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Cenozoic prograding sequences of the Antarctic continental margin: a record of glacio-eustatic and tectonic events

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ABSTRACT

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Sedimentary sections up to 6–14 km thick lie beneath many areas of the Antarctic continental margin. The upper parts of the sections contain up to 6 km of Cenozoic glacial and possibly non-glacial sequences that have prograded the continental shelf up to 85 km.

We describe the Cenozoic sequences using two general categories based on their acoustic geometries. Type IA sequences, which account for most prograding of the Antarctic continental shelf, have complex sigmoidal geometries and some acoustic characteristics atypical of low-latitude margins, such as troughs and mounds lying parallel and normal to the shelf edge and high velocities (2.0–2.6 km/s) for flat layers within 150 m of the seafloor. Type IIA sequences, which principally aggrade the paleoshelf, lie beneath type IA sequences and have mostly simple geometries and gently dipping reflections.

The prograding sequences are commonly located near the seaward edges of major Mesozoic and older margin structures. Relatively rapid Cenozoic subsidence has occurred due to the probable rifting in the Ross Sea, thermal subsidence in the Antarctic Peninsula, and isostatic crustal flexure in Wilkes Land. In Prydz Bay and the Weddell Sea, prograding sequences cover Mesozoic basins that have undergone little apparent Cenozoic tectonism.

Grounded ice sheets are viewed by us, and others, as the principal mechanism for depositing the Antarctic prograding sequences. During the initial advance of grounded ice, the continental shelf is flexurally overdeepened, the inner shelf is heavily eroded, and gently dipping glacial strata are deposited on the shelf (i.e. type IIA sequences). The overdeepened shelf profile is preserved (a) during glacial times, by grounded ice sheets episodically crossing the shelf, eroding sediments from onshore and inner shelf areas, and depositing sediments at the front of the ice sheet as outer shelf topset-banks and continental slope foreset-aprons (i.e. type IA sequences), and (b) during interglacial times, like today, by little or no clastic sedimentation on the continental shelf other than beneath retreated ice shelves lying far from the continental shelf edge. Ice streams carve broad depressions across the shelf and carry abundant basal sediments directly to the continental shelf edge, thereby creating trough-mouth fans and sheet-like prograding sequences (i.e. type IA sequences).

Numerous acoustic unconformities and multiple overcompacted layers within the prograding sequences suggest major fluctuations of the Antarctic Ice Sheet. The available drilling and seismic interpretations provide the following history: (1) Cenozoic ice sheets have existed in places near the continental shelf since middle to late Eocene time. (2) A grounded Antarctic ice sheet first expanded to the continental shelf edge, with probable overdeepening of the outer shelf, in late Eocene to early Oligocene time in Prydz Bay, possibly in early Miocene time in the Ross Sea, and at least by middle Miocene time in the Weddell Sea. (3) The relative amounts of shelf prograding and inferred ice-volume variations (and related sea-level changes) have increased since middle to late Miocene time in the eastern Ross Sea, Prydz Bay, and possibly Weddell Sea.

Our analysis is preliminary. Further acoustic surveys and scientific drilling are needed to resolve the proximal Antarctic record of glacio-eustatic, climatic, and tectonic events recorded by the prograding sequences.

Introduction

Antarctica is well known for its thick ice sheet, but is not commonly recognized as a continent whose margins are underlain by up to 14 km of sedimentary strata that include outer shelf prograding sedimentary sequences up to 6 km in thickness (Fig. 1). Like other margins of Gondwana, the sedimentary deposits of the Antarctic margin are of varied tectonic and depositional origin. Seismic reflection surveys have been conducted on the Antarctic margin since the early

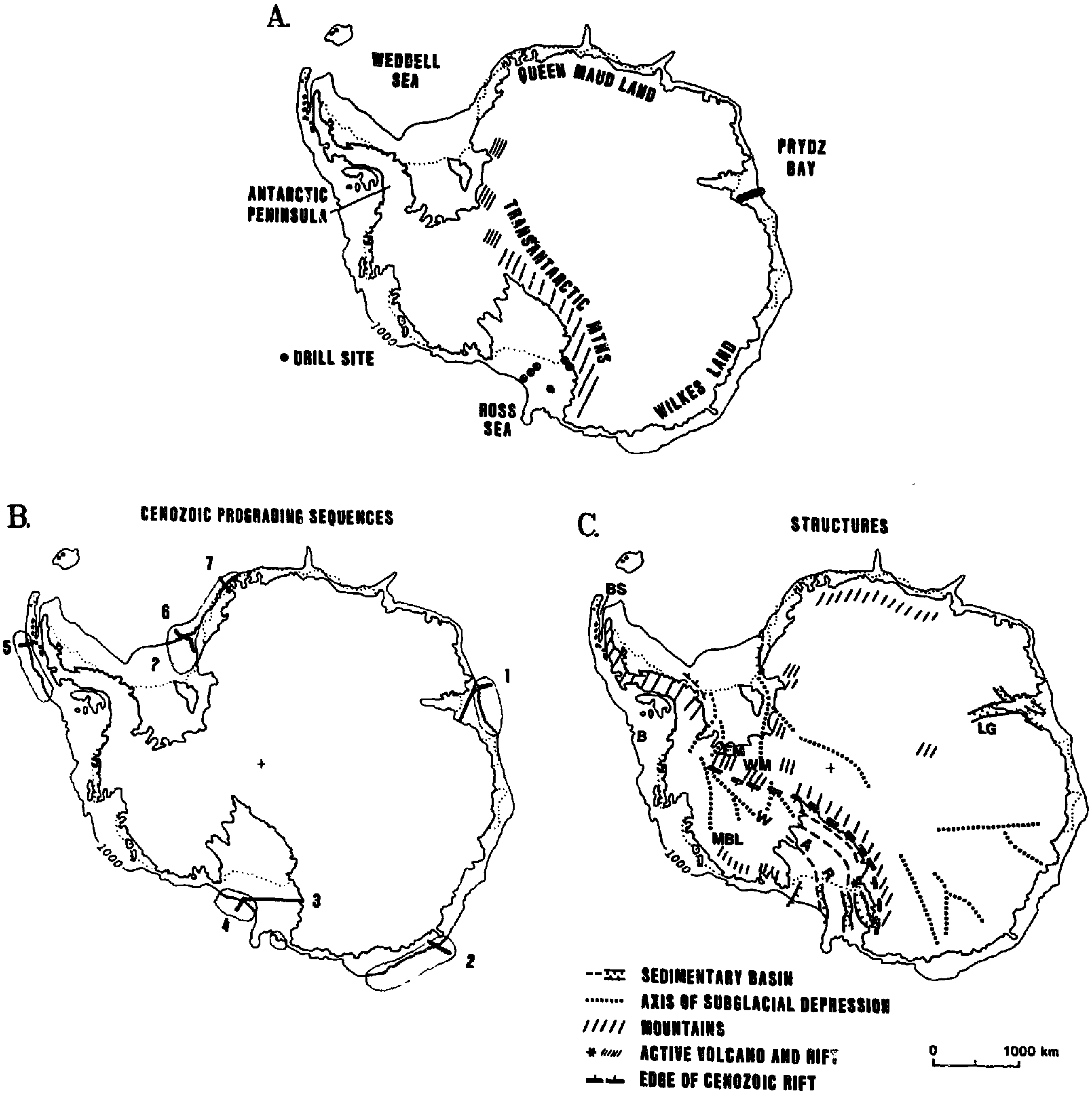


Fig. 1. Index maps of Antarctica. (A) Locations of geographic features and drill sites on the continental shelf. (B) Known locations of Cenozoic prograding sequences beneath the outer continental shelf (stippled) and of profiles shown in Figs. 2 and 3. (C) Locations of major exposed and sub-ice topographic and structural features. Mountains: *EM* = Ellsworth; *WM* = Whitmore. Structures: *LG* = Lambert Graben; *WAR* = West Antarctic Rift. Geographical names: *B* = Bellingshausen Sea; *BS* = Bransfield Strait; *MBL* = Marie Byrd Land.

1970s (e.g. Houtz and Davey, 1973), but the structure and distribution of Antarctica's sedimentary sequences and underlying basement rocks have been resolved only since 1977, with the collection of more than 120,000 km of multichannel seismic reflection (MCS) data (e.g. Hinz and Krause, 1982; Hinz and Block, 1984; Stagg, 1985; Eittreim and Smith, 1987; Cooper et al., 1987a, Larter and Barker, 1989; Behrendt, 1990).

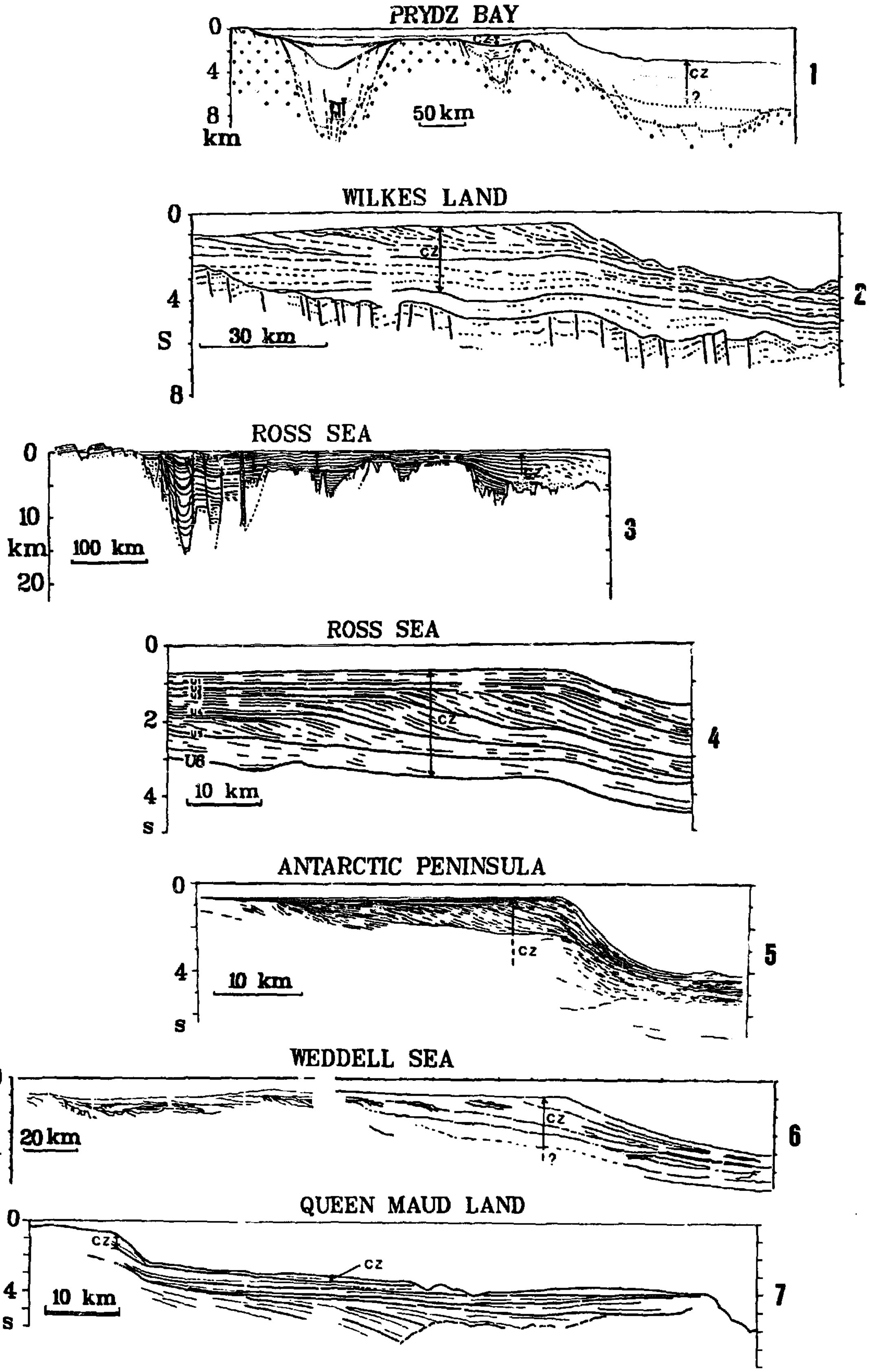
The MCS data across many parts of the Antarctic continental shelf from the Weddell Sea eastward to Marie Byrd Land show features characteristic of passive margins, such as layered sedimentary and possible volcanic deposits filling deep fault-bounded rift structures, widespread intermediate-depth unconformities, and overlying prograding sedimentary sequences that are bounded by unconformities (Fig. 2; Hinz and Krause, 1982; Hinz and Block, 1984; Wannesson et al., 1985; Cooper, 1989a; Wannesson, 1990; Cooper et al., 1991a). In other areas, which once may have been active margins (i.e. Marie Byrd Land eastward to the Weddell Sea), basement structures are not well imaged, yet widespread unconformities and shallow prograding sedimentary sequences are also common (Larter and Barker, 1989; Anderson et al., 1990, 1991). In many areas, up to 85 km of the outer continental shelf is underlain by prograding sedimentary sequences (Fig. 2).

The age and lithology of sub-seafloor deposits beneath the Antarctic continental margin are poorly known because these deposits have been drilled only at a few sites on the shelf and slope (Fig. 1). They are principally Cenozoic glacial marine strata (Hayes and Frakes, 1975; Barker et al., 1988; Barron et al., 1989; Barrett, 1986, 1989); Mesozoic and early Cenozoic pre-glacial sedimentary rocks have rarely been recovered and have been sampled at only a few localities onshore and *in-situ* at only six sites beneath the Antarctic margin [Wilkes Land (Domack et al., 1980), Weddell Sea (Barker et al., 1988; Anderson et al., 1983a), and Prydz Bay (Barron et al., 1989)]. However, recycled microfauna suggest that rocks of these ages occur beneath large areas of the margin (Truswell, 1983; Truswell and Drewry, 1984; Webb, 1990). The prograding sedimentary sequences may be mostly of glacial origin and younger than late Eocene to early Oligocene age (about 40 Ma), based on ages from ODP Site 742 near the acoustically defined base of the Prydz Bay prograding sequences (Cooper et al., 1991a) (see below).

Major structures related to the extensional breakup of Gondwana (e.g. Transantarctic Mountains, West Antarctic Rift, Lambert Graben, Figs. 1 and 3) lie beneath the thick ice sheets that cover Antarctica. These structures act as topographic barriers and conduits that partly control the flow of ice (and entrained sediment) from inland areas to the continental shelves. Prograding sedimentary sequences of the outer continental shelf lie near the outlets of these major topographic–structural depressions (e.g. Weddell Sea, Prydz Bay, Ross Sea). However, such sequences also lie near the fronts of topographic depressions that cannot be as readily related to Cenozoic rift structures (e.g. Wilkes Land and most of the Antarctic Peninsula). Nevertheless, the evolution of the offshore prograding sedimentary sequences appears to be closely tied to episodic movements of grounded ice, which are related to changes in sea level, out to the continental shelf edge (e.g. Barrett, 1989; Cooper and Webb, 1990; Bartek et al., 1991).

In this paper, we shall discuss the general characteristics of Cenozoic prograding sequences beneath the Antarctic continental margin and the many factors that are believed to control their deposition. Some of these factors are common to all continental margins, and others affect only glaciated polar margins. Consequently, the complex sigmoidal acoustic stratigraphy of the prograding sequences has general similarities with low-latitude, non-glaciated margins, but also has other acoustic features that we speculate to be unique to polar margins with formerly extensive grounded ice sheets.

We concentrate on the Prydz Bay and Ross Sea regions as examples of Mesozoic and Cenozoic rift margins which have been affected by Cenozoic glaciation, and which exemplify the acoustic characteristics of the Cenozoic prograding sequences found elsewhere around Antarctica. These two regions of Antarctica are the only places where drilling has been carried out on the continental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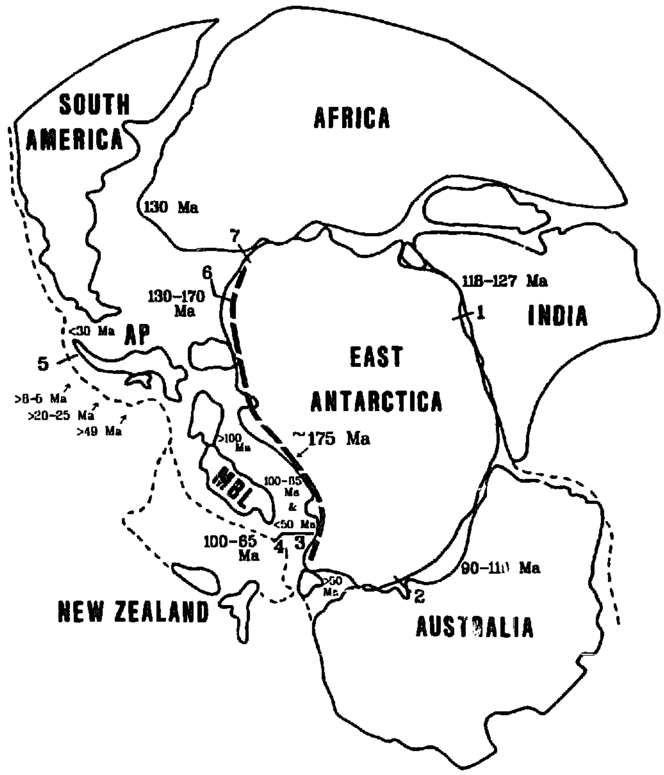


Fig. 3. Reconstruction of Gondwana prior to initial breakup (ca. 175 Ma) showing approximate times of initial rifting to breakup, approximate times of subduction along the Antarctic Peninsula (*AP*), and locations of profiles (modified from Lawver et al., 1991). Dashed line shows position of possible Jurassic rift marked by dolerite sills and intrusions (Thomson, 1983). Episodic rifting is postulated between Marie Byrd Land (*MBL*) and East Antarctica.

shelf to help constrain the ages of and the mechanisms that generated these sequences.

We, and others, believe that the prograding sequences record a proximal history of the waxing and waning of the Antarctic Ice Sheet (e.g. Cooper and Webb, 1990). These sequences may also hold the answer to the question of whether the 3rd order ($\sim$1 m.y.) cycles in the global sea-level curves of Haq et al. (1987) and Haq (1991) are caused by fluctuations in the volume of the Antarctic Ice Sheet. This paper is a brief introduction to the complex and poorly understood Cenozoic stratigraphy of the Antarctic continental margin. The answer to the above question awaits detailed syntheses of existing international data sets and acquisition of new geophysical and scientific drilling data (e.g. Cooper and Webb, 1990).

Previous studies

Seaward-dipping reflectors in the sedimentary sections beneath the Antarctic margin have been noted since the 1970s (e.g. Houtz and Davey, 1973; Hayes and Davey, 1975; Davey, 1985), but only with recent drilling data (e.g. Barrett, 1989; Barron et al., 1989) have they been closely linked to likely glacio-eustatic processes. Seaward-dipping strata of likely glacial origin have been delineated beneath the mid- to outer shelf in reflection profiles from ice-free areas:

In the Weddell Sea: (Hinz and Krause, 1982; Okuda et al., 1983; Elverhoi and Maisey, 1983; Haugland et al., 1985; Kuvaas and Kristoffersen, 1990; Ivanov and Kamenev, 1990; Traube and Rybnikov, 1990).

Fig. 2. Interpretive profiles across the Antarctic continental margin showing general configuration of likely Cenozoic (*CZ*) prograding sedimentary sections that commonly bury older sedimentary units and structures beneath many parts of the margin. Some vertical scales are in time and some in depth. Profiles are from (1) Leitchenkov et al. (1990), (2) Wannesson (1990), (3) Cooper et al. (1991), (4) Hinz and Block (1984), (5) Larter and Barker (1989), (6) Haugland et al. (1985) and (7) Hinz and Krause (1982).

Along the Antarctic Peninsula: (Kimura, 1982; Larter and Barker, 1989; Anderson et al., 1990, 1991; Gamboa and Maldonado, 1990; Jeffers and Anderson, 1990).

In the Ross Sea: (Hayes and Davey, 1975; Hinz and Block, 1984; Sato et al., 1984; Cooper and Davey, 1987; Hinz and Kristoffersen, 1987; Cooper, 1989b; Berger et al., 1989; Ivanov and Kamenev, 1990; Bartek and Anderson, 1990; Zayatz et al., 1990; Cooper et al., 1991b; Bartek et al., 1991).

Along Wilkes Land: (Tsumuraya et al., 1985; Wannesson et al., 1985; Eittreim and Smith, 1987; Wannesson, 1990).

In Prydz Bay: (Stagg, 1985; Mizukoshi et al., 1988; Leitchenkov et al., 1990; Ivanov and Kamenev, 1990; Cooper et al., 1991a).

Other seismic examples are given in Cooper and Webb (1990). Several mechanisms have been proposed for the dipping strata, including fluvial marine deltas (Hinz and Block, 1984), till deltas (Alley et al., 1989), and diamict aprons (Hambrey et al., in press). Many investigators ascribe such features to rapid deposition of glacial debris at the grounding line of moving ice sheets (Fig. 4; Drewry and Cooper, 1981; Anderson et al., 1982; Hinz and Krause, 1982; Barrett, 1989; Larter and Barker, 1989; Bartek et al., 1991; Hambrey, 1991; Hambrey et al., 1991, in press).

Antarctic Cenozoic glacial history

The following brief commentary on Antarctic Cenozoic glaciation is based on the reviews of Webb (1990, 1991) and Wise et al. (1991). We shall discuss several other aspects of glaciation more thoroughly in a later section.

Our knowledge of Cenozoic Antarctic glacial history is derived principally from offshore deposits because few Cenozoic sedimentary deposits are found on Antarctica. The onshore and continental shelf deposits indicate that (1) large ice sheets covered Antarctica at least from late Oligocene–early Eocene time (38–45 Ma), and local glaciation is recorded in the Antarctic Peninsula in middle Paleocene to early Eocene time, and (2) during the past 36 m.y. there has been a complex history of glacial–interglacial events supporting the likely existence of major marine incursions into and across the continent as the Antarctic Ice Sheet receded (Figs. 5 and 6).

The abyssal basins and oceanic plateaus around

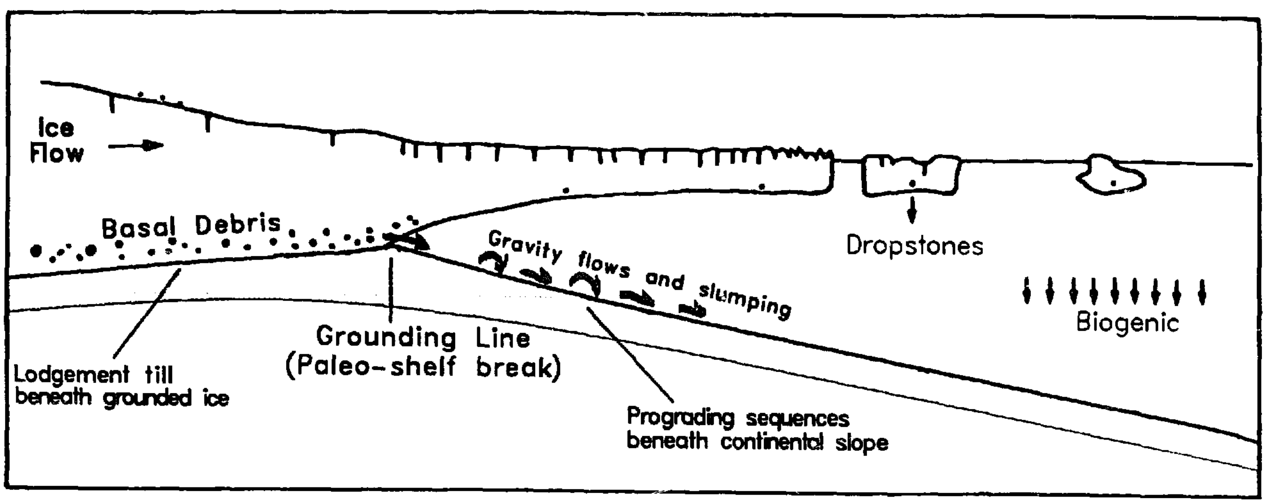
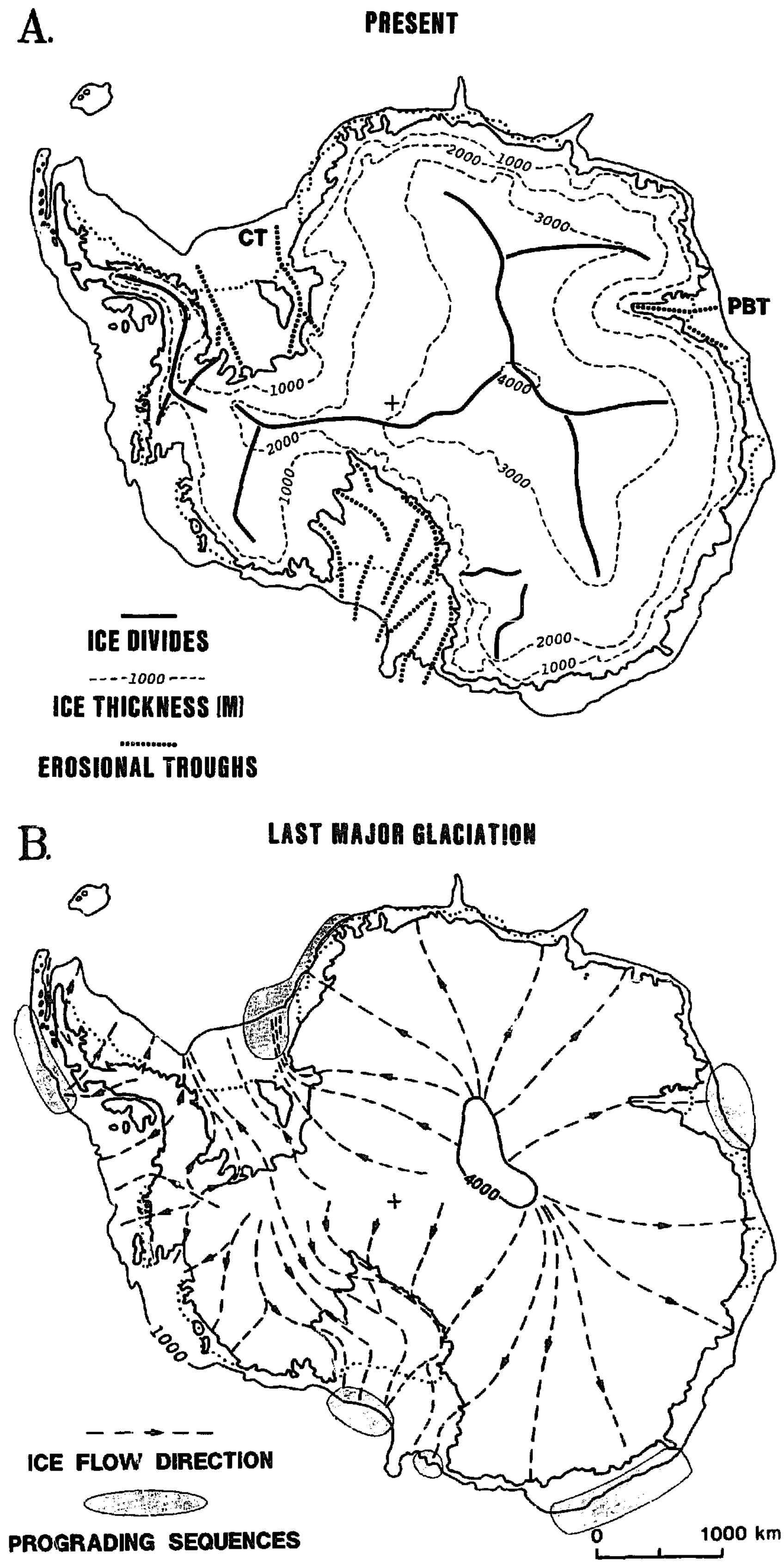


Fig. 4. Ice-sheet depositional model (modified from Hambrey et al., 1991). Sediments are eroded and carried from continent and inner shelf areas, principally as basal debris in grounded ice sheets, and are deposited on the continental shelf as lodgement tills and on the continental slope as marine diamictons.

Fig. 5. Maps of the Antarctic Ice Sheet. (A) Present: thickness of onshore ice and locations of offshore ice shelves (small dots), ice divides and large sub-ice and offshore topographic depressions (modified from Drewry, 1983). (B) Last major glaciation: inferred flow directions for ice to the continental shelf edge (from Denton et al., in press). Cenozoic prograding sequences lie near the inferred outlet of former grounded ice sheets draining interior Antarctica. CT=Crary trough; PBT=Prydz Bay trough.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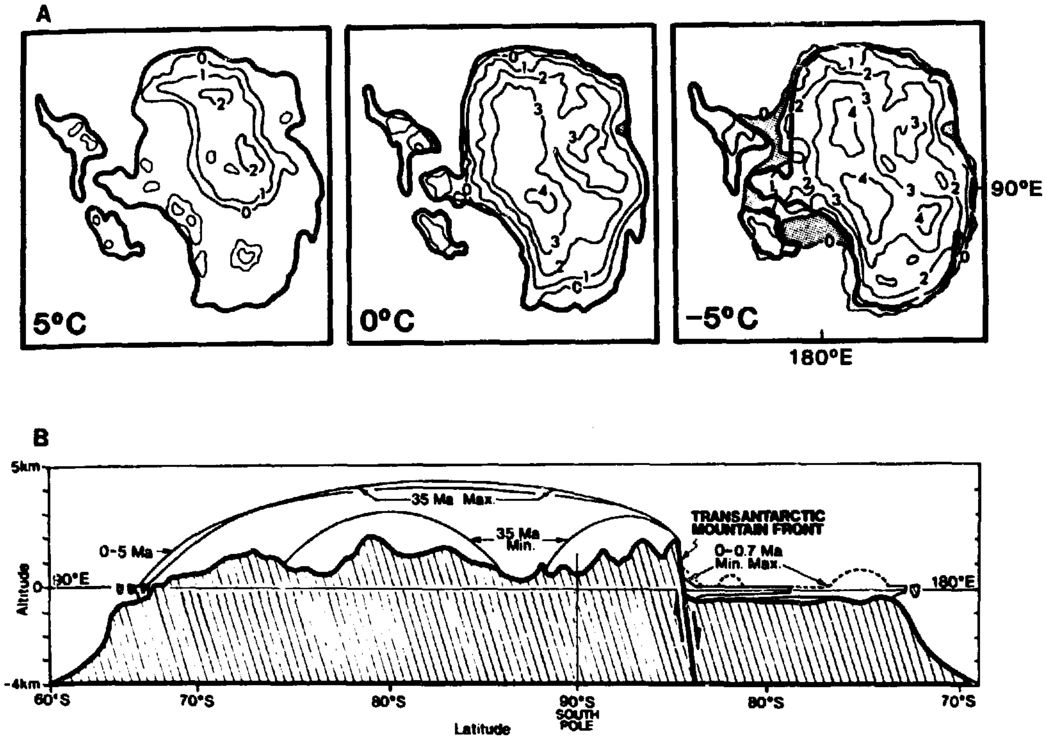


Fig. 6. Maps and profile of the Antarctic Ice Sheet (from Barrett et al., 1989). (A) Equilibrium states for three annual sea-level temperatures. Ice thickness is in kilometers and ice shelves are shaded (from Oerlemans (1982). (B) Cross section comparing the inferred size of Antarctic ice sheets in Oligocene time with those since the Miocene (modified from Robin, 1988).

Antarctica record a different history that suggests major buildup of ice during the Neogene, but with limited glaciation extending back to earliest Oligocene and possibly Late Cretaceous times (Kennett, 1982). The different histories derived from the proximal and distal ocean records reflect the different data bases, techniques and assumptions used to estimate the volume and size of the Antarctic Ice Sheet (e.g., Barrett, in press). However, we believe that the most direct and reliable record of Cenozoic Antarctic glaciation is held within the prograding sedimentary sequences that lie beneath the continental margins and beneath former interior marine basins [see also Mercer (1983), Webb et al. (1984) and McKelvey et al. (1991)].

Processes controlling Antarctic Cenozoic prograding sequences

The construction of Antarctica's prograding sequences is controlled, in part, by the eustatic, tectonic and depositional processes that are commonly ascribed to low-latitude (i.e. non-glacial) continental margins, and additionally by glacial processes presently found only in Antarctica (Fig. 7). Although geometric similarities exist between low-latitude and Antarctic continental margins (Hinz and Block, 1984; Bartek et al., 1991), the evolutionary terminology for prograding sequences on low-latitude margins (e.g., Vail et al., 1977; Haq et al., 1987; Wilgus et al., 1988) may not be fully applicable to Antarctica and other polar margins, especially at times when depositional (and lithospheric) processes are controlled largely by massive grounded ice sheets covering the overdeepened continental shelf. In our discussion we shall briefly review the processes associated with all continental margins and then describe the processes that are found only in Antarctica.

All continental margins

Cyclic sedimentary sequences of low-latitude continental margins are controlled principally by subsidence, tectonics, sediment supply and eustacy (e.g.,, Wilgus et al., 1988; Fig. 7). Sediment loading

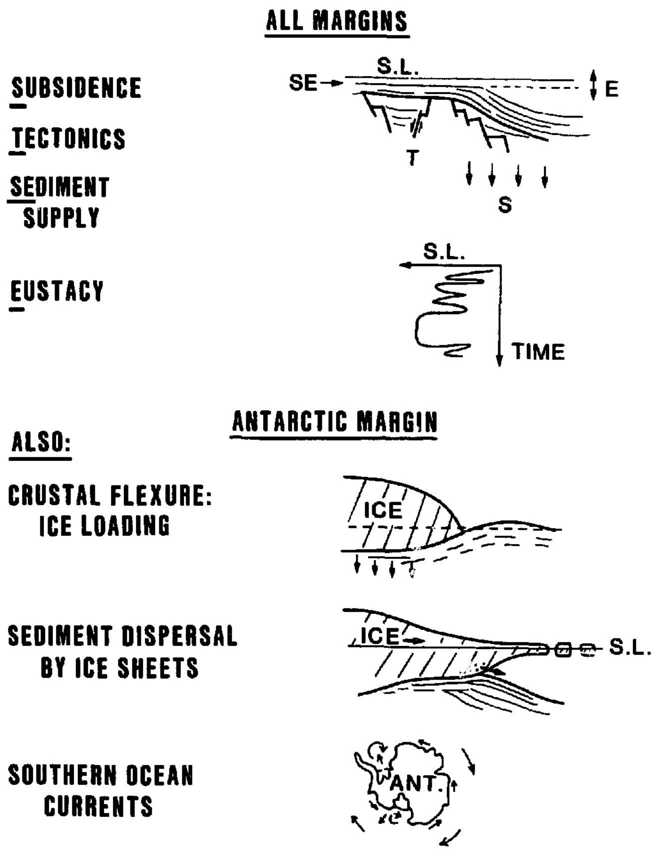


Fig. 7. Cartoons illustrating the different processes affecting the deposition of sedimentary sequences on continental margins. Some processes are unique to Antarctica and other glaciated margins.

and compaction are also factors (Reynolds et al., 1991). Continuous eustatic changes of up to 300 m over the past 100 m.y., and 70–150 m during 1 to 5 m.y. intervals, have been proposed and attributed to variations in the volume of polar ice sheets and ocean basins. Where water depths are generally less than 100 m on the continental shelves, the eustatic changes have resulted in major shifts in depositional environments and the superposition of distinctly different sediment types (Vail et al., 1977; Haq et al., 1987).

The cyclic sequences are best recorded on passive continental margins where tectonism and subsidence are usually slow and predictable, relative to the rate of eustatic change (e.g., Falvey and Mutter, 1981). Passive margin subsidence can be attributed to a rapid pulse of initial crustal thinning, followed by thermal cooling in which cooling rate decreases with time, and to an isostatic adjustment for sediment loading that depends on sediment supply (Falvey and Mutter, 1981; Bond and Kominz, 1988). The degree to which a passive margin progrades (i.e. builds outward) and aggrades (i.e. builds upward) depends upon the rate at which sediment is supplied to the margin, relative to the rate of tectonic subsidence. Prograding occurs when sediment is supplied faster than can be accommodated in shelf basins (e.g., Vail et al., 1977; Wilgus et al., 1988). Sedimentary sequences on tectonically active margins respond to the same basic processes, as described above. But, the rates of tectonic displacements are commonly high and variable, leading to greater diversity in sequence geometries.

Antarctic continental margin

The crustal framework of the Antarctic continental margin, with the possible exception of the Ross Sea and the greater Antarctic Peninsula areas, formed mostly prior to the deposition of the Cenozoic prograding sedimentary sequences under consideration. The oldest age of these sequences is unknown. We assume, for this paper, that these sequences are no older than late Eocene to early Oligocene age (i.e. ca. 36–40 Ma) because rocks within this age range were recovered at ODP Site 742 from near the base of the glacial prograding sequences in Prydz Bay (Barron et al., 1989).

During the late Paleozoic(?) to early Cenozoic breakup of Gondwana, basement rocks now underlying the continental margin were progressively rifted from the Weddell Sea eastward to the Ross Sea (Fig. 3). The ensuing passive margins subsided and shelf basins were filled with many kilometers of preglacial (i.e. pre-36–40 Ma) sediments, probably by processes similar to those that have acted on present low-latitude passive margins.

Since about 36–40 Ma, i.e. during "glacial times", the Antarctic continental margin has also been subjected to other processes, which are either unique to Antarctica or are common only to glaciated polar margins:

(1) *Lithospheric processes* related to the loading and disruption of the Antarctic lithosphere caused by the Antarctic Ice Sheet. Although not unique,

we also discuss active Cenozoic tectonism under this heading.

(2) *Ice-sediment processes* related to the distribution of sediment by the Antarctic Ice Sheet.

(3) *Isolation processes* (e.g., oceanographic and meteorological) related to the unique isolated position of Antarctica within the Southern Ocean.

These processes have resulted in some acoustic geometries within the Antarctic prograding sequences that we believe differ from those of low-latitude prograding sequences. Many of these processes are interrelated, and our classification of "processes" is not exclusive. The following discussion of these processes is necessarily brief due to the extensive available literature (e.g., Walcott, 1973; Kennett, 1982; Drewry, 1983; Anderson et al., 1983a; Deacon, 1984; Webb et al., 1984; Jacobs, 1985; Drewry, 1986; Dowdeswell, 1987; Robin, 1988; Jacobs, 1989; Anderson and Molnia, 1989; Webb, 1990; LeMasurier, 1990; Lawver et al., 1991; Hambrey et al., 1991, in press; Denton et al., in press).

Lithospheric processes

We recognize two Antarctic lithospheric processes that are partly responsible for forming pathways along which ice and sediment are transported to and across the continental shelf, and for controlling the 200–1200 m water depths of the depositional environments on the continental shelf. The two processes are: (1) active Cenozoic tectonism, and (2) loading and unloading of the Antarctic lithosphere by the Antarctic Ice Sheet (i.e. crustal flexure and isostacy).

Cenozoic tectonism is known or suspected in Antarctica principally in the West Antarctic Rift and on the Antarctic Peninsula. The West Antarctic Rift is a major ice-filled and partly sediment-covered series of basement and topographic depressions (up to 2540 m deep), that extends from the Ross Sea through Antarctica to the Bellingshausen coast and includes parts (and perhaps all) of Marie Byrd Land (LeMasurier and Thomson, 1990; Behrendt and Cooper, 1991; Tessensohn and Worner, 1991). The rift is bordered partly by the Transantarctic Mountains, which form the tectonic boundary between East and West Antarctica, and partly by the Whitmore and Ellsworth Mountains (Fig. 1). These rift-shoulder mountains have topographic elevations of 1500–4000 m, and have been uplifted episodically since Late Cretaceous time (Fitzgerald et al., 1987; Stern and Ten Brink, 1989; Behrendt and Cooper, 1991). Their uplift has long been suspected as one factor in initiating and controlling Cenozoic Antarctic glaciation (David and Priestly, 1914; Gould, 1931; Webb, 1990; Behrendt and Cooper, 1991).

The West Antarctic Ice Sheet covers most of the West Antarctic Rift and is part of a major ice drainage from the Antarctic interior across the Transantarctic Mountains and to the eastern Ross Sea (Drewry and Cooper, 1981; Denton et al., in press). One of the largest Cenozoic prograding sedimentary bodies in Antarctica fills the eastern Ross Sea at the former seaward mouth of this ice drainage system (Hinz and Block, 1984; Cooper et al., 1987a, 1991b; Bartek et al., 1991). The tectonic and depositional significance of this body has only recently been recognized (Hinz and Block, 1984; Cooper et al., 1991b; see below). Another major prograding sedimentary body lies beneath the Weddell Sea embayment at another major outlet of the West Antarctic Ice Sheet (Haugland et al., 1985; Kuvaas and Kristoffersen, 1991).

The Antarctic Peninsula has had a complex Cenozoic history of progressive ocean-ridge subduction, outer margin subsidence, and back-arc basin development (see Dalziel and Elliot, 1982). Thick prograding sedimentary sequences have been deposited across much of the offshore areas, and are well preserved on the once active and now passive outer margin (Fig. 2 (section 5); Larter and Barker, 1989, 1991; Anderson et al., 1983a, 1991).

Ice sheet loading: The Antarctic margin has an uncommon, reversed bathymetric profile with water depths of up to 1200 m adjacent to the continent, shallowing to 200–400 m at the shelf edge (Vanney and Johnson, 1985). Large bathymetric variations in this profile occur at deep shore-parallel and cross-shelf troughs that have probably been eroded by glaciers (Anderson et al., 1983a; Vanney and Johnson, 1985), but the average water depth for the shelf exceeds 400 m. This overdeepening is probably due to several factors (Ten Brink and Cooper, 1990), one of which is flexural loading, or warping of an elastic crust, at the front of thick grounded ice sheets (Walcott, 1970, 1972; Drewry, 1983).

Walcott (1970, fig. 10) estimates that an 1800 m thick terrestrial ice sheet, with dimensions similar to those of the present Antarctic ice sheet, would produce a proglacial depression 155 m deep and a forebulge 18 m high at a distance of 280 km from the front of the ice sheet. Or, if the same ice sheet were on a 100 m deep continental shelf, the trough would reach 255 m below sea level. Following withdrawal of the ice sheet, the land surface would rebound at about 6–8 cm/yr, with 50% of the rebound completed in about 2000 yrs. Walcott's depth and rebound-time estimates suggest that flexure alone cannot explain the present shelf depths. Grounded ice sheets are believed to thicken rapidly at their grounding line, where sediment is mostly dispersed (Drewry, 1986). Consequently, flexural loading during times when grounded ice sheets move across the continental shelf to the shelf edge may greatly affect depositional depths of prograding strata.

Two other factors are probably partly responsible for the great depth and shoreward-deepening bathymetric profile of the shelf: (1) prolonged low sedimentation rates relative to thermal subsidence of the shelf, especially during interglacial periods like the present (Domack and Anderson, 1983), and (2) erosion and redistribution of sediments from the inner shelf eroded troughs to the outer shelf depositional banks via grounded ice sheets during glacial maxima (Anderson et al., 1983a; Hambrey, 1991). This redistribution of sediments and gradual upward building of the outermost shelf is an important and probably common feature of the Antarctic continental shelf.

Ice-sedimentation processes

In the present interglacial period, the Antarctic Ice Sheet covers 98% of onshore areas and reaches a maximum thickness of about 4800 m (Fig. 5A). Offshore areas are covered principally by floating or seasonal ice that is not grounded, except over shallow banks. During glacial periods, however, the ice sheet may reach a thickness of over 5000 m, cover all land areas, and be grounded across and to the edge of the continental shelf (Figs. 5B and 6; Robin, 1988). Ice has often been the principal transport agent for supplying sediment to the continental shelf and directly to the the continental shelf edge (Kellogg et al., 1979; Anderson and Molnia, 1989). We note three ice processes that have been important, during glacial and interglacial periods, to the development of the Antarctic prograding sequences: (1) erosion by ice, (2) sediment transport by ice and (3) sediment deposition from ice.

Erosion by ice: Erosion by grounded ice sheets is greatly dependent on the glacial thermal regime, which controls the amount of water at the base of ice sheets (Carey and Ahmad, 1961; Alley et al., 1989). Most sediments now reaching the continental shelf areas from the continental interior are believed to be eroded by glaciers and ice streams that move over a water-saturated bed of till (Shabtai and Bentley, 1987; Alley et al., 1989). Hundreds of meters of erosion has occurred beneath parts of the onshore Antarctic ice sheet, especially within major structural embayments, such as the Ross and Weddell embayments and Lambert Graben, and within present and former temperate glacial valleys, as along the Antarctic Peninsula and Transantarctic Mountains (Robertson et al., 1982; Anderson et al., 1983a; Haugland et al., 1985; Barrett, 1989; Hambrey, 1991).

In the offshore, erosion by former grounded ice sheets, like those now onshore, has resulted in up to 1400 m deep troughs that lie parallel and perpendicular to the shelf edge around Antarctica (Vanney and Johnson, 1985). The deepest troughs lie in front of onshore structural embayments, such as the Lambert Graben, and along the sediment–basement contact at the edge of major mountains and sedimentary basins, as along the Crary and Prydz Bay troughs; Fig. 5A). Local channels also cut into the prograding sequences of the outer shelf. Most erosion and deposition of outer shelf sequences has apparently occurred along the nearly flat base of a moving grounded ice sheet (discussed below) and to a lesser extent locally by grounded icebergs (review in Anderson and Molnia, 1989; Larter and Barker, 1989; Hambrey et al., 1991, in press).

Sediment transport by ice: The present Antarctic ice sheet transports most sediment from interior regions to the continental shelf in a basal debris layer, which is up to 15 m thick and contains 5–8% sediment (Anderson et al., 1983a; Alley et al., 1989). Where the ice sheet flows into the sea, most of this debris melts out close to (i.e. within a few kilometers) of the grounding line through basal melting (Hambrey et al., 1991, in press). A much smaller amount of debris is blown or falls onto glacier surfaces, to be later buried in the glacier ice, and this can be transported by icebergs for hundreds or thousands of kilometers before melting releases it to settle to the seafloor.

Grounded ice sheets have previously moved across the continental shelf transporting sediment from upstream inner shelf shelf areas to the outermost shelf (Anderson et al., 1983a). This process has partly enhanced overdeepening of the shelf and has further increased the sediment supply to the outermost shelf. Thus, during past glacial periods, large volumes of sediment may have been carried to the prograding sequences beneath the outer shelf by grounded ice sheets, but during interglacial periods, like today, little terrigenous sediment is carried to the outer shelf by ice (Larter and Barker, 1989).

Sediment deposition from ice: Sediment can be deposited from ice only by melting. As most of the debris carried by the Antarctic Ice Sheet is in the basal debris layer, it is likely that most of the glacial sediment on the Antarctic continental shelf is released from that source either subglacially or from shelf ice or icebergs close to the grounding line. In some cases, such as the fast-moving ice streams of Marie Byrd Land, the ice moves by basal shear deformation of unconsolidated sediments (Engelhardt et al., 1990), with debris being dragged along the base of the ice stream to the grounding line (Alley et al., 1989). At the grounding line the sediment is melted out into a subglacial delta with geometries probably like those observed in the prograding sequences described here. Sediment melted from icebergs is deposited and widely redistributed (Anderson et al., 1983a) and cannot explain these geometries.

Lodgement tills and diamictons derived from the base of former grounded ice sheets are found in seafloor and drill cores from all parts of the continental shelf (Hayes and Frakes, 1975; Anderson et al., 1983a; Barron et al., 1989; Barrett, 1986, 1989). These sediments have highly variable physical and engineering properties (velocity, density and strength), as determined from core sample and downhole logging measurements (Hayes and Frakes, 1975; Edwards et al., 1987; Barron et al., 1989; Solheim et al., 1991). Downhole logging and seismic measurements at ODP Site 739 into the prograding sequence underlying Prydz Bay show clear evidence of high-velocity (1.9–2.6 km/s), overcompacted massive diamictons that are flat-lying and overlie more normally compacted, lower velocity stratified diamictons that dip steeply seaward (Cooper et al., 1991a; Cochrane and Cooper, 1991). The flat-lying deposits at Site 739 show several overcompaction events (Solheim et al., 1991), and are believed to have been deposited beneath the front of a moving grounded ice sheet (Fig. 4; Hambrey et al., in press).

Isolation processes

Antarctica is surrounded and isolated by the unrestricted flow of the Southern Ocean. The global meteorologic and oceanographic effects of this isolation have been recognized as potential contributors to the initiation and continuance of Antarctic Cenozoic glaciation (Kennett, 1982; Webb, 1990). Antarctica attained its present deep-water isolation with the opening of the Drake Passage at about 20 to 26 Ma (Barker and Burrell, 1977) and the earlier pull-away of the South Tasman Rise at about 40 Ma (Hinz et al., 1990). Seaways may have existed, however, through and around parts of interior Antarctica, such as in the Ross–Weddell embayment and Wilkes Basin at other times during the Cenozoic (Fig. 6; Webb, 1990).

We shall discuss two isolation processes that probably affected the depositional environments of the prograding sequences: (1) wind-driven currents and (2) Antarctic Bottom Water.

Wind-driven currents affect Antarctic sediment distribution across the outer margin. The dominant wind direction adjacent to Antarctica is easterly. These winds cover the shelf and mid-slope areas and produce the East Wind Drift, a water-column current that moves from east to west and controls the principal movement of Antarctic icebergs (Deacon, 1984). These currents directly influence bottom sediments only on shallow banks less than 300 m deep (Anderson et al., 1983a). Directly north of the mid-slope region where the Antarctic Divergence occurs, winds are westerly and warmer deep waters upwell. Here, greater productivity and more rapid melting of icebergs occurs, as is evident from larger concentrations of biogenic and ice-rafted sediment in the surface sediments of these areas (Anderson, 1972; Anderson et al., 1983a).

Geostrophic currents within deep waters around Antarctica move along the shelf and upper slope removing and transporting fine-grained sediments to the lower slope and rise (Anderson et al., 1983b). The Weddell Sea Gyre, with a clockwise current, is an example that has extensively eroded former slope deposits to form major regional unconformities in the Weddell Sea (Orheim and Elverhoi, 1981; Hinz and Krause, 1982). These currents are documented in areas where major prograding sequences exist, as in the Ross Sea, Weddell Sea and Wilkes Land, and are likely to be a factor affecting depositional environments.

Antarctic Bottom Water is cold, dense water that forms beneath ice shelves on the Antarctic continental shelf and sinks along bathymetric gradients into the adjoining abyssal basins. The greatest production is in the Weddell Sea, but production is also noted in the eastern Ross Sea, Wilkes Land and the Antarctic Peninsula regions (Deacon, 1984; Jacobs, 1985, 1989). The effect, if any, on the prograding sedimentary sequences is unknown. However, the general regions where bottom water is now produced and moves as contour currents on the continental shelf, and possibly continental slope (Jacobs, 1989), correspond to those areas of the Weddell and Ross Sea where extensive prograding sequences occur. This coincidence may, however, simply be fortuitous.

Seismic reflection data

General characteristics

For this discussion, we shall describe Cenozoic sedimentary sequences observed beneath at least five segments of the Antarctic continental margin using two general categories based on acoustic geometry largely near the present and paleocontinental shelf edges (Fig. 8):

Type IA: Sequences that mostly, but not exclusively, prograde the paleocontinental shelf, building it outward. Type IA sequences have complex geometries, and include some features that may occur only on polar margins with grounded ice sheets.

Type IIA: Sequences that mostly, but not exclusively, aggrade the paleocontinental shelf, building it upward. Type IIA sequences have less complicated geometries that are common on low-latitude, non-glaciated margins.

Our categories are neither unique nor exclusive, and the distinction between type IA and IIA sequences is sometimes subtle, especially in low-resolution seismic data that commonly have severe multiple reflection noise.

The Cenozoic sedimentary sections of the Antarctic margin are composed principally of type IA sequences. Hence, extensive prograding, with relatively minor aggrading, of the outer continental shelf is common, except in the eastern Ross Sea (Fig. 2). Type IIA sequences are less common and, in our categorization, occur below type IA sequences. We identify type IIA sequences in the eastern Ross Sea, probably in Prydz Bay, and possibly in Wilkes Land. All sequences are bounded by unconformities or disconformities, sometimes widespread. We suspect that the upward transition from type IIA to IA sequences, which is most readily observed at the paleoshelf edges, may signal an important change in sediment depositional processes on the continental margin, and be related to an increase in the extent of Antarctic glaciation, as discussed below.

Antarctic Cenozoic type IA sequences

Type IA sequences are illustrated in the literature for most areas of the Antarctic outer continental shelf where seismic reflection data have been recorded (e.g., Figs. 2, 9, 10A and 11). Type IA sequences also occur locally beneath nearshore banks (Barrett, 1986, 1989) and probably at the grounding line of major ice shelves (Alley et al., 1989). We show examples from the eastern Ross

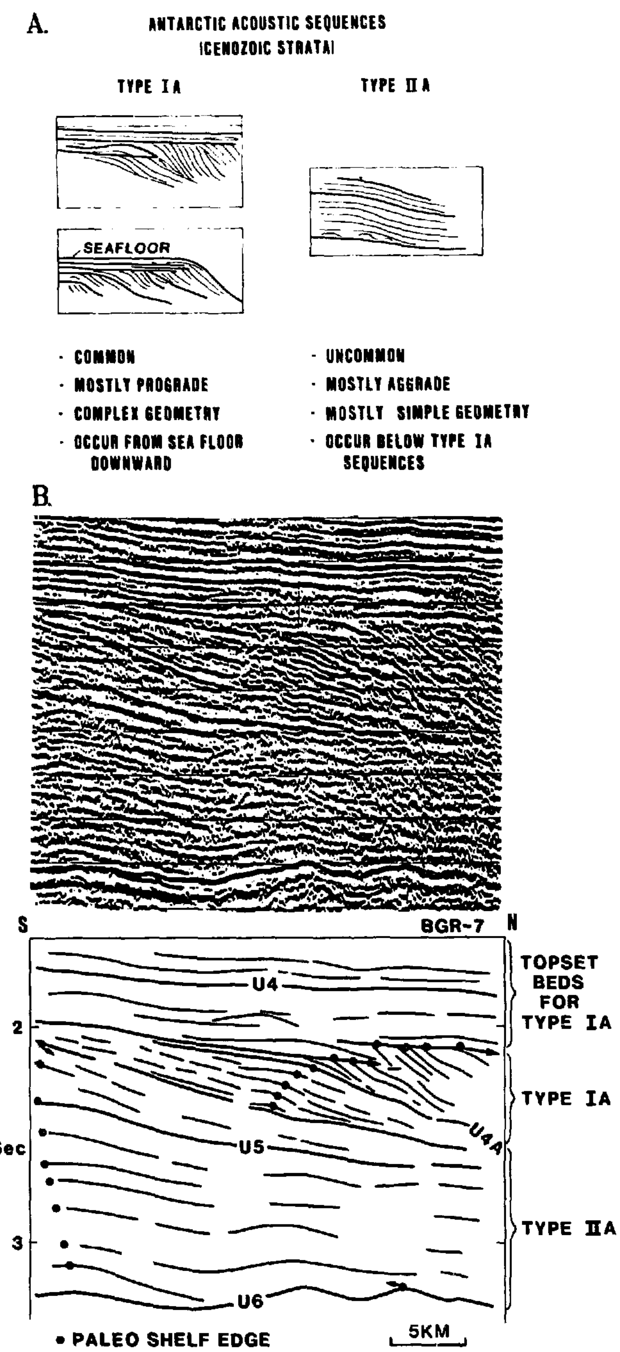


Fig. 8. Our initial characterization of Cenozoic sequences beneath the Antarctic continental margin. (A) General characteristics of type IA and IIA sequences (*A* is for Antarctica). (B) Some typical geometries within deeply buried type IA and IIA sequences. Transitional-IIA sequences occur between unconformities U4A and U5. Type IIA sequences may result from nonglacial to early glacial depositional processes on normal water depth shelves. Type IA geometries may be caused principally by full-glacial depositional processes on overdeepened shelves that have grounded ice sheets extending to the paleo-continental shelf edges. See Fig. 12A for location.

Sea and Prydz Bay where a wide range of type IA geometries exists.

Antarctic type IA sequences are most commonly characterized by strongly prograding strata that have relatively thin topset beds and steeply dipping (4–15°) foreset beds, particularly on the paleo-upper continental slope near the paleoshelf edges. Internally, the sequence geometry can be complex. Reflections within the sequences have strong and variable amplitudes that are often disrupted locally, but can sometimes be traced over tens of kilometers in the topset and foreset beds (Fig. 10 and 11). The dips of foreset beds may differ between adjacent sequences (Fig. 9), and foreset reflections downlap seaward onto underlying type IA and IIA sequences (Fig. 11).

Topset beds in type IA sequences can be acoustically complex. These beds are typically thin (0–50 m thick) and are strongly eroded over many parts of the shelf. In a few areas, however, such as the eastern Ross Sea and parts of Prydz Bay, stacked topset strata have built extensive outer shelf banks up to several hundred meters thick (Fig. 11; Cooper et al., 1991a). These banks are, in turn, eroded to give the seafloor part of its present bathymetric relief. The outer shelf banks are up to 50–100 km wide (Fig. 12; Bartek et al., 1991), and commonly the constituent topset strata thicken away from the axis of major ice-erosion channels now crossing the continental shelf (e.g., Cooper et al., 1991; Fig. 12). Strong layered reflections, which are mostly flat, mark the topset surfaces, which can be either interlayered disconformities within thick bank deposits or abrupt angular unconformities cutting into older foreset strata (Fig. 10A).

Reflections from the interlayered disconformities and unconformities within type IA topset beds sometimes terminate abruptly, and these termination points can move landward and seaward going upsection near the paleo-shelf edges (Figs. 10 and 11). Local mounded reflections may occur directly above and below the terminations. In areas where topset beds have been eroded, the resultant angular unconformities result in abrupt seaward shifts (up to 25 km) of the apparent paleo-continental shelf edge (Fig. 11).

On shelf seismic lines that are parallel to the paleo-shelf edge, acoustically diffuse lens-shaped

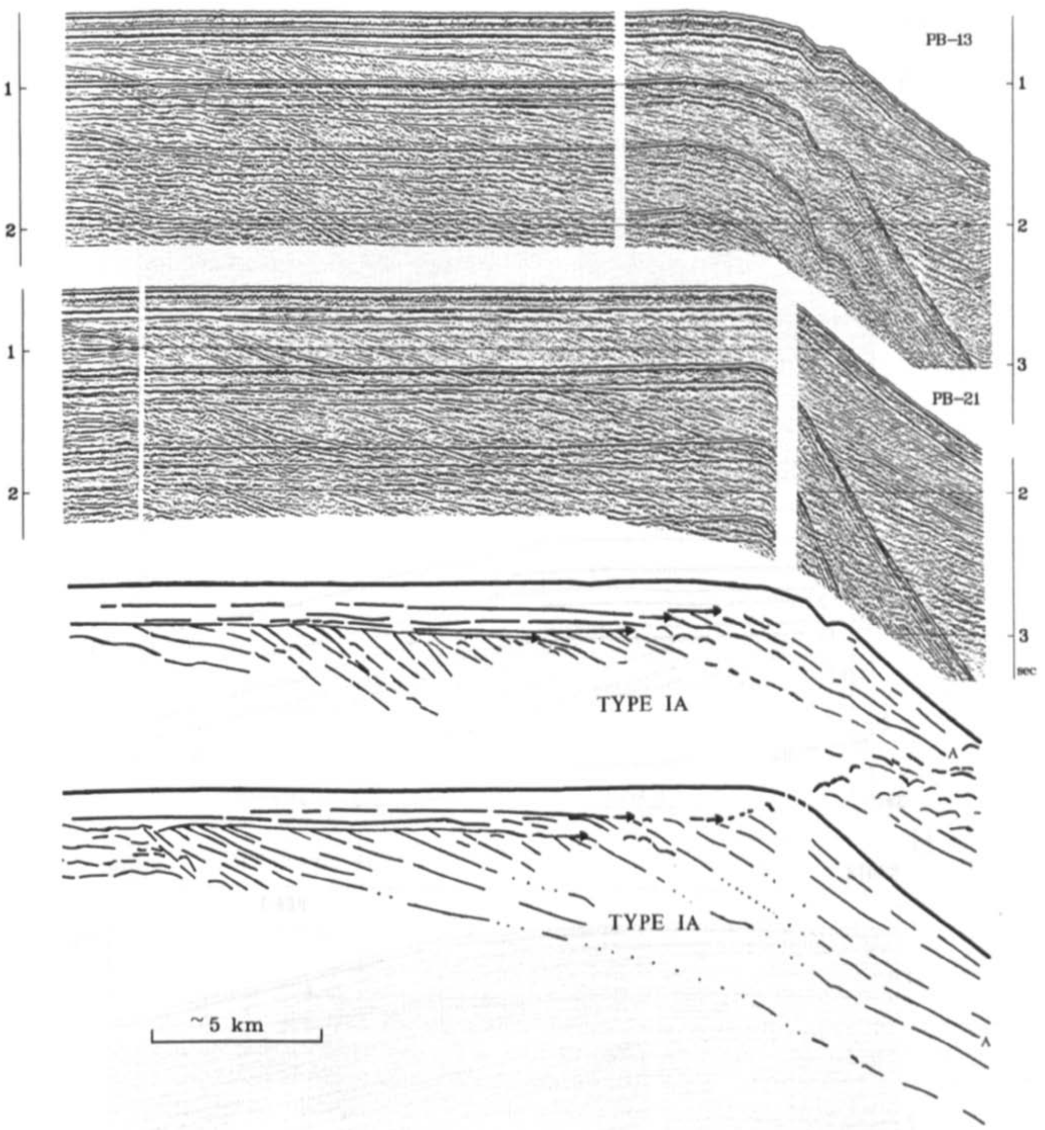


Fig. 9. Seismic data and line drawings showing examples of type IA sequences in Prydz Bay. Here, as is common around Antarctica, type IA sequences have thin, stacked topset strata (up to 250 m thick) with high-amplitude reflections from high-velocity beds (up to 2.6 km/s). Foreset strata are highly eroded and have variable dips at unconformities. Mounded reflections occur near preserved paleo-shelf edges. Arrows denote inferred extent of former grounded ice sheets to paleo-shelf edges. A is a regional reflector (see text). The water depth at the shelf edge is about 400 m. See Fig. 15 for location.

bodies up to 70–80 km wide and up to 150 m thick occur within the type IA sequences (Bartek, 1989; Kuvaas and Kristoffersen, 1991; Cooper et al., in press). These bodies cross the shelf and are similar in width to the large ice streams within the Ross embayment (Alley et al., 1989) and to large present-day glacially carved troughs (Kuvaas and Kristoffersen, 1990; Hambrey, 1991), and are wider than fluvial entrenched valley systems of non-glaciated margins (Bartek et al., 1991).

Some geometric features of type IA sequences may be common in Antarctica, and other polar margins, but missing, or rare in prograding sequences from low-latitude margins:

(1) Areal reflection geometries indicating deposition from a "line source" at the continental shelf edge, such as the front of an ice sheet, rather than from a point source such as a submarine canyon (Haugland et al., 1985; Cooper et al., 1991a).

(2) Mounded reflections, with locally steep dips,

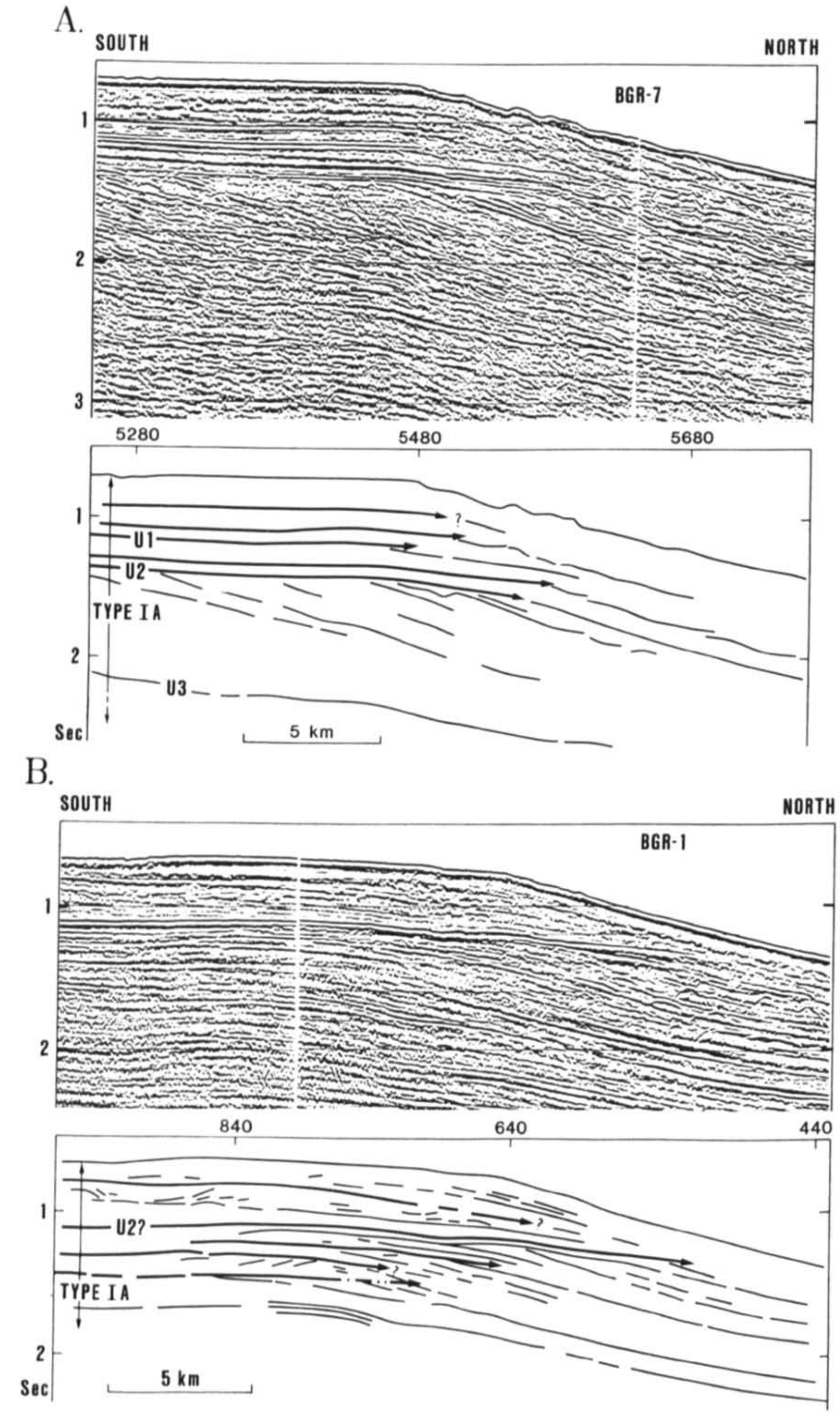


Fig. 10. Seismic data and line drawings showing examples of type IA sequences beneath the outer shelf in the Ross Sea. Topset strata are uncommonly thick with many channel and mound features. High-amplitude topset reflections commonly terminate abruptly near inferred paleo-shelf edges. Relative dips of foreset reflections are variable. Arrows denote inferred extent of former grounded ice sheets to paleo-shelf edges. Water depths at the shelf edge are about 500 m. See Fig. 12 for location.

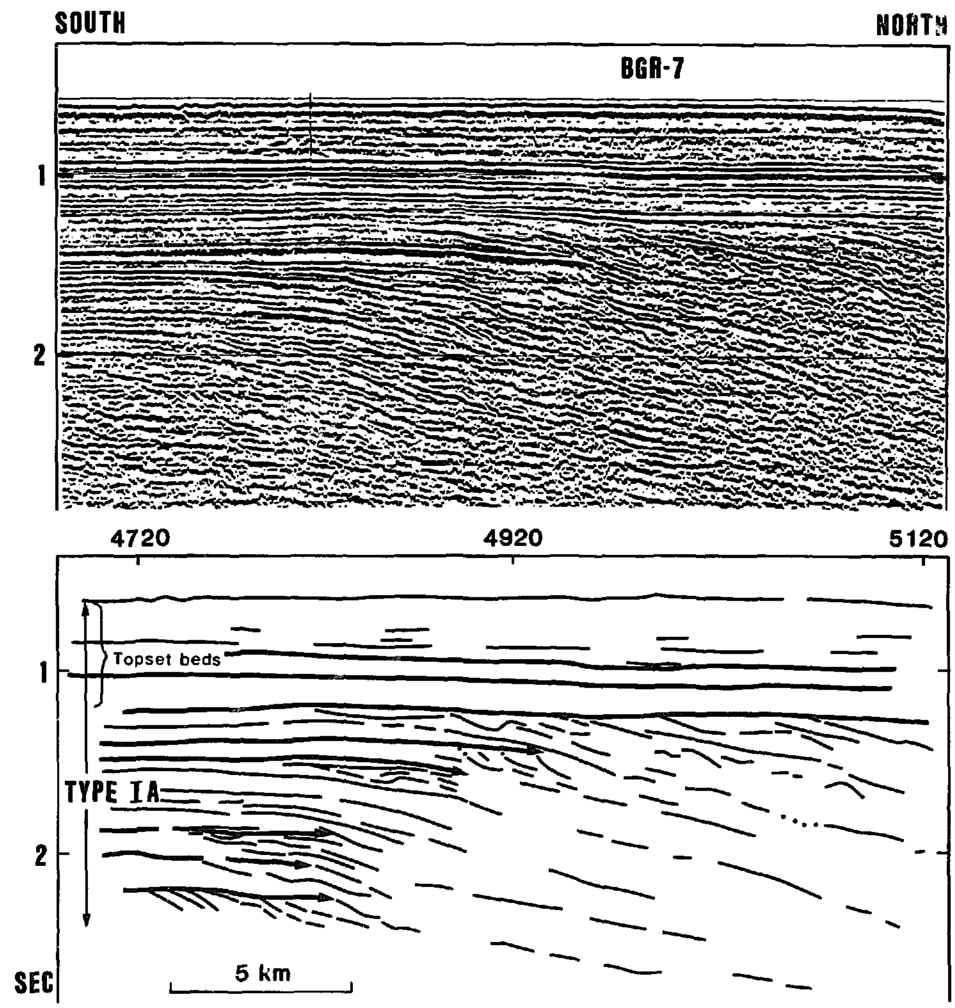


Fig. 11. Seismic data and line drawing showing examples of type IA sequences beneath a mid-shelf area of the eastern Ross Sea and directly seaward of Fig. 8. Like the outer shelf (Fig. 10), high-amplitude topset reflections commonly terminate near paleo-shelf edges, where mounded and channel reflections also occur. The relatively large-scale aggradation of type IA sequences here (e.g. SP 4820–4920) is uncommon elsewhere around Antarctica; extensive apparent prograding (e.g. SP 4920–5120) is most common. Arrows denote inferred extent of former grounded ice sheets to paleo-shelf edges. See Fig. 12 for location.

from small buried "ridges" lying directly above and below the abrupt seaward end (paleo-shelf edge) of a high-amplitude topset reflection; such features may be, respectively, lift-off moraines (King and Fader, 1986; Vorren et al., 1989; Fig. 8, 10 and 11) and diamictite aprons (Kuvaas and Kristoffersen, 1990; Hambrey et al., in press) from a previously grounded ice sheet.

(3) On survey lines parallel to the paleo-shelf edge, the presence of acoustically diffuse and broad lens-shaped reflection bodies that areally lead to the prograding sequences (Bartek, 1989); these may be channels that once carried grounded ice (and entrained sediment) to the shelf edge (King and Fader, 1986; Bartek et al., 1991).

Antarctic type IIA sequences

Type IIA acoustic sequences are relatively uncommon in Antarctic sedimentary sections of possible Cenozoic age beneath the outer continental shelf, but their geometry is relatively simple. We identify the type IIA sequences by (a) relatively gently dipping and continuous reflections that principally aggrade the buried paleo-shelf edges, (b) the absence of abrupt angular unconformities and

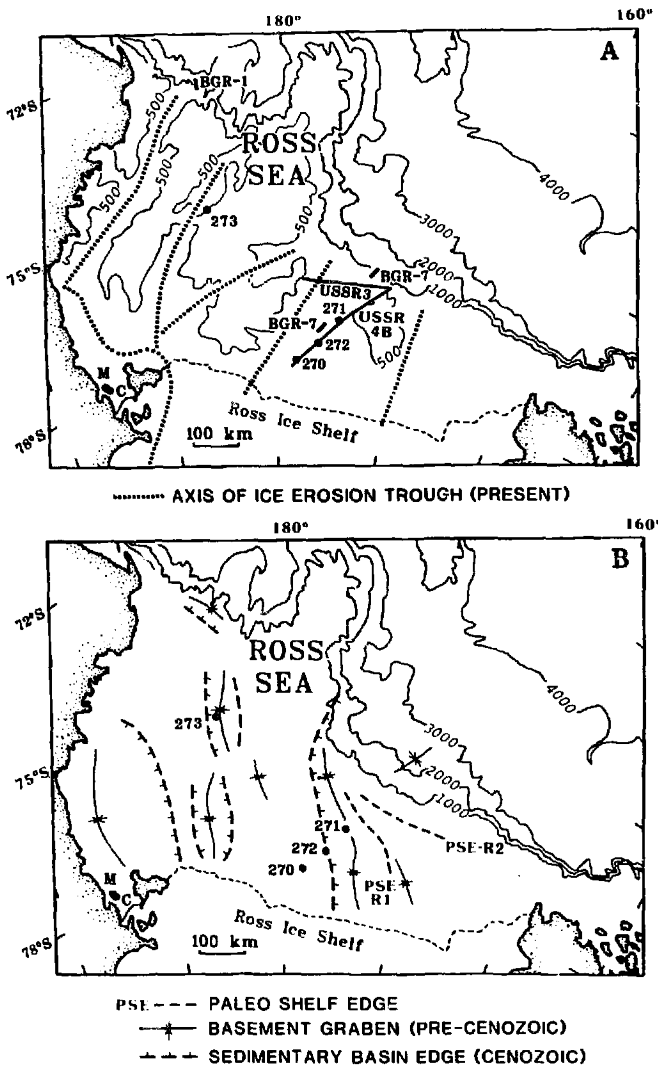


Fig. 12. Index maps of the Ross Sea. (A) Bathymetry and location of drill sites and seismic lines. Water depths increase landward from about 500 m at the shelf edge to nearly 1000 m in front of the Ross Ice shelf. (B) Edges of sedimentary basins, axes of structural lows, and paleo-shelf edges for unconformities U3 (PSE-R1) and U5 (PSE-R2) in Fig. 13. See Fig. 1 and 5 for regional continuation of structural and ice erosion features.

terminated topset reflections near the paleo-shelf edges such as those found in type IA sequences, (c) the downlap of overlying type IA strata onto these sequences, and (d) the truncation of the type IIA sequences by overlying type IA sequences.

The best examples of Cenozoic type IIA sequences are in the eastern Ross Sea, but they probably also exist beneath the Prydz Bay and possibly the Wilkes Land margins (Fig. 2) and elsewhere. These sequences may be more common, but are difficult to identify at great depths because of the severe multiple noise that is common in seismic profiles across the Antarctic continental margin. Also, drilling has not revealed the age of sequences with type IIA geometries, as in Wilkes Land. Sequences that are transitional between type IIA and IA are observed in the eastern Ross Sea (Fig. 8) and in Prydz Bay (described below), and these sequences have greater geometric complexity than type IIA.

Interpretations of type IA and IIA sequences

The explanation for the stratal geometries that we describe as type IA and IIA is uncertain and is currently being widely debated (e.g., Cooper and

Webb, 1990). The apparent similarity of these geometries (for some investigators) with prograding sequences on low-latitude margins has led to the interpretation that they are fluviomarine delta lobes composed of glacial sediments derived from icebergs, coastal currents, and possibly Antarctic bottom waters (e.g., Hinz and Kraus, 1982; Hinz and Block, 1984; Hinz and Kristoffersen, 1987). Other workers, however, including ourselves who see some important geometric (and geologic) differences from low-latitude margins, see a different interpretation evolving. A currently favored interpretation is that the acoustic geometries that we call type IA are caused principally by sediment deposition from grounded ice sheets that have episodically advanced regionally seaward to the continental shelf edge since late Paleogene time; most deposition occurs directly from the base of the ice sheet onto the upper continental slope (Haugland et al., 1985; Larter and Barker, 1989; Barron et al., 1989; Cooper and Webb, 1990; Hambrey et al., 1991, in press; Cooper et al., 1991a, b; Bartek et al., 1991; Kuvaas and Kristoffersen, 1991).

Although type IA sequences are widely seen, the detailed areal stratigraphy of these sequences is still poorly known. Where sampled by offshore drilling beneath the Ross Sea and Prydz Bay continental shelves, Antarctic type IA and IIA sequences are composed exclusively of Cenozoic glacially derived deposits younger than middle to late Eocene age. Our interpretations, and those previously published, regarding how ice-volume changes in Antarctica may relate to sea-level fluctuations, remain mostly untested by sampling and scientific drilling.

Below, we show seismic reflection profiles from the Ross Sea and Prydz Bay, where the Antarctic type IA and IIA sequences have been drilled to further illustrate their characteristics and variabilities.

Ross Sea

The Ross Sea lies at the seaward end of the major physiographic and structural embayment separating East and West Antarctica (Fig. 1). The Ross Sea is inferred, mostly from seismic reflection and nearby geologic data, to have had an episodic rift history extending from possibly early Mesozoic time to the present (Davey et al., 1982; Cooper et al., 1987a, b). Two principal rift episodes have been proposed: (1) early-rift downfaulting of the large basement grabens throughout the Ross Sea in Early–Late Cretaceous time and (2) late-rift volcanism and development of the Terror Rift (western Ross Sea) and subsidence of the eastern Ross Sea during the Eocene to the present (Cooper et al., 1987a, 1991b).

The eastern and western parts of the Ross Sea differ structurally (Davey, 1987; Cooper et al., 1987b). The western Ross Sea is incised by narrow sediment-filled early-rift grabens, and the most western one has late-rift and active volcanism and downfaulting (Fig. 2 (section 3) and 12). The eastern Ross Sea is also underlain by narrow early-rift grabens filled with possible late Mesozoic and early Cenozoic strata (Cooper et al., 1991b) that lie below a deeply buried regional unconformity (U6; Hinz and Block, 1984; Fig. 2, section 4). Regional subsidence of the eastern Ross Sea since at least late Oligocene time (Hinz and Block, 1984) and possibly Eocene time (Cooper et al., 1991b) has resulted in a broad trough that is filled with thick prograding sedimentary deposits (Fig. 2, section 3). Hinz and Block (1984) defined five glaciofluvial delta sequences in this area (Fig. 2, section 4) that they suggested, based on DSDP drilling, to have been deposited since late Oligocene time (Hayes and Frakes, 1975). Sato et al. (1984) documented a similar geometry.

We suggest a modified interpretation for the eastern Ross Sea type IA and IIA sequences, based on recent seismic and drilling results in Prydz Bay (Cooper et al., 1991a) and on regional syntheses of Ross Sea and Antarctic drilling (Barrett et al., 1989; Hambrey et al., in press; Barron et al., 1991). The seismic data in the eastern Ross Sea (line USSR-4B, Figs. 13 and 14) illustrate at least nine type IA sequences (above unconformity U5) beneath the mid- to outer continental shelf. We believe that the complex type IA geometries of these paleo-shelf edge reflections are caused by sediments derived from the base of a grounded ice sheet extending to the continental shelf edge, and by the glacial depositional processes described by

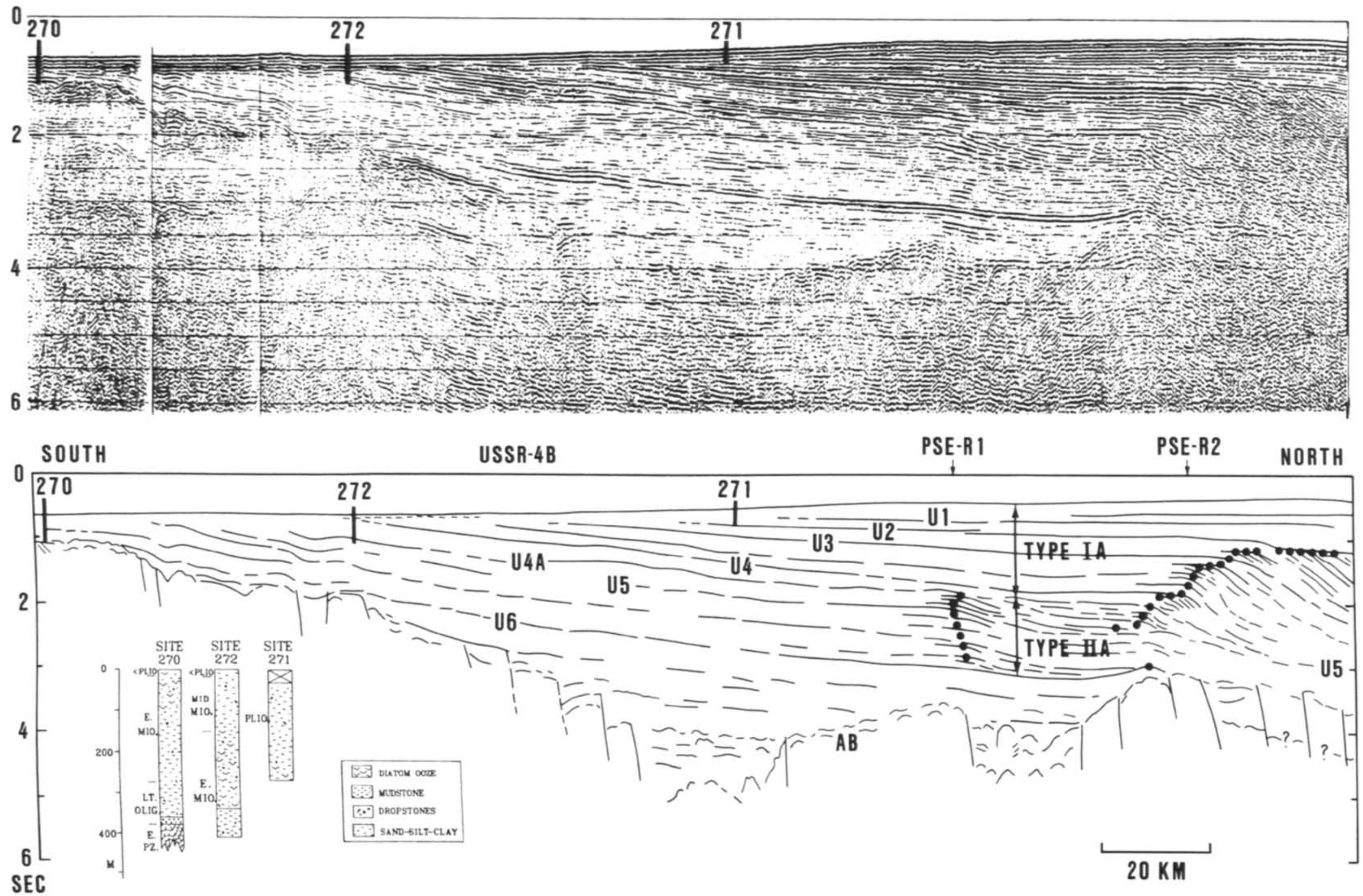


Fig. 13. Seismic data and line drawing showing Cenozoic type IA and IIA sequences (i.e. above U6) along line USSR-4B in the eastern Ross Sea. U1 to U6 are unconformities of Hinz and Block (1984). All sequences above U6 are of likely glacial origin (Hayes and Frakes, 1975). We suggest that the transition from type IIA to type IA sequences (see also Fig. 8) indicates a major early Miocene change in glacial depositional environments at the paleo-shelf edge. Type IIA sequences may result from distal grounded ice sheets on a normal depth shelf and type IA from ice sheets grounded out to the shelf edge on an overdeepened shelf. See Fig. 12 for location.

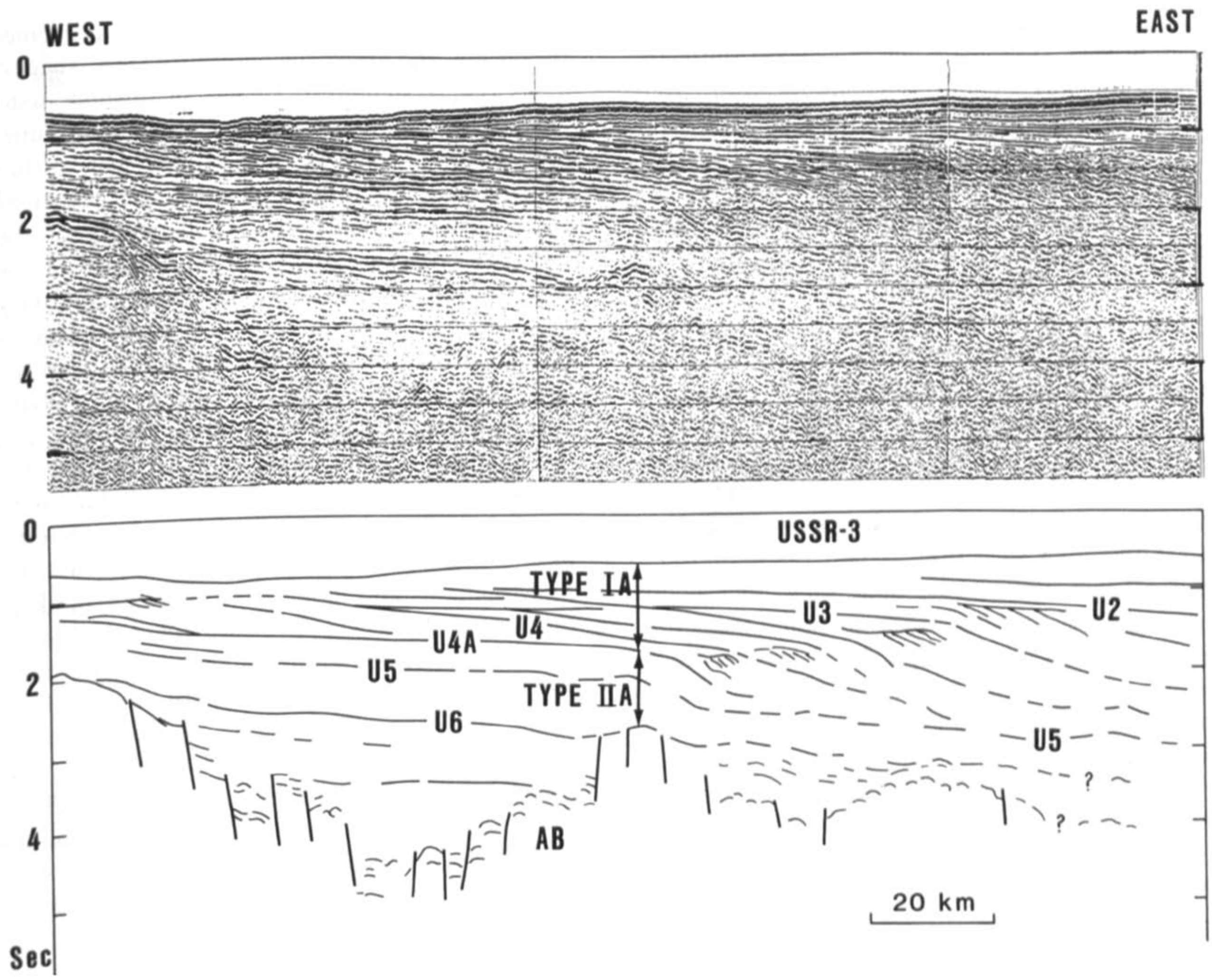


Fig. 14. Seismic data and line drawing showing Cenozoic type IA and IIA sequences (i.e. above U6) along line USSR-3 in the eastern Ross Sea. Extensive prograding and erosion of paleo-foreset strata occurs almost exclusively in type IA strata above U4A. See Fig. 13 for explanation and Fig. 12 for location.

Hambrey et al. (1991, in press) based on drilling in Prydz Bay.

We also suggest that the change in acoustic geometry of the paleo-shelf edge reflections from type IIA to IA (Fig. 8) signals the onset of major glacial advances of grounded ice (and entrained sediment) to the continental shelf edge. Grounded ice sheets may have existed previously, however, on an overdeepened inner shelf (Barrett et al., 1989; Bartek and Anderson, 1990). The initial advance of grounded ice sheets toward the shelf edge may also explain the "transitional" geometries upsection between type IIA and IA. The transition is marked by an abrupt seaward and downward shift of the paleo-shelf edges shortly after U5 time (Fig. 8) and could be caused by (1) a lowering of sea level due to ice buildup, and deposition of lowstand deposits seaward of the shelf edge, or (2) overdeepening of the continental shelf by the flexural loading of the advancing ice sheet, and deposition of distal glacial sediments on the overdeepened outer shelf and upper slope. We favor the overdeepening explanation based on comparison with Prydz Bay, where a stratigraphic sequence that may be equivalent was drilled at ODP Site 742 and ODP Site 739 (discussed below), and showed an upsection and seaward deepening of glacial depositional environments. Further support for overdeepening of the continental shelf on a long-term basis is the lack of any major downward shifts in the paleo-shelf edges above unconformity U4A (Fig. 13). Such downward shifts are commonly observed in lowstand deposits from normal depth, low-latitude continental shelves (Vail et al., 1977; Wilgus et al., 1988; Bartek et al., 1991).

Good acoustic evidence is found in the Ross Sea for erosion and deposition from likely grounded ice sheets (Hayes and Davey, 1975; Anderson et al., 1984; Karl et al., 1987; Bartek et al., 1991; Fig. 10B, 13 and 14), which we suspect has occurred on an overdeepened shelf, like today. Type IA sequences show clear evidence, not seen at great depth (i.e. below U5), for abrupt seaward-prograding, truncated foreset strata, and thick depositional banks composed of topset strata beneath the outer shelf (Fig. 11 and 13). Within the upper 1.0 s, the seismic data show that many sequences are cut by a strong erosional unconformity that truncates foreset strata. The thin topset strata, which are only about 0.1–0.2 s thick, may record cycles of grounded ice-sheet advance, lift-off by floating, and retreat (Fig. 9 and 10). Such cycles are similar to those proposed by Larter and Barker (1989) for the Antarctic Peninsula, by Hambrey et al. (1991, in press) for Prydz Bay, by Bartek et al. (1991) for the eastern Ross Sea, and by Kuvaas and Kristoffersen (submitted) for the Weddell Sea.

The general NE–SW orientation of Cenozoic glacial features in the Ross Sea differs from the N–S orientation for Mesozoic(?) early-rift basement horsts and grabens that lie below unconformity U6 (Fig. 12 and 13; Cooper et al., 1987a, 1991b). Cenozoic glacial banks and troughs, which indicate the direction of ice flow (Hughes, 1973), generally lie parallel to the likely Cenozoic coastline along the Transantarctic Mountains of the western Ross Sea (Behrendt and Cooper, 1991). Beneath the eastern Ross Sea, the paleo-shelf edge for type IA sequences at U3 time, and probably earlier at U4A time, lies normal to this flow direction (Fig. 12 and 13). However, the paleo-shelf edge at U5 time, although a less well-defined feature, more closely follows the orientation of underlying basement structures.

The different orientations of the paleo-shelf edges for U3 (U4A) and U5 times (PSE-R1 and PSE-R2, Fig. 12), and the different acoustic geometries below and above unconformity U4A, suggest to us that a significant change in offshore sedimentation patterns occurred at about U4A time, or in the early Miocene (Fig. 13). We suspect that offshore glaciation and sedimentation increased and was accompanied by overdeepening of the shelf in the eastern Ross Sea. Subsidence of the eastern Ross Sea at this time appears relatively rapid, compared to the central Ross Sea, and is the likely result of differential extension within the West Antarctic Rift (Fig. 1; Cooper et al., 1991b).

In general, the glacial erosional troughs and their cogenetic depositional banks parallel the trend of buried, or glacially exposed, basement rises where there is evidence of Cenozoic tectonic or erosional uplift, such as along the Transantarctic Mountains and along parts of Iselin Bank in the central Ross Sea. Evidence for older broad troughs now buried within the sedimentary section are observed in high-resolution seismic data (Karl et al., 1987; Bartek et al., 1991), and these troughs may have been the conduit for ice streams that once extended across the shelf (Bartek et al., 1991). The orientation of the buried troughs and banks is not known, but thick (up to 800 m) outer shelf topset banks have been deposited since Pliocene time (i.e. above unconformity U2, Fig. 13).

Prydz Bay

Prydz Bay lies at the seaward end of the onshore, 700 km long Lambert Graben, and is underlain by a NE–SW sediment-filled structural basin. (Fig. 1; Stagg, 1985; Cooper et al., 1991a). The Lambert Graben is a late Paleozoic(?) and Mesozoic rift structure that now drains nearly 20% of the East Antarctic Ice into the Amery Ice Shelf, and previously across Prydz Bay to the outer shelf edge (Fig. 15; Hambrey, 1991). ODP Leg 119 drilled a transect of holes across Prydz Bay and recovered a diverse suite of Paleozoic(?) and Mesozoic non-marine and Cenozoic glacimarine rocks (Fig. 16; Barron et al., 1989). Prydz Bay is a Paleozoic(?) to Mesozoic rifted margin, like the formerly adjacent India (Fig. 3), that is now mantled on its seaward edge by thick type IA sequences of early Oligocene and younger glacial marine

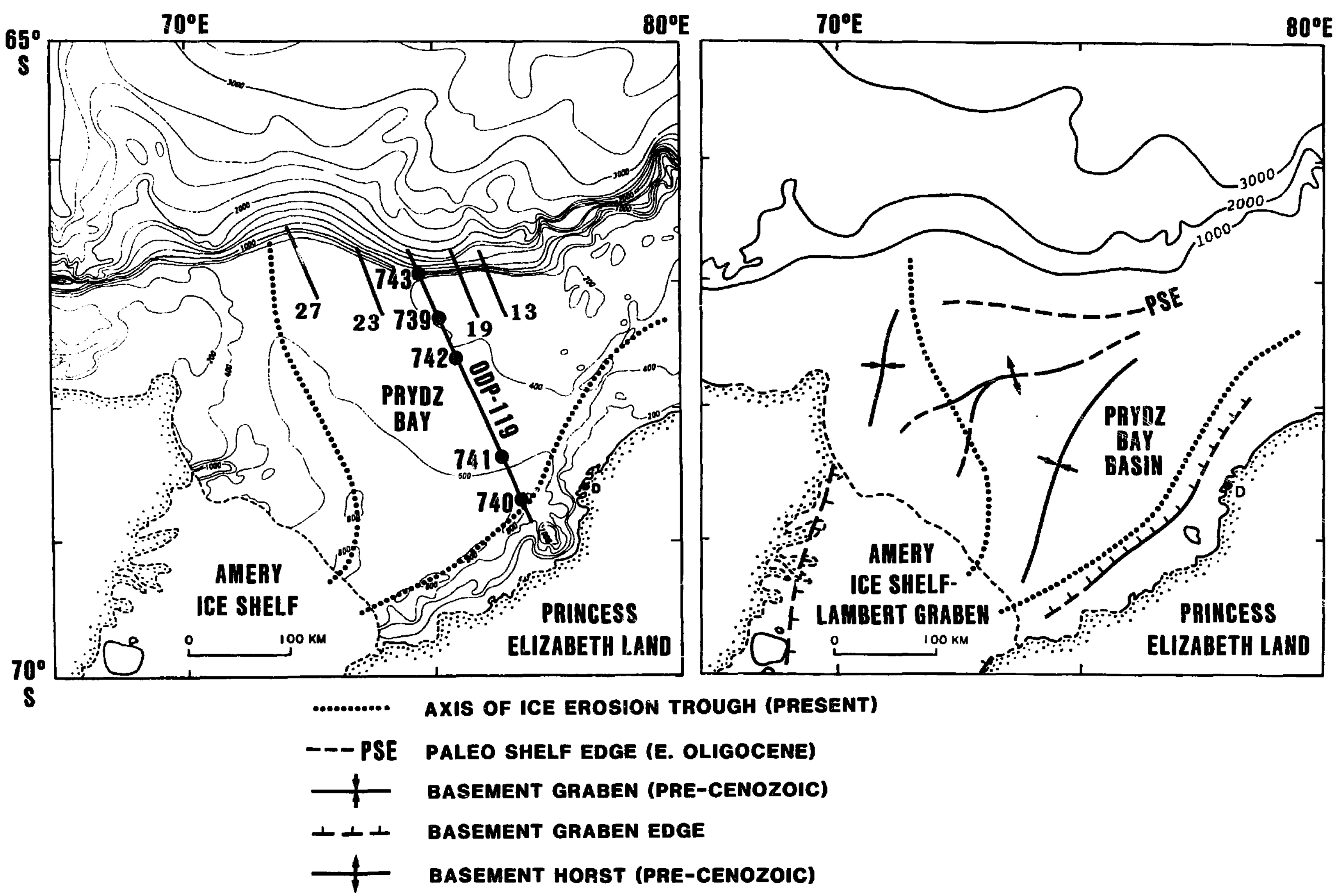
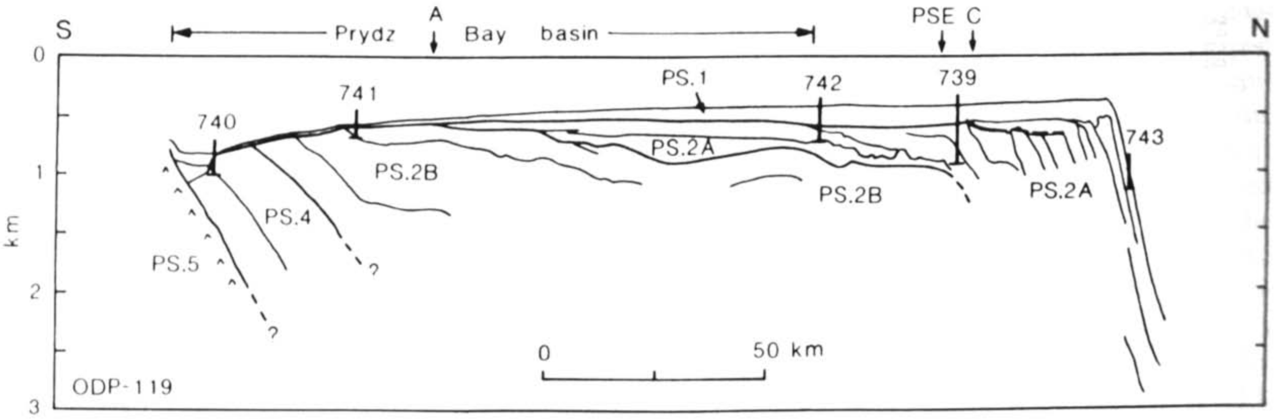


Fig. 15. Maps of Prydz Bay. (A) Bathymetry and locations of profiles cited in the text. (B) Locations of major Mesozoic and older basement structures and the early Oligocene paleo-shelf edge (*PSE*) in Prydz Bay. PSE and PSE-1 (Figs. 17 and 18) are equivalent. Lines ODP-119 and PB-21 are the same.



GLACIAL: PS.1 and PS.2A - late Eocene —early Oligocene to Holocene marine diamictite and minor diatomaceous sediment
NON-GLACIAL: PS.2B - Early Cretaceous nonmarine sandstone and siltstone
PS.4 - Permian(?) to Mesozoic red siltstone and sandstone
PS.5 - Precambrian(?) metamorphic rocks
PS.6 - Mesozoic(?) or Precambrian intrusive or metamorphic rocks

Fig. 16. Transect across Prydz Bay showing locations of ODP Leg 119 drill sites and ages and compositions of acoustic units PS.1 to PS.4 that were drilled (from Cooper et al., 1991a). The unconformity between PS.2B and PS.2A is the inferred preglacial to glacial boundary. See Fig. 17 for description of type IA and IIA sequences and Fig. 15 for transect location.

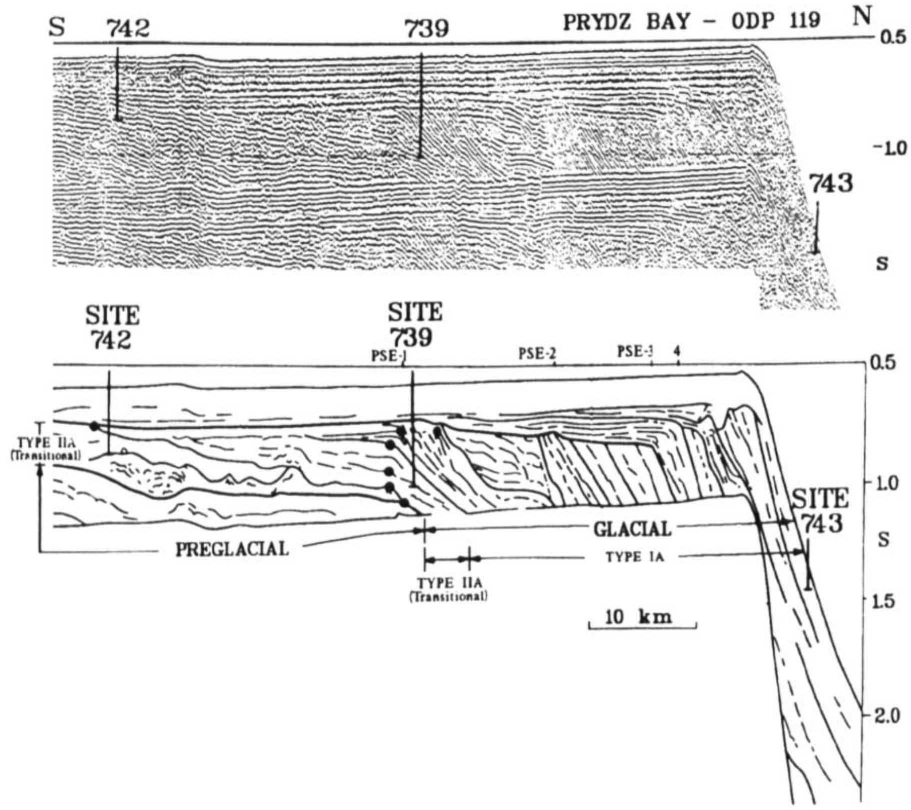


Fig. 17. Seismic data and line drawing across the type IA and IIA (transitional) glacial sedimentary sequences underlying the outermost shelf beneath Prydz Bay (modified from Cooper et al., 1991a). The upsection change from type IIA to IA geometries generally corresponds to a transition from early glacial conditions (i.e. ice sheets of limited extent) to full glacial conditions (i.e. ice sheets grounded to the paleo-shelf edge) based on inferences from rock facies, downhole logging, and technical engineering data at Sites 739 and 742 (see text). See Fig. 15 for location.

strata (Leitchenkov et al., 1990; Cooper et al., 1991a).

Seismic reflection profiles across the Prydz Bay outer continental shelf illustrate at least seven sequences seaward of PSE-1 (Figs. 9, 17 and 18) that we consider to be type IA sequences. Landward of PSE-1, buried sequences below flat-lying acoustic unit PS.1 (Figs. 16 and 18) are similar to the "transitional" sequences that are observed in the Ross Sea between type IIA and IA sequences (Fig. 8B), based on the gently dipping, aggrading reflections and on the likely seaward and downward shift in our inferred paleo-shelf edges (Fig. 17). The transition is not as clear here as in the Ross Sea, partly because the Prydz Bay sequences have greater internal structure and poorer reflection continuity than in the Ross Sea. Also, in Prydz Bay, the transitional, possibly type IIA sequences [i.e. type IIA (transitional)] lie directly on the inferred unconformity between pre-glacial (unit PS.2B) and glacial (unit PS.2A) rocks, but in the Ross Sea a thick section of type IIA sequences lies above the inferred pre-glacial to glacial unconformity (i.e. U6, Fig. 13).

Seismic tie lines are inadequate for accurately mapping the areal geometry of individual sequences in Prydz Bay, yet the general features, such as paleo-shelf edges and major dip changes, can be traced across several profiles for up to 200 km (Fig. 15; Cooper et al., 1991a). Dips of the individual sequences are variable, and at ODP Sites 739 and 742 the sequences generally strike parallel to the paleo-shelf edges. Foreset reflections can sometimes be traced seaward beneath the overlying sequences to where they either downlap onto deeper unconformities or continue into the Southern Ocean Basin, such as reflector A of Mizukoshi et al. (1988) (Fig. 18).

Acoustic sequences that were sampled beneath the outer continental shelf at ODP Sites 739, 742

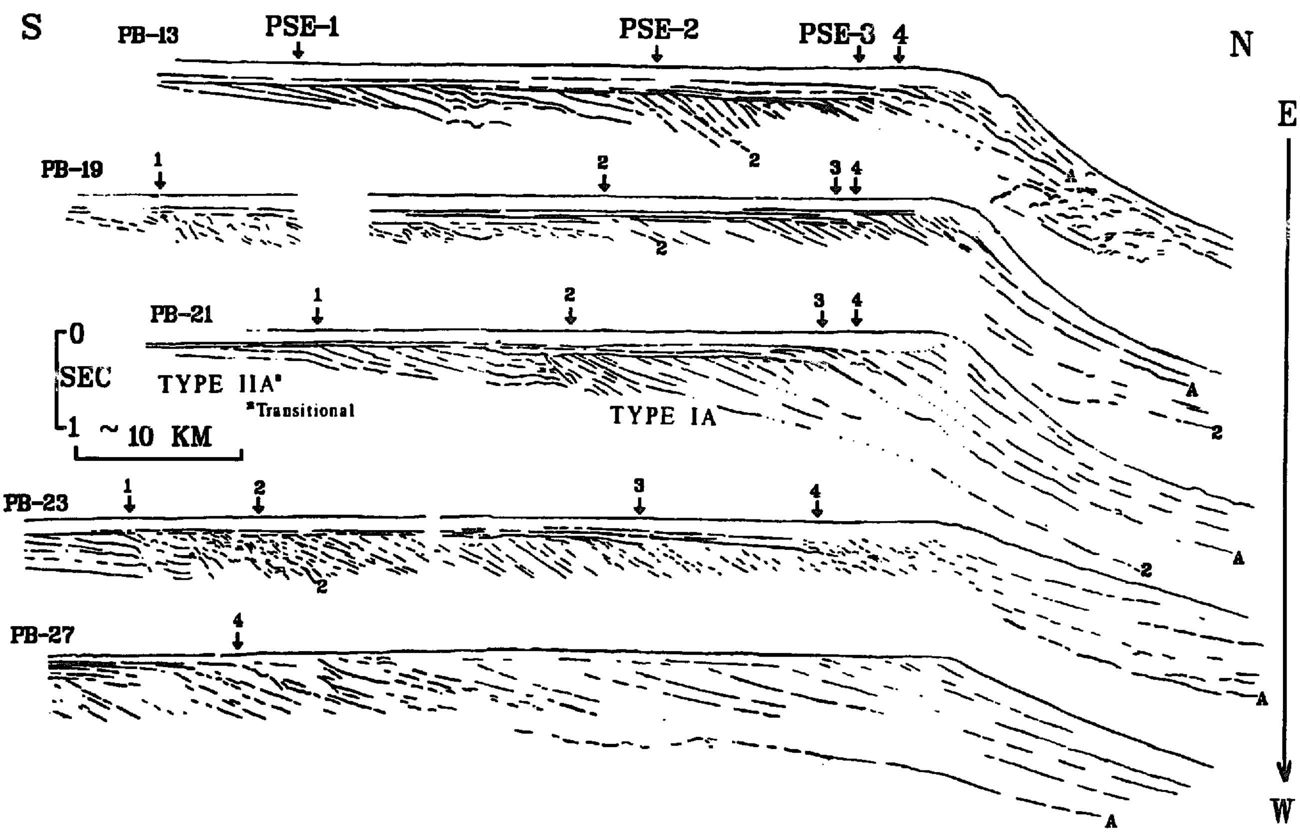


Fig. 18. Line drawings of five seismic profiles across the outer Prydz Bay shelf illustrating the type IA glacial sequences seaward of PSE-1 and type IIA (transitional) sequences landward. The correlations of paleo-shelf edges (*PSE*) are based on similarities in acoustic geometry because only limited tie lines are available. Reflector A is a regional reflector from Mizukoshi et al. (1988). Seismic data for part of lines PB-13 and PB-21 are shown in Fig. 9, and profile locations are shown in Fig. 15.

and 743 are composed of glacial deposits, mostly diamicts (Barron et al., 1989; Hambrey et al., 1991) ranging in age from middle(?) Eocene to the present (Barron et al., 1991). The compositions and ages of nearly all sequences seaward of Site 739 (i.e. type IA) are unknown, but are presumed to be of glacial origin and younger than early Oligocene in age (Cooper et al., 1991a). The possibility of interbedded non-glacial deposits cannot be excluded.

Our boundary between type IA and the type IIA (transitional) sequences in Prydz Bay corresponds to the apparent unconformity at about 315 m below seafloor at ODP Site 739 (Fig. 19) separating overlying, more steeply dipping reflections from underlying, gently dipping reflections. Distinct changes in lithology (e.g., clay content) and downhole logging measurements (e.g., velocity gradients) are also noted at or very close to this unconformity (Barron et al., 1991; Cooper et al., 1991a; Fig. 19). Also, overlying reflections downlap onto the unconformity.

Most diamicts sampled from the seaward-dipping parts of the type IA and type IIA (transitional) sequences (i.e. below flat-lying unit PS.1; Fig. 16) are interpreted as water-lain tills and proximal glacimarine sediments deposited near or at the front of a grounded ice sheet (Hambrey et al., in press). Contorted glacial sediments of glacimarine/lacustrine and possibly fluvial origin were sampled about 100 m above the inferred base of the glacial section and from the top of a regionally distorted acoustic unit (i.e. from a type IIA (transitional) sequence). The ODP lithologies and acous-

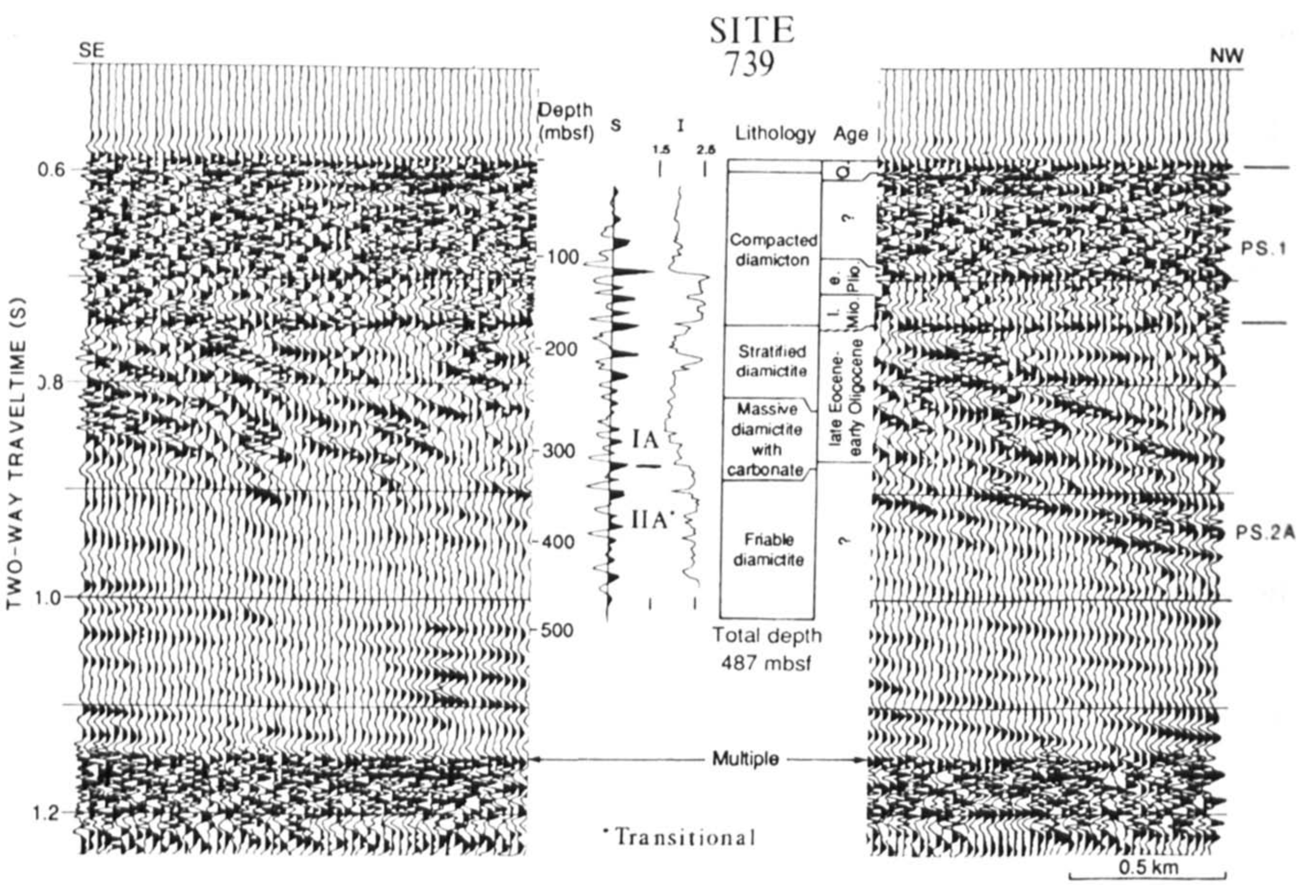


Fig. 19. Seismic reflection profile, drill core data, downhole logging velocities (*I*, km/s), and synthetic seismic trace (*S*) for ODP Site 739 in 422 m of water on the outer Prydz Bay shelf (modified from Cooper et al., 1991a). Distinct acoustic, logging, and lithologic changes occur at our boundary between type IA and IIA sequences. Flat, high-velocity (2.6 km/s) topset beds and multiple overcompaction events (Solheim et al., 1991) indicate episodic grounding of ice sheets at this site since at least early Oligocene time. See Fig. 15 for location.

tic stratigraphy are consistent with (1) ice sheets advancing onto, and probably deepening, the continental shelf with deposition of transitional-IIA sequences starting in mid- to late Eocene time, and (2) grounded ice sheets reaching the continental shelf edge, with deposition of type IA sequences by earliest Oligocene time (Cooper et al., 1991a; Hambrey et al., in press).

The age of type IA sequences beneath most of the outer Prydz Bay margin is unknown from drilling, and is not presently well constrained by seismic reflection data (Cooper et al., 1991a). Relatively soft sediments of late Miocene age were drilled at Site 739 from a 4 m thick layer about 35 m above a major angular unconformity, between units PS.1 and PS.2A (Fig. 16 and 19) underlying the outer shelf. We infer from this sediment age and from tracing the unconformity to point PSE-3 on the outer shelf (Fig. 17 and 18) that the shelf edge has differentially prograded across Prydz Bay since late Miocene time, this progradation ranging from about 6 km in the east (line PB-21) to about 50 km in the west (line PB-27). The inferred age for PSE-3 is uncertain, however, because the thin late Miocene layer cannot be definitively traced in seismic data for more than 1 km seaward from the drill site (Cooper et al., 1991a).

Systematic along-strike variations are observed within type IA sequences regarding dips of foreset beds and or paleo-continental slope surfaces, and regarding the thickness of the topset strata that form outer shelf banks. The foreset dips decrease westward (Fig. 18) on approaching the trough-mouth fan that lies directly seaward of the wide glacial erosional channel that presently extends N–S and landward across the continental shelf to the Amery Ice Shelf (Fig. 15). This channel and the fan may have been the principal outflow of grounded ice and the depositional site for entrained sediment during all glacial maxima (Hambrey, 1991; Hambrey et al., in press). The topset bed thickness also decreases westward toward the axis of this glacial erosional channel, suggesting to us that the topset strata have been eroded, or fewer topset strata have been deposited, by the moving ice formerly filling the channel.

Glacial and structural features observed in Prydz Bay seismic data commonly have different orientations, as in the Ross Sea. The regional trend of the present and paleo-shelf edges of the type IA sequences in Prydz Bay diverges by about 45° from the trend of underlying basement structures (Fig. 15). Also, the broad ice-erosional channel that crosses the shelf begins along the crest of, and then cuts across and erodes into, the basement ridge that forms the seaward edge of the Paleozoic(?) to Mesozoic Prydz Bay Basin (Shelestov and Alyavdin, 1987; Cooper et al., 1991a). The large divergence between the Cenozoic glacial trends and Mesozoic structural trends beneath the shelf indicates that factors other than rift tectonism, such as lithospheric flexure, paleoceanography and deposition–erosion, have greater influence on the direction of major ice flow and subsequent deposition of entrained sediment into the thick type IA sequences of the outer continental margin. Some deep glacial erosional troughs around Prydz Bay do, however, follow basement trends, in particular the deep erosional trough lying adjacent to the coast and along the southern edge of the Prydz Bay Basin (Fig. 15).

Discussion

Structural and depositional controls

Cenozoic sedimentary sequences beneath the Antarctic outer continental margin appear strongly affected by the depositional and erosional actions of former grounded ice sheets. These former ice sheets followed topographic depressions that formed by tectonism, erosion and deposition. Some Paleozoic(?) and younger structures, such as the Lambert Graben, Transantarctic Mountains, Ross and Weddell embayments, and Antarctic Peninsula mountain valleys, etc., have controlled the flow directions of these ice sheets. Yet, other large structures, such as Mesozoic shelf basins, have commonly exerted little control on the direction of ice flow, especially offshore. Hence, the orientation and location of erosional channels and depositional banks leading to, and overlying, the Cenozoic prograding sequences commonly differs from those of underlying shelf basin structures, as in Prydz Bay and the Ross Sea (Fig. 12 and 15). The trends of paleo-shelf edges for the Cenozoic that have prograded the continental shelf outward by up to 85 km are mostly the same as those for older shelf-edge structures, with the possible exception of the eastern Ross Sea where early and late Cenozoic trends of paleo-shelf edges may differ by up to 45° (Fig. 12).

Former topographic gradients, which controlled the movements of ice and entrained sediment, were apparently different from those of today in some places around Antarctica, probably due to the different effects of ice loading and tectonic subsidence. The ubiquitous acoustic geometries (i.e. type IA) of likely Cenozoic sedimentary sequences lying beneath the Antarctic continental margin have some acoustic similarities with relatively high-energy lowstand deposits on non-polar margins (Kuvaas and Kristoffersen, 1991; Bartek et al., 1991). However, we believe that the type IA sequences were deposited from grounded ice sheets, especially in view of (1) the present overdeepening of the continental shelf, which we suspect also occurred during, and possibly throughout, earlier glacial times (see above), (2) the high acoustic velocities (up to 2.6 km/s, Fig. 19) observed within 200 m of the seafloor (Haugland et al., 1985; Larter and Barker, 1989; Cooper et al., 1987b, 1991a; Cochrane and Cooper, 1991), and (3) the geotechnical evidence for multiple overcompaction events in glacial sediments recovered in Prydz Bay drill cores (Solheim et al., 1991).

The type IA geometries from passive margin segments of the Weddell Sea, Dronning Maud Land, Prydz Bay and Wilkes Land (i.e. large-scale progradational with generally thin topset strata and many unconformities) are consistent with generally slow Cenozoic subsidence of the outer shelf, following margin breakup in Mesozoic time, and with an intermittant but abundant supply of sediment. The sediment was probably carried to the shelf edge by grounded ice sheets that moved across and flexurally depressed (i.e. overdeepened) the continental shelves. The Wilkes Basin and parts of the Wilkes Land margin may also have been further downwarped due to the Cenozoic uplift of the adjacent Transantarctic Mountains (Stern and Ten Brink, 1989).

Two areas of the Antarctic margin, the Antarctic Peninsula and eastern Ross Sea, have type IA sequences with relatively well preserved topset beds. These regions are suspected to have undergone rapid Cenozoic subsidence, relative to the rate of sediment being supplied to the outer shelf (Hinz and Block, 1984; Larter and Barker, 1989; Cooper et al., 1991b). Here, like elsewhere around Antarctica, grounded ice sheets are suspected to have been the principal agent for carrying sediment to the shelf edge during later parts of the Cenozoic. The time of first ice sheet grounding to the shelf edge in these two regions is believed to be late Miocene to Pliocene time in the Antarctic Peninsula (Larter and Barker, 1989, in press) and middle Oligocene time in the eastern Ross Sea (Bartek et al., 1991).

We suggest, however, that the first grounding of ice sheets at the continental shelf edge in the eastern Ross Sea occurred in early Miocene time, based on the first clear indication of type IA sequences at the paleo-shelf edge at this time (i.e. type IA sequences above the early Miocene unconformity U4A; Fig. 13). When determining the regional extent and timing of grounded ice sheets the drilling data in the Ross Sea are equivocal, but the data indicate that ice sheets were probably nearby, at least locally in the Transantarctic Mountains, by early Oligocene time (Barrett, 1975; Barrett et al., 1989). Savage and Ciesielski (1983) suggest that glacial units at DSDP Site 272 indicate open-water conditions in the early Miocene, but Bartek et al. (1991) suggest that these units could also have been deposited by grounded ice sheets. Drilling along the western margin of the Ross Sea (Fig. 1) shows evidence of intermittently grounded ice sheets in late Oligocene and early Miocene time. This ice may have come from local glaciers originating in the Transantarctic Mountains and from continental ice sheets (Barrett et al., 1989). Further, Leckie and Webb (1983) note that benthic foraminifera in glacial deposits at DSDP Site 270 indicate that this site was overdeepened (water depth 300–500 m) by early Miocene time.

The uncommonly good preservation (for Antarctic margins) of type IA and IIA sequences in the eastern Ross Sea, particularly before Pliocene time, can most easily be explained by the relatively rapid subsidence of the eastern Ross Sea during the inferred Eocene and younger late-rift period

TABLE 1

Estimates of the amount of Cenozoic progradation of the Antarctic continental margin

Margin segment	Distance prograded	Time of progradation	Ref. and Fig.
Eastern Ross Sea	85 km	middle Miocene to present	1, Fig.2
Wilkes Land	80 km	probably Cenozoic	–, Fig.2
Prydz Bay (seaward of ODP site 739)	22 km	early Oligococene to late Miocene	2, Fig.17
	up to 50 km	late Miocene to present	2, Fig.18
Weddell Sea	about 40 km	late Oligocene to middle Miocene	3, –
	35 km	middle Miocene to present	3, –
Antarctic Peninsula	about 20 km	late Miocene to present	4, Fig.2

References: 1 = Hinz and block (1984); 2 = Cooper et al. (1991a); 3 = Kuvaas and Kristoffersen (1991); 4 = Larter and Barker (1989)

(Cooper et al., 1987a; 1991b). Intense prograding of the paleo-shelf edges that occurs at and above unconformity U2 (Fig. 13) signals likely large fluctuations in the amount of grounded ice and entrained sediment reaching the outer shelf since late Miocene time (Fig. 13; Bartek et al., 1991).

Large-scale progradation of type IA sequences during the late Cenozoic is observed on most parts of the Antarctic margin. The greatest progradation appears to have occurred since middle Miocene time, but earlier progradation may also have been quite significant (e.g., Prydz Bay and Weddell Sea). The amounts and timing of progradation around Antarctica are not well known from direct geologic sampling of the prograding sequences, but estimates (Table 1) suggest great variability, which probably indicates local geologic, tectonic and glaciologic controls during periods of local and continent-wide glaciation. We, and others, attribute this extensive prograding to the diachronous and episodic buildup and movement of ice across the shelf, and deposition of entrained sediment principally at the continental shelf edge. The extensive progradation of the Antarctic continental margin since the middle Miocene is also observed on many low-latitude continental margins (Bartek et al., 1991), suggesting a direct link between the buildup of grounded ice sheets on the Antarctic continental shelf and global sea-level fluctuations (Larter and Barker, 1989, 1991; Bartek et al., 1991), as discussed below.

We suspect that type IIA and IIA (transitional) geometries, which are most easily observed at paleo-shelf edges, result from depositional environments that are different from those causing type IA geometries. In Prydz Bay, drilling at ODP Sites 739 and 742 indicates that the gently dipping, aggrading, and sometimes distorted acoustic sequences [i.e. type IIA (transitional)] could be distal shelf facies of glacial deposits derived from grounded ice sheets situated on the continental shelf (Hambrey et al., in press; Cooper et al., 1991a). The paleo-water depths in Prydz Bay are unknown from fossil evidence (Barron et al., 1991). Glacial deposits at the base of ODP Site 742 indicate an upsection change from glacilacustrine to glacimarine and water-lain tills indicating possible deepening of the shelf (Hambrey et al., in press). However, the amount and timing of this deepening cannot be determined (Hambrey, pers. commun., 1990). In the eastern Ross Sea, the type IIA sequences at the early Miocene paleo-shelf edge (Fig. 13) are almost 180 km from ODP Site 270, and the drilling is of little help in explaining these type IIA acoustic geometries. We suspect, from analogy to Prydz Bay and from regional arguments for late Oligocene grounded ice sheets in the Ross Sea (Barrett et al., 1987, 1989; Bartek et al., 1991), that the change from type IIA to IA geometries at the paleo-shelf edge (Fig. 13) signifies a major transition in glacial depositional environments from those associated with distal grounded ice sheets (type IIA) to those with proximal ice sheets (type IA).

Erosion of the inner continental shelf by grounded ice has apparently been an important factor in (1) controlling ice sheet movements, (2) supplying sediment to the outer shelf sequences, and (3) preserving the overdeepened profile of the continental shelf. As noted above, deep and relatively narrow ice-erosion troughs are found beneath the inner shelf. These troughs commonly cut into the sedimentary section near and along the sediment–basement contact on basin flanks (Haugland et al., 1985; Cooper et al., 1991a). The trough locations may be partly controlled by basin-edge faults (Stagg, 1985) and by uplift of eroded basement rocks (Tingey, 1985). Several kilometers of sediment may have been eroded from these troughs and from other troughs that now lie buried landward beneath massive grounded ice sheets, such as under the Lambert Glacier in Prydz Bay (Hambrey, 1991).

In Prydz Bay, the innermost edge of the Cenozoic progradational glacial strata begins on the flanks of one of these troughs (point *A*, Fig. 16), suggesting that sediment removed from the trough has been recycled to the outer shelf sequences (Cooper et al., 1991a). As the volumes of the troughs on the continental shelf are small, relative to those for the prograding sequences, large amounts of sediment must also have been eroded from interior areas of Antarctica. The timing of first erosion of the inner shelf trough in Prydz Bay is unknown, but is thought to have occurred during the early stages (i.e. early Oligocene) of continental shelf glaciation (Cooper et al., 1991a; Hambrey et al., in press). A similar early Oligocene age is postulated for initial deposition of the large trough-mouth fan in front of the major Crary Trough in the Weddell Sea, also by the East Antarctic ice sheet (Kuvaas and Kristoffersen, 1991).

The reversed and overdeepened bathymetric profile of the Antarctic continental shelf is today partly a manifestation of the thickness and distribution of the type IA sedimentary sequences beneath the shelf and continental slope. Although several factors may contribute to this unusual profile, as noted above, recent seismic reflection data from the margin (Fig. 2) and flexural model studies of the Prydz Bay margin (Ten Brink and Cooper, 1990) and elsewhere (Reynolds et al., 1991) suggest to us that two factors may be most important in preserving this reversed shelf profile once it has been established: (1) the distribution, thickness and shape of type IA topset beds on the continental shelf resulting from erosion and deposition by episodic movement of grounded ice sheets across the shelf, and (2) sediment loading and compaction of the outer margin caused by type IA topset and foreset beds, which have been deposited principally during glacial maxima and only minimally during interglacial times, like today. The initial depression of the shelf by ice-sheet loading and by erosion of deep shelf troughs may be adequate to cause overdeepening of the shelf during early glacial advances (Walcott, 1970, 1972; Drewry, 1983; Anderson et al., 1983a), but this is not adequate for explaining the preservation of today's interglacial overdeepened shelf profile (Anderson et al., 1983b). Further studies based on new seismic data are needed to confirm this explanation for long-term and probably continuous (i.e. since initial glaciation) overdeepening of the Antarctic continental shelf.

Eustatic controls

Several investigators have noted that the acoustic geometries for known and inferred Cenozoic glacial deposits beneath the Antarctic continental margin have general similarities to those from non-polar margins where fluctuations in sea level have periodically subareally exposed and submerged the continental shelf (Hinz and Krause, 1982; Hinz and Block, 1984; Haugland et al., 1985; Bartek et al., 1991). In most places around Antarctica, where sequences (i.e. type IA) are highly eroded, the reflection geometries are rather like the lowstand deposits of Vail et al. (1977) and Wilgus et al. (1988). However, as discussed above, several acoustic characteristics of the Antarctic glacial sequences, which have been tested by Prydz Bay drilling, cannot be easily explained by invoking low-latitude depositional processes (e.g., Vail et al., 1977), especially on paleo-continental shelves that have been inferred to have been overdeepened.

Although detailed acoustic characteristics differ on polar margins with grounded ice sheets and non-polar margins, the underlying tenet that sea-level fluctuations have affected, and are recorded in, the sedimentary sequences from the two types of margins is valid:

(1) The growth and decay of major Antarctic ice sheets grounded to the continental shelf edge will result in an absolute fall and rise of sea level respectively

(2) The grounded ice sheets that may formerly have covered parts to all of the Antarctic continental shelf are sensitive to relative changes in sea level. Grounded ice sheets will float off the seafloor with rising sea level and advance seaward with falling sea level (Drewry and Cooper, 1981; Larter and Barker, 1989; Hambrey, 1991; Hambrey et al., 1991, in press; Kuvaas and Kristoffersen, submitted).

(3) During periods of increased ice (lowered sea level), grounded ice sheets may advance across the continental shelf, eroding sediments and depositing them on the shelf as topset beds and on the continental slope as foreset beds.

Different geometric guidelines must be applied to the Antarctic continental shelf, where water depths are too deep (400–1400 m deep) to expose the shelf to subaerial erosion during periods of lowered sea level, such as outlined by Haq et al. (1987, 1988). Rather, in Antarctica, erosion may have been principally by grounded ice sheets during glacial maxima and lowered sea levels. Some stratigraphic relationships that occur on low-latitude margins, such as the sharp geologic contrast between overlying highstand and underlying lowstand deposits, would not be expected following sea-level changes on the overdeepened Antarctic continental shelf. Sedimentary sequences from the Antarctic margin that record fluctuations in sea level are interpreted from the CIROS-1 drill site near the coast in the western Ross Sea (Fig. 1; Barrett et al., 1987, 1989) and from ODP drill sites on the Prydz Bay shelf (Barron et al., 1991).

As noted above, there are few geologic data to establish the length of time that (i) the Antarctic continental shelf has been overdeepened, like today, and (ii) the time period over which it has not been subaerially exposed by fluctuations in sea level. Seismic data may provide the only indirect evidence for the initial shelf overdeepening, which would probably result in a major change in depositional environments. The transition from type IIA to type IA acoustic sequences at the paleo-shelf edges, if it accurately signals the first inferred advance of grounded ice to the shelf edge and beginning of polar-margin depositional processes, would also provide the timing of the first glacial overdeepening of the shelf. Based on the earliest inferred ages for type IA sequences, the initial occurrences of overdeepening were probably diachronous around Antarctica (Fig. 2, Table 1).

For all continental shelf depths and all parts of the Antarctic margin, seismic data show widespread unconformities throughout the Cenozoic sedimentary sections, and these unconformities may be related to sea-level changes. Hinz and Kristoffersen (1987) proposed a circum-Antarctic stratigraphic concept that stated that unconformities in the inferred Cenozoic and older sedimentary sections in the Ross Sea, Weddell Sea and Wilkes Land are coeval and are related to circum-Antarctic paleoceanographic and structural events, such as initiation of the circum-Antarctic current and uplift of the Transantarctic Mountains, etc. Reflections cannot, however, be traced between these areas because of intervening ice. Consequently, the concept can be tested only by drilling, which to date has shown greater regional complexity than predicted (Barron et al., 1989; Barrett, 1989; Barker et al., 1988).

The Hinz and Kristoffersen (1987) stratigraphic concept does not directly tie all Antarctic depositional sequences and unconformities to either grounded ice sheets (e.g., our type IA sequences) or sea-level fluctuations. Rather, they suggest that sea-level changes during glacial times may be partly responsible for depositional and erosional processes (e.g., iceberg deposition, shelf currents, etc) that formed the prograding sedimentary sequences beneath the continental shelf. We see such processes as possible explanations for our type IIA sequences. Bartek et al. (1991) interpret the Hinz and Block (1984) Cenozoic unconformities U1 to U6 in the eastern Ross Sea as being caused by the grounding of ice sheets to the continental shelf edge since middle to late Oligocene time. They further speculate that the ice-volume changes responsible for the Cenozoic unconformities and prograding sequences in the eastern Ross Sea are also responsible for the global sea-level changes commonly inferred from Cenozoic prograding sequences beneath continental margins from around the world. Larter and Barker (1989, 1991) also suggest that ice-volume changes, which they infer from unconformities on the Antarctic Peninsula continental shelf, are related to sea-level fluctuations that have controlled the episodic grounding of the West Antarctic Ice sheet. All authors caution, however, that there are insufficient drilling data to adequately resolve the inferred link between depositional processes on the continental margin, ice-volume changes, and sea-level fluctuations.

Although we generally concur with the ideas of Bartek et al. (1991), we differ in our interpretations of the times at which grounded ice sheets first reached the continental shelf edge in the eastern Ross Sea. We suspect that this occurred in early Miocene time, rather than in middle Oligocene time, based on our interpretation of the seismic stratigraphy described above. However, at present we do not know what effect, if any, the first occurrence of grounded ice sheets at the continental shelf edge, and the likely related overdeepening of the continental shelf, may have had on the magnitude of global sea-level fluctuations. Also, we reiterate our observation that the geometries of sedimentary sequences beneath the mid- to outer shelf areas of the eastern Ross Sea, which were used by Bartek et al. (1991) to derive global sea-level models, differ regionally from other parts of the Antarctic continental shelf in terms of the great thickness of topset strata and good preservation of aggradational strata in type IA sequences (Fig. 2). This regional difference may not be as important to the general correlations of acoustic geometries, and associated depositional processes, between different segments of the Antarctic margin (Fig. 2 and 4) as it is to a global assessment of sea-level changes based on interpretations of Antarctic acoustic geometries characterized solely by those in the eastern Ross Sea (Bartek et al., 1991).

Nevertheless, sea-level variations should be recorded and at least locally preserved in sequences from the overdeepened Antarctic continental shelf (Fig. 20). In pre-glacial and early glacial times, when the continental shelf was not overdeepened, the sedimentary sequences (i.e. type IIA) should resemble highstand and lowstand deposits of low-latitude margins (Vail et al., 1977). When the shelf was overdeepened, as we suspect that it was during much of the Cenozoic, the sedimentary sequences on the shelf (i.e. type IA) would be entirely of marine origin, either glacial or interglacial. The greatest changes in acoustic geometries due to variations in sea level would be seen along the coastline or ice-grounding line and at the continental shelf edge, especially during glacial maxima when grounded ice extended to this shelf edge (Fig. 20). Extensive prograding of several segments the Antarctic sequences (i.e. type IA) since the middle Miocene has been suggested, as noted above, supporting the general concept that there is a direct link between the buildup of grounded ice sheets on the Antarctic continental shelf and global sea-level fluctuations (Larter and Barker, 1989; Bartek et al., 1991).

Summary and conclusions

We recognize two categories in the acoustic geometry of the thick Cenozoic sedimentary sequences beneath the Antarctic continental margin (Fig. 2 and 8): (1) Type IA sequences that we believe are deposited principally by grounded ice sheets on overdeepened continental shelves (identical sequences are not likely on low-latitude nonglaciated margins), and (2) type IIA sequences that we suggest were deposited possibly by non-glacial and open-marine glacial processes on normal depth shelves; such sequences do occur on low-latitude margins.

The ubiquitous occurrence of type IA sequences, particularly beneath the outer continental shelf, suggests to us that grounded ice sheets have strongly controlled the location and depositional environments of the thick offshore prograding sedimentary sections since at least middle Cenozoic time (Fig. 20). These grounded ice sheets probably extended to the continental shelf edge many times, as is evident from the numerous unconformities within the type IA sequences and from the drilling

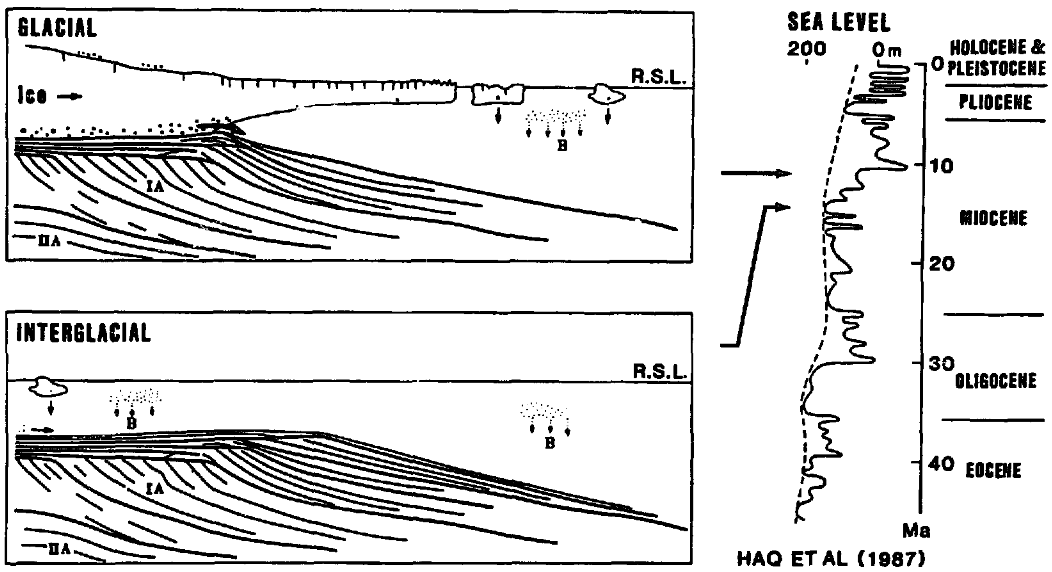


Fig. 20. Grounded ice-sheet model proposed to explain the origin of type IA and IIA sequences beneath the Antarctic margin. Such ice-sheet fluctuations are probably closely linked to sea-level changes. Type IIA sequences may be deposited during non-glacial and early glacial periods on a normal depth continental shelf, whereas type IA sequences may be deposited during glacial maxima periods on an overdeepended shelf and at the front of grounded ice sheets that extend to the paleo-shelf edge. Little sediment is deposited during interglacial periods. See also Fig. 4. *B* = biogenic debris; *R.S.L.* = relative sea level.

evidence of numerous high-velocity near-seafloor layers (Cochrane and Cooper, 1991) and multiple overcompaction events (Solheim et al., 1991).

The major prograding sequences lie near the outlets of topographic depressions (Fig. 1 and 5) that are due partly to post-early Cenozoic glacial erosion and lithospheric processes, such as rifting in the Ross Sea, the West Antarctic Rift and the Antarctic Peninsula–Bransfield Strait (Cooper et al., 1987a, b, 1991b; Behrendt and Cooper, 1991; Barker and Dalziel, 1983), crustal flexure from rifting and ice loading in the Wilkes Basin and Wilkes Land (Stern and Ten Brink, 1989), and post-active-margin thermal subsidence in the Antarctic Peninsula (Larter and Barker, 1989). Cenozoic subsidence and preferential ice-sheet erosion of late Mesozoic and older rifts in the Weddell Sea and Prydz Bay has partly controlled the location of extensive prograding sequences at these sites. The orientation of glacial erosional features commonly differs from that of underlying pre-Cenozoic structures.

Cenozoic fluctuations in the size of the Antarctic Ice Sheet onto and across the continental shelf probably affected: (1) water depths of the shelf (now overdeepened) due to ice loading and crustal flexure, (2) glacial erosion (largely from the continental interior and inner shelf areas) and glacial deposition (largely to the outer shelf and continental slope prograding sequences) during glacial maxima, (3) clastic sediment deposition (now low) on the continental shelf and slope during interglacial times, and (4) sediment compaction (inferred to be large) beneath grounded ice sheets. We suspect that these fluctuations, and related processes, preserved the overdeepened bathymetric profile of the Antarctic continental shelf following the initial advance of the ice sheet to the continental shelf edge. Such a profile is uncommon in low-latitude continental margins.

Sea-level variations (100–150 m) commonly attributed to fluctuations in the Antarctic Ice Sheet would not subaerially expose an overdeepened continental shelf (now 400–1400 m deep), thereby leading to stratigraphic sequences that are different from those found on low-latitude, normal depth shelves. However, sea-level variations are widely believed to affect the distribution of Antarctic grounded ice (e.g., Larter and Barker, 1989; Hambrey et al., 1991, in press) and related progradational sequences. The progradational sequences beneath the outer continental shelf are likely to

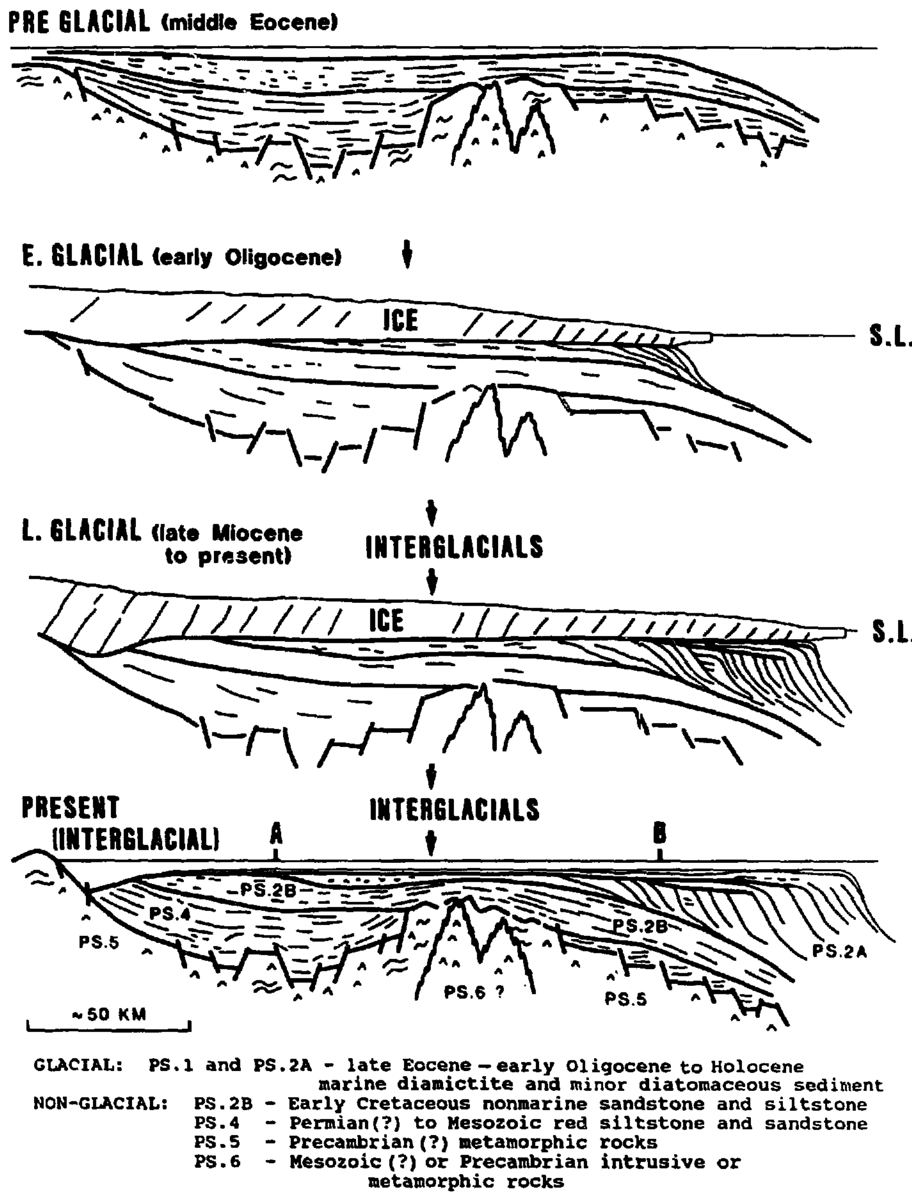


Fig. 21. Interpretive model for evolution of the glacial sequences (unit PS.2A) beneath the Prydz Bay outer continental shelf based on the grounded ice-sheet model in Fig. 4 and 20. After initial advance of grounded ice to the shelf edge, and likely overdeepening of the shelf, in the early Oligocene, the type IA and IIA (transitional) sequences within units PS.1 and PS.2A were deposited by numerous advances (glacial periods) and retreats (interglacial or non-glacial periods) of grounded ice sheets across the continental shelf. *A* and *B* are reference points mentioned in text. Modified from Cooper et al. (1991a).

contain a detailed record of major ice-volume and sea-level changes.

At present, the timing and synchroneity of glacial events around Antarctica are poorly understood. However, the available continental shelf drilling and acoustic data are consistent with the following scenario:

The initial major expansion of the East Antarctic grounded ice sheets seaward to the continental shelf edge (and the first appearance of type IA sequences) occurred in Prydz Bay by late Eocene to early Oligocene time (Barron et al., 1989). The precursors to this event in Prydz Bay included (1) likely gradual overdeepening of the continental shelf by inner shelf erosion and flexural loading by the advancing ice sheet and (2) deposition of gently dipping and acoustically distorted early glacial strata (type IIA transitional sequences) on the continental shelf and in front of the ice sheet (Fig. 17 and 21). Remnants of these early glacial strata cover nearly 120 km of the mid-shelf areas (i.e. between points *A* and *B* in Fig. 21)

Grounded ice sheets existed near the Ross and Weddell Sea by early Oligocene time (Barrett et al., 1987, 1989; Wise et al., 1991). In the Ross Sea, grounded ice sheets may have been on parts of the continental shelf in middle to late Oligocene time (Bartek et al., 1991). We suspect, however, based on the transition from type IIA to IA seismic sequences, that the first grounding of the ice sheets out to the shelf edge, and the initial overdeepening of the outer continental shelf, occurred in early Miocene time (Fig. 13).

Grounded ice sheets probably extended to the continental shelf edge episodically since middle Cenozoic time on at least five broad segments of the Antarctic margin. These ice sheets deposited abundant sediments at the continental shelf edge (i.e. type IA sequences) and prograded the shelf by up to 85 km (Table 1). The timing and amount of progradation vary around Antarctica, probably due to regional geologic, tectonic and glaciologic factors. Progradation has been greater since middle Miocene time, and probably increased further in latest Cenozoic time (Fig. 13).

Our summary and preliminary analysis of the prograding sequences of the Antarctic margin clearly underlines the need for further acoustic surveys and scientific drilling studies of the Antarctic continental shelf. To this end, we strongly encourage the cooperative Antarctic studies, such as ANTOSTRAT (Antarctic Offshore Acoustic Stratigraphy; Cooper and Webb, 1990), ANTALITH (Antarctic Lithospheric Studies; Dalziel and Zimmerman, 1989) and SEARISE (Bindschadler, 1990), that are now being undertaken by the international community.

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