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Keel depths of modern Antarctic icebergs and implications for sea-floor scouring in the geological record

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Abstract

Icebergs affect the geological record through the scouring or ploughing action of their keels, reworking sediments where they contact the sea floor. Satellite radar-altimeter measurements of the surface elevation of the marine margins of the Antarctic Ice Sheet are inverted to produce ice-thickness values. These data are assumed to represent the keel depths of newly-calved icebergs. There are three peaks in the frequency distribution of Antarctic iceberg–keel depths: (a) about 140–200 m in thickness, associated with smaller ice shelves fringing Antarctica (e.g. Wilkins, George VI and Brunt ice shelves); (b) about 250–300 m, from large ice shelves (e.g. Ross, Ronne, Amery); and (c) approximately 500–600 m, from fast-flowing ice-sheet outlet glaciers (e.g. Pine Island and Dibble glaciers), together with the Filchner Ice Shelf. Controls on iceberg keel depth are ice thickness at the grounding line, creep thinning and the rate of melting/freezing at the floating ice-shelf base with distance to the calving margin. The distribution of sea-floor scours produced by iceberg-grounding is dependent on the changing form and flow of the glaciers and ice sheets from which icebergs are derived and on berg drift tracks and melt rates. Sea-floor evidence for icebergs with particularly deep keels has sometimes been interpreted to indicate the presence of extensive former ice shelves. However, the largest modern Antarctic ice shelves do not produce large numbers of icebergs thicker than about 350 m, implying that sea-floor scours found in deeper water are unlikely to be an indicator of extensive past ice-shelf development. Iceberg scours found at water depths in excess of about 500 m in the geological record are, instead, probably indicative of either: (a) the former presence of fast-flowing ice-sheet outlet glaciers; (b) ice shelves fed from major interior basins where lateral spreading is constrained topographically; or (c) a pulse of large icebergs released during major deglaciation or ice-sheet collapse.

Keywords: Antarctica; icebergs; iceberg keel depths; ice shelves; satellite radar altimetry; sea-floor scouring

1. Introduction

Icebergs are produced at the marine margins of glaciers and ice sheets. They affect the geological record through the release of ice-rafted debris and by the scouring or ploughing action of their keels where they contact the sea floor. The morphology of icebergs is controlled by the dynamics of the parent ice mass and the nature of the interaction between the ice margin and the adjacent ocean. The horizontal dimensions of icebergs and their distributions and drift tracks in the polar seas have been investigated, in part to obtain evidence on the rate of mass loss from the parent ice sheet (e.g. Morgan and Budd, 1978; Orheim, 1985; Wadhams, 1988; Jacobs et al., 1992; Long et al., 2002). Much less is known, however, about the

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below-water extent of iceberg keels (cf. Dowdeswell et al., 1992), yet this is of significance in high-latitude waters as a result of the disturbance and reworking of sea-floor sediments (e.g. Woodworth-Lynas et al., 1985; Barnes and Lien, 1988; Dowdeswell et al., 1993, 1994; Syvitski et al., 1996; O'Brien et al., 1997), and the potential hazard to sea-floor engineering structures (e.g. Woodworth-Lynas and Guigné, 1990).

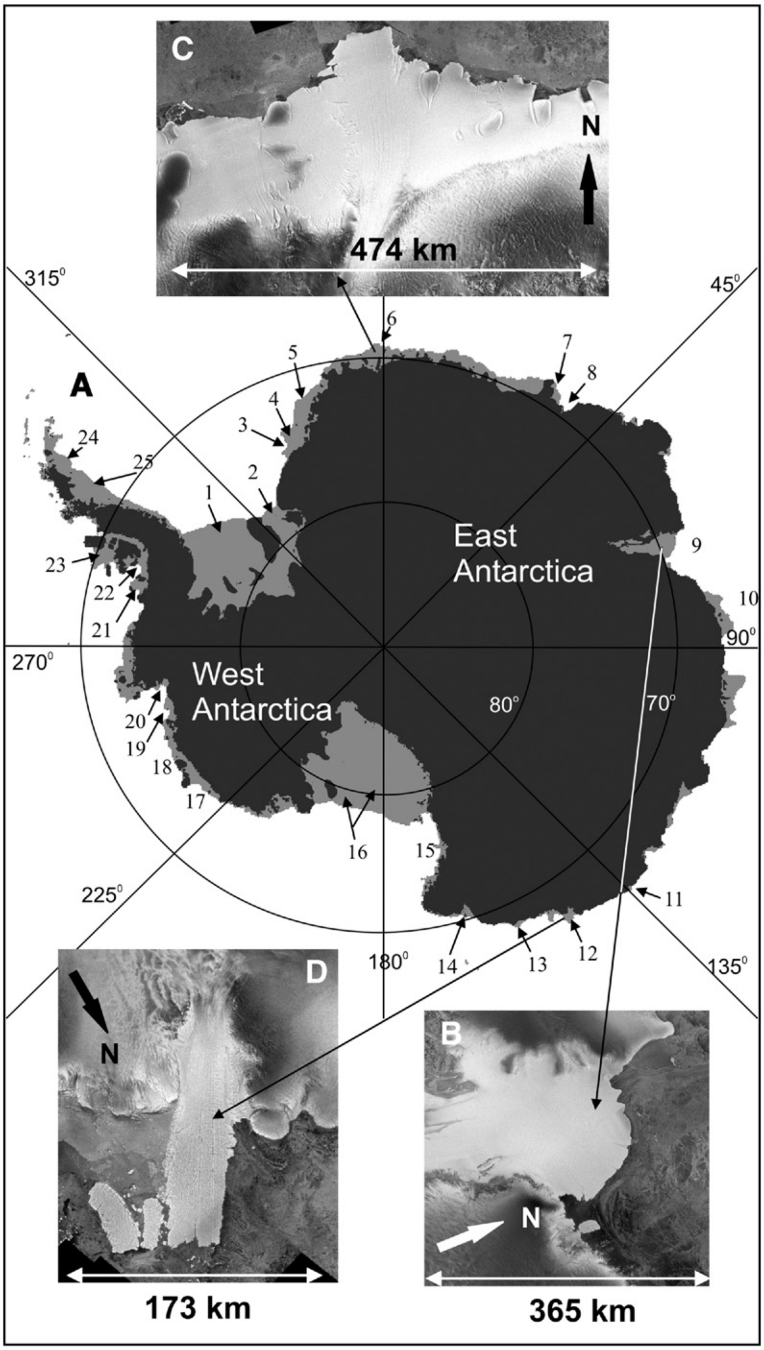

Fig. 1. (A) Location map of Antarctica, with the areas defined as floating ice using satellite radar-altimeter data shown as light grey. The areas of the floating ice-sheet margin where we have derived iceberg keel depths from the thickness of the ice margin are numbered. The length of the floating margin measured in each numbered area is given in Table 1. Radarsat images of different types of floating ice margin are also shown (Jezek, 1999; Jezek and RAMP Product Team, 2002). (B) Amery Ice Shelf. (C) Fimbulisen, showing the ice shelf and protruding fast-flowing outlet glacier. (D) Mertz Glacier. The images are located in (A).

In this paper, we discuss the modern glaciological settings in which large Antarctic tabular icebergs of varying keel depths are produced (Fig. 1). About 55% of the seaward margin of the Antarctic Ice Sheet is floating, with about 14,000 km of the coastline made up of ice shelves and a further 4000 km of outlet glaciers and ice streams with floating tongues (Drewry et al., 1982). The ice sheet, which is the largest in the world at almost 14 million km^2, also exhibits a variety of ice-dynamic settings around its floating margins, including both fast- and slow-flowing portions (Fig. 1). We use satellite radar-altimeter measurements of the surface elevation of the floating marine margins of the Antarctic Ice Sheet in order to calculate the below-water thickness of icebergs (Bamber and Bindschadler, 1997). We then consider the significance of these observations of iceberg–keel depths in the context of sea-floor scouring in the geological record. In particular, we demonstrate that different types of modern floating ice margin produce icebergs of a characteristic keel depth and show how this can be applied to the interpretation of the depth-distribution of ancient scour marks in order to reconstruct the likely palaeo-ice sheet setting.

2. Methods: ice-margin elevations and thickness

The marginal ice-sheet thickness values used here are derived from a digital elevation model (DEM) of the surface of the Antarctic Ice Sheet (Bamber and Bindschadler, 1997). The DEM was produced from over 40×10^6 measurements of ice-surface elevation, which were retrieved from the 336-day geodetic phase of the ERS-1 satellite radar altimeter between April 1994 and March 1995. The vertical accuracy of these surface elevations is better than ± 1 m in areas where slopes are very low, such as floating ice shelves (Bamber and Bentley, 1994). These surface-elevation data were gridded at a 5-km resolution to produce the DEM (Bamber and Bindschadler, 1997) and combined with the OSU-91a geoid (Rapp et al., 1991) to produce orthometric heights relative to mean sea level.

For our work, information on ice-surface elevations was taken from around the floating margins of the ice sheet. Floating ice was identified from a combination of US Geological Survey maps and by estimating the minimum ice thickness necessary for hydrostatic equilibrium based on the observed freeboard and an empirical relationship between ice thickness and freeboard obtained in a previous study (Bamber and Bentley, 1994). If the ice thickness (obtained from the BEDMAP data set; Lythe and Vaughan, 2001) was greater than 1.1 times the value required for flotation, the ice margin was assumed to be grounded. The areas of floating ice are shown in Fig. 1. Surface-elevation values for estimating the calving-front freeboard were obtained from the next grid cell (i.e. 5 km) upstream from the ice edge to reduce possible errors induced by boundary effects at the ice margin itself.

The floating margins of the ice sheet were then subdivided manually into 25 geographical areas, representing major fast-flowing units and ice shelves. Thus, for example, the floating tongues of Pine Island, Thwaites and Mertz glaciers, and the Ronne, Filchner, Amery and Ross ice shelves, are each defined as separate areas (Fig. 1). Taking this approach, with its consideration of varying ice dynamics around the Antarctic margin, allows ice thickness to be estimated as a mean value across each of the 25 segments. Statistics for ice-keel depths are then calculated for each segment of the floating-ice coastline (Table 1).

Table 1
Iceberg keel depths for 25 areas at the floating margins of the Antarctic Ice Sheet

Location (Fig. 1)	$\bar{x}$ (m)	SD (m)	max (m)	min (m)	Length (km)
1. Ronne Ice Shelf	269	72	380	126	550
2. Filchner Ice Shelf	562	38	638	502	185
3. Brunt Ice Shelf	139	10	163	120	95
4. Stancomb-Wills Glacier	265	19	295	235	55
5. Riiser-Larsen Ice Shelf	243	30	325	178	400
6. Fimbul Ice Shelf	274	56	465	188	285
7. Lutzow-Holm Bay Shelf	190	22	220	144	90
8. Shirase Glacier	335	35	387	297	35
9. Amery Ice Shelf	301	34	352	248	150
10. West Ice Shelf	242	48	311	179	70
11. Dibble Glacier	500	84	595	382	30
12. Mertz Glacier	209	56	298	92	115
13. Cook Ice Shelf	386	85	521	245	95
14. Rennick Glacier	154	28	195	116	45
15. Drygalski Ice Tongue	268	38	316	212	25
16. Ross Ice Shelf	255	52	316	113	610
17. Getz Ice Shelf (East)	577	54	678	467	110
18. Getz Ice Shelf (West)	381	26	437	344	140
19. Thwaites Glacier	312	55	432	232	210
20. Pine Island Glacier	496	83	633	414	60
21. Stange Sound Shelf	276	23	310	243	75
22. George VI Ice Shelf (R)	143	8	154	136	20
23. Wilkins Ice Shelf	153	27	185	95	105
24. Larsen A Ice Shelf *	191	16	217	161	105
25. Larsen B Ice Shelf *	286	18	326	240	255

*Now broken up, but in existence at the time of ERS-data acquisition.
R is the Ronne Entrance to the George VI Ice Shelf.
The calculated mean ($\bar{x}$), standard deviation (SD) and maximum (max) and minimum (min) keel depth are shown, together with the total length of coastline in each area.

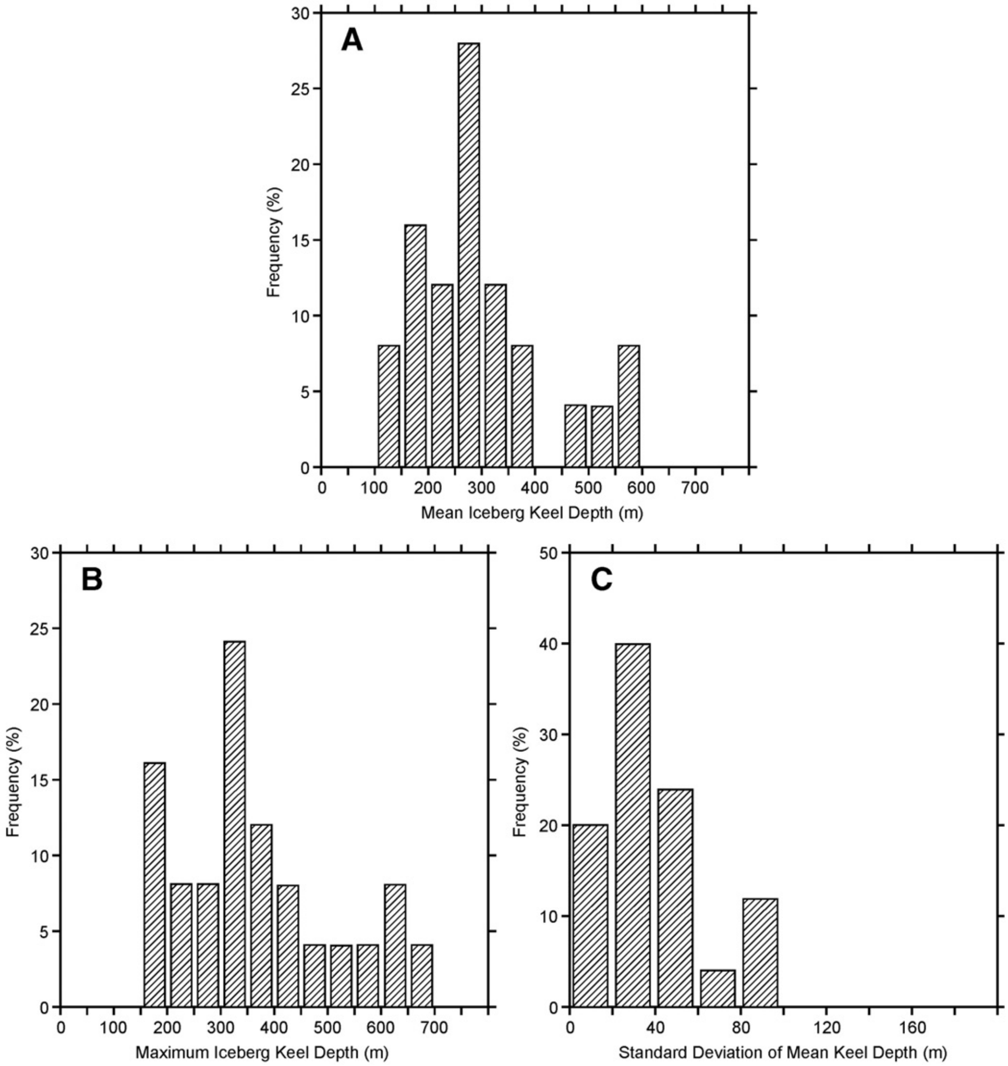

Fig. 2. The frequency distribution of iceberg–keel depths, produced at the floating margins of Antarctica, derived from marginal ice thickness. The number of individual measurements is 787, representing the number of 5 km-wide grid cells on the almost 4000 km-long floating-ice margin for which we have satellite radar-altimeter data. (A) Mean iceberg keel depth is given for each of the 25 areas into which the margin is divided (Fig. 1). (B) Maximum iceberg–keel depth in each area. (C) Standard deviation of iceberg–keel depths about the mean for each area. Note that the graphs do not represent the frequency distribution of icebergs in Antarctic waters, as this requires a knowledge of ice flux from each area and the rate of iceberg melting.

Ice-surface elevation measurements are inverted to produce ice-thickness values (Bamber and Bentley, 1994; Vaughan et al., 1994). Ice thickness (h_i) and iceberg–keel depth (h_k) are derived from the following equations:

$$h_i = 9.615(e - \delta) \quad (1)$$

$$h_k = h_i - e \quad (2)$$

where e is the surface elevation with respect to the geoid and δ is a density-correction factor for the layer of firn at the surface, which has a density less than that of solid ice. The value of 9.615 is based on the ratio between the densities of solid ice and sea water, assuming that water density is invariant around the Antarctic coast. For the Ross Ice Shelf, δ was found to be 15.5 m (Bamber and Bentley, 1994). This value was used for the Ross, Filchner–Ronne and Amery ice shelves. The value of δ depends, however, on the absolute thickness of the firn layer and its relative thickness with respect to the thickness of the underlying solid ice. For the smaller ice shelves and calving tongues, the value of δ will, therefore, be less than 15.5 m. This is especially the case for Antarctic Peninsula ice shelves, which lie near the northern limit of the continent and experience the warmest temperatures.

To overcome this problem, we have derived a simple parameterization of δ, based on latitude and surface elevation. The value of δ is a function of the densification rate of the firn layer which, in turn, is related primarily to surface temperature. Surface temperatures were obtained from ECMWF re-analysis data covering the last 40 years. Another possibility would have been to use passive microwave estimates of the number of melt days (Fahnestock et al., 2002). This would only be suitable, however, for ice shelves on the Antarctic Peninsula, where melting is an important process in the densification of the firn layer. The mean summer (December, January, and

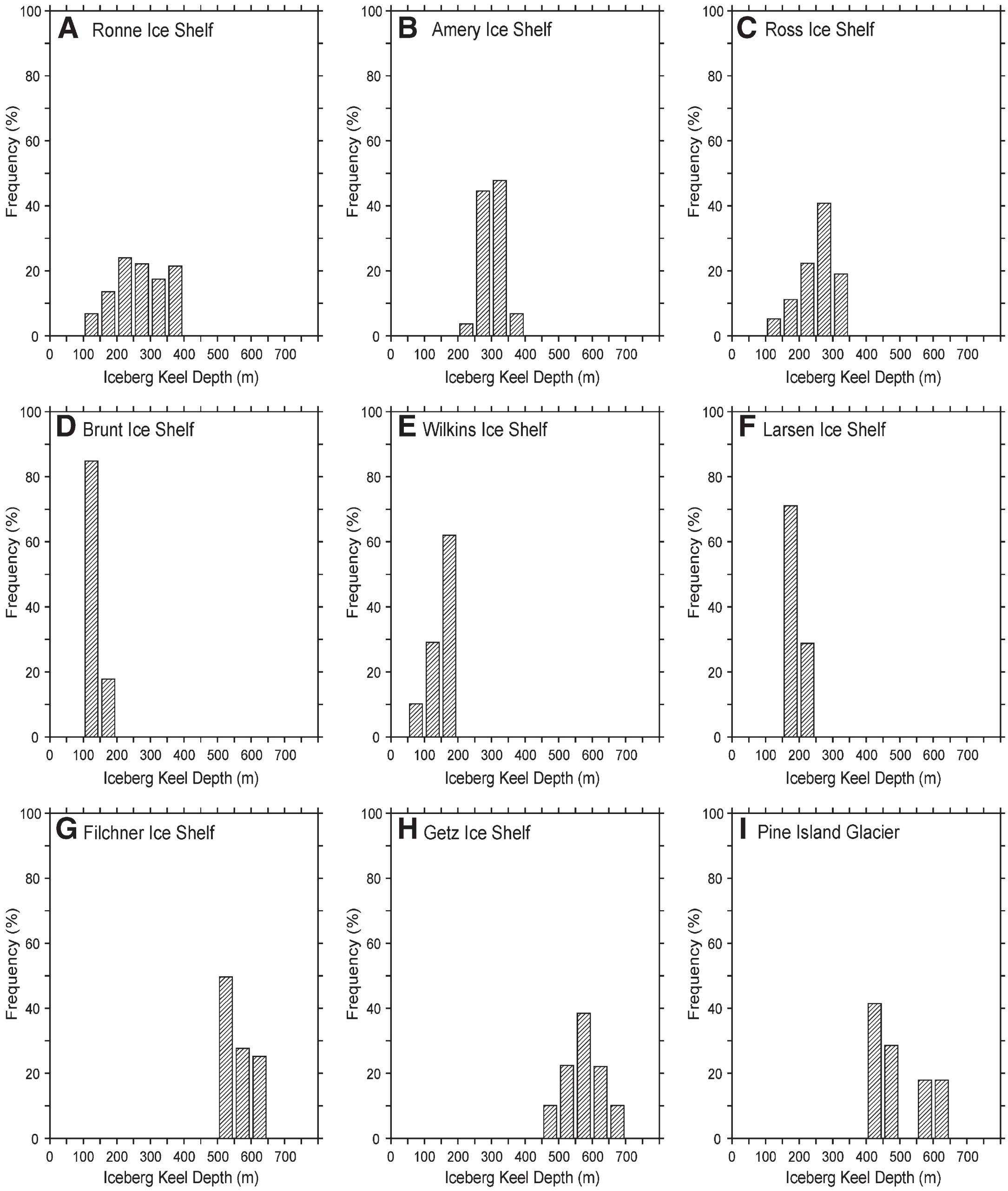

Fig. 3. The frequency distribution of iceberg–keel depths for a number of areas around the margin of Antarctica (located in Fig. 1). (A) Ronne Ice Shelf. (B) Amery Ice Shelf. (C) Ross Ice Shelf. (D) Brunt Ice Shelf. (E) Wilkins Ice Shelf. (F) Larsen A Ice Shelf. (G) Filchner Ice Shelf. (H) Getz Ice Shelf, East. (I) Pine Island Glacier.

February) data were used to parameterize δ, assuming a maximum value of 15.5 m for the Ross Ice Shelf (Bamber and Bentley, 1994) and a minimum of 1 m for the northern, former Larsen A Ice Shelf. The mean summer temperatures for these two regions were 260 and 267 K, respectively, and δ is assumed to vary linearly between these two values. Errors in the values of ice thickness derived using the above methods are about ten times the error in surface elevations, and Vaughan et al. (1994) calculated errors in ice thickness of about 70 m for the Ronne–Filchner Ice Shelf, using a similar approach.

To test our method, we have compared our satellite radar altimeter-derived values of ice thickness (Eq. (1)) with ice-thickness data derived from airborne radio-echo sounding investigations of the floating tongues of Thwaites and Pine Island glaciers in West Antarctica (Lythe and Vaughan, 2001). Agreement between the two sources is good, with differences in ice thickness between the data sources for profiles along the two floating tongues of 15±38 m for Thwaites Glacier and 18±29 m for Pine Island Glacier. In addition, we obtained a thickness of 151 m compared with 170 m measured at a single location on the Brunt Ice Shelf (Doake et al., 2001).

Finally, we assume that the icebergs produced from the Antarctic Ice Sheet are initially the same thickness as the values calculated for the ice-marginal DEM grid cells using Eq. (1). This is a reasonable assumption, since most of the icebergs of kilometre-scale length or greater calved from the Antarctic Ice Sheet are tabular in form (Fig. 1D), and are therefore relatively stable with respect to possible overturning, at least initially after calving (Kristensen et al., 1982).

3. Results: keel depths of Antarctic icebergs

The frequency distributions of the keel depths of icebergs produced from the floating margins of the Antarctic Ice Sheet and for which we have surface-

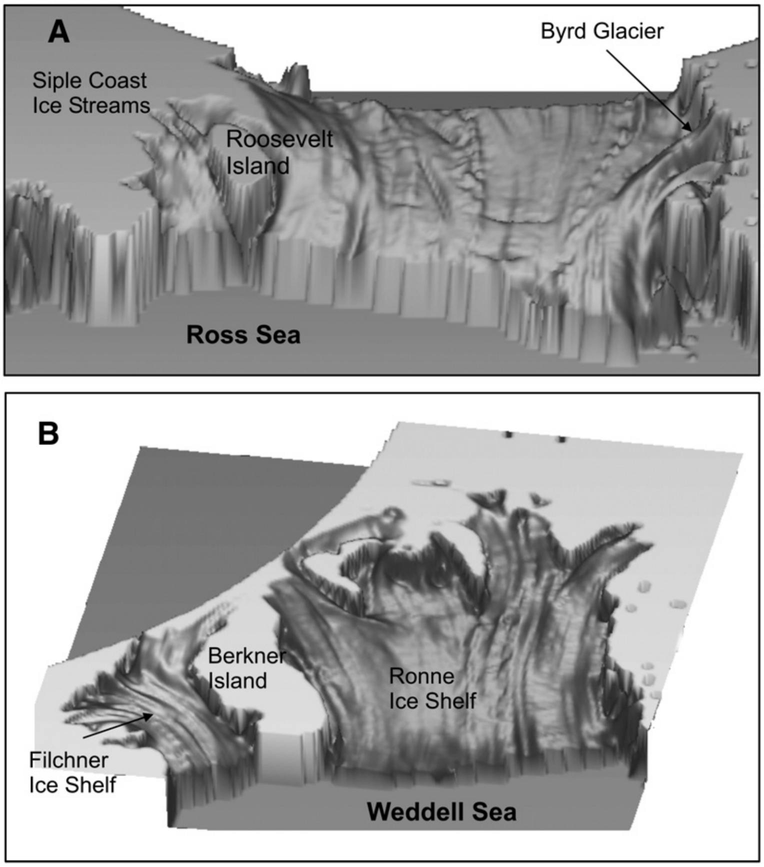

Fig. 4. Shaded isometric plots of the surface elevation of Antarctic ice shelves derived from ERS-1 radar-altimeter data. (A) Ross Ice Shelf. (B) The Ronne and Filchner ice shelves, separated by Berkner Island. The elevation range plotted is from 0–200 m above sea level. The sea-surface is set to 0 m so that the heights of the floating ice margin represent the actual freeboard. Note that each of these large ice shelves has thicker and thinner portions, with the higher-elevation and thicker linear zones usually related to greater mass flux from fast-flowing outlets of large interior drainage basins.

altimetric data of suitable quality (about 20% of that margin) are shown in Fig. 2. These areas of the ice-shelf front are free from contamination by local areas of grounding and are predominantly fed by active ice streams or outlet glaciers. This subset was identified by manual inspection of the thickness data combined with the RAMP mosaic (e.g. Fig. 1B–D) and balance flux estimates (Bamber et al., 2000). It represents about 30% of the grid points initially identified as floating ice fronts.

The mean iceberg keel depth, averaged over a window of between 4 and 122 grid cells (20 and 610 km of coastal length) for 25 segments along the coast, and maximum depth for a single grid cell in each window, are illustrated in Fig. 2A and B. The frequency distribution of standard deviations about the mean depth is also shown (Fig. 2C). The values on which these frequency distributions are based are given in Table 1, and reference numbers to geographical locations on the Antarctic coast are found in Fig. 1.

There are three peaks in the frequency distributions of mean and maximum iceberg–keel depth for the parts of the Antarctic margin that we investigate: at 150–200 m, around 250–300 m, and between 550 and 600 m for mean depth (Fig. 2A); and 150–200 m, 300–350 m and 600–650 m for maximum keel depth (Fig. 2B). The major Antarctic ice shelves, the Ronne, Amery and Ross, produce tabular icebergs that are calculated to have mean keel depths of 269±72 m, 301±34 m and 255±52 m, respectively (Fig. 3A–C). The relatively high standard deviations about the mean keel depth for these large ice shelves is explained largely by the thicker ice present along flowlines derived from fast-flowing outlet glaciers, such as Byrd Glacier, and the active Siple coast ice streams (Fig. 4A).

The relatively large Larsen B (which broke up in 2002), Fimbul and Riiser-Larsen ice shelves also produce icebergs with mean keels of 286±18 m, 274±56 m and 243±30 m (Table 1). The data for Larsen C, however, were too noisy to obtain reliable values. A transect across the coastal margin of Fimbulisen also demonstrates that thicker ice is associated with a fast-flowing outlet glacier, while the less thick and slower-flowing ice shelf produces mean iceberg keels of about 255 m (Fig. 5).

The smaller fringing ice shelves, such as those around parts of the Antarctic Peninsula, appear to have lower ice thicknesses, based on our mapping of ice-shelf front elevations. The former Larsen A Ice Shelf (which broke up in 1999), on the east side of the Antarctic Peninsula, and the Wilkins and George VI ice shelves to the west, produce icebergs with mean keel depths of between about 140 and 190 m (Fig. 3E–F). The Brunt Ice Shelf, fringing the Coats Land coast of East Antarctica, produces bergs with a mean keel depth calculated at 139±10 m (Fig. 3D). The size-frequency distributions for these smaller ice shelves are also rather narrow (Table 1), reflected in standard deviations about the mean keel depth of between about 10 and 30 m (Fig 3D–F).

By contrast, two types of modern glaciological setting appear to produce icebergs with the largest keel depths, of between about 450 and a maximum of about 700 m (Fig. 2A–B). First, several fast-flowing outlet glaciers of the Antarctic Ice Sheet, for example Pine Island and Dibble glaciers, produce icebergs of about 500 m in mean keel depth (Fig. 3I). Radio-echo measurements of ice thickness on Pine Island Glacier confirm that it is about 500 m thick near its seaward margin (Crabtree and Doake, 1982; Corr et al., 2001). Second, the Filchner Ice Shelf appears to calve icebergs that are significantly

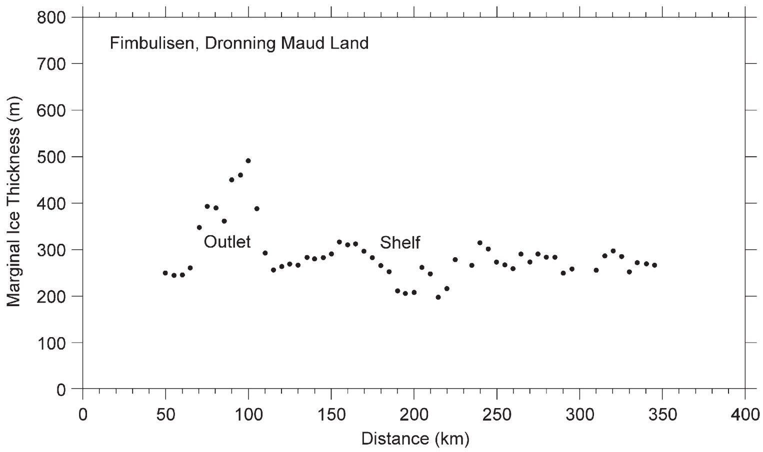

Fig. 5. A profile of radar-altimeter derived ice thicknesses along the margin of Fimbulisen in Dronning Maud Land, East Antarctica (Fig. 1A,C). The fast-flowing floating tongue of the outlet glacier is demonstrated to be thicker than the surrounding ice shelf.

thicker than those from either the other major ice shelves filling embayments in the Antarctic continent, or the smaller fringing ice shelves (Fig. 3G). A mean iceberg keel depth of 562 m is calculated for the Filchner Ice Shelf, with a standard deviation along the ice front of 38 m. This marked difference between the thickness of ice on the Filchner and Ronne ice shelves, east and west of Berkner Island, respectively, has been reported previously by Vaughan et al. (1994). It is illustrated by shaded isometric plots of the surface elevation of each of these ice shelves in Fig. 4B.

4. Discussion

4.1. Controls on iceberg–keel depth

Important glaciological controls on iceberg–keel depth are the thickness of ice at the grounding line, where ice becomes decoupled from its bed to form floating ice shelves, and the distance from the grounding line to the calving outer margin of the ice shelf. The initial thickness of an ice shelf is set by the thickness of the ice at the grounding line. Ice shelves will then spread out and thin under their own weight, by a process known as creep thinning (Robin, 1979; Sanderson, 1979). Thus, ice shelves will be relatively thick where deep ice is present at the grounding lines and/or their floating length is short. Where an ice shelf is relatively unconstrained laterally, the ice spreads out and thins more rapidly by creep and, especially if this is combined with relatively thin ice at the grounding line, then the ice shelf will be relatively thin.

A further, oceanographic control on ice-shelf thickness is the nature and rate of melting and freezing at the floating ice-shelf base (e.g. Robin, 1979; Jenkins and Doake, 1991; Rignot, 1998, 2002; Joughin and Padman, 2003). Behaviour at this interface will depend on the ocean-water temperature and circulation pattern, and on the gradient of the lower surface of the ice shelf that influences the pressure-dependent value of the melting point. In general, basal melting of several metres per year is likely close to the grounding line, because even water at the sea-surface freezing temperature contains sufficient sensible heat to melt glacier ice here as a result of depression of the freezing point with increasing pressure (Jacobs, 1989). The cold meltwater layer produced will tend to reduce further melting as this less dense water flows up the seaward-sloping ice-shelf base, and the freezing temperature–pressure relationship will also encourage freezing rather than melting at shallower sub-ice shelf locations (Foldvik and Kvinge, 1974; Jenkins, 1991). Renewed melting is likely to take place close to the ice-shelf front (Robin, 1979; Joughin and Padman, 2003).

Making this argument geographically specific, the large Ross, Ronne and Amery ice shelves, for example, are fed by drainage basins of 10^6 km^2 in area (Vaughan et al., 1999; Bamber et al., 2000) and ice that is often over 1000 m thick at the grounding line (e.g. Jenkins and Doake, 1991). However, flowlines from the grounding line to the Ross Sea, Weddell Sea and Prydz Bay coasts are several hundred kilometres long, providing ample time for basal melting and creep thinning to take place in the wide embayments in which these ice shelves are found. Icebergs calved from these three ice shelves are, on average, 250–300 m thick (Table 1). The Filchner Ice Shelf is an exception to this pattern, and is significantly thicker, at about 560 m, than the adjacent Ronne Ice Shelf because the large flux of ice, derived from a huge inland drainage basin (1,670,000 km^2) flowing mainly through Support Force Glacier (Vaughan et al., 1999), is constrained from creep thinning by Berkner Island to the west and the mountains of Coats Land to the east (Fig. 4B).

By contrast, some of the fast-flowing outlet glaciers of Antarctica are also over 1000 m thick at the grounding line (McIntyre, 1985; Rignot, 1998, 2002) and are fed from large interior drainage basins (10^5 km^2). However, they have flowlines from the grounding line to the floating-ice margin of tens rather than hundreds of kilometres. Dibble and Pine Island glaciers are examples. These outlets tend to calve thicker icebergs from about 450–650 m in keel depth. In the northern hemisphere, the terminus of Daugaard–Jensen Gletscher, a fast-flowing outlet glacier in East Greenland, floats for only 10–20 km beyond the grounding line and has also been observed to calve large numbers of icebergs with keel depths of 500–600 m (Dowdeswell et al., 1992).

Finally, the relatively thin fringing ice shelves tend to be derived from smaller ice-sheet drainage basins (10^3 to 10^4 km^2), and airborne radar investigations show that ice is often only a few hundred metres thick at the grounding line (e.g. Doake et al., 1998). Furthermore, some of those ice shelves on the western, Amundsen and Bellingshausen Sea margin of the Antarctic are affected by relatively warmer water than the bulk of the Antarctic coast, and thinning is also accelerated by basal melting (Potter and Paren, 1985; Jacobs et al., 1996; Jenkins et al., 1997). Icebergs less than about 200 m in thickness are, therefore, typically calved from the margins of the smaller fringing ice shelves (Fig. 3D–F).

4.2. Implications for sea-floor scouring by iceberg keels

Scouring or ploughing of the sea floor by iceberg keels is a common process on high-latitude continental shelves. It takes place where keels reach through the full depth of the water column to impinge on the sediments below (e.g. Barnes and Lien, 1988; Woodworth-Lynas et al., 1991; Dowdeswell et al., 1993; O'Brien et al., 1997; Anderson, 1999; Syvitski et al., 2001). The morphological features resulting from this process are known as sea-floor scours, plough-marks, furrows or gouges (Fig. 6). Individual scours produced by large icebergs are often kilometres in length, tens of metres wide and metres deep (e.g. Woodworth-Lynas et al., 1985). Sets of scours generally have an irregular pattern on the sea floor, reflecting the wandering drift tracks of individual icebergs. In addition, sediments beneath the sea floor can be deformed and reworked by the stresses induced by the iceberg keels (Woodworth-Lynas and Guigné, 1990). Iceberg scours have been observed using side-scan sonar, swath bathymetry and penetration echo-sounding systems on many high-latitude continental shelves. The process of ice-keel scouring reworks shelf stratigraphy and, therefore, the sedimentary record of past environmental change may be turbated and largely obliterated to the maximum depth of iceberg–keel grounding (e.g. Vorren et al., 1989; Dowdeswell et al., 1994) (Fig. 6).

Numerous examples exist of iceberg scours in water depths up to about 450 m. However, in deeper waters, reports of iceberg scouring become less common. Scours are found, for instance, at water depths of at least 500 m on Mertz Bank, off Wilkes Land in East Antarctica (Barnes and Lien, 1988), to 650 m in the Ross and Bellingshausen seas (Anderson, 1999; Ó Cofaigh et al., 2005; Mosola and Anderson, 2006) and the Scoresby Sund fjord system and adjacent shelf of East Greenland (Dowdeswell et al., 1993), and to 700 m in the Southwest Approaches to Britain (Belderson et al., 1973). At the extreme limit, ancient scour marks have been described, for example, from water depths of 450 to 850 m on the Yermak Plateau, north of Svalbard (Vogt et al., 1994; Crane et al., 1997). Antarctic continental shelves tend to deepen inshore, due to long-term deep erosion and isostatic loading by the ice sheet (Anderson, 1999). However, iceberg scours are not generally observed below about 500–600 m even in inner-shelf settings where water may reach over 1000 m in depth (e.g. Anderson, 1999; Wellner et al., 2001; Ó Cofaigh et al., 2002, 2005; Evans et al., 2006).

Although the marine geological and geophysical evidence for icebergs with particularly deep keels has sometimes been interpreted to indicate the presence of extensive former ice shelves (Vogt et al., 1994), our data on the keel dimensions of Antarctic icebergs suggest that relatively deep-keeled bergs are produced in several settings (Table 1; Fig. 3). The first source is from major

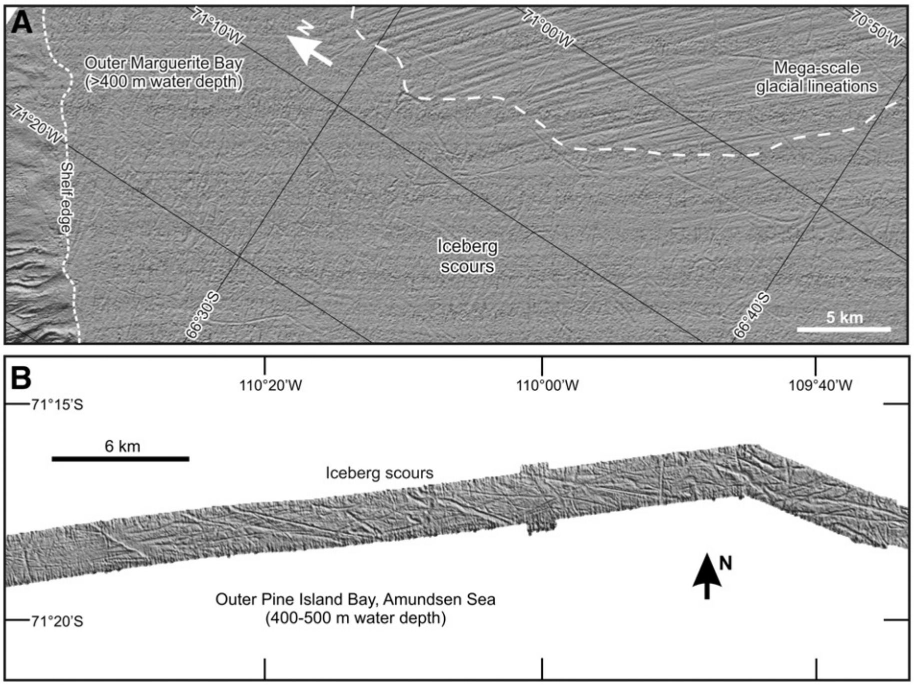

Fig. 6. Scour marks on the sea floor produced by the action of deep iceberg keels. (A) Outer part of Marguerite Bay, Antarctic Peninsula. (B) Amundsen Sea shelf offshore of Thwaites Glacier in Pine Island Bay. Both datasets were collected from the RRS *James Clark Ross* using a Kongsberg Simrad EM-120 swath bathymetry system.

ice-sheet outlet glaciers (Fig. 3I), where deep-water grounding lines, relatively short floating tongues and high mass-flux act to produce icebergs with the deepest keels we have observed (cf. Dowdeswell, 1995). This is the case not only for our dataset of modern Antarctic icebergs (Figs. 2 and 3; Table 1), but also for detailed observations on the freeboard and keel depth of icebergs derived from a modern fast-flowing outlet glacier in East Greenland (Dowdeswell et al., 1992).

A second ice-dynamic setting in which relatively deep-keeled icebergs are produced is where ice flow from large interior drainage basins is constrained by a mechanical barrier. An example is the way in which Berkner Island and the mountains of Coats Land restrict the horizontal spreading of ice flowing into the Filchner Ice Shelf, Antarctica (Fig. 4B). This produces a relatively thick ice shelf, and icebergs in excess 500 m in thickness are calved from it (Fig. 3G).

Many high-latitude and some mid-latitude continental shelves were inundated by ice extending to the shelf edge under full-glacial conditions (e.g. Anderson et al., 2002; Svendsen et al., 2004), and scouring of these shelves by iceberg keels is most likely a deglacial and interglacial phenomenon. An additional setting for the production of large numbers of relatively deep-keeled tabular icebergs is therefore during major deglaciations following full-glacial maxima. Changes in water depth during ice-sheet deglaciation of high- and mid-latitude continental margins are, however, often much more complex than the simple adjustment from a full-glacial sea level about 120 m lower than today (Andrews, 1970; Fairbanks, 1989). Thus, in addition to glaciological controls on iceberg–keel dimensions, factors affecting the maximum depth of ice-keel scouring will include not only global eustatic sea-level rise but also local sea-level adjustments to regional isostatic and tectonic effects (e.g. Fleming et al., 1998; Lambeck and Purcell, 2001; Lambeck and Chappell, 2001).

In addition, the collapse of a major ice-sheet drainage basin, not necessarily linked to global deglaciation, would also yield large numbers of icebergs, as indicated by the North-Atlantic ‘Heinrich’ layers (Bond et al., 1992; MacAyeal, 1993; Dowdeswell et al., 1995). If calved from outer-shelf settings, these icebergs would be likely to have keels of less than about 500–600 m, since water depths close to the shelf edge seldom exceed these values.

In general, most modern Antarctic ice shelves produce few icebergs thicker than about 350 m (Fig. 3), because creep thinning and strong basal melting tend to reduce ice-shelf thickness with distance from the grounding line. We conclude, therefore, that iceberg scours found at palaeo-water depths in excess of 450–500 m in the geological record are an indicator of either: (a) the presence of fast-flowing ice-sheet outlet glaciers in palaeo-ice sheets; (b) a high-flux and mechanically-constrained ice shelf setting where creep thinning is relatively ineffective; or (c) a pulse of production of deep-keeled icebergs during major continental-shelf deglaciation or the collapse of a large ice-sheet drainage basin.

5. Conclusions

- The dimensions of Antarctic icebergs are controlled by the dynamics of the parent ice sheet and the interactions between the ice margin and marine waters. About 18,000 km of the Antarctic coast is made up of floating ice (Drewry et al., 1982). We have calculated marginal ice-thickness for 25 areas of Antarctica by inverting a large data set of satellite radar-altimeter measurements of ice-surface elevation (Fig. 1).
- The frequency distribution of marginal ice thickness and, hence, of mean Antarctic iceberg–keel depths, has three peaks: at 150–200 m, around 250–300 m, and 550 and 600 m (Fig. 2A). The smaller ice shelves (10^3 km^2) fringing, for example, the Antarctic Peninsula, usually calve relatively thin icebergs and account for the peak at 150–200 m in ice-keel depth (Fig. 2A). The major Antarctic ice shelves (10^5 km^2), the Ronne, Amery and Ross, along with the relatively large (10^4 km^2) former Larsen B, Fimbul and Riiser-Larsen ice shelves, produce tabular icebergs centred on the 250–300 m range (Fig. 3). A third peak, relating to the thickest icebergs, is produced by some fast-flowing outlet glaciers along with the topographically-constrained Filchner Ice Shelf.
- Especially where fast-flowing outlet glaciers have flowlines from the grounding line to the floating-ice margin of tens rather than hundreds of kilometres, they tend to calve icebergs from about 450–650 m in keel depth. In addition, very few modern icebergs, from either Antarctica or Greenland, have keel depths of more than about 650 m (Figs. 2 and 3; Dowdeswell et al., 1992).
- Most modern Antarctic ice shelves do not produce large numbers of icebergs thicker than about 350 m (Fig. 3), because creep thinning and strong basal melting reduce ice-shelf thickness with distance from the grounding line. This implies that sea-floor scours found in deeper water are unlikely to be an indicator of extensive ice-shelf development.
- Scours at depths in excess of 450–500 m in the geological record suggest either: (a) the presence of fast-flowing ice-sheet outlet glaciers in palaeo-ice sheets; (b) a high-flux and mechanically-constrained

ice shelf setting where creep thinning is relatively ineffective; or (c) the calving of large icebergs during major deglaciation or the collapse of individual ice-sheet drainage basins.

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References

Anderson, J.B., 1999. Antarctic Marine Geology. Cambridge University Press. 289 pp.

Anderson, J.B., Shipp, S.S., Lowe, A.L., Wellner, J.S., Mosola, A.B., 2002. The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history: a review. Quat. Sci. Rev. 21, 49–70.

Andrews, J.T., 1970. A Geomorphological Study of Post-glacial Uplift with Particular Reference to Arctic Canada, vol. 2. Institute of British Geographers. Special Publication.

Bamber, J.L., Bentley, C.R., 1994. A comparison of satellite-altimetry and ice-thickness measurements of the Ross Ice Shelf, Antarctica. Ann. Glaciol. 20, 357–364.

Bamber, J.L., Bindschadler, R.A., 1997. An improved elevation data set for climate and ice sheet modelling: validation with satellite imagery. Ann. Glaciol. 25, 439–444.

Bamber, J.L., Vaughan, D.G., Joughin, I., 2000. Widespread complex flow in the interior of the Antarctic Ice Sheet. Science 287, 1248–1250.

Barnes, P.W., Lien, R., 1988. Icebergs rework shelf sediments to 500 m off Antarctica. Geology 16, 1130–1133.

Belderson, R.H., Kenyon, N.H., Wilson, J.B., 1973. Iceberg plough marks in the northeast Atlantic. Palaeogeogr. Palaeoclimatol. Palaeoecol. 13, 215–224.

Bond, G., Heinrich, H., Broecker, W., Labeyrie, L., McManus, J., Andrews, J., Huon, S., Jantschik, R., Clasen, S., Simet, C., Tedesco, K., Klas, M., Bonani, G., Ivy, S., 1992. Evidence for massive discharges of icebergs into the North Atlantic ocean during the last glacial period. Nature 360, 245–249.

Corr, H.F.J., Doake, C.S.M., Jankins, A., Vaughan, D.G., 2001. Investigations of an ‘ice plain’ in the mouth of Pine Island Glacier, Antarctica. J. Glaciol. 47, 51–57.

Crabtree, R.D., Doake, C.S.M., 1982. Pine Island Glacier and its drainage basin. Ann. Glaciol. 3, 65–70.

Crane, K., Vogt, P.R., Sundvor, E., 1997. Deep Pleistocene iceberg ploughmarks on the Yermak Plateau. In: Davies, T.A., Bell, T., Cooper, A.K., Josenhans, H., Polyak, L., Solheim, A., Stoker, M.S., Stravers, J.A. (Eds.), Glaciated Continental Margins: An Atlas of Acoustic Images, pp. 140–141.

Doake, C.S.M., Corr, H.F.J., Rott, H., Skvarca, P., Young, N.W., 1998. Breakup and conditions for stability of the northern Larsen Ice Shelf, Antarctica. Nature 391, 778–780.

Doake, C.S.M., Corr, H.F.J., Jenkins, A., 2001. Polarization of radio waves transmitted through Antarctic ice shelves. Ann. Glaciol. 34, 165–170.

Dowdeswell, J.A., 1995. Comment on: “Deep Pleistocene iceberg plowmarks on the Yermak Plateau: sidescan and 3.5 kHz evidence for thick calving ice fronts and a possible marine ice sheet in the Arctic Ocean”. Geology 23, 476–477.

Dowdeswell, J.A., Whittington, R.J., Hodgkins, R., 1992. The sizes, frequencies and freeboards of East Greenland icebergs observed using ship radar and sextant. J. Geophys. Res. 97, 3515–3528.

Dowdeswell, J.A., Villinger, H., Whittington, R.J., Marienfeld, P., 1993. Iceberg scouring in Scoresby Sund and on the East Greenland continental shelf. Mar. Geol. 111, 37–53.

Dowdeswell, J.A., Whittington, R.J., Marienfeld, P., 1994. The origin of massive diamicton facies by iceberg rafting and scouring, Scoresby Sund, East Greenland. Sedimentology 41, 21–35.

Dowdeswell, J.A., Maslin, M.A., Andrews, J.T., McCave, I.N., 1995. Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North Atlantic sediments. Geology 23, 301–304.

Drewry, D.J., Jordan, S.R., Jankowski, E., 1982. Measured properties of the Antarctic Ice Sheet: surface configuration, ice thickness, volume and bedrock characteristics. Ann. Glaciol. 3, 83–91.

Evans, J., Dowdeswell, J.A., Ó Cofaigh, C., Benham, T.J., Anderson, J.B., 2006. Extent and dynamics of the West Antarctic Ice Sheet on the outer continental shelf of Pine Island Bay, Amundsen Sea, during the last deglaciation. Mar. Geol. 230, 53–72.

Fahnestock, M., Abdalati, W., Shuman, C., 2002. Long melt seasons on ice shelves of the Antarctic Peninsula: an analysis using satellite-based microwave emission measurements. Ann. Glaciol. 34, 127–133.

Fairbanks, R., 1989. A 17,000 year glqacio-eustatic sea level record: influence of glacial melting rate on the Younger Dryas Event and deep ocean circulation. Nature 342, 637–642.

Fleming, K., Johnson, P., Zwartz, D., Yokoyama, Y., Lambeck, K., Chappell, J., 1998. Refining the eustatic sea-level curve since the Last Glacial Maximum using far and intermediate field sites. Earth Planet. Sci. Lett. 163, 327–342.

Foldvik, A., Kvinge, T., 1974. Conditional stability of sea water at the freezing point. Deep-Sea Res. 21, 169–174.

Jacobs, S.S., 1989. Marine controls on modern sedimentation on the Antarctic continental shelf. Mar. Geol. 85, 121–153.

Jacobs, S.S., Helmer, H.H., Doake, C.S.M., Jenkins, A., Frolich, R.M., 1992. Melting of ice shelves and the mass balance of Antarctica. J. Glaciol. 38, 375–387.

Jacobs, S.S., Hellmer, H.H., Jenkins, A., 1996. Antarctic ice sheet melting in the southeast Pacific. Geophys. Res. Lett. 23, 957–960.

Jenkins, A., 1991. A one-dimensional model of ice shelf–ocean interaction. J. Geophys. Res. 96, 20671–20677.

Jenkins, A., Doake, C.S.M., 1991. Ice–ocean interaction on the Ronne Ice Shelf. J. Geophys. Res. 96, 791–813.

Jenkins, A., Vaughan, D.G., Jacobs, S.S., Hellmer, H.H., Keys, J.R., 1997. Glaciological and oceanographic evidence of high melt rates beneath Pine Island Glacier, West Antarctica. J. Glaciol. 43, 114–121.

Jezek, K.C., 1999. Glaciological properties of the Antarctic ice sheet from RADARSAT-1 synthetic aperture radar imagery. Ann. Glaciol. 29, 286–294.

Jezek, K., RAMP Product Team., 2002. RAMP AMM-1 SAR Image Mosaic of Antarctica. Fairbanks, AK: Alaska Satellite Facility, in Association with the National Snow and Ice Data Center, Boulder, Colorado. Digital media.

Joughin, I., Padman, L., 2003. Melting and freezing beneath Filchner-Ronne Ice Shelf, Antarctica. Geophys. Res. Lett. 30, 1477. doi:10.1029/2003GL016941.

Kristensen, M., Squire, V.A., Moore, S.C., 1982. Tabular icebergs in ocean waves. Nature 297, 669–671.

Lambeck, K., Chappell, J., 2001. Sea level change through the last glacial cycle. Science 292, 679–686.

Lambeck, K., Purcell, A.P., 2001. Sea-level change in the Irish Sea since the Last Glacial Maximum: constraints from isostatic modelling. J. Quat. Sci. 16, 497–506.

Long, D.G., Ballantyne, J., Bertoia, C., 2002. Is the number of Antarctic icebergs really increasing? EOS, Trans. Am. Geophys. Union 83 (471), 474.

Lythe, M.B., Vaughan, D.G., 2001. BEDMAP: a new ice thickness and subglacial topographic model of Antarctica. J. Geophys. Res. 106, 11335–11351.

MacAyeal, D.R., 1993. Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic's Heinrich events. Paleoceanography 8, 775–784.

McIntyre, N.F., 1985. The dynamics of ice-sheet outlets. J. Glaciol. 31, 99–107.

Morgan, V.I., Budd, W.F., 1978. The distribution, movement and melt rates of Antarctic icebergs. In: Husseiny, A.A. (Ed.), Proceedings of the First Conference on Iceberg Utilization for Freshwater Production. Iowa State University, pp. 220–228.

Mosola, A.B., Anderson, J.B., 2006. Expansion and rapid retreat of the West Antarctic Ice Sheet in eastern Ross Sea: possible consequences of over-extended ice streams? Quat. Sci. Rev. 25, 2177–2196.

Ó Cofaigh, C., Pudsey, C.J., Dowdeswell, J.A., Morris, P., 2002. Evolution of subglacial bedforms along a paleo-ice stream, Antarctic Peninsula continental shelf. Geophys. Res. Lett. 29, GL014488. doi:10.1029/2001.

Ó Cofaigh, C., Larter, R.D., Dowdeswell, J.A., Hillenbrand, C.D., Pudsey, C.J., Evans, J., Morris, P., 2005. Flow of the West Antarctic Ice Sheet on the continental margin of the Bellingshausen Sea at the Last Glacial Maximum. J. Geophys. Res. 110, B11103. doi:10.1029/2005JB003619.

O'Brien, P.E., Leitchenkov, G., Harris, P.T., 1997. Iceberg plough marks, subglacial bedforms and grounding zone moraines in Prydz Bay, Antarctica. In: Davies, T.A., Bell, T., Cooper, A.K., Josenhans, H., Polyak, L., Solheim, A., Stoker, M.S., Stravers, J.A. (Eds.), Glaciated Continental Margins: An Atlas of Acoustic Images, pp. 228–231.

Orheim, O., 1985. Iceberg discharge and the mass balance of Antarctica. Glaciers, Ice Sheets and Sea Level: Effect of a CO_2-induced Climatic Change: Report DOE/EV/60235-1, 210-215. U.S. Dept. of Energy, Washinghton, DC.

Potter, J.R., Paren, J.G., 1985. Interaction between ice shelf and ocean in George VI Sound, Antarctica. In: Jacobs, S.S. (Ed.), Oceanology of the Antarctic Continental Shelf. American Geophysical Union, Antarctic Research Series, vol. 43, pp. 35–58.

Rapp, R.H., Wang, H., Pavlis, N.K., 1991. The Ohio State 1991 Geopotential and Sea Surface Topography Harmonic Coefficient Models. Report, vol. 410. Department of Geodetic Science and Surveying, Ohio State University.

Rignot, E.J., 1998. Fast recession of a West Antarctic glacier. Science 281, 549–551.

Rignot, E.J., 2002. Mass balance of East Antarctic glaciers and ice shelves from satellite data. Ann. Glaciol. 34, 217–228.

Robin, G. de Q., 1979. Formation, flow and disintegration of ice shelves. J. Glaciol. 24, 259–272.

Sanderson, T.J.O., 1979. Equilibrium profile of ice shelves. J. Glaciol. 22, 435–460.

Svendsen, J.I., Alexanderson, H., Astakhov, V.I., Demidov, I., Dowdeswell, J.A., Funder, S., Gataullin, V., Henriksen, M., Hjort, C., Houmark-Nielsen, M., Hubberten, H.W., Ingólfsson, O., Jakobsson, M., Kjær, K.H., Larsen, E., Lokrantz, H., Lunkka, J.P., Lyså, A., Mangerud, J., Matioushkov, A., Murray, A., Möller, P., Niessen, F., Nikolskaya, O., Polyak, L., Saarnisto, M., Siegert, C., Siegert, M.J., Spielhagen, R.F., Stein, R., 2004. Late Quaternary ice-sheet history of Northern Eurasia. Quat. Sci. Rev. 23, 1229–1271.

Syvitski, J.P.M., Lewis, C.F.M., Piper, D.J.W., 1996. Paleoceanographic information derived from acoustic surveys of glaciated continental margins: examples from eastern Canada. In: Andrews, J.T., Austin, W., Bergsten, H., Jennings, A.E. (Eds.), Late Quaternary Paleoceanography of the North Altantic Margins. Geological Society, London, Special Publication, vol. 111, pp. 51–76.

Syvitski, J.P.M., Stein, A.B., Andrews, J.T., Milliman, J.D., 2001. Icebergs and the sea floor of the East Greenland (Kangerlussuaq) continental margin. Arct. Antarct. Alp. Res. 33, 52–61.

Vaughan, D.G., Sievers, J., Doake, C.S.M., Hinze, H., Mantripp, D.R., Pozdeev, V.S., Sandhager, H., Schenke, H.W., Solheim, A., Thyssen, F., 1994. Subglacial and seabed topography, ice thickness and water column thickness in the vicinity of Filchner–Ronne–Schelfeis, Antarctica. Polarforschung 64, 75–88.

Vaughan, D.G., Bamber, J.L., Giovinetto, M., Russell, J., Cooper, A.P.R., 1999. Reassessment of net surface mass balance in Antarctica. J. Climate 12, 933–946.

Vogt, P.R., Crane, K., Sundvor, E., 1994. Deep Pleistocene iceberg plowmarks on the Yermak Plateau: sidescan and 3.5 kHz evidence for thick calving ice fronts and a possible marine ice sheet in the Arctic Ocean. Geology 22, 403–406.

Vorren, T.O., Lebesbye, E., Andreassen, K., Larsen, K.-B., 1989. Glacigenic sediments on a passive continental margin as exemplified by the Barents Sea. Mar. Geol. 85, 251–272.

Wadhams, P., 1988. Winter observations of iceberg frequencies and sizes in the South Atlantic Ocean. J. Geophys. Res. 93, 3583–3590.

Wellner, J.S., Lowe, A.L., Shipp, S.S., Anderson, J.B., 2001. Distribution of glacial geomorphic features on the Antarctic continental shelf and correlation with substrate: implications for ice behavior. J. Glaciol. 47, 397–411.

Woodworth-Lynas, C.M.T., Guigné, J.Y., 1990. Iceberg scours in the geological record: examples from glacial Lake Agassiz. In: Dowdeswell, J.A., Scourse, J.D. (Eds.), Glacimarine Environments: Processes and Sediments. Geological Society, London, Special Publication, vol. 53, pp. 217–223.

Woodworth-Lynas, C.M.T., Simms, A., Rendell, C.M., 1985. Iceberg grounding and scouring on the Labrador Continental Shelf. Cold Reg. Sci. Technol. 10, 163–186.

Woodworth-Lynas, C.M.T., Josenhans, H.W., Barrie, J.V., Lewis, C.F.M., Parrott, D.R., 1991. The physical processes of seabed disturbance during iceberg grounding and scouring. Cont. Shelf Res. 11, 939–951.