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RECONSTRUCTION OF THE ROSS ICE DRAINAGE SYSTEM, ANTARCTICA, AT THE LAST GLACIAL MAXIMUM

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ABSTRACT. We present here a revised reconstruction of the Ross ice drainage system of Antarctica at the last glacial maximum (LGM) based on a recent convergence of terrestrial and marine data. The Ross drainage system includes all ice flowlines that enter the marine Ross Embayment. Today, it encompasses one-fourth of the ice-sheet surface, extending far inland into both East and West Antarctica.

Grounding lines now situated in the inner Ross Embayment advanced seaward at the LGM (radiocarbon chronology in Denton and Marchant 2000 and in Hall and Denton 2000a, b), resulting in a thick grounded ice sheet across the Ross continental shelf. In response to this grounding in the Ross (and Weddell) Embayment, ice-surface elevations of the marine-based West Antarctic Ice Sheet were somewhat higher at the LGM than at present (Steig and White 1997; Borns *et al.* 1998; Ackert *et al.* 1999). At the same time, surface elevations of the East Antarctic Ice Sheet inland of the Transantarctic Mountains were slightly lower than now, except near outlet glaciers that were dammed by grounded ice in the Ross Embayment. The probable reason for this contrasting behavior is that lowered global sea level at the LGM, from growth of Northern Hemisphere ice sheets, caused widespread grounding of the marine portion of the Antarctic Ice Sheet, whereas decreased LGM accumulation led to slight surface lowering of the interior terrestrial ice sheet in East Antarctica.

Rising sea level after the LGM tripped grounding-line recession in the Ross Embayment, which has probably continued to the present day (Conway *et al.* 1999). Hence, gravitational collapse of the grounded ice sheet from the Ross Embayment, accompanied by lowering of the interior West Antarctic ice surface and of outlet glaciers in the Transantarctic Mountains, occurred largely during the Holocene. At the same time, increased Holocene accumulation caused a slight rise of the inland East Antarctic ice surface.

Introduction

The Ross ice drainage system includes the portion of the Antarctic Ice Sheet that flows into the marine Ross Embayment (Figs 1, 2). This system encompasses about one-fourth of the surface of the present-day ice sheet. Inland ice divides delimiting this drainage system are 5700 km long and feature numerous domes and saddles. Ice flow from the high East Antarctic plateau converges into outlet glaciers that pass through the Transantarctic Mountains (TAM) into either the Ross Ice Shelf or the Ross Sea. The marine-based West Antarctic Ice Sheet (WAIS) drains through major ice streams into the Ross Ice Shelf. Thus interior flowlines from both East and West Antarctica converge on outlet glaciers or ice streams that enter the marine Ross Embayment.

The changing configuration of grounded ice in the Ross Embayment during the last glacial hemicycle affords unique insights about the past and future behavior of the marine-based WAIS, and hence on the cause of late Holocene sea-level rise (Conway *et al.* 1999; Fleming *et al.* 1998). These data also constrain the Antarctic contribution to global sea-level lowering at the last glacial maximum (LGM), as well as the Antarctic role in sharp pulses of sea-level rise following the LGM (Fairbanks 1989; Fleming *et al.* 1998). Moreover, modern Antarctic Bottom Water (AABW), a major factor in global thermohaline circulation, is fed by dense shelf water formed on Antarctic continental margins, particularly in the Ross and Weddell Embayments. Grounding-line advance to the continental shelf edge would remove this important contribution of dense shelf water at the LGM, as pointed out by Kellogg (1987). The implication is that the modern mixture of AABW must have evolved as ice-sheet grounding lines receded from Antarctic continental shelves in late-glacial and Holocene time.

The Ross ice drainage system includes the sites of the Byrd, Taylor Dome, Vostok, and Dome Circe ice cores (Fig. 1). Details of the Taylor Dome record are given in Steig *et al.* (2000). The combined records from all these ice cores are critical in determining the mechanisms that couple paleoclimatic changes in the polar hemispheres. An important as-

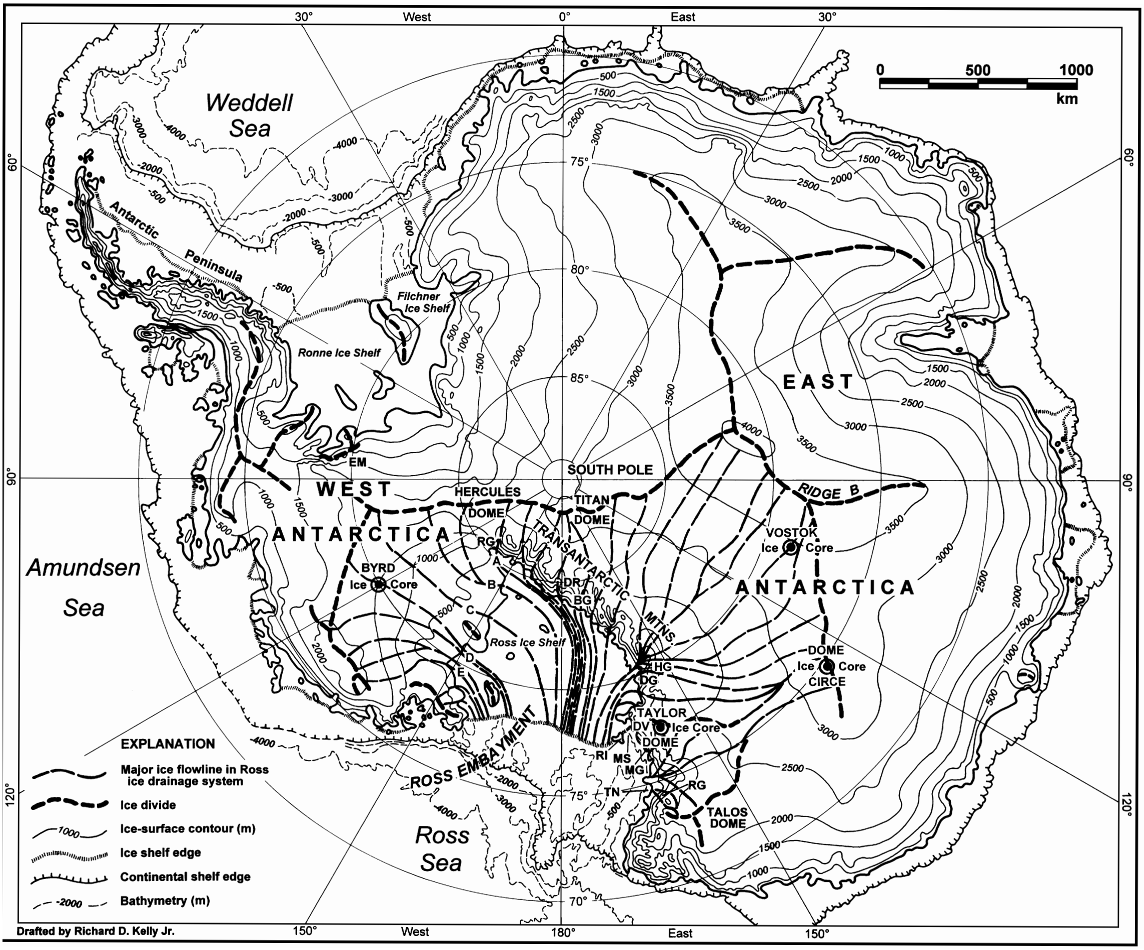
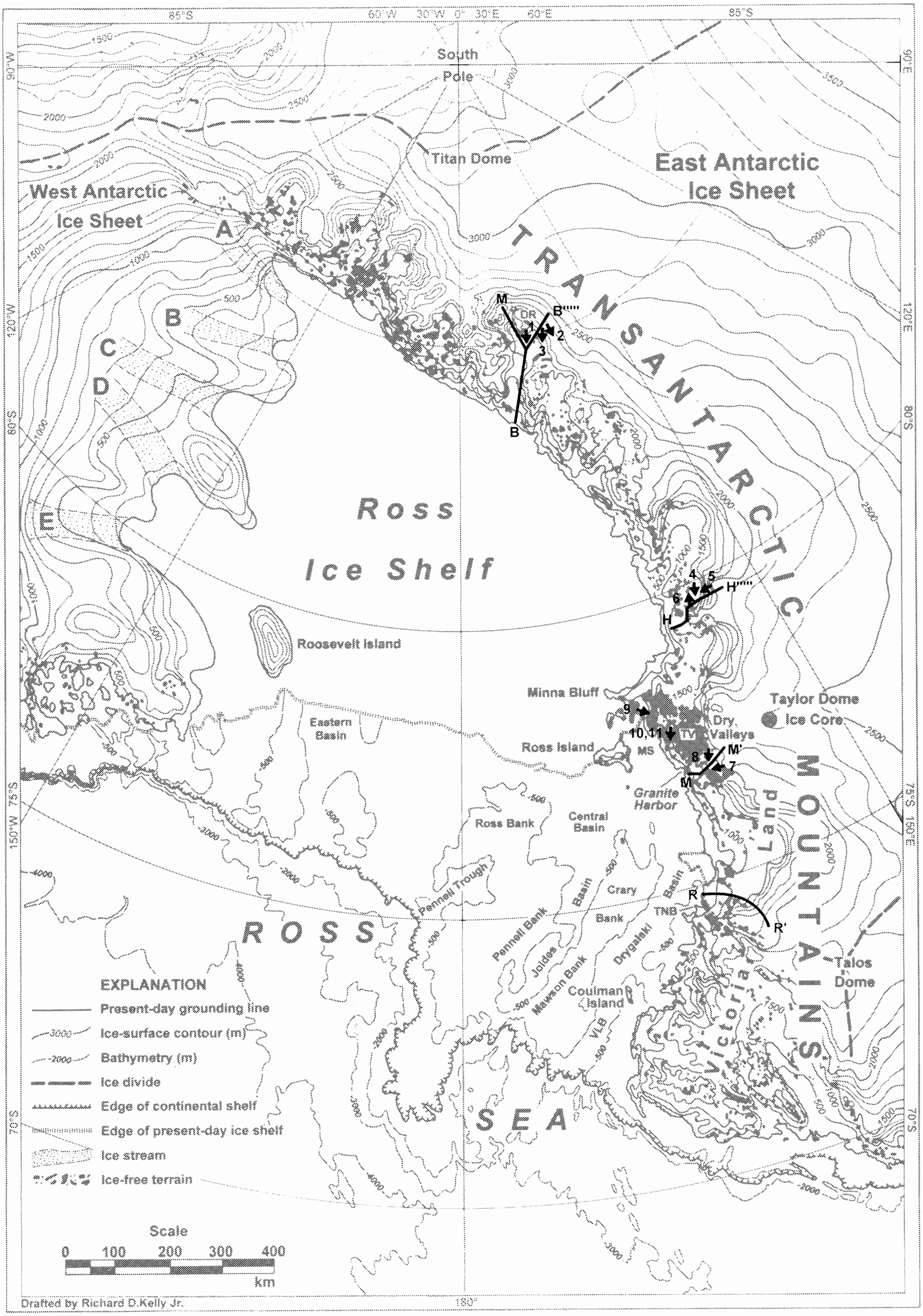


Fig. 1. Index map of Antarctica and the Ross ice drainage system. Ice flowlines are shown only for the Ross ice drainage system, which is defined as the portion of the Antarctic Ice Sheet that flows into the Ross Embayment. A, B, C, D, and E are West Antarctic ice streams. RG, Reedy Glacier; BG, Beardmore Glacier; DR, Dominion Range; HG, Hatherton Glacier; DG, Darwin Glacier; DV, Dry Valleys; RI, Ross Island; MS, McMurdo Sound; MG, Mackay Glacier; TN, Terra Nova Bay; EM, Ellsworth Mountains. Adapted from Drewry (1983).

Fig. 2. Present-day Ross Embayment. The Ross Embayment is part of the West Antarctic intra-continental rift system that extends inland beneath the marine-based West Antarctic Ice Sheet, forming basins up to 2500 m below sea level. The Transantarctic Mountains (TAM) form a high shoulder of this asymmetric rift system. West Antarctic ice drains into the Ross Embayment largely through five major ice streams (A, B, C, D, E). The present-day grounding line separates this inland grounded marine ice from the floating Ross Ice Shelf, which extends northward to a calving front that trends east– west across the Ross Embayment at the latitude of Ross Island. The terrestrial East Antarctic Ice Sheet is dammed inland of the TAM and drains through outlet glaciers into the Ross Embayment. South of McMurdo Sound these glaciers feed the Ross Ice Shelf, with grounding lines now close to the TAM front. North of McMurdo Sound they flow directly into the Ross Sea, locally feeding floating ice tongues and small ice shelves. Only in the McMurdo Sound region do East Antarctic outlet glaciers fail to pass through the TAM. The result is that the Dry Valleys, the eastern foothills of the Royal Society Range, and the coastal terrains alongside McMurdo Sound feature extensive ice-free areas. The continental shelf north of the Ross Ice Shelf ranges from 250 m to 1200 m deep, with an average depth of more than 500 m (Barnes and Lien 1988; Davey 1994). The seafloor features a series of northeast–southwest trending banks and ridges, separated by foredeepened troughs that contain basins. VLB, Victoria Land Basin; MS, McMurdo Sound; TNB, Terra Nova Bay; TV, Taylor Valley; DR, the Dominion Range. Arrowheads show the locations of the photographs in Figs 10a (1), 10b (2), 10c (3), 10d (4), 10e (5), 10f (6), 10g (7), 10h (8), 10i (9), 10j (10), and 10k (11). The dark lines with letters refer to longitudinal profiles of outlet glaciers in Figs 5–8.


South Pole
Titan Dome
East Antarctic Ice Sheet
West Antarctic Ice Sheet
TRANSANTARCTIC MOUNTAINS
Ross Ice Shelf
Roosevelt Island
Minna Bluff
Ross Island
Taylor Dome Ice Core
Dry Valleys
Granite Harbor
Eastern Basin
Ross Bank
Central Basin
Pennell Trough
Pennell Bank
Joides Basin
Crary Bank
Drygalski Basin
Mawson Bank
Coulman Island
TNB
VLB
Victoria Land
Talos Dome
ROSS SEA
EXPLANATION
Present-day grounding line
Ice-surface contour (m)
Bathymetry (m)
Ice divide
Edge of continental shelf
Edge of present-day ice shelf
Ice stream
Ice-free terrain
Scale
0 100 200 300 400
km
Drafted by Richard D.Kelly Jr.

sumption in interpreting these cores is that interior ice-surface elevations have not changed greatly during the course of the last glacial cycle. Direct measurement of former interior ice-sheet elevations could come from the total gas content of the ice itself (Lorius *et al.* 1984, 1985), but the validity of this technique has been questioned (Paterson and Hammer 1987). An analysis of isotope profiles at Byrd Station indicates an ice-surface lowering of 200 to 400 m in the last 4500 years (Steig and White 1997). The LGM reconstruction presented here forms an alternative control on the limits of surface-elevation changes at the ice-core sites, as well as on the varying distances of the sites from Antarctic coastlines during the last glacial hemicycle.

The configuration of grounded and floating ice in the Ross Embayment has been of interest for more than a century and a half. In 1841 and 1842, Ross (1847) mapped the calving front of the Ross Ice Shelf, or the 'Great Icy Barrier'. Scott (1905, Vol. 2, pp. 422–425) postulated that 'when the Southern glaciation was at a maximum …. the Great Barrier was a very different formation from what it is today …. the huge glacier, no longer able to float on a sea of 400 fathoms, spread out over the Ross Sea, completely filling it with an immense sheet of ice'. Scott (1905, Vol. 2, p. 425) further concluded that, during recession, the ice sheet became buoyant and broke away gradually so that the '…. Barrier is the remains of the great ice-sheet'. David and Priestley (1914, Fig. 46, Pl. XCV) also postulated northward expansion so that at the time of maximum glaciation the surface of the grounded Ross Ice Barrier (now called the Ross Ice Shelf) reached heights of 1000 ft (305 m) in McMurdo Sound and filled the Ross Sea with a grounded 'great ice sheet' for at least 320 km north of the current Barrier edge. In sharp contrast, Debenham (1921) interpreted glacial deposits in the McMurdo Sound region in terms of local ice expansion.

In recent years this difference of opinion has continued to be expressed in a long series of reconstructions of LGM ice extent in the Ross Embayment (e.g. Hollin 1962; Denton and Armstrong 1968; Mercer 1968, 1972; Denton *et al.* 1970, 1971, 1975, 1989a, 1991; Denton and Borns 1974; Kellogg *et al.* 1979, 1990, 1996; Drewry 1979; Stuiver *et al.* 1981; Colhoun *et al.* 1992; Shipp *et al.* 1999). These reconstructions depict a wide variety of possible LGM configurations, ranging from little difference compared with the modern situation all the way to complete grounding across the continental shelf.

There is now a convergence of terrestrial data (Denton *et al.* 1971, 1989a,b; Stuiver *et al.* 1981; Bockheim *et al.* 1989; Orombelli *et al.* 1990; Hall and Denton 2000a,b; Hall *et al.* 2000; Denton and Marchant 2000; Dochat *et al.* 2000) and marine core data (Shipp *et al.* 1999; Domack *et al.* 1999a, b) with regard to the configuration of grounded ice in the Ross Embayment at the LGM. Based on this combined information, we present here a revised reconstruction of the Ross ice drainage system at the LGM.

Terrestrial evidence

Figure 3 shows our reconstruction of the Ross Sea ice drainage system at the LGM. Fig. 4 gives details for the Ross Embayment portion of this system. Former longitudinal profiles of outlet glaciers that flow from the East Antarctic Ice Sheet through the TAM into the Ross Embayment afford important geologic controls for these reconstructions. Beardmore Glacier (Figs 5, 10a–c) is on a flowline that projects inland to Titan Dome. Hatherton Glacier (Figs 6, 10d–f) is alongside Byrd Glacier, which drains ice from Dome Circe. Mackay Glacier (Figs 7, 10g) originates at Taylor Dome, situated just inland from the Dry Valleys. Reeves Glacier (Fig. 8) flows seaward from Reeves Névé in northern Victoria Land.

These outlet glaciers were selected because they are among the few in the TAM with late Quaternary lateral drift sheets that delineate former longitudinal profiles. The drift sheets were differentiated in each area by surface morphology, surface boulder weathering, and soil development (staining and cohesion depths, solum thickness, morphogenetic salt stage, and weathering stage). Denton *et al.* (1989b) gave details for Beardmore Glacier, Bockheim *et al.* (1989) for Hatherton Glacier, and Orombelli *et al.* (1990) for Reeves Glacier.

A distinctive lateral gravel drift sheet with little weathering and with well-preserved morphology is the common factor that links the behavior of these outlet glaciers. This deposit is named Beardmore drift at Beardmore Glacier (Denton *et al.* 1989b), Britannia I and Britannia II drifts at Darwin/Hatherton Glaciers (Bockheim *et al.* 1989), and Terra Nova drift at Reeves Glacier (Orombelli *et al.* 1990). It is unnamed alongside Mackay Glacier. These drifts are easily recognized because of their distinctive light color, caused by numerous well-preserved sandstone or granite clasts. In sharp contrast, the adjacent older drifts are darker in color be-

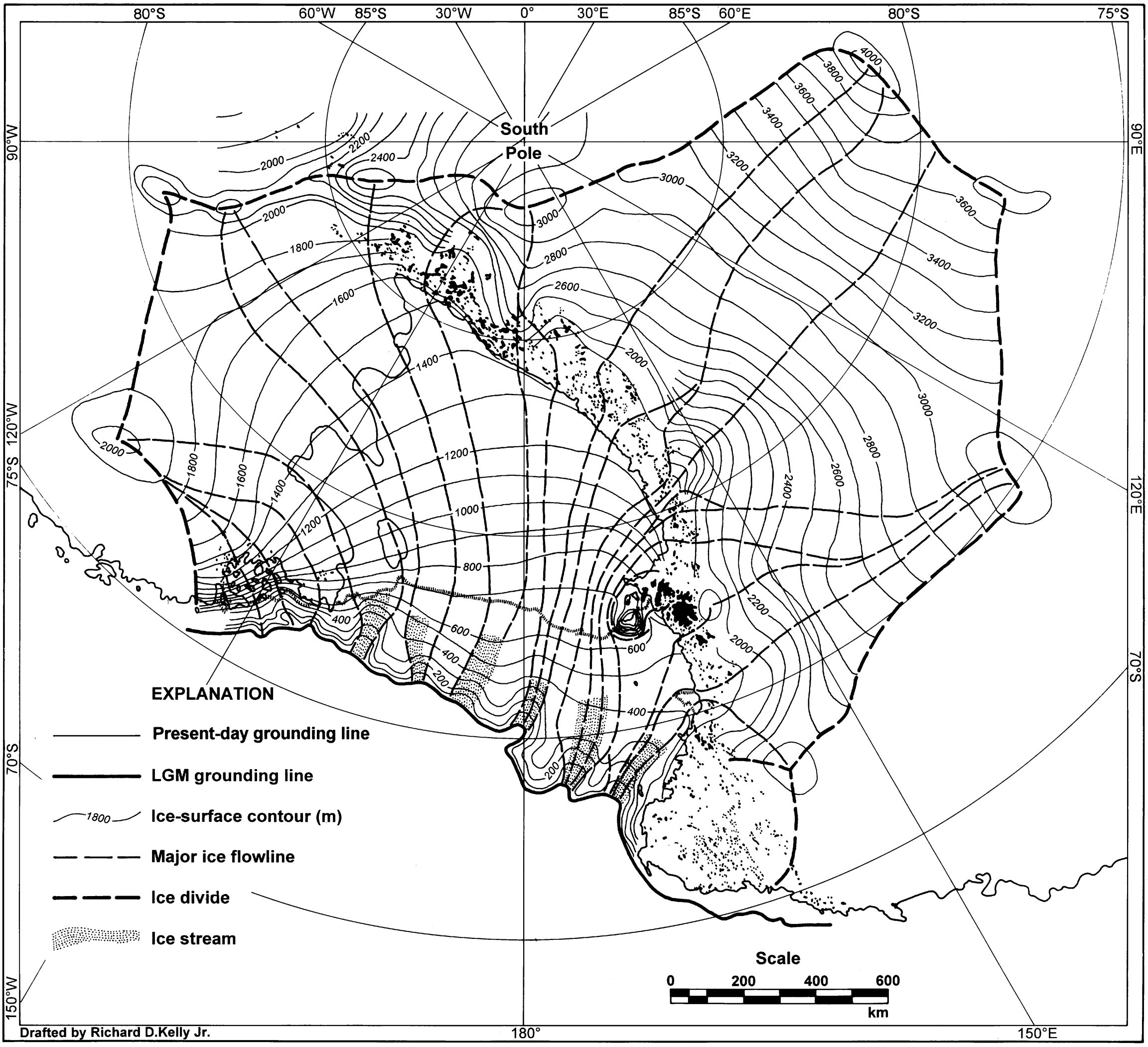


Fig. 3. Reconstruction of the entire Ross ice drainage system at the LGM. The geologic basis for this reconstruction is given in the text.

cause of an abrupt increase in the dolerite/sandstone ratios across the boundaries between drifts (Figs 10b–f).

The little-weathered, light-colored gravel drifts alongside these outlet glaciers all contain a high percentage of striated and bruised surface clasts. These drifts are unconsolidated and loose, comprise gravel with a sandy matrix, have little soil development, and feature high concentrations of surface boulders. They all show well-preserved morphology including ice-cored hummocky terrain, moraine ridges with openwork gravel, perched boulders, and kettles with small lakes. Older drifts have more soil development, greater surface boulder weathering, very few perched boulders, and more subdued surface morphology compared with the fresh gravel drifts. Such obvious differences between these drift sheets are common along the length of the TAM. Therefore, as a first approximation, we assume that the fresh gravel drifts alongside the outlet glaciers are of about the same age.

On the basis of the upper limits of the distinctive young gravel drift sheets, former longitudinal pro-

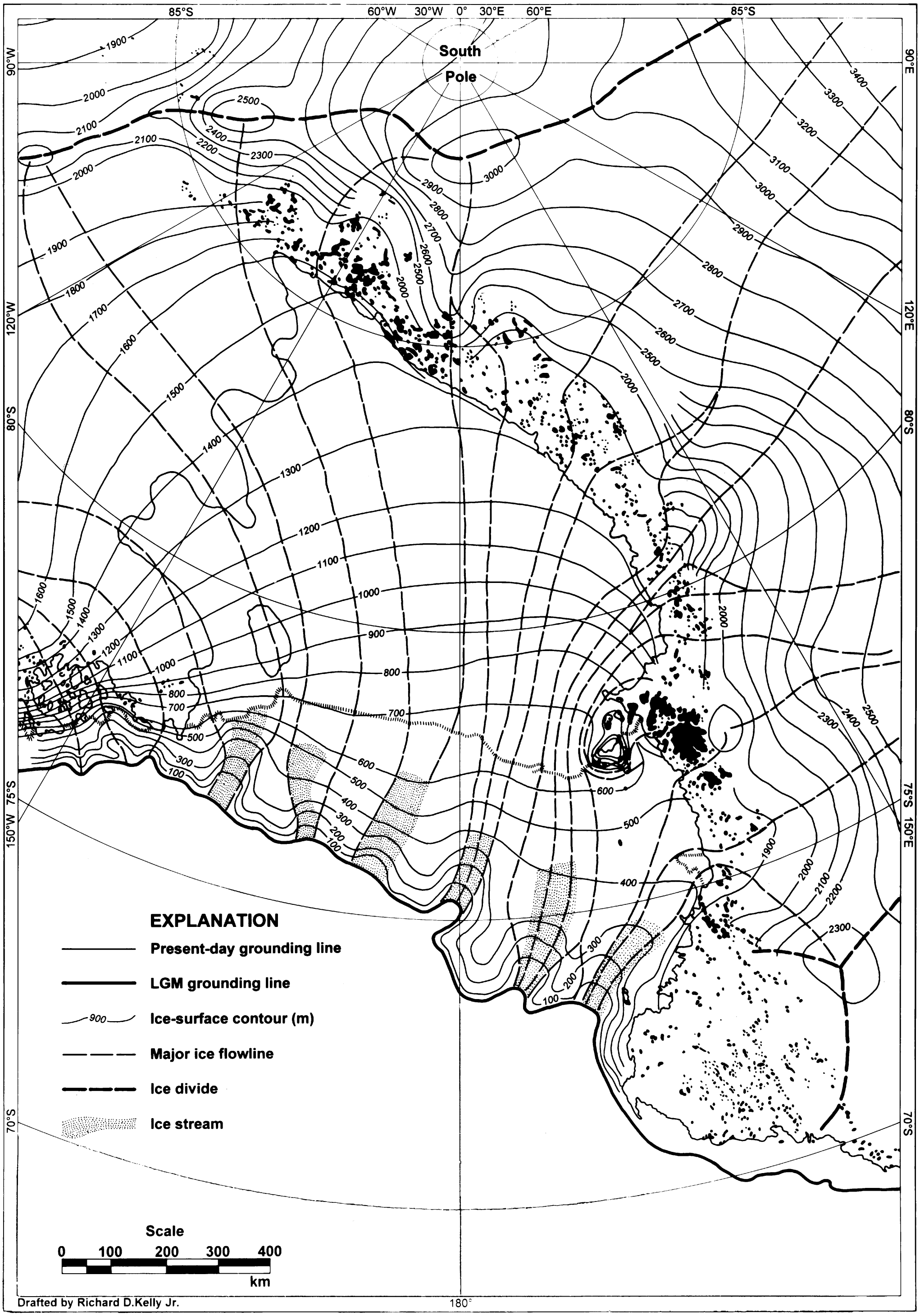


Fig. 4. Reconstruction of the grounded ice sheet in the Ross Embayment at the LGM. The geologic basis for this reconstruction is given in the text.

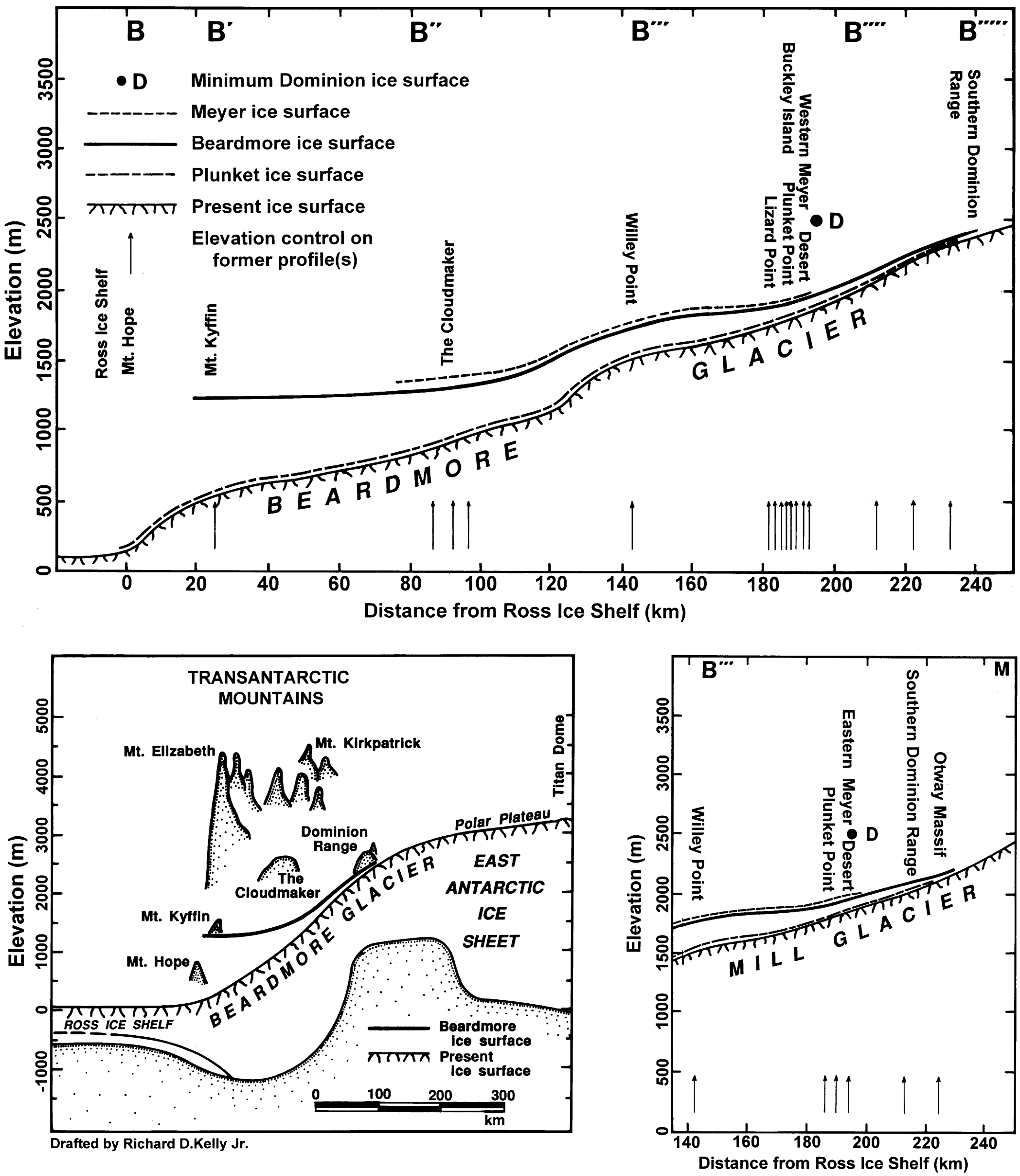


Fig. 5. Present and former (LGM) longitudinal profiles of Beardmore and Mills Glaciers. The positions of profiles B–B^{11111} and B^{111}–M are shown in Fig. 2. The Beardmore ice surface represents the LGM. See Denton *et al.* (1989b) for details.

files were reconstructed for Beardmore (Fig. 5), Hatherton/Darwin (Fig. 6), Mackay (Fig. 7), and Reeves (Fig. 8) Glaciers. We argue below that the thickening of outlet glaciers represented by these profiles dates to the LGM. It is important to note that all these former profiles exhibit the same fundamental characteristics: namely, they show only slight surface-level rise at the glacier heads near the polar plateau, but substantial rise at the glacier mouths near the Ross Ice Shelf or Ross Sea. The leading interpretation is that such former longitudinal profiles reflect a damming effect from ice-

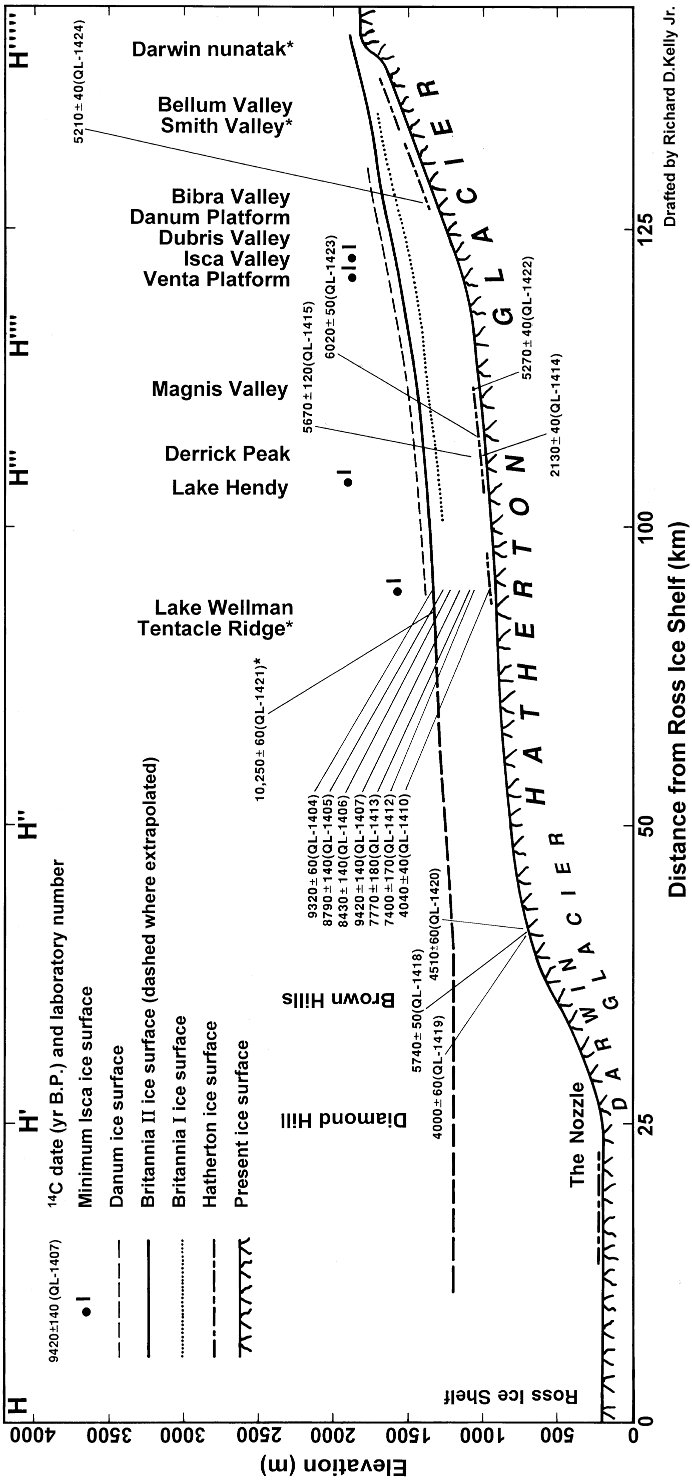


Fig. 6. Present and former (LGM) longitudinal profiles of Hatherton Glacier. Position of profiles is shown as H–H[11111] in Fig. 2. The Britannia II ice surface represents the LGM. For descriptions of the other ice surfaces and the radiocarbon samples, see Bockheim *et al.* (1989).

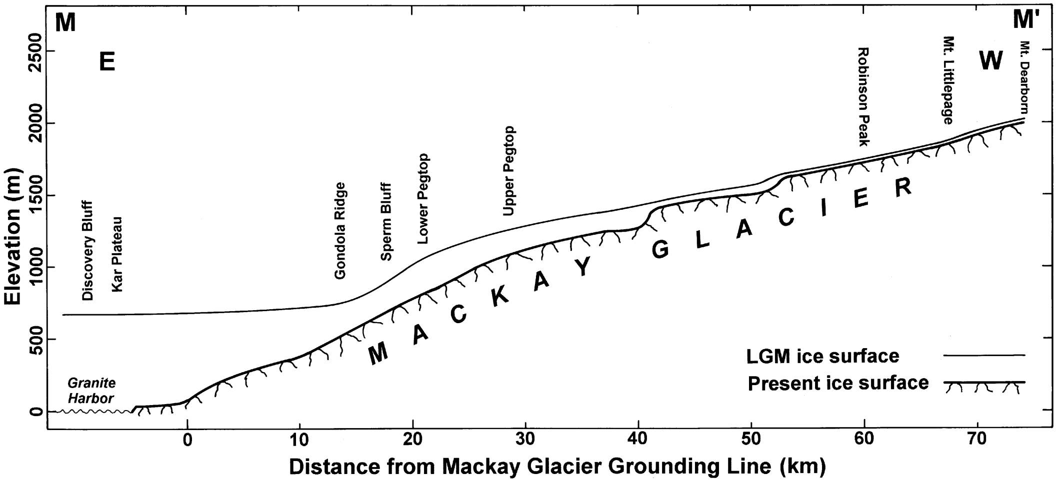


Fig. 7. Present and former (LGM) longitudinal profiles of Mackay Glacier. Position of profiles is shown as M–M[1] in Fig. 2.

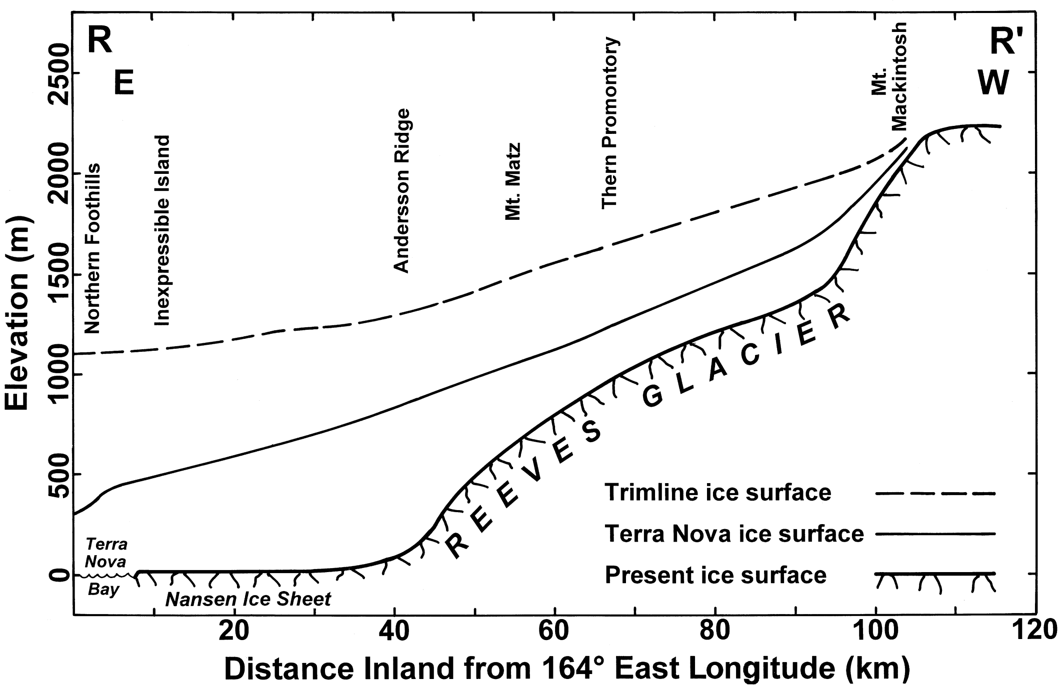


Fig. 8. Present and former longitudinal profiles of Reeves Glacier. Position of profiles is shown as R–R[1] in Fig. 2. Terra Nova surface represents LGM. See Orombelli *et al.* (1990) for details on trimlines.

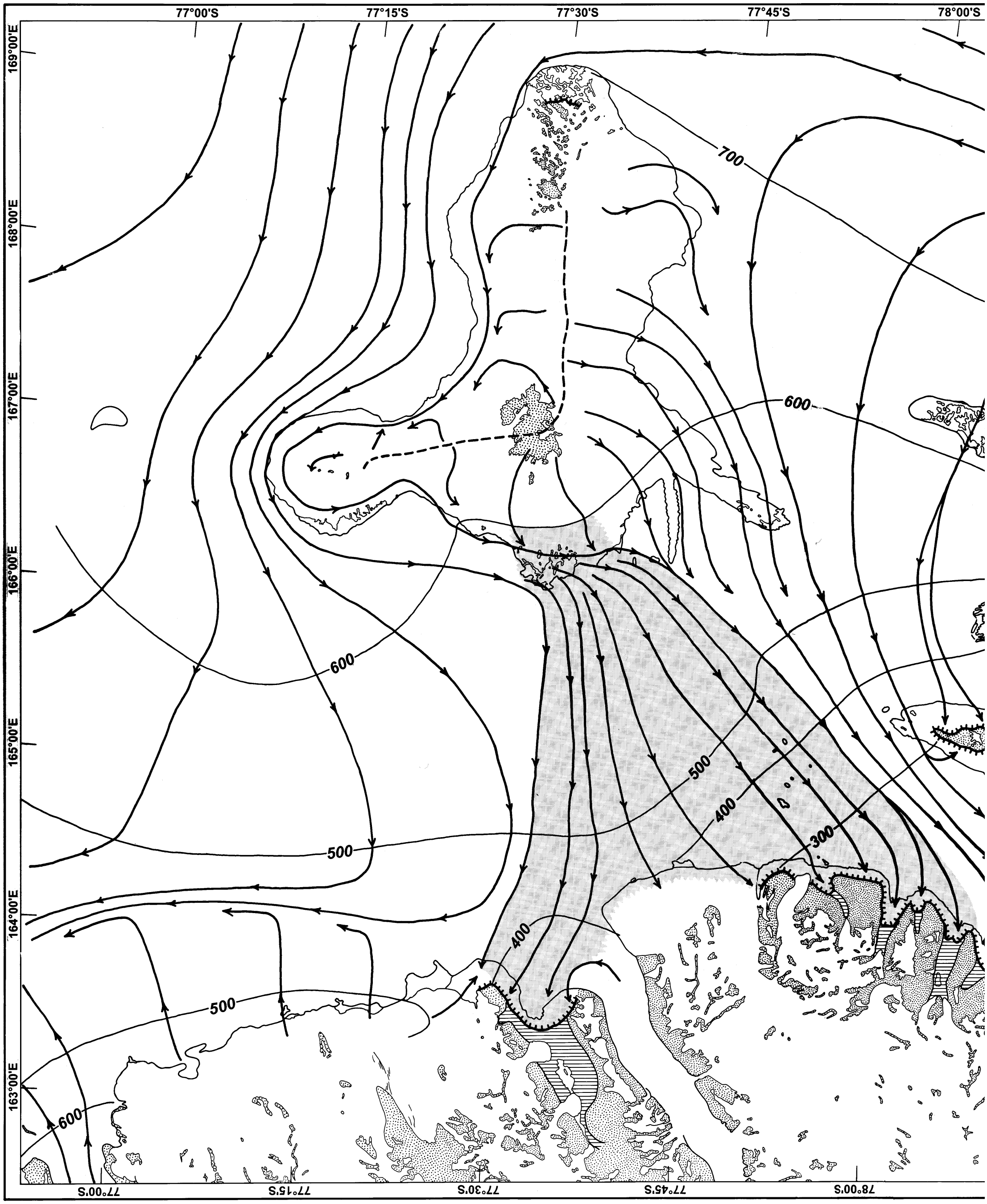
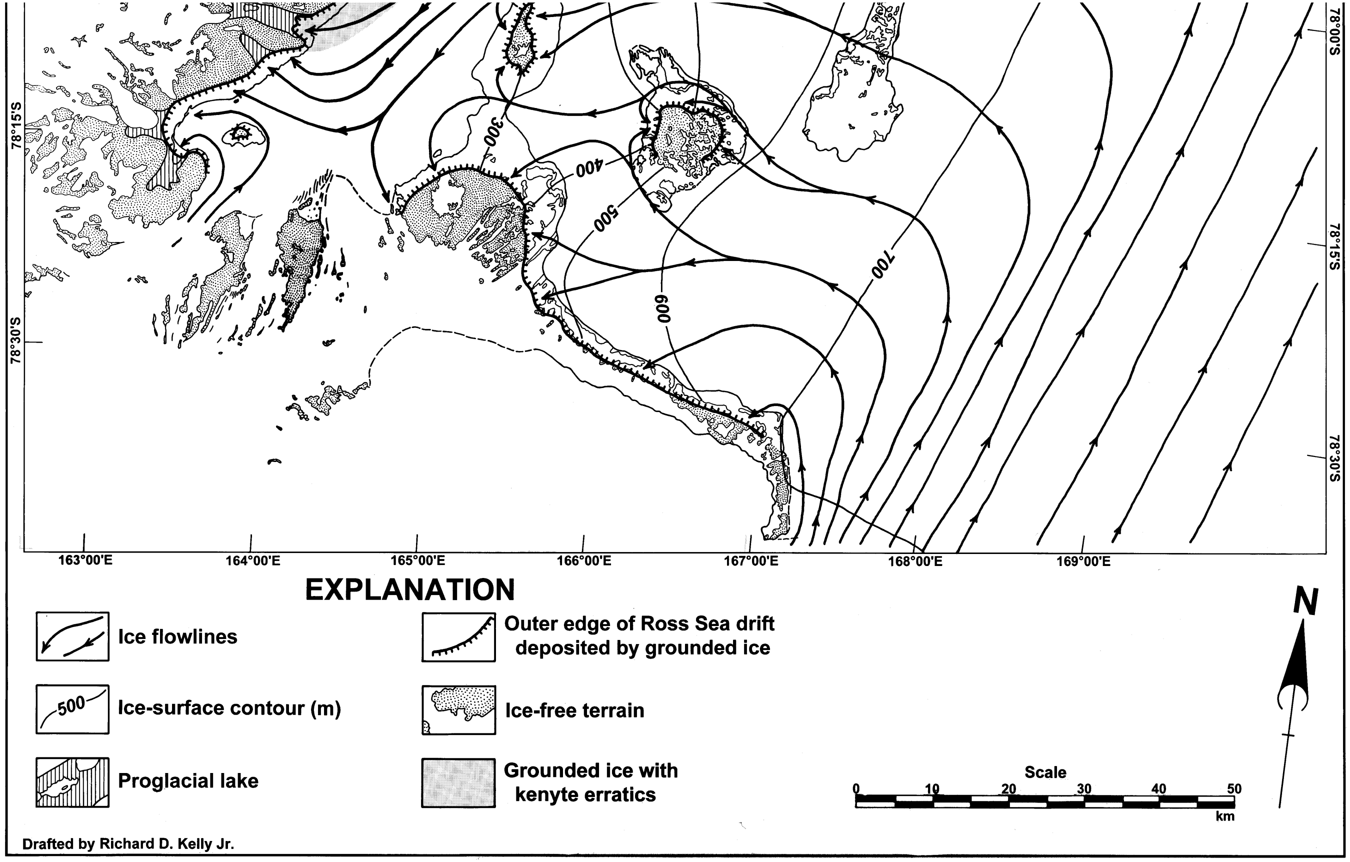


Fig. 9. Flowlines and surface contours of the grounded ice sheet that deposited Ross Sea drift in the McMurdo Sound region at the LGM. See Denton and Marchant (2000) for the geologic background for this reconstruction.


78°00'S
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168°00'E
169°00'E
EXPLANATION
Ice flowlines
Outer edge of Ross Sea drift deposited by grounded ice
500
Ice-surface contour (m)
Ice-free terrain
Proglacial lake
Grounded ice with kenyte erratics
Scale
0 10 20 30 40 50
km
N
Drafted by Richard D. Kelly Jr.

Fig. 10a. Beardmore Glacier. View is north from Plunket Point. Location of photograph is given in Fig. 2.

Fig. 10b. Beardmore drift alongside Beardmore Glacier on west side of the Dominion Range. Here the abrupt upper limit of Beardmore drift marks the LGM position of Beardmore Glacier at 60 m above the present-day ice level. See Fig. 2 for location and Denton *et al.* (1989b) for details.

sheet grounding in the Ross Embayment. The alternative explanation of increased ice flow due to higher precipitation on the polar plateau is less plausible, except possibly for northern Victoria Land. Based on the profiles, the maximum potential elevation rise of the ice surface on the polar plateau would be 35–40 m inland of Beardmore Glacier and 100 m inland of Hatherton and Darwin Glaciers. From drift geometry, Denton *et al.* (1989b) argued that the Dominion Range ice cap near the polar plateau was retracted when Beardmore Glacier stood at its expanded profile, implying lower precipitation on the Dominion Range and the adjacent polar plateau at that time. If this inference is correct, then the minor elevation change on the polar plateau inland from Beardmore Glacier would depend on the interplay between the grounding effect and the lowered precipitation and temperature. Decreased precipitation could result from lower air temperature and/or more distant open water.

A similar situation occurred at Taylor Dome on the polar plateau just inland of the Dry Valleys (Fig. 2). Ice from this dome feeds both Mackay Glacier (which flows into the Ross Sea at Granite Harbor) and Taylor Glacier (which terminates on land in Taylor Valley) (Drewry 1982). Even though both drain Taylor Dome, these glaciers showed out-of-phase behavior during the LGM. Mackay Glacier responded in the same fashion as other outlet glaciers that flow into the Ross Embayment, thickening greatly in its lower portions and only slightly in its upper reaches. Since the LGM, Mackay Glacier

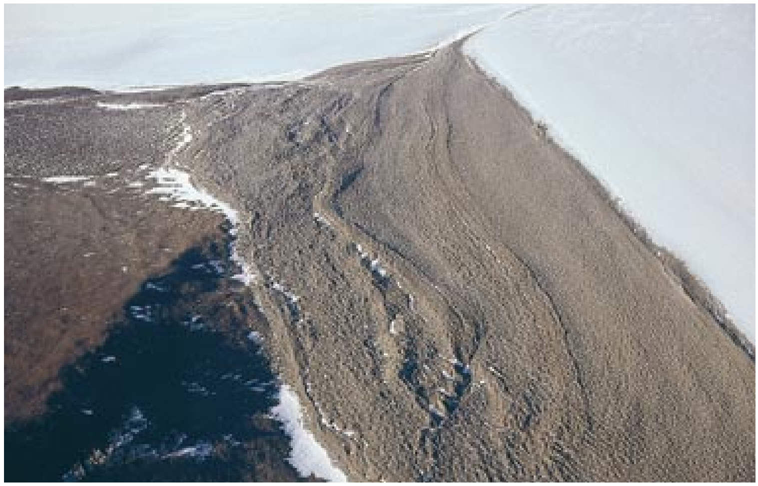

Fig. 10c. Beardmore drift at Lizard Point beside upper Beardmore Glacier. Here the sharp upper limit of Beardmore drift marks the level of Beardmore Glacier at the LGM (60–80 m above the modern ice surface). See Fig. 2 for location and Denton *et al.* (1989b) for details.

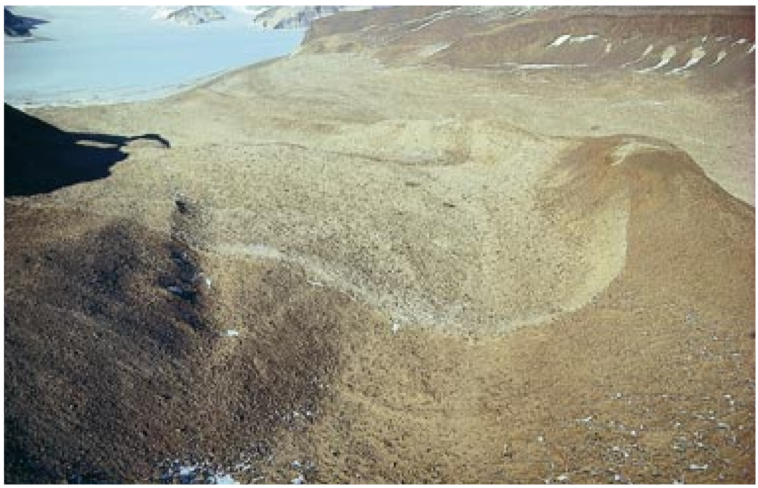

Fig. 10d. Britannia I and II drifts (light colored) in Dubris Valley alongside Hatherton Glacier (background). The sharp outer limit of the light-colored drift here marks the LGM position of Hatherton Glacier. Location of photograph is given in Fig. 2. See Bockheim *et al.* (1989) for details.

Fig. 10e. The outer limit of Britannia II drift (light colored) in Dubris Valley alongside Hatherton Glacier. Location of photograph shown in Fig. 2. See Bockheim *et al.* (1989) for details.

Fig. 10f. The outer limit of Britannia drift (light colored) in Isca Valley alongside Hatherton Glacier. Location of photograph is given in Fig. 2. See Bockheim *et al.* (1989) for details.

Fig. 10g. Mackay Glacier. View is eastward down the glacier toward the Ross Sea. See Fig. 2 for location of the photograph.

has dropped to its present level. In contrast, Taylor Glacier was less extensive than now during the LGM and has since advanced so that it currently occupies its maximum Holocene position (Denton *et al.* 1989a). This behavior of Taylor Glacier is identical to that of independent alpine glaciers in the Dry Valleys. The likely cause of this out-of-phase behavior is that Mackay Glacier was dammed, and hence thickened, by the grounded ice sheet in the Ross Sea, with lowering to its present-day longitudinal profile as the ice sheet retreated from the Ross Embayment. In contrast, Taylor Glacier was not susceptible to the damming effect because it terminated on land. As a result, it fluctuated in concert with local alpine glaciers, probably controlled by precipitation changes dictated by the growth and dissipation of grounded ice in the Ross Sea.

A possible exception to this pattern occurs in northern Victoria Land. Just as with outlet glaciers farther south, Reeves Glacier also responded to ice-sheet grounding in the western Ross Sea by altering its profile (Fig. 8). However, a reasonable case can be made that local ice caps and mountain glaciers expanded slightly at the same time as the profile of Reeves Glacier rose (Orombelli *et al.* 1990). Steig *et al.* (2000) attributed this slight northern Victoria Land glacier expansion, which contrasted so markedly with glacier behavior in southern Victoria Land, to northward displacement of low-pressure systems by the ice sheet grounded in the Ross Embayment at the LGM.

For reasons given in Denton *et al.* (1989a,b), Bockheim *et al.* (1989), and Orombelli *et al.*

Fig. 10h. The outer limit of unnamed, fresh gravel drift deposited by an expanded Mackay Glacier at the LGM. Photograph by David R. Marchant. Location of photograph is given in Fig. 2.

Fig. 10i. Aerial view of Ross Sea drift on the west coast of McMurdo Sound near the mouth of Miers Valley in the eastern foothills of the Royal Society Range. The drift is largely ice cored, and a moraine ridge usually marks its outer limit. See Fig. 2 for location of photograph.

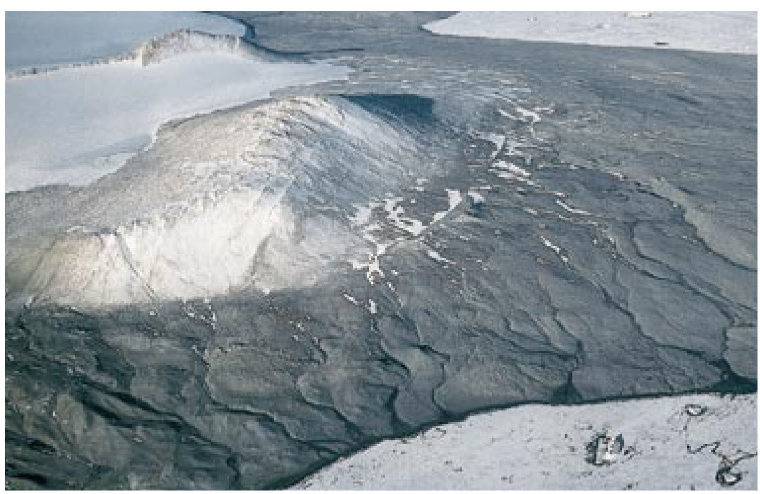

Fig. 10j. Aerial view of Ross Sea drift on the flank of Hjorth Hill alongside easternmost Taylor Valley. See Fig. 2 for location of photograph.

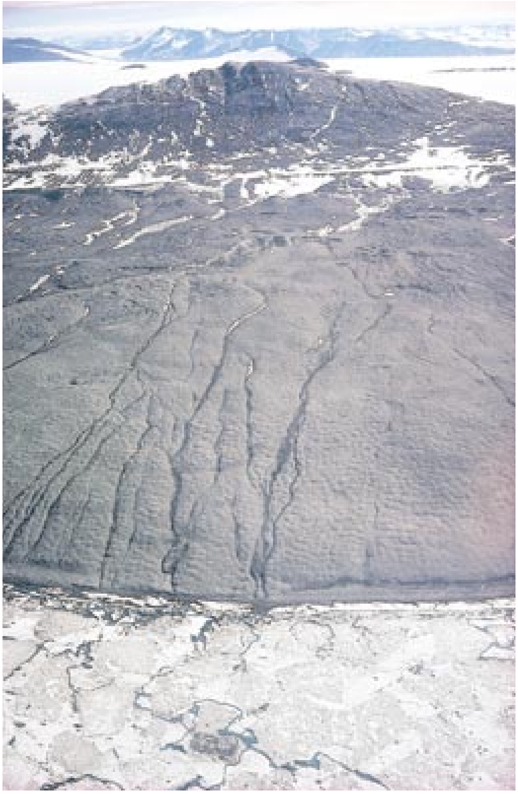

Fig. 10k. Aerial view of Ross Sea drift on the flank of Hjorth Hill alongside easternmost Taylor Valley, See Fig. 2 for location of photograph.

(1990), we correlate Beardmore, Britannia, and Terra Nova drifts with Ross Sea drift (Figs 10i–k), deposited on terrain adjacent to McMurdo Sound by the grounded Ross Sea ice sheet (Stuiver *et al.* 1981; Denton and Marchant 2000; Hall and Denton 2000a) (Fig. 9). Weathering and morphological characteristics, as well as available radiocarbon dates, are compatible with this interpretation. For example, AMS radiocarbon dates of individual shell fragments reworked into (and hence affording maximum dates of) Ross Sea drift at Cape Bird alongside McMurdo Sound and Terra Nova drift alongside Terra Nova Bay are 25,620–28,160 ^{14}C yr BP (Orombelli *et al.* 1990; Dochat *et al.* 2000). Numerous radiocarbon ages in the McMurdo Sound region show that Ross Sea drift dates to the LGM (Denton and Marchant 2000; Hall and Denton 2000a). Finally, minimum radiocarbon ages are similar (10,000–6000 ^{14}C yr BP) for Britannia, Ross Sea, and Terra Nova drifts (Bockheim *et al.*1989; Orombelli *et al.* 1990; Hall and Denton 2000a,b).

The correlation of Ross Sea drift in the McMurdo Sound region with gravel drifts alongside TAM outlet glaciers gives the terrestrial framework for reconstructing a grounded ice sheet in the Ross Embayment at the LGM. The former profiles of outlet glaciers afford surface elevations along the TAM front (uncorrected for postglacial rebound) for this grounded ice sheet of 1300 m (Beardmore), 1100 m (Hatherton), 600 m (Mackay), and 360 m (Reeves). The McMurdo Sound region is the only locality along the TAM front where East Antarctic outlet glaciers terminated inland, leaving the coast susceptible to incursions of landward-flowing grounded ice from the Ross Embayment. This ice flowed westward from the Ross Sea through McMurdo Sound to terminate against the headlands of the TAM or in lobes that plugged the eastern ends of ice-free valleys. The details are given elsewhere in this issue (Denton and Marchant 2000; Hall and Denton 2000a; Hall *et al.* 2000). Such westward flow into the sound implies a widespread grounded ice sheet in the Ross Embayment. Erratics were carried into the embayment by the dammed outlet glaciers farther south in the TAM embayment. The upper limits of such TAM erratics in Ross Sea drift on eastern Minna Bluff and Ross Island show that the surface elevation of grounded ice bypassing McMurdo Sound was 637–710 m (Denton and Marchant 2000).

Marine evidence

The Ross Sea floor features northeast–southwest trending banks, ridges, and troughs (Fig. 2). Mawson, Crary, and Pennell Banks are prominent in the western Ross Sea. These banks are 40–130 km wide, show steep sides, and have flat tops that are within 250 m of present-day sea level. The relief from bank tops to adjacent trough floors is 170–200 m. The banks chiefly comprise gently inclined Miocene strata truncated at trough edges, with only thin, isolated surface patches of younger sediments (Domack *et al.* 1999b). Narrow ridges (35–60 km wide), rather than broad banks, characterize the central and eastern Ross Sea. The ridge-to-trough relief is 60–135 m, and internal sediments display massive to chaotic geometry. The ridges have erosional bounding surfaces and some are capped by sediments with hummocks (Shipp *et al.* 1999).

The two distinct troughs in the western Ross Sea include the Victoria Land–Drygalski Basins and the JOIDES–Central Basins, respectively (Fig. 2). These troughs are 45–65 km wide. Both are fore-deepened, with shelf-break depths of about 500 m and inland depths of more than 900 m. The fore-deepened troughs in the central and eastern Ross Sea are 600–700 m deep, and 150–240 km wide.

Streamlined features are widespread in the troughs (Shipp *et al.* 1999). Lineations (up to 20 km long), interpreted to be subglacial flutes and furrows, are parallel to trough axes. Their average trend is N28°E (Shipp *et al.* 1999), but they diverge around sea-floor highs. Crest-to-crest spacing is 300–650 m. Teardrop-shaped highs on the floor of the Eastern Basin of the central Ross Sea trend N38°E–N40°E and taper seaward; individual highs are 0.5–0.75 km wide, 3–8 km long, and 40–100 m in relief (Shipp *et al.* 1999). They appear to have a core of bedrock and a thin mantle of sediment. They are interpreted as drumlins (Shipp *et al.* 1999). Finally, arcuate grooves on the sea floor of the outer shelf are inferred to be iceberg furrows (Shipp *et al.* 1999).

The distribution of lineations and of widespread basal till (seismic facies 4a of Shipp *et al.* 1999) resting on a glacial erosion unconformity define the outer limit of grounded ice in the Ross Embayment at the LGM, as depicted in Figs 3 and 4 (Shipp *et al.* 1999). In the trough including the JOIDES and Central Basins, patchy deposits of basal till (seismic facies 4a) overlie an erosional unconformity and are themselves fluted. Seismic facies 4a thickens seaward to a large grounding-line wedge at the latitude of Coulman Island. This wedge is similar to till tongues attributed by King and Fader (1986) and King *et al.* (1991) to the accumulation of basal glacial debris at the grounding line of a marine ice sheet, accompanied by progradation. Ice-proximal glaciomarine sediments (seismic facies 2 and 3 of Shipp *et al.* 1999), which extend 70 km seaward of the Coulman Island grounding-zone wedge, exhibit a fluted surface. Hence, it is inferred that the grounded ice sheet advanced across the wedge to the outer edge of the fluted glaciomarine sediments at the LGM (Shipp *et al.* 1999).

Sea-floor lineations show that the Mawson–Crary and Pennell–Ross Banks deflected seaward ice flow. The location of the LGM grounding line is based on the patchy distribution of seismic facies 4a on the bank tops. Downward-lapping wedges of seismic facies 4a (basal till) on the bank edges indicate that grounded ice rises remained on the banks after retreat of grounding lines from the troughs (Shipp *et al.* 1999). The LGM grounding line is located within Pennell Trough on the basis of the distribution of seismic facies. In the central Ross Sea the distribution of basal till, of flutes and rock drumlins, and of slope gullies together suggest that the LGM grounding line was situated at the outer edge of the continental shelf (Shipp *et al.* 1999).

From glacial lineations, Shipp *et al.* (1999) inferred that grounded ice flowed around banks and then converged into troughs and basins as ice streams with accentuated flow. Laterally shifting ice streams accreted sediment on the sides of ridges in the central Ross Sea. Ice streaming is also suggested by distinctive mineralogic signatures of sediments within individual troughs (Shipp *et al.* 1999). Thin and slow-moving ice occurred on the banks.

Glaciological background

Reconstructing the pattern of surface elevations and the flow regime for ice entering the Ross Embayment during the LGM (Figs 3, 4 and 9) draws upon (1) field observations from glacial deposits that mark former ice elevations and that can be traced to source areas, thereby deducing ice-flow directions, and (2) computer models that produce these features from known boundary conditions and from the dynamics of ice sheets. Ideally, both approaches are combined, with the field evidence from (1) providing the constraints for (2). This can be done for much of the Ross Embayment because of the terrestrial and marine data discussed above. However, for the inland ice portion of the Ross ice drainage system, we assume that ice elevations and flowlines from the interior of the East and West Antarctic Ice Sheets were not greatly different at the LGM from present-day conditions. This assumption is consistent with drift limits mapped alongside outlet glaciers that pass through the TAM (Figs 5–8) and with the geologic evidence in the McMurdo Sound region (Hall *et al.* 2000; Higgins *et al.* 2000a,b; Denton and Marchant 2000).

Ice elevations and flowlines in the Ross Embayment, as reconstructed from glacial (Denton *et al.* 1989a,b; Bockheim *et al.* 1989; Orombelli *et al.* 1990; Denton and Marchant 2000; Hall *et al.* 2000; Hall and Denton 2000b) and marine (Shipp *et al.* 1999) geology, allow two models for advance of the Antarctic Ice Sheet onto the continental shelf in the Ross Embayment (Kellogg *et al.* 1996). In the first model, present-day outlet glaciers from East Antarctica and ice streams from West Antarctica advanced with the expanding grounded ice, until they occupied subglacial troughs in the Ross Embayment that today extend from these outlet glaciers and ice streams to the edge of the Antarctic continental shelf (Bentley and Jezek 1981). In the second model, when lowered sea level caused grounding of the Ross Ice Shelf in the Ross Embayment, the bed became frozen so that the ice shelf acted like a giant plug that shut down present-day outlet glaciers and ice streams. Only near the edge of the continental shelf were ice streams able to develop in the submarine troughs. As sea level rose after the LGM, ice first became afloat in these troughs. The ice streams retreated along the troughs to their present-day positions as East Antarctic outlet glaciers and West Antarctic ice streams. The Ross Ice Shelf formed because the grounding line retreated faster than the calving front. For simplicity, we call these two models for the LGM the active ice model and the passive ice model. Both models apply if the Ross Embayment had ice streams in its outer part and a frozen bed in its inner part.

In reconstructing the flow regime during the LGM, we made use of ice-flow directions known from glacial and marine geology, along with present-day flow directions in the interior East and West Antarctic Ice Sheets. Using these flow directions, we reconstructed interior ice-sheet elevations along flowlines using equation 3.25 from Hughes (1998). This equation gives the surface slope for interior sheet flow, for stream flow in outlet glaciers and ice streams, and for shelf flow on the Ross Ice Shelf. Variables in the equation include parameters in the flow and sliding laws of glacial ice, the surface accumulation rate, the bed topography, the variable widths of ice flowbands, the side shear stress along outlet glaciers, ice streams and laterally confined ice shelves, the surface strain rates (if known), the ice thickness and velocity across ice-sheet grounding lines (known or calculated), the thawed fraction f of the bed when sheet flow dominates, and the ratio of basal water pressure P_W to ice overburden pressure P_I when stream flow dominates. In the active ice model, P_W/P_I is the dominant variable because stream flow dominates. In the passive ice model, f is the dominant variable because sheet flow dominates. When ice becomes afloat, $P_W/P_I=1$. When the bed is frozen under grounded ice, $f = 0$.

The basic idea in the model is that, as a frozen bed thaws, basal meltwater spreads out over the bed as a thin film which wets the low places first and the high places last as f increases from zero to one, with the bed being a mosaic of frozen and thawed patches for intermediate values of f. As a thawed bed freezes, the first frozen patches occur on high places and these expand until the low places are also frozen. This is the essence of sheet flow. Once the entire bed is thawed in a given section of an ice flowband, any further basal melting can no longer increase the area of the bed that is wet, so it must increase the thickness of the basal water layer. As the basal water layer thickens, it will progressively drown bedrock bumps that project into the basal ice, beginning with the smaller bumps and continuing until it drowns bumps of a size that controls the ability of ice to slide over bedrock. When this controlling bump size is drowned, ice is effectively uncoupled from the bed so that the overburden of ice is virtually entirely supported by the basal water pressure. During this transition, therefore, $f = 1$ and P_W/P_I increases from nearly zero when basal meltwater merely wets the bed, to nearly one when basal meltwater effectively drowns the controlling bedrock bumps. This is the essence of stream flow. Further thickening of the basal water layer results in shelf flow, for which resistance to flow is afforded by the sides of an embayment that contains the ice shelf and by islands or shoals within the embayment that anchor the ice shelf to specific pinning points where it is grounded locally.

In our application of this model to ice flowing into the Ross Embayment during the LGM, we took as the ice-sheet grounding line the farthest seaward limit of grounded ice mapped by Shipp *et al.* (1999), based on the geological evidence discussed previously. Shipp *et al.* (1999) identified paleo-ice streams in the three westernmost troughs on the continental shelf, and postulated that paleo-ice streams occupied the two easternmost troughs as well. Therefore, we reconstructed ice streams in all five troughs by letting P_W/P_I increase linearly from one at the ice-shelf grounding line to nearly zero at the heads of the ice streams, taken as the first major constriction of the troughs behind the grounding line. Ice elevations in the remainder of the Ross Embayment were reconstructed by adjusting thawed bed fraction f until ice elevations along flowlines were compatible with ice elevations along the TAM at sites depicted in Figs 5–9. This required that ice grounded in the Ross Embayment was on a mostly frozen bed at the LGM, as shown in fig. 3.25 in Hughes (1998). We must add, however, that this was a steady-state reconstruction. Our model also would allow a wholly frozen bed if the steady-state assumption were relaxed and the ice were allowed to thicken over time until it attained the known elevations at the LGM. This would have been the passive ice reconstruction. Because our reconstruction here includes ice streams on the outer continental shelf and sluggish sheet flow over a largely frozen bed on the inner continental shelf, it represents a combination of the active and passive ice models.

The East Antarctic outlet glaciers and the West Antarctic ice streams at the LGM had a much gentler surface slope, so the underlying bed may have been partly thawed. In that case, stream flow would have been replaced by sheet flow, and sheet flow would have continued all the way to the ice divides in East Antarctica and West Antarctica. We made this assumption in our reconstructed ice elevations along flowlines entering the Ross Embayment. We assumed that the thawed fraction of the bed along these flowlines was much the same at the LGM as it is today, because ice elevations were much the same. The present-day distribution of the thawed basal fraction *f* for sheet flow was determined by Wilch and Hughes (2000). Their methodology was discussed in Hughes (1998, pp. 67–77), where maps of the percentage thawed bed were presented in figs 3.14 and 3.15. Our assumption that basal thermal conditions have not changed substantially since the LGM in these interior regions is based on the fact that ice elevations are only weakly dependent on surface accumulation rates, so even a major reduction of accumulation rates at the LGM would result in only minor reductions of ice elevations over the Antarctic interior. Such elevation reductions would be offset by the damming effect of ice grounded on the outer continental shelf, so there was probably only minor net change in interior ice elevations.

Based on our reconstruction of ice elevations and flowlines at the LGM for Antarctic ice draining into the Ross Embayment, subsequent deglaciation in the embayment would have been triggered by rising sea level, especially the rapid rise after 12,700 ^{14}C yr BP (Fairbanks 1989). Rising sea level would have tripped the grounding-line instability postulated and quantified by Weertman (1974) and Thomas (1977), and applied to deglaciation of the Ross Embayment by Thomas and Bentley (1978) and Stuiver *et al.* (1981). Retreat of the grounding line by this instability mechanism requires that the ice streams retreat at the same rate to maintain the low stream-flow surface slope near the grounding line. Otherwise the low stream-flow surface would be converted into the steep sheet-flow surface and grounding-line retreat would stop. Ice-stream retreat at a rate fast enough to match or exceed that of the grounding line requires that the heads of ice streams retreat at these same or faster rates. In terms of ice dynamics, the head of an ice stream is the surface inflection point where the concave surface produced by stream flow is replaced by the convex surface produced by sheet flow. This is the point of maximum surface slope, and therefore is where the basal shear stress reaches a maximum value. The rate at which frictional heat is generated at the bed is determined by the product of the basal ice velocity and the basal shear stress. Therefore, basal heat production is greatest at the heads of ice streams and should cause there the greatest rate of basal meltwater production that tends to uncouple ice from the bed, and thereby convert sheet flow to stream flow. If that basal meltwater cannot be entirely discharged downstream (perhaps because sea level produces too much back-pressure), the remainder must be forced upslope, thereby causing the inflection line to retreat and the ice stream to lengthen. Robin and Weertman (1973) proposed a glacial surge mechanism by which this could happen. Hughes (1975) applied their mechanism to ice streams in the Ross Embayment to explain how gravitational collapse could continue even after sea level stabilized. Bindschadler (1996) showed that this retreat mechanism may still be lengthening some West Antarctic ice streams. Conway *et al.* (1999) postulated from geologic evidence that, in fact, grounding-line retreat has been ongoing in the Ross Embayment in the last 7000 ^{14}C yr in the absence of significant sea-level forcing (Fig. 11).

Radiocarbon chronology

The age of the reconstruction in Figs 3 and 4 is difficult to establish accurately. Radiocarbon ages of marine organic matter must routinely be corrected by 1300 ^{14}C yr, the antiquity of dissolved carbon in Southern Ocean seawater (Berkman and Forman 1996). But the major problem comes from the scarcity of carbonate fossils in marine sediment cores. Even where they are found, the carbonate remains may not be in place because of glacial reworking, sediment slumping, or furrowing by icebergs (Andrews *et al.* 1999). As a result, the marine chronology is based on AMS radiocarbon analysis of the total organic content (TOC) of sediment from deep-sea cores. Most of the dated TOC is particulate or-

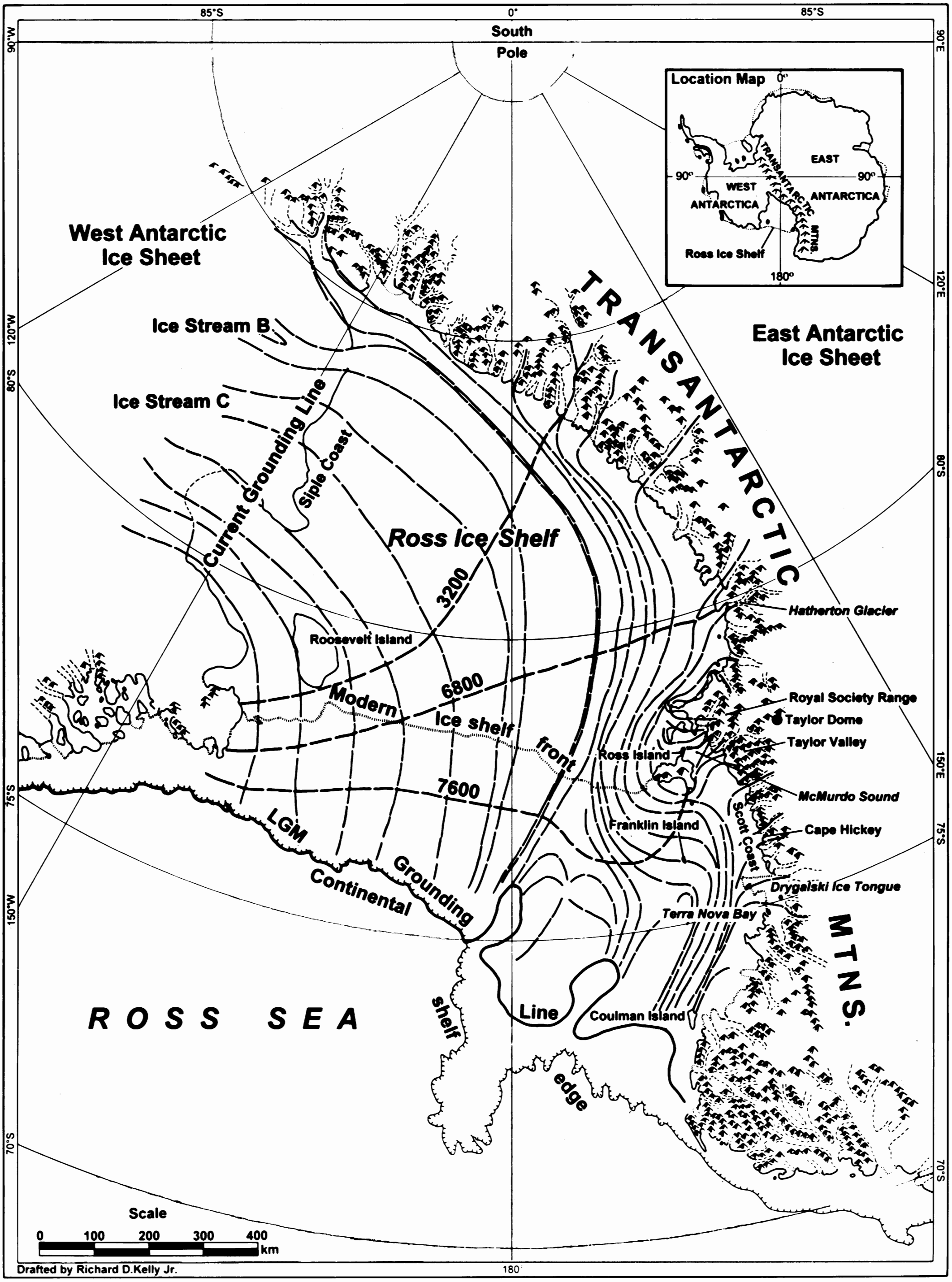

Fig. 11. Holocene grounding-line recession in the Ross Embayment. Adapted from Conway *et al.* (1999). Numbers refer to the timing of granding-line retreat in calender years.

ganic matter. In Antarctic sediment cores it is difficult to distinguish reworked from autochthonous organic material, either visually or chemically. Thus it is difficult to assess the reliability of radiocarbon dates of TOC in the marine sediment cores. For example, Domack *et al.* (1999a) concluded that the old dates of TOC from the till and ice-shelf units in the marine sediment cores are unreliable, probably because only a small fraction of the low organic content consists of autochthonous material. On the other hand, Domack *et al.* (1999a) estimated that radiocarbon dates of TOC from organic-rich portions of diatomaceous mud-and-ooze units in selected cores were not seriously affected by sediment remobilization. These mud-and-ooze units accumulated on top of siliciclastic muds after grounding-line and ice-shelf retreat. The radiocarbon dates from these units, although they are largely in stratigraphic order, must each be corrected by the age, at individual core sites, of the organic matter at the modern sediment–water interface. Thus a different correction factor, rather than an overall standard correction, was applied at each core site (Andrews *et al.* 1999), ranging from 2000 to 3800 ^{14}C yr. These variations in correction factors presumably originated from carbon fractionation during phytoplankton blooms, vital effects within surface phytoplankton, and reworking of detrital organic particles (Domack *et al.* 1999a). Taken at face value, the corrected radiocarbon dates from marine sediment cores indicate that open marine conditions existed within both the southern JOIDES Basin and the northern Drygalski Trough by 11,000 ^{14}C yr BP (Domack *et al.* 1999a). Open marine conditions had progressed southward to Granite Harbor in the western Ross Sea by 6500 ^{14}C yr BP.

An independent chronology for southward recession of the grounding line in the western Ross Sea comes from radiocarbon dates of organic material both in raised marine deposits and in glaciolacustrine sediments that were deposited in ice-dammed lakes in the McMurdo Sound region. The most reliable dates are of marine shells and of algae in deltas of ice-dammed lakes (Hall and Denton 2000a,b). Less reliable are dates of penguin remains and of algae in glaciolacustrine sediments deposited on proglacial lake floors. For reasons not yet understood, some penguin remains from Cape Bird (northwestern Ross Island) and Dunlop Island (southern Victoria Land coast) afford radiocarbon dates that are anomalously old and internally inconsistent when compared to independent chronologic and morphologic data (Dochat *et al.* 2000; C. Baroni 1998, oral communication). The dates of algae that grew on proglacial lake bottoms may need a reservoir correction due to relict dissolved inorganic carbon introduced into the lake by direct inflow from floating glacier margins (Doran *et al.* 1999). This potential problem does not apply to near-shore microbial mats, including those in deltas fed by meltwater streams (Doran *et al.* 1999).

The available radiocarbon dates of samples from the Terra Nova Bay region of northern Victoria Land appear in Orombelli *et al.* (1990, table 1). Dates of samples from raised beaches and marine deposits along the southern Victoria Land coast appear in Hall and Denton (2000b, tables 1–3). Radiocarbon dates of fossil algae in and near easternmost Taylor Valley are given in Hall and Denton (2000a, tables 1–9). Other dates of samples from Taylor Valley are in Stuiver *et al.* (1981), Denton *et al.* (1989a), and Higgins *et al.* (2000). Radiocarbon dates of samples from Cape Bird appear in Dochat *et al.* (2000, table 1). Those from elsewhere in the McMurdo Sound region are listed in Denton and Marchant (2000, table 2).

Altogether, there are now more than 450 radiocarbon dates from deposits above present-day sea level that apply to the glacial history of the western Ross Sea. The following major conclusions arise from this chronology.

– Maximum ages of 24,320 ^{14}C yr BP to >49,000 ^{14}C yr BP for ice-sheet grounding in the western Ross Sea come from retransported marine shells in drift alongside Terra Nova Bay (Orombelli *et al.* 1990), at Cape Bird on Ross Island (Dochat *et al.* 2000), at Cape Royds/Cape Barne on Ross Island (Denton and Marchant 2000), and in eastern Taylor Valley (Hall and Denton 2000a). Reworked mollusc shells in drift at Cape Barne gave a ^{234}U/^{230}Th date of 120,000 yr BP (Stuiver *et al.* 1981). Anorthoclase crystals from kenyte bedrock that has been subglacially scoured and dissected by grounded ice yielded ^{40}Ar/^{39}Ar dates of 88 ± 3 kyr to 91 ± 2 kyr at Cape Barne, 73 ± 5 kyr at Cape Royds, and 32 ± 6 kyr at Cape Evans on Ross Island (Esser *et al.* 1994).
– Grounded ice in McMurdo Sound, fed by westward inflow from the Ross Embayment, was sufficiently thick to block the mouth of Taylor Valley and dam Glacial Lake Washburn to a high level in the valley as early as 23,800 ^{14}C yr BP (Denton *et al.* 1989a). High lake levels recurred

until as late as 10,400 ^{14}C yr BP (Hall and Denton 2000a). Grounded Ross Sea ice was still sufficiently thick at the mouth of Taylor Valley as late as 8700 ^{14}C yr BP to dam a lake with a floating ice cover that transported debris, including kenyte erratics, westward across the valley-mouth threshold (78 m elevation) (Hall and Denton 2000a). Hence the pattern of westward flow of grounded ice around Ross Island and across McMurdo Sound depicted in Fig. 9 remained intact until at least this date. Grounded ice was still present in the mouth of Taylor Valley until after 8340 ^{14}C yr BP, the age of the youngest lacustrine delta that requires an ice dam. The general picture of a high-level ice-dammed proglacial lake during the LGM is replicated by the radiocarbon chronology of Lake Trowbridge in Miers Valley, farther south on the west coast of McMurdo Sound (Clayton-Greene *et al.* 1988).

– Radiocarbon dates of moraines add detail to the general outline afforded by proglacial Lake Washburn. Grounded Ross Sea ice formed the Hjorth Hill moraine, which marks the LGM extent near Taylor Valley, between about 12,700 and 14,600 ^{14}C yr BP (Hall and Denton 2000a). As late as shortly after 12,000–12,500 ^{14}C yr BP, westward-flowing grounded ice terminated at the threshold moraine at the mouth of Taylor Valley (Hall and Denton 2000a). Ice still terminated near the Hjorth Hill moraine at 10,700 ^{14}C yr BP, but began to thin rapidly after 10,445 ^{14}C yr BP (Hall and Denton 2000a). Farther south in the McMurdo Sound region, grounded Ross Sea ice still stood at the maximum LGM moraine near Blue Glacier as late as 12,330 ^{14}C yr BP, but had receded considerably by 9490 ^{14}C yr BP (Stuiver *et al.* 1981; Denton and Marchant 2000).

– Shipp *et al.* (1999) argued that the first recession from the LGM grounding-line position occurred in the western Ross Sea, while the ice sheet farther east remained grounded near the outer edge of the continental shelf. That southward retreat of the grounding line reached McMurdo Sound shortly before 6500 ^{14}C yr BP is indicated by the ages of shells from raised marine deposits in the mouth of Taylor Valley (Hall and Denton 2000a), by the shape and radiocarbon chronology of relative sea-level curves for the southern Victoria Land coast (Hall and Denton 1999, 2000b), by the chronology of deglaciation of eastern Taylor Valley (Hall and Denton 2000a), by the ages of marine shells in the McMurdo Ice Shelf (Stuiver *et al.* 1981; Kellogg *et al.* 1990; Denton and Marchant 2000) and in McMurdo Sound (Licht 1996), and by the age of an algal mat near sea level along the west coast of McMurdo Sound (Stuiver *et al.* 1981; Denton and Marchant 2000).

– The chronology for grounding-line recession from the LGM position (Figs 3 and 4) southward to McMurdo Sound is less certain. The shape and radiocarbon chronology of the relative sea-level curve from Terra Nova Bay in northern Victoria Land (Orombelli *et al.* 1990) are similar to that along the southern Scott Coast (Hall and Denton 2000b). Moreover, the oldest dates of shells preserved in raised marine sediments and ice shelves in these two areas are nearly identical (Stuiver *et al.* 1981; Kellogg *et al.* 1990; Orombelli *et al.*1990; Denton *et al.* 2000; Hall and Denton 2000a,b). Taken in isolation, these data imply rapid deglaciation of the western Ross Sea from Terra Nova Bay south to McMurdo Sound shortly before 6500 ^{14}C yr BP. However, radiocarbon dates of TOC from core KC 31 suggest that grounding-line recession had cleared Drygalski Trough several thousand years prior to this time (Domack *et al.* 1999a). Moreover, there are a few old dates of penguin remains on the coast of the western Ross Sea south of Terra Nova Bay that possibly indicate earlier deglaciation (Baroni and Orombelli 1994), if they are reliable.

– Grounding-line migration southward through the inner Ross Embayment to the Siple Coast (Fig. 11) was dated at three locations: McMurdo Sound (shortly before 6500 ^{14}C yr BP), the mouth of Hatherton and Darwin Glaciers at the TAM front (shortly before 6000 ^{14}C yr BP), and Roosevelt Island, an ice dome surrounded by the Ross Ice Shelf (3200 cal. yr BP) (Stuiver *et al.* 1981; Bockheim *et al.* 1989; Conway *et al.* 1999; Hall and Denton 1999). These results indicate that most of the grounding-line recession occurred in middle to late Holocene time. The reconstruction of this retreat history in Conway *et al.* (1999) shows a swinging-gate pattern of deglaciation, with the grounding line hinged north of Roosevelt Island until 3200 cal. yr BP. Conway *et al.* (1999) concluded that current grounding-line recession along the Siple Coast may be the manifestation of ongoing long-term retreat underway since the early Holocene. The implication is that the grounding line may continue to recede. Hence, deglaciation of the West

Antarctic Ice Sheet may be largely responsible for the 3–5 m rise of eustatic sea level in about the last 6000 ^{14}C yr BP (Fleming *et al.* 1998).

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