THE ROSS SEA

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Abstract

We present budgets of carbon and nitrogen for the Ross Sea, Antarctica. The novelty of this study consists in estimating both vertical water-column to sedimentary fluxes as well as horizontal exchanges due to water mass lateral transport between the continental shelf and the open ocean. To this end, we relied upon a large data set combining results from ten Italian oceanographic cruises as well as from literature data, in order to have the largest possible data set. The results presented here have largely benefited from consulting several US-JGOFS studies (AESOPS, 1996-98 cruises) and from international collaborations, the ROAVERRS project (1996-98). The quantities thus calculated can be referred to a budget of an average productive season lasting from November to February.

The deep layer, the most relevant for the continental shelf pump, releases carbon to the deep ocean as dissolved inorganic carbon (937 Gmol), dissolved organic carbon (13 Gmol) and particulate organic carbon (7 Gmol). As to particulate organic nitrogen, less than 1 Gmol is released to the deep ocean, but the nitrogen balance becomes strongly positive considering the 19 Gmol imported from the open ocean.

Our estimates indicate that during an average austral summer, the amount of matter recycled through biological and sedimentary processes within the Ross Sea is almost one order of magnitude higher than the amount exchanged with the open ocean. Within the upper layer 3486 and 523 Gmol of carbon and nitrogen, respectively, are incorporated into biomass, of which about 50 % are later exported to the deep layer. Since the burial is negligible (less than 0.5 %), considering the upper and deep layer together, we estimate that about 3155 Gmol of carbon and 473 Gmol of nitrogen are channelled to the higher trophic levels and to the dissolved organic pools. Thus, grazing becomes the most important way of export of the Ross Sea.

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1. Introduction

The present work attempts to provide a budget of carbon and nitrogen in the Ross Sea. These budgets include downward fluxes from the upper productive layer to the continental shelf sediments and cross-slope transfers to the Southern Ocean over a period lasting from late spring (November)

to summer (February). To this end, we relied upon a large data set combining results from ten Italian oceanographic cruises as well as from literature data (see specific references reported in the text) in order to obtain the largest possible data set. The results presented here have considered Italian cruises carried out from 1987 - 2001 within the Italian National Antarctic Research Program (PNRA) and have benefited from several US-JGOFS studies (AESOPS, 1996-98 cruises) and the ROAVERRS project (1996-98). Previous estimates considered water column and sedimentary fluxes of carbon, nitrogen and silica over the Ross Sea continental shelf (Nelson et al., 1996; Smith and Asper, 2000; Smith et al., 2003, 2006) and the whole Pacific sector of the Southern Ocean (Nelson et al., 2002), elucidating major aspects of their cycling, such as surface-layer production and regeneration, vertical transport, benthic regeneration and seabed accumulation. For the Ross Sea, three areas were considered in the budget, occupying its southwestern, southeastern and northern portions. The preferential preservation of silica over carbon was highlighted (Nelson et al., 1996).

Our study aims to integrate previous reports with data from stations and moorings covering a larger area of the Ross Sea continental shelf. Further, our estimates consider not only the vertical water column and sedimentary fluxes but also the horizontal exchanges due to water mass lateral transport between the continental shelf and the open ocean.

The comparison of our results with those previously reported has allowed to check their overall agreement with already existing functional schemes of the Ross Sea ecosystem as well as to infer some further considerations.

2. Environmental Setting

Geomorphology and sedimentary characteristics
The Ross Sea is a roughly triangular-shaped portion of the Antarctic continental shelf, located in the Pacific sector of the Southern Ocean. It extends between Cape Colbeck, at 158°W, and Cape Adare, at 170°E. Southward, at about 78.5°S, it is bordered by a continentally-based sheet of freshwater ice, the Ross Ice Shelf, which extends over nearly half the continental shelf and is about 250 meters thick on its northernmost side (Jacobs and Comiso, 1989). There, the ice calves to form icebergs that normally flow north along the Victoria Land coast.

The Ross Sea has tortuous bottom topography, with depths ranging from 250 to 1200 m and averaging 500 m. The continental shelf slopes towards the continent and is more rugged and deeper on the western side, where the Drygalski basin (which is the deepest of the Ross Sea) reaches a depth of 1200 m. This irregular morphology results from the erosion of the seafloor by outlet glaciers, that have created narrow transverse troughs and is further accentuated by the presence of the McMurdo Sound complex of volcanic islands and seamounts. The central continental shelf alternates shallow banks (~ 300 m) and deep basins (> 500 m), characterized by an elongate shape and oriented north to northeast. The eastern side is bounded by several small depressions, with depths as great as 900 m (Challenger and Little America Basins). Poorly sorted coarse sediments (residual glacial-marine) cover the tops of the banks, while compound glacial marine sediments consisting of terrigenous detritus, essentially due to glacial transport, are confined along the western coast (Anderson et al., 1984; Dunbar et al., 1985). A narrow belt of diatomaceous compound glacial-marine sediments marks the passage to siliceous mud and ooze, that are the prominent product of present-day sedimentation in the shelf basins (Langone et al., 1998; Brambati et al., 1999; Brambati et al., 2002a, 2002b). On the outer shelf, biogenic detritus of calcareous origin covers the seafloor (Anderson et al., 1984; Taviani et al., 1993).

Dynamical and hydrographical features
A prominent feature constraining the Ross Sea dynamics is the Antarctic Circumpolar Current (ACC), which moves west to east around Antarctica and interacts with several water masses along

its path. ACC carries the Circumpolar Deep Water (CDW), which reaches the Ross Sea continental shelf break, moves upward and intrudes onto the continental shelf at some specific locations. CDW strongly influences the thermohaline circulation of the basin, as it is the only water mass which provides heat to the shelf waters; interacting with the shelf waters it forms the modified-CDW (MCDW), the location of which is determined by a subsurface temperature maximum and a dissolved oxygen minimum (Jacobs et al., 1985; Locarnini, 1994; Jacobs and Giulivi, 1998, 1999; Budillon et al., 2000). CDW and shelf waters are generally separated by a front – the Antarctic Slope Front (ASF) – which is supposed to be determined by topography and with a high biological productivity (Jacobs, 1991).

A major factor of Ross Sea dynamics is the waxing and waning of seasonal ice cover. The area north of the Ross Ice Shelf and east of Ross Island has less days of ice cover per year than the surrounding areas, and it is thus known as the Ross Sea polynya. The polynya appears to be formed due to katabatic winds blowing off the continent and expands rapidly in November towards north (Zwally et al., 1985). In Terra Nova Bay (TNB), near 75°S, off the Victoria Land coast, a smaller coastal polynya is maintained by the combined action of katabatic winds and the presence of the Drygalski Ice Tongue, which prevents the northward drifting of pack ice (Bromwich and Kurtz, 1984; Jacobs et al., 1985). This region is the site of formation of the High Salinity Shelf Water (HSSW), produced by continuous freezing of surface seawater and brine rejection, under the influence of katabatic winds (Kurtz and Bromwich, 1983, 1985; Van Woert, 1999; Budillon and Spezie, 2000; Fusco et al., 2002). HSSW spreads northwards and southwards close to the bottom on the western side of the Ross Sea, following the axis of the Drygalski basin. The northward branch reaches the continental shelf break, where it mixes with the MCDW to form deep and bottom waters (Jacobs et al., 1970; Gordon and Tchernia, 1972; Rodman and Gordon, 1982; Budillon et al., 1999, 2006; Bergamasco et al. 2004, Gordon et al. 2004). The southward branch is supposed to move beneath the Ross Ice Shelf, where it interacts with the basal ice to produce Deep Ice Shelf Water (DISW) (Jacobs et al., 1989). DISW is colder than the surface freezing point, fresher and less dense (Jacobs et al., 1985, 1989; Jacobs and Comiso, 1989; Trumbore et al., 1991; Locarnini, 1994; Bergamasco et al., 2002). Its core is located at intermediate depths (from 300 to 500 m) and it moves northward along the Challanger Basin (Budillon et al., 2002; Rubino et al., 2002). On reaching the continental shelf break, DISW mixes with MCDW, hence giving a further contribution to the formation of deep and bottom waters that spread into the Southern Ocean. Because of these processes, HSSW and DISW are the most important Ross Sea shelf waters, which play an important role in the ventilation of the deep layers. The export of dense shelf water is obviously balanced by less dense (warmer and fresher) MCDW intrusions.

Jacobs and Giulivi (1998, 1999) and Budillon et al. (2003) provided a synthesis of hydrology and water mass distribution in the Ross Sea and Gordon et al. (2000) described the seasonal variability in the area.

Biological production in the Ross Sea

The Ross Sea is a heterogeneous ecosystem, characterized by a large variability in space and time of macro and micronutrients (Catalano et al., 2000; Gordon et al., 2000; Fitzwater et al., 2000; Sedwick et al., 2000; Smith et al., 2003, 2006), which exhibits the highest productivity of the Southern Ocean (Arrigo et al., 1998). Annual primary production has been estimated to be about four times greater than elsewhere in the Southern Ocean, with large seasonal variations in biomass accumulation and production, and less pronounced interannual variability (Nelson et al., 1996).

The four subsystems of the Ross Sea (ice-covered waters, polynyas and their marginal ice zones, ice-free waters during full summer, and coastal waters) contribute differently to pelagic production. Maximum primary production occurs in late spring and decreases sharply in summer in the central region, when this area is defined as a high nutrient low chlorophyll (HNLC) zone. The low production in offshore areas has been ascribed to iron limitation (DiTullio and Smith, 1996;

Sedwick and DiTullio, 1997), as well as to light limitation due to the deep vertical mixing (Nelson and Smith, 1991; Sakshaug et al., 1991; Saggiomo et al., 2002).

Phytoplankton populations are alternatively dominated by two major algal taxa: diatoms and prymnesiophyceans, in particular *Phaeocystis antarctica* (Smith and Asper, 2001). Both groups occur in spring, whereas diatoms definitely dominate summer assemblages. Prymnesiophytes occur particularly in the southcentral Ross Sea (Smith and Gordon, 1997; Arrigo et al., 1999; Saggiomo et al., 2000). The phytoplankton community composition has a strong influence on biogeochemical processes and the prevalence of one or the other algal group has a major impact on trophic pathways and carbon drawdown (Arrigo et al., 1999; Sweeney et al., 2000; Smith and Asper, 2001). Attempts to relate the dominance of either diatoms or prymnesiophyceans to environmental conditions (such as mixed layer depth, downwelling irradiance, macro- and micro-nutrient concentrations etc.) have been inconclusive and the mechanisms responsible for the specific composition of blooms have not been clarified (Goffart et al., 2000; Smith and Asper, 2001; van Hilst and Smith, 2002).

In the Ross Sea the distributions of organic matter and algal biomass show similar patterns, with higher concentrations in stations where photosynthetic processes are more intense, however, high heterotropic biomass and activity may be observed in ice free waters after ice melting (Nelson et al., 1989; Fabiano et al., 2000; Saggiomo et al., 2000; Gardner et al., 2000; Smith and Asper, 2001; Smith et al., 2006). A large amount of particulate and dissolved organic matter, deriving from primary production, is available to consumers of the higher trophic levels in the upper water column also entering in the detritus food web by means of the microbial loop (Azam et al., 1991; Riebsell et al., 1991; Gonzales, 1992).

In the Ross Sea, particulate organic carbon seems to satisfy the organic C requirement of the deep marine biota representing the main organic fuel of the biological pump (Ducklow et al., 2000) and ruling the mechanism of deep-sea remineralization (Azzaro et al., 2006) in spring-summer period. However, degradation processes change the concentration and composition of particulate and dissolved matter with depth. Phenomena mediated by heterotrophic micro-organisms (Carlson et al., 2000; Fabiano et al., 2000; Misic et al., 2002) and by physical and chemical processes (Mopper et al., 1995; Qian et al., 2001), influence the composition of organic material reaching the sediments (Dayton, 1990; Sayles et al., 2001; Povero et al., 2001).

3. Methods

Water mass, carbon and nitrogen exchanges across the continental slope

Lateral exchanges across the continental slope were estimated by quantifying advective transport of materials by water masses moving between the continental shelf and the open ocean. Hydrological data were processed by standard methods (UNESCO 1983, 1988). Temperature and salinity were used to extract the hydrological sub-sets corresponding to MCDW, HSSW and DISW from the general data set, following the criteria established by Jacobs et al. (1985), Trumbore et al. (1991) and Budillon et al. (2003). For each water mass weighted mean values were calculated for dissolved oxygen (O_2), dissolved inorganic carbon (DIC), inorganic nitrogen (DIN = NO_3+NO_2), dissolved and particulate organic carbon (DOC and POC) and particulate organic nitrogen (PON).

To estimate biochemical fluxes we must first quantify the water mass exchange between the continental shelf and the "open sea". We thus calculated the baroclinic transport using the hydrographic data. Low pass filtered current-meter data measured adjacent to the shelf break (Picco et al., 1999; Van Woert et al., 2003) were then used to resolve the intrinsic uncertainty of the geostrophic method for determining the absolute velocity field. Such procedure is, obviously, dependent on the reference values used and introduces an uncertainty in the mass flux estimation, which is difficult to assess. To minimize such uncertainty, we calculated the total transport along the shelf break and adjusted the reference levels to balance the out-going (primarily DISW and

HSSW) and in-going (MCDW) fluxes. To validate our results, the heat transport associated to water mass exchanges was also considered. As mentioned above, the major source of heat (and salt) for the continental shelf region is the CDW coming from the ACC. The CDW intrudes into the shelf region where it mixes with the shelf waters. The heat budget was thus calculated, considering the advective transport due to HSSW, DISW and MCDW at the continental shelf break. Clearly, this budget must differ from zero and the unbalanced amount should be compensated by the heat loss at sea surface over the whole Ross Sea continental shelf. Our estimation resulted in a value of -79 W m^{-2}, which is in very good agreement with a previous estimate (-96 W m^{-2}), based upon the analyses of the European Centre of Medium-Range Weather Forecast (Budillon et al., 2000).

Vertical budgets of carbon and nitrogen
The carbon and nitrogen budgets were constructed assuming the downward settling of particles to be the only exchange between the upper and deep layers. To date, models which take into account in the Ross sea also the convective vertical flux were considered unsatisfactory. Our assumption implies that convective vertical mixing between the upper and the deep layers is negligible, which seems reasonable due to the stratification that usually establishes in summer between the warmer and less salted surface water and the subsurface winter-water layer.

Several independent parameters were used to calculate fluxes of carbon and nitrogen along the water column and down to the sediment: gross phytoplankton primary production (GPP), carbon dioxide production rate (CDPR), drawdown of inorganic nitrogen (ΔDIN). Fluxes of particulate organic carbon (POC) and nitrogen (PON) were measured by means of sediment traps and, in a few cases, by ^{234}Th deficit, whereas burial rates were obtained from core samples.

All other values have been derived from these parameters according to the relationships and ratios shown in Table 1.

To assemble the integrated vertical budget, the Ross Sea was divided into 18 sub-regions, delimited by two degrees latitude and five degrees longitude (Fig. 1, 2). Sub-regions 1 and 2 were not computed in the area of the Ross Sea as they are mainly located beyond the continental shelf and were hence considered as belonging to the Southern Ocean domain. The periods of sampling extended from November through February (hereafter called summer) of different years.

Fig. 1

Fig. 2

For all stations within each sub-region, summer vertical fluxes were calculated and the mean value was multiplied by the area of the sub-region. In a few cases, no data were available for a given sub-region and values were hence taken from nearby areas or sites with similar characteristics. Vertical fluxes for the whole Ross Sea were obtained by summing the contributions of the 16 sub-regions covering the continental shelf, excluding sub-regions 1 and 2.

Gross Primary production
Primary production measurements were performed according to *in situ* simulated conditions for 3-4 hours around noon local time, in different areas and seasons over the last 12 years (for details see Saggiomo et al., 2002). In summer 1990, a time series of daily sampling was carried out in two stations of a coastal area of Victoria Land (Terra Nova Bay). During austral spring 1994, sampling was conducted for spatial and temporal variability in primary production in the Ross Sea polynya along a transect from Ross Ice Shelf to Cape Adare along 175°E. In January and February 1996 and 2001, two cruises were conducted in open and coastal water of the Ross Sea and primary production measurements were performed in different areas of the Ross Sea. In addition, during the 1996 and 2001 campaigns, photosynthetic parameters were obtained through P *vs* E experiments at several stations and at different depths of the water column.

Altogether, 218 stations were sampled at seven depths within the photic zone (0.1 % of incident irradiance) (Fig. 2b). Photosynthetic pigments were analyzed in all stations and primary production measurements were conducted at 98 stations. Further, 180 P vs E experiments were performed; average photosynthetic parameters have been defined for each of the Ross Sea areas and depth layers.

Total primary production in the Ross Sea (sub-regions 3-18) was estimated from November to February on the basis of *in situ* measurements (98 stations) and, for 120 other stations, by means of the equation established for the Ross Sea (Saggiomo et al., 2002). The equation considers the photosynthetic parameters for each area and depth layer and the mean daily irradiance and biomass concentration for each layer. An excellent correlation was found between measured and calculated primary production values for those stations where both *in situ* measurements and P vs E experiments were performed.

Net community production (NCP) was then calculated as the primary production minus the rate of carbon remineralization carried out by means of microbial community respiration (Tab. 1).

Electron Transport System activity (ETSa), Oxygen Utilisation Rates (OUR) and Carbon Dioxide Production Rates (CDPR)

Microbial community respiration evaluates oxidation of organic carbon both as DOC and POC, providing an overall estimate of inorganic carbon surplus in the marine system. Hence, it is of considerable importance for the understanding of the fate of organic matter especially within the deep layer. Moreover, since oxygen consumption is stoichiometrically related to carbon remineralization, it estimates the carbon remineralization as Carbon Dioxide Production Rate (CDPR).

Microbial community respiration is usually not measured directly in seawater samples, particularly in oligotrophic and deep-water environments, owing to their low rates. Indirect estimates have been obtained on large scale geochemical mass balance of oxygen and organic matter fields or *in vitro* oxygen consumptions (Aristeguì et al., 2005). The use of enzymatic indicators to estimate "potential" respiration is gaining acceptance due to their sensitivity, especially for meso- and bathy-pelagic waters (del Giorgio, 1992; Robinson and Williams, 2005). Electron Transport System activity (ETS) has therefore proved to be a useful direct tool to estimate the respiratory rates. In our study case, ETS measurements were performed according to Packard (1971, 1985) as modified by Kenner and Ahmed (1975), in 322 samples collected in 49 stations along the entire water column.

ETS was converted into oxygen utilisation rates (OUR, μl O2 h^{-1} dm^{-3}) by using the depth dependent factors currently adopted in recent oceanographic literature. Such extrapolation is questionable for data interpretation, but no more than those of other commonly used rate process techniques (del Giorgio, 1992). A OUR/ETS ratio value of 1 for the upper layer (Vosjan and Nieuwland, 1987; Aristeguì et al., 2002) and of 0.086 for the deeper layers (Packard et al.1988) were adopted by us. The error associated in the ETS assessment of respiration is about 30 % (Packard et al., 1988; Aristeguì and Montero, 1995).

Both the euphotic and aphotic OUR were converted into CDPR, by the equation:

$$\text{CDPR } (\mu g \text{ C } h^{-1} dm^{-3}) = (\text{OUR x } 12/22.4) \times (122/172)$$

where 12 is the C atomic weight, 22.4 the O_2 molar volume and 172/122 the Takahashi oxygen/carbon molar ratio (Takahashi et al., 1985). Depth-integrated euphotic and aphotic activities were calculated by means of the trapezoidal method.

Few data on ETS-derived respiration exists for the Antarctic surface waters (Martinez and Estrada, 1992; Aristegui and Montero, 1995; Crisafi et al., 2000). To our knowledge, with the exception of Azzaro et al. (2006), direct measurements of marine respiration from the deep Antarctic marine environments are still lacking.

Ducklow et al. (2000) performed respiratory rate measurements in the Ross Sea by estimating total Dark Community Respiration (DCR) from *in vitro* oxygen utilization in surface sea-water samples. The respiration rates determined were higher than our findings but comparable results will occur if the gap of 38 % between ETS and DCR is considered as established by Packard and Williams (1981).

Nutrient drawdown and remineralization
The summer nutrient deficit with respect to winter nutrient concentrations in the upper mixed layer (generally less than 100m) was calculated using the nutrient data from about 300 stations. They were sampled from late November to early February of three different years, covering a large portion of the Ross Sea, (Fig. 2a). The ΔDIN (winter-summer) deficit (Tab. 1) was estimated by the difference between the integrated quantities of DIN in the upper layer in winter and in summer. Winter nutrient concentrations were theoretically reconstructed from the concentration of DIN recorded in each station immediately below the upper mixed layer and considered as a memory of the winter values (Catalano et al., 1997).
We used a C:N ratio of 6.7 (Sweeney et al., 2000; Tab. 1) to calculate the equivalent inorganic carbon deficit ($\Delta DIC_{(deficit)}$) from ΔDIN (winter-summer). To obtain the total deficit for the entire Ross Sea (Gmol summer^{-1}) we calculated the average deficit (mmol m^{-2} summer^{-1}) for each Ross Sea shelf sub-region and then adding the values from the 16 sub-regions.
The integrated NCP value in the upper layer ($\Delta TOC_{(NCP)}$) was used to estimate the net total carbon assimilation into biomass ($\Delta POC_{(NCP)}$) and into the DOC pool ($\Delta DOC_{(NCP)}$). As suggested by Carlson et al. (2000) for the dissolved organic pool, we considered DOC to be 11% of NCP and a C/N ratio of 6.2 was used to calculate DON surplus from the DOC surplus (Table1).
In the deep layer (>100m), the vertical flux of POC and PON incoming from the upper 100 m has been considered as the substrate both for DIC and DIN surpluses produced by means of the CDPR of the microbial community and for the export toward higher trophic levels (eHTL) plus the dissolved organic pool (Tab. 1). The vertical flux was called $\Delta POC_{(drawdown)}$ and $\Delta PON_{(drawdown)}$ (Tab. 1) if derived from ΔDIN(winter-summer) deficit or $\Delta POC_{(export)}$ and $\Delta PON_{(export)}$ (Tab. 1) if derived from sediment traps (see below).
To calculate the DIN produced by microbial respiration ($\Delta DIN_{(OUR)}$), a DIN/O$_2$ ratio of 0.09 was applied to convert Oxygen Utilisation Rates (OUR) to the DIN surplus produced (Tab. 1) (Bender et al., 2000; Anderson and Sarmiento, 1994).
The lateral transport of matter due to advection was calculated separately for the upper and deep-water layers (UL and DL, respectively; Table 1).

Table 1 – *Definitions and list of independent variables used for calculating the derived parameters*

Estimates of sinking fluxes of particulate C and N from DIN drawdown and sediment traps
In addition to the evaluation of POC and PON export from the upper layer derived from DIN drawdown (Nelson et al., 2002), as described in the previous paragraph, the sinking flux of POC were also derived from sediment traps. Downward fluxes of biogenic particles were estimated by considering the data collected at 18 mooring sites (Fig. 2 c), deployed on the Ross Sea continental shelf since 1983 by several Italian and US research projects (DeMaster et al., 1992; Dunbar et al., 1998; Accornero et al., 1999, 2003; Collier et al., 2000; Langone et al., 2000, 2003; Accornero and Manno, 2005). Nine moorings were deployed for one year and the other nine moorings for a period of two to six years. The mean annual flux in each sub-region was estimated by averaging the mean annual fluxes of the moorings within the sub-region, independent of the duration of deployment. For each sub-region, the mean annual flux was multiplied by the total surface area to obtain the sub region-integrated annual flux.

Burial rates

During the last 15 years, several estimates of organic carbon accumulation on the bottom of the Ross Sea have been performed (DeMaster et al., 1996; Langone et al., 1998). Furthermore, information on organic carbon burial rates have also been provided by DeMaster et al. (1992), Hillfinger (1995), Frignani et al. (1998), Nishimura et al. (1998), Cunningham et al. (1999), Domack et al. (1999), Andrews et al. (1999), Brambati et al. (2002). From a total of 40 cores we compiled a list of sediment accumulation and burial rates of organic carbon and nitrogen (Fig 2d). All core chronologies were based on radiocarbon ages performed on carbonate or acid-insoluble organic matter. In most cases, the authors gave burial rates of biogenic components. Else, we calculated burial rates by combining data of sediment accumulation rates and of organic carbon and nitrogen concentrations from different studies carried out within the same Ross Sea sub-region.

The occurrence of coarse sediments on top of the banks (Anderson et al., 1984; Dunbar et al., 1985; Langone et al., 1998) implies the presence of strong currents, as measured at the shelf break (Picco et al., 1999), and, therefore, of erosion and/or no-deposition. We thus assumed that present-day biogenic sedimentation occurs only in areas deeper than 400 m. Total burial was calculated for such areas, which account for as much as 85% of the total Ross Sea continental shelf area (3.74×10^5 km^2 vs. 4.47×10^5 km^2).

4. Results and discussion

Water mass exchange and carbon and nitrogen budget between Ross Sea and open ocean
The estimates of the cross-slope transport along the Ross Sea continental margin are reported in Table 2. Budgets for the water mass exchange are expressed in Sv (1 Sv = 10^6 m^3 s^{-1}), while biogeochemical fluxes are in Gmol, both daily (Gmol day^{-1}) and for the entire period (Gmol summer^{-1}, i.e 120 days)

Table 2 - *Water mass exchange and carbon and nitrogen budgets between the Ross Sea and the Southern Ocean for the upper layer (depth <100 m) and for the deep layer (depth >100 m). The water mass exchange is reported in Sverdrup (Sv), budgets in Gmol both daily and for the entire period (120 days). Positive values are referred to incoming fluxes from the open ocean and negative to outgoing fluxes from Ross Sea.*

The final result of water mass exchange equaled -0.04 Sv, indicating a substantial equilibrium of the water exchange between the Ross Sea and the Southern Ocean. A loss of 0.04 Sv, likely due to the ice-water advection, seems quite reasonable. This approach does not consider the convective mixing between upper and deep layer and the application of a model, capable to account also for this phenomenon, will be the next challenge.

Recently Dinniman et al. (2003) used a high resolution three-dimensional numerical model to study the cross-shelf exchange in the Ross Sea. The annual mean cross-shelf exchanges of AABW (-0.66 Sv) and CDW (0.37) appear to be underestimate as recognized by the authors. Several model limitations can explain such deficiency, i.e. the climatological conditions used to force the model (which are temporally and spatially smoothed); the presence of intense polynyas in the Ross Sea which are not included in the model and the limitation of the imposed ice coverage. More recently Gordon et al., (2004) estimated the contribution Drygalski basin to the outflow of HSSW to the open ocean using CTD and LADCP measurements. In this way, they calculated a 0.5 to 0.8 Sv export of western Ross Sea shelf water considering a 250 m thick benthic layer. Such results are in good agreement with those obtained in the present work (0.65 Sv of HSSW, Table 2), especially considering the different approaches and methodology used, giving a further confirmation of the consistence of the estimation here reported.

Due to the water exchange across the continental slope, the Ross Sea as a whole exports a significant amount of dissolved inorganic carbon, 420 and 937 Gmol for upper and deep layer

respectively. The complexity of the equilibrium between the solubility pump and the biological pump and the presence of sea-ice for large part of the year are likely the cause of this result, despite the intense primary production which occurs in the upper layer during summer.

In contrast, a large amount of dissolved inorganic nitrogen is imported from the ocean, both within the upper and deep layer (50.7 and 18.6 Gmol, respectively). Finally, as expected, Ross Sea is a net source of particulate and dissolved organic matter: 140.8 Gmol of DOC are exported to the upper ocean during summer. Unfortunately, the measurements of dissolved organic nitrogen were too few and did not allow for calculating the dissolved organic nitrogen budget.

Vertical flux of carbon and nitrogen on the Ross Sea continental shelf

The vertical exports from the upper layer to the deep layer and to the sediment are summarized in Table 3 and in figures 3A,B and 4A,B. They are corrected for the lateral advection of carbon and nitrogen reported in Table 2.

To calculate the vertical budgets, we choose to start from gross primary production for two reasons. First, GPP considers the greatest carbon mass and, second, GPP measurements were available in a large number of stations with a good spatial and temporal coverage. Four fold differences in annual production were found between different areas (from 51 to 201 g-C $m^{-2}y^{-1}$). Highest production values occurred in the southern sectors; in the coastal area west of 165° E, from Terra Nova Bay to the Ross Ice Shelf, values ranged from 177 to 201 g-C $m^{-2}y^{-1}$. In offshore waters south of 74° S, values ranged from 90 to 180 g-C $m^{-2}y^{-1}$ (AVG=129, STD=31). North of 74°S variations were less pronounced, from 51 to 74 g-C $m^{-2}y^{-1}$ (AVG=61, STD=9).

Nelson et al. (1996) estimated 142 g-C $m^{-2}yr^{-1}$ considering six months of production (October to March) over an area of $3.3 \ 10^5$ km^2. They divided the Ross Sea into three sectors, not including the shelf break. In our case we considered an area of $4.5 \ 10^5$ km^2 as the sum of sub-regions 3-18 (Fig. 1) and a period of four months (November through February). Our estimate of the total gross primary production was 48.8 Tg-C or 4068 Gmol with an average of 109 g-C m^{-2}.

Table 3 - *Summary of biogeochemical fluxes of carbon and nitrogen on the Ross Sea continental shelf between the upper layer (depth <100 m) and the deep layer (depth >100 m) and between the deep layer and bottom*

Fig. 3

Fig. 4

Respiratory rates resulted low in such summer budget, showing slow oxidation rates of organic matter. Such finding corroborates the build-up of both particulate and dissolved organic matter that could remain available for the higher trophic levels or could be oxidized microbially during winter. Our estimates clearly do not include winter measurements nor reflect annual steady state conditions. It is likely that the importance of remineralization is greater on a year-round time scale, since respiratory efficiency has a lesser seasonal trend than primary production in Southern Ocean. The same can be assumed for the benthos metabolism, whose maintenance over longer periods can be hypothesized on charge of accumulated labile organic matter reservoir.

A seasonal production of 3239 Gmol of POC surplus (Tab. 3) is equivalent to 93 % of the net community production (3486 Gmol). This value matches very well the 94% found by Sweeney et al. (2000) in the outer part of the Ross Sea.

POC export to the deep layer calculated from DIN drawdown ranged between 1285-1681 Gmol of carbon (Tab. 3). The wide range in POC export depends upon to which extent DIC lateral losses (Tab. 2) were considered as resulting from the consumption of new production. (Tab. 3). This also implies that 40-52 % of the seasonally produced POC surplus is exported to the deep layer (Fig. 3A) that is in good agreement with the estimate of 50 % proposed by Nelson et al. (2002).

Subtracting the DOC surplus and POC export from the NCP value, we suggest that between 1558 and 1954 Gmol of carbon in the upper layer were exported to the higher trophic levels (i.e. 48 to 60 % of the POC surplus or 45-56 % of the NCP).

This pattern is significantly modified if the POC export is evaluated from sediment trap data (Fig. 3B), instead of DIN drawdown. In spite of the sediment trap data are referred to the whole year, they account for a vertical input of only 98.7 Gmol of POC into the deep layer (i.e. 3% of POC surplus in the upper layer), thus 3140 Gmol of carbon should be exported to higher trophic levels (i.e. 97 % of the POC surplus or 90 % of the NCP). This estimation reasonably agrees with the value of 133 Gmol of carbon, i.e. the 3.4% of POC production, calculated by Nelson et al. (1996) on the basis of data obtained by three mooring lines, two of which placed in the most productive area of the Ross Sea. It is noteworthy how this result is very different from that calculated on the basis of the DIN drawdown.

Sediment trap measurements yielded mean annual fluxes of biogenic materials ranging from 13.3 to 585.0 mmol m^{-2} for organic carbon and from 2.9 to 71.4 mmol m^{-2} for nitrogen. The highest carbon and nitrogen fluxes occurred in the south-central region (sub-regions 12 and 13). The integrated fluxes for the Ross Sea at 200 m (mean depth of sediment traps) were 98.7 Gmol for POC and 12.3 Gmol for PON. Pretty surprisingly, the POC export estimated in this study by sediment trap measurements over the whole basin is also in very good agreement with similar estimations performed over more limited areas and/or time periods (POC export versus gross primary production: 3.5%, Dunbar et al., 1998; 0.7-3%, Accornero et al., 2003; POC export versus POC stock in the upper water column: $\leq$ 3%, Smith and Dunbar, 1998; 2.3%, Asper and Smith, 1999; 0.5-1.5%, Accornero et al., 2003).

The sinking flux of POC was also derived from ^{234}Th deficits (Buesseler et al., 1992; Cochran et al., 1995) and from POC concentrations using the estimates of Cochran et al. (2000) and Langone et al. (1997) as recalculated by Buesseler et al. (2001). We calculated the ratio between NCP and ^{234}Th-derived export for those few stations where the ^{234}Th deficit was measured by us and applied these values to all stations where NCP was determined. During spring, POC export, as derived from ^{234}Th deficiency, was 8-10 % of NCP and ranged between 24 and 68 % in summer, therefore close to the estimate of 40-52 % calculated from DIN drawdown.

Considering our sediment trap results, they are consistent with other similar previous and synoptic experiments carried out in the Ross Sea, but, at the same time, they are one order of magnitude lower than the carbon and nitrogen exports estimated from DIN drawdown (Tab. 3) and from the few ^{234}Th determinations. The same disagreements were found by Nelson et al. (1996; 2002) and have been already pointed out by several other authors (Cochran et al., 2000; Collier et al., 2000; Sweeney et al., 2000).

The reason for such a large discrepancy in not thoroughly elucidated, although it appears now clear that sediment traps moored at relatively shallow depths (~ 200 m and anyhow < 1,000 m) substantially undercollect the actual flux of sinking POC (Yu et al., 2001; Scholten et al., 2001), possibly by as much as one order of magnitude in the Ross Sea (Fleisher and Anderson, 2003). This might be due to the ineffectiveness of moored sediment traps at collecting aggregates, which are particularly abundant in the Ross Sea (Asper and Smith, 2003) in the upper mesopelagic zone, while, deeper in the water column, the repackaging of these aggregates by zooplankton grazing may convert them into rapidly sinking particles, more efficiently collected (Yu et al., 2001)

The difference between the estimations based on DIN drawdawn (upper 100 m) and sediment trap measurements (average depth: 200 m) may also be at least partially due to the different thickness of the water layer considered: substantial remineralization and grazing can occur between 100 and 200 m depths. Remineralization has been shown to be important down to several hundreds of meters in the water column (Lee et al. 1998), as suggested by the general quasi-exponential decrease of POC flux and O_2 consumption between 100 and 500 m depth (Jenkins and Goldman, 1985; Martin et al., 1987).

In contrast to the carbon, the nitrogen budget was incomplete due to two major shortcomings: the data on dissolved organic nitrogen were too few and there are no data as to the denitrification rate in the Ross Sea.

Keeping these limits in mind, the nitrogen budget (Tab. 3) indicates a PON surplus of 483 Gmol from the NCP and 239 or 12 Gmol (if calculated from DIN drawdown or sediment traps, respectively) are exported to the deep layer (Fig. 4A,B). As a result, 244 Gmol (51 % of PON surplus) or 471 Gmol (98 % of PON surplus) are channelled to the higher trophic levels in the upper layer.

In the deep layer, 8 Gmol of PON are mineralized by respiration, hence 229 Gmol (as derived from DIN drawdown, Fig. 4A) or 3 Gmol (from sediment trap data, Fig. 4B) of nitrogen should be exported to the higher trophic levels and to the DON pool.

Carbon and nitrogen burial

The average burial rates of biogenic components ranged from 3.8 to 343.8 mmol C m^{-2} y^{-1} and from 0.9 to 33.7 mmol N m^{-2} y^{-1}. The lowest rates occurred in the northwestern portion of the Ross Sea, close to the shelf break. Maximum burial rates of all biogenic components were recorded in the southwestern sub-regions (11 and 12), close to Ross Island and along the coast in some protected bays. Biogenic sedimentary accumulation was relatively high also in the Joides basin (sub-region 4). Altogether, sub-regions 4, 11 and 12 accounted for about 80% of total organic carbon and nitrogen accumulation on the Ross Sea continental shelf.

The annual amounts of organic carbon and nitrogen accumulating on the seabed of the Ross Sea were 14.4 Gmol and 1.4 Gmol, respectively (Tab. 3). The organic carbon estimates were about twice the values provided by DeMaster et al. (1996), which may be partially explained by the different Ross Sea continental shelf area considered in the two studies (4.5×10^5 km^2 *vs* 3.3×10^5 km^2 for this and the DeMaster et al. study, respectively). Furthermore, the better spatial distribution of samples of this study provides a better approximation for the entire Ross Sea, in particular by considering the high sedimentary accumulation of carbon and nitrogen that occurred in the Joides basin.

As a result, only a very small portion (<0.5 %) of the POC resulting from the net community production is buried into the sediment. 63 Gmol are converted into carbon dioxide by microbial respiration and 1201-1597 Gmol or 15 Gmol of carbon are exported to higher trophic levels and to the DOC pool in the deep layer (Tab. 3), depending upon which flux of POC one prefers to consider: that from DIN drawdown (Fig. 3A) or that from sediment traps (Fig. 3B).

5. Final considerations

Our estimates are referred to an average austral summer (Fig. 3A,B and Fig. 4A,B) including the loss or gain of matter due to the water exchange across the continental slope (Tab.2) both in the upper and deep layer. The results are summarized in Table 3 and are representative of the summer pulse due to the enhancements of the biological processes. Some amounts of organic matter can thus accumulate, triggered by the high summer primary productivity. Assuming that a multi-annual derivative is not present, some of these accumulations, DOC for instance, should become equal to zero if integrated on the whole year. Unfortunately, we lack of over-winter measurements, when the productive processes are very low whereas others, like the water exchanges with the ocean, microbial degradation, grazing and burial, are still working. That said, our estimates indicate that the amount of matter recycled within the Ross Sea during summer is almost one order of magnitude higher than the amount exchanged with the ocean. The deep layer, the most relevant for the continental shelf pump, releases carbon to the deep ocean as DIC, DOC and POC (937, 13 and 7 Gmol, respectively) and contribute to oxygenate it with 1068 Gmol. As regards PON, less than 1

Gmol is released to the deep ocean, but the nitrogen balance becomes strongly positive considering the 19 Gmol of DIN imported from the open ocean (Tab 2).

The burial of C and N into the sediment is negligible as compared to the particulate matter produced in the upper layer (Fig. 3A,B; 4A,B), and a large export of matter (about 3155 Gmol of C and 473 Gmol of N, adding together upper and deep layers) is due to the incorporation into the higher trophic levels and partially (only for the deep layer) into the dissolved organic pool. This finding seems quite reasonable taking into account the high biomass supported by the primary production in the Ross Sea and in the Antarctic Seas in general.

A further consideration concerns the uptake of carbon dioxide into the upper layer in the Ross Sea during summer. The POC export from the upper to the deep layer can be also considered an estimate of the efficiency of the biological pump. It is noteworthy how this estimate results considerably different (40-52% instead of 3% of the POC surplus produced by primary production) according to the calculation based on the DIN drawdown or the sediment traps.

As a conclusion, our budget of carbon and nitrogen in the Ross Sea certainly suffers of some important approximations such as the lack of a convective mixing model and the lack of data on dissolved organic nitrogen and on nitrogen loss due to denitrification. However, our aim was to make a first step to provide carbon and nitrogen budgets in a way as comprehensive as possible by merging biogeochemical data obtained using a wide variety of experimental approaches with data on water mass circulation on the Ross Sea shelf.

References

Accornero A, Bergamasco A, Monaco A, Tucci S (1999). Particle fluxes at the edge of the Ross Ice Shelf: the role of physical forcing. In: Spezie G, Manzella GMR (eds.). The Oceanography of the Ross Sea - Antarctica. Springer-Verlag Italia, Milano, pp 178-195

Accornero A, Manno C (2005). Annual patterns of vertical fluxes in the Ross Sea: a tool to reconstruct seasonal biogeochemical dynamics in the water column. 3rd International Conference on the Oceanography of the Ross Sea (Antarctica), Venezia, 10-14 Ottobre 2005

Accornero A, Manno C, Arrigo KR, Martini A, Tucci S (2003). The vertical flux of particulate matter in the polynya of Terra Nova Bay: results from moored sediment traps (1995-1997). Part I. Chemical constituents. Antarctic Science, 15(1):119-132

Anderson JB, Brake CF, Myers NC (1984). Sedimentation on the Ross Sea continental shelf, Antarctica. Mar. Geol., 57:295-333

Anderson LA, Sarmiento JL (1994). Redfield ratios of remineralization determined by nutrient data analysis. Global Biogeochem. Cycles, 8:65-80

Andrews JT, Domack EW, Cunningham WL, Leventer A, Licht KJ, Jull AJT, DeMaster DJ, Jennings AE (1999). Problems and possible solutions concerning radiocarbon dating of surface marine sediments, Ross Sea, Antarctica. Quat. Res., 52:206-216

Arìsteguì J, Agustì S, Middelburg JJ, Duarte CM (2005). Respiration in the mesopelagic and bathypelagic zones of the oceans. In: Del Giorgio PA, Williams Pjle B (eds). Respiration in aquatic ecosystems. Oxford University Press, pp:181-205

Arìsteguì J, Denis M, Almunia J, Montero MF (2002). Water-column remineralization in the Indian sector of the Southern Ocean during early spring. Deep-Sea Res. I, 49:1707-1720

Aristeguì J, Montero MF (1995). Plankton community respiration in Bransfield Strait (Antarctic Ocean) during austral spring. J. Plankt. Res., 17:1647-1659

Arrigo KR, Robinson DH, Worthen DL, Dunbar RB, DiTullio GR, VanWoert M, Lizotte MP (1999). Phytoplankton community structure and the drawdown of nutrients and CO_2 in the Southern Ocean. Science, 283:365-367

Arrigo KR, Worthen DL, Schieber B, Lizotte MP (1998). Primary production in Southern Ocean waters. J. Geophys. Res., 103:15587-15600

Asper VL, Smith WO Jr. (1999). Particle fluxes during austral spring and summer in the southern Ross Sea, Antarctica. J. Geophys. Res., 104:5345-5360

Asper VL, Smith WO Jr. (2003). Abundance, distribution and sinking rates of aggregates in the Ross Sea Polynya. Deep-Sea Res. I, 50:131-150

Azam F, Smith DC, Hollibaugh JT (1991). The role of microbial loop in Antarctic pelagic ecosystem. In: Sakshaug E, Hopkins CCE, Øritsland NA (eds). Proceedings of the Pro Mare Symposium on Polar Marine Ecology, Trondheim, Polar Res., 10:239-243.

Azzaro M, La Ferla R, Azzaro F (2006). Microbial respiratory activity in the aphotic zone of the Ross Sea (Antarctica). Mar. Chem., 99:199-209

Bender ML, Dickson M, Orchardo J (2000). Net and gross production in the Ross Sea as determined by incubation experiments and dissolved O2 studies. Deep-Sea Res. II, 47:3141-3158

Bergamasco A, Defendi V, Budillon G, Spezie G (2004). Down flow observations near Cape Adare shelf break. Antarctic Science, 16(2):199-204

Bergamasco A, Defendi V, Meloni R (2002). Some dynamics of water outflow from beneath the Ross Ice Shelf during 1995/96. Antarctic Science, 14(2):74-82

Brambati A, Corradi N, Finocchiaro F, Giglio F (2002a). The position of the grounding line in the Joides Basin: an interpretation based on Sedimentological and geochemical Data. In Gamble J.A., Skinner D.N.B., Henrys S., e R. Lynch (Editors). Antarctica at the close of a Millenium. Royal Society of New Zealand Bulletin 35:365-372.

Brambati A, Fanzutti GP, Finocchiaro F, Melis R, Pugliese N, Salvi G, Faranda C (1999). Some palaeoecological remarks on the Ross Sea shelf, Antarctica. In: Faranda FM, Guglielmo L, Ianora A (eds.), Ross Sea Ecology. Springer-Verlag, Berlin Heidelberg New York, pp 51-61

Brambati A, Melis R, Quaia T, Salvi G (2002b). Late Quaternary climatic changes inthe Ross Sea area, Antarctica. In: Gamble J.A., Skinner D.N.B., Henrys S., Lynch R. (Editors) Antarctic Earth Sciences at the Close of a Millennium: Proceedings Volume 8th International Symposium on Antarctic Earth Sciences: Royal Society of New Zealand Bulletin, 35:359-364.

Bromwich DH, Kurtz DD (1984). Katabatic wind forcing of the Terra Nova Bay polynya. J. Geophys. Res., 89:3561-3572

Budillon G, Cordero Gremes S, Salusti E (2002). On the Dense Water Spreading off the Ross Sea shelf. J. Mar. Systems, 35(3-4):207-227

Budillon G, Fusco G, Spezie G (2000). A study of surface heat fluxes in the Ross Sea (Antarctica). Antarctic Science, 12:243-254

Budillon G, Pacciaroni M, Cozzi S, Rivaro P, Catalano G, Ianni C, Cantoni C (2003). An Optimum multiparameter mixing analysis of the shelf waters in the Ross Sea. Antarctic Science, 15(1):105-118

Budillon G, Salusti E, Tucci S (2006). The evolution of density currents and nepheloid bottom layers in the Ross sea (Antarctica). J. Mar. Res., (in press).

Budillon G, Spezie G (2000). Thermohaline structure and variability in the Terra Nova Bay Polynya (Ross Sea) between 1995-98. Antarctic Science 12(4):501-516

Budillon G, Tucci S, Artegiani A, Spezie G (1999). Water masses and suspended matter characteristics of the western Ross Sea. In: Faranda FM, Guglielmo L, Ianora A (eds.), Ross Sea Ecology. Springer-Verlag, Berlin Heidelberg New York, pp 63-81

Buesseler KO, Bacon MP, Cochran JK, Livingstone HD (1992). Carbon and nitrogen export during the JGOFS North Atlantic Bloom Experiment estimated from ^{234}Th:^{238}U disequilibrium. Deep-Sea Res., 39:1115-1137

Buesseler KO, Ball L, Andrews J, Cochran JK, Hirschberg DJ, Bacon MP, Fleer A, Brzezinski M (2001). Upper ocean export of particulate organic carbon and biogenic silica in the Southern Ocean along 170°W. Deep-Sea Res. II, 48:4275-4297

Carlson CA, Hansell DA, Peltzer ET, Smith WO Jr, (2000). Stocks and dynamics of dissolved and particulate organic matter in the southern Ross Sea, Antarctica. Deep-Sea Res. II, 47:3201-3225

Catalano G, Benedetti F, Predonzani S, Goffart A, Ruffini S. Rivaro P, Falconi C (2000). Spatial and temporal patterns of nutrient distributions in the Ross Sea. In: Faranda FM, Guglielmo L, Ianora A (eds.), Ross Sea Ecology. Springer-Verlag, Berlin Heidelberg New York, pp 107-120

Catalano G, Povero P, Fabiano M, Benedetti F, Goffart A (1997). Nutrient utilisation and particulate organic matter changes during summer in the upper mixed layer (Ross Sea, Antarctica). Deep-Sea Res., 44:97-112

Cochran JK, Barnes C, Achman D, Hirschberg DJ (1995). Thorium-234/Uranium-238 disequilibrium as an indicator of scavenging rates and particulate organic carbon fluxes in the Northeast water polynya, Greenland. J. Geophys. Res., 100:4399-4410

Cochran JK, Buesseler KO, Bacon MP, Wang HW, Hirschberg DJ, Ball L, Andrews J, Crossin G, Fleer A (2000). Short-lived thorium isotopes (^{234}Th, ^{228}Th) as indicators of POC export and particle cycling in the Ross Sea, Southern Ocean. Deep-Sea Res. II, 47:3451-3490

Collier R, Dymond J, Honjo S, Manganini S, Francois R, Dunbar R (2000). The vertical flux of biogenic and lithogenic material in the Ross Sea: moored sediment trap observations 1996-1998. Deep-Sea Res. II, 47:3491-3520

Crisafi E, Azzaro F, La Ferla R, Monticelli L (2000). Microbial biomass and respiratory activity related to the ice-melting upper layers in the Ross Sea (Antartica). In: In: Faranda FM, Guglielmo L, Ianora A (eds.), Ross Sea Ecology. Springer-Verlag, Berlin Heidelberg New York, pp 171-180

Cunningham WL, Leventer A, Andrews JT, Jennings AE, Licht KJ (1999). Late-Pleistocene-Holocene marine conditions in the Ross Sea, Antarctica: evidence from the diatom record. The Holocene, 9:129-139

Dayton PK (1990). Polar Benthos. In: Smith WO Jr (ed.). Polar Oceanography, Part. B, Academic Press, London, pp 637-676

del Giorgio PA (1992). The relationships between ETS (electron transport system) activity and oxygen consumption in lake plankton – a cross system calibration. J. Plankton Res., 14:1723-1741

DeMaster DJ, Dunbar RB, Gordon LI, Leventer AR, Morrison JM, Nelson DM, Nittrouer, CA, Smith WO Jr (1992). Cycling and accumulation of biogenic silica and organic matter in high-latitude environments: the Ross Sea. Oceanography, 5:146-153

DeMaster DJ, Ragueneau O, Nittrouer CA (1996). Preservation efficiencies and accumulation rates for biogenic silica and organic C, N and P in high-latitude sediments: the Ross Sea. J. Geophys. Res., 101:18501-18518

Dinniman MS, Klinck JM, Smith WO Jr (2003). Cross-shelf exchange in a model of the Ross Sea circulation and biogeochemistry. Deep-Sea Res. II, 50:3103–3120

Di Tullio GR, Smith WO Jr (1996). Spatial patterns in phytoplankton biomass and pigment distributions in the Ross Sea. J. Geophys. Res., 101:18467-18477

Domack EW, Jacobson EA, Shipp S, Anderson JB (1999). Late Pleistocene-Holocene retreat of the West Antarctic Ice –Sheet system in the Ross Sea: Part 2-Sedimentologic and stratigraphic signature. Geol. Soc. Am. Bull., 111(10):1517-1536

Ducklow HW, Dickson ML, Kirchmann DL, Steward G, Orchardo J, Marra J, Azam F (2000). Constraining bacterial production, conversion efficiency and respiration in the Ross Sea, Antarctica, January-February, 1997. Deep-Sea Res. II, 47: 3227-3247

Dunbar RB, Anderson JB, Domack, EW, Jacobs SS (1985). Oceanographic influences on sedimentation along the Antarctic continental shelf. In: Jacobs S.S. (ed.), Oceanology of the Antarctic Continental Shelf: Ant. Res. Series, AGU, 43:291-312

Dunbar RB, Leventer AR, Mucciarone DA (1998). Water column sediment fluxes in the Ross Sea, Antarctica: Atmopheric and sea ice forcing. J. Geophys. Res., 103:30741-30759

Fabiano M, Povero P, Misic C (2000). Spatial and temporal distribution of particulate organic matter in the Ross Sea. In: Faranda FM, Guglielmo L, Ianora A (eds.), Ross Sea Ecology. Springer-Verlag, Berlin Heidelberg New York, pp 135-149

Fitzwater SE, Johnson KS, Gordon RM, Coale KH, Smith WO Jr (2000). Trace metal concentrations in the Ross Sea and their relationship with nutrients and phytoplankton growth. Deep-Sea Res. II, 47:3159-3179

Fleisher MQ, Anderson RF (2003). Assessing the collection efficiency of Ross Sea sediment traps using ^{230}Th and ^{231}Pa. Deep-Sea Res. II, 50: 693-712

Frignani M, Giglio F, Langone L, Ravaioli M, Mangini A (1998). Late-Quaternary fluctuations of organic carbon and biogenic silica fluxes to sediments of the Northwestern Ross Sea (Antarctica). Annals of Glaciology, 27:697-703

Fusco G, Flocco D, Budillon G, Spezie G, Zambianchi E (2002). Dynamics and variability of Terra Nova Bay Polynya. Marine Ecology PSZNI, Supplement I:201-209

Gardner WD, Richardson MJ, Smith WO Jr. (2000). Seasonal patterns of water column particulate organic carbon and fluxes in the Ross Sea. Deep-Sea Res. II, 47:3423-3449

Goffart A, Catalano G, Hecq JH (2000). Factors controlling the distributions of diatom and *Phaeocystis* in the Ross Sea. J. Mar. Systems, 27:161-175

Gonzales HE (1992). The distribution and the abundance of krill faecal material and oval pellets in the Scotia and Weddel Seas (Antarctica) and their role in particle flux. Pol. Biol., 12:81-91

Gordon AL, Tchernia P (1972). Waters of the continental margin off Adelie coast, Antarctica. In Hayes DE (ed.), Antarctic Research Series, Antarctic Oceanology II: The Australian–New Zealand sector. American Geophysical Union, pp 56-69

Gordon AL, Zambianchi E, Orsi A, Visbeck M, Giulivi CF, Withworth III T, Spezie G (2004). Energetic plumes over the western Ross Sea continental slope. Geophysical Research Letters, 31,L21302, doi:10.1029/2004GL020785

Gordon LI, Codispoti LA, Jennings Jr JC, Millero FJ, Morrison JM, Sweeney C (2000). Seasonal evolution of hydrographic properties in the Ross Sea, Antarctica 1996 – 1997. Deep-Sea Res. II, 47:3095-3117

Hillfinger M (1995). Chronology of the Late Pleistocene-Holocene deglaciation in the Ross Sea. B.A. thesis, Clinton, N.Y. Hamilton College, 59 pp.

Jacobs SS (1991). On the nature and significance of the Antarctic slope front. Mar. Chem., 35:9-24

Jacobs SS, Amos AF, Bruchhausen PM (1970). Ross Sea oceanography and Antarctic Bottom Water formation. Deep-Sea Res., 17:935-962

Jacobs SS, Comiso JC (1989). Sea ice and oceanic processes on the Ross Sea continental shelf. J. Geophys. Res., 94(C12):18195-18211

Jacobs SS, Fairbanks RG, Horibe Y (1985). Origin and evolution of water masses near the Antarctic continental margin: evidence from H218O/H216O ratios in seawater. Antarctic Research Series, 43:59-85

Jacobs SS, Giulivi CF (1998). Interannual ocean and sea ice variability in the Ross Sea. Antarctic Research Series, 75:135-150

Jacobs SS, Giulivi CF (1999). Thermohaline data and ocean circulation on the Ross Sea continental shelf. In: Spezie G, Manzella GMR (eds.). Oceanography of the Ross Sea – Antarctica. Springer-Verlag Italia, Milano, pp 3-16

Jacobs SS, Gordon, AL, Ardai JL Jr. (1989). Circulation and melting beneath the Ross Ice Shelf. Science, 203: 439-442

Jenkins WJ, Goldman JC (1985). Seasonal oxygen cycling and primary production in the Sargasso Sea. J. Mar. Res., 43:465-491

Kenner RA, Ahmed SJ (1975). Measurement of electron transport activities in marine phytoplankton. Marine Biology, 33:119-128

Kurtz DD, Bromwich DH (1983). Satellite observed behavior of the Terra Nova Bay polynya. J. Geophys. Res., 88:9717-9722

Kurtz DD, Bromwich DH (1985). A recurring atmospherically-forced polynya in Terra Nova Bay. In: Jacobs SS (ed.). Oceanology of the Antarctic Continental Shelf. Antarctic Research Series, 43:177-201

Langone L, Frignani M, Labbrozzi L, Ravaioli M (1998). Present-day biosiliceous sedimentation in the NW Ross Sea (Antarctica). J. Mar. Systems, 17:459-470

Langone L, Frignani M, Cochran JK, Ravaioli M (1997). Scavenging processes and export fluxes close to a retreating seasonal ice margin (Ross Sea, Antarctica). Water, air, soil pollution, 99:705-715

Langone L, Frignani M, Ravaioli M, Bianchi C (2000). Particle fluxes and biogeochemical processes in an area influenced by seasonal retreat of the ice margin (Northwestern Ross Sea, Antarctica). J. Mar. Systems, 27:221-234

Langone L, Dunbar RB, Mucciarone DA, Ravaioli M, Meloni R, Nittrouer CA (2003). Rapid sinking of biogenic material during the late austral summer in the Ross Sea, Antarctica. In Biogeochemistry of the Ross Sea, AGU, Antarctic Research Series Monograph, G. DiTullio and R. Dunbar (eds.). 78:221-234

Lee C, Murray DW, Barber RT, Buesseler KO, Dymond J, Hedges JI, Honjo S, Manganini SJ, Marra J, Moser C, Peterson ML, Prell WL, Wakeham SG (1998). Particulate organic carbon fluxes: compilation of results from the 1995 US JGOFS Arabian Sea process study. Deep-Sea Res. II, 45:2489-2501

Locarnini RA (1994). Water masses and circulation in the Ross Gyre and environs. Dissertation for the PhD, Texas A&M University, 87 pp

Martin JH, Knauer GA, Karl DM, Broenkow WW (1987). VERTEX: carbon cycling in the northeast Pacific. Deep-Sea Res., 34:267-285

Martinez R, Estrada M (1992). Respiratory electron transport activity of microplankton in the Weddell Sea during early spring: influence of the ice cover and the ice edge. Polar Biol., 12:275-282

Misic C, Povero P, Fabiano M (2002). Ectoenzymatic Ratios In Relation To Particulate Organic Matter Distribution (Ross Sea, Antarctica). Microbial Ecology, 44:224-234

Mopper K, Sarpal RS, Kieber DJ (1995). Protein and humic substanc fluorescence of dissolved organic matter in antarctic sea water. Antarctic Journal of the U.S., 30(5)

Nelson DM, Anderson RF, Barber RT, Brzezinski MA, Buesseler KO, Chase Z, Collier RW, Dickson M, Francois R, Hiscock MR, Honjo S, Marra J, Martin WR, Sambrotto RN, Sayles FL, Sigmon DE (2002). Vertical budgets for organic carbon and biogenic silica in the Pacific sector of the Southern Ocean, 1996-1998. Deep-Sea Res. II, 49:1645-1674

Nelson DM, DeMaster DJ, Dunbar RB, Smith WO Jr (1996). Cycling of organic carbon and biogenic silica in the Southern Ocean: Estimates of water-column and sedimentary fluxes on the Ross Sea continental shelf. J. Geophys. Res., C8:18519-18532

Nelson DM, Smith Jr WO (1991). Sverdrup rivisited: critical depth, maximum chlorophyll levels, and the control of Southern Ocean productivity by the irradiance-mixing regime. Limnol. Oceanogr., 36:1650-1661

Nelson DM, Smith WO Jr, Muench RD, Gordon LI, Sullivan CW, Husby DM (1989). Particulate matter and nutrient distributions in the ice-edge zone of the Weddel Sea: relationship to hydrography during late summer. Deep-Sea Res., 36:191-209

Nishimura A, Nakasone T, Hiramatsu C, Tanahashi M (1998). Late Quaternary paleoenvironment of the Ross Sea continental shelf, Antarctica. Annals of Glaciology, 27:275-280

Packard TT (1971). The measurement of respiratory electron transport activity in marine phytoplankton. J. Mar. Res., 29:235-244

Packard TT (1985). Measurement of respiratory electron transport activity of microplankton. In: Jannasch HW, Williams PJ-LeB. (eds), Advances in Aquatic Microbiology. Academic Press, New York, Vol. 3, pp 207-261

Packard TT, Denis M, Rodier M, Garfield P (1988). Deep-ocean metabolic CO_2 production: calculations from ETS activity. Deep-Sea Res., 35: 371-382

Packard TT, Williams PJ-LeB (1981). Rates of respiratory oxygen consumption and electron transport in surface sea water from the Northwest Atlantic. Oceanol. Acta, 4:351-358

Picco P, Amici L, Meloni R, Langone L, Ravaioli M (1999). Temporal variability of currents in the Ross Sea (Antarctica). In: Manzella GMR, Spezie G (eds) Oceanography of the Ross Sea", Springer-Verlag Italia, Milano, pp 103-117

Povero P. Chiantore M., Misic C., Budillon G., Cattaneo-Vietti R. (2001). Pelagic-benthic coupling in Adelie Cove (Terra Nova Bay, Antarctica): a strongly land forcing controlled system. Polar Biol., 24:875-882

Qian J, Mopper K, Kieber DJ (2001). Photochemical production of the hydroxyl radical in Antarctic waters. Deep-Sea Res. I, 48:741-75

Riebsell U, Shloss I, Smetacek V (1991). Aggregation of algae released from melting sea ice: implications for seeding and sedimentation. Polar Biol., 11:239-248

Rodman M, Gordon A (1982). Southern Ocean bottom waters in the Australia - New Zealand sector. J. Geophys. Res., 87:5771-5778

Robinson C, Williams PJ (2005). Respiration and its measurement in surface marine waters. In: Del Giorgio PA, Williams Pjle B (eds). Respiration in aquatic ecosystems. Oxford University Press, pp 181-205

Rubino A, Budillon G, Pierini S, Spezie G (2002). A model for the spreading and sinking of the Deep Ice Shelf Water in the Ross Sea. Antarctic Science, 15(1), 25-30

Saggiomo V, Carrada GC, Mangoni O, Marino D, Ribera d'Alcalà M (2000). Ecological and physiological aspects of primary production in the Ross Sea. In: Faranda FM, Guglielmo L, Ianora A (eds.), Ross Sea Ecology. Springer-Verlag, Berlin Heidelberg New York, pp 247-258

Saggiomo V, Catalano G, Mangoni O, Budillon G, Carrada GC (2002). Primary production processes in ice-free waters of the Ross Sea (Antarctica) during the austral summer 1996. Deep-Sea Res. II, 49:1787-1801

Sakshaug E, Slagstad D, Holm-Hansen O (1991). Factors controlling the development of phytoplankton blooms in the Antarctic Ocean - a mathematical model. Mar. Chem, 35:259-271

Sayles FL, Martina WR, Chaseb Z, Andersonb RF (2001). Benthic remineralization and burial of biogenic SiO_2, $CaCO_3$, organic carbon, and detrital material in the Southern Ocean along a transect at 1701 West. Deep-Sea Res. II, 48:4323–4383

Scholten JC, Fietzke J, Vogler S, Rutgers van der Loeff M, Mangini A, Koeve W, Waniek J, Stoffers P, Antia A, Kuss J (2001). Trapping efficiency of sediment traps from the deep eastern North Atlantic: the ^{230}Th calibration. Deep-Sea Res. II, 48(10):2383-2408

Sedwick PN, DiTullio GR (1997). Regulation of algal blooms in Antarctic shelf waters by release of iron from melting sea ice. Geopys. Res. Lett., 24:2515-2518

Sedwick PN, DiTullio GR, Mackey DJ (2000). Iron and manganese in the Ross Sea, Antarctica: seasonal iron limitation in Antarctic shelf waters. J. Geophys. Res., 105:11321-11336

Smith WO Jr, Asper VA (2000). A balanced nitrogen budget of the surface layer of the southern Ross Sea, Antarctica. Geophys. Res. Lett., 27:2721-2724

Smith WO Jr, Asper VA (2001). The influence of phytoplankton assemblage composition on biogeochemical characteristics and cycles in the southern Ross Sea. Deep-Sea Res. I, 48:137-162

Smith WO Jr, Dinniman MS, Klinck JM, Hofmann E (2003). Biogeochemical climatologiesin the Ross Sea, Antarctica: seasonal pattern of nutrients and biomass. Deep-Sea Res. II, 50:3083-3101

Smith WO Jr, Dunbar RB (1998). The relationship between new production and vertical flux on the Ross Sea continental shelf. J. Mar. Systems 17:445-457

Smith WO Jr, Gordon LI (1997). Hyper productivity of the Ross Sea (Antarctica) Polynya. Geophys. Res. Lett., 24:233-236

Smith WO Jr, Shields AR, Peloquin JA, Catalano G, Tozzi S, Dinniman MS, Asper VA (2006). Interannual variations in nutrients, net community production, and biogeochemical cycles in the Ross Sea. Deep-Sea Res. II, in press

Sweeney C, Hansell DA, Carlson CA, Codispoti LA, Gordon LI, Marra J, Millero FJ, Smith WO Jr, Takahashi T (2000). Biogeochemical regimes, net community production and carbon export in the Ross Sea, Antarctica. Deep-Sea Res. II, 47:3369-3394

Takahashi T, Broecker WS, Langer S (1985). Redfield ratio based on chemical data from isopycnal surfaces. J. Geophys. Res., 90:6907-6924

Taviani M, Reid DE, Anderson JB (1993). Skeletal and isotopic composition and paleoclimatic significance of Late Pleistocene carbonates, Ross Sea, Antarctica. Journal of Sedimentary Petrology 63:84–90

Trumbore SE, Jacobs SS, Smethie WM Jr (1991). Chlorofluorocarbon evidence for rapid ventilation of the Ross Sea. Deep-Sea Res., 38:845-870

UNESCO (1983). Algorithms for computation of fundamental properties of seawater. Technical Papers in Marine Science, pp 44-53

UNESCO (1988). The acquisition, calibration and analysis of CTD data. A report of SCOR WG 51. Technical Papers in Marine Science, pp 54-59

van Hilst CM, Smith WO Jr (2002). Photosynthesis/irradiance relationships in the Ross Sea, Antarctica, and their control by phytoplankton assemblage composition and environmental factors. Mar Ecol Prog Ser 226:1-12

Van Woert ML (1999). Wintertime dynamics of the Terra Nova Bay polynya. J. Geophys. Res., 104(C49):7753-7769

Van Woert ML, Johnson ES, Langone L, Worthen DL, Monaghan AJ, Bromwich DH, Meloni R, Dunbar RB (2003). The Ross Sea circulation during 1990's. In: DiTullio G, R. Dunbar R (eds.). Biogeochemistry of the Ross Sea. American Geophysical Union, Antarctic Research Series, 78:5-34.

Vosjan JH, Nieuwland G (1987). Microbial biomass and respiratory activity in the surface waters of the east Banda Sea and the northwest Arafura Sea (Indonesia) at the time of the southeast monsoon. Limnol. Oceanogr., 32:767-775

Yu EF, François R, Bacon MP, Honjo S, Fleer AP, Manganini SJ, Rutgers van der Loeff MM, Ittekot V (2001). Trapping efficiency of bottom-tethered sediment traps estimated from the intercepted fluxes of ^{230}Th and ^{231}Pa. Deep-Sea Res. I, 48:865-889

Zwally HJ, Comiso JC, Gordon AL (1985). Antarctic offshore leads and polynyas and oceanographic effects. In Jacobs S.S. (ed.). Oceanology of the Antarctic continental shelf, Antarctic Research Series 43, AGU, Washington, pp 203-226

Figure captions

Fig. 1 - Bathymetric map of the Ross Sea and the division in sub-regions used for the calculations of mean fluxes (see text). BTN, Terra Nova Bay; DB, Drygalski Basin; CB, Crary Bank; MB, Maweson Bank; JB, Joides Basin; IB, Iselin Bank; PB, Pennell Bank; McMS, McMurdo Sound; GCB, Glomar Challenger Basin; LAB, Little America Basin.

Fig. 2 –Sampling sites visited in different cruises over the period 1987-2001 a) CTD and nutrient hydro-casts, b) primary productivity, c) moorings, d) sediment cores

Fig. 3 - Diagram of the flux of carbon between the upper layer and the deep layer as derived from A) DIN drawdown; B) sediment traps. The budgets are balanced for the lateral advection (ul - upper layer, dl - deep layer). All quantities are in Gmol.

Fig. 4 - Diagram of the flux of nitrogen between the upper layer and the deep layer as derived from A) DIN drawdown; B) sediment traps. The budgets are balanced for the lateral advection (ul - upper layer, dl - deep layer). All quantities are in Gmol

Table 1 – *Definitions and list of independent variables used for calculating the derived parameters*

Upper Layer (0 - 100 m)

ΔGPP	Summer gross phytoplankton production (measured)
$\Delta CDPR$	Summer carbon dioxide production due to community respiration (measured)
$\Delta TOC_{(NCP)} = \Delta GPP - \Delta CDPR$	Summer total organic carbon surplus due to the net community production (NCP)
$\Delta TOC_{(NCP)} = \Delta POC_{(NCP)} + \Delta DOC_{(NCP)}$	Carbon partitioning of the NCP
$\Delta TON_{(NCP)} = \Delta PON_{(NCP)} + \Delta DON_{(NCP)}$	Nitrogen partitioning of the NCP
$\Delta DIC_{(UL)}$	DIC advection calculated from the flux across the Ross Sea continental margin (Tab. 2)
$\Delta DOC_{(UL}$	DOC advection calculated from the flux across the Ross Sea continental margin (Tab.2)
$\Delta POC_{(UL)}$	POC advection calculated from the flux across the Ross Sea continental margin (Tab. 2)
$\Delta DIN_{(UL)}$	DIN advection calculated from the flux across the Ross Sea continental margin (Tab. 2)
$\Delta PON_{(UL)}$	PON advection calculated from the flux across the Ross Sea continental margin (Tab. 2)
$\Delta DOC_{(surplus)} = 0.11 * \Delta TOC_{(NCP)} + \Delta DOC_{(UL)}$	Summer DOC surplus coming from the partitioning of the NCP
$\Delta POC_{(surplus)} = \Delta TOC_{(NCP)} - \Delta DOC_{(surplus)} + \Delta POC_{(UL)}$	Summer POC surplus coming from the partitioning of the NCP
$\Delta DIC_{(deficit)} = \Delta DIN(winter-summer)*6.7 + \Delta DIN_{(UL)} * 6.7$	Calculated DIC deficit from observed ΔDIN deficit between winter and summer integrated values
$\Delta POC_{(drawdown)} = \Delta DIC_{(deficit)} + \Delta DIC_{(UL)}$	Summer POC drawdown from the upper layer deriving from DIN deficit (new production)
$\Delta POC_{(export)}$	Export of POC by sediment traps (measured)
$\Delta PON_{(surplus)} = \Delta POC_{(surplus)}/6.7 + \Delta PON_{(UL)}$	Summer PON surplus due to the NCP
$\Delta DON_{(surplus)} = \Delta DOC_{(surplus)}/6.2$	Summer DON surplus calculated from the ΔDOC surplus
$\Delta DIN_{(deficit)} = \Delta DIN(winter-summer) + \Delta DIN_{(UL)}$	Observed DIN deficit between winter and summer integrated values
$\Delta PON_{(drawdown)} = \Delta DIN_{(deficit)}$	Summer PON drawdown from the upper layer calculated from DIN deficit (new production)
$\Delta PON_{(export)}$	Export of PON by sediment traps (measured)
$\Delta C_{(eHTL)} = \Delta POC_{(surplus)} - \Delta POC_{(drawdown)} - \Delta POC_{(UL)}$	Pool of Carbon exported to the higher trophic levels according to DIN deficit (new production)
$\Delta N_{(eHTL)} = \Delta PON_{(surplus)} - \Delta PON_{(drawdown)} - \Delta PON_{(UL)}$	Pool of Nitrogen exported to the higher trophic levels according to DIN deficit (new production)

Deep Layer (100 m - bottom)

$\Delta POC_{(drawdown)}$ or $\Delta POC_{(export)}$	Summer export of POC from the upper layer either from DIN drawdown or from sediment traps
$\Delta POC_{(burial)}$	Burial of particulate carbon derived from sediment cores (observed)
$\Delta DIC_{(CDPR)}$	Summer carbon dioxide production due to microbial respiration (measured)
$\Delta POC_{(DL)}$	POC advection calculated from the flux across the Ross Sea continental margin (Tab. 2)
$\Delta PON_{(DL)}$	PON advection calculated from the flux across the Ross Sea continental margin (Tab. 2)
$\Delta PON_{(drawdown)}$ or $\Delta PON_{(export)}$	Export of PON from the upper layer either from DIN drawdown or from sediment traps
$\Delta PON_{(burial)}$	Burial of particulate nitrogen calculated from sediment cores (observed)
$\Delta DIN_{(OUR)} = 0.09 * \Delta O2_{(OUR)}$	Summer DIN surplus due to community respiration
$\Delta(DOC + C_{(eHTL)}) = \Delta POC_{(drawdown)} + \Delta POC_{(DL)} - \Delta POC_{(burial)} - \Delta DIC_{(CDPR)}$	Pool of Carbon converted in DOC or exported to the higher trophic levels
$\Delta(DON + N_{(eHTL)}) = \Delta PON_{(drawdown)} + \Delta PON_{(DL)} - \Delta PON_{(burial)} - \Delta DIN_{(OUR)}$	Pool of Nitrogen converted in DON or exported to the higher trophic levels

Table 2 - *Water mass exchange and carbon and nitrogen budgets between the Ross Sea and the Southern Ocean for the upper layer (depth <100 m) and for the deep layer (depth >100 m). The water mass exchange is reported in Sverdrup (Sv), budgets in Gmol both daily and for the entire period (120 days). Positive values are referred to incoming fluxes from the open ocean and negative to outgoing fluxes from Ross Sea.*

Mean Fluxes (Austral Summer) Import(+) / Export(-)	Water Mass Exchange Sv	ΔO2 Gmol	ΔDIC Gmol	ΔDIN Gmol	ΔDOC Gmol	ΔPOC Gmol	ΔPON Gmol
UPPER LAYER (depth < 100 m)							
AASW							
outcoming from the Ross Sea (AASW)	-2.72	-75.0	-527.8	-6.43	-11.0	-0.160	-0.018
incoming from the "open sea"	2.72	73.4	524.3	6.86	9.9	0.129	0.013
Daily - Upper Layer total budget		-1.6	-3.5	0.42	-1.2	-0.031	-0.005
Summer (Nov- Feb) - Upper Layer total budget		**-188**	**-420**	**50.7**	**-141**	**-3.73**	**-0.56**
DEEP LAYER (depth > 100 m)							
MCDW							
Drygalski and Joides Basins (T>-0.6 °C)	1.14	22.0	225.8	3.20	4.1	0.217	0.017
Challanger Basin (T>-1.3 °C)	0.34	7.4	66.6	0.93	1.2	0.072	0.011
HSSW							
Cape Adare	-0.65	-16.6	-129.1	-1.70	-2.4	-0.149	-0.015
Joides Basin	-0.17	-4.4	-34.3	-0.45	-0.6	-0.040	-0.004
Challanger Basin	-0.07	-1.7	-13.2	-0.18	-0.2	-0.021	-0.002
DISW							
Challanger Basin (T<-1.95 °C)	-0.62	-15.7	-123.8	-1.65	-2.3	-0.134	-0.010
Daily - Deep Layer total budget		-8.9	-7.8	0.16	-0.1	-0.055	-0.003
Summer (Nov- Feb) - Deep Layer total budget		**-1068**	**-937**	**18.6**	**-13**	**-6.58**	**-0.36**
ROSS SEA							
Daily - total budget		-10.5	-11.3	0.58	-1.3	-0.086	-0.008
Summer (Nov- Feb) - total budget		**-1256**	**-1357**	**69.4**	**-154**	**-10.31**	**-0.92**

Table 3 - *Summary of biogeochemical fluxes of carbon and nitrogen on the Ross Sea continental shelf between the upper layer (depth <100 m) and the deep layer (depth >100 m) and between the deep layer and bottom*

CARBON

Upper layer (0 - 100 m)	Gmol
Phytoplankton gross primary production (GPP)	4068
Carbon dioxide production due to community respiration (CDPR)	582
Total organic carbon surplus due to net community production (NCP)	3486
Particulate organic carbon lost across the continental edge (POC ul)	4
Particulate organic carbon surplus due to net community production (POC)	3239
Dissolved organic carbon surplus due to net community production (DOC)	243
Particulate organic carbon export to deep layer (according to DIN drawdown), max	1681
Particulate organic carbon export to deep layer (according to DIN drawdown), min	1285
Range of carbon export toward higher trophic levels (eHTL) according DIN drawdown	1558 - 1954
Particulate organic carbon export to deep layer according to sediment traps	99
Carbon export toward higher trophic levels (eHTL) according to sediment traps	3140
Deep layer (100 m - bottom)	**Gmol**
Carbon dioxide production due to community respiration (CDPR)	63
Particulate organic carbon lost across the continental edge (POC dl)	7
Range of DOC and carbon exported to higher trophic levels (eHTL+DOC) by DIN drawdown	1201 - 1597
DOC surplus and carbon export to high trophic levels (eHTL+DOC) by sediment traps	15
Burial into sediment	**Gmol**
POC burial	14

NITROGEN

Upper layer (0 - 100 m)	Gmol
Particulate organic nitrogen surplus due to net community production (PON)	483
Particulate organic carbon lost across the continental edge (POC ul)	1
Dissolved organic nitrogen surplus due to net community production (DON)	39
Total organic nitrogen surplus due to net community production (TOC)	523
DIN drawdown, calculated from winter-summer values, or DIN export to deep layer	239
Nitrogen export toward higher trophic levels (eHTL) according to DIN drawdown	244
Particulate organic nitrogen export (according to sediment traps)	12
Nitrogen export toward higher trophic levels (eHTL) according to sediment traps	471
Deep layer (100 m - bottom)	**Gmol**
Dissolved inorganic nitrogen surplus due to the community respiration	8
Particulate organic carbon lost across the continental edge (POC dl)	<1
DON plus nitrogen export to high trophic levels (eHTL+DON) according to DIN drawdown	229
DON plus nitrogen export to high trophic levels (eHTL+DON) according to sediment traps	2
Burial into sediment	**Gmol**
PON burial	1

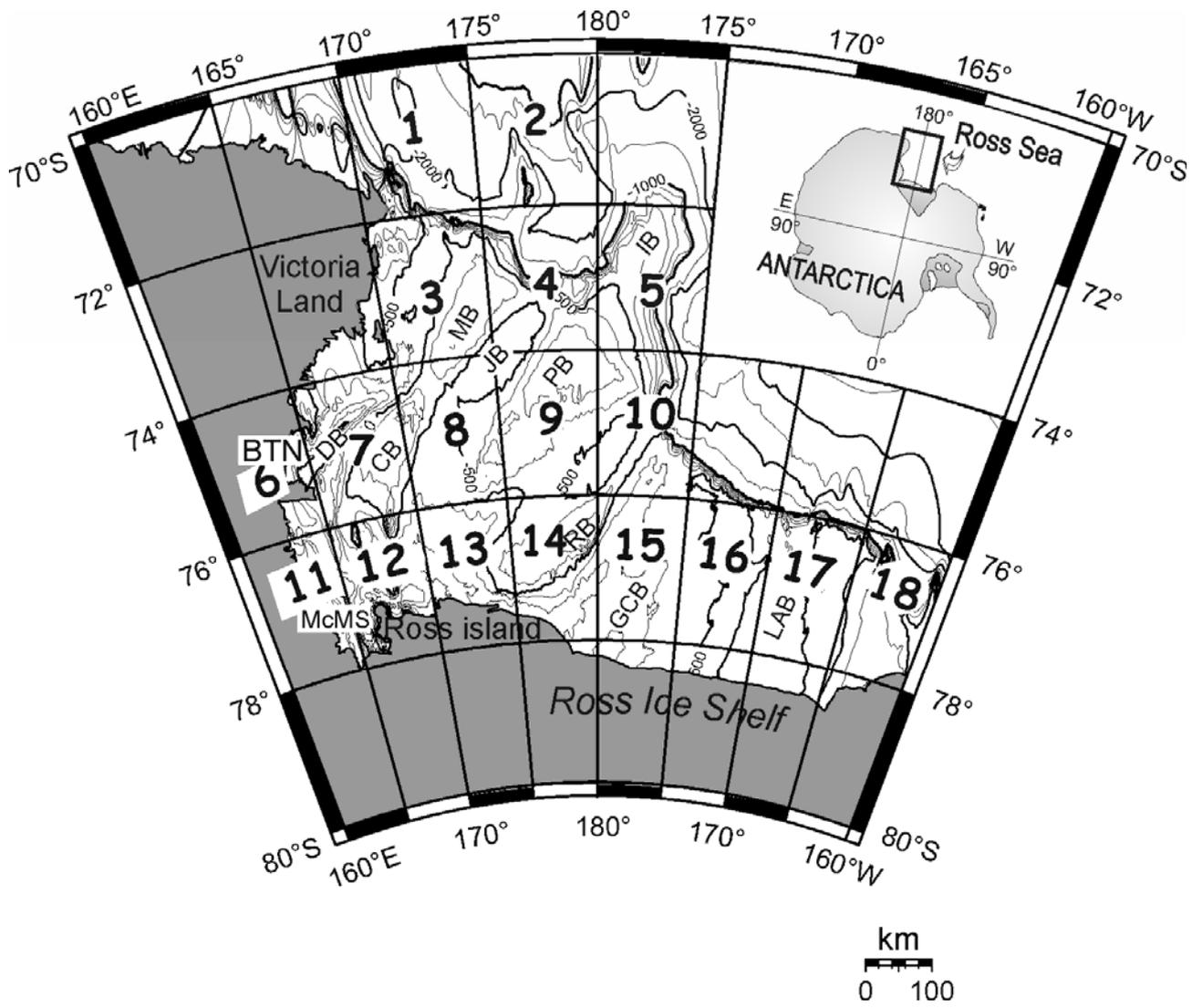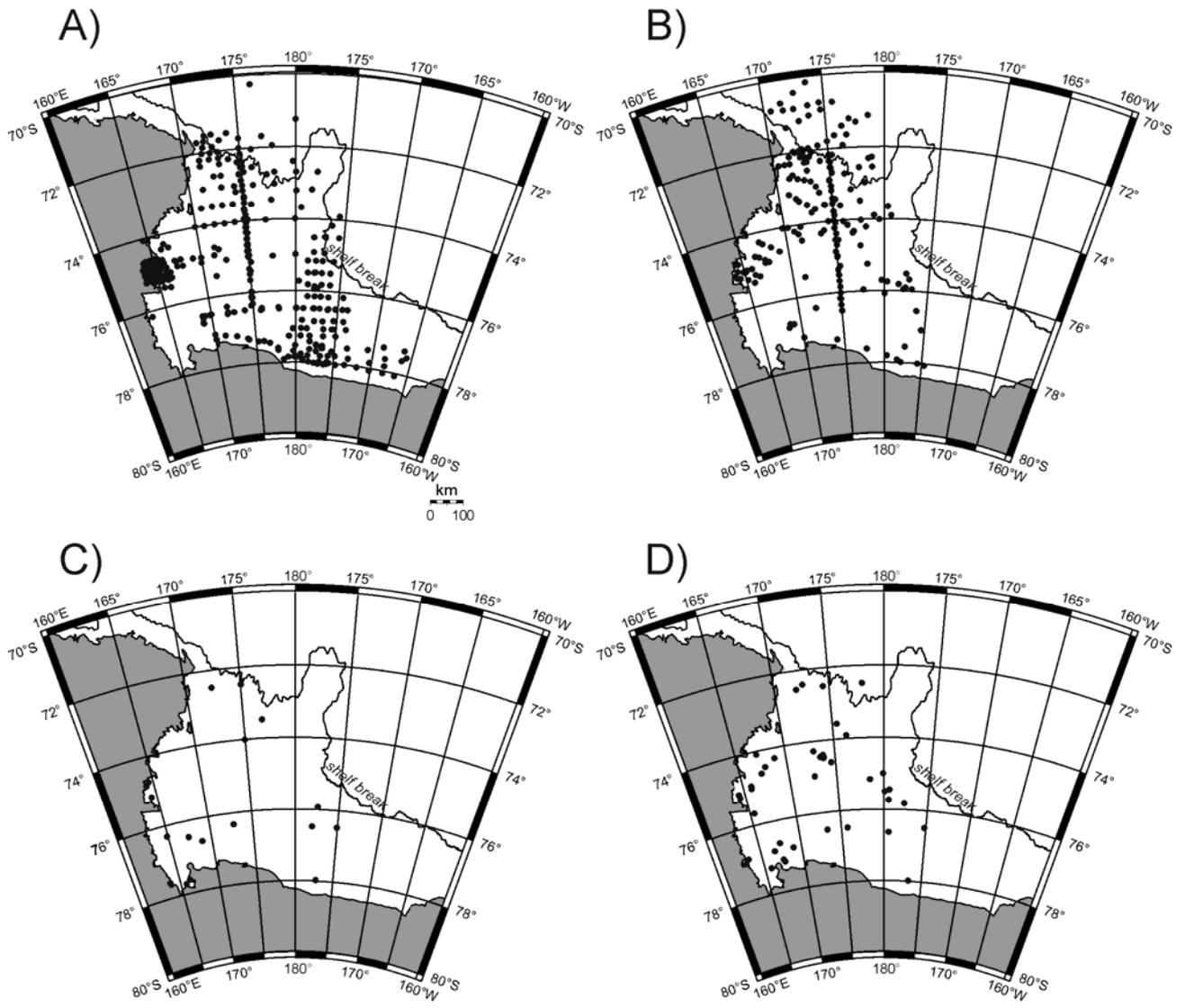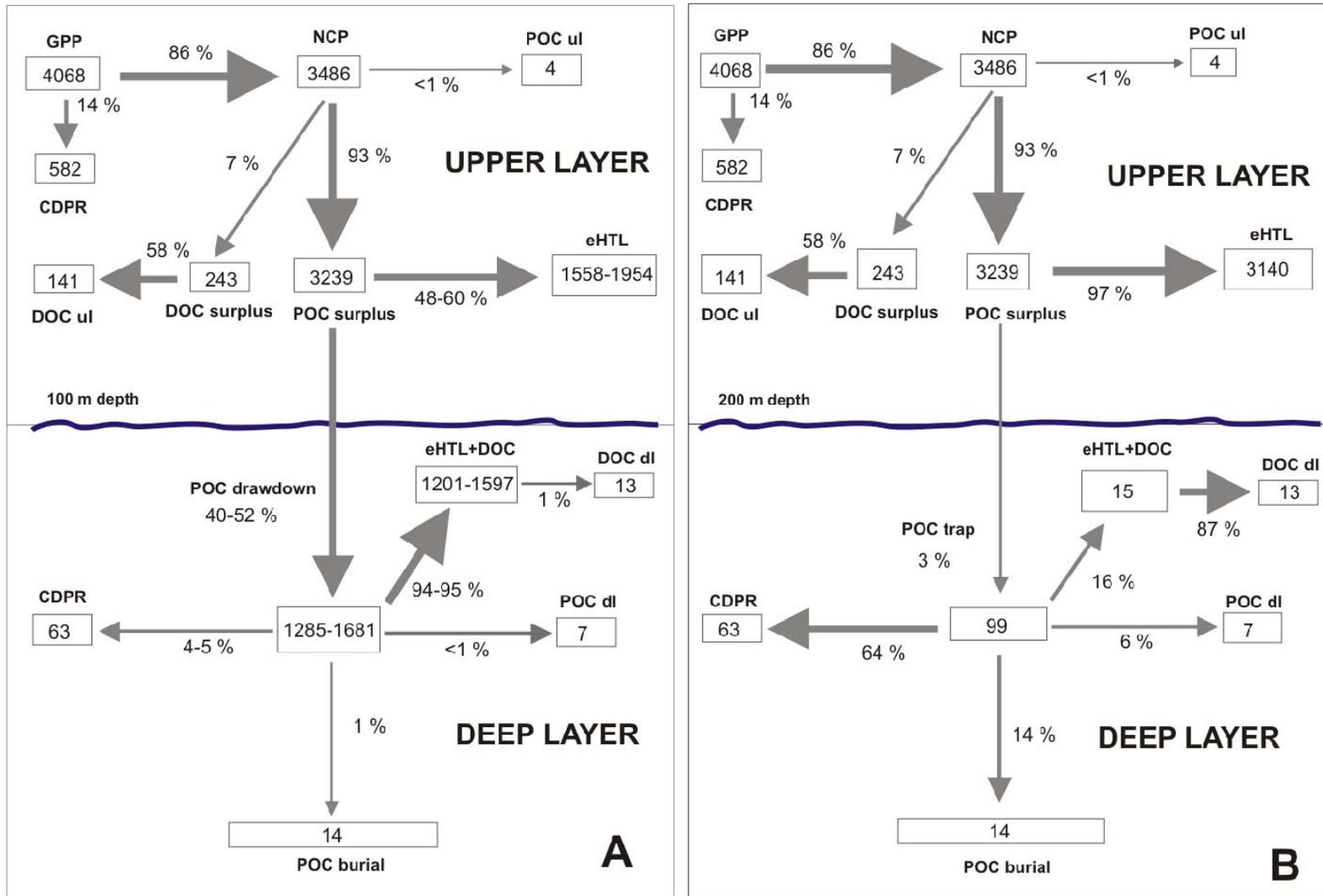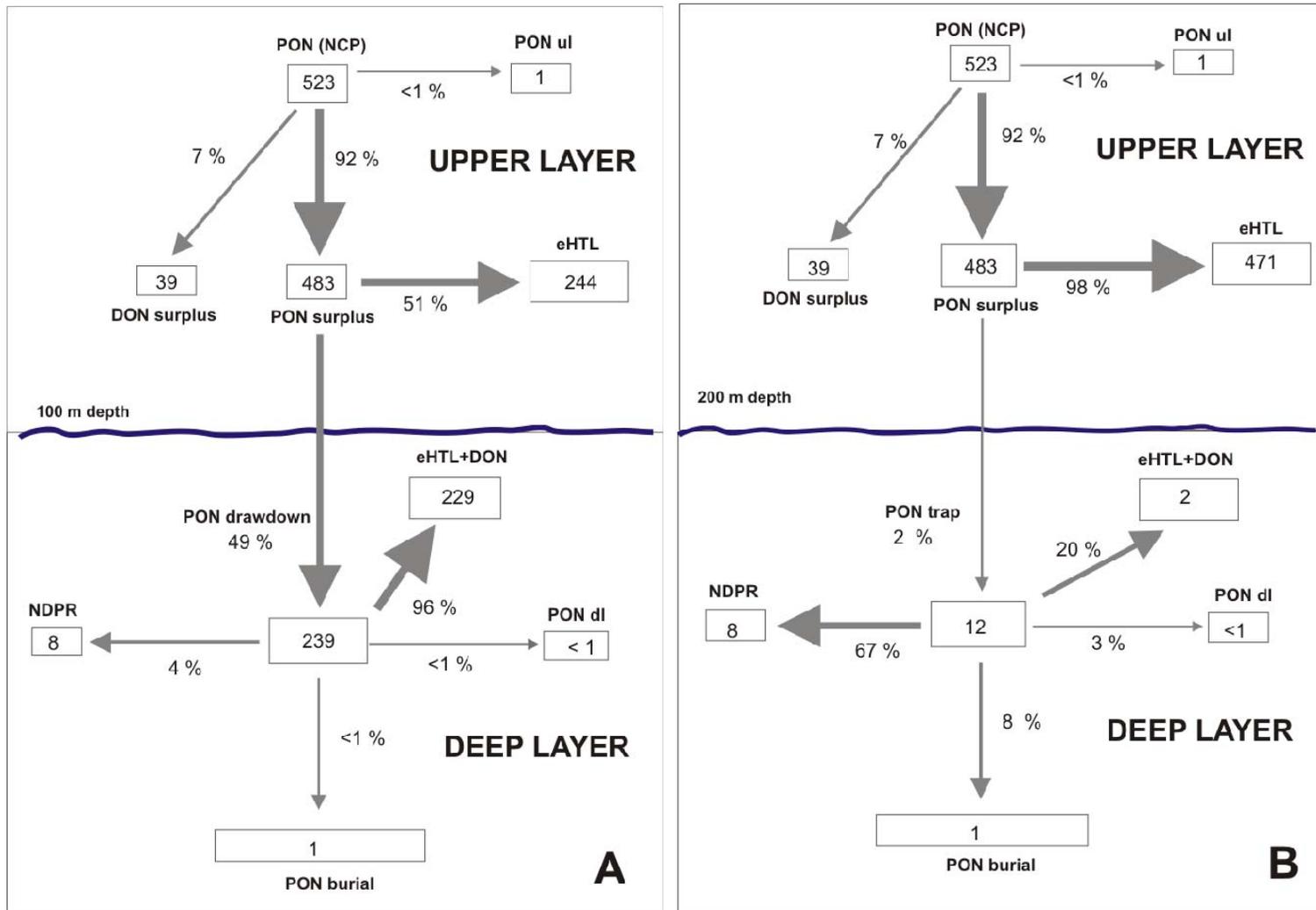Fig 1

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Fig. 2

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Fig. 3 A,B

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Fig. 4 A,B

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