First survey of Antarctic sub–ice shelf sediments reveals mid-Holocene ice shelf retreat

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ABSTRACT

The retreat of five small Antarctic Peninsula ice shelves in the late 20th century has been related to regional (possibly anthropogenic) climate warming. We use the record of ice-rafted debris (IRD) in cores to show that the Prince Gustav Channel ice shelf also retreated in mid-Holocene time. Early and late Holocene-age sediments contain IRD derived entirely from local ice drainage basins, which fed the section of ice shelf covering each site. Core-top and mid-Holocene (5–2 ka) sediments include a wider variety of rock types, recording the drift of far-traveled icebergs, which implies seasonally open water at the sites. The period when the Prince Gustav ice shelf was absent corresponds to regional climate warming deduced from other paleoenvironmental records. We infer that the recent decay cannot be viewed as an unequivocal indicator of anthropogenic climate perturbation.

Keywords: Antarctic, Holocene, ice-rafted debris, ice shelves.

INTRODUCTION

Small ice shelves east of the northern Antarctic Peninsula (Prince Gustav Channel, Larsen Inlet, Larsen-A; Fig. 1) have been retreating during the period of historical observations (since 1843; Cooper, 1997). The rate of retreat increased in the late 1980s (Skvarca et al., 1999), and in 1995 a major breakout of Larsen-A and the Prince Gustav Ice Shelf occurred (Rott et al., 1996). During this period, ice shelves west of the peninsula were also disintegrating (Vaughan and Doake, 1996). The retreat appears to have been caused by regional climate warming over recent decades (King, 1994), but it is not known if this reflects global climate change. Whether the retreat was an event unique in the present interglacial period, or whether advance and retreat of these small ice shelves has occurred repeatedly in response to natural Holocene climate change, may have a bearing on the climate sensitivity of the Antarctic Peninsula ice sheet. A variety of evidence from fjord, lacustrine, and terrestrial glacial sediments from the same region suggests that the mid-Holocene may have been a period of relative warmth (Ingolfsson et al., 1992, 1998; Björck et al., 1996; Domack and McClennen, 1996; Shevenell et al., 1996; Jones et al., 2000). We investigated sediments deposited beneath the former ice shelves for evidence of subglacial or open marine deposition.

British Antarctic Survey cruise JR48 (February–March 2000) was the first scientific survey of the area formerly occupied by the Larsen-A and Prince Gustav ice shelves, though bathymetric data were available for part of the Larsen-A area (del Valle et al., 1998). Bathymetric and 3.5 kHz acoustic profiler data showed Prince Gustav Channel to be 600–800 m deep, with an acoustic drape of sediment several meters thick overlying a smooth to hummocky surface. In cores, the drape corresponds to marine and glacial-marine facies (mainly silty clay and clayey mud with 15%–25% diatoms; Fig. 2). The underlying acoustically opaque sediment includes diamicts deposited near an ice-shelf grounding line. We also cored on the continental shelf beyond the historical ice margin (Fig. 1B). By comparison with the shelf west of the Antarctic Peninsula (Pudsey et al., 1994), the drape was initially thought to represent deposition for much of the Holocene, the diamicts dating from the last glacial maximum and the ensuing deglaciation.

The nature of deposition beneath floating ice shelves is poorly known because of the lack of opportunities for observation, though sedimentation models have been proposed on the basis of the facies succession in cores (e.g., Domack et al., 1998). The absence of biogenic material has been considered an important criterion for sub–ice shelf sediments in bays or fjords (Domack et al., 1995); however, in the case of a

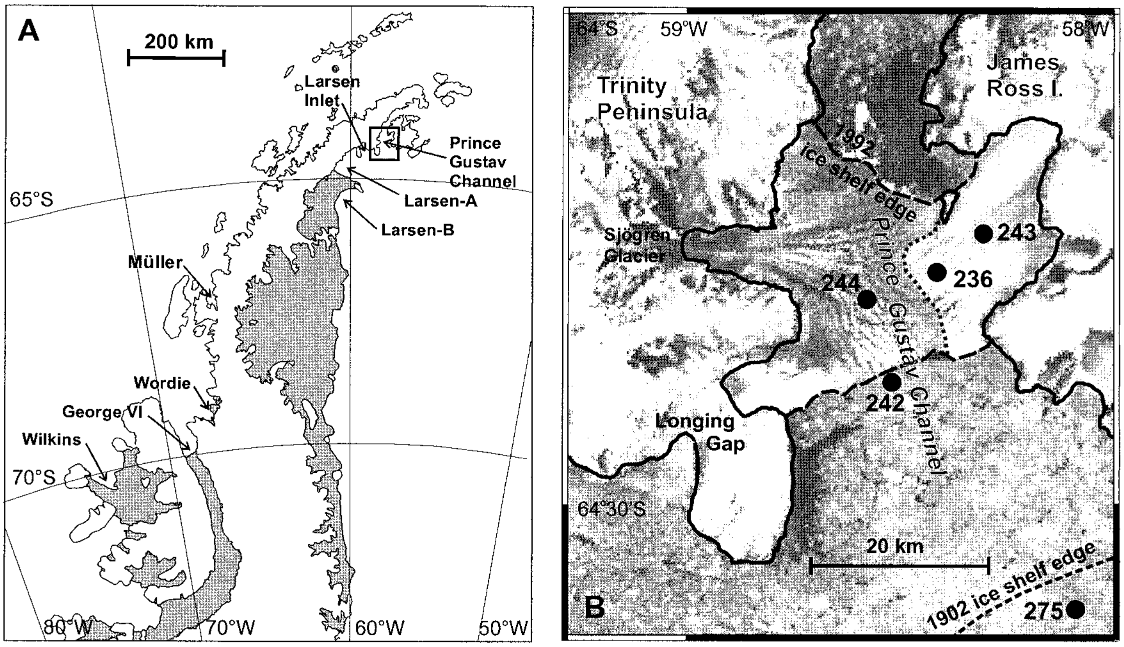

Figure 1. A: Antarctic Peninsula, showing 1999 extent of ice shelves (gray) and position of image in B. Named ice shelves have retreated in recent decades (Vaughan and Doake, 1996), Wilkins and eastern part of Larsen-B within past few years. B: Satellite image of Prince Gustav Channel (October 1992), showing northern and southern ice shelf margins (long-dash lines) and convergence of ice flowing from Trinity Peninsula and from James Ross Island (dotted line; Rott et al., 1998). Short-dash line marks 1902 ice shelf edge (Cooper, 1997). Circles mark core sites.

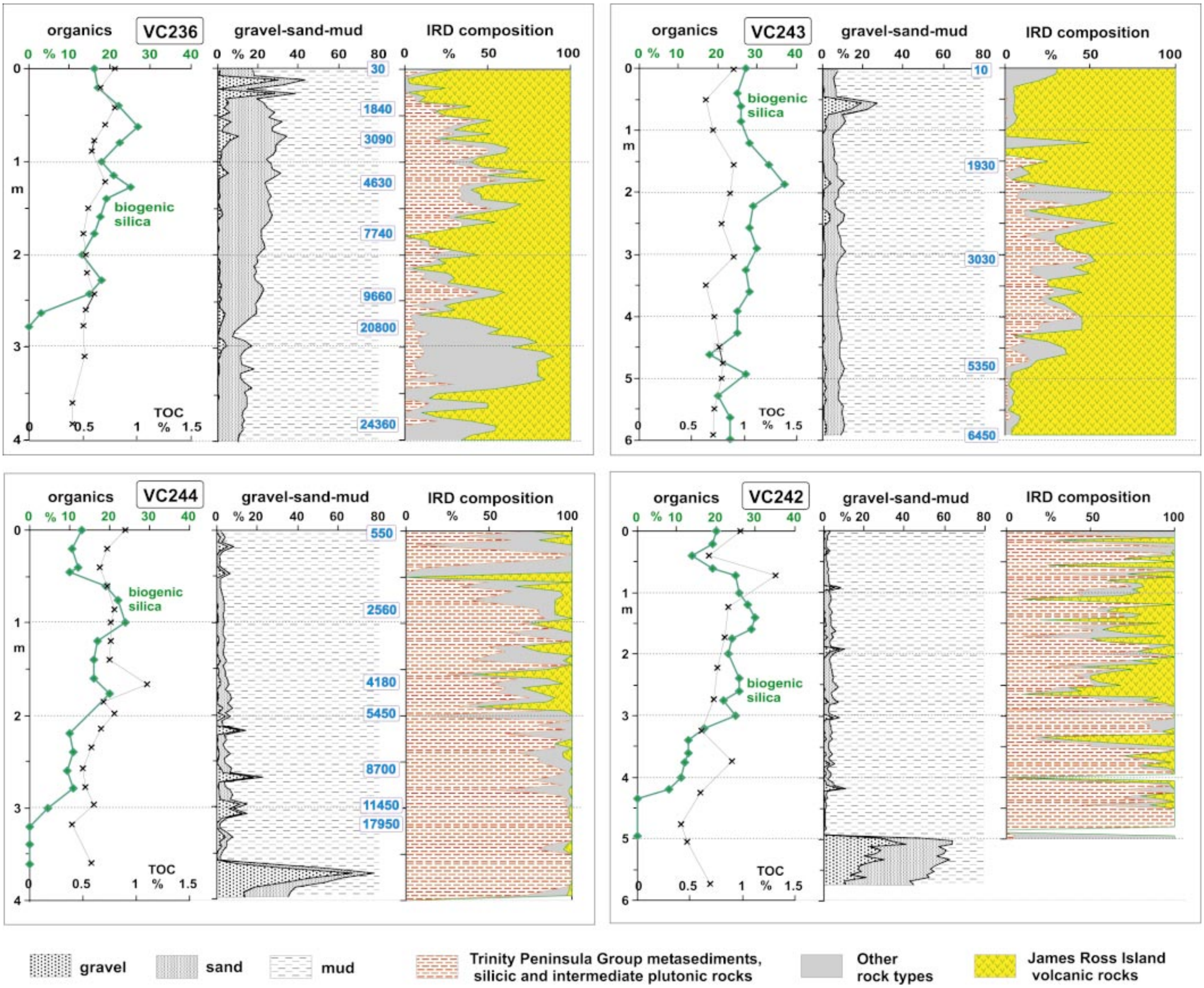

Figure 2. Organic content, sediment texture, and IRD composition for cores in Prince Gustav Channel (location in Fig. 1B). Biogenic silica (mainly diatom frustules; diamond symbols, plotted on 0%–40% scale) occurs throughout core 243 and in upper 2.5–4.3 m of other cores. Diatom assemblage is restricted, dominated by *Chaetoceros* spp. resting spores. Total organic carbon (TOC) (x symbols, on 0%–1.5% scale) decreases down each core from about 0.7% to 0.5% (0.9%–0.7% in VC243); it does not show sharp decrease at base of diatom-bearing interval. $CaCO_3$ values (not shown) are mainly <0.5% and show no trends. Sediments are silty clays with low proportion of gravel and sand, except for diamicts at base of cores 242 and 244 and relatively high sand content in core 236. IRD composition is discussed in text. Boxed blue numbers are AMS ^{14}C ages (see Table 1 and Fig. 3).

glacier tongue or an ice shelf floating over a deep channel, biogenic material can be advected from adjacent areas of seasonally open water, so its absence is not diagnostic of ice at the site. The cores in Prince Gustav Channel contain diatoms throughout the upper 2.5–6 m; total organic carbon (TOC) forms 0.4%–0.8% of the sediment (Fig. 2).

A more useful criterion is the presence of locally derived or exotic ice-rafted debris (IRD). Icebergs can drift through pack ice, but they will not penetrate even a thin floating ice shelf (Reece, 1950). In Prince Gustav Channel, we cored at sites formerly beneath parts of the ice shelf flowing from James Ross Island or from Sjögren Glacier, Trinity Peninsula (Fig. 1B; Rott et al., 1998). The geology of these areas, east and west of Prince Gustav Channel, is very different (Fleming and Thomson, 1979). The identification of IRD derived only from the local ice drainage basins, or from farther afield, can thus yield information on the presence of an ice shelf or of open water with seasonal pack-ice cover. We have therefore measured the amount and type of IRD downcore, comparing the rock types with specimens held by the British Antarctic Survey.

METHODS

The cores (5–6 m long) were obtained using the British Geological Survey Vibracorer. Diatom abundance was determined by point-counting smear slides, and TOC was measured in a LECO induction furnace. For IRD abundance and composition, each core was sampled as continuous 5 cm or 10 cm slabs. The samples were wet sieved at 63 μm, the sand fraction was dried and weighed, then dry sieved at 1 and 2 mm. The fraction coarser than 1 mm (most likely to be ice transported, rather than eolian or current transported) was examined under a binocular microscope, and the proportion of each petrographic type was estimated.

RESULTS

The following rock types were identified: (1) James Ross Island Volcanic Group: lavas, palagonite tuffs, and hyaloclastite breccias, distinguished by their fresh, glassy appearance and the presence of fresh olivine. The wide variety of lavas present includes vesicular and feldspar-phyric types (Nelson, 1966). These rocks occur only on James Ross Island and small areas farther north and northeast. (2) Trinity Peninsula Group metasedimentary rocks, mainly dark gray phyllites and metagraywackes with quartz veins, and some higher-grade metasedimentary rocks (Aitkenhead, 1965). (3) Silicic and intermediate plutonic rocks: granites and granodiorites (Aitkenhead, 1965; Leat et al., 1995). Single, unstrained quartz and feldspar grains were included in this category.

Metasedimentary and plutonic rocks are present in situ only on Trinity Peninsula, though erratics have been reported from James Ross Island (Ingolfsson et al., 1992). Other rock types, which were counted separately but are not distinguished in Figure 2, include poorly cemented lithic sandstones derived from the Cretaceous of James Ross Island and Sobral Peninsula (review in Hathway, 2000), altered volcanic rocks of the Antarctic Peninsula Volcanic Group (Jurassic; Weaver et al., 1982), and dark siltstones and sandstones similar to the Jurassic Botany Bay beds (Bibby, 1966). Black shales of the Jurassic Nordenskjöld Formation (Whitham, 1993) were not seen. Sobral Peninsula and the Jurassic outcrops lie west and southwest of the study area (Fig. 1A).

The petrographic data reveal a very clear signal of changing IRD source areas downcore (Fig. 2). Each core-top sample (recent, open-water sedimentation) contains a mixed IRD assemblage. The upper parts of VC236 and VC243, formerly beneath ice derived from James Ross Island, are dominated by fresh James Ross Island volcanic rocks. Phyllites and plutonic rocks from Trinity Peninsula increase in abundance downcore, peaking at 1.2 m in VC236 and 3–4 m in VC243. A second, smaller peak of Trinity Peninsula IRD occurs at 2.4 m in VC236, and the assemblage of mainly sedimentary and altered volcanic rocks from 2.7 to 3.5 m probably came from southwest of Prince Gustav Channel. Below 3.5 m in this core, IRD is extremely sparse. VC242 and VC244, formerly beneath ice flowing from Sjögren Glacier, contain mainly phyllites and metagraywackes of the Trinity Peninsula Group, with silicic and intermediate plutonic grains from the intrusions at the head of Sjögren Glacier (Aitkenhead, 1965). However, grains of James Ross Island volcanic rocks are present from 0.5 to 2.7 m in VC242 and from 0.5 to 1.9 m in VC244. An additional core from beyond the historical ice shelf limit south of James Ross Island (VC275, Fig. 1B; detailed data not illustrated) contains a diverse assemblage of IRD throughout the upper 1.5 m of diatom-bearing silty clay.

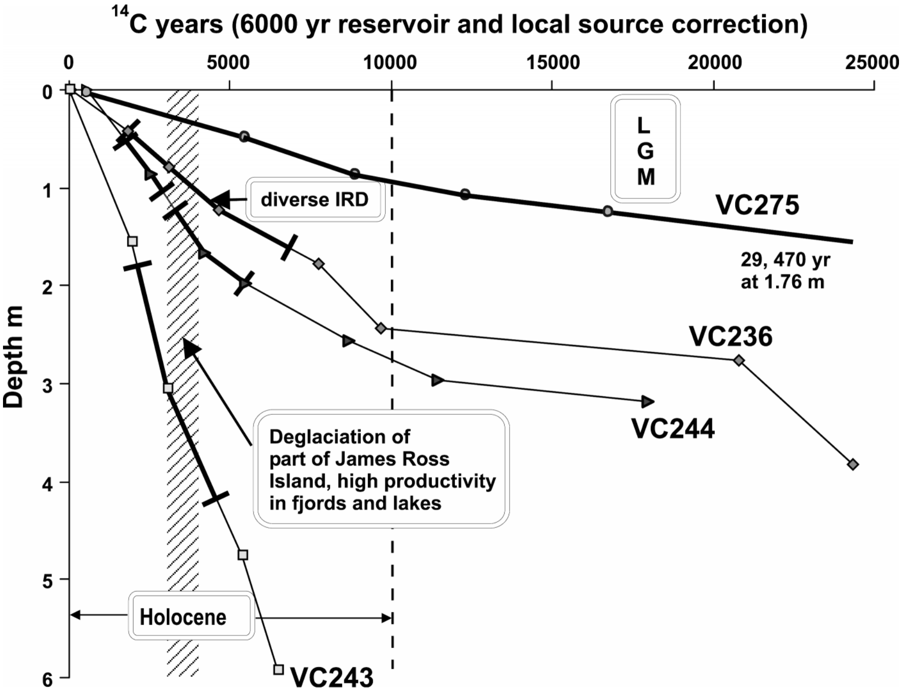

Figure 3. Age-depth plots for cores 236, 243, 244, and 275. These ages have been corrected for Antarctic marine reservoir effect (1300 yr) and local source effect of 4700 yr to bring all the core-top ages close to zero (Table 1). Thickened parts of each curve mark intervals of diverse IRD in cores. VC275 contains diverse IRD throughout.

The cores were ^{14}C-dated by accelerator mass spectrometry on the acid-insoluble organic carbon fraction, which is thought to be derived largely from marine phytoplankton, mainly diatoms. Each core yielded a good age-depth progression (Table 1[1], Fig. 3), though the high core-top ages of >6000 yr were unexpected. The Antarctic marine reservoir effect averages 1300 yr (Berkman et al., 1998). Subtracting this from the core-top dates gives ages of 4710–5250 yr for surface sediment (see Fig. 3). Several explanations are possible for this. There may have been virtually no deposition at the sites for the past 5000 yr, and the diatom-bearing sediments represent the interval from mid-Holocene to ~16 k.y. ago. This is inconsistent with all the other paleoenvironmental data from the region. Loss of surface sediment may have occurred during coring, or older material may have been mixed up-core by glacial or marine processes. The similarity among the core-top ages, combined with the smooth age-depth progression downcore and one order of magnitude variation in sedimentation rates, makes this unlikely (Andrews et al., 1999). There may be a local carbon reservoir effect related to the presence of old glacial meltwater during the phytoplankton growing season (Domack et al., 1989). Glacial ice on the northern Antarctic Peninsula is at most a few thousand years old (Thompson et al., 1994), so this would have only a small effect. Most likely, the sediments contain old carbon from some of the source rocks—i.e., the Cretaceous sediments on James Ross Island (0.5%–1% TOC; Macdonald and Butterworth, 1990), the Jurassic at Longing Gap (0.7%–2.4% TOC; Whitham, 1993), or the Trinity Peninsula Group metasedimentary rocks (<0.5% TOC).

Some 44% of the carbon in the samples would need to be of ''infinite'' age to produce a 4700 yr source correction. It is not known whether the total 6000 yr correction has changed significantly through the Holocene; the smooth age-depth curves in Figure 3, plus the apparent correspondence of the diatom-bearing intervals to the Holocene when a 6000 yr correction is used, imply that it has not. If the diatom-bearing interval in VC242 (0–4.3 m) also corresponds to the Holocene, the peaks of IRD from James Ross Island again lie in the mid-Holocene in this core (Fig. 2). The ages of more than 16 000 yr obtained from nondiatomaceous sediments in cores 236 and 275 may not be reliable, because we cannot be confident about the carbon source.

[1]GSA Data Repository item 2001090, AMS ^{14}C dates on bulk organic carbon in cores from Prince Gustav Channel and south of James Ross Island, is available from Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, editing@geosociety.org, or at www.geosociety.org/pubs/ft2001.htm.

CONCLUSIONS

Four cores in Prince Gustav Channel, from sites formerly covered by floating ice shelves, each contain an interval with IRD from the opposite side of the channel. This could be explained by either major changes in the position of the ice convergence (Fig. 1B) or the disappearance of the ice shelves so that icebergs carrying debris from different sources could drift freely over the sites. Only the latter is consistent with the apparent synchroneity, rather than succession, of the mixed assemblages of IRD (about 5–2 ka; Fig. 3) and with the presence of rock types known only from farther south, in the same samples.

Seasonally open marine conditions were established at about 11 ka, in or sufficiently near to Prince Gustav Channel for diatom frustules to be advected to the sites. A floating ice shelf may have persisted for several thousand years, its subsequent melting coinciding with large-scale deglaciation of northern James Ross Island at 6–7 ka (Ingolfsson et al., 1992) and with the suggested disappearance of the George VI Ice Shelf at $\sim$6500 yr (or 5800 yr if a 1300 yr reservoir correction is used; Clapperton and Sugden, 1982). Colder conditions after about 1.9 ka allowed the ice shelf to reform, until its retreat during recent decades. The inferred period of open water in Prince Gustav Channel compares well with the timing of climate warming deduced from marine, lacustrine, and terrestrial sites elsewhere around the Antarctic Peninsula (Fig. 3; Ingolfsson et al., 1992, 1998; Björck et al., 1996; Domack and McClennen, 1996; Shevenell et al., 1996; Jones et al., 2000). We have demonstrated that at least one small Antarctic ice shelf has varied in extent in response to natural Holocene climate forcing, confirming that such ice shelves are sensitive indicators of regional climate change (Vaughan and Doake, 1996). However, we should not view the recent decay as an unequivocal indicator of anthropogenic climate perturbation.

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