

RECORD OF HOLOCENE PALAEOCLIMATE CHANGE ALONG THE ANTARCTIC PENINSULA: EVIDENCE FROM GLACIAL MARINE SEDIMENTS, LALLEMAND FJORD

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(with six text-figures and one plate)

SHEVENELL, A.E., DOMACK, E.W. & KERNAN, G.M., 1996 (xi): Record of Holocene palaeoclimate change along the Antarctic Peninsula: Evidence from glacial marine sediments, Lallemand Fjord. In Banks, M.R. & Brown, M.J. (Eds): *CLIMATIC SUCCESSION AND GLACIAL HISTORY OF THE SOUTHERN HEMISPHERE OVER THE LAST FIVE MILLION YEARS*. *Pap. Proc. R. Soc. Tasm.* 130(2): 55-64. <https://doi.org/10.26749/rstpp.130.2.55> ISSN 0080-4703. Department of Geology, Hamilton College, Clinton, New York, USA 13323.

In light of recent warming and environmental changes observed on the Antarctic Peninsula, an increased knowledge of regional palaeoclimatic trends may provide an improved understanding of the expected response of the Antarctic glacial, oceanic and biotic systems to continued warming. Sedimentologic and geochemical analyses of a 5.5 m long, high-resolution sediment core (PD92 GC-1), collected in Lallemand Fjord, represent the most detailed record of Holocene climate change, to date, in Antarctica. Grain size, smear slide analysis, magnetic susceptibility and total organic carbon content (TOC) were measured. One radiocarbon date establishes a chronology for the base of the core. Correlation with the upper portion of core GC-1 with other cores collected in the fjord is based upon carbon stratigraphy, nine radiocarbon analyses and ^{210}Pb data. Deglaciation of Lallemand Fjord is believed to have occurred prior to 8000 yr BP, followed by a period of open marine conditions with variable extent of sea ice (variable TOC content) between 8000 and 2700 ^{14}C yr BP. A climatic optimum is recognised between 4200 and 2700 yr BP. Around 2700 yr BP, a decrease in TOC and diatom abundance reflects the formation of more extensive and seasonally persistent sea (fast) ice. The Müller Ice Shelf, now present in the fjord, advanced approximately 400 years ago, coincident with the Little Ice Age. These results indicate environmental variability throughout the Holocene that was consistent across most portions of the maritime Antarctic Peninsula. Surprisingly, the timing of climate transitions correlates with Northern Hemisphere "T-Events" and ice-core data from Greenland, indicating the possibility of coherent climate variability in the Holocene, at least for the high latitudes.

Key Words: Holocene, palaeoclimate, Lallemand Fjord, Antarctic, glacial marine.

INTRODUCTION

Until recently, little documented evidence existed of a Hypsithermal event in the Antarctic region (Domack *et al.* 1991, Domack 1995). This lack of recognisable evidence has been attributed to the absence of outcrop exposures suitable for study, yet the recent analyses of ice and marine sediment cores have suggested such an event did occur (Domack 1995). In light of the environmental changes believed to be influenced by present atmospheric warming in the Antarctic Peninsula region (AP; Jones *et al.* 1993), the identification and documentation of a Hypsithermal event in Antarctica could provide researchers with an improved understanding of the expected response of the Antarctic glacial, oceanic, and biotic systems to continued regional warming (Leventer *et al.*, in press).

MARINE SEDIMENT RECORD

Evidence from marine sediments suggests that a middle Holocene Hypsithermal event exists in East Antarctica, as well as in offshore regions of the AP (Domack *et al.* 1991, Pudsey *et al.* 1994, Leventer *et al.*, in press). Sediments from three deep troughs located off the East Antarctic coast (the Mertz/ Ninnis Trough, the Dumont d'Urville Trough and the Amery Depression) provide a record of a "synchronous mid Holocene response of regional outlet glaciers" (Domack *et al.* 1991). A major facies change from glacier proximal to open marine conditions has been recognised and dated to between 4300 and 3800 yr BP (Domack *et al.* 1991). (All

ages are expressed as corrected radiocarbon years BP and include unproven assumptions regarding the temporal variability of the Antarctic reservoir effect.) The late Holocene dates obtained for this environmental change correspond in time with the decline of the Northern Hemisphere Hypsithermal event and the onset of the Neoglacial (Pielou 1991, Dwyer *et al.* 1996).

Studies of the sedimentary record of the AP suggest a pattern of sedimentation believed to be indicative of climate change following the Hypsithermal. On the AP shelf, marine sediment records typically indicate the response of ice-shelf and ice-sheet systems to climate change during the late Pleistocene to mid Holocene. Cores collected from the continental shelf, west of Anvers Island, contain a decrease in the coarse sediment fraction above 2 m depth, believed to represent a facies shift from glacial sedimentation to open marine sedimentation (Pudsey *et al.* 1994). Radiocarbon dates of material collected near the facies transition indicate an age of 13 000 to 12 500 years for this facies change. A peak in biogenic sedimentation, as evident from diatom analysis, may have occurred from 6400 to 3000 yr BP (Pudsey *et al.* 1994). Hence, open marine conditions prevailed on the AP continental shelf during the mid Holocene (Pudsey *et al.* 1994, Leventer *et al.*, in press).

Marine records from the Palmer Deep, inshore of the region studied by Pudsey *et al.*, contain a transition occurring at 6 m in core PD92-30 (as described by Leventer *et al.*, in press). An age of approximately 2700 yr BP has been proposed for the transition based upon calculated sedimentation rates in the region. Below 6 m, an increase

in productivity has also been noted indicating more open marine (warmer) conditions prior to 2700 yr BP. Assuming the estimated age of the transition is accurate, the authors believe that the decline of productivity above 6 m represents the transition from the warmer climates of Hypsithermal, to those of the Neoglacial (Leventer *et al.*, in press).

In addition, trends of increasing productivity with depth have been documented in the sedimentation records from fjord environments on the AP. Sediments in Lallemand Fjord (LF; figs 1, 2) have shown maximum total organic carbon (TOC) values, believed to be indicative of higher regional productivity, at depth (Stein 1992, Domack *et al.* 1995, Domack & McClennen 1996). The TOC maximum has been estimated to have occurred before 2000 yr BP (a minimum estimate; Stein 1992, Domack *et al.* 1995). The trends in first order TOC content of three cores (PD90 KC 72, PD90 KC 75 and PD92 KC 5), collected from the glacier proximal to distal environments in LF, exhibit an increase with depth. However, these cores are not long enough to determine if an absolute Holocene TOC maximum exists. PD92-2 GC-1 is a longer core, that may provide an extended record of biogenic sedimentation in the region and may lead to the documentation and resolution of a TOC maximum (climate optimum) for the Holocene in LF.

SETTING

LF is a large embayment located on the west coast of the AP at 67°S, 66°W (figs 1, 2). It is unusual in that it contains the northernmost ice shelf on the western coast of the AP and is, therefore, today a polar fjord. The persistence of the Müller Ice Shelf, despite the changing regional climate, is believed to be directly attributed to the protection offered by the fjord. Without pinning on an ice rise, the ice shelf would most likely disintegrate, as have the Wordie and parts of the George VI ice shelves found further south (figs 1, 2); (Domack *et al.* 1995, Stein 1992, Doake & Vaughan 1991, Vaughan & Doake 1996). Seasonal sea-ice coverage has a great deal of interannual variability. In most years, the fast ice does not break out of LF until late in the summer season.

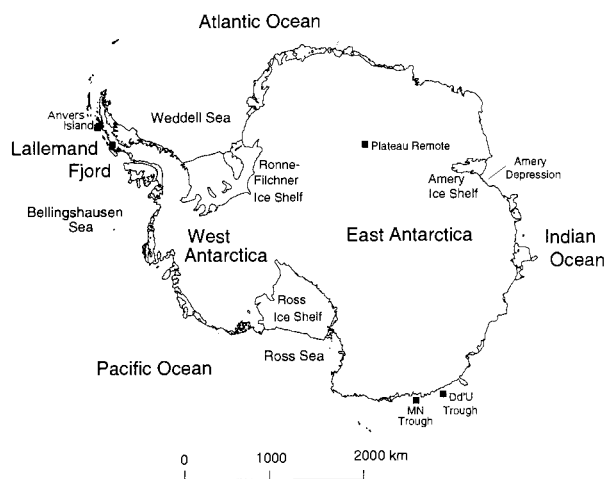


FIG. 1 — Location of Lallemand Fjord in Antarctica along with other features mentioned in the text (MN = Mertz/Ninnis Trough, Dd'U = Dumont d'Urville Trough).

During unusual years, the fjord may be free of sea ice as early as December. The persistent sea ice is an important factor, since it limits the amount of light available to phytoplankton during the spring and summer. Hence, the protected environment of LF provides an optimum environment for studying regional palaeoclimatic trends during the Holocene, as they may have influenced sea-ice and ice-shelf extent.

METHODS

Gravity Core 1 (GC-1) was obtained by the science party of project S-285 in LF (fig. 2) during cruise 92-2 of the RV *Polar Duke* (19 March–15 April 1992). The 5.5 m long by 70 mm diameter core was collected, using a modified piston core apparatus in a water depth of 630 m. The removal of the trigger and piston device from the core apparatus allowed for the collection of long gravity cores. It is believed that the sediment–water interface was recovered for GC-1; however, detailed physical property measurements, such as water content, were not conducted.

Physical Properties

Sediment samples, taken in 1995 from GC-1 at 0.1 m intervals, were analysed with the Hamilton College Malvern Mastersizer E particle-size analyser (using laser diffraction) to obtain volume per cent measurements of fine/medium sand, coarse silt, med/fine silt and clay. Samples of sediment were placed in a ultrasonic Calgon bath and completely suspended for five minutes to disperse clay floccules prior to analysis. The samples were then suspended in an ultrasonic

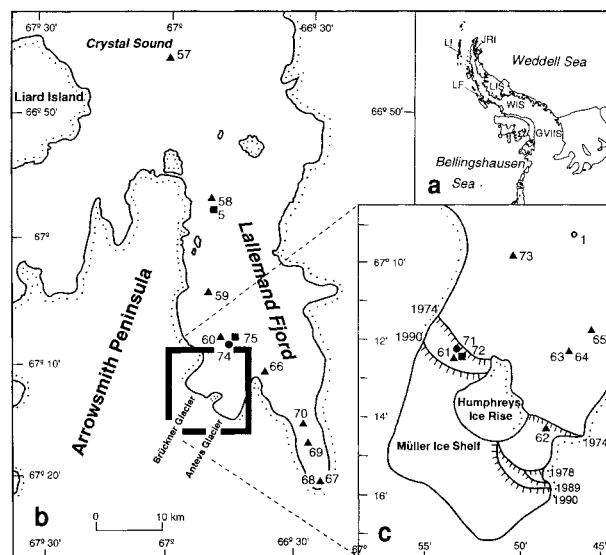


FIG. 2 — (A) General map of the Antarctic Peninsula showing the location of Lallemand Fjord = LF, Larsen Ice Shelf = LIS, Wordie Ice Shelf = WIS, LI = Livingston Island, JRI = James Ross Island, and George VI Ice Shelf = GVIIS. (B) Detail of Lallemand Fjord, Antarctica, showing sediment core and surface grab sampling stations: ▲ = surface grabs, ● = piston cores, ■ = kasten cores, and ○ = gravity core. (C) Detail of region proximal to the Müller Ice Shelf with location of sampling stations and calving line positions since 1974 (modified after Domack *et al.* 1995).

bath with a mechanical stirrer and analysed to determine the volume percentages of specified grain sizes. (For complete procedure see LoPiccolo 1996).

Weight per cent sand analyses were also done on GC-1 in the autumn of 1995. Samples of 3–5 g of sediment were placed in 50 ml beakers and dried for 24 h in an oven at 40°C. Upon removal from the oven, samples were cooled and weighed to obtain a total dry weight for the sample. The sample was then covered with 50 ml of a Calgon solution and stirred to suspend the sediment and encourage the disintegration of any floccules. After 24 h, each sample was wet sieved through a 63 µm sieve. Sediment which did not pass through the sieve was dried for another 24 h and reweighed. The weight obtained represented the weight per cent sand for the sample.

The Malvern is unable to analyse particles greater than 600 µm in diameter; therefore, a gravel grain count was conducted for the particle fraction in excess of 2 mm. Data were obtained, using X-ray radiographs of the archive half of GC-1 taken in 1994. The X-rays (pl. 1) were analysed at 50 mm intervals down-core, and particles greater than 2 mm in diameter were counted within each 50 mm interval, to obtain information regarding the concentration (grains/50 mm core length) of the coarsest grain-size fraction of the core (after Grobe 1987).

Whole core magnetic susceptibility for GC-1 was measured at 50 mm intervals, using the Barrington MS-2B Core Sensor. Magnetic Susceptibility (MS) values are expressed as $\times 10^{-6}$ CGS.

Composition

Smear slides were taken at seven intervals and were point-counted for 300 particles, first for all major constituents and then again for the ratio of pennate to centric diatoms out of 300 diatoms frustules counted. Counting groups included mineral grains, centric diatoms, pennate diatoms, *Chaetoceros* resting spores and diatom fragments.

Geochemical Analyses

During the spring and summer of 1995, 55 sediment samples from GC-1 were analysed at Hamilton College for TOC content, using the LECO induction furnace with attached LECO WR-12 carbon determinator. Each sample was placed in a separate 100 ml glass beaker and completely mixed with about 25 ml of 2 N hydrochloric acid to dissolve any calcium carbonate. After 24 h, the acid was carefully decanted with a syringe, using special caution not to interrupt the sediment–acid interface. Each sample underwent a series of de-ionised water washings, during which 10 ml of de-ionised water was placed in each beaker and the sediment was totally suspended. The samples were allowed to settle for 24 h and then the supernatant liquid was decanted so as not to disturb the sediment interface. The washing procedure was repeated five times. After the last decantation, the samples were placed in an oven at 106°C, for 24 h, to remove moisture. Calcium carbonate contents are very low for these samples (less than 1%) so corrections for CaCl and hydration were not conducted during the analyses. Each dry sample was removed from its beaker and gently ground using a mortar and pestle. Between 0.4 and 0.6 g of dry sediment was weighed and placed in a ceramic crucible in

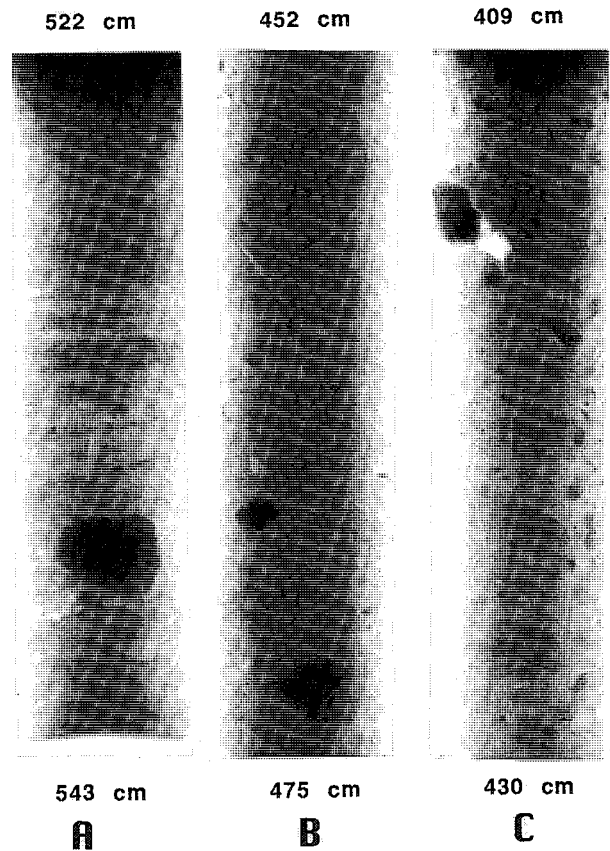


PLATE 1

X-ray radiographs of three intervals of core GC 1, illustrating (A) laminated facies of the lower portion of the core, (B) the scaphopod sample from which a radiocarbon date was obtained, and (C) typical pebbly mud interval. In B, the obliquely inclined angle of the shell and the presence of the light grey feeding burrow visible at the anterior of the shell strongly suggest *in situ* preservation.

preparation for burning. The LECO furnace is designed to minimise sources of error; however, drift may occur during a series of burns. Corrections for drift are made before, during, and after the samples have been burned. Standards were burned, as described by Franceschini (1995), and these results were corrected to account for instrument drift. In all cases, drift was observed to be linear versus time. Accuracy was determined to be $\pm 0.005\%$ TOC

Radiocarbon Dating

A scaphopod (*Dentalium*) shell sample was chosen for radiocarbon analysis, based upon evidence suggesting *in situ* preservation (pl. 1B). An undisturbed sample is essential to obtain an accurate radiocarbon age for the associated sediment interval depth. Hence, the determination of *in situ* preservation is necessary to rule out any reworking of the sediment and/or shell, which would produce a false radiocarbon date.

The scaphopod class (of Mollusca) is characterised by conical (tusk-shaped) shells, open at both ends. Members of the class are benthic selective feeders and dwell in the upper 0.1 m of the marine sediment. They burrow into the sediment at an oblique angle with the aid of a muscular

foot, usually until only the uppermost portion of the shell protrudes at the sediment–water interface (pl. 1B; Gainey 1972). Upon completion of burrowing, the animal constructs a feeding cavity by probing and packing the surrounding sediment with this foot. The action results in a conical cavity, slightly larger than the opening for the foot, at the anterior end of the animal (pl. 1B). The scaphopod brings foraminifera into this cavity from the surrounding sediment and ingests them with its proboscis (Gainey 1972). Although the feeding mechanisms of the class are variable, the action of burrowing and the construction of some type of feeding cavity are universal for the class (P. Reynolds, pers. comm.).

Examination of the X-ray radiograph from 4.52–4.75 m (pl. 1B) reveals several indications of *in situ* preservation. The 40 mm scaphopod shell is oriented in an obliquely inclined position, typical of a scaphopod in feeding position. In addition, a 10 mm long, light grey, conical region exists at the anterior end of the shell (pl. 1B). The orientation of the scaphopod shell, a lack of coarse sediment infill within the shell, coupled with the presence of a feeding burrow at the anterior end of the shell (interpreted by P. Reynolds, pers. comm.), strongly suggest *in situ* preservation of the sample. The possibility of the three factors existing in reworked sediment is “highly unlikely” (P. Reynolds, pers. comm.).

The scaphopod was removed from a depth of 4.56–4.60 m and sent to Oxford University’s tandem accelerator mass spectrometer facility, where a radiocarbon analysis was conducted via GEOCHRON Laboratories (GX-20530-AMS). An uncorrected radiocarbon age of 9358 ± 70 yr BP was obtained, using the Libby half life (5570 years). The age was corrected for a $\delta^{13}\text{C}$ value of $+0.5\text{‰}$. The interpretation of radiocarbon dates from the Antarctic is complex, due to unusually low ^{14}C concentrations in Antarctic waters, as well as the geographical variation of water mass circulation (Gordon & Harkness 1992). Living organic material has been found to exhibit anomalously “old” ^{14}C dates, a phenomena often referred to as the Antarctic Reservoir Effect. Therefore, measured radiocarbon dates must be corrected to account for the elevated ages (low ^{14}C activity) seen in recent materials, in order for accurate down core ages to be obtained (Gordon & Harkness 1992). The reservoir correction has an estimated value of 1300–1250 yr BP for the AP (as suggested by Gordon & Harkness 1992, Domack 1992).

It should be noted that biogenic material (calcite) from near-surface sediments in LF has been dated to 1900 yr BP, indicating that the local reservoir effect in this fjord may be much older than the 1300 yr BP correction. However, due to the absence of a radiocarbon date for living organic material in LF, a standard value of 1300 years has been used as the reservoir correction for core GC-1. Using a corrected date of 8058 yr BP for the scaphopod shell, the projected sedimentation rate for core GC-1 is 0.57 mm/yr. This is roughly half of the sedimentation rate documented for cores KC 72 and 75 (1.2 and 1.3 mm/yr respectively). However, it should be noted that a single radiocarbon date has been used in the extrapolation (rate determination) in GC-1. The use of a single date assumes that the sedimentation rate was constant at this site throughout the period represented by the core.

RESULTS

Core GC-1 was described as a grey to grey-olive green silty-pebbly mud with scattered ice-rafted debris in the upper 5.15 m; and a laminated grey silty mud and medium to fine-grain sand from 5.15 to 5.5 m (fig. 3, pl. 1A).

The depositional record, graphically represented by plotting each of the parameters versus core depth, extends over a 9700-year period (fig. 3). The distribution of each parameter can be used to evaluate environmental variability over time. Total organic carbon and magnetic susceptibility values relate to sediment type, recording biogenic and terrigenous (respectively) input to the fjord system (Domack & Ishman 1992). Logically, the values of the two parameters should be inversely related; that is, high rates of biogenic sedimentation exist with relatively low magnetic susceptibility values (Domack & Ishman 1992). Volume per cent grain-size data and gravel grain counts represent a distribution of particle sizes present in the core (fig. 3).

Gravity Core 1 was obtained from a position roughly 8 km from the calving front of the Müller Ice Shelf (fig. 2). Generally, GC-1 contains relatively high magnetic susceptibility values ($> 340 \times 10^{-6}$ CGS), low percentages of organic carbon content ($< 0.4\%$), and almost continuous, coarse, ice-rafted sediment (averaging 10–15 grains/50 mm interval; fig. 3).

Grain Size

GC-1 consists of 75 to 90% clay and fine–medium silt-sized particles, based upon the volume per cent data. These percentages are slightly higher than the percentage of fine grains found in KC 72 and lower than those in KC 75 (Domack *et al.* 1995) and reflect the position of the cores within the fjord; coarser sediments are observed closer to the ice shelf (KC 72) and finer sediments are characteristic of more open fjord settings (KC 75; Stein 1992, Domack *et al.* 1995). Core GC-1 is located directly between KC 72 and KC 75. The silt and clay fraction of GC-1 remains relatively constant ($> 75\%$) until 5 m and then decreases to approximately 60% fines. The coarse silt and sand fraction represents 10 to 15% of the particle distribution from 0 to 5.5 m. High percentages (20–25%) of coarse silt and sand were measured at 1.9 m and 5.0–5.5 m; while low percentages ($< 10\%$) were measured at 0.5 m, 2.9 m, 3.7 m and 4.8 m. Volume per cent sand and weight per cent sand exhibit slight variations with depth, as expected, though excellent agreement in the two methods exists between 5.5 and 5.0 m (Shevenell 1996).

Ice-rafted debris (IRD) abundance, indicated by the number of gravel grains (> 2 mm diameter) per 50 mm interval, is variable, yet continuous throughout the core (fig. 3). There is an increasing trend from the top of the core (< 10 grains/50 mm interval) to 4.15 m, where a peak (58 grains/50 mm interval in gravel abundance) is present. X-ray radiographs of GC-1 show a concentration of large, angular gravel grains at a depth of 4.15 m (pl. 1C). From 4.2 m to the base of the core, the concentration of gravel decreases. It is interesting to note that, where relatively large numbers of gravel grains exist, extremely low percentages of sorted sand are present. This trend is most visible from 3 m to the base of the core (fig. 3). Domack *et al.* (1995) noted a similar pattern and interpreted the sediments with the most gravel clasts as having been deposited distal to the Müller Ice Shelf.

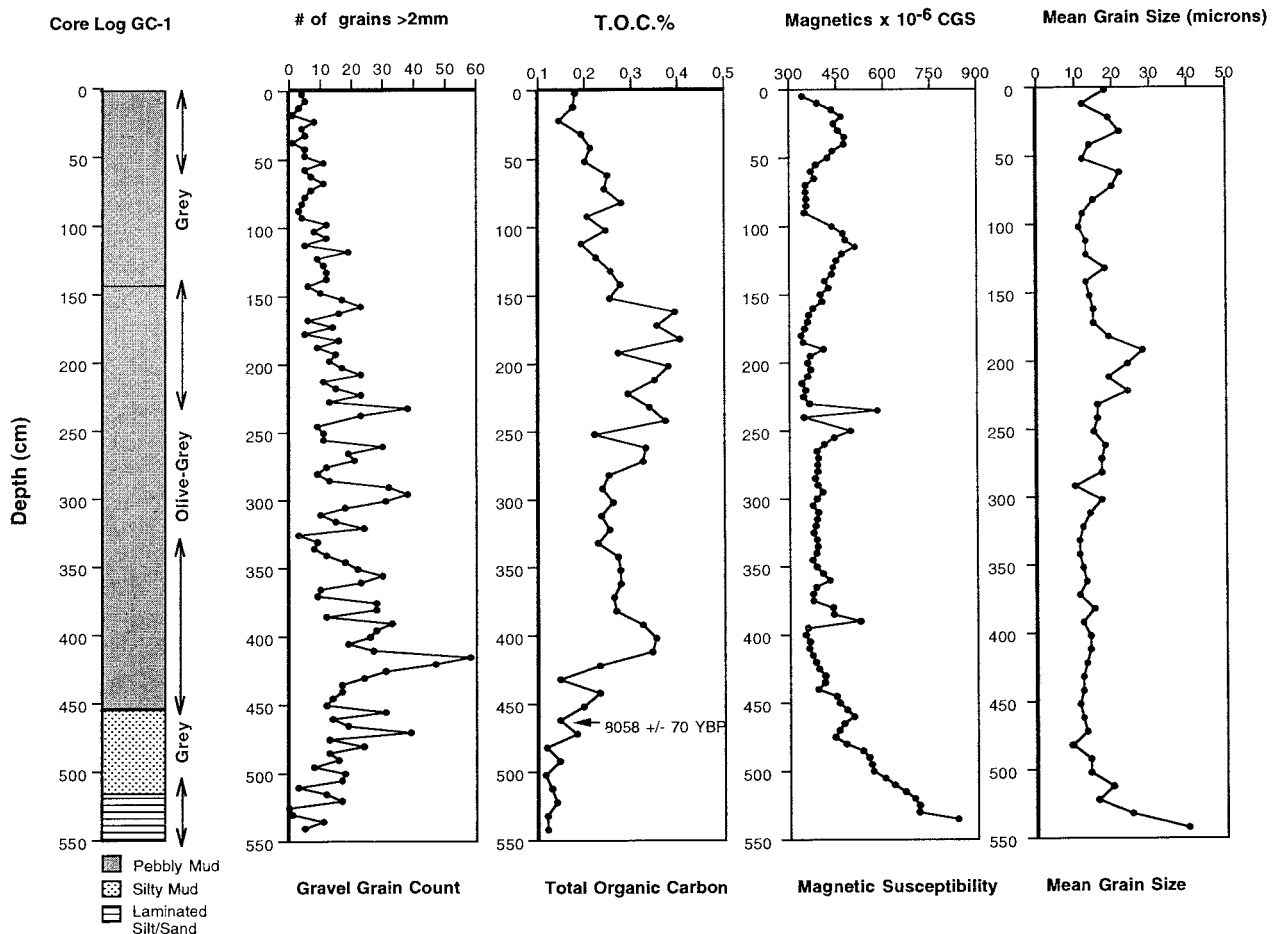


FIG. 3 — A visual core log, gravel grain count, TOC %, magnetic susceptibility and mean grain size versus depth for samples taken from Gravity Core 1.

Total Organic Carbon

The TOC contents in AP cores are useful for palaeoenvironmental information, because of the lack of terrestrial sources of plant debris and reworked organic particulates. This means that the TOC content in Antarctic marine sediments can be more closely related to productivity in the marine realm as opposed to the Arctic. TOC content in modern sediments of LF is low (<0.4%) and reflects the polar climate and persistent sea-ice coverage (Domack *et al.* 1995). The content in GC-1 is generally lower than that in KC 5 and KC 75, and higher than that in KC 72 (fig. 4). Again, the variations in sediment type directly reflect the geographic variation in sedimentation in the fjord with respect to the glacial margin (fig. 4).

The TOC content for GC-1 generally increases to a depth of 1.6 m, although minima were noted at 0.3 m and 1.2 m. At 0.3 m, the content is comparable to minima found in KC 72 and KC 75, which have been interpreted as Little Ice Age intervals (fig. 4; Stein 1992, Domack *et al.* 1995). At 1.6 m, it increases dramatically (0.4%); high TOC content (0.25 to 0.42%) exists in core GC-1 from 1.6 m to 2.6 m. Within this interval, a core maximum of 0.42% was measured at 1.8 m. From 1.8 m to 2.6 m, peak percentages decrease slightly. Within this unit, minima exist at 1.9 m (0.26%), 2.3 m (0.28%) and 2.5 m (0.21%); however, the relatively low values in this unit are high with respect to the TOC content for the entire core.

Following the period of high TOC content, it drops to an average value for the core and remains relatively stable (0.25–0.30%) from 2.8 m to 3.8 m (fig. 5). A higher content was measured around 4.0 m with values (0.33–0.36%) comparable to those within the period of peak biogenic sedimentation for the region. At 4.1 m it decreases, and this general trend continues to the base of the core; the content of biogenic material in the lowest unit of the core appears to be minimal.

No notable correlation was observed between TOC content and medium to fine silt indicating that the preserved carbon signal has not been masked by changes in the fine-grained terrigenous input (Shevenell 1996).

Magnetic Susceptibility

An interesting pattern of MS exists in GC-1. Extremely high values (340 to 850×10^{-6} CGS) indicate the importance of terrigenous input, and a large magnetite component to the overall pattern of sediment flux in LF. Recent magnetic values are low. Values decrease from 0.35 to 0.95 m and then increase to a peak at 1.15 m, decrease to a minimum (340×10^{-6} CGS) at 1.8 m and then remain relatively stable (400×10^{-6} CGS) until 4.0 m, where they begin to increase to a maximum value for the core at 5.35 m (850×10^{-6} CGS). Peaks exist at 1.9 m (420×10^{-6} CGS), 2.4 m (630×10^{-6} CGS), and yet another at 3.9 m (530×10^{-6} CGS) (fig. 3).

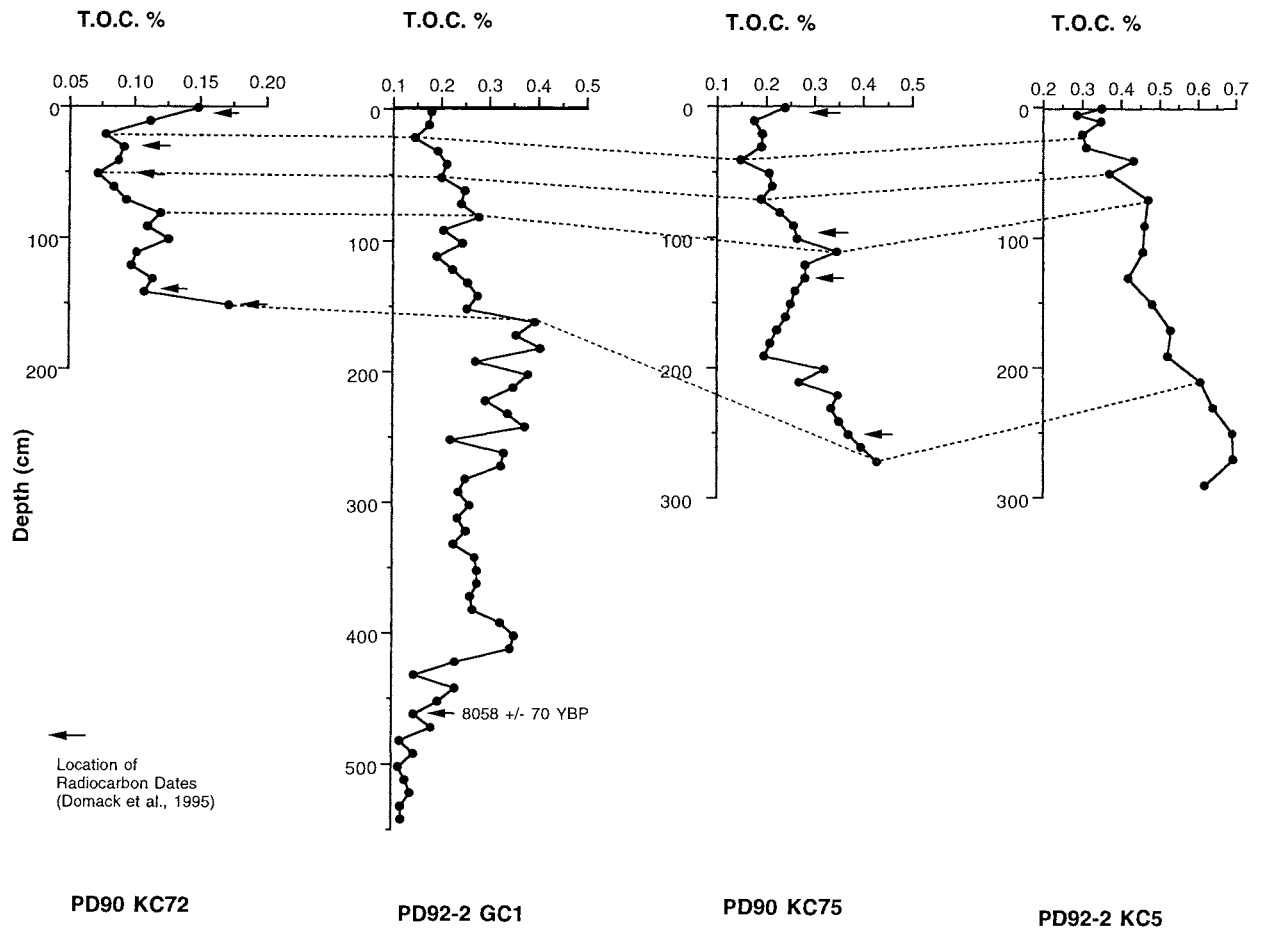


FIG. 4 — TOC % versus depth for samples taken from kasten cores 72, 75 and 5 and Gravity Core 1. Dashed lines connect time synchronous depositional horizons based upon the preserved carbon signal.

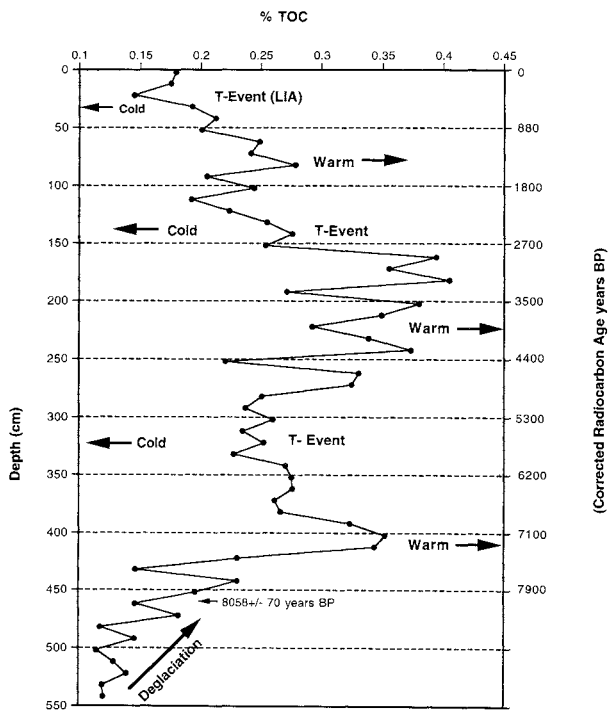


FIG. 5 — TOC % versus depth for samples taken from Gravity Core 1, with a corrected radiocarbon chronology (yr BP) derived from the extrapolated sedimentation rate. Global T-events and warm climatic events are indicated.

MS and volume per cent sand data do not seem to be related, except from 4 m to the base of the core, where the increasing values seem to be directly related. The coarse silt-size fraction is likely to be the carrier of the MS signal. Peaks in MS at 2.4 m and 3.9 m directly relate to peaks in TOC at the same depths.

Smear Slide Analysis

Further support for conclusions drawn from the MS and TOC data is provided by smear slide analysis performed for the percentages of *Chaetoceros* resting spores, pennate diatoms, centric diatoms, diatom fragments and mineral grain components (fig. 6A, B). The smear slide data demonstrate significant variability in particle composition, in support of the observed variation in TOC content (fig. 5). Mineral grains dominate the samples in the lower and the upper portions of the core, while the middle portions of the core are dominated by diatom particles. Breakdown of the biogenic fraction reveals that the *Chaetoceros* resting spores are overly abundant in the high carbon intervals. In contrast, diatom fragments are abundant, where mineral grains are also abundant and where the TOC content is low.

MS lows are characterised by higher than normal abundances of *Chaetoceros* resting spores (Leventer *et al.*, in press). The minimum magnetic value (340×10^{-6} CGS) at 1.8 m corresponds to the maximum percentage of

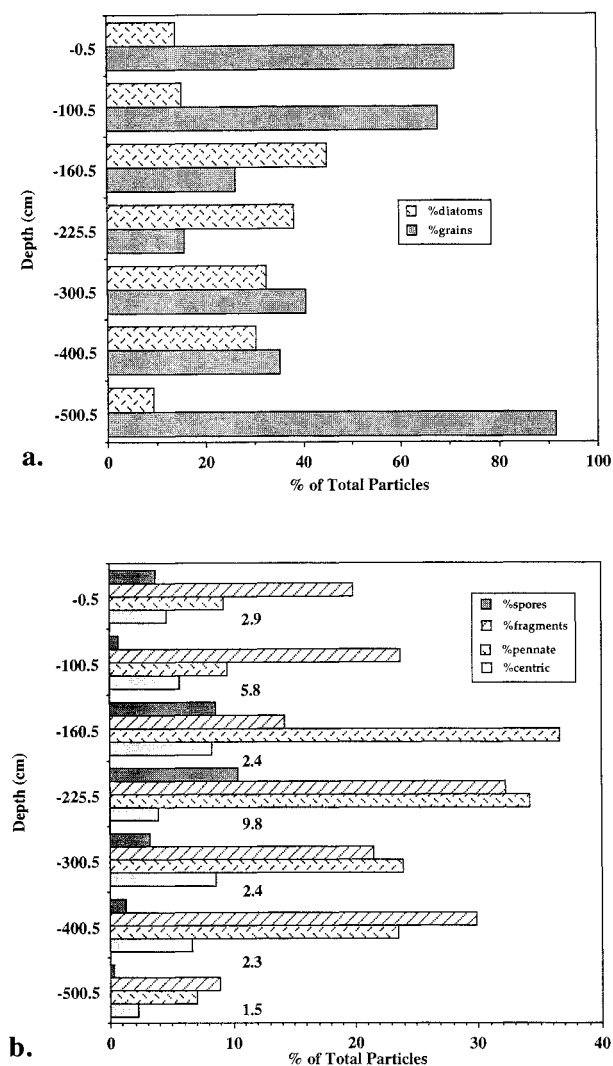


FIG. 6 — Smear slide analysis (> 300 particles counted) of selected depth horizons (— cm below top of core) in core GC-1, illustrating (A) mineral grains versus diatom particles; (B) *Chaetoceros* resting spores, diatom fragments, pennate diatoms, and centric diatoms. The ratio of pennate versus centric diatoms is also indicated for each sample interval in B and is based upon a separate 300 particle count of only whole frustules.

Chaetoceros resting spore per sample (fig. 6B). In the fossil record, high concentrations of *Chaetoceros* resting spores have been interpreted as indicating very high primary production (Donegan & Schrader 1972, Leventer 1991, Stockwell 1991, Grimm & Gill 1994). The results of the GC-1 smear slide analysis are reflective of this hypothesis (fig. 6B). The percentages of diatoms and *Chaetoceros* resting spores are at their maxima between 1.6 and 2.26 m. At this depth interval, the TOC levels are at their peak and the MS level is at its minimum. Between 1.6 and 2.26 m the diatoms are at their maximum concentration (40 to 50%) and the grains are at their minimum concentration (18–28%). The increase of diatom preservation from 5.01 to 1.6 m (fig 6A) is reflected in the increase of TOC levels for that same period. At 1.5 m, the diatom content precipitously decreases, about 2700 yr BP.

The high ratio of pennate to centric diatoms in the high TOC interval (about 4000 yr BP; figs 5, 6B) is consistent

with the higher pennate concentrations found in sea ice (Clarke *et al.* 1984), thus suggesting blooms are related to sea-ice meltwater layers (sea-ice marginal zone). In contrast, the TOC maximum interval (at about 3000 yr BP) is marked by a low ratio of pennate to centric diatoms but elevated concentrations of *Chaetoceros* spores (figs 5, 6B). This suggests that production during the TOC maximum interval, while elevated, was perhaps related to open water rather than sea-ice marginal processes.

The diatom fragment concentration is significantly higher during periods of low TOC (fig. 6B). Between 0 and 1.0 m and at 4.0 m, the fragment concentration exceeds 20% and dominates the biogenic fraction. During high to average TOC levels, the fragment concentration is generally lower and is always less than that of the whole pennate forms. The low TOC values suggest extensive sea ice during these depositional periods and infringement of glaciers along the periphery of LF, consistent with reworking of diatoms (fragmentation). This reworking dilutes the deposition of pelagic productivity in the system, which is suppressed by a persistent sea-ice regime.

Carbon Chronology

The chronology for GC-1 is based upon radiocarbon analysis of an *in situ* scaphopod sample, collected from a depth of 4.57 m from which a sedimentation rate of 0.57 mm/yr was obtained, using a reservoir correction of 1300 years. The rate was extrapolated to the surface. Assuming the core top represents the sediment–water interface, the core has a sedimentation rate of roughly half of those obtained by Stein (1992) and Domack *et al.* (1995) for cores KC 72, KC 75 and KC 5 (1.2–1.3 mm/yr). The discrepancy in sedimentation rates must be addressed; based upon the high resolution of GC-1, a sedimentation rate of 0.57 mm/yr seems quite accurate, and excellent correlations of the TOC signal can be made between cores 72, 1 and 75 (fig. 4).

Two important assumptions have been made and cannot be ignored with regard to the chronology obtained for GC-1. First, the sedimentation rate has been assumed to be constant for the entire depositional history of GC-1. Second, the initial rate has been calculated using a reservoir correction of 1300 years (as proposed for the AP). Due to the unavailability of a radiocarbon date for living organic material in the fjord, however, an older reservoir correction for LF has been proposed, based upon radiocarbon dating of near surface sediment (Domack *et al.* 1995). For the purposes of this paper, the chronology is assumed to be accurate, until evidence is found which either provides a definitive reservoir correction for LF or revises the constant sedimentation rate for the core.

DESCRIPTION AND INTERPRETATION OF GC-1 SEDIMENTARY UNITS

Three distinct palaeoenvironments may be inferred for LF throughout the Holocene. A number of parameters exhibit strong correlations and represent distinct palaeoenvironments in the fjord. A division of the core, based upon similar sedimentologic trends, yields three distinct units; the lower (5.5–4.55 m), middle (4.55–1.45 m), and the upper (1.45–0 m) units reflect ice (glacier) proximal, open marine and ice-shelf influenced depositional environments (fig. 3).

The Lower Unit

The lower unit (5.5–4.55 m) of GC-1 is a laminated grey silty mud and medium to fine-grained sand unit (5.5–5.15 m) overlain by a structureless grey silty mud (5.38–4.55 m). TOC content and gravel frequency generally increase up core and are inversely related to the trends of magnetic susceptibility and mean grain size (fig. 3).

The fining upward sequence represents a gradual shift from an ice (glacier) proximal to open marine depositional environment, indicative of fjord deglaciation. Laminated terrigenous silts and medium to fine-grained siliclastic sands typify proximal glaciomarine sediments in subpolar fjord settings, due to terrigenous sedimentation from meltwater influx. However, laminated silts and sands are a minor component of polar fjords, because of low meltwater input from the glacial system. Therefore, as deglaciation occurred in LF during the early to mid Holocene (prior to 8000 yr BP), meltwater input to the fjord system became more significant, resulting in coarse, laminated ice proximal deposits (pl. 1A).

As the glacial system receded towards the fjord heads, the percentage of coarse silt and sand decreased, due to settling of coarse material at the ice edge. Sedimentation at this time was gradually dominated by the settling of medium to fine-grained terrigenous silts, biogenic material and abundant iceberg rafting, a regime indicative of more open marine conditions (fig. 3). Increased preservation of the biogenic component (TOC content) is associated with the deglaciation of the fjord, as represented by the inverse relationship between magnetic susceptibility and TOC content trends in the upper section of the unit. Although TOC content increases slightly in the unit, the trend probably reflects a combination of the decreased dilution of biogenic material in the fjord system. Based upon the ^{14}C chronology obtained for LF, deglaciation neared completion by 8000 yr BP, as ice proximal sedimentation gave way to distal open marine (fjord) deposition.

The Holocene commenced with a period of warm temperatures, which has been documented in ice-core records from the Antarctic (Cias *et al.* 1991) and has most recently been reflected in ice-core data from Greenland (O'Brien *et al.* 1995). The lower unit reflects the deglaciation of LF (fig. 5) during the warm early Holocene (prior to 8000 yr BP) and correlates with a climatic warm period observed in ice core records from Greenland (O'Brien *et al.* 1995). The distinct sedimentation record present in the lower unit of GC-1 is characteristically different from that of the modern Müller Ice Shelf system, indicating that sedimentation was influenced by a different type of glacial system. The deglaciation observed in LF in the early Holocene could represent the continued retreat of the grounding line of the AP Ice Sheet. In contrast to Pudsey *et al.* (1994), the higher resolution record of GC-1 places the deglaciation of LF and the inner continental shelf prior to 8000 yr BP.

The Middle Unit

The middle unit (4.55–1.45 m) of GC-1 is a grey-olive green to light olive grey silty mud, with scattered IRD (fig. 4). TOC content is generally high (0.2–0.4 %), with maxima between 4.1 and 3.8 m, and between 2.7 and 1.6 m. The core maximum was measured at 1.8 m. Low magnetic

susceptibility ($350\text{--}400 \times 10^{-6}$ CGS) and mean grain size (12 μm) remain consistent throughout the unit, with the exception of two maxima. The IRD fraction generally decreases throughout this unit (fig. 3).

As the glacial system receded from LF, open marine (ice distal) conditions prevailed between 7980 and 2540 yr BP (fig. 5). Higher percentages of biogenic material and lower percentages of terrigenous input are typical of open marine depositional environments (fig. 6). Generally, increased incident light penetration, due to the absence of glacial ice, encourages plankton photosynthesis and, thus, higher levels of primary productivity in the stable, meltwater-induced, low salinity surface layer of the water column. Likewise, as the glacial front retreats, the terrigenous input decreases.

Given that the magnetic susceptibility and mean grain size remain relatively consistent, fluctuations in the TOC content within the middle unit probably represent variations in productivity, governed by sea-ice extent and duration in LF during the middle Holocene (fig. 4). Assuming that higher productivity results from an increased amount of incident light penetration, lower TOC content in the open marine environment may reflect extensive seasonal ice coverage, resulting in decreased productivity, while a higher content may reflect plankton blooms associated with minimal sea-ice coverage and increased meltwater input. IRD analysis further corroborates this interpretation, as the TOC content minima have generally higher amounts of IRD in the sediment (fig. 3). We also observe an abundance of *Chaetoceros* resting spores in this interval, consistent with elevated productivity (Leventer *et al.*, in press).

Based upon the ^{14}C chronology for GC-1, a period of open marine deposition occurred between 7980 and 2540 yr BP; sea-ice minima (defined by high TOC content) occurred in LF between 7200 and 6700 yr BP and between 4700 and 2700 yr BP. Minima in sea-ice extent correspond with climatic intervals found in the GISP2 ice core from Greenland (O'Brien *et al.* 1995). For instance, the period of low TOC content between 3.8 and 2.8 m (6700–4900 yr BP) is interpreted as a period of extensive sea-ice cover in the fjord and corresponds to a cold "T-Event" most recently documented in ice-core records from Greenland (6200–5000 yr BP; O'Brien *et al.* 1995). The higher content in the middle unit reflects the increased abundance and preservation of biogenic material indicative of an open fjord environment (8000–2700 yr BP).

The Upper Unit

The upper unit of GC-1 (0–1.45 m) is a light olive grey to grey silty mud, with scattered IRD (fig. 4). TOC content and IRD fraction decrease up core in this unit. Magnetic susceptibility and mean grain size are variable and generally higher than observed in the middle unit. Magnetic susceptibility values decrease from 0–0.35 m as does mean grain size (0.1–0.3 m; fig. 3).

The upper unit of GC-1 displays similar characteristics to KC 72 and KC 75, interpreted by Stein (1992) and Domack *et al.* (1995) as an ice-shelf influenced depositional environment (fig. 4). A decreasing trend of TOC content in the lower section of the unit reflects the changing depositional environment, from that of an open marine environment with minimal sea-ice coverage to that typical of a colder (fast ice or ice shelf) influenced environment.

Hence, some climatic and/or oceanographic event induced cooling and the formation and slow advance of the Müller Ice Shelf, culminating at around 400 yr BP (the Little Ice Age).

The siliciclastic content increases as biogenic components decrease with encroachment of the ice-shelf system (fig 6). In addition, increases in mean grain size and coarse silt and sand directly correspond to peak magnetic susceptibility value, indicative of an ice proximal setting. Input of coarse-grained sediment (sand) is probably aeolian in origin, due to the debris-poor nature of ice-shelf systems; for a more complete discussion of aeolian sedimentation, see Stein (1992) and Domack *et al.* (1995). It is interesting to note the decreased magnitude of the terrigenous sediment input with the advance of the ice shelf as compared to the early Holocene deglaciation of LF (figs 4, 6). The sedimentology of the upper unit reflects a typical ice-shelf depositional system, with minimal IRD and coarse-grained terrigenous material. In contrast, the glacial system, which receded from LF in the early Holocene, had an increased coarse sediment fraction, indicative of a glacial system (a calving tidewater margin) different from that of the modern Müller Ice Shelf system.

The upper unit of GC-1 displays similar sedimentologic and geochemical characteristics to Kasten Cores 72, 75, and 5. TOC content in all four cores increases with depth, with maxima at the base of cores 72 and 75 and at 1.8 m in GC-1 and 2.6 m in KC 5. Due to their short length, correlation of TOC maxima for LF may not be defined using cores 72, 75 and 5; however, the extended record of sedimentation in GC-1 defines a TOC maximum for the fjord at 3200 yr BP and a transition to a colder setting dominated by fast ice at ~2700 yr BP. The decline in TOC content associated with the formation of semipermanent fast ice around 2700 yr BP corresponds to another global "T-Event" (3000–2000 yr BP), recently documented in the GISP2 ice core (O'Brien *et al.* 1995). TOC content minima at 0.5 m, 0.3 m, 0.7 m, and 0.6 m in cores 72, GC-1, 75 and 5 (respectively), reflect the maximum extent of the Müller Ice Shelf during the LIA (roughly 400 yr BP) and are associated with the LIA "T-Event" (600–100 yr BP) observed in the Northern Hemisphere (O'Brien *et al.* 1995).

Increasing TOC content at the base of cores 72, 75 and 5 can be assumed to correlate with the increase in TOC content at the base of the upper unit of GC-1. A discrepancy in the dates of this transition arises from the variable sedimentation rates in cores 72, 75 and 5. However, the transition from open marine to fast ice conditions in GC-1 (2700 yr BP) directly correlates to a similar palaeo-environmental transition present in core PD92-2 30, which has been dated to 3000–2700 yr BP (Leventer *et al.*, in press). The dates of the transition in GC-1 and core 30 agree with the 2700 yr BP date obtained from lake sediments on Livingston Island and James Ross Island (fig. 2), suggested by Björck *et al.* (1991, 1996) to represent the end of a climatic optimum for the AP.

CONCLUSIONS

The sedimentology and geochemistry of Gravity Core 1 establishes three palaeoenvironments for LF, Antarctica, during the Holocene; each palaeoenvironment has been

inferred from specific sedimentologic units present in the core. The lower unit (5.5–4.55 m) of GC-1 reflects the deglaciation of the fjord and is characterised by coarse ice proximal terrigenous deposits and a diluted biogenic component. The sediment in the middle unit (4.55–1.45 m) is characteristic of an open fjord environment, with a relatively high biogenic component, low terrigenous component and variable IRD. Variation in TOC content of the unit is explained by the decline of productivity, based upon the extensive seasonal sea ice present in the fjord; particularly between 6700 and 4900 yr BP. The geochemistry of the upper unit (1.45–0 m) correlates with Kasten Cores 72, 75 and 5, previously documented by Stein (1992) and Domack *et al.* (1995) as reflecting the formation and fluctuation of the Müller Ice Shelf in the latest Holocene. The core's ^{14}C chronology, based upon the well-constrained AMS date on scaphopod calcite (see earlier), extends throughout the entire Holocene. A period of maximum sea-ice extent appears to have begun in the fjord about 2700 yr BP (the onset of the Neoglacial).

The high resolution sedimentologic record of Gravity Core 1 represents the most complete maritime record of Holocene climate change, to date, in the Antarctic. Results indicate extensive environmental variability throughout the Holocene in climatically sensitive LF, inferred as reflecting climatic variability. A palaeoclimatic reconstruction for the region suggests that a climatic warming in the early Holocene caused the deglaciation of the fjord. Warm temperatures gave way to colder temperatures which caused seasonally variable sea-ice coverage in the deglaciated fjord between 6700 and 4900 ^{14}C yr BP. Another climatic warm period, reflected by high productivity, occurred in the late middle Holocene between 4900 and 2700 yr BP. The formation of the Müller Ice Shelf around 400 yr BP reflects the colder climates of the late Neoglacial. Based upon this palaeoclimatic interpretation, climatic variability may be inferred for the AP, making the definition of a specific "Hypsithermal" event difficult. However, the overall cooling trend observed in LF for the middle to late Holocene is consistent with recent ice-core data from East Antarctica (Plateau Remote, Mosely-Thompson 1996). A similar, time-synchronous record of Holocene climatic variability was documented in the GISP2 ice core (O'Brien *et al.* 1995). Both the Northern Hemisphere ice-core record (GISP2) and the Southern Hemisphere sedimentary record (GC-1) correlate with climatically cold triple events ("T-Events" or minima in solar irradiance; Stuiver & Braziunas 1989) previously defined for the Northern Hemisphere (fig. 5). The documentation of a similar climatic variability in the polar regions of the Southern and Northern Hemispheres is important in terms of establishing a high-latitude pattern of variable climate change in the Holocene; this enables a shift of focus towards establishing an increased understanding of the causes of regional climatic variability. Understanding the natural variability of global and regional climatic cycles in the Earth's recent past may assist researchers in understanding the response of the Earth's systems to the warmer temperatures documented in the era of "Global Warming". An additional AMS date from the sea-ice maxima in the middle unit of core GC-1, as well as additional micropalaeontologic (diatom and foraminifera) analysis, may aid in solidifying the palaeoclimatic chronology of this important region.

ACKNOWLEDGEMENTS

This research was supported by the Office of Polar Programs (Climate and Ocean Sciences, and Earth Sciences), National Science Foundation, through the Research in Undergraduate Institutions program (grants OPP-8915977) and Research Experience for Undergraduates (OPP-941853 to Hamilton College). We wish to thank the officers and crew of the RV *Polar Duke* and the support staff of Antarctic Support Associates for all their help, while in the field. The assistance of Mr Matt Curren and Dr Thomas Janecek of the Antarctic Marine Geology Research Facility at Florida State University is also acknowledged, as well as the helpful comments of two anonymous reviewers.

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