J. Geomag. Geoelectr., 45, 403–422, 1993

Magnetic Signals from an Ocean Eddy

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(Received September 30, 1992; Revised February 10, 1993)

The movement of sea water in the magnetic field of Earth causes a process of motional electromagnetic induction. As a result, electric currents flow both in the sea water, and in the material of the seafloor.

There has been long-standing interest in observing the magnetic fields of such motional induction, especially at periods longer than a day; that is, longer than the main ocean tides. Conditions appear to have been favourable for such observations during an experiment which took place in 1983/84 on continental Australia and the floor of the adjacent Tasman Sea (the Tasman Project of Seafloor Magnetotelluric Exploration, or Tasman experiment). The particular circumstances were the passage of an active ocean eddy across a line of seafloor magnetometers and electrometers; the presence on the seafloor of a 1 km pile of porous (and thus electrically-conducting) sediments; and the simultaneous observation of magnetic fluctuations at sites remote from the eddy.

Using remote-reference techniques, the magnetic signals due to the passage of the eddy are clearly identified, and are of some 30 nT in amplitude. These combine with the corresponding seafloor electric field signals, of some 30 μV$\cdot$m^{-1} in amplitude, to give seafloor magnetic to electric signal ratios of 1.14 $\pm$ 0.19 nT$\cdot\mu$V$^{-1}\cdot$m. Applying basic theory for quasi-steady rectilinear flow to this figure gives a seafloor conductance value of 910 $\pm$ 150 S, and so a "leakage value" of 0.06 $\pm$ 0.01. Both these latter figures agree with estimates for the same quantities calculated independently, from knowledge of the Tasman seafloor sedimentary column; an agreement which supports the application of quasi-steady theory to the present case.

The result has an application to oceanography, in allowing an estimate of the integrated velocity, or barotropic flow, above a station where such seafloor electric and magnetic observations are made. The estimate of seafloor conductance, and the implication that conductivity must greatly decrease below the sediments, is a geological result of value, and suggests the similar application of such measurements in other situations. As a case history, there are lessons to be learned from the present instance regarding the operation of ocean-bottom geomagnetic observatories, especially in areas where there are active ocean movements and high seafloor conductances.

1. Introduction

The physical principles of electromagnetic induction by moving media are well established. The ocean is an electrical conductor, and waves, tides and currents cause it to move in the steady magnetic field of Earth. Thus, measurement of the movement of seawater should be possible by observing its motional induction.

However, generally such measurements are difficult, and a major necessary step to enable them to be made has been the development of suitable instrumentation. Recent achievements in

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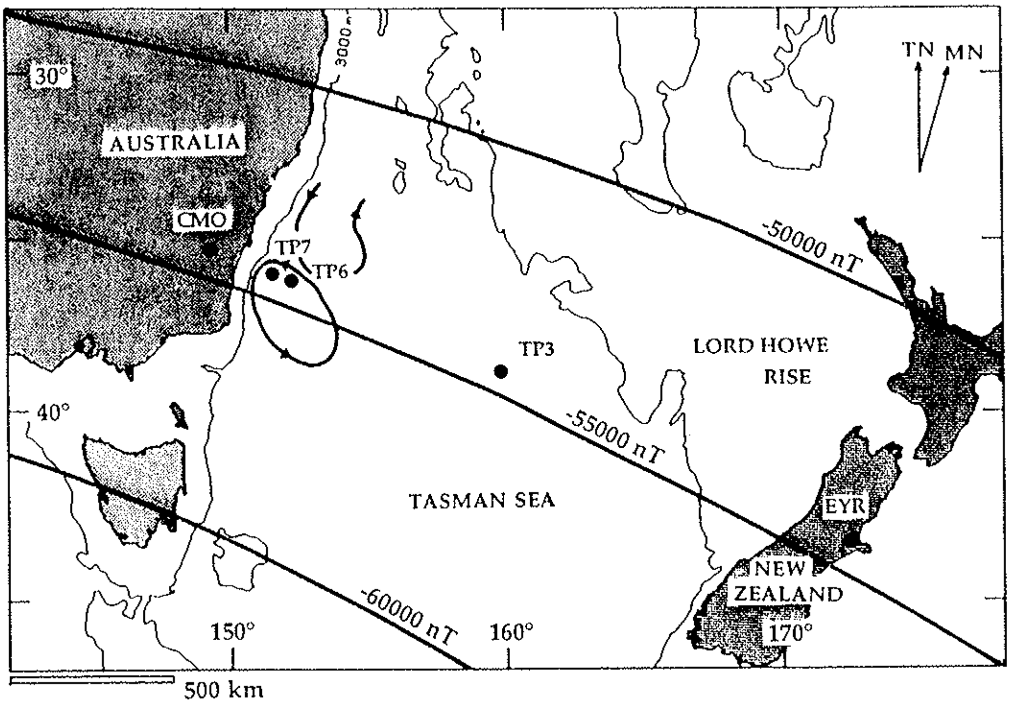

Fig. 1. Map of observing sites 3, 6 and 7 (marked TP3, TP6 and TP7) in the Tasman Sea between Australia and New Zealand, the Canberra Magnetic Observatory (CMO), and the Eyrewell Magnetic Observatory (EYR). Also shown are contours of equal intensity for the vertical component of Earth's magnetic field (from PARKINSON (1983) after LANGEL et al. (1980)). The character of the East Australian Current is shown by streamlines in the vicinity of sites 6 and 7. The loop open to the north indicates (as at mid-December 1983) the meander which developed to form the warm-core ring shown (as at the end of February 1984) by the closed loop (MULHEARN et al., 1988).

this field are reviewed by FILLOUX (1987). As a consequence of these recent achievements, there is now the potential for a major advance in the production of oceanographic information from electric and magnetic measurements (e.g. LARSEN and SANFORD, 1985; CHAVE and FILLOUX, 1985; BINDOFF et al., 1986; LILLEY et al., 1986; BINDOFF, 1988; CHAVE et al., 1989; LUTHER et al., 1991; SEGAWA and TOH, 1992; LARSEN, 1992).

Generally, with the results reported so far, ocean movement at periods longer than the tides has been evident in, and deduced from, measurements of the marine electric field. The associated magnetic signals have been too weak for detection. This paper now presents an example where, due to "leakage" electric currents in the seafloor sediments, the magnetic signals are sufficiently strong to be identified and demonstrated directly.

As a result of determining a seafloor magnetic to electric field ratio, information is obtained also on the seafloor electrical conductance. This information agrees with independent estimates of the conductance based on knowledge of the seafloor sediments.

The particular ocean current is that of a warm-core ring or eddy in the East Australian Current of the Tasman Sea, which is part of the western boundary current of the South Pacific Ocean (HAMON, 1965; NILSSON and CRESSWELL, 1981; OLSEN, 1991). The observations were made during the Tasman Project of Seafloor Magnetotelluric Exploration, or "Tasman experiment" (FILLOUX et al., 1985; LILLEY et al., 1989). During the same period, the Australian Coastal Experiment (FREELAND et al., 1986) made a range of oceanographic and meteorological observations along the region of continental shelf and slope crossed by the Tasman experiment instrument line. The sites relevant to the present paper are shown in Fig. 1.

Different aspects of the Tasman experiment have been described in earlier papers, and the present paper concentrates on a single subject not yet treated: the search for the magnetic field signal of an ocean current. The ocean current is a mesoscale eddy, the development of which as judged from satellite and other evidence is described by MULHEARN *et al.* (1986, 1988), and the electric-field signals of which are described by LILLEY *et al.* (1986); this latter paper also includes a satellite, infra-red, false-colour image of the eddy. The possible presence of such a magnetic field signal in the Tasman experiment data was first suspected at the time of data reduction, and noted by FERGUSON (1985).

2. Theory

The theory required for this paper is basically as developed by SANFORD (1971), following the work of LONGUET-HIGGINS *et al.* (1954). Note that a recent treatment of the theory of motional induction by moving seawater is by CHAVE and LUTHER (1990), and an informative discussion is given by SANFORD *et al.* (1990). The important elements are noted here.

For rectilinear ocean movement as a function of depth only, as indicated in the upper part of Fig. 2, and denoting the ocean velocity $v_y(z)$ by v, the electric current flow $J_x(z)$ by J, the electric field E_x by E, the steady vertical magnetic field B_z by B, the magnetic field $b_y(z)$ due to the motional induction by b, and the local electrical conductivity by σ, then Ohm's law for a moving medium may be expressed

$$J = \sigma(E + vB). \tag{1}$$

The horizontal electric field E is constant from sea surface to seafloor and there is variation of the horizontal electric current J with depth, corresponding to the variation with depth of velocity v. Figure 2 illustrates this situation for two particular hypothetical cases, showing that the electric current in the lower part of the ocean is generally the return current from that driven where the motion is occurring in the upper part of the ocean; and that if the seafloor is electrically conducting, then part of the return electric current flow will occur in the seafloor. This phenomenon is known as "seafloor leakage", and is critical in the occurrence of seafloor magnetic fields.

Then, integrating Eq. (1) over the depth of the water column, gives for the case of uniform ocean conductivity and zero seafloor leakage (case I in Fig. 2)

$$0 = \sigma E \int_{-H}^{0} \mathrm{d}z + \sigma B \int_{-H}^{0} v \mathrm{d}z \tag{2}$$

where integration of the electric current J yields zero due to the requirement of return current balance. Thus using the notation

$$\overline{v} = \frac{1}{H} \int_{-H}^{0} v \mathrm{d}z \tag{3}$$

Eq. (2) above gives

$$\overline{v} = -E/B. \tag{4}$$

In Eq. (3) $\overline{v}$ is the ocean barotropic velocity, so that Eq. (4) is a remarkably simple expression for this quantity, especially as B is well-known everywhere on the surface of Earth (and, for the Tasman Sea, is contoured on Fig. 1).

In the event that the seafloor material is not entirely resistive, so that leakage through it occurs (as in case II of Fig. 2), Eq. (1) is now integrated to the sea surface from the bottom of the return current flow. Again the integral of the electric current will yield zero, and otherwise

$$0 = E \int_{-H-h}^{0} \sigma \mathrm{d}z + \sigma_1 B \int_{-H}^{0} v \mathrm{d}z \tag{5}$$

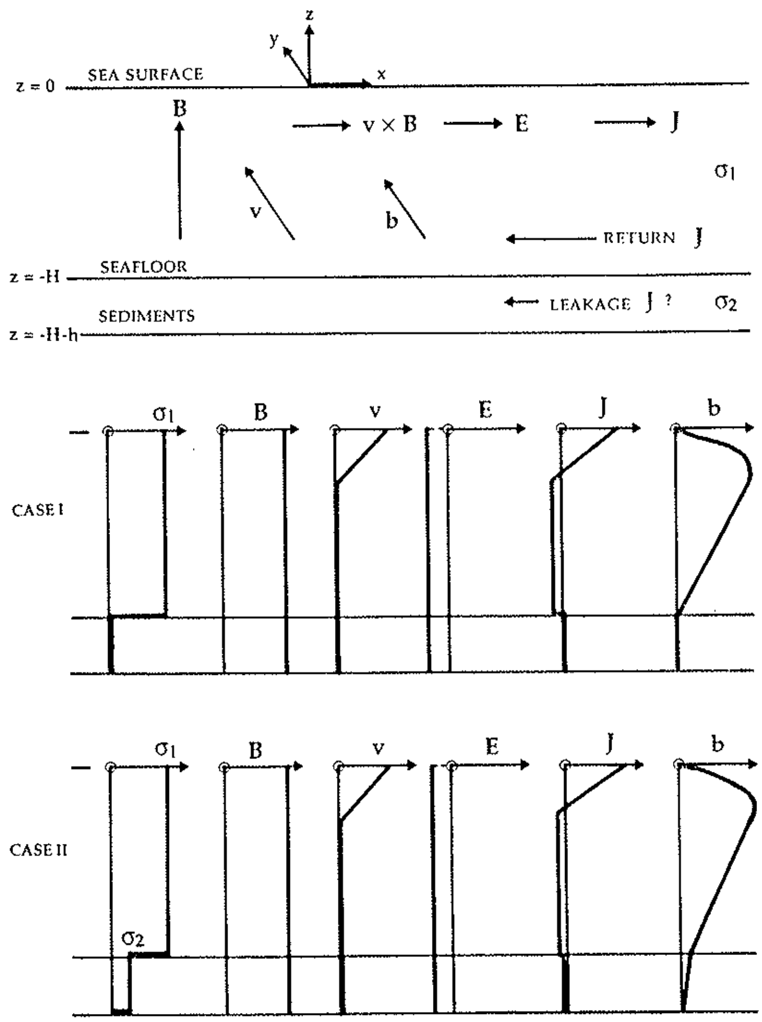

Fig. 2. The geometry and notation for motional induction by seawater, with results for two hypothetical examples. While intended to be qualitatively instructive, the profiles in the figures are in fact plotted to scale. The values of σ_1, σ_2, B and v_{max} are 3.2 S·m^{-1}, 1.0 S·m^{-1}, 55000 nT and 2 m·s^{-1} respectively. For case I, the maximum values attained by E, J and b are -14 μV·m^{-1}, 3.1×10^{-4} A·m^{-2} and 226 nT respectively, and case II is plotted to the same scale. Note that the presence of seafloor conductance in case II causes leakage current in the sediments, and a non-zero seafloor magnetic field.

as v is zero in the seafloor layer. Thus

$$0 = E(h\sigma_2 + H\sigma_1) + \sigma_1 BH\overline{v}$$

so

$$\overline{v} = -\frac{E}{B}(1+\lambda) \tag{6}$$

where

$$\lambda = \frac{h\sigma_2}{H\sigma_1} \tag{7}$$

and a "leakage" term λ, the ratio of the sediment to ocean conductances, enters as a correction in the expression for the barotropic flow. Sometimes (as in CHAVE and LUTHER, 1990) the leakage is quantified by a parameter C, where

$$C = \frac{1}{1+\lambda}. \tag{8}$$

The magnetic field b in Fig. 2, due to the electric current flow J is given generally by

$$b = \frac{1}{2}\mu_o\left(-\int_{-H-h}^{z} J\mathrm{d}z + \int_{z}^{0} J\mathrm{d}z\right) \tag{9}$$

for z in the range $-H-h < z < 0$ (and outside this range it is zero). Here μ_o is the magnetic permeability of seawater, taken as that of free space.

Figure 2 shows, for cases I and II, the profiles of b with depth, thus computed. Note the significant signal in the water column, the character of which is related to the profile of the water velocity.

For the purposes of this paper, note particularly the absence of a seafloor magnetic field b_s in case I, and its presence in case II. This seafloor magnetic field, using Eq. (9), is given by

$$b_s = -\mu_o\sigma_2 hE \tag{10}$$

so that the seafloor conductance, $\sigma_2 h$, is given directly by

$$\sigma_2 h = -b_s/(\mu_o E). \tag{11}$$

Further, the leakage factor λ may now be expressed

$$\lambda = \frac{-b_s}{\mu_o E\sigma_1 H} \tag{12}$$

and this value used in Eq. (6) for estimating barotropic flow.

When the theory is expanded to allow for conductivity variations in the ocean, so that σ may be a function of z, integration of Eq. (1) gives, in place of Eq. (5),

$$0 = E\int_{-H-h}^{0} \sigma\mathrm{d}z + B\int_{-H}^{0} \sigma v\mathrm{d}z \tag{13}$$

so that then a quantity $\overline{v}*$ (using Sanford's notation) is determined by the ratio E/B, that is

$$\overline{v}* = -E/B \tag{14}$$

where

$$\overline{v}* = \frac{\int_{-H}^{0} \sigma v\mathrm{d}z}{\int_{-H-h}^{0} \sigma\mathrm{d}z} \tag{15}$$

is the vertically averaged and sea-water conductivity weighted water velocity. The effect of the conductivity weighting on such determinations of barotropic velocity is discussed by CHAVE and LUTHER (1990) and shown generally to be small.

The matter of conductivity-weighting is not pursued further here as it does not affect the primary subject of this paper, the seafloor magnetic fields caused by motional induction. To see this, expand Eq. (9) for the seafloor

$$\begin{aligned} b_s &= \frac{1}{2}\mu_o\left(-\int_{-H-h}^{-H} (\sigma E + \sigma vB)\mathrm{d}z + \int_{-H}^{0} (\sigma E + \sigma vB)\mathrm{d}z\right) \\ &= \frac{1}{2}\mu_o E\left(-\sigma_2 h + \sigma_1 H + \frac{B}{E}\int_{-H}^{0} \sigma v\mathrm{d}z\right) \end{aligned}$$

where σ_1 and σ_2 are now the depth-averaged ocean and seafloor conductivities,

$$= \frac{1}{2}\mu_o E\left(-\sigma_2 h + \sigma_1 H - \int_{-H-h}^{0} \sigma \mathrm{d}z\right) \quad \text{using Eq. (13)}$$
$$= -\mu_o \sigma_2 h E \tag{16}$$

as before, making the point that the conductivity-weighting of velocity does not affect the b_s/E ratio for seafloor magnetic fields.

For the phenomenon of ocean motional induction, shown idealized in Fig. 2, the electric field E is due to the ocean movement only, and is otherwise zero. There are no other known sources of significant E at periods longer than several days in the ocean, steady or time-varying.

When measuring the magnetic field due to motional induction, however, two additional sources of signal become important. One source is of ionospheric origin, and enters the ocean by electromagnetic induction. The high frequency part of this ionospheric signal is attenuated by the seawater, but the low frequency part penetrates to the seafloor; its removal from the observed data will form the major exercise of Section 5 below.

The other source of spurious signal in measuring the magnetic field b is the steady horizontal component of Earth's main magnetic field. Although Earth's field is known everywhere in three components with sufficient accuracy that the appropriate value of B can be used in Eqs. (4) and (14) without introducing error, the b_s signals expected to be generated according to Eq. (10) are much smaller than the magnitude of Earth's main field, and generally in fact will be less than the accuracy with which the magnitude of Earth's main field is known. Thus the "steady-field" horizontal magnetic component is unlikely to be known sufficiently well at a seafloor site for it to be thus removed from measured data.

A quite different point is that seafloor magnetometers may be variometers, recording relative changes with time of the magnetic field, rather than absolute instruments recording relative to a known zero level.

Thus, with the ionospheric contribution removed, a general datum b_t of the seafloor magnetic field will comprise two components, b_m of the main steady field, and b_s due to motional induction. The separation of b_t into its two components $b_t = b_m + b_s$ is not possible; however two readings at different times, b_t^1 and b_t^2, may be expanded as

$$b_t^1 = b_m + b_s^1,$$
$$b_t^2 = b_m + b_s^2$$

and so may be differenced to give the change Δb_s in b_s,

$$b_t^1 - b_t^2 = b_s^1 - b_s^2 = \Delta b_s.$$

A possibility which is evident from the theory of this section, and which is supported by the conclusions of this paper (but not otherwise taken advantage of here), is that times of zero observed electric field are also times of zero seafloor magnetic field b_s, and so may be used to mark zero levels of this latter quantity. The theory is for steady-state movement of seawater, and changes which cause such zeros must be slow or "quasi-static".

3. Calculation of Ocean and Seafloor Conductances for the Tasman Sea

The electrical conductances of the ocean and of the seafloor which arise in the theory of the preceding section can be calculated for the Tasman Sea, using oceanographic data and the results of extensive marine geological and geophysical studies of the seafloor. Such calculations are carried out in this section.

3.1 Ocean conductance

The ocean conductance value is the integral of the conductivity profile of the water column at a particular site. In the present case a value of 3.27 $S{\cdot}m^{-1}$ has been taken for the average conductivity through the general water column (BULLARD and PARKER, 1970), and this value multiplied by the known water depth at each site (4836 m at site 6, and 4840 m at site 7), to give conductance values of 15800 S for both sites 6 and 7.

3.2 Seafloor conductance

Seafloor conductances are based on knowledge of the sediments on the Tasman seafloor, assuming that beneath the sediments the material is more resistive and effectively an insulator. To estimate the sediment conductance, five steps have been followed.

One-way seismic reflection travel-time values for the sites have been taken from the map of SCHNEIDER and PACKHAM (1983), discussed in SCHNEIDER (1985). The values are 0.55 s for site 6, and 0.6 s for site 7. The sediment type is described as detrital silt and mud (SCHNEIDER and PACKHAM, 1984).

Next, these travel-times have been converted to sediment thicknesses, using the relationships of HAMILTON (1985), and taking into account the sediment type of detrital silt and mud. In particular, the relationship in S.I. units for seismic velocity V in terms of thickness h,

$$V = 1501 + 1.151h$$

gives travel times by integration. This process gives a sediment thickness of 1100 m for site 6, and 1250 m for site 7.

Porosities for the sediment layer have then been calculated using the data of HAMILTON (1976). The porosity values obtained are 42% for site 6, and 40% for site 7. These porosity values are vertical averages since, as shown by HAMILTON (1976), the porosity changes by a factor of approximately two to three in the upper 1 km of sediment. The difference between the average porosity at sites 6 and 7 is due to different thicknesses of sediment and thus a different amount of compaction at the sites.

From the porosity values, electrical conductivities have been determined (for saturation by sea water of conductivity 3.16 $S{\cdot}m^{-1}$, the conductivity of seawater at 1°C, $35^0/_{00}$ concentration, and 4000 m depth), using the relationship of BOYCE (1968) as discussed by BULLARD and PARKER (1970). The resultant values are 0.69 $S{\cdot}m^{-1}$ for site 6, and 0.64 $S{\cdot}m^{-1}$ for site 7.

Seafloor conductances have then been determined multiplying sediment thickness (as determined) by sediment conductivity (as just described) to give values of 760 S for site 6, and 805 S for site 7. Hence, for both sites

$$\lambda = 0.05$$

and

$$C = 0.95.$$

These values of conductance, λ and C, are consistent with others calculated for comparable geological environments. They will now be compared with estimates determined experimentally, using seawater motional electromagnetic induction.

4. The Seafloor Electric and Magnetic Data

The data to be examined, recorded in the Tasman experiment, have been reduced particularly by FERGUSON (1985, 1988). The instruments and their methods of recording are as described by FILLOUX (1987). A most important matter for the present paper is that of instrument drift, the slow change of baseline with time, which may occur for a variety of reasons.

In the case of Filloux seafloor instrumentation, the electric and magnetic instruments have quite different characteristics of drift. The electric instruments, due to the incorporation in their design of a salt-bridge switching device, are not vulnerable to the electrochemical potentials of their electrodes in seawater; in fact the combined potential of an electrode pair is monitored and recorded as a time series by the system. Free of electrode effects the electric field data are measured against a background field which is generally zero. The data are thus ideal for oceanographic purposes, which require stability (and lack of drift) over long time scales.

In contrast, seafloor magnetometers are based necessarily on different principles. The magnetic sensors themselves (whatever their design) may drift with time, especially as an instrument comes to thermal equilibrium with the seafloor temperature. In addition, the magnetometers are measuring fluctuations against a much larger steady background field, the main magnetic field of Earth. This circumstance means that any mechanical settling or change of orientation of the instrument on the seafloor will generally introduce a drift component into the measurement of fluctuations.

Such drifts, for Filloux magnetometers, are customarily reduced by fitting to the data empirical curves, comprised of one or more exponential decay terms, of different time constants. For the analysis of recorded data at higher frequencies, as in seafloor magnetotelluric studies (e.g. FILLOUX, 1977, 1980; FERGUSON *et al.*, 1990) such a procedure is satisfactory and produces high-quality data. As a procedure it may be less satisfactory for phenomena of longer time scales, such as ocean currents, but it is considered controlled in the analysis below, as will be demonstrated.

Also, in the analysis to follow, data from the Canberra Magnetic Observatory have been used as a reference source. These data are recorded following traditional observatory practice, and are taken to be drift free.

The Filloux seafloor data are recorded at 2 minute intervals (more exactly at 32 data points per hour), so that they observe the highest frequencies of natural ionospheric magnetic fluctuations which penetrate to the Tasman seafloor. In the present paper, relatively slow or long time-scale phenomena are studied (consistent with the theory in Section 2 being for steady-state, or zero frequency). Thus the time series presented and plotted below are daily means of the recorded data, taken every 6 hours. The adoption of daily means suppresses, in the data, signals of tidal origin, and those of the magnetic daily variation, which otherwise would be strong to the point of being dominant (as shown for example in FERGUSON, 1988 and LILLEY *et al.*, 1989).

5. Search for a Magnetic Signature

5.1 Subtraction of site 3 records from those of sites 6 and 7

As shown in Fig. 1, site 3 is mid-way between Australia and New Zealand; it is also remote from the activity of the East Australian current, and is in the same depth of ocean water as sites 6 and 7. FERGUSON (1988) demonstrated a high degree of coherence between the magnetic fluctuation records of the Canberra and Eyrewell magnetic observatories, in Australia and New Zealand respectively (see Fig. 1), as would indeed be expected for two stations along almost the same parallel of geomagnetic latitude, in mid-latitudes.

Hence (except that it is itself affected by some ocean movement, and uncorrected instrument drift) site 3 should give a record of the geomagnetic activity of ionospheric origin recorded at site 6 (and across the Tasman Sea generally, for uniformity of water depth and seafloor conductivity structure across this region). Thus subtracting a site 3 record from the corresponding site 6 record, point by point, should give a site 6 record of ocean movement (and uncorrected drift) only. The same process should also apply to site 7.

This procedure is carried out in Figs. 3 to 6, for the two sets of horizontal component data at the two sites 6 and 7. The data are daily means, plotted at six-hourly intervals, and joined together to give a continuous trace. The bottom trace in each of these figures is thus an estimate

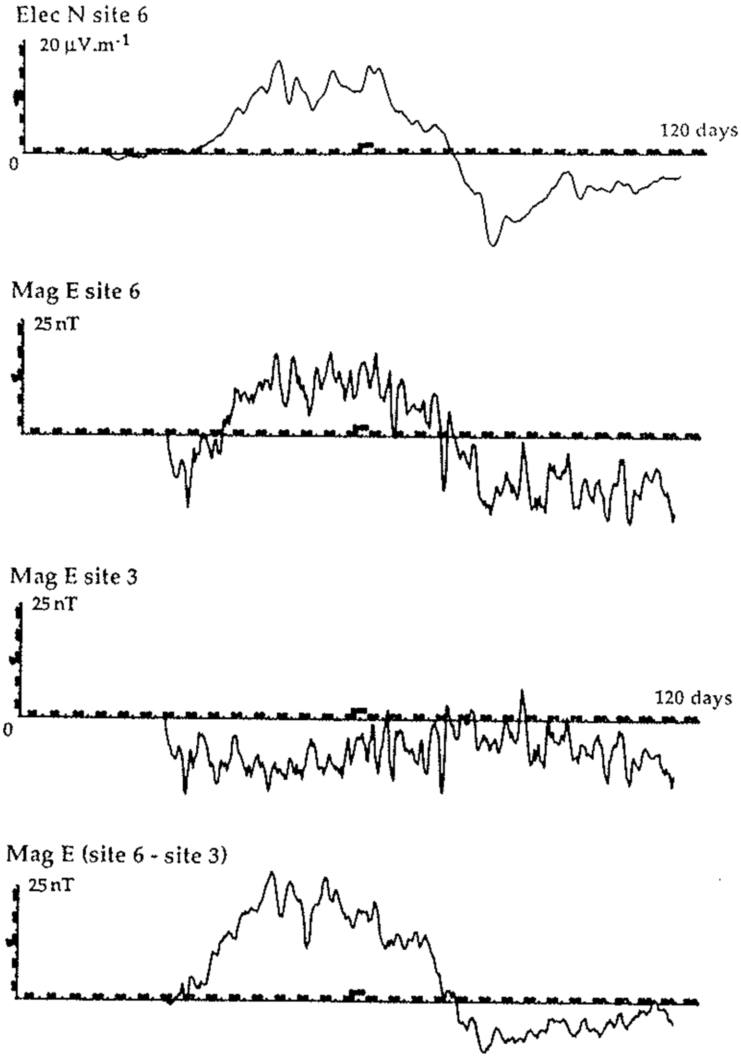

Fig. 3. Records for site 6. From top to bottom, the profiles show (i) the electric field signal, north component, (ii) the magnetic field signal, east component, (iii) the magnetic field signal, east component, site 3, and (iv) the result of subtracting (iii) from (ii), to give an estimate of the oceanic magnetic signal at site 6, which (according to the theory of Section 2 applied to slow or quasi-static data) should be similar to (i).

of that magnetic field component at the appropriate site due to ocean movement (and remaining instrument drift) only. Extending the results of Section 2 from steady to "quasi-static" or slow variation with time indicates that such magnetic signals should show the same time variation (scaled by some factor and generally with a different baseline) as the orthogonal seafloor electric field, which is shown in the top trace of each figure. The theory also predicts that positive electric north should accompany positive magnetic east, while positive electric east should accompany negative magnetic north. The matches of the upper and lower traces in each figure, and the observation of the sign rule just stated, indicates that seafloor magnetic field signatures have indeed been observed. The match between magnetic and orthogonal electric is not only at the longest time scales evident, of lengths tens of days, but also at shorter time scales, of length several days.

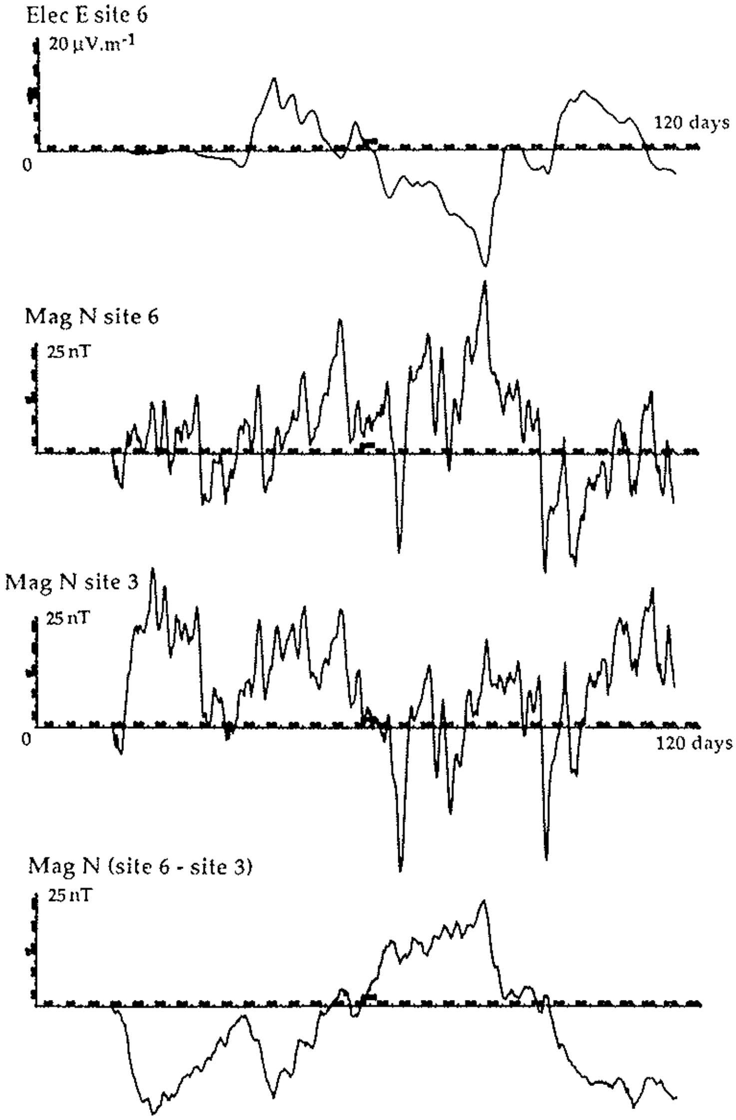

Fig. 4. As for Fig. 3, but site 6 electric east and magnetic north components. Note, relative to Fig. 3, the greater strength in the magnetic north ionospheric signal, relative to the magnetic east.

5.2 Use of a transfer function algorithm

The correlation between magnetic and orthogonal electric signals on the seafloor has been determined quantitatively in the four cases just discussed by using a transfer function algorithm. A seafloor magnetic field component from either site 6 or site 7 is regarded as an output resulting from two inputs: the same component of magnetic field as recorded at Canberra Magnetic Observatory; and the orthogonal component of electric field as recorded at the site. Canberra Magnetic Observatory data are used in this instance as being free of drift, and free of the influence of any local ocean movement.

The algorithm used is standard from the Numerical Algorithms Group (NAG library algorithm G13BEF) and will not be described in detail here. The output series y_t, $t = 1, \ldots, T$ is assumed to be the sum of (unobserved) components $z_{i,t}$ which are due respectively to the inputs $x_{i,t}$, $i = 1, \ldots, m$. Thus

$$y_t = z_{1,t} + \cdots + z_{m,t} + n_t$$

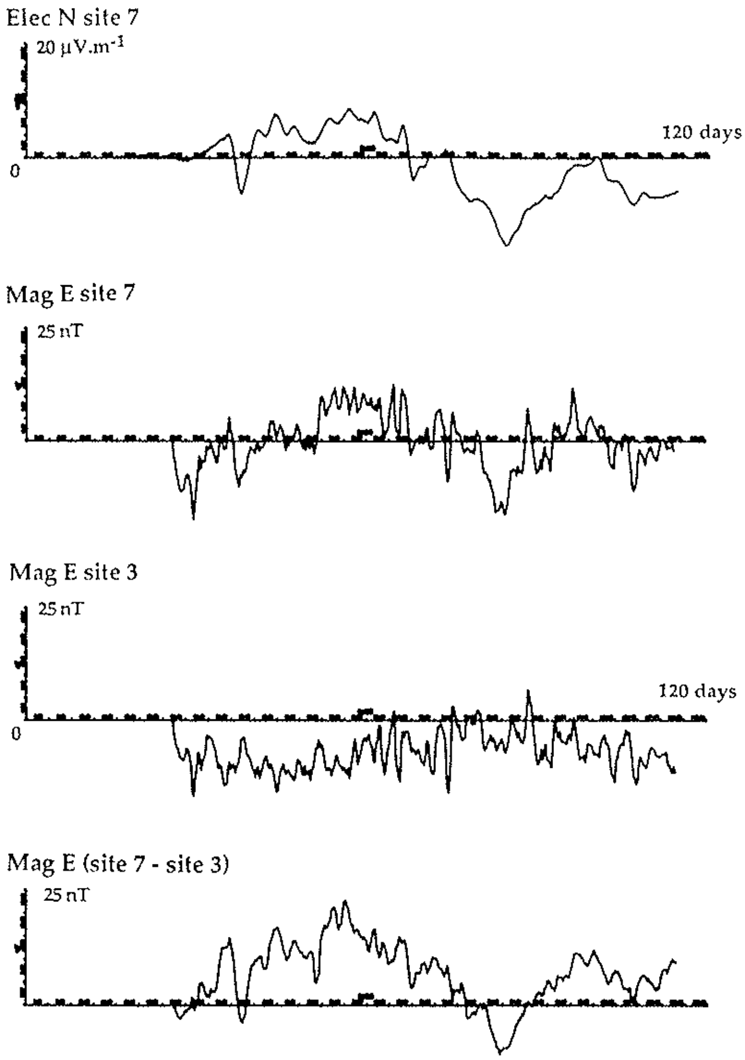

Fig. 5. As for Fig. 3, but site 7 electric north and magnetic east components.

where n_t is the error, or output noise component.

In the present instance, consistent with the theory of Section 2 being for steady or quasi-steady flow, the $z_{i,t}$ are taken to be simple regression coefficients of simple inputs; that is,

$$z_{i,t} = w_i x_{i,t}.$$

Three slightly different estimation criteria are possible (non-linear least squares, exact likelihood and marginal likelihood) and all give the same results in the present instance.

The results of the application of the algorithm to the four cases of two components at each of sites 6 and 7 are shown graphically in Figs. 7, 8, 9 and10. The graphs can be read as showing that the magnetic field (top line) when regarded as the output from two inputs (given in the second and third lines) requires them precisely as scaled in the fourth and fifth lines, with the sixth line as minimized residual, remaining from the signal in the first line.

The residuals are considered to show, in their long time-scale behaviour, uncorrected drift in the magnetic records. There are certain purposes for which this information would be useful. In their short time-scale behaviour the residuals show some morphology recognizable as similar to that of the observatory records. A reduction of this latter part of the residual would be unlikely

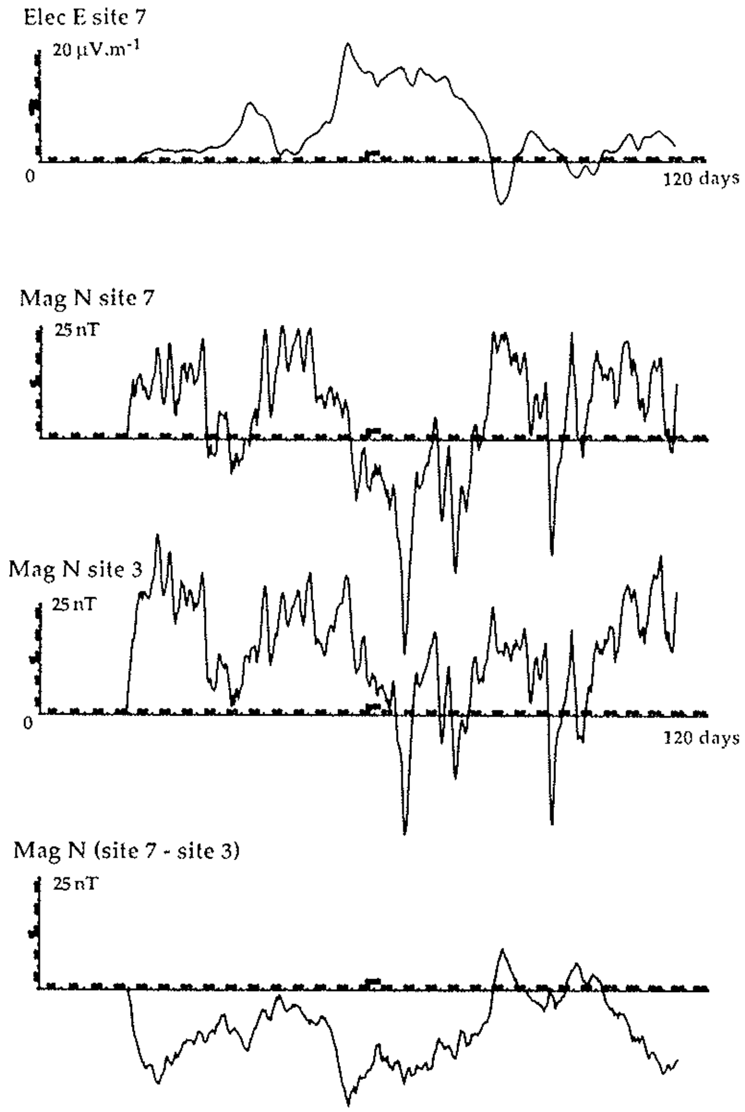

Fig. 6. As for Fig. 3, but site 7 electric east and magnetic north components.

to significantly change the values of the transfer functions to be used below, as these relate the seafloor magnetic field to the seafloor electric field (and not to the observatory magnetic field).

6. Interpretation and Discussion

The four cases of two components each at sites 6 and 7 give four b_s/E determinations, which are listed in Table 1. While in principle the method might distinguish between different seafloor conductances at sites 6 and 7, and could (with an expanded transfer function procedure) determine anisotropy at an individual site, in view of the errors determined for the b_s/E values it is judged appropriate here to calculate from them together a mean value for the conductance of that region of the Tasman seafloor; it is relevant also that the width of the eddy is greater than the separation of sites 6 and 7. This mean value, as shown in the table, gives a leakage factor of

$$\begin{aligned}\lambda &= b_s/\mu_o\sigma_1 HE \\ &= 0.057 \pm 0.009\end{aligned}$$

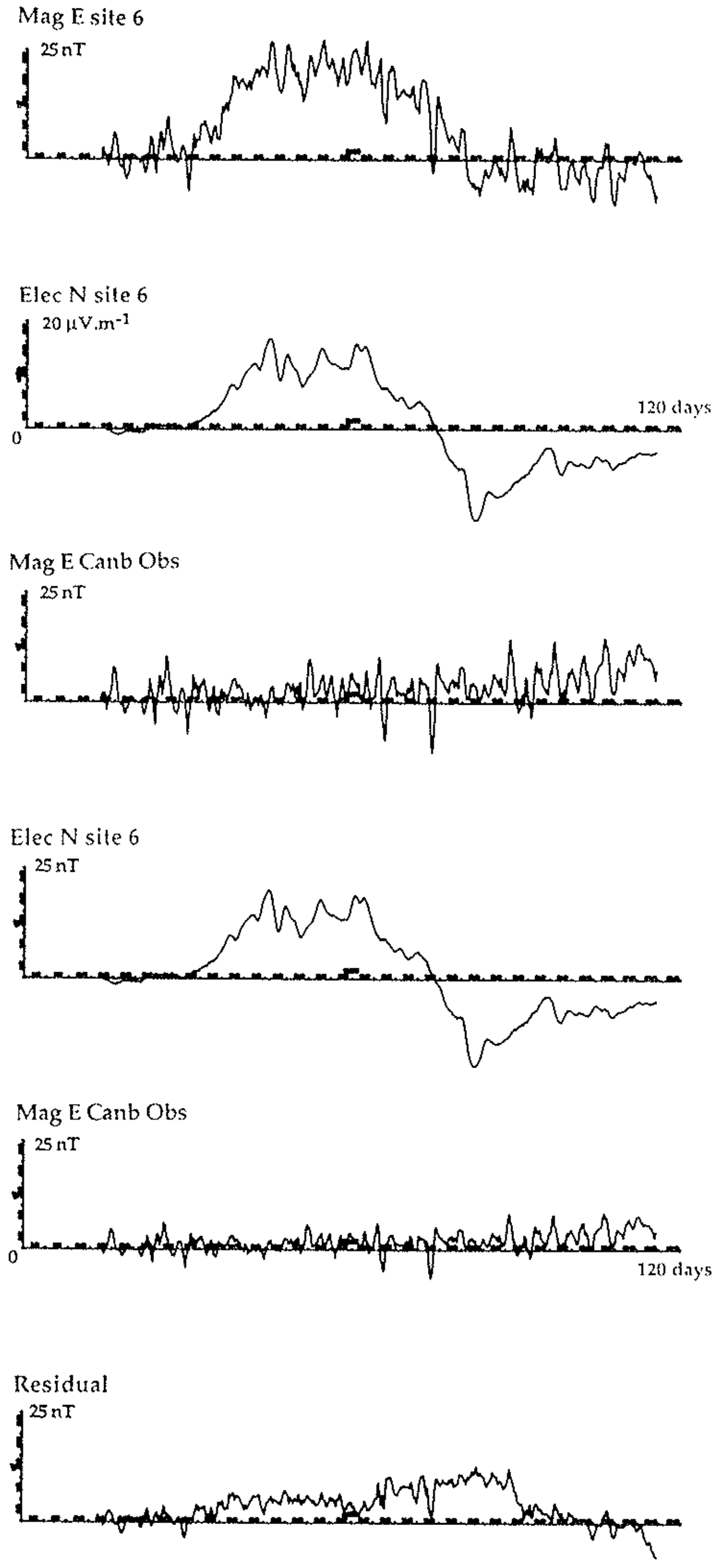

Fig. 7. Transfer function determination for site 6, as described in the text, taking the top line (Mag E) as output, and the next two lines (Elect N and CMO Mag E) as inputs. The fourth and fifth lines are these two inputs scaled as required, and the sixth line is the residual then remaining from the output in line 1.

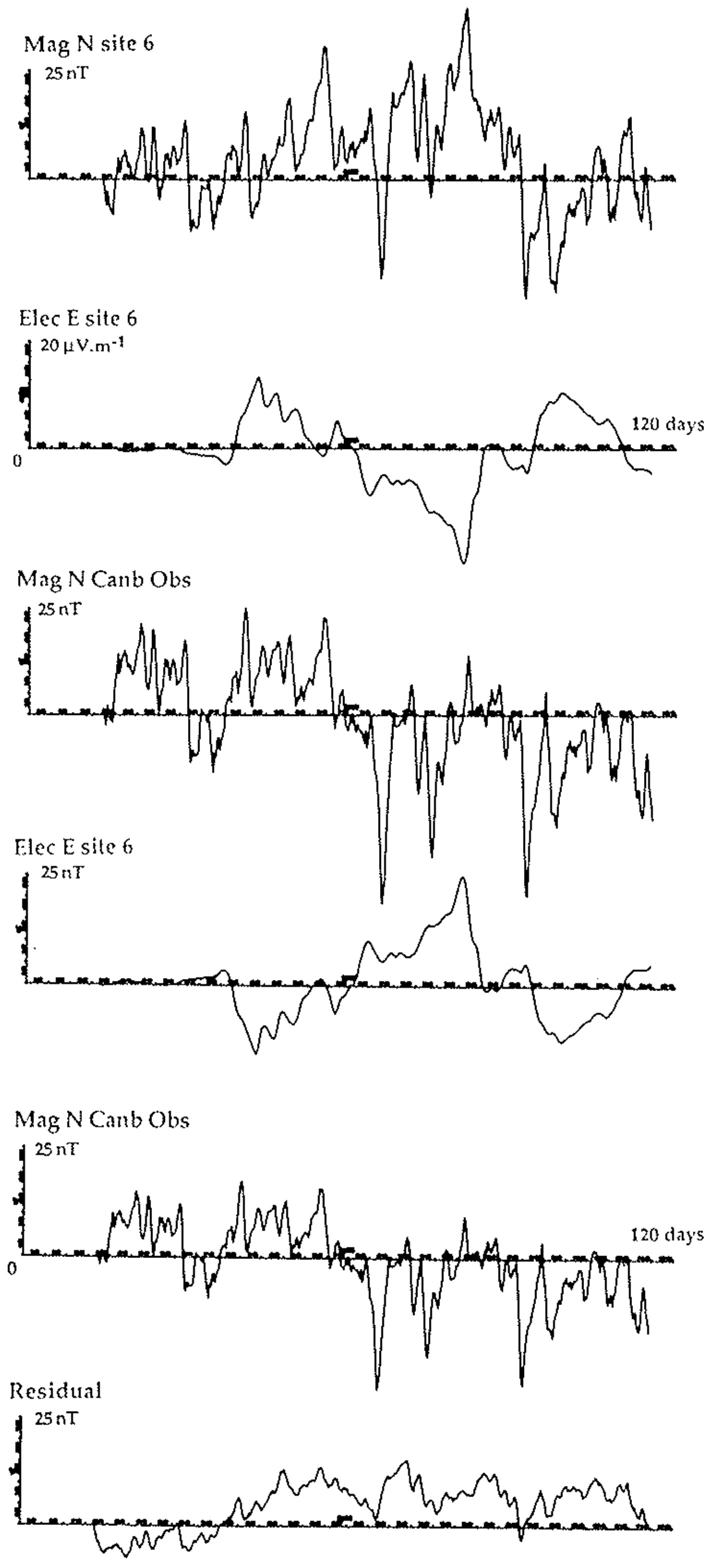

Fig. 8. As for Fig. 7, but Mag N and Elect E components at site 6.

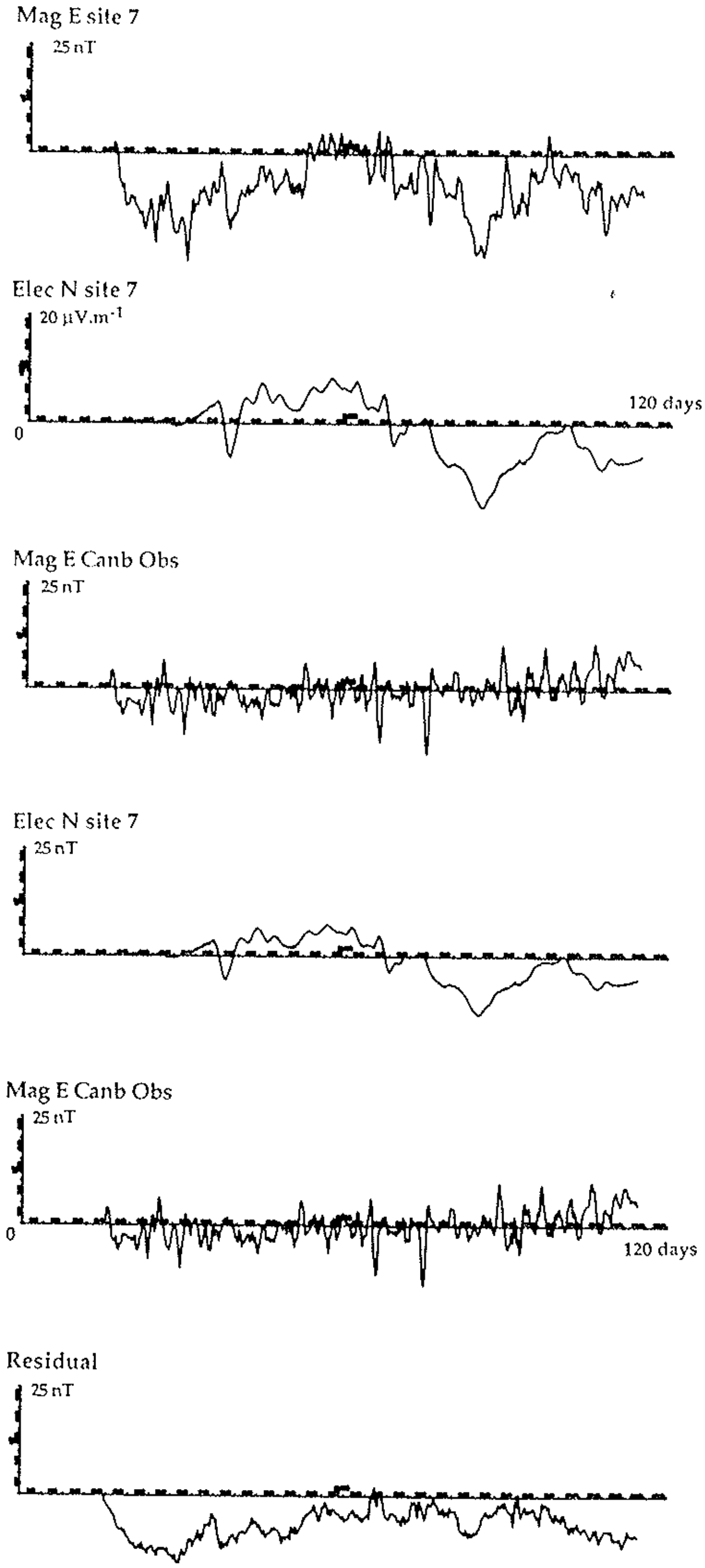

Fig. 9. As for Fig. 7, but Mag E and Elect N components at site 7.

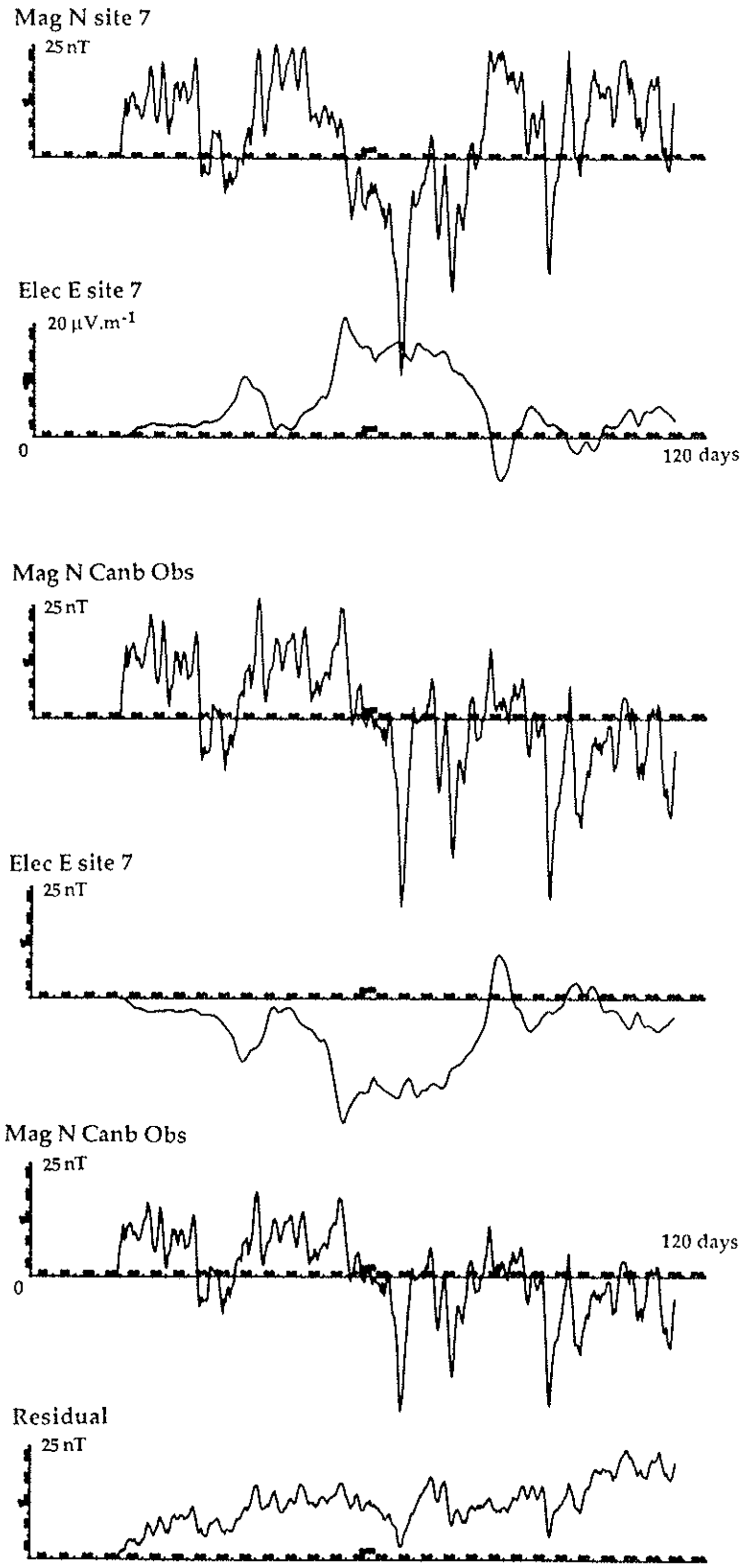

Fig. 10. As for Fig. 7, but Mag N and Elect E components at site 7.

Table 1. Values of b_s/E determined as transfer functions, electric to magnetic fields (units: $\mathrm{nT} \cdot \mu\mathrm{V}^{-1} \cdot \mathrm{m}$).

	Site 6	Site 7
East Magnetic, North Electric	1.20 ± 0.02	0.87 ± 0.03
North Magnetic, East Electric	1.18 ± 0.04	1.31 ± 0.04

Mean of all values = 1.14 ± 0.19

and also, directly (dividing the mean b_s/E value by μ_o), an estimate of the seafloor conductance of

$$b_s/\mu_o E = 910 \pm 150\ S.$$

Having thus determined a seafloor conductance electromagnetically, it is appropriate to return to Section 3 and compare the seafloor conductance values determined there. As the values for sites 6 and 7 were 760 S and 805 S respectively, the agreement is gratifying, giving confidence both in the various steps upon which the conductance value determinations have been based and the application of the theory of Section 2. Similarly the leakage factor determined electromagnetically here agrees pleasingly with the value of 0.05 in Section 3.

In two earlier papers on the Tasman experiment, LILLEY *et al.* (1986) and MULHEARN *et al.* (1988), barotropic flow was estimated on the basis of electric field data only, and a leakage of 0.1 was taken. It is now seen that that value was an overestimate, and that all barotropic flow and transport values determined were too large by a factor of 1.04. That is, transport values in those papers should now be corrected by multiplication by a factor of 0.96. This correction is generally not significant, and the determination above of 0.06 for the Tasman Sea supports the general world-wide leakage value of 0.1 suggested by SANFORD *et al.* (1990). The present case history also indicates the utility of sediment-based calculations for estimating λ.

7. Conclusions

7.1 General

The paper set out to seek magnetic signals due to the motional induction of seawater over time scales of days, and several factors indicate this objective has been achieved. These factors are the obvious correlation of the seafloor electric field signals (known to be of oceanic origin) with the appropriate orthogonal magnetic field signal in Figs. 3, 4, 5 and 6, and the consistency of this correlation when measured quantitatively, and as tabulated in Table 1. Secondly, the amplitude of the magnetic signals is of the order which would be expected from the elementary theory in Section 2, and from the estimates (based on other evidence) of the seafloor conductances in Section 3. Indeed, the agreement between the predicted and observed conductances and leakage factors gives confidence in the applicability of the theory of Section 2, even to an eddy in which the water motion is typically cylindrical rather than rectilinear. The reason for this validity is likely to be that the radius of the cylindrical motion is generally much greater than the water depth; the eddy is of very low aspect ratio.

There have been particular circumstances which have allowed this demonstration of the magnetic field of ocean currents. The active ocean movement, in this case due to the occurrence of a mesoscale eddy, has been of primary importance. A second important factor has been the pile of conducting sediments of the Tasman Abyssal Plain, forming a conducting ocean floor allowing leakage. Thirdly, the different horizontal scale lengths of the oceanic and ionospheric magnetic source-fields have been important in their separation. It has been important also, of course, to have remote reference magnetic data, as from the Canberra Magnetic Observatory, and Tasman site 3.

7.2 Oceanography

For oceanographic purposes, this case history has enabled an estimate of the depth-integrated velocity in the eddy, by providing an observed λ value for sites 6 and 7. The stability of the ocean floor is such that these sites are now "calibrated", and future observations at them of magnetic field changes only could be used to give changes in the oceanic depth-integrated velocity overhead.

There is also encouragement in the use of magnetometers to measure oceanic flow in other places; though more generally the possibility of three-dimensional electric current flow patterns must be remembered, with geometry more complicated than as shown in Fig. 2. The model of Fig. 2 has been taken for the radial section of an eddy in part because an ideal eddy has a cylindrical symmetry, which other ocean motions may not share.

Figure 2 also suggests the interest and challenge of measuring the ocean magnetic profile with depth.

7.3 Seafloor conductance

Regarding marine geology, the example described above has given an estimate of the seafloor electrical conductance, and confirmed the method of calculating this quantity from known sediment information. The model is supported of an increase in resistance of several orders of magnitude in the seafloor material below the sediments. This result is in accord with sediments resting on seafloor crystalline basalt.

The method of determining the seafloor conductance has used electric power generated by the ocean eddy, to sound the ocean floor. Compared to continental electrical sounding methods, which typically rely on ground "point-contact" at various electrodes, the electrical coupling of the ocean eddy with the seafloor by ocean water everywhere is quite ideal.

7.4 Ocean-bottom geomagnetic observatories

At the present time the major question is being addressed of supplementing the world-wide geophysical observatory network (seismic and geomagnetic) with seafloor stations, with which communication by existing seafloor cables may be feasible (WALKER, 1991; BUTLER *et al.*, 1992). For such geomagnetic observatories, the separation of oceanic signals from ionospheric signals is of prime importance. The example in the present paper shows the benefit in such an analysis of also having seafloor electric data (as at long periods these data will be almost entirely oceanographic), and also a remote magnetic station (either land or seafloor) to use for correlation, and the determination of the part of the seafloor signal of ionospheric origin.

If the purpose of such a seafloor observatory is not oceanographic, then there should be preference for sites of low seafloor conductance. Also, in this case, active ocean regions should be avoided where possible.

7.5 An homogeneous disc dynamo

Finally, recording both the electric and magnetic fields of the mesoscale eddy suggests viewing it as a disc rotating in a medium of the same material, with isotropic electrical conductivity (except at the sea surface). While it is thus a dynamo of very low "magnetic Reynolds Number", it is nevertheless an unusual example in nature of kinematic dynamo action. The mechanism can be viewed as a toroidal velocity (that of the disc) acting on a poloidal magnetic field (the vertical component of Earth's steady magnetic field) to produce a toroidal induced magnetic field (as observed at the seafloor).

An even more ideal homogeneous dynamo may be a submerged eddy, such as the "Meddies" which form in the Atlantic Ocean from water from the Mediterranean Sea (ARMI *et al.*, 1989; BORMANS and TURNER, 1990). Such submerged eddies are not at an air-sea interface, as are surface eddies such as those in the East Australian Current.

Regarding dynamo action, a Meddy should exhibit an extra mode of induction, that of the east-west velocity component cutting the horizontal (north-south) Earth's main magnetic field, to give an east-west dipole induced field at the centre of the Meddy. Unlike the toroidal induced field first referred to above, and detected on the seafloor, this dipole induced field will not be contained in the good conductor (seawater plus sediments), but will permeate space. It is intended to give more attention to such dynamo aspects of ocean eddies in a subsequent paper.

Many people have contributed to this paper in comments, discussion and encouragement. One of us (F.E.M.L.) benefitted from the enthusiasm of J. C. Larsen during a visit to Seattle several years ago, and A. D. Chave, T. B. Sanford and D. S. Luther, and others, contributed valuable comments at the IUGG Meeting in Vienna in 1991. R. W. Griffiths and M. Bormans are thanked for discussions on eddies and Meddies. The Australian Geological Survey Organisation (formerly the Bureau of Mineral Resources) is thanked for the data from Canberra Magnetic Observatory. Especially N. L. Bindoff laid the ground for the present work in his doctoral research at the Australian National University on the seafloor electric fields due to oceanic movement.

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