Seasonal patterns of water column particulate organic carbon and fluxes in the Ross Sea, Antarctica

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Abstract

The standing stock of particulate organic carbon (POC) was determined during five cruises in the Ross Sea in 1996 and 1997 and compared with primary production of carbon measured in short-term 14C-incubations and the flux of organic carbon collected in moored sediment traps. POC concentrations were estimated from transmissometer profiles that were calibrated with discrete POC bottle samples from each cruise. The mean standing stock of POC integrated to a depth of 100 m and averaged along a 330 km transect at 76.5°S in mid-October (early spring) was only 240 mmol C m⁻², but more than doubled to 560 mmol C m⁻² 10 days later. By mid-January (summer) the standing stock had increased by an order of magnitude to ~ 5300 mmol C m⁻², but dropped to 3500 mmol C m⁻² one week later. By late April (autumn), the standing stock was only 200 mmol C m⁻². The following spring the standing stock increased from 700 mmol C m⁻² in late November to 2200 mmol C m⁻² in early December. Despite the high standing stock in the photic zone in summer, 1997, little POC was collected in the moored sediment traps until late summer (February-March) when the traps showed an increase in POC and silica flux. A three-fold increase in POC flux occurred in autumn (March-April) dominated by pteropods, but the standing stock of POC in the photic zone at that time was very low. Light-scattering sensor data suggest that, although present in all seasons, aggregates were most abundant in autumn and were distributed throughout the water column. These aggregates may have temporarily stored POC and provided food support for a pteropod population that died and settled into the traps in March-April. Still, the trap POC...
flux was only 5% of the peak standing stock. Resuspension and lateral advection of recently settled organic matter from a nearby topographic high may explain the larger flux measured in the deep sediment traps, a flux that continued into winter. © 2000 Elsevier Science Ltd. All rights reserved.

1. Introduction

A general circulation model indicates that the Southern Ocean may significantly affect and be affected by global climate change (Sarmiento et al., 1998), suggesting that it is important to understand basic oceanographic processes in this area and to monitor environmental changes as possible indicators of global climate change. The Ross Sea, one of the major shelf seas around Antarctica, has been shown to have large seasonal variations in surface productivity and biogenic matter concentrations in the water column as a result of extremely intense seasonal phytoplankton blooms (Smith and Nelson, 1985; DeMaster et al., 1992; Sullivan et al., 1993; Arrigo and McClain, 1994; Smith et al., 1996). These blooms occur in seasonally stratified waters and consist of two major types (Arrigo et al., 1998). The first is dominated by the haptophyte Phaeocystis antarctica and is commonly found in the south-central Ross Sea. The second is dominated by diatoms, which commonly bloom in the western and eastern portions of the southern section (DiTullio and Smith, 1996).

Concentrations of biogenic material generated by these blooms greatly exceed the mean concentrations found in other areas off the continental shelf of Antarctica (Sullivan et al., 1993; Comiso et al., 1993). Phytoplankton blooms occurring over much of the southern Ross Sea (Arrigo and McClain, 1994) result in marked removal of nutrients and carbon dioxide (Sweeney et al., 2000). Measurements of the flux of organic carbon from surface waters to depth in the Ross Sea also show large seasonal and spatial variations (Dunbar et al., 1998).

Flux from surface waters appears to be a function of the phytoplankton assemblage, which in turn strongly controls the food web structure (Smith and Dunbar, 1998). Hence, in areas where P. antarctica dominates, export from the euphotic zone appears to be regulated by the sinking of large aggregates, whereas in regions dominated by diatoms, the flux is dominated by fecal pellets produced by mesozooplankton grazing. It has been suggested that the spatial determinant of assemblage composition is mixed-layer depth (Arrigo et al., 1999), but others (Smith et al., 2000; Sweeney et al., 2000) have suggested that additional factors are also important in determining the community structure of the surface layer.

One of the objectives of the US Southern Ocean JGOFS study (Antarctic Environment and Southern Ocean Process Study — AESOPS) is to obtain increased seasonal resolution for all carbon pools and to conduct simultaneous measurements of primary production, standing stocks of POC throughout the entire water column, and particle fluxes at the same location over seasonal time scales in the Ross Sea. Results of the AESOPS primary productivity and phytoplankton biomass measurements have been analyzed by Smith et al. (2000). For this analysis we increased the vertical resolution of
POC estimates by calibrating optical profiles of beam attenuation with discrete POC samples (Gardner et al., 1993; Gundersen et al., 1998; Claustre et al., 1999) in order to increase the sampling density for POC. In addition, we compare the change in standing stock of POC with the fluxes measured by sediment traps moored nearby. The enigma is that while the increase and decrease in POC standing stock follows the climatological data, the summer flux measured by sediment traps in this program (Collier et al., 2000) was much lower than previously reported (Dunbar et al., 1998; Asper and Smith, 1999). Furthermore, the maximum trap flux measured by Collier et al. (2000) occurred during winter after the Ross Sea was covered with ice. It appears, however, that this anomalous flux pattern of lower-than-average POC flux in summer and higher than average POC flux in the autumn was recorded in other traps deployed in the Ross Sea during the same time (Collier et al., 2000; R. Dunbar, personal communication). We examine the POC and other sets of data to look at seasonal trends to search for possible explanations of the anomalous trap fluxes.

2. Methods

2.1. Ross Sea cruises

During 1996 and 1997 we collected data during four process cruises covering three seasons on the R.V. I.B Nathaniel B. Palmer as part of AESOPS (Table 1; Fig. 1). Sampling was concentrated along a transect consisting of eight stations in the southern portion of the Ross Sea along 76°30’S, in a region that was expected to be accessible during all cruises. In addition to the four AESOPS cruises, data were collected along the 76°30’S line during a cruise of the ROAVERRS program (Research on Ocean — Atmosphere Variability and Ecosystem Response in the Ross Sea), as well as many other stations around the Ross Sea. The first AESOPS cruise (NBP96-04A) was timed to occur during the very early spring at the beginning stages of spring phytoplankton growth. There was still extensive ice cover, but frequent open areas occurred as well (Smith et al., 2000). The ROAVERRS cruise (NBP96-06) occurred in early summer when growth rates were high and surface biomass was increasing. The third cruise (NBP97-01) was during austral summer when biomass was high, with no ice coverage along the transect. Cruise four (NBP97-03) sampled during the...
Fig. 1. Map of sampling stations in the Ross Sea. MS6 and MS7 are sediment trap mooring sites of Collier et al. (2001). B is a sediment trap mooring site reported in Dunbar et al. (1998). F is a sediment trap mooring site reported in Accornero et al. (1999). Only the 400, 500, 1000, 2000 and 3000 m contours are included.

autumn–winter transition when ice coverage was complete and surface biomass was low. The fifth AESOPS cruise (NBP97-08) occurred the following spring during the period of highest productivity, but before biomass had reached peak levels, the same period as the 1996 ROAVERRS cruise. Ice coverage was initially high during this cruise but decreased rapidly.

During each cruise a SeaTech transmissometer and light-scattering sensor (LSS) were interfaced with a SeaBird 911+ CTD to make hydrographic and optical profiles at each station. The CTD rosette included 24 10-l Niskin bottles to collect water samples for nutrients and biomass. Samples for particulate organic carbon (POC) were collected in the upper 200 m on each cruise.

2.2. Analytical methods

All particulate carbon concentrations were determined according to JGOFS protocol (JGOFS, 1996) by filtration of known volumes through combusted (450°C for 2 h)
25-mm Whatman GF/F filters under low (< 0.5 atm) vacuum. Samples were drained from the Niskin bottles into polyethylene bottles and filtered using Gelman filtration funnels. The bottles were inverted on the filtration racks to prevent aerosol contamination. The filters were rinsed with ca. 5 ml of 0.01N HCl in seawater added prior to the completion of the filtration to remove inorganic carbonates, after which they were placed in combusted glass tubes, capped with combusted aluminum foil, and dried at 60°C. Time from sample collection to filtration end was generally ca. 1 h or less. Replicates from the same Niskin bottle varied by ± 5%. Filtration volumes ranged from 0.050–2.0 l and were adjusted according to the particulate load in the water. Moran et al. (1999) have demonstrated that small-volume POC concentrations may be overestimated at low concentrations (1–2 μmol C l⁻¹) due to absorption of DOC onto filters. However, most of our samples had high concentrations of POC (up to 100 μmol C l⁻¹), so the magnitude of this error relative to the particulate signal was small. POC determined directly as above and from TOC–DOC measurements (Carlson et al., 2000) were highly correlated in this high-POC system. Blanks were combusted filters that had been immersed in seawater and treated with acid. The total signal of blanks generally represented less than 2% that of a sample from the euphotic zone. All samples were returned to the laboratory and processed using a Carlo-Erba Model 252 elemental analyzer. Acetanilide was used as a standard.

Beam attenuation was determined using a SeaTech 25-cm pathlength transmissometer (660 nm) interfaced with a SeaBird SBE911+ CTD (see Gardner et al., 1993, 1995 for details). Beam attenuation is the sum of attenuation due to water, particles and colored dissolved organic matter (CDOM); thus, \( c = c_w + c_p + c_{CDOM} \). At 660 nm \( c_{CDOM} \) is generally negligible (Bricaud et al., 1981); \( c_w \) is constant and can be subtracted from all values of \( c \) to obtain \( c_p \). All CTD data were routinely binned into 2-dB intervals.

Attenuation due to particles (\( c_p \)) has been shown to be linearly correlated to the particulate matter (PM) concentration (Zaneveld, 1973; Baker and Lavelle, 1984; Gardner et al., 1985, 1995; Moody et al., 1987). Thus, a regression of beam attenuation versus particulate matter concentration yields a linear equation of the form

\[
 c_p = K \times [PM].
\]

In the surface waters of the Ross Sea almost all of the particulate matter is biogenic. Therefore, a good linear correlation between \( c_p \) and POC could be expected, as has been obtained in other areas (Gardner et al., 1993; Gundersen et al., 1998). Thus,

\[
 c_p = K \times [POC].
\]

Using this equation makes it possible to estimate the POC concentrations from the optical parameter beam \( c \) at the same vertical resolution of the CTD measurements (to 1–2 m), which is much greater than is practical by discrete bottle sampling. \( K \), the slope of the regression, should be empirically determined for each region or season investigated (Baker and Lavelle, 1984; Gardner et al., 1985).

Concentrations of particulate organic carbon were correlated with the calibrated transmissometer data. Model II linear regressions, where deviations between data points and the regression line are minimized in both the \( x \) and \( y \) dimensions rather
Fig. 2. Relationship of beam \( c_p \) and POC concentration. The Model II linear regression line is for the summer cruise (NBP9701; \( \text{POC} = 635^*c_p; r^2 = 0.818 \)). The late autumn data covered such a small range that Model II linear regression for summer was used to calculate POC concentrations. See Table 2 for regressions during each cruise.

Table 2
Slopes and intercepts of \( c_p \) to POC correlations for each cruise

<table>
<thead>
<tr>
<th>Cruise</th>
<th>Slope</th>
<th>Intercept</th>
<th>( R^2 )</th>
<th>( n )</th>
<th>POC range</th>
</tr>
</thead>
<tbody>
<tr>
<td>NBP96-04A</td>
<td>0.02028</td>
<td>0.0128</td>
<td>0.7104</td>
<td>234</td>
<td>0.79–10.94</td>
</tr>
<tr>
<td>NBP96-06</td>
<td>0.033</td>
<td>0.0427</td>
<td>0.9247</td>
<td>69</td>
<td>1.80–81.50</td>
</tr>
<tr>
<td>NBP97-01</td>
<td>0.01896</td>
<td>0.0996</td>
<td>0.8176</td>
<td>320</td>
<td>3.35–108.13</td>
</tr>
<tr>
<td>NBP97-03</td>
<td>0.006*</td>
<td>0.0105</td>
<td>0.0647</td>
<td>101</td>
<td>1.20–5.91</td>
</tr>
<tr>
<td>NBP97-08</td>
<td>0.02316</td>
<td>0.0477</td>
<td>0.8777</td>
<td>339</td>
<td>0.62–155.75</td>
</tr>
</tbody>
</table>

*A slope and intercept from NBP96-01 was used to estimate POC because of the low correlation for the data.

than just \( x \) or \( y \), were used because there was uncertainty in both variables. The regressions yielded relationships between attenuation and POC (Fig. 2, Table 2)

\[
[\text{POC}] = \left(1/K\right)*c_p. \quad (3)
\]

The slopes and intercepts in Table 2 were used to estimate the POC concentrations throughout the Ross Sea. There were areas of the Ross Sea where diatoms were the dominant phytoplankton and others were \( \text{Phaeocystis} \) dominated. There was no difference in the correlation between POC and \( c_p \) between these areas, so all data for a cruise were combined. Our slopes were also close to those measured in the Arabian Sea (Gundersen et al., 1998) and Equatorial Pacific (Claustre et al., 1999). Integrations of POC were made using 2-m resolution, so that a 100-m integration used 50 points.

While the transmissometer measures low-angle forward scattering and absorption at 660 nm, the LSS measures the backward scattering of light at 880 nm. Because the
LSS signal does not include absorption, it is less sensitive to light-absorbing, chlorophyll-rich particles in surface waters, but the LSS is more sensitive to changes in particle size. Both instruments detect the occurrence of large particles, such as aggregates, which are manifested as spikes in the raw, unaveraged data. The sensing area of the LSS varies with particle concentration, but is on the order of 4 l in clear water, whereas the volume of the transmissometer beam is 44 cm$^3$. Because of the larger sensing volume, spikes in the LSS data more accurately reflect the presence of aggregates than spikes in the transmissometer data. In the routine data smoothing using SeaBird’s SeaSoft program for all data collected through the CTD, spikes are deleted or diminished in magnitude. For the subset of data shown in this paper we plotted the raw, unaveraged, unbinned LSS data to demonstrate the relative abundance of aggregates during each cruise.

3. Results

3.1. Temporal variations in POC standing stock

During each cruise in the Ross Sea, one or more transects were completed along 76°30’S (Fig. 1). The standing stock of POC (determined from the beam $c_p$ profile and Eq. (3)) was integrated for the upper 100 m at each station along the transect (Fig. 3). Scales on all plots are uniform so comparisons can be made directly. The integrated standing POC stock from our calibrated continuous profiles agrees well with the integrations calculated from the discrete bottle POC values (Smith et al., 2000). Differences largely result because the two averages do not always include the same stations. Smith et al. (2000) averaged the integrated standing stock over successive one-week intervals during the cruise whereas our integration was computed repeatedly over a specific transect (76°30’S). Our optical data provide an assurance that the low vertical resolution of features from the discrete bottle POC data is not a major liability for the integration schemes employed here or in Smith et al. (2000).

As expected, the POC standing stocks were low during the very early spring and late autumn (NBP 96-04a and 97-03), averaging about 200 mmol C m$^{-2}$, but ranged from 3000 to 7000 mmol C m$^{-2}$ at individual stations in the summer (NBP97-01) and 1000–5000 mmol C m$^{-2}$ in the austral spring of 1996 (NPB96-06) and 1997 (NPB97-08). POC stocks at a given station during a single cruise varied two-fold between successive transects (one-two week separation) during austral summer 1997. During austral spring (November–December) of 1997, when primary productivity was high and biomass was increasing rapidly, temporal differences in POC stocks were nearly an order of magnitude between transects 3 weeks apart at the same station.

Advection, as well as growth and losses, must be considered when interpreting these results. We used an average of the entire 76°30’S line to compensate for potential advection of material within the Ross Sea, but in reality, the maximum standing stock nearest the traps (178°W; Fig. 1) was between 2000 and 3000 mol C m$^{-2}$, about half the maximum value to the west.
Fig. 3. Integrated standing stock of POC in the top 100 m. Each point is derived from a single transmissometer profile calibrated with the data shown in Fig. 2 and Table 2.
Table 3
Mean change in POC and DOC concentrations, net community production (NCP), primary production (PP) and trap flux between cruises or transects

<table>
<thead>
<tr>
<th>Carbon Flux (mmol C m$^{-2}$ d$^{-1}$)</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Changes/rates between early spring and summer cruises</strong></td>
<td></td>
</tr>
<tr>
<td>POC buildup during 87 days between cruises</td>
<td>65.0 ± 13</td>
</tr>
<tr>
<td>DOC buildup between early cruises</td>
<td>11 ± 2</td>
</tr>
<tr>
<td>NCP to 200 m between early cruises</td>
<td>81.6 ± 66.7</td>
</tr>
<tr>
<td>PP during last week of early spring cruise</td>
<td>73.8 ± 16.2</td>
</tr>
<tr>
<td>PP during first week of summer cruise</td>
<td>43.7 ± 27.3</td>
</tr>
<tr>
<td>Sediment trap C$_{org}$ flux after early spring at 206 m</td>
<td>0.2</td>
</tr>
<tr>
<td><strong>Changes/rates between early summer and late autumn cruises</strong></td>
<td></td>
</tr>
<tr>
<td>POC loss during 84 days between cruises</td>
<td>39.5 ± 13</td>
</tr>
<tr>
<td>PP during late autumn cruise</td>
<td>1 ± 0.5</td>
</tr>
<tr>
<td>DOC loss between cruises</td>
<td>18</td>
</tr>
<tr>
<td>NCP loss to 200 m between cruises</td>
<td>40.5 ± 36.9</td>
</tr>
<tr>
<td>Sediment trap C$_{org}$ flux range at 206 m between cruises</td>
<td>1.5–1.7</td>
</tr>
<tr>
<td>Sediment trap C$_{org}$ flux range at 465 m between cruises</td>
<td>1.6–2.9</td>
</tr>
<tr>
<td><strong>Changes/rates after late autumn cruise</strong></td>
<td></td>
</tr>
<tr>
<td>Sediment trap C$_{org}$ flux range at 206 m</td>
<td>0.12–5.0</td>
</tr>
<tr>
<td>Sediment trap C$_{org}$ flux range at 465 m</td>
<td>0.05–10.0</td>
</tr>
<tr>
<td><strong>Changes/rates during spring, 1997 cruise (Nov–Dec)</strong></td>
<td></td>
</tr>
<tr>
<td>POC buildup during ~14 days between transects</td>
<td></td>
</tr>
<tr>
<td>1 and 4 of cruise</td>
<td>69.8 ± 65</td>
</tr>
<tr>
<td>PP range during spring cruise</td>
<td>46–87</td>
</tr>
</tbody>
</table>

3.2. **Comparisons of POC and DOC production/loss, primary productivity, net community production and trap POC fluxes**

We calculated the mean rate of POC production (or loss) by dividing the increase (or decrease) in mean POC standing stock in the upper 100 m between cruises (or transects) by the number of days between midpoints of a transect’s occupation (Table 3). Primary productivity rates and POC fluxes measured by sediment traps at 206 and 485 m during appropriate periods of time are included in Table 3 (Smith et al., 2000; Collier et al., 2000). Fig. 4 contains the mean and standard deviation of the POC standing stock along each transect as well as the POC fluxes measured by the three sediment traps at the MS7 site (Fig. 1). The standing stock of POC between 100 and 200 m was calculated in the same manner as for the upper 100 m to examine the possible sequestration of POC in that depth range (Fig. 4).

Our estimates of POC production ($58.5 ± 19 \text{ mmol C m}^{-2} \text{ d}^{-1}$) were similar to primary production of carbon from short-term $^{14}$C-based incubations by Smith et al., 2000) for the early spring (first week of November: $73.8 ± 16 \text{ mmol C m}^{-2} \text{ d}^{-1}$) and
Fig. 4. The standing stock of POC from 0–100 m and 100–200 m measured during each process cruise (Fig. 3) and the POC fluxes measured by the sediment traps at MS7 (Collier et al., 2001). Each dot represents the mean of a transect of stations along 76°30′S. Bars represent one standard deviation about the mean POC of each transect in Fig. 3.

summer (mid-January: 43.7 ± 27 mmol C m\(^{-2}\) d\(^{-1}\)). Primary productivity probably continued at closer to the higher rate between mid-November and early January to produce the POC standing stock measured in mid-January. Although we have no productivity measurements for that period in 1996, productivity measurements from mid-November to mid-December in 1997 ranged from 46 to 87 mmol C m\(^{-2}\) d\(^{-1}\) (Smith et al., 2000). During mid-January, 1997 the mean standing stock of POC decreased from 5300 to 3500 mmol C m\(^{-2}\) over a period of 14 days (Fig. 4). This is a mean daily loss of ca. 130 mmol C m\(^{-2}\) d\(^{-1}\), which agrees with the carbon export calculations of Sweeney et al. (2000).

Mean POC along the transect increased in the upper 100 m more than five-fold between early November and the end of December, but less than two-fold between 100 and 200 m (Fig. 4). Only 10% of the increase in the upper 100 m settled to the 100–200 m region suggesting that most of the POC produced either remained in surface waters, was remineralized, or settled below 200 m. During the next three weeks the 0–100 m POC stock increased by 2430 mmol C m\(^{-2}\) (78%) whereas the 100–200 m POC stock increased 740 mmol C m\(^{-2}\) (256%). The POC increase between 100 and 200 m was equivalent to one-third of the increase in the upper 100 m during that time, suggesting that much more POC was being exported during that time.

Between the austral summer (January) and autumn (April) cruises, surface-layer POC concentrations decreased to background values. Productivity was negligible in
the late autumn (Table 3; Smith et al., 2000). The loss of POC over this time period averaged 40 mmol C m$^{-2}$ d$^{-1}$.

The buildup of DOC in surface waters between early spring and summer was 1800 mmol C m$^{-2}$ (Carlson et al. 2000), or 11 mmol C m$^{-2}$ d$^{-1}$. By late autumn (April) DOC declined to winter values, which represents a loss of 18 mmol C m$^{-2}$ d$^{-1}$.

Sweeney et al. (2000) calculated the net community production (NCP) integrated to 200 m along the transect for each cruise based on the deficit in total CO$_2$ and predicted gas exchange. The average daily NCP between early spring and summer ($82 \pm 67$ mmol C m$^{-2}$ d$^{-1}$) matched the primary production during the last week of the spring cruise ($76 \pm 16$ mmol C m$^{-2}$ d$^{-1}$) and the buildup of POC between early spring and summer ($65 \pm 13$ mmol C m$^{-2}$ d$^{-1}$) quite well (Table 3). Between the summer cruise and late autumn cruises, NCP decreased from 7.3 $\pm$ 2.2 to $3.9 \pm 0.9$ mmol C m$^{-2}$, or $41 \pm 37$ mmol C m$^{-2}$ d$^{-1}$ over 84 days. This closely matches our estimates of $40 \pm 13$ mmol C m$^{-2}$ d$^{-1}$ decrease in POC over the same time period. Sweeney et al. estimated the total carbon export over the bloom (early spring to late autumn) to be $3.8 \pm 1.0$ mol C m$^{-2}$.

Sediment traps were deployed at 206 and 481 m on a mooring at 76°33’S, 178°01’W and another sediment trap was deployed at 465 m on a separate mooring 4 km to the south (Collier et al., 2000). Fluxes measured in traps from early December to mid-January (Fig. 4) were very low both at 206 and 465 m ($0.2$ and $0.1$ mmol C m$^{-2}$ d$^{-1}$, respectively). When the POC standing stock in surface waters decreased in mid-February, the POC flux increased to about $1.5$ mmol C m$^{-2}$ d$^{-1}$ at 206 m and $3--3.5$ mmol C m$^{-2}$ d$^{-1}$ in the deeper traps. There was a 2–3 week delay between the peak in POC standing stock and the February peak in trap fluxes. However, the major peak in POC flux occurred in April–May when the flux reached $5$ mmol C m$^{-2}$ d$^{-1}$ at 206 m and $10$ mmol C m$^{-2}$ d$^{-1}$ in the deeper traps. By this time the standing stock of POC had declined to winter background levels of $200$ mmol C m$^{-2}$. The mean decrease in the standing stock of POC was $40$ mmol C m$^{-2}$ d$^{-1}$ between January and April, compared with a February trap POC flux of $1.5$ mmol C m$^{-2}$ d$^{-1}$.

Therefore, of the material leaving the 0–100 m POC standing stock, only about $3.8\%$ was collected in the trap at 206 m. This increased to $12.5\%$ per day in April, a period when ice cover was nearly complete. The February peak in trap POC was associated with a peak flux of silica (not shown), while the April trap POC peak was associated with a peak in CaCO$_3$ flux from pteropods (Collier et al., 2000). The POC flux remained high into July in the deeper traps, but the 2-month collection interval of traps during this period made temporal resolution of events very poor. By October the trap POC flux was below $0.1$ mmol C m$^{-2}$ d$^{-1}$. Nearly identical POC fluxes were measured in both deep traps on moorings separated by 4 km (Fig. 4).

### 3.3. Seasonal abundance of aggregates in the Ross Sea

A detailed examination of the data from the transmissometer and LSS provides valuable insights into the relative abundance of aggregates or large plankton at
different times of the year. Transmissometer profiles were used in this study to estimate POC concentrations. The transmissometer signal comes primarily from particles $< 20 \mu m$ in diameter in the open ocean (Pak et al., 1988; Chung et al., 1996). Larger particles interrupt the light path in the same way as small particles, but they are far less abundant than small particles. When large particles enter the beam, they appear as a spike in the signal, but the size of particles or aggregates causing the spikes cannot be determined. The volume of the transmissometer beam is 0.0441, so with one aggregate per liter, chances are only one in 25 that an aggregate will be encountered. As aggregate abundance increases, so will the number of spikes. The light-scattering sensor (LSS) is more useful than the transmissometer when searching for evidence of aggregates because it “illuminates” a large tear-drop-shaped lobe of water ($\sim 4 \text{ l}$, but volume decreases with increasing particle concentration), so chances of encountering aggregates are much greater. Asper (1997) showed that in the Ross Sea there could be tens of aggregates per liter in mid-water during the summer. With a volume of several liters, the LSS signal is likely to include many aggregates.

The SeaSoft algorithms used in normal processing of CTD data have spike removal routines that eliminate most of the spikes caused by large particles. Averaging data into 2-m intervals further decreases the signal of aggregates. Even with this spike removal and averaging, we noted the highest abundance of spikes in the transmissometer and LSS data during the April cruise (NBP97-3), so we reprocessed CTD casts from near the MS7 trap site for each cruise using no spike removal or bin averaging of the data (Fig. 5).

A comparison of raw LSS profiles at the sediment trap mooring site from each cruise shows important similarities and differences (Fig. 5). The LSS profiles did not have their zero values calibrated, so absolute minima should not be used for comparison of background concentrations between cruises, but the voltage is linearly correlated with particle concentration in the same way that beam $c_p$ is correlated with POC concentrations. Each profile in Fig. 5 is typical of the degree of spikiness observed during the entire cruise, and all data are plotted at the same scale for ease of comparison. We have observed spikes in LSS and $c_p$ profiles in other places where camera images have shown aggregates to be abundant (Gardner et al., 1999; Gardner et al., 1999, 2000; Walsh et al., 1999). Spikes in the LSS data likely indicate depths containing an abundance of aggregates, not the presence of a single aggregate.

Both the beam attenuation and LSS signals indicated that little material was present in surface waters during the very early austral spring and late autumn (NBP96-4A and NBP97-3), whereas material was abundant in surface waters during summer (NBP97-1) and spring (NBP97-8). Bottom nepheloid layers contained substantial amounts of material during each period. The most important difference among the LSS profiles is the number and magnitude of “spikes” recorded during each period. Although the concentration of small particles was low during April (NBP97-03), the spike-filled optical records suggest aggregates were more abundant during April than other seasons, even though the transmissometer indicated there was very little material in the water.
Fig. 5. LSS profiles at the MS7 sediment trap site. Station and cast numbers (e.g. 03-02) are identified for each profile. The spikes are indicative of the presence of aggregates, with the greatest abundance during late autumn. Note the presence of a bottom nepheloid layer with abundant spikes during each season.

4. Discussion

4.1. POC standing stocks and production

The surface concentrations and integrated standing stock of particulate organic carbon estimated from the beam attenuation profiles reflected both the stocks and temporal cycles previously seen in the Ross Sea either directly (Smith et al., 1996; DiTullio and Smith, 1996; Asper and Smith, 1999) or remotely (Comiso et al., 1993). For example, surface-water POC concentrations reported for January, 1990 and 1992 were 14–45 and 9–24 μmol l⁻¹ (Smith et al., 1996), similar to the mean we found during January–February, 1997. Similarly, mid-January values in the upper 150 m reported by Asper and Smith (1999) were 340–3400 mmol C m⁻², again similar to the standing stocks we observed in the upper 100 m. Thus, our findings of POC concentrations and dynamics are similar to those found in previous years, and suggest that the seasonal cycle of particulate organic carbon is characterized by three stages. The first is a period of rapid increase, where phytoplankton are growing rapidly, biomass
is accumulating, and losses are extremely low. The second phase is one in which biomass increases are low or near zero. This may result from net increases (that is, phytoplankton growth rates) being smaller (potentially as a result of trace-metal limitation; Sedwick and DiTullio, 1997; Smith et al., 2000), combined with greater losses due to grazing and enhanced sinking rates. The third phase is characterized by declining POC concentrations. During this period the loss rates exceed growth rates and results in the observed decreases in POC standing stocks.

A year-long deployment of a transmissometer at 230 m at the edge of the ice shelf (Fig. 1) by Accornero et al. (1999) confirms this same general pattern, but shows high background concentrations and episodic periods of even higher concentrations at depth through August. The data were reported in percent transmission and were much lower (higher particle concentration) than any values observed at 230 m in our profiles on any cruise. Their values were closer to those we observed in surface waters.

4.2. Comparison of JGOFS trap fluxes with other flux measurements

Asper and Smith (1999) measured particle fluxes along 76°30′S using surface-tethered floating arrays of traps deployed for 1–3 days in spring and summer (mid-November to mid-January) in 1994 and 1995/1996. Organic carbon fluxes at 200 m ranged from 7–12 mmol C m⁻² d⁻¹ and accounted for a loss of 0.8–2.3% per day of the standing stock of POC in the upper 150 m during the deployments. These flux values were much greater than the 0.1–1.5 mmol C m⁻² d⁻¹ measured in 1997 at 206 m for the same time of year (Collier et al., 2000; Fig. 4). Similarly, the 0.8–2.3% per day of the standing stock of POC being collected by the floating traps was also much greater than our calculation of 0.02–0.06% per day of the POC standing stock arriving in the traps in January. By April (late autumn) the trap POC flux increased to 2.5% per day of the POC standing stock as a result of both increased fluxes and decreased standing stock.

At a site ~ 70 km east of MS7 (Fig. 1), Dunbar et al. (1998) measured POC fluxes up to 13.3 mmol C m⁻² d⁻¹ at 230 m in February–March of 1990 and 8.7 mmol C m⁻² d⁻¹ in 1991, just prior to the seasonal onset of ice cover. These values are 4–9 times the flux observed by Collier et al. (2000) during that period in 1997. At the edge of the ice shelf (78°S, 177°W) Accornero et al. (1999) measured fluxes at 433 m of 0.3–1.0 mmol C m⁻² d⁻¹ in February–March, 1995 and 0.6–1.2 mmol C m⁻² d⁻¹ between December 1995 and January 1996. These fluxes are less than those measured by Collier et al. (2000) for that period in 1997, but the water was moving north and west from under the ice shelf — a region of very low productivity — and could account for the low trap fluxes. The February–March increase in POC flux in all years was driven by a large maximum in silica flux, consisting mostly of diatoms (Dunbar et al., 1998; Accornero et al., 1999; Collier et al., 2000).

Since little of the material was being collected in the traps in the spring and summer in 1997, we looked for other reservoirs or sinks of carbon. Using the beam attenuation-derived POC data, we calculated that the 100–200-m depth interval contained 1275 mmol C m⁻² during January, but only 168 mmol C m⁻² in April. While this is
much less than the standing crop in surface waters, it is equivalent to 13.2 mmol C m\(^{-2}\) d\(^{-1}\) averaged over the entire time between the summer and late autumn cruises. Some remineralization must occur between 100 and 200 m, but it is arguably at a much lower rate than in the upper 100 m. To account for the entire flux measured with the trap at 206 m in March and April (1.5–5 mmol C m\(^{-2}\) d\(^{-1}\)) requires the collection of 22% of the POC sequestered between 100–200 m. While this helps explain the high April flux, it does not explain the anomalously low trap POC fluxes in February–March.

Unlike most areas of the world ocean, the buildup of DOC during the bloom period in the Ross Sea was not as large as the buildup of POC (Carlson et al., 2000). Although DOC increased during the spring bloom, it was largely consumed by late summer (Table 3; Carlson et al., 2000) so the DOC pool cannot be viewed as a potential sink via winter mixing as in subtropical waters (Ducklow et al., 1995).

Nelson et al. (1996) calculated that 90–95% of the POC exported from the euphotic zone decomposes at depths between 50 and 250 m. The floating trap measurements of Asper and Smith (1999) show a 40–75% decrease in flux between traps at 50 m (a depth they state was substantially below the mixed-layer depth) and 200 m, which is less of a decrease in flux than that calculated by Nelson et al. (1996). The Nelson et al. estimate was from the base of the euphotic zone (depth not listed in their paper) to 250 m and was measured by moored traps of a design similar to that used by Collier et al. (2000) in the AESOPS study (Dunbar et al., 1998).

This degree of recycling or decrease in flux with depth is of the order expected in the open ocean, but there traps are often moored 1000 m below the surface, giving more depth and time for recycling. Accounting for this rapid recycling or loss in the Ross Sea is not simple because the top trap was at 206 m. Caron et al. (2000) found the phytoplankton mortality rate to be unmeasurable in 28 of 34 experiments conducted during the four seasonal cruises of this program in the Ross Sea. The average mortality rate when measurable was 0.14 d\(^{-1}\), which is small compared to maximal observed short-term growth rates of ca. 0.6 d\(^{-1}\). Macrozooplankton grazing was also low (M. Dagg and J. Urban-Rich, personal communication, 1999). Microbial turnover rates were substantial and contributed to the recycling of organic carbon in surface waters and as particles settled (Ducklow et al., 2000). However, grazing measurements were restricted to the euphotic zone, so we know little about the rates of feeding and recycling below the euphotic zone.

Advection of surface waters also must be considered when evaluating temporal variations. Like elsewhere, blooms in the Ross Sea are not uniformly ubiquitous (Arrigo et al., 1999), so advection of water from one area to another can cause apparent increases or decreases in POC concentrations at a given geographic site. One of the reasons we chose to average POC standing stock along the entire 70°30’S transect rather than just at the trap mooring site was to include the influence of patchiness in our estimates. Asper and Smith (1999) found that their surface-tethered trap arrays always drifted north during the spring, suggesting that lower POC productivity near or under the ice shelf might yield lower fluxes at the MS7 trap. A current meter at 220 m on the trap mooring at MS7 showed that water moved NW (R. Collier, OSU, personal communication). Jaeger et al. (1996) found currents at a site
south of our transect line to be moving east at 240 m in 1990, but SSE in 1991. At the same site Jaeger et al. (1996) measured water at 740 m moving NE in 1990 and ENE in 1992. In 1995 Accornero et al. (1999) found that the average current moved between W and N at both 244 and 390 m at the edge of the ice shelf (77°S, 177°W). Thus, the currents are highly variable in time and space in the basin and ridge topography of the Ross Sea, complicating analyses of particle sources and sinks.

An important question is why the traps in the open Ross Sea collected substantially less material during January and February 1997 than in other years when the 1997 POC standing stock was of comparable magnitude to other years. Furthermore, independent estimates of carbon export suggested higher fluxes than measured by the traps. Measuring the removal of $^{234}$Th from surface waters, Cochran et al. (2000) estimate the POC flux at 100 m to be $7.1 \pm 5$ mmol C m$^{-2}$ d$^{-1}$ on 19 January and $30.4 \pm 7.5$ mmol C m$^{-2}$ d$^{-1}$ on 1 February. Sweeney et al. (2000) have constructed a carbon budget based on measurements of total CO$_2$ and TOC, and estimate carbon export down to 200 m to be $3.8$ mol C m$^{-2}$ between November 1996 and April 1997. The net carbon flux measured by traps at 206 m from early spring to late autumn was $0.29$ mol C m$^{-2}$ (Collier et al., 2000). The peak POC mean standing stock to 100 m was $5.3$ mol C m$^{-2}$, so only $5.5\%$ of the peak stock was collected in traps at 206 m. Our estimate of POC loss to 100 m along the transect between January and April was $3.3$ mol C m$^{-2}$, but about half that much if we consider only the station at the trap site. If we include the trap data to the end of September, the flux was $0.48$ mol C m$^{-2}$, still much less than predicted by other methods.

One might question whether the traps functioned properly. The traps in question had electronic mechanisms to assure that the cups turned at the correct time. If the timer mechanism failed and, for example, the collection cups were opened for uniform periods of time, the highs and lows of fluxes would be even more extreme, plus the annual mean flux would still be low. Furthermore, the traps at 206 and 465 m were of slightly different design (OSU and ParFlux) and had different timing mechanisms, requiring that independent systems malfunctioned in the same way to yield similar fluxes, which would be highly unlikely. Furthermore, the trap used at 206 m by Collier et al. was of the same design and manufacturer (OSU) as the traps moored by Dunbar et al. (1998), which eliminates questions about differences in fluxes being due to trap design. Lastly, the traps at 465 and 481 m were on separate moorings 4 km apart. The remarkably close agreement in the POC fluxes (Fig. 4) shows that the flux pattern was reproducible over modest spatial scales. The trap at MS6 near the edge of the shelf break to the northwest also showed a similar flux history and composition. Dunbar (personal communication) has compiled sediment trap flux data for the Ross Sea and observed considerable interannual variability. Interannual variability elsewhere has been shown to be as much as 10-fold for given annual periods in the Norwegian Sea (Wassmann, 1998). Most important, during 1997 Dunbar had traps moored to the southwest of the traps of Collier et al. and measured fluxes of POC that were very similar to those measured by Collier for the same seasons with anomalously low fluxes in spring and summer and a larger peak in autumn (Collier et al., 2000). Thus, interannual variability may account for the anomalous fluxes of Collier et al. (2000),
but it does not explain the > 12-fold discrepancy in carbon export calculated by Sweeney et al. (2000).

Another possible explanation for the low fluxes is hydrodynamic bias of the trapping process (Gardner, 2000). Bale (1998) reported that funnel traps collected only 10% of the material collected by cylindrical traps in the North Sea when currents averaged 30–40 cm s\(^{-1}\). He reported a few field deployments where the mean current was 10–20 cm s\(^{-1}\), but the data were compiled and “corrected” in such a way that it is not possible to determine whether undercollection occurred at the lower velocities. For the Ross Sea traps, however, flow past the traps was almost always less than 20 cm s\(^{-1}\) and was generally less than 10 cm s\(^{-1}\), which is within the range that should not cause significant hydrodynamic artifacts for cylindrical traps, but conical traps have not been tested as rigorously (Gardner et al., 1997). Furthermore, laboratory experiments on the efficiency of sediment traps have not tested the efficiency of collecting aggregates, and because this environment has an abundance of aggregates (Asper, 1997), undersampling may still be a problem. Aggregates could be shredded in the turbulence generated around and inside of traps, thus altering their settling characteristics (Gardner and Zhang, 1997). Baldwin et al. (1998) reported that funnel traps undercollected aggregates in a field deployment. On top of all of these caveats, however, one must note that all of the moored trap fluxes measured by Dunbar and colleagues have used funnel traps, so any environmental adjustments applied to the fluxes of Collier et al., also would apply to those of Dunbar and colleagues. The traps used by Asper and Smith (1999) were cylinders.

One carbon export pathway that might be missed by sediment traps results from vertical migration of zooplankton (Longhurst and Harrison 1988; Dam et al., 1995). For vertical migration to affect the flux of carbon differently in 1997 than in other years requires a different food web structure in different years. It appears that there may be aspects of the carbon system or our instruments that we do not fully understand.

4.3. Pteropods and the winter peak in flux

There were similarities and differences in the composition and magnitude of seasonal fluxes measured in 1990–1991 by Dunbar et al. (1998), in 1995 by Accornero et al. (1999) and in 1997 by Collier et al. (2000). In 1991 the major POC flux at 230 m was in February, reaching 8.7 mmol C m\(^{-2}\) d\(^{-1}\). There was a secondary POC peak flux in April–June, 1991 (2.5 mmol C m\(^{-2}\) d\(^{-1}\)). The February peak was associated with a peak in silica flux, whereas the April–June peak was associated with a peak in CaCO\(_3\) flux that was dominated by pteropods (R. Dunbar, personal communication). Measurements of POC concentrations were made in 1990 and 1992 (Smith et al., 1996), but standing stocks were not reported for comparison with the trap fluxes. In 1995 the peak POC flux at the ice shelf edge was in January–February (0.3–1.0 mmol C m\(^{-2}\) d\(^{-1}\)). There was no substantial flux peak in April–June that year, but a peak in percent POC of the sample occurred during that time period and was associated with an input of empty pteropod shells (Accornero, 1999). In 1997 Collier et al. (2000) measured a POC flux in February of only
1.5–3.5 mmol C m$^{-2}$ d$^{-1}$, but it reached 5–10 mmol C m$^{-2}$ d$^{-1}$ in April–June. Again, the February peak was associated with a peak in silica flux while the April–June peak was associated with a peak in CaCO$_3$ dominated by aragonite, suggesting the presence of pteropods. Microscope examination revealed the sample was dominated by the pterpod *Limacina helicina* (Collier et al., 2000). The February peak in POC trap flux occurred as the standing stock of POC in the upper 100 m was decreasing, but the April–June peak in trap POC flux was collected when the standing stock of POC in the upper 100 m was at an annual low (Fig. 4). In 1983–1984 Dunbar (1984) deployed a single cup trap for a year and found that the carbonate fraction of the trap (50% by weight) was dominated by pteropods, but no seasonal determinations of flux or composition could be made.

Is the flux of pteropods real or an artifact? *Limacina helicina* is the same pterpod species that Harbison and Gilmer (1986) found was preferentially collected in floating sediment traps. They suggested that these pteropods should be categorized as “swimmers” and removed from trap analyses because their escape mechanism when physically disturbed is to drop their mucous feeding web and sink or swim downward; otherwise they remain nearly neutrally buoyant when their mucous feeding web is deployed. Therefore, one might reasonably speculate that the abundance of pteropods in the Ross Sea traps was an artifact and masked the true trap fluxes. However, the preferential collection of pteropods noted by Harbison and Gilmer (1986) was in surface-tethered traps floating at 25 m, where the pteropods were most abundant. They argued that at greater depths where living pteropods were absent, they would not be collected preferentially in the traps as swimmers.

Very few data are available about the distribution and life cycle of pteropods in high latitudes because of wintertime inaccessibility. The only data available on the depth distribution of pteropods in the Ross Sea comes from net tows to 100 and 300 m in McMurdo Sound (the south-westernmost portion of the Ross Sea). These results showed that concentrations of *Limacina helicina* were slightly higher in the upper 100 m than those averaged through 300 m (Foster, 1987), but the vertical resolution of that study was poor, the study limited in time (Nov.–Dec., 1985), and restricted in space. Net collections during austral winter on cruise NBP97-03 in the Ross Sea indicated an abundance of zooplankton in the water (M. Huntley, personal communication), but quantification of the seasonal changes in zooplankton biomass have not been completed. In the Southern portion of the Weddell Sea *Limacina helicina* comprised 17% of the zooplankton in the upper 300 m in January and February (Boysen-Ennen et al., 1991).

The precise depth distribution of *Limacina helicina* in the Ross Sea is unknown. If they are confined to the upper 150 m (as in the Arctic, Kobayashi, 1974), pteropods should not be regarded as swimmers when found in traps below that depth. Indeed, Collier et al. (2000) suggest that most of the pteropods had died in a seasonal population crash and settled into the traps, rather than being collected as “swimmers” that actively migrated or were swept into the trap by advective currents. Dunbar (1984) also suggested that his measured pteropod flux was real. Collier et al. noted that 30% of shells were empty, while others were in varying stages of decay. Because the preservative formalin was used in the sample cups rather than a poison, it is likely that
the organisms remained close to the state in which they entered the traps. However, with a settling velocity on the order of 1 cm s\(^{-1}\) for large, empty shells (Byrne et al., 1984), there is little time for the organisms to decay. In other areas the naked pteropod *Clione limacina* feeds specifically on *Limacina helicina*, pulling them out of their shells and consuming them whole (Lalli and Gilmer, 1989). Foster (1987) reported no *Clione* in his McMurdo Sound sampling, but Accornero (personal communication) observed *clione* in the Western Ross Sea.

A possible explanation of this peak in CaCO\(_3\) and POC fluxes is that it represents an autumn population crash of pteropods in the Ross Sea that occurred after most of the POC in the water column was utilized. Perhaps the empty shells resulted from species-specific feeding by *clione* or other plankton, and the partially decayed pteropods perished and sank. Most of the pteropods were small (0.4–0.9 mm), mature adults (Collier et al., 2000). Kobayashi (1974) estimated the lifespan of *Limacina helicina* to be 18–24 months in the central Arctic Ocean, which suggests there must be some food under or within the ice during the winter because there is no evidence that pteropods utilize stored lipids in a manner similar to copepods or krill (Lalli and Gilmer, 1989). *Clione* have been maintained in laboratory tanks for up to a month without food (Lalli and Gilmer, 1989), but little is known of the seasonal cycle of plankton in the Ross Sea. Although standing stocks of POC based on bottle and transmissometer data appeared to be at background levels by late April, is it possible that sufficient POC occurred in the abundant aggregates throughout the water column (Fig. 5) to provide enough carbon for *L. helicina* to survive until April.

It has been argued that *Phaeocystis* is not quickly grazed by most zooplankton, so *Phaeocystis*-derived material is likely to settle without repackaging (DiTullio and Smith, 1996). Dunbar et al. (1998) cited studies showing that *P. antarctica* settle quite slowly (relative to diatomaceous material), and Smith and Dunbar (1998) suggested that the lag between maximum biomass and flux rates was due to the slower sinking rates in regions dominated by *P. antarctica*. Because they lack silica or CaCO\(_3\) skeletons, their density would likely be less than aggregates dominated by diatoms (Alldredge and Silver, 1988). If just a small percentage of the aggregates were neutrally buoyant, they could retain POC in the water column for some time and be a food source for pteropods, which generally feed on phytoplankton (Lalli and Gilmer, 1989).

Note that the trap fluxes at this time represent < 2.5% per day of the standing stock of POC in surface waters, so an increase in trap flux requires only a slight change in the percentage of surface POC reaching the traps. The lag between peaks in surface-water POC biomass and the February POC flux was ca. 15 days, but the time lag to the larger POC flux in April was ca. 100 days. For particles to settle 100 m in 15 days requires a settling speed of only 7 m d\(^{-1}\) and about 25 m d\(^{-1}\) to reach the 465 m trap. The 100-day lag requires settling speeds of 1 m d\(^{-1}\) for the shallow trap and about 4 m d\(^{-1}\) for the deeper traps. Although 100 days is a long time to maintain aggregates in the water column, it is on the order of the 6–10 m d\(^{-1}\) settling speed for aggregates estimated by Smith and Dunbar (1998). Reduced settling also provides more time to achieve the 90–95% recycling of POC in surface waters calculated by Nelson et al. (1996) or the 40–75% measured by Asper and Smith (1999).
4.4. Aggregates and fluxes

A study of aggregates in the spring and mid-summer was made in the Ross Sea by Asper (1997) using an aggregate camera on an ROV in 1994 and in a profiling mode in 1995. He determined that the abundance of aggregates ranged from 30 to 230 l\(^{-1}\) in surface waters as the bloom declined. The abundance substantially decreased below 40 m to about 15–201 \(l^{-1}\) by 200 m. Asper (1997) also deployed a settling chamber with a time-lapse camera system that was used to determine the in situ size and settling velocity of aggregates. Aggregates imaged ranged from 0.8 to 3.2 mm in diameter and exhibited settling velocities of 5–115 m d\(^{-1}\) (mean of 60 m d\(^{-1}\)) during two deployments in November 1994 and 25–230 m d\(^{-1}\) (mean 135 m d\(^{-1}\)) during January 1996. Dunbar et al. (1998) measured settling velocities of aggregates and fecal pellets and found them to settle at 20–1500 m d\(^{-1}\), with a mean of 200 m d\(^{-1}\). Asper’s settling rates were on the low end of the settling velocities measured by Dunbar et al. (1998), but the latter measurements were dominated by compact fecal pellets and remolded aggregates extracted from preserved samples collected by sediment traps. Trap samples would be biased towards rapidly sinking particles and the dewatering of aggregates sitting in a trap for months would undoubtedly increase their sinking velocities. In either study turbulence encountered upon entering the trap or Asper’s in situ settling chamber also could affect aggregate size and density.

It is important to note, however, that in both studies only those aggregates that were actively settling would enter and be measured in the sinking chamber or collected in the trap. Asper et al. (1992) speculated that some aggregates may not settle at all. Such aggregates are neutrally buoyant and make ideal recycling sites of organic matter via grazing and microbial remineralization. Alldredge and Gotschalk (1989) measured the density of aggregates to be as little as 10\(^{-5}\) kg m\(^{-3}\) greater than the surrounding seawater, making settling rates very low regardless of the size of aggregates. Although Alldredge and Gotschalk (1989) found a weak correlation between aggregate size and settling velocity, Azetsu and Johnson (1992), Asper (1997) and Diercks and Asper (1997) found no correlation in their studies.

Aggregate abundance was not measured in the Ross Sea during the 1997 cruise with large-particle sensors. However, as explained in Results (Section 3.3), the abundance of spikes in the LSS data represents the relative abundance of large particles in the water. Unfortunately, we cannot calibrate the LSS data in terms of size or abundance of particles because there are too many degrees of freedom in the size/abundance/location/sensing volume spectrum. We suggest, however, that most of the spikes are aggregates rather than zooplankton. Walsh et al. (1997) found that < 5% of the aggregates ( > 500 \(\mu\)m) in seawater might be zooplankton based on concurrent enumeration of aggregates and zooplankton using an in situ aggregate camera and plankton nets, and therefore concluded that aggregates are the numerically dominant form in that size class. A calculation from plankton tows made in McMurdo Sound by Foster (1987) yields 0.1 zooplankton > 0.2 mm l\(^{-1}\), which is orders of magnitude lower than the 15–230 aggregates l\(^{-1}\) > 0.5 mm observed by Asper (1997).

Although we lack highly resolved, quantitative seasonal data for the abundance of aggregates, it appears that among our sampling seasons, aggregates were most
abundant throughout the water column around the time when POC flux peaked (April). The aggregate abundance measured by Asper (1997) was during January, which is comparable to our summer period (NBP97-01). This suggests there were more than 15–230 aggregates l\(^{-1}\) in the water column during April. We suggest that many of these aggregates were in large part not settling (or settling extremely slowly) and were responsible for maintaining a small percentage of the seasonal production in the water column, which in turn served as a late autumn food source for pteropods. In autumn as the ice cover redeveloped, POC was depleted, the pteropod population crashed and the increased flux was recorded in the traps. This is consistent with an increase in the CaCO\(_3\) flux at this time, as well as a decrease in silica flux.

Another factor potentially influencing the collection of pteropods is the formation of brine as sea water freezes. The sinking of dense brine plays a major role in the overturning of water during autumn. The reaction of pteropods to the cascading brine and the potential for the brine to carry the pteropods (and other material) downward are unknown. The downward cascading of particle-laden brine could partially explain the periodic increases in beam attenuation observed by Accornero et al. (1999) at 230 m. Regardless, the peak in POC flux, while notable compared with other times of the year, remains a small percentage of the loss of POC from surface waters.

4.5. Resuspension and advection

Resuspension and advection also must play a role in the accumulation of carbon within Ross Sea sediments. Dunbar et al. (1998) found that mass fluxes generally decreased between their traps at 230 and 519 m (50 m above bottom). However, the data of Collier et al. (2000) generally showed an increase in flux from 206 to 465 m. The most likely cause of an increase in flux near the seafloor is resuspended sediment (Gardner et al., 1985; Gardner and Richardson, 1992). Field experiments show evidence of sediment erosion when current velocities one meter above the bed exceed 8–12 cm s\(^{-1}\) in fine-grained marine sediment with 2–3% organic carbon (Young and Southard, 1978). Newly arrived phytodetritus aggregates can be resuspended at free-stream flows as low as 9 cm s\(^{-1}\) (7 cm s\(^{-1}\) at 1 m above seafloor; Lampitt, 1985). Aggregates with pteropods might require a slightly higher velocity for resuspension.

As noted earlier, a current meter on the trap mooring at MS7 showed water moving NW at 5–15 cm s\(^{-1}\) at 220 m during this study (R. Collier, OSU, personal communication), and the currents often were sufficient to resuspend phytodetritus from the seafloor, assuming the currents extended to the seafloor. Both the LSS profiles (Fig. 5) and beam \(c_\rho\) profiles (Fig. 6) show the presence of bottom nepheloid layers during every cruise. The complicated bottom topography makes it difficult to determine actual near-bottom flow. However, there is a steep ridge some 50–100 km west of the trap site and a plateau to the east (Fig. 6). At the trap site the beam \(c_\rho\) profiles show a fine-particle bottom nepheloid layer along the transect during every cruise (Figs. 5 and 6), with the nepheloid layer occasionally extending above the bottom traps (465 and 481 m). Furthermore, Coale et al. (in prep.) found that particulate iron concentrations increased from 8 nmol kg\(^{-1}\) at 400 m to 56 nmol kg\(^{-1}\) at 500 m and 96 nmol kg\(^{-1}\) at 550 m during April 1997, suggesting resuspension and advection of
bottom sediments to at least 500 m. We do not know whether an aggregate nepheloid layer (Gardner and Walsh, 1990) may have extended higher into the water column, but the abundance of spikes in the LSS profiles increased near the bottom during at least three of our four cruises (Fig. 5). V. Asper (personal communication, 1999) found significant increases in the abundance of aggregates in the bottom 30 m during a 1995/6 summer cruise. Studies of sedimentation rates in the Ross Sea suggested that 50% of the material accumulating in the troughs of the Ross Sea (where the Collier et al. traps were located) was transported laterally (DeMaster et al., 1996). Therefore, although we cannot identify the geographic source of the resuspended material more than to say it is most likely from higher ground, it is clear that resuspension and advection play a role in the redistribution of carbon and other components of the sediments in the Ross Sea and account for the increase in flux in the bottom traps of Collier et al. (2000).

Previously published and unpublished sediment trap data show that the increase in pteropod and POC flux in April occurs in others years (e.g. 1991: Dunbar et al., 1998), though the event of 1997 is larger than that of 1991. Existing data suggest that timing of the productivity cycle of the Ross Sea is fairly predictable (DeMaster et al., 1992; Nelson et al., 1996), but the magnitude of fluxes appears more variable (Dunbar et al., 1998 and personal communication, 1999).

5. Summary

The mean standing stock of POC along 76°30’S increased from 240 mmol C m⁻² in early austral spring to 5300 mmol C m⁻² in mid-summer, which implies a minimum

Fig. 6. Water depth along 76°30’S and nepheloid layer thickness. Inverted triangles and triangles indicate the maximum and minimum height of the small-particle nepheloid layer observed during all process cruises. The T’s indicate the depth at which traps were deployed by Collier et al. (2001).
daily POC accumulation rate of 65 mmol C m$^{-2}$ d$^{-1}$. This production estimate is close to the measured primary productivity rate measured in short-term incubations, suggesting that loss processes during this period were negligible. POC concentrations reached their seasonal maximum in mid-January, and declined significantly (by 34%) during a two-week period (from January 18 to February 1). POC standing stocks continued to decrease with time, and reached background, winter values by April. The decrease in POC was consistent with the carbon export calculated by Sweeney et al. (2000), which was > 12-fold higher than carbon fluxes measured with sediment traps. The Collier et al. (2000) traps may have under-collected aggregates, but other traps in the area yielded similar fluxes. The carbon export averaged along the entire 76°30’S line was greater than the export in the trap vicinity (~ factor of 2).

Sediment trap fluxes at 206, 465 and 481 m were only 0.2 mmol C m$^{-2}$ d$^{-1}$ until January 1997, suggesting that nearly all of the carbon fixed in surface waters remained there or was recycled above 206 m during that time. An increase in the measured sediment trap POC flux in February–March was associated with diatoms, but the largest POC flux occurred in April at 206 m and continued on through June at the deeper traps. This late autumn peak was associated with a carbonate flux composed of pteropods. Although concentrations of small particles were very low during that time, aggregates were abundant during April and may have provided food for a pteropod population. The flux of POC and pteropods may have resulted from a massive population crash when food became depleted. The increased flux in the near bottom traps most likely resulted from resuspension and lateral advection of material from nearby ridges or high ground.

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