INDUCTION IN THE EARTH'S CRUST: OBSERVATIONAL METHODS ON LAND AND SEA

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Abstract. During the last twenty years, the measurement of the Earth's electromagnetic field has been greatly improved. Old devices, such as suspended magnet variometers or the inductive sensors, are now up to date, thanks to the use of negative feedback techniques. Very sensitive instruments, like the superconducting quantum interference device (SQUID) are now available for geophysical applications. The extensive use of large arrays of magnetometers has permitted the study of the distribution of telluric flow inside the Earth's crust and the delineation of large, deep conductivity anomalies. We also have a better knowledge of the distribution of the resistivity at depth, even under the ocean bottom. But we are still very far from understanding how the telluric currents are induced in the Earth, especially at very low frequencies.

Résumé. Depuis vingt ans la mesure du champ électromagnétique naturel s'est beaucoup améliorée. D'anciens instruments comme les variomètres à aimant suspendu ou les capteurs inductifs ont été remis en honneur grâce à l'utilisation de techniques nouvelles, comme la contre-réaction de champ. D'autres appareils, extrêmement sensibles, tels que les SQUID (superconducting quantum interference device) sont maintenant disponibles pour des applications géophysiques. L'usage intensif de réseaux de magnétomètres a permis d'étudier la distribution des courants telluriques en profondeur et de mettre en évidence d'importantes anomalies de conductivité. Nous avons aussi une meilleure connaissance de la répartition de la résistivité à grande profondeur y compris sous les océans. Mais nous restons très ignorants des mécanismes qui régissent l'induction des courants telluriques à l'échelle de la terre entière, particulièrement aux très basses fréquences.

1. Introduction

Induction phenomena, produced inside the earth by the variations of the external magnetic field, are presently extensively studied by geophysicists. To this end, geophysicists have developped various methods which can be used on land, off shore, in the air and even inside the earth, in boreholes. Some of these methods are old but, when correctly up-dated, can still be usefull. Other methods have recently been developed involving, for example, the use of a SQUID. It is not possible, even in a review paper, to describe all the devices and all the methods used to measure the electromagnetic field or its components. I shall not, for instance, describe measurements in boreholes which, most of the time, are just adaptations of surface methods. Despite its interest, I shall not describe the SQUID which is extensively documented in an article by Dr Fischer in this issue. Finally I shall mainly describe the principle of the devices, rather than the technical details except when they are important to understand the way devices operate.

The measurement of the Earth's electric field is performed by measuring a difference of potential which is often very small (1 μV or less) but this measure is only simple in principle. In practice, it is often very difficult because of the presence of various noises

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which can be larger than the desired signal (noise of the input amplifier, probes, interference noise etc.).

Important technical improvements have recently been made concerning the amplification of weak signals, but it is difficult to remove the other sources of noise, especially when the measurements are performed off shore.

Devices designed to measure the magnetic field or its components involve a large number of physical principles and this leads to a wide number of different sensors. The first measurements of the transient field have been performed with suspended magnet variometers and with induction sensors. These two kinds of devices still give valuable information thanks to some improvements such as the field feed-back. For some particular applications, they are not outclassed by more recent devices. This is particularly true as far as variometers are concerned. Concerning absolute measurements it is better to use magnetic resonance devices. The use of this kind of device is compulsory when one needs to move the sensors quickly, as, for example, when establishing a magnetic map from a boat or a plane. Other devices are designed to measure very weak fields or weak magnetic gradients. Among these the SQUID is used more and more despite its high price and the problems associated with the use of liquid helium. Finally, the fluxgate is a kind of directional magnetometer, light, compact, useful for various applications requiring neither high performances nor good stability with time.

All modern devices except those designed to measure high frequency phenomena, are generally equipped with a digital output which makes the recording of the data and their processing easier. Some of the devices are equipped with microprocessors which make it possible to process the data in real time, in the field.

The methods based on the measurement of the electric or magnetic field are less numerous than the devices. The best known is the magnetotelluric method which gives good results in the case of horizontally layered structure or when the frequency of the electromagnetic wave is sufficiently high ($>$ 1 Hz). Different kinds of geomagnetic sounding, the most recent being the differential geomagnetic sounding method, make it possible to point out conductivity anomalies in the crust.

During the last 20 yrs, improvements have been mainly due to the recording techniques and computer progresses which allow one to process a very large number of experimental data.

The effort made to improve the signal to noise ratio have been mainly concerned with the development of more and more efficient mathematical algorithms. Additionally, the origins and the physical properties of the noise began to be analyzed, so that it could be better removed. In this respect, the method developped by Gamble (1979) is interesting, for it allows one to reduce the influence of the instrumental noise and to point out the parasitic signals. Despite this progress there is still a lot to do before we can understand how the currents are induced and flow within the earth's crust.

2. Electric Probes

Measuring the variable potential difference which appears between two electrodes is not difficult at high frequencies, above 10 to 20 Hz for instance. The electronic industry supplies at present low-noise and high-input-impedance amplifiers, perfectly adapted to the measurement of a few microvolts. It is more or less what one can get practically by using telluric lines, a hundred meters long. In this case the nature of the electrodes is not very important.

Let us quote an interesting technique due to Dupis and Guineau, (Guineau, 1975). The electric sensor is made of two copper plates ($S = 0.2\ m^2$), a few meters apart, isolated from the ground by a thin sheet of rubber. When the device slides on the ground, it is possible to get, at high frequencies, a continuous recording of the electric field, thanks to the capacitive connection of the probe with the ground.

On the contrary in the field of low or very low frequencies the measurement is much more difficult. Two electrodes buried in the earth form an electric battery whose electromotive force varies with many factors such as temperature, dampness or physical composition of the earth. The smallest variation of one of these parameters leads to an interfering signal, often much more important than the potential difference to be measured. In this case one uses preferably impolarizable electrodes made of a metal in contact with a solution of one of its own salts; the electromotive force of polarization does not disappear completely but it becomes much less important, of about one millivolt only. The electrochemical relations between the earth and the electrode are so complex that researches in order to improve the quality of the measures of the electric field are very often empirical.

One of the first electrodes made on this principle used the couple Ag-AgCl proposed by Filloux as early as 1967. It is still very widely used at the present time and is manufactured semi-industrially. One can also use the combination of cadmium-chloride with cadmium, or copper with copper sulphate or one can use a calomel electrode. Petiau and Morat (1977) compared the noise of different kinds of electrodes as a function of the frequency of the field to be measured. At 0.01 Hz the electrodes Ag-AgCl have a noise between 6 and 20 times weaker than most of the materials used (Figure 1). The best results with regard to long time stability are obtained using the couple lead-chloride, lead (Figure 2). The drift is of a few millivolts per year and the stability versus temperature is also very good (less than 50 $\mu V\ {}^\circ C^{-1}$). In the field it seems difficult to reduce the random drifts below 2 or 3 $\mu V\ hr^{-1}$ even with buried electrodes.

It is perhaps not necessary to have better performances. Indeed experience shows that the value of the measured electric field varies with the position of the station, even over very short distances. This effect is due to local distorsions of the current lines flowing in the shallow inhomogeneous layers of the earth. This does not forbid an accurate measurement of the electric field, but limits the physical meaning of the results. When the topographical noise becomes larger than the instrumental noise, it is no longer necessary to try to improve the precision of the measurements. In the sea, this problem is less disturbing, for the medium is relatively much more homogeneous. On the other hand random variations of temperature and salinity create signals to which Ag-AgCl electrodes are very sensitive.

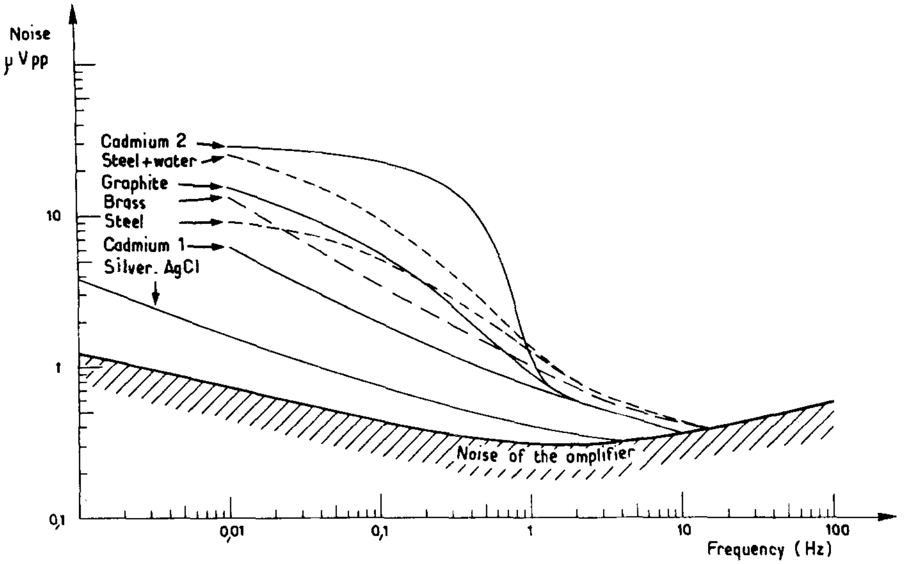

Fig. 1. Noise figure of different kinds of electrodes (from Petiau and Morat, 1977).

The 'Water shopping', imagined and realized by Filloux, remains the better device to eliminate the low frequency noise of the electrodes. One can measure by this means the natural variations of the telluric field using lines of only some meters. When longer lines are used (1 km), and for higher frequencies (1 Hz for instance) the signal to noise ratio is large enough and it becomes possible to amplify directly the measured signal (Cox, 1978).

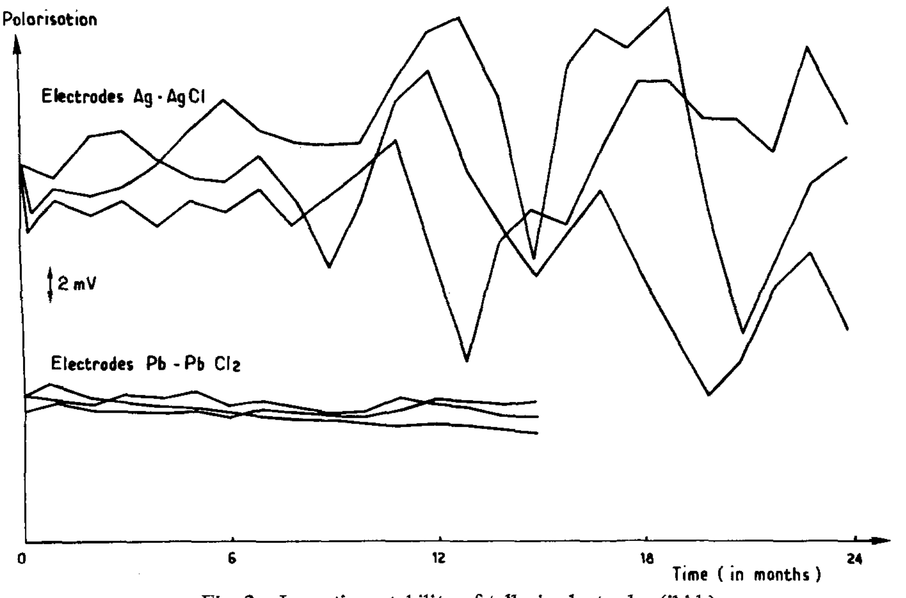

Fig. 2. Long time stability of telluric electrodes (ibid.).

3. Magnetic Sensors

The measurement of the magnetic field, or rather of its time dependent part – the only part of interest regarding induction studies – can be performed using very different kinds of sensors. Stuart (1972) and Serson (1973) have presented reviews of the devices used. They can be classified in five groups:

Nuclear magnetic resonance or optical pumping devices measure the modulus of the total field independently of its direction. They are absolute sensors, for the measured signal (the gyromagnetic frequency) is linked to the magnetic field by nuclear or atomic parameters and does not depend on the experimental conditions. These magnetometers, particularly those using free precession, have a long term stability much superior to other devices having a comparable sensitivity. They are very good observatory sensors. Cancelling some components of the Earth's magnetic field, it is possible to use these magnetometers as directional variometers. Conversely, it is possible to take advantage of their insensitivity to orientation, to use them on boat or plane, to make measurements on a wide scale. Nevertheless they are not very useful for the study of the transient electromagnetic field for they can easily measure only the component ΔF of the transient field parallel to the total field F. The other components can only be obtained by using compensating coils of large size, which make these sensors difficult to handle in the field. The component ΔF may nevertheless be used indirectly in some induction problems.

Suspended magnet variometers were the first devices used to study the transient magnetic field and are still widely used. They can be divided into two groups.

First, those for which the magnetic couple exerted on the magnet is compensated by a couple a different origin, for instance the torque of a wire. This principle is used in Askania variographs, still widely used and in the devices designed by Bobrov (1971) or Gough and Reitzel (1967). This last one, cheap and easy to build, has allowed researchers, to use large arrays of magnetometers and thus measure the field simultaneously in a great number of stations.

As the magnetic moment of the magnet and compensating torque vary with temperature according to different laws, such magnetometers have fairly large thermal drifts depending on the ambient thermal conditions. It is then necessary, either to control the device thermostatically, or to bury it, or to try to cancel the drift by that of another element, sensitive to temperature. The use of optical devices amplifying the rotation of the magnet makes these instruments difficult to handle and their photographic recording system does not permit easy data processing.

A second group includes variometers in which the variations of the natural field H are compensated by those – equal and opposite – of a field H^* created near the magnet by an electronic feedback system. The magnet is submitted only to the residual couple $\mathfrak{M}(H\text{-}H^*)$, extremely weak, and its direction in space is almost invariable. The small rotations due to the field $H\text{-}H^*$ are measured with a very sensitive detector and they are used for producing the feedback field after amplification. This kind of feedback device has several advantages with respect to other devices operating in open loop. The sensitivity, for instance, is constant in the whole passband of the variometer. Besides, the measure of the field H is reduced to that of the current which produces the field H*. It is an electrical signal easy to record either analog or digitally. Finally, the equilibrium position of the magnet does not depend on the value of its magnetic moment and the temperature drift is much weaker than in the previous group.

In the first devices the detector consisted of two photoelectric cells measuring the rotation of a light beam reflected by a mirror clamped to the magnet. (Figure 3). Jolivet (1964) and Adam and Major (1967) built such devices. Moreover, these variometers were made insensitive to shocks and vibrations by immersing the magnet in a liquid of proper density. At least in the case of the Jolivet magnetometer, this technique has not proved to be successful. Indeed, very small thermal gradients can produce convection currents in the liquid and cause large drift. Mosnier (1970) developed a comparable device but the magnet was not immersed in a liquid and he has obtained better performances (noise less than 0.02 nT for periods of several minutes).

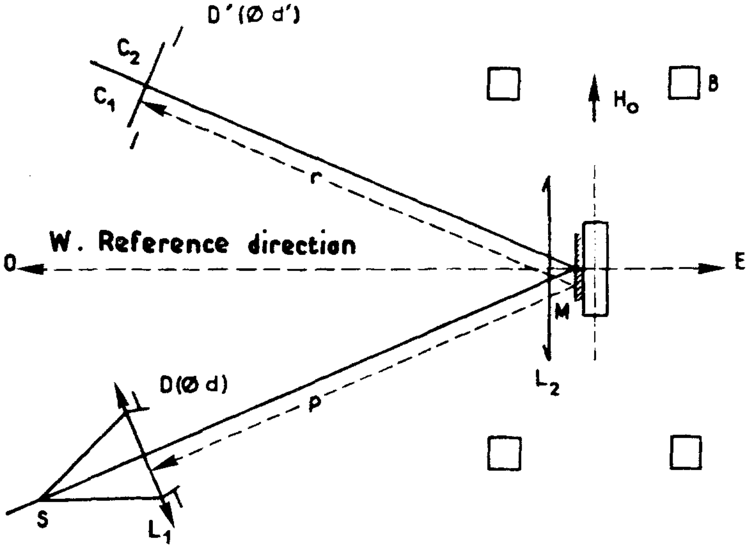

Fig. 3. Schematic diagram of a *D* variometer with a negative feedback field.

The variometers using a photoelectric feedback are generally characterized by a high electrical consumption. It can be reduced by replacing the photoelectric detector by a capacitive one (Mosnier and Yvetot, 1976), (Figure 4). The power consumption is, in this case, of only 0.2 W. The dimensions and the weight of the sensor are also reduced but its performance is worse than that of the previous one. The thermal stability is rather good (0.1 nT $^{\circ}C^{-1}$) but thermal gradients must be avoided for they can create large drifts. These drifts are probably due to thermo electrical currents flowing in the metallic part of the variometer. Important improvements should be attainable in this domain.

Suspended magnet variometers can be used on the sea bottom at the price of some complications due to the fragility of the suspension wires and to the impossibility of orientating the devices once immersed. Filloux (1967) has built an original declinometer in which the magnet orientates itself along the magnetic meridian before being suspended.

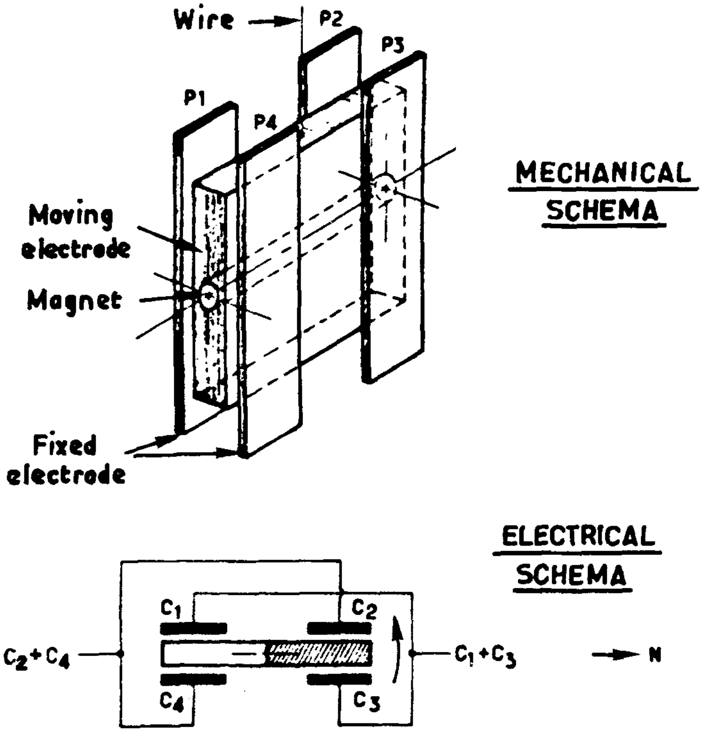

Fig. 4. Mechanical and electrical schemas of a capacitive detector. (from Mosnier and Yvetot, 1977).

Mosnier and Durand (1978) built a submarine device including two orthogonal sensors associated with an automatic device compensating the earth magnetic field. The data are transmitted to land by a radio buoy (Figure 5). The two N-S and E-W components of the field are then deduced by means of an analogue computer. The stability reaches 0.1 γ for periods of 10 to 15 min.

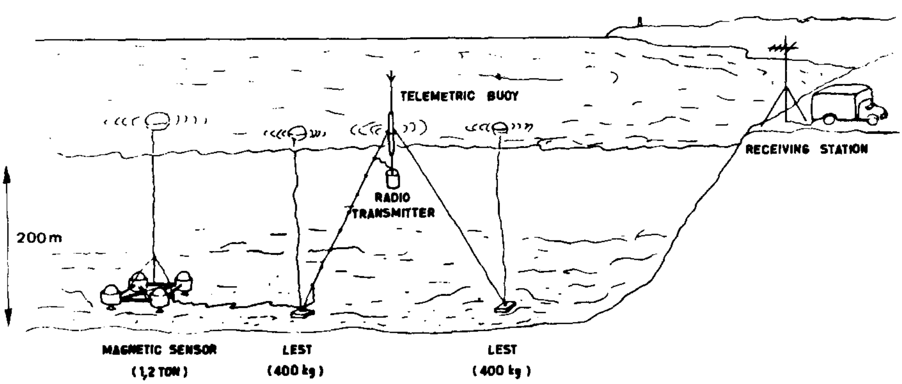

Fig. 5. Suspended magnet variometer and telemetric buoy, used for sea bottom measurements (from Durand and Mosnier, 1977).

A third group of variometers comprised devices with saturable cores, using the hysteresis of ferromagnetic materials. If such a material is placed inside a coil and is excitated by a sinusoidal magnetic field H, of frequency f, one obtains at the coil terminals, a voltage which contains only odd harmonics. The presence of a field H_0, added to the field H, creates a signal of frequency $2f$ which cancels out and has its sign changed when H_0 goes across 0. Using this signal, it is possible to build a feed-back system to cancel out the field H_0. Most of the fluxgates work according to this principle, to within some alternatives (Figure 6). Gordon and Brown, (1972) gave a rather full report on this kind of device. A recent one is due to Primdahl (1979).

Variometers with saturable cores are compact, reliable and rather sensitive. Their most important drawback is their thermal drift, the variations of temperature changing the electrical properties of the core. Morton and Suckau (1976) realized a device stable enough to be used in the Australian observatories of the Antartic. Comparable devices equip the French stations in the same area. Other kinds of fluxgates are used for submarine or space applications (Mac Pherron *et al.*, 1975) Several industrial companies sell equipment measuring the three components of the field with an accuracy of 0.1 to 1 nT and thermal drifts of about 0.1 to 0.2 nT $^{\circ}C^{-1}$.Most of them are equipped with a numerical output, allowing an easy recording of the transient field. On the whole, these devices are well adapted to the study of induction phenomena, except for very low frequencies where thermal drifts are particulary disturbing.

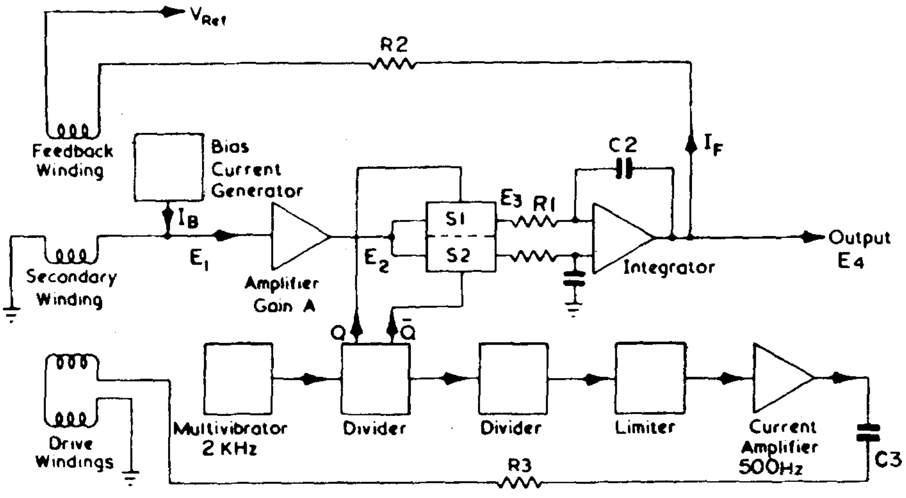

Fig. 6. Typical block diagram of a fluxgate variometer (from Morton and Suckau, 1976).

A fourth group of devices comprises induction sensors. Once more, their principle is very old. The variations of the magnetic field H induce, in a coil having a highly permeable ferromagnetic core, an electromotive force proportional to dH/dt. In order to obtain $H(t)$ the coil must be attached to an integrating circuit. Devices built on this principle differ from one another in the way one tries to correct some unavoidable drawbacks, of which the main ones are:

– a decrease of the sensivity with frequency which makes the use of inductive sensors below 10^{-3} Hz very difficult.

– the existence of a resonant frequency of the sensor which creates modulus and phase variations depending on the frequency of the field to be measured.

The use of feedback flux (Figure 7) (Clerc, 1971) has been an important improvement. Once more, the principle lies in the compensation of the natural variations of the field by a current produced by an electronic feedback system and flowing in an additional coil of some tens of turns. The main coil is used only as a null detector and its characteristics are no longer important. So, with a single coil, it is possible to obtain a flat passband from 3.10^{-4} Hz to several hundreds Hertz. Towards the high frequencies one can reach 30 kHz with ferrite cores and 300 KHz with air coils. The noise level is 10^{-4} nT Hz$^{-1/2}$ towards 1000 Hz.

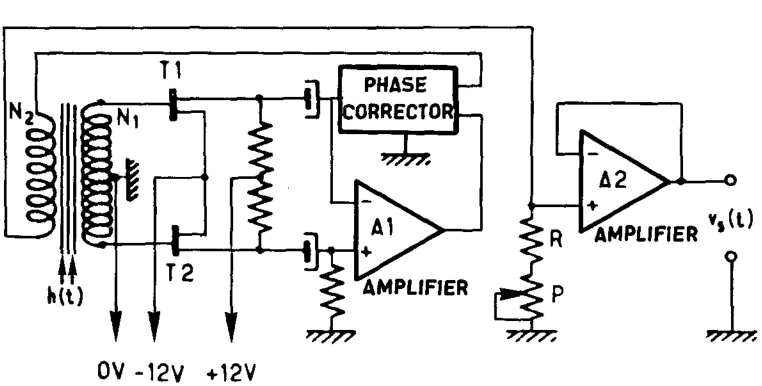

Fig. 7. Schematic diagram of an inductive variometer with a negative feedback (from Clerc, 1970).

Like fluxgates, different kinds of induction variometers are developed industrially for very different purposes. They are reliable devices with a low power consumption, the sensitivity of which can be very good when they are associated to good quality amplifiers. Their optimum frequency domain is usually above 1 Hz. For low frequencies they are surpassed by the SQUID and even by suspended magnet variometers. If one accepts the loss of a certain amount of sensitivity, one can miniaturize them so that they can be used in rockets or satellites. Other models are used for measuring the field in boreholes. The changes with time of the magnetic properties of the cores sometimes causes a modification in their characteristics. It is advisable to calibrate these instruments periodically in amplitude and phase. This can be done with a calibration station allowing the creation of uniform, controlled fields in a large enough volume. Otherwise one can use the method proposed by Zambrevsky *et al.* (1980).

The SQUID (Superconducting quantum interference device) forms the basis of a fifth group of variometers which are in fact the only actual improvement realized for over 10 yrs in the measure of weak geomagnetic fields. The principle and the performance of these sensors – which have been realized under different forms – are reviewed by Dr Fischer. I will not detail them here except to mention that the need for liquid helium is a serious drawback but does not forbid their use for field experiments. Dinger *et al* (1977) have even developed a submarine SQUID, operating to a depth of 100 m. From a technical

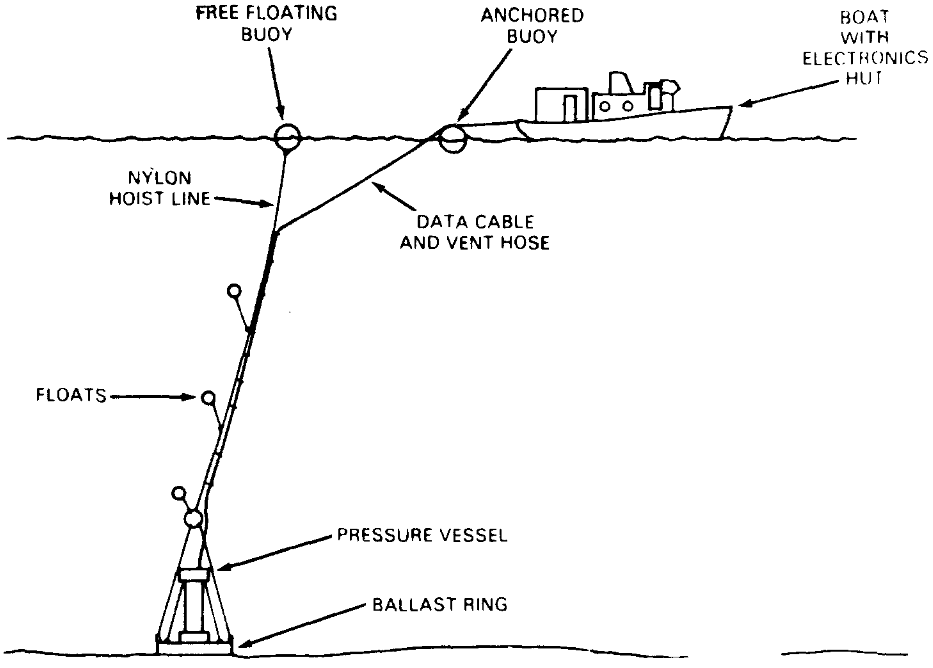

Fig. 8a. Ocean bottom superconducting variometer: Schematic diagram of the operating configuration (from Dinger *et al.*, 1977).

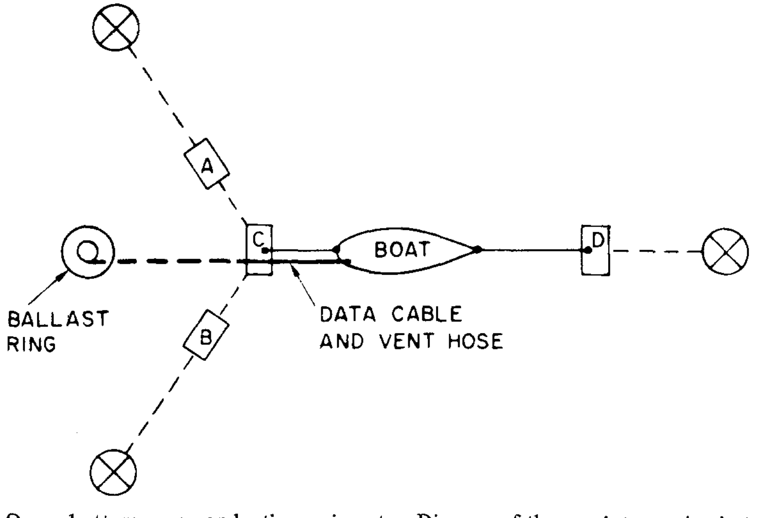

Fig. 8b. Ocean bottom superconducting variometer: Diagram of three point mooring in taut position.

point of view, their use is still complicated for one needs a boat at a fixed place and another one moving around. There are still some practical problems to be solved as, for instance, the evacuation of gas under pressure; Nevertheless it can be hoped that small liquefiers working in closed circuit will soon be available. The SQUID would then become operational for submarine researches, even at great depth.

In addition to previously described devices which are frequently used for measures in the field, other apparatus can also be used for studying weak magnetic fields.

Thin shell magnetometers are very similar to fluxgate ones. An exciting field – of frequency f – applied on the sensor creates a secondary sinusoidal field. In the presence of a slowly variable field a $2f$ frequency signal appears. This last signal can be used in a feedback loop. The noise level is of 0.05 nT $Hz^{-1/2}$ for a frequency of 1 Hz and decreases with frequency. This kind of variometer cannot be used for very low frequency fields but can be used for rapid variations.

Magnetodiodes use the effect of magnetoconcentration which modifies the electric properties of a semi-conducting junction. These devices are cheap and small, but their sensitivity is still insufficient for geophysical applications.

The performances of the variometers we have just quoted are largely variable and often excellent. Nevertheless one can wonder whether – as in the case of the measure of the electric field – there exists a limit beyond which it is useless to improve these performances. This question is of some importance particularly with regard to the SQUID for their theoretical sensitivity is at least 100 times better than that of other sensors.

The noise of a variometer depends on three factors:

– The noise of the device itself, which is mainly of electronic origin.

– The noise due to mechanical movements of the sensor in the Earth's magnetic field.

– The electromagnetic noise which is a signal of no interest for the experimenter.

Concerning the SQUID, the first of these three noises is very weak (10^{-5} nT $Hz^{-1/2}$). If one can eliminate the two other noises, one can get high performance. This is easily obtained when two SQUIDS are fixed in a rigid support and when they are used as a gradient-meter. In these circumstances, the mechanical vibrations and all the field which is homogeneous in a volume equal to that of the sensor are completely cancelled out. This is the reason why superconducting gradient-meters are decisively better than all the other known devices.

Now if one considers an isolated SQUID or a gradient-meter of great length, in which the sensors are mechanically independent, one must take account of mechanical vibrations of each variometer which cannot be totally avoided, for instance those produced by the helium boiling in the cryostat. Performances are then less good.

The vibrations and the electronic noise of the device are instrumental noises. One can hope to reduce their influence on the final result by techniques of mathematical processing of the signal. The solutions adopted by Anav *et al.* (1976) and Gamble *et al.* (1979) are based upon the hypothesis of the noise incoherency and of the signal coherency. It is then possible to improve the signal-to-noise ratio considerably by means of appropriate statistical processing.

But this does not apply to electromagnetic noise which is coherent to a great extent but with a spatial distribution generally different from that of the required signal. The techniques of correlation are not sufficient to eleminate it. All things considered, one must admit that there is a 'topographical noise' of the magnetic field which has the same origin as that observed in measurements of the electric field. It is less important, for magnetic effects are not very sensitive to local variations of resistivity but it is often greater than the device sensitivity. Thus, beyond a certain degree of precision, the result loses its physical meaning.

4. Observational Techniques

The electric and magnetic fields measured with the devices we have just described can be used in different ways to determine the distribution of resistivity inside the earth and under the sea.

For instance, one can measure simultaneously at one station the electric field and the magnetic field and study the relation between E, H, ω (pulsation of the electromagnetic wave) and ρ. This is the magnetotelluric method of Cagniard and Tikhonov.

One can also measure, at two or several stations, only one of the electric or magnetic fields (usually the second one) and look for a relation between H_1, H_2 ..., ω and the geometrical and electrical parameters of the layers. It is the principle of different telluric and magnetic soundings, the most well known being 'Geomagnetic deep sounding' (GDS).

Contrary to the magnetotelluric method which yields its best results in the case of stratified media, magnetic sounding applies only to inner anomalies of conductivity. Under these conditions one must use at least two stations – one of them located outside the anomalous area will be used as a reference station. The GDS is based upon the idea that the normal transient field has no vertical component. If a vertical component is observed, it is necessarily produced by a discontinuity of conductivity. One tries then to establish a time independent relation between the vertical component Z_a (anomalous) at one point and the horizontal components North-South, H_N, and East-West, D_N, at the normal reference station. The method, developed from Weise's and Parkinson's works, rests broadly on those of Rikitake (1953) and Schmucker (1964) who introduced the use of transfer functions. These are complex frequency-dependent coefficients A and B such that, in the frequency domain, the anomalous field Z_a can be expressed as:

$$Z_a = AH_N + BD_N .$$

To map a region quickly enough one uses simultaneously a great number of stations from which it is concluded that arrays of magnetometers such as those developed by Gough are interesting.

For the past 10 yrs this technique has given information on the circulation of telluric currents and has made it possible to realize what a complicated question it is.

Among recent studies let us quote those of Richards *et al.* (1979) in the Rhinegraben, Cochrane and Wright (1977) at Newfoundland, (Figure 9) Bailey *et al.* (1978) in the state of Maine (U.S.A.) and Woods and Lilley (1979) in Central Australia. Gregori (1980) compared the different methods of magnetic sounding using the Z component and showed that they were broadly equivalent to one another (see also Parkinson and Jones, 1979).

One can try to use the horizontal components H_a and D_a instead of the vertical

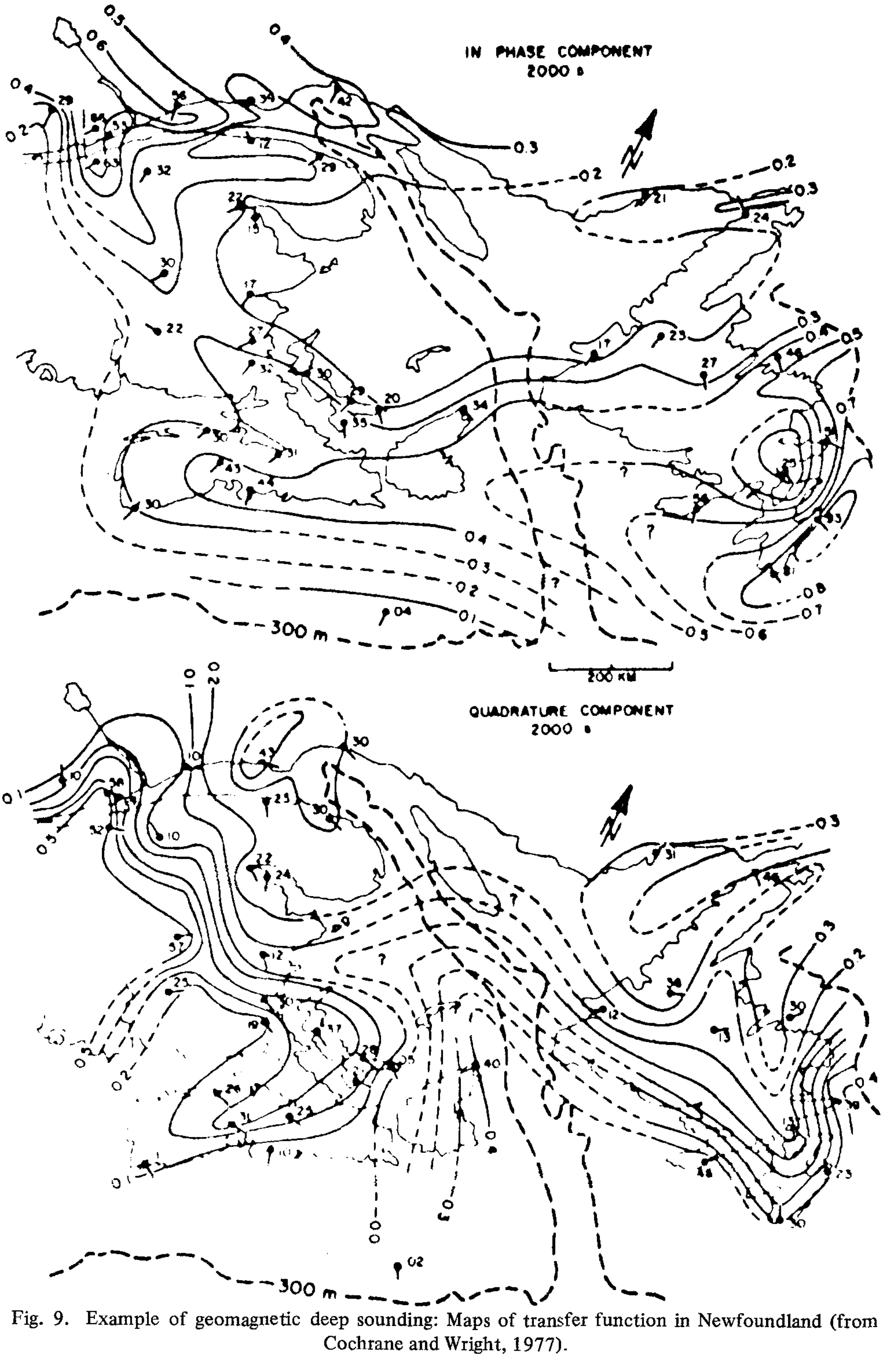

Fig. 9. Example of geomagnetic deep sounding: Maps of transfer function in Newfoundland (from Cochrane and Wright, 1977).

component Z_a but it is then necessary to separate the normal components H_N and D_N present in all the stations. If one assumes that H_a and D_a are zero at the reference station and that H_N and D_N are uniform, one has only to calculate the difference between two homologous components measured, one in the anomalous area and the other outside, in order to obtain the anomalous transient field alone. Thus one makes a 'differential magnetic sounding'. This method developed by Babour and Mosnier (1973, 1977), is not yet widely used except in France; it has the disadvantage of requiring a large amount of equipment if one wants to measure the field in a sufficient number of stations at one and the same time. On the other hand it makes a more precise localization of the anomalous areas, and gives information on the local orientation of the structures. Recent studies were concerned with the Rhinegraben (1978) and particularly with the Pyrenees where the GDS enabled one to locate an exceptionally important anomaly (Babour *et al.* 1976, 1977) (Figure 10).

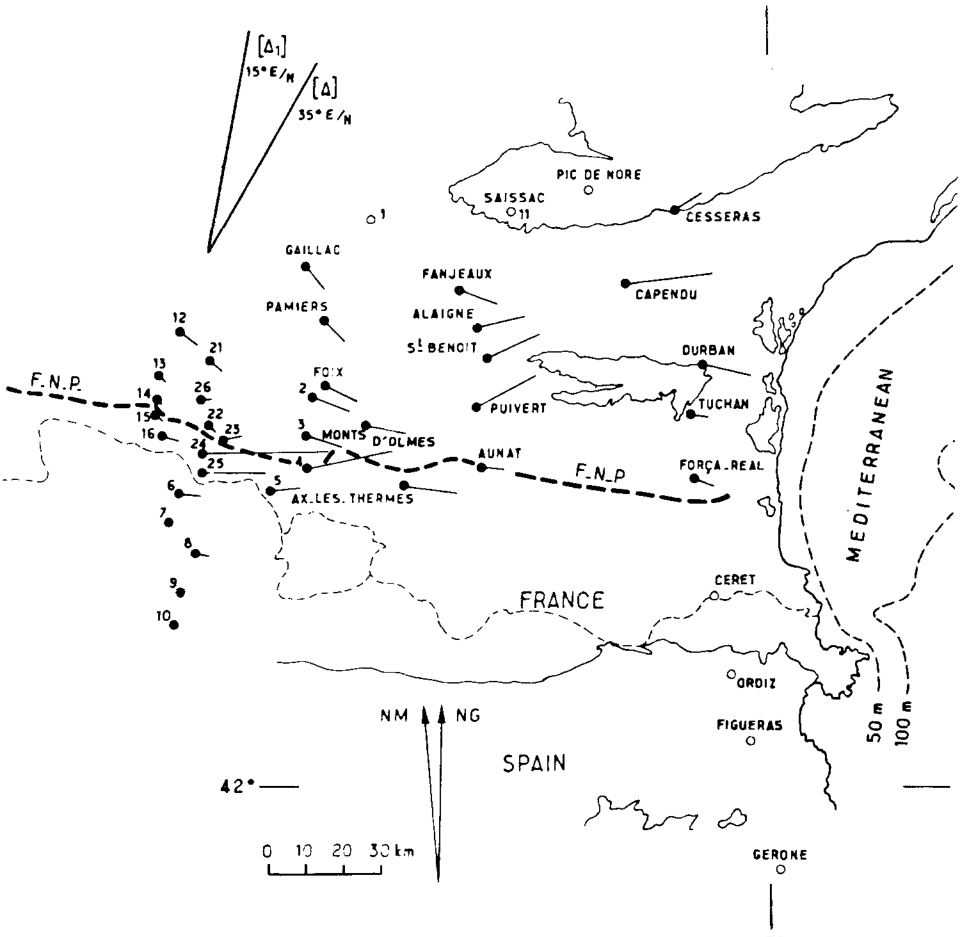

Fig. 10. Example of differential geomagnetic sounding: Map of the telluric flow in the eastern part of the Pyrenees. (from Babour *et al.*, 1977).

From a practical point of view and in order to perform synchronous measurements, a telemetric link is used between every station and the reference station. It is also possible to use independent stations piloted by very stable clocks. Theile and Luehr (1976) have been able to compare the transient field in two distant stations by using an ordinary telephone connection. Le Mouel *et al.* (1977) have used a variant of the GDS in using only Askania variometers deprived of any synchronization system. The synchronization is performed later on, in the laboratory, using easily recognizable magnetic events as reference marks. The accuracy so obtained is middling but the results are still acceptable if the anomaly is large (Galdeano *et al.*, 1980).

Aeromagnetic surveys with proton or optical pumping variometers allow one, to a certain extent, to study also induction phenomena. Indeed, when flightlines cross, the field values measured at the intersection must be the same. To get rid of the influence of the field variations during the flight, one subtracts from the raw value the amplitude of the transient field measured at a reference station. In spite of this precaution, differences are often found between values observed at the same point, these differences being larger than the errors in measuring. Le Borgne and Le Mouel (1975) use these residual differences that they attribute to inductive phenomena, to point out current channelling in the Atlantic ocean and in the Mediterranean sea. Gregori and Lanzerotti (1979) propose to improve their method by changing the law of interpolation between the references and the station of measure. As far as I am concerned I do not think that the results so obtained are satisfactory for several reasons, the most important of which is that resonance magnetometers are almost insensitive to the East-West component of the field. So they cannot point out a North-South anomalous telluric circulation, which is enough to alter interpretations.

On the other hand they are very interesting when applied to the method called 'Sounding by magnetic gradient'. Let us suppose that one measures the total field F at the surface and at the bottom of the sea. Far from every anomaly the variation of the vertical component Z is zero. The time variation $F(t)$ is then equal to $H(t) \times \cos I$, where $H(t)$ is the horizontal transient field North-South and I the inclination. The ratio of the values of $F(t)$ measured at the surface and at the bottom is a function of the conductances of the sedimentary layer located under the ocean that one can then calculate. This method seems to have been developed exclusively in the Soviet Union (Trofimov and Fonarev, 1972; Sochel'Nikov and Chzhu, 1978). The shielding effect produced by seawater, which is conductive, considerably reduces the amplitude of the rapid variations of the external field (Filloux 1977), which seems to limit the implications of the method towards high frequencies. In this band one can try to use the magnetic field produced by surface waves. Theoretically such a signal should not reach the bottom of the ocean if it is at great depth. Cox *et al.* (1978) have shown that, in fact, interference between acoustic waves propagating in the sea could create movement in the water and consequently a magnetic field induced at a frequency twice that of the sea waves. So the electromagnetic sounding of the bottom is possible towards 0.2 Hz. At the other end of the spectrum one can use sea currents. Experiments made in the Atlantic ocean, show that currents induced at these frequencies hardly penetrate the sedimentary substratum. This method is nevertheless used in the Soviet Union (Lejbo 1979; Sochel'Nikov 1979) particularly in the Artic Ocean and the sea of Barents.

Finally let us quote, although it is not strictly speaking an experiment of induction, an attempt to determine the conductivity of the bottom of the oceans thanks to an artifical current source. If an alternating voltage, the frequency of which is about 1 Hz, is applied to two electrodes connected to the sea floor, electromagnetic waves do not propagate in water (because of the skin effect) but they can propagate in sediments which are much more resistive. One achieves a kind of dipole sounding similar to the one made on land for prospecting; this makes it possible to calculate a medium conductivity of the substratum. This method has been proposed by Cox (1978) and, nearly at the same time, by Shaub (1978). The experiment was made during a campaign in the eastern Pacific (Cox *et al.*, 1980) and allows one to measure conductivities of 0.004 S m^{-1} much less than those determined from samples (0.03 S m^{-1}) (Figure 11). This result is interesting for it helps to explain current exchanges between ocean and mantle, but a lot of work is still needed to get a clear idea of the conductivity distribution at the bottom of the seas.

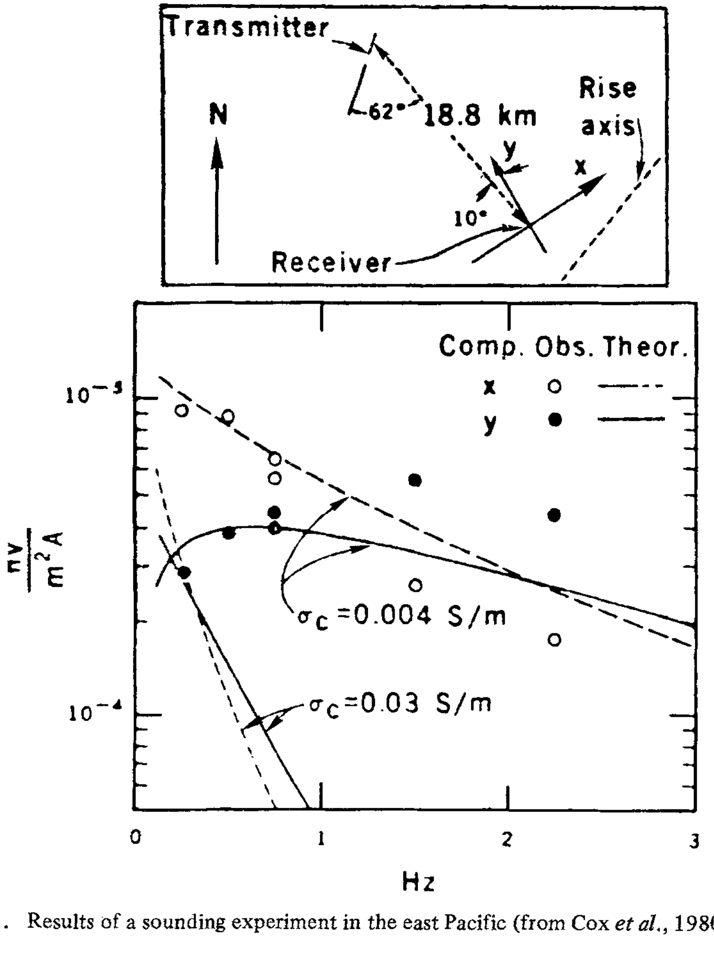

Fig. 11. Results of a sounding experiment in the east Pacific (from Cox *et al.*, 1980).

5. Conclusion

To end with, two questions must be asked.

(1) Is there a kind of device or a method of observation better than all the others to study the natural electromagnetic field.

(2) What kind of progress can be hoped for in this domain during the next few years?

The answer to the first question is: No. All the variometers we have just described have different characteristics. They have been conceived with a view to special applications to which they are more particularly adapted. Many studies use several instruments at one and the same time (Richards, 1978). One cannot really consider that one kind of device only, not even the SQUID, is adequate to solve all the problems. The same holds true in respect of methods which very often appear to be complementary.

On the other hand it is important that instruments and techniques should be compared to one another as often and as carefully as possible. It is only in that case that researchers will be able to exchange their experimental information and to enter upon international research programs of some value. Considerable progress is still to be made in this domain.

Concerning the second question, all I can do is to express my opinion. I do not think that the important progress one can hope for in the study of natural induction phenomena during the next decade will come from an improvement in the methods of observation. Those we already have can give a sufficiently precise idea of the distribution of the electromagnetic field. But there are still two important points to be studied.

The first one is the structure of the primary exciting field. It is obvious that its assimilation to a plane wave is a gross simplification, at least in the study of very low frequency phenomena. In order to interpret experimental data correctly, it is necessary to know the distribution of sources, which supposes a very serious effort in the experimental domain. Bannister and Gough (1978) studied on several occasions, using arrays of magnetometers, the shape of the ionospheric currents flowing in the polar areas during magnetic substorms. There are works devoted to the study of the equatorial electrojet. Schmucker and Weidelt (1976) examined the influence of a variation of the amplitude of the primary field with latitude. But there is still a lot to do before having a valid model of the transient field on the scale of the Earth.

The second point on which I think considerable progress can be made is the understanding of the mechanisms responsible for the distribution of the field induced at a planetary scale. For several years the importance of 'edge effects' due to current exchanges between the crust and the mantle has been realised (Ranganayaki *et al.*, 1980) and new theories are developed concerning global induction models nearer reality than present models. It is possible that this will lead to the introduction of new experimental methods, but I think that it should be better to deal with this problem by creating great international programmes of observation.

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