

## Particulate matter and nutrient distributions in the ice-edge zone of the Weddell Sea: relationship to hydrography during late summer

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**Abstract**—Previous estimates of the marginal ice zone's quantitative contribution to biogeochemical cycles and annual productivity in the Southern Ocean may be conservative because of assumptions that phytoplankton blooms are associated only with actively retreating ice edges. Observations during March 1986, near an almost stationary ice edge in the northwestern Weddell Sea, revealed very low geostrophic currents, no appreciable horizontal gradients in temperature or salinity and no significant net melting or freezing in the ice-edge region. Vertical stratification within the upper 50 m was evident throughout the study area, and resulted primarily from prior melting of pack ice. In contrast with previous observations in marginal ice zones, the distribution of phytoplankton biomass showed little correlation with the meltwater field; here, significant horizontal biomass gradients occurred in an area where vertical stability was almost uniform laterally and both elevated biomass and diminished nutrient levels extended well below the pycnocline. Absolute levels of chlorophyll were modest (generally  $<0.7 \mu\text{g l}^{-1}$ ), but the highest particulate carbon, nitrogen and biogenic silica values ( $>10$ , 1.5 and  $3.5 \mu\text{mol l}^{-1}$ , respectively) were similar to those found previously during a spring phytoplankton bloom at the Weddell–Scotia Sea ice edge. The mean mole ratio of biogenic silica to organic carbon within the particle assemblage was 0.44, which is very high for surface seawater and about three times higher than that typically found in pure diatom cultures. Therefore, despite relatively low chlorophyll levels, the ice edge remained a localized maximum in biogenic particulate matter at least through March and this material was unusually rich in silica. Combined, these physical and biological data show that elevated phytoplankton biomass in the ice-edge zone can persist well past the time when net melting stops and the ice edge becomes stationary.

### INTRODUCTION

In polar and subpolar oceans that experience seasonal advance and retreat of pack ice, phytoplankton blooms in the ice-edge zone are a major component of the annual cycle of biogenic particle production. In all marginal ice zones studied to date, phytoplankton biomass levels have been substantially higher than those observed either under the ice or seaward of the region affected hydrographically by ice melt, e.g. the Bering Sea (ALEXANDER and NIEBAUER, 1981), the Ross Sea (SMITH and NELSON, 1985), Fram Strait (SMITH *et al.*, 1987) and the Weddell–Scotia Sea (NELSON *et al.*, 1987). Despite major

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differences in hydrographic conditions, nutrient concentrations, surface irradiance and biological features among ice-edge systems in different geographic locations, these past studies have indicated that vertical stability within the upper 100 m is essential for the growth and accumulation of biogenic material and may be the single most important factor regulating these processes. This appears to be particularly true in the Southern Ocean, where major nutrient concentrations are rarely, if ever, low enough to limit phytoplankton growth. Because the density of seawater is controlled almost entirely by salinity at low temperatures, ice melt creates a mesoscale region of vertical stability with high phytoplankton biomass and productivity. This mesoscale bloom has a significant impact on the annual productivity, trophic dynamics and biogeochemical cycles of the entire Southern Ocean (SMITH and NELSON, 1986; AINLEY *et al.*, 1986; NELSON and SMITH, 1986). However, major uncertainties remain as to these blooms' duration and magnitude, especially during the late summer/early autumn period, from which virtually no data have been available.

A previous study in the Weddell-Scotia Sea in spring showed that the ice-edge zone was the site of localized maxima in chlorophyll *a*, and particulate carbon, nitrogen and biogenic silica, and that these increased levels correlated spatially with the density structure (NELSON *et al.*, 1987). Mixed-layer depths in that study ranged from about 20 to 150 m, and Chl *a* concentrations ranged from  $<0.2 \mu\text{g l}^{-1}$  under the ice and  $<0.5 \mu\text{g l}^{-1}$  at stations with deep mixed layers  $>300$  km from the ice edge to values typically  $>5 \mu\text{g l}^{-1}$  within the bloom. The highest Chl *a* levels observed were  $>30 \mu\text{g l}^{-1}$  at one station near the ice (W. O. SMITH, unpublished data). Particulate carbon, nitrogen and biogenic silica showed similar distributions. A satellite image of pigment concentration confirmed these distributions, and also substantiated the shipboard observation that there was significant along-ice variation in particulate matter concentrations (NELSON *et al.*, 1987). These observations in the Weddell-Scotia Sea, taken together with a summer study of the ice-edge zone in the Ross Sea (SMITH and NELSON, 1985; WILSON *et al.*, 1986; NELSON and SMITH, 1986), indicate that the growth and accumulation phase of phytoplankton blooms in the marginal ice zone can be initiated early in the spring (no later than late October) and continues at least through early February, following the ice edge as it retreats. However, understanding of the degradation phase of this feature (its timing and its causal mechanisms) has been hampered by a shortage of late summer and autumn data on appropriate spatial scales.

In March 1986 we conducted a study in the marginal ice zone of the western Weddell Sea as part of the AMERIEZ (Antarctic Marine Ecosystem Research at the Ice-Edge Zone) project. Our goal was to determine the magnitude of any ice-edged phytoplankton bloom and to assess its dependence on the mesoscale density structure, circulation patterns and nutrient fields. We also sought to obtain the first data on the degradation phase of an ice-edge bloom in late summer so that estimates of annual productivity of Southern Ocean marginal ice zones (e.g. SMITH and NELSON, 1986; SMITH *et al.*, in press) could be made more accurate. We report here the distributions of temperature, salinity, nutrients and biogenic particulate matter that characterized the Weddell Sea ice-edge zone in March, and compare these distributions with the conditions observed in this and other ice-edge zones during austral spring and summer.

The study area extended westward from the deep basin of the central Weddell Sea onto the continental rise off the eastern coast of the Antarctic Peninsula (Fig. 1). Bottom depths in the eastern portion of this region exceeded 4500 m and shoaled to  $<3500$  m at

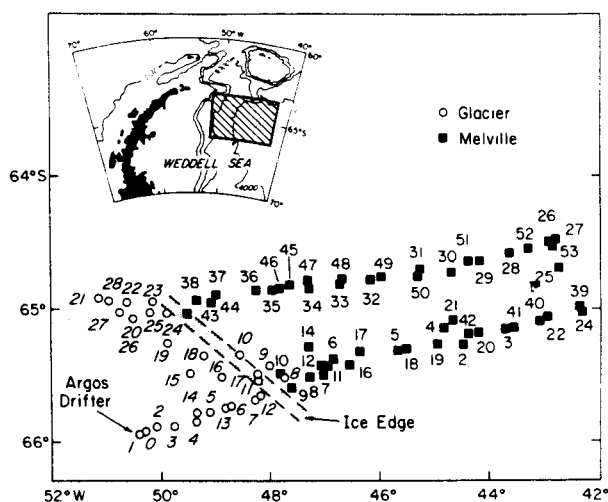


Fig. 1. Study area and station locations. An Argos-tracked buoy was deployed on a large ice floe at the location shown.

the westernmost stations. The region is situated more than 100 km east of the continental slope of the Antarctic Peninsula and more than 200 km south of the shoal region (<3000 m) and extends eastward from the tip of the Antarctic Peninsula. It lies more than 200 km south of the considerably shallower banks surrounding the South Orkney Islands. The westernmost portion of the study region is typically covered with multi-year pack ice throughout the year, whereas the eastern portion is ice-free in summer (approximately January to March) and partially to completely ice-covered for the remainder of the year (ZWALLY *et al.*, 1983).

Most previous physical oceanographic research in the Weddell Sea has consisted of regional-scale analyses of circulation and water-mass structure based upon relatively widely spaced temperature and salinity data. These analyses, integrated through physical reasoning with the surface wind field, have established the presence of a wind-driven, basin-wide cyclonic gyre in the Weddell Sea (e.g. GORDON *et al.*, 1978, 1981; DEACON, 1979). Our study area is believed to be characterized by a general northward flow, which comprises the western limb of this gyre. The water column is nearly homogeneous, indicating that the currents associated with the gyre are primarily barotropic; the geopotential surface height above the 1000 db level shows only very weak baroclinic surface currents (GORDON *et al.*, 1978). Closure of the wind-driven gyral circulation requires a western boundary current with a northward transport of order 60 Sv, vertically integrated current speeds of about  $8 \text{ cm s}^{-1}$ , and a current width of about 225 km at  $65^\circ\text{S}$  (GORDON *et al.*, 1981).

#### METHODS

We collected data aboard the R.V. *Melville* and the U.S. Coast Guard icebreaker *Glacier* from 7 to 27 March, 1986. The *Melville* was restricted to open-water locations, while the *Glacier* worked primarily in areas of significant ice cover (Fig. 1). Movement of

the ships was co-ordinated so that on two occasions we obtained nearly synoptic east-west transects through the marginal ice zone, with sampling completed in <46 h. One transect crossed the southern portion of the study area from 7 to 9 March (*Melville* Stas 2–10; *Glacier* Stas 1–7), and the second, the northern portion from 25 to 27 March (*Melville* Stas 43–53; *Glacier* Stas 24–28). Each transect was approximately 400 km in length. Additional stations were located along primarily these two main transect lines (Fig. 1). On 4 March a Polar Research Laboratory Argos-tracked buoy was deployed through a hole bored into a large multi-year ice floe (about 200 m in horizontal dimension and 2 m thick) in the extreme southwest corner of the study area (Fig. 1). Several position fixes were obtained per day throughout the month of March.

At each station we collected vertical profile data on temperature, salinity, nutrient concentrations ( $\text{NO}_3^-$ ,  $\text{NO}_2^-$ ,  $\text{NH}_4^+$ ,  $\text{Si}(\text{OH})_4$  and  $\text{HPO}_4^{2-}$ ), Chl *a*, particulate carbon, nitrogen and biogenic silica. Temperature and salinity data were obtained using a Neil Brown Mark IV CTD system on each ship. We made a total of 115 CTD casts, with nearly all extending to 500 m and more than half to 1500 m. The CTD system aboard *Glacier* was calibrated before and after the cruise at the Northwest Regional Calibration Center (Bellevue, Washington). The *Melville's* system was calibrated at Texas A&M University after the cruise. Temperature and salinity calibrations were obtained about every third cast, using a Niskin bottle and reversing thermometers aboard the *Glacier* and a rosette sampler on the *Melville*. The lowering rate of the CTD was limited to  $0.5 \text{ m s}^{-1}$  through the upper 150 m to optimize vertical resolution and minimize problems related to sensor lag time (such as spiking through the relatively strong vertical temperature and conductivity gradients present within this depth interval). Simultaneous casts taken when the ships were separated by only a few hundred meters at the start and end of the cruise indicate that the data from the two instruments agree to within  $0.01^\circ\text{C}$  and  $0.01\text{‰}$ .

We collected 10 l seawater samples at nine depths in the upper 150 m at each station. The upper seven depths corresponded to the approximate depths of 100, 50, 30, 15, 5, 1, and 0.1% of surface irradiance, as estimated from lowerings of a submersible quantum meter and a Secchi disk. The others provided equal sample spacing between the 0.1% isolume and 150 m. Samples from stations occupied at night were taken at depths selected from the most recent daylight station. Samples taken in areas of significant ice cover were collected at depths chosen from the most recent irradiance profile in an open lead or polynya. Each sample was drawn into a 20 l polypropylene carboy fitted with a spigot 2 cm above the bottom. Subsamples for individual analyses were drawn from this carboy. This sampling procedure ensures that rapidly sinking particles are adequately represented in all samples, as the carboys were routinely agitated during subsampling (CALVERT and McCARTNEY, 1979).

Duplicate samples for nutrient analysis were gently syringe filtered through  $0.6 \mu\text{m}$  pore-size polycarbonate membrane filters (Nuclepore, Inc.), frozen and stored for later analysis. One complete set of frozen samples from both ships were returned to the U.S. on each ship. Maximum–minimum thermometers placed in the freezers aboard both ships confirmed that the samples were returned without thawing. Samples were then thawed in a tepid water bath and analysed on a four-channel, continuous flow colorimetric analyser using methods adapted from those of ATLAS *et al.* (1971). Separate samples for ammonium analyses were filtered in a similar manner and analysed immediately at sea, using a modification of the method of SOLORZANO (1969). Standard

deviations for a single analysis were 0.4  $\mu\text{M}$  for nitrate and silicic acid, 0.03  $\mu\text{M}$  for phosphate, 0.013  $\mu\text{M}$  for nitrite and 0.07  $\mu\text{M}$  for ammonium.

Chlorophyll *a* analyses were performed on board by filtering 1.0 l seawater samples through Gelman A/E glass fiber filters. These filters were then homogenized in 90% acetone, extracted and centrifuged. The fluorescence of the extract before and after acidification was measured with a Turner Designs fluorometer which had been calibrated with commercially purified Chl *a* preparations.

Particulate carbon and nitrogen were measured by filtering 1.2 l of seawater through precombusted (450°C for 2 h) Whatman GF/C glass fiber filters. Each filter was rinsed with 5–10 ml of 0.01 N HCl in filtered seawater to remove any inorganic carbonates, then folded, placed in precombusted glass vials, dried at 60°C and returned to the University of Tennessee for analysis on a Perkin-Elmer model 240B elemental analyser. Corrections for absorption of dissolved organic carbon and nitrogen by the filter were made by subtracting values from a second filter placed beneath the primary filter and processed identically. These corrections were typically small, but not negligible; e.g. they averaged 6.7% for nitrogen. Approximate limits of detection were 0.5 and 0.1  $\mu\text{mol l}^{-1}$  for carbon and nitrogen, respectively.

Samples (1.2 l) for analysis of biogenic particulate silica were vacuum filtered through 0.6  $\mu\text{m}$  Nuclepore filters which were then folded, dried at 60°C in covered plastic Petri dishes and returned to Oregon State University for analysis. Biogenic silica was dissolved by NaOH digestion (PAASCHE, 1973), and the resulting solution analysed for dissolved reactive silicate by the acid-molybdate method of STRICKLAND and PARSONS (1972). To maximize sensitivity the sample and reagent volumes were reduced so that 4.0 ml of acidified molybdate solution and then 6.0 ml of metol/sulfite reducing reagent were added to neutralized digestion solutions whose volume was 10 ml. Absorbance at 810 nm was measured in low-volume (*ca* 9 ml) 10 cm spectrophotometer cells (Hellma, Inc.). Analysis of reagent blanks, filter blanks and low-level standards in our laboratory indicates an analytical precision of  $\pm 5\%$  and a detection limit of 10–15  $\text{nmol l}^{-1}$  for this method.

Bacteria were enumerated by DAPI (4,6,6-diamidino-2-phenylindole 2 HCl, Sigma) stained specimens collected on 0.2  $\mu\text{m}$  pore size Nuclepore filters, as described by GROSSI *et al.* (1984). Direct counts were performed using a Zeiss research microscope equipped with an ultraviolet light source (HBO Osram, 50W) and appropriate filters. Bacterial biovolumes were estimated from cell dimensions (ZIMMERMAN, 1977) and carbon biomass derived from biovolume using a conversion factor of 18.3  $\text{fmol C } \mu\text{m}^{-3}$  (BRATBAK and DUNDAS, 1984).

## RESULTS

The edge of the pack ice was sharply defined within the study area in March 1986, with the entire transition from open water to >80% ice cover generally occurring within 10 km or less (Fig. 1). Also, the ice edge maintained an almost stationary position throughout our 3 week sampling period. Both of these properties of the ice frontal zone were very different from those we had observed in the Weddell–Scotia Sea during austral spring, when the ice-cover gradient was much more diffuse and the entire ice-edge system was retreating southward at a mean rate of  $>5 \text{ km day}^{-1}$  (COMISO and SULLIVAN, 1986).

The upper water column was well stratified in salinity and density throughout the study area (Fig. 2). There was an upper mixed layer 30–50 m deep at every station occupied. Temperature in this layer varied from freezing (about  $-1.83^{\circ}\text{C}$ ) beneath the ice to warmer than  $-0.5^{\circ}\text{C}$  toward the eastern end of the study region. Surface salinity was between 33.3 and 33.9‰, with the lower values in the more northerly section and underlying the ice edge. A sharp halocline and resulting pycnocline underlay the upper mixed layer, with salinities increasing to about 34.4‰ at approximately 50 m. Salinity increased gradually with depth below this halocline. In contrast to the vertical salinity distribution, temperature in the open-water portion of the study area was lowest in a layer centered at about 80 m, in which temperatures were colder than  $-1.5^{\circ}\text{C}$ . Temperature increased with depth below this layer, with the strongest gradients between 100 and 200 m. The upper mixed layer comprises the Weddell Surface Water layer, which varies seasonally in its temperature and salinity characteristics. The underlying cold core is the Weddell Winter Water, a remnant from the preceding winter's cold convective layer. Beneath the Winter Water, temperature and salinity increase with depth in the upper portion of the Weddell Warm Deep layer. A statistically rigorous discussion of these water types has been presented by CARMACK (1977).

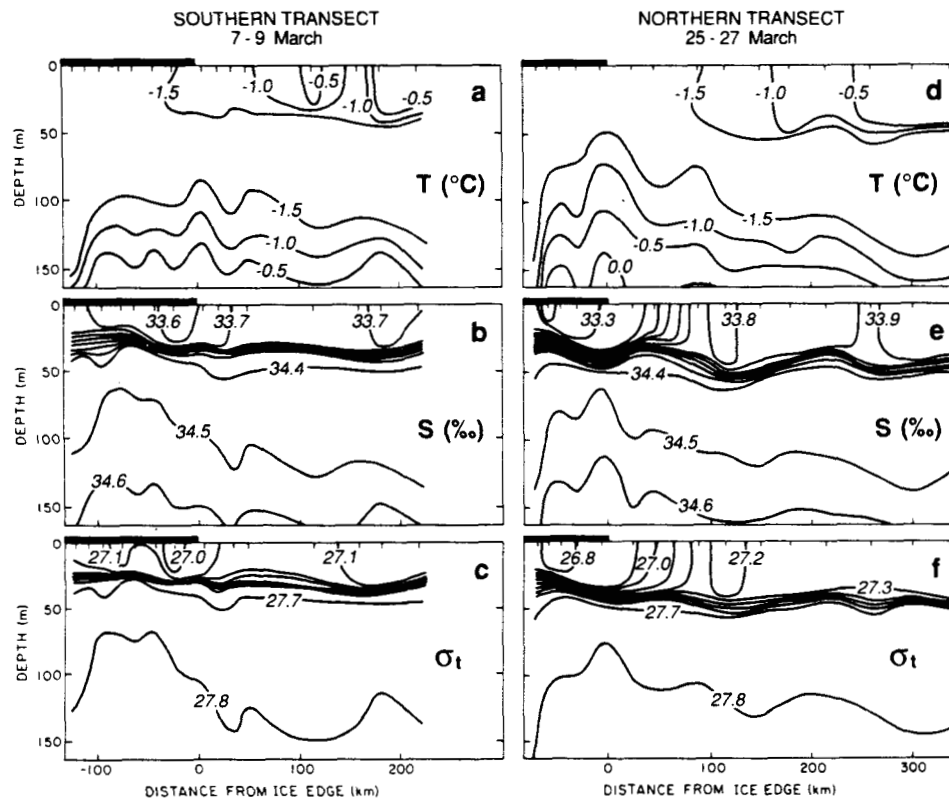


Fig. 2. Distributions of temperature, salinity and density ( $\sigma_t$ ) in the upper 150 m along the quasi-synoptic transects occupied on 7–9 March (a–c) and 25–27 March (d–f). The bar in the upper left of each panel denotes heavy (>80%) ice cover.

The dynamic topography of the surface relative to 500 db indicates that baroclinic circulation of the upper 500 m was confined to a barely significant eastward, then northward, surface drift in the western portion of the study area (Fig. 3). The surface dynamic topography relative to 1500 db (not shown) exhibits virtually the same pattern. These observations are consistent with past conclusions that the regional flow is primarily barotropic. The large multi-year ice floe which was instrumented for drift measurements near the start of the field program showed almost no net meridional drift and little net zonal drift during March (British Antarctic Survey, unpublished data).

The gross features of the vertical distributions of temperature and salinity did not appear to change significantly during March, as seen by comparing distributions early and late in the cruise (Fig. 2). The later, more northerly section showed lower salinities in the upper layer and a somewhat deeper upper mixed layer than the earlier southern transect. Air temperatures were sub-freezing for all but 4 days of the cruise, and the ice edge was in water colder than  $-1.0^{\circ}\text{C}$ , so it is unlikely that significant melting occurred during March. The location of the ice edge, as monitored both from the cruise vessels and by the Navy-NOAA ice forecast center using satellite data, remained effectively unchanged throughout the study, consistent with the absence of either net melting or net freezing. It is likely that the depressed salinities farther north represented a remnant of previous meltwater accumulation along the ice edge. Little variation occurred below the Surface Water layer, as temperature and salinity remained unchanged during the course of our observations. An elevation of the isohalines (and isopycnals) in the vicinity of the ice edge persisted through the cruise. While this feature may suggest upwelling, surface winds during the cruise were neither persistent enough nor in the proper direction to induce upwelling.

There was a consistent maximum in phytoplankton biomass centered about 150 km seaward of the ice edge (Figs 4a-d and 5a-c). Phytoplankton biomass maxima often have been observed in open water 50-200 km seaward of the ice edge in other areas (e.g. ALEXANDER and NIEBAUER, 1981; SMITH and NELSON, 1985; NELSON *et al.*, 1987). However, these previous studies have shown a high degree of spatial correlation, both vertically and laterally, between elevated phytoplankton biomass and low-salinity surface

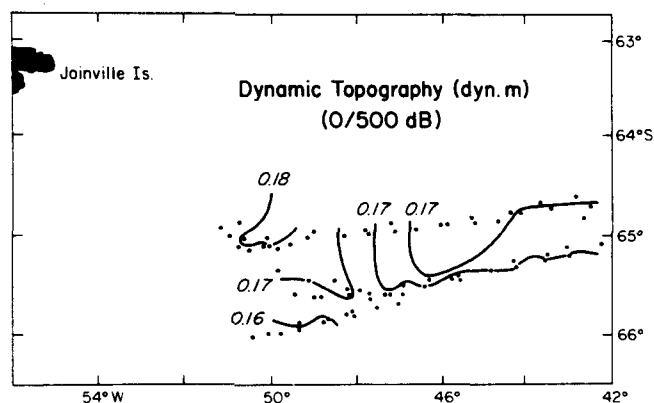


Fig. 3. Dynamic topography (in dyn m) relative to the 500 db surface within the study area. Almost no mean baroclinic flow is detectable.

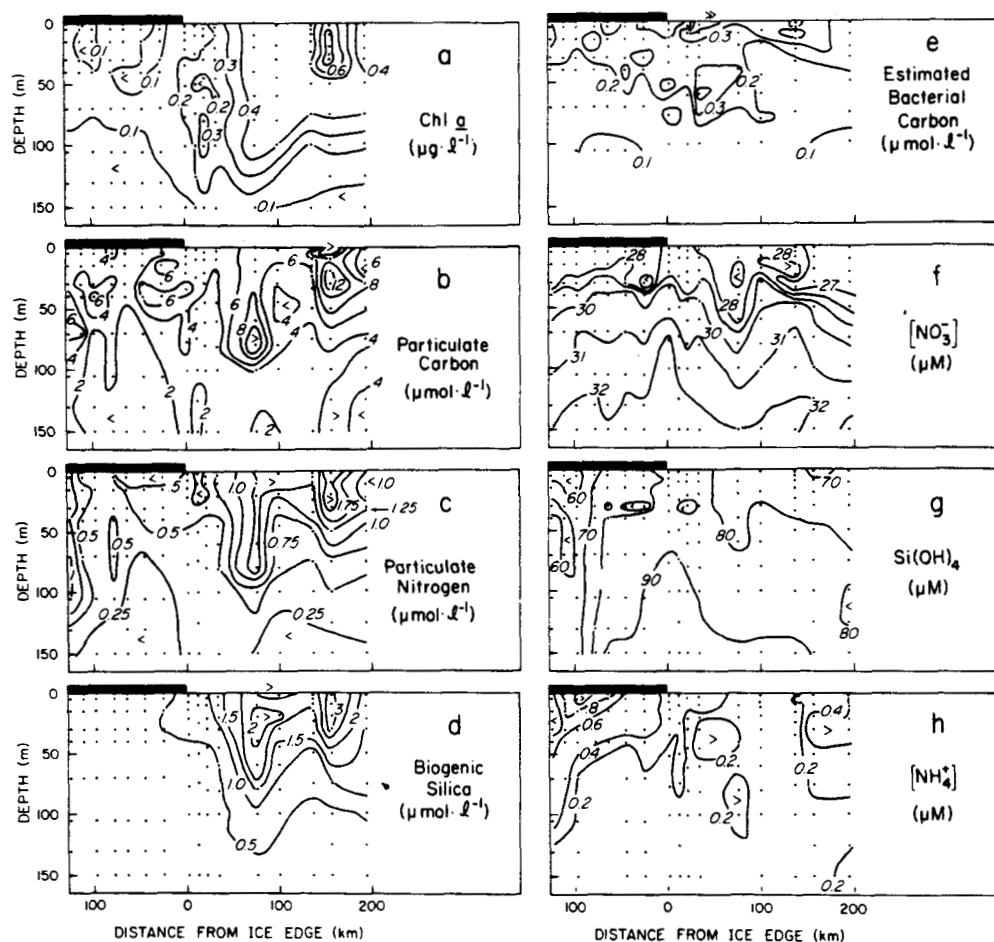


Fig. 4. Sections of phytoplankton biomass and nutrient concentrations in the upper 150 m along the transect across the southern portion of the study area, 7-9 March. (a) Chl *a*; (b) particulate carbon; (c) particulate nitrogen; (d) biogenic silica; (e) estimated bacterial carbon; (f) nitrate; (g) silicic acid; (h) ammonium. The bar in the upper left of each panel denotes heavy (>80%) ice cover.

water derived from ice melt. This was not the case in the Weddell Sea ice-edge zone during March 1986. All phytoplankton biomass parameters showed distinct mesoscale maxima 100-150 km in iceward-seaward extent (Figs 4a-d and 5a-c), while the hydrographic structure of the upper 100 m was dominated by a strong halocline/pycnocline at 30-50 m that was present throughout the ice-edge region and whose depth and intensity were nearly uniform laterally (see Fig. 2). In addition, all phytoplankton biomass parameters show distributions in which high biomass was not restricted to the stable surface layer, but instead penetrated into and through the main pycnocline.

There were two distinct relationships between the biogenic particle fields and the ice edge itself: the two biomass parameters whose only significant source is phytoplankton growth (i.e. Chl *a* and biogenic silica), decreased to very low levels at the ice edge and



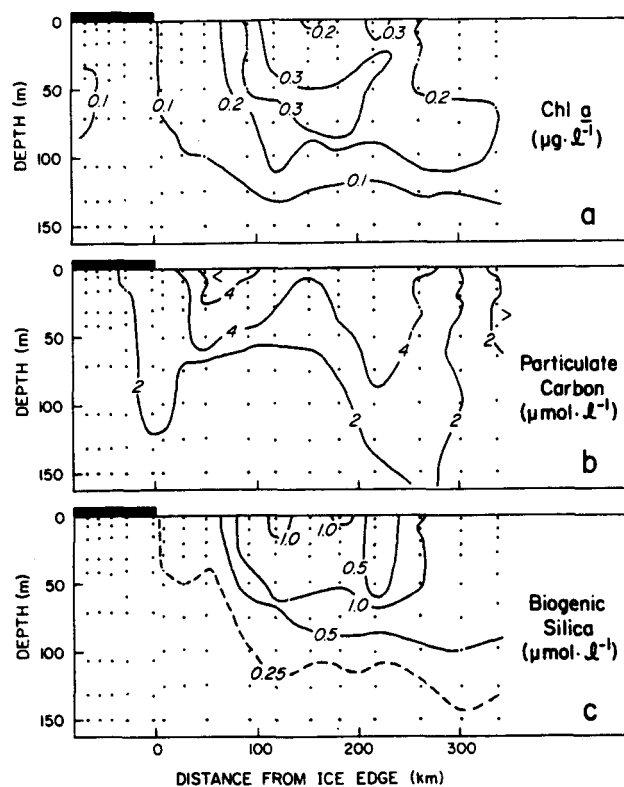


Fig. 5. Sections of selected phytoplankton biomass parameters in the upper 150 m along the transect across the northern portion of the study area, 25–27 March. (a) Chl *a*; (b) particulate carbon; (c) biogenic silica.

remained low throughout the ice-covered portion of the system (Figs 4a, d and 5a, c). In contrast, particulate carbon and nitrogen, which include both autotrophic and heterotrophic biomass as well as any organic detritus, did not decrease sharply at the ice edge and at times exhibited maxima in ice-covered waters (Figs 4b, c and 5b).

Chlorophyll *a* concentrations seldom exceeded  $0.7 \mu\text{g l}^{-1}$ , even in the core of the phytoplankton biomass maximum. Many previous studies of phytoplankton in ice-edge regions have reported Chl *a* concentrations  $>5 \mu\text{g l}^{-1}$ , with occasional values  $>20 \mu\text{g l}^{-1}$  (e.g. ALEXANDER and NIEBAUER, 1981; SMITH *et al.*, 1985; SMITH and NELSON, 1985; NELSON *et al.*, 1987). We have previously reported the spatial dimensions of ice-edge phytoplankton blooms in the Southern Ocean by defining “bloom” chlorophyll concentrations as those exceeding  $1.0 \mu\text{g l}^{-1}$  (NELSON and SMITH, 1986; NELSON *et al.*, 1987). By that definition, the much less intense maximum in Chl *a* observed in the ice-edge region during March 1986 was not strong enough to be characterized as a phytoplankton bloom.

In contrast to the relatively low Chl *a* levels, the highest particulate carbon concentrations exceeded  $10 \mu\text{mol l}^{-1}$  (Fig. 4b). These values are about half as high as those measured within a fairly intense (Chl *a*  $> 5 \mu\text{g l}^{-1}$ ) ice-edge bloom in the Weddell–Scotia Sea in the spring of 1983 (NELSON *et al.*, 1987). Biogenic silica concentrations were

generally  $>1.5 \mu\text{mol l}^{-1}$  and occasionally  $>3.5 \mu\text{mol l}^{-1}$  within the biomass maximum sampled in March (Fig. 4d). These values actually exceeded those measured in the spring 1983 ice-edge bloom (NELSON *et al.*, 1987).

Bacterial biomass generally decreased with depth, and the bacterial carbon distribution was similar to that of all phytoplankton biomass parameters in that high values penetrated well into, and beneath, the main pycnocline (Fig. 4e). However, unlike Chl *a* and biogenic silica, bacterial carbon was quite high in the surface layer in ice-covered waters. The estimated bacterial carbon concentrations (generally  $0.2\text{--}0.4 \mu\text{mol l}^{-1}$ ) ranged from about 2 to 10% of the total particulate carbon.

The phytoplankton biomass maximum was associated with localized minima in the concentrations of nitrate and silicic acid (Fig. 4f, g) as well as phosphate (not shown). This spatial correlation between phytoplankton maxima and nutrient minima is quite common in marginal ice zones of both northern and southern polar oceans (e.g. SMITH *et al.*, 1985; NELSON and SMITH, 1986; NELSON *et al.*, 1987) and indicates that the nutrient field is affected strongly by *in situ* uptake by phytoplankton. Most significantly, the high-biomass waters within and beneath the pycnocline also exhibited nitrate and silicic acid minima (Fig. 4). Since the temperature and salinity sections provide no evidence of vertical mixing through the pycnocline in these areas (Fig. 2), the nutrient sections suggest that active uptake of nitrate, phosphate and silicic acid by the phytoplankton extended through the pycnocline to at least 75 m, and perhaps deeper, in high-biomass areas seaward of the ice edge.

Surface waters were consistently richer in ammonium than were subsurface waters, both under the ice and in open-water areas (Fig. 4h). Maximum concentrations ( $0.8\text{--}0.9 \mu\text{M}$ ) occurred under the ice, whereas open-water stations exhibited ammonium concentrations no higher than  $0.6 \mu\text{M}$ . None of the ammonium concentrations measured approached those observed in subsurface maxima in the Ross Sea (BIGGS, 1982) or the Fram Strait marginal ice zone (SMITH *et al.*, 1985).

## DISCUSSION

### *Physical conditions*

During March 1986 the area we examined in the western Weddell Sea was characterized by negligible baroclinic circulation. The ice floe that had been instrumented so that its motion could be tracked using the Argos system was almost motionless through this period, suggesting that total as well as baroclinic flow was small. GORDON *et al.* (1981) suggested a hypothetical western boundary current, flowing northward along the Antarctic Peninsula, in order to close the cyclonic circulation in the Weddell Gyre. Their computed width for this boundary current was 225 km. Assuming that those computations are correct, our data, collected 100–500 km east of the continental slope of the Antarctic Peninsula, would have included the eastern portion of this current. It is difficult to reconcile the very weak observed circulation with the concept of a western boundary current, and it may be that the actual current is narrower or has lower speeds, at least in late summer.

A lack of significant current activity in the study area is consistent with the observed vertical temperature and salinity structure. Small steps (on the order of meters in vertical scale) occurred at all stations in the upper Warm Deep Water layer from about 100 to 180 m (HUSBY and MUENCH, 1988). Similar features have been reported for the same

general area by MIDDLETON and FOSTER (1980), who attributed them to double-diffusive processes. Below about 180 m, nearly half of the stations showed temperature and salinity steps with vertical scales exceeding 100 m. These also were reported for the Weddell Sea further east by FOSTER and CARMACK (1976), who attributed them to a combination of double diffusion and the cabelling instability. The widespread occurrence and persistence of such features suggest that lateral current shear and related turbulent mixing are small, and that the distributions of temperature and salinity may be controlled primarily by vertical, rather than horizontal processes.

The hydrographic structure of the upper layers near the ice edge was more uniform horizontally than has been observed in other ice-edge regions. The Bering Sea and Greenland Sea ice edges are accompanied by major oceanic frontal structures (MUENCH and SCHUMACHER, 1985; PAQUETTE *et al.*, 1985). The spring Weddell–Scotia Sea ice edge was associated with large mesoscale lateral variations in water characteristics (NELSON *et al.*, 1987). In each of these cases, the lateral variations were strongly influenced by regional oceanic circulation features. For example, the East Greenland Sea ice edge coincides with the East Greenland Polar Front, a mesoscale ocean circulation feature which interacts with, and exerts a strong control over, the ice edge. The complexity of the Weddell–Scotia Sea ice edge is forced by mesoscale oceanic features associated with the strong currents through the Drake Passage. No such features were observed in the northwestern Weddell Sea. The upper mixed layer extended, with little variation in depth, from well beneath the ice to more than 300 km seaward of the ice edge. Low salinity fronts within this layer (Fig. 2) may have been associated with localized circulations, but these features were not studied in detail.

#### *Biomass distributions*

When the particulate matter distributions in the two sections are compared, the overall spatial patterns of all parameters are similar (i.e. maxima seaward of the ice edge and a decrease under the ice; see Figs 4 and 5). However the first, and more southerly, transect exhibited maximum levels of all biomass parameters about twice as high as those observed on the second. The first transect (Fig. 4) was sampled 2–3 weeks earlier, and was about 80 km farther south, than the second. It is possible that the differences we observed between the two transects resulted from biological removal processes that were degrading the bloom with time. However, EL-SAYED and TAGUCHI (1981) have found marked north–south trends in productivity and biomass in the Weddell Sea, over distances of several hundred km. A similar meridional distribution of biomass appears to have existed during the time of our study. Satellite ocean-color data show that the ice-edge phytoplankton biomass maximum we observed between 64° and 65°30'S extended southward to approximately 72°S and exhibited its highest chlorophyll concentrations well to the south of our study area (COMISO *et al.*, 1987). Data on vertically integrated Chl *a* within our study area also provide some indication of a north–south gradient in phytoplankton biomass that persisted through March (Fig. 6). Considering this meridional biomass gradient in the context of the larger-scale synoptic view provided by satellite data suggests that our study area was located near the northern limit of an extensive ice-edge phytoplankton bloom that was significantly more intense 300–800 km to the south.

The observed lack of close correlation, either vertically or horizontally, between the upper-layer density structure and the distribution of biogenic particulate matter was

unexpected, and differs sharply from previous observations. In most earlier studies of marginal ice zones, any observed increase in phytoplankton biomass was strongly related to vertical stratification (e.g. ALEXANDER and NIEBAUER, 1981; SMITH and NELSON, 1985), although in some situations the previous light regime of the water masses at the ice edge (NELSON *et al.*, 1987) or nutrient concentrations (SMITH *et al.*, 1985) also had influenced the development of the ice-edge bloom. In the Southern Ocean, where nutrients are almost always present at non-limiting concentrations, vertical stability has been implicated as the major factor in controlling primary productivity (e.g. SAKSHAUG and HOLM-HANSEN, 1984; SMITH and NELSON, 1986). In March of 1986 the entire marginal ice zone of the northwestern Weddell Sea exhibited strong vertical stability in the upper 30–50 m, and all measured nutrient concentrations were high enough to have supported vigorous phytoplankton growth. A phytoplankton biomass maximum was present, but was both less intense and less correlated spatially with the physical structure than has been the case in ice-edge phytoplankton blooms at other times of year. This implies that some factor other than vertical stability and nutrient availability was playing a major role in limiting the intensity and spatial extent of the bloom. Possibilities include:

*Light limitation.* Total daily irradiance in March ranged from 5.8 to 41 Ein m<sup>-2</sup> day<sup>-1</sup>, with a mean of 16.0. This is considerably less light than has been measured in other ice-edge regions during phytoplankton blooms. For example, the mean integrated daily irradiance in the Weddell–Scotia Sea during November 1983 was 25.7 Ein m<sup>-2</sup> day<sup>-1</sup>, and that in the western Ross Sea during late January/early February 1983 was approximately 70. SooHoo *et al.* (1987) reported spectral irradiance data and photosynthesis vs irradiance response curves which suggest that phytoplankton photosynthetic rates were limited by light through 20–80% of the euphotic zone in our study area. Thus it is possible that the magnitude of the bloom we observed was limited by light availability. However, light limitation provides no obvious explanation of the lack of spatial correlation between phytoplankton biomass and vertical stability (Figs 2 and 4) or for the appearance of large ice-edge blooms well to the south of our study area (COMISO *et al.*, 1987).

*Micronutrient availability.* While biologically utilizable forms of nitrogen, phosphorous and silicon were present in high concentrations in the surface layer throughout the study area, the concentrations and chemical speciation of trace metals and other potentially limiting micronutrients were not measured. Recently, particular emphasis has been given to the possibility of iron limitation in the Southern Ocean, because of the absence of identifiable sources of iron to the water column (MARTIN and FITZWATER, 1988). A strongly stratified surface layer, late in the productive season, may be an excellent place to find depletion of iron or other micronutrients to levels that limit phytoplankton growth if limitation of this sort does in fact occur in the ice-edge zone.

*Grazing or other removal processes.* It is possible that the Weddell Sea ice-edge zone in March 1986 was no less favorable an environment for phytoplankton growth than were other marginal ice-zone systems that exhibited high phytoplankton biomass. Biomass-specific rates of <sup>14</sup>C and <sup>15</sup>N uptake by phytoplankton in our study area (SMITH and NELSON, 1987; NELSON and SMITH, 1987) were comparable to those in more intense phytoplankton blooms, lending support to the idea that this area was characterized by phytoplankton growth no less vigorous than that occurring in ice-edge zones in spring and summer. If this is the case, then the lower biomass levels may have resulted from intensified grazing pressure or other removal processes. Standing stocks of most herbivorous macrozooplankton appear to have been too low to exert major control over the

phytoplankton biomass (HOPKINS and TORRES, in press), but microzooplankton biomass ranged from about 20 to 60% of the estimated phytoplankton biomass and showed a positive spatial correlation with Chl *a* (GARRISON and BUCK, 1987). This estimated ratio of microzooplankton biomass to phytoplankton biomass is about four times as high as that observed in the spring 1983 ice-edge bloom in the Weddell–Scotia Sea (GARRISON and BUCK, 1987), so it is likely that grazing by microzooplankton played a greater role in controlling phytoplankton biomass in the ice-edge zone during March 1986 than has been the case in some other ice-edge systems.

Another possible removal term is grazing by krill; net tows and acoustic sampling during our cruise provide no evidence of high krill abundance in open-water areas of the northwestern Weddell Sea during March (DALY *et al.*, 1987), yet gut-content analysis of seals and seabirds in the study region indicates that krill comprised a major portion of the diet of these very abundant carnivores (e.g. AINLEY *et al.*, in press). Clearly the use of krill as a major food item does not necessarily imply high krill stocks. Nonetheless, the direct and indirect sampling methods do lead to somewhat different conclusions regarding the abundance of krill within the study area. Juvenile krill were observed in high abundance on the under surface of ice floes (DALY *et al.*, 1987), and it is possible that the combination of large standing stocks of krill and low phytoplankton biomass resulted in the ammonium concentration maximum observed under the ice (Fig. 4h). If krill also were present in high concentrations seaward of the ice edge, they could have exerted significant control on the phytoplankton biomass.

We cannot assess any one of these possibilities in detail on the basis of data collected in 1986. It appears, however, that some factor, or combination of factors, other than water-column stability and macronutrient availability was instrumental in controlling both the spatial distribution and the overall level of phytoplankton biomass in the ice-edge zone of the northwestern Weddell Sea during the late summer/early autumn of 1986.

#### *Particulate matter composition*

The biogenic particle field we sampled in March was very different in its chemical composition from that observed in the Weddell–Scotia Sea during the austral spring (NELSON *et al.*, 1987). In that system, Chl *a* concentrations were about an order of magnitude higher than those reported here, particulate carbon and nitrogen about twice as high, and biogenic silica nearly the same. The elemental ratios and carbon:chlorophyll ratios observed in spring were similar to those of compositionally normal phytoplankton (NELSON *et al.*, 1987). The ratios encountered during this study (Table 1) were highly unusual for particulate matter in the oceanic surface layer, but similar to those observed in the ice-edge zone of the western Ross Sea during January–February 1983 (SMITH and NELSON, 1985). During that study the bloom was very dense, with all biomass parameters approximately 10 times higher than those we report here. The high overall biomass levels within the Ross Sea bloom and the anomalous composition of that material combined to produce biogenic silica concentrations greater by a factor of three than any which had been observed previously in surface seawater (NELSON and SMITH, 1986). If the compositional ratios reported here are not greatly biased by detrital material, this is the second reported observation of a biomass maximum in the marginal ice zone of the Southern Ocean that is unusually rich in organic carbon and exceptionally rich in biogenic silica per unit chlorophyll. These direct observations of biogenic particulate matter that is rich in silica are consistent with previously reported large-scale patterns of vertical nutrient

Table 1. Mean values of selected phytoplankton biomass parameters in three marginal ice zones of the Southern Ocean

Parameter	Weddell-Scotia Sea November 1983*	Western Ross Sea January-February 1983†	Western Weddell Sea March 1986‡
Chlorophyll <i>a</i> ( $\mu\text{g l}^{-1}$ )	4.0	2.9	0.38
Particulate carbon ( $\mu\text{mol l}^{-1}$ )	10.6	33.2	3.6
Particulate nitrogen ( $\mu\text{mol l}^{-1}$ )	1.5	5.6	0.46
Biogenic silica ( $\mu\text{mol l}^{-1}$ )	1.4	21.7	1.6
Carbon:chlorophyll mass ratio	31.8	138	114
Carbon:nitrogen mole ratio	7.1	5.9	7.8
Silica:carbon mole ratio	0.13	0.65	0.44

\*Data from NELSON *et al.* (1987).

†Data from NELSON and SMITH (1986).

‡Present study; The data set includes all values from the surface to the 1% light depth at locations seaward of the ice edge.

concentration gradients in the upper ocean, which suggest that the Southern Ocean is the only geographically extensive region in which the net removal of silicic acid from the surface layer significantly exceeds that of nitrate (KAMYKOWSKI and ZENTARA, 1985). High ratios of biogenic silica to organic carbon in the phytoplankton also may help explain the observation that the Southern Ocean exhibits very high accumulation rates of siliceous sediments (DEMASTER, 1981; LEDFORD-HOFFMAN *et al.*, 1986) in spite of relatively low photosynthetic productivity (HOLM-HANSEN *et al.*, 1977; SMITH and NELSON, 1986).

In the Ross Sea ice-edge bloom very little organic or siliceous detritus was observed microscopically (SMITH and NELSON, 1985),  $^{14}\text{C}$  autoradiography indicated that nearly all physically intact cells were photosynthetically active (WILSON *et al.*, 1986), and the relative uptake rates of carbon, nitrogen and silicon closely paralleled the observed, highly anomalous, elemental ratios of the suspended particle assemblage (NELSON and SMITH, 1986). These lines of evidence all suggest that the unusual elemental composition of the suspended particulate matter reflected the composition of the phytoplankton rather than a detrital influence. In the present study we have less evidence on the relative proportions of living and detrital particulate matter. Bacterial biomass made only a small (2–10%) contribution to the total particulate carbon pool (Fig. 4b, e), but protozoan microzooplankton biomass may have comprised up to 30% of the total suspended particulate carbon at some stations (GARRISON and BUCK, 1987). Carbon:nitrogen ratios were similar to those observed in phytoplankton cultures growing vigorously with excess nitrate (PARSONS *et al.*, 1984), implying that living phytoplankton comprised a high fraction of the particulate organic matter. Finally, microscopic examination of preserved whole water samples did not indicate large amounts of either organic or siliceous detrital material, and cell counts from selected stations within the bloom indicate that the diatom *Dactyliosolen tenuijunctus* dominated the phytoplankton assemblage, both numerically and on a biovolume basis (K. R. BUCK, personal communication). This diatom was observed to have very little of the volume within the frustule occupied by photoplasm and to contain very little autofluorescent material, including Chl *a*. This cell structure of a dominant diatom within the study area is consistent with our measurements of high biogenic silica levels per unit Chl *a*. From the several lines of evidence detailed above we conclude that the particulate matter within the biomass

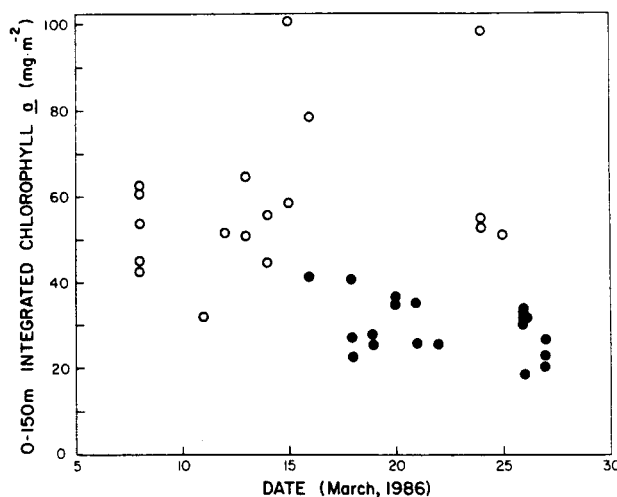


Fig. 6. Integrated Chl *a* content of the upper 150 m at open-water locations, as a function of time during March 1986. Open circles denote stations occupied along the more southerly transect line; solid circles represent stations along the more northerly line. The mean Chl *a* content for all southerly stations was  $50.7 \text{ mg m}^{-2}$  (S.D. = 22.6) while the mean for all northerly stations was  $26.9 \text{ mg m}^{-2}$  (S.D. = 8.2).

maximum probably consisted mostly of living phytoplankton cells, with a smaller but significant fraction of living microzooplankton, and that the phytoplankton was unusually rich in silica in comparison with its organic matter and pigment content.

The approximate amount of detrital particulate material can be estimated in two other ways. Figure 7 presents the vertically integrated particulate carbon and biogenic silica content of the upper 150 m, plotted as a function of vertically integrated Chl *a*. When the regressions are done point-by-point on data from individual depths rather than using vertically integrated biomass, the relationships are similar (Table 2). We have illustrated the integral regressions in Fig. 7 because the very large number of data points from individual depths result in dense clouds of points from which individual values cannot be discerned. To a first approximation the value of the *y*-intercept in these regressions is an estimate of the amount of detrital, or at least non-chlorophyllous, organic or siliceous material present at stations exhibiting low Chl *a* concentrations and the slope of the regression line estimates the carbon:Chl *a* or silica:Chl *a* ratio of the phytoplankton assemblage. The *y*-intercept of  $217 \mu\text{mol C m}^{-2}$  for vertically integrated data implies an average of about  $1.5 \mu\text{mol l}^{-1}$  detrital C in the upper 150 m. Biogenic silica shows a higher degree of spatial correlation with Chl *a* than does POC in both the point-by-point and station-by-station regressions (Table 2), and in both cases the *y*-intercept for biogenic silica is statistically indistinguishable from zero. This tends to confirm the conclusion derived from microscopic examination that there was very little detrital siliceous material in the study area. Taken together, these regressions suggest that the silica:carbon mole ratio of the living material (i.e. after correcting for detrital carbon) was at least (4.15/8.07) or 0.51. This is four times as high as the mean silica:carbon ratio found for diatoms in nutrient-replete cultures (BRZEZINSKI, 1985) and comparable to the very high ratio observed in the Ross Sea (SMITH and NELSON, 1985). Thus these regressions suggest a

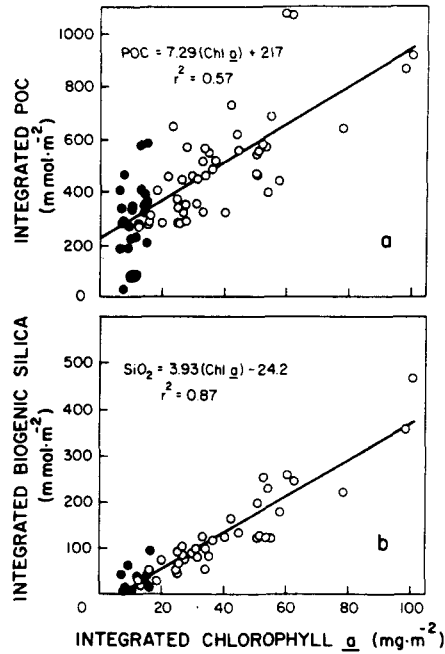


Fig. 7. Integrated particulate organic carbon (a) and biogenic silica (b) content of the upper 150 m, as a function of vertically integrated Chl *a*. Closed circles represent ice-covered locations, open circles areas with little or no ice cover. Linear regression lines (see Table 2) are also shown.

Table 2. Statistical parameters of linear regressions of particulate organic carbon and biogenic silica vs Chl *a* in the ice-edge zone of the northwestern Weddell Sea during March 1986

	Particulate organic carbon	Biogenic silica
(a) From regression of individual data points:		
Slope (mol g Chl $a^{-1}$ )	8.07	4.15
Equivalent mass ratio	96.8	249
Intercept at zero Chl <i>a</i> ( $\mu\text{mol l}^{-1}$ )	1.70	-0.12
Correlation coefficient ( $r^2$ )	0.55	0.81
(b) From regression of vertically integrated data:		
Slope (mol g Chl $a^{-1}$ )	7.29	3.93
Equivalent mass ratio	87.5	236
Intercept at zero Chl <i>a</i> ( $\text{mmol m}^{-2}$ )	217	-24.2
Correlation coefficient ( $r^2$ )	0.57	0.87

minimum detrital carbon value on the order of  $1.5 \mu\text{mol l}^{-1}$ , negligible siliceous detritus and living phytoplankton with unusually high carbon:chlorophyll and silica:carbon ratios.

A second estimate of detrital carbon can be made by summing the estimated carbon content of the identified biomass pools that would be collected by a 10 l sampling bottle (phytoplankton, microzooplankton and bacteria) and subtracting that total from the measured particulate carbon concentration (ROBINSON *et al.*, 1987). These estimates of detrital carbon range from 1.2 to  $3.3 \mu\text{mol l}^{-1}$ , in reasonably good agreement with the  $>1.5 \mu\text{mol l}^{-1}$  estimate derived from the carbon:Chl *a* regressions. The organic carbon



mass-balance approach also indicates that the amount of detrital organic material varied much less with distance from the ice edge than did phytoplankton biomass. Thus detritus appears to have been a major component of the suspended particle assemblage beneath the ice and for several kilometers seaward of the ice edge but the biomass maximum centred 50–150 km seaward of the ice edge contained a much higher fraction of living material.

#### *Seasonal duration of ice-edge phytoplankton blooms*

The growing body of data on phytoplankton biomass in the marginal ice zone of the Southern Ocean indicates that ice-edge blooms are quite common spatially and quite persistent seasonally. Two previous sets of observations from the Weddell Sea have indicated that such blooms begin no later than October or November (JENNINGS *et al.*, 1984; NELSON *et al.*, 1987), and data reported here show elevated biomass levels in the ice-edge zone through late March. If this is generally true, then ice-edge blooms persist much later into the summer/autumn season than does active retreat of the ice edge. Ice-edge phytoplankton blooms may have even greater seasonal duration than these data indicate. Preliminary observations from the eastern Weddell Sea aboard the F.S. *Polarstern* in 1986 have shown elevated chlorophyll levels in the ice-edge region during late August (DIECKMANN, 1987), a time of year when the ice edge in the Weddell Sea is at its northern limit (*ca* 58°S) and thus receiving much more light than would be the case at more poleward locations. Those observations combine with ours to show that there are at least some areas of elevated phytoplankton biomass in the marginal ice zone of the Southern Ocean from at least late August to late March, a much longer time than has been considered in seasonal productivity estimates for the Southern Ocean. Thus, recent estimates of the importance of ice-edge blooms in the overall annual productivity of the Southern Ocean (SMITH and NELSON, 1986; SMITH *et al.*, in press) must be conservative because they consider only retreating ice edges. Since these estimates attribute 35–50% of the total annual primary productivity of the Southern Ocean to phytoplankton blooms in the ice-edge regions, our observation of elevated biomass persisting through March underscores the major quantitative importance of the ice-edge region in the ecological and biogeochemical cycles of the Southern Ocean.

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