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*The Holocene* 2010; 20; 395 originally published online Feb 22, 2010;
DOI: 10.1177/0959683609353430

Research paper

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The Holocene
20(3) 395–404

DOI: 10.1177/0959683609353430
http://hol.sagepub.com
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**Abstract**

Climatic and oceanographic changes in the South Orkney Plateau, western Antarctica, during the last 8600 years are reconstructed from a 525 cm long gravity core based on sedimentological, geochemical and diatom analyses. The core sediments are composed mostly of light greyish olive diatomaceous silt and mud with a few diatom ooze laminae in the basal part. The core can be divided at 350 cm into two units (4800 cal. yr BP): the lower unit is characterized by variable total organic carbon (TOC) content and higher $CaCO_3$ content, and the upper unit is characterized by higher TOC and lower $CaCO_3$. The content of biogenic silica varies similar to TOC content in the lower unit but does not increase in the upper unit despite increased TOC. The variations in the organic matter composition and the amount of carbonate suggest that Scotia Sea water had been dominant in the study area prior to 4800 cal. yr BP. Warmer conditions during the middle Holocene are also supported by the fewer sea-ice diatom taxa and a more sub-polar form of *Eucampia antarctica* in the lower unit. The increased sea ice and decreased influence of Scotia Sea water in the upper unit reflect climate cooling occurred at 4800 cal. yr BP.

**Keywords**

Antarctic, climate change, Holocene, palaeoclimate, palaeoceanography, South Orkney Plateau

## Introduction

The Antarctic Peninsula is warming very rapidly, with a temperature increase during the last century more than five times the global mean (Vaughan *et al*., 2003). To clarify whether this rapid warming is a part of natural variability or an effect of human activity, we must understand how the natural climate system of the Antarctic Peninsula works. Reconstruction of past climate changes, especially those of the near past, is a key strategy to understand the climate variability of the Antarctic Peninsula.

Knowledge of Holocene climate variability in Antarctica and the Southern Ocean is not as well established as in the Northern Hemisphere owing to difficulties in obtaining palaeoclimate archives, the rare preservation of foraminifera in the Southern Ocean sediments and the limited number of available proxies. Deuterium records from 11 Antarctic ice cores indicate a widespread early-Holocene climate optimum between 11.5 and 9 ka, followed by a minimum about 8 ka (Masson *et al*., 2000). Diatom abundances and assemblages in the Palmer Deep on the western Antarctic Peninsula (Ocean Drilling Program site 1098) indicate the occurrence of a ‘deglaciation phase’ between 13 and 11 ka and ‘climate reversal’ between 11 and 9 ka (Sjunneskog and Taylor, 2002; Taylor and Sjunneskog, 2002).

Holocene climate reconstructions from sediment core records from the western Antarctic Peninsula and the Southern Ocean generally suggest warmer middle-Holocene and cooler late-Holocene conditions. The diatom assemblage in Lallemand Fjord on the western Antarctic Peninsula indicates a sea-ice minimum between 7.9 and 3.9 ka (Taylor *et al*., 2001). The mid-Holocene climate optimum (~9–4 ka) and subsequent late-Holocene Neoglacial were identified in the Palmer Deep on the western Antarctic Peninsula by variations in diatom proxies (Sjunneskog and Taylor, 2002; Taylor and Sjunneskog, 2002) and magnetic parameters (Brachfeld and Banerjee, 2000; Brachfeld *et al*., 2002). Minimum sea ice and warm water conditions occurred in the Maxwell Bay, South Shetland Islands, between 8.2 and 5.9 ka (Milliken *et al*., 2009). In Western Bransfield Basin, diatom analysis pointed to a shorter mid-Holocene climate optimum spanning from 6800 to 5900 cal. yr BP (Heroy *et al*., 2008). On the western Antarctic Peninsula in Neny Fjord, climate cooling occurred between 6 and 4.5 ka (Allen *et al*., 2007). Marine palaeoclimatic evidence from the South Atlantic Sector of the Southern Ocean indicates climate cooling in that area at approximately 5 ka, which accompanied surface temperature cooling, sea ice advance and an increase of ice-rafted debris (IRD) (Hodell *et al*., 2001). Sea-ice variations based on diatom assemblages also confirm that a winter sea-ice expansion comparable with the present was only established after the middle Holocene in the Atlantic Sector of the Southern Ocean (Gersonde and Zielinski, 2000). The annual trend of sodium ion fluxes in the snowpack in the Dome Fuji region of East Antarctica, which is controlled by the sea-ice extent in the southern Indian Ocean and Weddell Sea, suggests that the onset of climate cooling took place about 5.4 ka (Iizuka *et al*., 2008).

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Received 2 February 2009; revised manuscript accepted 14 September 2009

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However, the timing and mode of the mid-Holocene climate optimum seems to have varied depending on the locality. The record of IRD from Prince Gustav Channel in the east of the northern Antarctic Peninsula shows that the climate optimum occurred at 5–2 ka when the Prince Gustav Channel ice shelf retreated and seasonally open water existed (Pudsey and Evans, 2001). Diatom record from Lallemand Fjord provides evidence of a bimodal warm period at ~3–5 ka as well as at ~8 ka (Shevenell *et al.*, 1996). The diatom abundance from the Palmer Deep indicates a short warm period 4.8–4.4 ka following a cool phase ~6.5–4.7 ka (Sjunneskog and Taylor, 2002). To understand the complexity of the Holocene climate of the Antarctic Peninsula, it is necessary to acquire as many palaeoclimate data as possible from different parts of the peninsula and adjacent regions.

The South Orkney Plateau is a continental fragmentation constituting part of the South Scotia Ridge (Figure 1). Holocene climate change around the South Orkney Plateau has rarely been studied (e.g. Herron and Anderson, 1990; Jones *et al.*, 2000). The deposition of postglacial marine sediment was deduced to be prior to 6–7 ka, based on radiocarbon dates for diatomaceous muds overlying poorly sorted muds in piston cores from the western slope of the South Orkney Plateau (Herron and Anderson, 1990). Terrestrial records during the last 5.7 ka from Signy Island, South Orkney Islands, suggest a climate optimum at ~3.8–1.3 $^{14}$C ka (Jones *et al.*, 2000).

In this paper, we present an 8.6 ka palaeoceanographic history of the northwestern part of the South Orkney Plateau. The relatively high sedimentation rate (about 61 cm/ky) and the lack of erosional surfaces in the studied core make it suitable for palaeoclimate research. The core site is located between the Southern Boundary of the Antarctic Circumpolar Current (ACC) and the Antarctic Slope Front at present (Orsi *et al.*, 1995; Heywood *et al.*, 2004; Figure 1); thus the core is suited for the study of the oceanographic response to climate changes. The study area was probably not grounded by the South Orkney Ice Cap during the last glaciation (Herron and Anderson, 1990). Herron and Anderson (1990) suggested that the sea-ice conditions over the plateau have not changed much since the retreat of the ice cap and ice shelf, considering the homogeneity of the uppermost diatomaceous mud. Geochemical and diatom analyses in this study, however, show that the climatic and oceanic conditions over the plateau were variable during the Holocene.

## Material and methods

Gravity core GC02-SOI03 (47°00′W, 60°22′S, 786 m water depth, core length 525 cm) was retrieved from northwestern part of South Orkney Plateau, western Antarctica, during the 2002/2003 Korea Antarctic Research Program (KARP) cruise of the R/V *Yuzhmorgeologiya* (Figure 1). The core was measured for

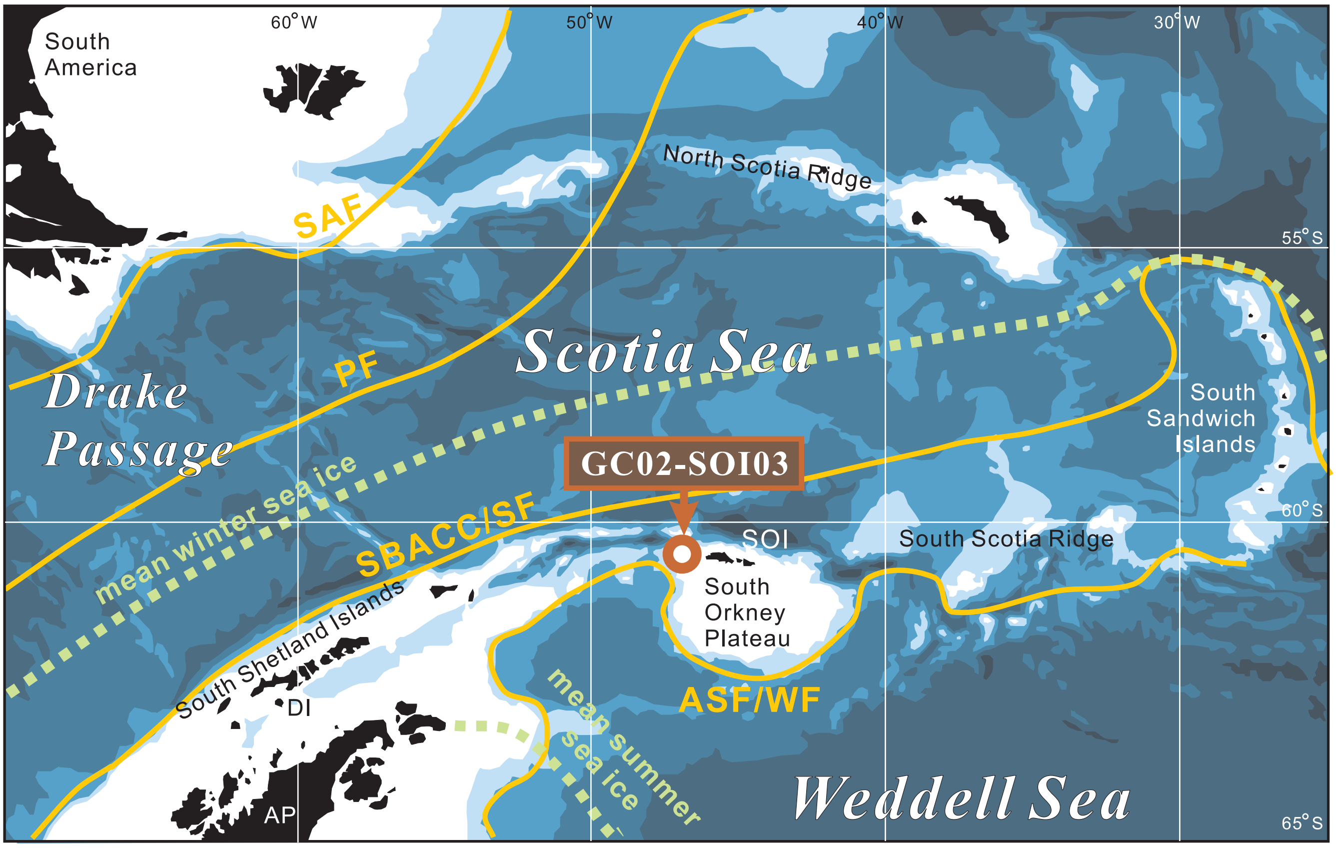


**Figure 1.** Map showing the core location, fronts of the ocean current system and bathymetric features in and around the Scotia Sea. Water depth coutours are shaded in 1000 m steps. SAF, Subantarctic Front; PF, Polar Front; SBACC/SF, Southern Boundary of the Antarctic Circumpolar Current and Scotia Front; ASF/WF, Antarctic Slope Front and Weddell Front (Orsi *et al.,* 1995; Heywood *et al.,* 2004). Mean winter and summer sea ice extents are according to the Naval Oceanography Command Detachment (1985). AP, Antarctic Peninsula; DI, Deception Island; SOI, South Orkney Islands

**Table 1.** Accelerator mass spectrometry radiocarbon ages from core GC02-SOI03 sediments. Uncorrected radiocarbon age from Bak *et al.* (2007).

| Depth (cmbsf) | Lab code | $^{14}C$ age ± 1σ | $\delta^{13}C$ (‰) (yr BP) | Corrected and calibrated age (cal. yr BP) | Material |
|---|---|---|---|---|---|
| 0 | NZA17341 | 2909 ± 45 | –25.1 | 0 | total organic carbon, from box core BC02-SOI03 |
| 102 | NZA18572 | 4572 ± 40 | –25.5 | 1217 | total organic carbon |
| 202 | NZA18573 | 5629 ± 45 | –25.0 | 2437 | total organic carbon |
| 302 | NZA18574 | 6769 ± 55 | –24.6 | 3820 | total organic carbon |
| 402 | NZA18575 | 8455 ± 55 | –24.7 | 5936 | total organic carbon |
| 502 | NZA18576 | 10 542 ± 70 | –25.0 | 8093 | total organic carbon |

magnetic susceptibility, described for visible sedimentary structure, and x-rayed. Samples were taken every 4 cm for grain size analyses and water content. The size distribution of grains larger than 4φ (63 μm) was determined by wet sieving. Finer fractions were analysed on a Micrometrics Sedigraph 5100. Another set of 1 cm thick samples were collected at 2 cm intervals for carbon and nitrogen analyses. Total carbon and nitrogen contents were analysed using a Flash EA 1112 element analyser. Total inorganic carbon content was analysed for the 2N HCl-dissolved portion using a UIC 5030 coulometer and multiplied by 8.333 to determine carbonate content. Total organic carbon (TOC) content was determined by subtracting total inorganic carbon from total carbon content. A third set of samples was collected every 10 cm to determine biogenic opal content following the method of Mortlock and Froelich (1989). Mineralogy of the core sediment was examined by x-ray diffraction analysis of seven bulk sediment samples. The samples were scanned from 3 to 50° 2θ with a scan speed of 5° 2θ/min under Ni-filtered CuKα radiation. The radiocarbon AMS dating technique was used for the TOC from six sediment samples to determine the age and sedimentation rate (Table 1).

Radiocarbon dating of Antarctic waters and marine organisms has yielded anomalously old ages as well as highly variable ages according to sampling sites (Gordon and Harkness, 1992). Therefore, measured radiocarbon ages must be corrected using an appropriate reservoir age. We obtained the radiocarbon age of sea-floor sediment from the top of a box core retrieved from the same site as SOI-03 to prevent uncertainty related with possible loss of the core top. Measured radiocarbon ages were corrected using the core top age as the reservoir age and then converted to calibrated calendar ages using the CALIB 5.0 program (Stuiver and Reimer, 1993). Sample ages were determined by interpolation between calibrated ages, assuming a constant sedimentation rate between dated horizons.

Samples for diatom analysis were collected at 8 cm intervals for the upper 500 cm, and more samples were taken thereafter to reflect changes in sediment facies. Quantitative diatom slides were prepared by the settling technique of Battarbee (1973). At least 400 diatom valves were counted from each slide at 800× to 1000× magnification using Y-IM 50i Nikon light microscope. Relative abundances of diatoms were re-counted excluding *Chaetoceros* resting spores, which dominate the diatom assemblage in all samples and make it difficult to read the record of the other diatom species. Bak *et al*. (2007) reported the abundances of diatom species from the core. In this study, the abundances of some indicator species, i.e. *Fragilariopsis curta*, *F. cylindrus* and *Eucampia antarctica*, were used for palaeoclimate reconstruction.

# Results

## *Age and sedimentation rate*

The radiocarbon age for the sea-floor sediment from the South Orkney Plateau is 2909±45 yr BP. This old age of the core top may be attributed to the large and highly variable reservoir effect of Antarctic waters (Gordon and Harkness, 1992) and/or the possible input of older, recycled organic matter. In Antarctic marine sediments, the input of terrestrial organic matter is usually considered insignificant (e.g. Venkatesan and Kaplan, 1987), but some old carbon may incorporate into sediments, as seen in the discrepancy between radiocarbon ages of sediment-trap and core-top sediments from the Jane Basin, northern Weddell Sea (Pudsey and King, 1997). Radiocarbon ages for the SOI-03 core continuously increase with depth (Figure 2), implying that input of old carbon, if any, has been minimal or has not varied much during the deposition of the core sediment. The narrow range of $\delta^{13}C$ of dated organic matter, all within the range of –25±0.5‰ (Table 1), also supports the relatively homogenous composition of organic matter. The core-top age was subtracted from measured radiocarbon dates as a correction for the reservoir effect and possible old carbon input. On the basis of corrected and calibrated ages, the core represents deposition during the last 8600 years, with an average sedimentation rate of 61 cm/ky. This rate is about 20 times higher than those observed from the southern slope of the South Orkney microcontinent (Frank *et al*., 1995) and about five to ten times higher than those from the Scotia Sea (Pudsey and Howe, 1998). This high sedimentation rate makes it possible to reconstruct palaeoenvironmental changes of the South Orkney region during the Holocene in a relatively high resolution. The sedimentation rate decreases with depth (Figure 2).

## *Core description*

Core GC02-SOI03 is dominated by light greyish olive (10Y 5/2) diatomaceous silt to mud with a few diatom ooze laminae in the basal part (Figure 3). The silt and clay size fraction constitutes more than 90% of the sediments, and the mean grain size is 6.9 φ (Figure 3). The core is mostly composed of silt, according to Folk and Ward's (1957) scheme, but it becomes finer grained below 440 cm, with less silt and more clay (Figure 3) and is classified as mud. Grains larger than 2 mm are rare, implying only a minor contribution of ice-rafted materials to the study area. The upper part of the core (above 330 cm) is mostly composed of massive light greyish olive diatomaceous silt. X-radiography shows that extensive bioturbation obscured stratification in most of the

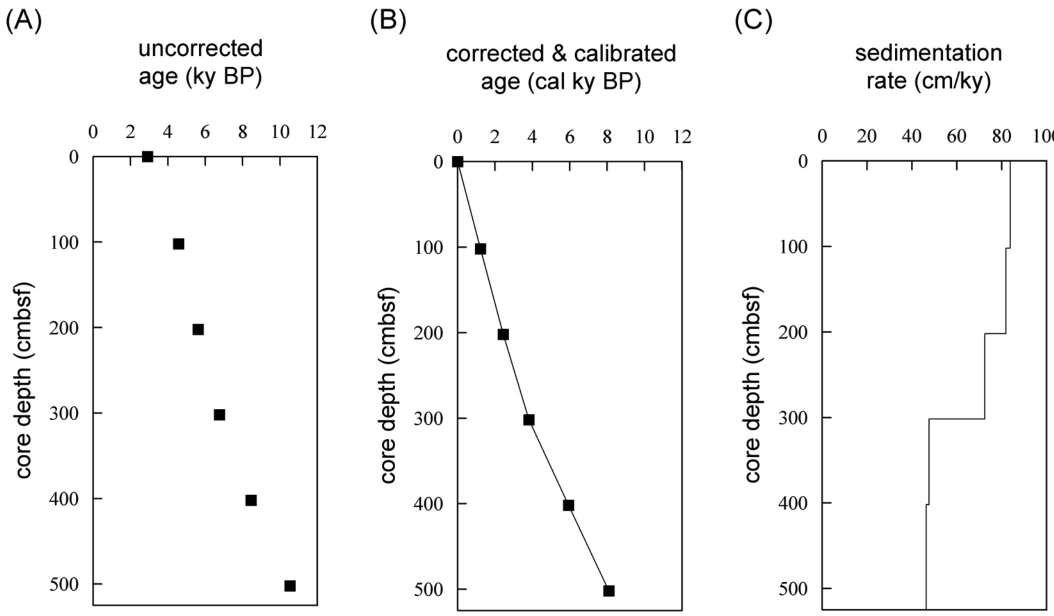


**Figure 2.** (A) Uncorrected radiocarbon data, (B) our age model based on reservoir-corrected and calibrated radiocarbon data, and (C) sedimentation rate for GC02-SOI03. Radiocarbon data according to Bak *et al.* (2007)

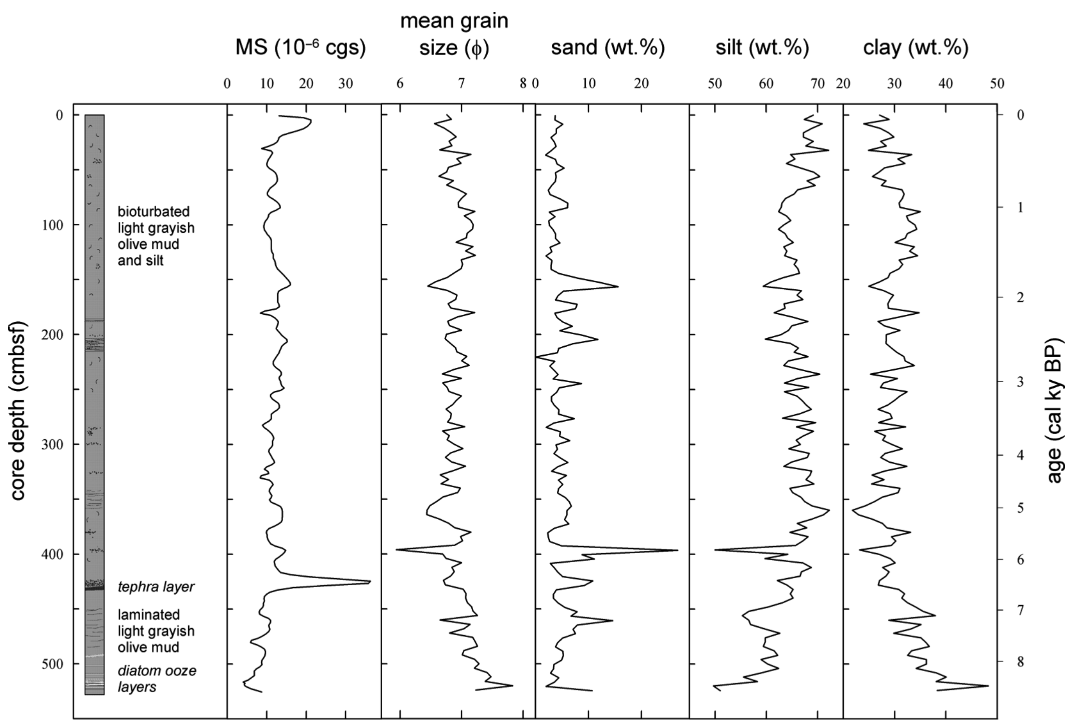


**Figure 3.** Magnetic susceptibility (MS) and grain size distribution for GC02-SOI03

interval (Figure 4A). Stratification is often preserved below 330 cm. Between 330 and 450 cm light greyish olive diatomaceous silt with crude stratification can be seen (Figure 4B). Below 450 cm, the core is composed of stratified light greyish olive mud (Figure 4C). Moderate yellow (5Y 7/6) and dusky yellow (5Y 6/4) laminations with monospecific diatoms are abundant below 506 cm (Figure 4D). The ooze layers are intercalated with medium dark grey (N4) mud between 520 and 525 cm. Diatom species constituting the layers are *Corethron criophilum*, *Eucampia antarctica* var. *recta* and *Rhizosolenia styliformis*. Herron and Anderson (1990) found diatom ooze layers underlying the diatomaceous mud unit at 155–165 cm below the sea surface in core 85-12 from the southwestern slope of the South Orkney Plateau. Sedimentation in the study area seems to be three times higher than that in the southwestern slope area, assuming that the diatom ooze layers are correlated with those from this study.

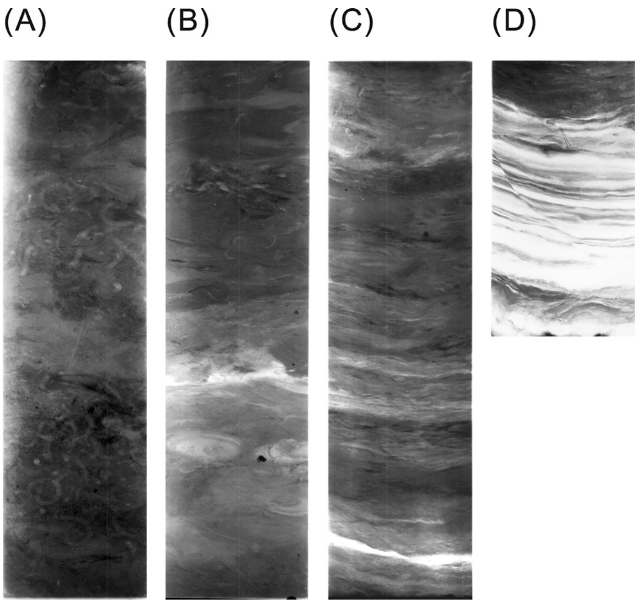


**Figure 4.** X-radiograph images of core GC02-SOI03. (A) 60–90 cm, (B) 360–390 cm, (C) 480–510 cm, (D) 510–525 cm

A tephra layer was observed at 423–429 cm (about 6400–6500 cal. yr BP), distinguished by its prominent magnetic susceptibility (MS) peak (Figure 3). Observation of the layer under an optical microscope revealed a high concentration of basaltic ashes, which was probably responsible for the high MS values. The prominent MS peak associated with this tephra layer will help in the identification of correlative layers in the vicinity. The age of the layer is estimated to be close to the 6.5 ka age of the Vega Drift ash layer in the northern Prince Gustav Channel, Weddell Sea, which was derived from Deception Island (Keller *et al*., 2003). Detailed geochemical evidence is required to determine whether these ash layers truly are correlated, as the study area is much farther from Deception Island than Prince Gustav Channel is (63°50′S; 58°15′W). Microscopic observation of sediments near the top of the core, which also record higher MS values, indicates the presence of both silicic and basaltic ashes.

X-ray diffraction analysis on bulk sediment samples reveals little variation in mineralogy (Figure 5). Quartz, plagioclase, illite/mica and chlorite were detected in all samples. Amphibole was identified in some samples. The abundance of plagioclase and presence of amphibole are consistent with the predominance of volcanic rocks in the northern part of the Antarctic Peninsula region. Probably because of the amorphous nature of the silica, opal peaks were not identified from the x-ray diffraction diagrams even though the core sediments contain 20 to 30 wt.% of biogenic silica.

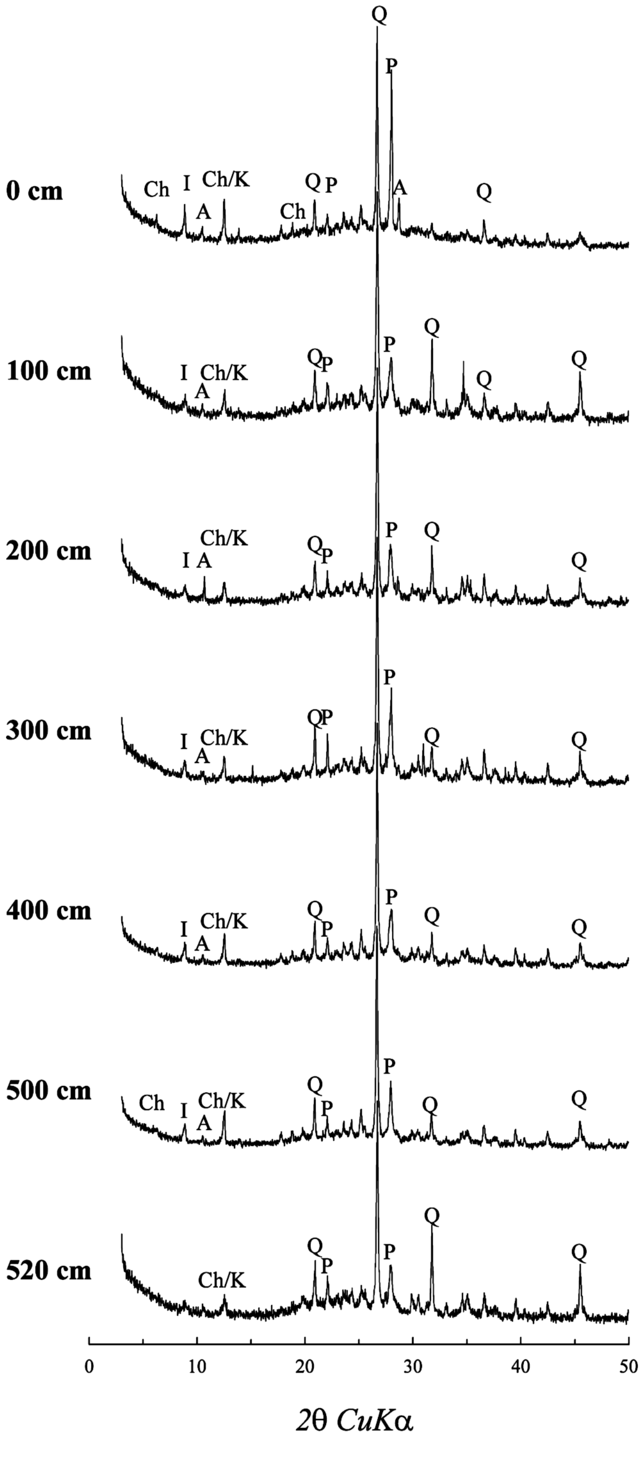


**Figure 5.** X-ray diffraction diagram of powdered bulk sediment samples of core GC02-SOI03. A, amphibole; Ch, chlorite; Ch/K, chlorite and/or kaolinite; I, illite; P, plagioclase; Q, quartz

## *Total organic carbon, calcium carbonate, and biogenic silica contents*

Core GC02-SOI03 was divided into two units based on TOC and $CaCO_3$ contents (Figure 6): low and variable TOC and high $CaCO_3$ content in the lower unit (350–525 cm; 8600 to 4800 cal. yr BP) and high TOC and low $CaCO_3$ content in the upper unit (0–350 cm; 4800 cal. yr BP to present).

The TOC contents of the sediment samples range from 0.56 to 1.07 wt.% with a mean of 0.81 wt.% (Figure 6). TOC content is generally higher in the upper unit (0.69–0.96 wt.%, mean 0.84 wt.%) than in the lower unit (0.56–1.07 wt.%, mean 0.76 wt.%), although the highest TOC contents are found in diatom ooze layers from the lowermost part of the lower unit (Figure 6). Because the sedimentation rate in the upper unit is higher than that in the lower unit (Figure 2), the TOC accumulation rate in the upper part should still be higher than that in the lower part. The ratio of total organic carbon to total nitrogen (C/N) does not vary much and has a mean of 8.8 (Figure 6), indicating the dominance of marine organic matter in the sediment (Stein, 1991). TOC contents in sediments generally increase with decreasing grain size, but the

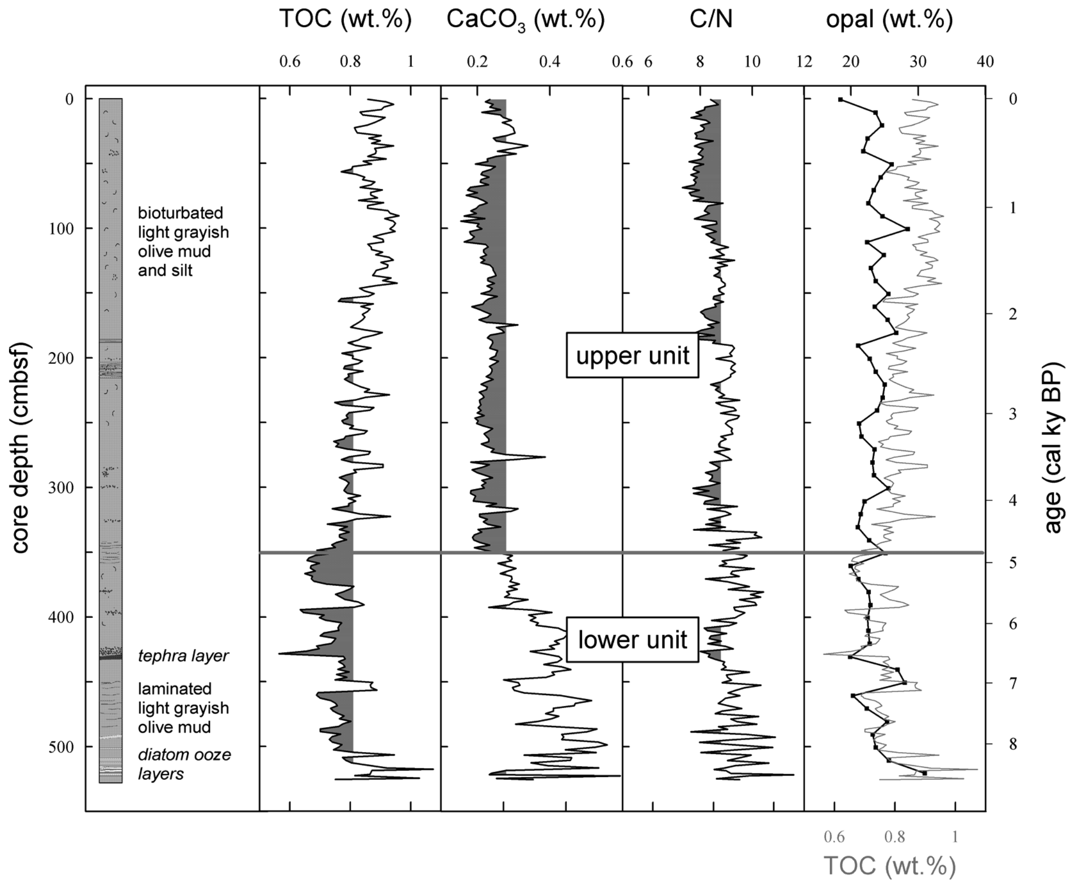


**Figure 6.** Total organic carbon content (TOC), $CaCO_3$ content, TOC/total nitrogen ratio (C/N), and biogenic silica content (opal) of core GC02-SOI03

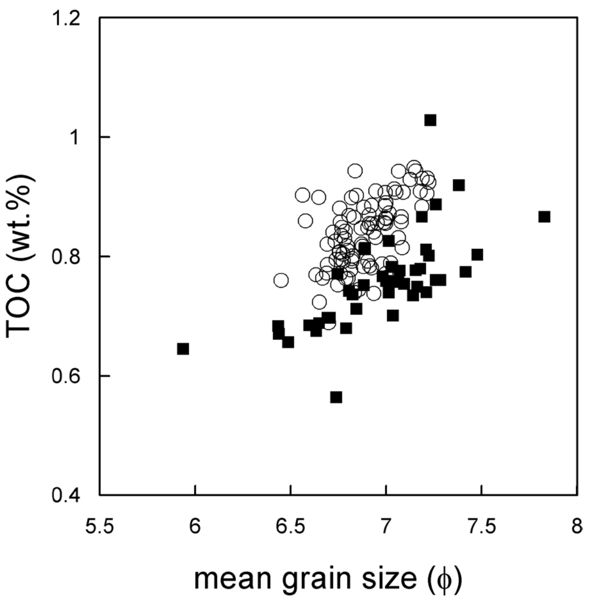


**Figure 7.** Total organic carbon (TOC) content versus mean grain size for the lower (squares) and upper (circles) units of core GC02-SOI03. TOC content increases with decreasing grain size for both units. The TOC contents of the upper unit samples are higher than those of lower unit samples with comparable grain size ranges

higher content of TOC in the upper unit compared with the lower unit cannot be attributed to the decreased grain size; the diagram of TOC content versus mean grain size shows that the TOC contents of the upper unit samples are higher than are those of the lower unit samples of comparable mean grain size (Figure 7). Considering the TOC accumulation rate, C/N ratio and TOC-size correlation, the increased TOC content in the upper unit is most likely the result of increased marine production and/or preservation.

Contents of $CaCO_3$ range from 0.15 to 0.59 wt.%, with a mean of 0.28 wt.% (Figure 6). Microscopic observation of sand-sized grains revealed few carbonate grains and the absence of detrital carbonate grains; only a few robust, partially dissolved benthic foraminiferal grains were found in the lower unit. The $CaCO_3$ contents in the upper unit (0.15–0.39 wt.%, mean 0.23 wt.%) are lower than in the lower unit (0.23–0.59 wt.%, mean 0.38 wt.%), and no biogenic carbonate grains were observed. The low $CaCO_3$ content of the core and the presence of few partially dissolved biogenic carbonate grains in the lower unit imply that the $CaCO_3$ content of the core was controlled by dissolution rather than carbonate production.

The core sediments contain 18.4–30.9 wt.% of opal, with a mean of 23.5 wt.% (Figure 6). Despite the increased TOC content in the upper unit, opal contents in the lower unit (19.9–30.9 wt.%, mean 23.6 wt.%) are similar to those in the upper unit (18.4–28.4 wt.%, mean 23.5 wt.%). Variation of opal content in the lower unit is similar to that of the TOC content (Figure 6), suggesting the dominance of opal-producing organisms (mainly diatoms) in the organic carbon sources. This synchronous variation of opal and TOC content was not observed in the upper unit (Figure 6). The decoupling of opal and TOC contents in the upper unit implies that the contribution from organisms other than diatoms increased, and as a result diatoms no longer played a dominant role in controlling TOC content.

## Diatom analysis

The diatom abundance of the core ranges from 10 to 20 million valves per gram dry sediment except in diatom ooze layers, and the diatoms are dominated by *Chaetoceros* resting spores (CRS) (Bak *et al*., 2007). High abundances of CRS are found in the marginal ice zone where the influx of fresh water and high concentration of nutrients during spring melt provide optimal conditions for *Chaetoceros* (Crosta *et al*., 1997). Abundant CRS in the core thus indicates that the study area has been under the influence of sea ice during the last 8800 years.

*Fragilariopsis curta* and *F. cylindrus* are common in the marginal ice zone, with their relative abundance increasing southward and reaching the largest values of more than 60% in coastal areas of the Weddell Sea (Gersonde and Zielinski, 2000). Gersonde and Zielinski (2000) noted that the relative amount of *F. curta* and *F. cylindrus* in the diatom assemblage is mainly controlled by the distribution of Antarctic sea ice and more than 3% for these species in the diatom assemblage indicates the presence of winter sea ice. The relative abundance of *F. curta* and *F. cylindrus* in the core ranges between 2 and 29% (mean 10%) and is higher in the upper unit than in the lower unit (Figure 8). In the lower unit, *E. antarctica* is more abundant than in the upper unit, and the subpolar form of *E. antarctica* predominates (Figure 8). The subpolar form of *E. antarctica* (*E. antarctica* var. *antarctica*) is characterized by its asymmetric girdle view, whereas the polar form of *E. antarctica* (*E. antarctica* var. *recta*) is symmetric in girdle view (Fryxell, 1989; Fryxell and Prasad, 1990). However, an *E. antarctica*-rich ooze layer at 517 cm is composed of the polar form of *E. antarctica* only. In the upper unit, abundance of *E. antarctica* is lower, and the polar form predominates in the population (Figure 8).

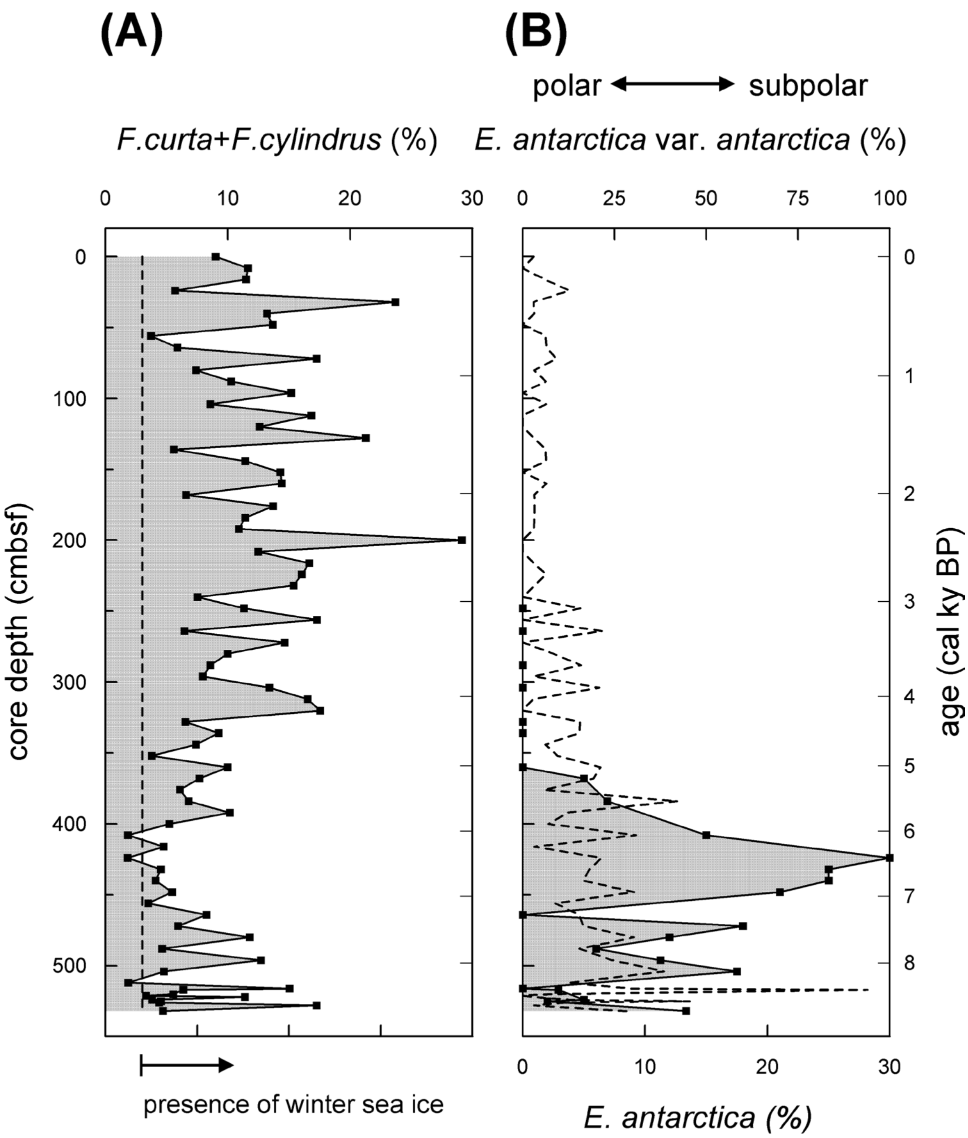


**Figure 8.** Distribution of diatom species indicating palaeoenvironmental changes. (A) Relative abundance of sea-ice indicator diatoms *Fragilariopsis curta* and *F. cylindrus* in the diatom assemblage. Relative abundance of more than 3% for *F. curta* and *F. cylindrus* indicates the presence of winter sea ice (Gersonde and Zielinski, 2000). (B) Relative abundance of *Eucampia antarctica* in the diatom assemblage (dashed line) and relative abundance of *E.antarctica* var. *antarctica* within *E. antarctica* (solid line). Diatom abundance data are from Bak *et al.* (2007)

## Discussion

Sediment deposited prior to 8200 cal. yr BP (below 506 cm) from the South Orkney Plateau is characterized by the presence of a few diatom ooze layers and highly variable TOC and carbonate contents. Diatom ooze layers underlying diatomaceous glacial-marine sediment were also reported from a core from the southwestern slope of the plateau (Herron and Anderson, 1990), suggesting that the environmental condition that caused the deposition of the diatom ooze layers may have prevailed over the region. In the sediment below 506 cm, the abundance of *F. curta* and *F. cylindrus* is variable but slightly higher than in the overlying sediment, and *E. antarctica* is dominated by the polar form (Figure 8), implying more sea-ice cover and a colder condition in the plateau prior to 8200 cal. yr BP. Whether this indicates a climate reversal event as in the Palmer Deep at 11.0–9.0 ka (Taylor and Sjunneskog, 2002) or continued warming and deglaciation following the last glaciation could not be determined in this study because of the lack of an older sediment record.

The mid-Holocene climate optimum occurs in the South Orkney Plateau record at 8200–4800 cal. yr BP. The decreased abundance of *F. curta* and *F. cylindrus* and the increased subpolar form of *E. antarctica* suggest a less extensive winter sea-ice cover on the plateau during the climate optimum (Figure 8). The lowest abundance (about 3%) of *F. curta* and *F. cylindrus* between 400 and 456 cm explicitly displays the retreat of the Weddell Sea marginal ice zone and the active Weddell/South Orkney Plateau communication, suggestive of the warmest climatic condition occurring during 7100–5900 cal. yr BP. This is comparable with the shorter mid-Holocene climate optimum identified from the Bransfield Basin at 6800 to 5900 cal. yr BP (Heroy *et al*., 2008).

Since 4800 cal yr BP, the differences in organic and geochemical proxies with lower relative abundance of the subpolar form of *E. antarctica* (Figures 6 and 8) may indicate a return to a colder climate. The timing agrees with most marine palaeoenvironmental records of Neoglacial onset based on uncalibrated radiocarbon ages (Ingólfsson and Hjort, 2006, and references therein).

Analyses of the TOC, $CaCO_3$ content, and diatom assemblage revealed a contrast between the upper and lower units, implying that a significant environmental change occurred at 4800 cal. yr BP. The increased TOC, despite the higher sedimentation rate and coarser grain size in the upper unit, likely reflects increased marine production during the later part of the Holocene. Increased marine production is often attributed to climate warming, but diatom analysis of the core disputes the warmer late Holocene in the study area and instead indicates the presence of more sea ice during the deposition of the upper unit (Figure 8). The relationship between opal and TOC contents implies a change in the composition of organic matters from diatom-dominated to mixed at the boundary between the lower and upper units (Figure 6). Therefore, increased TOC in the upper unit seems to reflect an increase of organic matters other than diatoms.

The study area is located in the Weddell–Scotia Confluence (WSC) region (Gordon, 1967) between the Southern Boundary of the ACC (Scotia Front) and the Weddell Front (Figure 1). The WSC region separates Weddell Sea water in the south from eastward-flowing Scotia Sea water in the north. Weddell Sea water in the south is chemically distinguished from Scotia Sea water by its high silicate concentration (Prego, 1991; Leynaert *et al*., 1991; Dafner *et al*., 2003). The phytoplankton community is dominated by diatoms in the Scotia Sea, but the number of flagellates increases southward, and flagellates are more abundant than diatoms in the Weddell Sea (Bak *et al*., 1992; Schloss and Estrada, 1994; Socal *et al*., 1997). The molar ratio of biogenic silica to particulate organic carbon is also higher in the Scotia Sea (Leynaert *et al*., 1991). Assuming that the difference in organic composition can be detected in sediments, sediment under the influence of Scotia Sea water would have a stronger diatom signal than would sediment under the influence of Weddell Sea water, which is observed in the lower unit. It is thus concluded that the differences in organic and geochemical proxies between the lower and upper units were induced by a replacement of the main water regime influencing the deposits from Scotia Sea water to Weddell Sea water. This shift seems to have accompanied the climate change from warm to cold conditions; during the warmer period (8600–4800 cal. yr BP), the Scotia Sea water intruded into the study area, but the intrusion weakened from 4800 cal. yr BP with the onset of the Neoglacial condition in the South Orkney Plateau.

The $CaCO_3$ content is less than 0.5 wt.% throughout the core owing to the poor preservation of carbonate facies in the study area (Figure 6). A distinct drop of the $CaCO_3$ content is observed at the boundary between the lower and upper units, indicating a more corrosive water condition during the deposition of the upper unit. The coincidence of the timing of decrease in $CaCO_3$ content, increase in TOC content, and change in diatom composition seems to be due to the increased influence of the Weddell Sea water. Anderson (1975) recognized four major bottom water masses in the Weddell Sea: saline Shelf Water, fresh Shelf Water, Warm Deep Water and Antarctic Bottom Water. With the exception of fresh Shelf Water, which flows westward off the eastern continental shelf of the Weddell Sea, these Weddell bottom waters dissolve $CaCO_3$. Surface sediments from the Scotia Sea are also subject to carbonate dissolution, except samples from northern sites near South America (Pudsey and Howe, 1998). The overall carbonate-dissolving nature of the bottom waters in the surrounding area explains the generally low $CaCO_3$ content (less than 0.6 wt.%) of the core sediments. The coring position obtained from 786 m water depth is under the Warm Deep Water (WDW) observed along the South Scotia Ridge at 200–1500 m water depth (Naveira Garabato *et al*., 2002). The water is characterized by a mid-depth potential temperature maximum in the Weddell Sea. After the onset of the Neoglacial condition, the glacial regime probably increased with an expansion of sea ice. From a palaeoceanic point of view, the relative intensification of WDW from the Weddell Sea seems to have promoted the dissolution of carbonate deposits of the South Orkney Plateau. If supported by more data on $CaCO_3$ content from surface sediments across the WSC, the $CaCO_3$ content of the confluence region may make a good indicator for the influence of Weddell Sea water.

## Conclusions

The mid-Holocene climate optimum in the South Orkney Plateau occurred between 8200 and 4800 cal. yr BP. The dominance of Scotia Sea water during the climate optimum was inferred from TOC, opal and $CaCO_3$ contents. The warmest period in the South Orkney Plateau during the last 8600 years seems to have occurred between ~7000 and 6000 cal. yr BP based on diatom analysis. Climate cooling occurred at 4800 cal. yr BP, marked by an increase of sea ice and the decreased influence of Scotia Sea water. The study area has mostly been under the influence of Weddell Sea water since then.

## Acknowledgements

We would like to thank H.S. Moon, S.M. Jeong and other staff of KOPRI for their assistance in the laboratory. We are also grateful to the crew of the R/V *Yuzhmorgeologiya*. We acknowledge the comments from anonymous reviewers. This work was supported by a Korea Polar Research Institute project (PE09010).

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