Chapter 4
Is Oceanic Heat Transport Significant in the Climate System?

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4.1 Introduction

It has long been believed that the transport of heat by the ocean circulation is of importance to atmospheric climate. Circulation of the Atlantic Ocean warms and moistens western Europe, the argument goes, and, because of the pivotal role of the Atlantic/Arctic region, also affects global climate (e.g., Stommel 1979). Indeed, major oceanographic field programs have been launched by many nations, based on this premise. In the US, NOAA issues quarterly assessments of subtropical North Atlantic meridional heat transport (http://www.aoml.noaa.gov/phod/soto/mht/reports/index.php). Estimates from ocean observations show the annual-mean, northward heat transport by the global circulation to decrease by about 1.5 pW (10^{15}W) between latitudes 25° N and 50° N, with nearly 1 pW of that within the narrow Atlantic sector alone (see Bryden and Imawaki 2001, who estimate the uncertainty of individual section heat transports at 0.3 pW). This effect of the oceanic meridional overturning forces an enormous upward flux of heat and moisture in subtropical latitudes, providing a significant fraction of the zonally integrated atmospheric northward energy flux (which peaks at between 3 and 5.2 pW, as discussed further below). Occurring dominantly in wintertime, oceanic warmth and moisture energize the Pacific and Atlantic storm tracks. Combined action of atmosphere and ocean carries this energy northward, with great impact on all facets of high-latitude climate.

Northern Atlantic climate hovers in the midst of debates over the dynamical origins and impacts of the global oceanic meridional overturning circulation (MOC), and its contribution to the coupled atmosphere–ocean system. Still, the

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complexity of the shared process of heat and fresh water transport and exchange by ocean and atmosphere continues to belie oversimplified ‘conveyor belt’ images. Does ‘Bjerknes compensation’ occur, under which decadal variability the ocean and atmosphere components of meridional heat transport compensates, one for the other, leaving an unchanging top-of-atmosphere radiation field (e.g., van der Swaluw et al. 2007; Dong and Sutton 2005)?

Taken as a whole, the current debate over the most fundamental principles of the MOC demonstrates how much physical oceanography has yet to learn about its sector of the climate system and how important that sector is. Some of the implied questions currently under debate are:

Is the MOC pushed by buoyancy forcing or pulled by mixing induced by winds and tides (e.g., Toggweiler and Samuels 1995; Wunsch et al. 2004)?

Does deep overturning, the shallow overturning or the lateral wind-driven gyres dominate the meridional transport of heat (e.g., Boccaletti et al. 2005; Talley 2003)?

Where are the pathways of upwelling in the global scheme of the MOC (e.g., Sarmiento et al. 2004; Hallberg et al. 2006)?

How are the cycles of heat and fresh water transport coupled, and how is their dynamical impact on the MOC measured (e.g., Stommel and Csanady 1980)?

Do the zonally integrated overturning streamfunction and its thermal analogues adequately measure the MOC, or can a more penetrating definition be made by analyzing transport across sections and transformation within boxes, on the potential temperature/salinity plane (e.g. Lumpkin and Speer 2000; Fox and Haines 2003; Marsh et al. 2006; Bailey et al. 2005)?

What is the relative importance of the Southern Ocean and the northern Atlantic sinking regions (e.g., Toggweiler and Samuels 1995)?

Some things are not in doubt:

- The existence of ‘maritime climates’ downwind of the major oceans
- The oceanic moisture source for the entire atmosphere
- The contribution of latent heat associated with this moisture to the heating of the atmosphere
- The presence of storm tracks over the northern Pacific and Atlantic, which channel atmospheric meridional transports of heat and freshwater in these sectors
- The presence, movement and impact on atmospheric climate of sea-ice, in response to atmosphere and ocean circulation and temperatures (see Rhines 2006, for a non-technical discussion)

At the most basic level we are reminded that the ocean is the *dominant* global reservoir of mean thermal energy, water, carbon, anthropogenic thermal energy and is a *significant* reservoir of anthropogenic carbon, primary biological production and respiration. The imprint of physical circulation on the global distribution of ecosystems is widely apparent, and Schmittner (2005) argues that major disruption of the Atlantic MOC will greatly impact ecosystems and global productivity.

Northern Atlantic climate itself involves several nested questions:

- What is the impact of oceanic heat storage on warming the wintertime atmosphere?

- Does northward heat transport by the ocean circulation greatly increase this warming?
- What are the secondary effects through the cryosphere, of the ocean circulation?
- What is the impact of ocean circulation on the climatological mean, the seasonal cycle, and the decade-to-century variability?
- What is the level of dynamical feedback between ocean and atmosphere in the wintertime Atlantic storm track?
- What is the effect of oceanic heat transport on the development of individual cyclonic systems?
- While such general questions can be mind-numbing they come alive when made specific:
- What keeps the Barents and Labrador Seas ice free?
- What caused the 1920s–1930s warming that engulfed the northern Atlantic Ocean and atmosphere and affected ecosystems widely?
- Is explosive cyclogenesis responding to the Gulf Stream front?
- Will the global MOC weaken significantly in the next few decades, as the majority of IPCC climate models predict?
- What are the dominant fresh-water pathways and their impact on deep water formation in the subpolar Atlantic?
- Will the widely predicted (again, by the mean of many IPCC climate models) predominantly zonal bands of precipitation change under global warming and cause a greatly wetter western Europe, great freshening of the northern Atlantic and Arctic, and stronger drought in the subtropical regions of descent in the atmospheric MOC?

Observations required to address these many questions have historically been sparse. Direct and indirect measurement of oceanic heat flux and storage requires time- and space series that have only gradually approached adequate resolution and sustained duration. Fortunately, promising new technologies bearing on thermodynamics are now available. There are multiple ways to constrain ocean–atmosphere heat transport and exchange with the atmosphere, through air–sea flux measurements and bulk-formulas based on wind speed and temperature difference, observations of atmospheric lateral flux, of top-of-the atmosphere radiant flux, of water-column lateral transport, and of water-column heat storage and of regions of water-mass formation and sinking. Such direct and indirect methods are summarised by Bryden and Imawaki (op. cit.). Of particular note is the recent capability provided by ARGO float hydrography and satellite altimetry, which together measure the steric and dynamic height of the oceanic water column (Willis et al. 2003; Hadfield et al. 2007). Inference of air–sea heat exchange by ingenious dynamic use of veering of the thermal wind velocity with depth is also promising (the 'cooling spiral' of Stommel 1979). By contrast key sources, sinks and transport pathways of fresh water for the ocean circulation and key sites of global ocean upwelling are far less well observed.

Sophisticated analyses of observations and their assimilation into climate models are aimed not only at refining the numbers, but they also can tell us whether our most basic picture of the workings of the MOC are correct. Two 'back of the envelope' calculations suggest the importance of the Atlantic MOC in the poleward transport of heat and freshwater.

First, heat transport: 16 Sv (Sverdrups or megatonnes second^{-1}) of mass transport (call it F_m) in the Atlantic MOC with a temperature difference of 15 K between upper, northward and deep, southward flowing branches yields a meridional heat transport, $\rho C_p \Delta\theta F_m$, of amplitude 1.0 pW ($1.0 \times 10^{15}$ W), which is comparable with results from both direct and indirect methods (e.g., Bryden and Imawaki 2001; ρ is density, C_p the specific heat capacity at constant pressure, $\Delta\theta$ the potential temperature difference).

Second, fresh-water transport: the same schematic 16 Sv of Atlantic oceanic mass transport is more saline in its northward, upper ocean flow and less saline in its deep Equatorward flow. This difference implies the low latitude evaporation and high-latitude precipitation and runoff which balance a compensating poleward atmospheric fresh-water flux.. The global, east–west integrated value of the water vapor transport, northward in atmosphere, returned southward in the oceans, is fairly convincingly estimated to have a peak value of 0.8 Sv. at about 40° N latitude, representing roughly 2 pW of latent heat transport (e.g., Trenberth and Caron 2001). Wijffels (2001) describes oceanographic determination of the southward return flow of 0.8 Sv of fresh water (riding on top of the net Arctic throughflow communicated through Bering Strait). The Atlantic fraction of this flux is estimated to be roughly 0.4 Sv of fresh water transport difference between 10° N and 50° N. A simple 'box-model' MOC would have a net fresh water transport ½ $(\Delta S/S)F_m$, where ΔS is the salinity difference between upper and deep branches of the flow. Observed ΔS of order 1 psu relative to a mean of 35 psu would support a transport of only 0.23 Sv fresh water, seemingly smaller than observed. Yet the Atlantic also exports moisture westward to the Pacific in the Trade Winds. LeDuc et al. (2007) cite 0.13–0.37 Sv of fresh-water jumping over Central America, which helps to explain this discrepancy. The horizontal-gyre component of the Atlantic circulation above the thermocline also contributes to the equatorward 0.8 Sv of oceanic fresh water transport. This simple reasoning is an example of transport played out on the potential temperature plane, first exploited by Stommel and Csanady (1980).

Some of the relatively new observational resources applicable to these questions are:

- Satellite radiation observations, for example, the ERBE and CERES sensors begun in 1984, and infrared sea-surface temperature measurements, recently extended to long-wave bands which see through cloud cover (AMSR-E sensor).
- Satellite altimetry by NASA Topex/POSEIDON/JASON instruments and European Space Agency instruments since 1992 providing global coverage of sea-surface height, which has a strong contribution from ocean water-column heat storage.
- Satellite scatterometer surface wind-fields, applicable to air–sea momentum and heat fluxes.

- Steady improvement in atmospheric circulation reanalyses providing essential detail for evaluating atmospheric and, as a residual oceanic, heat and moisture fluxes.
- Enhanced ocean observation programs (e.g., WOCE, RAPID and ASOF) targeting key ocean sections with repeated high-resolution hydrography.
- The ARGO float program, now approaching its goal of 3,000 drifting, hydrographic profiling floats in the world ocean.
- Robotic gliders directed to survey key hydrographic sections, boundary currents and convection zones.
- The historic data base of XBT and hydrographic temperatures, mined to reconstruct ocean heat storage time-series (e.g., Levitus et al. 2005).
- Ocean surface flux moorings.
- Deep-sea moorings in key boundary currents providing semi-quantitative mass transports.

  Chemical tracer programs, especially CFCs, tritium, radiochemical effluents, carbon and standard nutrients and oxygen, providing quantitative estimates of some of the most difficult elements of the global circulation, particularly the global upwelling sites, diapycnal mixing rates, long-distance boundary current transports, and formation of water masses in 'stable' gyre centers. Tracers also measure air–sea interaction rates in their own way, and can constrain heat- and freshwater exchange across the sea surface.

4.2 The Contribution of the Atlantic Ocean Circulation to Wintertime Climate

The importance of ocean circulation to atmospheric climate has been challenged by Seager et al. (2002), hereinafter 'SO2', in their paper, 'Is the Gulf Stream responsible for Europe's mild winters?' While centering attention on the mild climate of Europe, their work, if correct would have greater consequences. They argue that:

(i) Only a small portion of the total northward heat transport north of 40° N, is accomplished by the ocean in comparison with the atmospheric heat transport.
(ii) Oceanic heat storage is local, with the summer's heating of the mixed layer being the dominant source of wintertime oceanic heat release to the atmosphere, with little contribution from oceanic heat transport.
(iii) Fresh-water transport coupled with heat transport can be neglected.

Here we show that while (i) is true it is misleading, (ii) is based on an analysis which is in error due to comparison of ocean heat transport and surface heat loss on different timescales. (iii) they have missed the most important climate interaction of all.

One could hardly argue against the persuasive reasoning provided by SO2 that the maritime climate maintains the temperature contrast between North America and Europe. However, their conclusions regarding the impact of the oceanic heat transport could be taken to mean that oceanic heat transport has no significant consequence for the climate in Europe and elsewhere, beyond a minor warming of 0–3 °C.

We address points (i)–(iii) in order.

(i) Satellite radiation measurements combined with atmospheric observations assimilated into models give estimates of total, atmospheric and oceanic meridional heat flux. One recent analysis elevates the atmospheric contribution somewhat, the atmospheric transport peaking between 4 and 5 PW (4 to 5 $\times 10^{15}$ W), while the ocean transport peaks at about 2 PW (Trenberth and Caron, op. cit.). However, as Bryden and Imawaki op. cit. emphasize (using transport estimates of Keith 1995), the meridional heat flux is comprised of *three* nearly equal (in amplitude) contributions from latent- and sensible heat flux (the latter known as dry static energy flux) in the atmosphere and sensible heat flux by the ocean. Latent heat is fresh water (2.4 pW per Sverdrup), and its transport is an intrinsically coupled ocean/atmosphere mode. Keith's transports, or the more recent transports, quantitatively similar, by Trenberth et al. op. cit., plotted against latitude show the dominance of subtropical ocean evaporation (typically 1.5 m year^{-1}) in driving the global system; this activity lies poleward of the transition from tropical Hadley circulation to the latitude of the midlatitude, eddy-driven jet stream. Each of the three modes of meridional energy transport has peak amplitude of roughly 2 pW, with the latent-heat mode carrying 0.8 Sv of fresh-water northward, mirrored by equatorward ocean transport. The moisture/latent heat pump of the Atlantic storm track is a crucial part of maritime climate. It is ignored in the thermodynamic discussion of SO2. The Trenberth and Caron, op. cit. discussion uses ERBE top-of-atmosphere radiation data and atmospheric observations/assimilation. Although few error estimates are presented, the occurrence of large heat flux divergence over land is suggestive of significant error.

Wunsch (2005) argues that in fact ERBE radiation observations add significant uncertainty, and provides an error analysis. By taking the ocean observations of heat transport and calculating the atmospheric heat transport as a residual, his analysis revises downward the atmospheric heat transport in the northern hemisphere. The maximum atmospheric transport now averages 4.1 pW (ranging between 3 and 5.2 pW at one standard deviation). Wunsch's analysis also gives a greater ocean transport at high northern latitudes than do Trenberth and Caron.

(ii) If the ocean (here, the Atlantic Ocean) participated in climate only through local, seasonal heat storage and release in the shallow mixed layer, then calculations of ocean circulation would be unnecessary for climate models. Indeed, SO2 state in their abstract that "..the majority of heat released during winter from the ocean to the atmosphere is accounted for by the seasonal release of heat previously absorbed and not by ocean heat flux convergence." This conclusion follows from their comparison of the *annual mean* oceanic heat transport convergence with the *wintertime* release of heat at the sea surface, the latter being much larger. Both are inferred using climatological mean surface heat flux fields for the Northern Atlantic, developed from the COADS ship observation dataset by da Silva et al. (1994). Values they estimate (averages north of 35° N) are 37 W m^{-2} heat convergence by the ocean circulation, vs 135 W m^{-2} wintertime heat release from surface observations. This is based on the estimate of 0.8 PW northward oceanic heat transport at 35° N.

The air/sea flux affecting oceanic water-column heat balance includes downward short-wave radiation (corrected for albedo related reflection), net long-wave radiation, sensible heat flux and latent heat flux. The air/sea flux affecting the atmosphere differs from this by the downward short-wave solar radiation, which heats the ocean but does not cool the atmosphere. Thus the maps of air–sea heat flux that matter for the atmosphere show much larger numbers than those we are familiar with, for the ocean. However we want to compare the air/sea heat flux with that of a mixed-layer-only, climatologically steady world, in which no lateral heat transport is allowed, and in this mixed-layer ocean the annual average flux vanishes. Thus to consider the non-seasonal, non-local heat storage and forcing of the atmosphere, the full air–sea heat flux including solar radiation is the relevant field.

Let us assume all these numbers in the paragraphs above are accurate. A model of the annual cycle would include year-round northward heat flux by the Atlantic circulation, *together with its release to the atmosphere in a few winter months*. During summer warming none of this deep heating escapes to the atmosphere. We thus should be comparing the time-averaged heat-flux convergence by the ocean circulation, *multiplied by the ratio 12/(number of months of wintertime heat loss)*, with the upward heat flux at the sea surface observed during those winter months, or else simply annualize all the fluxes. The details depend upon the vertical distribution of the north–south heat advection (referenced to the late winter mixed-layer temperature). Using an estimate that half of the transport lies deeper than 100 m (above which depth most of the local, seasonal heating is trapped), suppose we release that heat in 3 winter months and release the other half from the upper 100 m during 6 months of the year. The surface heat flux during winter becomes augmented by a factor $12/6 \times ½ + 12/3 \times ½ = 3$. Multiplying 37 W m^{-2} from the SO2 estimate by 3 gives 111 W m^{-2}, enough to account for much of the observed winter upward heat flux at the sea surface (135 W m^{-2}). This argument shows that oceanic heat advection is plausibly important in warming the atmosphere in winter. Note with a linear model of heat storage in a mixed-layer-only ocean, SO2's procedure would be correct, for the laterally advected heat would be 'available' to the atmosphere in all seasons. The point is that much of it is in fact sheltered below the seasonal mixed layer during the warm months.

The same, or even more dramatic, result follows if we take air–sea heat flux climatology, with monthly surface heat flux averaged in the Atlantic north of 25° N, and integrate the flux with respect to time, Fig. 4.1. Start in spring, when the net surface flux changes sign and begins to warm the upper ocean; then integrate forward. In regions with annual average heat flux that is zero or upward, the integral

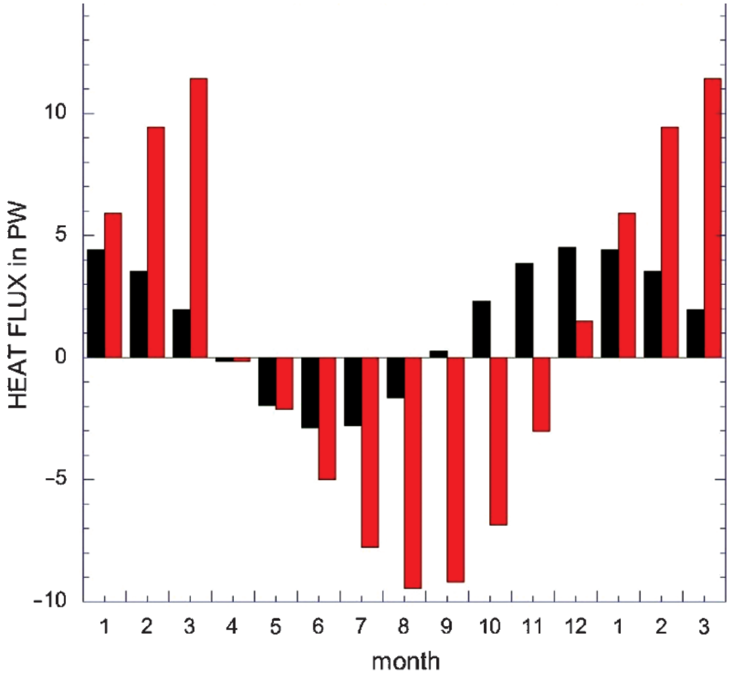

Fig. 4.1 Using the da Silva et al. (1994) air–sea heat flux estimates (black bars) we integrate forward in time (red bars), averaging over the Atlantic north of 25° N. When the integral returns to zero, the local, seasonal heating has been removed by autumnal cooling. On average, by early December the local heat source is exhausted and for the remainder of the winter oceanic warming of the atmosphere relies on heat imported by the ocean circulation. Positive values indicate heat loss from the ocean to the atmosphere. The NOC1.1a heat-flux climatology gives a very similar picture. The ordinate labels refer to the black bars and should range from -1.0 pW to 1.0 pW

will eventually come back to zero, indicating that the locally stored summer's heating has been removed by cooling from above. After this date, continuing upward heat flux must have been imported by the ocean circulation. Averaging north of 25° N latitude (and keeping north of the zero-mean air–sea heat flux line) in the Atlantic we see that by early to mid-December, the locally stored heat is exhausted, and excavation of imported heat dominates the rest of the winter. The geographical distribution of the year days is shown in Figs. 4.2a for NOC/SOC climatology

Fig. 4.2 (continued) subpolar Pacific. Deep red regions (year days >400) the heat balance is local, without significant lateral advection by the ocean circulation. Contour interval: 20 days. (b) Annual mean air–sea heat flux felt by the oceans (short-wave radiation, long-wave radiation, sensible- and latent-heat fluxes), from NOC1.1a data. The total upward heat flux felt by the atmosphere is this map without the net downward short-wave radiation, hence with much larger upward flux. This figure, however, represents the non-local heating of the atmosphere owing to the ocean circulation. Maximum values exceed 150 Wm^{-2} in the Sargasso Sea where roughly 0.5 pW of upward heat flux occurs in winter. Contour interval: 20 Wm^{-2}, zero contour bold black. (c) North Atlantic surface heat flux annual cycle ($W\ m^{-2}$) against year-day at two longitudes: 60° W (cyan) and 65° W (yellow), plotted from the Equator to 60° N. The curves with strongest upward (negative) wintertime heat flux in winter are in the Gulf Stream extension, ~40° N. The integrals of these curves produce the year-day when local seasonal heat storage is exhausted (Fig. 4.2a)

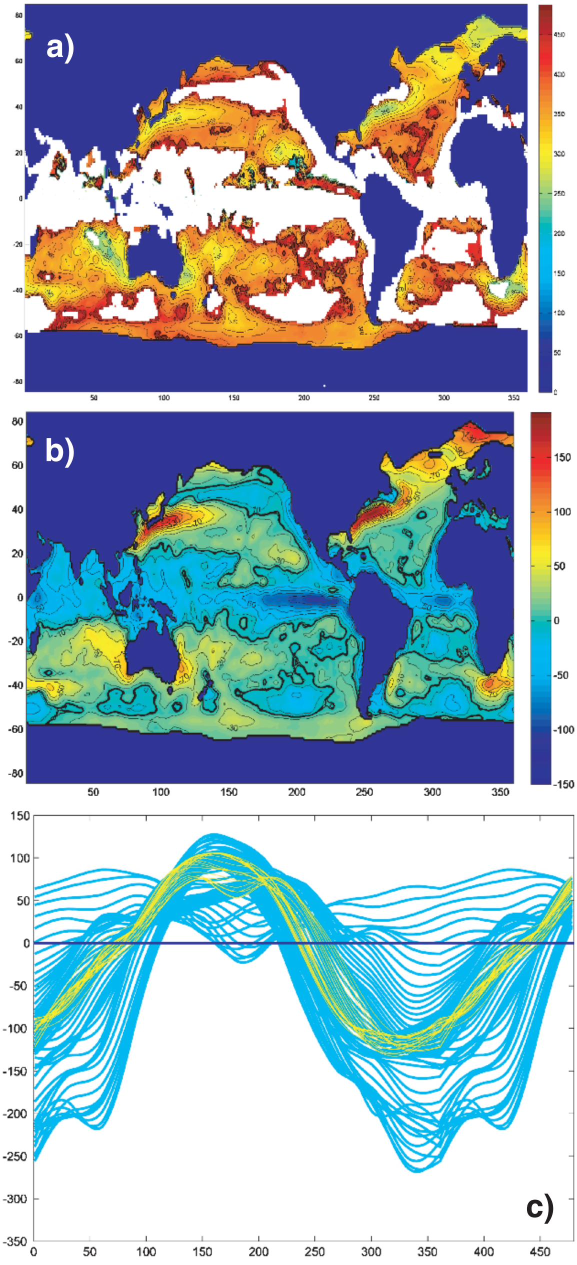

Fig. 4.2 (a) Year-day when local seasonal ocean heat storage has been exhausted by winter cooling. The northern and western Atlantic, Barents, Nordic and Labrador Seas fall in the range, day 225–350. (In the southern hemisphere 180 is subtracted from the year day so that seasonal color pattern is the same as in the northern hemisphere.) Thereafter in much of the late fall and through the winter, warming of the atmosphere by the ocean depends on imported heat flux by the lateral ocean circulation. White regions of annual-mean downward heat flux are never exhausted by wintertime cooling, and heat is exported from them by the ocean circulation. Based on NOC1.1a data. Regions of strong effect of ocean circulation on the atmospheric heat budget appear both east and west of Australia, in the Kuroshio and broadly in the subtropical Pacific, a small region of the

(Grist and Josey 2003, now termed NOC1.1a flux climatology). daSilva/Levitus (1994) climatology yields very similar results (not shown). The Gulf Stream/Sargasso Sea region shows the strongest effect of heat advection by the circulation, with early exhaustion (by September) of the locally stored heat. Yet in a band extending northeastward to the Nordic and Barents Seas, heat flux convergence by the ocean circulation supplies as much heat as does local seasonal heat storage. Plots of the ratio of mean convergence of oceanic lateral heat transport, divided by mean downward solar radiation, show the same northwest Atlantic region, where the contribution from the ocean circulation is significant. Generally speaking, there are large areas of the world ocean within the upward-mean-heat-flux regions, in which the locally stored seasonal heating is insufficient to provide more than half of the upward fall/winter heat flux. The net annual heat transport at the sea surface from NOC1.1a data is shown in Fig. 4.2b, where we include the solar short-wave radiation. The total oceanic warming of the atmosphere omits this term and hence is much larger. It includes contributions from both non-local advection of heat and local re-remission of some of the previously gained solar energy in the form of longwave, latent and sensible heat loss. The construction of Fig. 4.2a is perhaps made clearer by looking at the annual cycle of oceanic heat balance at two longitudes, Fig. 4.2c. Here the deep negative values correspond to the Gulf Stream extension region, where wintertime heat loss exceeds 300 W m^{-2}.

These ideas are all subject to accuracy of the consensus oceanic heat transports, and analysis of air–sea heat flux feedbacks due to the ocean circulation-induced SST. Improvement will occur when water-column heat storage observations become numerous enough. Indeed, wherever winter mixed layers exceed 50 –100 m in depth, we infer that ocean circulation is important, because seasonal surface heating cannot mix down deeper than this, even with the aid of the winds. This is a strong argument for sustained time-series observations of temperature and salinity as can be provided by floats, gliders and moorings. Several parallel arguments given in the SO2 paper, and a similar one given by Wang and Carton (2002) suffer from the same logical error pointed out here, for example when geographical distribution of winter air–sea heat flux is compared with annual-mean heat convergence by the ocean circulation.

SO2 remark also that the wintertime poleward heat transport in their calculation is much reduced in mid-latitude, and attribute this to southward transport in the shallow wind-driven Ekman layer. Other estimates of Ekman heat transport do not support such a large effect, and it is more likely that what they are seeing is the huge (0.5 pW) upward heat flux in the Gulf Stream/Sargasso Sea region which dominates Figs. 4.2 in subtropical latitudes. This upward heat flux reduces the wintertime poleward ocean heat transport, and is a part of the essence of our argument.

There are subjective elements in the model simulations of SO2. With suppressed ocean circulation their models show surface winter temperature changes of 6–12 °C over much of northern Eurasia, reaching 21 °C in Scandinavia. The average temperature change north of 35° N is 6 °C in their GISS-model. We would call these changes ‘large’; yet SO2 argue that they have “little impact”. The great differences apparent between their two simulations (one with a non-dynamical ice model, the other without any ice model) remind us of the complexity and uncertainty of coarsely resolved climate model results, when so many critical high-latitude and upper ocean physical processes are under-represented. And, more to the point, surgical removal of oceanic heat transport has other implications (iii, below).

While this has been a discussion of mean and seasonal wintertime heat balance, some aspects apply also to decadal and secular variability. The warming of northern Asia associated with greenhouse forcing, yet partially associated with strong positive phase of the North Atlantic Oscillation in the early 1990s, is shown by Thompson and Wallace (2001) to involve zonal heat advection: we take this to be a sign of Atlantic oceanic heating penetrating farther eastward over Asia.

(iii) The SO2 model experiments use oceanic mixed layer models which are only governed by heat exchange with the atmosphere and by (diagnosed) heat transport related to the oceanic MOC. This type of model ignores fresh-water flux and fresh-water transport which are known to play an important role in inhibiting heat release from the ocean and determining sinking regions of the meridional overturning circulation (MOC) at subpolar and polar latitudes: Too much fresh water at the surface stabilizes the water column and sea ice can form, changing fundamentally the seasonal cycle of heat exchange between the ocean and atmosphere. On spatial scales beyond the convective regions, the fresh water cycle and heat transport are coupled globally in the atmospheric latent heat flux, and through the thermohaline circulation as was first discussed by Stommel and Csanady (1980). This coupling is played out on the θ–S plane, which is the fundamental ‘phase plane’ of physical oceanography (e.g., Bailey et al. 2005). It is summarised in maps of integrated buoyancy, integrated from the surface downward (essentially upside-down dynamic height), which we can call ‘convection resistance’, C_R:

$$C_R(x,y,z_1) = g\int_{z_1}^{0} (\sigma_0(x,y,z) - \sigma_0(x,y,0))dz$$

where σ_0 is surface-referenced potential density, g is gravitational acceleration, and z is vertical coordinate. This quantity shows the amount of buoyancy that must be removed by air–sea interaction in order to convectively mix the water column to a depth z_1. Maps and sections of C_R (Bailey et al. 2005), Fig. 4.3, can be split into its respective salinity and temperature components, assuming an approximately linear equation of state. These maps illustrate how much influence over water-mass formation is provided by thin upper-ocean layers with low salinity. In the Labrador Sea, for example, Hátún et al. (2007) argue that fresh water advected off the west Greenland boundary currents and continental shelf control the geographic distribution of deep convection in winter. Similarly, Häkkinen et al. (2007a) map, for the Greenland–Norwegian seas, the contributions to upper ocean density from salinity and temperature, show-

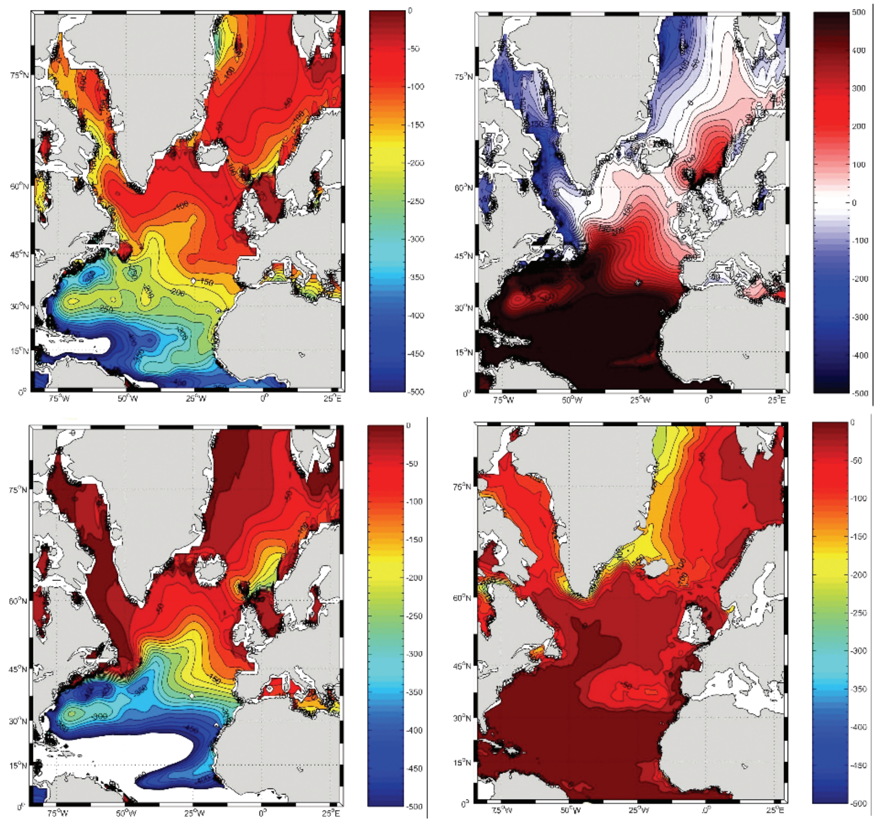

Fig. 4.3 Convection resistance, C_R, in the North Atlantic, showing the integrated density anomaly, relative to the sea surface in winter, integrated from the surface to 500 m depth. *Upper left*: total C_R in units of ppt m; *Lower left*: contribution of thermal stratification to C_R; *lower right*: salinity contribution to C_R; *upper right*: the *difference* between thermal and haline contributions to C_R. In blue regions of the upper right panel salinity stratification dominates, while in red regions temperature stratification dominates buoyant stability of the water column; for shallower depths, z_1 (not shown) upper ocean low-salinity layers more extensively dominate the northwest Atlantic

ing the strong imprint of surface fresh water advection near Greenland, and temperature advection near Norway. These distributions of upper ocean buoyancy control where deep convection occurs in winter, and hence where water-mass formation occurs; yet they are not likely to be modeled well by current climate models.

The global hydrologic cycle has a familiar pattern of high precipitation and runoff at high northern latitudes, evaporation in subtropical oceans, and narrow bands of evaporation and precipitation associated with the ITCZ. A net flux of fresh water from high northern latitudes to the low latitude evaporation sites is needed, even after river pathways are accounted for. The thermohaline MOC provides the return circuit for atmospheric vapor transport. In the North Pacific, low salinity stabilizes the surface layer of the subpolar gyre and there are no truly deep sinking regions. A shallow salinity minimum guided and subducted by the wind-driven Ekman transport, reaches toward the tropics. Yet much of the excess precipitation seems to escape through the Arctic (with the Bering Strait throughflow carrying low-salinity Alaskan coastal current water as well as water from midocean, Woodgate et al. 2006). The robust MOC in the Atlantic illustrates how the θ–S diagram couples the heat- and fresh-water transports, and involves both subpolar and Arctic water-mass transformations. The pioneering study of Stommel and Csanady (1980) gave simple two-degree of freedom illustrations of the nature of these coupled transports. They estimated the northward mass transport of salty waters and the compensating mass transport of less salty deep water using the observational estimates of heat and fresh water transport and water mass properties for the latitudes 40–45° N.

We wish to develop the 'back-of-the-envelope' calculation in the introduction, and reiterate the conclusions of Stommel and Csanady to show that heat transport and fresh water transport are intimately coupled. Removal of only one of them renders the problem meaningless. We consider a two-layer box model of the polar and subpolar oceans bounded by the Bering Strait and 45° N, using information of the 'known' mass fluxes at the surface, river runoff and at the Bering Strait. From the conservation of salt and fresh water we can diagnose the overturning to satisfy the equilibrium conditions and at the same time diagnose the heat transport when the upper-lower layer temperature difference is given. The following computation is done using the definitions of Wijffels et al. (1992) for fresh water and salt transports. For simplicity we assume densities to be 1,000 kg m^{-3} for the ocean and river and P–E fluxes. The inflow (Vbe) of the Bering Strait is 0.8 Sv with salinity (Sbe) 32.5 ppt. P–E flux over the area from the Bering Strait to 45° N is about 0.1 Sv and the runoff (R) from land in the same region amounts to about 0.19 Sv. At 45° N we want to solve the average flow (Vo) and the baroclinic transport V (all velocities are defined positive southward). The upper layer Atlantic salinity (Sa) is 35.3 ppt and the bottom layer salinity (Sb) is 34.9 ppt. The conservation equations for salt and fresh water are:

$$\text{Salt: Vbe Sbe} = (-V + Vo/2)\,Sa + (V + Vo/2)\,Sb$$

$$\text{Fresh water: Vbe}\,(1-Sbe) + R + P\text{–}E = (-V + Vo/2)\,(1-Sa) + (V + Vo/2)\,(1-Sb)$$

Substituting the above values in the conservation equation gives, for Vo and V, 1.09 Sv and 30.65 Sv, respectively. This simple scheme illustrates the thermohaline nature of the fresh water redistribution, where the northward mass transport in the upper layer is 29.04 Sv and the southward transport in the bottom layer is 30.11 Sv. If the temperature difference between the upper and lower layer is 8 °C, the northward heat transport would be about 0.9 PW at 45° N, which is close to the current estimates of ocean heat transport at 40° N representative of the present climate (e.g., Bryden and Imawaki 2001). Thus based solely on conservation of salt and fresh water, with hydrographic data we can diagnose the overturning and the associated heat transport to satisfy the equilibrium conditions when the various fresh water fluxes of the present climate are given. This traditional overturning picture of meridional heat transport does conflict with recent arguments suggesting that the deep branches of the MOC are unimportant for transporting heat (Boccaletti et al., 2005); their argument continues to depend on a particular choice of reference temperature, which has little effect on the arguments given here.

We have arrived at the crux of the problem not considered in the numerical experiments of SO2. Removal of the oceanic heat transport due to the thermohaline circulation means also that the redistribution of the fresh water is blocked which in the real world would lead to accumulation of fresh water at the high latitudes. The lack of the thermohaline circulation intensifies freshening because no salt is transported northwards. Fresh water accumulation will eventually build an extensive sea ice cover north of 40° N and influence the seasonal uptake of heat in the ocean. This is consistent with the paleo-records showing that periods of extensive ice cover over the high latitude ocean, and over the European and North American continents, were associated with weak production of North Atlantic deep water (Boyle and Keigwin 1982, 1987) and thus a weak thermohaline circulation. So in fact during the height of the last glaciation, the maritime effect was reduced to a minimum, and the temperature gradient across the Atlantic vanished.

In summary, accounting for the fresh water accumulation at the high latitudes alters significantly the picture suggested by climate models that would neglect the oceanic MOC: It is the existence of the oceanic heat transport that allows the maritime effect to operate in the northern North Atlantic and to create a milder European climate than in the North America; without the heat transport, ice would likely extend over much greater areas of ocean and land. Since the northward heat transport and southward fresh water transport in the Atlantic are strongly tied together, removing oceanic heat transport influences the climate and atmospheric circulation in ways that are not possible to simulate with a simple mixed layer model coupled to an atmospheric model. This also suggests that use of this type of model with a fixed oceanic heat transport (today's climate) is not suitable to describe climatic states where the thermohaline circulation is expected to change significantly from the present, as might happen for instance in doubled CO2 scenarios where the fresh-water input at high latitudes can increase by 40% or more (Manabe and Stouffer 1994). The signature of oceanic heat transport is deep convective mixing in winter, which accesses energy well below the ~50–100 m penetration of local summertime warming. Improved global mapping of winter mixed-layer depth using ARGO, XBT lines and other water column observations should go far toward identifying these regions.

Removal of one piece of a complex machine (here, the oceanic heat transport) can have unforeseen consequences. We have pointed out some, and there may be others, such as effects on cloudiness, atmospheric standing waves and storm tracks. The conclusion of SO2 that the particular climate feature of interest, the warming of western Europe, is 'fundamentally caused by the atmospheric circulation interacting with the oceanic mixed layer', and thus 'does not require a dynamical ocean' is flawed in the three aspects described above.

4.3 Atmosphere: Ocean Fluxes of Heat and Freshwater

4.3.1 Currently Available Estimates

Estimates of the ocean–atmosphere fluxes of heat and freshwater are available from a number of sources. Gridded monthly mean surface heat flux datasets were first produced from voluntary observing ship and buoy meteorological observations using a bulk formula approach (e.g., Bunker 1976; da Silva et al. 1994; Josey et al. 1999). More recently, atmospheric model reanalyses have provided an alternative, widely used source of flux estimates, the two principal datasets being the NCEP/NCAR (Kistler et al. 2001) and ECMWF reanalyses (Uppala et al. 2005). Attempts are now also being made to produce flux datasets by applying the bulk formula approach to combinations of reanalysis and satellite based meteorological fields (Yu and Weller 2007). Indirect estimates of the net air–sea heat flux have also been obtained using residual techniques that employ top-of-the atmosphere radiative flux measurements from satellites and estimates of the atmospheric flux divergence from reanalyses (e.g., Trenberth and Caron 2001).

Precipitation estimates are also available from the reanalyses and, for 1979 onwards, from satellite observations. The Global Precipitation Climatology Project Version-2 (GPCPV2) Monthly Precipitation Analysis dataset (Adler et al. 2003) incorporates precipitation measurements from satellite and rain gauges which are merged in an analysis that retains the best features of each dataset. The resulting dataset is independent of the reanalyses and is the leading satellite/rain gauge based set of precipitation fields currently available. Note, however, that there remain large differences between the various precipitation datasets and thus the freshwater flux field is more poorly determined than the net heat flux.

Significant differences exist between the various datasets in many regions of the ocean and these reflect the difficulty in obtaining accurate estimates of the fluxes. Specific problems include:

(i) Poor sampling in regions away from the major shipping lanes (e.g., the high latitude North Atlantic (see Josey et al. 1999, Fig. 4.2; see also Gulev et al. 2007).
(ii) Uncertainty over the values of the transfer coefficients which appear in the bulk formula for the sensible and latent heat fluxes (although significant progress has been made with the development of the COARE algorithm, Fairall et al. 2003).
(iii) Differences in the spatial and temporal averaging methods used to produce the gridded flux fields.
(iv) Poor representation of clouds in the atmospheric models used for the reanalyses which can have a major impact on shortwave and longwave flux estimates (e.g., Cronin et al. 2006).

A detailed review of flux estimation techniques and associated sources of error is provided in the report of the WMO/SCOR Working Group on Air–Sea Fluxes (WGASF 2000).

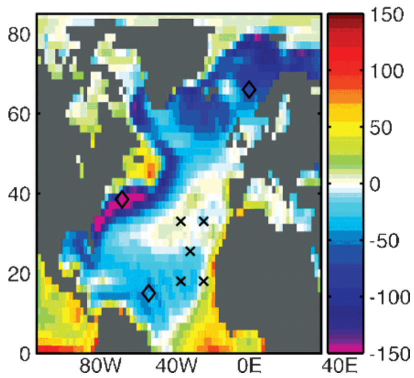

Fig. 4.4 Annual mean heat flux field from NCEP reanalysis for the period 1949–2001, in units of Wm^{-2}. Also shown are the presently maintained flux reference sites (black diamonds) and the NTAS and CLIMODE moorings and OWS M, and the earlier Subduction Experiment flux buoy array (crosses)

The main features of the net heat flux field in the North Atlantic are common to each of the various datasets currently available and are illustrated in Fig. 4.4, which shows the annual mean field from the NCEP/NCAR reanalysis (Kistler et al. 2001) for the period 1949–2001. In particular, there is strong net heat loss over the Gulf Steam region, of order 150 Wm^{-2} in the annual mean, with winter month averages (not shown) up to 400 Wm^{-2}, and a transition to heat gain at more southerly latitudes. The Nordic and Labrador Seas also experience strong cooling, but it should be noted that these regions are poorly sampled and thus the values there may be biased low as the reanalysis surface flux estimates are reliant to a certain extent on the assimilation of surface observations. Furthermore, they have difficulty (because of the relatively coarse spatial resolution in the atmospheric models employed) in representing small spatial scale features such as the central Labrador Sea and Greenland tip-jet that may be key to fully understanding the location and processes by which deep ocean convection occurs (e.g., Hátún et al. 2007; Lilly et al. 1999, 2003; Pickart et al. 2003). Use of subsurface observations of water-column heat storage are promising for the future, as ARGO, repeat hydrography lines and glider-based hydrography become more plentiful.

A measure of the uncertainty in the net heat flux field is provided by Fig. 4.5 which shows the variation with latitude in the North Atlantic of the zonal mean net heat flux for five recent climatological datasets: NCEP/NCAR, ECMWF, Trenberth residual (Trenberth and Caron 2001), NOC1.1a (formerly termed the adjusted SOC climatology, Grist and Josey 2003) and adjusted UWM/COADS (da Silva et al. 1994). There is some dispersion between the datasets with typical differences at the 20–30 Wm^{-2} level in the zonal annual mean; these differences are likely to be further amplified when monthly means for specific locations are considered. In the absence of high quality independent flux measurements it has not been possible to firmly establish the reasons for these differences and thereby narrow the gap between the different estimates. However, there is now the prospect for significant progress on this front as a result of the increasing number of moorings within the surface flux reference site array (also shown on Fig. 4.4). Of particular interest is the CLIMODE mooring deployed in November 2005 at 38.5° N, 65° W, which samples the strong

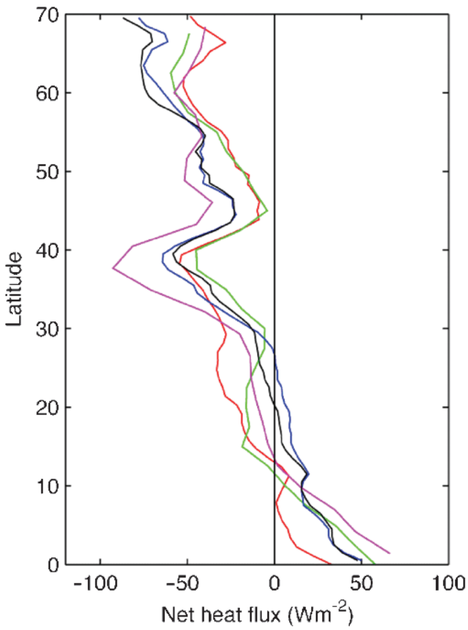

Fig. 4.5 Zonally averaged annual mean net heat flux in the North Atlantic for five different surface flux climatologies: NCEP/NCAR (red), ECMWF (green), Trenberth residual (magenta), NOC1.1a (black) and UWM/COADS adjusted (blue)

heat loss region towards the western boundary of the Atlantic which is a major source of uncertainty (Josey et al. 1999). Detailed analysis of the flux time series from this and other reference sites in the next few years is expected to firmly establish the causes of uncertainty (biased flux algorithms, differences in analysis procedures, sampling issues) and ultimately lead to more accurate flux estimates.

4.3.2 Evaluation Methods

Given the differences between the gridded flux datasets discussed above, and the advent of hybrid products obtained through various combinations of reanalysis, satellite and ship fields (Yu and Weller 2007; Large and Yeager 2004) together with flux fields from ocean synthesis (e.g., Stammer et al. 2004), a common method of evaluation is needed to provide a means by which their accuracy can be compared and potential biases identified. To this end, a set of guidelines for evaluation of flux products has recently been developed (Josey and Smith 2006). Previous studies have been limited by the availability of high quality reference observations which comprise both

(i) Local measurements of the fluxes from research buoys/vessels.
(ii) Large-scale constraints, principally estimates of heat and freshwater transports across hydrographic sections, from which regionally averaged fluxes can be inferred.

However, there has been a significant increase in the number of reference observations in recent years as noted above which will enable significant progress towards a more accurate picture of ocean–atmosphere interaction. Specific examples of flux evaluations using the limited amount of data available to date are now discussed.

(a) Comparisons with Local Flux Reference Data

Renfrew et al. (2002) found that in the Labrador Sea, NCEP overestimates the sensible and latent heat fluxes by 51% and 27%, respectively. They ascribed these biases to an inappropriate choice for the roughness length formula in the NCEP reanalysis under large air–sea temperature difference and high wind speed conditions. Thus, they were able to extend conclusions drawn from an analysis in a specific region to provide an indication of biases that are likely to arise in other regions experiencing similar conditions (e.g., the Gulf Stream and Kuroshio in winter).

(b) Evaluation Using Ocean Heat Transports

Hydrographic estimates of the heat and freshwater transport typically along zonal sections may be used to identify biases in net air–sea heat flux and net evaporation datasets by comparison with the climatologically implied property transport. Grist and Josey (2003) carried out such an evaluation of the heat transport for various datasets and Fig. 4.6 is an updated version of their Fig. 4.9a. Agreement within the error bars is typically obtained in the North Atlantic but the implied heat transport for the ECMWF dataset becomes unrealistically high in the South Atlantic. Further insight is obtained by considering regional differences using section pairs, as discussed by Grist and Josey (2003), which reveals that there is an underestimate of the ocean heat gain in the Tropical Atlantic in the ECMWF reanalysis. With a bit of oceanic chauvinism we point out that classic hydrographic ocean observations first analyzed by Hall and Bryden (1982) at 24° N in the Atlantic so accurately portrayed the MHT of the ocean that the atmospheric scientists were forced to re-evaluate the atmospheric MHT upward by almost 50%.

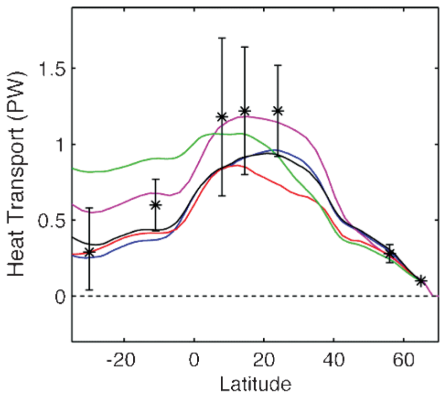

Fig. 4.6 Climatologically implied ocean heat transport (in pW, 10^{15} W) in the North Atlantic for five different surface flux climatologies: NCEP/NCAR (red), ECMWF (green), Trenberth residual (magenta), NOC1.1a (black) and UWM/COADS adjusted (blue). Hydrographic estimates are indicated by star symbols with error bars

4.3.3 Specific Mid-High Latitude Regions

(a) North Atlantic Subpolar Gyre. The subpolar gyre has shown significant decadal variability in both gyre strength and the salinity of the major water masses. Long-term freshening of the gyre from the 1960s through to the mid-1990s as part of a wider pattern of change in the freshwater balance of the Atlantic is now well documented (e.g., Curry et al. 2003). This freshening has recently been linked to increases in net precipitation over the ocean, river input, ice attrition and glacial melt over a broader domain including the Arctic Ocean (Peterson et al. 2006). The change in precipitation is driven partly by the multidecadal upward trend in the North Atlantic Oscillation (NAO) although detailed analysis of the eastern gyre region has shown that the second mode of sea level pressure, the east Atlantic Pattern also plays a significant role (Josey and Marsh 2005). A few regions of intense deep convection have been studied with dedicated observations over many years (e.g., Lilly et al. op. cit.). Gyre fluctuations have been inferred from satellite altimetry and hydrography (Häkkinen et al. 2004; Hátún et al. 2005).

The freshening trend appears to have partly reversed over the last decade from about 1995 onwards as the water entering the Nordic Seas from the gyre has become more saline (Hátún et al. 2005). This increase in salinity may reflect an increase in the amount of subtropical gyre water being advected north as a result of a weakening of the subpolar gyre (Häkkinen et al. 2004, 2007a, b).

(b) Nordic Seas. Decadal variability in the surface forcing of the Nordic Seas is particularly difficult to quantify given the lack of observations in this region. The major influence is likely to be the NAO as observations of deep convection in the Greenland Sea show a strong anticorrelation with the NAO (Dickson et al. 2000). Convective activity was particularly strong in the 1970s during which time the NAO was predominantly negative leading to enhanced heat loss in the Greenland Sea. The relative roles of heat loss and wind stress in controlling deep convection and subsequent variability of the deep outflows to the Atlantic remains to be fully established. Grist et al. (2007) find from a coupled model analysis variations in the Denmark Strait transport of up to 30% in response to Greenland Sea heat flux variability. However, other studies have suggested that variations in the wind field are the dominant factor controlling the overflow (e.g., Biastoch and Käse 2003).

4.4 Conclusion

Evaluation of the meridional transports of heat and fresh-water in the ocean, using several independent means, suggests that fundamentals of Earth's climate are indeed responsive to the ocean circulation, in the mean and seasonally, and likely (though not discussed here) at decadal time-scales. The degree of active feedback between ocean and atmosphere in each case is still controversial, yet will be refined by rapidly improving models and observations. The upward heat flux at the sea surface, in places reaching wintertime averages of hundreds of W m^{-2}, and exceeding the net solar radiation at the surface (annual mean of order 100 W m^{-2}, and far weaker in winter), is a significant contribution to atmospheric climate. Both decadal variability and persistent global warming have the potential to alter these heating and moistening patterns greatly. Indications of an increasing hydrologic cycle are already documented (e.g., Liu and Curry 2006; Curry et al. 2003).

The imprint of oceanic upward heat flux, and its enhancement by the ocean circulation, on the atmosphere is so apparent in diagnosed diabatic heating maps for the atmosphere (e.g., Held et al. 2002), in the existence of ice-free ocean at high latitude, like the Norwegian, Barents and Labrador Seas, and in the general warmth and moisture content of the storm track winds that it is inconceivable that climate models could neglect it. Meridional energy-transport is a strongly interacting collaboration of warm, moist storm track winds and warm underlying ocean. The fresh water cycle, coupled with the heat flux, involves massive subtropical evaporation from the oceans, followed by poleward transport in the storm track circulations, demonstrably concentrated over the oceanic sectors of the northern hemisphere, and finally by precipitation at high latitude. Indeed, the 3 km ice-mountain of Greenland is a living record of this Atlantic storm-track transport. Were the dynamical ocean to be replaced by a thin mixed layer, some aspects of seasonal heat and local evaporative forcing would remain. It is difficult to believe, however, that the intense warming and moistening of the northward moving air masses would continue, nor would the distributions of deep convection and water mass formation, nor the geography of sea-ice cover, unless those model mixed layers were artificially forced to mimic the true surface conditions of the ocean. Papers like SO2 stimulate us to observe more accurately the vertical structure of energy and fresh-water transport, and, just as important, to move toward descriptions of the ocean circulation through transports across sections and transformation within ‘boxes’, played out on the θ–S plane.

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