

MODELLING OF NATURAL SOURCES OF MAGNETOSPHERIC ORIGIN IN THE INTERPRETATION OF REGIONAL INDUCTION STUDIES: A REVIEW

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Abstract. Traditional interpretation of regional induction studies relies on the assumption that the external inducing field is quasi-form. My review is devoted to as clear as possible a definition of the limitations of this assumption, which is not always justified. Therefore, I will first systematically describe the actual current systems responsible for induction at high, middle or low-latitudes, and then review the means presently available for taking their localised nature into consideration when interpreting surface electromagnetic data. Realistic source models will be presented for all latitudes.

Résumé. Les méthodes géophysiques traditionnelles basées sur l'induction électromagnétique régionale utilisent toutes l'hypothèse de base suivant laquelle le champ externe inducteur est quasi-uniforme. Puisque cette hypothèse n'est pas toujours justifiable, ma revue a pour but de définir ses limites aussi clairement que possible. Pour ce faire, je décrirai d'abord systématiquement les véritables systèmes de courants responsables de l'induction à hautes, moyennes ou basses latitudes, pour passer ensuite aux méthodes permettant de prendre en considération leur étendue limitée lors de l'interprétation d'observations de surface. Des modèles réalistes de sources seront présentés pour toutes latitudes.

1. Introduction

The problem of assessing the effects of the localized nature of most natural sources on the interpretation of regional induction studies has plagued geomagneticians for many years. Much hand-waving has been done – the dimensions of the external field being somehow ‘large enough to justify the plane wave limit’ or the bizarre behavior of some observations being ‘possibly due to the closeness of a localised source’ – however much theoretical work has also been initiated, most of it being based on the relative importance of some ‘typical’ dimension of the source field and of some sort of ‘attenuation depth’ characteristic of the region considered. More specifically, the magnitude of the source ‘dominant’ horizontal wave number $k = 2\pi/\lambda$ is compared to that of $(\omega\sigma\mu_0)^{1/2}$ for a stationary source or to that of $(|\omega + \mathbf{k} \cdot \mathbf{v}| \sigma\mu_0)^{1/2}$ for a source travelling with a uniform velocity \mathbf{v} . Of course, neither the conductivity distribution of the region, here loosely represented by the symbol σ , nor the dimensions of the source, are known a-priori, and therefore various astute techniques have to be developed to infer both of them from ground based measurements alone.

The general solution of an induction problem necessitates the knowledge of what is commonly known as the TE and TM modes of the source, i.e., of that part of the external current system which only flows parallel to the surface of the conductor (TE)

and of that part which is closed in current loops perpendicular to the conductivity discontinuity (TM). In the one-dimensional (1D) problem, i.e., if the earth can be treated as a horizontally layered half-space, the TE and TM modes are decoupled. All induction is due to the TE mode, the TM mode consisting of currents impinging vertically down on the earth in what Dmitriev and Berdichevsky (1979) call a 'galvanic' manner. The amount of current flowing inside a 1D earth because of that galvanic interaction is negligible compared to the amount of current induced by the TE mode (e.g., Dmitriev and Berdichevsky, 1979; Vanyan and Butkovskaya, 1980) and is thus usually totally ignored. Since it is common practice to consider the earth as laterally homogeneous on a global scale, it is also common practice to assert that surface magnetic observations alone do not allow the discrimination of the three-dimensional (3D) nature of the source from an equivalent system of currents flowing on a spherical shell concentric with the earth.

Indeed, it is true that in regions where the earth can be treated as a horizontally stratified conductor, all induced currents, regardless of the actual geometry of the source, flow in horizontal planes and, thus, an ionospheric 'equivalent' current system is sufficient to represent the fields relevant to the induction process. However, even in this simple case, an exact representation of the equivalent current system is not often easy to achieve either in the wave number or space domain, and, at short periods (T less than a few hours) is only performed satisfactorily at mid-latitudes. The definite source effects observed in the equatorial and auroral regions at these periods are traditionally represented by the effects of ionospheric electrojets which, if they give reasonably good estimates in many cases (in particular in the frequency space domain), can also be quite misleading when one has to deal with high-latitude current systems able to develop in surges on horizontal scales as small as a hundred km.

It is also true that even when the region of interest includes lateral discontinuities in conductivity (either 2 or 3D structures), the problem of induction is usually solved numerically by determining the local perturbation (due to the anomalous domain) to the field that would be observed over a horizontally layered earth, i.e., that even though the solution of the general problem of induction does not allow the decoupling between TE and TM modes, in practice, approximate solutions can usually be obtained by evaluating the source term over a normal structure, thus by considering an equivalent ionospheric source over a 1D earth.

But how can an equivalent local ionospheric source be defined without knowledge of what the real source might be? By spherical harmonics analysis? By Fourier spectral decomposition? One soon finds out that these approaches are very seldom satisfactory, especially at high-latitudes. For that reason, and because of the tremendous amount of information that has recently been gathered about external current systems from a variety of observations other than ground-based magnetic observations, I have decided to begin this review by presenting models as simple and as comprehensive as possible of the actual sources involved in regional induction work. I will then proceed to the representation of those sources, as conceived by the

induction community, and review the various methods that have been developed to take their localised nature into account over a horizontally stratified conductor. I will not consider the extension to laterally heterogeneous earth models since this would require putting at least as much weight on the numerical techniques than on the modelling of the sources themselves (see Kaikkonen, 1986). Work in that respect has been pursued by several authors, e.g., Hibbs and Jones (1973, 1978), Hughes and Wait (1975a, b), Quon *et al.* (1979), and Hermance (1984).

2. Basic Techniques of Observation of External Current Systems

2.1. MAGNETIC OBSERVATIONS AT SATELLITE HEIGHTS

Satellite-borne magnetometers should, in principle, provide a global magnetic mapping of the earth's environment. However, the high relative velocity of the spacecraft, combined with the small scale size of the systems of interest, enforces sampling windows of the order of 10–20 s for hundred km scale sizes, thus providing nothing more than snapshots of the in-situ magnetic field. The patching together of such snapshots has led to the well-known statistical distribution of permanent large scale field-aligned currents in high latitude regions. Figure 1 illustrates the flow of those currents as they move into and out of the auroral ionosphere, roughly defining two rings of sheet current associated with magnetospheric convection and present in ionosphere under almost any external conditions. 'Region 1', the more poleward and primary ring, consists of current flowing into the ionosphere on the dawnside and out on the duskside whilst 'region 2', created with a time lag of the order of an hour to provide shielding of the lower latitudes, has the opposite polarity. Note that the rings move equatorwards during disturbed times. Typical average current strengths (based on TRIAD observations) are often of the order of at least a million A in region 1, whereas for region 2 currents typically range from 0.5 million A during quiet periods to several million A for disturbed times (e.g., Nopper and Carovillano, 1978). These observed values are consistent with the parameters obtained by joint ground-based and satellite magnetic observations of specific events (i.e., Sulzbacher *et al.*, 1980, 1982; Rostoker *et al.*, 1982; Rostoker and Mareschal, 1982).

Satellite observations of pure polar cap field-aligned currents also exist but cannot be averaged as easily as the region 1 – region 2 'Birkeland' currents since they are highly dependent on the direction as well as on the intensity of the interplanetary magnetic field (IMF). Quiet days have been analysed which don't show much departure from the double ring system already defined (e.g., Zanetti *et al.*, 1983; Burrows *et al.*, 1984), yet days of extremely northward IMF have displayed a third ring poleward of region 1 and of opposite polarity (e.g., Iijima, 1984). Pure polar cap field-aligned currents are so variable that any 'statistical distribution' must be given in terms of specific IMF configurations (e.g., Friis–Christensen, 1984).

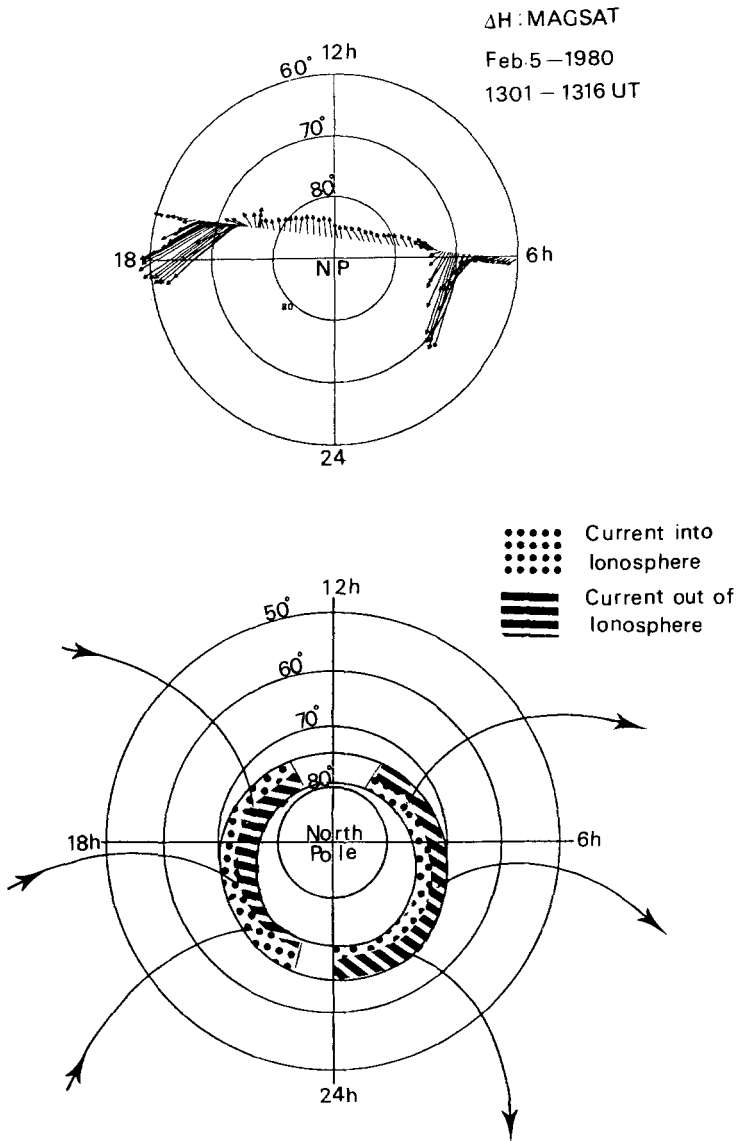


Fig. 1. Example of residual horizontal magnetic field along a satellite orbit – MAGSAT in this case – (top) and the schematic average field-aligned current distribution obtained from such observations – the complex noon and midnight regions being deliberately deleted – (bottom). (Adapted from Burrows *et al.*, 1984, and Rostoker, 1979. The coordinates are eccentric dipole latitude and magnetic local time).

Mid- and low-latitude observations of field-aligned currents by distant satellites are basically non-existent; this of course is partly due to the geometry of the magnetic field in those regions. However the equatorial electrojet has been observed, for instance, from the low orbiting satellites POGO (e.g., Oni, 1973; Onwumechili and Agu, 1982), and COSMOS 321 (e.g., Vanyan *et al.*, 1975). At this point it is

interesting to note that low altitude satellites provide useful information on ionospheric as well as field-aligned currents, but are contaminated by induced contributions. These have to be simulated in the interpretation of data, usually by the effect of a perfect conductor placed at some depth below the earth's surface.

2.2. RADAR TECHNIQUES

Radar techniques basically allow the observation of plasma irregularities, and eventually of drift velocities, in the ionosphere. They can be classified into two main types depending on whether they measure the incoherent or the coherent backscatter.

In the incoherent scatter approach, electrons, struck by very high frequency signals (50 to 1000 MHz, power of the order of several MW) emit an echo back individually (and independently of one another). From the distribution of these individual waves, it is possible to infer the ion distribution and all related parameters. In particular, the Doppler shift can be determined and the ion drift velocities can be obtained leading eventually to an evaluation of ionospheric currents. A very comprehensive review of the application of this method to the definition of various ionospheric electrodynamic parameters is given by Blanc (1979). The calculations necessarily require some averaging and simplifying assumptions, but at least the ionospheric current defined in this manner is not contaminated by either field-aligned or induced effects! The incoherent scatter technique is currently being used in all major latitude ranges (e.g., Chatanika, Alaska; Millstone Hill, U.S.A.; Saint Santin, France; EISCAT, Northern Scandinavia; Arecibo, Puerto Rico; Jicamarca, Peru). In conjunction with ground based magnetograms, it has helped define both the auroral (e.g., Kamide and Brekke, 1975) and equatorial (e.g., Crochet *et al.*, 1976) electrojets as well as the *Sq* system (e.g., Mazaudier and Blanc, 1982).

The coherent backscatter technique, on the other hand, is mainly used to define instabilities usually collocated with the electrojet. In this technique it is the signal emitted by the radar itself (same frequency range, but power transmitted of the order of several kW) that is reflected by the bulk of the drifting irregularity (see Greenwald, 1979, for a general presentation of the technique). The simplest type of measurement shows that the amplitude of the backscatter signal is often proportional to the electrojet current density (e.g., Baumjohann *et al.*, 1978; Mareschal *et al.*, 1979; Uspensky *et al.*, 1983), whilst dual radar systems (e.g., STARE in Scandinavia) permit the determination of the electric field at different horizontal ranges, i.e., for any 2D structure (e.g., Greenwald, 1979). This information, coupled with information obtained from ground based magnetometers, permits the definition of electrojets and of field-aligned currents (Baumjohann *et al.*, 1980; Baumjohann and Kamide, 1981; Andre and Baumjohann, 1982).

2.3. OTHER TECHNIQUES

The best information on purely external systems is, of course, given by radar observations coupled to satellite data and ground-based magnetometers (e.g., Robinson, 1984). However, it is certain that the use of other techniques (rockets,

TABLE I

Observational methods which contribute to studies of the magnetospheric currents
(after Akasofu, 1984)
(J_{\parallel} (J_{\perp}): current density parallel (perpendicular) to \mathbf{B})
(Σ : ionospheric conductivity tensor)

	Measured quantity	Inferred quantity
<i>Ground-Based</i>		
Magnetic obs.	ΔB	$J_{\parallel}, J_{\perp}, E, \Sigma$
Radar obs.	Ion drift speed Electron density	Σ, J_{\perp} (ionosphere), J_{\parallel}
<i>Satellites and Rockets</i>		
Magnetic obs.	ΔB	J_{\parallel}, J_{\perp}
Electric obs.	E	
Ion drift speed obs.	Ion drift speed	E
Particle flux obs.	Particle fluxes	J_{\parallel}, J_{\perp}
<i>Balloon</i>		
Electric field obs.	E	E (ionospheric)

ballons, etc.) in conjunction with these can only improve our understanding of external systems (see Richmond (1983) for a clear and succinct review of those systems related to our understanding of ionospheric electrodynamics). Examples of multi-instrumental studies are now very common (e.g., Pellinen *et al.*, 1982) but their detailed description is beyond the scope of this review. The parameters that they measure, and their relation to current modelling, are summarized in Table I.

3. The Definition of the Ionospheric Conductivity Tensor

At this point, it is worth clarifying the definition of the ionospheric conductivity tensor since the different directional conductivities play a major role in defining ionospheric current concentrations. A simple exposé of the subject can be found in Parkinson (1983).

Basically, the ionospheric conductivity in a given direction will depend upon the local plasma density and upon the directional mobility of the ions and electrons that it contains. It is customary to talk of a 'direct' conductivity, i.e., the conductivity in a direction parallel to both \mathbf{E} and \mathbf{B} which can only exist if there is a voltage drop along magnetic field-lines and which generates, *a priori*, field-aligned currents. If $\mathbf{E} \perp \mathbf{B}$, one defines the 'Pedersen' conductivity σ_P corresponding to current flowing in the direction of \mathbf{E} (and thus \perp to \mathbf{B}). Note that since $\mathbf{J} \cdot \mathbf{E} > 0$, Pedersen currents involve energy dissipation and therefore must have a generator of either magnetospheric or atmospheric origin. The 'Hall' conductivity σ_H regulates the current flowing at right angles to both the electric and magnetic fields. In this case $\mathbf{J} \cdot \mathbf{E} = 0$, and these currents are not dissipative in the ionosphere. Because of

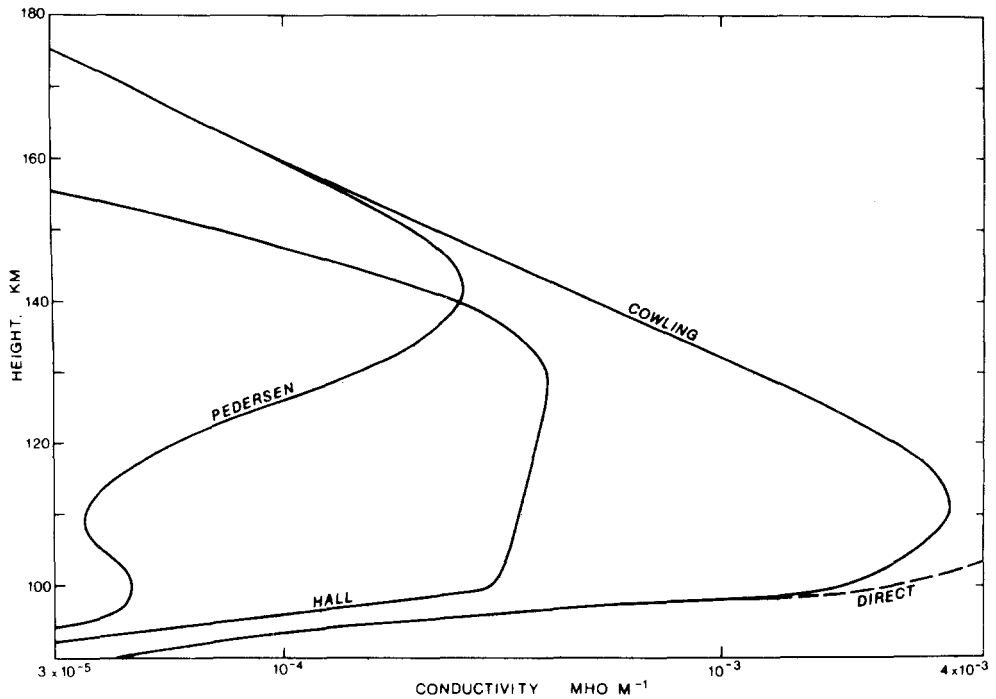


Fig. 2. Direct, Pedersen, Hall and Cowling conductivities based on a typical day-time ionosphere (Parkinson, 1983, p. 242).

horizontal discontinuities in conductivity, a polarization E -field can be created which is orthogonal to both \mathbf{E} and \mathbf{B} , and the flow of current is, in that case, regulated by what is called the 'Cowling' conductivity, σ_c , where $\sigma_c = (\sigma_H^2 + \sigma_P^2)/\sigma_P$ in the case of the electrojet. Figure 2, taken from Parkinson (1983), gives typical daytime ionospheric conductivities based on calculated distributions of plasma densities and ion-neutral particle collisions at mid-latitudes. Such models are not valid within the electrojets themselves, where the conductivity must be defined empirically from radar and rocket measurements of energetic electrons (see e.g., Reiff, 1984, for a recent review of models of auroral conductivities and Forbes, 1981, for an excellent review of ionospheric parameters including conductivity models for the equatorial electrojet). The problem of defining night-time conductivities at mid-latitudes (e.g., Richmond, 1983, and references therein) will not be treated here.

4. Definition of the Basic Current Systems Regulating the Diurnal and Quasi-Regular Variations

Magnetic variations at all latitudes exhibit effects of diurnal variations on which are superimposed a variety of disturbance fields. A pure diurnal variation is, by definition, essentially due to DC magnetospheric currents fixed in inertial space, and

thus give rise to apparently oscillating fields at a given geographical position as the earth rotates within the magnetosphere. However, observed diurnal variations are never pure but reflect the combined effects of various current systems, some of which are affected by magnetospheric and/or interplanetary conditions. These are what I will call the ‘quasi-regular’ variations of which the Sq field at mid-latitudes, the electrojet field in the equatorial regions and the convection field at high-latitudes are good examples. All other variation fields, which are dependent on the IMF conditions and/or are intrinsically temporal will be considered in Section 5.

4.1. THE MID-LATITUDE VARIATIONS

The Sq system is the major contributor to the quasi-regular variations at geomagnetic latitudes below 50° . What is usually thought of as its major component is the effects of atmospheric winds displacing charged particles across geomagnetic field lines, and thus generating electric fields and currents where the ionospheric conductivity is large enough, i.e., in the dayside ionosphere (see e.g., Matsushita, 1967, for an exhaustive review of the Sq system, and Richmond, 1979, for a review of the dynamo theory). However, it is important to realize that field-aligned currents in the high-latitude regions provide an important coupling between magnetospheric and ionospheric processes, and that calculations based on TRIAD observations yield, even at quiet times, electric fields at all latitudes which are comparable in strength to those produced by dynamo wind models and which have non-diurnal characteristics (e.g., Nopper and Carovillano, 1979; Mishin *et al.*, 1979). These introduce definite temporal characteristics to the Sq system. Nevertheless, most of the current still appears to be concentrated in the ionosphere E -region, and the ‘real’ Sq system might thus be quite similar to the equivalent 2D representation given in Figure 3. Clearly, its essentially global character makes it of little direct interest to this review except through the vital role that it plays in defining the equatorial, and, to a lesser extent, the polar, diurnal variations.

4.2. THE LOW-LATITUDE VARIATIONS

The existence of the daytime equatorial electrojet can essentially be explained by an extension of the Sq electric field to the equatorial regions where \mathbf{E} is basically parallel to the equator. In those region, \mathbf{B} is virtually horizontal, and thus an effective Cowling channel is created along the (dip) equator with its characteristic enhanced conductivity and currents. Much has been written on the morphology of the equatorial electrojet (e.g., Onwumechilli, 1967; Fambitakoye, 1973, 1976; Fambitakoye and Mayaud, 1975, 1976a, b) whose regular variations, S_R^E , are superimposed on the regular global Sq (S_R^P) along the magnetic dip equator. The basic points of all relevant observations, as well as of the major models that have been constructed prior to the early 1980’s to represent its surface magnetic signature, are given in an extremely clear and thorough review by Forbes (1981) where all references of interest can be found. Note that the equatorial electrojet current displays longitudinal gradients, with regional and seasonal variations. Such variations are

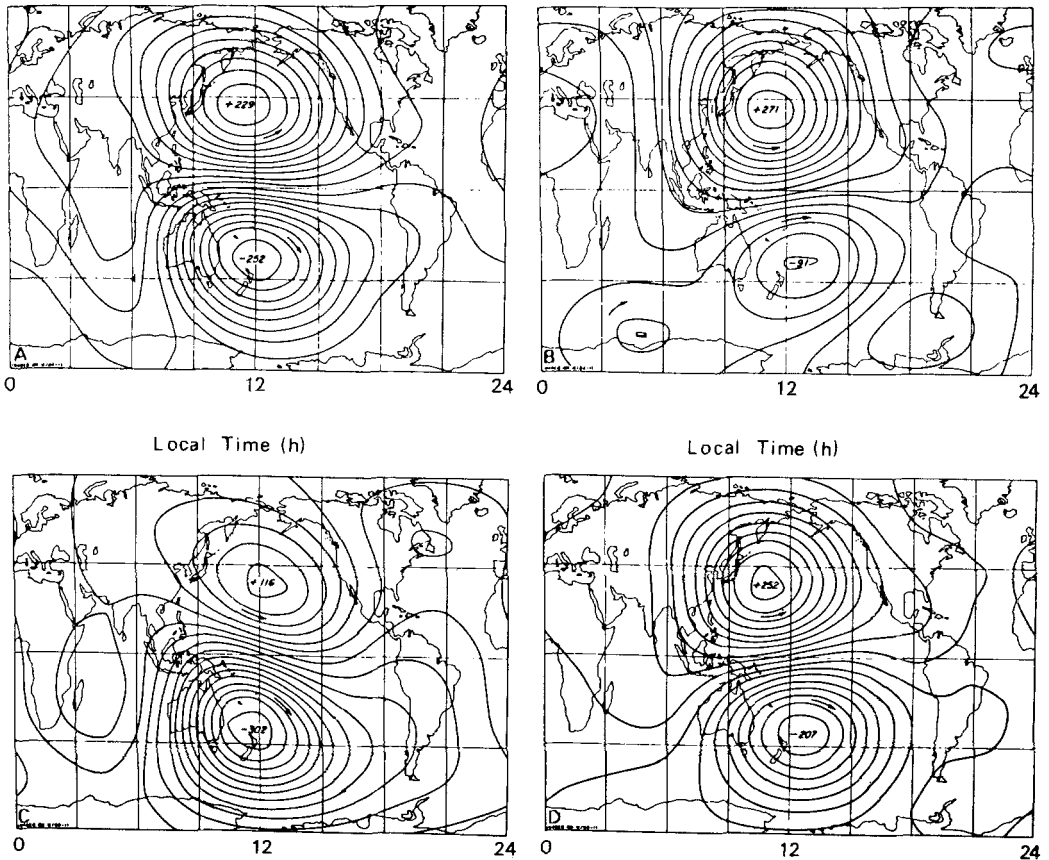


Fig. 3. Examples of current functions corresponding to the S_q field for March equinox, northern summer, September equinox and southern summer (clockwise, starting on top left). The contour intervals are approximately 25 kA. (Adapted from Parkinson, 1983, pp. 272–273).

known to be strongly influenced by the asymmetric way in which the S_q vortices intrude on the equatorial regions in the northern and southern hemispheres and must therefore be reproducible by any realistic electrojet model. A good model must also be able to generate the disturbance fields associated with the electrojet, in particular the ‘counter electrojet’, i.e., that narrow band of westward current frequently appearing on either sides of the dip equator in the morning (0600–0800 LT) and afternoon (1400–1800 LT) sectors (Mayaud, 1977), without, of course, requiring enormous amounts of computer time to reproduce the surface magnetic observations.

Richmond’s (1973a, b) model, which is based on measured ionospheric electric fields and neutral wind velocities, belongs to that class. It considers the variation in ionospheric conductivity with latitude and its related generation of vertical currents (e.g., Untiedt, 1967). In this model, it is the curvature of the magnetic lines of force that define the N–S extent of the electrojet, and no restriction is made on current

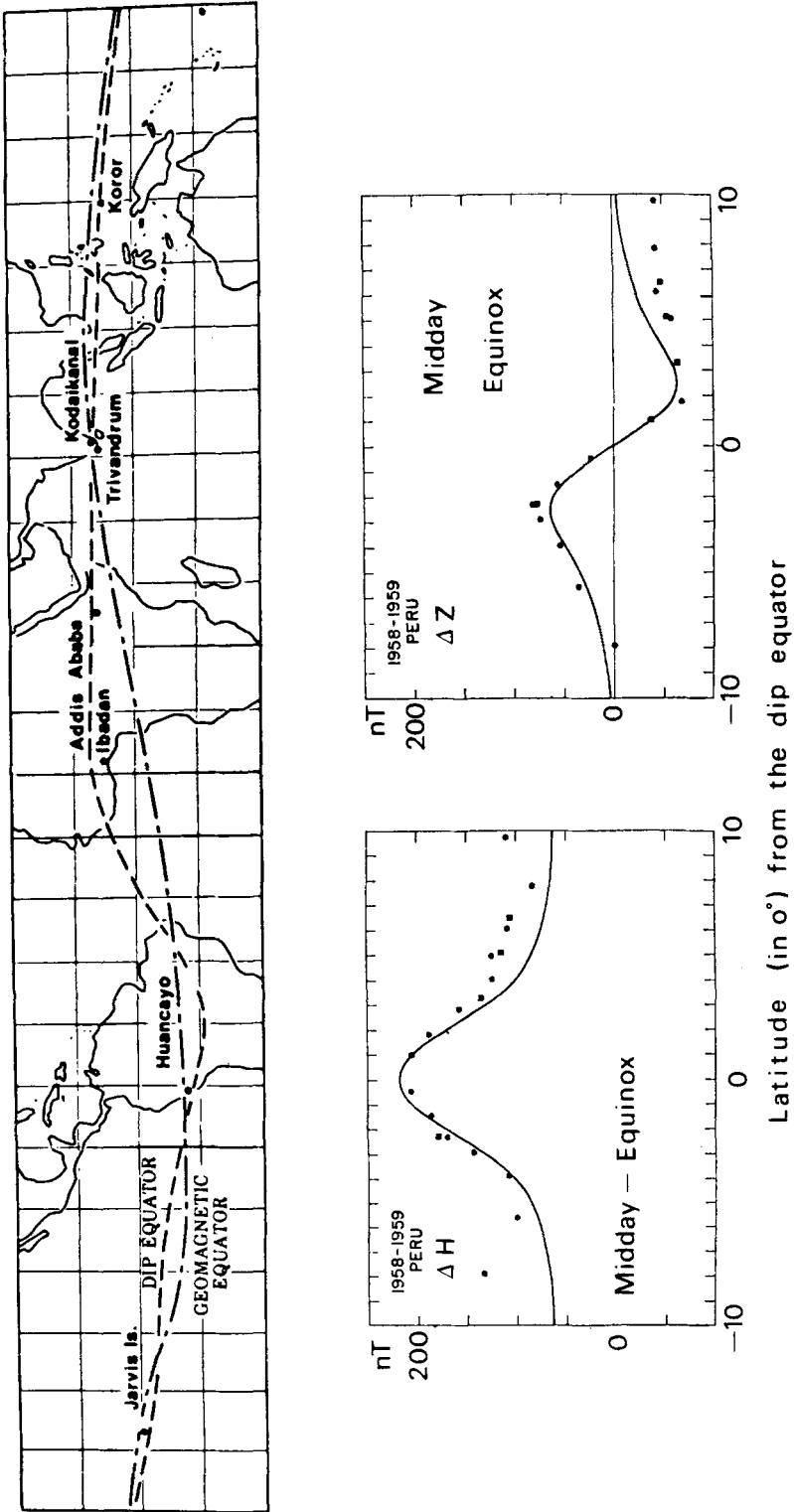


Fig. 4. Definition of the dip equator (top) and comparison between observed (Forbush and Casaverde, 1961) and computed (Richmond, 1973a) equatorial electrojet surface fields (bottom). (Adapted from Gupta, 1973 and Richmond, 1973b).

flow (ionospheric, along the field lines or in the vertical magnetic meridian plane) other than to prevent flow across the boundaries between conducting and non-conducting regions. Figure 4 illustrates the accuracy with which Richmond's model reproduces the surface magnetic variations from measured ionospheric parameters (see also Fambitakoye *et al.*, 1976; Gagnepain *et al.*, 1977). The model can also simulate two-stream and gradient drift plasma instabilities within the electrojet as well as the westward electric field responsible for the afternoon counter-electrojet (e.g., Marriott *et al.*, 1979), i.e., the major disturbance variations observed at low-latitudes.

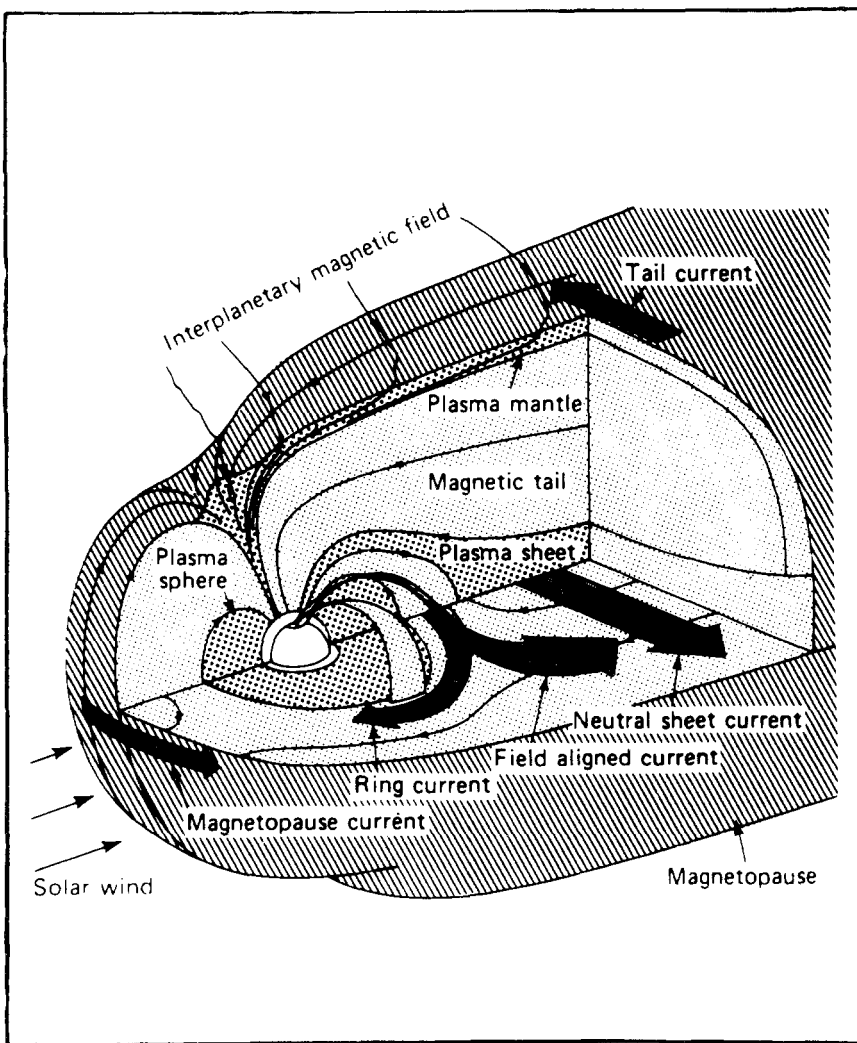


Fig. 5. A schematic view of the magnetosphere. (Cover page of EOS Trans. 65, January 31, 1984).

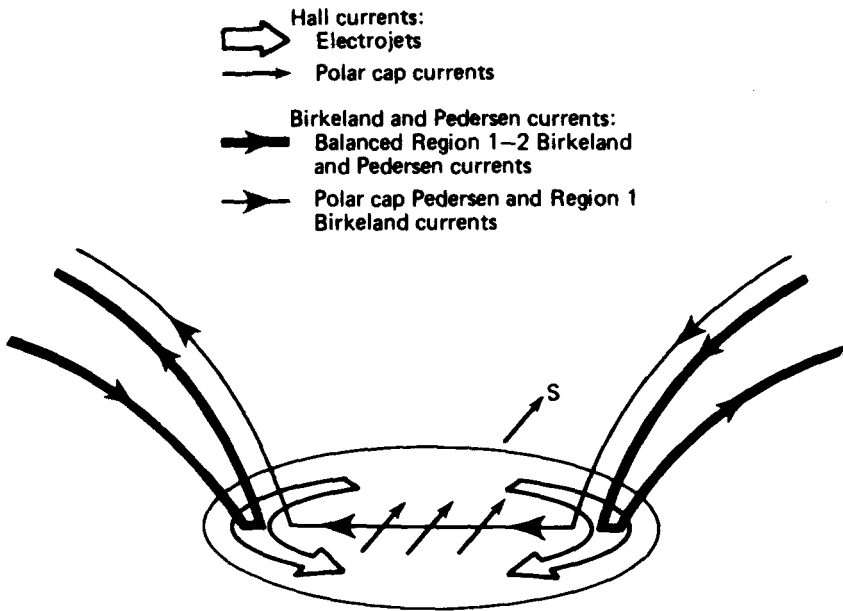


Fig. 6. Possible interaction between field-aligned currents, convection electrojets and cross-polar cap current flow in the sunlit polar ionosphere (Zanetti *et al.*, 1983).

4.3. THE HIGH LATITUDE VARIATIONS

The problem of extending the quasi-regular Sq variations to the polar regions is exacerbated by the ever present effect of various current flows generated by different physical processes mapping via field lines to different parts of the magnetosphere. For instance, the current flow in the nightside auroral oval (basically the convection and substorm electrojets) is governed by processes taking place in the plasma sheet, whilst most of the current flowing near the polar cusps and in the dayside polar cap is connected to processes taking place in the magnetopause (see Figure 5) (e.g., Rostoker, 1979; Baumjohann, 1983; Baumjohann and Friis-Christensen, 1984). It is in the high latitude regions that the separation on physical grounds between quasi-regular and irregular – or disturbed fluctuations – is most artificial, and it is accordingly rather arbitrarily that I have decided to consider in this section only the S_q^p and convection electrojets system.

Indeed, in these regions, the Sq system has been traditionally complemented by a 2D equivalent system called the S_q^p consisting of currents crossing the polar cap towards the sun and closing at lower latitudes. Many aspects of the generation mechanism of the real S_q^p system remain elusive, but its relation to solar wind velocity is now widely acknowledged (e.g., Mishin *et al.*, 1979; Baumjohann and Friis-Christensen, 1984). Since this system appears to be independent of IMF conditions, it has been attributed to viscous interaction between the solar wind and magnetospheric field lines. This interaction does not seem to be restricted to closed field-lines which means either, in Baumjohann and Friis-Christensen's words, "that

the viscously driven ionospheric currents create polarization charges in the ionosphere (due to conductivity gradients between the polar cap and auroral zone) which are accompanied by a dawn-dusk potential difference” (and thus creates the sunward Hall current across the polar cap as modelled by Nagata and Kokubun, 1962; Tvasaki and Nishida, 1967; Mareschal, 1974), “or that viscous interaction does not only take place on closed low-latitude boundary layer field-lines but also on open lines in the high-latitude regions”. All Hall currents (which is that part of the real current system coming closest to its equivalent representation) are always associated with accompanying Pedersen and ‘Birkeland’ – or field-aligned – currents. Consequently, the real S_q^p currents are probably part of a 3D system connected to the auroral (or convection) electrojets possibly via the region 1/region 2 field-aligned currents.

A simple model of the possible interaction, illustrated by Zanetti *et al.*(1983), is reproduced in Figure 6. Within the electrojets themselves, the ionospheric current density \mathbf{J} is given by

$$\mathbf{J} = \sigma_P \mathbf{E}_\perp + \sigma_H \frac{\mathbf{B} \times \mathbf{E}_\perp}{|\mathbf{B}|},$$

where \mathbf{E}_\perp is the component of the electric field orthogonal to \mathbf{B} . The conductivity in the auroral zone being primarily due to precipitation of electrons energized in the magnetosphere, the conductance has a minimum in the morning sector (7–10 S) and a maximum in the midnight sector (10–20 S) (e.g., Baumjohann, 1983). The electric field pattern always reflects the large scale magnetospheric convection, \mathbf{E}_\perp being primarily poleward in the region of the eastward electrojet (afternoon local time) and equatorward in the region of the westward electrojet (morning sector). It follows that the E–W electrojet current flows are basically Hall currents, whilst the flow connecting region 1 and region 2 Birkeland sheets across the electrojets is due to Pedersen currents. Across the polar-cap, a dawn to dusk Pedersen current, fed by region 1 field aligned currents, may be superimposed upon the simple sunward Hall current first modelled in early studies.

It must be clear, at this point, that the quasi-regular variations defined here are parts of larger physical phenomena. However, all that probably matters in terms of modelling high-latitude localised sources for induction studies, is to know that most current systems above 55° of geomagnetic latitude, regardless of their intrinsic sophistication or level of disturbance, can be represented by various combinations of the two basic current loops shown in Figure 7 (after Kisabeth and Rostoker, 1977). Obviously, the important input parameters are the local ionospheric electric fields and the ionospheric conductivities. From these, the relative importance of Pedersen to Hall currents can be estimated, and thus the relative contributions of the two loops assessed. Then their contributions to the surface magnetic field can be evaluated using Biot–Savart’s law, and an adequate combination of elementary loops found to reproduce the ground magnetic observations. The modelling techniques presented by Kisabeth and Rostoker (1977), Hughes and Rostoker (1979), Rostoker and

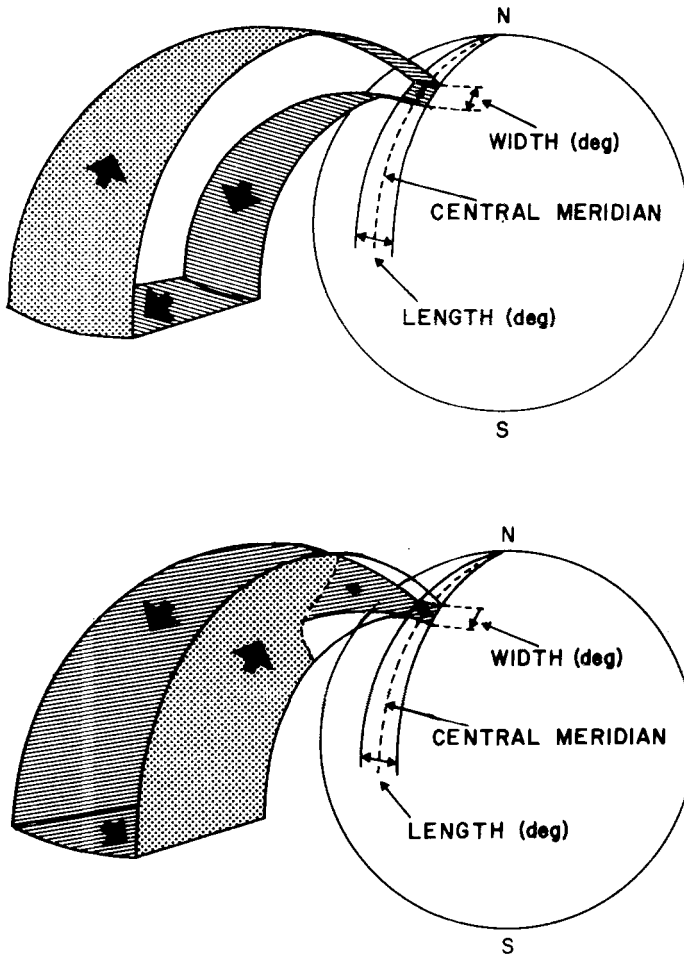


Fig. 7. Models of three-dimensional current loops always present in the high-latitude regions (Kisabeth and Rostoker, 1977). The E-W(N-S) loop is basically Hall (Pedersen) current flow.

Hughes (1979) and Kisabeth (1979) are all based on this principle and differ only in details. A good example of application of this type of modelling is given in Figure 8a where 'net' (i.e., unbalanced) downward field-aligned currents leaking from the ring current around local noon have to feed the eastward and westward electrojets and eventually diverge upwards in the region of the Harang discontinuity (a region of westward electric field in the pre-midnight sector) under the form of net Birkeland current (i.e., more 'up' than 'down') (Rostoker *et al.*, 1982). The simplified model used for the calculation is given in Figure 8b whilst Figure 9 shows the agreement between observed (along the Alberta line of stations) and calculated field values.

Other methods of modelling the high-latitude magnetic variations exist, however they usually simplify the geometry of the problem by reducing the earth to a flat

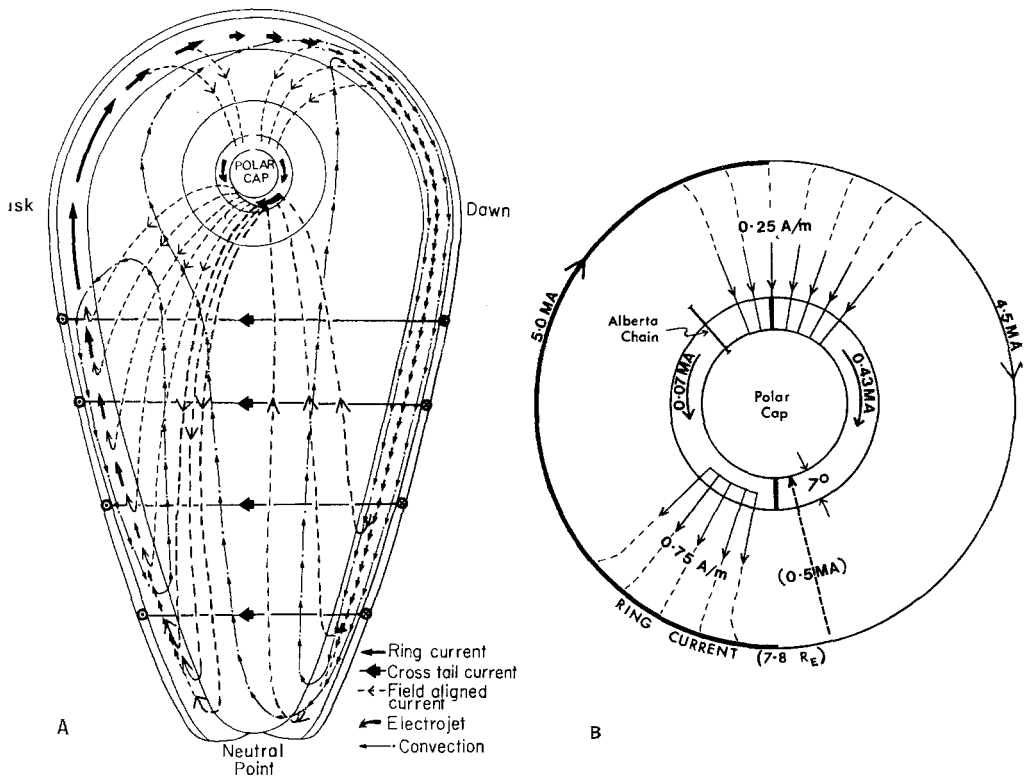


Fig. 8. (a) A schematic representation of various magnetospheric currents (see text); (b) Their simulation with a near-Earth model (Rostoker *et al.*, 1982).

surface and/or the field-aligned currents to vertical lines (e.g., Kamide *et al.*, 1981). Such models can be extremely useful on very localised bases, and often have the advantage of being more flexible in their ability to represent the ionospheric conductivity than the 'Kisabeth models'. However, they neglect completely the effect on mid-latitude magnetic observations of current flowing along high-latitude curved field lines (which produce contributions of the order of 15% to the H component of a typical 'bay'), as well as some of the edge effects of Birkeland sheets in the auroral zones (large D components), a problem discussed in some detail by Mareschal (1981).

5. The Disturbance Variation Fields

5.1. DEFINITION

Whether observed at high or low latitudes, disturbed fields always correspond to enhanced magnetospheric activity, which, in itself, is impossible to predict but which often has the effect of intensifying pre-existing current systems. Most disturbance fields can thus be simulated accurately with the techniques described in the previous

OBSERVATIONS

MODEL

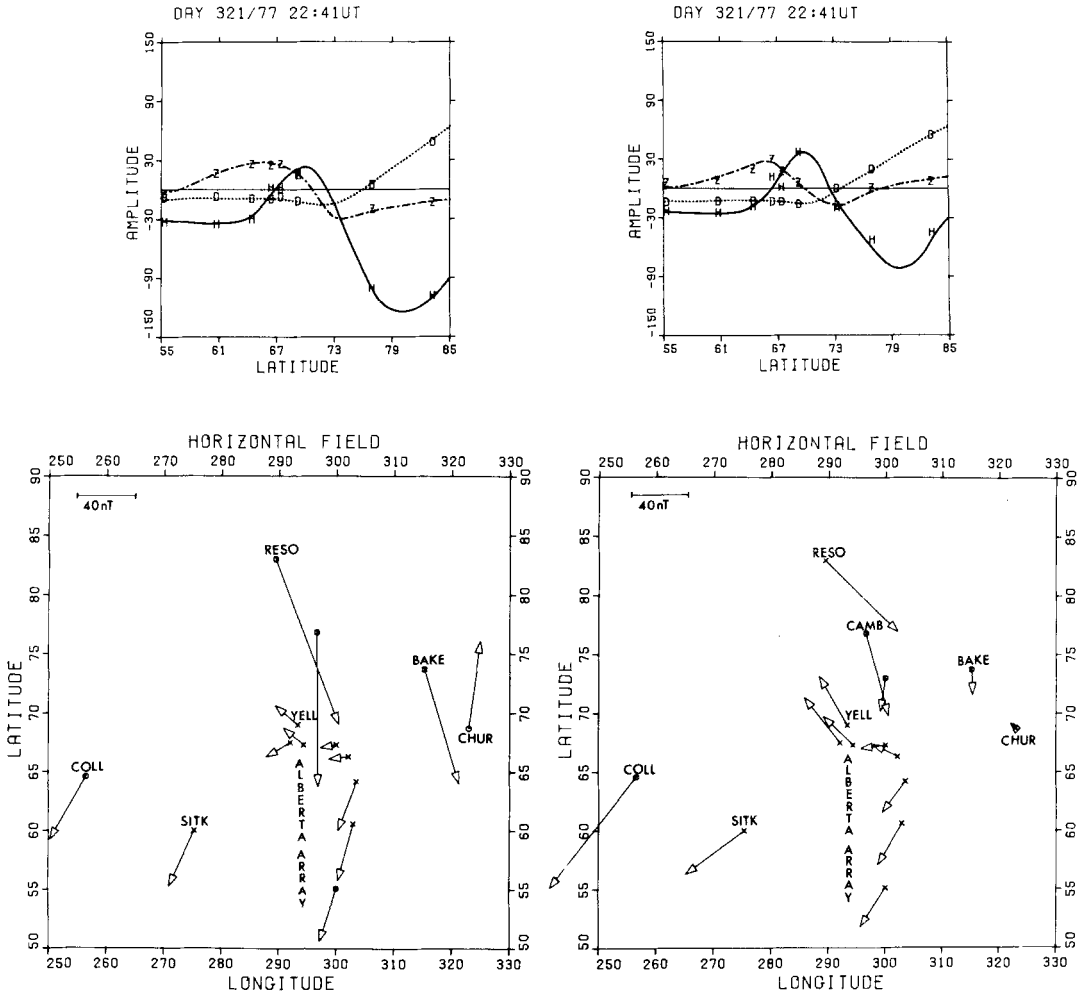


Fig. 9. Comparison between observed and computed high-latitude magnetic field components. The calculations are based on the numbers given in Figure 8b and use the models presented in Figure 7 (Rostoker *et al.*, 1982).

section, even though, in this case, their variations are intrinsically temporal. For instance, night-time high-latitude currents such as those associated with substorms (Figure 10), discrete or break-up aurorae, westward travelling surges and eastward travelling omega bands (Figure 11 and Baumjohann, 1983) can all be modelled with the current loops presented in Figure 7. Regular and irregular micropulsation fields, whether localised or not, and despite the fact that their generation mechanisms are still controversial, can, in some cases, be modelled with the same technique. Indeed,

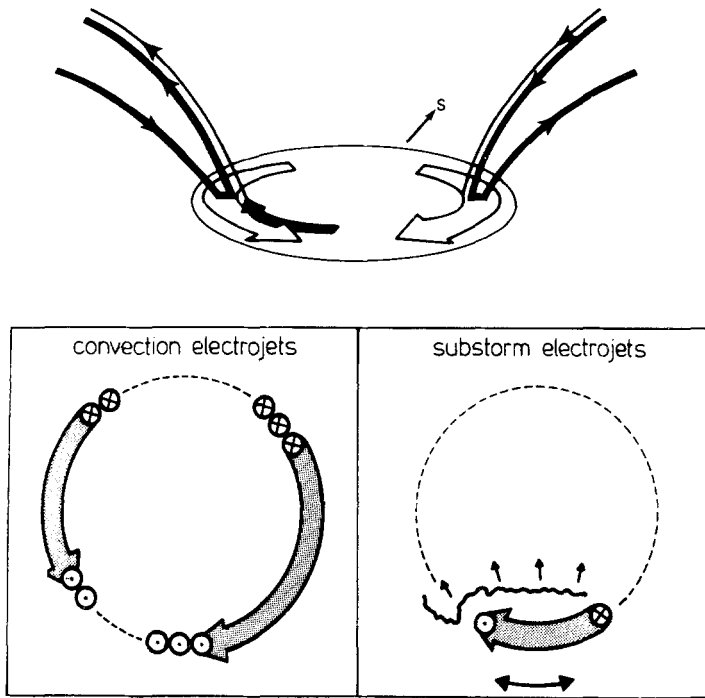


Fig. 10. Schematic representation of the intrusion of a substorm current system into the convection electrojets system for a dark ionosphere. ⊗ represents field-aligned currents into the ionosphere, ⊙ out of it. (Adapted from Baumjohann, 1983 and Zanetti *et al.*, 1983).

the Pc4's and Pc5's have ground magnetic signatures corresponding to hydromagnetic wave resonances in the magnetosphere later affected by their transmission through the ionosphere (e.g., Southwood, 1983; Southwood and Hughes, 1983; Glassmeier, 1984) but which also appear to be reproducible by the direct vibrations of the Hall current system driven by the wave in the ionosphere (e.g., Hughes and Southwood, 1976b), i.e., by the vibrations of a 3-D East-West loop such as that shown in Figure 7. Samson and Rostoker (1981) have used this type of model to reproduce the morphological characteristics of daytime Pc5's in terms of night-time substorm activity. Along the same line, Lam and Rostoker (1978) had been able to show spatial horizontal wave variations in the Pc5's range on scale lengths of the order of 100 km, and thus at least comparable to typical skin depths. Irregular pulsations, such as the Pi2's, have also been shown to vary on small horizontal scale lengths which can be correlated to changes in field-aligned currents with an auroral surge and thus can be modelled in terms of current circuits such as Kisabeth's (e.g., Pashin *et al.*, 1982; Samson and Rostoker, 1983). Note that Southwood and Hughes (1978) definitely showed how large vertical components could be induced in geomagnetic pulsation signals at high-latitudes simply because of the localised nature of the source.

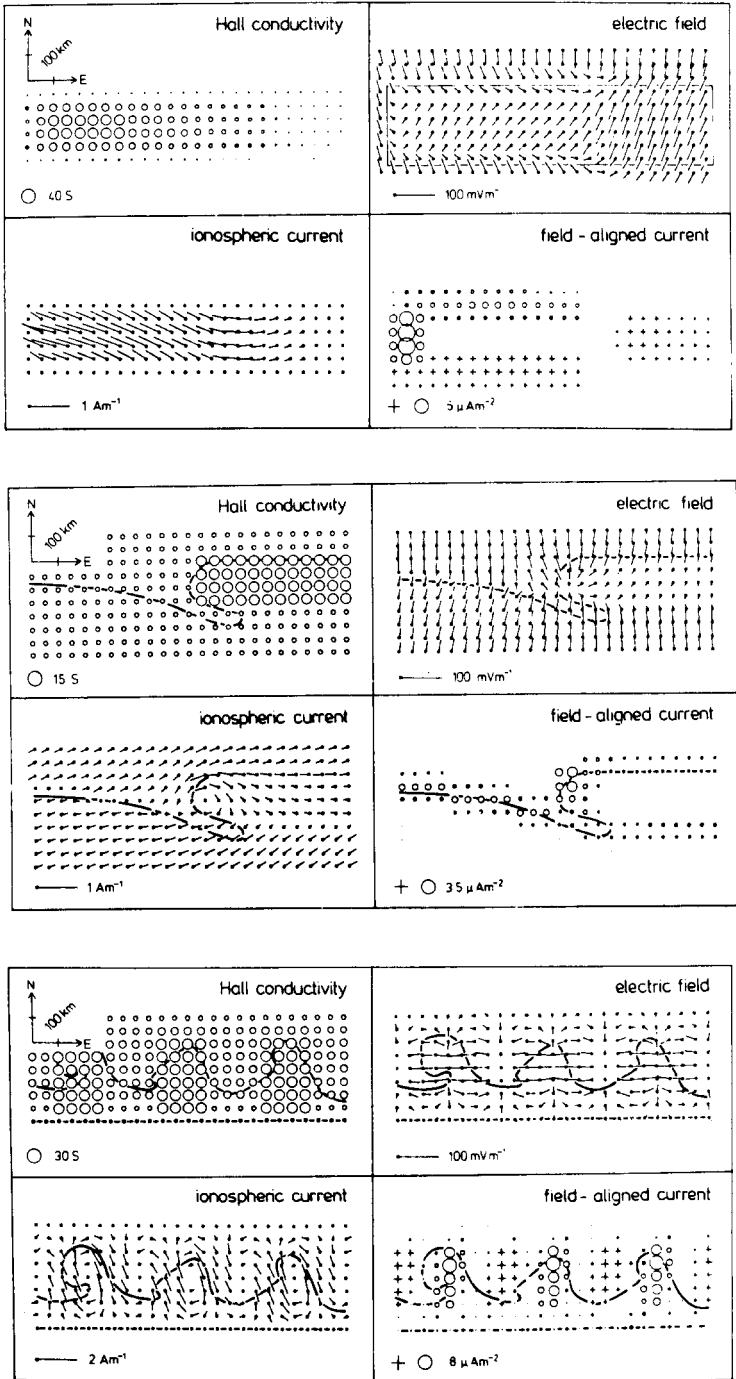


Fig. 11. Signatures of a break-up aurora (top), a westward travelling surge (middle) and an eastward travelling Ω band (bottom). The field-aligned currents are represented as in Figure 10 (Baumjohann, 1983).

Secondary magnetospheric effects at mid- and low-latitudes, such as those due to the intrusion of atmospheric winds produced by Joule heating of the auroral ionosphere, can be simulated with the atmospheric dynamo model (e.g., Blanc and Richmond, 1980) and/or the equatorial electrojet model developed by Richmond (1973a). The currents driven by reconnection, which have a geometry very similar to that of the S_q^p system but are highly dependant on IMF conditions, can also be modelled at high-latitudes with the 3D current loops presented in Figure 7 as well as included in Richmond's model of the equatorial regions via the calculation of relevant electric field (e.g., Nopper and Carovillano, 1979). These currents, which can be observed at high-latitude via the third ring of Birkeland currents mentioned in Section 2 (Iijima, 1984) and in the equatorial regions as the counter electrojet mentioned in Section 4, typically accompany the storage of current in the magnetospheric tail prior to a substorm (e.g., Baumjohann and Friis-Christensen, 1984). They reinforce the S_q^p – electrojet system during negative B_z (southward IMF), and are then traditionally called DP2, disappear for $B_z = 0$, and generate a 'reversed' two cells current system for positive B_z , i.e., the westward electric field commonly observed in low latitude regions as already mentioned in the description of Richmond's model.

Many global models of near-earth electrodynamics are now being developed (e.g., Nopper and Carovillano, 1979), most of which might be of little interest to the person only interested in defining local sources. However, it is interesting to note the models of Nisbet (1982) and Bleuler *et al.* (1982) which describe currents, Joule heating, etc., generated by field-aligned currents directly in terms of the geophysical indices (K_p , AL ...) well known to the solid earth geophysicist.

5.2. SOME ORDERS OF MAGNITUDE

The auroral electrojets flow within the auroral zone whose latitudinal distribution is fairly well defined by the location of region 1/region 2 Birkeland sheet currents. The zone can be modelled as an oval, always crossing the local midnight meridian at lower geomagnetic latitudes than the noon meridian (Figure 1). Typical locations of the electrojet central latitude can vary from 68° for quiet times to 58° under disturbed conditions at local midnight, with corresponding noon values ranging from 75° to 65° . Typical electrojet widths vary from 200 km to as much as 2000 km during highly disturbed times but often exhibit smaller scale internal lateral variations, especially during phases of successive enhancement; for magnetic local times 0500–0800 (i.e., within the westward electrojet sector) and 1600–1800 (eastward electrojet), Burrows and his group have found from recent MAGSAT observations that the ratio of E–W to N–S current densities was typically of the order of 1.5 to 2.5 for moderately disturbed conditions ($1_0 < K_p < 3^+$) but was surprisingly low (0.5 to 1.0) for quieter conditions (Burrows, pers. comm.). Note that the most commonly used ratio of σ_H/σ_P in modelling high latitude currents is 2.0. Typical Hall conductances around local noon range from 5 to 10 S, and from 10 to 20 S around midnight. Typical auroral electric fields are of the order of 20 mV m^{-1} during quiet times and

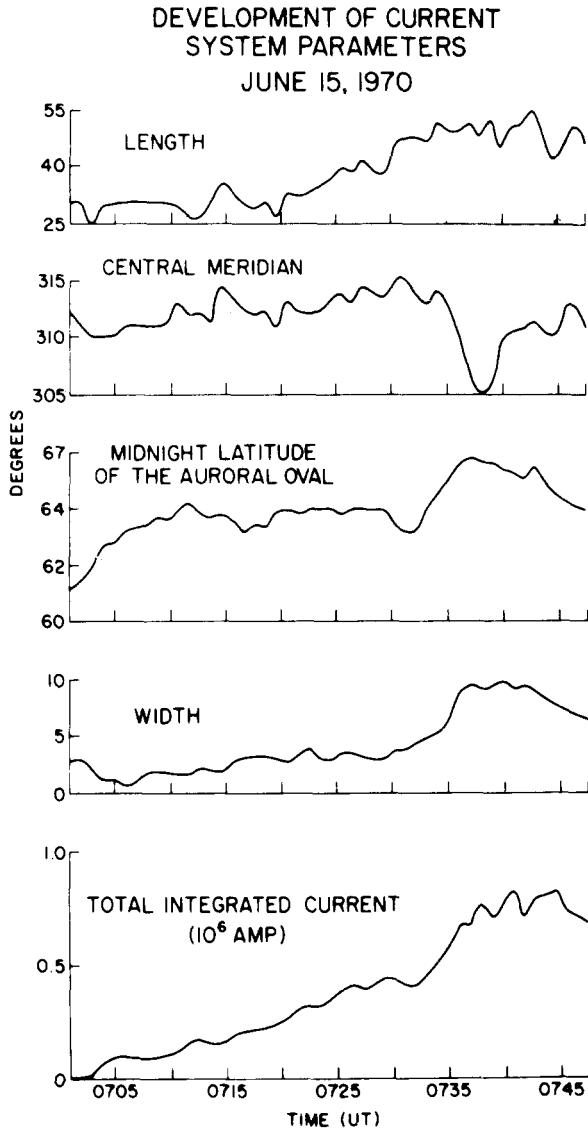


Fig. 12. Example of development of some of the parameters of a substorm current system obtained by linear inversion of surface magnetic observations with models such as those defined in Figure 7 (Mareschal, 1975).

50 mV m^{-1} for active geomagnetic conditions. Total current intensities within an auroral electrojet can vary from a few hundred thousand to several million A. Figure 12, after Mareschal (1975) (also in Kisabeth, 1975) illustrates the development of the parameters of a particular substorm current system over a little less than an hour.

The equatorial electrojet is roughly centered on the dip equator, but varies in width and intensity with geographical location (e.g., Onwumechilli and Agu, 1982). A

width of the order of 700 km can be taken as typical, with current densities regularly decreasing away from the equator. Typical sheet current densities would be of the order of 300 A km^{-1} at the dip equator but would already be reduced to 40 A km^{-1} at latitudes of 3° , generating total currents of the order of several tens of thousands of A. The morning counter-electrojet can consist of temporary reversals in current flow on either sides of the equator within relatively thin bands (about 200 km) and of sheet current densities of the order of a few tens of A km^{-1} . Once again, irregularities and flow reversals due to magnetospheric coupling via the high-latitude field-aligned currents introduce latitudinal variations on horizontal scale lengths which can certainly be as small as 100 km.

Note that since both the equatorial and auroral electrojets basically flow within the ionospheric *E*-layer, i.e., at an altitude of approximately 110 km, any of their irregularities varying on horizontal scale lengths shorter than 100 km will be severely attenuated through the atmosphere (e.g., Hughes and Southwood, 1976a, b) and thus does not need to be considered as potential localised sources.

6. Regional Induction Studies and the Concept of Localized Sources

6.1. DEFINITION OF THE RESPONSE FUNCTIONS IN THE WAVE-NUMBER DOMAIN

Regional induction studies are concerned with defining particular distributions of sub-surface conductivity in restricted areas. They are based on the definition of various response functions which depend on the characteristics of the source as well as of the substratum. Methods of interpretation were first developed on the principle that the inducing field was effectively uniform over the region of interest, which, in turn, could be reduced to a layered half-space. Progress in interpretation and modelling techniques has led to the consideration of the restricted horizontal extent of most natural sources (as well as of lateral conductivity discontinuities within the substratum).

Estimates of source effects in the wave number domain rely on the fact that any surface field can be decomposed into a series of elementary sinusoidal spatial undulations over a locally flat earth. The only parameters relating the flat earth to the external field are then the series of tangential wave numbers necessary to either synthesize this field from elementary sources or to represent its typical horizontal scale length (e.g., Price, 1950; Wait, 1953). If $k = \sqrt{k_x^2 + k_y^2}$ denotes the modulus of the tangential wave number vector \mathbf{k} , the response functions usually considered are the magnetotelluric impedance

$$Z(\omega, \mathbf{k}) = -E_x(\omega, \mathbf{k})/H_x(\omega, \mathbf{k}) = E_x(\omega, \mathbf{k})/H_y(\omega, \mathbf{k})$$

the geomagnetic impedance

$$Z'(\omega, \mathbf{k}) = kH_z(\omega, \mathbf{k})/(k_x H_x(\omega, \mathbf{k}) + k_y H_y(\omega, \mathbf{k}))$$

and the ratio of internal to external parts

$$S(\omega, \mathbf{k}) = H_{\text{hor}}^i(\omega, \mathbf{k})/H_{\text{hor}}^e(\omega, \mathbf{k}) = -H_z^i(\omega, \mathbf{k})/H_z^e(\omega, \mathbf{k}).$$

Over a layered earth, these are interrelated through the inductive scale length $C(\omega, k)$ as

$$Z(\omega, k) = i\omega\mu_0 C(\omega, k) \quad (1)$$

$$Z'(\omega, k) = ikC(\omega, k) \quad (2)$$

and

$$S(\omega, k) = (1 - kC(\omega, k))/(1 + kC(\omega, k)). \quad (3)$$

The conversion into the equivalent apparent resistivity is also given by

$$\rho_a = \omega\mu_0 |C(\omega, k)|^2$$

(e.g., Schmucker, 1973). Note that the field components depend on the vector \mathbf{k} (intensity and direction), while their ratios are functions of $|k|$ only.

6.2. VARIATION OF THE RESPONSE FUNCTIONS WITH k

If k is very close to zero (physically, of course, no induction will take place if $k = 0$, i.e., if the source is a true plane wave), the relations defined above reduce to the traditional magnetotelluric impedance expression

$$Z(\omega, 0) = i\omega\mu_0 C(\omega, 0),$$

geomagnetic relation

$$Z'(\omega, 0) \approx 0$$

(which simply indicates that the 'normal field' has no Z component at the surface of a stratified conductor), and the internal/external ratio

$$S(\omega, 0) \approx 1.$$

Note that since k is never exactly zero, one can also, in principle, always use gradients of the EM field to infer the conductivity of the substratum, the only constraint on such methods being their practical feasibility. The most common of these is the magnetovariation technique, which is based on the relation that exists between the vertical component and the horizontal gradients of the surface magnetic field in the frequency-space domain, i.e.,

$$C(\omega, 0) \approx Hz(\omega, x, y)/(\partial Hx(\omega, x, y)/\partial x + \partial Hy(\omega, x, y)/\partial y) \quad (4)$$

when $k|C(\omega, 0)| \rightarrow 0$, but k is finite.

If k is not very small, it is essential to determine when the relation $\rho_a = \omega\mu_0 |C(\omega, k)|^2$ (or equivalent) can be used with confidence to define the apparent resistivity of the substratum. Indeed, if $C(\omega, k)$ tends towards (true $\rho_a/\omega\mu_0$)^{1/2} when $k|C(\omega, 0)|$ is very small, it also tends towards $1/k$ when $k|C(\omega, 0)|$ is very large, i.e. in this case is totally independent of the characteristics of the conductor and leads to aberrant apparent resistivities.

Mathematically, it is easy to see that the magnetotelluric and magnetovariation relations can be treated in similar ways to estimate their contamination by source

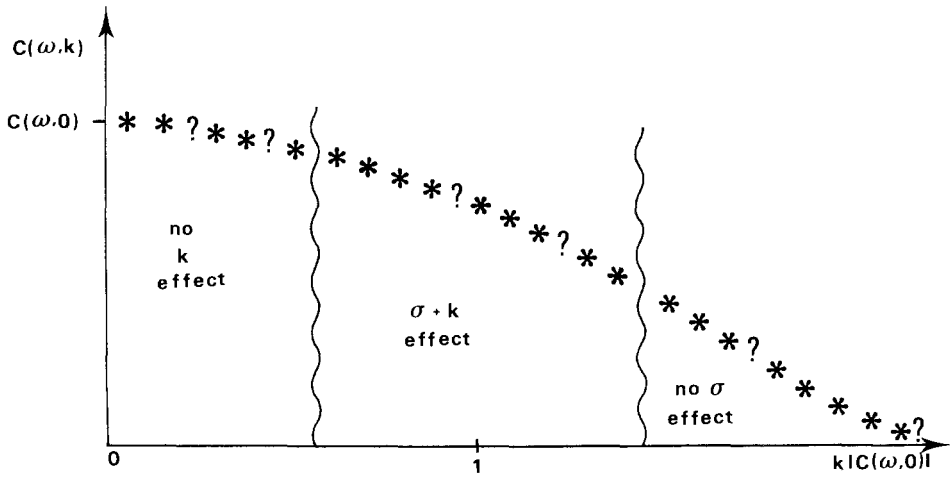


Fig. 13. Schematic representation of the variation of a response function with wave-number and of its regions of different relative dependency on k and σ .

effects since neither (1) nor (4) formally depends on k . On the other hand, both the geomagnetic and internal/external relations are explicitly functions of k and thus render this estimation more difficult to perform. This explains why most quantitative evaluations of source effects concern the magnetotelluric or magnetovariation methods. In principle, the whole point of such an exercise consists in defining the shape of the curve relating a response function (C , or Z , or eventually S) to k (or to $k|C(\omega, 0)|$ to work in dimensionless units). Indeed, as $k|C(\omega, 0)|$ will vary from zero to infinity (or, at least, to that point where $1/k = 100 \text{ km}$ since the signal will be seriously attenuated beyond this threshold), three different regions will be defined, i.e., a region in which ρ_a is unaffected by k , a region of mixed σ and k effects, and a region of k dependence alone (see Figure 13). As long as that curve will remain 'flat', i.e., as long as $\partial C(\omega, k)/\partial k \approx 0$, the source contamination in the evaluation of an apparent resistivity will be negligible.

In fact, very little work on the functional relation 'response function = $f(k)$ ' appears yet to have been achieved for or from real sources. Several 'correction formulae' to the traditional magnetotelluric impedance relation have long been obtained in terms of specific k values (e.g., Wait, 1954; Price, 1962, 1967; Srivastava, 1965) and have been applied recently to the calculation of $C(\omega, k)$ for simple electrojet models (e.g., Beamish, 1980; Jones, 1980). Several elegant mathematical treatments covering the problems of induction by non-uniform sources of given structure have also been presented (e.g., Weaver, 1973; Rae, 1982), but the wave number content of real sources is still usually rather poorly defined. An interesting evaluation of that 'predominant' value of k from direct magnetic field measurements has been attempted by Oni and co-workers in Nigeria during the last decade (e.g., Oni, 1972; Oni and Agunloye, 1974) and has recently been summarised by Oni (1984). Unfortunately, the limited number of stations used by the authors precludes

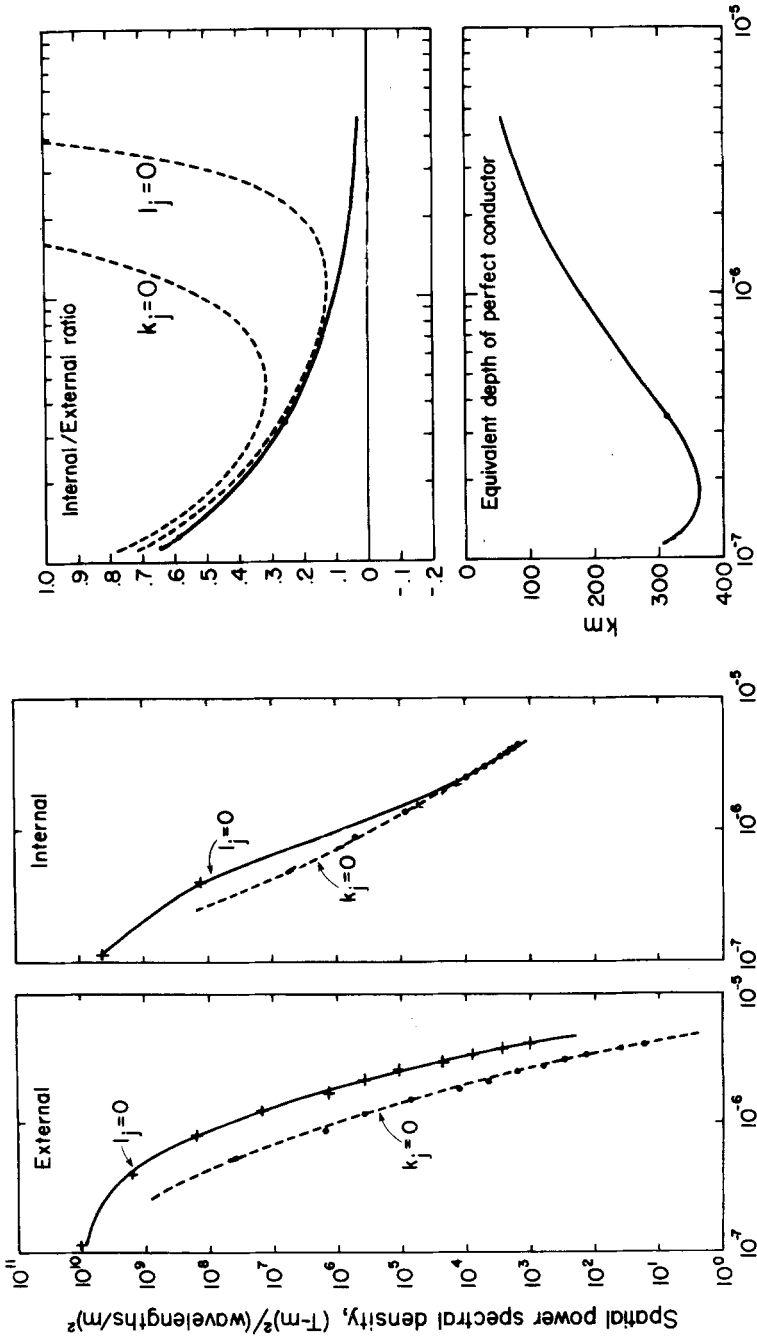


Fig. 14. Example of three-dimensional separation of a substorm field in parts of internal and external origin. The curves labelled $k_j = 0$ (or $l_j = 0$) correspond to two-dimensional separations where only North-South (or East-West) wave-numbers are allowed (see text). (Adapted from Richmond and Baumjohann, 1984).

the definition of reliable power spectra in the k -domain.

A formal definition of the wave number content of the source field can be approached in two manners: either by model fitting to the observations (assuming that the earth is a perfect conductor) or by separation of these observations in parts of internal and external origin. The model fitting approach has only been used successfully in moderately disturbed conditions, i.e., when the source field could be basically simulated by the field of a simple current system whose geometry varied only very slowly with time (e.g., Horning *et al.*, 1974; Mareschal, 1975 (from which Figure 12 was taken); Fambitakoye and Mayaud, 1976a; Oldenburg, 1976; Bannister and Gough, 1977, 1978; Gough and Bannister, 1978). Even though the depth to the equivalent perfect conductor has sometimes been adapted to the frequency content of the source in order to better define the external field (e.g., Mareschal, 1976; Mareschal and Kisabeth, 1977), the results are rather disappointing by their forced simplicity and have not been used to define the k content of the source.

Separation into internal/external contributions in view of defining the external current system has only been achieved with limited success too because of the problems associated with the limitation of the region of observation and of the necessary interpolation of field data from one station to the other. In that respect, it is worth mentioning the recent effort of Richmond and Baumjohann (1984) who developed a technique of separation for magnetometer array data leading to the estimation of the source wave number spectrum with a quantification of the errors involved in interpolating data between the stations, and separating and mapping the magnetic variations. Richmond and Baumjohann applied their method to a data set recorded in Scandinavia which sampled the development of an eastward electrojet followed by a reversal associated with the intrusion of a westward jet. Figure 14 illustrates the variations of spatial power densities for the external and internal magnetic potentials with the N–S and E–W wavelength content of the source. It also illustrates the variations of $S(\omega, k)$ with wave number and the corresponding variations in the depth of the perfect conductor that would produce induced currents equivalent to those just obtained. The wavelengths involved vary from 10000 to 100 km. Note how clearly the neglect of wave number in one direction (i.e., k_x or $k_y = 0$) affects the response function.

This approach is still based on a mathematical interpolation of data, but, at least, shows an attempt to quantify the errors. In principle, the most satisfying solution would consist of evaluating the variation of a response function with k directly from the observations, assuming that these were recorded over a layered region of known conductivity distribution $\sigma(z)$. That conductivity distribution could be obtained from the ratio E_y/H_x measured as a function of ω at one remote point where the source field is almost 'uniform'. This is probably feasible in the equatorial regions where the source behaves in a much more constraint manner than at high-latitudes and where the assumption that its geometrical characteristics do not vary much with time is justifiable. Then, from the sole knowledge of $H_x(\omega, x)$ along a profile crossing the electrojet, the external part $H^e x(\omega, x)$ could be obtained by convoluting $H_x(\omega, x)$

with an operator constructed from $\sigma(z)$. This does not seem to have been done yet (because of the difficulty in finding stable telluric fields?). Rather than pursuing wave number definitions, new approaches to the problem of source contamination have been developed, with, in particular, theoretical and modelling work in the frequency-space domain. It is those advances that will be examined in the following two sections.

7. On the Spatial Characteristics of Acceptable Source Fields in Traditional Magnetotelluric and Magnetovariation Work

The relations between various field components in the frequency-wave number domain defined in Section 6a are extremely elegant by their simplicity. Unfortunately, they require transformations from space to wave number which are generally impractical since most observations are not carried out over the full width of the source field even when it is localised. The only way of using an unrestricted representation of the response functions (e.g., $C(\omega, k)$) for all wave numbers is to work in the frequency-space domain. This approach leads to convolution integrals over kernels that usually tend towards zero beyond a certain distance range from the point of observation. Schmucker (1970, 1980) defined the kernel corresponding to the magnetotelluric relation for a 1D source from

$$Ex(\omega, y) = i\omega\mu_0(N(\omega, y) * Hy(\omega, y)) \quad (5)$$

where

$$N(\omega, y) = (1/\pi) \int_0^\infty C(\omega, k) \cos ky dk$$

and

$$C(\omega, 0) = \int_{-\infty}^{+\infty} N(\omega, y) dy.$$

The equivalent relations for a 2D source are

$$Ex(\omega, x, y) = i\omega\mu_0(G(\omega, x, y) ** Hy(\omega, x, y)) \quad (6)$$

where

$$G(\omega, x, y) = (1/4\pi^2) \iint_{-\infty}^{+\infty} C(\omega, \mathbf{k}) e^{i\mathbf{k}\cdot\mathbf{r}} dk_x dk_y$$

with

$$C(\omega, 0) = \iint_{-\infty}^{+\infty} G(\omega, x, y) dx dy.$$

Clearly, if the horizontal magnetic field is uniform enough over distance ranges given by the inductive scale length $C(\omega, 0)$, then the convolution integrals (5) and (6) reduce to multiplications of the horizontal field measured at the station by $i\omega\mu_0 C(\omega, 0)$, i.e., to the traditional magnetotelluric relation (Tikhonov, 1950; Cagniard, 1953). But Dmitriev and Berdichevsky (1979) and Berdichevsky *et al.* (1981) have recently shown that the relation remained true for a much wider class of surface fields. Indeed, when these are represented in terms of power series of the spatial coordinates, the magnetotelluric and magnetovariation relations take the

form of series respectively with even (MT) and odd (MV) space derivatives of the horizontal surface field, i.e.,

$$E_x(\omega, y) = i\omega\mu_0(I_0 Hy(\omega, y) + \frac{I_2}{2!} \frac{d^2 Hy}{dy^2}(\omega, y) + \dots)$$

and

$$H_z(\omega, y) = I_0 \frac{dHy(\omega, y)}{dy} + \frac{I_2}{2!} \frac{d^3 Hy}{dy^3}(\omega, y) + \dots$$

for the 2D source, and

$$E_x(\omega, x, y) = i\omega\mu_0(I_0 Hy(\omega, x, y) + \frac{I_2}{2!} \Delta Hy(\omega, x, y) + \dots)$$

and

$$H_z(\omega, x, y) = I_0 \left[\frac{\partial H_x}{\partial x}(\omega, x, y) + \frac{\partial Hy}{\partial y}(\omega, x, y) \right] + \frac{I_2}{2!} \left[\frac{\partial}{\partial x} \Delta H_x(\omega, x, y) + \frac{\partial}{\partial y} \Delta Hy(\omega, x, y) \right] + \dots$$

with

$$\Delta = \partial/\partial x^2 + \partial/\partial y^2$$

with the 3D source. In those expressions,

$$I_n = \int_{-\infty}^{+\infty} N(u) u^n du \quad (\text{so } I_0 = C(\omega, 0)).$$

The fundamental implication of such developments is that if over the distances to be considered in (5) (or (6)), the earth can be considered as a layered conductor and the horizontal magnetic field components vary linearly, the traditional (i.e., zeroth order approximation) magnetotelluric and magnetovariation relations are valid. The magnetovariation relation remains valid even if the second degree space derivative are non-zero since only the third order term must disappear to satisfy Equation (4). Dmitriev and Berdichevsky (1979) conclude that the diameter of the region surrounding the station over which these field constraints must be respected are under 100 to 200 km when a sediment cover with thickness up to 3 km is studied and increased to 300 to 500 km for investigations of fairly well conducting formations of the upper mantle lying as deep as 100 km (periods up to 10000 sec).

If the horizontal fields do not satisfy the conditions of linearity (or second order variations), correction terms have to be evaluated, which can be extremely difficult to do (via the calculation of the kernel N) if the source is not of simple geometry. A rapid check on the importance of the term in I_2 is possible by using the

approximations that $\partial H_x/\partial x + \partial H_y/\partial y = H_z/C(\omega, 0)$ and that over a uniform half-space, $I_2 = C(\omega, 0)^3$. It is evident then that the correction is significant for the MT relation only if $|(C(\omega, 0)/2)\partial H_z/\partial y|$ (or $|(C(\omega, 0)/2)\partial H_z/\partial x|$) is significant when compared to $|H_y|$. (For the MV relation, it is the ratio of $|(C(\omega, 0)/2)(\partial^2 H_z/\partial x^2 + \partial^2 H_z/\partial y^2)|$ to the horizontal field gradients that is important). It is not the presence of 'normal' vertical magnetic components that will prevent the use of traditional MT but the existence of significant gradients in Hz. Note also that the relations will automatically fail if the source field geometry varies with the time interval that is Fourier transformed!

Jones (1980, 1982), has systematically used these criteria in his induction studies of Scandinavia and, for instance, applied surface fitting by second order polynomials to substorm events recorded in the region of Kiruna with the large array of Koppers *et al.* (1979) in order to interpret his magnetovariation data. However, neither Beamish and Johnson (1982), who used linear approximation to their variation data recorded in Northern England (they had very few stations), nor Camfield (1981) who used the magnetovariation technique in Quebec with a relatively large array of magnetometers and fitted his data with third order polynomials, succeeded in obtaining consistent response functions. The problem that Beamish and Johnson encountered might be due to the highly heterogeneous substratum that they were studying since, by using linear interpolation, they were minimizing the contributions of third order derivatives in the least-square sense, and thus, in principle, reduced source effects. Even though the region that Camfield was studying is tectonically active and has a lateral conductivity discontinuity following a straight fault for about 200 km, it had been shown by magnetotelluric experiments to be truly 1D on either side of the discontinuity. Hermance (1984) recently showed that under such conditions, a field originally polarized parallel to the strike would only be perturbed in the close vicinity of the lateral heterogeneity, but that at modest distances from the contact, the response was more influenced by the source dimensions and what is beneath the observer than by what is at some lateral distance away. Since substorm fields happen to be highly polarized in a direction roughly parallel to the discontinuity studied by Camfield, it might be worth estimating the correction term in I_2 to $C(\omega, 0)$ and verifying whether or not the response improves.

8. Modelling in the Frequency-Space Domain

When the spatial gradients used in magnetovariation work are obtained by least-square fitting of a second order polynomial, at least six stations recording simultaneously are required to constrain the solution. Even if the curl free condition is imposed on the surface magnetic field, five stations are still necessary (e.g., Jones, 1980). If a campaign involves only a small number of magnetometers that are moved around, an application of the criteria described in the previous section is not possible. (Equivalent constraints apply to magnetotelluric investigations with a limited number of stations). As a result, modelling in the frequency-space domain can play

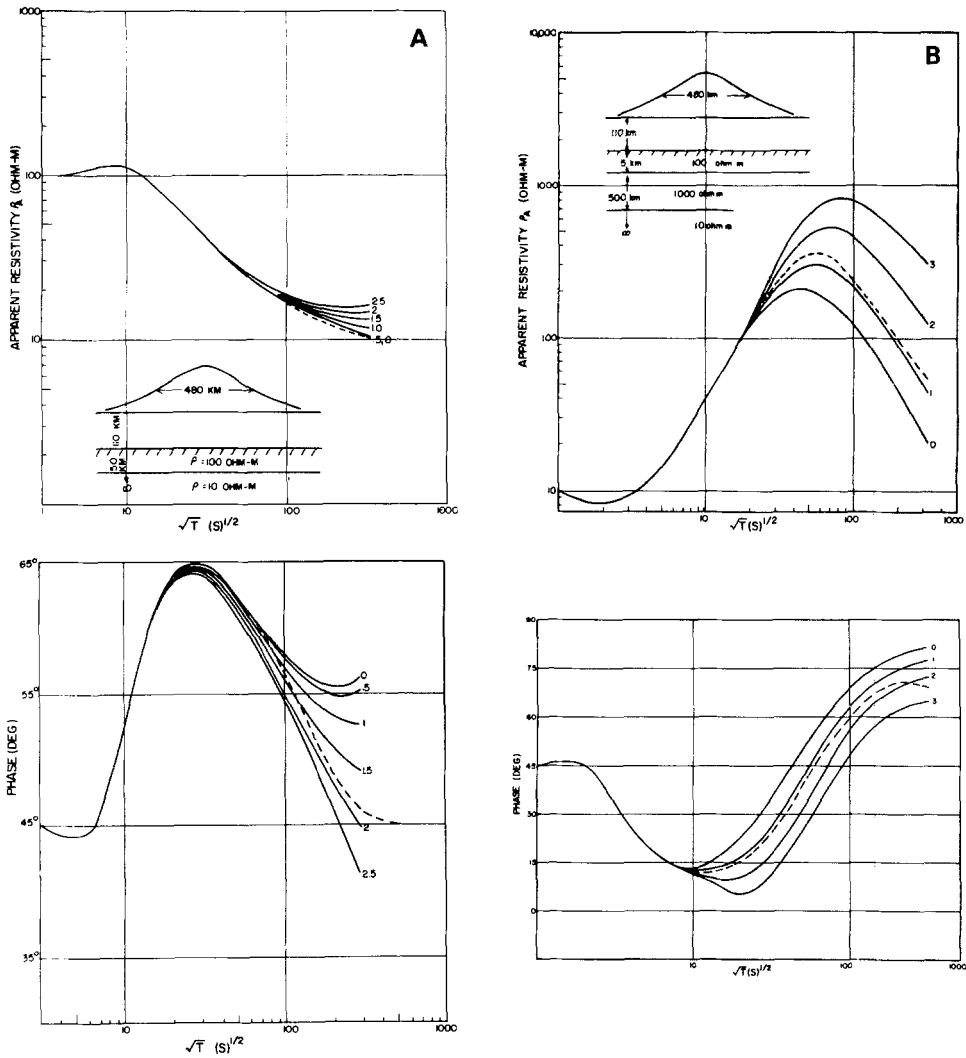


Fig. 15. Apparent resistivity and phase variation curves for a source electrojet and model earth as defined in inserts on (A) and (B). The numbers on the curves refer to the position of the observer with respect to the center of the electrojet in units of 240 km. The dashed lines give the traditional MT responses (Peltier and Hermance, 1971).

a useful role in defining regions and conditions under which the traditional MT or MV interpretation is acceptable.

Several authors have used modelling to define the response of a given substratum to simple sources such as line or band electrojets, and to evaluate their spatial variation with distance from the source (e.g., Hermance and Peltier, 1970; Peltier and Hermance, 1971; Hutton, 1969, 1972; Hutton and Leggeat, 1972; Oni and Alabi, 1972; Hughes and Wait, 1975b; Osipova, 1983; Kao, 1984). The representation of the source is based on the synthesis of undulations smoothly ranging from $k_x = -\infty$

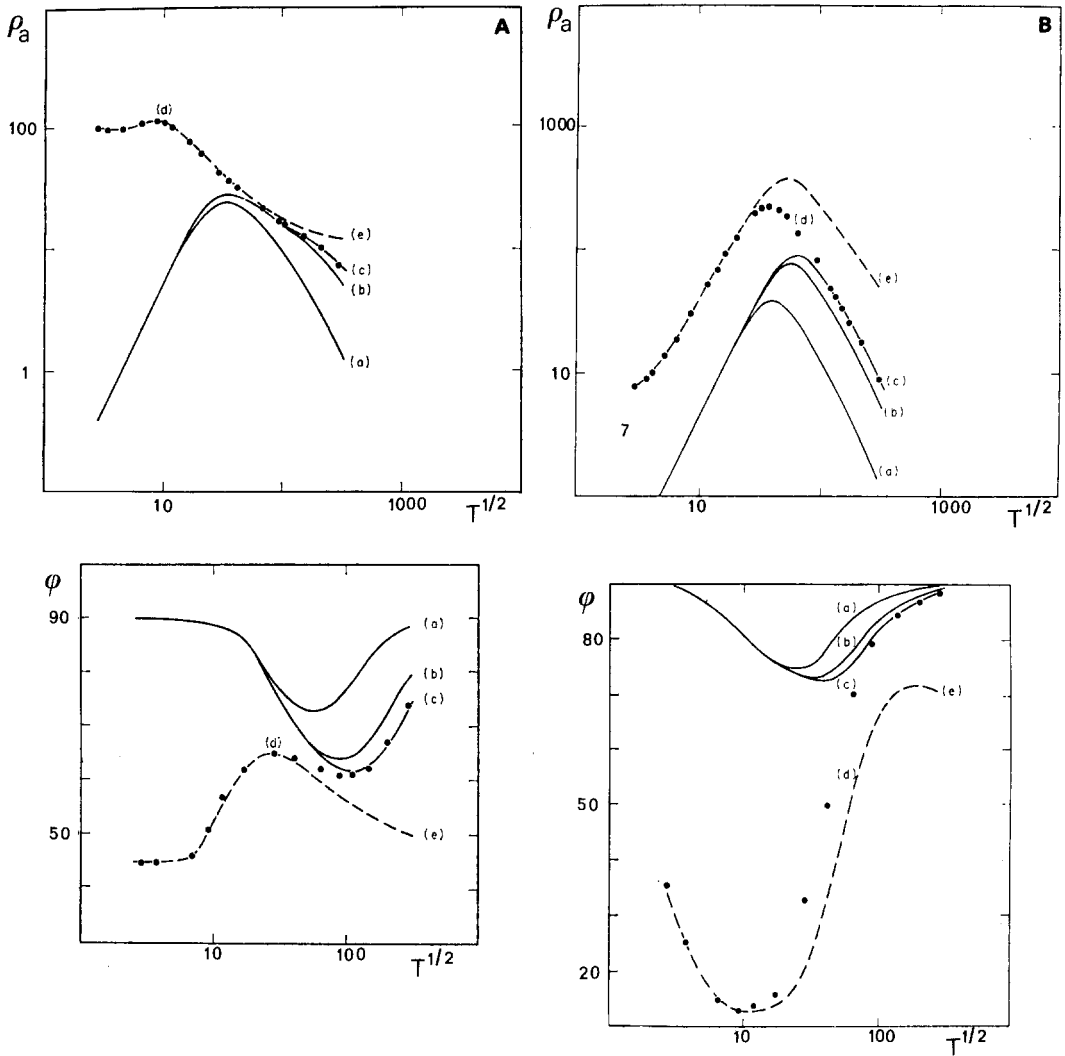


Fig. 16. Apparent resistivity and phase variation curves obtained over the same earth models as those defined in Figure 15 (A) and (B) but for electrojets of finite length (a, b, c). Curves d) correspond to Peltier and Hermance's Electrojet while curves e) give the traditional MT response (MKS units; Hutton, 1972).

to $k_x = +\infty$ and thus does not show any dominant effect of specific values of k on the definition of the apparent resistivity. The results are clearly quite consistent with intuitive arguments based on source versus skin depth scale lengths. They show, for instance, that the source effect increases with the resistivity of the subsurface (e.g., Figure 15, after Peltier and Hermance, 1971) which, of course, indicates that conventional techniques of interpretation will be worse in continental shield areas than in tectonically active regions (e.g., Iceland). For a given substratum, source effects also increase with period at a given location but eventually decrease with

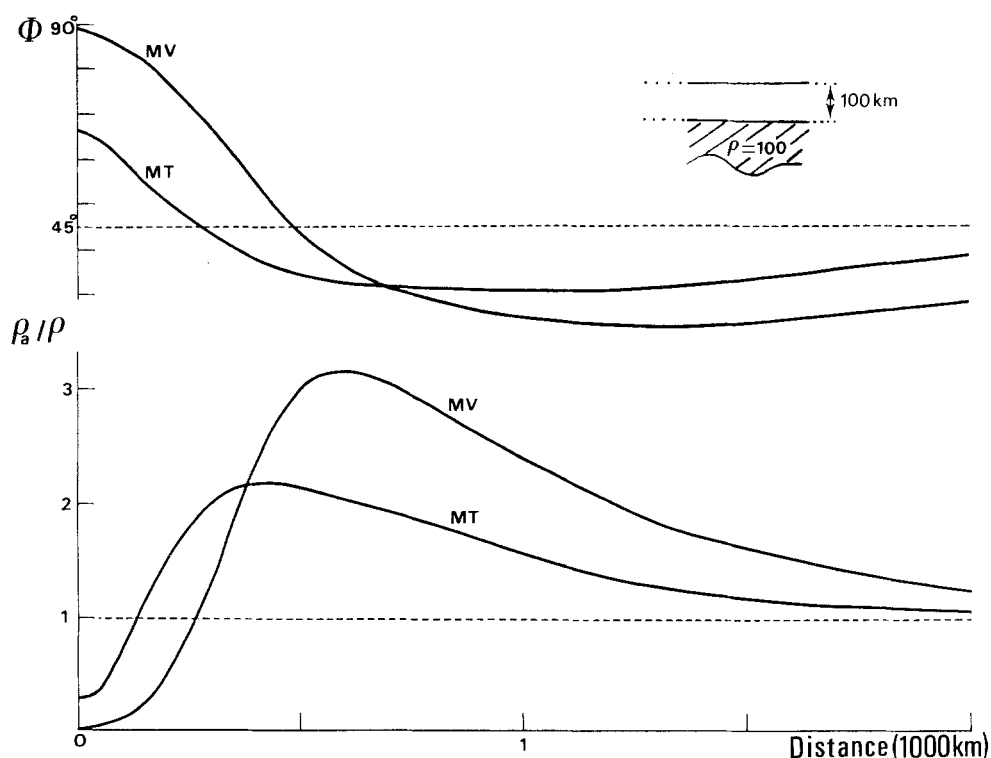


Fig. 17. Difference in phase and apparent resistivity curves calculated for the model shown in insert from either the magnetotelluric (MT) or the magnetovariation (MV) relation. The responses are given in terms of the position of the observer with respect to the line electrojet (Kao, 1984).

distance from the electrojet. Note that magnetotelluric measurements made at long periods directly under the electrojet tend to underestimate the resistivity and overestimate the phase, while the reverse is true past the edges of the electrojet (see Figure 15). Since a source which is bounded both in latitude and longitude does not include the low wave numbers that does an infinite electrojet, its calculated impedance should differ from that of the plane wave source over a larger range of frequencies than indicated by the model of Peltier and Hermance. This effect is illustrated in Figure 16 by Hutton (1972) who modelled electrojets containing longitudinal gradients in an attempt to represent the actual equatorial electrojet. Note that with Hutton's particular model, the magnetotelluric impedance is independent of location with respect to the electrojet. Sources of very restricted dimensions have also been considered by Osipova (1983) who clearly showed the deterioration of the traditional apparent resistivity curve with decrease in the source size (ranging from a uniform field to an ionospheric dipole source via an ionospheric current arc). That author also calculated curves of apparent resistivity for electrojet and dipole sources for the two types of geoelectric sections observed in the Baltic shield, i.e., a normal section for which resistivity decreases smoothly with depth and a section with a conducting layer at a depth of 20–30 km overlying the normal

conductivity profile. The apparent resistivity vertically beneath the dipole source and above the resistive section is much smaller than would be predicted by the Cagniard approach beyond periods of the order of 10 sec. For such an extreme case of source effects, the dipole and uniform sources curves only give similar results at distances of the order of 1200 km from the source (see also Shaub, 1979, 1980). (Note, however, that very concentrated dipole sources at ionospheric height do not have much physical significance). Finally, Kao (1984), recently showed that impedances obtained from MV seemed to be more affected by the geometry of the source (here, a line current) than impedances determined by MT (see Figure 17 – could this be paralleled to the apparent higher rate of failure in MV than MT work at high-latitudes?) and recommended the use of impedance averaging over several stations or several sources in order to minimize source effects. Note that the problem of contamination of the data by source fields is much less important at low than at high-latitudes since, by recording at night, one can always get rid of the influence of the equatorial electrojet.

9. What About Response Functions That Are Explicitly Dependent on k ?

Because of its intrinsic dependence on the wave number k , the geomagnetic response function $Z'(\omega, k) = k\text{Hz}(\omega, k)/\mathbf{k} \cdot \mathbf{H}$ is seldom used to estimate the conductivity structure of a region (e.g., Porath *et al.*, 1971; Camfield and Gough, 1975). Its variation with station location in the auroral and equatorial regions has basically been used to estimate the spatial extent of non-uniform source field effects and possibly their dominant wavelength (e.g., Hermance and Garland, 1968; Oni and Agunloye, 1974; Mareschal, 1981; Kao, 1984; Rastogi *et al.*, 1984).

If field gradients are not considered, arrays of magnetometers are usually used only in order to evaluate the amount of current induced in a region with lateral discontinuities in conductivity. For that purpose, the array variation fields are usually separated into normal and anomalous parts. The normal part is supposed to represent the magnetic contribution of the real external current system plus that of the currents that it would induce in a layered substratum typical of the region. The anomalous part is the remainder of the surface field and should in principle only be due to lateral discontinuities in the region's conductivity. The simplest way of achieving this separation is to consider that (a) one station of the array is located over the 'normal' structure and that (b) the source field variations are on scale lengths larger than the dimensions of the array. Then the response of the reference station can be subtracted from the observations of all the other stations and the separation is effectively accomplished (e.g., Schmucker, 1970; Babour *et al.*, 1976; Babour and Mosnier, 1979). However, since the reference station cannot in any way be proven to be normal, the separation can lead to unstable normal fields (e.g., Dragert and Clarke, 1977). Another method, applicable to tight arrays, is to define the normal field by least square fitting of planes to the real and imaginary parts of the Fourier spectra of the measured field components (e.g., Ingham *et al.*, 1982). The worst

method, in terms of source field bias, consist of ignoring the source's geometry, i.e., in assuming that the normal part of the vertical variations is negligible compared to the anomalous vertical field ($Z'(\omega, k) = Z'(\omega, 0) = 0$), and that it does not correlate with either of the normal horizontal components, which, in turn, are much larger than the corresponding anomalous components.

Transfer functions between vertical and horizontal components are established and can be used to define either induction arrows or induction ellipses which, in turn, give a qualitative definition of the location of the induced currents. Since transfer functions only give qualitative information, it is not possible to assess quantitatively their amount of disturbance by localised sources. However, statistical methods can be used to define the treatment of data that will minimize those source effects. These are based on observations of systematic bias either on induction arrows or on transfer function estimates, at a variety of locations and for a variety of source periods (e.g., the observations of Anderson *et al.*, 1976, 1978; Beamish, 1980; Camfield, 1981, for pulsation periods, as well as those of Beamish, 1979; Gough and de Beer, 1980; Camfield, 1981, for substorm periods). Ensemble averaging is always recommended. For instance, Beamish (1980) recommends averaging mid-latitude pulsations of periods less than 500 sec over as much as 20 days in an effort to remove high night-day variations. Beamish (1979) also showed that the mid-latitude contribution of the source field to the definition of the transfer functions of course increased with latitude and period, but could be minimized by weighting the transfer function estimates obtained from individual records with the vertical field partial coherence. Gough and de Beer (1980) recommend selecting only events which have dissimilar H_x and H_y magnetograms and whose $H_x * H_y$ cross correlations peak at time shifts distributed in large negative and large positive values. Jones (1981), translated all such recommendations in terms of statistical frequency analysis, i.e., in terms of minimization of confidence intervals. However, more quantitative estimates of the amount of source contamination in geomagnetic depth sounding can only be achieved through the evaluation of how much current is induced in a specific substratum by a given source, i.e., by estimating the influence of the wave number k on the response function $S(\omega, k)$. Unfortunately, apart from the work of Forbush and Casaverde (1961) on the equatorial electrojet, the separation in view of defining the internal contribution of a local surface field has never been pursued. When separation techniques are applied on that scale, they refer only to the anomalous part of the regional variation field and only attempt to verify its internal origin and thus to get rid of potential source effects in a very round about manner.

Note that the separation of diurnal variations is, in principle, much easier to achieve (Ducruix *et al.*, 1977; Bhattacharya, 1981), even though they lead to relative scale sizes problems in the equatorial regions. Indeed, there, $|C(\omega, 0)|$ is of the order of 400 to 600 km (over continents) while $1/k$ can be taken as being of the order of 200 to 400 km. Obviously, one stands in the range of the curve defined in Figure 11 where ϱ_a is heavily contaminated by its k dependance, which probably explains the

long standing argument on whether (and where) the equatorial electrojet generates induced fields (e.g., Carlo *et al.*, 1982).

10. Discussion and Conclusion

Is source modelling with regards to induction studies worth bothering about any longer? My first reaction would be to answer negatively, at least as far as modelling along the lines reviewed in Section 9 is concerned. I do not deny the fact that modelling with simple sources has definitely helped to visualise the spatial and temporal effects of the non-uniformity of pseudo ionospheric sources on various response functions nor that it has encouraged optimism in particular in the field of magnetotellurics at high-latitudes.

However, it seems to me that modelling, for all its practical help in those simple problems, could yet achieve its most valuable contribution, that is, become the link between global and local studies. Indeed, only realistic three-dimensional source models can generate the local surface fields that should then be treated in the frequency wave number space to define an almost satisfactory functional relation between response function and wave number. Furthermore, three-dimensional source models could be used to help clarify the problem of channelling, that is, the definition of currents that have been induced in a certain region by a certain source and are channelled to another region subjected to quite different source behaviour. If nothing else, this might help the selection of better locations for permanent magnetic observatories.

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