Heat and Freshwater Fluxes through the Nordic Seas

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ABSTRACT

The major exchanges of volume, heat, and freshwater between the Arctic Ocean and the World Ocean occur through the Nordic seas. Here is presented the northernmost estimate for the oceanic transport of these properties that is derived from a set of hydrographic and direct current measurements, using a lowered acoustic Doppler current profiler, between the Greenland and European coasts. By applying box inverse methods to a synoptic section from the summer of 1999, a heat transport of $0.20 \pm 0.08$ PW toward the Arctic and a freshwater transport of $0.10 \pm 0.05$ Sv ($1$ Sv = $1 \times 10^6$ m$^3$ s$^{-1}$) away from the Arctic are calculated, with a likely additional freshwater transport on the order of $0.05$ Sv near the Greenland coast. Uncertainties associated with how representative the section is of the seasonal mean are included in the error analysis. Large depth-independent components in the currents throughout the section, including the Atlantic inflow, are observed. The increase (decrease) in the heat transport resulting from an increase (decrease) in the transport of this inflow is $0.033$ PW Sv$^{-1}$, and this is the dominant source of uncertainty in the solution. Therefore, determining only depth-dependent transports is unlikely to be sufficient when measuring heat transport in the region. The overturning components of the section heat and freshwater transport are $0.15 \pm 0.07$ PW and $0.04 \pm 0.02$ Sv, respectively. From the horizontal transport of layers within the section, a densification of $4.3 \pm 2.5$ Sv of waters north of the section is predicted, to densities greater than the boundary between inflow and outflow waters between the Atlantic Ocean and the Nordic seas.

1. Introduction

Vertical mixing in the Nordic seas and the Arctic Ocean is an integral part of the global thermohaline circulation and is dependent on horizontal exchanges of heat and freshwater through the region (e.g., Schlictholz and Houssais 1999a; Aagaard and Carmack 1989). The loss of heat to the atmosphere drives convection at several locations in the region. Because of river runoff and excess precipitation, the Arctic is a major source of freshwater. This is important in a region in which salinity has a large influence in determining density gradients; in many areas of convection, notably the Greenland Sea, surface salinity can be too low for deep convection to occur even when the surface is cooled to the freezing point (Pawlowicz 1995). As a result, exchanges of freshwater between the atmosphere/cryosphere and the ocean are a key component in the thermohaline circulation.

The dominant surface feature of the Nordic seas is a cyclonic circulation; warm, saline water, occupying approximately the upper 700 m of the ocean, flows through the Norwegian Sea in the Norwegian Atlantic Current (NAC) on the eastern boundary, with a flow of cold freshwater away from the Arctic in the East Greenland Current (EGC), which follows the Greenland continental slope (Fig. 1). The Atlantic inflow, which forms the NAC, and other exchanges between the Nordic seas and the North Atlantic Ocean are reviewed in detail by Hansen and Østerhus (2000). The NAC, whose transport varies with the North Atlantic Oscillation (NAO; Mork and Blindheim 2000), loses much of its heat to the atmosphere before flowing toward the Arctic through the Barents Sea and Fram Strait (in the West Spitsbergen Current). Mauritzen (1996a,b) presented a circulation scheme in which densification due to this heat loss was the most important form of deep-water formation in the region. A substantial portion of water in the NAC recirculates in Fram Strait. The remainder continues around the Arctic Ocean and exits the Arctic with the EGC, undergoing large changes in water properties in this circuit. Rudels et al. (1999) review this circulation around the Arctic Mediterranean, a circulation that they describe as the “Arctic Circumpolar Boundary Current.”

The EGC is recognized to have a strong depth-independent component (Woodgate et al. 1999), and recent moored current data indicate that the West Spitsbergen Current does also (Fahrbaeh et al. 2001). Woodgate et al. (1999) observed a transport of the EGC at
Fig. 1. Circulation scheme for the Nordic seas, according to Østerhus et al. (1996), with the position of CATS section 1 and station numbers: 1000-m (light) and 3000-m (heavy) isobaths are shown, as well as warm surface (light), cold surface (dark), and deep (dashed) currents.

75°N that is greater than at Fram Strait (~79°N) or Denmark Strait (~67°N), suggesting that it merges with the cyclonic Greenland Sea gyre. The Greenland Sea gyre has historically been a site of deep convection to near the bottom. It is ideally suited to deep-water formation because of weak stratification, the doming of isopycnals at its center (Rhein 1996), and the formation of sea ice in winter (Pawlowicz 1995). However, convection in the region has decreased since the 1970s (Dickson et al. 1996). Deep water formed in the Greenland Sea is advected to the Norwegian Sea (Blindheim 1990). Relatively saline deep-water inflows through Fram Strait from the Arctic Ocean (Aagaard et al. 1991) also contribute to the deep-water properties.

The deep North Atlantic is ventilated by several outflows from the Nordic seas. A mixture of intermediate water from the Greenland Sea and recirculated Atlantic Water in the EGC forms a major constituent of Denmark Strait Overflow Water (Strass et al. 1993), which is an important source of North Atlantic Deep Water. Waters from the Iceland, Norwegian, and Greenland Seas contribute to overflow waters between Iceland and the Faroes (Hansen and Østerhus 2000; Mauritzen 1996b). The dense outflow is composed largely of a mixture of Norwegian Sea Arctic Intermediate Water and Norwegian Sea Deep Water; a decrease in salinity and density of the outflow since the 1970s suggests that the latter fraction has gradually become much smaller (Turrell et al. 1999).

Although the horizontal exchanges of heat and freshwater in the region are known to be important processes, large-scale estimates of oceanic transports of heat and freshwater north of the Greenland–Scotland Ridge are rare. We use the terms quasi-heat and quasi-freshwater transports to describe calculations in which there is nonzero volume transport; it is common practice to quote these transports relative to a given temperature or salinity, frequently a local mean value. Quasi-heat transports in the NAC in the southern Norwegian Sea have been estimated by several authors (e.g., Pistek and Johnson 1992; Skagseth and Orvik 1999; Mork and Blindheim 2000), and Schlichtolz and Houssais (1999b) estimated both quasi-heat and quasi-freshwater transport in Fram Strait. However, estimates for the breadth of the Nordic seas, through which all of the major exchanges of volume and heat, and most exchanges of freshwater, between the Arctic and the World Ocean occur, have been limited to budget calculations (e.g.,
Here, we calculate the heat and freshwater transport between Norway and Greenland from a 1999 hydrographic section. Simultaneous direct current measurements from a lowered acoustic Doppler current profiler (LADCP) provide a good estimate of the bottom current, and the measured shears are used for comparison with geostrophic shears. Box inverse methods are used to determine transports into and away from the Arctic. The datasets provided by the CTD and the LADCP are presented in section 2. Section 3 contains a description of the inverse model and the choice of constraints that were used. In section 4, we describe the circulation features inferred from the inversion. In section 5, we present heat and freshwater flux estimates and discuss the implications for the heat and freshwater budget of the Arctic region. The conclusions are summarized in section 6.

2. Data

The data discussed in this paper were collected between 26 July and 4 August 1999 during the Circulation and Thermohaline Structure—Mixing, Ice and Ocean Weather (CATS-MIAOW; Bacon and Yelland 2000) expedition of the RRS James Clark Ross, cruise number 44, on CATS section 1. Hydrographic and current data were gathered for 52 stations between 77°18′N, 13°58′W, ~120 km from the Greenland coast, and 64°48′N, 10°07′E, ~35 km from the Norwegian coast (Fig. 1). The accuracy of the CTD was 0.002°C in temperature and 0.002 in salinity (less accurate in regions of large vertical salinity gradients). Current data were collected using an LADCP operating in both water-tracking and bottom-tracking modes. The LADCP data are discussed in section 2b.

a. Hydrography and water masses

At the time of the CATS cruise, the Arctic front was a clearly defined feature in the water properties of the Nordic seas (Fig. 2). Between stations 28 and 29, 38 km apart, there is a decrease in surface temperature of 1.6°C and in surface salinity of 0.12. Figure 3, a θ/S plot of the section, shows that the properties of water masses within the section converge at a potential temperature θ of ~0.8°C and a salinity S of ~34.905, which corresponds to a depth of 1500 m. Above this density level, water to the southeast of the Arctic front is warmer and more saline than water to the northwest, and in the upper ocean the front delineates the boundary between water masses of Atlantic origin and those of Arctic or polar origin. For the purposes of the inversion, water masses are defined by potential density and station pair (Tables 1 and 2).

The 35.0 isohaline, commonly used to define the lower boundary of Atlantic water (AW) in the Norwegian Sea (Hopkins 1988; Orvik et al. 2001), is typically at a depth of 650 m in the Lofoten Basin, corresponding to a value of θ of 3°C, rising to between 250 and 450 m (θ = 4.2°C) over the continental slope (Fig. 2). The salinity reaches a maximum of ~35.31 near the surface at station 15. Over the Norwegian shelf, there is a layer of Norwegian Coastal Water (NCW), freshened by river runoff, with a mean thickness of 75 m (defined with a lower boundary at the 35.0 isohaline) and a minimum salinity of 31.6 at the surface at station 3. The maximum temperature is 12.96°C at the same point.

The ~100-m-thick surface layer in the Greenland Sea is warmer and fresher than the underlying waters, except for near the Greenland shelf, where the surface-intensified EGC transports cold, very fresh, Polar Water (PW) away from the Arctic Ocean, probably with a large contribution from icemelt on this path. The minimum salinity of this layer is 30.0 at the surface, and the minimum temperature is ~1.72°C at a depth of ~70 m. Below PW, Return Atlantic Water (RAW) forms a warm, saline (S > 34.9) layer that has a maximum thickness of ~400 m within 50 km offshore of the shelf break of Greenland.

Very small vertical density gradients exist in Greenland Sea Arctic Intermediate Water (GSAIW), which forms the water mass between ~100 and ~1500 m in the center of the Greenland Sea (considerably thinner near the Greenland shelf) with core properties of S = 34.86 and θ = ~0.8°C. In the Norwegian Sea, the salinity minimum (S ~ 34.89, θ = ~0°C) at ~1000 m (~700 m over the Norwegian continental slope) is Norwegian Sea Arctic Intermediate Water (NSAIW; Hansen and Østerhus 2000), which is probably advected from the Iceland and Greenland Seas (Blindheim 1990). During the early 1980s, Blindheim observed salinity minima in the region that were shallower and more saline than in the CATS section. Consistent with Hopkins (1988) and Hansen and Østerhus (2000), we use the 34.90 isohaline in defining the boundary of the Arctic Intermediate Waters.

We call the water between AW and NSAIW (salinity of 34.9–35.0) Lower Atlantic Water (LAW). This water may have been formed by diapycnal mixing of AW and NSAIW; by recirculation of AW in the Lofoten Basin, similar to that observed by Read and Pollard (1992) in the Norwegian Basin in a water mass that they called Norwegian North Atlantic Water (NNAW); or may be advected NNAW.

The deep 34.90 isohaline is found at a lesser potential density in the Norwegian Sea than in the Greenland Sea (Fig. 3), although the underlying water shows no large-scale variation in properties between the basins. In the Norwegian Sea, this underlying water mass, formed by mixing of Greenland Sea Deep Water (GSDW; the densest water mass in the Nordic seas) and the deep Arctic outflows, is called Norwegian Sea Deep Water (NSDW), even though it is not formed there. In the Greenland Sea, we describe it as Modified Greenland Sea Deep Water (mGSDW), because it is probably formed by the mixing of GSDW with saline inflows from the Arctic...
Fig. 2. Vertical distribution of (a) potential temperature (°C) and (b) salinity (psu) in the CATS section. Shading for temperature is <0° (white), <3° (light), and >3°C (dark). Contour spacing is 0.1°C in the deep-ocean panel and 1°C in the upper panel, except for the dashed -0.5°C contour in the Greenland Basin. Shading for salinity is <34 (white), <34.9 (light), and >34.9 (dark). Contour spacing is 0.01 in the deep-ocean panel and 0.1 in the upper-ocean panel, except for the dashed 34.88 contour in the Greenland Basin.
FIG. 3. Plots of $\theta$-S for 2-dbar CTD data on CATS section 1: (a) all observations more saline than 34.4 psu and colder than 10°C, with contours of potential density relative to the surface; (b) deep water, with contours of potential density relative to 2000 dbar. Gray and black points represent observations to the southeast and northwest of the Arctic front, respectively.

TABLE 1. Isopycnal surfaces for the boundaries between layers used in the inverse model and shown in Fig. 5, with transports toward the Arctic and uncertainties in the inverse solution.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Max density</th>
<th>Transport (Sv)</th>
<th>Uncertainty (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$\sigma_0 = 26.99$</td>
<td>-0.17</td>
<td>0.49</td>
</tr>
<tr>
<td>2</td>
<td>$\sigma_0 = 27.65$</td>
<td>3.28</td>
<td>2.35</td>
</tr>
<tr>
<td>3</td>
<td>$\sigma_0 = 27.879$</td>
<td>2.32</td>
<td>1.58</td>
</tr>
<tr>
<td>4</td>
<td>$\sigma_0 = 28.000$</td>
<td>-1.74</td>
<td>1.11</td>
</tr>
<tr>
<td>5</td>
<td>$\sigma_1 = 32.772$</td>
<td>-2.34</td>
<td>0.85</td>
</tr>
<tr>
<td>6</td>
<td>$\sigma_1 = 32.804$</td>
<td>-1.11</td>
<td>1.33</td>
</tr>
<tr>
<td>7</td>
<td>$\sigma_1 = 37.448$</td>
<td>-0.71</td>
<td>1.14</td>
</tr>
<tr>
<td>8</td>
<td>$\sigma_2 = 41.985$</td>
<td>0.57</td>
<td>2.47</td>
</tr>
<tr>
<td>9</td>
<td>$\sigma_2 = 41.985$</td>
<td>-0.11</td>
<td>2.84</td>
</tr>
</tbody>
</table>

Ocean. The boundary between GSAIW and mGSDW, where the density gradient is dominated by salinity changes, in the Greenland Sea probably marks the limit of winter convection over recent years, whereas the smooth transition in the Norwegian Sea is probably caused by mixing of NSAIW and NSDW over longer timescales. It would be misleading to describe this latter transitional Norwegian Sea water mass as Lower Arctic Intermediate Water, which is identified by a maximum in salinity and temperature in the southern Greenland Sea and parts of the Norwegian Sea (Swift and Aagaard 1981) that we do not observe; we describe it as Upper Norwegian Sea Deep Water (uNSDW).
Table 2. Water masses in the section, defined by potential density and station pair [typical values of $S$ (psu) and $\theta$ (°C) at the boundaries are included as a guide], with transports toward the Arctic and uncertainties in the inverse solution. Water mass 11 includes water masses 11a and 11b, and so the volume of the subdivided water masses are given in parentheses. The uncertainty in the transport of AW is given in parentheses because the transport of AW + NCW is constrained in the sensitivity study.

<table>
<thead>
<tr>
<th>Water mass</th>
<th>Layers</th>
<th>Stations</th>
<th>Description</th>
<th>% volume</th>
<th>Transport (Sv)</th>
<th>Uncertainty (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 NCW</td>
<td>1</td>
<td>3–10</td>
<td>Runoff-freshened Norwegian coastal water, $S &lt; 35.0$</td>
<td>0.28</td>
<td>0.59</td>
<td>0.28</td>
</tr>
<tr>
<td>2 AW</td>
<td>1</td>
<td>10–29</td>
<td>Atlantic Water in Norwegian Sea, $S &gt; 35.0$ (°C &gt; 3)</td>
<td>12.1</td>
<td>6.93</td>
<td>(2.20)</td>
</tr>
<tr>
<td>3 LAW</td>
<td>4</td>
<td>3–29</td>
<td>Lower Atlantic Water, properties between those of AW and NSAIW ($34.9 &lt; S &lt; 35.0$)</td>
<td>4.2</td>
<td>1.33</td>
<td>0.98</td>
</tr>
<tr>
<td>4 GSW</td>
<td>2–4</td>
<td>29–39</td>
<td>Greenland (Sea) Surface Water</td>
<td>1.6</td>
<td>−0.14</td>
<td>0.91</td>
</tr>
<tr>
<td>5 PW</td>
<td>1–3</td>
<td>39–54</td>
<td>Polar Water in Greenland Sea, $S &lt; 34.7$</td>
<td>1.6</td>
<td>−2.18</td>
<td>1.08</td>
</tr>
<tr>
<td>6 RAW</td>
<td>4</td>
<td>39–54</td>
<td>Return Atlantic Water in Greenland Sea, $S &gt; 34.9$ (lower boundary)</td>
<td>1.5</td>
<td>−2.84</td>
<td>0.98</td>
</tr>
<tr>
<td>7 NSAIW</td>
<td>5</td>
<td>9–29</td>
<td>Norwegian Sea Arctic Intermediate Water, $S &lt; 34.9$ (0.6 $\theta &gt; -0.5$)</td>
<td>5.2</td>
<td>0.83</td>
<td>1.31</td>
</tr>
<tr>
<td>8 GSAIW</td>
<td>5–6</td>
<td>29–49</td>
<td>Greenland Sea Arctic Intermediate Water, $S &lt; 34.9$</td>
<td>20.3</td>
<td>−5.53</td>
<td>4.11</td>
</tr>
<tr>
<td>9 uNSDW</td>
<td>6</td>
<td>9–29</td>
<td>Mixture of NSAIW and NSDW (upper NSDW)</td>
<td>7.8</td>
<td>1.25</td>
<td>1.99</td>
</tr>
<tr>
<td>10 NSDW</td>
<td>7, 8</td>
<td>9–29</td>
<td>Norwegian Sea Deep Water, $S = 34.908$ ($-0.7 &lt; \theta &lt; -1.0$)</td>
<td>20.4</td>
<td>3.66</td>
<td>5.38</td>
</tr>
<tr>
<td>11 mGSDW</td>
<td>7, 8</td>
<td>29–49</td>
<td>Modified Greenland Sea Deep Water, properties identical to NSDW, mixture of GSDW and Arctic outflows</td>
<td>15.5</td>
<td>−3.79</td>
<td>3.71</td>
</tr>
<tr>
<td>11a mGSDWc</td>
<td>7</td>
<td>42, 44, 45*</td>
<td>High-salinity mGSDW due to Canadian Basin Deep Water, $S \sim 0.004$ greater than ambient water</td>
<td>(0.28)</td>
<td>−0.68</td>
<td>0.35</td>
</tr>
<tr>
<td>11b mGSDWc</td>
<td>7, 8</td>
<td>30–36</td>
<td>High-salinity mGSDW due to Eurasian Basin Deep Water, $S \sim 0.001$ greater than ambient water</td>
<td>(6.3)</td>
<td>−3.14</td>
<td>3.82</td>
</tr>
<tr>
<td>12 GSDW</td>
<td>9</td>
<td>29–49</td>
<td>Greenland Sea Deep Water, $S &lt; 34.908$ (°C &lt; −1.0)</td>
<td>9.6</td>
<td>−0.11</td>
<td>2.84</td>
</tr>
</tbody>
</table>

* Note that station 42 is northwest of station 43.

All water within the density range of mGSDW/NSDW (Tables 1 and 2) has a salinity between 34.906 and 34.913 (typically 34.908), except for over the continental slopes. Even over the slopes, no water meets the definition of $S > 34.92$ used by Schlichtholz and Houssais (1999b) for Canadian Basin Deep Water (CBDW) and Eurasian Basin Deep Water (EBDW), the saline outflows from the Arctic Ocean, and by Dickson et al. (1996) for Arctic Ocean Deep Water (AODW; the generic term for these water masses). However, their influence is observed; there is a core of high salinity close to the seafloor at 1300–2500 m on the Greenland Slope and a more diffuse region between 1500 m and the bottom over the northwest side of Mohn Ridge. The former of this is found in the lighter part of the NSDW range and is associated with a salinity enhanced by typically 0.004 (up to 0.01) relative to the ambient water. The salinity maximum, CBDW, has been observed before (Aagaard et al. 1991). The high salinity signal over Mohn Ridge covers a much greater area, but the enhancement in salinity is typically 0.001, less than the accuracy of the salinometer. It is unlikely that this water mass has recirculated in the Greenland Sea gyre because the deep currents in the region were southward (Fig. 4) and the water mass reaches to denser layers than CBDW; there is most likely an outflow of EBDW from the adjacent Arctic Ocean over Mohn Ridge. We term these water masses mGSDWc (mGSDW Canadian) and mGSDWc (mGSDW Eurasian), respectively (Table 2).

The densest water mass in the section is GSDW, found below ~2500 m in the Greenland Sea (Fig. 2), which has been undergoing change since the 1950s. Bönisch et al. (1997) described a change from the classical definition of $-1.29 < \theta < -1.0^\circ C$ and $34.88 < S < 34.90$ to $\theta = -1.149^\circ C$ and $S = 34.899$ in 1994 (mean between 2000 m and bottom), with cooling in the 1970s and warming afterward. We observe a continuation of this increase in temperature and salinity, with bottom properties of $\theta = -1.15^\circ C$ and $S = 34.90$ and a mean in GSDW (as defined in Tables 1 and 2 and shown in Fig. 2) of $\theta = -1.109^\circ C$ and $S = 34.904$.

b. LADCP data

The LADCP continuously gathered short overlapping profiles, and the data were processed using software...
developed at the University of Hawaii (Firing et al. 1998). Velocities introduced by the motion of the package were removed by differentiating velocities. The resulting water-tracking shear profiles were integrated over the cast to yield upcast and downcast velocities sorted into 20-m-depth-bin profiles (Bacon and Yelland 2000). The upcast and downcast profiles were averaged to give the final depth-dependent profile. [We use the terms “depth-independent” and “depth-dependent,” rather than “barotropic” and “baroclinic,” to describe the bottom velocity (or transport for a barotropic water column moving at the bottom velocity) and the depth-dependent velocity (or transport) relative to the bottom, respectively.]

The depth-independent velocity at each station was the mean velocity, measured both from the upcasts and the downcasts, in bottom-tracking mode; the length of these profiles was typically 200 m, with 16-m-depth bins. The tidal current, obtained from the Arctic Ocean tide model of Kowalik and Proshutinsky (1994), was then removed. The amplitude of the tidal current in the model was typically 3 cm s$^{-1}$ near the Norwegian coast, decreasing to 0.5 cm s$^{-1}$ in the deep water and 1.5 cm s$^{-1}$ near the Greenland coast. Woodgate et al. (1999) observed an amplitude of 4 cm s$^{-1}$ near the Greenland coast, suggesting that the model underestimates velocities here. However, the effect of the tidal correction on the results of the inversion is minimal (section 5).

Figure 4 shows the cross-section (northeastward) and along-section (southeastward) tide-corrected LADCP velocity; shading represents northeastward and southeastward flow in the respective panels. Contours are at 5 cm s$^{-1}$ intervals.

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**Figure 4.** Vertical distribution of (a) cross-sectional and (b) along-sectional tide-corrected LADCP velocity; shading represents northeastward and southeastward flow in the respective panels. Contours are at 5 cm s$^{-1}$ intervals.
velocity field. Circulation southeast of the Arctic front is complex, with northward flow near the coast (stations 3–6), between the shelf break and approximately the 1500-m isobath (stations 10–12), and offshore of the foot of the Norwegian continental slope (stations 19–21), and a return southward flow between these features. This produces a similar circulation pattern on the slope to that observed by Orvik et al. (2001). Throughout most of this region, sloping isopycnals (Fig. 5) imply geostrophic shear of the same sign as the LADCP field. However, the depth-independent component dominates over the depth-dependent component. Exceptions are at the foot of the Norwegian continental slope, where the deep current is toward the Arctic, and the upper-ocean flow is away from the Arctic (Fig. 4a). Within the Lofoten Basin, there is a strong cyclonic circulation. This circulation does not appear to be associated with topography, at least on its northwestern boundary, and from this synoptic section it is uncertain whether it is a transient or a permanent feature.

Over Mohn Ridge, at the Arctic front (stations 27–30), there is an eastward surface jet (Fig. 4); below ~1000 m, flow is southward. In the Greenland Basin (stations 36–44), a generally cyclonic circulation is observed, existing at all depths except the bottom 1000 m in the deepest part of the Greenland Basin, where there is southward flow over much of the basin. The core of the southward-flowing EGC is on the continental slope, between the shelf break and the 3000-m isobath (stations 42–47). The mean cross-section velocity over the water column here is 11 cm s$^{-1}$, reaching a maximum of 25 cm s$^{-1}$ near the surface. At most stations on the Greenland continental slope, there are large shears near the bottom, where flow is weaker or rotated anticlockwise, possibly because of bottom Ekman forcing. Currents over the Greenland shelf (stations 49–54) are variable, but the mean flow is southward.

### 3. The box model inversion

#### a. Application of the inverse method

The CATS section and the Arctic landmasses form a closed box, except for known transports through the Bering Strait and the Canadian Archipelago. It is initially assumed that the circulation is representative of the summer mean state and that the depth-dependent velocity field is given by geostrophic balance, determined from the CTD data. This is preferable to the direct use of LADCP data, which is likely to be affected by high-frequency ageostrophic variability. There is a small contribution from Ekman transport in the layer in contact with the surface, derived from the Hellerman and Rosenstein (1983) annual mean wind stress climatological dataset. For a conservative property in a closed box, the depth-dependent property transport field $\mathbf{d}$ is given by

$$\mathbf{A}\mathbf{b} = \mathbf{d},$$

where $\mathbf{A}$ is a matrix of the area $\times$ property concentration field and $\mathbf{b}$ is the depth-independent velocity vector. The unknown in this equation, the depth-independent velocity vector, is solved for through singular value decomposition (Wunsch 1996). The data were inverted using version 4.2 of Dobox (Morgan 1995), software previously used by, for example, Sloyan and Rintoul (2000) and Naveira Garabato et al. (2002). The model contains nine layers (Table 1), 51 columns (the 51 station pairs), and 51 unknowns (the depth-independent velocity at each station pair). The transport in bottom triangles, where adjacent stations extend to differing depth, is estimated by extrapolating to the bottom the properties at the deepest common depth. Layer transports are not conserved, because it would be unreasonable to assume zero net transport between layers in a highly convective region. As a result, the full rank solution was used.
FIG. 6. Initial velocity field in the inversion; shading represents northeastward flow. Contours are at 5 cm s$^{-1}$ intervals. Geostrophy is used to determine the depth-dependent component of the field, and the depth-independent component is derived from LADCP data.

Because the problem is highly underdetermined, the final depth-independent velocity field necessarily depends greatly on the initial estimate of the depth-independent velocity field. This was derived from the bottom-tracked LADCP velocity for each station pair, determined by averaging the mean bottom-tracked velocity of the two stations. For each station pair, the bottom velocity was further adjusted to produce a profile that best represents the entire water column. This was achieved by finding the offset that provided a geostrophic profile with best least squares fit to the water-tracked LADCP shear. The upper 200 m of the water column were ignored in this adjustment because of surface ocean ageostrophic processes. The initial velocity field is shown in Fig. 6.

The bottom velocity at each station pair was assigned an uncertainty. Higher uncertainties (lower weighting) were attributed to shallow station pairs, which have greater tidal components and fewer observations in the vertical because of shorter profiles. Higher uncertainties were also allocated where there were significant differences between geostrophic and cross-section LADCP vertical shears, particularly if these differences resulted in a large adjustment to the depth-independent current, and where there were significant differences between the LADCP profiles at the two adjacent stations. The ascribed uncertainty was approximately 2 times as great for stations with the greatest uncertainty as for those with the lowest uncertainty.

b. Constraints

Because we are inferring seasonal transports from a one-time hydrographic section that is likely to be affected by high-frequency variability, it is necessary to ensure, as far as is possible, that a seasonally representative circulation is derived from the inverse model. Therefore, known time-mean circulation features are constrained. The three constraints applied are the conservation of volume, the transport of AW + NCW, and the transport of the EGC. The latter two constraints are not strict a priori constraints but are applied only for the error analysis, as discussed below.

1) NET VOLUME TRANSPORT ACROSS THE CATS SECTION

In the Nordic seas there is high convective activity. It is therefore inappropriate to apply a constraint of zero or near-zero net cross-sectional mass transport within density layers that are remote from the surface and/or are enclosed by bathymetry, as is common in regions where diapycnal fluxes can be more easily estimated (e.g., Macdonald 1998; Ganachaud et al. 2000). We instead apply full-depth volume conservation, with net transport toward the Arctic across the CATS section equal to the net transport away from the Arctic through the Bering Strait and the Canadian Archipelago. Although this is true in the long term if the Arctic is in equilibrium, Ganachaud (1999) found that volume is not conserved over short timescales, because of large baroclinic variability [5 Sv (1 Sv = 10$^6$ m$^3$ s$^{-1}$) for a general circulation model simulation at 36$^\circ$N in the Atlantic Ocean]. Therefore a large uncertainty must be attached to this constraint.

Hopkins (1988) estimated the Bering Strait inflow toward the Arctic at 1.5 Sv and the Canadian Archipelago outflow from the Arctic at 2.1 Sv. Coachman and Aagaard (1988) observed a mean Bering Strait inflow of 0.8 Sv (with a 0.3-Sv annual cycle peaking in
summer), a result used in the box-model study of Robitaille et al. (1995) and the inversion of Bacon (1997). Roach et al. (1995) observed a mean Bering Strait transport of 0.83 Sv between 1990 and 1994 and a peak transport of over 1.3 Sv in July. Rudels (1986) obtained a net outflow through the Canadian Archipelago of 1.0 Sv. Aagaard and Carmack (1989) and Bacon (1997) used a value for the Canadian Archipelago outflow of 1.7 Sv in budget calculations for the Arctic, but Bacon suggested that this may be an overestimate. The net mean transport toward the Arctic is therefore likely to be close to zero, and the uncertainty in the value is small when compared with the uncertainty in net volume conservation on short timescales.

We assign a net volume transport toward the Arctic of 0 ± 5 Sv. In the initial velocity field (Fig. 6), there is a net transport of 1.4 Sv away from the Arctic, so the constraint of zero net transport in the inversion results in a mean modification in cross-section velocity of 0.04 cm s⁻¹ toward the Arctic. The estimate of uncertainty due to high-frequency variability, assumed to be equal to that estimated by Ganachaud (1999) at 36°N, is probably pessimistic because the Nordic seas are a much smaller basin than the North Atlantic. Conservation of volume is preferable to conservation of salt because of significant variability in salinity within and between the Bering Strait and the Canadian Arctic Archipelago. Furthermore, the volume gain north of the CATS section due to river runoff and excess precipitation, estimated as 0.13 Sv for the Arctic and 0.04 Sv for the Nordic seas (Aagaard and Carmack 1989), is small relative to the uncertainty. We then calculate freshwater transport away from the Arctic through the Nordic seas as the amount of freshwater necessary to produce the apparent salt transport toward the Arctic.

2) The inflow of Atlantic water through the Norwegian Sea

The closest historical section in the Norwegian Sea to the CATS section is the Svinøy section, which runs from 62°N on the Norwegian coast to 64°40'N, 0° at a depth of ~2800 m (Mork and Blindheim 2000) and which crosses the NAC. Using data from 1978 to 1996, Mork and Blindheim found that the summer geostrophic transport, referenced to 1000 m, is positively correlated with the February–April NAO index. The appropriate Gibraltar–Iceland NAO index for the CATS cruise in August of 1999 was 0.50 (Jones et al. 1997, updated to present by T. J. Osborn, Climate Research Unit, University of East Anglia), a midphase value, so the inflow is unlikely to have been in an extreme state in 1999. The mean summer transport calculated by Mork and Blindheim was 5.0 Sv, with most of the northward transport in two cores above the 600–800- and 1400–2500-m isobaths, respectively. Although this estimate includes water masses that do not fall within our definition of AW, the contribution of such water masses is probably small. However, whereas they only calculate the depth-dependent transport, only the depth-independent transport can be modified to satisfy the constraints in inverse models. Furthermore, because the CATS LADCP data show strong depth-independent transport toward the Arctic throughout the southeastern Lofoten Basin and over the continental slope (Fig. 4), the total transport of the Atlantic inflow is probably significantly greater than 5 Sv. By combining current-meter observations with hydrographic data, Orvik et al. (2001) obtained an annual transport for the shallower Svinøy branch of 4.2 Sv, with a summer minimum of 2.0 Sv, and suggest a yearly mean Atlantic inflow of 7.6 Sv; however, direct current observations are not available from the length of the Svinøy section.

Comparable estimates of the Atlantic inflow exist between Iceland and Scotland, which has been extensively investigated and reviewed, most recently by Hansen and Østerhus (2000). This is not an ideal location for comparison, because AW mixes with other water masses in the Norwegian Sea, as well as undergoing exchanges with the atmosphere, leading to ambiguity in the definition of AW. Furthermore, there may be significant exchange of AW across the Arctic Front between Iceland and the CATS section (Fig. 1); the Denmark Strait inflow passes to the north of Iceland and into the Norwegian Sea, and a branch of the Atlantic inflow passes to the south of the section. However, we would expect to obtain a broadly similar volume transport for the inflow. The historical Atlantic inflow estimates reviewed by Hopkins (1988) range from 2.0 to 4.5 Sv, apart from an estimate of 8.0 Sv by Worthington (1970), based on budget calculations. However, many of these estimates did not cover the entire inflow or were otherwise limited, and Hopkins proposed an inflow between Iceland and Scotland of 8.35 Sv, plus a transport of 0.6 Sv through the Denmark Strait. In a box-model study of the Arctic Mediterranean, Robitaille et al. (1995) used the estimates of Aagaard and Carmack (1989) to prescribe a transport of AW of 6.1 Sv in the Norwegian Sea. Hansen and Østerhus (2000), considering various results but ignoring purely geostrophic calculations, gave a “preliminary estimate” of 7.0 Sv, plus 1 Sv through Denmark Strait.

The value of 6.7 Sv for the transport of Atlantic water in the CATS section in the initial velocity field, subject only to the zero volume flux constraint, is in close agreement and within the range of these estimates. This is slightly increased to 6.9 Sv by the zero volume transport constraint. The additional 0.6 Sv of runoff-freshened water that we observe in the Norwegian Coastal Current (NCC) compares well to the value imposed by Robitaille et al. (1995) of 0.7 Sv. We therefore constrain the transport of AW plus the NCC to be unchanged from the solution with zero net transport (~7.5 Sv). This has no effect on the best estimate but is relevant in the error analysis. We ascribe an uncertainty of ±2.2 Sv in this value. The range in the above estimates is for the inflow...
over the Iceland–Scotland Ridge; further uncertainties due to exchanges within the Nordic seas and variability in the inflow are considered in this uncertainty estimate.

3) **The East Greenland Current**

The EGC transports polar waters along the Greenland continental slope and into the Atlantic. As with the Atlantic inflow, historical transport estimates, reviewed by Hopkins (1988), were small (on the order of 2 Sv) because of assumptions made about the depth-independent component or the exclusion of deeper water, except one estimate of 31.5 Sv by Worthington (1970). Schlichtholz and Houssais (1999a) found a geostrophic depth-dependent transport of 5 Sv at 79°N from an inverse modeling study of Fram Strait. Fahrbach et al. (2001), using 2 yr of current-meter observations at the same latitude, obtained a mean total transport of the EGC of 13.7 ± 1.7 Sv. At the latitude of the CATS section, there is southward transport in the Greenland Sea gyre adjacent to the EGC, which results in a greater southward transport near the Greenland shelf than in Fram Strait.

The EGC crosses the CATS section at 75°–76°N, close to the moored current array at ~75°N discussed by Woodgate et al. (1999). For their best estimate of the eastward extent of the EGC (~40 km east of the mooring located at the foot of the continental slope) they obtained a mean for 1994–95 of 21 Sv, with a minimum of 11 Sv in September and a typical summer value of ~15 Sv. Choosing a location for the eastward boundary of the EGC 40 km farther east or 29 km farther west resulted in a transport that was 4 Sv greater or 5 Sv smaller, respectively. They also used current-meter data from 1987 to 1994 to estimate a mean for these years, excluding the easternmost 30 km, and in 1994–95 the EGC was slightly weak.

The uncertainty associated with locating the eastern boundary of the EGC in the CATS section is even greater than in locating the boundary in the moored arrays used by Woodgate et al. (1999). Station 41 is ~60 km southeast of the foot of the continental slope (the CATS section was perpendicular to the continental slope). This location for the boundary provides an estimate that is as consistent with that of Woodgate et al. (1999) as is possible. Applying the zero volume transport constraint only, the transport of the EGC northwest of this boundary is 16.4 Sv. Using station 42 as the boundary, at the foot of the slope (station 42 is northwest of station 43), the transport is 9.8 Sv. Using station 40, the limit of southward transport in the Greenland Sea and 100 km southeast of the foot of the slope, the transport is 24.9 Sv. The choice of the western boundary is of lesser importance, because the transport over the shelf is small (1.2 Sv), and station 54 (the westernmost station) is used throughout this paper.

The injection of sulfur hexafluoride (SF₆) into the Greenland Sea on the σ₀ = 28.049 (σ₀ is potential density minus 1000 kg m⁻³, referenced to n-thousand-dbar pressure surface) surface in August of 1996 (Watson et al. 1999) provides an alternative method of tracking the boundary of the EGC with the Greenland Sea gyre. Figure 7 shows the vertical distribution of SF₆ observed on the CATS section (M.-J. Messias and A. J. Watson 2001, personal communication). There is a gradual decrease in SF₆ concentration between station 41, near the gyre center, and station 42, 60 km to the northwest. There is an abrupt drop in column-integrated SF₆ concentration (centered at station 44) between stations 42 and 45, a distance of 20 km apart. (Because there are sloping isopycnals at the location, the sharp horizontal density gradient exaggerates this feature in the figure.) In terms of the source of the waters, this is the most appropriate location for the boundary between the EGC and the Greenland Sea gyre (for this season), although there is significant isopycncal mixing between the two currents.

However, it is desirable to ensure that the EGC transport in the CATS section is consistent with the seasonally averaged estimate of Woodgate et al. (1999). We therefore define the EGC with an eastern boundary at station 41 when constraining the inverse model. Because the uncertainties associated with the eastern boundary definition are much greater than the difference between
the initial CATS transport and that of Woodgate et al., it would be inappropriate to constrain transport in the inversion to equal that measured by Woodgate et al. Instead, we apply the same approach used in constraining the Atlantic inflow and constrain the EGC to be unchanged from the zero-net-transport solution ($\sim 16.4 \text{ Sv}$). Quantifying an uncertainty is problematic, because past estimates at Fram Strait and Denmark Strait are of limited value here and the main sources of uncertainty are due to the definition of the boundary and the choice of a seasonally representative value; a large uncertainty of $\pm 6 \text{ Sv}$ is chosen.

c. Method of error analysis

The effects of two kinds of uncertainties in the flux estimates are investigated: 1) uncertainties in the initial velocity field and 2) uncertainties in the inversion. Because observed currents may not represent the mean flow, random depth-independent cross-sectional currents with a mean velocity of zero and a standard deviation of $5 \text{ cm s}^{-1}$ were added to each station pair to form alternative initial fields. The large value of $\pm 5 \text{ cm s}^{-1}$ was chosen because large high-frequency variability has been observed in the NAC (Orvik et al. 2001) and in the EGC (Woodgate et al. 1999), but it may be an overestimate for other parts of the section. The standard error in these outputs was used as the uncertainty in the initial depth-independent velocity field. The effects of doubling or removing the tidal correction and of excluding the surface Ekman component from the inversion were also investigated but were small.

Uncertainties in the output because of uncertainties in the constraints were investigated by constraining the model with adjusted values for each constraint, in turn, at the limit of its uncertainty (Table 3). The effect of uncertainties in the station pair weighting on the results of the inversion was investigated by removing the weighting and was found to be negligible. The total uncertainty is the quadratic sum of all the above uncertainties and is given for volume transports in Tables 1 and 2 and for heat and freshwater transports in Table 3.

4. Circulation features deduced from the inversion

Figure 8 shows the column-integrated transport in the best estimate from the inversion. Table 3 shows the transport, with uncertainties, of the layers and water masses. In both the Greenland and Lofoten Basins, there is cyclonic circulation. The transport of the Greenland Sea gyre, defined as the transport toward the Arctic between stations 40 and 35, is $19 \pm 10 \text{ Sv}$, but the true transport of the gyre might be greater because the section may not have crossed its center. Because net transport is away from the Arctic between station 35 and the Arctic front (station 29), this limb of the gyre existed over a smaller area than that suggested by the climatological circulation scheme in Fig. 1. There is a similar transport of the limb flowing away from the Arctic, $17 \pm 7 \text{ Sv}$, if its boundary is at station 44, derived from the SF$_6$ distribution. The transport of the EGC with its boundary at this station is $7.6 \pm 3.8 \text{ Sv}$. The transport of the cyclonic circulation in the Lofoten Basin, defined as the transport away from the Arctic between stations 21 and 23, is $15 \pm 11 \text{ Sv}$. There is also a cyclonic circulation on a larger scale; the total transport southeast of the Arctic front (stations 3–29) is $15 \pm 6 \text{ Sv}$ toward the Arctic, with an equal return transport northeast of the Arctic front. The large uncertainties in these transports are mostly due to the uncertainty of $5 \text{ cm s}^{-1}$ ascribed to the initial depth-independent field due to high-frequency variability from the seasonal mean. Therefore, they represent uncertainties in the long-term transport rather than in the observed synoptic transports.

Because there is zero net volume transport across the section, the net transport of upper-ocean waters toward
the Arctic must be equal to the net transport of deep waters away from the Arctic. This provides an estimate of the rate of densification to the north of the section. A frequently used definition for overflow water into the Atlantic is $\sigma_0 > 27.8$ (Dickson and Brown 1994; Hansen and Østerhus 2000). The net transport of water above this level and, therefore, the densification of water to below this level in the Arctic and northern Nordic seas are $4.3 \pm 2.5$ Sv, slightly lower than current estimates for the overflow into the Atlantic, such as $5.6$ (Dickson and Brown 1994) and $6$ (Hansen and Østerhus 2000) Sv. This may be due to densification south of the CATS section in the Atlantic inflow; the lower boundary of Atlantic Water in the section is $\sigma_0 = 27.879$. If this is nominated as the boundary, $5.4 \pm 2.5$ Sv is converted. This is approximately the maximum in the overturning streamfunction.

The net transport away from the Arctic in layers 7–9, $0.2 \pm 2.2$ Sv, is indistinguishable from zero. The top of layer 7 ($\sigma_1 = 32.804$) is, at present, approximately the maximum density of overflow water through the Faroe–Shetland Channel (Turrell et al. 1999) and denser than Denmark Strait Overflow Water (Strass et al. 1993). Therefore, any such transport would be balanced by a net diapycnal transport between the CATS section and the Greenland–Scotland Ridge.

The schematic in Fig. 1, adapted from Østerhus et al. (1996), suggests that the Atlantic inflow splits into four branches before reaching the CATS section, one of which circulates cyclonically to the south of the section.
Fig. 9. Cumulative section (a) quasi-heat and (b) quasi-freshwater transport, relative to the section mean properties (temperature and salinity), summed from zero at the Greenland coast. The solid line shows total transport, and the dashed line shows the depth-dependent part of the flow.

Candidates for the other three branches may be identified in the output from the inversion: 1) above the Norwegian continental slope, 2) in the Lofoten Basin, and 3) above the Mohn Ridge slope. The total transport of AW in branch 1 (inshore of station 15) is 5.3 Sv, in branch 2 (between stations 18 and 21) it is 4.3 Sv, and in branch 3 (between stations 24 and 29) it is 5.0 Sv. These branches are labeled in Fig. 8. Jets of recirculating water to the northwest of branch 1 and branch 2, with transports of 3.0 and 4.7 Sv, respectively, result in the net transport of AW toward the Arctic, defined as \( \sigma_0 < 27.88 \), of 6.9 Sv.

5. Heat and freshwater fluxes

a. Solutions from the inversion

Net heat and freshwater fluxes are calculated for zero net volume transport across the section using the temperature and salinity fields, the velocity field in the inverse solution, and mean values for in situ density and specific heat capacity of 1034.5 kg m\(^{-3}\) and 3990 J kg\(^{-1}\), respectively. The calculated net heat flux was 0.20 ± 0.08 PW toward the Arctic, and the freshwater flux was 0.10 ± 0.05 Sv away from the Arctic.

Figure 9 shows the cumulative quasi-heat and quasi-
freshwater transports from Greenland to Norway. Because cumulative volume transport is nonzero except at the Norwegian coast, transports are given relative to the section mean properties (0.270°C, 34.905). Also shown are the depth-dependent components of the heat and freshwater transport, totaling 0.09 PW and 0.06 Sv, respectively. Therefore, in the Nordic seas, calculations based only on the depth-dependent component are likely to yield an underestimate of heat and freshwater transport.

The Norwegian Coastal Current (stations 3–10) and the branch of the Atlantic inflow following the Norwegian slope (stations 10–15), both of which have large depth-independent components, are important in the poleward transport of heat, with a combined quasi-heat transport of 0.18 PW. The depth-dependent component dominates near the Arctic front (0.10 PW at stations 23–29), where there is a flow reversal between warm surface waters and cold, southward-flowing, deep waters. There are large quasi-heat transports in the Greenland Sea gyre: 0.07 PW in the southward-flowing limb (stations 40–44) and 0.09 PW away from the Arctic in the northward-flowing limb (stations 36–40). Therefore, the Greenland Sea gyre transports a small net amount of heat (0.02 PW) away from the Arctic.

The majority of the quasi-freshwater transport (0.07 Sv, relative to the section mean salinity) occurs between stations 50 and 54 on the Greenland shelf. Although volume transports away from the Arctic in this area are small, the extremely low salinity (between 30 and 34) of the surface water results in a large southward freshwater transport. In addition, there is probably significant freshwater transport inshore of station 54. It is likely that volume transport decreases in the shallower water near the coast but that there are lower salinities because of glacial meltwater. If the mean quasi-freshwater transport per unit of distance across the shelf is extrapolated into the region between the westernmost end of the section and the coast, there is an additional quasi-freshwater transport of 0.05 Sv, resulting in a net freshwater flux away from the Arctic of 0.15 Sv. Because we observe poleward transport of freshwater near the Norwegian coast, this equatorward transport is probably partly negated by the Norwegian Coastal Current inshore of station 3.

Table 2 shows the relative importance of various sources of uncertainty in the total uncertainty of 0.08 PW in heat transport and 0.05 Sv in freshwater transport. The sensitivity of the heat transport to the imposed transport of the Atlantic inflow plus the NCC is linear, equal to 0.033 PW Sv⁻¹, and dominates other uncertainties. The freshwater flux is relatively insensitive to the value of the inflow. An uncertainty of 6 Sv in the EGC results in an uncertainty of only 0.02 Sv in the freshwater transport. Uncertainties in the representation of the seasonal mean circulation by the initial field are the greatest source of uncertainty in the freshwater flux (0.04 Sv). An uncertainty of 5 Sv in the net volume transport has a small but nonnegligible effect on net freshwater flux.

Because there is no net volume transport, the net heat flux can be broken down into overturning (baroclinic) and horizontal (gyre) components by the equation

\[
\int_{-h}^{0} \int_{0}^{-H} v(x, z)H(x, z) \, dx \, dz = \int_{-h}^{0} \int_{0}^{-H} \bar{v}(z)\bar{H}(z)L(z) \, dz + \int_{-h}^{0} \int_{0}^{-H} v'(x, z)H'(x, z) \, dx \, dz,
\]

where \(v\) is the velocity; \(H\) is the product of temperature, specific heat capacity, and density; overbars indicate averages over a given density surface; \(L\) is the horizontal length of that surface; \(h\) is the mean bottom depth of a given station pair, \(v' = v - \bar{v}\); and \(H' = H - \bar{H}\) (Bryden and Imawaki 2001). The freshwater flux can similarly be separated into components. Choosing potential density as the \(z\) coordinate, we obtain overturning and horizontal components of 0.15 ± 0.07 and 0.05 ± 0.02 PW, respectively, for heat transport and 0.04 ± 0.02 and 0.07 ± 0.04 Sv, respectively, for freshwater transport. (Where nonzero volume transport is introduced in the error analysis, we ensure that there is no contribution from this transport by using the temperature or salinity relative to the section mean value.)

b. Comparison with previous estimates

(Quasi-)heat transport estimates for the entrance to the Nordic seas have been made by several authors. Hopkins (1988) combined estimates for the channels between the Atlantic Ocean and the Arctic Mediterranean, including small contributions from the Bering Strait and the Canadian Archipelago, and obtained a total quasi-heat transport of 78.9 Sv °C (0.32 PW) toward the Arctic. A quasi-heat transport of 0.28 PW (66.6 Sv °C) was due to the Atlantic inflows between Iceland and Scotland. Simonsen and Haugen (1996) suggested that the preferred heat transport estimate is 0.30 PW between Greenland and Scotland, of which 0.22–0.25 PW is lost to the atmosphere in the Nordic seas, leaving 0.05–0.08 PW to enter the Arctic. Bacon (1997), by the
inversion of two hydrographic sections between Greenland at 60°N and Ireland (−54°N) and with the use of ADCP current data in the EGC, obtained a value of 0.28 ± 0.06 PW. Defining Atlantic Water as having a salinity greater than 35.1, Hansen et al. (1999) obtained a quasi-heat transport of 0.11 and 0.13 PW for the Iceland–Faroe Ridge and the Faroe–Shetland Channel, respectively, by use of direct hydrographic and current measurements. Hansen and Østerhus (2000) updated the estimate for transport between the Faroes and Shetland to 0.14 PW. There is also a small contribution from Irminer Atlantic Water through Denmark Strait, estimated to be 0.015 PW by Hopkins (1988) and 0.025 PW by Hansen and Østerhus (2000).

Farther north, in the Svinøy section (~62°–65°N), Mork and Blindheim (2000) obtained an average quasi-heat transport of the Norwegian Atlantic Current of 0.135 PW between 1978 and 1996. Pistek and Johnson (1992) obtained a similar value (approximately 0.1 PW when calculated relative to 0°C) in the same region, from satellite altimetry and hydrographic data, with a minimum in summer. Combining current-meter data with hydrographic data from the Svinøy section between April 1995 and June 1999, Skagseth and Orvik (1999) obtained a typical transport for the “Norwegian Atlantic Continental Slope Current” of 0.19 PW between the 490- and 990-m isobaths on the Svinøy section. They calculated a summer minimum that was 65% of the winter maximum and observed very little interannual variability among the three years considered (1996-98). They used a narrower definition of Atlantic Water (θ > 5°C, broadly corresponding to S > 35) than did Mork and Blindheim (θ > 1°C). The quasi-heat transport of the Atlantic inflow in the CATS section relative to 0°C is 0.17 ± 0.08 PW (the additional quasi-heat transport associated with freshened water in the Norwegian Coastal Current is 0.03 ± 0.01 PW). However, the depth-dependent component, comparable to the estimates of Pistek and Johnson (1992) and Mork and Blindheim (2000), is only 0.05 PW. The heat transport associated with the Norwegian slope, comparable to the estimate of Skagseth and Orvik, is 0.12 ± 0.07 PW.

Under the assumption that temporal variability does not affect comparisons, there are three possible contributory factors to the lower (quasi-)heat transport estimates in the southern Norwegian Sea than between Iceland and Scotland. The first is that the three Norwegian Sea estimates from the literature (Pistek and Johnson 1992; Skagseth and Orvik 1999; Mork and Blindheim 2000) are limited in length and there may be further heat transport that is associated with the Atlantic inflow. This is likely to be an important factor in the Skagseth and Orvik (1999) estimate, although their estimate is the largest, but not in the other estimates, which include transport associated with the Arctic front. The second factor is that the estimates of Pistek and Johnson (1992) and Mork and Blindheim (2000) include only the depth-dependent geostrophic component. This method leads to an underestimate of heat transport over the Iceland–Scotland Ridge (Hopkins 1988; Hansen and Østerhus 2000), and our results suggest that it does in the Norwegian Sea also. The third factor is actual heat loss to the atmosphere between the two locations. Hopkins (1988) estimates that over one-half of the additional heat carried by the Norwegian Atlantic Current is lost to the atmosphere before entering the Arctic Ocean or the Barents Sea or recirculating. The CATS section is approximately 2 times as far from the Faroe–Shetland Channel as the Svinøy section; therefore, it is not surprising that we obtain significantly lower comparable heat transport estimates than those calculated in the Svinøy section.

Our total heat transport estimate (0.20 ± 0.08 PW) suggests that one-third of the heat entering the Nordic seas is lost to the south of the CATS section, and that this is mostly due to heat loss in the Atlantic inflow. The heat loss between the Greenland–Ireland section (Bacon 1997) and the CATS section is 0.08 ± 0.10 PW. This represents a mean summer estimated heat flux to the atmosphere of 40 ± 50 W m⁻² between these locations. This compares with an annual mean in the Southampton Oceanography Centre (SOC) climatological data of 50 W m⁻² (Josey et al. 1998). (There is a bias in the oceanographic data: both datasets were collected in summer.) A heat loss of 0.20 ± 0.08 PW to the north of the section represents a mean summer heat flux to the atmosphere of 20 ± 8 W m⁻² in the northern Nordic seas and Arctic Ocean. Large areas of the Arctic Ocean are absent from the SOC data, and so comparisons here are impossible.

Some studies of the Atlantic inflow suggest a summer minimum in heat transport: Orvik et al. (2001) observed a 2:1 ratio in quasi-heat transport between winter and summer in the eastern branch of the NAC in the Svinøy section. Pistek and Johnson (1992) and Skagseth and Orvik (1999) observed a similar seasonal signal in the same region. However, the depth-dependent component observed by Mork and Blindheim (2000) revealed a maximum in winter and a minimum in spring; the range was 1.1 PW. Over the Iceland–Scotland Ridge, there is little variation in volume transport (Hansen and Østerhus 2000). The seasonality of the quasi-heat transport west of the Arctic front is not known; therefore it is uncertain what the annual heat transport across the CATS section is in comparison with the summer value.

Estimates by Aagaard and Carmack (1989, their Table 1) suggest oceanic freshwater transport of 0.11 Sv from the Nordic seas to the Atlantic relative to a salinity of 34.93, plus a contribution of 0.02 Sv from ice export. Hopkins (1988) deduced a freshwater transport of 0.20 Sv away from the Arctic between Greenland and Scotland. Bacon (1997) estimated a transport of 0.19 Sv between Greenland and Ireland and suggested a freshwater gain of 0.17 Sv in the Arctic when freshwater transports from the Bering Strait, the Canadian Archipelago, and ice export were included. Our estimate, 0.10
± 0.05 Sv plus an estimated 0.05 Sv on the Greenland continental shelf, suggests that the ocean gains 0.05 Sv (with an uncertainty of at least 100%) of freshwater between the CATS section and the Bacon (1997) section. Precipitation minus evaporation in the Nordic seas, which are divided approximately in half by the CATS section, is estimated to be 0.025 Sv (Aagaard and Carmack 1989). We estimate the quasi-freshwater transport of the Norwegian Coastal Current, which has an ultimate source in Scandinavia, to be 0.02 Sv, and there is a likely contribution from Greenland glaciers. Martin and Wadhams (1999) show a decrease in sea-ice flux between 76° and 71°N on the order of 0.01 Sv (in 1994), suggesting the input of this volume of freshwater because of melting of sea ice. Therefore, there is an expected freshwater input into the ocean between the CATS section and the Greenland–Scotland Ridge on the order of 0.05 Sv (equivalent to ~700 mm yr⁻¹ of net precipitation over the region). If it is assumed that there is an additional transport of 0.05 Sv inshore of the CATS section on the Greenland shelf, and if the oceanic freshwater flux between Greenland and Scotland is approximately 0.2 Sv, then our value for the freshwater flux of 0.10 Sv agrees with this estimate.

The relationship between summer freshwater flux and the annual mean flux is unclear. The EGC, which dominates the freshwater flux, is typically approximately 2 times as strong in winter than in summer (Woodgate et al. 1999), which might suggest greater freshwater fluxes in winter; however, it is uncertain whether the same seasonal cycle exists on the shelf where most of the freshwater transport occurs. Sea-ice production is on the order of 0.05 Sv less in summer than outside of summer (Martin and Wadhams 1999). This fact adds further complications: increased freshwater transport in sea ice may be related to decreased freshwater transport in the liquid ocean.

### 6. Conclusions

It is reassuring that the circulation in the initial field is consistent with previous estimates. The only adjustment made by the inverse model in the best estimate is a mean increase (decrease) of 0.04 cm s⁻¹ throughout the section of the current into (away from) the Arctic. It is therefore possible that the large-scale circulation at the time of the CATS expedition represents the mean summer circulation. Uncertainty in the assumption that the CATS section was representative is included in the uncertainties in the transport estimates.

There is a large-scale cyclonic circulation of 15 ± 6 Sv in the Nordic seas, with cyclonic circulation features of 15 ± 11 and 19 ± 10 Sv, respectively, in the Lofoten and Greenland Basins. The net transport of waters as light as the Atlantic inflow in the CATS section is 4.3 ± 2.5 Sv. The maximum in the streamfunction, at σ₀ = 27.879, is 5.4 ± 2.5 Sv, suggesting an overturning circulation north of the section with a transport equal to this value.

The CATS section encompasses the major routes of heat exchange with the Arctic; we calculate a net value of 0.20 ± 0.08 PW. The sensitivity of the heat transport estimate to the volume transport of the Atlantic inflow is 0.033 PW Sv⁻¹. The overturning component of the heat flux is 0.15 ± 0.07 PW. If we assume that the summer heat transport is representative of the summer mean, then the heat loss to the atmosphere north of the section is 20 ± 8 W m⁻². Comparison with other studies suggests a mean heat flux to the atmosphere of 40 ± 50 W m⁻² south of the CATS section and north of a pair of sections between Greenland and Ireland (Bacon 1997). The depth-independent component is of the same order as the depth-dependent component, contributing 0.11 ± 0.08 PW. There is a quasi-heat transport, relative to the section mean temperature, of 0.18 ± 0.07 PW over the Norwegian shelf and Norwegian slope and a transport of 0.10 ± 0.03 PW associated with the Arctic front.

The freshwater transport through the section is 0.10 ± 0.05 Sv out of the Arctic, with a likely additional transport on the order of 0.05 Sv on the Greenland shelf. The overturning component of the freshwater flux is 0.04 ± 0.02 Sv. The depth-independent contribution to the transport is 0.04 ± 0.05 Sv. A quasi-freshwater transport of 0.08 ± 0.04 Sv occurs over the Greenland shelf, with a transport into the Arctic of 0.02 ± 0.01 Sv over the Norwegian shelf.

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