a notable exception is the array work of *Bennett and Lilley* [1973]). In the present review we draw attention in general only to those results that have some bearing on the coast effect and particularly its cause.

a. *Australasia*

The continental shelf is moderately narrow around the southern half of Australia, if Bass Strait is considered part of the continent. The GCE seems to be normal in the southwest and southeast [*Parkinson*, 1964; *Bennett and Lilley*, 1971; *Lilley and Bennett*, 1972]. The Otway anomaly, on the SW coast of Victoria, interferes with the normal GCE only locally. The shallow water of Bass Strait seems to have little effect on the direction of the variation field [*Lilley and Bennett*, 1972]. In general, the transfer functions A and B in (2) are almost real both in the SW and in the SE [*Everett and Hyndman* 1967*a*; *Bennett and Lilley*, 1972]. There is some evidence of inland current flow near the Darling fault, but the coast effect dominates.

An important feature noticed by *Everett and Hyndman* [1967*a*] is that although the amplitude of the transfer function decreases inland from both the west coast and the south coast, it decreases more slowly than in other environments, for instance, California and eastern Australia.

A 300-km-long profile extending from the edge of the continental shelf in South Australia [*White and Polatajko*, 1978] shows a normal and very consistent coast effect. Real induction arrows point consistently SSW. The inland decrease of the transfer function is very similar to that found in New South Wales by *Bennett and Lilley* [1971].

In the northern half of the continent the continental shelf is wider and joins Australia to Indonesia and New Guinea, isolating the deep Banda Sea. Results from Kuyper (near Jakarta) and Hollandia indicate a normal coast effect [*Parkinson*, 1964]. However, the effects reported from Port Moresby, Thursday Island, and Darwin suggest a greater flow of underground current than can be accommodated by the narrow and shallow water between Australia and New Guinea [*Parkinson*, 1963]. Results from Dili (eastern Timor) [*Chamalaun and White*, 1975] indicate a good conductor to the north, which is unlikely to be the isolated Banda Sea.

Results from northern Australia do not contradict the normal form of the GCE but indicate a more complex conductivity pattern, which is not surprising considering that the region is a plate boundary.

On the east coast of the North Island of New Zealand the coast effect is small and largely marked by the influence of a conductivity anomaly whose relation to the geothermal zone is obscure [*Midha et al.*, 1978].

The only other measurements from New Zealand known to the authors are those for Amberley Observatory in the South Island [*Lawrie*, 1965; *Parkinson*, 1964], which are normal for an east coast site.

b. *California*

California with its hinterland and nearby ocean constitutes one of the most thoroughly studied areas of the world. *Schmucker* [1970*a*] has described variation measurements extending along a traverse from southern California to Texas and a net covering most of California south of 40°N. *White* [1973*a, b*] has extended these measurements southward to central Baja California. Some of Schmucker's observations were made on islands on the continental shelf. As well as these, ocean floor measurements of both electric and magnetic fields have been made [*Cox et al.*, 1970; *Greenhouse*, 1972]. The continental shelf varies in width from 250 km in the south to less than 100 km in the north (near central California).

There is a typical GCE along the coast of California, with a somewhat greater transfer function in the north than the south. ΔZ decreases inland more rapidly than in western Australia, as is shown in Figure 1. Induction arrows throughout California tend to point WSW. Along the coast the real part of the transfer function does not vary much with frequency between 15- and 60-min period. The imaginary part is generally $\frac{1}{2}$ to $\frac{1}{4}$ of the real part and is more erratic and period sensitive.

The conductive San Joaquin valley (about 250 km inland) has a decided effect on induction arrows, especially at short periods.

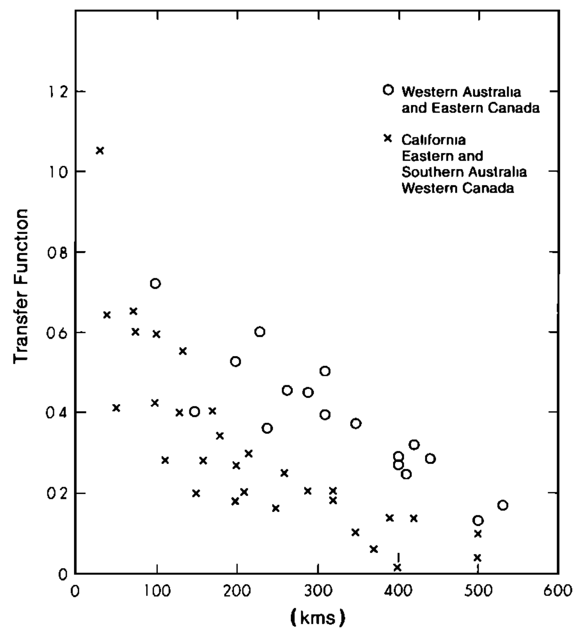

Fig. 1. Real transfer functions as a function of distance from the continental slope. Data are adapted from *Cochrane and Hyndman* [1970], *Everett and Hyndman* [1967*a*], *Hyndman and Cochrane* [1971], and *White and Polatajko* [1978].

The seabed stations of *Greenhouse* [1972] continue to show a coast effect relative to the continental slope, with almost real transfer functions.

The GCE continues along the Pacific coast of Baja California to 27°N although with diminished amplitude and a relatively stronger imaginary component of the transfer function. However, the region is dominated by a strong current concentration along the line of the Gulf of California [*White*, 1973*a, b*]. This line crosses Schmucker's southern traverse of stations between southern California and Texas with only a slight increase of ΔZ at Yuma. It is hard to escape the conclusion that the Gulf of California or some horizontal conductor below it is connected vertically to a highly conducting region.

c. *Western Canada*

The coast effect is normal at Tofino, on the Pacific coast of Vancouver Island [*Cochrane and Hyndman*, 1970]. However, the pattern inland is very complicated, partly because of a high-conductivity region under the Rocky Mountains [*Caner*, 1970]. At Tofino the imaginary part of the transfer function is small at a 30-min period but almost equal to the real part at 60 min. The rate of decrease inland seems to be at least as sudden as in California but is complicated by the nonuniform conductors inland.

d. *Northern Canada*

Two anomalous regions have been reported, one near the observatory at Mould Bay [*DeLaurier et al.*, 1974] and one near Alert [*Niblett et al.*, 1974]. The feature of Mould Bay itself and nearby sites on the same latitude is an attenuation of vertical field variations to 10% or less of the corresponding horizontal variations for all periods less than about 60 min. A site only 100 km north and about 150 km from the continental slope has a more or less normal coast effect. However, there is a complicated pattern at sites on the edge of the shallow water south and east of Mould Bay. It is difficult to assess what is the influence of the morphology of the primary field in this region. At a period of 160 min, all induction arrows in the neighborhood of Mould Bay point to the west, i.e., toward the nearest deep water. The Mould Bay anomaly is therefore a short-period phenomenon.

The Alert anomaly appears to have nothing to do with the ocean. There is a band of high conductivity running from the Lincoln Sea across Ellesmere Island probably at a midcrustal depth [*Praus et al.*, 1971].

e. *Eastern Canada and United States*

Induction studies in Newfoundland and the Maritime Provinces of Canada have been reported by *S. P. Srivastava and A. White* [1971], *Hyndman and Cochrane* [1971], *Cochrane and Hyndman* [1974], and *Bailey et al.* [1974]. Typical coast effects are found at Saint John's, on the eastern tip of Newfoundland, and at Dartmouth, on the coast of Nova Scotia, the imaginary part being about half the real part. In spite of the complication of current channeling in the Gulf of Saint Lawrence, the inland decrease is similar to that of the shield area of Western Australia (see Figure 1). The coast effect seems to persist almost halfway across Newfoundland to Deer Lake. Later work suggests that this may be affected by an inland anomaly [*Cochrane and Wright*, 1977]. Large short-period Z variations are found at Cape North, on the northern tip of Cape Breton Island, correlating with north-south variations, but with a comparatively large out of phase component.

This is probably due to channeling in Cabot Strait. Similar results have been obtained from Flowers Cove on the Strait of Belle Isle [*Cochrane and Wright*, 1977].

Perhaps the most significant results from this area are those of Sable Island, almost 200 km from the coast but near the edge of the continental shelf. The vertical component there is less than 10% of the horizontal component for all periods. Hence if the coast effects observed at Saint John's and Dartmouth were caused by seawater conductivity, they would have to be due to shallow water, which seems unlikely. A more likely cause is an underground high-conductivity region under the continental shelf. *Cochrane and Hyndman* [1974] suggest a conductive layer in the lower crust. No high heat flow has been found associated with the conductor. The coast effects at Saint John's and Dartmouth do not extend to diurnal variation periods [*S. P. Srivastava*, 1971].

Edwards and Greenhouse [1975] report a small coast effect on the coast of Virginia about 350 km from the continental slope. Again, a conducting lower crust is suggested, with the conductivity increasing to the west.

f. South America

Normal coast effects have been reported at Paramaribo, Surinam [*Parkinson*, 1962*a*], at Vassouras, Brazil [*Maeda et al.*, 1965], and on the coast of Chile [*Schmucker*, 1969; *Aldrich*, 1972]. Perhaps the most important South American results are those from Peru [*Schmucker*, 1969; *Aldrich*, 1972]. The coast effect is completely absent, or even reversed, over a long stretch of coastline from central to southern Peru. Instead, a similar effect is found inland in the region of the Andes. A two-dimensional model involving a perfect conductosphere of varying depth fits the field observed at substorm frequencies [*Greenhouse et al.*, 1973]. However, the fact that *Sq* is not affected limits the conductivity of the good conductor.

g. British Isles

A normal coast effect was found at Valentia, Ireland [*Parkinson*, 1962*a*]. An extensive network of stations throughout Ireland, Wales, England, and southern Scotland [*Edwards et al.* 1971] shows that this is confined to the western part of Ireland. The amplitude of vertical variations decreases eastward and is small in central Ireland. At longer periods (144 min) the decrease of ΔZ eastward is more gradual. A well-marked concentration of current crosses southern Scotland near Eskdalemuir. A similar effect was deduced for *Sq* by *Riddihough* [1967]. Except at sites on the west coast of Ireland, imaginary transfer functions are as large as real transfer functions. The shallow water around England and Wales does not seem to cause any obvious coast effect. At Sherborne (Dorset) there is a strong positive $Z:H$ correlation, but this appears to be a purely local effect. Similar inland anomalies in Devon are suggested by *Davey and Rosser* [1971]. Apart from this, and the Eskdalemuir anomaly, induction arrows in Great Britain are small. They are interpreted by *Edwards et al.* [1971] as being due to current flow in the Irish Sea, in the sea south of Ireland, in the North Sea, and across Scotland near Eskdalemuir.

The general absence of coast effects is confirmed by magnetotelluric work in the Irish Sea [*Jain and Wilson*, 1967], where a rather low conductivity was found. *Osemeikhian and Everett* [1968] found no definite coast effect at Malin Head (on the north coast of Ireland).

h. Mediterranean

A significant observation is that there is no coast effect on the shore of the Mediterranean Sea just north of the Pyrenees [*Babour et al.*, 1976]. This suggests that the Mediterranean is not large enough to support extensive eddy currents. However, channeling of current through the Strait of Bonifacio is indicated by a strong northerly induction arrow at Maddalena and (by indirect deduction) a southward arrow at Pertuzato [*Giorgi and Yokoyama*, 1967]. This is a short-period phenomenon for which the amplitude at 40-min period is only half of that at 10 min (see Figure 4*c*). *LeBorgne and LeMouel* [1975] have found evidence of channeling near the Strait of Gibraltar, but they believe that the current flow is through northern Morocco rather than through the strait itself.

There is a strong positive correlation between Z and H at Ponza Island (near Naples) and nearby sites on the SW coast of Italy [*Simeon and Sposito*, 1963]. This is also a short-period phenomenon and is almost certainly a local anomaly. No coast effect is found near Genoa.

An interesting phenomenon has been reported by *Babour and Mosnier* [1973]. The horizontal components of the variation field were measured at two sites 50 km apart near the coast of Brittany. The difference was found to be very sharply polarized and confined to a direction perpendicular to the coast. They conclude that the seawater cannot be the source of the effect and ascribe it to a comparatively highly conducting schist formation. Similar results were observed near Brest and near the east end of the Pyrenees, but it is not yet known how universal the effect is. Polarization of the direction of the horizontal component itself occurs at Alert [*Niblett et al.*, 1974] and, of course, in the vicinity of an electrojet.

i. Northern Europe

No coastal effect can be seen in the extensive results quoted by *Untiedt* [1970]. In fact, on Rügen Island, east of Jutland, the correlations are in the opposite sense [*Wiese*, 1962]. Small variations in the total intensity have been interpreted by *Fontaine et al.* [1965] as channeling in the English Channel.

j. Asia

There seems to be some coast effect on both the east and the west coast of India [*B. J. Srivastava et al.*, 1974], but because these reports are based on daily ranges, they are difficult to evaluate. More recent work by *Nityanander et al.* [1977] indicates normal coast effects at Alibag and Trivandrum, but Annamalainagar seems to be influenced by current channeling to the south.

Japan is probably the most intensively studied area in the world. Most of the work is summarized by *Rikitake* [1966, 1971]. The directions of the variation field are approximately those appropriate to a normal coast effect, but the induction arrows tend somewhat more to the south in central Honshu and more to the west in Kyushu [*Rikitake*, 1965]. The decrease inland is rapid, in conformity with the tectonic nature of the region.

The effect of the Japanese anomaly can be clearly seen at Miyake-Jima Island [*Honkura*, 1971], about 100 km from Honshu, where the transfer function increases with period up to 120 min. It is still observable, but with reduced amplitude, at Hachijo-Jima, almost twice as far from Honshu [*Honkura*, 1974]. This is in sharp contrast to the behavior on Sable Island, off the coast of eastern Canada. The effect on *Sq* is mainly a change of phase of Z [*Rikitake*, 1959].

A normal but rather small coast effect has been found in eastern Siberia on the coast of the Sea of Japan [*Nikiforova et al.*, 1978].

k. Islands

A definite coast effect ('island effect') has been observed on several islands. Induction arrows tend to point away from the island toward the sea. When the superimposed effect of the Japan anomaly is removed, this pattern has been found at Ooshima [*Sasai*, 1968] and Hachijo-Jima [*Honkura*, 1974]. In various places on Oahu, different phases are found for the Z component of Sq, the effect increasing with frequency to a complete reversal at a period of 1 hour [*Mason*, 1963]. The island effect on Hawaii has been evaluated and removed to determine conductivity as a function of depth [*Klein*, 1975; *Klein and Larsen*, 1978].

A typical island effect is found on Puerto Rico [*Elvers et al.*, 1965].

On the Kurile Islands a modified island effect is seen which amounts to an extreme case of current channeling, especially between Iturup (Etorofu) and Urup (Oruppu) [*Rokityansky*, 1975*a*, p. 217]. However, this anomaly may have a deep source [*Adam*, 1976, p. 698].

l. Inland Decrease

Several workers have reported that different rates of decrease inland are found in different geological environments. *Everett and Hyndman* [1967*a*], working on data acquired in Australia, resolved the arrows representing the real and imaginary parts of the transfer functions into components perpendicular to the coastline and compared these as a function of distance from the coast with similar arrows from the work of *Schmucker* [1963] for southern California. They found that the coast effect for southwestern Australia is considerably larger and decreases inland at a different rate compared with either California or southeastern Australia, though California and southeastern Australia agree well with one another. Figure 1 summarizes data on the inland decrease of the amplitude of the real transfer function.

4. Calculations of the Induction Effect of Seawater

Whether induction in seawater is the principal cause of the coast effect or not, it is important to be able to evaluate its influence. If it can be sufficiently well evaluated, then its effect can be subtracted from observations to give information about other sources of the induced field. Therefore a great deal of effort has gone into calculating the induction effect of the oceans. Three approaches to the problem have been tried: exact analytical solutions of simple conductivity distributions, iterative methods applied to a realistic model representing ocean conductivity, and local induction problems (flat earth approximations) solved by numerical operations over a grid of mesh points.

a. Global Calculations

Analytical solutions to induction in thin plates forming part of a sphere have been investigated mainly by *Ashour* [1971*a*, *b*, 1973]. Unless the model is kept reasonably simple, analytical calculation is impossible. On the other hand the model must be sufficiently realistic to be useful when it is applied to the real earth. The compromise used by *Ashour* [1971*b*] comprises a uniform external field, a hemispherical 'ocean' whose conductivity (integrated over depth) is a function of colatitude θ only and is given by

$$\sigma(\theta) = \sigma_0(1 + \alpha)(1 + \alpha \cos^2 \theta)^{-1} \cos \theta$$

where α is a parameter that can be chosen to vary the conductivity distribution. The conductivity varies smoothly to zero at the equator. If there is a discontinuity in surface conductivity, the field becomes infinite, but the average field over any finite region remains finite. The conductosphere is represented by a perfectly conducting concentric sphere with radius $0.875R$, at the surface of which the normal component of the field vanishes. Ashour makes use of the fact that the difference between the horizontal field components above and below the ocean equals the current density, which can be expressed in terms of the conductivity and the time rate of change of the vector potential.

$$[dA^{(i)}/dr]_{r=R+} - [dA^{(i)}/dr)]_{r=R-} = 4\pi\sigma(\theta)(d/dt)(A^{(i)} + A^{(e)}) \quad (3)$$

This equation is the analogue of an equation for the current function derived by *Price* [1949]. The vector potential may be expressed in terms of a harmonic series with coefficients B_n, and (3) results in an infinite system of linear equations in the coefficients B_n. These were solved by successive approximations for periods of 24 hours, 12 hours, and 1 hour. The shorter the period, the slower the convergence of the iteration process. Equation (3) takes no account of the electrostatic charges that result when current flows across isoconductivity surfaces [*Price*, 1964]. The symmetric case is still valid, but the asymmetric case requires modification [*Ashour*, 1974].

Rikitake [1961*a*] also tackled the problem of a hemispherical ocean. He considered a conductosphere of finite conductivity as well as a primary field and frequency appropriate to the diurnal variation. His conclusion was that only negligibly small vertical components could be produced by a hemispherical ocean.

The basis of the iterative methods was laid by *Price* [1949], who showed that given an approximation to the current function, a better approximation can be obtained by one of two iteration methods. The convergence of the methods depends on a parameter

$$\eta = (2n + 1)R(\mu_0 a\omega)^{-1}$$

where R is the surface resistivity, a the radius of the earth, ω the angular frequency, and n the degree of the spherical harmonics involved in the expansion of the current function. One method converges if the maximum value of η is <1, and the other converges if the minimum value of η is >1. Unfortunately, when actual values of the conductivity of seawater and land are substituted (at a typical substorm frequency), the values of the parameter η straddle unity and extend some orders of magnitude on each side of it. This is not so if only diurnal frequencies are considered. *Bullard and Parker* [1970] made use of this in the most successful calculation so far of the induction effect of the oceans. They considered only variations of 24-hour period, for which the parameter η is less than 1 for both ocean and land. Bullard and Parker pointed out that the Price scheme was equivalent to a Fredholm integral equation, and *Hewson-Brown et al.* [1973] used this fact to improve the range of convergence. A similar approach to the convergence of these iterations has been shown by *Hobbs and Brignall* [1976] to be applicable to nonsymmetric models and so should be capable of solving the general problem. A somewhat different approach to this problem has been taken by Zinger and Fainberg [see *Fainberg*, 1978].

b. Local Calculations

The local induction problem is studied by using a model of a semi-infinite conductor with a plane boundary. *Price* [1950] presented a general theory for electromagnetic induction in a plane earth by any inducing field. Since that time, much work has been done on plane earth problems. Both analytic and numerical methods have been used.

The first work in which the continent-ocean edge effect seems to have been studied explicitly was that by *Weaver* [1963]. Weaver's two-dimensional model was that of a semi-infinite conductor with a plane boundary consisting of two quarter spaces of different finite conductivity. In two-dimensional models of this kind, two separate types of solutions are encountered, and these are sometimes termed the H and E polarization cases, depending on whether the magnetic or electric field is parallel to the direction along which the model extends uniformly. Weaver considered both the H polarization and E polarization cases analytically. In the E polarization case it was necessary for him to use an approximate boundary condition requiring the horizontal magnetic field at the surface of the conductor to be constant. However, he was able to show, for this case, the increase in amplitude of the magnetic component normal to the surface of the conductor as the region of the discontinuity is approached, and he related this to the enhancement of the vertical magnetic component observed at coastal stations. A numerical approach to the quarter-space problem was taken by *Jones and Price* [1970], who showed that in the E polarization case, both the horizontal and the vertical components of the magnetic field vary across the vertical contact. *Weaver and Thomson* [1972] have subsequently improved on *Weaver's* [1963] work by using a perturbation technique proposed by *Mann* [1970] and have avoided in the E polarization case the use of the earlier approximate boundary condition so that better solutions of the semi-infinite quarter-space problem have been obtained.

Sloping and wedge models have also been considered for representing the ocean-continent interface. *Blake* [1970], *Geyer* [1972], and *Nabighian et al.* [1967] have considered analytical solutions for dipping contacts, and *Jones and Price* [1971] and *Tatrallyay and Jones* [1974] as well as others have used numerical methods for such models.

A more complicated model was studied by *Lines et al.* [1973]. It included lateral inhomogeneous structures at depth, which could correspond to variations in structure between oceanic and continental mantle. They undertook calculations for three models: (1) only ocean-continent structure, (2) only mantle-crust structure, and (3) ocean-mantle-crust structure. Of primary interest in the work was the comparison between models 1 and 3. From the results it was apparent that especially at longer periods the mantle discontinuities strongly affect the surface fields. A similar effect has been emphasized by *Ranganayaki and Madden* [1978]. For the model and frequencies studied, the coastal anomaly at longer period was strongly attenuated by the variation in mantle structure. It is apparent that for such a model a strong mutual coupling between oceanic and mantle currents exists and the net result is a decrease in the effect of the ocean alone. Although one can, for this particular set of models, separate the ocean and mantle effects, it is difficult to draw general conclusions relating to actual observations in field measurements.

A number of recent papers have considered the possibility that part of the current flowing in the oceans returns via the region of finite conductivity under the oceans. *Bailey* [1977] and *Nicoll and Weaver* [1977] obtained analytic solutions for a model in which a perfectly conducting ocean occupies half the $z = 0$ plane and the half space $z > 0$ has a uniform finite conductivity. Most of the current flow is confined to about 1.5 skin depths below the surface.

Brewitt-Taylor [1975, 1976] considered a similar model but with a finitely conducting ocean, a perfectly conducting conductosphere, and a horizontally layered finitely conducting space between. He concludes that the depth of the conductosphere has only a slight effect on the current distribution. Both of these models are two-dimensional, and the fields assumed were H polarization. The results are useful for magnetotelluric impedances but not for magnetic field studies.

Fischer et al. [1978] have solved the problem of a perfectly conducting ocean for E polarization. The conductivity under the ocean is considered to be the same as that under the continent, and no conductosphere is included in the model. The transfer functions near the coastline are not unlike those observed, but the predicted phase relationship between vertical and horizontal fields is not that observed.

A more general model is that used by *Vasseur and Weidelt* [1977], in which they consider a nonuniform conducting lamina overlying uniform horizontal layers. The problem deals with a one-dimensional half space perturbed by a conductivity anomaly in the lamina.

A complicated coastal model in three dimensions was considered by *Jones and Lokken* [1975] using a numerical approach. They found that effects of smaller-scale coastal features on measurements of a local nature may be pronounced.

Even though calculations and measurements of this kind have been made, no explicit conclusion concerning the relative contributions of seawater and mantle differences have been obtained. It may be that on such a scale the results depend greatly on the model parameters employed, and so general overall results are elusive. It is apparent that a definitive field or model experiment is required, and this is the direction that should be taken in future work of this nature.

5. EXPERIMENTAL MODELS

The solution of problems in electromagnetic induction by scale models is common in applied geophysics [*Strangway*, 1966]. A necessary condition of the scaling is that the magnetic Reynolds number $\mu\sigma L^2/t$ be preserved. It is also necessary that the ratios of both σ and μ separately (not just the product) between media be the same for the model as for the earth. Some effort has been made to apply these techniques to the seawater induction problem. Several models have used the flat earth approximation. This is legitimate if only local induction is involved, but when the importance of induction far from the site being studied is not known, such models are open to question.

a. Global Models

Global models have some disadvantages. To be accommodated in a laboratory, they must be of the order of 1-m diameter, which involves a reduction in linear dimensions of the order of 10^7. This makes it difficult to duplicate fine detail or shallow water. The advantage of global models is that the problems of return current loops and end effects are eliminated. Also, complicated source fields can be used.

Nagata et al. [1955] constructed spherical and hemispherical conducting sheet models which they used to verify numerical calculations.

The only other global model known to the authors is the one constructed by *Parkinson* [1963], known as an 'induction terrella.' It consists of an insulating sphere 0.56 m in radius to which are attached copper sheets cut and beaten into the shape of the deep oceans. Concentrically inside the sphere is an aluminium sphere of 0.504-m radius to duplicate the conductosphere. The ratios of conductivity and length on the terrella (t) and earth (e) are

$$\sigma_t/\sigma_e = 2.67 \times 10^7 \qquad L_t/L_e = 0.88 \times 10^{-7}$$

The ratio of frequencies is chosen to preserve the magnetic Reynolds number.

The source field is duplicated by a wire coil wound onto a hemispherical frame in the form of the current flow lines suggested by *Silsbee and Vestine* [1942]. This can be placed over any hemisphere of the terrella. As it is rotated about the pole the horizontal field at a fixed point assumes various orientations. By measuring the ratio of vertical to horizontal fields with small search coils it is possible to determine the transfer functions A and B.

The transfer functions indicated by the terrella for Kakioka, Valentia, and the coast of California are quite different from those observed at the actual sites. The influence of the ocean water seems to be almost negligible at these positions on the terrella. However, at several southern hemisphere sites the coast effect is quite well duplicated. This appears to be because the auroral electrojet part of the source field (where the primary vertical field is strongest) is over an ocean in the southern hemisphere and over land in the northern hemisphere. In fact, if the primary field coil is moved equatorward, so that the auroral electrojet is over the northern Pacific, the current density and vertical field near the coast of California increase greatly, as is shown in Figure 2.

Another interesting observation made with the terrella is that the transfer functions at Carnarvon and Brisbane (on the west coast and east coast of Australia, respectively) depend quite strongly on the conductance of the connection between the Indian and Pacific oceans north of Australia. This is shown in Figure 3.

These experiments indicate that (1) although the ocean water (treated as an insulated conductor) can influence the field considerably, it alone does not explain the observed coast effect and (2) when only insulated oceans are considered, the magnitude of the vertical component near a coastline is not controlled by the local horizontal field but rather by distant vertical fields.

It should be emphasized that although the ocean water was accurately modeled on the terrella and the primary field was probably reasonably close to that which occurs in nature (but see, for instance, *Bannister and Gough* [1978] on the primary field), the absence of a finite conductivity layer under and in contact with the oceans changes the whole nature of the model. Induction by local horizontal fields, which is inhibited on the terrella, may take place. In fact, the indications given by the terrella that induction by distant sources is important constitutes perhaps the strongest argument against an insulated oceanic layer as the site of the eddy currents responsible for the coast effect.

b. Local Models

Analytical and numerical solutions of local induction problems have been discussed in section 4*b*. Analogue modeling is another approach which has the advantage that it is not restricted to greatly simplified models but can be used to study complicated two- and three-dimensional problems. It is perhaps the most potentially productive means of studying local coast effect problems and has been used to good effect by Dosso and his colleagues as well as others. In a review paper of such studies of the coast effect, *Dosso* [1973] described the technique as well as the results for some complicated models of the continent-ocean interface. He pointed out that an underlying highly conducting step in coastal regions can play an

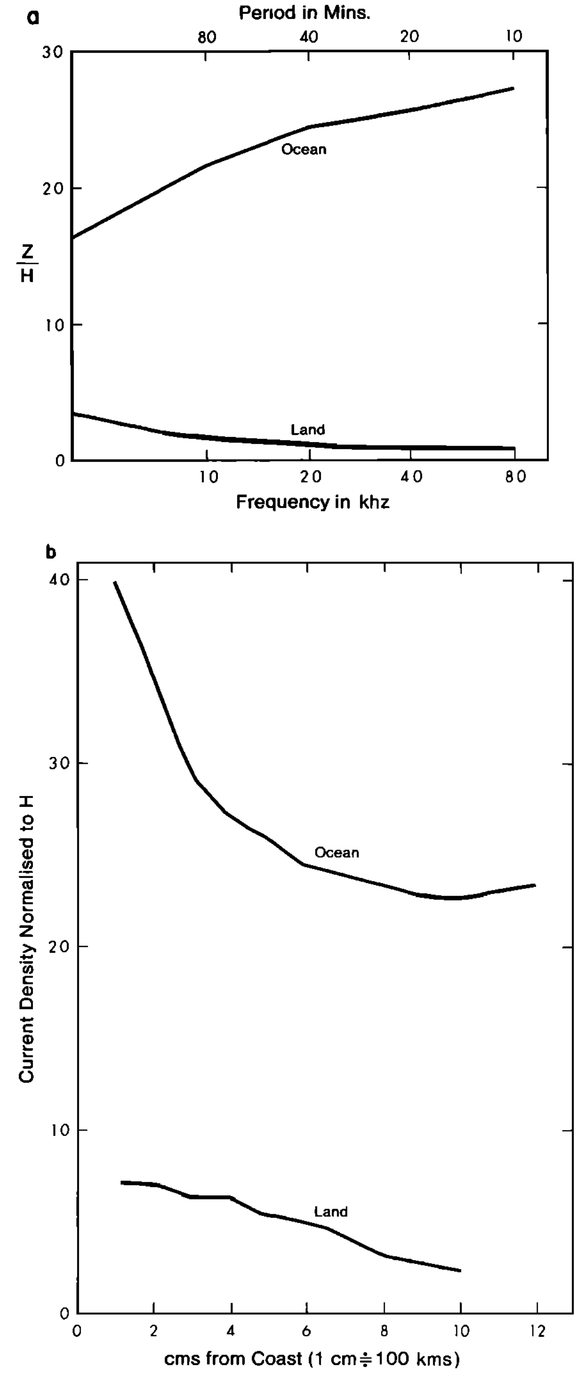

Fig. 2. Terrella results showing the effect of position of the primary current loop. The curve marked ocean was obtained with the center of the primary current loop over the Pacific Ocean, and that marked land with the center over North America. (a) Frequency spectrum of Z/H (in-phase part; 1 cm represents 100 km) inland. (b) Current density as a function of distance from the coastline.

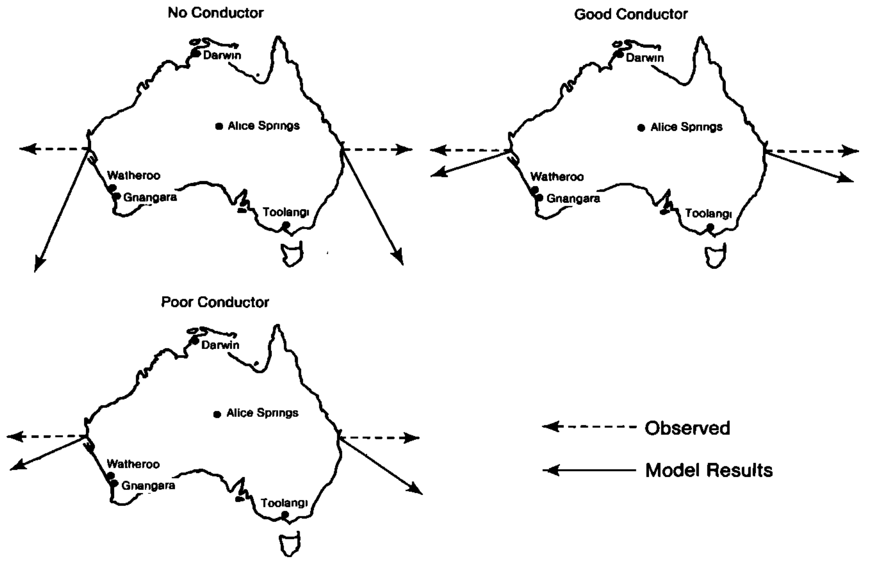

Fig. 3. Induction arrows at Carnarvon and Brisbane showing the effect of the conducting path between the Pacific and Indian oceans. The dashed arrows indicate the observed in-phase transfer functions. The solid arrows indicate terrella results.

important role in the coast effect. The location of this step relative to the coastline is also a critical factor in the nature of the coast effect.

In an experiment performed by *Launay* [1970] the coast effect was duplicated quite well with a model that included a conductor underneath the ocean. Instead of using a thick conductor to represent the conductosphere she inserted an image current system at twice the depth of the conductosphere. This technique, however, neglects the mutual inductance of the ocean and conductosphere. A model experiment to explain the coast effect recorded in Iceland was carried out by *Hermance* [1968]. The model indicates a greater effect than is observed, probably because no conductosphere was included. *Rikitake* [1961*b*] used a model that amounts to an analogue computer to determine the effect of the conductosphere. He found that the presence of a conductosphere greatly decreases the effect of the oceans. This is understandable considering the decrease in vertical component caused by the image currents. One of the first applications of scale models to electromagnetic induction was an attempt by *Roden* [1964] to account for the Japanese anomaly. He used a purely vertical primary field, no conductosphere, and a flat earth approximation. The distortion of the field caused by a hole in a copper sheet accurately simulating the coast of Japan and nearby Asia resembled the effect observed near Japan [*Rikitake and Honkura,* 1973]. However, no indication of the corresponding horizontal field was present in the model, so there is no way of normalizing the anomalous vertical field with the corresponding horizontal field.

6. Cause of the Coast Effect

a. Models to Explain the Coast Effect

Three possibilities exist to explain the coast effect: (1) induction in an effectively insulated set of oceans joined together by shallow water; (2) induction in a conductor of considerable vertical extent the top of which is the highly conducting ocean water; (3) induction in the deep crust and mantle, which have significantly different conductivity distributions under oceans and continents.

The observations may be summarized by saying that along most coastlines the transfer function arrow is perpendicular to the coastline and that Z/H (H being the horizontal component perpendicular to the coastline) is between 1.0 and 0.3. The fact that this is almost universal seems to favor possibility 1 or 2. But there exist a few places where this is not the case. The most outstanding is the coast of southern Peru. Also, near Honshu [*Rikitake,* 1971], seawater does not seem to be the controlling factor. Another example of the absence of a coast effect is Sable Island [*Cochrane and Hyndman,* 1974; *Hyndman and Cochrane,* 1971], which lies at the edge of the continental shelf near Newfoundland but has $Z/H < 0.1$ for all directions of H.

These examples, however, do not necessarily eliminate possibility 1. It is well known that at many inland locations, conductivity anomalies cause the vertical component to be large [e.g., *Gough,* 1973*a*], and sites such as southern Peru may be where the normal coast effect is interfered with by deep conductivity anomalies. This is particularly likely in tectonically active areas such as Peru and Japan.

Calculations of *Ashour* [1971*b*] as well as model experiments of *Roden* [1964] indicate that possibility 1 is capable of generating a sufficiently large Z/H. However, both of these tend to exaggerate the induced currents because of the abnormally high vertical component of the primary field. The terrella experiments (see section 5*a*) indicate only a weak influence of seawater in the northern hemisphere. It is hard to reconcile these results with possibility 1. Furthermore, the terrella indicates that local induction is unimportant in 1 and that eddy currents in the oceans in temperate latitudes are largely in the nature of channeled currents generated at higher latitudes. This is inconsistent with the high correlation between the vertical field and the local horizontal field observed at most coastlines.

Several investigators have examined their results at specific locations to distinguish between possibilities 1 and 3. *Bennett and Lilley* [1971] found that 1 cannot account for their observations but that it was necessary to postulate a shallower depth to the (assumed perfectly conducting) conductosphere under the ocean than under the continent. *Dosso* [1973] con-

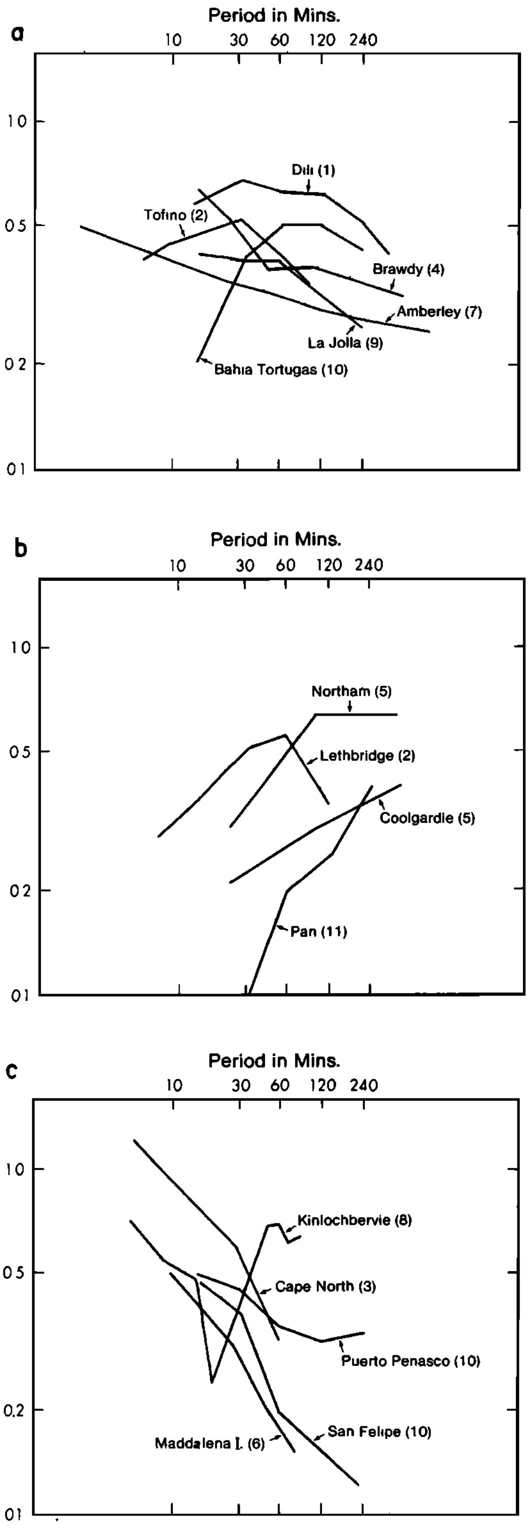

Fig. 4. Frequency spectra of real transfer functions for various locations. (*a*) Frequency spectra of real transfer functions for various coastal locations. (*b*) Frequency spectra of real transfer functions for various inland locations. (*c*) Frequency spectra of real transfer functions for various locations at which channeling is likely to be important. The data are taken from the following references, indicated by numbers after the place names: (1) *Chamalaun and White* [1975], (2) *Cochrane and Hyndman* [1970], (3) *Cochrane and Hyndman* [1974], (4) *Edwards et al.* [1971], (5) *Everett and Hyndman* [1967*a*], (6) *Giorgi and Yokoyama* [1967], (7) *Lawrie* [1965], (8) J. M. Sik (personal communication, 1978), (9) *White* [1973*a*], (10) *White* [1973*b*], (11) *White and Polatajko* [1978].

structed a model to represent the coast effect at Tofino, including an accurately shaped sea floor. He concluded that deep water cannot account for the observed coast effect. The calculations of *Richards* [1970] do not agree with observations of *Lilley and Parker* [1976] for the diurnal variation in Australia. On the other hand the results of *Greenhouse* [1972] near the coast of California could be explained by varying water depth, without any change in the depth to the conductosphere.

Certain features of the coast effect could be considered to be diagnostic. The three most obvious of these are the phase difference between the primary and secondary fields, the decrease in the anomalous field in the inland direction, and the frequency spectrum of Z/H.

The calculations of *Parker* [1968] for thin strips, as well as those of Ashour for hemispherical shells, predict large phase differences between the primary and secondary fields. Making the approximation that the vertical field is mainly secondary and the horizontal field mainly primary (or at least in phase with the primary field) this predicts a similar large phase difference between vertical and horizontal components, that is, the imaginary part of the transfer function should be larger than the real part. Only very rarely is this the case. The terrella results also indicate an almost real transfer function, although this aspect of the experiment requires more careful investigation. Considerations of phase do not favor possibility 1.

Three suggestions have been made to account for the varying rate of decrease of the transfer functions in the inland direction. *Everett and Hyndman* [1967*a*] associated it with the geology of the continent, a region of recent tectonic activity giving the most sudden decrease and a shield the most gradual. Later work such as that of *Bennett and Lilley* [1971] seems to agree with this suggestion, although very few shield areas have been investigated. *Launay* [1970] ascribed the different rates of inland decrease to varying depths to the conductosphere. *Roden* [1964] pointed out that the rate of inland decrease depends on the width of the ocean. The last two are based on models in which the eddy currents flow only in the oceans. The fact that Roden's relation does not apply is an argument against possibility 1 above. It is difficult to distinguish observationally between the explanation offered by Everett and Hyndman and that by Launay. The observed inland decrease can most easily be explained by possibility 3, but it is not inconsistent with 2.

If the eddy currents causing the coast effect flow exclusively in the ocean water, they can give rise only to a vertical field except very near the coast [*Bullard*, 1967]. Currents flowing at depth should cause an anomalous horizontal field perpendicular to the coast and in the same sense as that component of the primary field. Not many array studies have included a continental coastline. Perhaps the one most likely to show this effect is the array survey of *Bennett and Lilley* [1972]. This indicates very little change in the horizontal component across the array, which argues for a shallow location for the eddy currents.

The period spectra of most coastal sites are fairly flat over the period range normally observed (see Figure 4). However, the maximum value of the transfer function Z/H appears to be at a period of about 30 min. *Rokityansky* [1975*b*] applied a transform similar to the Kertz operator to a two-dimensional model containing an anomalous conductor. He derived a relation between the peak of the period spectrum and the product of the cross section and conductivity of the anomalous conductor. Applying this to the GCE he concluded that the ocean is sufficiently deep to explain the period spectrum and that mantle conductivity, such as possibility 3 above, would result in a period spectrum with a peak at much longer periods.

The peak of the period spectrum for the sites of Figure 4c is less than 10 min, as would be expected if they are dominated by current flowing in comparatively poor or shallow conductors.

The period spectrum derived from terrella results depends on the position of the primary field, as shown in Figure 2a. This is probably because of the different relative importance of channeling and local induction (what Rokityansky calls 'conductive' and 'eddy current' anomalies).

Hyndman and Cochrane [1971] identify four types of spectra which can be predicted by two-dimensional models. These models include vertical flow of current beneath the oceans. The similarity of these four types with observations in the Gulf of Saint Lawrence area seems to support local induction.

There is some controversy in the literature as to whether the eddy currents that cause the coast effect are induced by vertical or horizontal components of the primary field. *Bullard and Parker* [1970] have pointed out that in the ocean, insulated from the substratum, the induction effect by horizontal components is negligible because it causes vertical current loops. Therefore if seawater is the main cause of the coast effect, the inducing component of the primary field must be the vertical component. *Bailey and Edwards* [1976] also considered the vertical component of the primary field to be important. On the other hand, *Lilley and Parker* [1976] found that at diurnal variation frequencies the anomalous vertical component at Gnangarra (Western Australia) correlates better with the onshore horizontal components than with the normal vertical component. *Bennett and Lilley* [1973] stated that 'the existence of Parkinson vectors implies induction by horizontal variations or correlation between components of primary field.' They consider that such a correlation is unlikely. However, there are some strong constraints on the primary field. Not only are both the divergence and the curl zero, but the primary field is of external origin (together with a predictable image). In fact, given either the vertical or the north component of the primary field over the surface of the sphere it is possible to calculate the other two components and the complete image of the field. Not only this, but the current pattern of substorms or of diurnal variation is of a fairly well known and consistent form. Therefore polarization involving statistical relationships between the components of the primary field is not unlikely. Polarization of the horizontal field has been reported at Alert [*Niblett et al.*, 1974] and of the anomalous part by *LeBorgne and LeMouel* [1975]. However, it is felt (E. R. Niblett, personal communication, 1979) that the polarization at Alert is associated with the secondary or induced part of the horizontal field rather than the primary horizontal field.

b. *Seawater Induction*

In this section the effect of ocean water insulated from its substratum is considered. It is sometimes said that because the coast effect appears to be a dependence of Z on the horizontal components of the primary field, its ultimate cause must be induction by horizontal fields.

Calculations similar to those of Ashour, described in section 4a, show that this is not so. Let us consider a point just off a hemispherical ocean occupying the northern hemisphere. When the primary field is horizontal and toward the north (i.e., primary field parallel to the axis), the vertical field is about $0.9H_0$, and the horizontal field equal to $1.06H_0$ (B. A. Hobbs, personal communication, 1978). When the primary field is horizontal and toward the east, the vertical field is zero, and the horizontal field is about $1.3H_0$. Now if the primary field is horizontal and at an angle ϕ east of north, it can be considered to be a combination of these two fields. So the north, east, and vertical components are

$$X = 1.06H_0 \cos \phi$$

$$Y = 1.3H_0 \sin \phi$$

$$Z = 0.9H_0 \cos \phi$$

and (2) applies with $A = 0.85$, $B = 0$. Thus the site experiences a typical coast effect, although the primary field is purely horizontal at the site. The induction, of course, is caused by vertical fields at some distance from the site. The question is not whether seawater can produce a coast effect. It is whether the seawater alone does produce it, and this depends on the magnitude of the total flow of current in the oceans.

c. *Conductive Structure Under Oceans and Continents*

An important factor in looking for an alternative to seawater conduction to explain the coast effect is the distribution of conductivity with depth under the continents and oceans.

The global studies summarized in section 1 are mainly based on data from continental sites, so these results can be considered to represent continental conditions. A feature of all models suggested to explain these data is that they contain a low-conductivity zone extending either from the surface or immediately below a surface conducting layer to a depth of some hundreds of kilometers. In spite of the abundance of continental data a rather wide range of models has been suggested [*Banks*, 1972; *Parker*, 1970]. Unfortunately, many of the observatories supplying the data are affected by the coast effect. The inclusion of these observatories may have exactly the opposite effect to that which might be expected. The conductivity distribution is based on the complex ratio of vertical to horizontal components of diurnal variation fields. Stations near coastlines will have greater vertical components, not because the conductosphere is deeper but because it is not horizontal or, alternatively, because of the influence of seawater. A greater average vertical field results, and this indicates a depth to the conductosphere greater than the actual depth. This can be minimized either by basing the calculations only on very long period variations [*Banks*, 1972] or by making use of magnetotelluric observations in regions far from the coast. Magnetotelluric results suggest a shallower depth to the conductosphere than do magnetic results [*Niblett*, 1967]. Estimates of the conductivity structure beneath the oceans are based on a much smaller amount of data. *Filloux*'s [1967] data observed off the coast of California suggest a conductivity of the order of 10^{-1} mho m^{-1} at shallow depths but give little information at a depth below 100 km. *Greenhouse*'s [1972] work indicates a somewhat lower conductivity. As *Schmucker* [1973] has pointed out, Filloux's results indicate that induction in seawater is of less importance, and subocean currents of more importance, than was previously thought.

Cox et al. [1970] suggested a conductivity of about 0.5 mho m^{-1} at a depth of 25 km, with probably a low conductivity (less than 0.1 mho m^{-1}) above that level. Perhaps the most

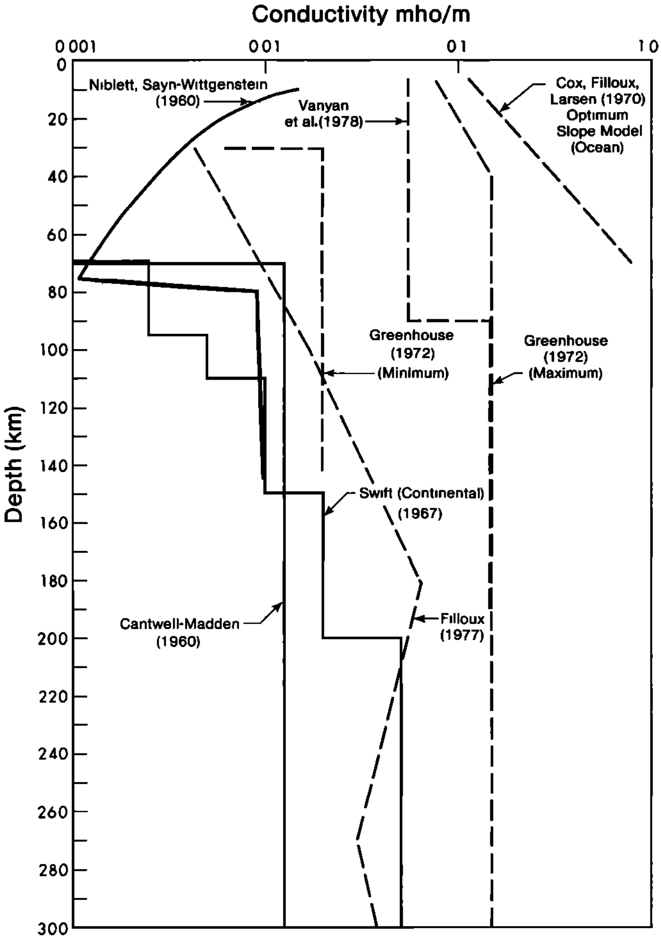

Fig. 5. Conductivity distribution from continental and oceanic measurements.

definite information comes from the work of *Filloux* [1977], who found that in the North Pacific the conductivity increases from about 10^{-2} mho m^{-1} at the sea floor to about 1 mho m^{-1} at a depth of about 400 km. The work of *Larsen* [1975] on the Hawaiian Islands indicates a conductivity rising monotonically with depth, except for a peak at 350 km, from a near-surface value of about 0.1 mho m^{-1}. A similar model was found by *Klein* [1975]. More recent work of *Klein and Larsen* [1978] on Hawaii indicates a monotonically increasing conductivity of 10^{-2} mho m^{-1} near the surface to 10^{-1} mho m^{-1} at a depth of 260 km. Unfortuantely, Hawaii is unlikely to be typical of oceanic locations. However, the low-conductivity upper layer common to all continental models is absent.

Poehls and Von Herzen [1976] carried out a geomagnetic deep-sounding survey on the floor of the Atlantic Ocean near Bermuda. Their interpretation is a conductivity of 0.1 mho m^{-1} to a depth of 120 km followed by a decrease to 0.025 mho m^{-1}. *Vanyan et al.* [1978] have combined these data with some of *Berdichevsky et al.* [1970] and suggest a model with a conductivity of 0.055 mho m^{-1} for the top 90 km (which they call the lithosphere) and 0.15 mho m^{-1} for the next 100 km. Apart from Poehls and Von Herzen's lower layer these conductivities are within a factor of 2 of Larsen's results. The probing in the Atlantic was not deep enough to confirm the peak at 350 km claimed by Larsen. Figure 5 summarizes the results as far as they are known at present.

In spite of the uncertainties in both continental and oceanic conductivity profiles it seems likely that there is a low-conductivity region under continents with conductivities significantly less than 0.01 mho m^{-1} which is either absent or much thinner under oceans. The difference may be very significant for the explanation of the coast effect.

d. Finite Conductivity Beneath the Oceans

The importance of flow of current in a conductor underneath and in contact with the oceans was first pointed out in papers given at the Second Workshop on Electromagnetic Induction in the Earth by Brewitt-Taylor and Bailey [*Nicoll and Weaver*, 1977]. It was again emphasized by C. S. Cox at the IAGA Assembly in Seattle (1977).

The conductivity of seawater is about 4 mhos m^{-1} and that of the continental lithosphere is of the order of 10^{-4} mho m^{-1}. It seems therefore natural to model the problem of induction in the oceans as if the oceans were insulated. The conductivity of the oceanic lithosphere, however, may be as high as 0.1 mho m^{-1}, so that insulated ocean approximation is not as good as might be thought.

Cox [1978] pointed out that vertical flow of current through the ocean floor is important if the suboceanic conductivity is more than a critical value and suggested that this critical value depends on the width of a 'boundary layer' at the edge of the ocean in which current flow is controlled by electrostatic charges on the coastline. Experiments with the terrella indicate a boundary layer of the order of 500 km wide, corresponding to a critical value of about 10^{-2} mho m^{-1}. The suboceanic conductivity may well be above this critical value.

If this is so, then the fundamental nature of the problem changes. Vertical current loops and induction by horizontal fields become possible. According to *Bailey* [1977], half of the current in the oceans could flow in the top 250 km of the suboceanic mantle (assuming a conductivity of 0.01 mho m^{-1}).

There is strong evidence for the vertical flow of induced currents in some places; for instance, the channeled current along the Gulf of California [*White*, 1973*a*] seems to have little influence on the line of stations established by *Schmucker* [1970*a*]. A vertical flow of current between the northern end of the Gulf of California and Yuma seems likely. An even more striking example of convergence of surface current flow is that described by *Garland* [1975] in eastern Canada.

Calculations on models including a conductivity region beneath the oceans have not received much attention to date. The difficulty in analytical and numerical solutions is that a two-dimensional model is of limited use. The E polarization produces no vertical currents, and so return flow through the lithosphere is excluded, and H polarization produces no vertical magnetic field and therefore no coast effect. This was pointed out by *Brewitt-Taylor* [1975], who went on to consider a three-dimensional perfectly conducting ocean partly capping a two-dimensional prism. He assumed that the lithosphere was a poor conductor and neglected self-induction. With these restrictions he calculated that there would be a vertical field of the correct order of magnitude to account for the coast effect.

Bailey [1977] obtained an analytic solution of the H polarization case with a uniformly conducting substratum below a perfectly conducting ocean. This gives a useful idea of the depth of current flow. Most of the current is confined to a depth of 2 skin depths, which is less than 200 km for a period of 1 hour. This model is not applicable where ocean lithosphere meets continental lithosphere (conductivity 10^{-4} mho

m^{-1}), but it does suggest that the oceanic lithosphere may furnish the return path for an appreciable fraction of the current and that it is not just a conducting path to the conductosphere.

e. The Source Field

So far we have not considered the effects of source field configurations. On a global scale the source field configuration has considerable effect on the overall currents in the earth and so on the currents near coastlines.

It is usual to consider the primary field at temperate latitudes as being horizontal. If we accept a source field configuration like that of *Silsbee and Vestine* [1942], the vertical field is small but not negligible. The fact that it is correlated with the equatorward component of the horizontal field [*Parkinson,* 1964] influences the coast effect. The very strong coast effects experienced on coastlines where the continent is nearer to the equator (e.g., Esperance, Hermanus, Paramaribo) is probably caused by this phenomenon.

f. Local Induction and Current Channeling

According to the local induction model, induced fields observed at a site can be explained by induction in the local conductors by fields measurable at that site. The alternative is that of eddy currents induced at a great distance and flowing near the site because of the configuration of conductors. These are usually known as channeled currents. The importance of such currents was pointed out by *Price* [1964]. There are several instances where channeling is most likely, such as the North American central plains anomaly [*Alabi et al.,* 1975], the Gulf of Saint Lawrence [*Cochrane and Hyndman,* 1974], the Gulf of California [*White,* 1973*a*], and Alert [*Niblett et al.,* 1974]. *Nienaber et al.* [1973] found anisotropic apparent resistivity near Victoria (British Columbia) which they ascribe to current channeling around Vancouver Island.

Some kinds of channeling seem to take place around the British Isles according to *Edwards et al.,* [1971], but it is not clear whether the shallow water of the North Sea and the English Channel constitutes the conductor. Similar channeling occurs across Scotland near Eskdalemuir. It is significant that there seems to be no observable current channeling through Bass Strait [*Lilley and Bennett,* 1972].

The question of whether the coast effect is nearer to the model of local induction or channeling has not been settled. All two-dimensional models tacitly assume local induction [*Gough,* 1973*a*], and the apparent success of many interpretations based on such models argues for some form of local induction. *Bennett and Lilley* [1974] and *Booth and Stuart* [1974] consider local induction to be important in interpreting the coast effect.

The terrella experiments show that if the model of insulated oceans as the course of induced currents (possibility 1 of section 6*a*) is maintained, the coast effect must be caused by some form of channeling in which the presence of the continents has a major effect on the whole current flow in the oceans.

g. Island Effects

An early investigation [*Parkinson,* 1962*b*] indicated that stations on islands behave quite erratically regarding the coast effect. More detailed studies [*Klein,* 1975; *Mason,* 1963; *Elvers et al.,* 1965] showed that the effect depends on the location of the station on the island. This suggested a sensitivity to fine details of the coastline that does not appear at many stations on the coasts of continents. The explanation of the strong coast effect even on small islands is tied up with the question of local and distant induction. In the case of small islands it seems certain that the effect that we see has its cause far away. The problem is that of an almost uniform current sheet being perturbed by the low conductivity of the island [*Price,* 1964]. The island effect is great not because the island is small but because the ocean is large. It is interesting to note that although small islands have well-developed coast effects, quite large inland seas do not. Examples are the Caspian [*Adam,* 1976, p. 652] and possibly the Mediterranean [*Babour et al.,* 1976].

h. Conclusion

Considering all the evidence available at present, the opinion of the authors is that the most likely cause of the coast effect is induction in large vertical loops involving the ocean water, the oceanic lithosphere, and possibly the conductosphere, i.e., possibility 2 of section 6*a*. Because of its higher conductivity the ocean water has a controlling influence on the form of current flow at sea level. This explains the high correlation of vertical field near the coastlines with the direction of the local horizontal field and the tight control exercised by the shape of coastlines on the direction of the induction arrows. It also explains why two-dimensional modeling has, in general, been successful in explaining observations.

Exceptions to the normal coast effect occur where the normal current flow is severely interfered with by abnormal underground conductivity, generally by abnormally high conductivity under the adjacent continent.

7. TECTONIC IMPLICATIONS

Conductivity in earth materials is controlled by three parameters: water content, temperature, and partial melting. Which of these three parameters is most important depends partly on depth. Electromagnetic observations often have poor depth resolution, which makes it difficult to distinguish between the three mechanisms. However, frequency dependence gives some information [*Rokityansky,* 1975*b*].

Shallow upper crustal anomalies are obviously controlled by the greater water content of more porous rocks. *Hyndman and Hyndman* [1968] stressed the importance of water in the lower crust also. They pointed out that at crustal depths, semiconductor conductivity is unimportant and partial melting of granitic material changes the conductivity only slightly.

Changes in conductivity are associated with the dilatancy that often occurs before rupture of rocks. Changes in transfer functions have therefore been used as predictors of earthquakes. The subject has been reviewed by *Rikitake* [1976] and by *Niblett and Honkura* [1978].

The relation between conductivity and temperature in semiconductor material (such as the mantle) is given [*Tozer,* 1959] by

$$\sigma(T) = \sum_i \sigma_i e^{-(T_i/T)}$$

where T is the absolute temperature and σ_i and T_i are constants, depending on the material. An interpretation of conductivity anomalies in terms of heat flow and mantle temperature has been given by *Uyeda and Rikitake* [1970]. They associate low ΔZ anomalies with high heat flow and directional anomalies, such as the GCE, with sloping isothermal surfaces.

Chan et al. [1973] have investigated the process of increased conductivity caused by partial melting. They consider that this mechanism can cause a greater increase in conductivity than can temperature alone. This might make conductivity critically sensitive to pressure in certain temperature ranges.

The GCE is consistent with the well-known idea that isotherms are shallower under the oceans than under the continents [*Mercier and Carter,* 1975]. *Sugimura and Uyeda* [1973] suggested that at island arcs, isotherms are shallower under the continent than under the ocean. If this is the case, and temperature is the controlling factor on conductivity, there should be a decreased, or even a reversed, GCE at island arcs. This is rarely observed but may explain the situation in Peru.

Little is known of the conductivity structure beneath ocean ridges. Observations on Iceland indicate a 5-km-thick good conductor (0.07 mho m^{-1}) sloping downward from the spreading axis to a depth of 20 km [*Beblo,* 1978]. *Bennett et al.* [1978] suggest a conductivity of as much as 0.1 mho m^{-1} at a depth of 50 km beneath the African rift system.

Coode and Tozer [1965] and also *Lambert and Caner* [1965] have pointed out that the seismic low-velocity zone (LVZ) is shallower under the oceans by about 70 km and have suggested that it might be a zone of abnormal conductivity as well as of abnormal elastic properties. The importance of partial melting as a control on conductivity supports this idea. However, depth resolution so far has not been good enough to link a conductivity change with the LVZ unamiguously. The ocean floor measurements of *Cox et al.* [1970] require a suboceanic conductivity 2 orders of magnitude greater than the continental conductivity at a depth of 40 km with a sharp transition at the continental slope. Pore water is unable to account for such an increase, and they suggest a temperature of 2000°K at 30 km under the ocean floor. This implies heat transfer by convection in the mantle.

The rate at which the GCE decreases inland appears to be connected with the geology of the adjacent continent [*Cochrane and Hyndman,* 1970]. An example of one extreme is the mobile belt of western Canada, where the GCE persists for only a short distance inland. The lower crust there is thought to be hydrated [*Caner,* 1970]. The other extreme is the shield of western Australia, where the GCE persists far inland and magnetotelluric results indicate low conductivity throughout the crust and upper mantle [*Everett and Hyndman,* 1967*b*].

Perhaps the most interesting regions are those at which the almost ubiquitous GCE is absent or abnormal. *Gough* [1973*b*] has discussed the situation in southern Peru and Japan. The presence of an inland subduction zone in Peru is undoubtedly associated with the absence of a normal GCE there. However, a complete explanation must include the reason that the GCE is normal on the coast of Chile [*Aldrich,* 1972], where there is also a subduction zone. The situation in Japan is much more complicated, with possibly three subduction slabs involved. The explanation of the Japanese anomaly probably involves a wedge-shaped insulator [*Rikitake and Honkura,* 1973]. A subduction zone may be responsible for the high transfer function at Dili (Timor) [*Chamalaun and White,* 1975], but observations in this region are still scarce. The absence of a GCE at Sable Island appears to be due to a shallow conductor of saturated porous rock in the crust [*S. P. Srivastava and A. White,* 1971].

The GCE might be diagnostic of tectonic conditions if, first, its cause in terms of conductivity distributions and, second, the mechanism of abnormal conductivity were unambiguously known. At present, neither of these is the case. However, like other electromagnetic induction studies, it can indicate regions of abnormal physical properties and can contribute to multitechnique studies.

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