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Magnetic Investigations on the Ekström Ice Shelf, Antarctica

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Abstract: Analysis of geomagnetic measurements carried out on the Ekström Ice Shelf in Antarctica during the overwintering campaign 1987 reveals a high reliability of the differential total intensity field determination as the error is only ± 5 nT. This was controlled by repeat station measurements and seems remarkable in such relatively high geomagnetic latitudes close to the Polar Electrojet. The measured values were separated from external variations and the normal field with data from the Georg-von-Neumayer observatory sited north of the Ekström Ice Shelf and IGRF 85 model data, respectively. The reduced geomagnetic differential total intensity field values are called „ΔT_{red}-anomalies". The results are: a prominent ΔT_{red}-anomaly in the center part and a positive coast parallel anomaly in the north of the Ekström Ice Shelf. Upward field continuation reveals a good agreement with a map derived from Soviet aero-magnetic measurements . Power spectra analysis and forward modelling calculations suggest a Curie-depth of about 12 to 18 km. True locations of the geomagnetic anomalies are calculated with the method of reduction to the magnetic pole.

Zusammenfassung: Während der Überwinterungskampagne 1987 wurden auf dem Ekström-Eisschelf geomagnetische Feldmessungen der Totalintensität durchgeführt. Wiederholungsmessungen haben gezeigt, daß die Bestimmung des Feldes auch in den relativ hohen geomagnetischen Breiten sehr genau möglich war. Die Meßwerte wurden von externen Feldanteilen gereinigt und mit IGRF 85, dem internationalen geomagnetischen Referenzfeld, von regionalen Anteilen befreit. Die so reduzierten Totalfeldwerte zeigen im Zentralteil eine deutliche negative und im nördlichen Bereich eine positive, dem Küstenverlauf folgende Anomalie. Eine Feldfortsetzung nach oben auf das Flugniveau einer später durchgeführten sowjetischen aeromagnetischen Befliegung zeigt eine gute Übereinstimmung. Aus Spektralabschätzungen und aus Modellierung kann eine Curie Tiefe von 12 bis 18 km abgeschätzt werden. Die Lage der geomagnetischen Anomalien wurde mit Verfahren zur Reduktion auf den Pol direkt über den verursachenden Störkörpern abgebildet.

INTRODUCTION

Access to any region of Antarctica for scientific exploration is not easy as the areas are generally covered with huge ice sheets and the continent is nearly inaccessible. During winter time the weather is not only cold and often stormy, but very cold and almost always stormy. Time for field work is limited and only 2% of Antarctica is accessible for geological surveys (no ice sheets). This is the reason why geophysical methods are one of the most important tools to gather information about the structure of the crust and upper mantle.

For logistic reasons, most scientific activity is concentrated on the rim of Antarctica in ice shelf areas which cover 11% of the whole continent. In the multitude of geophysical investigations magnetic measurements are most successful on ice shelves, gravity measurements are nearly impossible in areas close to the ice edge (swell induced movements into the ice shelf), seismic methods are very expensive and difficult.

In the following results of a magnetic survey are presented measured with proton precision magnetometers in the area around the Georg-von-Neumayer-Station (GvN) during an overwintering campaign. The geographical coordinates of GvN-Station (latitude φ, longitude λ) are:

$$\varphi = -70° \, 36' \, S, \ \lambda = -8° \, 22' \, W.$$

with geomagnetic coordinates:

$$\varphi_{mag} = -61° \, 14' \, S, \ \lambda_{mag} = 41° \, 28' \ W .$$

We could expect that the disturbances created by variations of the external geomagnetic field would not be too problematic for magnetic measurements and for the use of a base station for field separation. Figure 1 shows the position of the geomagnetic pole (DP dipole axis) and the aurora oval which is strongly associated with the Polar Electrojet (PEJ) of the southern hemisphere for normal magnetic activity at 6:00 a.m., 12:00 a.m., 6:00 p.m. and 12:00 p.m. - it touches the GvN-region only at night.

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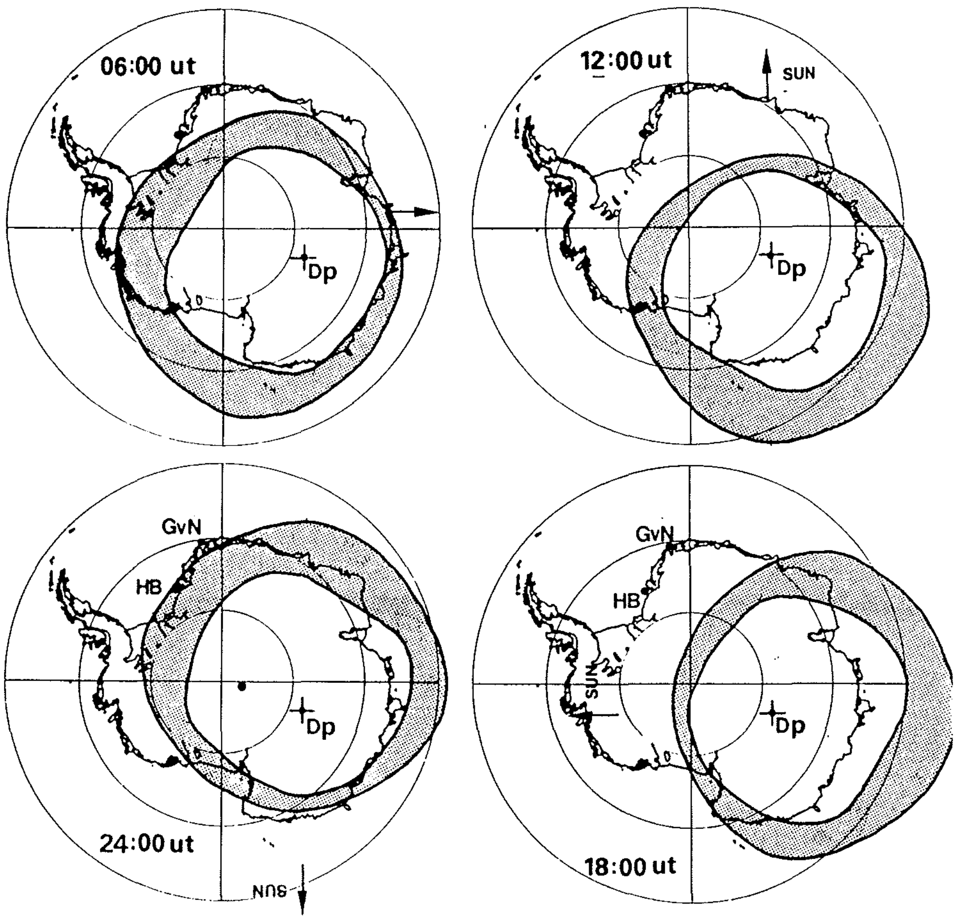

Fig. 1: Position of the Aurora Oval in the southern hemisphere as it changes position over a 24 hour period.

Abb. 1: 24-stündige Periode der Position des Polarlichtovals auf der Südhalbkugel.

After the separation of external field parts the inner geomagnetic normal field (sources in earth's liquid outer core) was separated from the regional/local field (sources in the crust). The latter was used for further interpretation based on potential theory: depth estimation of crustal sources, upward field continuation and reduction to the pole.

FIELD MEASUREMENTS

Several magnetic profiles (total length of 731 km, with 0.5 or 1 km spacing between individual measurement points) were measured during the overwintering campaign of 1987. Figure 2 shows the profiles in the area around the GvN-Station, the Atka Bay, the Ekströ̈m Ice Shelf (including a profile to the southern Kottas Mountains) and a profile to the Quar Ice Shelf. The logistics in such an inaccessible region are complicated and many restrictions have to be taken into consideration. First, the ice shelf is heavily crevassed, especially close to the grounding line and the ice shelf edges. Second, the weather conditions in the Antarctic usually prevent expeditions into the hinterland during overwintering.

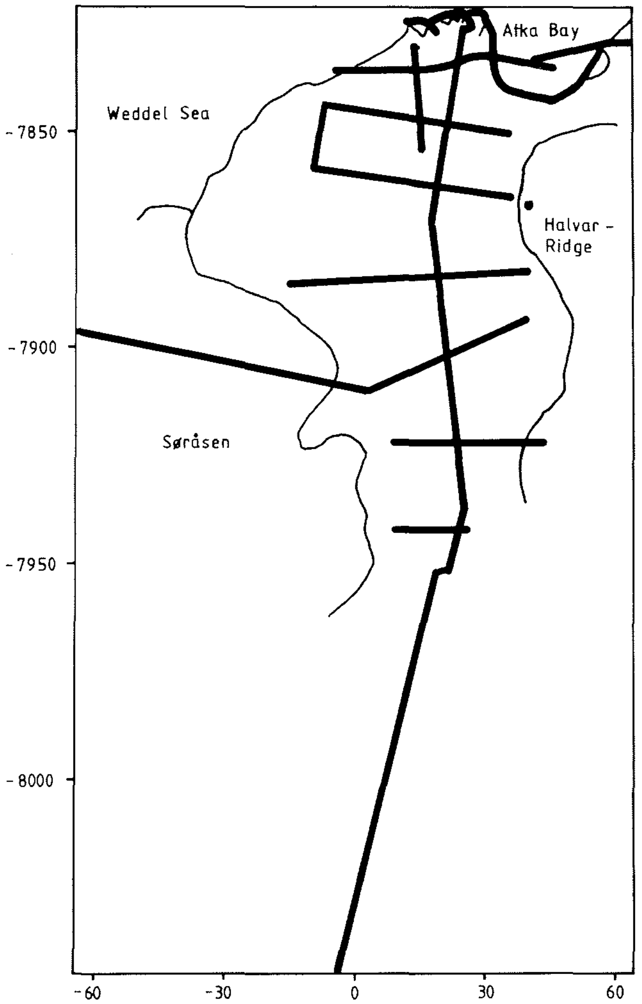

Fig. 2: Profile chart of the magnetic survey on the Ekström Ice Shelf, Antarctica, in 1987 in UTM-coordinates in km: (x,y) - (west/east from -9° W; south from equator).

Abb. 2: Profilplan der magnetischen Vermessung auf dem Ekström-Schelfeis, Antarktika, 1987 in UTM-Koordinaten /km: (x,y) - (westl./östlich von Bezugslänge -9° W; südlich des Äquators).

Therefore, most of the profiles follow traverse routes to seismological and geodetic stations in the vicinity of GvN-Station and the configuration of the profiles in Figure 2 is not an ideal regular network, but a better one would have included higher risk.

Several locations (geodetic and seismological stations, starting and end points of profiles and other prominent positions) were determined with satellite navigation systems (Transit, GPS) at different times between November 1986 and February 1988. These „fixed points" read an absolute accuracy of about ±25 m, depending on the applied navigation systems. They were usually marked by aluminum poles or geodetic triangles on the ice. As these points were not measured at the time the magnetic profiles were measured and the Ekström Ice Shelf is flowing differently in different areas, the coordinates of „fixed points" have to be interpolated for the time of actual magnetic measurement. The ice flow rates for different „fixed points" have been taken from the ice flow map (see Fig. 3). So, corrected starting and ending or crossing points of magnetic profiles were calculated for the actual measuring time. The actual equidistance of observation points, which were determined with the help of the speedometers of snow mobiles, were then interpolated in between these modified corrected points. The expected accuracy should be better than ±50 m (REIPRICH 1990). The accuracy of profile coordinates determined on sea ice is worse having to depend on compass azimuths and distances taken from speedometers.

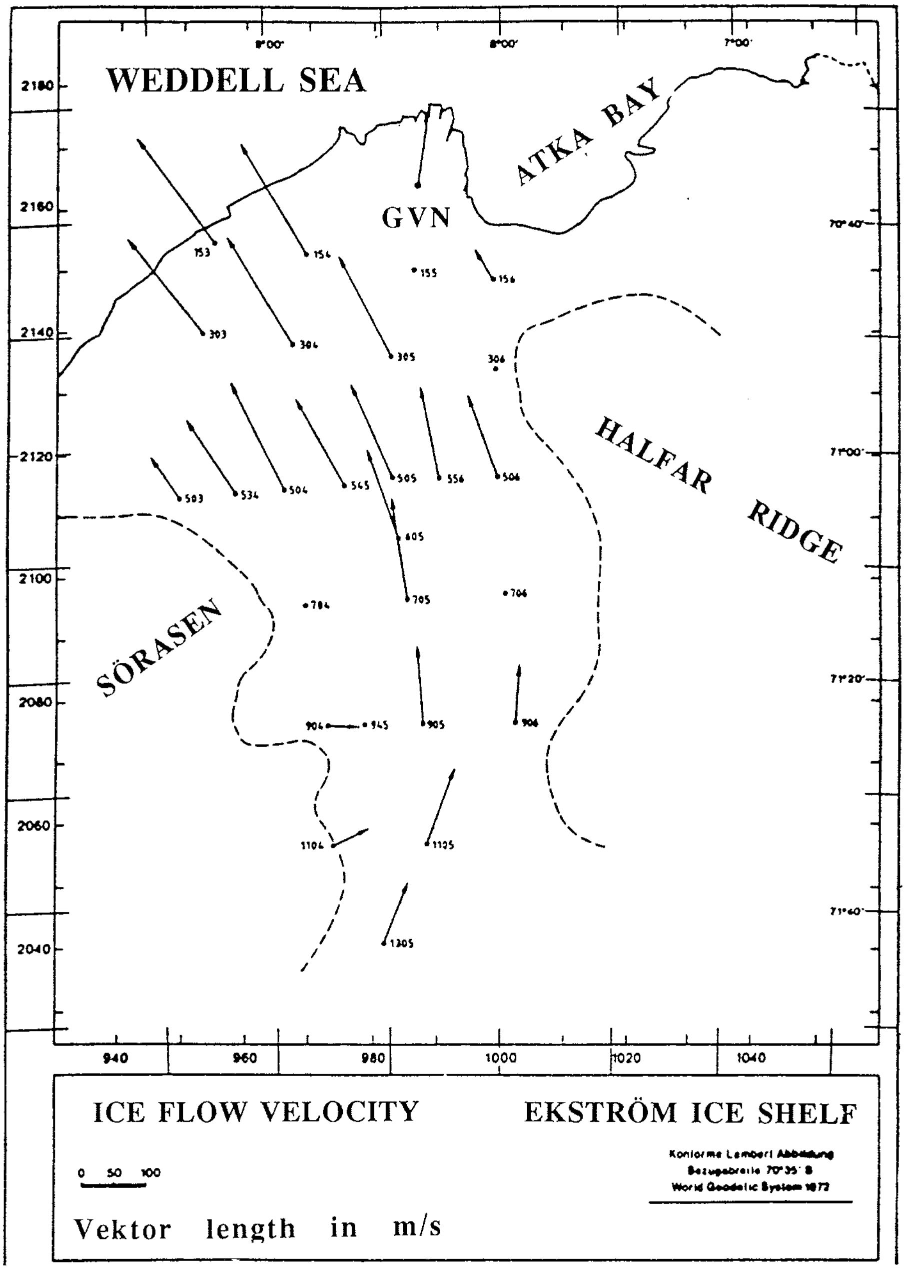

Fig 3: Ice flow velocities of discrete geodetic points on the Ekström Ice Shelf based on satellite navigation measurements (GPS and TRANSIT) in 1986/87 and 1987/88 (ERHARDT & STENGELE 1988).

Abb. 3: Fließgeschwindigkeit des Ekström-Schelfeises an diskreten geodätischen Meßpunkten, berechnet aus Messungen in 1986/87 und 1987/88 zu künstlichen Erdsatelliten (GPS und TRANSIT, aus ERHARDT & STENGELE 1988).

CALCULATION OF ΔT_{red}-ANOMALIES

Reductions

The aim of reducing magnetic survey data is to obtain the field anomlies arising from the crustal/sediment sources which are associated with variations of geological and geophysical properties of the crust. This is done in two steps:

1) separation of earth's external transient magnetic field from the inner field;

2) separation of the main part of the inner field, i.e. the slowly changing „normal field", from the constant „regional/local field" of crustal origin.

Separation of the External Field

The first step is simply done by calculating the difference ΔT between the magnetic total intensity values measured exactly at the same time at a survey point in the field and at a base station:

$$\Delta T = T_{field} + \partial T_{ext.\,field}(t) - [T_{obs} + \partial T_{ext.\,obs}(t)] \qquad (1)$$

$T_{field/obs}$ = static total intensity at survey field-point / at geomagnetic observatory;
$\partial T_{ext.\,field/obs}$ = transient external total intensity at survey field-point / at geomagnetic observatory.

If the transient external magnetic field variations are quasi-homogeneous in the investigation area, i.e. $\partial T_{ext.field}(t = t_i) = \partial T_{ext.obs}(t = t_i)$, the differential total intensity ΔT of a survey field point with respect to the GvN-observatory represents a quasi-static (due to secular variation) anomaly of the earth's inner magnetic field only: $\Delta T = T_{field} - T_{obs} \pm \partial T_{error}$.

A typical magnetogram of the GvN-observatory shows moderate Sq-variations with an amplitude of 30..40 nT and a 24-hours period which are sometimes disturbed by magnetic substorms, i.e. non-periodic events originating from solar-terrestrial interaction, e.g. sun spot activity. The amplitudes of the latter sometimes exceed ±500 nT (BRODSCHOLL 1988). But as the measurements were done shortly after a minimum of the 11-year periodicity of solar activity in 1987 the amplitudes rarely exceed ±250 nT (REIPRICH 1990). During the actual measuring periods magnetic disturbances were rare, but Sq-variations during the day when the aurora oval was far away from the Ekström region were regular (see Fig. 1 and also BRODSCHOLL & MILLER 1988). As repeat measurements on Ekström Ice Shelf revealed, the error made when reducing the external field parts in the data by simply calculating the differential total intensity ΔT (Eq. 1) was about ±5 nT or smaller, i.e. quasi-homogeneity of external Sq-variations can be assumed.

Separation ot the Normal Field

As second step the IGRF-normal-field value $T_{IGRF}(\varphi,\lambda)$ (International Geomagnetic Reference Field) was calculated for every survey field point and for the observatory and their difference $\Delta IGRF = T_{IGRF.field}(t = t_i) - T_{IGRF.obs}(t = t_i)$ was subtracted from the ΔT-value of the survey field point:

$$\Delta T_{red} = \Delta T - \Delta IGRF \qquad (2)$$

The ΔT_{red}-value thus represents a real static regional/local magnetic field value.

The calculations of the IGRF-normal-field values were carried out by expansion in spherical harmonic functions based on Gauss-coefficients and their first derivatives describing the secular variation which was provided by IAGA 1985 (International Agency of Geomagnetism and Aeronomy) up to order and degree 10.

In consideration of not only the normal field's secular variation but also the ice flow, especially the northward drift of the GvN-observatory at an average rate of 160 m/year, calculation of normal field data $T_{IGRF.obs}(t)$ was based on the observatory's actual position. Figure 4 shows the change of measured monthly mean values of total intensity and of the calculated normal field values at the GvN-observatory. It reveals an interesting feature:

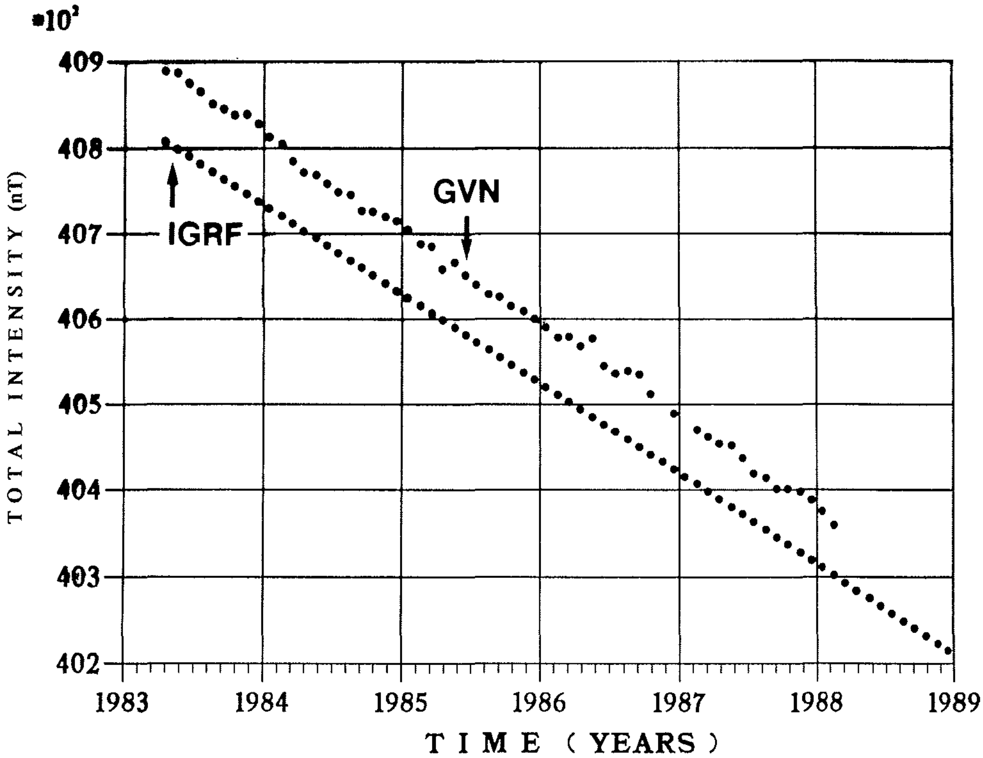

Fig. 4: Secular variation of total magnetic intensity at GvN-observatory: The measured monthly mean values are compared with the calculated IGRF 85 model data.

Abb. 4: Säkularvariation der Totalintensität des Erdmagnetteldes. Im GvN- Observatorium gemessene Monatsmittelwerte im Vergleich zu den theoretischen Werten des IGRF 85-Modells.

the difference T_{obs} - $T_{IGRF,obs}$ decreased from 85 nT (1983) to 60 nT (1988). Although the GvN-station moved 800 m northward in five years the small horizontal normal field gradient to the north of about -8 nT/km is insufficient to explain this, but as GvN-station is located on a magnetic local field anomaly with a local horizontal gradient of about -30 nT/km (BRODSCHOLL 1988), a decrease of 25 nT in T_{obs} - $T_{IGRF,obs}$ is not surprising. It demonstrates the stability of IGRF-model data and the accuracy of magnetic total intensity measurements.

By the way, similar systematic trends were found in repeat measurements at survey field points, i.e. small changes in ΔT-data of 1..3 nT were in accordance with ice flow rates and magnetic horizontal gradients. This leads to the idea of carrying out ice flow determination by „magnetic geodesy".

ΔT_{red} - ANOMALIES ON EKSTRÖM ICE SHELF

Figs. 5a and 5b demonstrate the effect of normal field separation which was dramatic in north-south direction but much smaller in west-east direction.

To study the geomagnetic regional/local field not only along profiles, the ΔT_{red}-data were gridded, i.e., interpolated on a regular grid consisting of equidistant nodal points (2 x 2 km or 1 x 1 km). Two different interpolation routines were used („quadratic" and spline interpolation, search radius 10..30 km). Although they produced charts with only small deviations, one should keep in mind that errors made by interpolation are relatively large between the profiles and very small along them. Figure 6a gives an anomaly map of the whole area.

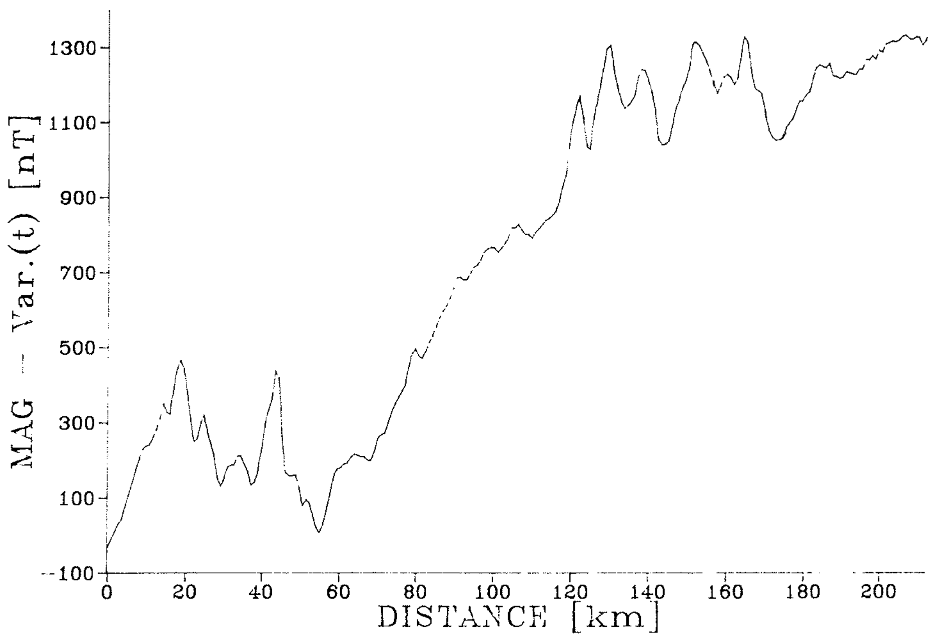

Fig. 5a: Profile of the geomagnetic differential total intensity ΔT (before separation of the normal field) from Georg von Neymayer station to the southern Kottas Mountains.

Abb 5a: Profil der variationskorrigierten magnetischen Totalintensität ΔT (vor der Normalfeldreduktion) von der Georg-von-Neumayer-Station in Richtung der südlichen Kottasherge.

Fig. 5b: Profile of the reduced total intensity ΔT_{red} (after separation of the normal field) from Georg von Neumayer station to the southern Kottas Mountains.

Abb 5b: Profil der reduzierten Totalintensität des Erdmagnetfeldes ΔT_{red} (nach Normalfeldreduktion) von der Georg-von-Neumayer-Station in Richtung der südlichen Kottasherge.

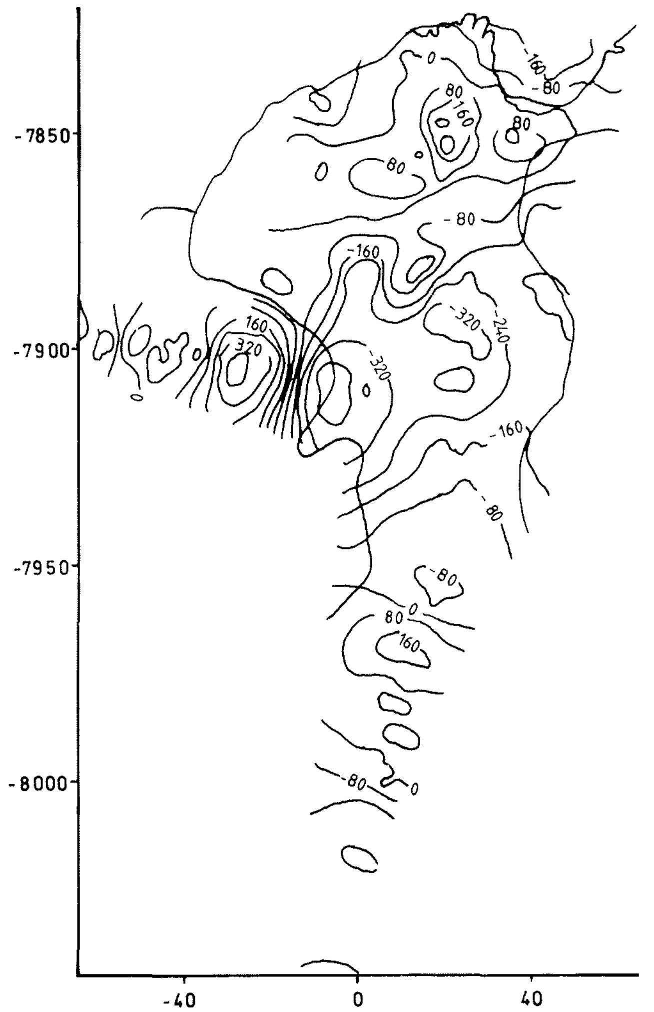

Fig. 6a: Map of the reduced total intensity ΔT_{red} over the area of investigation including the Søråsen and the southern hinterland of the Ekström Ice Shelf.

Abb. 6a: Magnetik-Karte der reduzierten Totalintensität ΔT_{red} im gesamten Untersuchungsgebiet einschließlich des Søråsens und des südlichen Hinterlandes.

Figure 6b shows a magnetic map of only the Ekström Ice Shelf where the reliability of interpolation is better. A view from the SSE direction into the three dimensional ΔT_{red}-anomaly-"mountains" is given in Figure 6c. In the center region of the map a prominent negative magnetic anomaly is found, surrounded by a high positive one in the west and a smaller positive anomaly in the north running parallel to the coast line. Such coast parallel anomalies are often seen in regions of passive continental margins. The western high positive anomaly in the Søråsen area which decreases further west (see Fig. 6a) may indicate a higher position of magnetic basement . This however is highly speculative without additional geophysical information.

Upward Field Continuation

As aero-magnetic measurements at a flight level of 2.5 km have been carried out by Soviet expeditions in the same area, upward field continuation was done by 2D-Fourier transformation of the interpolated data into the spatial wave-number-domain applying a transform function and then inverse Fourier transformation using a software package(TOEDT, personal communication). HENDERSON (1970) showed that ΔT-anomalies determined by (cheap and fast) proton precision measurements may be continued directly without intermediate steps.

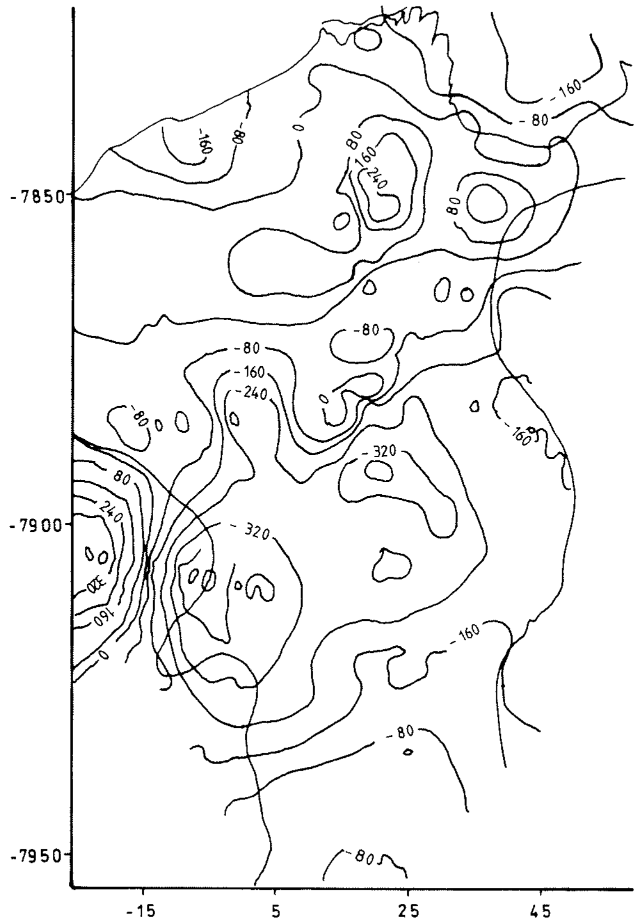

Fig. 6b: Map of the reduced total intensity ΔT_{red} on Ekström Ice Shelf in UTM coordinates/km: (x,y) - (west/east from -9° W, south from equator).

Abb. 6b: Magnetik-Karte der reduzierten Totalintensität ΔT_{red} des Ekström-Schelfeises in UTM-Koordinaten /km: (x,y) - (westl./östlich von Bezugslänge -9° W, südlich des Äquators).

Figure 7 shows the result of 2.5 km upward field continuation (which has the same effect as low-pass-filtering). The map is in good agreement with the Soviet aero-magnetic measurements.

Reduction to the Pole

Unless the field inclination is 90°, the source of a geomagnetic anomaly is not exactly below the anomaly itself due to the multipole character of the magnetic field. Thus the amount and direction of the field vector is dependent on latitude, and therefore the amplitude, sign and position of an anomaly relative to the geological source (the „disturbing body") changes with magnetic latitude.The so-called „reduction to the pole" corrects for this effect. The procedure converts the actual inclination to the pole inclination (I = 90°). This corresponds to an inducing field including a magnetization vector pointing perpendicular to the surface of the earth.

The anomalies reduced to the pole are sometimes called „pseudo-gravimetric" as this reduction is based on the Poisson-Eötvös-Equation which gives the link between the magnetostatic potential U of a homogenously magnetized body and its gravity potential V if the body has a constant density:

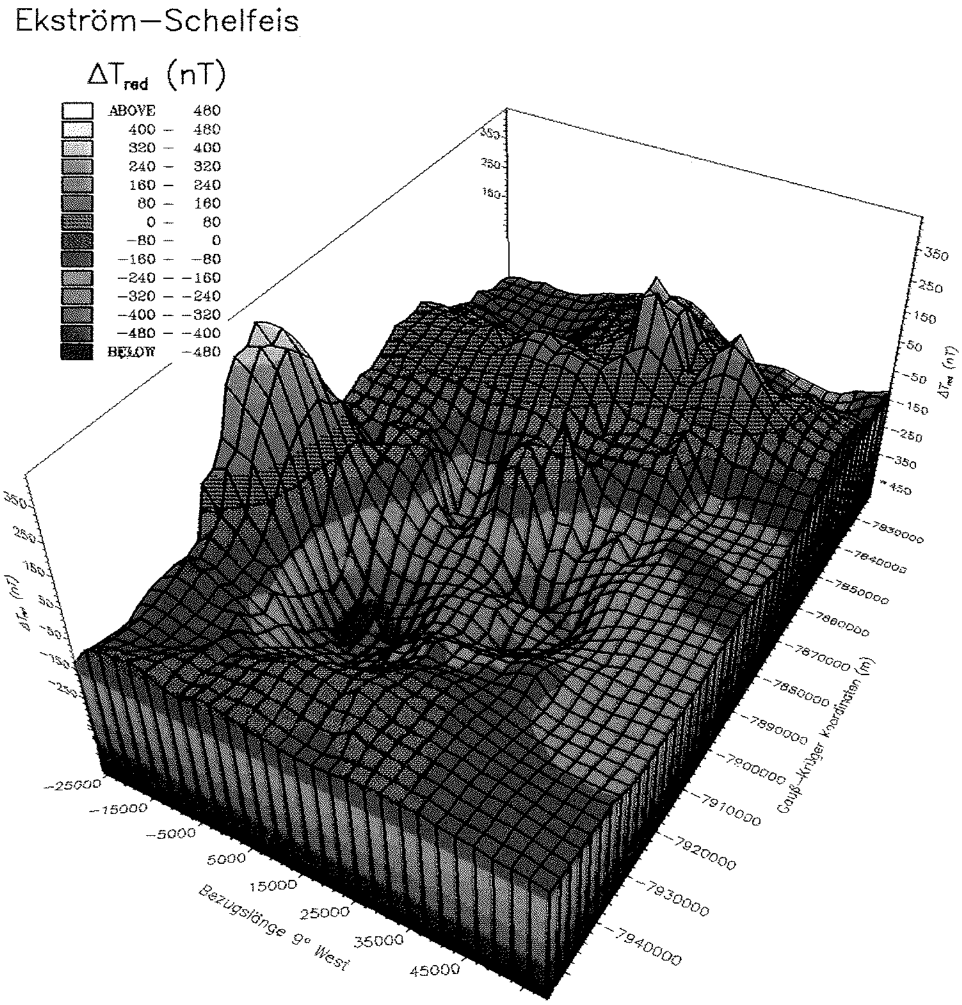

Fig. 6c: 3D-map of the reduced total intensity ΔT_{red} on the Ekström Ice Shelf, seen from SSE direction.

Abb. 6c: 3D-Magnetik-Karte der reduzierten Totalintensität ΔT_{red} des Ekström-Schelfeises, Blickrichtung von SSE.

$$U(\vec{r}) = -\frac{1}{4\pi f\sigma}\cdot\vec{M}\cdot\nabla V(\vec{r})$$

f – gravity constant
σ – density
$\vec{M}$ – magnetization vector

Differentiation to the vertical direction ($\partial/\partial z$) leads to the total intensity reduced to the pole ΔT_{pol} (see e.g. MILITZER, LINDNER & SCHEIBE 1984):

$$\Delta T_{pol} = -\frac{\mu_0 M}{4\pi f \sigma} \cdot \frac{\partial g}{\partial z} \qquad (3)$$

μ_0 – permeability

$\frac{\partial g}{\partial z}$ – vertical gravity gradient

The calculation of equation 3 was done in the spatial wave number domain (TOEDT 1984) after 2D-Fourier transformation [$\Delta T_{red}(x,y) \rightarrow S(u,v)$; u,v = wave numbers in x,y direction].

A magnetic anomaly is the result of two vectors: the induced and the remanent magnetization (M_{ind}, M_{rem}):

$$\vec{M} = \vec{M}_{ind} + \vec{M}_{rem} = \chi \cdot \vec{H} + \vec{M}_{rem}$$

χ – magnetic susceptibility; $\vec{H}$ – magnetizing force .

Usually continental rocks are characterized by induced magnetization and marine rocks have a relatively large remanent magnetization. The reduction to the pole depends on the validity of the following conditions:
1) Homogeneity of magnetization, i.e. spatial constancy of remanent magnetization and susceptibility; also the induced and remanent inclination and declination angles have to be known;
2) Isotropy of the susceptibility;
3) A quasi-homogenous normal field in the area of interest;
4) The magnetic anomaly is small relative to the normal field ($\Delta T_{red} << T_{IGRF}$)

The last two conditions are certainly satisfied, but it is impossible to check wether the first two assumptions hold, as unfortunately, no samples of rock are available. So the results are speculative.

For an initial guess of the location of disturbing bodies two cases have heen calculated:

1) $I_{ind} = I_{rem} = -62°$, $D_{ind} = D_{rem} = -12°$
2) $I_{ind} = -62° \neq I_{rem} = -28°$, $D_{ind} = D_{rem} = -12°$

I – inclination, D – declination, $_{ind}$ – induced, $_{rem}$ – remanent

The first case represents the actual local field direction. The results are shown in Figure 8a. The anomalies, e.g. the postive coast parallel anomaly, are displaced about 10 km to the SSE after reduction to the pole (compare Fig. 8a with Fig. 6b). The idea behind the second choice of a remanent inclination $I_{rem} = -28°$ was that the area of Ekström Ice Shelf, a passive continental margin, might have been much closer to the magnetic equator when the Jurassic continent of Gondwana broke into new continental pieces. The position of the interesting area was once towards South East Africa (former position of Ross Orogen, continuation of Ellsworth Orogen; FÜTTERER 1986) and the geological bodies creating the magnetic anomalies had heen effusive rocks cooling below Curie temperature 60 million years ago or earlier at this magnetic latitude

Assuming different inclination for the remanent and the induced field, the structure of the anomalies becomes more different by reduction to the pole, and they, e.g., the positive coast parallel anomaly, are displaced to SSE more than 20 km, depending on spatial wave length (see Fig. 8b and Fig. 6b).

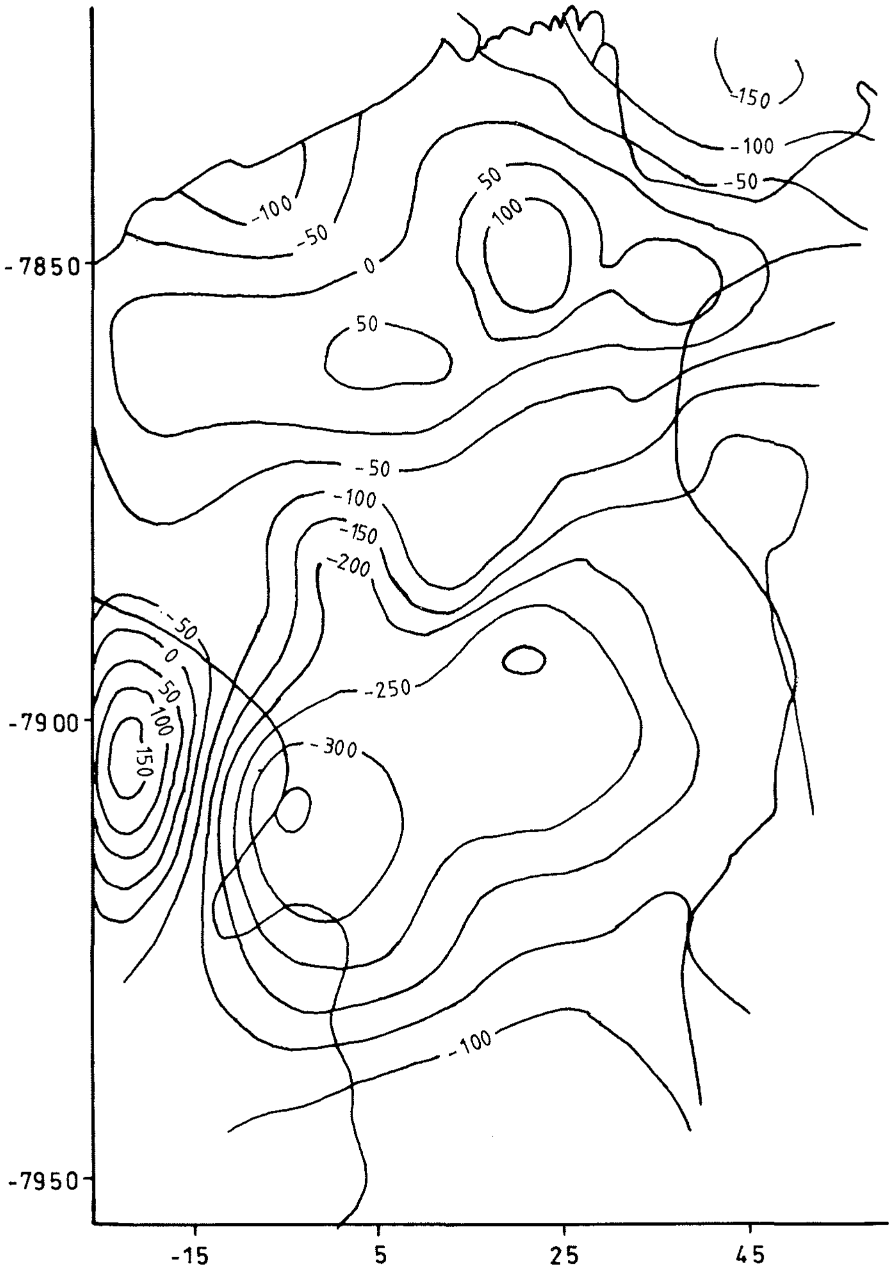

Fig. 7: Upward field continuation of the reduced total intensity ΔT_{red} on Ekström Ice Shelf, altitude 2.5 km.

Abb. 7: Feldfortsetzung der reduzierten Totalintensität ΔT_{red} auf dem Ekström-Schelfeis in eine Höhe von 2.5 km.

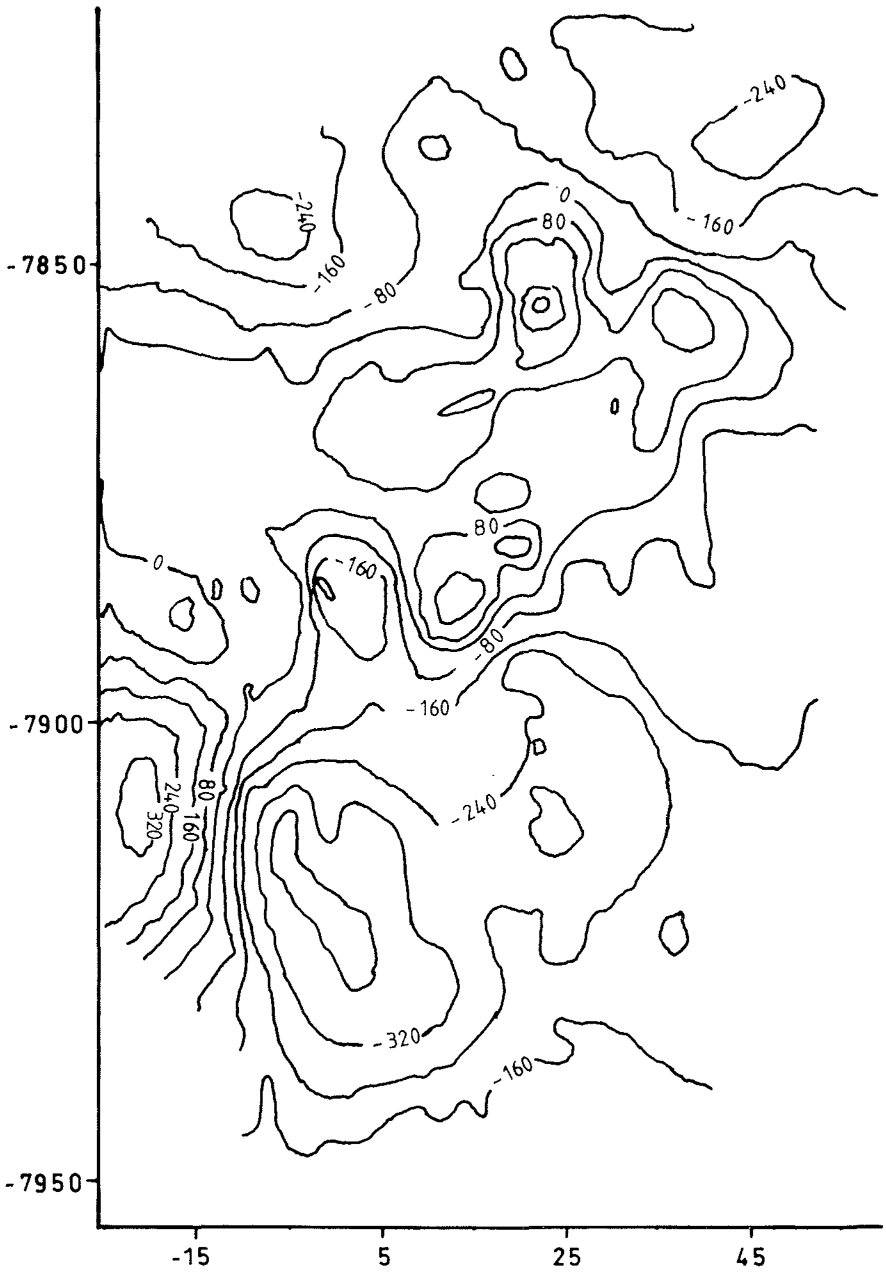

Fig. 8a: Reduction to the pole ΔT_{pol} of the reduced total intensity on Ekström Ice Shelf, equal induced and remanent magnetization: $I_{ind} = I_{rem} = -62°$, $D_{ind} = D_{rem} = -12°$.

Abb. 8a: Polreduktion ΔT_{pol} der reduzierten Totalintensität auf dem Ekström-Schelfeis, induzierte gleich remanente Magnetisierung: $I_{ind} = I_{rem} = -62°$, $D_{ind} = D_{rem} = -12°$.

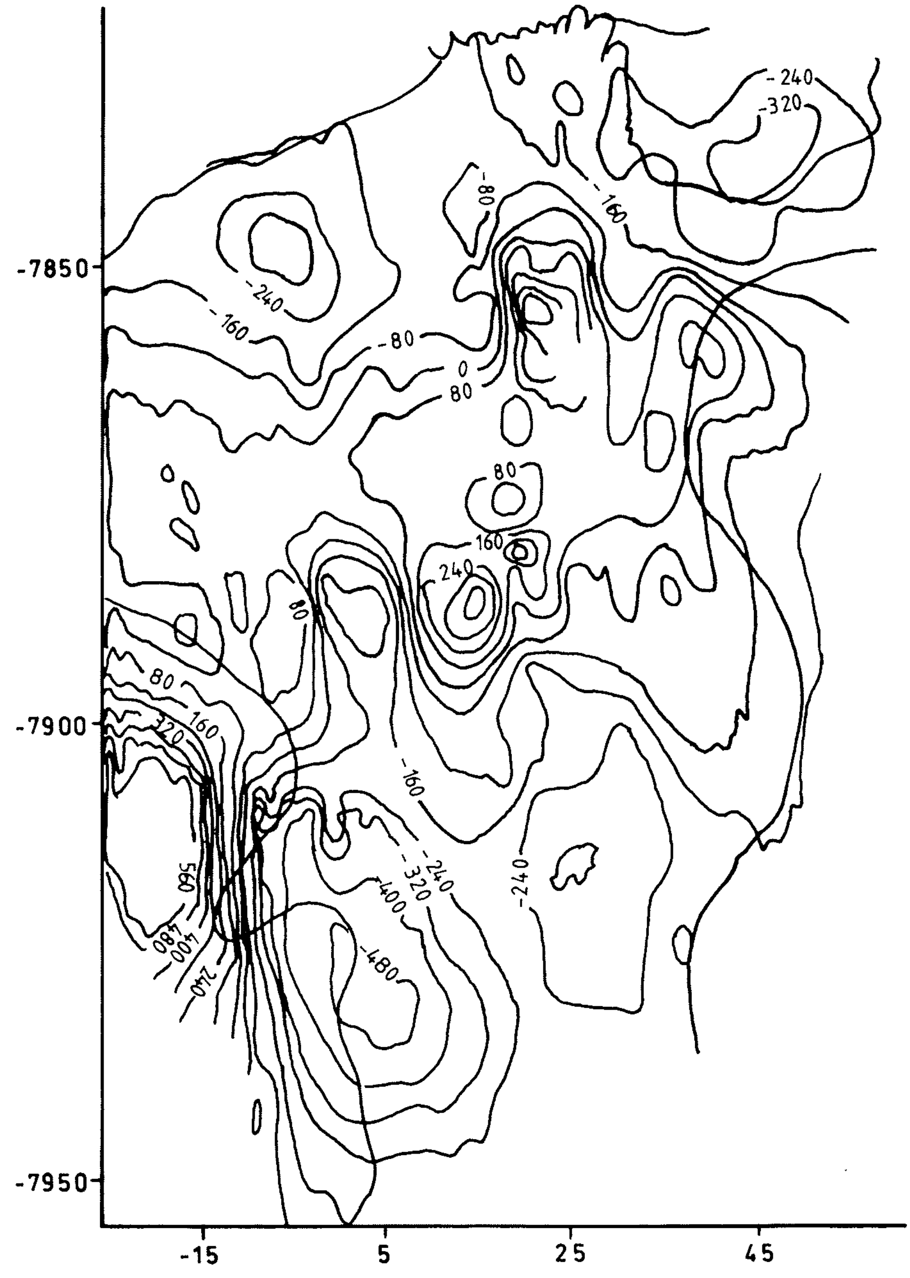

Fig. 8b: Reduction to the pole ΔT_{pol} of the reduced total intensity on Ekström Ice Shelf, different induced and remanent magnetization: $I_{ind} = -62°$, $I_{rem} = -28°$, $D_{ind} = D_{rem} = -12°$.

Abb 8b: Polreduktion ΔT_{pol} der reduzierten Totalintensität auf dem Ekström-Schelfeis, ungleiche induzierte und remanente Magnetisierung: $I_{ind} = -62°$, $I_{rem} = -28°$, $D_{ind} = D_{rem} = -12°$.

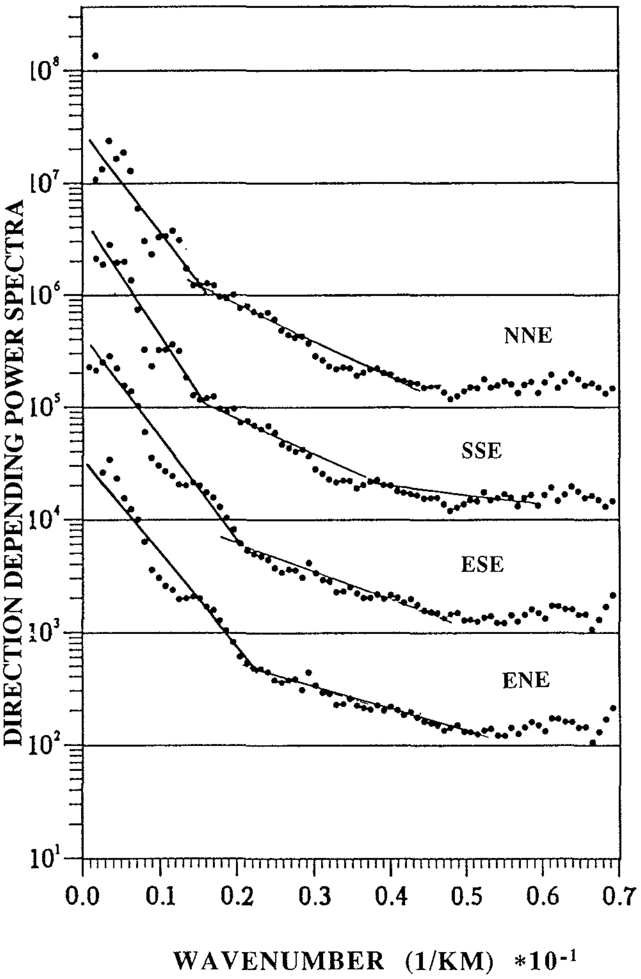

Fig. 9: Direction dependent power density spectra of reduced total intensity ΔT_{red} on the Ekström Ice Shelf: log E as a function of the spatial wave number.

Abb. 9: Richtungsabhängiges Energiedichtespektrum der reduzierten Totalintensität ΔT_{red} auf dem Ekström-Schelfeis: log E als Funktion der Wellenzahl.

DEPTH ESTIMATION OF DISTURBING BODIES

Although reduction to the pole is speculative due to the lack of information about induced and remanent magnetization of crustal sources, the two-dimensional spectrum S(u,v) of their anomaly field alone contains information which is sufficient for reliable depth estimations. As the multipole magnetic field decreases fast with increasing distance to a source (for the dipole part ~ $1/r^3$) shallow structures are characterized by high, and deep structures by low spatial wave numbers.

Figure 9 shows the power spectra $E(u,v) = S(u,v)^{2}$ calculated for different directions, beginning from the center of the ΔT_{red}-anomaly map, to NNE, SSE, ESE and ENE. Log E is plotted as a function of wave number k (spectra in ESE, SSE and NNE direction are shifted by the factors 10, 10^2 and 10^3 to show them all in one plot). If this function can be approximated by straight lines (in the semi-logarithmic diagram), the sources can be interpreted as block-like disturbing bodies and the slope of one straight line is a measure of the depth h of the upper edge of a block-like body (SPECTOR & GRANT 1970). So the depth h can be calculated by the formula

$$E(k) = \text{const } L \cdot e^{-4\pi kh} \quad \text{(LINDNER \& SCHEIBE 1977).} \qquad (4)$$

The spectra are approximated by two or three (SSE) dominant straight lines.

The steepest slopes of the first lines are in the „red" part („red spectrum", i.e. most of the energy created by deepest bodies with low wave numbers / long wave length) and the corresponding depths are h_1 = 14 - 18 km. This is about 13 - 17 km below the sea bottom as ice thickness of the Ekström Ice Shelf plus water depth is about 1 km. And it means that Curie depth, where magnetization is zero as rocks are too hot, in the region exceeds 13 - 17 km. The depths of block-like structures were also determined by forward modelling of the long north-south profile crossing the whole Ekström Ice Shelf to the southern hinterland (see Figs. 2 and 5). SPIESS (pers. comm. 1990) found Curie depths of about more than 12 - 18 km which is in good agreement with spectral depth estimations.

The lines in the middle part of the spectra in Figure 9 have gentle slopes corresponding to depths of h_2 = 4 - 6 km, i.e. 3 - 5 km below sea bottom. The third line approximating the spectrum in SSE-direction yields h_3 = 2 km, i.e. the upper edge of the disturbing body is only 1 km below sea bottom. So the sediment cover (almost no significant magnetization) cannot be very thick and there may be geologic features like sills and dykes close below sea bottom. A first seismic investigation in the central part of the Ekström Ice Shelf 15..30 km south of GvN (pers. comm. DEGUTSCH 1989) revealed dipping reflectors up to the sea bottom.

„White noise" (slope $\approx$ 0) is seen in the spectra at wave numbers $k \geq 0.5$ km^{-1} and $\lambda \leq 2$ km, respectively. But most energy is distributed at long spatial wave lengths: $\lambda_1(h_1) \approx$ 250..65 km, $\lambda_2(h_2) \approx$ 65..20 km. So, although the spectra were determined from interpolated data with gaps between the profiles of more than 10 km, but survey point density of 0.5..1 km on the profiles, the applied method seems to be reliable enough for depth estimations.

CONCLUSIONS

On Antarctic ice shelves, which are normally accessible for geophysical studies only with great efforts, relatively cheap and fast investigations of magnetic total intensity carried out with proton magnetometers are very worthwhile for studying the continental margin of Antarctica. They can be seen as a continuation of marine measurements to aeromagnetic surveys to get more local details of the morphology of the reduced magnetic field.

It is necessary to run base station measurements during the field campaign to control the variations and the inhomogeneity of the external field. Repeat station measurements displaced up to 90 km far from the GvN-observatory yield that the error of the reduced differential total intensity is smaller than 5 nT.

Upward field continuation is possible in the spatial wave number domain, based on differential total intensity data only. It is worthwhile to compare terrestrial with aero-magnetic measurements and to filter out the more regional field parts from the local and/or regional anomalous field.

Reduction to the pole gives information about the true location of the geological features creating the anomaly, but the applied parameters depend on the unknown geological history and are therefore very speculative. In general the anomalies are displaced at least 10..20 km south to south-east, depending on the wavelength and the difference between the direction of actual induced and former remanent magnetization. If the latter had equal direction, the anomaly map reduced to the pole is very similiar to the normal reduced one (except for the southward shift of anomalies position). ARKANIHAMED (1988) calculated similiar poleward tended shifts of positive anomalies on the east coast of Canada.

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