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Correlation between frozen ground thickness measured in Antarctica and permafrost thickness estimated on the basis of the heat flow obtained from magnetotelluric soundings

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Abstract

A comparative study of the results of vertical electrical soundings (VES) and magnetotelluric soundings (MTS) carried out at Seymour and James Ross Islands (NE of Antarctic Peninsula) is conducted. Frozen ground thickness, estimated by VES, is compared with permafrost thickness estimates, obtained using: (a) steady geothermal heat flow, inferred from the depths of conductive layers in the crust and upper mantle estimated by MTS, and (b) mean annual air temperatures (MAAT). At Seymour Island, 250 m of permafrost thickness, in equilibrium with an inferred heat flow of 72 mW/m^2 (corresponding to a geothermal gradient of 0.037 °C/m) and with an MAAT of −9.4 °C, is estimated for the upper terrace (about 200 m a.s.l.). This value is consistent with the frozen ground thickness (200 m in the upper terrace) previously estimated by VES. Magnetotelluric soundings carried out in the northwestern region of James Ross Island (volcanic island with recent activity) suggest a magma chamber with top at about 7 km depth in the crust. From the depth of this conductive body, a heat flow of 145 mW/m^2 and a geothermal gradient of 0.074 °C/m are estimated, suggesting 67 m of permafrost thickness in the upper terraces (35 m a.s.l.) of this island in equilibrium with the estimated steady heat flow and an MAAT=−5°C. The frozen ground thickness estimated by VES lies between 40 and 45 m for the same terrace. The differences between frozen ground thickness and permafrost thickness in both islands could be attributed to a cryopeg, according to the high salinity beneath the frozen ground suggested by MT soundings. Although studies indicate that a climatic warming process is currently taking place in the Antarctic Peninsula, results of this study suggest that, until approximately 1980, the region had stable geological and paleoclimatic conditions during a lapse of time long enough to reach an approximately steady temperature profile in the subsoil. Around 2000 years are estimated for Seymour Island and 140 years for the northwestern region of James Ross Island as the minimum periods of time during which the soil surface was free of ice. The magma chamber, possibly located in the northwestern region of James Ross

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Island, could have an important role in controlling permafrost thickness, conductance below the frozen ground and deglaciation.

Keywords: Permafrost; Antarctica; Heat flow; Magnetotellurics

1. Introduction

The islands Seymour and James Ross are located in the Weddell sea, northeastern region of the Antarctic Peninsula (Larsen Basin, Antarctica), at about 64°S and 57°W (Fig. 1). Magnetotelluric soundings (MTS) were carried out in Seymour Island in the summers 1978–1979 and 1979–1980 (Fournier et al., 1980, 1987, 1989, Del Valle et al., 1982, Pomposiello et al., 1988, Fournier, 1994) and in James Ross Island in 1992 (Mamaní et al., 1998) to describe the electric resistivity variation with depth in the crust and upper mantle, and infer the geological structure of the Larsen Basin. According to the range of periods of the natural electromagnetic field measured in these studies (0.1 s and higher), only the crust from depths of about 600 m and more, and the upper mantle could be studied, but not the permafrost. Therefore, in the frame of these studies, only the thickness of the Larsen basin, its stratigraphic configuration and the presence of a conductive layer (possible asthenosphere) in the upper mantle were determined. However, the MTS gave indirect evidence about a very conductive (saline) layer present beneath the frozen ground of these islands (Fournier et al., 1987, Mamaní et al., 1998).

In 1982, Corte (1982) made the first inventory and interpretation of periglacial features and processes in Seymour Island and made the first gross estimation of permafrost thickness. In the following years, geophysical studies particularly devoted to measure frozen ground thickness were conducted in the region: in Seymour Island, in the summer of 1987–1988 (Buk, personal communication), and in Seymour and James Ross Islands in 1989–1990 (Fukuda et al., 1992). In these studies, vertical electrical soundings (VES) were carried out using Schlumberger and Wenner electrode configurations (Kunetz, 1966).

VES were carried out by McGinnis et al. (1973) in the Dry Valley (Antarctica). According to these

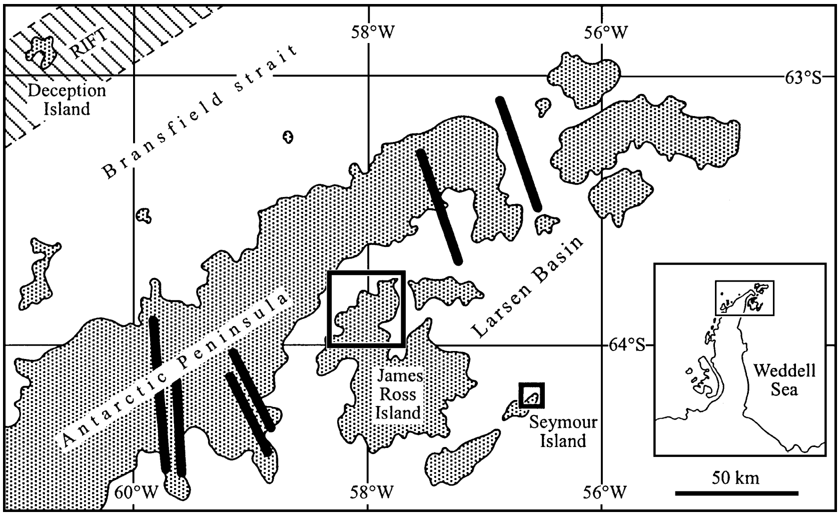

Fig. 1. Simplified map of the Antarctic Peninsula region showing the Seymour and James Ross Islands, located in the Larsen Basin. The frames indicate the zones where MTS studies were carried out (after Fournier et al., 1980 and Mamaní et al., 1998). Heavy solid lines denote fracture zones, after Del Valle and Rinaldi (1992). A rift is present at the Bransfield Strait.

authors, resistivities lower than 1000 Ω m may indicate unfrozen sedimentary rocks, which could be below 0° C (permafrost), however, because of the high salt concentration in the pore fluids. Resistivities between 1000 and 10,000 Ω m are indicative of bedrocks of frozen ground conditions in unconsolidated materials, and resistivities higher than 10,000 Ω m may indicate frozen basement rocks. Electrical resistivities of about 4100 Ω m were also measured by Fournier et al. (1986) at the Vallecitos rock glacier (approximately 3600 m a.s.l.) using VES (Frontal Cordillera, Andean Range, Mendoza, Argentina, 33°S) to obtain presence and thickness of mountain permafrost. Near the same site, however, Barsch and King (1989) and Trombotto et al. (1999) interpreted the presence of permafrost with resistivities of much higher values, between 30,000 and 50,000 Ω m in the first case and more than 8500 Ω m in the second case. This wide spectrum of resistivities might depend on the chosen sites, which have differences in the granulometry and consolidation of the sediments or the genesis of the ice.

In the present work, a correlation and comparative study of VES and MTS results is made. From the depths of conductive layers determined in the crust and upper mantle by MTS, heat flows in steady state and geothermal gradients are estimated. The permafrost thicknesses then estimated, using the geothermal gradients and the mean annual air temperatures (MAAT), are correlated with the frozen ground thicknesses estimated by VES in order to infer whether or not the permafrost, present at Seymour and James Ross Islands, is near equilibrium with the related physical parameters, and therefore, whether paleoclimatic and geological changes occurred in the recent past.

The VES and MTS results obtained by different authors at Seymour and James Ross Islands from 1979 to 1992 will be described and analyzed.

2. Methodology

The VES is a geophysical exploration method that allows us to estimate frozen ground thickness. “An electrical sounding, or vertical resistivity profile, consists of a succession of apparent resistivity measurements made with an increasing electrode separation, the center of the configuration and its orientation remaining fixed” (Kunetz, 1966, p. 50). Two pairs of electrodes are used: two electrodes (AB) to create an electrical field in the ground, and two others (MN) to measure the difference of electrical potential between two points on the ground surface. From the measured values of electrical current and potential, the apparent electrical resistivity is calculated for each electrode separation. The apparent electrical resistivity variation as a function of electrode separation is a measure of the electrical resistivity variation with depth.

The magnetotelluric sounding is a method for studying the electrical conductivity of the Earth’s interior, based on the skin-effect and using the Earth natural electromagnetic field. This field is supposed to be formed (Tikhonov-Cagniard model) by plane electromagnetic waves normally incident on a plane-layered Earth. By measuring the ratio of the electric to magnetic horizontal components of this field (impedance) at a physical point on the Earth’s surface and over a wide range of periods, the electrical resistivity variation with depth can be evaluated using the skin-effect (the larger the period of the electromagnetic wave, the lower its attenuation in the subsoil). For a complete description of the MTS method, see for example Kaufman and Keller (1981), Rokityansky (1982), among many others.

3. Study area

The study area is conformed by the islands Seymour and James Ross, next to the northeastern border of the Antarctic Peninsula in the Larsen Basin (Fig. 1). It is an area of Antarctic continuous permafrost (Trombotto, 1991).

The Antarctic Peninsula is principally composed of Mesozoic granitic rocks intruded into a Permo-Triassic metamorphic basement (Ramos, 1999). The northeastern extreme of this peninsula has a block structure. Four major fracture zones have been observed in this region (Fig. 1), transversal to the edge of the Mesozoic folding of the Antarctic Peninsula (Del Valle and Rinaldi, 1992). An important rift structure at the Bransfield Strait (NW of the Antarctic Peninsula, Fig. 1) is also present, where high heat flow values (about 220 mW/m^2) were registered (Ghidella et al., 1997). The Larsen Basin is located east of the Antarctic Peninsula, between the eastern border of this peninsula in the west and the islands James Ross and Seymour in the east (Fig. 1). This basin is formed by Jurassic and Cretaceous rocks covered by alkaline basalts (Pliocene–Pleistocene) (Ramos, 1999). The geology of the islands is presented in Fig. 2.

Seymour Island is located at 64°15′S and 56°45′W (Fig. 3), about 100 km east of the Antarctic Peninsula (Fig. 1). Tertiary strata crop out in its northern section (Sobral, Cross Valley and La Meseta Formations) (Figs. 2 and 3), centered about a meseta at 200 m a.s.l., and apparently arranged in a broad syncline (Fig. 3) (Elliot et al., 1975). According to these authors, this Tertiary sequence is about 500 m thick. Upper Cretaceous rocks crop out in the southern region of Seymour Island (Fig. 3) (López de Bertodano Formation), and similarly to the Tertiary sediments at the northern area, they are of a very

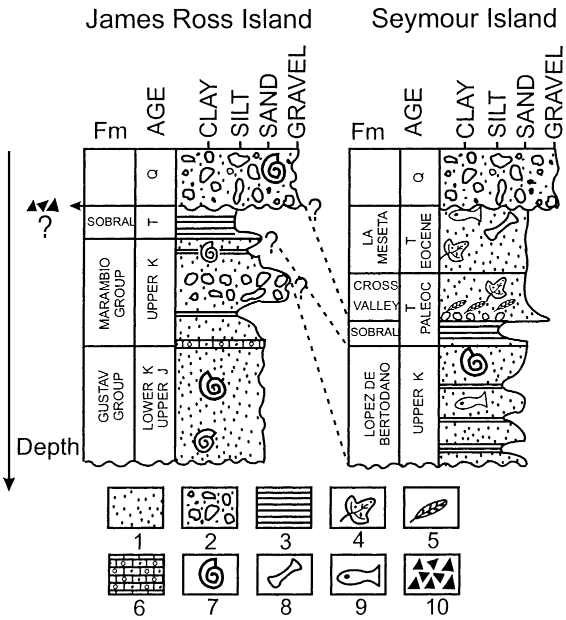

Fig. 2. Schematic stratigraphic sections of Seymour and James Ross Islands. (1) Sandstone, (2) Conglomerate, Gravel, (3) Tuff, Siltstone, Glauconite, (4) Flora, (5) Coal, (6) Calcareous sandstone concretions, (7) Marine invertebrate fossils, (8) Terrestrial and coastal fossils, (9) Marine vertebrate fossils, (10) Volcanic rocks. J, K, T and Q denote Jurassic, Cretaceous, Tertiary and Quaternary, respectively. "Fm" denotes Formation. Without vertical scale.

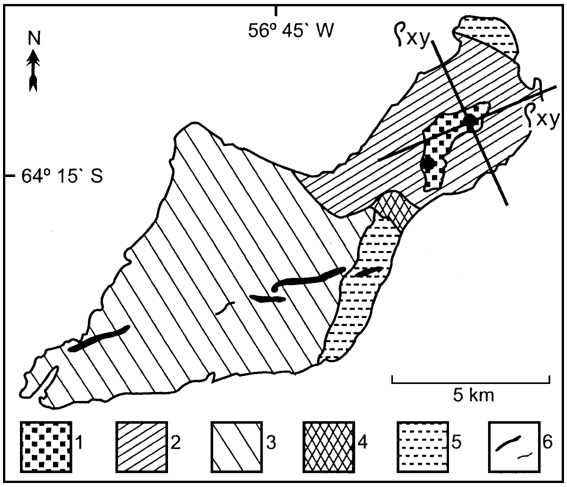

Fig. 3. Simplified geologic map of Seymour Island showing: (1) Quaternary glacial deposits, (2) "La Meseta" Formation, (3) "Lopez de Bertodano" Formation, (4) "Cross Valley" Formation, (5) "Sobral" Formation and, (6) Basaltic dikes, (after Zinsmeister, 1982). Black circles indicate MTS locations. Heavy solid lines indicate principal directions of MTS (N23°W and N67°E), after Pomposiello et al. (1988).

porous unconsolidated nature. According to Zinsmeister (1984), the sheltered location of Seymour Island—east of James Ross Island—is responsible for the little snow it received, and consequently this island is free of permanent ice. The unconsolidated nature of the rocks and the absence of vegetation on Seymour Island have caused a badland topography, typical of desert regions. Eruptive rocks (dikes) outcrop mainly in three areas, affecting the Cretaceous formations. Important fault systems (with ENE–WSW and ESE–WNW strikes) are also present on this island (Rinaldi et al., 1978). Five Quaternary terraces with erratics were identified at 1–2, 4, 18, 35 and 200 m a.s.l. (Zinsmeister, 1980). According to Zinsmeister (1980), the first four are clearly recognized as marine terraces, while the fifth (meseta at about 200 m a.s.l.) could have been formed by glacial erosion because of its erratics. A cryoplanation, however, was also suggested as the origin of the upper terrace (Corte, 1982).

James Ross Island is located at 64°S and 58°W (Figs. 1 and 7). It is a Plio-Pleistocene volcanic island with recent activity on its eastern border (Strelin and Malagnino, 1992) with about 70% of the surface covered by glaciers, except its northwestern region and some areas near the coast (Rabassa et al., 1982). Upper Cretaceous–Lower Tertiary sediments (Rabot and López de Bertodano Formations, Marambio Group) outcrop on its southeastern border (Marenssi et al., 1992), while Cretaceous deposits (Gustav Group) are exposed on the western border forming a belt about 50 km long (Medina et al., 1992).

The main geomorphological features of James Ross Island are a result of glacial and periglacial processes. The ice moves in a radial pattern from the Haddington and Dalinger ice domes (Fig. 7), covering almost the entire surface of the island. During glacial and interglacial periods, several marine terraces were formed as a consequence of glacieustatic and combined tectonic movements. A minimum of seven marine terraces is observed at altitudes of: 0.75–1, 2–3, 4–6, 10–14, 20–22, 30–40 and 100 m a.s.l. (Strelin and Malagnino, 1992), while only four terraces at 32–35, 21–24, 10–17 and 3–5 m are mentioned by Fukuda et al. (1992) in the region of Croft Bay (Fig. 7). According to the Quaternary geological history of this region, a major portion of the present ice-free regions of Seymour, James Ross and Shetland del Sur islands lost much of their ice covers approximately 6000–5000 years ago (Hjort et al., 1994).

Air temperatures were measured at Seymour and James Ross islands at different times and by different authors. A 10-year temperature record (1971–1980), obtained at Seymour Island (Estadísticas Meteorológicas, 1986), results in an MAAT of -9.4 °C in the upper terrace (200 m a.s.l.). Aristarain and others drilled 17 shallow cores during summer seasons 1976–1978–1979–1981 on the ice cup in the central part of James Ross Island (Dalinger and Haddington Domes, Fig. 7) (Aristarain et al., 1987). From these wells, 15 snow temperature values—corresponding to a 10-m depth—were measured, with an average value of -13 °C corresponding to an average altitude of 1465 m a.s.l. This temperature could be considered close to the MAAT corresponding to this altitude and date in James Ross Island. The MAAT, thus estimated, is consistent with that obtained in 1995–1996 by Sone and Strelin (1997) in Riscos Rink at 450 m a.s.l. (-7.5 °C, western coast of James Ross Island) if the altitudinal lapse-rate of 0.58 °C/100 m—determined in Aristarain et al. (1987)—is used to correct both temperatures to the same altitude. It is important to mention that, according to Skvarca (1993), a rapid and continuous recession of the northern end of the Larsen Ice Shelf was produced in the last decades, probably in response to an Antarctic Peninsula regional climate warming.

4. Permafrost in Seymour Island

The first experience with the magnetotelluric (MT) method in Seymour Island was carried out by the Instituto Antártico Argentino in its northern section (upper terrace) (Fig. 3), during the summer of 1978–1979 (Fournier et al., 1980). This sounding was repeated, in the following summer, about 2 km SW of the previous location, extending the natural electromagnetic field record to lower periods to study the basin at lower depths (Del Valle et al., 1982; Fournier et al., 1987, 1989; Fournier, 1994), and obtaining results consistent with the first study. Induction coils and a flux-gate magnetometer were used in these studies for measuring the geomagnetic variation field, and lead electrodes for measuring the electric field of the Earth. Both field were recorded in two horizontal and orthogonal directions using chart recorders (Fournier et al., 1980). The objective of these investigations was not to study the permafrost on the island but its stratigraphic sequence, the thickness of the Larsen Basin and the crust and upper mantle structures of the region. As the natural electromagnetic field was only characterized at periods higher than 0.1 s., the Larsen Basin was only described from a depth of about 600 m and greater. In both studies, two apparent electrical resistivity curves corresponding to NS and WE magnetic directions were determined from manual processing on chart recorders (Fig. 3 shows the locations of the MTS studies). From the 1D modeling of the resistivity curve corresponding to the magnetic NS direction, the thickness of the Larsen Basin was estimated to be about 6 km and an intermediate conductive layer (ICL) in the upper mantle (possible asthenosphere) with an average depth of about 74 km was also estimated (Fournier et al., 1980, 1987, 1989; Fournier, 1994).

A preliminary computer-processing of these MT data was published in 1988 (Pomposiello et al., 1988). This processing was made only considering results with a coherency higher than 0.85, and gave a pair of principal directions orientated N67°E and N23°W (Fig. 3). The pair of orthogonal horizontal directions in the MT site, along which, the maximum anisotropy in electrical resistivity is obtained, is called "Principal directions". An average skew of 0.30 was also obtained in this processing, indicating regional tectonics close to a 3D structure. Fig. 4 shows, from 0.1 s of period and higher, the principal apparent resistivity curves (ρT) obtained. The lateral anisotropy in resistivity shown by these curves indicates the presence of "distortions" in one or both curves, i.e. they are shifted and/or modified in shape. The distortions in the MT curves are produced by lateral inhomogeneities in the electrical resistivity distribution in the Earth, which make the telluric current circulation pattern to become deformed. The electric charges and current channelings thus occurring in the subsoil are a source of an "anomalous" electromagnetic field, which is the cause of distortions. For further details, please consult Berdichevsky and Dmitriev (1976) and Rokityansky (1982), among others. Therefore, previous to carrying out 1D modeling of MT resistivity curves (formal interpretation), it is necessary to consider its possible distortions in order to obtain a "normal curve" (ρN), i.e. an MT curve without distortions.

The examination of ρT curves in Fig. 4 suggests a static galvanic distortion (static shift) in one or both curves, i.e. a galvanic distortion without curve deformation. This kind of distortion is produced by a local inhomogeneity near the surface. The geological structure of the northern part of Seymour Island seems to be complex, with elements consistent with the orientations of the principal directions. In fact, as previously mentioned, the Quaternary and Tertiary sediments present in the MTS site were apparently deposited in a broad syncline (Fig. 3). The

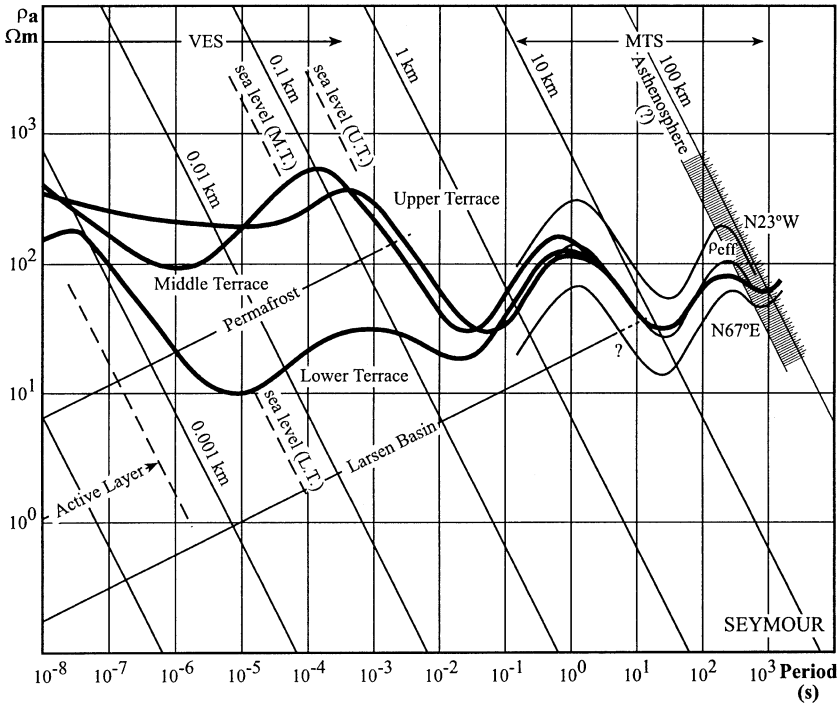

Fig. 4. Apparent electrical resistivity curves as function of the period of the Earth's electromagnetic field, possible to be obtained in Seymour Island's terraces by MTS. MT curves from 10^{-8} to 10^{-4} s of period were obtained from VES results (after Fukuda et al., 1992) by using a 1D MT modeling program. Curves (thin solid lines) from 0.15 s of period and higher are MT curves with distortion (ρT) corresponding to upper terrace, and N23°W, N67°E denote principal directions (after Pomposiello et al., 1988). ρeff is the average curve (considered without distortions, ρN) used for interpretation. Heavy solid curve from 1 s of period and higher is the model response of ρeff 1D modeling. No field data are available between 10^{-4} and 0.15 s of period. A conductive layer is detectable in the upper mantle, between 100 and 1000 s of period (possible asthenosphere). The high resistivity of soil because of freezing, and the high conductivity below the frozen ground are clearly visible. U.T., M.T., and L.T. mean upper, middle and lower terraces, respectively.

region also presents a block structure where four major fracture zones have been observed. These megafaults (Fig. 1) have orientations in agreement with one of the principal directions shown in Fig. 3. Basaltic dikes are also present at this island, parallel to the Antarctic Peninsula cordillera and the Bransfield Strait rift, which are consistent with the other principal direction (Fig. 3). In addition, interpretations of seismic studies in the Larsen Basin suggest a possible fault parallel to the Antarctic Peninsula, in front of Seymour Island (Strelin, 1994). This feature suggests that Seymour Island could be an elevated block of the basin (Del Valle et al., 1993, p. 140). All these tectonic elements form a complex tectonic structure where, with only two MTS in the region, it is impossible to determine which curve ρT is shifted in Fig. 4. Therefore, the average curve (ρeff on Fig. 4) will be used for interpretation as the best approximation to ρN. 1D modeling of this curve (Fig. 5) shows a conductive layer between 5 and 11 km depth suggesting a thickness of the Larsen Basin, in this location, around 11 km depth. However, taking into account the presence of a near rift structure (Bransfield Strait) with presumably high heat flow in the region—as will be analyzed in the present study—it is also possible to speculate that the bottom of the Larsen Basin could not be visible by MTS because of the low resistivities of the crystalline basement in its upper section. A similar situation appears in James Ross Island (close to Seymour Island, Fig. 1) where it was impossible to estimate the thickness of the Larsen Basin, apparently because of the thermal structure of the basement (see James Ross study below).

The ρeff curve (Fig. 4) also shows with some uncertainty, between 100 and 1000 s of period, an intermediate conductive layer (ICL) with top at about 70-km depth (upper mantle) (Fig. 5). However, considering that these MTS were carried out at an island, a probable "coast effect" on data must be analyzed. According to Rokityansky (1982), p. 308: "The coast effect is understood as an anomalous geomagnetic variation field due to concentration of the currents induced in the sea water, where conductivity is generally higher than in the rocks composing the coast and seafloor". Usually, this effect is notable only in the vertical component (Z) of the geomagnetic variation field, and it depends, in general, on the shape of the coast, the depth of the sea next to the coast and the distance from the study site to the coast. In the present case, bathymetric information (Keller et al., 1985) shows depths of the

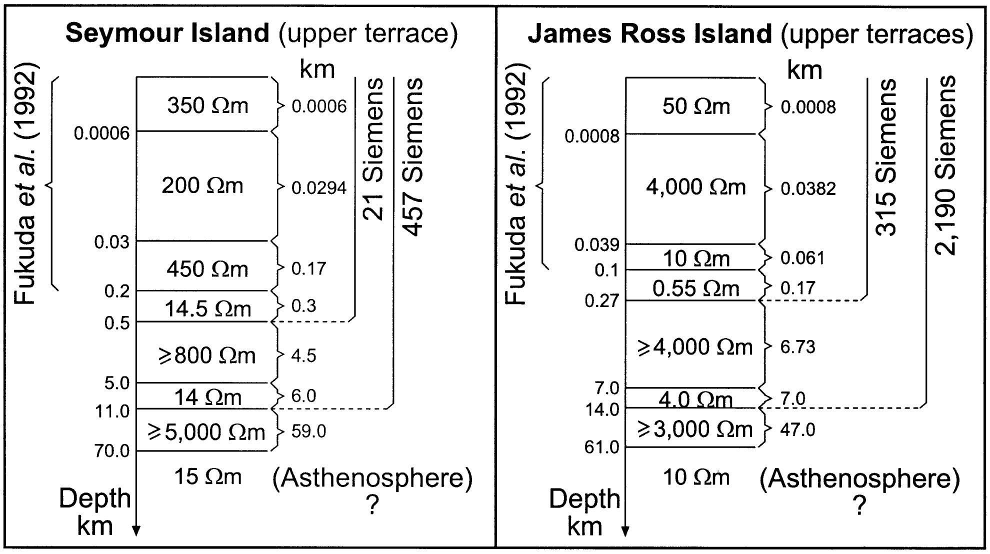

Fig. 5. One-dimensional MT modeling results corresponding to normal (ρN) curves estimated in upper terraces of Seymour and James Ross Islands. This figure shows the electrical resistivity variation with depth in both islands. VES interpretations after Fukuda et al. (1992) are included. Longitudinal conductances are given in S (Ω^{-1}) [longitudinal conductance (in S)=thickness (in m)/electrical resistivity (in Ω m)].

Weddell sea surrounding Seymour Island of less than 400 m up to a distance of about 100 km from the island, where the sea becomes rapidly deeper. According to this characteristic of the seafloor near Seymour Island, the coast effect will not be important in the horizontal electric field parallel to the coast. In addition, this ICL is also observable with similar characteristics (at about 60 km depth) in the MTS carried out in James Ross Island (Mamaní et al., 1998). Therefore, this conductive layer is considered as real in the region. Its low resistivity (in the present case about 15 Ω m) is attributed to partial melting of rocks (see e.g., Feldman, 1976, Rokityansky, 1982 p. 243). On the other hand, its depth (determined by MTS) is correlated with the regional heat flow: the higher the heat flow, the shallower the ICL (Adám, 1976a, 1978).

The first attempt to measure the permafrost electrical resistivity in Seymour Island was made in the 1978–1979 field campaign when, using the VES method and distances between current emission electrodes of 10 and 20 m (Schlumberger configuration), two apparent resistivity values were obtained in the upper terrace: 380 and 110 Ω m, respectively (Fournier et al., 1980). A rough interpretation of these data suggests a reduction in permafrost resistivity between a depth of 3–30 m approximately. This feature was then confirmed by subsequent VE soundings (Fukuda et al., 1992).

The first rough estimation of permafrost thickness in Seymour Island was reported by Corte (1982). As a first approximation, a geothermal gradient of 0.03 °C/m—considered as normal for continental areas—was used, as well as an MAAT of −10 °C. Assuming a homogeneous medium and permafrost in equilibrium (Lachenbruch, 1968), a thickness of 300 m was estimated. The active layer was calculated (1989–1990) to range between 0.35 and 0.47 m.

A permafrost geophysical study in Seymour Island was carried out in the summer of 1987–1988 (Buk, personal communication) using VES with Schlumberger electrode configuration. Frozen ground thicknesses of 180–198 m were estimated in the upper terrace (200 m a.s.l.), 120 m in the middle (50 m a.s.l.) and 50 m in the lower terraces (5 m a.s.l.), respectively, with maximum electrical resistivities recorded of about 1000–2300 Ω m. Another VES survey was conducted in the summer of 1989–1990 to study the permafrost of this island. In this case, a Wenner electrode configuration was used (Fukuda et al., 1992) obtaining the following estimates of the frozen ground thickness: 200 m in the upper terrace, 105 m in the middle and 35 m in the lower terrace. The maximum resistivities recorded were of 720–800 Ω m. In these studies, a section with lower resistivity could be identified in the permafrost, which was considered to be a part of the frozen ground (Fukuda et al., 1992). In a general context, the thickness of the active layer at Seymour Island may be estimated to range between 0.35 and 0.60 m.

Fig. 4 shows the frozen ground resistivity, according to Fukuda et al. (1992) data, but converted to MT apparent resistivity curves (between 10^{-8} and 10^{-4} s of period) using a 1D MT modeling program. The idea behind this conversion is to build—from the VES and MTS data—only one apparent resistivity curve to better appreciate the resistivity variation with depth in the whole thickness of the basin. Although VES and MTS methods may provide not exactly the same electrical resistivities if the subsoil is inhomogeneous in resistivity, Fig. 4 shows the MT curves likely to be obtained at the different terraces, taking into account the Fukuda et al. (1992) results and the MTS results corresponding to the deepest section of the Larsen Basin, lower crust and mantle. The same deep resistivity structure was assumed in Fig. 4 for the different terraces like that obtained by MTS in the upper terrace. An important reduction in resistivity is visible in Fig. 4 for the frozen ground at the lower terrace, surely produced by the influence of the saline marine water. Slight reduction in resistivity is also shown in the frozen ground corresponding to the upper and middle terraces. On the other hand, although there are no field data between 10^{-4} and 10^{-1} s of period (Fig. 4), the presence of a very conductive layer (saline) beneath the frozen ground is clear. The evidence arises because of the low apparent resistivities obtained in the MT soundings in spite of the high resistivities of frozen ground near the surface (in Fournier et al., 1987, this conductive layer had already been suggested and analyzed). At James Ross Island, a similar situation occurs (Fig. 9) (Mamaní et al., 1998). It is also important to note that if there were, in the range of period between 10^{-4} and 10^{-1} s, thin (resistive) frozen ground sections under the frozen ground measured at Seymour Island, intercalated into the mentioned very conductive layer, according to MT theory these hypothetical resistive horizons could be undetectable by MT because of their thickness in relation to their depth. As a rough estimation: "a resistive layer can be identified when its thickness is comparable to, or larger than, the total thickness of all the overlaying layers" (Rokityansky, 1982, p. 187). As an example, if there were a resistive (frozen) layer at 250-m depth, with a thickness of about 20 m, this layer would not be detectable by MTS.

Quantitative connections were deduced from a high number of geoelectrical and geothermal data published in the monograph: "Geoelectric and Geothermal Studies, East Central Europe and Soviet Asia" (Adám, 1976b), which correlate regional surface heat flows with depths of conductive layers in the crust and upper mantle determined by MTS (Adám, 1976a, 1978).

For the first conductive layer in the upper mantle (intermediate conductive layer, ICL), the following empirical relationship was obtained:

$$h = 155q^{-1.46}$$

(Adám, 1976a, 1978) where: h: depth of conductive layer (in km); q: heat flow (in HFU, 1 HFU=1 μcal/cm^2 s).

This conductive layer (asthenosphere) is considered to be produced by partial melting of basic rocks (Feldman, 1976). Although it is not clear, according to Adám (1976a), if Adám's correlations characterizing a limited region in the world can be applied to other regions, the use of this equation would allow us to obtain, at least, a rough estimation of the regional heat flow at Seymour Island. Using the previous relationship, a heat flow of 1.724 HFU $\cong$ 72 mW/m^2 (1 HFU $\cong$ 42 mW/m^2) is estimated, considering the ICL with an estimated top at 70-km depth (according to MTS results). This value should only be taken as an approximation to the heat flow in steady state, likely to be measured in the region. A rift is present at the Bransfield Strait, located 180 km northwest from Seymour Island (Fig. 1), where heat flows higher than 220 mW/m^2 were measured (Ghidella et al., 1997). High heat flows are common in other rift areas in the world, e.g., in the Baikal rift zone, heat flows of about 80–130 mW/m^2 or higher were reported (Lysak, 1976).

Using the Fick law (correlation between conductive heat flow and temperature gradient), the temperature variation with depth in the permafrost of Seymour Island can be estimated. As an approximation, we use an average thermal conductivity taken from laboratory studies (Sawada and Ohno, 1985) corresponding to a frozen sand sample with a density of 0.99 g/cm^3 and 94% of water saturation. This sample gave in the laboratory an average thermal conductivity of 1970 mW/m °C and a thermal diffusivity of 0.99×10^{-6} m^2/s (Sawada and Ohno, 1985, pp. 54–55). Using the thermal conductivity mentioned and a heat flow of 72 mW/m^2, the thermal gradient estimated is 0.037 °C/m.

A 10-year temperature record in Seymour Island (1971–1980) (Estadísticas Meteorológicas, 1986) gives an MAAT of −9.4 °C in the upper terrace. Taking into account this temperature and the estimated temperature gradient, a permafrost thickness of about 250 m is obtained for the upper terrace, based on the Lachenbruch's (1968) model. This estimate should be considered as an approximation to the permafrost thickness in equilibrium with the estimated steady heat flow and the MAAT. The presence of a cryopeg as part of the permafrost below the frozen ground (basal cryopeg) is possible according to MTS, which indicate the presence of a saline conductive layer below the frozen ground. The cryopeg would not be detected with high resistivity because it is an unfrozen ground according to its salinity. The cryopeg is an unfrozen layer in a zone of permafrost or cryotic zone with a temperature under 0 °C but with a depression at the freezing point (see Multilanguage Glossary of Permafrost, 1998).

Therefore, the permafrost thickness in equilibrium (250 m) is consistent with the frozen ground thickness estimated by VES (200 m in the upper terrace). The difference could be attributed to the presence of about 50 m of cryopeg below the frozen ground (Fig. 6). This result suggests stable geological and climatic conditions for a period of time long enough to reach a subsoil temperature profile near the equilibrium.

5. Permafrost in James Ross Island

Three MTS were carried out in the northwestern region of James Ross Island (Brandy Bay and Hidden

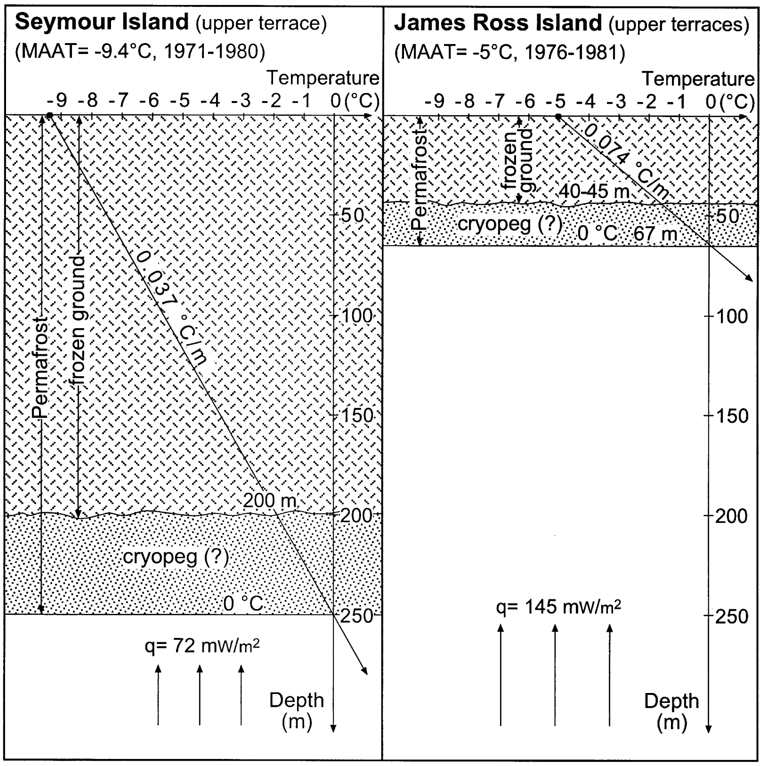

Fig. 6. Estimates of permafrost thickness in Seymour and James Ross Islands, in equilibrium with the geothermal heat flow (estimated from the depths of conductive layers in the crust and upper mantle as inferred by MTS), and the mean annual air temperature (MAAT) (Lachenbruch, 1968). Frozen ground thickness shown in this figure are from Fukuda et al. (1992).

Lake, Fig. 7), free of ice cover, during the Antarctic summer expedition of 1991–1992 (Mamaní et al., 1998). The studied region is located about 100 km from the Bransfield Strait rift. Fig. 8 shows the apparent resistivity curves obtained. From the ρN curves estimated in BYB1 and HDL (Fig. 8), an average ρN curve is estimated (curve from 1 s of period and higher in Fig. 9) that will be used in the present analysis. In observing this curve, and its 1D modeling (Fig. 5), the presence of two conductive zones is clear: one located between 7 and 14 km depth in the crust, with 4 Ω m of resistivity (and about 1800 S of longitudinal conductance), and another with top at about 60-km depth in the upper mantle, with about 10 Ω m of resistivity (Fig. 5).

Some arguments make us think that the first conductive zone suggests a possible magma chamber in the crust:

(a) A magma chamber can be detected by MTS depending principally on the resistivity of the host rocks. It is detectable on the "longitudinal curves" (curve corresponding to the principal direction parallel to regional strike) as inductive effects produced by current channelings, mainly when the magma chamber is bidimensional (Newman et al., 1985). In the present case, the conductive zone is principally evidenced as inductive effects in the longitudinal curves (ρ||) (Fig. 8).
(b) The high longitudinal conductance of this "layer" (about 1800 S, Fig. 5) and the volcanic nature of James Ross Island (Plio-Pleistocene age) with recent activity support the idea of a magma chamber.
(c) The proximity of the Larsen Basin border (Antarctic Peninsula) suggests that the cause of this conductive zone is not related with the basin but with deeper basement sections.

As at Seymour Island, a conductive layer at about 60-km depth is also observable in James Ross Island

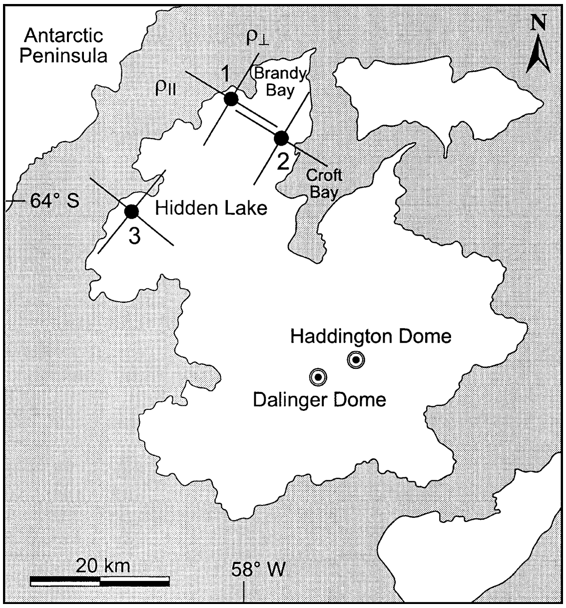

Fig. 7. James Ross Island map showing the sites (1, 2 and 3) where MTS were carried out (Mamaní et al., 1998). Solid lines denote principal directions of soundings. ρ|| and ρ⊥ denote longitudinal and transversal MT curves, respectively, i.e. MT curves along and across the regional geological strike (after Mamaní et al., 1998). 1, 2 and 3 correspond to BYB1, BYB2 and HDL sites in Fig. 8.

(Figs. 5 and 9), indicating a lithospheric thickness attenuation in the region. For both islands, this conductive layer in the upper mantle is suggested by MTS with some uncertainty in its depth because of its deep location.

Four VES were carried out, in the austral summer of 1989–1990, in the NW region of James Ross Island by Fukuda et al. (1992). These works were carried out on four terraces: upper terraces (32–35 and 21–24 m a.s.l.), middle terrace (10–17 m a.s.l.) and lower terrace (3–5 m a.s.l.). A frozen ground thickness lower to that of Seymour Island was estimated: 40–45 m in upper terraces, and 3.40 and 5.80 m for the middle and lower terraces, respectively (Fukuda et al., 1992).

The active layer in James Ross Island has a thickness of approximately 0.70–1.45 m. The maximum value, mentioned, was determined at 2 m a.s.l., and corresponds to an average of five measurements.

Fig. 9 shows—like Fig. 4 for Seymour Island—VES data converted to MTS curves, using a 1D MT modeling program. Comparing Figs. 4 and 9, a higher conductance (about 300 S) below the frozen ground in James Ross Island than in Seymour Island (only about 20 S) is obvious. This difference could be attributed to higher temperatures in the crust in James Ross Island (possibly because of the magma chamber), and differences in soil salinity. Frozen ground thickness is also very different in both islands: in fact, if the magma chamber really exists, it should increase the heat flow at the surface. Like in Seymour Island, it is possible to use an Adám's relationship to estimate the regional heat flow. In this

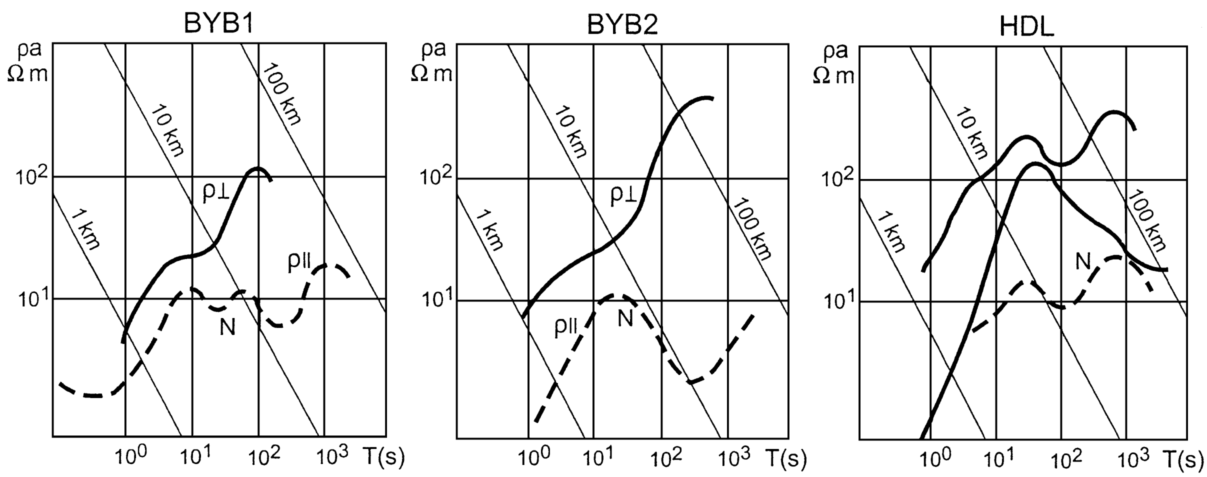

Fig. 8. Magnetotelluric curves obtained in James Ross Island (Mamaní et al., 1998). ρ|| and ρ⊥ indicate longitudinal and transversal curves, respectively. Dashed lines denote normal curves (N) estimated. (Reproduced from Mamaní et al., 1998 with modifications.)

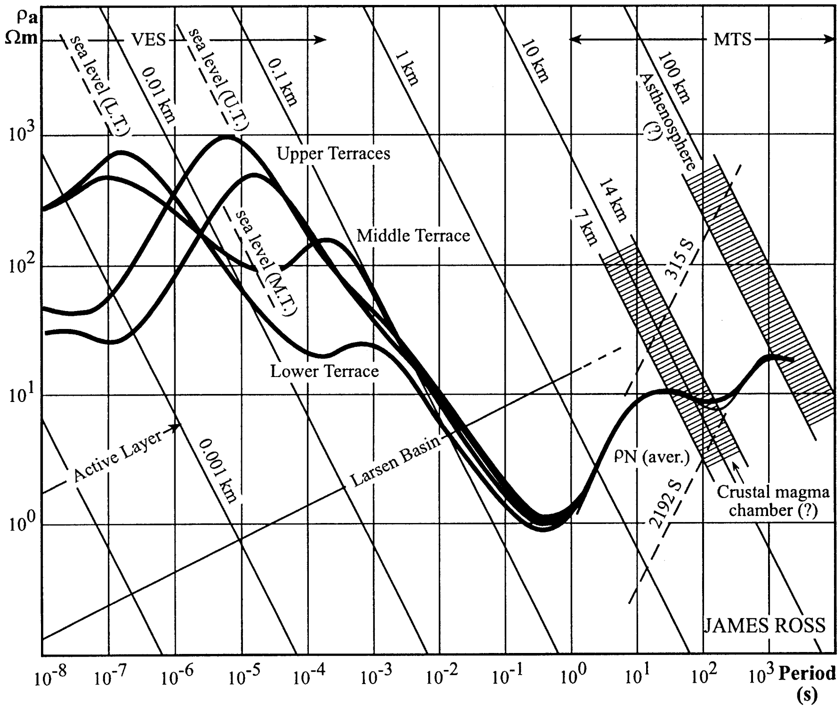

Fig. 9. The same explanation as in Fig. 4 but applied to James Ross Island. VES results converted to MTS curves, between 10^{-8} and 10^{-4} s of period, are from Fukuda et al. (1992). The MT curve, between 1 and 1000 s of period, is an approximate average curve (ρN aver.) obtained from the normal curves estimated in BYB1 and HDL sites (Fig. 8). This curve was used for 1D modeling. The freezing near surface and a very important conductive (saline) layer below the frozen ground are clearly observable. A conductive layer (possible magma chamber) is present in the crust.

case, as the conductive body is seated in the crust, we use:

$$h = 35q^{-1.30}$$

(Adám, 1976a, 1978) where h and q have the same meanings as in the Seymour Island study.

Considering the top of this conductive body (suggested by MTS) at about 7-km depth, we estimate a heat flow q=3.449 HFU ≅ 145 mW/m^2 as the heat flow likely to be measured in steady state. Considering, for the frozen ground, the same thermal conductivity used in Seymour Island (1970 mW/m °C), a thermal gradient of 0.074 °C/m is estimated for the NW region of James Ross Island.

During four summer campaigns, between 1976 and 1981, 17 holes were drilled on the ice cup, (in the neighborhood of Dalinger and Haddington Domes, Fig. 7) between depths of 8 and 31 m (Aristarain et al., 1987). In these studies, 15 temperature measurements, between 10- and 14-m depth, were obtained from the wells using conventional thermometers and thermistors. These values were then corrected to 10-m depth (Aristarain et al., 1987). From these 15 corrected values—and altitudes a.s.l.—it is possible to estimate an average temperature of −13 °C corresponding to an average altitude of 1465 m a.s.l. This temperature can be considered as the MAAT corresponding to 1465 m a.s.l. and to the 1976–1981 period in James Ross Island. Using an air temperature gradient of 0.58 °C/100 m—estimated in Aristarain et al. (1987)—it is possible to estimate an MAAT of −5 °C corresponding to an altitude of about 34 m a.s.l. (upper terrace in Fukuda et al., 1992).

Two temperature profiles covering the first 20 m in depth were also obtained in two locations on the ice cup (Aristarain et al., 1987). From these profiles, a geothermal gradient of 0.067 °C/m in the central ice cup of James Ross Island can be estimated. This value is consistent with the estimation carried out in the present work (0.074 °C/m) corresponding to the northwestern part of this island. This agreement supports the idea of a high heat flow in James Ross Island and the presence of the magma chamber suggested by MTS.

Considering an MAAT=−5 °C at 35 m a.s.l. (upper terraces in Fukuda et al., 1992), and a geothermal gradient of 0.074 °C/m, a permafrost thickness of about 67 m is estimated in the northwest region of this island, considering a homogeneous medium, following the Lachenbruch's (1968) model, (Fig. 6). Like in Seymour Island, the permafrost thickness estimated in James Ross Island should be taken as an approximation to its thickness in equilibrium with the steady heat flow and the MAAT considered. The estimated value (67 m) is consistent with the frozen ground thickness estimated by VES (40–45 m in the upper terrace) (Fukuda et al., 1992), if we consider that about 25 m of cryopeg are present below the frozen ground. The presence of a cryopeg is very probable taking into account the high soil salinity inferred from the high conductance below the frozen ground (Fig. 9). Results suggest—like in Seymour Island—that the permafrost is approximately near equilibrium with the present geological and climatic conditions (until 1980 approximately).

The minimum lapse of time during which the ground surface must have been free of ice in order to approach the steady heat flow in both islands could be estimated by the following equation (e.g., Lliboutry, 1982, p. 164):

$$t = h^2/\chi$$

where: t: time (s); h: permafrost thickness (m); χ: thermal diffusivity (m^2/s).

Using the mean thermal diffusivity of 0.99×10^{-6} m^2/s, corresponding to the sample of sand examined by Sawada and Ohno (1985), p. 54, estimates of about 2000 years for Seymour Island and 140 years for James Ross Island are obtained as the minimum lapse of time after deglaciation on each island.

6. Discussion and conclusions

Because of the saline soil and the consequent possible presence of a cryopeg below the frozen ground, permafrost thickness is difficult to estimate, i.e. subsoil at temperature below 0 °C, using electrical resistivity measurements. These surveys can estimate the base of the frozen ground with some uncertainty, but the position of the 0 °C mean annual temperature isotherm is difficult to infer because the cryopeg would not represent a notable change in resistivity in relation to the subsoil at a positive temperature. In the present work, a comparison and analysis have been made between actual frozen ground thickness—obtained by VES—and estimates of permafrost thickness likely to be present if the system were in equilibrium. These estimations were made using approximate values of steady heat flow inferred from the depths of conductive layers in the crust and upper mantle determined by MTS. Although the equilibrium permafrost thickness thus estimated may include some uncertainty, its comparison with the actual frozen grounds—corresponding to both islands—seems to suggest a state approaching equilibrium. This result means that the soil should have been uncovered of ice and with approximately stable climatic and geological conditions for a period of time long enough to approach equilibrium.

Summarizing, the following conclusions can be drawn:

(1) The permafrost on the islands Seymour and James Ross seems to be constituted by a frozen ground with a cryopeg underneath (basal cryopeg). In its relative thickness, seawater seems to have an influence: all sediments and rocks above sea level are frozen, and the cryopeg would be present below sea level. The influence of seawater would be determined basically by its salinity, and also by its having a higher temperature than air.

(2) Although MTS data might be not completely accurate at high periods, consistent MTS results obtained in both islands show a conductive layer in the upper mantle (possibly the asthenosphere) suggesting a lithosphere thickness of about 60–70 km in this region. This thickness attenuation is consistent with the presence of the rift at Bransfield Strait.

(3) MTS carried out at James Ross Island suggest the presence of a magma chamber in the northwestern region of this island, located in the crust between 7- and 14-km depth, with a longitudinal conductance of 1800 S. The associated high heat flow

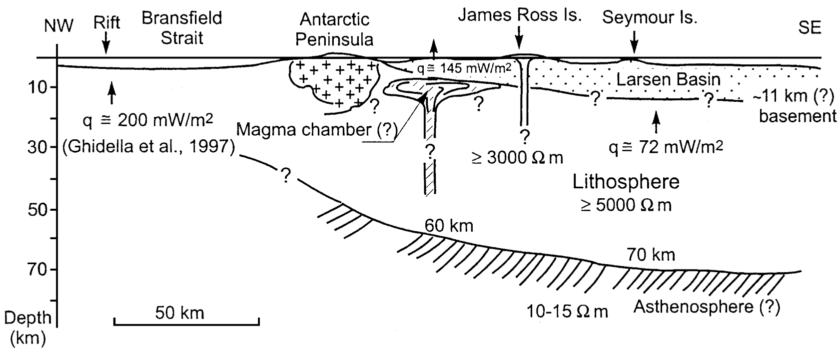

Fig. 10. Schematic cross section in the studied area showing a hypothetical geological interpretation of this region, according to the results of the present study.

seems to have a strong influence in determining the permafrost thickness and perhaps the deglaciation of this part of James Ross Island.

(4) It is difficult to give an estimation of the thickness of the Larsen Basin by MTS. In the northwestern part of James Ross Island, the possible high electrical conductivity present in the upper basement, presumably produced by the magma chamber, could be screening the bottom of this basin. In Seymour Island, the MTS seem to suggest a thickness of about 11 km for this basin with a longitudinal conductance of about 450 S. However, distortion effects on the MT curves in Seymour Island make this thickness estimate not completely reliable.

(5) The heat flows in the studied region are about 72 mW/m^2 in Seymour Island and 145 mW/m^2 in the northwestern region of James Ross Island. The geothermal gradients are estimated in 0.037 °C/m for Seymour Island and 0.074 °C/m for James Ross Island. This last value is consistent with that estimated from the data in Aristarain et al. (1987) (0.067 °C/m, in the central ice cup).

(6) Permafrost at Seymour Island (at 200 m a.s.l.) and James Ross Island (at 34 m a.s.l.) is suggested to be close to equilibrium with the regional steady heat flow and the present climatic (between about 1971 and 1980) and geological conditions. Thickness of 250 m for Seymour Island and 67 m for James Ross Island are obtained as estimates of permafrost thickness in equilibrium. Estimates of about 2000 years for Seymour Island and 140 years for James Ross Island are obtained as the minimum lapse of time after deglaciation. However, temperature records and Larsen Ice Shelf studies, carried out over the last decades, strongly suggest a climatic warming in the Antarctic Peninsula region. Therefore, at present, permafrost at both islands would be in an unbalanced condition.

The profile in Fig. 10 shows a plausible geological interpretation for the study area, according to the obtained geophysical results.

A first preliminary sketch of this work was presented at the 31st International Geological Congress held in 2000 (Brazil), where as a conclusion, we advanced the idea that the permafrost in both islands seemed to be away from the equilibrium with the present heat flow and the MAAT (the estimated permafrost thickness was smaller than the measured frozen ground thickness). That result was a consequence of using −6.6 °C as the MAAT (obtained by Fukuda et al., 1992 in lower terrace at Seymour Island) instead of −9.4 °C corresponding to upper terrace, and a preliminary depth for the ICL of only 50 km.

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