Vertical Redistribution of Oceanic Heat Content

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ABSTRACT

Estimated values of recent oceanic heat uptake are on the order of a few tenths of a W m$^{-2}$, and are a very small residual of air–sea exchanges, with annual average regional magnitudes of hundreds of W m$^{-2}$. Using a dynamically consistent state estimate, the redistribution of heat within the ocean is calculated over a 20-yr period. The 20-yr mean vertical heat flux shows strong variations in both the lateral and vertical directions, consistent with the ocean being a dynamically active and spatially complex heat exchanger. Between mixing and advection, the two processes determining the vertical heat transport in the deep ocean, advection plays a more important role in setting the spatial patterns of vertical heat exchange and its temporal variations. The global integral of vertical heat flux shows an upward heat transport in the deep ocean, suggesting a cooling trend in the deep ocean. These results support an inference that the near-surface thermal properties of the ocean are a consequence, at least in part, of internal redistributions of heat, some of which must reflect water that has undergone long trajectories since last exposure to the atmosphere. The small residual heat exchange with the atmosphere today is unlikely to represent the interaction with an ocean that was in thermal equilibrium at the start of global warming. An analogy is drawn with carbon-14 “reservoir ages,” which range from over hundreds to a thousand years.

1. Introduction

A large recent literature (e.g., Levitus et al. 2001; Barnett et al. 2005; Easterling and Wehner 2009; Meehl et al. 2011; Lyman et al. 2010; Chen and Tung 2014) has discussed the rates of oceanic heat uptake from the atmosphere and particularly whether inferences of “missing” heat can be explained by warming of the deep ocean. Estimates of the patterns of oceanic heat exchange with the atmosphere (e.g., Stammer et al. 2004) have an annual average range of hundreds of W m$^{-2}$ of both signs, with much larger seasonal extremes. Finding and understanding a residual oceanic warming even as large as 1 W m$^{-2}$ with useful accuracy requires considerable insight into the distribution and physics of the exchanges. A recent paper (Wunsch and Heimbach 2014, hereafter WH14) used a 20-yr-duration oceanic state estimate (described briefly below) to infer an oceanic uptake of heat on the order of 0.2 ± 0.1 W m$^{-2}$ (formal error only). Noting only that the value is a very small residual of far larger space and time variations, the purpose of this present paper is not to defend or modify that estimate of the residual, but instead to understand the gross mechanisms by which the ocean, having absorbed or yielded heat to the atmosphere, redistributes that heat internally. Regional variations of heating and cooling and
of consequent heat transports are much larger in magnitude than the residuals, and their physics are far more robust than are their small integrated differences.

Among other goals, we seek a better understanding of the WH14 conclusion that parts of the abyssal ocean appear to have been cooling over the last 20 years. Furthermore the ocean, far from being a passive reservoir filled and emptied by the atmosphere, is a dynamically active, turbulent element of a coupled system. Understanding of the patterns of heat exchange require depiction of the basic mechanisms by which heat is removed from, or added to, the layers in active exchange with the atmosphere. Determining those patterns is one of the goals here.

2. Basis of the estimate

Results here are derived from an Estimating the Circulation and Climate of the Ocean (ECCO) state estimate, labeled version 4, release 1, which can be understood as the result of the least squares fit of the MITgcm (Adcroft et al. 2004), with sea ice and mixed-layer submodels to about O(10^3) observational points [see Wunsch and Heimbach (2013) for a review]. Lagrange multipliers are used (Wunsch and Heimbach 2007) to enforce the model equations in such a way that basic conservation rules (heat, freshwater, momentum, energy, etc.) are satisfied globally and locally in the interval 1992–2011. The estimate is based upon the model in a configuration with 50 layers and a spatial resolution of 0.25°–1° in latitude and 1° in longitude.

WH14 provided much more detail about the state estimate, including the central test of skill lying with the model–data misfits. Given the large number of data constraints, a full description is a long story. Figures 4 and 5 in WH14, for example, show the CTD temperature misfits. While some large misfits do appear, they are localized in nature and are not visible on the larger scales.

The state estimate was confined to the interval after 1992 when, with the flight of the TOPEX/Poseidon altimeter satellite, the advent of the World Ocean Circulation Experiment (WOCE), and the deployment of the Argo array and other instrument types, quasi-global coverage of the ocean existed. Thus, a decadal time scale of averaging and of change is accessible. On the other hand, the ocean has a long memory, and its behavior over much longer periods will be reflected in the starting estimate of the system initial conditions. Because initial conditions are part of the control vector of the state estimate, they are adjusted as part of the procedure of rendering the model consistent with the time-evolving and time-average data. In more conventional calculations, often with forward models, a usual assumption is that the ocean was in thermal equilibrium with the atmosphere at the initiation of anthropogenic warming and other disturbances. To the extent that the assumed initial state is regionally or globally too warm or too cold, heat will enter and move through the system in ways that are potentially misleading.

The continued relative sparsity of abyssal ocean temperature measurements means that the state estimate is not completely immune to the same problem. Nonetheless, to the extent the system is consistent with all of the data for 1992–2011, and with the adjusted initial conditions, possible unexpected behavior (specifically, regions of inferred cooling) are physically and dynamically acceptable and interpretable.

3. The vertical movement of heat

The goal here is a description of the oceanic movement of heat in the vertical dimension. That discussion can only take place in the context of the corresponding vertical movement of mass (volume). Thus, appendix A provides a brief description of the vertical mass (volume) balances in the state estimate. References already cited sketch the lateral movements of mass and heat in the state estimate.

The net vertical ocean heat flux $H_v$ below 200 m involves advective $H_a$ and diffusive $H_d$ terms. Penetrating solar radiation appears important in the upper 200 m. Advective terms consist of the Eulerian transport, $\rho C_p w T$, and the bolus transport, $\rho C_p w^* T$, where $\rho$ is the density of the seawater; $C_p$ is the specific heat of seawater at constant pressure; $T$ is temperature; $w$ is the Eulerian vertical velocity; and $w^*$ is the bolus velocity, which is eddy induced and is calculated following Gent and McWilliams (1990). Note that the advective heat flux is calculated with a constant reference temperature of 0°C. The diffusive heat flux can be divided into a term, $-k_v \partial T/\partial z$, where $k_v$ is the total vertical diffusivity, and a second term representing part of the contribution of the isopycnal diffusivity in the vertical direction. For a detailed description of the calculations, see appendix B.

a. Global distributions

Apart from geothermal heating at an average rate of about 0.1 W m$^{-2}$ (e.g., Pollack et al. 1993; Davies 2013), which is not represented in the state estimate, the ocean receives and loses heat at and near the sea surface. Figure 1 shows maps of the mean and the temporal standard deviation from monthly values of the 20-yr net ocean–atmosphere heat exchange from the state estimate. Generally speaking, and as is conventional, the ocean receives heat at low latitudes and loses it at high latitudes. Many detailed deviations occur in structures that are related to the major ocean currents, such as the Kuroshio, the Gulf Stream, and the Antarctic Circumpolar Current.
The strongest ocean–atmosphere heat exchange (>100 W m$^{-2}$) occurs in the tropical regions, especially the eastern tropical Pacific Ocean, the Gulf Stream, and the Kuroshio, as well as a limited region in the high-latitude North Atlantic Ocean. At the same time, the largest temporal variations of the ocean–atmosphere heat exchange also appear in the Gulf Stream and the Kuroshio. Except for the tropical regions, a major portion of the ocean is dominated by temporal variations.

Over much of the ocean, the 20-yr mean values are not obviously statistically significant. The existence of spatial patterns of heat exchange of one sign means that this statement about insignificance must be used guardedly: without evaluation of the spatial covariance of the misfit components, a rigorous estimate of significance is not possible. But ratios of the mean to the standard deviation at single points or over small regions are often indistinguishable from zero. Therefore, obtaining reliable
temporal mean values of the ocean–atmosphere exchange, which are much smaller than the magnitudes of the temporal variations in a major portion of the ocean, is a challenge.

The horizontal spatial pattern of the vertical heat redistribution in the ocean interior differs from that of the air–sea net exchange, consistent with the ocean circulation being dynamically active. Figure 2 displays the time means \( \bar{H}_v \) and the standard deviations \( \sigma_v \) of the 20-yr net vertical heat flux, \( H_v(\lambda, \phi, z, t) \) at a few sample depths. (Independent variables are longitude, latitude, depth, and time.) The magnitude and spatial patterns of the time-mean vertical heat flux (left column) show strong variations in the vertical direction. Near the surface (~50 m), excluding the penetrating solar radiation, heat is primarily carried upward in both the tropical and high latitudes and downward in the subtropical regions. The magnitude of \( \bar{H}_v \) reaches 200 W m\(^{-2}\) near the surface in the tropical regions. With increasing depth, \( \bar{H}_v \) in the tropical and subtropical regions decreases significantly, with the strongest values mainly occurring at high latitudes, particularly the North Atlantic and the Southern Ocean. These spatial patterns also vary with increasing depth: in the Pacific and Indian Ocean the dominant structure is a change from a downward net transport in the upper ocean to an upward one in the deep ocean. In the Atlantic Ocean, a downward pattern exists from the surface to the abyssal ocean. Maintenance of a quasi-steady temperature field in the ocean requires strong lateral transports that are not being
described here but which have been discussed at length in the papers cited.

Figure 2, depicting $\sigma_v$, shows that regions of large values (e.g., the tropical Indian Ocean) are not collocated with regions of large $H_v$ (e.g., the Southern Ocean). The largest values of $\sigma_v$ appear in the tropical regions at all depths, consistent with deep-penetrating strong temporal variations of the vertical heat flux in the tropical regions. Because similar temporal variations in the tropical regions are not apparent in the ocean–atmosphere heat exchange (see Fig. 1), these fluctuations are a consequence of internal ocean dynamics. The deep western North Atlantic also displays strong temporal variations, likely associated with the deep western boundary current, as well as in regions of strong convection. A weak but detectable relationship between the enhanced temporal variations in the abyssal ocean and large-scale topographic features is consistent with topographic induction of variability there.

Frequency spectral estimates from a few representative regions confirm that ocean processes on a variety of time scales contribute to the temporal variance. While the annual cycle is a nearly universal component of geographically varying fractional importance, spectra of the vertical heat flux from different regions of the global ocean show clear differences in both form and value. For instance, spectra of the vertical flux in the tropical regions are “blue,” with a relatively large high-frequency variance, but in other regions they are generally “white.” Excess high frequencies at depth at low latitudes are roughly consistent with expectations of the relative efficiency of vertical baroclinic Rossby wave propagation near the equator (e.g., Gill 1982). For a summary description of the frequency spectra of $H_v$, see appendix C. The dynamics of the temporal variations of $H_v$ are important and will be examined elsewhere.

The ratio $\Pi_v/\sigma_v$ for the 20-yr net vertical heat flux at a few sample depths is shown in Fig. 2 (right column). In the very upper ocean, the estimates of the 20-yr mean of the heat flux are significant (in the sense of relatively large mean to standard deviation ratio) on the equator and in the subtropical ocean basins. In the deep ocean, however, estimates of the 20-yr mean are only significant in the Southern Ocean and in a limited region in the North Atlantic Ocean. Overall, the large values of $\sigma_v$ and the comparatively short records imply that reliable estimates of global averages and integrals must be inferred indirectly, as in the present calculations.

Values and structures of $H_v$ below 200 m are governed by two processes: vertical advection and vertical diffusion. Figure 3 shows net $\Pi_v$, advective $\Pi_a$, and diffusive $\Pi_d$ vertical heat fluxes as well as their corresponding temporal standard deviations $\sigma_v$, $\sigma_a$, and $\sigma_d$ at around 700 m. Visually, the spatial patterns of $\Pi_v$ and $\Pi_a$ are strikingly similar at this depth, with both showing positive (upward) heat transport on the equator, downward transport in the subtropical basins, and strong values at high latitudes, particularly the North Atlantic and the Southern Ocean. The standard deviations $\sigma_v$ and $\sigma_a$ display almost identical spatial patterns, with the largest values appearing in the tropical regions as well as in the North Atlantic. The values for $\Pi_d$ contribute little to $\Pi_v$, being generally very small and showing downward transport in a major portion of the ocean. Exceptions exist in the Southern and North Atlantic Oceans, where strong upward values appear. Relatively large temporal variances of $H_d$ appear in the Southern and the North Atlantic Oceans but are numerically significantly smaller than those for $H_a$. Note the different range of the colorbar for $\sigma_d$ in Fig. 3. Patterns at 700 m generally persist into the deep and abyssal oceans, consistent with dominance of vertical advection and a coupling of upper and lower ocean heat budgets.

b. Global integrals

Despite the overall weakness of the diffusion process, the global integral of the diffusive heat flux roughly balances the advective vertical heat transport in an “abyssal recipes” form (Munk 1966), but reinterpreted as a global integral (Munk and Wunsch 1998). The global spatial averages ($\langle H_v \rangle$, $\langle H_a \rangle$, and $\langle H_d \rangle$) are shown in Fig. 4. Except for the upper 200 m, where penetrating solar radiation is important, $\langle H_v \rangle$ is close to zero, indicating the required near-balance between the advective and diffusive heat fluxes. Note that in the present study, $\langle H_a \rangle$ consists of the Eulerian-mean and the eddy-induced terms, both of which are generally larger than $\langle H_d \rangle$ and cancel each other about above about 4000 m. Similarly, $\langle H_d \rangle$ also includes two terms, the diapycnal diffusion and the nonnegligible vertical projection of isopycnal diffusion [see Eq. (B3)], the latter of which is particularly pronounced at high latitudes. This suggests that the abyssal recipes can be further interpreted as a balance between residuals of stronger dynamical processes.

Below about 1000 m, $\langle H_v \rangle$ is upward and is on the order of 0.1 W m$^{-2}$. The change of the heat content of a particular water column is determined by the convergence of heat transport passing through its upper and lower surfaces. Thus, in this estimate, a small part of the observed ocean warming in the upper 2000 m arises from a heat flux from below, and the ocean below 2000 m is cooling. Because of the strong annual cycle, the monthly values are assumed not independent. We then define the uncertainty of the net heat flux as $\sigma/\sqrt{n}$ where $n = 20$. The global average of the net vertical heat flux
shows a varying uncertainty estimate in the deep ocean, from about 1 W m$^{-2}$ at 100 m, to about 0.1 W m$^{-2}$ at 1000 m and 0.03 W m$^{-2}$ at 3000 m. Below about 100 m and above 3500 m, the global average of the advective vertical heat flux is negative, implying a net downward heat transfer in that depth range. The average $\langle H_d \rangle$ is, however, positive at all depths. As shown in Fig. 3, the upward diffusive heat flux mainly occurs in the high-latitude North Atlantic as well as the Southern Ocean and is due to the vertical projection of isopycnal diffusion. Uncertainties from temporal variations in the estimates of $\langle H_d \rangle$ in the deep ocean are mainly associated with the advection term rather than with diffusion: the temporal variation of $H_y$ is dominated by $H_a$; $H_d$ is both small and temporally stable in the deep ocean (see appendix D).

4. Discussion

A perhaps surprising result of the current calculation is that the global lateral average of $\Pi_y$ shows a small upward transport in the deep ocean (below 1000 m) on the order of 0.1 W m$^{-2}$ even without the presence of the geothermal driving. Evidently, if the deep ocean in 1992 was slightly warmer than its equilibrium value, no physical contradiction exists with an upward movement.
Global average cooling in the deep ocean conflicts with some previous ocean heat content estimates (e.g., Balmaseda et al. 2013) but is consistent with the long thermal memory of the ocean and with other recent studies (e.g., Durack et al. 2014; Llovel et al. 2014). All existing estimates of the deep ocean states, including this present one, are based on very limited in situ observations in the deep ocean, and the uncertainties are large. Furthermore, upper ocean warming may have been generally underestimated: Any bias errors in the initializing state rendering the upper ocean warmer than is correct would produce such an underestimate. Note the historical emphasis on measurements of the relatively warm North Atlantic Ocean and the tendency for shipborne observations to focus on lower latitudes generally, particularly in winter (see, e.g., Fig. 2 of Atkinson et al. 2014).

An upward heat transport in the deep ocean may appear to be in conflict with the widespread idea that a large portion of the extra heat added to the Earth system in the past decades should be transported into the deep ocean (e.g., FAQ 3.1, Fig. 1 in Stocker et al. 2013). That inference is based on the assumption that the ocean was in equilibrium with the atmosphere before any extra heat entered. When interpreting measurements of the ocean heat content, it is often assumed that the disturbances arise only from the recent past. However, as emphasized by Wunsch and Heimbach (2014) and the present analysis, the long integration times in the ocean circulation imply an observed response involving the time history of the circulation over hundreds of years, at least.

Although it is a very crude measure and easy to misinterpret, note, for example, that the radiocarbon “age” of midlatitude surface water is about 400 yr, exceeding 1000 yr at high southern latitudes and high northern Pacific ones (e.g., Bard et al. 1994). (Ages are best interpreted more fundamentally as the logarithm of the...
carbon-14 concentration in seawater relative to the atmosphere.) If interpreted literally, such durations are adequate for the fluid both to have been exposed to very different atmospheric conditions and to have undergone complex exchanges within the ocean itself. In principle, an out-of-equilibrium ocean can be warming the present atmosphere—if its current surface thermal properties were set in the remote past (in addition to the geothermal forcing). Times for the surface ocean to equilibrate with atmospheric radiocarbon are on the order of a decade (e.g., Williams and Follows 2011), longer than for thermal equilibration with the top few meters but nonetheless far shorter than the time that would have been required to produce surface radiocarbon equilibrium. As with the radiocarbon reservoir values, surface thermal properties are an amalgam of recent local atmospheric forcing and the history of the three-dimensional ocean circulation itself.

Between vertical advection and diffusion, the former is more important in determining the spatial patterns of the vertical heat transport and its temporal variations. In other words, obtaining reliable vertical velocity estimates is crucial for understanding regional vertical heat transports. Vertical diffusion, on the other hand, is not as important as vertical advection regionally but is of equal importance in the global integral. Abyssal recipes (Munk 1966; Munk and Wunsch 1998) balance demonstrably works in terms of the global integral, while being violated on a regional basis. As shown in appendix B, the advection contribution can be divided into its Eulerian-mean and eddy-induced parts. The diffusion term is the sum of diapycnal diffusion and the nonnegligible vertical projection of isopycnal diffusion. Fully deciphering the dynamics of the ocean vertical heat transport requires a deep understanding of these processes and the relationships between them. Because this paper focuses on the description of the ECCO estimates, and examining the dynamics of these processes is important in itself, we leave the detailed dynamical analyses to the future.

Because the temporal variability of $H_v$ is a strong function of position, including depth, uncertainty estimates of its temporal mean also vary greatly. As shown in Fig. 4, $\langle \tilde{H}_v \rangle$ shows widely varying uncertainties at different depths. These uncertainty estimates are based only upon the estimated temporal variability, ignoring any unknown systematic errors. The uncertainty at 100 m is around 1 W m$^{-2}$, which is an order of magnitude larger than the desired accuracy of global ocean heat content measurements (<0.1 W m$^{-2}$). However, below about 1000 m, uncertainties become smaller than 0.1 W m$^{-2}$. Thus, assuming a perfect scenario where no systematic estimation errors exist, in a major portion of the ocean, heat uptake accuracies can be achieved in the range of 0.02–0.1 W m$^{-2}$ over a 20-yr period. However, systematic errors associated with data and models are unavoidable. Recall too that this particular state estimate does not resolve the eddy field; hence, the variability values are a lower bound.

The complex vertical redistribution of heat and the clear variation in governing physics have major implications for the design of an observing system capable of producing estimates of understanding the oceanic heat budget at the level of 0.1 W m$^{-2}$ or better. Different types of measurements are needed for ocean regions with different governing physics. Noisy regions will require different data than quiet ones. When using the available historical measurements to estimate the global mean values, measurements from regions with different temporal variances must be differently weighted. The maps presented in this study can serve as preliminary references for that purpose.

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APPENDIX A

Vertical Volume Balance in ECCO v4

Vertical transport of oceanic water involves upwelling and downwelling in complex spatial patterns (Fig. A1), with volume conservation requiring that the spatial integrals be equal to the net volume flux associated with net freshwater input. The pattern of the vertical velocity is similar to that of the heat flux (Fig. 2). Basinwide integrals (Fig. A2) are simpler than the complicated variations seen within the oceanic interior. In the Pacific Ocean, the integrals show water moving primarily upward. In the Atlantic Ocean, except in a shallow layer near the surface, the gross water mass movement is downward in both the expected high-latitude regions and also in a major portion of the remainder of the ocean basin. The Indian Ocean has upward volume transport below about 1300 m but is dominated in the upper 1300 m by a downward volume transport, the opposite to at least one schematic of the meridional overturning circulation (Talley 2013). In the Southern Ocean (south of 30°S), net sinking of water mass occurs above about 500 m and below about 2000 m in the Pacific sector and the Indian sector. The Atlantic sector is dominated by a net upward movement over nearly the whole water column.
FIG. A1. The 20-yr mean of the residual vertical velocity ($w + w^*$) in m s$^{-1}$ at four sample depths. Results at around (top)–(bottom) 50, 700, 2000, and 3000 m. Positive and negative values stand for upward and downward velocities, respectively. The spatial pattern of the vertical velocity is clearly similar to that of the net vertical heat flux (Fig. 2), again indicating the important role of the vertical advection in the vertical redistribution of ocean heat content.
APPENDIX B

Calculation of the Advective and Diffusive Vertical Heat Fluxes in ECCO v4

The net vertical heat flux $H_v$ below 200 m consists of advective $H_a$ and diffusive $H_d$ terms. The $H_a$ term consists of two parts: the Eulerian-mean transport $\rho C_p w T$ and the eddy-induced transport $\rho C_p w^* T$, where $\rho$ is the density of the seawater; $C_p$ is the specific heat of seawater at constant pressure; $T$ is temperature; $w$ is the Eulerian vertical velocity; and $w^*$ is the bolus velocity.

Here, the advective heat flux is calculated with a constant reference temperature of 0°C. The Eulerian vertical velocity $w$ is diagnosed from volume continuity:

$$ w = - \int_0^z \mathbf{v}_h \cdot \mathbf{v}_h dz', \quad (B1) $$

where $\mathbf{v}_h$ is the horizontal velocity vector and $\mathbf{V}_h$ is the horizontal gradient operator. The eddy-induced bolus velocity $w^*$ is parameterized following Gent and McWilliams (1990) as

$$ w^* = - \mathbf{V}_h \cdot \left( \kappa_{gm} \frac{\mathbf{V}_h \rho}{\rho_z} \right), \quad (B2) $$

where $\kappa_{gm}$ is an empirical coefficient that is adjusted through the least squares fitting process to minimize the model–data misfits (G. Forget et al. 2015, submitted to Geosci. Model Dev.) and ranges from 130 to 2750 (the smallest 5% and largest 5% of $\kappa_{gm}$ are excluded).

The total vertical diffusive heat flux is defined as

$$ H_d = - \rho C_p \left[ k_{wx} \frac{\partial T}{\partial x} + k_{wy} \frac{\partial T}{\partial y} \right]_{rhs1} + (k_{wz} + k_z) \frac{\partial T}{\partial z} \right]_{rhs2}, \quad (B3) $$
where horizontal gradients (rhs1) are solved explicitly, whereas the vertical gradient (rhs2) is solved numerically via an implicit scheme; the “explicit” term and $k_w \partial T/\partial z$ are contributions of the isopycnal diffusion to the vertical transport of temperature (Redi 1982); and $k_z$ includes three components: the background diffusivity, which is adjusted through the adjoint process, the part parameterized using GGL90 (Gaspar et al. 1990), and the part related to the convective instability.

**APPENDIX C**

**Frequency Spectra of the Net Vertical Heat Fluxes in a Few Representative Regions**

For discussions of adequate temporal sampling and direct calculations from data, the spectral description of the elements of $H_v$ is essential. Frequency spectra of the regional averages of $H_v$ from a few representative
regions are estimated using the multitaper spectral estimation method and are shown in Fig. C1. The most pronounced feature is that an annual peak appears in all the estimated spectra, with a greatly spatially varying amplitude, and is mainly associated with the variation of the vertical velocity. Except for this ubiquitous annual peak, the forms of the spectra show significant spatial variation in the relative proportions of low- (<1 cycle per year) and high-frequency (>1 cycle per year) bands.

A few examples are summarized as follows:

In the eastern tropical Pacific, except in the very upper layers, the temporal variance of the vertical heat flux is the largest in the global ocean. The high-frequency band produces spectra that are generally blue, with some red noise behavior at the lowest frequencies. Spectra of similar form also appear in the tropical Indian Ocean. The relatively strong high-frequency variability could be associated with the fast response of the tropical ocean to the wind forcing (e.g., Gill 1982).

Vertical heat fluxes in the western boundary currents also show large temporal variance. For example, in the Kuroshio, in addition to the annual peak, its harmonics are also clearly visible. Spectra vary with depth, particularly at low frequencies. In the upper 1000 m, values in the low-frequency band are almost comparable to those in the annual cycle. In the deep ocean, however, the low-frequency band is relatively suppressed, with most of the fluctuations associated with the western boundary currents.

Frequency spectra of the vertical heat flux in the North Atlantic are distinctive. Except for peaks around the annual cycle and a few of its harmonics, spectra are flat in the frequency band above 1 cycle per year. Values in the low-frequency band show an increase toward the lowest frequency in the middle of the water column but a decrease in the upper and the abyssal ocean. The observed intensification of low-frequency variability in the middle of the water column appears associated with movement of the North Atlantic Deep Water.

For regions with small temporal variance of $H_v$, such as the Southern Ocean and subtropical ocean basins, in addition to the ubiquitous annual peak, large values also appear in the low-frequency band. In contrast to other examined regions, where low-frequency values vary vertically, the Southern Ocean and subtropical ocean basins show intensified low-frequency values in the whole water column.
APPENDIX D

Temporal Variations of the Global Averaged Net, Adveotive, and Diffusive Vertical Heat Fluxes

Figure D1 shows the time series of the global averages of the net $H_u$, advective $H_a$, and diffusive $H_d$ heat fluxes at 700 m. The temporal variation of the global average of the net vertical heat flux is dominated by a strong annual cycle and has a standard deviation of 0.6 W m$^{-2}$, which is an order of magnitude larger than the 20-yr mean of the net vertical heat flux (0.03 W m$^{-2}$) at the same depth. Removing the annual cycle and its first harmonic reduces the standard deviation of the time series to about 0.4 W m$^{-2}$, which is still significantly larger than the temporal mean value. Temporal variation of the global average of the advective vertical heat flux is almost identical to that of the net vertical heat flux, showing the importance of the advective term in the net vertical heat flux, not only in the mean value but also in the temporal variation. For the diffusive vertical flux, with the exception of a few extremes between 1992 and 1996, no clear temporal variation appears, producing a temporally stable contribution from the vertical diffusive processes in the deep ocean—at least as represented in the present dynamical representation.

REFERENCES


