Seasonal heat and freshwater budgets of the upper ocean in the north-east Atlantic

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SUMMARY

Heat- and freshwater-content changes were calculated from two repeated sections at 45°N, 18°W and 48°N, 26°W, surveyed during the Vivaldi cruise of the research ship RRS Charles Darwin in May 1991. Best estimates of the air–sea heat-fluxes agreed within error with the observed heat-content changes, suggesting that diabatic forcing dominated the heat budget. This was confirmed by a numerical model which showed that Ekman and geostrophic advection and eddy diffusion together contributed less than 20 W m⁻² to the heat budget.

However, the model also showed that, near the oceanic polar front under westerly winds, Ekman advection of cold, fresh water southwards over the front can cause -40 W m⁻² and 4 mm d⁻¹ heat- and freshwater-content changes in the upper 53 m, compared to air–sea fluxes of typically 67 W m⁻² and less than 1 mm d⁻¹. The Ekman advection contributed little to the model’s density-budget as the advective cooling and freshening had opposing effects. However, an air–sea interaction model showed that the advective surface-cooling may be reduced because it triggers an extra air–sea heat flux and the remaining advective freshening then makes an important contribution to the density budget.

KEYWORDS: Air–sea interaction Heat fluxes North Atlantic Current Ocean model World Ocean Circulation Experiment

1. INTRODUCTION

Air–sea heat and freshwater fluxes are needed as boundary conditions for ocean- and atmosphere-circulation models. The heat fluxes may be estimated using the bulk aerodynamic and climatological formules together with sea-surface meteorological observations. However, the errors in these estimates arising from uncertainties in the coefficients used in the formules (Isom and Hasse 1987) are thought to be 40 W m⁻². In long-term climate-prediction, the effect of this uncertainty may be an order of magnitude larger than climate changes caused by doubling atmospheric CO₂ (Houghton et al. 1990) and therefore one of the goals of the World Ocean Circulation Experiment (WOCE) is to reduce the errors in the estimated air–sea heat fluxes.

To achieve this, it is useful to have an independent method of calculating the fluxes. One possibility is to use hydrographic repeated sections such as those surveyed during the Vivaldi cruise of the research ship RRS Charles Darwin in the north-east Atlantic in spring 1991. If heat is stored locally in the ocean, as proposed by Gill and Niler (1973) and as demonstrated for the north-east Atlantic by Prangsma et al. (1983), it is possible to infer the air–sea heat-fluxes from observed heat-content changes, thus providing an independent estimate of the fluxes to compare with bulk-formulaire results. The Vivaldi repeated sections were especially useful for this purpose because they were 300 km long (larger than the eddy scale) and contained roughly 75 profiles. Therefore, the horizontally averaged estimates of heat-content change obtained from them had low standard-errors.

However, near the oceanic polar front and North Atlantic Current, advection or diffusion may be important in the heat or freshwater budgets. Despite this, air–sea heat or freshwater fluxes can still be determined provided the contribution of these other processes is calculated. During the Vivaldi 1991 cruise, a high-resolution systematic survey of the hydrography of the 1600 × 1800 km area surrounding the repeated sections was made and these data were used to initialize a numerical model of the heat and freshwater content of

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the area during May 1991 which estimated the contribution of advection and diffusion to the heat and freshwater budgets.

The model showed that air–sea heat fluxes dominated the heat budget at both the repeated sections, and its predictions of heat-content change were in agreement with those observed when best estimates of the air–sea heat fluxes were used.

The model also showed that, farther north near the North Atlantic polar front, Ekman flow can become important in the heat budget and, when winds are westerly, dominate the freshwater budget because of the resulting southward flow of sub-polar water over the sharp temperature and salinity gradients at the front. The effect of the Ekman flow on density in the model was small because of the opposing effects of advective cooling and freshening. However, a simple air–sea interaction model was used to show that the local surface-temperature anomaly caused by the advection may be reduced by feedbacks with the air–sea heat fluxes. The remaining advective freshening then contributes significantly to the density budget near the front.

2. Observations

(a) The Vivaldi 1991 cruise

The Vivaldi cruise in spring 1991 was a contribution by the United Kingdom to the WOCE. The north-east Atlantic was surveyed between 25 April and 8 June along the track shown in Fig. 1. The area surveyed, referred to here as the Vivaldi area, extended between 39°N and 54°N meridionally, and zonally between the mid-Atlantic Ridge (30°W to 34°W) and the European-shelf break (10°W). High resolution (one profile every 4 km) hydrographic profiles were collected continuously using the Sea Soar system: a towed-undulating Conductivity–Temperature–Depth (CTD) instrument. A series of 36 deep-CTD casts were made in a regular grid pattern spaced 333 km latitudinally and 300 km longitudinally. Their positions are shown in Fig. 1.

![Figure 1. The Vivaldi cruise track. The southern half of the pattern was surveyed first, from east to west, then the northern half from west to east. Note the two zonal repeated sections in the centre. The positions of CTD casts are marked with solid circles.](image-url)
(b) Hydrography

The heat content $H$ of the ocean from a depth $z = -D$ to the sea surface was defined as

$$H = \rho C \int_{-D}^{0} \theta(z) \, dz,$$

where $\theta$ was the potential temperature at depth $z$ in °C, $\rho$ was the mean density and $C$ the specific heat capacity of sea water. The freshwater content of a water column from a depth $z = -D$ to the ocean surface was defined as

$$F = \int_{-D}^{0} \left( \frac{38}{S(z)} - 1 \right) \, dz,$$

where $S(z)$ was the salinity at depth $z$ on the practical salinity scale. The reference value of 38 was chosen so that the freshwater content was positive definite. Therefore, the heat and freshwater contents are given hereafter relative to 0 °C and 38 respectively. The choice of reference values can be arbitrary as only changes in heat and freshwater content are considered.

Using (1) and (2), the heat and freshwater content of the roughly 2500 Sea Soar profiles collected during the cruise were calculated within six layers bounded by the depths 0, 29, 53, 77, 101, 277 and 453 m. These layers were chosen to provide more vertical resolution near the surface, to include a separate mixed layer 29 m deep and to fit in with the hydrographic data which was gridded vertically at 8 m intervals from 5 to 501 m depth. Data below 453 m were often lacking so this depth was chosen as the lower limit.

According to Bauer et al. (1991) seasonal heat- and freshwater-content changes in spring are restricted to the upper 60 m and indeed the heat content in the upper two layers (upper 53 m) measured during the cruise contained systematic temporal variations. Before these data were used to initialize the model for 25 April, they were modified by subtracting from the heat content of each profile, within the upper two layers, the estimated diabatic heat-input since the start of the cruise, on 25 April, using the Bunker Atlas climatology. The heat- and freshwater- (uncorrected) content values for all the profiles were then interpolated horizontally within each layer using an optimum interpolation scheme with an e-folding length of 500 km for the correlation scale. This scale was larger than the 300 km spacing between fronts in this area.

Figure 2 shows the horizontal distribution of temperature (corrected for 25 April) and salinity (from 25 April to 8 June) of the mixed layer (from the surface to 29 m depth) derived from the Sea Soar data. In the north-west, the position of the oceanic polar front can be inferred from the closer spacing of the isotherms. It forms a boundary between cold sub-polar water to the north-west and warmer subtropical water to the south and east. The northward gradient in temperature across the front was $-1.6$ K deg$^{-1}$. The front is also apparent in Fig. 2(b) as a boundary between the fresh (low salinity) water to the north-west and saline water to the south and east. The northward gradient in salinity across the front was $-0.29$ psu$^*$ deg$^{-1}$. The vertical gradient of temperature over the upper 53 m was 0.02 K m$^{-1}$ near the front. The vertical salinity-gradient was positive, characteristic of North Atlantic central water, south of the front at 51°N and negative to the north.

* The practical salinity unit (psu) is defined in terms of conductivity but, to a good approximation, sea water with a salinity of 1 psu has one gram of salts dissolved in each kilogram of sea-water solution.
Figure 2. The upper 29 m, interpolated from Vivaldi 1991 cruise Sea Soar data: (a) temperature (°C), (b) salinity (psu).

(c) The repeated sections

The two repeated sections extended 300 km zonally at 48°N (see Fig. 1), the west section between 24°W and 28°W and the east section between 16°W and 20°W. The west section was surveyed on 12 and 29 May 1991 (with a 17-day interval between surveys) and the east section on 3 May and 5 June (a 34-day interval).

Figure 3 shows the temperature and salinity changes for both the repeated sections as a function of depth. The changes were averaged over the 300 km zonal length of the sections (each made up of about 75 repeated profiles). As a result of the averaging, the effects of mesoscale eddies on the results should be reduced. The standard errors of the horizontally averaged temperature and salinity changes were 0.06 K and 0.01 psu.

Figure 3(a) shows an increase of temperature at the surface of up to 1 K over 17 days at the west section and 1.2 K over 34 days at the east section. At the west section there was
a decrease in temperature below the seasonal thermocline (60 m deep) which increased with depth, reaching 0.4 K over 17 days at 400 m. This cooling may have been caused by the arrival of a cold-core eddy at the west section, but the cooling may also have been caused by a localized upwelling, a possibility discussed in section 3.

The heat- and freshwater-content changes were calculated for each repeated profile along the section and these results were then averaged. In the upper 53 m, the heat-content changes were 183 ± 8 W m⁻² for the west section and 118 ± 3 W m⁻² for the east section. The errors given are the standard errors of the averages. If these heat-content changes were caused solely by air–sea heat fluxes then these fluxes were larger than is typical for an average May according to the Bunker Atlas (Isemer and Hasse 1987) which gives an estimate of 67 W m⁻² for this position.

Figure 3(b) shows that, in the upper 53 m at both repeated sections, there was a net evaporation, although below this depth (especially at the west section) there was a decrease in salinity associated with the cooling at depth (see Fig. 3(a)). The salinity changes in the upper 53 m imply, if advection and diffusion may be neglected, a latent-heat flux of −76 and −27 W m⁻² at the west and east sections respectively.

### 3. Formulation of the model

A numerical model was developed to predict the variation of the heat and freshwater content of the seasonal thermocline (upper 53 m) in the north-east Atlantic during May 1991 and to determine the relative importance of the processes causing heat- and freshwater-content changes. The Eulerian rate of change of temperature is given by

\[
\frac{\partial T}{\partial t} = \frac{1}{\rho C} \frac{\partial Q}{\partial z} - \mathbf{u} \cdot \nabla_h T - \mathbf{u}_g \cdot \nabla_h T - w_e \frac{\partial T}{\partial z} - w_g \frac{\partial T}{\partial z} + \nabla \cdot (K \nabla T),
\]

where the terms on the right-hand side represent the vertical divergence of air–sea heat-fluxes, Ekman and geostrophic advection, vertical Ekman pumping and geostrophic advection and eddy diffusion of heat with coefficient \( K \). The \( \nabla \) and \( \nabla_h \) represent three-dimensional and horizontal spatial derivatives respectively. The \( \mathbf{u} \) and \( w \) are the horizontal and vertical velocities and the subscripts \( e \) and \( g \) refer to the Ekman and geostrophic flow.
We assumed that the Ekman pumping was negligible on monthly timescales following Bauer et al. (1991) who noted that the line of zero wind stress curl passes through the Vivaldi area. The Ekman-pumping speed was derived from the divergence of the horizontal Ekman flows calculated as discussed in section 4(c) and was at most 8 m month\(^{-1}\), which would cause only about 5 W m\(^{-2}\) heat-content change in the upper 53 m, given the vertical temperature-gradient from section 2(b). Typical Ekman pumping-rates in this area are less than 2 m month\(^{-1}\) (Gill 1982).

To satisfy the continuity equation, the vertical geostrophic flow was neglected. To estimate the error incurred by this assumption, and to investigate the importance of any localized upwelling, the upward flow implied by the changes in density at 400 m at the west repeated section was calculated and found to be 3 × 10\(^{-5}\) m s\(^{-1}\). Assuming the vertical flow was zero at the surface, and linearly interpolating, gives a flow in the upper 53 m of 2 × 10\(^{-6}\) m s\(^{-1}\), which may have caused a heat-content change of 3 W m\(^{-2}\). The west section was 300 km from the polar front and North Atlantic Current, so frontal upwelling from this source should be absent. The vertical flow could not be calculated using the \(\omega\)-equation method (see Leach 1986) because the Vivaldi hydrographic sections were too widely spaced.

We consider two layers in the ocean, the first from the surface to a depth of \(h_1\) and the second between the depths \(h_1\) and \(h_2\), and integrate (3) with depth for each layer. In the model, the depth \(h_2\) was 53 m and (according to Thompson (1976)) all but 3% of the downward short-wave radiation is absorbed above this depth. Therefore, the downward short-wave radiation was assumed to be zero at depth \(h_2\). Adding the integrals of (3) for each level, and taking the upper layer to represent both the Ekman layer and the mixed layer, we have

\[
\int_{-h_2}^{0} \frac{\partial T}{\partial t} \, dz = \frac{Q}{\rho C} - \int_{-h_2}^{0} \mathbf{u}_e \cdot \nabla_h T \, dz - \int_{-h_2}^{0} \mathbf{u}_g \cdot \nabla_h T \, dz - \int_{-h_2}^{0} \mathbf{u}_g \cdot \nabla_h T \, dz + \int_{-h_2}^{0} \nabla \cdot (K \nabla T) \, dz, \tag{4}
\]

where \(Q\) is the air–sea heat flux through the sea surface. We assumed that \(\mathbf{u}_e\) and \(\mathbf{u}_g\) had no vertical shear within each layer. The calculations of geostrophic flow using the Vivaldi hydrographic data (see subsection 4(b)) indicated that the shear between 29 and 53 m depth, for example, was 0.1 cm s\(^{-1}\) or 2% of the surface flow of 6 cm s\(^{-1}\). Using (1) for the heat content of the upper \((H_1)\) and lower \((H_2)\) layers and (4), we obtain

\[
\rho C \int_{-h_2}^{0} \frac{\partial T}{\partial t} \, dz = Q - \mathbf{u}_e \cdot \nabla_h H_1 - \mathbf{u}_{g1} \cdot \nabla_h H_1 - \mathbf{u}_{g2} \cdot \nabla_h H_2 + K \nabla^2 (H_1 + H_2). \tag{5}
\]

The Ekman and mixed-layer depth were assumed to be constant with time. To test this assumption, the mixed-layer depth was calculated for every Sea Soar profile collected during the Vivaldi cruise. The base of the mixed layer was chosen to be the depth where the density was 0.125 kg m\(^{-3}\) greater than its value at 5 m depth. Averaged over the duration of the cruise, this depth was 26 m with a standard deviation of 21 m and a standard error of 0.4 m. There was no significant trend in the mixed-layer depth after the first few days when there was a shallowing of the layer. Therefore, using this assumption, we have an expression for the rate of heat-content change

\[
\frac{\partial (H_1 + H_2)}{\partial t} = Q - \mathbf{u}_e \cdot \nabla_h H_1 - \mathbf{u}_{g1} \cdot \nabla_h H_1 - \mathbf{u}_{g2} \cdot \nabla_h H_2 + K \nabla^2 (H_1 + H_2). \tag{6}
\]
Similarly, for the rate of change of freshwater content $F$ given by (2), we obtain

$$
\frac{\partial (F_1 + F_2)}{\partial t} = -Q_{E-P} - u_x \cdot \nabla_h F_1 - u_{y1} \cdot \nabla_h F_1 - u_{y2} \cdot \nabla_h F_2 + K \nabla^2 (F_1 + F_2)
$$

(7)

where $Q_{E-P}$ is the excess of evaporation over precipitation at the ocean surface.

Equations (6) and (7) were solved using a $z$-coordinate numerical model for the upper 53 m of the ocean in the north-east Atlantic, covering the same area as the Vivaldi cruise (see Fig. 1). The area was bounded on the west and east by lines drawn 900 km west and 700 km east of the 20°W meridian. This made it possible to impose a cartesian grid over the earth’s curved surface. The volume was bounded to the north and south by the 39°N and 55°N latitude lines. The 1600 km $\times$ 1800 km model grid had a horizontal resolution of 100 km, with 18 and 16 grid points respectively in the meridional and zonal directions. With a time step of one day, the Courant–Friedrichs–Lewy number was 0.1, assuming a maximum flow of 10 cm s$^{-1}$. For further details of the numerical model see McCulloch (1995). The model was initialized for 25 April using the temperature and salinity fields as shown for the model’s upper layer in Fig. 2(a,b) derived from the Vivaldi Sea Soar profiles. The high resolution of the data (4 km) meant that the heat and freshwater content at every grid point was calculated from about 25 profiles. Advection was performed using the Flux-Corrected Transport Algorithm of Zalesak (1979) and Dietachmayer (1986) and the eddy diffusion coefficient $K$ was given a value of 2000 m$^2$ s$^{-1}$.

4. FORCING FIELDS FOR THE MODEL

(a) Air–sea fluxes

The air–sea heat and freshwater (evaporation minus rainfall) fluxes for May 1991 used in the model were obtained from the operational numerical weather prediction model of the Meteorological Office. They were interpolated spatially onto the model grid and temporally for every day of the model run. Sensible-heat fluxes were not available but are typically less than 10 W m$^{-2}$, whereas net heat-fluxes for May are between 67 and 154 W m$^{-2}$ (see below).

The air–sea freshwater fluxes averaged over the period of the Vivaldi cruise are shown in Fig. 4(a). Evaporation was largest in the south-east, because of the warm, dry tradewinds. In the north-west, near the polar front, rainfall roughly balanced evaporation.

Figure 4(b) shows the net air–sea heat flux for the same period which increased towards the north-west. Since the air–sea heat fluxes are likely to introduce the dominant error into the model it is useful to compare these fluxes with other available estimates. The Bunker Atlas (Isenmer and Hasse 1987) includes monthly average air–sea heat fluxes derived using the bulk aerodynamic formulas together with sea-surface meteorological observations spanning the years 1941 to 1972. As a result, the fluxes estimated for each month apply only to an average month over this period. The Bunker Atlas (revised) estimate of net air–sea heat flux for May in the centre of the Vivaldi area was significantly lower than the fluxes shown in Fig. 4(b) at 67 W m$^{-2}$.

A second estimate, this time specifically for May 1991, was calculated using the bulk formulas by P. K. Taylor (personal communication 1993) from sea-surface and meteorological observations made during the Vivaldi cruise. These fluxes averaged 154 W m$^{-2}$ over the duration of the cruise and agree more closely than the Bunker values with the atmospheric-model fluxes, suggesting that during May 1991 the air–sea heat fluxes were significantly higher than in a typical May. Table 1 shows the components of all three estimates of air–sea heat flux at the west and east repeated sections. Also shown are
Figure 4. Net air–sea fluxes estimated by the Meteorological Office for May 1991 over the Vivaldi area in mm d⁻¹ and W m⁻²: (a) fresh water; (b) heat.

<table>
<thead>
<tr>
<th>Data Set</th>
<th>Short wave</th>
<th>Long wave</th>
<th>Sensible</th>
<th>Latent</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bunker Atlas</td>
<td>185(186)</td>
<td>-35(-36)</td>
<td>-10(-8)</td>
<td>-77(-72)</td>
<td>63(70)</td>
</tr>
<tr>
<td>Vivaldi</td>
<td>211(211)</td>
<td>-22(-22)</td>
<td>-6(-6)</td>
<td>-29(-29)</td>
<td>154(154)</td>
</tr>
<tr>
<td>Met. Off.</td>
<td>196(205)</td>
<td>-32(-32)</td>
<td></td>
<td>-25(-38)</td>
<td>139(135)</td>
</tr>
<tr>
<td>FW budget</td>
<td></td>
<td></td>
<td></td>
<td>-76(-27)</td>
<td></td>
</tr>
</tbody>
</table>

The table presents four independent estimates of the air–sea heat flux (W m⁻²) for the west (and east, in brackets) repeated sections (rows one to four) and the observed heat storage from Vivaldi Sea Soar data. FW denotes freshwater.
latent-heat fluxes calculated using observed freshwater-content changes from the Vivaldi repeated sections in the upper 53 m (see Fig. 3(b)) and the total heat-content change of the upper 53 m from the repeated-section temperature data. According to Weather (1991), the atmospheric pressure anomaly over the north-east Atlantic was +12 hPa during May 1991. High pressure is associated with less cloud (and so higher short-wave radiative heat flux into the ocean) and lower wind speeds (thus lower turbulent heat loss from the ocean). In Table 1, in agreement with this, the short-wave heat gain was higher and the latent-heat losses were lower in the two estimates for May 1991.

The agreement between the Vivaldi and atmospheric-model fluxes (Table 1, rows 2 and 3) and the agreement of these estimates within error bars (typically 40 W m⁻² for the fluxes) with the observed heat-content changes (Table 1, row 5) support the validity of the atmospheric-model fluxes and their use in the model. The agreement between these fluxes and the observed heat-content change also suggests that air–sea heat fluxes dominated the heat budget of the upper 53 m. The numerical model that was described in section 3 was used to verify this conclusion (see section 5). In the next subsection, the Ekman and geostrophic flow fields used to estimate the advection of heat and freshwater in the model are discussed.

(b) Geostrophic flow

The geostrophic flow in the Vivaldi area in May 1991 was calculated using the dynamic method

\[ \mathbf{u}(z) = -\frac{g}{\rho_f} \int_{-L}^{z} \hat{\mathbf{k}} \times \nabla \rho \, dz, \]  

where \( \mathbf{u} \) is the flow vector at a depth \( z \) and \( \hat{\mathbf{k}} \) is the vertical unit vector. The Coriolis parameter \( f \) varied with latitude and the level of no motion \( L \) was taken as 1500 m depth. In a sensitivity study, a change in the level of no motion from 1500 m to 2000 m altered the predicted heat-content change of the model by (typically) 1 W m⁻² in the upper 53 m. The geostrophic flow at 453 m depth was calculated using the Vivaldi cruise deep-CTD data that extended to the sea floor. The high-resolution Sea Soar data were then used to calculate the geostrophic flow above 453 m relative to that depth and this flow was then added to the CTD-derived flow.

Figure 5 shows the geostrophic flow calculated as above for the upper 29 m over the Vivaldi area. The direction of flow is shown by the arrows and the speed by the arrow length and the contours (isotachs). The fastest flow is 6 cm s⁻¹ eastwards in the west at about 51°N, which corresponds to the usual position of the North Atlantic Current. The flow turns northwards at 20°W. The volume transports in the upper 453 m through the west, north, east and south edges of the Vivaldi area respectively were 16 Sv eastwards, 12 Sv northwards, 0.5 Sv eastwards and 2 Sv southwards (1 Sv = 10⁶ m³ s⁻¹). This implies that 75% of the eastward flow into the Vivaldi area is transported northwards and there is only a small recirculation southwards back into the subtropical gyre.

(c) Ekman flow

The Ekman flow over the Vivaldi area in May 1991 was computed using monthly-averaged sea-level pressures derived from the same atmospheric model that provided the air–sea heat and freshwater fluxes. The winds were calculated assuming a geostrophic balance but also taking into account the Ekman spiral in the lower atmosphere following Thompson et al. (1983) who suggested a rotation of the wind 16° to the left and a reduction in speed by 30% in the Vivaldi area. The wind stress on the ocean surface was calculated
Figure 5. The geostrophic flow (cm s$^{-1}$) in the upper 29 m, calculated using the dynamic method from Vivaldi Sea Soar and deep-CTD data. Arrows show the flow direction and contours, and length of arrows indicate flow speed.

correcting for the underestimation caused by the use of monthly-averaged winds, again as in Thompson et al. (1983). The Ekman flow $u_e$ was then calculated using

$$u_e = -\frac{k \times \tau}{\rho_e f h_e}, \quad (9)$$

where $\rho_e$ is the density of the Ekman layer, $\tau$ the wind-stress vector and $h_e$ the Ekman layer depth.

Figure 6(a) shows that the Ekman flow in May 1991 converged on the centre of the Vivaldi area due to the high-pressure system positioned there. Figure 6(b) shows a typical Ekman flow pattern for May derived from Bunker Atlas pressure-data. In May 1991, the Ekman flow was directed eastwards along the oceanic polar front, whereas during a typical May it is southward over the relatively sharp temperature- and salinity-gradients across the front. In the latter case, Ekman flow is likely to have a much larger effect on the heat and freshwater budgets. To satisfy the formulation of the model, the Ekman flow was modified to eliminate horizontal divergence. This altered the southward flow in the west section, for example, where the flow was largest, by only 0.4 cm s$^{-1}$ or 8%.

5. RESULTS OF THE MODEL

(a) Validation

Figure 7 shows the observed upper-ocean temperature and predictions of it by the model, plotted against the number of days into the Vivaldi cruise. After 35 days the model was overestimating the mixed-layer temperature by about 1.3 K. This difference could have been caused by a plausible error of 50 W m$^{-2}$ in the air-sea heat fluxes. A similar graph is not shown for salinity because the spatial variations in salinity are much larger than the temporal changes.
(b) At the repeated sections

Table 2 shows the heat budget of the upper 53 m of the ocean predicted by the model at the west and east repeated sections (columns 1 and 3). The errors shown in columns 2 and 4 were determined from a sensitivity study of the model. The error bars of the air–sea heat fluxes were calculated following Hsiung (1986) who suggested a 20% error from each of the four components of the flux. The total predicted heat-content change is shown in row 5 and the contributions of each of the processes to that total are shown in rows 1 to 4. The observed heat-content changes of the upper 53 m derived from the Vivaldi repeated-section data are shown in row 6.

The air–sea heat fluxes dominated the model’s heat-budget at both positions in May 1991 and advection and diffusion together contributed less than 20 W m$^{-2}$. The contribu-
tion of Ekman advection was small in this instance because the Ekman flow was parallel to isotherms at the west section and small at the east section (see Fig. 6(a)). The predictions of heat-content change by the model for both sections agreed within error bars with those observed. The discrepancy was 21 W m\(^{-2}\) at the west section and 40 W m\(^{-2}\) at the east section.

Table 3 shows the freshwater budget of the model at the repeated sections. The contribution of the air–sea freshwater-fluxes was smaller than those of advection and diffusion in the freshwater budget at both repeated sections although the errors on the air–sea fluxes are perhaps as much as 2 mm day\(^{-1}\) according to Pollard and Pu (1985). The predicted freshwater-content changes agreed with those observed, given the large error bars, at both sections. The freshwater budget was dominated by diffusion and air–sea fluxes (because of their large uncertainty).

\(\text{(c) At the polar front}\)

The results discussed above are for May 1991. The large positive pressure-anomaly for this month indicates that the Ekman flow (Fig. 6(a)) may be atypical and the Ekman flow derived from the Bunker Atlas indicates that the flow is usually southwards at the oceanic polar front (Fig. 6(b)). To investigate this possibility, the model was rerun using

<table>
<thead>
<tr>
<th>TABLE 2. HEAT BUDGET AT THE REPEATED SECTIONS.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Process</td>
</tr>
<tr>
<td>------------------------</td>
</tr>
<tr>
<td>Air–sea heat flux</td>
</tr>
<tr>
<td>Geostrophic flow</td>
</tr>
<tr>
<td>Ekman flow</td>
</tr>
<tr>
<td>Eddy diffusion</td>
</tr>
<tr>
<td>Total</td>
</tr>
<tr>
<td>Observed</td>
</tr>
</tbody>
</table>

The table indicates the contribution of processes to the model's heat budget (W m\(^{-2}\)) and observed heat-content change at the west and east repeated sections during spring 1991.
TABLE 3. Freshwater budget at the repeated sections.

<table>
<thead>
<tr>
<th>Process</th>
<th>West</th>
<th>Error</th>
<th>East</th>
<th>Error</th>
</tr>
</thead>
<tbody>
<tr>
<td>E-P</td>
<td>−0.07</td>
<td>±2.0</td>
<td>−0.29</td>
<td>±2.0</td>
</tr>
<tr>
<td>Geostrophic flow</td>
<td>−0.33</td>
<td>±0.1</td>
<td>−0.23</td>
<td>±0.2</td>
</tr>
<tr>
<td>Ekman flow</td>
<td>0.15</td>
<td>±0.2</td>
<td>0.17</td>
<td>±0.1</td>
</tr>
<tr>
<td>Eddy diffusion</td>
<td>0.86</td>
<td>±0.4</td>
<td>−0.43</td>
<td>±0.1</td>
</tr>
<tr>
<td>Total</td>
<td>0.61</td>
<td>±2.1</td>
<td>−0.77</td>
<td>±2.1</td>
</tr>
<tr>
<td>Observed</td>
<td>−2.03</td>
<td>±1.0</td>
<td>−1.06</td>
<td>±0.4</td>
</tr>
</tbody>
</table>

The table indicates the contribution of processes to the model's freshwater budget (mm d⁻¹) and observed freshwater content change at the west and east repeated sections during spring 1991.

Figure 8. Components of budgets along a section from 39°N, 27°W to 55°N, 29°W in W m⁻² and mm d⁻¹: (a) heat; (b) freshwater. Note, air-sea fluxes are not shown.


Figure 8 shows the advective and diffusive components of the heat and freshwater budgets predicted by the model for the upper 53 m along a meridional section 300 km from the western boundary of the model during May between 39°N, 27°W and 55°N, 29°W. The errors incurred at the boundary on the model's western edge were advected eastward by the geostrophic flow at 6 cm s⁻¹ (see Fig. 5) and would take over 50 days to reach this meridional section, longer than the 35-day model run. Figure 8(a) shows that, near the polar front at 51°N, Ekman advection caused −40 W m⁻² heat-content change, twice as large as both the geostrophic advection and diffusion there, although still smaller than the air-sea heat fluxes of 67 W m⁻² (not shown). At this position, the Ekman flow was directed southwards (see Fig. 6(b)), across the sharp temperature-gradient at the front.

Figure 8(b) shows the contributions to the freshwater budget along the same section. At the polar front, Ekman advection dominated the freshwater budget causing a freshwater-content increase of 4 mm d⁻¹. These changes were larger than the contribution of the air-sea freshwater-fluxes (not shown) which are typically less than 1 mm d⁻¹.

The mechanism behind the Ekman advective cooling and freshening predicted by the model is illustrated in Fig. 9. This schematic shows the isotherms of the north-east Atlantic Ocean both at the ocean surface and in a section down through the ocean. The prevailing
westerly winds along the oceanic polar front induced Ekman flow southwards across the front in the model, bringing colder, fresher water to the west section. This cross-frontal flow did not occur in May 1991 as the high-pressure area (discussed in subsection 4(a)) was centred on the Vivaldi area causing southerly winds in the west which induced Ekman flow eastwards along the polar front. This accounts for the small predicted effect of Ekman advection for May 1991 in Tables 2 and 3.

These results show that, if the winds are westerly at the oceanic polar front, air–sea heat-fluxes and Ekman advection can both be important in the heat budget. In the freshwater budget at the front, Ekman advection can make the dominant contribution. The difference between the importance of Ekman flow in a typical May and in May 1991 implies that the heat and freshwater budgets are sensitive to variations in the direction and strength of the wind near ocean fronts.

The model results suggest that during summer, when vertical buoyancy-driven mixing is absent (as assumed in the model), cold fresh water spreads southwards over the polar front in the mixed layer over the warmer, more saline, waters of the subtropical gyre. This spreading has been observed since it was reported by Neumann (1940) and observations of it were discussed by Bauer et al. (1991). For reasons discussed in section 7, the latter came to a different conclusion from that reached here and suggested that the fresh water originated farther west near Newfoundland and was advected eastward into the north-east Atlantic by the geostrophic flow. However, in their study, they calculated horizontal advection terms using heat contents relative to a (varying) reference temperature at 60 m depth and this may have led to errors, especially near the polar front where temperatures at 60 m vary horizontally.
6. Density Changes

Recent studies (for example Marshall et al. 1993) have discussed the importance of the buoyancy (or density) budget of the upper ocean in determining annual subduction patterns over the North Atlantic. Therefore, in this section, the contribution of the dominant processes in the heat and freshwater budgets (as determined above) to the density budget are discussed. The emphasis is on density $\rho$, but buoyancy $B$ can be defined as $B = -\rho g$. Assuming initially that, as shown by the model in subsection 5(b), the temperature and salinity of the upper ocean depend primarily on air–sea heat and freshwater fluxes, the rate of change of density of a water column in the upper ocean may be approximated as

$$\frac{\partial \rho}{\partial t} = -\frac{\alpha Q_H}{Ch} + \frac{\rho \beta S_0^2}{38h} Q_{E-P},$$

(10)

where $Q_H$ and $Q_{E-P}$ represent the air–sea heat- and freshwater-fluxes, the coefficients $\alpha$ and $\beta$ are the thermal expansion and haline contraction coefficients respectively, $h$ is the column's depth (53 m) and $S_0$ its salinity. Equation (10) was solved numerically for a water column at 51°N and 24°W at the polar front. The initial temperature and salinity of the column were derived from Vivaldi hydrographic data for May 1991 at this position. The Bunker Atlas heat-fluxes were used, the freshwater fluxes were derived from the Bunker latent-heat fluxes, and rainfall was neglected.

Figure 10. Predicted evolution of density (kg m$^{-3}$) within the upper 53 m of the ocean at the oceanic polar front (51°N, 24°W) taking into account fluxes and flows: Bunker Atlas air–sea heat and freshwater fluxes (solid line); Bunker fluxes and southward Ekman flow over the front (dashed line); Bunker fluxes, southward Ekman flow and perturbations to the Bunker fluxes predicted by an air–sea interaction model (dashed line with stars). The model predicts density changes increased by 60% over those caused by the Bunker fluxes alone, because of feedbacks between the Ekman advection and air–sea heat and freshwater fluxes.

Figure 10 shows the evolution of the density of the column over 60 days during June and July. The density decreased under an average air–sea heat flux of 80 W m$^{-2}$. The evaporation rate of 1.3 mm d$^{-1}$ and the resulting increase in salinity were too small to reverse the effect of higher temperatures on density.

The results of subsection 5(c) showed that Ekman advection of both heat and freshwater can be important at the oceanic polar front. To assess its effect on density, a southward flow ($u_e$) over the front was included in the model

$$\frac{\partial \rho}{\partial t} = -\frac{\alpha Q_H}{Ch} + \rho \alpha u_e \cdot \nabla T + \frac{\rho \beta S_0^2}{38h} Q_{E-P} - \rho \beta u_e \cdot \nabla S,$$

(11)
where the second and fourth terms on the right-hand side represent the Ekman advection of heat and fresh water. The southward flow imposed was 3.2 cm s$^{-1}$, corresponding to the Bunker Atlas value for May (see Fig. 6(b)). The meridional gradients in temperature and salinity were given in subsection 2(b).

Figure 10 shows the evolution of density over the two-month period, including the effect of Ekman advection, which caused a heat-content change of $-85$ W m$^{-2}$ (thus cancelling the diabatic heating), and a freshwater-content change of 7 mm d$^{-1}$. In conclusion, the effect of Ekman advection on density was only about 20% that of the air–sea heat fluxes (in Fig. 10 the solid and dashed lines are close) because the advective cooling and freshening tended to cancel each other in their effect on density.

7. AIR–SEA INTERACTION

In this section, the sensitivity of the conclusion of section 6 is tested by including the feedback between the Ekman flow, which reduces the sea-surface temperature, and the air–sea heat and freshwater fluxes which are allowed to deviate from their original Bunker values (which were derived using typical sea-surface temperatures) in response to the anomaly. The density was assumed to vary as

$$\frac{\partial \rho}{\partial t} = -\frac{\alpha (Q_{H} + Q'_{H})}{Ch} + \rho \alpha \mathbf{u}_e \cdot \nabla T + \frac{\rho \beta S_{0}^{2}}{38h} (Q_{E-P} + Q'_{E-P}) - \rho \beta \mathbf{u}_e \cdot \nabla S,$$

where $Q'_{H}$ represents the perturbation of the climatological (Bunker) heat fluxes caused by the local sea-surface temperature anomaly (caused by the Ekman flow) and $Q'_{E-P}$ is the associated perturbation to the climatological rate of evaporation. It was assumed that the effect of the advective sea-surface temperature anomalies were not included in the original Bunker fluxes (whose derivation made use of observed sea-surface temperatures) because the anomalies are confined to a narrow strip along the front whereas the Bunker values were smoothed over larger scales, and the anomalies may occur only under favourable winds (i.e. not every month).

An air–sea-interaction model of Williams (1988) was used to derive values of $Q'_{H}$ and $Q'_{E-P}$ from the predicted sea-surface temperature. This model assumes that the sea-surface temperature anomaly does not alter the local air-temperature and humidity at 10 m height or cloud cover as the air passes over the anomaly too quickly. Figure 8(a) shows that Ekman advection dominated the heat budget over a region 400 km wide. With a wind speed of 10 m s$^{-1}$, the air would pass over the anomaly in 12 hours. Values of the humidity of the air at sea level, needed for calculations of the modified latent-heat flux and evaporation-rates, were calculated from the modified sea-surface temperature using an empirical relation (Gill 1982) between temperature and water vapour pressure. McCulloch (1995) used a simpler air–sea-interaction model for the same analysis.

Figure 10 shows the density changes predicted incorporating the air–sea-interaction model. The decrease of density, or the buoyancy input, over the two months was enhanced by about 60% over that solely due to the Bunker air–sea heat and freshwater fluxes. This occurred because the sea-surface temperature reduced by the Ekman advection led to a reduction (from the Bunker Atlas values) of latent- and sensible-heat loss, and of longwave heat loss, by about 30, 20 and 5 W m$^{-2}$ respectively. The short-wave heat input was unaffected and the result of the heat by the ocean that reduced the impact of the original Ekman cooling from $-85$ to $-30$ W m$^{-2}$. The Ekman cooling was reduced but the Ekman freshening remained, thus causing a large buoyancy input. The reduced sea-surface temperature also reduced evaporation, thus enhancing the freshening and its buoyancy input.
by 8%. These results show that the conclusion reached in section 6, that Ekman advection
near the front has only a small effect on surface density, may be incorrect. When air–sea
interaction is considered, Ekman advection can contribute significantly (relative to air–sea
heat fluxes) to the density or buoyancy budget. Thus, whenever westerly winds force cold,
fresh water across an oceanic front, there may be a large buoyancy gain south of the front.

The occurrence of such feedbacks may explain some of the observations of Bauer et al.
(1991), who documented the southward, surface spreading of cold, fresh water across
the polar front similar to that discussed here. They noted that an Ekman flow southward
over the polar front could account for about half of the freshening over summer near
the front, but also produced more advective cooling than was observed. To balance their
heat and freshwater budget, they decided that the freshening must be caused primarily
by eastward geostrophic advection of freshwater from near Newfoundland. Consequently,
they increased the eastward geostrophic flow in their calculations and also adjusted their
diabatic forcing. However, the effect of local Ekman advection on the heat budget may be
reduced, as discussed above, because of air–sea interaction (effectively because the air–sea
heat flux increases to offset the advective cooling near the front). Therefore, the heat and
freshwater budget of Bauer et al. (1991) may be satisfied more simply by including the
effect of air–sea interaction.

8. CONCLUSION

A numerical model has been used to show that air–sea heat fluxes dominated the
heat budget in May 1991 in the north–east Atlantic. This conclusion agrees with that of
Gill and Niller (1973) and suggests that observations of seasonal heat–content changes
from repeated hydrographic sections could be used to infer the air–sea heat fluxes more
accurately than is possible using the bulk formulas. However, the model also showed
that near the oceanic polar front under westerly winds, Ekman advection was important
in the heat budget and dominated the freshwater budget as cold, fresh water was forced
southwards across the front. The Ekman advection contributed little to the model’s surface-
density budget as the advective cooling and freshening had opposing effects. However,
an air–sea-interaction model showed that the advective surface cooling may be reduced
because it triggers an extra air–sea heat flux, and the remaining advective freshening then
makes an important contribution to the density budget.

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