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The Eastern Ross Sea continental shelf during the Cenozoic: implications for the West Antarctic ice sheet development

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Abstract

The present-day bathymetric profile in the Ross Sea, as in other regions around the Antarctic margin, is deepening landward and shows unusually high water-depths: up to 1000 m in the inner shelf. These two features are the product of multiple ice sheet advances and retreats on the continental shelf. In this paper, we present a reconstruction of paleo-bathymetric profiles of the Eastern Ross Sea throughout the Cenozoic. The evolution of the sea-floor morphology from shallow and seaward dipping to the present-day configuration gives new insights into the understanding of the West Antarctic Ice Sheet (WAIS) history in this sector. Paleo-bathymetric profiles have been calculated by applying a reverse post-rift modelling, starting from a cross-section derived from multichannel seismic data. The post-rift reverse modelling includes: sediment decompaction, isostatic compensation after removing and recovering sediments of the post-rift thermal subsidence. The major uncertainty in our model is due to the paucity of stratigraphic constraints for the late Miocene and Pliocene sequences that prevents precise values of paleowater-depth being estimated. Nevertheless, major changes in the shape of the continental shelf and slope throughout the Cenozoic can be recognised, and mark some critical steps in the Ross Sea evolution. (1) Pre-Miocene: the Eastern Ross Sea was a deep structural basin bordered to the west by areas (e.g., the Central High) outcropping the sea level and hosting valley glaciers or small ice caps. A continental shelf edge was not clearly developed yet, the eastern flank of the Central High appeared as an inclined ramp, dipping towards the ocean. (2) Early to middle Miocene: tectonic subsidence gradually produced a marine flooding over most of the pre-Miocene sub-aerial areas. A continental shelf, slope and rise are gradually delineated. The shelf profile was seaward dipping and not yet overdeepened. The geometry of the depositional sequences is mainly determined by eustacy, tectonic and sediment supply. (3) Starting from Late Miocene (likely from 10 Ma to at least 4 Ma) the bathymetric profile evolved progressively from seaward to landward dipping and reached an overdeepened configuration, very similar to the present-day profile. Depositional and erosional processes over the continental shelf were largely controlled by ice streams.

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Outcropping of large parts of the continental shelf during the early Cenozoic has important implications on the volume of the West Antarctic Ice Sheet. At that time, the WAIS contribution to the eustatic fluctuations was most likely much larger than today.

Keywords: Antarctic; Cenozoic; continental shelf; Ross Sea; seismic stratigraphy; backstripping; rifting; paleobathymetry

1. Introduction

Despite the important role that the Antarctic criosphere plays in many global processes, its history is still poorly known and largely debated. The Antarctic ice sheet today is formed by two main systems: the East Antarctic (EAIS) and the West Antarctic ice sheet (WAIS) (Fig. 1). The first represents 90% of the total ice volume and is mainly grounded above sea level. The second is smaller and mainly grounded below sea level. These variations largely determine the different behaviour of the two systems: the WAIS has a less significant ice volume, but is more sensitive to climate and sea level variations and more dynamic as compared to the EAIS.

The Ross Sea that is one of the principal drainage areas for the WAIS (Fig. 1) is part of a wide rifting system, the Western Antarctic Rift System (WRS), that has been evolving since the 120 Ma break-up between Australia, New Zealand, and Antarctica (LeMasurier, 1990; LeMasurier and Rex, 1989; Behrendt et al. 1991; Lawver et al., 1991). The Ross Sea is bounded to the west by the Transantarctic Mountains (TAM), which represent the highest rift shoulder for the WRS and gradually isolated the Ross Sea during the Cenozoic from the direct influence of the EAIS. Consequently, the Eastern Ross Sea is a key area for studying the WAIS dynamics, because the WAIS waxing and waning shaped the morphology of the Eastern Ross Sea shelf directly and ruled its sedimentation history.

Ten Brink et al. (1995) modelled the relationship between strata geometry, morphology and ice sheet dynamics. They demonstrated that many advances and retreats of the ice sheet grounding line, to and from different locations across the continental shelf, produces erosion on the inner shelf and deposition on the outer shelf. These processes are principally responsible for the typical landward-deepening of the bathymetric profile observed in the Eastern Ross Sea and in many areas around the Antarctic continental margins (Cooper et al., 1991a).

Seismic profiles have been widely adopted to reconstruct the glacial history of the Ross Sea, but conclusions reached by different authors can be

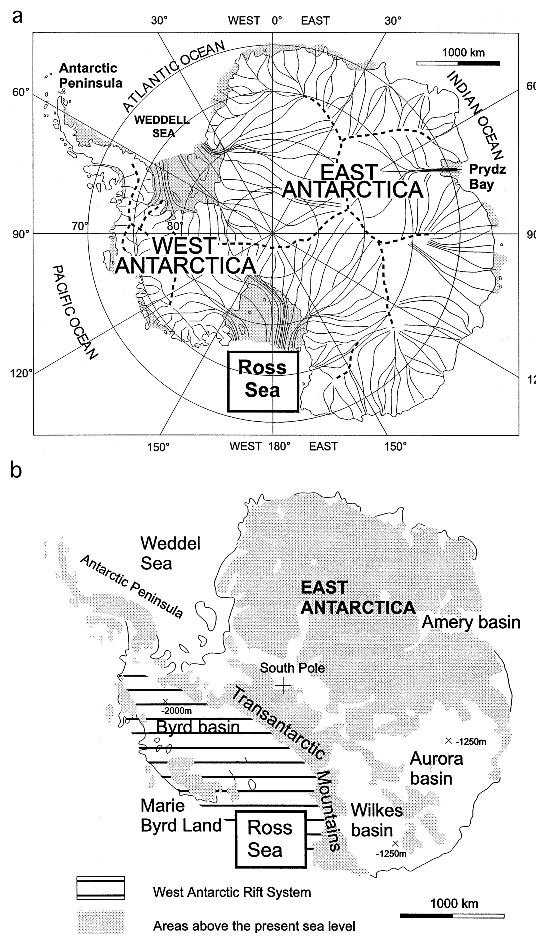


Fig. 1. (a) Ice flow directions in the East and West Antarctic Ice Sheet and ice shelves (shaded areas) (Drewry, 1983). (b): Sub-ice topography of Antarctica (after Drewry, 1983, modified by Webb, 1990). Shaded areas are above present-day sea level; spot values denote bedrock depth in metres below present-day sea level (Bentley, 1991). Major present-day depressions in West Antarctica, such as Byrd Basin and Ross Sea, are the combined results of tectonic (West Antarctic Rift System), sub-ice erosion and ice loading.

slightly different, mainly because of the lack of stratigraphic control and univocal characterisation of glacio-marine acoustic facies. Drilling sites like CIROS-1, MSSTS-1, Cape Roberts-1 and 2, and DVDP, are all concentrated in the westernmost area of the Ross Sea (Fig. 2), close to the TAM, and therefore their stratigraphic record is influenced mainly by local glaciers and/or by the EAIS (Barrett, 1989). Geological samples that recorded the WAIS history in the Ross Sea have been recovered by DSDP leg 28 (Hayes and Frakes, 1975) (Fig. 3). Unfortunately, data from DSDP leg 28 are too sparse and uncertain to constrain without doubt the age and the evolution of the glacial regime in the Ross Sea (Hayes and Frakes, 1975; Leckie and Webb, 1983, 1986; Savage and Ciesielski, 1983; Hambrey and Barrett, 1993). The stratigraphic record is, in fact, affected by a large hiatus or lack of data because of poor recovery in significant moments of the Antarctic criosphere evolution, like, for example, the period between 14 and 4 Ma or the entire Pleistocene. Moreover, there is still uncertainty about the depositional environment: it is not clear, for example, whether the early Miocene diamictites found in DSDP 270 (Hayes and Frakes, 1975) were deposited below and/or near a large, and polar, ice sheet or a small, and temperate, ice cap (Balshaw, 1981; Hambrey and Barrett, 1993). The occurrence of subglacial or ice-proximal sediments in the Eastern Ross Sea at DSDP 270, quite far from the present coastline, induced some authors to assume that the ice sheet was already large in size and, therefore, probably polar in the early Miocene. High resolution seismic data confirmed the ice-proximal or subglacial origin

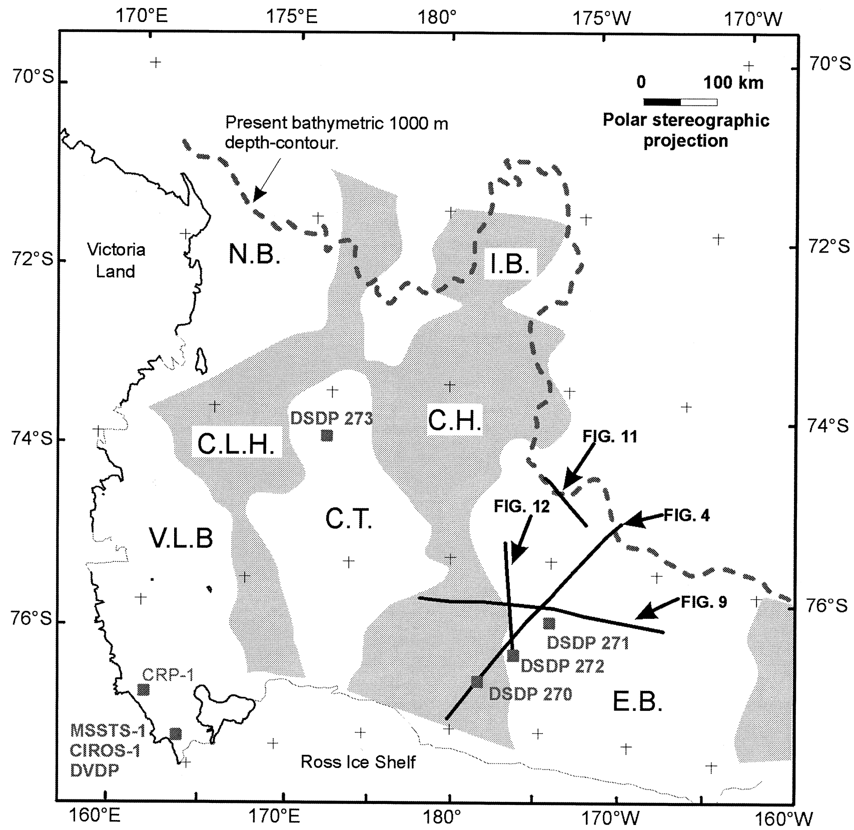


Fig. 2. Ross Sea structural highs (shaded areas) and location of the seismic lines shown in the paper. Leg 28 DSDP sites are also shown. Location of the BGR80-07 seismic profile is also shown across the Eastern Basin. N.B. is the Northern Basin; I.B. is the Iselin Bank; C.L.H. is the Coulman High; C.H. is the Central High; V.L.B. is the Victoria Land Basin; C.T. is the Central Trough; E.B. is the Eastern Basin.

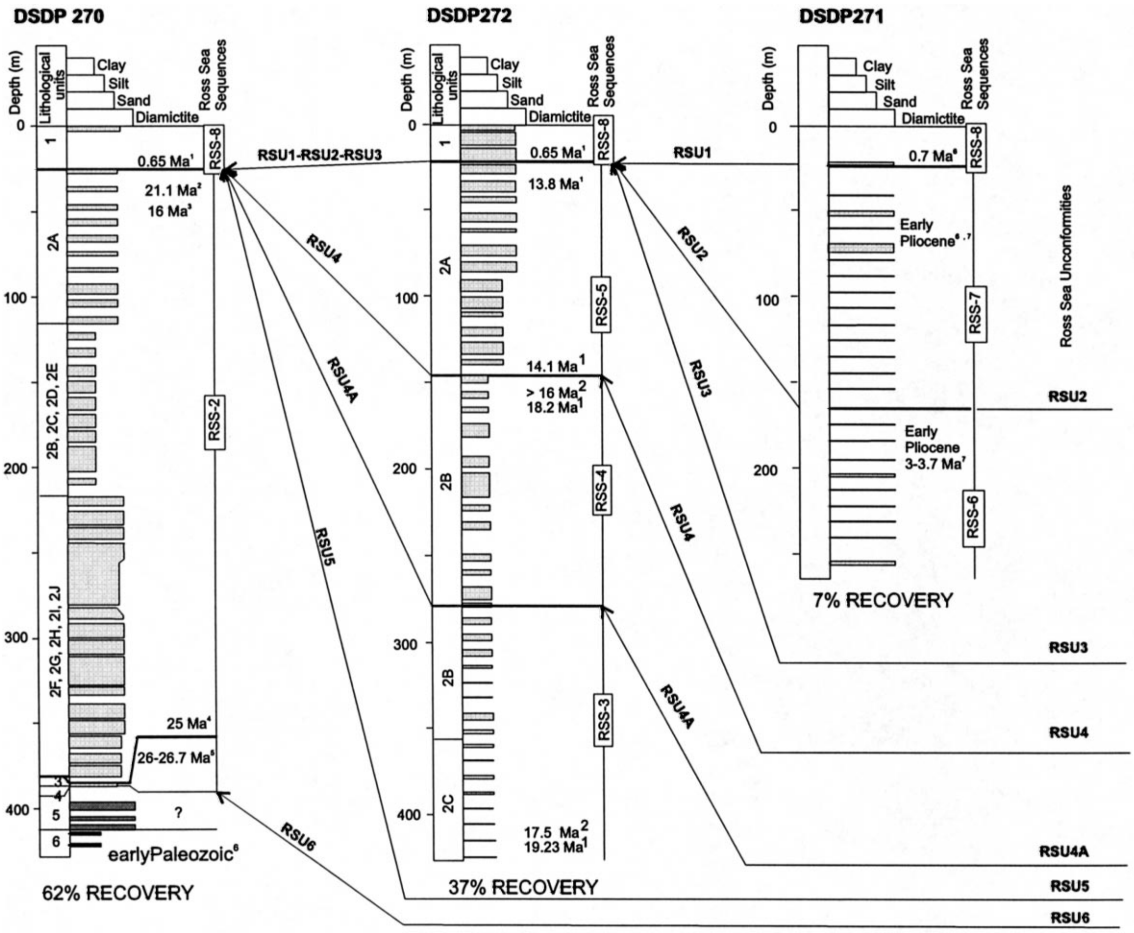


Fig. 3. Correlation chart of the RSS and RSU with existing drill sites data. See Fig. 2 for drill sites locations. Site 270 ages are from: 1 = Hayes and Frakes (1975); 2 = Steinhauff and Webb (1987); 3 = Allis et al. (1975); 4 = D'Agostino and Webb (1980), Leckie and Webb (1986); 5 = McDougall (1977); and 6 = Ford and Barrett (1975). Site 272 ages are from: 1 = Savage and Ciesielski (1983) and 2 = Leckie and Webb (1986). Site 271 ages are from: 6 = Chen (1975) and 7 = Ciesielski (1975). Lithologic descriptions are from Hayes and Frakes (1975) and Hambrey and Barrett (1993).

of these deposits (Anderson and Bartek, 1992). Moreover, sparse refraction seismic data, showing anomalous, high and irregularly trending velocities in early Miocene sequences, have been interpreted as indications for grounded ice reaching up to the continental shelf edge (Cooper et al., 1991a; Cochrane et al., 1995). The transition from generally aggrading seismic sequences (type II) to prograding sequences (type I) was interpreted as the onset of a large sized grounding ice sheet across the continental shelf already in the early Miocene (Cooper et al., 1991a).

Recent compilations of multichannel seismic data (Brancolini et al., 1995a,b; De Santis et al., 1995), integrated with high resolution seismic and drilling data, have depicted a different scenario for the early Miocene. At that time, DSDP 270 site was located on the flank of a structural high (Central High, Fig. 2), which most likely hosted valley glaciers or little terrestrial ice caps. The rest of the Ross Sea was characterised by deep structural basins (e.g., the Eastern Basin) filled with ice-distal glacio-marine sediments, with no clear evidence of grounding ice.

This scenario is consistent with evidence of a temperate glacial environment found in Eocene–Miocene sediments from terrestrial and other marine areas of the Ross Sea (Barrett, 1989; Jiang and Harwood, 1992; Cape Roberts Scientific Team, 1998).

In conclusion, different reconstructions for the evolution of the Eastern Ross Sea continental shelf can be roughly subdivided into two categories:

(a) The early fully glacial setting (Cooper et al. 1991a; Anderson and Bartek, 1992): development of an ice sheet during the late Oligocene and/or the early Miocene;

(b) The late fully glacial setting (Hinz and Block 1984; De Santis et al., 1995; Brancolini et al., 1995a,b): major development of a polar ice sheet during the late Miocene–early Pliocene.

Our contribution to this discussion consists of a quantitative analysis of a cross-section in the depocenter of the Eastern Ross Sea. Geometric patterns and morphologic profile variations of the continental shelf are reconstructed throughout the Cenozoic period.

2. Procedure and data

Starting from present-day cross-sections, it is possible to reconstruct the post-rift development and paleo-bathymetric profiles of a basin, with a four-step operation:

(a) Removing each sedimentary unit, starting with the youngest and progressively decompacting the lower ones (decompaction);

(b) Correcting the depth of each unit for the isostatic uplift after having removed the upper sediment and water load (isostatic compensation);

(c) Correcting for long-term, global, eustatic sea level changes;

(d) Recovering the post-rift thermal subsidence.

This procedure, which is repeated for each sequence so as to produce a reverse model of the basin, uses the ‘‘backstripping’’ technique (Steckler and Watts 1978; Kusznir et al., 1995; Roberts et al., 1998). The backstripping technique has been generally used to obtain information on one spot about the tectonic subsidence that affected different locations of a basin, where paleo-bathymetric and sediment properties information are known from well sites. The objective of our work is to obtain paleo-bathymetric 2-D profiles taking into account the subsidence history of the Eastern Ross Sea from known literature, and the paleodepth information from DSDP site 270.

Drill site 270 (Fig. 3) was particularly useful in constraining paleowater-depth, at least in one spot of the Eastern Basin, because it reached the basement and recovered Paleozoic metamorphic rocks (Ford and Barrett, 1975), overlain by: (1) a regolith layer interpreted as sub-aerial paleosoil (Hayes and Frakes, 1975); (2) shallow-water glauconitic sandstone (10-50 m bsl, Leckie and Webb, 1983) of late Oligocene age (26 Ma, McDougall, 1977); (3) an early Miocene (26–?21 Ma, Steinhauff and Webb, 1987) glacio-marine section that, based on foraminifera assemblage, has been deposited in a gradually deepening water environment (from subaerial to shallow water and finally to relatively deep water, at least to 300–500 m, Leckie and Webb, 1983).

The backstripping procedure has been applied to a cross-section, lying along the depocenter of the Eastern Basin and intersecting DSDP site 270, obtained from the multichannel seismic line BGR80-07 (Hinz and Block, 1984) (Fig. 4). The seismic line BGR80-07 has been depth-converted by using a set of average velocity functions from multichannel seismic velocity analysis. Backstripping of line BGR80-07 has been achieved by using the software (‘‘Flexural Decompaction’’), written by Kuznir N. and distributed by Badley Earth Sciences (Computing Limited). For the post-rift reverse reconstruction, this software assumes the uniform stretching model of McKenzie (1978).

2.1. Decompaction

One of the parameters used to calculate sediment decompaction is the sediment porosity versus depth. The glacio-marine sequence at DSDP site 270 has an average porosity of 0.4 (Barrett and Froggart, 1978). The same porosity values have been found in similar glacio-marine sediments, recently drilled at Cape Roberts-1 site (Cape Roberts Scientific Team, 1998). Subglacial, overcompacted diamictite forming late Miocene–Pleistocene topset beds at Prydz Bay ODP leg 119 have a porosity of ca. 0.18, less than 100 m below the sea-floor (bsf) (Solheim et al., 1991a). Such low porosity values at shallow depth imply a high compaction coefficient, interpreted as being

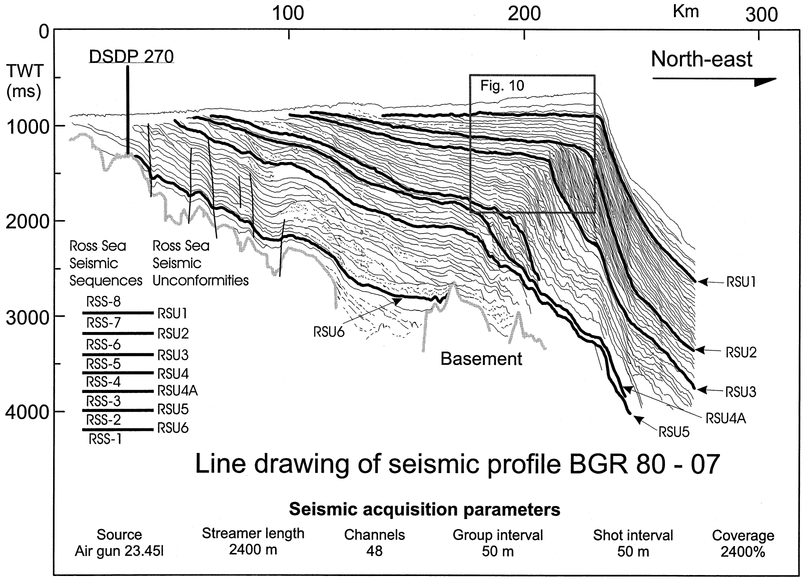


Fig. 4. Seismic profile BGR80-07 and depth converted line drawing (from Brancolini et al., 1997; Lovo, 1997). Unconformities depth and correlation with the drill sites are from ANTOSTRAT (1995). See Fig. 2 for track line location.

related to grounding ice and strong erosion (Solheim et al., 1991b).

The porosity values obtained from DSDP site 270 refer only to the lower part of the Miocene sedimentary cover. For the rest of the sedimentary section, and for a location nearer to the paleo-shelf edge, density/depth curves can be empirically derived from stack and refracted velocity data on line BGR80-07 (Fig. 5). For this purpose, several velocity profiles along the BGR80-07 line have been

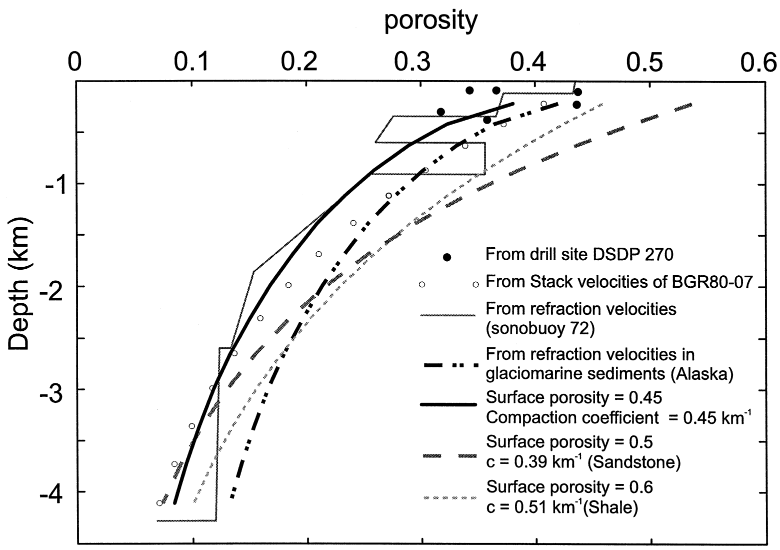


Fig. 5. Porosity–depth plot obtained from stack (this work) and refraction velocities (Cochrane et al., 1995). Porosity data from drill site DSDP 270 (Froggart and Barrett, 1978) are also plotted. Similar porosity/depth curve for glacial marine sediments in the Gulf of Alaska (Bruns and Carlson, 1987) and for sand and clay from the North Sea wells (Sclater and Christie, 1980) are also shown.

analysed. We adopted the relation between seismic velocity and rock bulk density (ρ_b) by Gardner et al. (1974) (see also Table 2-21 in Dobrin and Savit, 1988; with a power relation of 0.25). This relation is mainly based on laboratory measurements of samples. Porosity/depth curves can then be derived from density/depth curves.

Fig. 5 shows different experimental porosity/depth curves obtained from different velocity/depth functions. A porosity/depth curve obtained using our same empirical method for glacial marine sediments in the Gulf of Alaska (Bruns and Carlson, 1987) and one obtained from direct measurements in sediments (sand and clay) from North Sea wells (Sclater and Christie, 1980) are also shown. A curve with a surface porosity (φ_0) of 0.45 and a compaction coefficient (c) of 0.45 km^{-1} is obtained from stack and refraction velocities in the Eastern Basin. These values would suggest that no overcompacted sediments generally characterise these sequences. This curve matches with the porosity data from Barrett and Froggart (1978) for the most superficial sediments at DSDP site 270 quite well (Fig. 5).

Different values of surface porosity (φ_0) and compaction coefficient (c), keeping all the other parameters constant, were tested for all the sequences (Fig. 6). The results show that all the models are not significantly sensitive to the use of these

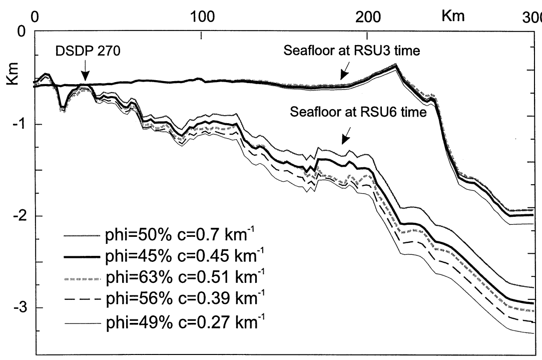


Fig. 6. Modelled sea-floor at RSU6 time (?30 Ma) and at RSU3 time (?10 Ma) using different values of surface porosity and compaction coefficient and keeping all the other factors constant: the stretching factor Beta = 2, the Effective Elastic Lithospheric Thickness $T_e = 0$ (local isostatic compensation), the rift age = 100. The values of surface porosity used are: 50%, 63%, 49%, 56%, 45%. The values of compaction coefficient used are: 0.7, 0.51, 0.27, 0.39, 0.45 km^1.

different values as input parameters, for the young sequences. In all cases, after backstripping and thermal reverse modelling, the paleo-sea-floor shows a seaward-deepening profile at the time of the Ross Sea Unconformities (RSU) RSU6, RSU5, RSU4A and RSU4 and a landward-deepening profile starting from the time of RSU3 (Fig. 6).

2.2. Isostatic compensation

The isostatic compensation allowed the correction of the paleodepth, taking into consideration the rebound caused by removing the overburied sedimentary and water load. This isostatic compensation depends on the strength of the lithosphere, and the thickness and weight of the removed sediments.

In the Ross Sea, high values of Lithospheric Effective Elastic thickness (T_e) of about 85 km, existent before the onset of rifting, have been calculated by Busetti et al. (this volume) using the concept of the level of necking the present crustal geometries, basement, and Moho depth. An early rift age of 100–85 Ma affecting the Ross Sea was also assumed in this model. Van der Beek et al. (1994) found values of T_e of 20 km in the Victoria Land basin at 60 Ma, while Stern and Ten Brink (1989) suggest a present average T_e of 19 ± 4 km for the entire Ross Embayment. However, they also noticed that these values were not very well-determined and there are probably strong lateral variations in rigidity within the embayment. Some of the scattering in values of T_e obtained by different authors can result from the use of different parameters and methods for its estimation (see a comprehensive discussion about this subject in Burov and Diament, 1995). A 25% of uncertainty is a quite realistic estimate for the accuracy of most data on T_e (Burov and Diament, 1995). McKenzie and Fairhead (1997) conclude that there is no evidence for T_e, in continental regions, being larger than 25 km (even in old shields), and that in areas of active extension a lower value is appropriate.

To evaluate different T_e effects on the backstripping, we made several tests on line BGR80-7. Fig. 7 illustrates the resulting sea-floor profile at the time of RSU6 (the oldest regional unconformity above the basement) for $T_e = 0$ (local compensation), 5, 30, 60, and 70 km respectively, assuming a rift age at 100–80

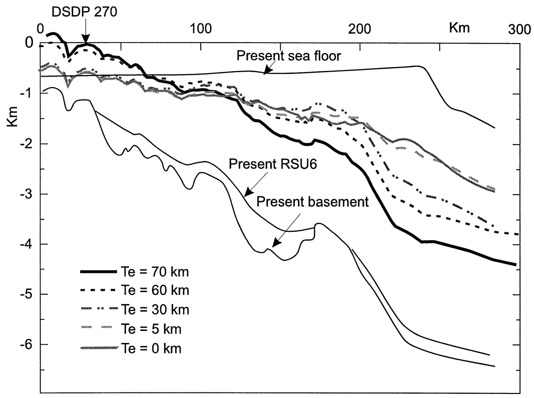


Fig. 7. The effect of isostatic response (assuming different values of Effective Elastic Lithospheric Thickness $T_e = 0$, 5, 30, 60, 70 km) to the removing of sedimentary load on shaping the RSU6 unconformity. The restoration was made by considering a rift age = 100 Ma and a post-rift age = 85 Ma; a Beta stretching factor = 2; surface porosity = 0.45 and compaction coefficient = 0.45 km^{-1}.

Ma and keeping all the other parameters constant. The long-term, global eustatic curve of Haq et al. (1987) was also considered. As a reference, we also plotted the present bathymetry, present RSU6 and present basement profiles. Differences between the various restorations showed an increase of flexural deflection with the lithosphere rigidity. The most evident effect of such a flexural deflection using different T_e values can be observed between 200 and 250 km from the left end of the backstripped section in Fig. 7, where a short wavelength change in sediment loading occurs across an abrupt shelf break. The morphology results are emphasized and the dip of the slope is much steeper in the cases of large T_e during backstripping.

Assuming a rift age at 100 Ma, the only test that allowed us to restore the sea-floor depth at sea level, at the time of RSU6, at DSDP site 270, was for $T_e = 70$ km. The same result was obtained in two other cases:

(1) For thinner T_e, but a younger rift age (e.g., $T_e = 55$ km and rift age = 60 Ma), keeping the other parameters constant. This case disagrees with previous studies that suggest a Mesozoic age for the Ross Sea rifting phase (see Section 2.3 below).

(2) Assuming that two rifting phases (each with Beta = 2) affected this area, at 100 and at 60 Ma. In this case, a thinned lithosphere ($T_e = 3$–5 km) was required. This last case is in disagreement with that by Busetti et al. (1999-this volume) that suggests that only one rifting phase affected the Eastern Ross Sea with Beta of 2–3 at 100–85 Ma.

Furthermore, drilling of strata below the RSU6 and modelling of present lithospheric thickness would provide direct validation of these different hypothesis. A significant result, using the existing data is that in all our tests, after backstripping and thermal reverse modelling, the paleo-sea-floor shows a seaward-deepening profile at the time of RSU6, RSU5, RSU4A and RSU4 and a landward-deepening profile starting from the time of RSU3.

2.3. Recovering the post-rift thermal subsidence: the rift age

There is indirect evidence of a Mesozoic Age, for the early opening phase of the Ross Sea. Marine magnetic anomalies in the oceanic crust, bordering the continental margin of the Ross Sea (Stock and Molnar, 1987; Grunov et al., 1991; Lawver et al., 1991, 1992; Lawver and Gahagan, 1994), date the initiation of sea-floor spreading in this sector and the break-up between the Eastern Ross Sea and its conjugate margin (the Campbell Plateau) at about 85 Ma. This age also corresponds to the onset of the spreading between Antarctica and Australia (Cande and Mutter, 1982; Lawver et al., 1992; Lawver and Gahagan, 1994), followed by the spreading between Antarctica and New Zealand at 72 Ma (Stock and Molnar, 1987) and the break-up between the Western Ross Sea and Tasmania dated at 40 Ma (Lawver et al., 1992). Paleomagnetic and geological observations made on Marie Bird Land (Di Venere et al., 1994; Lyendick et al., 1996) confirm that a wide stretching occurred between east and west Antarctica as well as within the west Antarctic microplates after 100 Ma.

Geophysical investigations indicate that the structural configuration of the Ross Sea is characterised by rifted basins separated by N–S trending ridges (Fig.2) (Hinz and Block 1984; Cooper and Davey, 1987; Cooper et al., 1991a,b; ANTOSTRAT, 1995). The lowest part of the Ross Sea sedimentary sequence, affected by major downfaulting of a mainly

N–S and NW–SE direction, is tilted and truncated by a regional unconformity (U6 of Hinz and Block, 1984 or RSU6 of ANTOSTRAT, 1995). Based on CIROS-1 drill and DSDP site 270, a late Oligocene age has been attributed to this unconformity (Brancolini et al., 1995a,b; De Santis et al., 1995). However, the age of the sediments below the RSU6, deposited during the tectonic phases that caused the opening of the Ross Sea basins, is still unknown. High acoustic velocity shown by strata below the regional unconformity RSU6 has been interpreted as consistent with lithified rocks probably of Mesozoic age (Cooper et al., 1991b), in agreement with the rift onset at 100 Ma. Alternatively, a younger than 100 Ma rift age should be considered for the opening of the Eastern Basin. Cenozoic tectonic activity widely affected the Ross Sea sequences in the Western and Central Ross Sea (Cooper and Davey, 1987; Salvini et al., 1997; Busetti et al., 1999-this volume). This phase was characterised by intense alkaline magmatism and a significant uplift of the TAM (around 60 Ma) (Fitzgerald, 1992), linked to major episodes of plate motion, direction and spreading rate change in the Southern Ocean (Weissel and Hayes 1977; Cande and Mutter 1982; Stock and Molnar 1987; Lawver et al., 1992; Salvini et al., 1997). On the basis of these observations and of our results, we believe that a Cenozoic rather than a Mesozoic major extension in the Eastern Basin should be considered. However, further drilling of strata below RSU6 are needed for establishing the rift age in the Eastern Ross Sea and to confirm new and previous hypotheses.

2.4. Recovering the post-rift thermal subsidence: the lithospheric stretching

The total lithospheric stretching produced by the Ross Sea rifting has been calculated by Busetti et al. (this volume), who suggested an average value of the Beta factor of ca. 2.5; values of Beta = 2 and 3 have been calculated in the Central High and Eastern Basin, respectively. An extended crust of 100% in the Ross Embayment was also suggested by Lawver and Scotese (1987) and Cooper et al. (1991b).

In our model, we assumed an average value of the Beta factor, because the software that we used accepts only one Beta value for the whole length of the section. Different tests were run, varying the Beta value equal to 1 (no stretching), 2 (stretching of 100%) and 3 (stretching of 200%), keeping the other values constant, to evaluate the difference between the Central High where Beta is less than 2, and the Eastern Basin where Beta is about 3.

In all the cases, after backstripping and thermal reverse modelling, the paleo-sea-floor shows a seaward-deepening profile at the times of RSU6, RSU5, RSU4A and RSU4 and a landward-deepening profile starting from the time of RSU3.

3. Seismic sequences configuration

Here, we summarise the main geometric features of the seismic sequences in the Eastern Basin on lines along the depocenter of the basin (e.g., line BGR80-07, Fig. 4). Criteria to distinguish Ross Sea Seismic Sequences 1 to 8 (from RSS-1 to RSS-8) have been described in detail in ANTOSTRAT 1995, the age of sequences is illustrated in Fig. 3. Fig. 8A–H illustrates a series of reconstructed sections showing the proposed evolution of the sea-floor profile and internal reflector geometry across the depocenter of the Eastern Basin. The reconstructed sections illustrate an interval spanning from RSU6 times to the present and were obtained from backstripping and thermal reverse modelling of BGR80-07, as discussed above, and assuming: a constant stretching Beta factor = 2 along the length of the line, a Lithospheric Effective, Elastic Thickness T_e = 70 km with rift age = 100–85 Ma, an initial sediment porosity of 0.45 and a compaction coefficient of 0.45 km^{-1}. The error in the paleowater-depth estimate for the sea-floor from 26 to 21 Ma can be evaluated on the basis of information at DSDP site 270. The error from 21 Ma to present cannot be quantified due to the erosion of the younger than 21 Ma section in site DSDP 270 and to the lack of trustworthy, bathymetric proxies in the other sites. During this period, we assumed that the tectonic deepening of the basement was gradual because there are no proofs of younger structures that would have caused further subsidence and/or uplift.

Fig. 8a shows the proposed reconstruction for the RSU6 time (> 26 Ma, ?30 Ma) at the end of the deposition of the RSS-1. RSU6 is a continuous reflector that onlaps on structural highs. At RSU6 times, a continental shelf edge was not clearly developed and the sea-floor deepened seaward. The sequence RSS-1 is characterised by sub-parallel and sub-horizontal strata that progressively fill small (tens of kilometers) asymmetrical basins. Locally, strata of RSS-1, are inclined, diverging along growth faults, and are truncated by the upper sequence boundary RSU6. One example of the U-shaped channels described by Brancolini et al. (1995a,b) and De Santis et al. (1995) is shown on the left of the section BGR80-07 (Fig. 8a), SW to DSDP site 270. These channels cut into the shallow sector of the Central and Coulman Highs, and are around 5 km wide, some tens of kilometers long, and have a relief of about 200–500 m. They were interpreted as being associated with glacial erosion from tidewater glaciers, although a fluvial origin cannot be discarded (De Santis et al., 1995).

Fig. 8b shows the proposed reconstruction for the RSU5 time (ca. 21 Ma) at the end of the deposition of the RSS-2. There is no stratigraphic information for the most inland part (ca. 50 km on the left side of the figure, dashed line) of the study section because of erosion occurred during and after RSU5. Even considering the subsidence load caused by the uppermost sediments that were present at RSU5 times, now removed by erosion, our model suggests that in this most inland part of the shelf, the sea-floor profile was ca. 100–200 m deep. Its landward inclination may be related to more recent erosion. The sea-floor on the rest of the shelf is deepening seaward and a continental shelf edge was not clearly delineated yet.

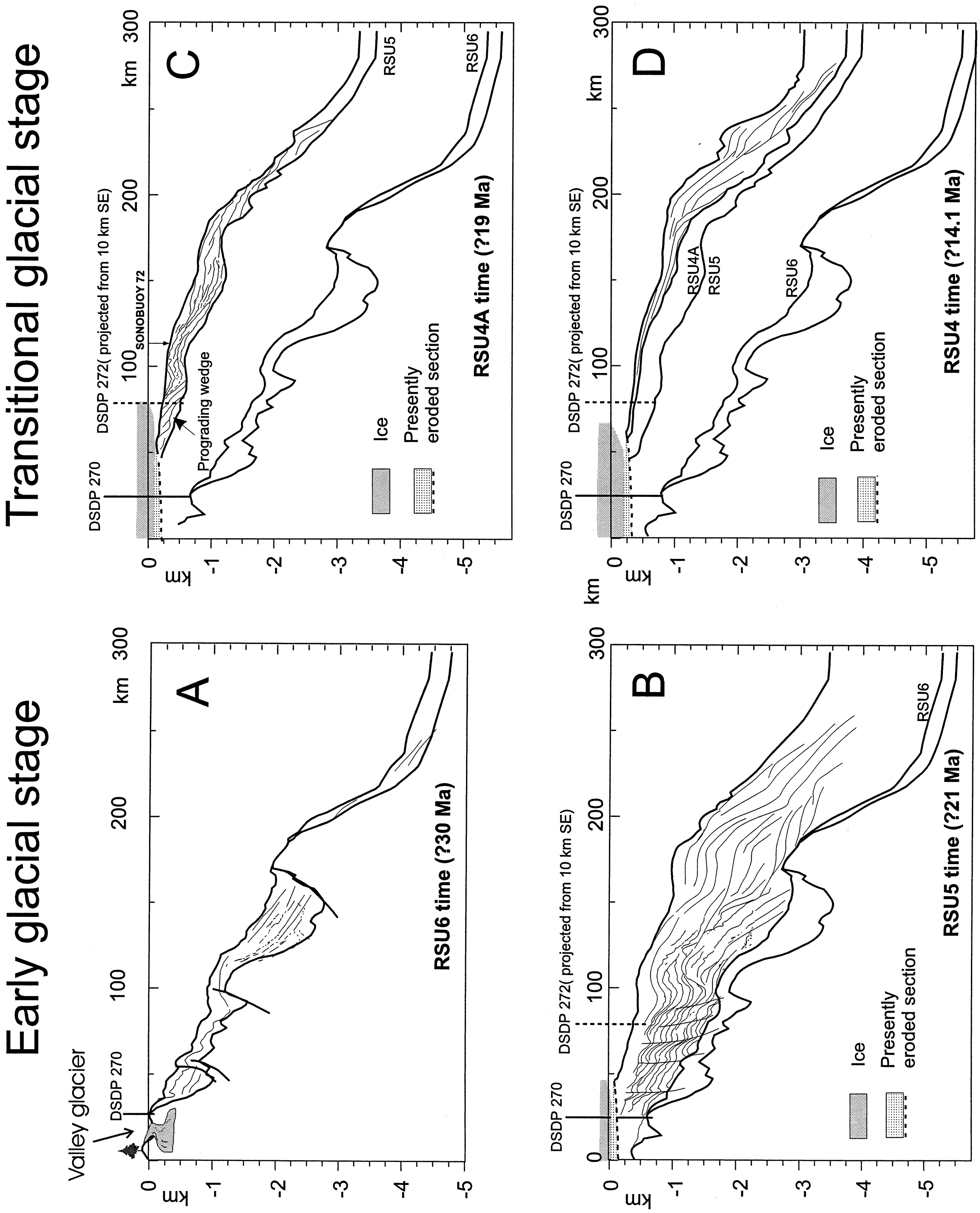
The internal configuration of this sequence is characterised by wedges (up to 300-m thick) of strata, mainly confined to the westernmost part of the basin near the Central High, consisting of gently eastward inclined strata and downlapping onto RSU6 (Fig. 8b). These strata appear folded locally and displaced by extensional faults with offset of tens of meters (Fig. 8b). This stratified unit is interfingered seaward with internally chaotic lenses (few tens of kilometers in length, up to 200 m thick) and internally stratified mounds lying on RSU6 (Fig. 8b). The uppermost part of the sequence RSS-2 is characterised by a large prograding and aggrading wedge with less evidence of extensional tectonics than in the upper part (Fig. 8b). According to sequence stratigraphic geometric patterns by Van Wagooner et al (1988), the seismic configuration of the uppermost unit of RSS-2 might resemble that of a high stand wedge, deposited above a low stand and a transgression system, and truncated by a major erosional surface (the RSU5 sequence boundary). The lithology of the lower Miocene section at DSDP site 270 may represent this kind of trend, with ice distal glacio-marine units (units 2A–E of Hayes and Frakes, 1975; Hambrey and Barrett, 1993; De Santis and Barrett, in prep.) deposited at a gradually deeper water-depth and ice distal setting (from 50 to 500 m bsl, Leckie and Webb, 1983) unconformably overlying a subglacial to ice proximal unit (units 2J–F of Hambrey and Barrett, 1993).

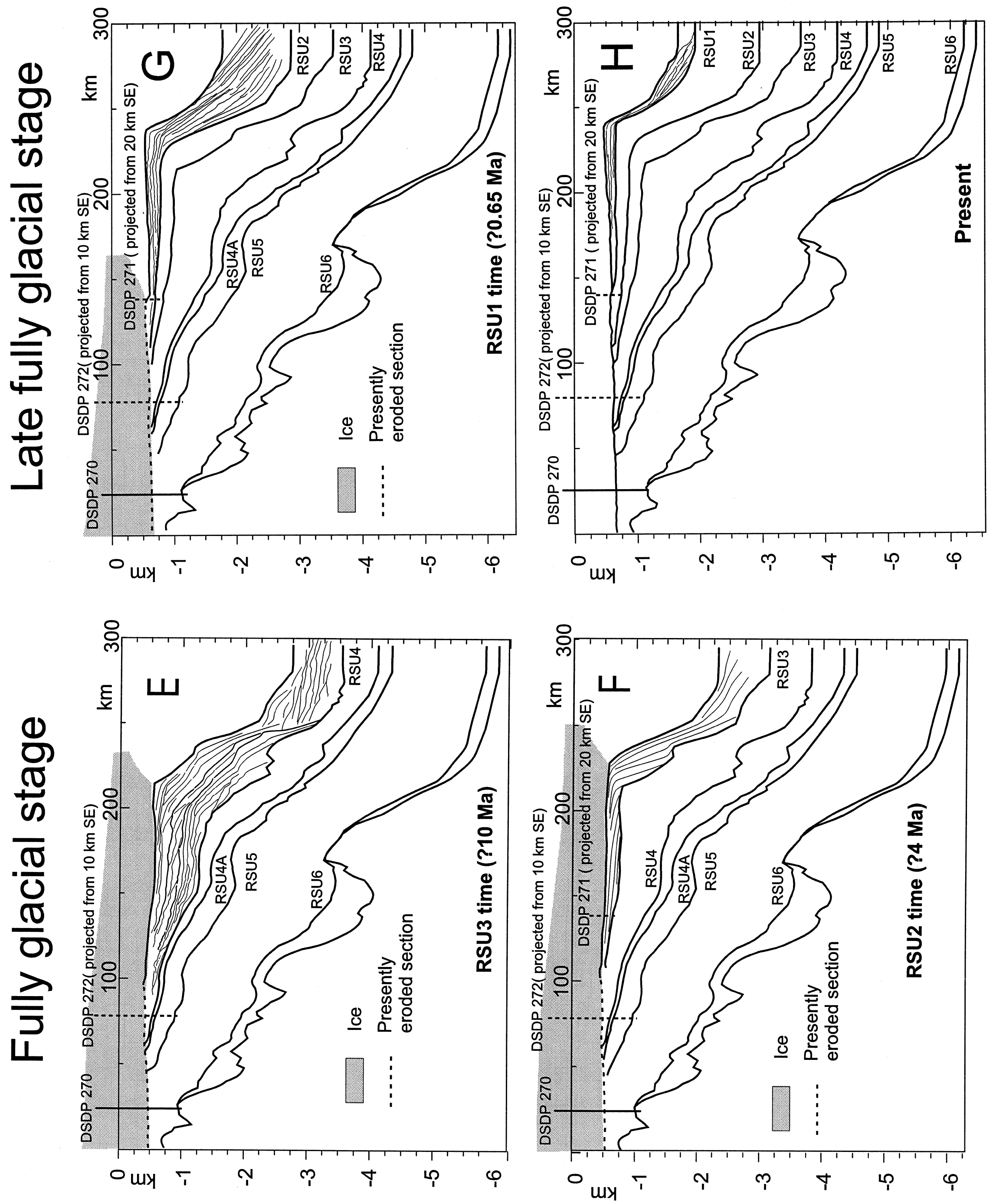
Fig. 8c shows the proposed reconstruction for the RSU4A time at the end of the deposition of the RSS-3. RSU4A is missing in the most inland part (ca. 50 km on the left side of the figure) because of the more recent erosion (dashed line). The landward inclination of the sea-floor in this part may be related to more recent events.

From about 50 to 100 km from the left end of the figure, the profile gradient appears to be gradually seaward-deepening, and the water-depth is ca. 200 m. Seaward, it is gradually steeper and deeper with a slope break at ca 200 km. RSU4A is a low-angle truncational surface with no clear evidence of large erosional features, typical of glacial troughs. RSU4A has no stratigraphical information. It must be younger than RSU5, and older than the base of DSDP site 272 (19.23–17.5–16 Ma depending on the interpretations, see Fig. 3).

One of the most evident features of the RSS-3 internal geometry is a wedge, about 300 m thick, between 50 and 100 km from the left end of the profile. (Fig. 8c), consisting of prograding foresets, downlapping on RSU5. This feature is onlapped by

Fig. 8. (A–H) Evolution of the sea-floor profile configuration (from ?30 Ma to the present) from modelling line BGR80-07 that runs across the depocenter of the Eastern Basin. Geometry of the internal strata is also qualitatively shown on the basis of the present relationship among strata dip, strata termination and sequence boundaries. Ice thickness is conjectural and not on scale. See text for explanation.


Early glacial stage
Transitional glacial stage
A
B
C
D
Valley glacier
DSDP 270
DSDP 272(projected from 10 km SE)
SONOBUOY 72
Prograding wedge
Ice
Presently eroded section
RSU4A
RSU5
RSU6
RSU6 time (?30 Ma)
RSU5 time (?21 Ma)
RSU4A time (?19 Ma)
RSU4 time (?14.1 Ma)
km
0
100
200
300
-1
-2
-3
-4
-5


Fully glacial stage
Late fully glacial stage
E
F
G
H
km
DSDP 270
DSDP 272(projected from 10 km SE)
DSDP 271 (projected from 20 km SE)
Ice
Presently eroded section
RSU3 time (?10 Ma)
RSU2 time (?4 Ma)
RSU1 time (?0.65 Ma)
Present
RSU1
RSU2
RSU3
RSU4
RSU4A
RSU5
RSU6

about 400 m of sub-parallel strata that are interfingered with mound-shaped deposits where the seafloor became steeper.

Fig. 8d shows the proposed reconstruction for the RSU4 time at the end of the deposition of the RSS-4. The possible age of RSU4 in the Eastern Basin (ca. 14.1 Ma) has been discussed by De Santis et al. (1995) and based on stratigraphic information from DSDP site 272 (Fig. 3). Due to the lack of knowledge about faunal associations at high latitude, confident paleowater-depth and environmental information from the drilled sediments is not available for the RSU4 time. Our model shows that the RSU4 sea-floor profile gradient is deepening seaward from ca. 250 to 700 m.

RSS-4 is more developed on the outer shelf than RSS-3, where it forms a fan made of prograding wedges. These wedges also have an aggrading component with toplap terminations preserved. There is no clear evidence of erosion, typical of a glacial trough, on seismic lines across the Eastern basin, like those described by Anderson and Bartek (1992) and De Santis et al. (1995) in the Central Ross Sea in the same time interval. The only RSS-4 stratigraphic record recovered at drill site DSDP 272 is made of ice-distal glacio-marine sediments deposited along an inclined surface (Hayes and Frakes, 1975; Hambrey and Barrett, 1993; De Santis et al., 1995).

Fig. 8e shows the proposed reconstruction for the RSU3 time at the end of the deposition of the RSS-5. This is the first unconformity that shows a sub-horizontal to gently landward-deepening sea-floor profile after backstripping and thermal reverse modelling of line BGR80-07. The average water-depth is ca. 500

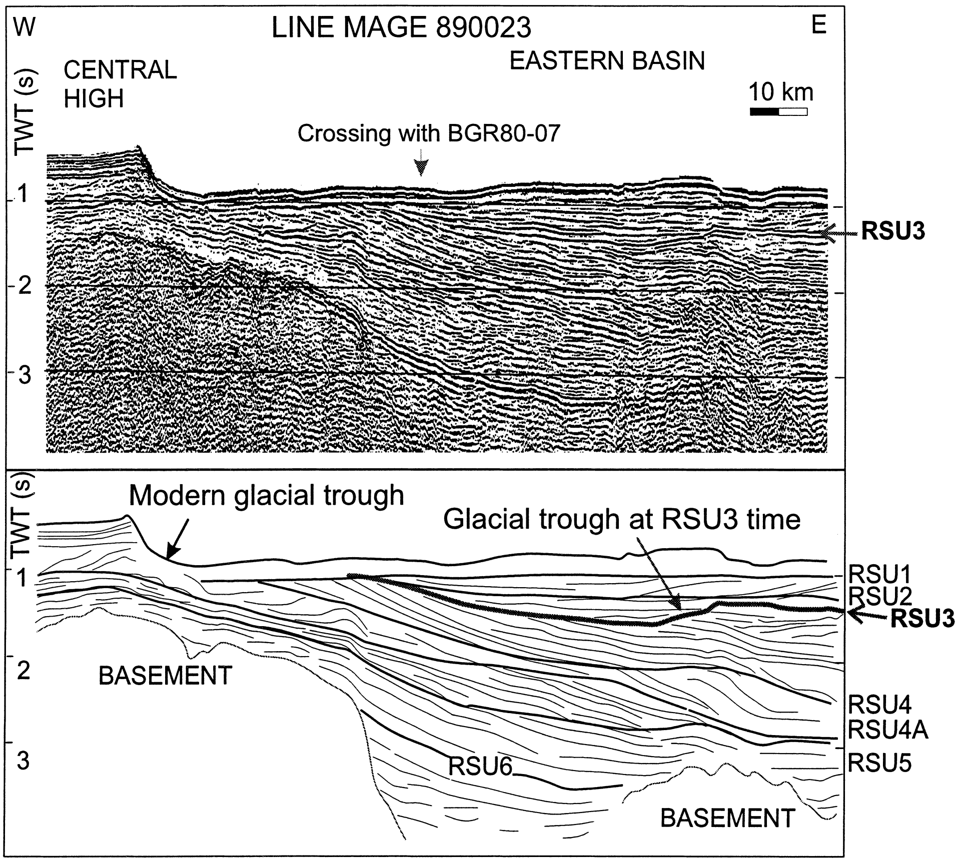


Fig. 9. East–west seismic line Mage 890023 showing a large glacial trough eroded at RSU3 time in the Eastern Basin (From De Santis et al., 1995). See Fig. 2 for track line location.

m up to the shelf edge. This configuration is confirmed by all tests performed using several realistic input parameters (e.g., Figs. 6 and 7). These water-depth values do not take into account the further sea-floor deepening that could have been caused by the load of a large and thick ice sheet, likely grounding on most of the shelf at the time of RSU3.

RSU3 is also an erosional surface with an irregular relief (ca. 400 m, assuming a velocity of 2.5 km/s) measured on strike lines in the Eastern Basin (Fig. 9), showing typical-size erosion associated with an ice stream glacial trough.

RSS-5 shows peculiar geometries compared to the older sequences. One of the most evident characteristics of RSS-5 are the large shelf margin fans, developed in the upper part of the sequence (Fig. 10), totally prograding the shelf edge by several tens of kilometers. These shelf margin fans pinch out toward the middle–inner shelf. Below the shelf margin fans, a unit composed of foresets prograding from the inner to the outer continental shelf was observed, downlapping onto sub-horizontal reflectors (Fig. 8e). Seismic profiles collected from the continental rise show that large, mound-shaped sedimentary bodies have been accumulated during the deposition of RSS-5 (Fig. 11) at the base of the continental slope. These bodies are stratified and eroded by truncational surfaces. The age of RSU3 is still unknown. It is comprised in a hiatus spanning from 13.8–3.7 Ma (Fig. 3).

Fig. 8f shows the proposed reconstruction for the RSU2 time at the end of the deposition of the RSS-6. The sea-floor profile clearly shows a sub-horizontal to landward-deepening gradient, from ca. 600 m on the inner shelf to less than 500 m at the shelf edge. These water-depth values do not take into account further sea-floor deepening that could have been caused by the load of a large and thick ice sheet grounding on the shelf, up to its outer edge, at the time of RSU2.

On the inner and middle continental shelf of the Eastern Basin, RSS-6 is very thin and is charac-

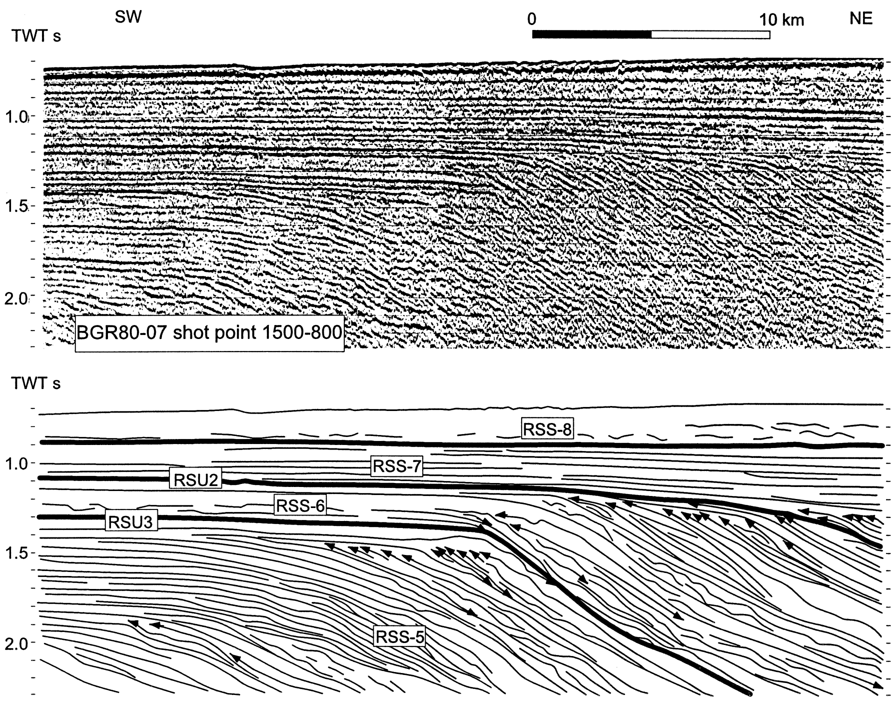


Fig. 10. Enlargement of line BGR80-07 showing sequences RSS-5, 6, 7, and 8, and their boundaries. RSS-5 shows a generally aggrading pattern of seismic reflectors. A prograding wedge, sharply truncated at the top by RSU3, clearly develops in a later stage of RSS-5 only. A prograding wedge is also well-developed within RSS-6 deposition at the continental shelf edge. (From De Santis et al., 1997b). See Fig. 4 for track line location.

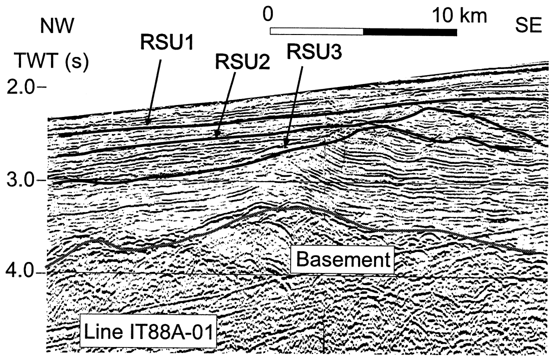


Fig. 11. Seismic line IT88A-01 from the ocean basin floor seaward of the Eastern Basin. See Fig. 2 for track line location.

terised by sub-horizontal or gently inclined reflectors, downlapping onto RSU3 from the inner to the outer shelf. Within RSS-6, large shelf margin fans developed at the shelf edge, sharply truncated by erosional unconformities (Fig. 10). Some offlaps of the truncational surface are preserved within topsets and allow us to distinguish different continental shelf edges. The youngest foresets are sharply truncated by RSU2 (Fig. 10). Total shelf edge progradation during the deposition of RSS-6 is estimated to be of about 50 km.

The age of RSU2 is still unknown. Correlation to DSDP site 271 and indirect correlation to DSDP site 273 in the Central Ross Sea both suggest an age of ca. 3–4 Ma (Fig. 3) (ANTOSTRAT, 1995; De Santis et al., 1995).

Fig. 8g shows the proposed reconstruction for RSU1 times (0.65 Ma, Hayes and Frakes, 1975) at the end of the deposition of the RSS-7. The sea-floor profile shows a gentle landward-deepening and it is very similar to the present configuration (Fig. 8h).

RSS-7, the sequence below RSU1, is thicker at the shelf edge, gradually thinning landward. In the Eastern Ross Sea, RSS-7 is an aggrading sequence made of sub-horizontal reflectors, onlapping onto RSU2. RSS-8 is the youngest sequence recognised on the MCS seismic data set, between RSU1 and the present sea-floor and is mainly aggradational and mostly restricted to the continental shelf.

No shelf margin fan developed (at least in the depocentre areas) and the shelf edge of the Eastern Basin did not prograde since the time of RSU2 (Fig. 8g and h) (Alonso et al., 1992; Anderson and Bartek, 1992). Backstepping asymmetric banks, up to 100 m high, were observed at the top of the most recent sequence (RSS-8) in the Eastern Basin (e.g., in Fig. 12) particularly in the depocenter of glacial troughs. On high resolution seismic data, these banks appear to be composed of transparent lenses (that thickens seaward) lying above thin wedges of chaotic to stratified facies with prograding foresets (Fig. 12).

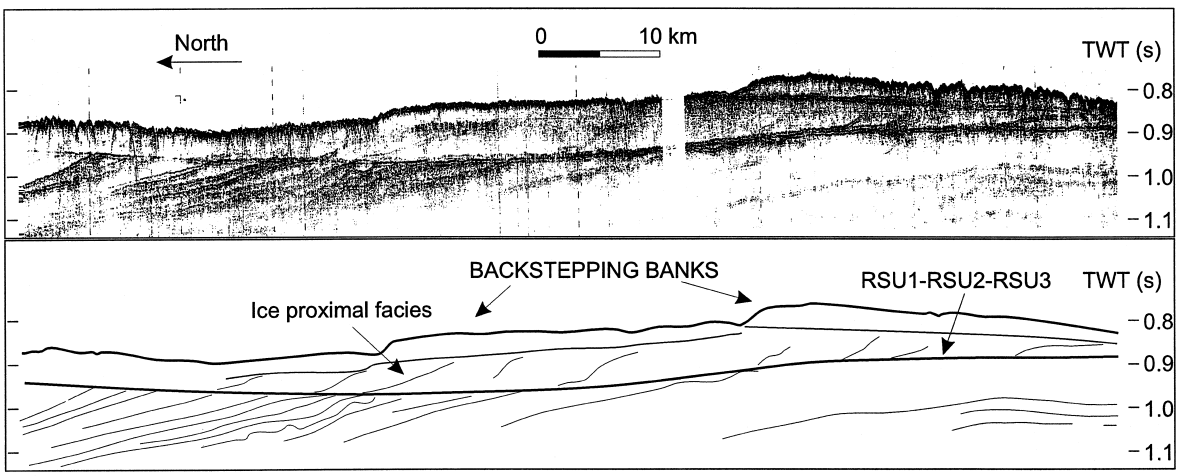


Fig. 12. High-resolution (Sparker) line Mage 890027 showing a prograding wedge, less than 5 km north of DSDP site 272 in RSS-5. Backstepping ridges within the most recent sequence RSS-8 are also shown. See Fig. 2 for track line location.

In the Eastern Ross Sea, at the mouth of a glacial trough shown by the RSU1 structural map (ANTOSTRAT, 1995), roughly corresponding to the present location of the extension of ice stream ‘‘C’’ (Anderson et al., 1992), where the continental slope is steep, slump structures occur and erosional scours are frequent at its base. Apparently, during the deposition of RSS-8, the rise and slope are sediment-starved, compared to the deposition occurring during the older sequences (Fig. 11).

4. Discussion

Four different stages can be recognised in the evolution of the Eastern Basin: (1) an early glacial stage (pre-Miocene–early Miocene time), when tidewater glaciers and small ice caps developed, while the basin was still subsiding and was influenced by marine sedimentation; (2) a glacial transitional stage (early–middle Miocene time), during which the basin was gradually filled with sediments and ice caps expanded onto the continental shelf; (3) a fully glacial stage (late Miocene–Pliocene time) during which a large and thick ice sheet deeply shaped the continental shelf up to the shelf edge; (4) a late fully glacial stage (Pliocene–present time) during which the continental shelf was clearly overdeepened and the sedimentation was reduced and was mainly influenced by glacial processes.

4.1. The early glacial stage (pre-Miocene–early Miocene time) of the Eastern Ross Sea continental shelf evolution (RSS-1, RSS-2)

The geometry and spatial distribution of RSS-1 into narrow, deep and still subsiding basins suggests that it was deposited during and immediately after the rift stage that opened the Ross Sea. Growth faults suggest active extension during the deposition of RSS-1.

U-shape valleys in RSS-1 cutting into the Central High and the Coulman High, were observed by Brancolini et al. (1995a) and De Santis et al. (1995). These features would suggest that tidewater glaciers likely developed onto sub-aerial and shallow water sectors, similar to those along the coast of the Victoria Land (Barrett, 1989; Cape Robert Scientific Team, 1998). The overall stratified internal geometry of RSS-1, with no evidence of erosional features, a large water-depth, and a seaward-deepening trend shown by the sea-floor profile, would suggest a non-glacial origin for most of the sequence.

Further stratigraphic ground truth is needed to understand the origin of the strong reflection and low angle truncational surface formed at the time of RSU6, although many hypotheses have been suggested. Hinz and Block (1984) believe that RSU6 was related to reworking and erosion processes due to strong currents, after the opening of the Drake Passage around 29 Ma. Anderson and Bartek (1992) suggested subglacial related processes as a possible explanation for the origin of RSU6. Reworking and overcompaction of diamictite layers under grounded ice-sheets could cause a large reflection coefficient, however, a similar effect can also be originated by a large discharge of ice-rafted debris above fine-grained sediments for instance, or by indurate, condensed surface formed in a starved basin above fine-grained sediments. We believe that there is enough evidence to discard a subglacial origin for RSU6 in the Eastern Basin, since it is quite unlikely that a large-sized ice sheet was grounding on the sea-floor to ca. 2000 m bsl without leaving any sign of erosion and deposition.

The RSS-2 is truncated by RSU5. Some authors suggested that such a truncation and the prograding wedges in RSS-2 might have been caused by subglacial erosion and deposition by a large and stable polar ice sheet grounding onto the continental shelf up to the shelf edge. Seismic facies analyses conducted on high resolution data demonstrated the existence of ice-proximal features into RSS-2, near the eastern flank of the Central High (De Santis et al., 1995). The vertical facies association observed at DSDP site 270 from ice proximal to ice distal, represents a glacio-marine record deposited near a fluctuating ice sheet, formed on the Central High, but the basins were filled with ice distal and marine deposits (De Santis et al., 1995). The rest of the Eastern Ross Sea was most likely influenced more by marine processes. The chaotic and mound-shaped units, that lie seaward of the stratified wedge of RSS-2 (Fig. 8b), probably represent deep-water, gravity flow deposits accumulated onto the lower continental slope. The extensional structures affecting RSS-2 strata on seismic sections, and the subsidence occurring at site 270 in the early Miocene (Leckie and Webb, 1983) justify such a slope instability and the occurrence of gravity flow processes along the eastern flank of the Central High. The gravity flow process was probably enhanced during sea level low standing and/or ice expansion episodes onto the shallower, inner parts of the continental shelf.

No major subglacial erosional features like those typically produced by grounding ice stream are observed and no overdeepened, reverse profile is shown by the sea-floor at the time of RSU5.

Therefore, these observations would present argumentation against a major and long-lasting ice sheet advance onto the outer continental shelf at the end of the time of RSS-2 (Fig. 8b). A subglacial origin for high amplitude reflectors like RSU6 and RSU5 can be accepted only for areas on and near the structural highs in likely subaerial or shallow water conditions. Sparse ice caps, likely land-based could have occurred onto structural highs on the Ross Sea continental shelf. This hypothesis is in agreement with stratigraphic evidence for a glacial temperate environment observed in late Oligocene–early Miocene sediments in other areas of the Ross Sea as well (Hayes and Frakes, 1975; Barrett, 1989; Cape Roberts Scientific Team, 1998). It is also consistent with reworked Miocene palynomorphs found in Quaternary sediments derived from interior West Antarctica (Truswell and Anderson, 1984; Truswell and Drewry, 1984).

To conclude, we suggest that the eastern sector of the Ross Sea was underlain by a subsiding basin bordered to the west by the Central High ridge that outcropped the sea level until late Oligocene time. The exposed ridge, as well as the TAM, hosted valley glaciers that calved at sea. In early Miocene times, the combined effects of eustatic changes and tectonic subsidence produced a relative, general rise in sea level with the onset of open marine conditions over most of the area. During the Miocene, the depositional environment was generally glacio-marine, although some sea level drops might have caused sporadic exposure of large portions of the continental shelf. During these periods, ice caps nucleated on the emerged and shallow-water area and influenced the depositional environment of the continental margin.

4.2. The glacial transitional stage (early–middle Miocene time) of the Eastern Ross Sea continental shelf evolution (RSS-3, RSS-4)

Cochrane et al. (1995) observed high and widely fluctuating refraction velocity values on top of RSS-3 and identified RSU4A as the first unconformity which has been clearly subglacially eroded by a large ice sheet over most of the continental shelf. The abrupt velocity fluctuations occurring above RSU4A are interpreted as caused by erosion, deposition and loading processes associated with intermittently grounded ice sheets. Cochrane et al. (1995) data are located in the inner–middle parts of the paleo-continental shelf at the time of RSU4A. One of these refraction velocity record sites, plotted on line BGR80-07, is located just seaward of a wedge observed in RSS-3 (sonobuoy 72 in Fig. 8c), consisting of prograding foresets. We interpret this feature as a glacial delta (Powell and Alley, 1997), formed below and near grounding-ice. The seaward-deepening sea-floor profile suggests that marine rather than subglacial processes dominated the outer continental shelf depositional environment during the RSS-3. The lack of prominent glacial valley erosion suggests that ice streams like the present ones were not developed in RSS-3 times in this sector of the Ross Sea. The seaward thinning of RSS-3 and the lack of thick sedimentary accumulation at the base of the slope, would also argue against a major glacial expansion onto the continental shelf. Unconformity RSU4A is onlapped by sub-parallel strata showing a generally aggrading trend. A shelf margin wedge developed at the end of RSS-4, truncated by RSU4 (Fig. 8d). The RSU4 sea-floor, similarly to RSU4A, does not show an overdeepened reverse profile and no major glacial troughs are observed on seismic lines across the Eastern Basin. Therefore, we believe that in the early–middle Miocene, the continental shelf in the Eastern Basin was occasionally and partially exposed above sea level, but no stable ice stream developed. The shelf margin fan prograding wedge was probably originated by marine and glacio-marine process rather than subglacial. The location of the grounding line in Fig. 8c and d is consistent with the ice distal depositional environment of sediments found at DSDP site 272 located ca. 10 km southeast of line BGR80-07, whose age ranges from ?19.23–?17.5 to 14 Ma (Fig. 3).

To conclude the scenario that we propose for the transitional stage of the Eastern Ross Sea evolution is that of a continental shelf still characterised by basins that were gradually filled with sediments. Most of the ice erosion and deposition still occurred in the inner shelf near the structural high. Only occasionally were ice caps able to grow and reach the outer continental shelf.

4.3. The fully glacial stage (late Miocene–Pliocene time) of the Eastern Ross Sea continental shelf evolution (RSS-5, RSS-6)

RSU4 is onlapped by sub-horizontal to gently seaward inclined strata that show an aggrading pattern (Fig. 10). This unit has been interpreted by Anderson and Bartek (1992) as deposited during conditions of high sea level. A correlation between acoustic facies and DSDP site 272 lithologic units (De Santis et al., 1997a) shows that the lower RSS-5 is characterised by sub-parallel, stratified reflectors, consisting of ice distal glacio-marine sediments. A prograding wedge made of ice proximal sediments downlap onto this stratified unit in RSS-5 at site DSDP 272 (De Santis et al., 1997a) suggesting an ice advance. Large shelf margin fans developed in the upper part of the sequence RSS-5 and large amounts of sediment were also discharged onto the continental slope and rise (Fig. 11) and accumulated in mound-shaped deposits. Similar sediment bodies were recognized in other parts of the Antarctic margin, interpreted as indicative of major glacial sediment transport and discharge to the continental margin (Escutia et al., 1997; Rebesco et al., 1997). RSU3, the upper boundary of RSS-5, shows clear evidence of glacial trough incision from the ice stream across the Eastern Basin continental shelf (Fig. 9).

Backstripped section BGR80-07 shows that at the end of RSS-5 the sea-floor has an overdeepened, sub-horizontal to gently reverse profile for the first time in the evolution of the continental shelf architecture in the Eastern Basin (Fig. 8e). This fact demonstrates that at that time, the advance of large and relatively stable ice streams across the continental shelf occurred.

A possible age for RSU3 could be 10 Ma. At this time, a major eustatic fall (Haq et al., 1987), following a significant cooling period (Miller et al. 1991), occurred. However, until stratigraphic data are collected, these are clearly speculative correlations.

The seismic sequence RSS-6 deposited above RSU3 has the typical characteristics of glaciated margin sequences, dominated by a large shelf margin fan, truncated by erosional surfaces (Fig. 10). The preservation of some offlap terminations below the erosional surfaces, bounding prograding foreset wedges, demonstrates that RSS-6 is the product of several ice advances towards the shelf edge, during which time the ice streams could have migrated laterally and/or have expanded differentially onto the continental shelf. Alonso et al. (1992) demonstrated the patchy distribution of several sub-sequences into RSS-6 using high resolution data. The preservation of offlap terminations suggests that deposition rather than erosion is the dominant process near the continental shelf edge. The accumulation of glacial sediment during several ice-advancing cycles enhanced the landward dip of the sea-floor profile, causing the shallowing of the outer continental shelf and the overdeepening of the inner sector.

To conclude, the scenario that we propose for the fully glacial stage of the Eastern Ross Sea evolution is that of a wide continental shelf that was covered and deeply shaped by a thick and large ice sheet. Backstripped, depth-sections show that the overdeepened profile of the sea-floor, typical of glaciated margins, was finally established in the Eastern Basin at the time of RSU3 (most likely in the late Miocene). Ice streams from WAIS developed then, they cut glacial troughs across the sea-floor up to the shelf edge, and discharged large amounts of subglacial material in trough-mouth fans and sedimentary drifts.

4.4. The late fully glacial stage (Pliocene–present time) of the Eastern Ross Sea evolution (RSS-7 and RSS-8)

Paleogeography and geometry at the shelf margin changed dramatically in the Eastern Ross Sea after RSU2. The continental shelf margin did not prograde long after RSU2 and only thin aggrading sequences

(RSS-7 and RSS-8) have been observed, mainly confined to the outer continental shelf. On high resolution data, these sequences appear to be mainly characterised by subglacial and ice-proximal facies (Alonso et al., 1992; Brancolini et al., 1995a,b; De Santis et al., 1995; Andersen, 1997; Bartek et al., 1997a,b; Shipp and Anderson, 1997a,b,c).

Bart (1997) showed that in the northwestern Ross Sea, no sediments have been deposited onto RSU2. Where this unconformity is exposed at the sea-floor (e.g., near the shelf edge of Joides Basin), piston cores sampled residual glacial marine deposits, implying lags associated with long-term exposure to bottom current (Anderson and Smith, 1989). The high amplitude and continuity of RSU2 in the basinal areas and the sedimentary thickness above RSU2 suggests that a possible long-term sediment starvation occurred at this time (Bart, 1997).

One explanation for the origin of backstepping banks observed on the continental shelf in the Ross Sea (Fig. 12), like in several areas of the Antarctic margin, (e.g., the Western Ross Sea, Bart, 1997; Weddell Sea, Anderson, 1997; Antarctic Peninsula, Vanneste and Larter, 1995; Prydz Bay, O'Brien et al., 1999), can be that they represent grounding line still-standing positions during the last ice sheet retreat from the outer continental shelf. Chronostratigraphic data from the continental margin reveal that, during the last glacial maximum, the ice sheet did not reached the continental shelf on many occasions (Licht et al., 1996; Domack et al., 1998). An increase in the frequency of eustatic cycles and climate changes, forcing the ice sheet to fluctuate, has been considered one of the possible causes for partial advance onto the continental shelf, low sediment transport and, consequently, sediment starving at the continental margin during the Plio–Pleistocene (Alonso et al., 1992; Anderson and Bartek, 1992). This explains the mainly aggradational geometry of the shelf margin sequence and the absence of large sediment accumulation on the slope, compared to the previous sequences also observed in the Eastern Basin (Alonso et al., 1992). The change from mainly progradational to aggradational geometry at the shelf margin, across unconformity such as RSU2, was discussed considering other factors:

(1) An increase of shelf width by progradation with time makes the probability of grounding line positions being less skewed toward the shelf edge (Ten Brink et al.,1995). Aggradation, rather than progradation, can be achieved by decreasing the input sediment flux with time or by decreasing the probability that the grounding line will reach the shelf edge. Ten Brink et al. (1995) demonstrated that a greater longevity of the grounding ice sheet event (and therefore, a larger transportation of sediment) does not favour a progradation rather than an aggradation pattern at the shelf edge.

(2) The reduction in continental rise sedimentation was caused by cooling. A greater sediment transport capacity during the Pliocene is interpreted as due to the warmer conditions and, therefore, to a different glacial regime of the Antarctic ice sheet. The average sedimentation rates on slope and rise would have been greater in the early Pliocene than during the late Pliocene and Pleistocene because of more extended and faster ice flow (hence sediment transport) across the shelf (Barker, 1995). This hypothesis is also supported by the recent sedimentation rate reported for Miocene–Pleistocene sequences by drilling the Antarctic Peninsula Margin (Barker et al., 1999).We believe that the transition to colder climatic conditions was not the only cause for the decrease in sediment accumulation at the shelf edge during the most recent Plio–Pleistocene sequences.

We suggest that the overdeepening of the continental shelf, which probably occurred in the late Miocene–early Pliocene, may have represented an important change in WAIS configuration and, therefore, in the geometry pattern of the margin architecture. We believe that starting from that time the WAIS became a truly marine ice sheet, grounding at a large water-depth and relatively more sensitive to climatic and eustatic fluctuations. Starting from this time, WAIS probably caused high frequency eustatic fluctuations, before the Northern ice sheet developed. It is likely that the WAIS was not always able to reach the shelf edge, as during the LGM and, therefore, no large amount of sediment was systematically transported and deposited at the shelf edge and slope since RSU2.

In short, we suggest that, at the beginning of the deposition of RSS-7 and 8 (in the late Miocene–early Pliocene), the WAIS reached its maximum extension on the continental shelf (RSU2 time) and built large shelf margin fans, causing progradation of the shelf edge for several kilometers. After then, the continental shelf became wide, deeply eroded and overdeepened. This peculiar paleogeography, combined with a decreased sedimentation rate, due to a change in a polar glacial regime, could have resulted in a generally sediment-starved environment.

5. Conclusions

Four different stages characterise the evolution of the Eastern Basin:

(1) An early glacial stage (pre-Miocene–early Miocene times), when tidewater glaciers and small ice caps developed, while the basin was still subsiding and was influenced by marine sedimentation. At that time, the Ross Sea was not characterised by a wide, overdeepened continental shelf, like today but by an archipelagos of islands separated by deep basins.

(2) A transitional stage (early–middle Miocene time), during which the basin was gradually filled with sediments, and ice caps expanded onto the continental shelf;

(3) A fully glacial stage (late Miocene–Pliocene times) during which a large and thick ice sheet deeply shaped the continental shelf up to the shelf edge. During this time interval, the Ross Sea continental shelf reached its present typical configuration: landward, overdeepened, incised by marine ice streams draining from the WAIS and EAIS.

(4) A late fully glacial (Pliocene–present time) stage during which the continental shelf was clearly overdeepened and the sedimentation was reduced and mainly influenced by glacial processes.

One significant conclusion of this work is that the WAIS fully developed onto the Eastern Ross Sea continental shelf only in the late Miocene–early Pliocene. Before that time, the WAIS was mainly terrestrial. This has an important implication on the contribution of the Antarctic ice volume to the Cenozoic global eustatic fluctuations and to the mechanism for the formation of deep and bottom masses: in the pre-Miocene, the WAIS contribution to eustatic fluctuations was probably much larger than today.

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