Glacial geology

A new mechanism for diatom emplacement and concentration in glacigenic deposits

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Recent studies report diatoms in a variety of glacigenic and subglacial deposits. Some interpretations of these occurrences are controversial and require ice-sheet collapse. Harwood (1986a) found freshwater diatoms in the base of the Greenland Ice Sheet. Similarly, diatoms were found in sediments of the Sirius Group (Kellogg and Kellogg 1984; Webb et al. 1984; Harwood 1986b), which crops out along the Transantarctic Mountains; on the surface of the east antarctic ice sheet at Elephant Moraine (Faure and Harwood 1990); and beneath the west antarctic ice sheet at Upstream B (Scherer 1991, 1993). Here, we summarize an alternative mechanism to account for the presence of diatoms in glacigenic sediments, one which does not require collapse of an ice sheet.

Assuming no major ice-sheet collapse, two independent processes could combine to deliver diatoms to subglacial sediments: eolian transport (onto the ice sheet) followed by glacial transport (to the base of the ice sheet or to the ice-sheet margin). Both freshwater and marine diatoms have been recovered from ice and snow samples on present-day ice sheets (Burckle et al. 1988; Ram, Gayley, and Petit 1988; Ram and Gayley 1991, 1994; Kellogg and Kellogg 1996). These occurrences are all due to surface winds carrying diatoms, and even diatomaceous sediment clasts, to central regions of Antarctica as well as to other ice sheets from sources beyond the ice-sheet margins. Kellogg and Kellogg (1996, in preparation) demonstrated that winds transport diatoms, opal phytoliths, and sponge spicules to the South Pole. Burckle et al. (1988) found diatoms in ice at Dome C (figure 1), whereas Burckle and Potter (1996) reported diatoms and diatomaceous clasts in cracks in Paleozoic and Mesozoic igneous and sedimentary rocks from Marie Byrd Land and the McMurdo Dry Valleys (figure 1).

Ice melting off the bed or along the margin of an ice sheet is ultimately derived from snow accumulation on the ice sheet. Diatoms deposited on an ice sheet by eolian processes are buried and trapped in the snow. Snow is compressed to ice and moves vertically and horizontally through the ice column along a flowband until it either reaches the ice bed or the ice margin, where it is melted off or otherwise removed by ablation (Hooke and Hudleston 1978; Paterson 1994; Gow and Meese 1996). Solid particles in the ice are also transported in this manner (figure 2). Thus, atmospherically transported diatoms have the potential to occur in reworked assemblages containing diatoms of different ages and habitats (Kellogg and Kellogg 1996). If basal melting causes the flowlines to intersect the glacier bed, the diatoms are delivered to the subglacial bed (figure 2*A*). Note that deposition takes place without collapse of the overlying ice sheet.

If flowlines do not move to the ice bed because it is frozen (figure 2*B*) or freezing (figure 2*C*), diatoms should crop out below the equilibrium line near the ice-sheet periphery. An

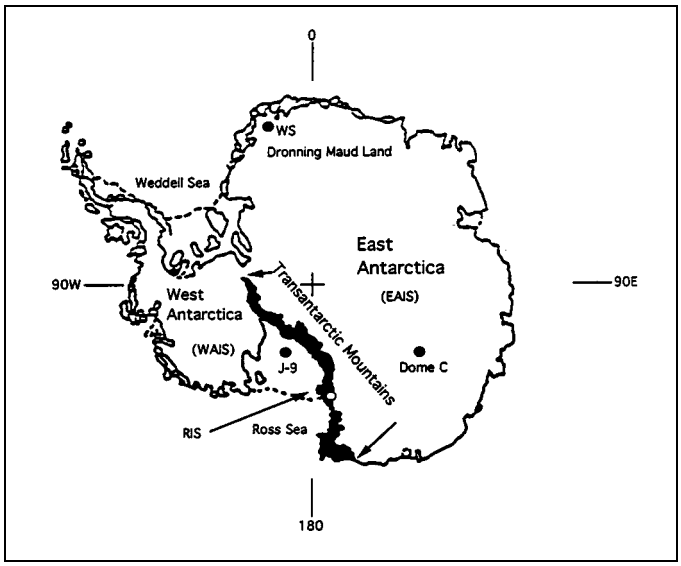

Figure 1. Map of Antarctica showing locations discussed in text. Abbreviations are as follows: J-9, Ross Ice Shelf site J-9; RIS, Ross Ice Shelf. The Wilkes-Pensacola Basin is located along the east side of the Transantarctic Mountains.

example might be Elephant Moraine in northern Victoria Land, where flowlines intersect the surface because of local surface ablation. If flowlines crop out at the ice-sheet margin, diatoms may accumulate in end moraines rather than ground moraines (basal tills) as in the melting-bed scenario. Our last example is of a marine-based ice sheet feeding an ice shelf (figure 2*D*). Beneath grounded portions of the ice sheet, basal melting should be accentuated under ice streams because of frictional heating. Thus, diatom deposition should be concentrated in ice-stream beds, as at Upstream B. If basal melting occurs beneath an ice shelf, flowlines intersecting the bed should release diatoms and other particles to the underlying water where they settle to the sea floor, contaminating other sedimentary components being deposited contemporaneously (e.g., Ross Ice Shelf Project site J-9).

These pathways for diatoms are described by a simplified model of a flowband using reasonable assumptions for the dome height, velocity field, surface accumulation rates, and ice hardness for various basal melting rates (Burckle et al. 1997). The model estimates residence times for diatoms deposited along the top surface of the flowband. At a basal melt rate of 0.01 meter per year, the trajectories of all flowlines but those nearest the margin intersect the bed; diatoms deposited near the dome reach the bed about halfway down the flowband. Larger values of basal melting lead to the diatoms reaching the bed even faster and closer to the point of origin.

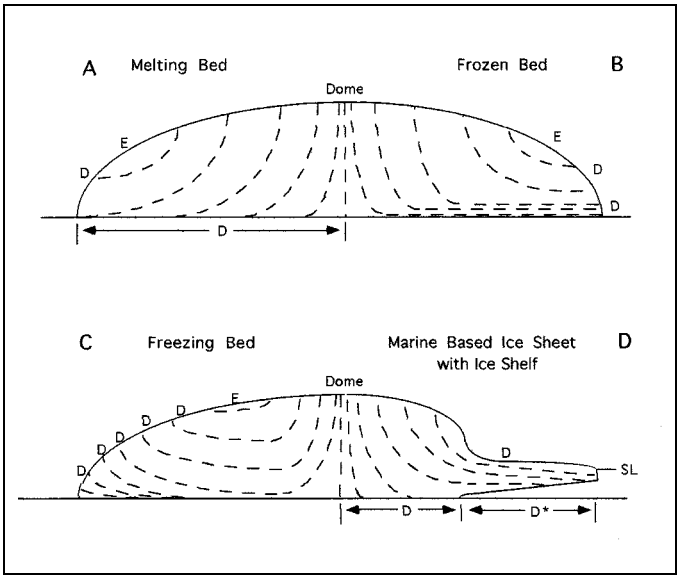

Figure 2. Model illustrating transport of diatoms (and other particles) deposited by winds on the ice-sheet surface. *A*. Ice sheet with basal melting. Note that diatoms (D) are deposited beneath the ice sheet in basal till. *B*. Ice sheet with frozen bed. Note that diatoms are carried to the ice-sheet margin where they may accumulate in end moraines. *C*. Ice sheet with freezing bed. Note that diatoms are carried to the surface at locations below the equilibrium line (E). *D*. A marine-based ice sheet flowing into an ice shelf. Note diatoms deposited beneath grounded ice as in A but also settling out beneath the ice shelf where released by basal melting (D*).

In view of these results, we suggest alternate interpretations of some reported diatom occurrences in glacially derived sediments in Antarctica (Burckle et al. 1997). We thank T. Hughes, J. Kleman, N. Reeh, and A. Stroeven for helpful comments during various stages of this study. Research was supported by grants from the National Science Foundation: OPP 92-20216 to Lloyd H. Burckle, OPP 94-16306 to Davida E. Kellogg, and EPSCoR grant R11-8922105 to Thomas B. Kellogg and James L. Fastook. Lloyd H. Burckle also acknowledges support from the Swedish Natural Science Research Council.

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Beryllium-10 in sediments beneath the west antarctic ice sheet: Hypotheses and assumptions

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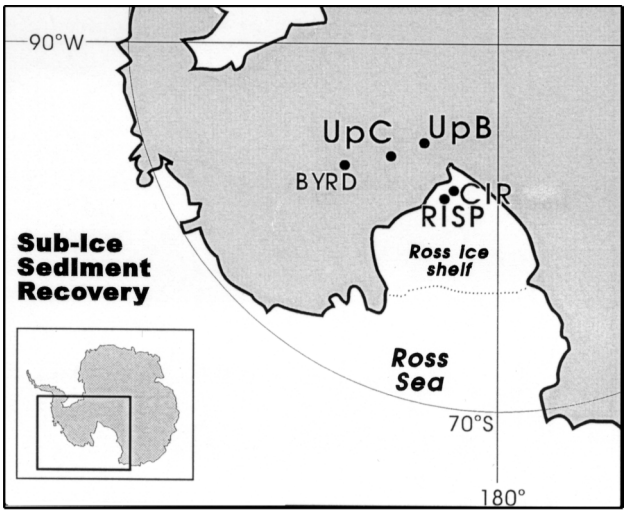
Sediments beneath the west antarctic ice sheet contain invaluable information regarding ice-sheet history and glacial processes. Samples of sediments have been recovered from beneath the west antarctic ice sheet at Upstream B and the nearby shear marginal zone and, most recently, from beneath Upstream C. Additional samples have been recovered from beneath Crary Ice Rise, Ross Ice Shelf Project site J-9, and basal debris from the ice core at Byrd Station (figure). We are exploring the potential of the cosmogenic isotope beryllium-10 (^{10}Be) in these subglacial sediments as a tracer of ice-sheet history and glacial processes. In this article, we outline our initial working hypotheses and assumptions.



Ross sector of west antarctic ice sheet showing sites of subglacial sediment recovery: Upstream B (UpB), Upstream C (UpC), the Ross Ice Shelf Project site J-9 (RISP), and the Crary Ice Rise (CIR).

The first researchers to use ^{10}Be in sediments from the interior of West Antarctica were Yiou and Raisbeck (1981), who studied samples from beneath the southern part of the floating Ross Ice Shelf. They found low ^{10}Be concentrations in core samples (less than 10^7 atoms per gram) and interpreted this finding as indicative of pre-Quaternary sediment in agreement with Harwood, Scherer, and Webb (1989). The surface layer, in contact with sub-ice shelf water, had a measurable concentration (approximately 10^8 atoms per gram), but even this was an order of magnitude lower than the concentrations in Quaternary marine sediment.

The cosmogenic isotope ^{10}Be (halflife 1.5 million years) is produced both in the atmosphere and *in situ* at the Earth's surface. *In situ* production is not significant in fine-grained submarine or subglacial sediments, so all ^{10}Be detected in the clay-sized fraction of subglacial sediments is assumed to be originally from atmospheric fallout. The ^{10}Be atoms become part of the sedimentary record by adsorbing onto available clay particles.

We discuss three potential sources of ^{10}Be to the west antarctic ice-sheet subglacial environment:

- direct release of ^{10}Be trapped within the ice, due to basal melting,
- marine waters advecting ^{10}Be atoms beneath a floating ice shelf, and
- residual ^{10}Be from past marine deposition in the currently ice-filled basin, which implies massive retreat or collapse of the marine ice sheet.

^{10}Be release by basal melting is insignificant where the ice sheet is frozen to the bed but could become significant where the basal melt rate is high, particularly in regions of high geothermal flux. We consider advection of ^{10}Be beneath a floating ice shelf at Upstream B and vicinity, deep within the west antarctic interior, to provide at best a very minor source of ^{10}Be because ice shelves this far south would likely be unstable and short-lived. We also consider the possibility that tidal pumping might introduce a small amount of ^{10}Be from sub-ice shelf water, across a broad coupling zone, although sub-ice shelf waters are likely depleted in ^{10}Be (Aldahan et al. in press). Sediments older than late Miocene will have experienced significant ^{10}Be decay, thus high residual ^{10}Be concentration in subglacial sediments must reflect contribution from post-Miocene deposition in an open marine environment, implying disintegration of the ice sheet, as reported by Scherer (1991).

To apply ^{10}Be successfully as a tracer for addressing questions of basal dynamics or ice-sheet history, it is important to understand the relative contributions of the three ^{10}Be sources described above. To realize this goal, we must further our understanding of glacier bed conditions, especially with regard to basal water production and flow. If a water conduit system or sheet flow removes excess water from the bed, then some fraction of the ^{10}Be released from basal melt will likely be flushed to the grounding line. Rapid basal flow seems to be the case at Upstream B, where borehole experiments demonstrate that water at the bed flows, at least locally, significantly faster than the ice (which itself flows at approximately 400 meters per year) (Engelhardt and Kamb 1997). In a basal sliding scenario, ^{10}Be might accumulate only if there is net clay deposition and no erosion.

Ice flow directly linked with till deformation in a several meters thick till layer would provide a mechanism for continuously adding and storing ice-derived ^{10}Be in the sediments.

Continuous mixing would distribute ^{10}Be throughout the active till layer. Consequently, a vertical ^{10}Be profile through the sediment column would reveal no stratigraphic variability. Detecting a basal melt signal would, however, require a very significant ^{10}Be addition, because the signal would become diluted through the sediment column. In fact, rapid till mixing by deformation processes would result in an absence of stratigraphic variability in ^{10}Be content, whether derived from relict sediments or basal ice melt.

We have calculated the anticipated input of ^{10}Be by basal melt, assuming a 1-millimeter-per-year basal melt rate and 10^5 ^{10}Be atoms per gram of melted ice, over the entire Holocene (last 10,000 years). This scenario yields a total ^{10}Be inventory of 10^8 atoms per square centimeter. With this inventory and continuous ^{10}Be accumulation for 10,000 years with no erosion of the surface sediment, the ^{10}Be concentration is expected to be about 10^7 atoms per gram of sediment in contact with the ice. If this amount is diluted through a 10-meter-thick deforming till devoid of residual ^{10}Be, then the average concentration should be less than 10^6 atoms per gram, or near the detection limit. These parameters would result in undetectable levels even if mixed through a 1-meter-thick till, but a signal should be detectable in a deformed layer of less than 1 meter.

The above calculations assume no erosion of the sediment column. Obviously, erosion at the bed and mixing with older sediments would significantly reduce the ice-derived ^{10}Be content of the sediments, just as higher basal melt rates would increase the ^{10}Be input. One may be tempted, for these calculations, to extend the sub-ice release time well beyond the Holocene, which in a closed system would lead to higher ^{10}Be content. Clear evidence has been found, however, of very significant erosion and removal of older subglacial debris in West Antarctica during Quaternary and earlier glaciations.

From the above calculations, we tentatively conclude that significant ^{10}Be concentration in sediments beneath the ice sheet is easier to explain by marine deposition in the west antarctic interior than by basal ice melt, although basal melt and flow conditions must be better constrained. An absence of ^{10}Be throughout the sediment column would reflect old sediment, derived from Miocene and older strata, with little contribution from younger deposits, and an undetectable contribution from basal ice melt or other sources. We will continue to develop these hypotheses as we analyze our ^{10}Be data from ice stream B, ice stream C, the Crary Ice Rise, the Ross Ice Shelf Project site J-9, and Ross Sea diamictons, in hopes of testing theories regarding ice-sheet history and glacial processes.

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Diatoms in subglacial sediments yield clues regarding west antarctic ice-sheet history and ice-stream processes

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Alley et al. (1987) suggested that rapid flow of ice stream B, West Antarctica, can be explained by active deformation of a layer of unconsolidated, water-saturated till, about 6 to 10 meters thick. Such a layer is recognized seismically (Blankenship et al. 1987). Hot-water boreholes at Upstream B have provided access to the bed for sampling of these sediments by piston coring and other methods. The sediments recovered are consistent with predicted physical properties, but active deformation of this layer has not been established. Engelhardt and Kamb (in press) report an *in situ* test that suggests that

basal sliding accounts for between 69 and 83 percent of ice-stream motion at this site and that till deformation may be restricted to the upper few centimeters, rather than the predicted thickness of as much as 10 meters.

Here we present "quantitative microstratigraphic" analysis of sediments recovered from various boreholes in the Upstream B area. We use absolute abundance of diatom fragments as sedimentary tracers. Diatoms provide excellent tracers for several reasons.

- They have known initial conditions (size, shape, approximate age, and environment of origin).
- Fragments as small as 1 or 2 micrometers can be recognized as being derived from diatoms (although fragments generally cannot be identified to the species level).
- New data on diatom comminution places constraints on physical conditions in the subglacial environment (R. Scherer, N. Iverson, and T. Hooyer unpublished data, manuscript in preparation).

Upstream B diatoms are too rare to perform accurate estimates of whole fossils, but abundance estimates of small diatom fragments (<5 micrometers) in the <250-micrometer fraction of these sediments are highly reproducible, using the method of Scherer (1994), despite the fact that the number of fragments generated from a single diatom can vary from fewer than 5 to more than several hundred.

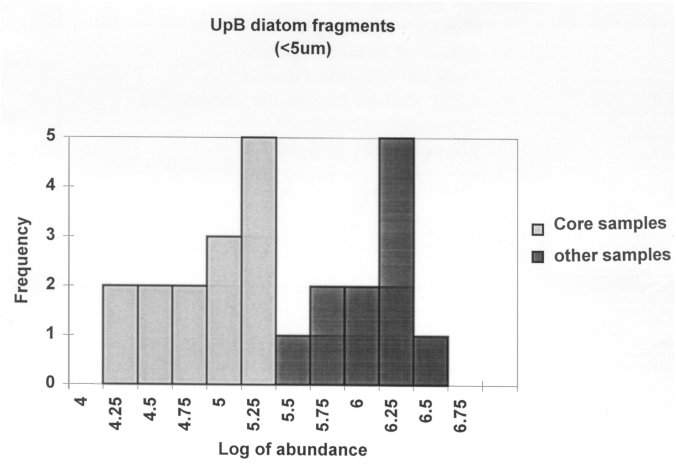
The data demonstrate a distinct difference between stratigraphic sediments recovered by piston coring and sediment samples that include the particles in closest proximity to the ice. The uppermost sediments contain a significantly higher concentration of diatoms and diatom fragments than those below. Core samples contain a mean concentration of 1.4×10^5 fragments per gram dry sediment (fr/g) (n=14), whereas samples that approximate the top of the sediment column, contain a mean of 1.8×10^6 fr/g (n=11), an order of magnitude higher (figure). These values are several orders of magnitude lower than "typical" Ross Sea glacial marine sediments, including subglacial sediments from Crary Ice Rise (Scherer et al. 1988) and Ross Ice Shelf Project site J-9 (Harwood, Scherer, and Webb 1989).

These results suggest that the till column, recovered by piston coring, is distinct from the sediments in direct contact with ice and the basal water system. Better preservation of diatoms in the uppermost sediments suggests that it has undergone less shearing than sediments that characterize the underlying weak but cohesive till. This inferred uppermost layer has never been recovered undisturbed, probably because of high water content and lack of cohesion, as well as disturbance due to drilling.

Miocene fossils strongly dominate the diatom assemblage of the upper unit, but several sediment samples have been found to contain rare Pliocene and Quaternary marine diatoms as well (including Scherer 1991; new data). These are found exclusively in the uppermost subglacial sediments, never in the diatom-poor core samples. Post-Miocene diatoms in these sediments have been interpreted as direct evidence of previous ice-sheet collapse (Scherer 1991, 1993).

Several lines of evidence argue for a subglacial origin, rather than an eolian source, for the post-Miocene diatoms. For the sake of brevity, we list three.

- A strong dominance of Miocene diatoms in the uppermost sediments, which are clearly of subglacial provenance (including diatomaceous clasts up to 0.5 centimeters), despite only trace occurrences of diatoms in the underlying till deposit, indicates that diatoms are transported from upstream source beds in a thin basal layer. This finding demonstrates that whole diatoms can be transported from upstream deposits.



Frequency plot of diatom fragment abundance (log scale). Light columns are core samples; dark columns are core tops and noncore samples. (um denotes micrometers.)

- Large-size Quaternary diatoms (many >60 micrometers) are as common as small diatoms, which are more easily transported by wind.
- No modern nonmarine diatom specimens have been found in Upstream B sediments, despite the fact that extant nonmarine diatoms greatly outnumber marine diatoms in filtered antarctic ice (Kellogg and Kellogg 1996; R. Scherer and S. Tulaczyk, unpublished data from ice stream B).

The above observations cannot be reconciled with a purely eolian source for the Quaternary diatoms.

Diatom data indicate that sediments in contact with the ice represent a distinctly different sedimentary unit than the underlying meters-thick diamicton recovered by piston coring. The uppermost sediments beneath Upstream B are either a very thin, waterlain, and actively deforming sediment layer, the favored view, or suspended sediment-rich water. We tentatively conclude, based on diatom and other data, that the (centimeters-thick) upper sedimentary unit is mobile, probably actively deforming, and the lower, thicker (meters-thick) unit is relict and not undergoing active deep deformation as described by Alley et al. (1987). The contact between the two sedimentary units is unconformable and probably an active erosion surface. The upper unit, therefore, may be analogous to the mobile drift layer described by Alley et al. (1987), but deformation is currently restricted to a very thin layer. The lower unit may have been formed by deep deformation, cer-

tainly the diatom data suggest significant shearing, but the deposit was probably formed under a different glacial regime than present conditions.

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High-resolution diatom record in bioturbated antarctic sediments

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As a part of a project studying late Holocene climatic changes in the Antarctic Peninsula area, the stratigraphic resolution of the diatom record from bioturbated marine sediments recovered by trigger and box coring is examined. We are analyzing absolute diatom abundance and the diatom assemblage as proxies for diatom paleoproductivity and paleoenvironments. Determining stratigraphic resolution in box and trigger cores, which may recover the sediment-water interface, will aid in correlation between cores, including tying trigger and box cores to the longer records of piston cores, which do not recover an unaltered sediment-water interface. Correlating these records will allow calibration of paleoceanographic records with modern conditions. This study includes analysis of two trigger cores from the Bransfield Strait (TC 86-7, TC 86-9) and two box cores from the Gerlache Strait (BC 86-79, BC 86-82) (table).

The bottom of the Bransfield and Gerlache Straits is oxygenated, and the diatomaceous sediments are bioturbated typically to a depth of approximately 10 centimeters (cm) (Harden, DeMaster, and Nittrouer 1992). The sediments accumulate in the basins at rates of between 1 and 3 millimeters (mm) per year (Harden et al. 1992). Our hypothesis is that bioturbation in these high-accumulation-rate cores acts as a "low pass" filter, smoothing minor variation and highlighting sub-

Coring sites, core length, and water depth (from Bryan 1992)

Core label	Sampling site	Latitude/ Longitude	Water depth (meters)	Core length (centimeters)
TC 86-7	Bransfield Strait	61°75.5/ 55°29.0	1,903	35
TC 86-9	Bransfield Strait	62°09.4/ 56°09.0	1,125	54
BC 86-79	Gerlache Strait	64°59.7/ 63°20.0	494	37
BC 86-82	Gerlache Strait	64°37.8/ 62°52.05	96	45

century-scale trends. This hypothesis holds only where bioturbation has been vigorous and mass sedimentation events absent. The samples, taken at 5-cm intervals down the cores, were prepared for identification and counting in the light microscope as described by Scherer (1994), allowing quantitative analyses of samples.

The slides were examined for total count of diatom valves, and also counted on a non-*Chaetoceros* basis, counting at least 400 valves excluding the genus *Chaetoceros*. Two separate counts are necessary because *Chaetoceros* resting spores dominate the assemblage, masking trends of other, less common species (Leventer et al. 1996). Several species of *Chaetoceros* are represented, but distinguishing between taxa in the light microscope is extremely difficult. To evaluate the stratigraphic resolution, the absolute abundance of diatoms was calculated, as well as ratios of different species. From the total count, the absolute abundance and a ratio of *Chaetoceros* resting spores vs. *Chaetoceros* vegetative cells (Cs/Cv) were calculated. From the *Chaetoceros*-free count, a ratio of sea-ice-related species (*Fragilariopsis curta* + *F. cylindrus*) versus a common neritic species (*Thalassiosira antarctica*) was calculated. These ratios are suggested to reflect the input of meltwater and a receding ice edge (Leventer et al. 1996).

The diatom assemblage preserved in the sediment is clearly dominated by *Chaetoceros* spp., making up between 80 and 98 percent of the assemblage (mean = 92 percent). The amount of vegetative *Chaetoceros* cells ranges from 1 to 16 percent (mean = 8 percent). On a *Chaetoceros*-free count, the diatom assemblages show some differences between the two straits, reflecting the different climatic/oceanic settings. The Gerlache Strait non-*Chaetoceros* assemblage is dominated by the neritic species *Thalassiosira antarctica,* with approximately 50 percent. The Bransfield Strait assemblage has approximately equal amounts of *Fragilariopsis curta, F. kerguelensis, Thalassiosira antarctica,* and *T. gracilis*, approximately 15–20 percent each, indicating a more diverse setting. The distribution of non-*Chaetoceros* species shows little variation downcore, and no obvious co-variation between species was found. Minor fluctuations occur, but often these cannot be observed in adjacent cores.

Of the four cores investigated, all but TC 86-7 show a general trend of increased diatom abundance downcore (figure 1). High concentrations of *Chaetoceros* resting spores in sediments traditionally have been interpreted as indicating high primary production (Leventer et al. 1996; Scherer 1992). The *Chaetoceros* abundance calculated from these cores [1×10^8 to 5×10^8 cells per gram dry sediment (c/gds)] compares well with those calculated by Leventer (1991), Scherer (1992), Zielinski and Gersonde (1997), and by Crosta, Pichon, and Labracherie (1997). Core TC 86-7 shows extremely high diatom abundance (>1×10^9 c/gds) at the 15-cm level. This peak is interpreted as a local mass-sedimentation event, which may not reflect decadal-scale trends. Because of this anomaly, this core is not included in the evaluation of the stratigraphic resolution of the well-mixed sediments, and it highlights the importance of careful examination of the cores as samples are selected.

The *Chaetoceros* ratio (Cs/Cv) also shows an increasing trend down the core (figure 2). In the samples from the Gerlache Strait (BC 86-79 and BC 86-82), increase in the Cs/Cv ratio is slightly out of phase, compared to the diatom abundance curve, whereas in TC 86-9 the ratio and valve abundance correlate well (figures 1 and 2). The ratio of (*F. curta* + *F. cylindrus*)/*T. antarctica* shows little or no variation in the samples from the Gerlache Strait, whereas the sample from the Bransfield Strait shows a downcore increase and good correlation with diatom abundance (figure 2).

Higher diatom abundance downcore and, coincidentally, higher Cs/Cv ratios are interpreted as indicating more favorable conditions for diatom growth. Lower diatom abundance today could also be influenced by increased terrigenous input, possibly caused by increased glacial melt.

With a stratigraphic sampling of only 5 cm, some fine-scale fluctuations in the diatom record can be observed. The signal from diatom abundance is

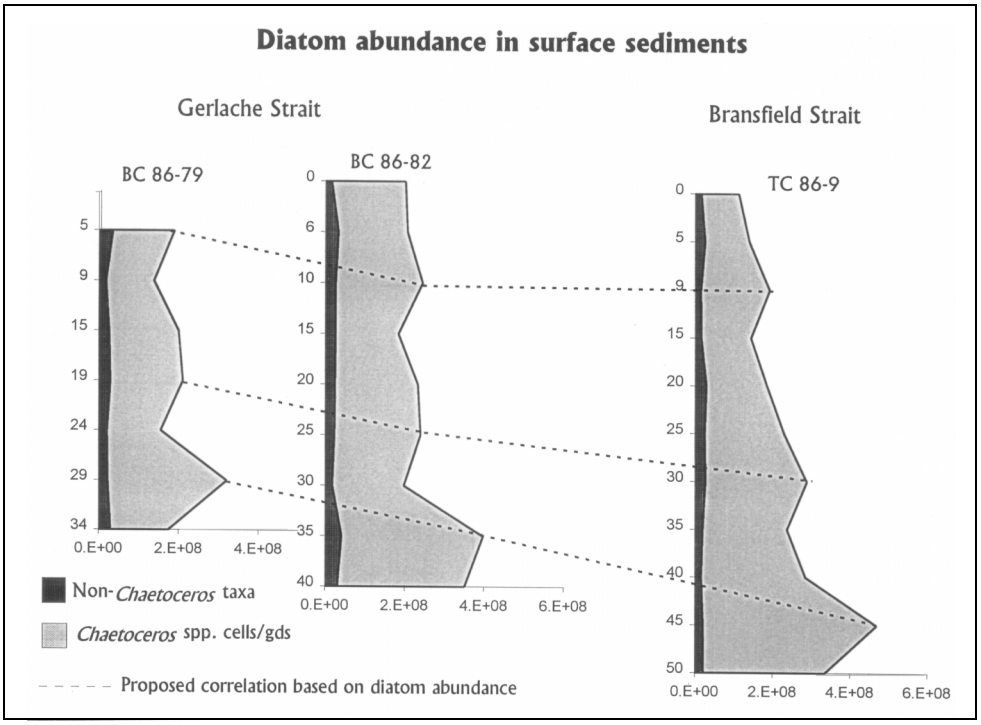

Figure 1. Diatom abundance, cells per gram dry sediment, and tentative correlation based on abundance.

distinct, despite bioturbation, suggesting a subcentury-scale trend. Tentative correlation between the Gerlache and Bransfield Straits, based on the diatom abundance, may be possible, if we exclude TC 86-7. Such a correlation implies a higher sediment accumulation rate in the Bransfield Strait than in the Gerlache Strait. This higher rate agrees with Harden et al. (1992), who estimated the accumulation rate in the northern Bransfield basin to be up to 3 mm/year, declining southward and about 2 mm/year in the Gerlache Strait.

The high resolution obtained from closely spaced sampling might be helpful in correlating trigger and box cores to piston cores and, in turn, enhancing the possibility of making correlations between piston cores. With better chronostratigraphy, correlation, and calibration with the modern sediment flux, the sediments of the Antarctic Peninsula will provide a detailed paleoclimate proxy record.

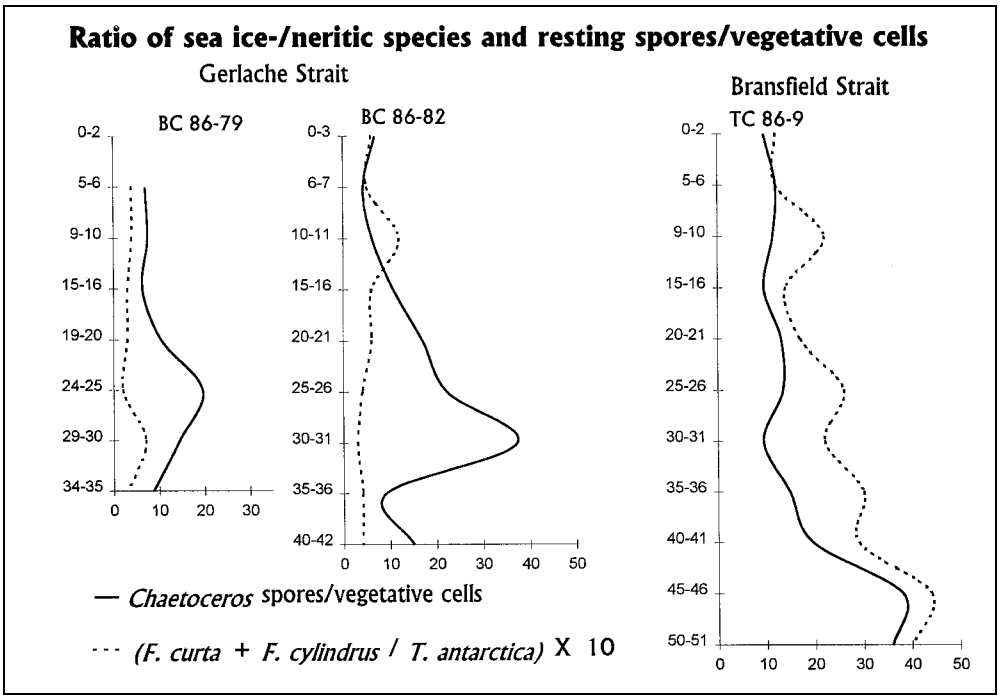

Figure 2. Downcore changes in ratio of sea-ice related species versus neritic species (enhanced by 10 for presentation purposes) and in ratio of *Chaetoceros* resting spores verus *Chaetoceros* vegetative cells.

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Replication of accelerator mass spectrometry carbon-14 dates on the acid-insoluble fraction of Ross Sea surface sediments

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The West Antarctic Ice Sheet Project (WAISP) is a multidisciplinary effort concerned with developing past, present, and future scenarios for the history of this marine-based section of the antarctic ice sheet (Bindschadler 1991, 1995). A major issue is how to develop comparable time-histories for a variety of proxy records including

- ice cores,
- terrestrial based glacial geological observations, and
- the records preserved in marine sediments from the western and east-central Ross Sea.

In particular, the problem with dating marine sediments around the antarctic continental margin is that, frequently, marine carbonates are scarce to nonexistent, and dates must be obtained on the acid-insoluble organic fraction. Such samples from the modern seawater/sediment interface give ages between 2,000 and 5,000 carbon-14 (^{14}C) years (Domack et al. 1989; DeMaster, Ragueneau, and Nittrouer 1996; Harris et al. 1996; Licht et al. 1996). Why is this? We can deduce that it is probably associated with the recycling of old carbon (Truswell and Drewry 1984), but geographical and temporal variations are not well constrained.

This short article is concerned with one problem in dating surface marine sediments. Several sources of variability might affect the radiocarbon date of surface marine samples (Andrews et al. in preparation):

- the net sediment balance at the site,
- the nature of the coring system (e.g., the true top was not recovered),
- differences in laboratory pretreatments,
- geographic variability in terms of inputs of old carbon, and
- spatial variability in the ocean reservoir.

Another concern, which is rarely if ever addressed, however, is within-sample variability: the problem of replicating a result from samples collected a few to tens of centimeters apart. Normally, we would obtain a date from a single sample and then assume that if we repeated the experiment, 95 percent of the time the resulting age would lie within 2 standard errors about the mean. For example, a date of 2,050±65 before present implies that we are 95 percent sure that the true age lies somewhere between 1,920 and 2,180 years ago.

Our experiment

Our data consist of three sets of three dates and a single acid-insoluble/marine carbonate pair (table). We used three surface samples from cruises *NBP*9606, collected by Leventer, and a set of samples from a core (*NBP*9501-39), sampled by Licht.

The samples were shipped to the Accelerator Mass Spectrometry (AMS) National Science Foundation Facility at the University of Arizona. Sample treatment was carried out at the AMS Facility and consisted of an acid pretreatment to remove any carbonate fractions. The sample is then washed with distilled water and dried. The samples are further processed as discussed in the next paragraph.

At the AMS Facility, all pretreated samples were combusted with copper oxide to make carbon dioxide (CO_2). The CO_2 gas is cleaned by passing the gas over zinc at room temperature and copper and silver at 600°C. The gas is then split and a portion reserved for stable-isotope measurement by mass spectrometry. The larger split is further processed to graphite using the methods described by Donahue, Jull, and Linick (1990). The graphite powder is pressed into a target holder and mounted in a 32-position carousel, which is placed in the ion source of the AMS. The calculation procedures are reported in Donahue, Linick, and Jull (1990).

Results

The nature of the general problem is identified immediately when comparing the marine carbonate and acid-insoluble fraction (aif) from *NBP*9606-38 (table). The brachiopod date of 1,065±45 is somewhat younger than the generally accepted ocean reservoir correction of 1,200 to 1,300 years before present (Gordon and Harkness 1992; Berkman and Forman 1996). This date, however, is nearly 2,500 years younger than the aif age of 3,580±50. An examination of the two surface-data sets (*NBP*9606-91 and -15) indicates that the maximum range in the reported surface ages is 170 years and 480 years, respectively. In *NBP*9606-15, the three samples have overlapping error bars at the 95 percent confidence level, but in *NBP*9606-91, the maximum difference of 480 years is larger than would be expected by chance, suggesting some lateral variability in sediment organic composition. Note that the isotopic carbon-13 ($\delta^{13}C$) values differ between *NBP*9606-15 and *NBP*9606-91 and *NBP*9501-39 (table). The three replicate samples from the core at a depth of 98–100 centimeters cluster tightly together and have a maximum range in quoted values of only 145 years.

In conclusion, our data suggest that some lateral variability in sediment organic composition exists and could add a small variation in sample age but that this variability is small

Radiocarbon dates

Core site (latitude, longitude, water depth) Lab number	Depth[a]	Material	Date error	$\delta^{13}C$‰
***NBP*9501-39 (74.474°S 173.474°E; 557 meters)**				
AA-25257	98–100[b]	aif[c]	6,556±60	–28.9
AA-25275	98–100	aif	6,520±55	–28.9
AA-25276	98–100	aif	6,665±55	–28.8
Weighted mean: 6,582±45				
***NBP*9606-15 (75.033°S 166.263°E; 939 meters)**				
AA-25277	Surface	aif	2,755±45	–23.3
AA-25278	Surface	aif	2,925±45	–24.4
AA-25279	Surface	aif	2,805±50	–23.9
Weighted mean: 2,830±52				
***NBP*9606-91 (75.498°S 178.328°E; 451 meters)**				
AA-25280	Surface	aif	3,580±50	–28.7
AA-25281	Surface	aif	3,345±50	–28.8
AA-25282	Surface	aif	3,825±50	–28.9
Weighted mean: 3,583±139				
***NPB*9606-38 (78.142°S 171.027°W; 562 meters)**				
AA-25283	Surface	aif	3,580±50	–26.7
AA-25284	Surface	Carbonate[d]	1,065±45	+.07

[a]In centimeters.
[b]Core depth approximate due to transportation problems.
[c]Core depth approximate due to transportation problems.
[d]Carbonate (brachiopod).

compared to the ^{14}C age of the surface sediments.

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