Geophysical Challenges in Using Large-Scale Ocean-Generated EM Fields to Determine the Ocean Flow

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For some time, oceanographers have used measurements of the small electromagnetic fields generated by the flow of the electrically conductive oceans through the Earth’s main magnetic field to infer values for the ocean flow velocities. An overview of the process of motional induction is given including a description of the global electromagnetic fields generated by a global model ocean. We describe how electromagnetic methods are currently used in oceanography and outline the most important challenges presently faced.

1. Introduction

The purpose of this paper is to give geophysicists an overview of motional induction and its role in oceanography and to describe the objectives and some of the current challenges. In oceanography, electromagnetic (EM) fields are measured in the ocean to infer estimates of ocean velocities. In other fields of geophysics, ocean EM observations are made to gain information about the conductivity and hence structure of geological formations.

While the goals of these studies are different, similar practical challenges are faced and there are benefits in a combined research effort. In perhaps the strongest example of this, a precise determination of ocean flow from EM observations requires estimates of ocean-bottom conductance. Such estimates can and have been made in oceanography using not only the EM observations but also knowledge of the oceanographic feature measured. These estimates might be utilized in geophysical applications. On the other hand, there is likely a substantial amount of data and geophysical expertise that could be more fully utilized in making these ‘corrections’ in oceanographic applications.

In Section 2 we give a brief description of the ocean circulation together with some plots showing the global circulation from an ocean circulation model. In Section 3 we give a brief description of motional induction for some idealized cases, and then show the calculated EM fields generated by the global model ocean circulation. In Section 4 we give a short review of how and why EM observations are made in oceanographic applications.

2. Description of Ocean Circulation

The ocean circulation is largely driven by atmospheric wind stresses and buoyancy forces. These result in regional differences in water density and sea surface height, which are the driving forces for the 3-dimensional ocean circulation. The dynamic inertia and the huge heat capacity of the ocean together with the atmospheric noise cause low-frequency variations in the climate system.

The direct effect of the wind forcing is typically confined to a surface ‘mixed layer’ which has a thickness ranging from several tens of meters below the sea ice and in coastal and equatorial upwelling areas while in the subtropics and particularly in midlatitudes strong winds can cause a mixed layer depth of many hundred meters. The convergence of wind-driven flow, however,
together with the continental boundaries and rotation effects creates pressure gradients and more complicated flow patterns that are not restricted to the mixed layer. As a brief description of the major ocean features, we will discuss Figs. 1 and 2 which show results for the steady component of the ocean circulation as computed by the OPYC general circulation model (these results and the OPYC model are further described in other work (Oberhuber, 1993a,b; Tyler et al., 1997)).

A system of polar and subpolar ocean gyres in each of the ocean basins reflects the wind stresses of corresponding polar and subpolar atmospheric systems. The poleward currents on the western sides of the ocean basins are generally intensified due to the Earth's rotation. These 'western boundary' currents can be extremely intense, having velocities exceeding a meter per second (in the mid-ocean, velocities may typically be 1–10 cm/s) and extending several kilometers to the sea floor. Well-known examples of western-boundary currents are the Florida Current and the Kuroshio (off Japan). A rather anomalous current system is the Antarctic Circumpolar Current (ACC) whose total water transport is twice that of the Florida Current. Unlike the western-boundary currents, the ACC remains intense and extends deeply over an extremely large global scale—a hemisphere essentially. This is likely due to the combination of strong westerly winds and the lack of continental boundaries to block the flow (though blocking might occur to some extent in shallow passages such as the Drake Passage between South America and Antarctica).

Aside from the dominant wind-driven horizontal ocean currents (and the more complicated secondary 3-D flows due to wind-driven pressure effects, etc.) another important agent effecting low-frequency ocean flow is the buoyancy forces due to differences in seawater density. The density of seawater increases with salinity and pressure (depth) and decreases with temperature. For a given depth, the salinity and temperature of a water parcel will usually determine whether the water is relatively heavy and will sink, or relatively light and will rise. The salinity and temperature effects can counteract; a dramatic example is the outflow of Mediterranean water through the Strait of Gibraltar. The Mediterranean entering the North Atlantic is relatively

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Fig. 1. Horizontal barotropic transport stream function as simulated by OPYC. Units are $10^9$ kg/s. Flow is counterclockwise around high.
Fig. 2. Meridional overturning transport stream function in the Atlantic (a) and Pacific (b) oceans as produced from OPYC. Units are $10^9$ kg/s. Flow is clockwise around high.
warm (light) but also relatively salty (heavy). The net effect is that the warm Mediterranean water spreads out into the Atlantic not over the surface but at an intermediate depth.

Low-frequency buoyancy-driven flow can be expected due to the continual external influences on seawater density. First, atmosphere-ocean heat fluxes affect the temperature of the surface waters, and regional differences in evaporation and precipitation create changes in the salinity. Additionally, during the formation of sea ice, brine (salt) is rejected, tending to increase the density of the surface waters. All of the factors above affecting seawater density can be important in creating buoyancy-driven flow. In particular, in certain key polar regions where dense cold surface water is made even denser with the addition of salt during the formation of sea ice, extremely dense water is created which sinks to great depths or even the sea floor and spreads out over the ocean basins through organized flow systems. The polar surface waters may be replaced by warmer water flowing poleward and the global process has been likened to an ‘oceanic conveyor belt’ (Broecker, 1987) which transports heat from the low latitudes to the polar regions.

A focus of oceanography and climate research is currently to understand these overturning processes and their potential contribution to climate changes since it appears that the heat transported in this way is of the same order as that transported by the atmosphere, and the overturning likely plays a role in establishing global climate. A model description of the ocean overturning can be seen in Fig. 2.

An important parameter in calculating the ocean-generated electromagnetic (EM) fields is the electrical conductivity of the oceans. In Fig. 3 we show the electrical conductivity near the sea surface, and in Fig. 4 we show a transect of the conductivity along 170°W (through the mid-Pacific and Bering Strait). The transect is indicative of most of the ocean in that we see

![CONDUCTIVITY (S/m); (Z=-20 m)](image)

Fig. 3. Ocean conductivity (S/m) at 20 m depth.
that outside of a thin surface layer, the conductivity of the ocean is quite uniform. Near the surface, the conductivity variations are largely due to water temperature variations, except in polar regions where temperature variations are often small and salinity variations (due to river runoff\(^1\), for example) can be large. When considering the ocean-generated EM fields, however, it is important to not discount the ocean-conductivity variations since these variations occur where the flow is also strongest—near the surface. The net poleward transport of heat by the ocean can also be associated with a net poleward transport of conductivity (see plots in Tyler et al., 1997).

It should be noted that the description of steady ocean circulation given in plots in this paper is that produced by a numerical model; although the realistic ocean will certainly be different, the model description given here is known to reliably produce the observed steady ocean features and we think that for the purposes here the model ocean will be very adequate. A bigger concern is that we will only be using a non-eddy-resolving numerical simulation of the ocean circulation. Since about half the oceanic kinetic energy is attributed to transient ocean features (e.g., eddies, meanders), the electromagnetic fields we will show are similarly an underestimate, and this should be kept in mind. It is also important to note that the ocean (and ocean-generated EM fields) varies on an extremely wide range of time-scales, from periods of a few seconds for gravity waves to the tectonic time scales describing the age of ocean basins and associated circulation features.

\(^1\)The Arctic Ocean receives a large disproportionate amount of river runoff. Also, because of the seasonality in the runoff and the restricted geographic connection of the Arctic basin with the other oceans, the Arctic experiences relatively large variations in salinity (and conductivity).
3. Description of Motional Induction in the Ocean

3.1 A cartoon of motional induction

Ocean motional induction is a process whereby oceanic kinetic energy is converted into EM energy and heat. This takes place due to the interaction between moving charges comprising currents generating the Earth's magnetic field and charges moving with the ocean flow. A theoretical treatment of motional induction and solutions for various idealized cases describing ocean flow appears in many earlier works and will not be extended here. (Some of the basic approximations and formulations are, however, presented with other material in Section 4.) In this section we present only a couple of cartoons to give the reader a quick description of a typical case.

In Fig. 5 we show a symmetric circulation of ocean flow (a gyre) confined to a thin mixed layer overlying a deep ocean. The background Earth's main magnetic field $\mathbf{F}$ is assumed uniform and we will neglect the 'self induction' due to products of the velocity with the ocean-generated magnetic field—hence, $\mathbf{B}$ in Ohm's law for a moving conductor

$$\mathbf{J} = \sigma(\mathbf{E} + \mathbf{u} \times \mathbf{B})$$

(1)

can be replaced by $\mathbf{F}$ (really, for this simple geometry the self induction vanishes exactly). As seen in the figure, induced electric currents move outward in the surface mixed layer and return in lower layers. A toroidal magnetic field flux is induced through these current loops.

In principle, since the mantle is not a perfect insulator, ocean-driven electric currents will reach down to the fluid core creating Lorentz forces tending to entrain the fluid. More generally, this coupling would tend to reduce the relative velocities between the core and ocean flows. These forces, however, are extremely small as seen in estimates (Tyler, 1995) and, following the opinion

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Fig. 5. Cartoon showing motional induction for a simplified case of a symmetric ocean surface gyre (confined to the mixed layer) flowing in a uniform main magnetic field. (Taken from Tyler, 1995.)
of several experts on core dynamics with whom we have spoken, should be of no consequence. If this is correct, then the presence of the core need only be considered in motional induction studies when treating ocean features having horizontal scales greater than about 1000 km since for these scales the ocean generated EM fields become sensitive to the high-conductive lower boundary. For example, calculations of large scale electric potential differences such as that to be discussed in Fig. 11 will likely require a representation of the core for increased accuracy. A core representation may be even more vital in calculations involving large-scale fluctuating ocean features.

Notice that in Fig. 5, the induced electric currents flow entirely in planes containing the vertical axis and the induced magnetic field is confined to the ocean. When less uniform cases are considered, however, we will in general have 3-D electric currents. For example, in Fig. 6 we have assumed the background main field increases in strength in the $y$-direction. In this case, we will have an effect similar to that shown in Fig. 5, but in addition we will also see electric currents closing in horizontal planes and magnetic fields reaching outside of the ocean as shown.

Observations and scaling analysis indicate that typically, the dominant EM fields behave like those shown in Fig. 5—electric currents flow in vertical planes and induce magnetic fields which are confined to the ocean. The horizontal-plane electric currents (which will also be referred to as 'nonlocal' electric currents in the rest of this paper), even if secondary in strength, are still very important for reasons which will be further elaborated in later sections.

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[^2]: R. T. Merrill, however, has suggested that such small forces might be important in giving a bias to the structure of normal and reversed states of the geodynamo. He warns, however, that recent work calls to question previous statistical evidence for a preferred structure.
3.2 Steady EM fields generated by global ocean circulation

To date, few studies have focused on calculating the electromagnetic fields induced by a prescribed realistic large-scale ocean circulation. All of these studies have assumed inertial-form EM equations. Stephenson and Bryan (1992) used a 2-D thin shell formulation to calculate the ocean-generated 2-D electric potential and 2-D electric current stream function due to coarse-resolution global circulation. Tyler et al. (1997) found global 2-D results in qualitative agreement with the Stephenson and Bryan study but showing additional smaller-scale features due to higher resolution. The Tyler et al. study also used an approximate set of EM equations to calculate the 3-D dominant EM fields within the ocean which are not resolved in 2-D formulations. Probably the highest accuracy has recently been attained by Flossadöttir et al. (1997) in a high-resolution study of the North Atlantic which attempted to include a realistic description of bottom conductance. The model used in this study had enough resolution to capture oceanic eddies and other transient effects. Intercomparison of results from the three studies suggest that the magnitudes of the calculated ocean-generated EM fields increase as increased spatial and temporal resolution of the model allows more of the oceanic energy to be included.

We can give a brief description of the patterns and magnitudes of the ocean-generated EM fields by presenting EM fields calculated from the prescribed ocean state illustrated in Figs. 1
Fig. 8. The eastward (a) and northward (b) components of the ocean-generated electric current density \( \text{A/m}^2 \) along 170\(^\circ\)W.
Fig. 9. The vertical component of the ocean-generated electric field (V/m) at a depth of 20 m.

Fig. 10. The vertical component of the ocean-generated electric field (V/m) along 170°W.
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Fig. 11. The electric potential (V) generated by the flow of the ocean through the Earth's magnetic field.

and 2. An example of the dominant ocean-generated magnetic field at a depth of 1 km is shown in Fig. 7, and a profile of the horizontal electric current density through the Pacific Ocean is shown in Fig. 8. An example of the vertical component of the ocean-generated electric field at a depth of 20 m is shown in Fig. 9 while a profile through the Pacific is given in Fig. 10. A further description of these fields together with an extensive set of plots can be found elsewhere (Tyler et al., 1997).

The horizontal components of the large-scale electric field due to global circulation can be characterized by a potential function \( \phi \) (Stephenson and Bryan, 1992; Tyler et al., 1997) as depicted in Fig. 11 (the formulation for the electric potential is described in the Appendix). The negative gradient of the electric potential function gives the electric field vectors. The magnitude of the horizontal electric field is shown on a log scale in Fig. 12. Finally, the depth averaged nonlocal electric currents can be characterized by a stream function \( \psi \) as shown in Fig. 13. These nonlocal electric currents give rise to magnetic fields reaching outside of the ocean, estimates of which are shown in Fig. 14. (Formulations for \( \psi \) and the technique for obtaining the magnetic fields outside of the ocean are given in the Appendix.)
4. Why Are Oceanographers Interested in Measuring the Ocean-Induced EM Fields and What Are the Challenges?

There is a long history of the measurement of ocean electric fields and their interpretation in terms of ocean currents. Oceanographic interest in this approach stems, in part, from the lack of alternatives for measuring ocean velocities and volume transports. Ocean currents are invariably undersampled. As a result variability is not well resolved and spatial averages are not necessarily accurate. A few dozen point measurements, such as from moored current meters, are used to estimate the vertical and lateral integral of the flow to obtain transport. Many modern methods have similar limitation in the amount of the water column sampled. For example, the Acoustic Doppler Current Profilers (ADCP) sample only the upper 200-400 m, about 5-10% of typical ocean depths. Moreover, shipboard methods require much time to sample a large region. Motional induction, on the other hand, produces nearly instantaneous results integrated over a large volume of the ocean. The instantaneous, spatially averaged characteristics are favorable compared with alternative oceanographic methods.

The traditional methods for measuring ocean currents, such as current meters and the dynamic method (i.e., relating tilts of density surfaces to ocean velocities), have been complemented by new methods based on tracking surface drifters and subsurface floats, acoustic Doppler returns (e.g., ADCP), long-range acoustic travel times (e.g., acoustic tomography), and slopes of the sea.
surface from satellite altimeters. The latter two methods have significant advantages of spatial averaging or rapid spatial coverage. However, each of the alternative methods has significant liabilities, such as cost, representativeness, and flawed principles of operation.

Motionally induced electric fields first were observed on submarine cables, especially those across the English Channel. The voltage from England to France was observed by grounding one end of the insulated submarine cable and measuring the voltages between the center conductor and the ground at the other end. Much of the early work was rather qualitative, such as noting that the voltages across the Channel contained fluctuations of tidal periods, with less effort devoted to determinations of the lower frequency or steady component. This was because little was understood of the theory of motional induction, how to interpret the voltages, the influence of bottom electrical conductance, how to correct for electrochemical induced voltage offsets at the respective grounds and how to record voltages for long periods.

The next application was to observe ocean electric fields between a pair of electrodes towed behind a vessel. The early measurements of Young et al. (1920) appear not to have been pursued until the work of von Arx (1950). By then a much better understanding of motional induction and the interpretation of ocean electric fields was put forth by Longuet-Higgins et al. (1954). More recently, the ocean-induced voltage differences across abandoned telecommunications cables have been exploited (Larsen and Sanford, 1985) and various EM instrumentation has been developed for measuring the ocean induced EM fields (see review by Filloux, 1987).
4.1 Practical uses of HEF in physical oceanography

Physical oceanographers persist in the measurement of the horizontal electric field (HEF) because the method has unique advantages which generally outweigh the disadvantages. The challenge of direct measurements of ocean waves, currents and transports (fluxes) requires accurate, frequent, long-duration, representative and affordable measurements. Secondly, quantities should be commercially produced and reported in near real time.

There are many alternatives to motional induction methods. Many are traditional and used extensively over many or all of the alternatives. But motional induction does have some significant advantages, such as
- simplicity
- representativeness
- passive, without need to produce fields
- inexpensive
- often available in real time.

There are disadvantages, some inherent requiring compensation and some incidental requiring improved equipment. Some of the disadvantages in using motional induction methods for physical oceanography are:
4.2 Interpretation of motionally induced HEF

Some of the advantages and disadvantages may become clearer in an examination of the simplest interpretation of ocean electric fields in terms of oceanic flows. The governing equation
for motional induction is Ohm's Law for a moving medium:

\[ \mathbf{J} = \sigma (-\partial_t \mathbf{A} - \nabla \phi + \mathbf{v} \times \mathbf{F}) \]  

(2)

where \( \mathbf{A} \) is the magnetic vector potential, \( \phi \) is the electrostatic potential, \( \mathbf{v} \) is the ocean velocity vector, \( \mathbf{F} \) is the Earth's magnetic field, \( \mathbf{J} \) is the electric current density and \( \sigma \) is the electrical conductivity of the sea water. The solution for oceanic flows which are shallow compared with their width and of low frequency, such that \( \partial_t \mathbf{A} \) is negligible, obeys

\[ \nabla \phi = F_z k \times \mathbf{v}^* - \mathbf{J}^*/\sigma \]  

(3)

(Sanford, 1971), where

\[ \mathbf{v}^* = \int_{-H}^{0} \frac{\sigma \mathbf{v} d\xi}{\int_{-H}^{0} \sigma d\xi}, \]  

(4)

with \( F_z = \) vertical components of Earth’s magnetic field, \( H = \) ocean bottom depth, \( H_s = \) lithospheric depth beneath which the electrical conductivity is assumed to vanish, and \( \mathbf{J}^* = \) electric currents of nonlocal origin. In most cases, we assume that \( \mathbf{J}^* \) is negligible. In 2-D flows, it is identically zero. In flows with downstream variations in velocity, depth or bottom electrical conductance, \( \mathbf{J}^* \) will arise to circulate in circuits closed in the horizontal plane. Basically, the currents arise from the spatial distribution of \( \nabla \cdot (F_z \mathbf{v}^*) \) (Sanford, 1971; Chave and Luther, 1990). This expression demonstrates that simple curvature of the flow, such as exhibited in an eddy or ocean vortex, does not produce \( \mathbf{J}^* \) electric currents. Such electric currents probably are small except in special regions where the topography changes rapidly and in flows which reverse direction in space, such as wave-like, tidal constituents.

It is convenient to substitute some expressions for integral quantities:

\[ \mathbf{v}^* = \frac{\sigma \mathbf{v}}{\sigma(1 + \lambda)} = \frac{\mathbf{v}}{1 + \lambda} + \frac{\sigma' \mathbf{v}'}{\sigma(1 + \lambda)} = \frac{\mathbf{v}(1 + \gamma)}{1 + \lambda} \]  

(5)

where the electrical conductance ratio, \( \lambda \), and the velocity-conductivity correlation (\( \gamma \)) are defined as

\[ \lambda = \frac{\int_{-H}^{0} \sigma d\xi}{\int_{-H}^{0} \sigma d\xi}, \]  

(6)

and

\[ \gamma = \frac{\sigma' \mathbf{v}'}{\sigma \mathbf{v}} \]  

(7)

where \( \sigma \) and \( \mathbf{v} \) have been expressed as the sum of a vertical average ('\( ' \)) and deviations ('\( ' \)).

A simple rearrangement of terms in Eq. (2) yields:

\[ \partial_t \mathbf{A} + \nabla \phi - \mathbf{v} \times \mathbf{F} = -\mathbf{J}/\sigma. \]  

(8)

A physical oceanographer is likely to measure \( \nabla \phi \) to obtain \( \mathbf{v}^* \) in order to infer \( \mathbf{v} \), the customary oceanographic estimate of depth averaged velocity. With \( \mathbf{v} \) one can estimate transport per unit width \( x \):

\[ T(x) = \int_{-H}^{0} \mathbf{v}(z) dz = Hv \]  

and volume transport normal to a specified interval, say 0-\( X \), is:

\[ Q(X) = \int_{0}^{X} T \times d\mathbf{x} = \int_{0}^{X} \int_{-H}^{0} \mathbf{v}(x, z) dxdz = \int_{0}^{X} Hv dz \]  

(9)
where $v_y$ is the velocity component normal to the $x$ direction. Some of the problems in applying this theory is that the expressions are approximations (good ones but still approximations) and there are geophysical parameters, such as $\lambda$, which are not well known, especially in shallow water. Thus, the challenge is to make the best corrections for $\lambda$ and $\gamma$ possible from the available information about the bottom electrical conductances from marine geophysical studies or from the induction method itself. Much of our interest in preparing this paper is to acquaint our marine geology and geophysics colleagues with the need in physical oceanography for this mutually interesting information about the electrical structure of the sea floor.

The above expressions contain terms which depend on information or assumptions about bottom electrical conductance, the product of some average electrical conductivity and its average thickness. The $\lambda$ term has not often been determined or estimated. One method we have used is to measure $v(z) - V^*$ with a free fall EM and acoustic Doppler velocity profiler (Sanford et al., 1985). The combination of these data produces $V$ and $V^*$. For example, the ratio of $v_x$ and $v_z$ is equal to $(1 + \gamma_x)/(1 + \lambda).$ Of the two quantities, $\gamma$ will in general be different in each horizontal direction but $\lambda$ should be the same if the conductivity of the seabed is not anisotropic. Conventional wisdom is that $\gamma$ is small and the ratio yields an estimate of $(1 + \lambda)^{-1}$, which leads directly to $\lambda.$ Luther and Chave (1993) have made estimates of $\gamma$ based on archived hydrographic data from which $v'(z)$ is computed based on the geostrophic assumption and $\sigma'(z)$ is computed from temperature and salinity profiles. They conclude that except for strong baroclinic currents with strong vertical shear and strong vertical gradients of $\sigma$ that the $\gamma$ contribution is small. A form of it, $\sigma^2v'/\sigma$, is less than 0.01 m s$^{-1}$. On the other hand, it can be much larger in the Gulf Stream. By this process, it is possible that a strongly depth-dependent current with no net vertically averaged flow (known as a baroclinic current) can produce an apparent depth-averaged component (known as a barotropic current).

A recent modeling study (Tyler et al., 1997) focused on estimating the sensitivity of ocean-generated EM fields on ocean bottom conductance. The global ocean EM fields were calculated for idealized cases assuming very high and very low ocean-bottom conductance and the results were compared. Over most of the ocean, the results were essentially identical, suggesting that most of the electric current circulates within the ocean and precise knowledge of the sediment conductance is not required to obtain satisfactory oceanographic information. However, in some cases of important flow in regions of shallow or highly variable topography, the calculated EM fields were sensitive to the bottom conductance assumptions and at a minimum, simple first corrections such as that using $\lambda$ above are required.

### 4.3 Implementations of HEF on drifting and towed platforms

The depth-averaged velocity, $V^*$, is what is usually determined from stationary instruments measuring the horizontal electric field (HEF). If the measurement circuit is moving, as when installed on a drifting float, the measurement is $v(z) - V^*$. Independent measurements of $v(0)$ at the sea surface can be made using GPS and LORAN-C or other electronic navigation methods. Determinations of $v$ at the depth $z$ can be made from acoustic Doppler methods (e.g., AVP) and acoustic tracking (e.g., RAFOS floats). Examples of current and future implementations are:

**TTM:** a towed HEF sensor measures one or more components of the apparent HEF vector: $v_s - V^*$, where $v_s$ is the surface velocity of the measurement platform (generally assumed to be identical to that of the towing vessel). Once corrected for this velocity and windage, the result is one or more components of $V^*$ (Sanford et al., 1995a).

**AVP:** an Absolute Velocity Profiler is an HEF system installed on a free-fall profiler which measures $v(z) - V^*$ from the surface to the seabed in depths up to 6000 m (Sanford et al., 1985). An acoustic Doppler system determines the absolute velocity of the profiler as it approaches within about 500 m of the sea floor.
XCP and AXCP: an eXpendable Current Profiler is similar to the AVP without the acoustic Doppler correction. That is, it measures $v(z) - v^*$. These observations are transmitted to the sea surface via thin insulated wire (Sanford et al., 1982).

EFF: the Electric Field Float is an EM RAFOS float which is an HEF system that has been installed on the neutrally buoyant drifting RAFOS floats (Rossby et al., 1986). As the EFF drifts at its equilibrium depth, say 500 m, and rotates in the presence of internal waves, it measures the HEF (Sanford et al., 1995b).

VEFF: a Vertical EFF is a vertical electric field (VEF) system being considered for installation on neutrally buoyant floats. The purpose of this measurement is to determine the overlying relative vorticity of the ocean. The VEF on a Lagrangian platform arises from electric currents needed to supply the radially outward horizontal electric currents produced by ocean relative vorticity above the float depth. The VEFF can not rely on rotation for chopping the electrodes; a solenoid operated system (i.e., water switch) will be added to separate ocean VEF from electrode offset.

4.4 The isolation of $J^*$

The separation of the nonlocal electric current density $J^*$ from the total electric current density is a great challenge. The approach is generally to either assume it is small or make simple computations to set an expected upper bound. As HEF methods expand to wider applications, often ones that are not in strongly 2-D ocean currents where $J^*$ is theoretically small, better ways to estimate this quantity are needed.

For ocean flows having horizontal scales much greater than their depth, the induced electric potential is largely independent of depth, and determining $J^*$ is largely equivalent to determining the depth averaged electric currents $\overline{J}$ (averaged between the sea surface and depth $H_s$). Also, in the neglect of time rates of change of the electric charge density, the electric current is nondivergent and $\overline{J}$ can be written in terms of a stream function $\psi$ (see the Appendix). This approach has been used to obtain information about the importance of the nonlocal electric currents in global ocean numerical models (Stephenson and Bryan 1992; Tyler et al., 1997). The focus of the Tyler et al. study, aside from studying the importance of ocean bottom conductance, as described above, was to show that using the ‘Sanford approximation’ a simplified set of equations could be used to easily produce the dominant ocean-generated electromagnetic fields without running a globally integrating model. A globally integrating model was, however, also run to justify this claim. From the latter results, the effects of nonlocal electric currents were estimated and it was found that these effects were generally not bigger and typically much smaller than those due to the bottom conductance. Hence corrections for nonlocal electric currents do not appear to help the oceanographic interpretations until reasonably realistic models of ocean bottom conductance are included (but see related discussion below in Conclusions section).

5. Conclusions

We have attempted to give a brief overview of the objectives and methods used in oceanographic motional induction studies and to describe some of the current challenges. Our view is that the expertise of colleagues in other areas of marine geology and geophysics could be increasingly utilized in facing these challenges. On the other hand, instrumentation developed for oceanographic use could be utilized in other applications such as the study of the conductivity of the sea floor.

Probably the highest current priority for the motional induction community is to gain a better understanding of how to correct for the nonzero conductivity of the sea floor. At a minimum, this entails having an estimate for the conductance of a layer below the ocean extending from the sea floor down to a depth potentially as great as the horizontal scale of the overlying ocean flow.
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(which could reach thousands of kilometers). Typically, however, it has been assumed that most of the ocean bottom conductance is due to a relatively thin and conductive sediment layer. In this case the conductance value would not be dependent on the scale of the flow. The need for this information is greatest for studies involving shallow regions or strongly depth independent flows such as the Florida Current or Antarctic Circumpolar Current.

Toward this goal, it would be extremely useful to create a global (digitized) description of ocean/sediment conductance. The conductance for the ocean layer has already been calculated and presented by Tyler et al. (1997). The conductance of the sediment layer might be estimated from geological data. Even if such a global description included areas of large uncertainty, such work could still be equally useful (provided the uncertainty levels were also described) since many of these uncertain areas may coincide with areas of the deep ocean for which precise estimates of bottom conductance are not needed for the oceanographic applications. Another approach might be to use additional instrumentation to gain an independent estimate of bottom conductance during the oceanographic measurements.

The next priority is probably to develop methods for understanding and estimating the influence of the nonlocal electric currents. As we have mentioned though, the global accuracy of calculations of the nonlocal electric currents will depend on the realism of the ocean bottom conductances assumed.

The order of priorities given above is rather general and descriptive of the global scale problem. For a regional observational study, it may be that either the bottom conductance or nonlocal effect, or both, may be unimportant in creating errors when calculating flow from the electric field observations. In fact, an important motivation for better understanding these two effects is so that we can better choose the regions where oceanographic EM methods are the most appropriate. That is, the oceanographic EM methods are simplest when no corrections are needed for the bottom conductance (e.g., in much of the deep ocean where the sediment conductance is a small fraction of the oceanic conductance (Tyler et al., 1997)) or nonlocal currents are zero.

At the next level of sophistication, corrections can be made for these effects using data and/or numerical results. Even when accurate numerical results are limited by insufficient knowledge of parameters such as the bottom conductance, the numerical simulations can provide limiting cases which give very useful information about the errors associated with the flow estimates.

The nonlocal electric currents should be better understood so that allowance for their effects in ocean electric field observations can be made. But this is not the only reason. Another important reason for studying the nonlocal electric currents is that these currents are related to and in many cases entirely responsible for the EM fields observable outside of the ocean and at satellite altitudes. A full evaluation of the potential for using EM measurements to remotely sense ocean variability will require a better understanding of these electric currents.

Appendix: Effects of Nonlocal Electric Currents

Equation for ocean-generated electropotential

A 2-D equation for the electric potential $\phi$ (which for steady or low-frequency cases describes the horizontal electric field by $E_H = -\nabla \phi$) can be obtained by taking the divergence of Ohm's law (1) divided by the conductivity $\sigma$ and integrated through the depth $h$ which is viewed as a thin layer including the ocean and underlying conductive sediments. Global calculations using this approach appear in two previous works (Stephenson and Bryan, 1992; Tyler et al., 1997) to which we refer the reader for further details. The electric potential equation is

$$\nabla \cdot \left( \Sigma \{ \nabla_H \phi - F_z \mathbf{u} \times \mathbf{n} \} \right) = 0,$$

(A.1)
where \( \Sigma = \int_0^h \sigma dz \) is the electrical conductance (S) of the layer \( h \), \( F_z \) is the vertical component of the Earth's main magnetic field, and \( \bar{u}^* = \Sigma^{-1} \int_0^h \sigma u dz \) is the conductivity-weighted depth-averaged ocean current velocity (m/s).

**Stream function for the nonlocal (depth-averaged) electric currents**

If we consider the ocean-induced electrical currents to be concentrated in a thin surface layer \( h \) comprising the ocean and conductive underlying sediments, we can write the horizontal components of the depth-integrated electrical current density as

\[
\int_{-h}^0 J_H dz \approx \Sigma (\nabla \phi + \bar{u}^* \times F_z). \quad (A.2)
\]

In steady, or even quasi-steady cases, \( J \) is non-divergent and a stream function \( \psi \) can be defined as

\[
\int_{-h}^0 J_H dz = -\nabla \psi \times \hat{n}, \quad (A.3)
\]

where \( \hat{n} \) is the unit vector normal to the sea surface and directed outward (up). By taking the curl of \( (A.3)/\Sigma \) and using Eq. (A.2) together with standard vector identifications, we can produce the following governing equation for \( \psi \):

\[
\nabla \cdot (\Sigma^{-1} \nabla \psi + \bar{u}^* F_z) = 0. \quad (A.4)
\]

Equation (A.4) was solved using extensions to the OPYC code. The solution technique is similar to that for the electric potential equation and has been described recently in other work (Tyler et al., 1997). The solution for \( \psi \) is shown in Fig. 13. The electric currents described by the streamfunction \( \psi \) can usually be viewed as second order compared to the dominant electric currents typically occurring in the regions of strong flow near the sea surface. These nonlocal electric currents are extremely important nonetheless for at least two reasons: first, before oceanographers can associate EM observations with ocean flow, the nonlocal electric currents must be estimated—even if simply to show that they can be neglected; second, the nonlocal electric currents are responsible for the magnetic fields reaching outside of the ocean. In principle, these nonlocal magnetic fields might be used to remotely sense ocean dynamics. The strengths of these magnetic fields are calculated in the next section.

**Ocean-induced magnetic fields outside ocean**

In this subsection, we will estimate the magnetic fields due to the ocean-induced electric current sources in the thin layer \( h \). In particular, it is interesting to estimate these fields at the altitudes at which magnetic satellite surveys are conducted.

We will assume here that the region above sea level is electrically insulating. Hence, there are no electrical currents (neglecting advection of spatial charge densities), and since in this region

\[
\nabla \times \mathbf{b} = \mu_0 \mathbf{J} = 0, \quad (A.5)
\]

we can write the magnetic field \( \mathbf{b} \) as the gradient of a potential \(-P\)

\[
\mathbf{b} = -\nabla P. \quad (A.6)
\]

Also, since magnetic fields are nondivergent (\( \nabla \cdot \mathbf{b} = 0 \)), the divergence of Eq. (A.6) gives the governing equation

\[
\nabla^2 P = 0 \quad (A.7)
\]
in the region above sea level. Natural boundary conditions for $P$ are used for the upper boundary ($P(r \to \infty) = 0$). The lower boundary condition is specified by $\psi$ in the following.

We can integrate Eq. (A.6) over the thin layer $h$ and use Eq. (A.3) to write

$$\delta b_H \approx \mu_o \nabla \psi$$  \hspace{1cm} (A.8)

where $\delta b_H = b_H(z = 0) - b_H(z = -h)$ is the difference in the magnetic field between the sea surface and depth $h$. The approximation sign in Eq. (A.8) is required because effects of the small vertical electrical currents within the layer have been neglected. For the purposes here, we can treat the layer $h$ as uniform, giving $b_H$ of equal and opposite values above and below the layer. Then the ocean-induced magnetic anomaly at the sea surface can be estimated as

$$b_H(z = 0) = \frac{1}{2} \mu_o \nabla \psi.$$  \hspace{1cm} (A.9)

Comparing Eq. (A.6) with Eq. (A.9) it appears sufficient to take as the lower boundary value for $P$:

$$P(z = 0) = -\frac{1}{2} \mu_o \psi.$$  \hspace{1cm} (A.10)

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