Magneto–Telluric Investigations in the Irish Sea and Southern Scotland

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Summary

The magneto–telluric field, in the spectrum 8 s to 100 min, has been recorded at four stations, Eskdalemuir (lat 55°17' N, long 3°11' W), Portpatrick (54°51' N, 5°5' W), Port Erin (54°5' N, 4°45' W) and Nefyn (52°56' N, 4°33' W). Records of $E$ were obtained simultaneously on two short cables orientated N–S and E–W at Eskdalemuir, the $H$ variations being obtained from the observatory records there. At the other stations $E$ was recorded on six cables (9–120 km long), five of which cross part or all of the Irish Sea. The measurements were made on each cable in turn. At each station the component of $H$ orthogonal to the cable was also measured.

Assuming a top layer of resistivity 3000 ohm-metres, the profiles at Eskdalemuir indicate a thickness for this layer of 12 km. Underlying this is a layer of resistivity 45 ohm-metres which extends to a depth of 30 km; beneath this the resistivity is about 2500 ohm-metres. These results do not depend critically upon the resistivity assumed for the top layer.

After corrections have been made for the estimated thickness and resistivity of the sediments in the Irish Sea, the results from five of the six long cables confirm the presence of a low resistivity (10–100 ohm-metres) layer in the middle or lower part of the crust.

On comparison of the magnetic field records with the magnetograms of Eskdalemuir observatory, it was found that, although the shape of the pulsations was similar at all stations, the amplitude ratios between stations showed considerable dependence upon the period of the oscillations. This can be explained largely by anomalies in conductivity near Nefyn and Port Erin. It appears unlikely that sedimentary basins in the Irish Sea can account for more than a small part of the anomalies.

The effect of finite dimensions of the source is discussed; corrections for a possible variation of the field dimension have been made and it is found that these do not materially affect the results.

Possible induction effects of the Atlantic Ocean have been considered and it is concluded that these have been negligible in this survey.

1. Introduction

The magneto–telluric method has been widely used in recent years (Niblett & Sayn-Wittgenstein 1960, Cantwell & Madden 1960, Rokityanski 1961, Fournier 1963, and others) to investigate the resistivity of the crust and upper mantle. Cagniard (1953) has developed the simple theory of the method for a uniform, oscillating field parallel
to the surface of the Earth, while Price (1962) has treated the more general case when
the source field is considered. In brief, the apparent resistivity $\rho_a$ is calculated from
the ratio of the amplitudes of simultaneous electric and magnetic disturbances. The
profile of $\rho_a$ as a function of the period $T$ of the field may then be compared with
appropriate theoretical standard curves to find the resistivity of each layer and the
depth of each interface.

These measurements have assumed particular importance in view of the depend-
ence of resistivity upon temperature and pressure (Coster 1948, Parkhomenko &
Bondarenko 1963, and others), and the possibility that some of the values obtained
so far, e.g. in Northern Germany (Bartels 1957) and in Northern Canada (Whitham
1963) are due to zones of high temperature in the crust or upper mantle.

The present investigation arose from examining the records of Bowden & Hughes
(1961), who had been measuring the flow of water in the Irish Sea. These records
of the electric potential at two cables (Nefyn–Howth and Portpatrick–Donaghadee,
see Fig. 1) showed large, tidally-generated fluctuations with smaller, short-period dis-
turbances superimposed. Bowden & Hughes pointed out that the latter disturbances
were due to Earth currents and were often similar in shape on the two cables. This
similarity also frequently extended to the magnetic field recorded at the Eskdalemuir,
Abinger, Hartland and Valentia observatories.

This suggested that there was a widespread uniformity of the telluric currents;
Bowden & Hughes then kindly put at our disposal their records, obtained over several
years, for analysis of the $E$ variations. The magnetic records of Eskdalemuir observa-
tory were used to obtain the variations in $H$. The results of this analysis were encourag-
ing, yielding estimates of the resistivity at depths up to about 50 km. These slow-run
records, however, limited the measurable spectrum of the electrical variations to
periods greater than 5 min, so after this trial analysis we extended the observations
to shorter periods and made measurements on more cables (Fig. 1). The component
of $H$ orthogonal to the cable was also measured, at one end of each cable, in order
that the crust and upper mantle should be sampled more nearly in the same place,
for $E$ and $H$.

Except at Eskdalemuir, $E$ (and the corresponding $H$ component) was measured in
only one direction at any one time. Determination of anisotropy was therefore
impossible (even at Eskdalemuir, where there were insufficient observations).
Although the spectrum was extended, the lower limit of the periods measured now
being 8 s, this was insufficient to find the resistivities of near-surface layers (the top
few km); these therefore have had to be assumed.

Some inherent limitations of the method may also be mentioned briefly. In
common with other electrical methods there is, in practice, ambiguity in the interpreta-
tion, even with the widest spectrum of frequencies, so that it is not possible to find a
unique resistivity–depth relation to fit the data. Cagniard’s theory is strictly applicable
only to a laterally uniform current sheet; this restriction is considered later. The
effect of the finite dimensions of the source is discussed in section 8; this can be import-
ant, but the theory is not yet sufficiently developed for general applicability to magneto-
telluric surveys and Cagniard’s method is used in our interpretation.

2. Apparatus

(a) Recording of the electric field

The observations in the Irish Sea were made on telephone cables (shown in Fig. 1)
which were kindly put at our disposal by the Post Office. The recording gear was
housed in the telephone repeater stations at Nefyn, Port Erin or Portpatrick, recordings
being made on each cable in turn. At the latter station recordings were also made
on the overland cable to Stranraer. An attempt was made to record on a cable from
Port Erin to Blackpool, but this proved impossible owing to high frequency noise.
Fig. 1. (a) Locations of cables and recording stations.
(b) Locations of observatories.
The methods of tapping the cables were identical with those used by Bowden & Hughes. At Eskdalemuir two cables belonging to the Atomic Weapon Research Establishment were used. These were 0.6 km in length, orientated N–S and E–W, and connected to Earth by steel plates 2 ft square.

The longer period fluctuations (greater than 3 min) were obtained in the manner of Bowden & Hughes, on a siphon-pen recording milliammeter, no amplification or filtering being required except at Eskdalemuir, where the variations were d.c. amplified. The short period oscillations were recorded on a modified Willmore seismic recorder, with a filter to exclude periods longer than 5 min; the natural period of the galvanometers being ½ s, the response was flat down to about ½ s. No amplification was necessary.

(b) Recording of the magnetic field

At each station the horizontal component of the magnetic field orthogonal to the direction of the cable was observed on an Askania Gf6 field balance, the eye-piece of the balance being replaced by a split photo-cell. The output from this cell was recorded directly on an Askania chopper-bar recorder for the long-period variations, a dot being put on the chart every 30 s. No filtering or amplification was required, since the longer periods usually had a much greater amplitude than the short periods. For the short-period variations, the output was amplified, filtered and recorded alongside the electric field.

The magnetometer, which had a natural period of 4 s, was 0.7 critically damped by sending a part of the amplified photocell output through a Helmholtz coil attached to the instrument. The response was thus flat for periods above 8 s, this setting the lower limit to the periods that could be analysed in this survey.

At Eskdalemuir the routine records obtained by the observatory were available throughout the spectrum 8 s to 1 h, for all components. Our investigations are therefore most complete at this station. When analysing the results from the other stations the Eskdalemuir magnetic records were also used, as explained in section 4b.

The spectra observed on the two sets of recorders overlapped between 3 and 5 min and at most stations there were enough fluctuations at these periods to ensure that the recorders were giving mutually consistent results.

All recorders were calibrated at the beginning and end of each record.
(c) Duration of recordings

Approximately the same amount of data was collected from each cable (see Fig. 2 for total data collected). For some cables sufficient data was obtained in a week of continuous recording. Longer recording, up to 4 weeks, was necessary during quiet periods and when the sensitivity of the apparatus was less, as at Eskdalemuir.

3. Nature of the records

Magnetic variations greater than about 1 γ were distinctly recorded, while telluric fluctuations as low as 0.1 mV/km could sometimes be detected, owing to the length of some of the cables. Occasionally there was ‘hum’ or other artificial disturbance on the magnetic records due to traffic, machinery, etc.; this was never seen on the telluric records, which thus assisted in the identification of true variations. The disturbances of the natural field were neglected if there was any suspicion of extraneous interference.

Most of the recorded variations were of Pt type (Jacobs & Westphal 1964), although some Pc and Pg as well as some isolated cycles were also recorded. The duration of each train of disturbance varied from a few to many tens of cycles. The amplitude and the period, as well as the ratio E/H often varied from cycle to cycle; each cycle was therefore considered separately. The periods of the recorded oscillations ranged from a few seconds to isolated waves of a few thousand seconds, but the majority fell into groups of 15–120 seconds and 300–1200 seconds (Fig. 2). There was a remarkable absence of activity in the period range of 150–250 seconds. The magnetic and telluric variations usually showed good correlation, but there was considerable scatter in the ratio of their amplitudes and in the difference of their phases, the latter sometimes exhibiting reverse polarity.

Seven very clear instances of ‘pearls’ were observed in the telluric field, the corresponding magnetic variations being too small to be detected. Six of these occurrences were in early morning hours, the seventh being after 21.00 h G.M.T. The frequency varied from 0.5 to 4 c/s and the amplitude from 0.2 to 0.5 mV/km. The interval between successive pearls ranged from a half to two minutes and the duration from ten seconds to two and a half hours. The pearls occurred with or without Pt’s of periods 30–60 seconds and of amplitude about five times that of the pearls.

4. Analysis of the data

(a) Determination of E/H

No attempt at harmonic analysis was made, but instances were selected on the charts when both E and H traces exhibited a fairly good sinusoidal wave-form. The ratio of amplitudes was measured, for each cycle, the values meaned for each frequency and the apparent resistivity obtained from Cagniard’s formula

$$\rho_a = 0.2T(E/H)^2,$$

where $\rho_a$ is in ohm-metres, $T$ in seconds, $E$ in mV/km and $H$ in gammas.

Coherency between E and H was tested statistically for some cases by finding the correlation coefficient between their amplitudes. Thus 120 events were selected in the range 28–32 s for the Port Erin–Ballyhornan cable which appears to be a typical example. The scatter was such that, for a given value of $H$ (the limits of $H$ being 1 and 10 γ), $E$ typically varied over the range 0.5–2.0 mV/km. However, despite this large scatter, a good correlation was found between $E$ and $H$, the correlation coefficient being 0.26 for $H < 5 \gamma$, and 0.38 for all points. These values are significant at 2% and 1% respectively. This suggests that there is no serious decrease in the coherency for the smaller amplitudes, as would be the case if ‘noisy’ events had been picked.
It would be possible to fit a line, passing through the origin, to a plot of \( E \) vs \( H \); in this way a value of \( E/H \) could be obtained. We have not adopted this approach since the larger values of \( E \) and \( H \) would then be given greater weight; this seems unjustifiable in the light of the correlation tests described above, and in any case is undesirable since the polarization or other characteristics of the variations may be correlated with amplitude.

The values of \( E/H \) were therefore simply meaned, for each frequency. The error of the mean naturally depended upon the amount of data collected at that frequency, but otherwise was found to be similar at all frequencies. Thus the scatter of \( E/H \), and therefore of \( \rho_a \), is independent of the period as well as the amplitude of the oscillations. In order to avoid congestion in Fig. 5, the standard error of the mean for each value of \( \rho_a \) is not shown, but may be judged from the scatter of the points.

The chief causes of the scatter are thought to be the anisotropy and the inhomogeneity of near-surface formations, the changes in the polarization ellipse of the inducing magnetic field and, for the longer periods in the Eskdalemuir profile, the dimensions of the source of the field. Much more work will have to be done to evaluate the relative importance of these causes.

(b) Lateral variations in the field

Cagniard's theory assumes a uniform horizontal current sheet, which is an idealization seldom realized in practice. Irregularities in the crust and upper mantle must imply non-uniformity in the current flow. It was thus desirable to test this assumption. In addition, the geometry of the cables and recording sites was peculiar to the experiment. Except at Eskdalemuir, the cables were some tens of kilometres in length and

![Image of histograms](https://academic.oup.com/gji/article-abstract/12/2/165/620841)
therefore a measurement of \( E \) effectively samples a large part of the crust and upper mantle. \( H \), by contrast, was measured at only one end of each cable. Some assurance is needed that \( H \) varied sufficiently slowly with horizontal distance that it could be matched with the value of \( E \).

A further point is that irregularities in the current sheet might have been largely smoothed out over the length of the cables. This possibility can be rejected on account of the good correspondence in wave-form usually found between \( E \) and \( H \). Moreover the scatter in \( E/H \) was found to be of the same order at Eskdalemuir as in the other profiles. It therefore seems that the length of the cables was no serious disadvantage in this exploratory survey.

The general uniformity of the current sheet over the whole area was tested for many events by comparing \( \frac{H_{sta}}{H_{Esk}} \), the value recorded at the station Portpatrick, Port Erin or Nefyn with the simultaneous value \( H_{Esk} \) recorded at Eskdalemuir, for which records were always available. From the N–S and E–W components there, the component \( H_{Esk} \) parallel to \( H_{sta} \) (i.e. at right angles to the cable) was computed. Owing to the nature of the Eskdalemuir records, this procedure could not be used for the short periods; in these cases the larger component at Eskdalemuir was measured. Except when there was a significant vertical component (when the event was in any case rejected), all simultaneous variations in \( H \) agreed well in period, duration and shape.

The ratio \( \frac{H_{sta}}{H_{Esk}} \), however, varied considerably at some of the stations. Fig. 3, compiled from some 2000 events, shows the scatter in this parameter for each station. As might be expected, this is least at Portpatrick, the station closest to Eskdalemuir. Here \( \frac{H_{sta}}{H_{Esk}} \) lies nearly always within the limits 0.7 and 1.5, the mean value being 1.00 for short periods and 1.11 for long periods, for both orientations of the magnetometer at Portpatrick. For the other stations the scatter is much larger and the mean is no longer 1.00 for short periods, though still close to 1.00 for long periods.
Fig. 4 shows the variation of $H_{stn}/H_{Esk}$ with period $T$, at each station. It can be concluded that, since the values of $H$ at Eskdalemuir and Portpatrick are very similar, Cagniard's theory can be applied to the cables at these stations with some confidence. At Port Erin and Nefyn, however, particularly at the latter station, the simple theory may be invalid at the shorter periods.

**Fig. 5.** The magneto–telluric profiles. The first place name is that of the magnetic station. ▲, Eskdalemuir, magnetic N–S; ▼, Eskdalemuir, magnetic E–W; ◆, Portpatrick–Stranraer; ○, Portpatrick–Stranraer, with June data; ▼, Port Erin–Ballyhornan; △, Port Erin–Cemaes Bay; □, Portpatrick–Donaghadee; ■, Portpatrick–Whitehead; ○, Nefyn–Howth; ———, Model A; ——–, Model B; ······, Model C.
5. The magneto-telluric profiles and their interpretation

Fig. 5 shows the values of $\rho_a$, plotted against $T$, for all profiles. These fall into three groups, Eskdalemuir, Portpatrick–Stranraer and the Irish Sea cables. We consider each group in turn.

(a) Eskdalemuir profiles

The two profiles are not identical, showing that the medium is anisotropic. Unfortunately, very few oscillations were recorded simultaneously in both components and it is not possible to evaluate the anisotropy factor.

The observed spectrum does not extend to sufficiently low periods to enable the surface resistivity to be found. However, the rocks in the top few kilometres, being of Palaeozoic or older age, possibly with granite intrusions, fall into the class of rocks examined experimentally by Brace, Orange & Madden (1965). They found that saturated rocks such as quartzite and granite have a resistivity of order 300 ohm-metres under zero pressure, the resistivity increasing rapidly with pressure.

Using Yungul's (1961) curves, three-layer models have therefore been fitted to the Eskdalemuir N–S profile (which more closely resembles his curves than the E–W profile), assuming a top layer of at least 300 ohm-metres resistivity. This can be done in many ways, but in every case it is necessary to postulate a middle layer of low resistivity. Depending upon the model chosen, the resistivity of the top layer ranges between 300 and 2000 ohm-metres, that of the middle layer between 8 and 88 ohm-metres and the thickness of this middle layer between 3 and 40 km. The depth to the lower interface lies between 18 and 44 km, beneath which the resistivity is high (greater than 1000 ohm-metres).

If the two profiles are combined, by taking the geometric mean of $\rho_a$ for each period, a model which fits the data yields again a low-resistivity middle layer. It is

![Fig. 6. The data for three of the profiles, corrected for anomalous values of $H_{sta}/H_{Esk}$. Symbols as in Fig. 5.](https://academic.oup.com/gji/article-abstract/12/2/165/620841/02 March 2019)
shown in Fig. 7 (model A). The resistivity of the top layer has here been taken as 3000 ohm-metres, a value close to the average for the top 10 km suggested by the data of Bruce et al., though as the rocks are presumably not saturated with brine, the resistivity may be even higher.

(b) Portpatrick–Stranraer profile

As expected, this profile is intermediate between Eskdalemuir and the other groups, the cable being overland, but with its ends near the sea. Salt-water infiltration therefore lowers the resistivity at short periods (small depths of penetration). The shape of the profile, in the range 40–200s, depends on whether preliminary data, collected over this part of the spectrum in 1963 June, is included or not. If not, there is clear evidence of a four-layer structure, shown in Fig. 7. With all data included the curve is flatter in this range, and it is then possible to fit a three-layer model, also shown. In either case it is found that there is a layer of low resistivity (40–70 ohm-metres) in at least the lower part of the crust. For this profile \( H_{\text{Esk}} \) was used to obtain \( \rho_b \).

(c) Irish Sea profiles

We first consider this group as a whole. In these profiles the resistivity is much less than for the other two groups, and it is necessary to allow for the effects of the sea and of porous sedimentary rocks. The average depth of the sea in the North Channel is 14 metres, while for the rest of the Irish Sea it is only about 60 metres. It can easily be shown that the sea itself has little effect for the periods considered here, merely decreasing \( \rho_b \) by about 10–15% for all periods.

The resistivity of saturated porous rocks has been estimated by Wyllie (1963) as \( \rho_0/\eta_p^2 \) where \( \eta_p \) is the porosity and \( \rho_0 \) is the resistivity of the fluid (0.25 ohm-metres for sea water). There is at present little information on the thickness of sedimentary basins in the Irish Sea, but plausible estimates may be made from gravity and seismic surveys (Bott 1964, Griffiths, King & Wilson 1961). Thus Bott suggests that the Manx–Furness basin may have a maximum depth of 2.7 km, if the density contrast between younger, porous rocks and Palaeozoic rocks is 0.4 g/cm\(^3\). This density contrast corresponds to a porosity of about 23%, which gives a resistivity of 5 ohm-metres. Near-surface sediments (the Layer 1 of Griffiths et al.) may have a density of about 2.1 g/cm\(^3\), a porosity of 35% and a resistivity of about 2 ohm-metres.
A Layer 1 of resistivity 1 ohm-metre, comprising the sea and porous sediments, underlain by a Layer 2 of resistivity 9 ohm-metres (typical of Triassic rocks), each layer of 1 km thickness, is therefore suggested. These figures, which are arithmetically convenient, probably given an upper limit for the average thickness of porous rocks over the length of any of the cables. Beneath these layers, rocks may be expected to have, for a few kilometres, the high resistivity of those discussed earlier.

It is now possible to compare the Irish Sea data with the other profiles. Many different models can be fitted to each of the profiles; the effect of a Layer 3, corresponding to the top layer at Eskdalemuir (300–3000 ohm-metres) is probably masked, although its presence is suggested by the Portpatrick–Whitehead data. The specimens of Brace et al. have a resistivity of at least 900 ohm-metres at 2 km depth. This value is therefore chosen for model B, the resulting plot (Fig. 5) yielding a curve which lies almost everywhere above the plotted data for the Irish Sea profiles. This shows that, for a considerable part of the crust under all the Irish Sea cables, the resistivity must lie well below this value.

The observed spectrum does not extend to sufficiently long periods to establish with certainty the existence of a high resistivity layer at a depth of order 35 km, as found at Eskdalemuir, but models which fit the Portpatrick–Whitehead data tolerably well and are consistent with the resistivity of surface sediments are shown in Fig. 7. As shown in section 4b, the Cagniard theory is applicable for this cable. In each of these models there is a middle layer of resistivity similar to that found at Eskdalemuir.

6. The anomalous profiles and their correction

Fig. 5 shows that all the Irish Sea profiles, except Portpatrick–Whitehead, have values of $\rho_a$ much less than those of model B, especially at the short periods where in some cases the resistivity is even less than that of salt-water. Another model, C, has therefore been constructed, in which the 900 ohm-metre layer of model B is replaced by a layer similar to the low-resistivity layer found on the land profiles. The model has a top layer of 0.1 km thickness (an upper limit) of sea water ($\rho = 0.25$ ohm-metres), underlain by 0.9 km of porous sediment ($\rho = 2$ ohm-metres), underlain by a semi-infinite medium of $\rho = 9$ ohm-metres. It is seen that the data for $\rho_a$ still fall well below this curve for periods less than about 50 s on three of the profiles and for periods less than about 300 s on the Portpatrick–Whitehead data.

These are the period ranges for which $H_{\text{sta}}/H_{\text{Esk}}$ is anomalously high (Fig. 4) on the Port Erin and Nefyn profiles. If $H_{\text{sta}}$ is replaced by $H_{\text{Esk}}$, on the assumption that the field at Eskdalemuir is normal and the fields at Port Erin and Nefyn anomalous, the re-calculated values of $\rho_a$ for these profiles now fall much closer to models B and C (Fig. 6). Even after this correction has been made it is clear that a large part of the crust has a resistivity considerably lower than 900 ohm-metres.

The profile for Portpatrick–Donaghadee cannot be corrected in this way, and here the low values of $\rho_a$ at short periods remain unexplained. However, at long periods this profile agrees well with Portpatrick–Whitehead. After all corrections have been made, this profile and Port Erin–Ballyhornan still have low values of $\rho_a$ for $T<40$ s; this suggests an anomalous zone near Ballyhornan and Donaghadee, but this is conjectural at present.

7. The anomalies in $H$ at Nefyn and Port Erin

Since the Nefyn and Port Erin profiles can be suitably corrected by changing $H$ alone, it follows that the $E$ variations are probably normal, the conductivity being anomalous only near the ends of the cables where $H$ was measured.
The possibility that these anomalies are due to unusually thick or porous sedimentary basins close to Nefyn and Port Erin is now considered. Griffiths et al. (1961) and Blundell, King & Wilson (1964) have shown that there is some 2 km thickness of sedimentary rocks, probably post-Palaeozoic, in Tremadoc Bay. Even if these rocks have an unusually low resistivity, it is hard to see how they could account for the anomaly at Nefyn, since the edge of this sedimentary basin is some 15 km from Nefyn. Assuming a uniform current sheet flowing at an average depth of 1 km, it is easily shown that the horizontal component of the magnetic field from a sheet at this distance would be less than 1/50 of the field from a similar sheet underlying the magnetic station. There is less information about the thickness of sediments to the west of Nefyn, but the aeromagnetic map of this area (H.M. Geological Survey 1964) shows many sharp anomalies, suggesting that the porous sedimentary rocks here are not so thick as in the Manx–Furness basin or in Tremadoc Bay. Considerable allowance for porous sediments has already been made in models B and C (in which the layers are taken as extending under the station), and it seems unlikely that the anomalously high values of $H$ at Nefyn can be due to sedimentary basins alone.

At Port Erin the station is close to large expanses of the Irish Sea and the anomaly is smaller than at Nefyn. However, the Bouguer anomalies found by Bott to the west of the Isle of Man are nearly all greater than +20 mgals, with little suggestion of a thick sedimentary basin here. The centre of the Manx–Furness basin is 60 km to the ENE. Despite the lack of information for the area to the south of the station it seems that here also it is unlikely that the anomaly in $H$ at short periods is entirely due to porous sediments.

The maximum value of $H_{\text{stat}}/H_{\text{Esk}}$ for Nefyn is 5-2, occurring at $T=20$ and $T=30$ s. This value is higher than, for example, the anomaly at Alert in N. Canada discussed by Whitham & Andersen (1962), which has a maximum value of about 1-8 occurring, like ours, at the shortest periods in their spectrum. However, these periods are 10 min–1 h at Alert and if the anomalies are compared at these periods it is seen that $H_{\text{stat}}/H_{\text{Esk}}$ is only $1.1-1.5$.

Whitham & Andersen discuss a model consisting of a horizontal conducting cylinder of infinite length embedded in a non-conducting medium. It can be shown that for this model the maximum value of the anomaly $H_{\text{stat}}/H_{\text{normal}}$ occurs when the cylinder is touching the surface of the Earth and when the station is directly above the axis of the cylinder; the anomaly is then approximately equal to 2.

It seems, therefore, that the anomalous zone at Nefyn cannot be modelled by a cylinder. The size and/or conductivity is not so large as that at Alert, on account of the different frequency response. There is no geological evidence for a conducting body near the surface at Nefyn.

The frequency response for the anomaly at Port Erin is not so well marked, and further speculation is unprofitable until more measurements have been made at both localities.

8. The effect of the source field

Price has considered the general case of magneto–telluric induction and has introduced a parameter $v$, where $2\pi/v$ may be taken as a measure of the linear dimensions of the source field. Cagniard’s theory applies only to the special case when $v=0$, that is when the field has an infinite wave-length. Price suggests that $v$ is most likely to lie between $1.57 \times 10^{-9}$ cm$^{-1}$ (with a wave-length equal to the circumference of the Earth) and $1.57 \times 10^{-8}$ cm$^{-1}$, the wave-length then being 4 times the height of the ionospheric currents. Using this more general inducing field he has derived the expression

$$E/\omega H = \theta^{-1} \cdot [\theta + v + (\theta - v) e^{-2\theta D}] / [\theta + v - (\theta - v) e^{-2\theta D}],$$
where \( \theta^2 = v^2 + 4\pi i w \sigma \) for a model in which the conductivity has a constant value \( \sigma \) down to depth \( D \) and is zero below this depth. He has calculated values for \( E/wH \), with \( v \) and \( D \) as variables, in the case when \( \sigma = 10^{-14} \text{ e.m.u.} \) and \( T = 2\pi / w = 100 \text{ s} \).

The quantity \( \theta E/i w H \) is non-dimensional and the calculations can thus be extended to other values of the parameters if \( \sqrt{(\sigma/T)} \) and \( v \) are changed by a factor \( k \) while \( D \) is changed by the factor \( 1/k \).

### Table 1

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<thead>
<tr>
<th>Period (s)</th>
<th>Ratio</th>
<th>Layering and Conductivity</th>
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<td>10</td>
<td>0.00</td>
<td>10^{-10} e.m.u.</td>
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<tr>
<td>100</td>
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<td>3160</td>
<td>2.00</td>
<td>(1 ( \Omega )-m)</td>
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</table>

Table 1 shows three special cases, calculated in this way, in which the parameters resemble those found in this survey. The first two models are simplifications of the deeper layering, while the third represents 1 km of sediments in the Irish Sea. \( E/wH \) has been computed, at certain periods, for the two likely limits of \( v \), \( v = 0 \) and \( v = 1.57 \times 10^{-7} \text{ cm}^{-1} \). It is found that increasing \( v \) leads, in general, to a decrease in \( E/wH \). The table shows the ratio of the maximum to the minimum values of \( E/wH \) corresponding to the two limiting values of \( v \). It is seen that an increase in \( v \) from 0 to \( 1.57 \times 10^{-7} \text{ cm}^{-1} \) leads to a large decrease in \( E/wH \) only in models I and III, at the longest periods. If, however, \( v \) has approached \( 1.57 \times 10^{-7} \text{ cm}^{-1} \) for any of the fields in this survey, the value of \( E/wH \) and thus of \( \rho_a \) will have been reduced at these longer periods, by one or more orders of magnitude. If \( v \) has varied widely over the range \( v = 0 \) to \( v = 1.57 \times 10^{-7} \text{ cm}^{-1} \) this will have increased the scatter in the data. The data have therefore been re-examined and there is some evidence of a larger scatter in \( E/wH \) for periods at Eskdalemuir greater than about 300 s. For the other profiles the scatter does not increase with period and these profiles are thought to be unaffected by the source dimensions.

Corrections have been made to the data for Eskdalemuir, using model I, in which the depth-integrated conductivity is close to that found in section 5. We assume that \( E/wH \) has, at each period, a value equal to the geometric mean of its limits; this entails multiplying the values of \( \rho_a \) by the figures shown in the table. The values of \( \rho_a \) are thus normalized to the case \( v = 0 \), when Cagniard’s theory is applicable. The corrections for \( T < 100 \text{ s} \) are negligible, but the effect of the larger corrections is to over-steepen the long-period branch of the profile beyond the theoretical maximum slope. It thus appears that these corrections are too great. However, the plotted data can now be fitted to Yungul’s curves (which assume a deep layer of infinite resistivity) more easily than before. If it is assumed that the thicknesses and resistivities of the top two layers are in approximately the same ratios as before, it is found that the top layer has a resistivity of 1600 ohm-metres and a thickness of 14 km, and is underlain
by a layer of resistivity 40 ohm-metres and thickness 13 km (model A', Fig. 7). The depth to the lower interface is 27 km, agreeing closely with model A.

If similar corrections are made to the data for the Irish Sea profiles, using the values suggested by model III in Table 1, it is found that the plotted data (Fig. 5) still fall below the curve for model B for all periods up to 2000 s. Thus, even if these corrections should be applied, the conclusions reached in section 5 remain unaltered. It is inferred that, in this survey, Cagniard's theory is applicable without serious error.

9. Coastal effects

Ashour (1965) has calculated the coast-line effect on rapid geomagnetic variations, assuming an inducing field either parallel or normal to the axis of a hemispherical ocean. For measurements on land, near the coast, the induced horizontal field is greatest when the inducing field is vertical. This case can be ignored for our purposes, since in these experiments events which contained a significant component of $Z$ were excluded. When the inducing field is horizontal, Ashour finds that the induced horizontal field near the coast is about 0.25 of an inducing field which is at right angles to the coast and 0.1 of an inducing field which is parallel to the coast. These figures can be regarded as upper limits, since the effect is reduced when the Earth's core is taken into account.

Records from Eskdalemuir, Valentia, Hartland and Abinger observatories (Fig. 1) have been examined for events at the long-period end of our spectrum, at which the effect of the Atlantic Ocean may be expected to be greatest. A total of about 280 comparisons were made between two or more of these observatories, as and when records were available; the periods ranged between 5 min and 1 h. The shape and duration of most of the events was similar at all four stations and there was no pronounced increase in amplitude with proximity to the ocean. Simultaneous events tended to have amplitudes that were equal at Hartland and Valentia, these being 2/3 of the amplitude at Eskdalemuir and 1 1/2 times that at Abinger.

Since the cables and stations used in the present survey are at similar distances (250-400 km) from the edge of the continental shelf it is concluded that the Atlantic Ocean has had no noticeable effect on our observations.

10. Conclusions

Variations in $H$ have been compared at the stations occupied in this survey. These suggest that Cagniard's theory is applicable to the data obtained in the northern part of the area examined and especially at Eskdalemuir. At this station assumptions about the resistivity of the top few kilometres are not critical and it seems clear that much of the middle and lower parts of the crust there must have a low resistivity (order of 10-100 ohm-metres). This conclusion is supported by the data from all the other profiles, except perhaps that at Nefyn.

At a depth of the order of 30 km the resistivity at Eskdalemuir increases to several thousand ohm-metres. The thickness of the crust under the Irish Sea and Southern Scotland has been found by seismic methods (Agger & Carpenter 1964) and it varies between 22 and 35 km. It is therefore possible that the large increase in resistivity found in our survey at about this depth may be associated with the Mohorovicic Discontinuity ('Moho'). Despite some twelve weeks of recording time there were insufficient events of large enough period (and therefore of depth of penetration) to detect the zone of low conductivity which has been found at depths of 70-150 km by many workers (Lahiri & Price 1939, Srivastava 1963, etc.).

The rather low resistivity found by us in the middle or lower parts of the crust has also been found elsewhere (Whitham & Andersen 1962, Whitham 1963) although most workers using the magneto–telluric method have found much higher resistivities.
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in the crust. The temperatures necessary to cause these anomalies by ionic semi-conduction are very high (about 800°C, Coster 1948), but no heat flow measurements appear to have been made in the area of the present survey (Benfield 1939). Runcorn (1955) has suggested that impurity semi-conduction is important in the crust and this process, with perhaps some local temperature variation, appears to be the best explanation for these anomalies in the crust. The resistivity found by us at the top of the mantle agrees well with the results of other magneto-telluric studies. The anomalous conductivities found at Nefyn and Port Erin await further investigation.

The effect of the finite dimensions of the source has been discussed and, although Price's two-layer model is a simplification, it is thought that possible changes in the source parameter are unlikely to have affected materially the main results of our survey. In future work, however, the source effect should be investigated more thoroughly, possibly with the method suggested by Price. He has shown that if the phase difference, as well as the amplitudes of $E$ and $H$, are measured, it is in principle possible to determine both $\nu$ and $\sigma(z)$.

Although variations in $E$ and $H$ were recorded in this survey at periods ranging down to 8 s and covering 8 or 9 octaves, this spectrum should be extended still further, by recording down to 0.1 s, if near-surface resistivities are to be obtained. It is also desirable to use two orthogonal cables and to record all three magnetic components.

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References


