HEAT FLOW AND TECTONICS OF THE WESTERN ROSS SEA

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Abstract: As a part of the 1990-91 sixth Italian Antarctic Program, a diverse marine geophysical investigation of the western Ross Sea area was undertaken. We studied the recent Cenozoic tectonic activity in the Victoria Land Basin (VLB) to help determine its mechanisms of extension and its relation to the uplift of the Transantarctic Mountains (TAM). The survey was mainly concentrated in the Drygalski Basin and in the Terror Rift, for which we produced bathymetric and structural maps at a 1:100,000 scale. Seismic data show active rifting associated with extensional tectonic and volcanic activity, confirming a tectonic origin for the entire VLB. Major unconformities, affecting the sedimentary sequence and recording the main ice waxing and waning cycles, have been traced from the seismic lines. Oblique extension in the Drygalski Basin, with axis oriented in NE-SW direction, seems to be a late Neogene tectonic phase, which deviates from the regional E-W extension in the VLB. We collected a consistent heat flow (HF) data set in the deepest part of VLB. Limits to the uncertainties in the HF measurements, with particular care to the last glacial event (grounding, erosion and retreat), have been estimated. The corrected HF in the Drygalski Basin (98–110 mW m^{-2}, $n = 2$) is higher than that in the Terror Rift (79–89 mW m^{-2}, $n = 9$). The new seismic and HF data indicate, as the Drygalski Basin may represent a propagating rift, that developed in a pull-apart scheme, associated with transtensional deformation across the TAM.

Key words: heatflow, tectonics, rifting, volcanism, Victoria Land Basin

Introduction

During the 1990–91 sixth Italian Antarctic Program, we collected single-channel (SC) high resolution and multi-channel (MC) reflection seismics, gravity, magnetics, bathymetry, coring and heat flow (HF) data in the western Ross Sea. The main targets of the *R/V OGS Explora* program during January 1991 were: the recent tectonic evolution of the Victoria Land Basin (VLB), connected to the broad and active tectonism affecting the Transantarctic Mountains (TAM), the sediment transport and the depositional processes in the western Ross Sea.

The rifting and uplift mechanisms responsible for the intense lithospheric deformation in the broad Ross Sea rift area (Tessensohn and Wörner, 1991; Behrendt *et al.*, 1991) include the TAM uplift, the VLB crustal thinning and subsidence (Fig. 1). Late Cenozoic volcanic rocks, exposed along the rift margins, suggest that volcanism in this region was rift-related (Behrendt *et al.*, 1991). The west Antarctic rift shoulder extends along the TAM for about 3000 km, from northern Victoria Land to the Ellsworth Mountains. Behrendt and Cooper (1991) inferred that the uplift of the TAM and of the entire rift shoulder was episodic and has probably been an order of magnitude faster (~1 km /m.y.), at times including the present, than the mean rate. Flexural-type features (Stern and ten Brink, 1989) run subparallel, on opposite sides, to the rift shoulder and can be recognized in the Wilkes Basin outer gravity low to the west and in the central Ross Embayment outer gravity high to the east (Stern and ten Brink, 1989). Both rifting and uplift mechanisms strongly suggest lateral heterogeneities and regional thermal anomalies in the crust and upper mantle. To evaluate the thermal conditions near the TAM-VLB transition, we have measured HF in the Drygalski Basin and Terror Rift, making a particular effort to select appropriate sites where reliable HF measurements could be obtained. During the *OGS Explora* cruise the following data were acquired:

- approximately 500 km of MC seismic lines across Iselin Bank;
- 1650 km of analogue SC seismic data in the central VLB;
- approximately 3000 km of bathymetry and gravity data and 2200 km of magnetic profiles in VLB;
- seven gravity cores (18.4 m of total sediment recovery);
- 12 HF measurements, using the Trieste University multipenetration probe.

The locations of the data collected in the Drygalski Basin and in the Terror Rift are shown in Figs. 2 and 3 respectively. Although part of our data has not yet been fully processed, the reliability of our conventional marine HF measurements in VLB can be evaluated as well as the other marine HF measurements in the periantarctic basins.

Bathymetry Data

We combined bathymetric echo sounding data from two (1989–90 and 1990–91) *OGS Explora* cruises in the western Ross Sea with the previous available data (Davey and Cooper, 1987; Programma Nazionale di Ricerche in Antartide, 1989) to make a new bathymetric chart. Uncorrected depths, with a sound velocity in water of 1450 m/s, have been plotted to a scale of 1:100,000 with a 50 m contour interval, for both Drygalski Basin (Fig. 2) and Terror Rift (Fig. 3). Discrepancies in depth among different data sets are within 5 m and the satellite Global Positioning System (GPS) navigation data guarantee accurate positioning

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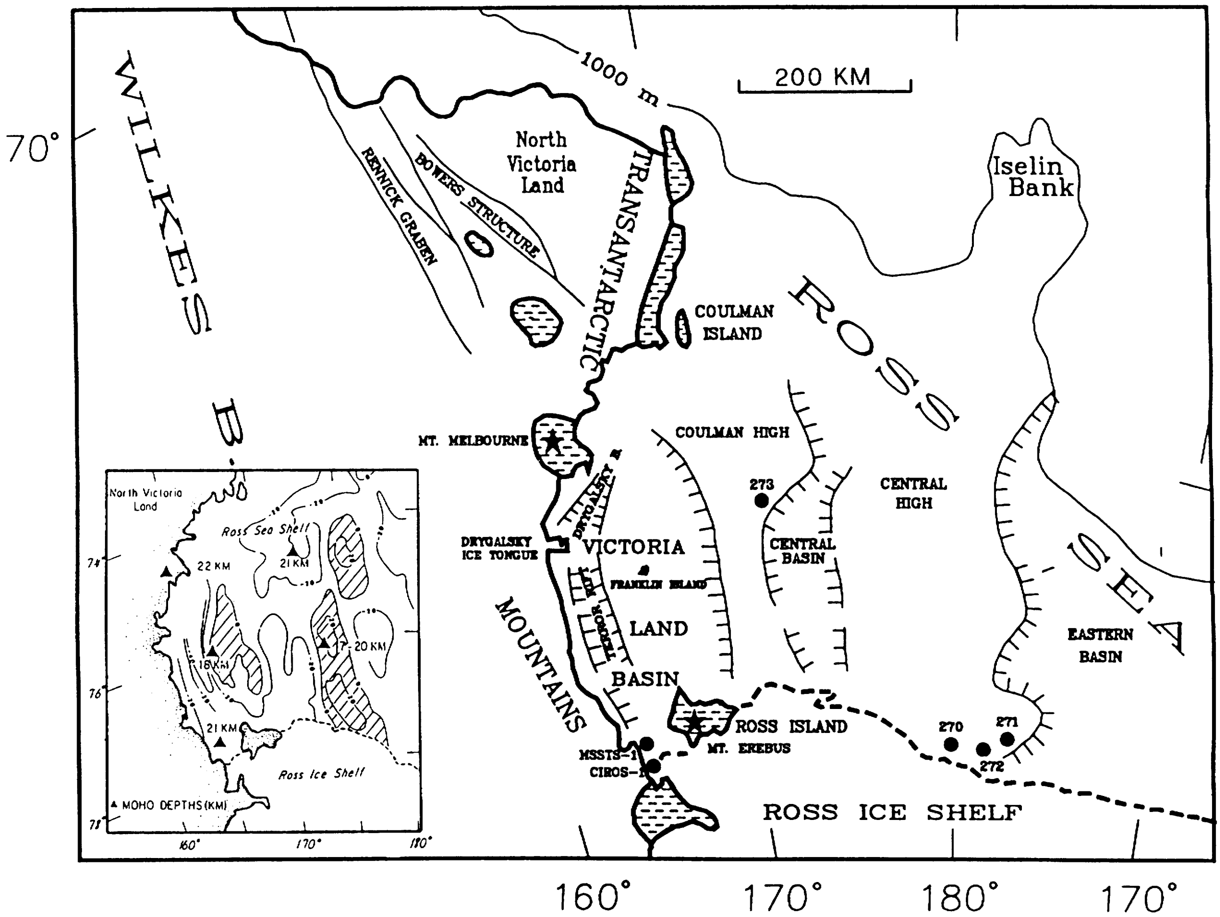


Fig. 1. Index map (modified from Behrendt *et al.*, 1991) of the Ross Sea region showing some structures and geographic locations discussed in the text. Late Cenozoic volcanic outcrops (dashed areas) and active volcanoes (stars) are indicated. Full dots indicate the locations of the drill holes. 1000 m bathymetric contour and the edge of the Ross Ice Shelf are also shown. The inset map (from Behrendt *et al.*, 1991) shows Moho depth estimates from seismic experiments in the Ross Sea. Simplified free air gravity contours at a 20 mgal interval are from Davey and Cooper (1987), and hatching denotes positive anomalies.

within 50 m.

From the bathymetric map shown in Fig. 2, the Drygalski Basin is a NE-SW asymmetric basin, deepening in the SW direction. Close to the Drygalski Ice Tongue (Fig. 1) a maximum depth of 1185 m has been observed. Rugged glacial topography characterizes the whole basin, which generally shows steeper and rougher flanks towards the coast than seaside. Breaks through the sea bottom, fault scarps and block faulting are the most prominent tectonic features.

The Drygalski Ice Tongue marks the approximate limit between the Drygalski Basin and the Terror Rift (Fig. 3). The latter clearly shows a N-S orientation that makes an angle of about 40° with the Drygalski axis and is deeper in the northern part. The eastern flank of the Terror Rift basin is bounded by a N-S volcanic ridge, well imaged in the magnetic signature (Bosum *et al.*, 1989), with a volcanic peak that rises up to 150 m below sea level.

Seismic Data

SC sparker lines, MC seismic lines from the 1990 *OGS Explora* cruise and USGS MC seismic data (Cooper *et al.*, 1987) have been studied. The OGS lines across the Drygalski Basin and Terror Rift (labelled as *IT90* in Figs. 2 and 3) are extremely useful for areal correlation as well as structural interpretation. Figure 4a shows line IT61 which runs across the Drygalski Basin, at about 75°S (labelled *IT90R61B* in Fig. 2) and Fig. 4b shows line IT63 across the Terror Rift, a little south of 76°S (labelled *IT90AR63* in Fig. 3).

Based on drilling sites (DSDP, CIROS-1 and MSSTS) and available seismic data, an attempt has been made to correlate the main seismic unconformities (Hinz and Kristoffersen, 1987) in the Terror Rift and Drygalski Basin (Figs. 4a and b). The more recent depositional sequences since middle Miocene, bounded by the unconformities U_1, U_2, U_3 and U_4 (which are recognizable in the Eastern Basin), are missing or extremely reduced in the VLB because of erosion and/or poor deposition. Unconformity U_5 (early Miocene) has been identified in the Terror Rift only, whereas the older ones are present in both basins. The maximum thickness of the Neogene sediments back to U_6 is less than 1 s in TWT and there are about 1.5 to 2.0 s of older sediments. The two seismic lines shown in Figs. 4a and b can be assumed as representative of the structural and depositional

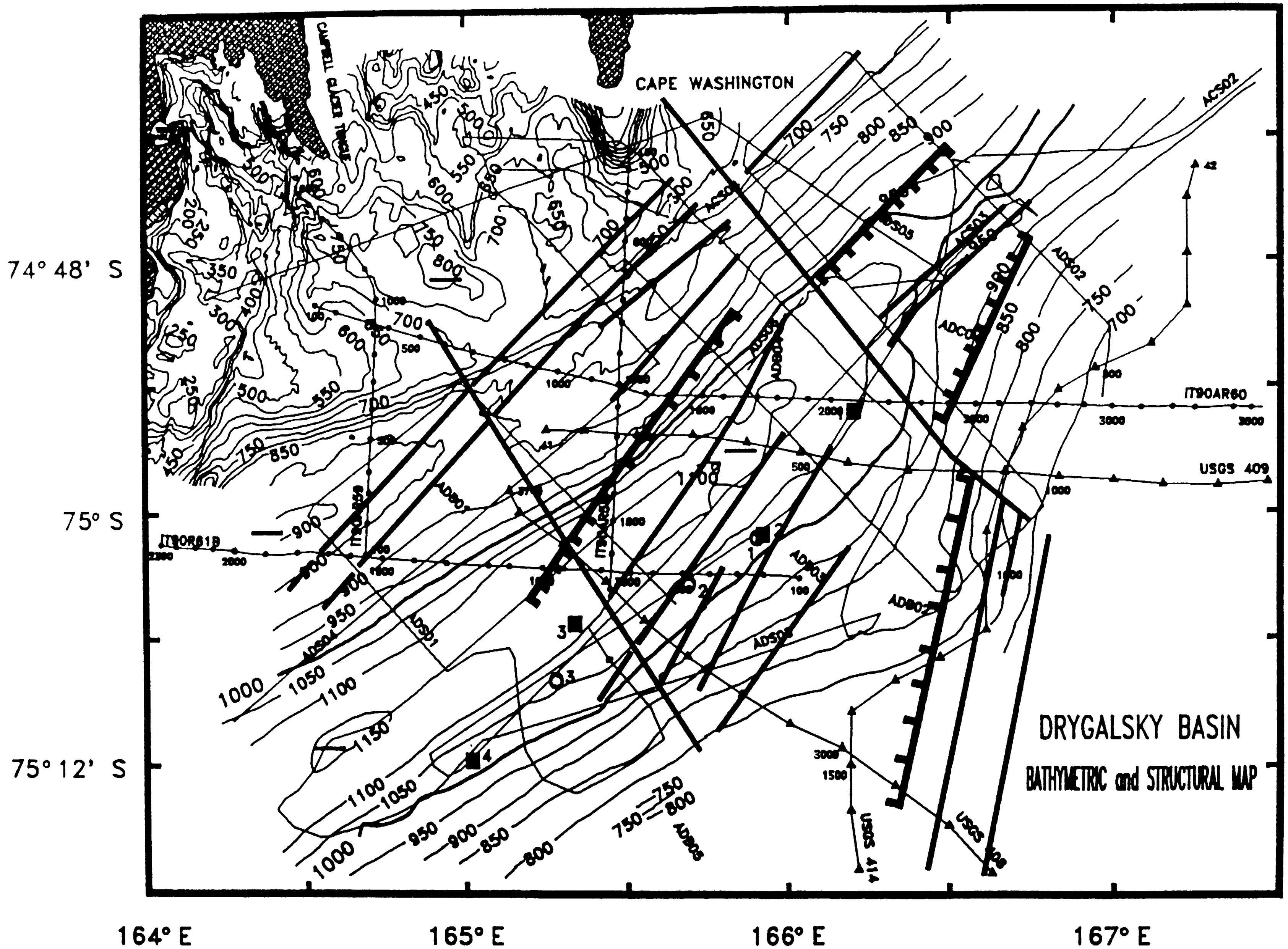


Fig. 2. Bathymetric and schematic structural map of the Drygalski Basin. Bathymetric contour interval is 50 m. The locations of USGS (lines with triangles) and OGS MC seismic lines (lines with full dots) and SC sparker lines (thin lines with no shot point marks) are shown. HF (open circles) and coring (solid squares) stations are also indicated. Major faults are indicated as thick lines; master faults are shown as thick toothed lines.

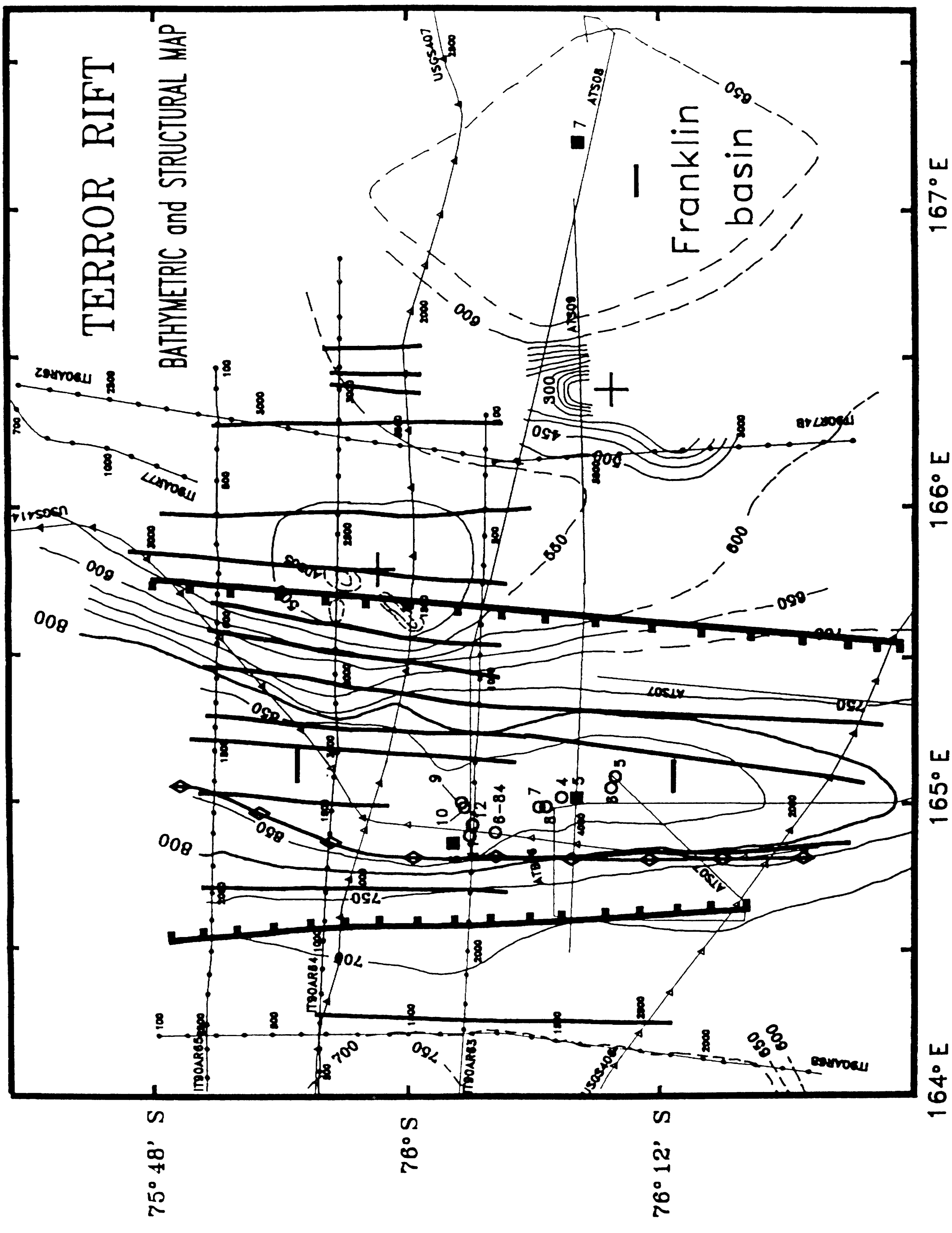
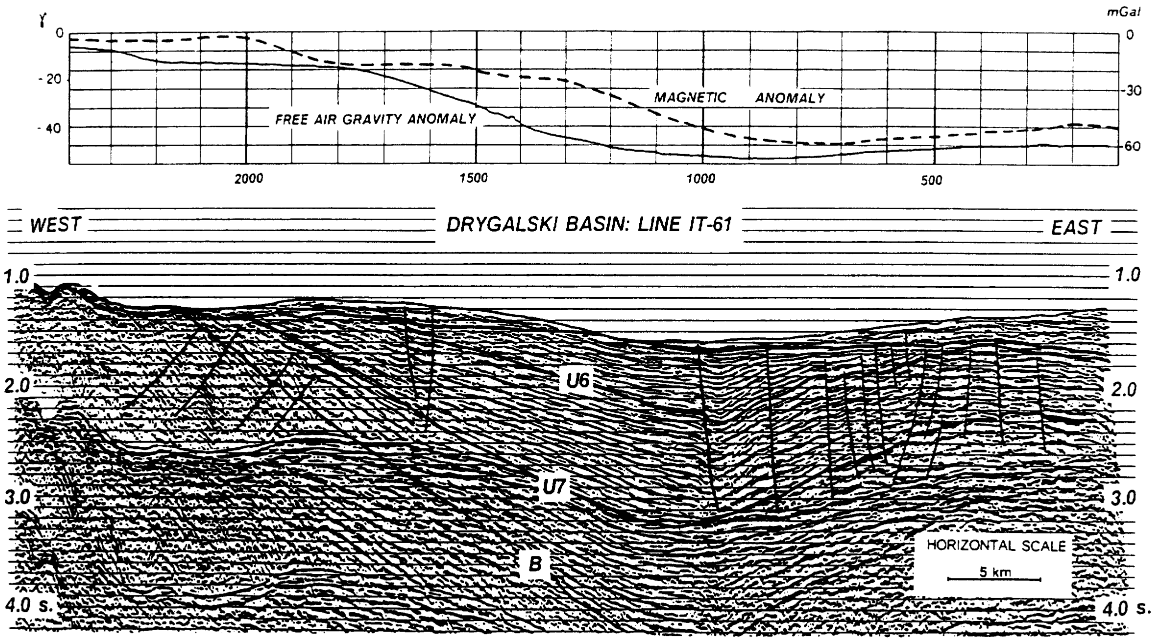
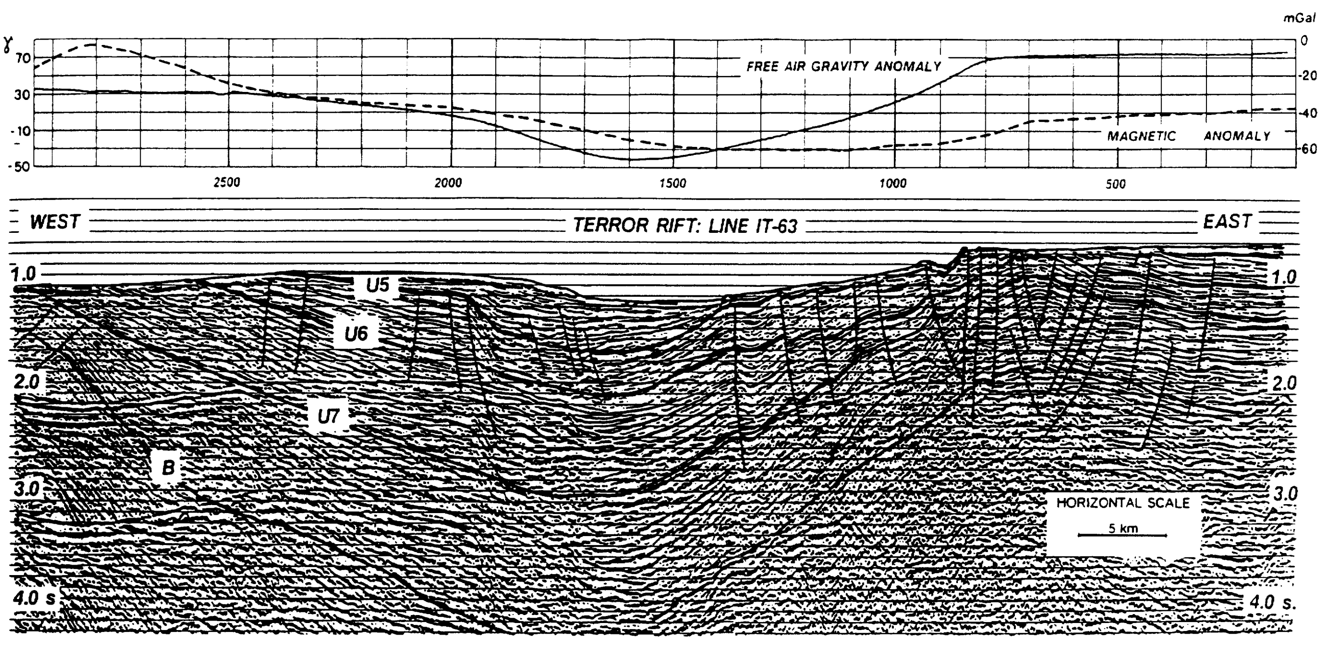


Fig. 3. Bathymetric and schematic structural map of the Terror Rift. The line with open diamonds represents the axis of the early Cenozoic basin (see Fig. 4b). Other symbols are as in Fig. 2.



(a)



(b)

Fig. 4. (a) MC seismic line IT-61 across the Drygalski Basin at about 75°S (location in Fig. 2 as line *IT90R61B*). The main unconformities (U6 and U7), faults and acoustic basement (B) are indicated. The gravity and magnetic anomalies are also shown. Arabic numbers on the horizontal scale are shot point numbers. (b) MC seismic line IT-63 across the Terror Rift at about 76°S (location in Fig. 3 as line *IT90AR63*).

relationships of the entire VLB. Well stratified and relatively subparallel units (about 1 s thick) underlie U_6 on the western flank of the basin. Beneath these units there is a major angular unconformity that represents the first tectonic extensional phase. The stratified and eroded older sequences below U_7 could be interpreted as part of the clastic deposits of the Beacon Supergroup.

Structural Map

Using the available seismic data, we produced schematic structural maps for the Drygalski Basin and Terror Rift (Figs. 2 and 3). Extensional and listric faults have been recognized on the seismic lines both in the Drygalski Basin and Terror Rift, which suggest a tectonic control on the evolution of the entire VLB. Its evolution seems to be controlled by extensional tectonic processes, including rifting, faulting, subsidence and locally associated magmatic and volcanic activity.

Three main extensional phases can be recognized in the central part of the VLB; they are well documented on the available MC seismic lines (Cooper *et al.*, 1987), as well as on the OGS lines (Figs. 4a and b). The oldest phase is probably associated with the Gondwana break-up (Mesozoic age), while the intermediate one is coeval with the TAM uplift and with the formation of the main Ross Sea depocenters (late Mesozoic-early Cenozoic). The most recent uplift phase (late Neogene) affects a narrow portion in the axial part of VLB. The master faults, bordering the basin and approximately defining the extent of the last extensional phase, are schematically shown in Figs. 2 and 3. One of the most evident and well established products of this recent extensional phase is the rift-related alkaline magmatism of the Cape Hallett-McMurdo province (Wörner *et al.*, 1989; Tessensohn and Wörner, 1991).

The last rifting event only partially follows the previous N-S structural trends. In the Drygalski Basin a clearly different orientation (NE-SW) can be observed (Fig. 2). Structural analysis of brittle fault arrays and dolerite dyke swarms in southern Victoria Land has shown the occurrence of similar fault patterns of Cenozoic age, indicating that there is a significant component of right-lateral shear displacement along the TAM (Wilson, 1992). Strike-slip lineations (with a NW-SE orientation) across the Drygalski Basin seem to accommodate the rifting-related stress field, which acts differently from the SW rifted to the NE locked area. This segment of offset rift seems bounded to the N by an E-W tectonic structure (just north of Cape Washington and parallel to the polar 3 anomaly) and to the S by an E-W lineation of structural weakness approximately at the latitude of the Drygalski Ice Tongue.

The Drygalski Basin may represent a propagating rift, developing in a pull-apart scheme, associated with transtensional deformation across the TAM (Wilson, 1992).

Heat Flow Data

Reliable HF measurements in deformational areas are particularly useful in deciphering the regional tectonic regime and the deep thermal conditions. The assessment of temperature at depth provides important constraints for geological and geophysical modelling. The eleven published HF measurements (Blackman *et al.*, 1987 and references therein) in the Ross Embayment indicate that there is greater than average HF in the central and western Ross Sea (96 ± 32 mW m^{-2} is the mean value, with standard deviation, of the nine measurements in that area). One of them (station 9), located in the northern part of the Drygalski Basin, has a HF of 73.3 ± 11.7 mW m^{-2} and a second one (station 6-84 in Fig. 3), located in the central part of the Terror Rift, has a HF of 63.3 ± 5.2 mW m^{-2}. During *OGS Explora* cruise 1991, three more measurements were taken in the Drygalski Basin, in water depths exceeding 1000 m, and nine in the Terror Rift, in water depths greater than 850 m. The measurements (Table 1) were taken near the axis of the basins and, if possible, along a MC seismic line. The Department of Naval Architecture, Marine and Environmental Engineering (Trieste University) multipenetration HF probe, with seven outrigger thermistors spaced along a 5.5 m long barrel and *in situ* thermal conductivity measurements, was used for the investigation. The *in situ* thermal conductivity values, measured with a continuous line heater in each thermistor prove, were compared to more than 300 needle probe (Von Herzen and Maxwell, 1959) thermal conductivity values measured on seven gravity cores collected in the area (Table 2). Locations of HF measurements and coring stations are shown in Figs. 2 and 3. Summary of HF data and coring stations is shown in Tables 1 and 2, respectively.

The average sea bottom temperature was approximately

Table 1. R/V OGS Explora Ross Sea Cruise January 1991: Heat flow data.

Station No.	Latitude South	Longitude East	Depth (m)	N[a]	D[b] (m)	Thermal Gradient (mK m^{-1})	Thermal Conductivity (W m^{-1} K^{-1})	Heat Flow (mW m^{-2})
						Drygalsky Basin		
01	75°01'15"	165°55'10"	1020	4	2.8	93.5 ± 4.6	1.334 ± 0.097	125 ± 6
02	75°03'22"	165°42'28"	1032	4	2.8	94.6 ± 2.1	1.278 ± 0.032	121 ± 3
03	75°07'49"	165°17'13"	1060	2	1.5	≥ 56.8	0.906 ± 0.015	≥ 51
						Terror Rift		
04	76°07'18"	165°07'18"	865	4	4.0	76.6 ± 6.6	1.307 ± 0.048	100 ± 9
05	76°09'50"	165°05'36"	870	4	3.9	83.1 ± 1.1	1.101 ± 0.028	91 ± 1
06	76°09'40"	165°04'19"	870	4	3.7	109.4 ± 2.8	0.895 ± 0.192	98 ± 3
07	76°06'17"	164°59'16"	885	3	3.4	71.9 ± 0.5	1.087 ± 0.036	78 ± 0
08	76°06'25"	164°59'09"	875	3	3.1	99.1 ± 4.4	1.042 ± 0.182	103 ± 5
09	76°02'36"	164°59'18"	875	5	3.7	91.4 ± 9.0	0.951 ± 0.013	87 ± 9
10	76°02'33"	165°00'14"	880	4	3.8	87.8 ± 3.9	1.140 ± 0.041	100 ± 5
11	76°02'54"	164°53'33"	885	4	3.8	88.1 ± 3.7	1.188 ± 0.021	105 ± 4
12	76°03'00"	164°54'54"	893	4	3.8	123.2 ± 23.5	0.960 ± 0.037	118 ± 23

[a] Number of reliable thermistors penetrated.
[b] Depth of penetration of lowermost thermistor.

Table 2. R/V OGS Explora Ross Sea Cruise January 1991: Gravity cores and needle probe thermal conductivity Measurements.

Station No.	Latitude South	Longitude East	Depth (m)	P[a] (m)	R[b] (m)	Thermal Conductivity Measurements No.	Mean Value (W m^{-1} K^{-1})
				Drygalsky Basin			
D01	74°55'55"	166°12'06"	1056	3.0	2.7	49	1.10 ± 0.32
D02	75°01'02"	165°56'30"	1041	1.8	1.2	21	1.43 ± 0.38
D03	75°05'21"	165°21'03"	1120	2.5	1.8	32	0.85 ± 0.18
D04	75°11'07"	165°01'31"	1055	3.0	2.5	47	1.29 ± 0.22
				Terror Rift			
T05	76°08'06"	165°07'51"	820	4.5	3.9	66	0.98 ± 0.10
T06	76°02'08"	164°52'03"	870	3.8	3.1	57	1.00 ± 0.28
				Franklin Basin			
F07	76°08'12"	167°13'30"	704	4.0	3.2	59	0.91 ± 0.20

[a] Estimated penetration.
[b] Length of recovered core.

−1.9°C for the Drygalski Basin and −1.5°C for the Terror Rift. Despite the coarse and stiff sediment, penetrations of the order of 3 m in Drygalski Basin and 4 m in the Terror Rift were achieved; we used the shallower interval temperature gradient to calculate the depth below the sea floor of the topmost thermistor that penetrated the sediments. The average tilt of the probe was in the range of 5–15°C.

Most of the temperature (T) versus depth (z) plots are remarkably linear (Fig. 5a). Average thermal gradients of 94 ± 1 mK m^{-1} (number of observations $n = 2$) and of 92 ± 16 mK m^{-1} ($n = 9$) have been observed in the Drygalski and Terror Rift, respectively. The average gradient in the Terror Rift is in good agreement with the value of 98 mK m^{-1} which was measured by Blackman *et al.* (1987). A higher T gradient (Fig. 5a) may sometimes be observed in correspondence of the change in the sediment structure (Fig. 5b) between the uppermost 1–2 m of poorly consolidated till (with a thermal conductivity $k = 0.7$–0.8 W m^{-1} K^{-1}) and the underlaying more compacted sediment ($k > 1$ W m^{-1} K^{-1}). A very good agreement has been obtained between the *in situ* ($n = 40$) and the needle probe ($n = 331$) thermal conductivity values (Fig. 5b). The thermal conductivity structure of the recovered sediment in the Drygalski Basin differs from that observed in the Terror Rift (Fig. 5b). A high conductivity layer (~2 W m^{-1} K^{-1}) about 0.5 m thick is located in the Drygalski Basin, between the uppermost one meter of very low conductivity soft sediment and the underlying more compacted sediment. The mean observed HF values are 123 ± 3 mW m^{-2} ($n = 2$) in the Drygalski Basin and 98 ± 11 mW m^{-2} ($n = 9$) in the Terror Rift.

The standard marine HF technique and data reduction procedure (Lawver *et al.*, 1991) were adopted. *In situ* equilibrium temperatures are estimated within ±0.002°C, while the *in situ* thermal conductivities are accurate to 3–5%. The HF was computed as a weighted mean of all the interval HFs calculated between each thermistor pair of adjacent probes. The thermistor spacing, of about 0.6 m, is clearly not adequate to obtain detailed temperature and thermal conductivity profiles with depth (Fig. 5b); nevertheless the above reduction technique, used in combination with the detailed (each 5 cm) thermal conductivity profile derived from the cores, allowed us to estimate the final HF value within an overall uncertainty of approximately ±10%.

Among the environmental disturbances that may affect the HF measured in the western Ross Sea, we took into account the effect of bottom water circulation, erosion and sedimentation and disturbances related to the ice dynamics during the last glacial advance. The full stage of this glaciation has been assumed at about 18,000 years ago according to Kang Jiancheng and Wen Jiahong (1991).

On the basis of water temperature values and water masses dynamics in the Ross Embayment (Jacobs *et al.*, 1985), we assumed that the present bottom water temperature (BWT) in the deepest part of VLB as constant with time. This assumption, which implies a negligible correction for the annual BWT variation, is also supported by the quite linear distribution observed in the T *vs.* z data.

The sedimentary history is strictly related to the advance/retreat of the Antarctic ice sheets (Alley *et al.*, 1990; Blankenship *et al.*, 1990). The thermal disturbances caused by the ice advancing, grounding, erosion and retreat are considered to be the most relevant and difficult ones to assess. We computed separately in a very simple model (Fig. 6) the effects of an instantaneous total glacial erosion, estimated in about 500 m, and of a glacial thermal pulse, of between −5 and −8°C (relative to the actual BWT), into the bottom sediment of the Drygalski Basin and Terror Rift. The correction procedure is similar to the one proposed by Blackman *et al.* (1987), although the values assumed for some parameters (as total erosion, bottom grounded ice temperature and timings of the glacial event) are quite different. According to Kang Jiancheng and Wen Jiahong (1991), we used the following time-steps (Fig. 6d): beginning of ice sheet grounding 21,000 years ago, resident time

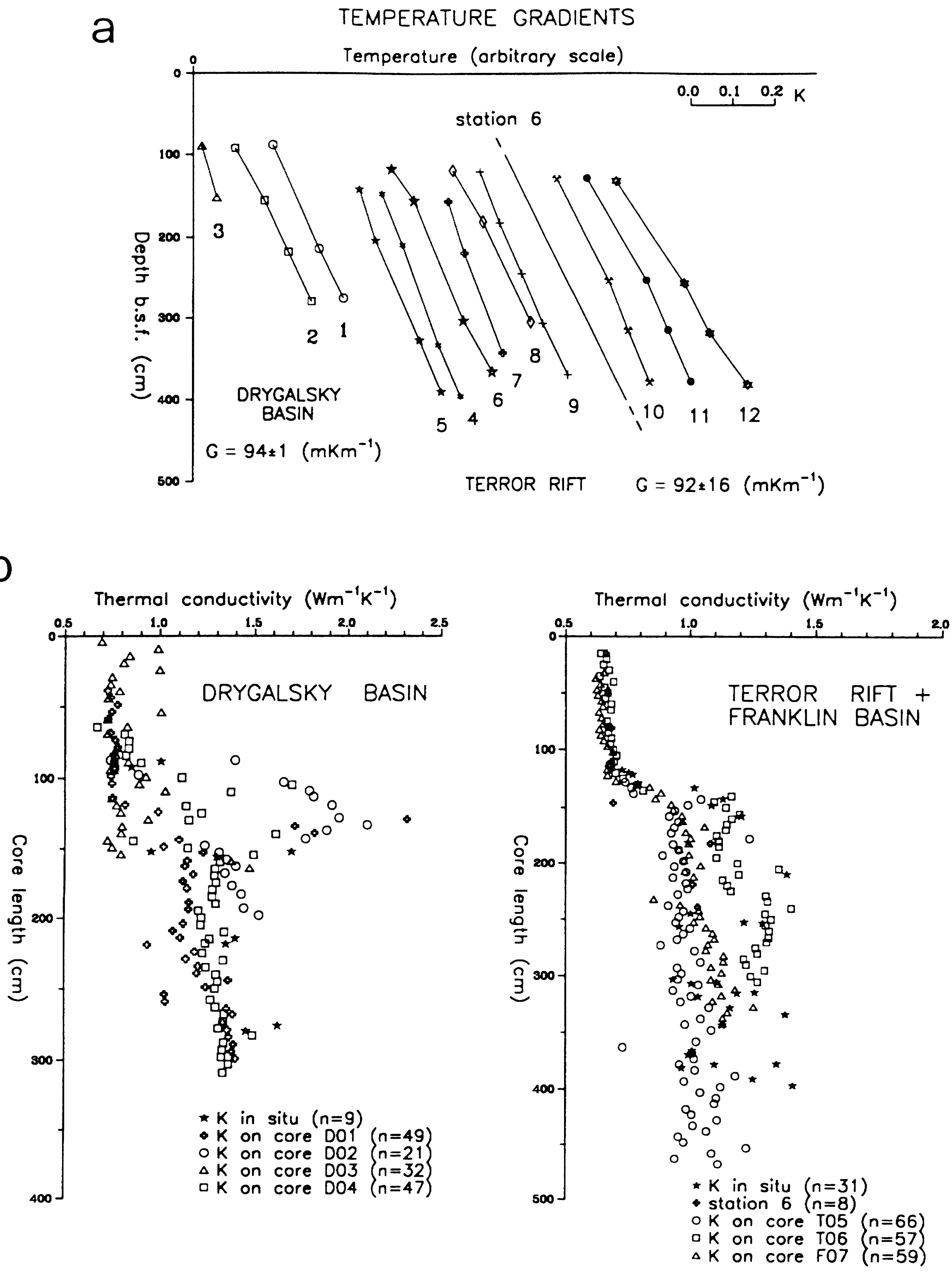
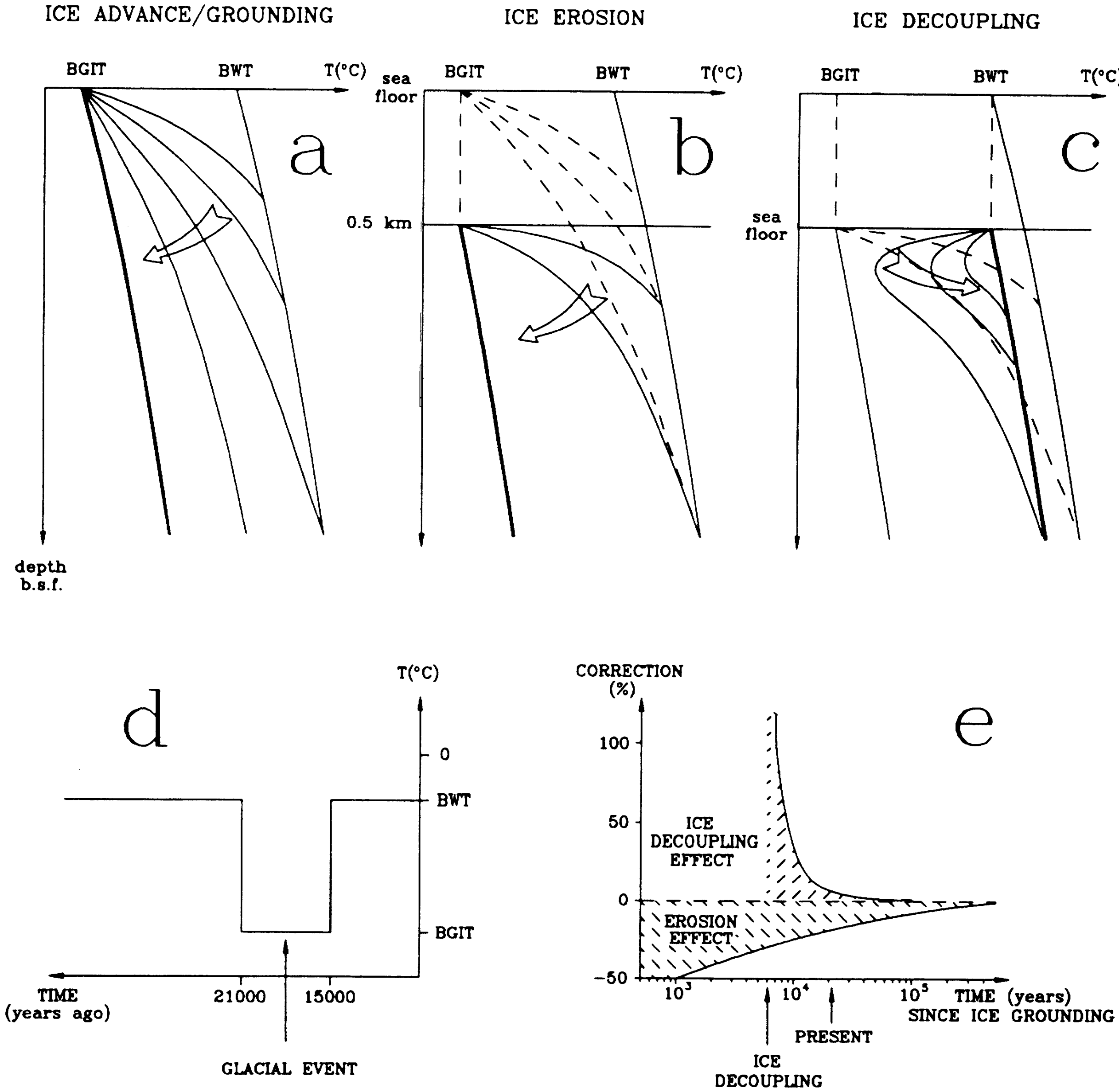


Fig. 5. (a) Thermal gradient plots of the three HF stations in the Drygalski Basin and nine in the Terror Rift. The average values for the two basins are also shown. The average thermal gradient of station 6-84 of Blackman *et al.* (1987) in the Terror Rift (see location in Fig. 3) is reported for comparison. (b) Plot of the *in situ* and needle probe thermal conductivity values for the Drygalski Basin, Terror Rift and Franklin Basin. Thermal conductivity values of station 6-84 (Blackman *et al.*, 1987) in the Terror Rift are also shown.



BGIT = bottom grounded ice temperature (<−1.9 °C)
BWT = bottom water temperature (~ −1.9 °C)

Fig. 6. Cartoon showing the thermal disturbances induced by a glacial event into the bottom sediments and the correction model applied. (a) Ice grounding lowers the BWT increasing of the thermal gradient near the sea floor. Transient temperatures are bounded by the two subparallel steady-state geotherms. (b) When grounded ice moves, erosion takes place. Sediment removal enhances the thermal disturbance induced by grounding, increasing even more the *T* gradient. Dashed geotherms represent *T* distribution before the beginning of erosion. 0.5 km of sediment erosion is assumed. (c) Ice decoupling from the sea bottom produces a thermal wave which propagates in depth showing, near the sea floor, an initial negative *T* gradient which slowly recovers back. Dashed geotherms represent *T* distribution before beginning of decoupling. (d) Time-step model for the glacial advance and retreat. (e) Percentage of correction as a function of time to be applied to the observed HF to account for erosion and thermal disturbances of the last glacial event in the Ross Sea.

of the ice sheet grounded on the bottom 6,000 years, beginning of ice decoupling from the bottom 15,000 years ago. Grounding and erosion (Figs. 6a and b) and bottom temperature changes by glacial decoupling from the sea bottom (Fig. 6c) have an opposite effect on the surface HF and so partially cancel each other. In VLB the glacial erosion, which tends to increase the observed HF value, has the most relevant and time-lasting effect, while the thermal pulse into the sea bottom dies away very rapidly and has now a weak effect. Figure 6e shows indicatively the percentage of correction as a function of time to be applied to the observed HF to account for erosion and thermal disturbances due to the last glacial event in the Ross Sea. According to this model, the HF values presently observed in the Drygalski Basin and in the Terror Rift need to be reduced by 10–20% approximately. The best corrected HF estimates become 98–110 mW m^{-2} for the Drygalski Basin and 79–89 mW m^{-2} for the Terror Rift. These values indicate a regional thermal anomaly (almost twice the normal HF) for the entire VLB, though they do not seem high enough to support present day rifting, as suggested by seismic data and volcanic activity. Likely this means that the corrections applied to the HF measurements could not be adequate. The large uncertainties in glacial erosion rates, in the timing of the glacial events as well as and in the sea bottom temperature changes will not be reduced without drilling in VLB. The difference between Drygalski Basin and Terror Rift HF could be accounted for by either local effects (for example, differences in erosion rates) or geological processes, such as more recent and/or intense crustal stretching in the Drygalski Basin than in the Terror Rift. On the basis of the available seismic, gravity and structural data we infer that the latter possibility is the most likely one.

Conclusions

Newly acquired *OGS Explora* echo-sounding and high-resolution seismic reflection data have been used to produce bathymetric and structural maps for the Drygalski Basin and the Terror Rift. Seismic data show active rifting associated with extensional tectonic and volcanic activity. In the Drygalski Basin we observe Neogenic oblique extension which appears to be episodic and propagating towards N-E.

We have not been able to evaluate Cenozoic and neotectonic rift structures (such as faults, basement grabens and volcanic intrusions) across the western Ross Sea active rift zone on the basis of HF, nor we have enough data to determine the relationship between HF and times of rifting. Nevertheless our data are in good agreement with previous studies (Blackman *et al.*, 1987 and references therein) and allowed us to estimate a better average HF value for the VLB. We therefore conclude that the HF in the western Ross Sea is about twice the normal one and it constitutes a regional feature related to the Cenozoic rifting/uplift processes.

The high HF in the VLB and the anomalous thermal conditions over the entire Ross Sea Embayment imply high temperatures at shallow depths mainly in the younger rift areas and could explain the uplift of the TAM by a thermal plume mechanism (Behrendt *et al.*, 1991). The extremely reduced crustal thickness (less than 10 km), the high tectonic subsidence, the rift-related volcanism affecting the VLB and the presence of flexural-type features (Stern and ten Brink, 1989) in the broad Ross Sea rift area are in favour of a thermal origin for the uplift of the TAM.

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