Journal of MARINE RESEARCH

Volume 60, Number 2

Physical and biogeochemical fluxes and net budgets in the subpolar and temperate North Atlantic

by Marta Álvarez¹, Harry L. Bryden², Fiz F. Pérez¹, Aida F. Ríos¹
and Gabriel Rosón³

ABSTRACT

A transoceanic hydrographic section across the North Atlantic Subpolar gyre from Vigo (northwestern Iberian Peninsula) to Cape Farewell (south of Greenland) was sampled in summer 1997 as part of the World Ocean Circulation Experiment program (WOCE A25, 4x cruise). The circulation pattern across the 4x section is diagnosed using inverse methods. The flow is constrained with measured mass transports at specific sites, while conserving mass and salt for the region north of the section and forcing the silicate flux to a reasonable value. The fluxes of physical (heat and freshwater) and chemical (nutrients and oxygen) properties are estimated and decomposed into barotropic, baroclinic and horizontal components. The heat transport amounts to 0.65 ± 0.1 PW poleward, with 54% and 45% of the flux due to the baroclinic (or overturning) and horizontal circulation, respectively. From the salt conservation, an equatorward freshwater transport of -0.4 ± 1.5 Sv is estimated, resulting from net precipitation plus runoff over the North Atlantic Ocean north of the section. The Subpolar gyre exports nutrients and oxygen southward toward the Subtropical ocean at rates of -50 ± 19 , -6 ± 2 , -26 ± 15 and -1992 ± 440 kmol s⁻¹ for nitrate, phosphate, silicate and oxygen, respectively. The main mechanism responsible for the nutrient transport is the overturning cell, whereas oxygen is mainly transported southward due to the large-scale horizontal circulation. Combining our fluxes with those from the 36N section (Rintoul and Wunsch, 1991) allows us to examine budgets of physical (heat and freshwater) and chemical (nitrogen and oxygen) properties for an enclosed area of the Subpolar and Temperate North Atlantic. The tentative nitrogen budget for the box between the 4x and the 36N sections suggests that the Temperate North Atlantic is exporting

1. Instituto de Investigaciones Marinas (CSIC), Eduardo Cabello No. 6, 36208 Vigo, Spain. **email:** malvarez@iim.csic.es

2. Southampton Oceanography Centre, Empress Dock, Southampton SO14 3ZH, United Kingdom.

3. Facultad de Ciencias, Universidade de Vigo, Campus Lagoas-Marcosende, 36200 Vigo, Spain.

organic nitrogen toward the Subtropical and the Subpolar provinces, which is consistent with indirect evidence.

1. Introduction

The importance of the North Atlantic Ocean in the context of the world ocean thermohaline circulation and in global climate change is well known (e.g., Broecker and Denton, 1989; Dickson and Brown, 1994; Schmitz, 1996). The northern North Atlantic is a key water mass formation region: the Denmark Strait Overflow Water (DSOW) and Iceland-Scotland Overflow Water (ISOW), and Labrador Sea Water (LSW) are formed by cooling, sinking and mixing and ultimately flow southward through the North Atlantic in the Deep Western Boundary Current (DWBC) as North Atlantic Deep Water. Hence, this lower limb of the “Conveyor Belt” constitutes a major pathway for the meridional transport at depth of heat, freshwater and chemical constituents, such as nutrients or oxygen. The compensating shallow northward transport is carried by the North Atlantic Current (NAC), an extension of the Gulf Stream north of Cape Hatteras.

The present work is based on observations along the southern boundary of the North Atlantic Subpolar Gyre made as part of the World Ocean Circulation Experiment (WOCE). The main goal of WOCE is to determine the large-scale oceanic circulation as a basis for developing and testing ocean circulation models and coupled climate-change models. Therefore, reliable estimates of heat and freshwater fluxes in the northern Atlantic Ocean are of particular importance. In addition, fluxes of nutrients and oxygen in this area provide valuable information about the processes controlling the biogeochemical cycles in the North Atlantic.

Here we present results on heat, freshwater, nutrient and oxygen fluxes across the southern boundary of the Subpolar North Atlantic gyre in a section from Vigo (northwestern Iberian Peninsula) to Cape Farewell (southern tip of Greenland) (Fig. 1). The cruise, called 4x because it repeated section 4 sampled during the IGY (Dietrich, 1969), was carried out in August 1997 on board the RRS *Discovery*. Data collection and processing procedures are given by Bacon (1998).

A major difficulty in obtaining reliable fluxes of any property is the determination of the velocity field. In this work we adjust an “initial guess” for the geostrophic velocity field with measurements of mass transports in the area and then we develop an inverse model using additional constraints in order to obtain our “best estimate” of the circulation. The main purposes of the present work are to examine the physical and chemical fluxes across the southern boundary of the Subpolar North Atlantic gyre, to compare them with the magnitude of their corresponding sources, to identify the mechanisms responsible for the transport of each property and to assess the overall budgets in the North Atlantic by combining our results with those from other sections in the area.

In this regard, several studies have addressed the magnitude of heat, freshwater and chemical properties in the North Atlantic, with special emphasis on the Subtropical gyre (Brewer *et al.*, 1989; Hall and Bryden, 1982; Lavín *et al.*, 1998; Martel and Wunsch, 1993;

Figure 1. Station locations from the 4x cruise with the 36N (solid line) and CONVEX (dashed line) sections from Rintoul and Wunsch (1991) and Bacon (1997), respectively. IAP stands for Iberian Abyssal Plain, CGFZ for Charlie-Gibbs Fracture Zone and EGC for East Greenland Current.

Rintoul and Wunsch, 1991; Roemmich and Wunsch, 1985) but few studies have been conducted in the northern North Atlantic (Bacon, 1997; Stoll *et al.*, 1996).

2. Components of the fluxes

The net transport of any property across the 4x section can be computed as:

$$T_{Prop} = \int_{Vigo}^{Farewell} \int_{-H}^0 v \rho_{S,T,P} Prop dx dz \tag{1}$$

where T_{Prop} is the property transport, calculated by integrating the product of the property concentration (Prop), the velocity orthogonal to the section (v) and the *in-situ* density ($\rho_{S,T,P}$) from Vigo to Cape Farewell over the entire water column (that is, from the surface, 0, to the bottom depth, $-H$). The triangular area remaining below the deepest common level at each pair of stations is treated separately: the transport in the bottom triangles is calculated by multiplying the velocity at the deepest common level by the bottom triangle area and a weighted average of each property in the bottom triangle.

The velocity field is assumed to be geostrophically balanced except for the wind-driven Ekman layer. For calculating the geostrophic velocity with the thermal wind equation, CTD data (recorded every 2 db for temperature, salinity and pressure) are smoothed to 20 db intervals. Chemical data obtained at bottle depths are linearly interpolated to 20 db

intervals so as to match the physical fields. We make the assumption that the property distributions below the upper 100–200 db are uninfluenced by the seasonal cycle and represent long-term distributions in order to obtain a climatological estimate of the mass, heat, freshwater and chemical transports from a single section. The upper Ekman layer will be treated separately, as explained below.

In order to understand the processes leading the transports, we have separated the geostrophic fluxes into components as introduced by Roemmich and Wunsch (1985), Bryden *et al.* (1991), Saunders and Thompson (1993) and Bryden (1993). The geostrophic transport of any property can be broken up into 3 parts, a barotropic term due to the net transport across the section; a baroclinic term due to the horizontally averaged vertical structure; and finally, a horizontal term due to the residual flow after the barotropic and baroclinic components have been subtracted, which is associated with the horizontal variations about the baroclinic profile. The barotropic transport is associated with the net mass transport through the section, whereas the baroclinic and horizontal have no net mass flow. The baroclinic and horizontal transports are, respectively, associated with the meridional overturning circulation and the large-scale gyre circulation including smaller scale eddies. To quantify these components, the orthogonal velocity, v , and each property, Prop, are separated into a section-averaged value ($\langle \bar{v} \rangle$ and $\langle \text{Prop} \rangle$, respectively), a baroclinic profile of zonally averaged values at each depth ($\langle v \rangle(z)$ and $\langle \text{Prop} \rangle(z)$), and the deviations from zonal averages (anomalies) for each pair of stations and depth ($v'(x, z)$ and $\text{Prop}'(x, z)$). Hence,

$$v = \langle \bar{v} \rangle + \langle v \rangle(z) + v'(x, z) \quad \text{Prop} = \langle \text{Prop} \rangle + \langle \text{Prop} \rangle(z) + \text{Prop}'(x, z). \quad (2)$$

The corresponding transports are calculated as:

1. Barotropic component:

$$\rho_{S,T,P} \langle \bar{v} \rangle \langle \text{Prop} \rangle \int L(z) dz \quad (3a)$$

2. Baroclinic component:

$$\int \rho_{S,T,P} \langle v \rangle(z) \langle \text{Prop} \rangle(z) L(z) dz \quad (3b)$$

3. Horizontal component:

$$\int \rho_{S,T,P} v'(x, z) \text{Prop}'(x, z) dz dx \quad (3c)$$

where $L(z)$ is the width of the section at each depth and $\int L(z) dz$ is the area of the section. Mass transport is expressed in Sverdrups ($10^6 \text{ m}^3 \text{ s}^{-1}$), so density is not introduced in its calculations. The heat transport is given in PetaWatts ($\text{PW} = 10^{15} \text{ Watts}$), so the specific heat capacity at constant pressure (C_p) is used in the temperature version of Eq. 3. The salt

fluxes are given in Mkg s^{-1} (10^6 kg s^{-1}) and finally, the transports of the chemical properties, nutrients and oxygen are given in kmol s^{-1} (10^3 mol s^{-1}). Throughout, we use the convention that positive fluxes refer to northward transports orthogonal to the section.

The Ekman transport of any property is calculated as:

$$T_{\text{Prop}}^{\text{Ek}} = \int \frac{\tau}{f} \rho_{S,T,p} \overline{\text{Prop}_{\text{Ek}}} dx \quad (4)$$

where τ is the cross-section wind stress component, f the Coriolis parameter and Prop_{Ek} is the mean value of each property in the Ekman layer, taken as 75 db. In the case of mass transport no density is included; for the heat transport C_p is included in the integral. In order to obtain a climatological mean of the circulation, we calculate the seasonal mean Ekman transport, and the corresponding annual mean. Seasonal means for the east and northward wind components were obtained from the Southampton Oceanography Centre (SOC) wind climatology (Josey *et al.*, 2002). As our cruise was performed in late summer in order to avoid any bias in the Ekman transport, we coupled the seasonal wind stress values with the mean seasonal values of each property in the upper 75 db taken from the World Ocean Atlas 1998 data set (Ocean Climate Laboratory, NOAA) to estimate seasonal values of property transports and then averaged the seasonal transports to obtain the annual average Ekman property transports.

3. Determination of the velocity field

a. Approximation

In this section we will describe the process for deriving the closest-to-reality estimate of the velocity field. This becomes our “initial guess” to be introduced in an inverse model with additional constraints to obtain the “best estimate” of the circulation. Because circulation schemes based on minimal adjustment velocities resulting from under-determined least-squares solutions are dependent on the initial choice of reference levels (Rintoul and Wunsch, 1991) careful attention must be paid to define our “initial guess.” Table 1 summarizes the evolution of the velocity field toward the “initial guess,” giving an overview of the adjustments or modifications applied and these are explained in detail in the following paragraphs.

The first geostrophic velocity field was calculated referenced to a Level of No Motion (LNM) situated at a constant pressure level equal to 3200 db. This LNM was taken following Saunders (1982) who states that for the eastern North Atlantic a deep LNM at 3200 db is satisfactory and cannot be distinguished from a shallower reference level given the errors in computing the geostrophic transport. The net mass transport across the section calculated with this first LNM was 66 Sv (Table 1), clearly an unrealistic overestimation. Next we proceeded to adjust the LNM everywhere except for the Iberian Abyssal Plain stations. A LNM at $\sigma_2 = 36.94 \text{ kg m}^{-3}$ (Fig. 2b) was applied in accordance with Bacon’s (1997) analysis where this LNM minimized the volume flux divergence in a box defined in

Table 1. Evolution of the velocity field toward the “initial guess” circulation. The evolution in five steps is described as changes to the Level of No Motion (LNM), barotropic adjustments for the whole section or for selective areas (Iberian Abyssal Plain (IAP), Charlie Gibbs Fracture Zone (CGFZ), and East Greenland Current (EGC)). The corresponding mass and salt transports are also shown, in Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$) and Mkg s^{-1} (10^6 kg s^{-1}).

LNM	Barotropic adjustment cm s^{-1}	Mass Tr. Sv	Salt Tr. Mkg s^{-1}
Press = 3200 db	0	66	2401
$\sigma_2 = 36.94$ except IAP	0	15	562
$\sigma_2 = 36.94$ except IAP	-0.13	0	11
	IAP $\mapsto 0.09$		
$\sigma_2 = 36.94$ except IAP	CGFZ $\mapsto 2.26$	2.13	91
	EGC $\mapsto -4.81$		
$\sigma_2 = 36.94$ except IAP	-0.02	-0.36	0

the northern North Atlantic. Then, the resulting net mass transport of 15 Sv was uniformly distributed across the section to make the net mass flux be zero. This uniform barotropic velocity added everywhere was -0.13 cm s^{-1} (Table 1).

Based on direct observations and reliable estimates of mass transport in specific areas crossed by the section, we adjusted the mass flux at three locations: the Iberian Abyssal Plain (IAP), the Charlie-Gibbs Fracture Zone (CGFZ) and the western boundary region within 110 km of Greenland in the East Greenland Current (EGC) (Fig. 1). Specifically, because the eastern North Atlantic constitutes a cul-de-sac for waters below 2000 m, we adjusted the velocity in the stations located over the IAP to have no mass transport below the $\sigma_2 = 37 \text{ kg m}^{-3}$ surface (Fig. 2b), which effectively coincides with the 2000 db pressure level. The second adjustment is in the western boundary region in the EGC, where we applied a barotropic correction over the whole water column within 110 km of Greenland to obtain a total EGC flow of -25 Sv , as proposed by Bacon (1997) from a combination of ADCP measurements and historical hydrographic data in the area. The third adjustment was made over the CGFZ to reproduce the deep transport value given by Saunders (1994) of -2.4 Sv below $\sigma_\theta = 27.8 \text{ kg m}^{-3}$ (Fig. 2b). The net volume and salt fluxes across the 4x section after these considerations are 2.13 Sv and 91 Mkg s^{-1} (Table 1).

Instead of the requirement for zero mass transport across the section, we consider the northern North Atlantic to be a closed basin, and hence, a better constraint should be the conservation of salt, which is not affected by the transport into or out of the region by Runoff, Precipitation and Evaporation ($R + P - E$). There are three areas where oceanic water can leave or enter the Arctic: through the Bering Strait (Coachman and Aagaard, 1989; Roach *et al.*, 1995), through the Canadian Archipelago (Fissel *et al.*, 1988) and as ice or water between Greenland and Europe. The larger terms are the Bering Strait flow into the Arctic (about 1 Sv) and the Canadian Archipelago Throughflow (about 1–2 Sv from the Arctic). In this study we assume the net flow between Greenland and Europe must be small

Figure 2. (a) Barotropic corrections (cm s^{-1}) obtained from the inverse model, with the selected solution as a solid line. (b) Vertical distribution of the geostrophic velocity field in cm s^{-1} across the 4x section. Note in Figure 2b the expanded view of the eastern boundary. Shaded areas indicate south or southwestward velocities. Geographic areas are highlighted: MAR stands for Mid-Atlantic Ridge and CGFZ for Charlie-Gibbs Fracture Zone. Note the different scale for the upper 1000 db. In both Figure 2a and 2b the upper axis shows some station positions, indicated in Figure 1.

($0 \pm 1 \text{ Sv}$), as in Bacon (1997), Fu (1981), Lavín *et al.* (1998), Martel and Wunsch (1993), Rintoul (1988) and Wunsch *et al.* (1983). With this assumption the previous salt transport (91 Mkg s^{-1}) can be compensated by a small uniform barotropic velocity over the whole section of -0.02 cm s^{-1} (Table 1).

Table 2. “Initial guess” transports across the 4x section. Units are Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$), PW (10^{15} W), Mkg s^{-1} (10^6 kg s^{-1}) and kmol s^{-1} (10^3 mol s^{-1}).

	Barotropic	Bottom Tr.	Ekman	Baroclinic	Horizontal	Final
Mass—Sv	5.06	−4.04	−1.38	0.00	0.00	−0.36
Heat—PW	0.11	−0.04	−0.06	0.41	0.27	0.69
Salt— Mkg s^{-1}	184.1	−147.1	−49.9	6.4	6.4	0.0
NO_3 — kmol s^{-1}	87	−67	−8	−78	4	−63
PO_4 — kmol s^{-1}	6	−5	−1	−6	−1	−7
SiO_4 — kmol s^{-1}	83	−55	−7	−179	53	−105
O_2 — kmol s^{-1}	1323	−1197	−384	−468	−1104	−1829

The velocity field obtained after this set of adjustments is considered as the “initial guess” and the corresponding transports are shown in Table 2. The salt transport is negligible so the mass transport (−0.36 Sv) denotes the $R + P - E$ balance, pointing to an excess of precipitation plus runoff over evaporation in the northern North Atlantic.

Over this “initial guess” circulation minor reference level velocity corrections will be applied to adjust the circulation and to satisfy a suite of constraints in an inverse model. The inverse method (Wunsch, 1978, 1996) determines a set of reference level velocities that satisfy given constraints on the flow; here the unknown velocities are found solving a system of linear equations using the Singular Value Decomposition (SVD).

Tréguer *et al.* (1995) estimated the riverine input of silicate into the world ocean to be $177 \pm 58 \text{ kmol s}^{-1}$. Considering, a total river flux into the world ocean of $39.7 \cdot 10^{18} \text{ cm}^3 \text{ y}^{-1}$ and $5.8 \cdot 10^{18} \text{ cm}^3 \text{ y}^{-1}$ into the Arctic and northern North Atlantic (Baumgartner and Reichel, 1975), the corresponding input of silicate into these basins is about 26 kmol s^{-1} . Thus, the southward silicate flux of -105 kmol s^{-1} obtained from the “initial guess” pattern of circulation (Table 2) seems significantly high, and we use the inverse method to constrain the silicate flux to be -26 kmol s^{-1} while conserving mass and salt (i.e., three constraints in total).

We are aware that silicate is not strictly conservative, since it is an important chemical requirement for certain biota such as diatoms and radiolaria. However, most of the biogenic silica is dissolved in the water column, because the ocean is highly undersaturated in silicate, and only about 3% of the biologically produced silica is buried in the sediments (Nelson *et al.*, 1995; Tréguer *et al.*, 1995). In this connection, treating silicate as conservative is congruous with the present understanding of biogenic processes.

In solving the system of linear equations with SVD, the three equations are scaled by their norm, or row scaled, i.e. the three constraints are equally important, so no additional weighting is introduced. The relative importance of the unknowns with respect to each other, or column scaling, can be treated in different forms as described for example in Wunsch (1996) and Holfort and Siedler (2001). The method we use is to treat the barotropic transports as unknowns and scale them by the square root of the total area

Table 3. “Best estimate” transports and corresponding errors across the 4x section. Units are Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$), PW (10^{15} W), Mkg s^{-1} (10^6 kg s^{-1}) and kmol s^{-1} (10^3 mol s^{-1}).

	Barotropic	Bottom Tr.	Ekman	Baroclinic	Horizontal	Final	Error
Mass—Sv	5.26	−4.28	−1.38	0.00	0.00	−0.40	1.48
Heat—PW	0.11	−0.04	−0.06	0.37	0.31	0.69	0.11
Salt— Mkg s^{-1}	191.3	−155.8	−49.9	5.8	7.8	−0.8	6.7
NO_3 — kmol s^{-1}	90	−71	−8	−66	4	−50	19
PO_4 — kmol s^{-1}	6	−5	−1	−5	−1	−6	2
SiO_4 — kmol s^{-1}	86	−57	−7	−112	62	−26	15
O_2 — kmol s^{-1}	1375	−1268	−384	−480	−1235	−1992	440

between station pairs, which is equivalent to weighting each station pair equally regardless of its area.

Finally, the constraints are not solved exactly because that can lead to an unrealistic circulation pattern. It must be taken into account that the constraints are approximations with a certain degree of uncertainty, both in the direct measurements and in the assumptions. In contrast, we search for the simplest consistent solution, minimizing the variance in the reference level velocities from their initial values. The degree of stringency in imposing the constraints is given by the rank of the solution matrix, i.e., the rank of the matrix with the new barotropic velocities. A high rank means that the system is exactly solved, but this can magnify the errors and produce unrealistically high and variable velocity adjustments. In contrast, with a low matrix rank the system is solved with a high degree of uncertainty and the solutions can be considered inaccurate. Therefore, we search for an optimal rank that satisfies the constraints rather precisely and introduces a relatively low noise in the final barotropic velocities.

In this case, we have selected the solution which takes into account the first two singular values or eigenvalues of the SVD (two out of three), which satisfy the constraints within 99.9% and introduce less noise than the exact solution, especially at the western and eastern limits of the section (Fig. 2a). The final barotropic adjustments have a mean value of -0.05 cm s^{-1} with a standard deviation of 0.13 cm s^{-1} . In summary, in order to decrease the initial flux of silicate (-105 kmol s^{-1} , Table 2) to -26 kmol s^{-1} the reference level velocity corrections increase the northward flux of high silicate water in the IAP area, where the highest silicate concentrations are reported (Álvarez *et al.*, 2002), and slightly decrease the northward transport across the rest of the section. The final velocities (Fig. 2b) and transports (Table 3) will be discussed in the following sections. An assessment of the sources of uncertainty of our calculations is detailed in the Appendix. The maximum error estimated for each property flux is also given in Table 3.

b. Flow field across the 4x section

In this section we focus upon the distribution of the best-estimate geostrophic velocity and the mechanisms responsible for the meridional circulation across the 4x section. The

absolute velocity field (Fig. 2b) exhibits a columnar nature, with alternating flows intensified in the upper 1000 db reflecting mesoscale eddies. Similar results are commonly obtained from inverse computations (e.g., Fú, 1981; Rintoul and Wunsch, 1991; Wunsch *et al.*, 1983).

Although not clearly differentiated in Figure 2b, a strong current carries water southward between 400–700 db along the Iberian coast with a speed between 30 and 40 cm s^{-1} . This flux can be ascribed to the variable and seasonally dependent Portugal Current (Arhan *et al.*, 1994; Krauss and Käse, 1984; Richardson, 1983). At about 70 km offshore appears a remarkable poleward slope current about 1000 db deep, the Portugal Coastal Countercurrent, with maximum values around 12 cm s^{-1} . This current has been described in spring as well as in winter by several authors (Ambar *et al.*, 1986; Frouin *et al.*, 1990; Haynes and Barton, 1990) and mainly carries Mediterranean Water northwards (Daniault *et al.*, 1994; Mazé *et al.*, 1997; Zenk and Armi, 1990). Northward velocities predominate in the rest of the Iberian Basin except for two mesoscale eddies about 500 db deep and 300 km wide located around stations 28 and 34 (Fig. 2b). These eddies have warmer temperatures in their cores than in the surroundings, revealing their anticyclonic rotation. Similar mesoscale features have been described by Arhan *et al.* (1994) in the Iberian Basin and called “winter intensified anticyclones,” which are likely formed from instabilities of the NAC farther north (Krauss, 1986; Krauss and Käse, 1984; Sy *et al.*, 1992). Below 2000 db in the IAP a cyclonic circulation is suggested in this basin as also observed by long-term current meter measurements (Dickson *et al.*, 1985) and also postulated by an inverse model for this basin (Paillet and Mercier, 1997).

The latitudinal band between 40N and 55N constitutes an area of great oceanographic interest because it is the transition between the North Atlantic Subtropical and Subpolar gyres. Observations confirm the permanent existence of a northern branch of the NAC located above the CGFZ; we estimate a maximum speed of 16 cm s^{-1} in the NAC over the CGFZ. Several current branches of less than 100 km wide can be distinguished above the western flank of the Mid-Atlantic Ridge (MAR) around stations 52 and 56, confirming the observations of Sy *et al.* (1992). On the eastern flank of the MAR, between stations 46 and 50 a strong cyclonic eddy is distinguished in the upper 800 db.

The CGFZ constitutes the passage of deep water from the northeast Atlantic into the western basin. This flow supplies a mixture of Iceland-Scotland Overflow Water and Eastern North Atlantic Deep Water (Álvarez *et al.*, 2002) ultimately to the DWBC, so it is an important branch of the southward return flow of the Atlantic meridional overturning circulation (Dickson and Brown, 1994; Schmitz and McCartney, 1993; Smethie and Swift, 1989). However, reversals of the flow, turning eastward, have been reported and related to interactions between the CGFZ throughflow and the NAC (Saunders, 1994; Schott *et al.*, 1999). As we forced the circulation to correspond to a climatological mean value, the deep flow in the CGFZ is directed southwestward with a maximum speed of -8 cm s^{-1} .

The cyclonic circulation in the upper 500 db of the Irminger Basin is resolved by the

model solution, with a northward flow above the western slope of the Reykjanes Ridge and a strong southward flow over the whole water column in the EGC region. Lying against the lower part of the Greenland slope, the bottom-trapped current associated with the DSOw flow below 2000 db has a maximum speed of -20 cm s^{-1} . This is a bit lower than the values reported by Dickson and Brown (1994) from a current meter array at the slope of Cape Farewell. In the surface layer, along the continental margin of Greenland, the EGC occupies the upper 800 db over the slope with a maximum speed of -48 cm s^{-1} .

As previously noted, the baroclinic or overturning flux is linked to the global Conveyor Belt, whereas the horizontal flux is ascribed to the large-scale gyre circulation. In this sense, the vertical profile of zonally integrated mass transport and the horizontal profile of vertically integrated mass transport across the 4x section are shown in Figure 3. Regarding the baroclinic structure (Fig. 3a), the overall pattern consists of a northward flux of 14.8 Sv in the upper layers. Below 3200 db a weak northward flow of 1.7 Sv occurs over the IAP. The upper northward flux is effectively compensated by a southward flux of -16.5 Sv between 1100 db and 3200 db, and this depth interval includes LSW and the two overflows crossing the section, DSOw and ISOW, that ultimately join together in the DWBC. Thus Figure 3a portrays the overturning circulation across the Subpolar gyre. In the same way as detailed in the Appendix, the estimated error in this overturning circulation is $-16.5 \pm 3.6 \text{ Sv}$. Higher variability was reported by Koltermann *et al.* (1999) from an analysis of three sections in the northern North Atlantic at three distinct periods of deep convection in the Labrador Sea, which they hypothesized would affect the size of the meridional overturning circulation.

Figure 3b shows the horizontally integrated mass transport accumulated from zero at the east (i.e., Vigo at the right). The northward flux in the eastern part of the section (Fig. 3b) is noisy due to eddy activity in the upper layers of the eastern North Atlantic (Paillet, 1999) (Fig. 2b). On the eastern flank of the MAR, from 1500 to 1700 km, a strong eddy is distinguished (see also Fig. 2b). The NAC signature is clearly evident in the steep transport increase to 28.4 Sv from about 1700 to 2500 km (stations 50 to 64). As described by Bacon (1997), the strength of the general circulation of the Subpolar gyre can be deduced from the southward flux between Cape Farewell and the center of the Irminger Basin, associated with the EGC and DSOw. In agreement with Bacon's analysis, we obtained a transport of -25.4 Sv southward over this region which effectively balances the northward flow in the NAC.

A northward flux of 2.2 Sv was obtained in the Iberian Basin for waters denser than $\sigma_2 = 37 \text{ kg m}^{-3}$. In the context of global circulation this flux corresponds with the 2 Sv of modified Antarctic Bottom Water (Schmitz and McCartney, 1993; Schmitz, 1996) flowing northward across the Iberian Basin to join the overflows. Regarding the deep flow in the CGFZ, a southwestward flow of -2.6 Sv for $\sigma_2 > 36.93 \text{ kg m}^{-3}$ is included in our circulation in agreement with Saunders' (1994) mean transport estimate.

Figure 3. Baroclinic (a) and horizontal (b) components of the mass transport across the 4x section. The horizontal transport is accumulated from zero at Vigo. The vertical integrated baroclinic transport down to 1000 db is 14.8 Sv, -16.5 Sv from 1100 to 3200 db and 1.7 Sv below 3200 db. ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$). The upper axis in Figure 3b shows station positions.

4. Property transports: Components and net budgets

a. Heat

A poleward heat flux of 0.69 ± 0.1 PW (see Appendix for the error assessment) was computed across the 4x section. Yet, this value is biased as the section was sampled in

summer. If we homogenize the upper 210 db with the temperature at 210 db to simulate deep winter mixed layers, the heat transport is reduced to 0.60 PW. Thus, the most likely annual mean for the heat transport across the 4x section would be 0.65 ± 0.1 PW poleward. However, in order to maintain the consistency of our exposition in this manuscript, we will concentrate on the results from the real data taken during the cruise.

As they conserve mass by definition, the baroclinic and horizontal heat fluxes contribute directly and, in this case, to almost all the northward heat transport (54 and 45%, respectively, Table 3). The barotropic component must be combined with the Ekman transport and the flux in the bottom triangles to determine the barotropic heat flux with mass conservation (0.01 PW, being only a 1% of the total, Table 3).

The baroclinic heat flux represents a northward transport of 0.37 PW (Table 3), which is mainly concentrated in the upper 1000 db of the water column (Fig. 4a), with a weak northward heat flow at intermediate levels. Thus, the overturning cell across the southern boundary of the Subpolar North Atlantic is compensating the heat loss to the atmosphere north of the section associated with water mass transformation. Figure 4b shows the accumulated horizontal heat transport from Vigo on the right, this plot is directly correlated with the accumulated horizontal mass transport (Fig. 3b) until the MAR (about 1700 km from Vigo), westwards they are inversely correlated. This is due to the fact that temperature anomalies ($\theta'(x, z)$) are mainly positive (warmer than the section mean) east of the MAR, whereas northwestward of the MAR they are negative (cooler than the section mean). These anomalies combined with the northward transport east of the MAR and southward transport west of the ridge shown in Figure 3b yield the along-section distribution of the horizontal heat fluxes. Figure 4c reveals the strong eddy activity in the eastern North Atlantic from Vigo to the Azores-Biscay Rise (about 1000 km offshore Vigo), but these eddies do not transport any heat. The eddy fields within the NAC and the EGC transport -0.14 PW and 0.35 PW, southwestward and northeastward, respectively. At first sight these results are unexpected due to the high northeastward and southwestward transport associated with the NAC and EGC, respectively. However, velocity anomalies are positive (higher than the section zonal mean) within the NAC and negative (lower than the section zonal mean) in the EGC. In the case of temperature, the NAC is associated with a thermal front separating warmer (positive temperature anomalies) southward from colder waters (negative anomalies) northward, whereas the EGC temperature anomalies are uniformly negative. The combination of the velocity and temperature anomalies renders a total eddy flux negative within the NAC and positive at the EGC. So, the horizontal circulation within the NAC contributes to cool the ocean north of the section and the EGC does the opposite, leaving warmer waters north of the section.

In the context of the North Atlantic meridional heat transports estimated by bulk formulae (Hastenrath, 1982; Isemer *et al.*, 1989; Semtner and Chervin, 1992; Trenberth and Solomon, 1994) and inverse models (Schiller, 1995; Schlitzer, 1993), our obtained value of 0.65 ± 0.1 PW poleward is within the ranges given by those authors. No direct comparison, however, is possible since our section is not zonal. According to Macdonald

Figure 4. Baroclinic (a and b) and horizontal (c and d) components of the heat and salt transports across the 4x section. The horizontal transport is accumulated from zero at Vigo. The insets in (a) and (b) show the zonally averaged profiles of temperature and salinity along with the corresponding section mean values. The baroclinic profile is the deviation from the mean value as a function of depth. The upper axis in Figure 4c and d shows station positions.

and Wunsch (1996); Rintoul and Wunsch (1991) and Sato and Rossby (2000) the heat transport across 36N is 1 ± 0.2 PW, 1.3 ± 0.2 PW or 1.4 ± 0.3 PW, respectively. Likewise, between Cape Farewell and Ireland (CONVEX section) 0.28 ± 0.06 PW (Bacon, 1997) is transported northward. In the light of these figures it seems that the heat loss to the atmosphere in the eastern North Atlantic between our section and that studied by Bacon (1997) is somewhat higher than we had expected.

A more detailed analysis can be performed. We have calculated the surface heat fluxes for the boxes bound by the 36N, the 4x and the CONVEX sections (Fig. 1), and compared

Table 4. Heat convergence, loss to the atmosphere, between sections in the North Atlantic Ocean, calculated from the SOC mean annual heat flux (Josey *et al.*, 1999) (*SOC heat flux*) and from hydrographic estimations (*Hydro. heat flux*) taken from Rintoul and Wunsch (1991), Bacon (1997) and this work, for the 36N, the CONVEX and the 4x sections (using 0.65 PW), respectively.

	Area (10 ¹² m ²)	SOC heat flux (W m ⁻²)	Hydro. heat flux (W m ⁻²)	Difference (W m ⁻²)
36N-CONVEX	9.6	-33	-106	-73
36N-4x	6.3	-24	-103	-79
4x-CONVEX	3.3	-9	-112	-103

them with the corresponding climatological mean values from the annual mean net heat flux given in the SOC climatology (Josey *et al.*, 1998). A disagreement between the climatological and hydrographic estimates (Table 4) is found. The climatological heat losses are considerably weaker than the hydrographic values. This discrepancy was also noticed by Josey *et al.* (1999) in a box comprised between 36N and the CONVEX sections and mainly ascribed to an underestimation of the heat loss over boundary currents due to a problem in the estimation of the latent heat flux. However, a greater disagreement between the two sets of calculations is localized in the eastern North Atlantic, a relevant area of winter convection where Subpolar Mode Water and Eastern North Atlantic Central Water are formed (McCartney and Talley, 1982; Pollard *et al.*, 1996).

If it is assumed that the North Atlantic is an open basin north of our section, the impact of a flux of about 0.8 Sv through the Bering Strait (Roach *et al.*, 1995) at a temperature of about 0°C (Coachman *et al.*, 1975) is negligible. Balancing this barotropic flux by an additional -0.8 Sv flowing across the section at 5.18°C would introduce an additional southwestward heat transport of only 0.02 PW. Likewise, balancing net precipitation plus runoff over the northern North Atlantic introduces a negligible southwestward heat flux of -0.008 PW.

b. Salt and freshwater

The baroclinic and horizontal salt fluxes are both northward across the 4x section and both resemble those of heat (Fig. 4). The overturning circulation transports salt toward the north in a two-lobe structure, within the upper 1200 db and between 1400 and 3300 db (Fig. 4b). The accumulated horizontal salt transport (Fig. 4d) shows that in the Iberian Abyssal Plain (up to station 34) there is low northward transport despite the strong eddy activity. The strongest horizontal fluxes are concentrated in the NAC and EGC systems, with opposite directions, -5.9 Mkg s⁻¹ and 10.4 Mkg s⁻¹, respectively.

Combining the thermohaline circulation mechanisms, the overturning circulation has a two-lobe structure with warmer and more saline (with respect to the section average) waters flowing northeastward in the upper 1200 db, mainly in the NAC system, returning southwestward colder and fresher at intermediate levels. Consequently, in both layers there is a northeastward transport of heat and salt. The vertical overturning circulation helps to

compensate for the air-sea heat loss and the net precipitation plus runoff over the northern North Atlantic. The horizontal thermohaline fluxes are associated with the cyclonic Subpolar gyre circulation: warmer, saltier waters flow northeastward in the NAC and return southwestward at colder temperature and lower salinity in the EGC. This Subpolar gyre circulation also helps to compensate for the air-sea heat exchange and net precipitation over the northern North Atlantic. The baroclinic and horizontal circulation each contribute about half of the total heat and salt transports across the 4x section.

The total barotropic salt transport (barotropic plus bottom triangle and Ekman transports) must counterbalance the northward salt transport due to the baroclinic and horizontal contributions in order to conserve salt for the region north of the 4x section. Here, the barotropic salt transport is -14.4 Mkg s^{-1} (Table 3), and almost all of this southward salt transport is due to a net southward mass transport of -0.4 Sv , which corresponds to the precipitation plus runoff north of the 4x section. Thus, in order to maintain the salt budget in the region, a southwestwards mass transport is required to compensate the northeastward baroclinic and horizontal salt flux contributions.

Assuming that the Arctic Ocean is a closed basin, salt must be conserved and thus, the salt flux across the section has been forced to be zero given the uncertainty in the flux calculations of $-0.8 \pm 6.7 \text{ Mkg s}^{-1}$ (Table 3). Therefore, the mass or volume imbalance corresponds to the net gain of freshwater at the ocean surface due to precipitation and runoff. According to our calculations, $-0.4 \pm 1.5 \text{ Sv}$ exits southwestward across our section, a value fortuitously equal to estimates of the total freshwater loss to maintain the strength of the meridional overturning circulation (Zaucker *et al.*, 1994). Of the 0.4 Sv , 0.18 Sv corresponds to the coastal runoff into the Arctic and the northwestern European coast (Baumgartner and Reichel, 1975), and about 0.22 Sv corresponds to net precipitation over the Arctic and northeastern North Atlantic confirming that precipitation dominates over evaporation at high latitudes (Beránger *et al.*, 1999; Schmitt *et al.*, 1989). The corresponding precipitation over the northeastern North Atlantic can be estimated from the SOC climatology atlas to be 0.09 Sv as an annual mean. Therefore, the net precipitation over the Arctic plus the sea-ice melting are about 0.13 Sv . The runoff into the Arctic is about 0.11 Sv (Baumgartner and Reichel, 1975). In consequence, the Arctic freshwater gain can be calculated to be 0.24 Sv . This estimation is similar to the direct estimate given by Aagaard and Carmack (1989) of 0.20 Sv , the freshwater balance performed by Bauch *et al.* (1995) of $0.21\text{--}0.26 \text{ Sv}$, or that calculated by oceanic budgeting given by Bacon (1997) of $0.17 \pm 0.06 \text{ Sv}$.

Until now, we have been considering the area north of the 4x section to be closed, implicitly assuming that the Bering Straits flow enters the North Atlantic through the Canadian Archipelago and not across the 4x section. If instead we assume that the flow through the Bering Strait carries $26.7 \pm 3.3 \text{ Mkg s}^{-1}$ of salt (Coachman and Aagaard, 1989) out of the Pacific into the Arctic and then into the northern Atlantic, to maintain salt conservation an equal amount of salt must be transported southwards across the 4x section

and through the Atlantic (Wijffels *et al.*, 1992). In our case, conserving mass across section, the volume flux (5) and the salt flux (6) can be written simply as:

$$w + T_B + T_{4x}^N = 0 \quad (5)$$

$$w0 + T_B S_B + T' S' + T_{4x}^N S = 0 \quad (6)$$

where w is the precipitation plus runoff minus evaporation budget north of the section, T_B is the transport through the Bering Strait (0.8 Sv) at a salinity (S_B) of 32.5, T_{4x}^N stands for the net or barotropic transport across the section with a mean salinity (S) of 35.10. $T' S'$ corresponds to the baroclinic and horizontal salinity transport for a zero net volume flux, equal to 13.12 Sv psu. From these equations we obtain a net freshwater input, w , of 0.32 Sv, similar to our previous estimate given the uncertainties in the calculations. Thus, the barotropic transport of salt ($T_{4x}^N S$) is -39.3 Sv psu southwestward, while the baroclinic and barotropic terms amount to 13.1 Sv psu northeastward, so the net salt flow through the 4x section amounts to -26.2 Sv psu ($\cong -27.1 \cdot 10^6 \text{ kg s}^{-1}$) southwestward, not different given the errors from the quantity required by Wijffels *et al.* (1992). Thus, including the Bering Straits flow across the 4x section adjusts the net P - E + R for the region north of the 4x section from 0.4 Sv to 0.32 Sv. The total freshwater flow across the section would amount to $-1.12 \cdot 10^9 \text{ kg s}^{-1}$ southwestward, which is the sum of the net freshwater input and the Bering Strait contribution $[(T_B + w)(1 - S/1000)]\rho$. This value is higher than that given by Wijffels *et al.* (1992), who integrated Baumgartner and Reichel's (1975) compilation of the air-sea freshwater exchange and runoff over the Atlantic from the reference transport through the Bering Strait. The disagreement is probably due to a low estimation of the evaporation minus precipitation by Baumgartner and Reichel (1975), because our value is similar to the freshwater fluxes given by da Silva *et al.* (1996) or Bernger *et al.* (1999).

c. *Nutrients and oxygen*

Only a few studies have dealt with the nutrient (nitrate, phosphate, silicate) and oxygen fluxes across oceanic sections (Brewer and Dyrssen, 1987; Ganachaud and Wunsch, 1998; Holfort and Siedler, 2001; Rintoul and Wunsch, 1991; Robbins and Bryden, 1994; Schlitzer, 1988). With the completion of the WOCE observational phase new estimates should be available soon. Based on our circulation across the 4x section, the Subpolar North Atlantic exports nutrients and oxygen southwards towards the Subtropical gyre (Table 3).

Nitrate and phosphate are exported southwestward across the 4x section at $-50 \pm 19 \text{ kmol s}^{-1}$ and $-6 \pm 2 \text{ kmol s}^{-1}$, respectively. Likewise, oxygen is transported southwestward at $-1992 \pm 440 \text{ kmol s}^{-1}$. The silicate transport ($-26 \pm 15 \text{ kmol s}^{-1}$) was imposed in order to constrain the circulation, as explained previously. The principal contributors to these transports are the baroclinic and horizontal circulations while the barotropic contributions are small (Table 3). The main mechanism contributing to the

southwestward flux of nutrients is the baroclinic or overturning circulation, whereas in the case of oxygen the principal contributor is the large-scale horizontal circulation. The total net barotropic transport (barotropic plus Ekman plus bottom triangle transports) contributes nitrate, phosphate, silicate and oxygen fluxes of only 11, 0, 22 and -277 kmol s^{-1} , respectively. The horizontal circulation drives a remarkable southwestward flux of oxygen while the horizontal component for nitrate or phosphate is practically zero and for silicate is northward helping to compensate for the high southward baroclinic silicate transport.

The baroclinic profiles of nitrate and phosphate (insets in Fig. 5a and b) are related with a slope of 14.3 ± 0.02 ($r^2 = 0.989$, $p < 0.001$), close to the $\Delta N/\Delta P$ Redfield ratio of 16 ± 1 (Anderson and Sarmiento, 1994). The upper 1000 db presents much lower concentrations of nitrate and phosphate (insets in Fig. 5a and b) than the section mean, this layer is flowing northeastwards (Fig. 3a), so the baroclinic transports are southwestwards (Fig. 5a and b). The intermediate layer indicates concentrations slightly above the section mean, which combined with the zonal transport fluxes at this layer yield a small southwestward baroclinic flux for both nutrients. At deeper levels where the concentrations are much higher than the mean, the southwestward transport is very small making limited contribution to the northeastward baroclinic flux. Consequently, the overturning circulation removes nitrate and phosphate from the region north of the 4x section mainly in the upper 1000 db, but with a small contribution centered at 3000 db, because it transports nutrient-depleted surface waters northeastwards and nutrient-rich intermediate waters southwestward, both contributing to the southwestward transports.

Regarding silicate, the baroclinic transport (Fig. 5c) also presents similar southwestward flux in the upper layer, but as well, a significant flux at intermediate levels partially compensated by the deep and bottom northeastward flow. So the overturning circulation removes silicate from the region north of our section within the upper 1000 db and in a layer centered at 3000 db.

The baroclinic structure of the oxygen transport (Fig. 5d) reveals a two-lobe distribution centered at 700 db and 2000 db at the extremes of the mean oxygen profile (inset on Fig. 5d): the upper minimum just below the euphotic zone and the deeper maximum ascribed to LSW. Thus, the upper waters flowing northeastward are oxygen poor, compared with the section mean value, whereas the intermediate waters flowing southwestward are oxygen rich. Both contribute to the export of oxygen from the Subpolar gyre.

The horizontal transports of nitrate and phosphate accumulated from the coast of Spain are similar (Fig. 5e and f) and the overall horizontal nutrient fluxes are practically zero (Table 3). The main absolute contributors to the horizontal transport of these nutrients are the NAC and EGC. The NAC is flowing northeastward (Fig. 3b) where nutrients mainly present positive anomalies (higher concentrations), balanced by the EGC, which is flowing southwestward also coupled with positive nutrients' anomalies.

Silicate is transported northeastward due to the large-scale gyre circulation (Table 3, Fig. 5g). In the eastern basin the flux is slightly southeastward but it then accumulates northeastward along the rest of the section, especially in the NAC (Fig. 5g). The highest

Figure 5. Baroclinic (a to d) and horizontal (e to h) components of the nitrate (NO_3), phosphate (PO_4), silicate (SiO_4) and oxygen (O_2) transports across the 4x section. The horizontal transport is accumulated from zero at Vigo. The insets in figures a to d show the zonally averaged profiles of the chemical properties along with the corresponding section mean values. The baroclinic profile of each property is the deviation from the mean value as a function of depth. The upper axis in Figure 5e to h shows station positions.

Figure 6. Vertical distribution of the oxygen anomalies ($O_2'(x, z)$ in $\mu\text{mol kg}^{-1}$) along the 4x section. The upper axis shows station positions.

anomalies in the silicate distribution are at about 3000 db. Thus the main horizontal cell is driven by deep high silicate waters flowing northeastward east of the MAR returning southwestward with a lower silicate concentration west of the MAR.

The vertical distribution of the oxygen anomalies ($O_2'(x, z)$) along the section is shown in Figure 6. It reveals the strong horizontal contrast in the oxygen content for waters between 500 and 1000 db. On the eastern side of the section centered at 1000 db, the low oxygen anomalies point to the influence of the oxygen-depleted Mediterranean Water (Álvarez *et al.*, 2002), whereas on the western end the influence of the high oxygen LSW is clearly discerned. The large-scale gyre circulation explains 62% of the total oxygen transport toward the Subtropical gyre, with the EGC being the main contributor (80%) to the total horizontal flow (Fig. 5h). In this area, oxygen rich waters (Fig. 6) are flowing southwestward (Fig. 2b) and, thus exporting oxygen from the Subpolar gyre.

d. Area-averaged biogeochemical balances

To put the former fluxes into context we can examine the basin-scale biogeochemical budgets for the regions north of the 4x section and for the box between the 4x and 36N sections (Fig. 1). Initially, we formulate the following nitrate budget for the Atlantic Ocean north of the 4x section assuming a steady-state:

Figure 7. (a) Nitrate (NO_3) and (b) organic nitrogen (Norg) plus ammonium (NH_4) budgets north of the 4x section and within the box comprised between the 4x and the 36N sections. All numbers are in kmol s^{-1} . Open horizontal arrows indicate river input, dark arrows indicate the fluxes across the sections, open squared arrows for the Arctic contribution, dots for the atmospheric input, open vertical arrows for sedimentation and crosses for the losses to the atmosphere. The NH_4 contributions are underlined. Processes over the shelves (sedimentation and atmospheric input) are presented as bold numbers.

$$\frac{\Delta \text{NO}_3}{\Delta t} = I - O + NP_{\text{NO}_3} = 0. \quad (7)$$

The inputs (I) minus the outputs (O) of nitrate must be balanced by the net production of nitrate due to biological activity (NP_{NO_3}). The temporal scale of this assumption can be roughly calculated as the flushing time of the ocean north of the section (total volume of the ocean north of the section divided by the outwards volume flux across the section), estimated here as about 10 years. Estimates provided by an atmospheric chemical model suggest an atmospheric deposition of nitrate in open ocean and coastal waters north of the 4x section of about 2 kmol s^{-1} and 2 kmol s^{-1} , respectively (Prospero *et al.*, 1996). According to Lipschultz and Owens (1996), nitrogen fixation north of the 4x section can be neglected. The river supply of nitrate north of the section without the Arctic contribution can be roughly estimated from the total input of nitrogen into the area, about 6 kmol s^{-1} (Howarth *et al.*, 1996) and from the percentage contribution of nitrate, about 26% according to Wollast (1993). This represents about 1.6 kmol s^{-1} of nitrate from river runoff. However, in estuaries and shelves, denitrification removes nitrate and Nixon *et al.* (1996) estimated an output of nitrate due to this process of about 5.7 kmol s^{-1} for the eastern boundary between 40–70N. The Arctic also contributes about 9.5 kmol s^{-1} (Anderson and Dyrssen, 1981; Anderson *et al.*, 1983). Figure 7a shows a schematic representation of the nitrate budget north of the 4x section. According to our circulation estimates $-50 \pm 19 \text{ kmol s}^{-1}$ of nitrate is exported across the section. Thus, this nitrate export must come from a net mineralization of organic nitrogen north of the 4x section producing $40.6 \pm 14 \text{ kmol s}^{-1}$ of nitrate (see Appendix for the error assessment), indicating that this region behaves as a net heterotrophic system, in agreement with the analysis by Schlüter *et al.* (2000) for the northern North Atlantic between 60–80N.

Next we consider the budget of organic nitrogen (dissolved, suspended and sedimented) also in steady state:

$$\frac{\Delta N_{\text{org}}}{\Delta t} = I - O - \text{Sed} + NP_{\text{org}} = 0. \quad (8)$$

where the inputs (I) minus the outputs (O) minus sedimentation (Sed) are balanced by biologically-mediated consumption or production of organic nitrogen (NP_{org}). In this balance we must also include the ammonium contribution as a reduced form of nitrogen. For inputs, atmospheric deposition represents about 13 kmol s^{-1} of organic nitrogen (Cornell *et al.*, 1995) and 4.3 kmol s^{-1} of ammonium (Duce *et al.*, 1991). River runoff introduces 2 kmol s^{-1} as ammonium and 2.4 kmol s^{-1} as organic nitrogen, with a total nitrogen input (nitrate, ammonium and organic nitrogen) equal to 6 kmol s^{-1} , as given by Howarth *et al.* (1996). The total mean contribution of nitrogen from the Arctic is 47 kmol s^{-1} (Michaels *et al.*, 1996). Given that 9.5 kmol s^{-1} come as nitrate and the negligible ammonium contribution (about 0.03 kmol s^{-1} , Gordeev *et al.*, 1996), most of the rest must be organic nitrogen. Organic nitrogen sedimentation in the open ocean north of the 4x section is very low, about 0.32 kmol s^{-1} according to Michaels *et al.* (1996). Likewise, the sedimentation in the shelves has been estimated to be 1.2 kmol s^{-1} (Wollast, 1998). We assume that only 20% of the organic nitrogen introduced by rivers is bioreactive (Wollast, 1993). If no accumulation of nitrogen is taking place in the ocean, the production of inorganic nitrogen must be balanced by the consumption of labile organic nitrogen plus ammonium and vice versa, thus NP_{org} would be roughly equal to $-40.6 \pm 14 \text{ kmol s}^{-1}$. Thus, we end up estimating an import of organic nitrogen plus ammonium from south of the 4x section roughly equal to $15 \pm 16 \text{ kmol s}^{-1}$ (Fig. 7b).

The Subpolar gyre ventilates the Subtropical region with an oxygen export of $-1992 \pm 440 \text{ kmol s}^{-1}$. The oxygen input from river runoff can be estimated assuming oxygen saturation and a river temperature of 10°C , corresponding to an oxygen concentration of $353 \mu\text{mol kg}^{-1}$. The river input into the ocean north of our section is 0.18 Sv , thus about 65 kmol s^{-1} of oxygen are introduced by rivers, quite a small quantity. A nitrate production of 40.6 kmol s^{-1} translates into an oxygen consumption of $431 \pm 149 \text{ kmol s}^{-1}$ using the $\Delta\text{O}_2/\Delta\text{N} = -170/16$ ratio of Anderson and Sarmiento (1994) since most of the reduced nitrogen is organic nitrogen with a very small quantity of ammonium. If no oxygen accumulation is taking place in the ocean north of the 4x section there exists an air to sea flux of $2358 \pm 460 \text{ kmol s}^{-1}$, equal to $4.6 \text{ mol m}^{-2} \text{ y}^{-1}$, in agreement with the range of fluxes predicted by Najjar and Keeling (2000) based on oxygen, wind speed and heat flux climatologies (Najjar and Keeling, 1997).

The air-sea thermally induced flux of oxygen is calculated according to Keeling *et al.* (1993):

$$F_{\text{TO}_2} = - \frac{\partial[\text{O}_2]_{\text{Sat}}^0}{\partial T} \frac{Q}{C_p} \quad (9)$$

where F_{TO_2} is the thermally driven oxygen flux, Q is the air-sea heat flux and C_p is the specific heat capacity, taken to be constant at $3993 \text{ J kg}^{-1} \text{ K}^{-1}$. The average variation of the oxygen saturation with temperature between 0 and 20°C is only $6 \mu\text{mol kg}^{-1} \text{ K}^{-1}$. From the heat transport across the 4x section Q is taken to be $0.65 \pm 0.1 \text{ PW}$ for the region north of the 4x section. Thus, this heat loss to the atmosphere drives an oxygen thermal influx of $977 \pm 150 \text{ kmol s}^{-1}$. This quantity can be considered as a lower limit, since cooling reduces the partial pressure of oxygen in water and mixing exposes oxygen-poor underlying waters, and both effects are additive, enhancing the oxygen flux into the ocean.

In summary, the Arctic-Subpolar region north of the 4x section behaves as a heterotrophic system, where nitrate is produced as a consequence of a net organic matter remineralization. A significant portion of this organic matter is imported from the temperate North Atlantic, south of the 4x section. Oxygen must be taken up from the atmosphere to accomplish the net remineralization. In addition, as a region of deep and mode water formation, the Arctic-Subpolar region takes up oxygen from the atmosphere as a result of the thermally driven flux. Thus, in the Arctic-Subpolar area both the solubility and biological pumps drive a remarkable oxygen uptake by the ocean, with a likely greater contribution from the solubility pump.

For the region south of the 4x section we will now use the Rintoul and Wunsch (1991) and Ganachaud and Wunsch (1998) transport results for the 36N zonal section in order to obtain sensible nitrate, organic nitrogen and oxygen budgets between the 4x and 36N sections (Fig. 1). Therefore, the following analysis will depend both on our analysis of the 4x section reported here and on the Rintoul and Wunsch's (1991) and Ganachaud and Wunsch's (1998) analysis of the 36N section. In this regard, Rintoul and Wunsch (1991) calculated a northward nitrate flux of $119 \pm 35 \text{ kmol s}^{-1}$ and a southward oxygen flux of $-2940 \pm 180 \text{ kmol s}^{-1}$ from a regional inverse model. Ganachaud and Wunsch (1998) using a global inverse model proposed a northward nitrate and a southward oxygen flux of $120 \pm 50 \text{ kmol s}^{-1}$ and $-3500 \pm 800 \text{ kmol s}^{-1}$, respectively. Combined with the 4x results presented here, these figures suggest a nitrate convergence of about $170 \pm 71 \text{ kmol s}^{-1}$ and an oxygen divergence ranging from 948 to 1508 kmol s^{-1} .

Nitrate and organic nitrogen (plus ammonium) budgets for the box comprised within the 4x and the 36N section were performed, using Eqs. 7 and 8, and the same assumptions and data sources as in the previous budgets (Fig. 7). We estimate a net consumption of nitrate of $162 \pm 56 \text{ kmol s}^{-1}$ for the region between the sections, and a organic nitrogen export southwards across 36N of $-154 \pm 66 \text{ kmol s}^{-1}$.

The difference between our estimate for the heat flux across the 4x section and that from Rintoul and Wunsch (1991) across 36N is $0.55 \pm 0.2 \text{ PW}$. Using Eq. 9 this heat loss to the atmosphere implies an oxygen thermal influx of $826 \pm 300 \text{ kmol s}^{-1}$ within the region between both sections, probably mainly captured within the Labrador Sea convection area. River runoff into the region contributes 26 kmol s^{-1} (estimated assuming oxygen saturation at a mean river temperature). The consumption of nitrate within the box implies an oxygen production of $1721 \pm 595 \text{ kmol s}^{-1}$ ($\Delta\text{O}_2/\Delta\text{N} = -170/16$, Anderson and Sar-

miento, 1994). Summing up, about 2350 kmol s^{-1} of oxygen are introduced within the box, exceeding the hydrographic oxygen divergence range ($948\text{--}1508 \text{ kmol s}^{-1}$) and pointing to a net oxygen out-gassing of $799 \pm 799 \text{ kmol s}^{-1}$ toward the atmosphere in this temperate area, in agreement with the results of Boyer *et al.* (1999).

Lacking any measurements of organic nitrogen either as dissolved or particulate, our conclusions concerning the biogeochemical behavior of the temperate North Atlantic between the 4x and the 36N sections as an organic nitrogen source remain speculative. However, indirect evidence supports our hypotheses. High levels of primary production are reported for the area (Antoine *et al.*, 1996; Berger, 1989; Longhurst *et al.*, 1995). This implies high concentrations of organic matter, particulate or dissolved, as inferred by Bishop (1989) using empirical functions relating a well measured quantity (ocean temperature, nutrients or light) with organic matter abundance. Moreover, given the low sedimentation rates of the area (Broecker and Peng, 1982; Jahnke, 1996; Lampitt and Antia, 1997) this organic matter is prone to be exported outward or recycled within the water column. If remineralized, there should be a transport of nitrate across 36N and/or the 4x sections, but neither the results of Rintoul and Wunsch (1991) nor ours suggest so. Moreover, given that there is no accumulation of nitrogen in the area (steady-state assumption) and only 1% of the organic matter produced reaches the deep-sea floor (Lampitt and Antia, 1997), the organic nitrogen produced cannot be recycled within the area and must be exported, thus, ascribed to the whole water column new production. In this sense, our 162 kmol s^{-1} of produced nitrogen can be translated into a new production rate of about $74 \text{ gC m}^{-2} \text{ y}^{-1}$, using the Anderson and Sarmiento (1994) $\Delta\text{C}/\Delta\text{N}$ ratio, representing 24% of the total primary production ($303 \text{ gC m}^{-2} \text{ y}^{-1}$) estimated by Longhurst *et al.* (1995) for the region between the sections. Unfortunately, these numbers cannot be compared with the estimations given in the literature which are principally defined for the euphotic zone (e.g. Berger, 1989; Campbell and Aarup, 1992; Eppley and Peterson, 1979).

On the other hand, oxidation of the organic nitrogen exported southward across 36N in its transit to the Subtropical gyre could provide the necessary regenerated nitrate going northward across 36N (Rintoul and Wunsch, 1991; Ganachaud and Wunsch, 1998), and/or flowing southward across 24N according to Lavín (1999). Additionally, this oxidation would explain the high OUR (Oxygen Utilization Rates) reported by Jenkins and Goldman (1985) in the subtropical area south of 36N. Finally, this southward flux of organic nitrogen would also support the hypothesis of a net lateral input of organic matter into the Subtropical gyre, in order to balance the estimated nitrate budget on the basis of radon measurements performed by Sarmiento *et al.* (1990). Moreover, our results also support one of the possible scenarios dealing with the transport of dissolved organic nitrogen in the North Atlantic proposed by Walsh *et al.* (1992) in which 170 kmol s^{-1} flow southward across 36N.

If the Bering Strait flow of -0.8 Sv (Roach *et al.*, 1995) is added barotropically across the 4x section flowing at the section mean value of each property, there would be additional southward nitrate, phosphate, silicate and oxygen transports of -14 , -1 , -13 and

-209 kmol s^{-1} , respectively. Given the range of uncertainty of the fluxes (Table 3) these contributions are relatively low.

5. Summary and concluding remarks

Hydrographic and chemical data from the WOCE A25 (4x cruise) allowed us to estimate the transport of physical and chemical properties across the southern boundary of the Subpolar North Atlantic Ocean. The “initial guess” for the circulation is obtained by modifying the geostrophic velocity field calculated for a fixed level of no motion with barotropic adjustments applied in specific areas in order to reproduce the closest-to-reality estimate of the velocity field, and with a final uniform velocity adjustment to conserve mass and salt, as we assume the North Atlantic Ocean north of the 4x section to be a closed basin. The ageostrophic wind-driven Ekman transport is calculated taking into account the seasonal variations of the wind stress, taken from the SOC climatology (Josey *et al.*, 2002), and of the properties in the Ekman layer, taken from the World Ocean Atlas 1998 data set.

Finally, over this “initial guess” an inverse method is implemented so as to force the silicate flux to be in agreement with the input of silicate north of the section, about 26 kmol s^{-1} according to Tréguer *et al.* (1995). Thus the “best estimate” of the circulation is obtained. The main sources of error affecting the mass and property transports are evaluated, revealing the significance of a proper spatial coverage to resolve the eddy scale.

The calculated “best estimate” of the circulation across the 4x section is consistent with the current knowledge about water mass transports in the Subpolar North Atlantic. The strength of the Subpolar gyre is -25.4 Sv comprising the southward transport of Denmark Strait Overflow Water (DSOW) within the East Greenland Current (EGC), which balances the northeastward transport of the North Atlantic Current (NAC) system of 27.7 Sv . Over the Iberian Abyssal Plain 2.6 Sv of waters with $\sigma_2 > 36.98 \text{ kg m}^{-3}$ flow northward, reflecting the 2 Sv of Antarctic Bottom Water entering the eastern North Atlantic basin across the Vema fracture (Schmitz and McCartney, 1993).

With the aim of understanding the processes leading the transports of the physical and chemical properties the geostrophic fluxes are decomposed into their barotropic, baroclinic and horizontal components following Bryden *et al.* (1991). The overturning circulation across the Subpolar gyre is evaluated as the integrated baroclinic flow of $-16.5 \pm 3.6 \text{ Sv}$ southward between 1100 and 3200 db, which are compensated by a northward flow in the upper 1100 db of 14.8 Sv and of 1.7 Sv below 3200 db.

The net heat flux across the section amounts to $0.65 \pm 0.1 \text{ PW}$ poleward. The overturning cell contributes 54% of the heat loss to the atmosphere due to water mass formation north of the section, with the remaining 45% ascribed to the horizontal circulation.

The combination of our results with those from Bacon (1997) and Rintoul and Wunsch (1991) for a section from Ireland to Cape Farewell and the 36N zonal section, respectively, allows us to make a direct estimate of the heat convergence between sections which reveals

a great discrepancy with the annual heat loss calculated from the SOC climatology, especially in the eastern North Atlantic Ocean.

As the salt transport across the section is practically negligible ($-0.8 \pm 6.7 \text{ Mkg s}^{-1}$), the mass imbalance, $-0.4 \pm 1.5 \text{ Sv}$ southwestward, is equivalent to an excess of precipitation plus runoff over evaporation over the North Atlantic Ocean north of the 4x section.

According to our estimates the North Atlantic Subpolar gyre exports nutrients and oxygen toward the temperate North Atlantic, at rates of -50 ± 19 , -6 ± 2 , $-1992 \pm 440 \text{ kmol s}^{-1}$ for nitrate, phosphate and oxygen, respectively. The main mechanism responsible for the nutrient fluxes is the overturning circulation removing nutrients from the region north of the 4x section, as it transports nutrient-poor surface water northeastward and nutrient-rich deeper water southwestward. In the case of oxygen, the principal contributor is the horizontal circulation due to the strong oxygen contrast in the 500 to 1000 db layer between the western and eastern ends of the section. High oxygen Labrador Sea Water flows south in the west while low oxygen older waters, including Mediterranean Water, flow north in the east.

Assuming that the nitrate budget is in steady state on time-scales of about 10 years, we evaluate the sources and sinks of nitrate within the region between the Bering Strait and the 4x section (Arctic-Subpolar region). Our findings point to a net heterotrophic system, producing nitrate at $40.6 \pm 14 \text{ kmol s}^{-1}$ as a result of a net remineralization of organic matter, which is partially imported from the temperate North Atlantic.

The poleward transport of nitrate across 36N (Ganachaud and Wunsch, 1998; Rintoul and Wunsch, 1991) and its equatorward transport across the 4x section cause a convergence of nitrate in the temperate North Atlantic Ocean. A tentative nitrogen budget suggest that this region behaves as a source of organic nitrogen, with a northward transport of organic nitrogen across the 4x section of $15 \pm 16 \text{ kmol s}^{-1}$ and a southward transport across the 36N section of $154 \pm 66 \text{ kmol s}^{-1}$. While this hypothesis will remain tentative until organic nitrogen measurements are done across the two sections, it is supported by indirect evidence. The high levels of primary production (Antoine *et al.*, 1996; Berger, 1989; Longhurst *et al.*, 1995) reported for the area between the sections imply high concentrations of organic matter (Bishop, 1989). Given low sedimentation rates (Jahnke, 1996) the organic matter must be exported, since no outwards nitrate transport is obtained either across the 4x or the 36N section. The oxidation of this southward-flowing organic nitrogen across 36N would also explain the high oxygen utilization rates reported for the Sargasso area (Jenkins and Goldman, 1985) and confirm one of the possible scenarios dealing with the transport of dissolved organic nitrogen in the North Atlantic proposed by Walsh *et al.* (1992).

The circulation scheme and transport results obtained across the WOCE A25, 4x section are consistent with both dynamical and biogeochemical principles. The computed fluxes of physical and chemical species provide valuable information about the behavior of the Subpolar North Atlantic on long time-scales. This information could be useful to constrain

global ocean circulation models and helpful to predict future climate changes. We emphasize the importance of inferring regional divergences/convergences of physical or chemical properties when combining results from different transatlantic sections. We have provided a tentative composite view for the nitrogen and oxygen budgets in the Subpolar and temperate North Atlantic Ocean.

Acknowledgments. We are grateful to Sheldon Bacon, Principal Scientist, to the captain, officers, and crew of RRS *Discovery* and to the science team for their hard work during the 4x cruise. The cruise was primarily supported by the Natural Environment Research Council (NERC) under the UK WOCE Community Research Programme. Spanish participation was funded by the Spanish CICYT project MAR-97-1622-E. The stay of M. Álvarez at SOC was financed by a “Beca de estancia no estranxeiro” and a “Beca Predoutoral” from the Xunta de Galicia. The authors would like to thank X. A. Álvarez-Salgado for valuable suggestions and comments, P. Saunders and S. Bacon for their help with the circulation, and S. Josey for aid with the SOC climatology. Constructive comments by two anonymous reviewers helped to improve the manuscript.

APPENDIX

In this section we try to estimate the maximum error in the fluxes and budgets.

Error estimation on the fluxes

The likely error sources are the precision of the measurements, their temporal variability, the spatial resolution of the sampling and errors on estimating the Ekman layer transports. Moreover, we must consider the sensitivity of our calculations to the imposed circulation on certain areas, the CGFZ, the EGC and the IAP, as well as the silicate constraint. Another source of error arises from considering the North Atlantic Ocean north of our section as a closed basin, disregarding the flow through the Bering Strait, so the impact of this assumption is assessed in the corresponding section dealing with each property transport.

Measurements errors. Errors due to the uncertainty in pressure, salinity and temperature determination can be considered negligible, not affecting the heat and freshwater transports (Holfort and Siedler, 2001). Nitrate, phosphate, silicate and oxygen values have a maximum error of $\pm 0.1 \mu\text{mol kg}^{-1}$, $\pm 0.05 \mu\text{mol kg}^{-1}$, $\pm 0.1 \mu\text{mol kg}^{-1}$ and $\pm 1 \mu\text{mol kg}^{-1}$, respectively. These errors are randomly distributed along the section and should be small compared to other errors, so we do not take them into account.

Temporal variability errors. We are aware of the decadal and interannual changes in the thermohaline characteristics of intermediate and deep waters within this oceanic area (e.g., Curry *et al.*, 1998; Dickson *et al.*, 1996; Read and Gould, 1992). Some studies concerned the analyses of repeated full-depth ocean-wide hydrographic sections in the northern North Atlantic (Bersch *et al.*, 1999; Kolterman *et al.*, 1999; Read and Gould, 1992; Sy *et al.*, 1997). Concretely, Kolterman *et al.* (1999) examined the decadal variability of the mass, heat and freshwater transports across a section comprising the latitude band between

Table A1. “Best estimate” final transports (Final) across the 4x section and flux errors due to the eddy resolution (Eddy), variations in the Ekman transport (Ekman), in the patterns of circulation across the section (Circ.), due to the inverse model settings (Inv.), the silicate constraint (Silicate) and Bottom Triangles (BT). Units are Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$), PW (10^{15} W), Mkg s^{-1} (10^6 kg s^{-1}) and kmol s^{-1} (10^3 mol s^{-1}).

	Final	Eddy	Ekman	Circ.	Inv.	Silicate	BT	Total
Mass—Sv	−0.40	1.25	0.14	0.02	0.04	0.01	0.02	1.48
Heat—PW	0.69	0.01	0.01	0.04	0.04	0.00	0.01	0.11
Salt—Mkg s^{-1}	−0.8	0.5	4.9	0.7	0.2	0.1	0.3	6.7
NO_3 — kmol s^{-1}	−50	2	1	4	6	2	4	19
PO_4 — kmol s^{-1}	−6	0	0	1	1	0	0	2
SiO_4 — kmol s^{-1}	−26	2	1	0	0	12	0	15
O_2 — kmol s^{-1}	−1992	31	38	152	180	24	15	440

43–48N, and suggested significant changes in the strength of the meridional overturning circulation, leading to decadal changes in the transport of heat and freshwater. However, it was beyond the scope of this paper to study the long-term variability of the transports across the 4x section, and we assume our data to be climatologically representative of the mean oceanic state.

Spatial resolution error. The 4x section comprises 89 stations separated by distances ranging from 5 km to a maximum of 55 km, which should be close enough to resolve the main eddy scale in the northern North Atlantic (50–100 km). On the other hand, the effect of reducing the spatial resolution was assessed by reducing to half the number of stations and calculating again the transports approximating the circulation field as explained in the corresponding section, constraining the salt flux to zero and the silicate transport to -26 kmol s^{-1} . Reducing the spatial resolution to half increases the heat transport from 0.69 to 0.71 PW, leading also to an increase in the southward nitrate, phosphate and oxygen transports of 18%, 5% and 6%, respectively. Moreover, changes are produced in the circulation field. The mass transport in the IAP for waters $\sigma_2 > 37 \text{ kg m}^{-3}$ increases from 2.2 to 2.6 Sv, in the EGC is reduced from -25.4 to -24.8 Sv and the transport in the CGFZ for waters $\sigma_\theta > 27.8 \text{ kg m}^{-3}$ is reduced from -2.6 to -2.1 Sv.

It is possible to make some estimate of the error incurred due to eddy activity. If the total eddy transport of each property has an error equal to the standard deviation (STD) of the eddy transports between station pairs. This STD is modified by a normally distributed random number, this modification is done about a hundred times to reflect the number of stations, and the STD of the net eddy flux for the complete section is calculated. We evaluate the error due to eddy activity as equal to this final STD. See Table A1 for the estimated eddy error contribution of each property to the total flux error.

Ekman error. Uncertainties in the Ekman transport arise from the differences in wind stress climatologies, the actual temporal variability in the wind stress and the Ekman layer depth.

The SOC wind stress climatology was used in this study as representative of the wind forcing during WOCE (Josey *et al.*, 2002). This climatology includes data from 1980 to 1993, and was calculated using a drag coefficient confirmed by recent observational analyses. Compared with the Hellerman and Rosenstein (1983) climatology in the North Atlantic Subpolar gyre, the SOC analyses doubles the strength of the Ekman pumping. This difference is ascribed to the different sampling periods and choice of drag coefficient (Josey *et al.*, 2002).

The Ekman layer depth was chosen to be 75 db for the whole section and we think that the error due to its variability is lower than that ascribed to the seasonal variation of wind stress and properties in the Ekman layer. In this sense, seasonal means for wind stress were calculated and then coupled with the mean seasonal values for each property in the upper 75 db taken from the World Ocean Atlas 1998-NOAA. The Ekman transport calculated from the annual mean wind stress is -1.36 Sv while the mean of the seasonal Ekman transports is -1.38 Sv. Likewise, the Ekman heat flux decreases by 30% when considering the seasonal variation of temperature, but the salt transport practically does not change. The nitrate, phosphate, silicate and oxygen Ekman transports decrease in a 21%, 32%, 55% and 6%, respectively, when using the 4x values, mainly because the surface concentrations of these chemical properties are biased to low summer values in the upper layer of our section. The Ekman mass transport was allowed to vary $\pm 10\%$ in order to calculate the uncertainty due to the Ekman transport (Table A1).

Circulation error. The sensitivity of the fluxes to changes in the circulation is assessed by varying our imposed mass transports in the CGFZ, the EGC and the IAP. Thus, in the CGFZ the mass flux for $\sigma_\theta > 27.8 \text{ kg m}^{-3}$ was allowed to change within -2.4 ± 0.5 Sv (Saunders, 1994); the mass flux was varied by -25 ± 5 Sv over the whole water column of the EGC (Bacon, 1997); and in the IAP by 0 ± 2 Sv for $\sigma_2 > 37 \text{ kg m}^{-3}$. Several sets of combinations were evaluated and adjusted for a zero salt and -26 kmol s^{-1} silicate transport across the section. The circulation error was calculated as the minimum minus the maximum transport of each property among all the combinations divided by two, thus, half of the range of the new transports. The circulation errors (Table A1) are considered to be due to changes in the main pattern of circulation across the 4x section and mainly ascribed to alterations in the strength of the EGC (-25 ± 5 Sv).

Inversion error. The sensitivity of the final fluxes to variations in the inverse model solution was assessed by adding to the three principal constraints (no change in the mass transport, salt conservation and a silicate flux equal to -26 kmol s^{-1}) another three, constraining the mass flux over the CGFZ, the IAP and east of Greenland to be -2.4 Sv, -25 Sv and 0 Sv, respectively. Thus, there is a set of 5 equations to be solved with SVD. The number of eigenvalues used to calculate the barotropic corrections, or matrix rank, expresses the importance of the set of equations over the initial state. The higher the matrix rank the better fulfilment of the constraints but greater deviations are introduced in the

initial velocity field. A system with 5 equations has 5 possible solutions, and therefore 5 sets of final fluxes can be calculated. The error due to the inverse model solution is approximated as previously, with half of the range of the new calculated transports, disregarding those obtained using just the first eigenvalue as it practically does not change the initial state. See Table A1.

Within the inverse model error we must consider the effect of a wrong estimation of the silicate river input or of the non-conservative nature due to biological activity. The silicate error is assessed allowing a variation of $\pm 15\%$ around the -26 kmol s^{-1} value, within the context of the usual SVD we performed. The range divided by two of the new fluxes is considered to be the error ascribed to the silicate flux. See Table A1.

Bottom triangles' error. The velocity and properties' distribution at the Bottom Triangles (BT) can be approximated in different ways: some authors do not consider the flow across the BT when the velocity field has been already designed to conserve mass (e.g., Lavín *et al.*, 1998); others assume that the velocity decreases linearly from the deepest common level to usually 1000 m below or to the bottom (e.g., Holfort and Siedler, 2001).

In this work the velocity at the BT is approximated to be the same as the velocity at the deepest common level of each pair of stations. Correspondingly, the BT properties are calculated as a weighted-mean taking into account the number of 20 db layers below the maximum common level:

$$\text{PropMean} = \frac{\sum_i^{nn} \text{Prop}(MCL + ii) * (nn + 1 - ii)}{nn * (nn + 1)/2}$$

where PropMean is the weighted mean of the property at the BT, MCL is the Maximum Common Level, Prop is the property concentration below the MCL at the deepest station within each pair, *nn* is the number of levels below the MCL and *ii* is the counter.

For bottom triangles, we estimate the average property in the bottom triangle as a triangle-weighted average of the property profile at the deepest station and we consider this estimate to have a negligible error for the BT property. Since the BT transport can only be smaller than the approximation considered here, we recalculated the circulation field in the same way but assuming no transport in the BT. The difference between the two approximations is taken here as the error due to the BT definition (Table A1). The small error due to BT variations is a result of the fact that with and without BT transport the circulation pattern is constrained similarly and that the section-averaged property is not very different from the BT property for most of the properties.

In general, we tried to assess the main sources of uncertainty in the estimated transports across the 4x section, giving the maximum contribution from each source of uncertainty. The eddy error arises as the main contributor in constraining the mass flux across the section. Uncertainty in the Ekman transport mainly affects the salt transport with a minor effect elsewhere. The largest uncertainties in the fluxes of heat, nitrate, phosphate and

oxygen arise from alterations in the circulation across the section while variations in the silicate constraint only slightly affect the final fluxes other than for silicate.

Error estimation on the budgets

The budgets of nitrate, organic nitrogen and oxygen performed in this work assume steady state on time-scales on the order of the flushing time for the considered region. A maximum error is assigned to every known term of the balance (e.g., fluxes across sections, denitrification, atmospheric inputs, sedimentation). For the fluxes, the error is taken from the bibliographic source or directly from our estimations. A normally distributed random number multiplies these maximum errors to be added to the initial values. Then, the inferred quantity is calculated. This procedure is repeated a hundred times, finally the standard deviation (STD) of the resulting set of fluxes is obtained. This STD corresponds to the error on the inferred variable in the budget equation.

REFERENCES

- Aagaard, K. and E. Carmack. 1989. The role of sea ice and other fresh water in the Arctic circulation. *J. Geophys. Res.*, *94*, 14485–14498.
- Álvarez, M., F. F. Pérez, A. F. Ríos, G. Rosón and D. Smythe-Wright. 2002. Mixing analysis in the North Atlantic Subpolar gyre. *Deep-Sea Res. I*, (submitted).
- Ambar, I., A. F. G. Fiúza, T. Boyd and R. Frouin. 1986. Observations of a warm oceanic current flowing northward along the coasts of Portugal and Spain during November–December 1983. *EOS Trans. Amer. Geophys. Union*, *67*, 1054.
- Anderson, L. A. and J. L. Sarmiento. 1994. Redfield ratios of remineralization determined by nutrient data analysis. *Global Biogeochem. Cycles*, *8*, 65–80.
- Anderson, L. G. and D. W. Dyrssen. 1981. Chemical constituents of the Arctic Ocean in the Svalbard area. *Oceanol. Acta*, *4*, 305–311.
- Anderson, L. G., D. W. Dyrssen, E. P. Jones and M. G. Lowings. 1983. Inputs and outputs of salt, freshwater, alkalinity and silicate in the Arctic Ocean. *Deep-Sea Res.*, *30*, 87–94.
- Antoine, D., J. M. André and A. Morel. 1996. Oceanic primary production, 2, Estimation at global scale from satellite (coastal zone color scanner) chlorophyll. *Global Biogeochem. Cycles*, *10*, 57–70.
- Arhan, M., A. Colin de Verdière and L. Mémary. 1994. The eastern boundary of the subtropical North Atlantic. *J. Phys. Oceanogr.*, *24*, 1295–1316.
- Bacon, S. 1997. Circulation and fluxes in the North Atlantic between Greenland and Ireland. *J. Phys. Oceanogr.*, *27*, 1420–1435.
- 1998. RRS *Discovery* cruise 230, 07-Aug-17 Sep 1997. Two hydrographic sections across the boundaries of the Subpolar Gyre: FOUREX. Southampton Oceanography Centre Cruise Report, No 16, 104 pp.
- Bauch, D., P. Schlosser and R. G. Fairbanks. 1995. Freshwater balance and the sources of deep and bottom waters in the Arctic Ocean inferred from the distribution H_2^{18}O . *Progr. Oceanogr.*, *35*, 53–79.
- Baumgartner, A. and E. Reichel. 1975. *The World Water Balance*. Elsevier, 179 pp.
- Berángier, K., L. Siefridt, B. Barnier, E. Garnier and H. Roquet. 1999. Evaluation of operational ECMWF surface freshwater fluxes over oceans during 1991–1997. *J. Mar. Syst.*, *22*, 13–36.
- Berger, W. H. 1989. Global maps of ocean productivity, in *Productivity of the Ocean: Present and Past*, W. H. Berger, V. S. Smetacek and G. Wefer, eds., 429–455.
- Bersch, M., J. Meincke and A. Sy. 1999. Interannual thermohaline changes in the northern North Atlantic 1991–1996. *Deep-Sea Res. II*, *46*, 55–76.
- Bishop, J. K. B. 1989. Regional extremes in particulate matter composition and flux: effects on

- the chemistry of the ocean interior, *in* Productivity of the Ocean: Present and Past, W. H. Berger, V. S. Smetacek and G. Wefer, eds., 117–137.
- Boyer, T., M. E. Conkright and S. Levitus. 1999. Seasonal variability of dissolved oxygen, percent oxygen saturation and apparent oxygen utilization in the Atlantic and Pacific oceans. *Deep-Sea Res. I*, **46**, 1593–1613.
- Brewer, P. G. and D. Dyrssen. 1987. Ocean chemical fluxes across 25°N in the Atlantic Ocean. Discussion paper prepared for the International GOFS Meeting, Paris, 17–20 February 1987, 25 pp.
- Brewer, P. G., C. Goyet and D. Dyrssen. 1989. Carbon dioxide transport by ocean currents at 25°N Latitude in the Atlantic Ocean. *Science*, **246**, 477–479.
- Broecker, W. S. and G. H. Denton. 1989. The role of ocean-atmosphere reorganisations in glacial cycles. *Geoch. Cosmoch. Acta*, **53**, 2465–2501.
- Broecker, W. S. and T.-H. Peng. 1982. Tracers in the Sea. Lamont-Doherty Geological Observatory, Columbia University, Palisades, NY, 690 pp.
- Bryden, H. L. 1993. Ocean heat transport across 24° latitude, *in* Interactions Between Global Climate Subsystems: The Legacy of Hann, G. A. McBean and M. Hantel, eds., Geophysical Monograph, **75**, 65–75.
- Bryden, H. L., D. H. Roemmich and J. A. Church. 1991. Ocean heat transport across 24°N in the Pacific. *Deep-Sea Res.*, **38**, 297–324.
- Campbell, J. W. and T. Aarup. 1992. New production in the North Atlantic derived from seasonal patterns of surface chlorophyll. *Deep-Sea Res.*, **39**, 1669–1694.
- Coachman, L. K. and K. Aagaard. 1989. Transports through Bering Strait: Annual and interannual variability. *J. Geophys. Res.*, **93**, 1535–15539.
- Coachman, L. K., K. Aagaard and R. B. Tripp. 1975. Bering Strait: The Regional Physical Oceanography. Univ. of Washington Press, Seattle, WA, 172 pp.
- Cornell, S., A. Rendell and T. Jickells. 1995. Atmospheric inputs of dissolved organic nitrogen to the oceans. *Nature*, **376**, 243–246.
- Curry, R. G., M. S. McCartney and T. M. Joyce. 1998. Oceanic transport of subpolar climate signals to mid-depth subtropical waters. *Nature*, **391**, 575–577.
- Daniault, N., J. P. Mazé and M. Arhan. 1994. Circulation and mixing of Mediterranean Water west of the Iberian Peninsula. *Deep-Sea Res. I*, **41**, 1685–1714.
- da Silva, A., C. Young-Molling and S. Levitus. 1996. Revised surface marine fluxes over the global oceans: The UVW/COADS data set. WCRP workshop on air-sea flux field for forcing ocean models and validating CGMS (European Centre for Medium-range Weather Forecasts, Reading, UK, 24–27 October 1995). White, G. (Ed.), WMO/TD-N, 762.
- Dickson, R. and J. Brown. 1994. The production of North Atlantic Deep Water: sources, rate and pathways. *J. Geophys. Res.*, **99**, 12319–12341.
- Dickson, R. R., W. J. Gould, T. J. Müller and C. Maillard. 1985. Estimates of the mean circulation in the deep (>2000 m) layer of the eastern North Atlantic. *Progr. Oceanogr.*, **14**, 103–127.
- Dickson, R. R., J. Lazier, J. Meincke, P. Rhines and J. Swift. 1996. Long-term coordinated changes in the convective activity of the North Atlantic. *Progr. Oceanogr.*, **38**, 241–295.
- Dietrich, G. 1969. Atlas of the Hydrography of the Northern North Atlantic Ocean, Conseil International Pour L'Exploration de la Mer Service Hydrographique, Charlottenlund Slot, Denmark, 140 pp.
- Duce, R. A., P. S. Liss, J. T. Merrill, E. L. Atlas, P. Buat-Menard, D. D. Hicks, J. M. Miller, J. M. Prospero, R. Arimoto, T. M. Church, W. Ellis, J. N. Galloway, J. Hansen, T. D. Jickells, A. H. Knap, K. H. Reinhardt, B. Schneider, A. Soudine, J. J. Tokos, S. Tsunogai, R. Wollat and M. Zhou. 1991. The atmospheric input of trace species to the world ocean. *Global Biogeochem. Cycles*, **5**, 193–259.
- Eppley, R. W. and B. J. Peterson. 1979. Particulate organic matter flux and planktonic new production in the deep ocean. *Nature*, **282**, 677–680.

- Fissel, D. B., D. D. Lemon, H. Melling and R. A. Lake. 1988. Nontidal flows in the Northwest Passage. Canadian Tech. Rep. Hydrography Ocean Series, 98, 143 pp.
- Frouin, R., A. F. G. Fiúza, I. Ambar and T. J. Boyd. 1990. Observations of a poleward surface current off the coasts of Portugal and Spain during winter. *J. Geophys. Res.*, 95, 679–691.
- Fu, L.-L. 1981. The general circulation and meridional heat transport of the subtropical South Atlantic determined by inverse methods. *J. Phys. Oceanogr.*, 11, 1171–1193.
- Ganachaud, A. and C. Wunsch. 1998. Large scale oceanic nutrient and oxygen fluxes. *International WOCE Newsletter*, 32, 12–15.
- Gordeev, V. V., J. M. Martin, I. S. Sidorov and M. V. Sidorova. 1996. A reassessment of the Eurasian River input of water, sediment, major elements and nutrients to the Arctic Ocean. *Amer. J. Sci.*, 296, 666–691.
- Hall, M. M. and H. L. Bryden. 1982. Direct estimates and mechanisms of ocean heat transport. *Deep-Sea Res.*, 29, 339–359.
- Hastenrath, S. 1982. On meridional heat transports in the world ocean. *J. Phys. Oceanogr.*, 12, 922–927.
- Haynes, R. and E. D. Barton. 1990. A poleward flow along the Atlantic coast of the Iberian Peninsula. *J. Geophys. Res.*, 95, 11425–11441.
- Hellerman, S. and M. Rosenstein. 1983. Normal monthly wind stress over the World Ocean with error estimates. *J. Phys. Oceanogr.*, 13, 1093–1104.
- Holfort, J. and G. Siedler. 2001. The meridional oceanic transports of heat and nutrients in the South Atlantic. *J. Phys. Oceanogr.*, 31, 5–29.
- Howarth, R. W., G. Billen, D. Swaney, A. Townsend, N. Jaworski, K. Lajtha, J. A. Downing, R. Elmgren, N. Caraco, T. Jordan, F. Berendse, J. Freney, V. Kudeyarow, P. Murcoch and Shu Shao-Liang. 1996. Regional nitrogen budgets and riverine N & P fluxes for the drainages to the North Atlantic Ocean: Natural and human influences. *Biogeochem.*, 35, 75–139.
- Isemer, H. J., R. S. Pickart, R. M. Hendry and W. J. Smethie, Jr. 1989. Fine adjustment of large-scale air-sea energy flux parameterizations by direct estimate of ocean heat transport. *J. Clim.*, 2, 1173–1186.
- Jahnke, R. A. 1996. The global ocean flux of particulate organic carbon: areal distribution and magnitude. *Global Biogeochem. Cycles*, 10, 71–88.
- Jenkins, W. J. and J. C. Goldman. 1985. Seasonal cycling and primary production in the Sargasso Sea. *J. Mar. Res.*, 43, 465–491.
- Josey, S. A., E. C. Kent and P. K. Taylor. 1998. The Southampton Oceanography Centre (SOC) ocean-atmosphere heat, momentum and freshwater flux atlas. Southampton Oceanography Centre Rp. 6, Southampton, United Kingdom, 30 pp. + figures.
- 1999. New insights into the ocean heat budget closure problem from analysis of the SOC air-sea flux climatology. *J. Clim.*, 12, 2856–2880.
- 2002. On the wind stress forcing of the ocean in the SOC climatology: Comparison with the NCEP/NCAR, ECMWF, UWM/COADS and Hellerman and Rosenstein datasets. *J. Phys. Oceanogr.*, (in press).
- Keeling, R. F., R. P. Najjar, M. L. Bender and P. P. Tans. 1993. What atmospheric oxygen measurements can tell us about the global carbon cycle. *Global Biogeochem. Cycles*, 7, 37–67.
- Kolterman, K. P., A. V. Sokov, V. P. Tereschenkov, K. Lorbacher and A. Sy. 1999. Decadal changes in the thermohaline circulation of the North Atlantic. *Deep-Sea Res. II*, 46, 109–138.
- Krauss, W. 1986. The North Atlantic Current. *J. Geophys. Res.*, 91, 5061–5074.
- Krauss, W. and R. Käse. 1984. Mean circulation and eddy kinetic energy in the eastern North Atlantic. *J. Geophys. Res.*, 89, 3407–3415.
- Lampitt, R. S. and A. N. Antia. 1997. Particle flux in deep seas: regional characteristics and temporal variability. *Deep-Sea Res. I*, 44, 1377–1403.

- Lavín, A. 1999. Fluxes, trends and decadal changes in the subtropical North Atlantic. PhD. dissertation, University of Santander, 222 pp.
- Lavín, A., H. L. Bryden and G. Parrilla. 1998. Meridional transport and heat flux variations in the subtropical North Atlantic. *The Global Atmos. Ocean Syst.*, *6*, 269–293.
- Lipschultz, F. and N. J. P. Owens. 1996. An assessment of nitrogen fixation as a source of nitrogen to the North Atlantic. *Biogeochem.*, *35*, 261–274.
- Longhurst, A., S. Sathyendranath, T. Platt and C. Caverhill. 1995. An estimate of global primary production in the ocean from satellite radiometer data. *J. Plankton Res.*, *17*, 1245–1272.
- Macdonald, A. and C. Wunsch. 1996. An estimate of global ocean circulation and heat fluxes. *Nature*, *382*, 436–439.
- Martel, F. and C. Wunsch. 1993. The North Atlantic circulation in the early 1980s—an estimate from inversion of a finite difference model. *J. Phys. Oceanogr.*, *29*, 283–383.
- Mazé, J. P., M. Arhan and H. Mercier. 1997. Volume budget of the eastern boundary layer off the Iberian Peninsula. *Deep-Sea Res. I*, *44*, 1543–1574.
- McCartney, M. S. and L. D. Talley. 1982. The subpolar mode water of the North Atlantic Ocean. *J. Phys. Oceanogr.*, *12*, 1169–1188.
- Michaels, A. F., D. Olson, J. L. Sarmiento, J. W. Ammerman, K. Fanning, R. Jahnke, A. H. Knap, F. Lipschultz and J. M. Prospero. 1996. Inputs, losses and transports of nitrogen and phosphorus in the pelagic North Atlantic. *Biogeochem.*, *35*, 181–226.
- Najjar, R. G. and R. F. Keeling. 1997. Analysis of the mean annual cycle of dissolved oxygen anomaly in the World Ocean. *J. Mar. Res.*, *55*, 117–151.
- . 2000. Mean annual cycle of the air-sea oxygen flux: A global view. *Global Biogeochem. Cycles*, *14*, 573–584.
- Nelson, D. M., P. Tréguer, M. A. Brzezinski, A. L. Leynaert and B. Quéquiner. 1995. Production and dissolution of biogenic silica in the ocean: Revised global estimates, comparison with regional data and relationship to biogenic sedimentation. *Global Biogeochem. Cycles*, *9*, 359–372.
- Nixon, S. W., J. Ammerman, L. Atkinson, V. Berousnsky, G. Billen, W. Boicourt, W. Boynton, T. Church, D. DiToro, R. Elmgren, J. Garber, A. Giblin, R. Jahnke, N. Owens, M. E. Q. Pilson and S. Seitzinger. 1996. The fate of nitrogen and phosphorus at the land-sea margin of the North Atlantic Ocean. *Biogeochem.*, *35*, 141–180.
- Paillet, J. 1999. Central water vortices of the eastern North Atlantic. *J. Phys. Oceanogr.*, *29*, 2487–2503.
- Paillet, J. and H. Mercier. 1997. An inverse model of the eastern North Atlantic general circulation and thermohaline ventilation. *Deep-Sea Res. I*, *44*, 1293–1328.
- Pollard, R. T., M. J. Griffiths, S. Cunningham, J. F. Read, F. F. Pérez and A. F. Ríos. Vivaldi 1991—A study of the formation, circulation and ventilation of eastern North Atlantic Central Water. 1996. *Progr. Oceanogr.*, *37*, 167–192.
- Prospero, J. M., K. Barrett, T. Church, F. Dentener, R. A. Duce, J. N. Galloway, H. Levy II, J. Moody and P. Quinn. 1996. Atmospheric deposition of nutrients to the North Atlantic Basin. *Biogeochem.*, *35*, 27–73.
- Read, J. F. and W. J. Gould. 1992. Cooling and freshening of the subpolar North Atlantic Ocean since the 1960s. *Nature*, *360*, 55–57.
- Richardson, P. L. 1983. Eddy kinetic energy in the North Atlantic from surface drifters. *J. Geophys. Res.*, *88*, 4355–4367.
- Rintoul, S. R. 1988. Mass, heat and nutrients fluxes in the Atlantic Ocean determined by inverse methods. Ph.D. thesis, MIT-WHOI Joint Program, 287 pp.
- Rintoul, S. R. and C. Wunsch. 1991. Mass, heat, oxygen and nutrient fluxes and budgets in the North Atlantic Ocean. *Deep-Sea Res.*, *38*, 355–377.
- Roach, A. T., K. Aagaard, C. H. Pease, S. A. Salo, T. Weingartner, V. Pavlov and M. Kulakov. 1995.

- Direct measurements of transport and water properties through the Bering Strait. *J. Geophys. Res.*, **100**, 18443–18457.
- Robbins, P. E. and H. L. Bryden. 1994. Direct observations of advective nutrient and oxygen fluxes at 24°N in the Pacific Ocean. *Deep-Sea Res. I*, **41**, 143–168.
- Roemmich, D. and C. Wunsch. 1985. Two transatlantic sections: meridional circulation and heat flux in the subtropical North Atlantic Ocean. *Deep-Sea Res.*, **32**, 619–664.
- Sarmiento, J. L., G. Thiele, R. M. Key and W. S. Moore. 1990. Oxygen and nitrate new production and remineralization in the North Atlantic Subtropical Gyre. *Global Biogeochem. Cycles*, **95**, 18303–18315.
- Sato, O. T. and T. Rossby. 2000. Seasonal and low-frequency variability of the meridional heat flux at 36°N in the North Atlantic. *J. Phys. Oceanogr.*, **30**, 606–621.
- Saunders, P. M. 1982. Circulation in the eastern North Atlantic. *J. Mar. Res.*, **40**, 641–657.
- 1994. The flux of overflow water through the Charlie-Gibbs Fracture Zone. *J. Geophys. Res.*, **99**, 12343–12355.
- Saunders, P. and S. R. Thompson. 1993. Transport, heat, and freshwater fluxes within a diagnostic numerical model (FRAM). *J. Phys. Oceanogr.*, **23**, 452–464.
- Schiller, A. 1995. The mean circulation of the Atlantic Ocean north of 30S determined with the adjoint method applied to an ocean general circulation model. *J. Mar. Res.*, **53**, 453–497.
- Schlitzer, R. 1988. Modelling the nutrient and carbon cycles of the North Atlantic, 1, Circulation, mixing coefficients and heat fluxes. *J. Geophys. Res.*, **93**, 10669–10723.
- 1993. Determining the mean, large-scale circulation of the Atlantic with the adjoint method. *J. Phys. Oceanogr.*, **23**, 1935–1952.
- Schlüter, M., E. J. Sauter, A. Schäfer and W. Ritzrau. 2000. Spatial budget of organic carbon flux to the seafloor of the northern North Atlantic (60°N–80°N). *Global Biogeochem. Cycles*, **14**, 329–340.
- Schmitt, R. W., P. S. Bodgen and C. E. Dorman. 1989. Evaporation minus precipitation and density fluxes for the North Atlantic. *J. Phys. Oceanogr.*, **19**, 1208–1221.
- Schmitz, W. J. Jr. 1996. On the World Ocean Circulation: Volume I, Some Global Features/North Atlantic Circulation. Woods Hole Oceanographic Institution Technical Report WHOI-96-03.
- Schmitz, W. J. and M. S. McCartney. 1993. On the North Atlantic Circulation. *Rev. Geophys.*, **31**, 29–49.
- Schott, F., L. Stramma and J. Fischer. 1999. Interaction of the North Atlantic Current with the deep Charlie Gibbs Fracture Zone throughflow. *Geophys. Res. Lett.*, **26**, 369–372.
- Semtner, A. J. and R. M. Chervin. 1992. Ocean general circulation from a global eddy-resolving model. *J. Geophys. Res.*, **97**, 5493–5550.
- Smethie, W. M. and J. H. Swift. 1989. The Tritium:krypton-85 age of Denmark Strait Overflow Water and Gibbs Fracture Zone Water just south of Denmark Strait. *J. Geophys. Res.*, **94**, 8265–8275.
- Stoll, M. H. C., H. M. van Aken, H. J. W. de Baar and C. J. de Boer. 1996. Meridional carbon dioxide transport in the northern North Atlantic. *Mar. Chem.*, **55**, 205–216.
- Sy, A., M. Rhein, J. Lazier, K. P. Kolterman, J. Meincke, A. Putzka and M. Bersch. 1997. Surprisingly rapid spreading of newly formed intermediate waters across the North Atlantic Ocean. *Nature*, **386**, 675–679.
- Sy, A., U. Schauer and J. Meincke. 1992. The North Atlantic Current and its associated hydrographic structure above and eastwards of the Mid-Atlantic Ridge. *Deep-Sea Res.*, **39**, 825–853.
- Tréguer, P., D. M. Nelson, A. J. van Bennekom, D. J. DeMaster, A. Laynaert and B. Quéguiner. 1995. The silicate balance in the world ocean: A reestimate. *Science*, **268**, 375–379.
- Trenberth, K. E. and A. Solomon. 1994. The global ocean heat balance: heat transports in the atmosphere and ocean. *Clim. Dynam.*, **10**, 107–134.
- Walsh, J. J., K. L. Carder and F. E. Müller-Karger. 1992. Meridional fluxes of dissolved organic matter in the North Atlantic Ocean. *J. Geophys. Res.*, **97**, 15625–15637.

- Wijffels, S. E., R. W. Schmitt, H. L. Bryden and A. Stigebrandt. 1992. Transport of freshwater by the oceans. *J. Phys. Oceanogr.*, 22, 155–162.
- Wollast, R. 1993. Interactions of carbon and nitrogen cycles in the coastal zone, *in* Interactions of C, N, P, and S Biogeochemical Cycles and Global Change, R. Wollast, F. T. Mackenzie and L. Chou, eds., NATO ASI Series, Vol. 14.
- 1998. Evaluation and comparison of the global carbon cycle in the coastal zone and in the open ocean, *in* The Sea, Vol. 14. K. H. Brink and A. R. Robinson, eds.,
- Wunsch, C. 1978. The general circulation of the North Atlantic west of 50°W determined from inverse methods. *Rev. Geophys. Space Phys.*, 10, 251–285.
- 1996. The Ocean Circulation Inverse Problem, Cambridge Univ. Press, 437 pp.
- Wunsch, C., D. Hu and B. Grant. 1983. Mass, heat, salt and nutrient fluxes in the South Pacific Ocean. *J. Phys. Oceanogr.*, 13, 725–753.
- Zaucker, F., T. F. Stocker and W. S. Broecker. 1994. Atmospheric freshwater fluxes and their effect on the global thermohaline circulation. *J. Geophys. Res.*, 99, 12443–12457.
- Zenk, W. and L. Armi. 1990. The complex spreading pattern of Mediterranean Water off the Portuguese continental slope. *Deep-Sea Res. I*, 37, 1805–1823.