primary source of meltwater pulse 1B. This conclusion is not strongly dependent on the ice sheet model.

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16 July 1999; accepted 9 September 1999

Past and Future Grounding-Line Retreat of the West Antarctic Ice Sheet

H. Conway,[1] B. L. Hall,[2,3] G. H. Denton,[2] A. M. Gades,[1] E. D. Waddington[1]

The history of deglaciation of the West Antarctic Ice Sheet (WAIS) gives clues about its future. Southward grounding-line migration was dated past three locations in the Ross Sea Embayment. Results indicate that most recession occurred during the middle to late Holocene in the absence of substantial sea level or climate forcing. Current grounding-line retreat may reflect ongoing ice recession that has been under way since the early Holocene. If so, the WAIS could continue to retreat even in the absence of further external forcing.

The grounding line of the WAIS has retreated nearly 1300 km since the Last Glacial Maximum (LGM) about 20,000 years before present (yr B.P.), when grounded ice in the Ross Sea Embayment extended almost to Coulman Island (*1–3*) (Fig. 1). Complete collapse of the WAIS would cause sea level to rise 5 to 6 m. Estimates of the present stability of the WAIS are hampered by uncertainties in the overall mass balance (*4*) and uncertainties concerning the dynamic response of the ice sheet to changes in sea level or climate. It is thought that it would take $\sim 10^4$ years for the WAIS to reach equilibrium after a perturbation (*5*), but accurate assessment is difficult because the dynamics of the present ice sheet is dominated by ice streams. Fast-flowing ice streams evacuate inland ice rapidly, but field evidence indicates that abrupt changes from fast to slow flow have occurred in the past (*6*). We look to the deglacial history of the WAIS for clues about its future. Below we present dates from three locations, southern Scott Coast, Hatherton Coast, and Roosevelt Island (Fig. 1), that resolve the Holocene deglaciation of the Ross Sea Embayment.

At the LGM, outlet glaciers that flowed through the Transantarctic Mountains and across the coast thickened substantially where they merged with grounded ice filling the Ross Sea Embayment. Only along the southern Scott Coast adjacent to McMurdo Sound, 450 km south of Coulman Island, did the Ross Sea ice sheet terminate on land in the mouths of ice-free Taylor Valley and dry valleys fronting the Royal Society Range (Fig. 1). This peculiar situation arose because only here did East Antarctic ice and alpine glaciers terminate well inland, leaving the coast susceptible to incursions of landward-flowing grounded ice at the LGM. Over 200 ^{14}C dates of lacustrine algae from proglacial lakes (*3*, *7*, *8*), dammed in these valleys by grounded ice, show that the Ross Sea ice sheet was close to its LGM position from at least 27,820 to 12,880 calendar yr B.P. (*9*).

The grounding line was still north of McMurdo Sound 9420 yr B.P.; this date corresponds to the youngest delta of a proglacial lake dammed in Taylor Valley by grounded Ross Sea ice (*3*, *8*). Two thousand years later, the McMurdo Sound region was free of grounded ice, based on two lines of evidence. First, molluscs recolonized the area after the

[1]Geophysics Program, University of Washington, Seattle, WA 98195, USA. [2]Institute for Quaternary Studies and Department of Geological Sciences, University of Maine, Orono, ME 04469, USA. [3]Department of Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, MA 02543, USA.

grounding line retreated; the oldest shell from the region dates to 7550 yr B.P. (Table 1). Second, seasonal open water is required to form beaches. Dates of shells, sealskin, and other remains incorporated within raised beaches along the southern Scott Coast (*10*) were used to construct a range of relative sea level (RSL) curves (Fig. 2). With an exponential shape typical of formerly glaciated regions undergoing isostatic rebound (*11*), the curves suggest that the highest raised beach (at Cape Ross) formed ~7500 yr B.P. Open marine conditions must have existed then. Therefore, the grounding line must have retreated past McMurdo Sound between 9400 yr B.P., when grounded ice dammed proglacial lakes, and ~7600 yr B.P., when molluscs and beaches indicate that marine conditions had been reestablished. The rapid uplift indicated by the RSL curves at ~7600 yr B.P. suggests that glacial unloading occurred late in this interval. Dates of samples from the 8-m raised beach at Cape Bird on Ross Island (*12*) and from Franklin Island 100 km farther north (*3*) all lie within this range of RSL curves, suggesting nearly simultaneous glacial unloading through the entire southern Scott Coast–McMurdo Sound region.

Hatherton Glacier is an outlet glacier draining East Antarctic ice through the Transantarctic Mountains south of McMurdo Sound. Its longitudinal profile was controlled by the thickness of grounded Ross Sea ice along the adjacent coast. At the LGM, the surface elevation of the lower section of the glacier rose 1100 m, whereas that of the upper segment remained virtually unchanged (*13*). Radiocarbon dates of algae from ice-dammed lateral lakes show that the central section of the glacier was still near the LGM limit (about 450 m above present ice level) at 10,560 and 11,950 yr B.P. (*13*). As late as 9420 yr B.P., the glacier was still at least 170 m above its present level. However, dates of algal samples adjacent to the current margin range from 2125 to 6800 yr B.P. (*13*), suggesting that Hatherton Glacier was at or close to its present level ~6800 yr B.P. Because Ross Sea ice no longer influenced the longitudinal profile of the glacier, the grounding line must have retreated past the Hatherton Coast ~6800 yr B.P.

Roosevelt Island, now an ice dome situated within the Ross Ice Shelf, is grounded about 200 m below sea level. Reconstructions of the WAIS suggest that the ice surrounding the island was probably grounded and about 500 m thicker during the LGM (*3*). It is likely that the present dome morphology evolved during deglaciation (*14*). We measured ice thickness, stratigraphy, accumulation rate, surface topography, and velocity on the island (*15*). Ice thickness at the summit, $H = 744 \pm 10$ m, and the accumulation rate, $\dot{b} = 0.18$ m year^{-1} ice equivalent, set the characteristic ice flow time scale $H/\dot{b}$ to ~4000 years. Our measurements of accumulation and the spatial pattern of internal layering show no evidence of scouring at the divide. Hence, the bumps in the layering beneath the divide (Fig. 3A) are probably caused by the special flow pattern found at ice divides, rather than by a local accumulation deficit (*16*, *17*). Before divide flow began, internal layers (isochrones) were essentially flat and featureless. The bumps in the layering today began to develop at the onset of divide flow. Our ice flow layer-tracking model (*18*)

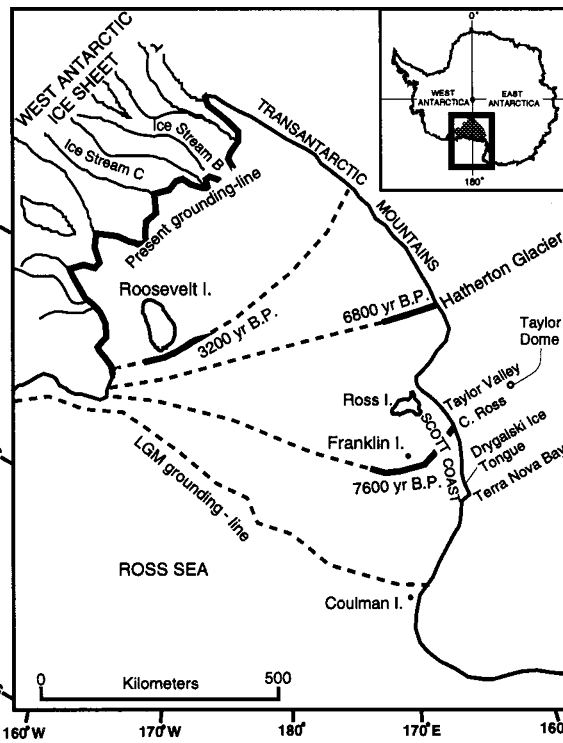

Fig. 1. Map showing dated locations used to resolve Holocene grounding-line retreat to its present position in the Ross Sea Embayment. Although the detailed structure of past grounding-line positions is unknown, dotted lines show the simplest grounding-line pattern consistent with the dates in the text.

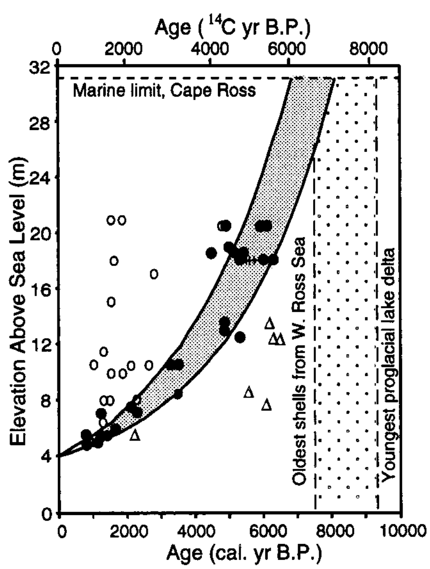

Fig. 2. Dates and elevations of organic material incorporated within raised beaches and deltas on the southern Scott Coast (*10*). Raised beaches are inferred to be former storm beaches that are 4 m above contemporary sea level (the elevation of modern storm beaches). Three samples (+) were from a delta, which would have formed at sea level; their elevation has been adjusted upward by 4 m to compare them with beach samples. These data, together with samples that date the raised beaches (●), were used to construct a range of exponential RSL curves (shaded band). Six samples from Kolich Point (△) are from sediments beneath the beaches and thus yield maximum ages, which must lie to the right of the curves. Samples from the tops of the beaches (○) are minimum ages, which must lie to the left of the curves. Uncertainties in elevation are less than ±2 m; uncertainties in age are smaller than the width of the samples. The speckled band gives the time of grounding-line migration past the southern Scott Coast.

Table 1. Ages of oldest postglacial shells collected from marine cores in the western Ross Sea Embayment north of Ross Island (*2*), from the McMurdo Ice Shelf (*3*, *32*), and from raised marine deposits along the Scott Coast (*10*).

Laboratory number	Age (^{14}C yr B.P.)	Age (Cal yr B.P.)
Shells from sediment cores		
AA-11876	6630 ± 60	7550
Shells from McMurdo Ice Shelf		
QL-1443	6450 ± 90	7420
Shells from raised marine sediments		
AA-26515	5540 ± 50	6300
AA-26500	5490 ± 55	6290
AA-26515	5480 ± 56	6280
AA-26515	5350 ± 60	6170
AA-26515	5275 ± 60	5990
AA-26515	5250 ± 97	5965

predicts that, for progressively more recent onset times, both the bump amplitude and the depth of its maximum today should be progressively smaller. The bump-amplitude profile also depends on the rate at which the ice cap is thinning, that is, the amount by which the downward surface velocity exceeds the accumulation rate. Ice at a particular depth in a thinning dome is older compared with ice in a dome of constant thickness because it has traveled farther from its higher point of origin. The bump amplitude in a thinning dome is larger because there has been more time to accumulate differential vertical strain between the divide and the flanks.

We used the distinctive signature of the bump-amplitude profile at the summit of Roosevelt Island to estimate when ice divide flow first started. The present thinning rate at the summit, calculated from the difference between the accumulation rate and the horizontal flux divergence of ice (*19*), is 0.06 to 0.11 m year^{-1}. The best model fit to the measured bump-amplitude profile, assuming constant thinning of 0.09 m year^{-1} and accumulation of 0.18 m year^{-1}, indicates that divide flow started 3200 yr B.P. (Fig. 3B). Histories that do not include long-term thinning within the range 0.06 to 0.11 m year^{-1} give poorer fits to the measurements; models with acceptable fits all require divide flow to begin 3000 to 4000 years ago. From glaciological and hydrostatic considerations, we suspect that divide flow started before the surrounding ice was floating (*14*, *20*). That is, the grounding line was still north of Roosevelt Island 3200 yr B.P.

These data, summarized in Fig. 1, show Holocene grounding-line retreat in the Ross Sea Embayment. We do not reconstruct recession from the LGM position to Franklin Island (our northernmost data point), because the timing of retreat in this region is not yet well resolved. For example, an envelope of RSL change constructed from bracketing ages of raised beaches at Terra Nova Bay (*21*) is in general agreement with the Scott Coast RSL curve (Fig. 2), implying that these two regions deglaciated within ~1000 years of each other. However, a few dates of penguin guano at Cape Hickey and near Terra Nova Bay suggest deglaciation as early as 13,800 yr B.P. (*22*). Ages derived from acid-insoluble organic material from marine sediments are also problematic (*23*). Recent work has shown that recycling of old carbon in the western Ross Sea has caused uncorrected ^{14}C ages to be much older than can be explained by the marine reservoir effect (*24*). Furthermore, the necessary correction varied from 1300 to 3200 years, depending on the region sampled. Such large uncertainties make it difficult to accurately resolve when this region deglaciated.

Regardless of the history north of Franklin Island, our reconstructions are consistent with a swinging-gate pattern of deglaciation (*3*) in which the grounding line was hinged north of Roosevelt Island until 3200 yr B.P. (Fig. 1). The grounding line was probably just north of Cape Ross 7600 yr B.P.; most recession in the Ross Sea Embayment took place during the middle and late Holocene. Additional support for this timing of deglaciation comes from evidence that trajectories of moisture-bearing storms reaching Taylor Dome were diverted while thick grounded ice occupied the Ross Sea Embayment but that the modern pattern was reestablished ~6000 yr B.P. (*25*).

This timing of deglaciation, after most of the sea-level rise from melting of Northern Hemisphere ice sheets had occurred, is consistent with studies of sea level change that suggest onset of Antarctic deglaciation after 11,000 yr B.P. (*11*, *26*). Given that advance and retreat of the WAIS generally have followed glacial-interglacial cycles (*3*, *8*), events such as rising sea level due to melting of Northern Hemisphere ice sheets or enhanced melting at Antarctic submarine margins likely triggered the onset of ice recession from the Ross Sea Embayment. However, once set in motion, recession to the Siple Coast continued throughout the Holocene in the absence of further external forcing. In fact, global sea level has risen 10 to 15 m since the grounding line retreated past the southern Scott Coast (*26*), but RSL in the region has decreased ~4 mm year^{-1} (*27*); other things being equal, this should cause grounding-line advance (*28*). However, ice sheet thinning (~27 mm year^{-1}) is much faster than RSL lowering, so the present grounding-line retreat is not surprising.

The grounding line today is about 900 km from McMurdo Sound, which implies average recession of about 120 m year^{-1} for at least 7500 years. Recent measurements indicate that grounding-line retreat is continuing at about the same rate; at Ice Stream C, it retreated about 30 m year^{-1} between 1974 and 1984 (*29*), whereas at Ice Stream B, it withdrew ~450 m year^{-1} over the past 30 years (*30*). There is no evidence to indicate that recession is slowing; it will likely continue (at least in the near future) because the current mass balance of the Ross Sea Embayment is strongly negative (*4*). Others have suggested that West Antarctica deglaciated completely in the past and, if the grounding line continues to pull back at the present rate, complete deglaciation will take about 7000 years (*31*).

We suggest that modern grounding-line retreat is part of ongoing recession that has been under way since early to mid-Holocene time. It is not a consequence of anthropogenic warming or recent sea level rise. In other words, the future of the WAIS may have been predetermined when grounding-line retreat was triggered in early Holocene time. Continued recession and perhaps even complete disintegration of the WAIS within the present interglacial period could well be inevitable.

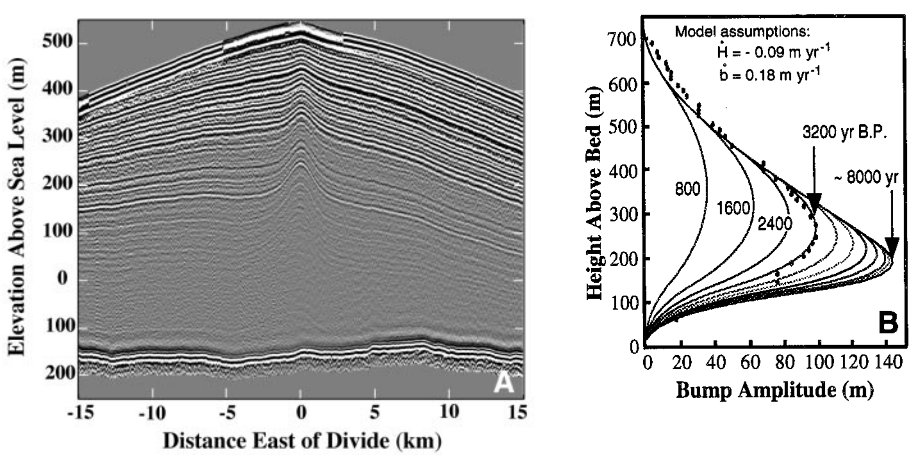

Fig. 3. (**A**) Stratigraphic section across Roosevelt Island measured by ice-penetrating radar. Vertical exaggeration is ~30:1. Reflecting horizons within the ice, assumed to be isochrones, are caused by variations in dielectric properties, mainly electrical conductivity. The bright reflector ~200 m below sea level is the bed. At the divide (0 km), ice thickness is 744 ± 10 m. Bumps in the layering beneath the divide have evolved because of ice divide dynamics since the divide formed. (**B**) Measured (*) variations of bump amplitude with depth at the summit today, compared with predicted amplitudes for various times of divide onset in the past, indicate that divide flow started 3200 yr B.P.

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4. The mass balance is strongly negative for active ice streams but positive for Ice Stream C, which stopped streaming about 150 years ago. The overall mass balance for the Ross Embayment is about -23 ± 15 km^3 $year^{-1}$, which corresponds to an average surface lowering of 27 ± 21 mm $year^{-1}$ [S. Shabtaie and C. R. Bentley, *J. Geophys. Res.* **92**, 1311 (1987)]. The large uncertainty arises because mass balance is the difference between two large numbers, each with large uncertainties; the few available point measurements of accumulation and thinning show large spatial variability.
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14. Our conceptual model invokes northward or westward flow of thick WAIS ice across Roosevelt Island during glacial times. During deglaciation, the thicker ice in the deep surrounding troughs underwent more rapid strain thinning than did the relatively shallow ice of Roosevelt Island, producing first a N-S ridge and then an isolated dome.
15. Variations of ice thickness and internal stratigraphy were measured from radio echo-sounding traverses with a center frequency of 7 MHz, which corresponds to a wavelength of ~20 m in ice. Surface velocities and topography were measured by repeat surveys of poles with the Global Positioning System. Accumulation rate was estimated from stratigraphic profiles in three 2-m snow pits and from beta activity of three 16-m cores analyzed by J. Dibb (University of New Hampshire). Spikes in beta activity represent horizons deposited in 1955 and 1965; the depth of the horizon, together with measurements of the firn density profile, was used to calculate the average accumulation rate.
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18. The model followed layers as they moved downward under the ice divide and under a flank. The two ice flow regimes were represented with three parameters: the downward velocity at the surface and two fractional heights above the bed, h_{divide} and h_{flank}, at which the downward velocity profile changes from quadratic to linear on the divide and flanks [W. Dansgaard and S. J. Johnsen, *J. Glaciol.* **8**, 215 (1969)]. A finite element model (*16*), constrained by our measured surface and bed topography, guided the choice of values for h_{divide} and h_{flank}.
19. Ice cap thinning results from an imbalance between accumulation and the downward velocity at the surface and was calculated from $\dot{H} = \dot{b} - \nabla \cdot Q$, where $\dot{H}$ is the thinning rate, $\dot{b}$ is the ice equivalent accumulation rate, and $\nabla \cdot Q$ is the horizontal flux divergence [L. A. Rasmussen, *J. Glaciol.* **31**, 115 (1985)]. The flux divergence comes from measurements of ice thickness H and the divergence of the horizontal surface velocity $\nabla \cdot U_S$; that is, $\nabla \cdot Q = \gamma H \nabla \cdot U_s$. The parameter γ varies from about 0.8 in the case of no sliding with flow dominated by bed-parallel shear to 0.6 as might occur beneath a nonsliding divide (*16*). Analysis of pole positions surveyed during the early 1960s [J. L. Clapp, *Univ. Wisconsin Res. Rep. 65-1* (1965)] indicates that $\nabla \cdot U_s$ at the summit of Roosevelt Island is $\sim 5.1 \times 10^{-4}$ $year^{-1}$, which, for $\dot{b} = 0.18$ m $year^{-1}$, implies thinning by 0.06 to 0.11 m $year^{-1}$.
20. Bedrock beneath Roosevelt Island is ~200 m below present-day sea level (Fig. 3), and the surrounding trough is about 500 m below sea level. If ice surrounding the island was floating (requiring ice <560 m thick, or less if sea level was lower), and the ice surface across the island was flat (no dome), then ice on the island would have to be ~260 m thick. This is unlikely because it requires thickening of ~460 m to achieve present morphology; this conflicts with the continuity analysis (*19*) and with the pattern of bumps beneath the divide (Fig. 3), both of which imply thinning.
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33. This work was funded by the NSF through Office of Polar Program grants 9118678, 9420648, and 9615347. Carbon-14 samples were processed at the NSF-Arizona AMS facility and at the University of Washington. We thank R. Ackert, B. Andersen, C. Bentley, J. Bockheim, M. Conway, R. Kelly, N. Lord, T. Lowell, N. Nereson, B. Overturf, B. Smith, M. Stuiver, and S. Wilson for valuable contributions. G.H.D. had the idea to study the future of the West Antarctic Ice Sheet by examining its past. B.L.H. and G.H.D collected and interpreted the geologic and ^{14}C data. A.M.G. and H.C. collected and interpreted the geophysical data from Roosevelt Island. E.D.W. conceived and implemented the internal layer evolution model. B.L.H. and H.C. wrote the report.

21 May 1999; accepted 23 August 1999

Tributaries of West Antarctic Ice Streams Revealed by RADARSAT Interferometry

Ian Joughin,[1]* Laurence Gray,[2] Robert Bindschadler,[3] Stephen Price,[4] David Morse,[5] Christina Hulbe,[3] Karim Mattar,[2] Charles Werner[1]

Interferometric RADARSAT data are used to map ice motion in the source areas of four West Antarctic ice streams. The data reveal that tributaries, coincident with subglacial valleys, provide a spatially extensive transition between slow inland flow and rapid ice stream flow and that adjacent ice streams draw from shared source regions. Two tributaries flow into the stagnant ice stream C, creating an extensive region that is thickening at an average rate of 0.49 meters per year. This is one of the largest rates of thickening ever reported in Antarctica.

The West Antarctic Ice Sheet, which would raise sea level by 5 to 6 m if it melted, has been a subject of intense glaciological study since doubts about its stability were first raised (*1*). Unlike the Greenland Ice Sheet and most of the East Antarctic Ice Sheet, much of the West Antarctic Ice Sheet is grounded below sea level and underlain by marine sediments. When saturated with water (*2*), these sediments may affect the dynamics of ice motion by allowing fast movement. Although the possibility of a catastrophic collapse of the ice sheet is under debate (*3*), field and satellite observations have established that substantial changes are occurring in West Antarctica (*4–6*), particularly in the ice streams.