An Observational Estimate of Inferred Ocean Energy Divergence

Kevin E. Trenberth and John T. Fasullo

e-mail: trenbert@ucar.edu

National Center for Atmospheric Research¹

P. O. Box 3000

Boulder, CO 80307

6 June 2007

J. Phys. Oceanogr

¹The National Center for Atmospheric Research is sponsored by the National Science Foundation.
Abstract

Monthly net surface energy fluxes ($F_S$) over the oceans are computed as residuals of the atmospheric energy budget using top-of-atmosphere (TOA) net radiation ($R_T$) and the complete energy budget tendency and divergence for the atmosphere ($\nabla \cdot F_A$). The focus is on TOA radiation from ERBE (February 1985 to April 1989) and CERES (March 2000 to May 2004) combined with results from two atmospheric reanalyses and three ocean datasets that enable a comprehensive estimate of uncertainties. An analysis of $F_S$ departures from the annual mean and the implied annual cycle in “equivalent ocean energy content” is compared with directly observed ocean energy content ($O_E$) and tendency ($\delta O_E/\delta t$) to reveal the inferred annual cycle of divergence ($\nabla \cdot F_O$). In the extratropics, $F_S$ dominates $\delta O_E/\delta t$ although supplemented by ocean Ekman transports that enhance the annual cycle in $O_E$. In contrast, in the Tropics, ocean dynamics dominate $O_E$ variations throughout the year in association with the annual cycle in surface wind stress and the North Equatorial Current. An analysis of the regional characteristics of the first joint Empirical Orthogonal Function (EOF) of $F_S$, $\delta O_E/\delta t$, and $\nabla \cdot F_O$ is presented, and the largest sources of uncertainty are attributed to $O_E$. The mean and annual cycle of zonal mean global ocean transports is estimated and compared with observations. Agreement is reasonable except in the North Atlantic, where transports from the transects are greater than the estimates presented here - a difference that exceeds the uncertainty bounds established herein.
1. Introduction

The oceans play a major role in moderating climate. In midlatitudes, energy absorbed by the oceans in summer is released to the atmosphere in winter, thus reducing the annual cycle in surface temperatures relative to those over land. Moreover, the advection of energy from the oceans also moderates the seasonal cycle over land where maritime influences prevail. Relative to the oceans, the atmosphere’s capacity to store energy is small and is equivalent to that of about 3.5 m of the ocean if their associated proportion of global coverage is considered. Because the main movement of energy in the land and ice components is by conduction, only very limited masses are involved in changes on annual time scales. Moreover, water has a much higher specific heat than dry land by about a factor of 4.5 or so. Accordingly, it is the oceans, through their total mass, heat capacity, and movement of energy by turbulence, convection and advection, that have enormous impact on the global energy budget, which can vary significantly on annual and longer time scales (Trenberth and Stepaniak 2004).

The net radiative flux \( R_T \) at the top–of–atmosphere (TOA) on longer than annual time scales is mostly balanced by transports of energy by the atmosphere and ocean, and local upwards surface energy fluxes \( F_S \) are largely offset by the ocean’s divergent energy flux \( \nabla \cdot F_O \), (Trenberth and Caron 2001). In contrast, for the annual cycle in midlatitudes the dominant ocean response is the uptake and release of energy, with changes in \( \nabla \cdot F_O \) being secondary (Jayne and Marotzke 2001). Here we firstly provide an estimate of the annual cycle in equivalent \( O_E \) based on \( F_S \) alone. The term “equivalent” is used because it neglects the influence of \( \nabla \cdot F_O \). While the sparseness of ocean observations is a major constraint on our understanding of its energy budget, to the extent that reliable estimates of both actual and equivalent \( O_E \) can be made, their difference
can be interpreted as $\nabla \cdot F_O$. Accordingly, we compute this residual for the mean annual cycle and annual mean.

The main source of information on $O_E$ comes from ocean temperature profiles compiled into datasets and analyzed as atlases. Previous analyses of $O_E$ include both regional (e.g., Moisan and Niiler 1998) and global estimates using the World Ocean Atlas (WOA) 1994 (Levitus 1984, 1987, Levitus and Antonov 1997) and WOA 2001 (Antonov et al. 2004). Recently, an update of these fields has been made available in the WOA 2005 (Locarini et al. 2006). These estimates start from the monthly analyses of the basic observations gridded into one-degree squares at standard depth levels and integrated in depth from the surface to 275 m. Despite the increasing volumes of data incorporated into successive versions of the WOA, over some parts of the ocean very few or no observations have been made at some times of the year (notably over the wintertime southern oceans), and consequently sampling errors are large so that the spatial patterns of $O_E$ (Levitus and Antonov, 1997) contain spurious noise. The recent availability of data from the ARGO floats will improve greatly the sampling of the southern oceans, however it will be several years before an extended data record is made available and the reliability of data processing remains an ongoing issue (Willis, et al. 2007).

Temperature anomalies in the ocean can penetrate below 275 m (Levitus and Antonov 1997) into the thermocline through subduction processes. Yan et al. (1995) and Moisan and Niiller (1998) found considerable sensitivity to their results in the North Pacific to the depth of integration, with best results in comparisons with $F_S$ for an integration to the depth of a fixed isotherm about 1°C colder than the coldest surface temperatures, which they refer to as the “wintertime ventilation isotherm”. $F_S$ is balanced by $\delta O_E/\delta t$ through efficient vertical mixing, however many other processes are neglected. For example, $\nabla \cdot F_O$ and large vertical motions across
the thermocline can considerably complicate the mixed layer budget. The analysis here therefore attempts a more complete budget analysis using physical constraints to infer divergences.

We use adjusted TOA radiation from Earth Radiation Budget Experiment (ERBE, Barkstrom and Hall 1982) for February 1985 to April 1989 and the Clouds and the Earth Radiant Energy System (CERES, Wielicki et al. 1996) retrievals, for March 2000 to May 2004, along with comprehensive estimates of the divergence of the vertically integrated atmospheric energy components, computed from both National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR, and hereafter NRA, Kistler et al. 2001) and European Centre for Medium Range Weather Forecasts 40-year reanalyses (ERA, Uppala et al. 2005). The annual cycle of the TOA radiation fluxes and atmospheric energy transports and divergences are given in Trenberth and Stepaniak (2003a, b; 2004). \( F_S \) is computed as a residual and allows the implied annual mean \( \nabla \cdot F_O \) to be inferred. The transports derived from this method have been shown to correspond quite well with direct observations from ocean hydrographic sections (Trenberth and Caron 2001).

Here we examine the annual cycle of \( F_S \) and compare it with direct observations of \( O_E \) from three sources including: the WOA 2005 (Locarini et al. 2006), the analysis of the Japanese Meteorological Association Version 6.2 (JMA, Ishii et al. 2006), and the Global Ocean Data Assimilation System (GODAS, Behringer and Xue 2004, Behringer 2006). We analyze in some detail the overall ocean energy budget and its agreement with fundamental global constraints in a range of estimates. We conclude that many of the results derived herein are dominated by physically real changes, thereby revealing new aspects of the annual cycle of ocean energy divergence. Nevertheless, residual errors are found to be significant in instances and their influence is quantified. Section 2 describes the data and the processing, section 3 presents the results, and section 4 discusses the results and their implications, and conclusions are drawn in section 5.
2. Data and methods

More detailed discussion of the datasets and methods is given in Fasullo and Trenberth (2007a), referred to hereafter as FT07, upon which the analysis of budget terms here is based. We use a prime to denote a departure from the annual mean.

a. Energy budgets

The dominant energy terms and balances for the vertically integrated atmosphere and ocean are presented briefly below. In the atmosphere,

\[ F_S = \nabla \cdot F_A - \partial A_E/\partial t - R_T \]  

where \( F_S \) and \( R_T \) have been previously defined, and \( \nabla \cdot F_A \) and \( \partial A_E/\partial t \) are the vertically integrated atmospheric total energy divergence and tendency respectively. Moreover, for the ocean, given \( F_S \) and \( O_E \), the divergence of ocean energy transport can be inferred based on

\[ \nabla \cdot F_O + F_S + \partial O_E/\partial t = 0, \]

and the ocean energy is approximated by the ocean heat content,

\[ O_E = \int T(z) \, C_w \, dz, \]

where \( z \) is depth, \( T \) is the ocean temperature, and \( C_w \) is the specific heat of sea water. The challenges in diagnosing the terms in (1) – (3) include obtaining high quality analyses of global observations with adequate sampling of the large temporal and spatial gradients of all terms.

b) Adjusted satellite retrievals

For terms in the TOA budget contributing to \( R_T \), adjusted satellite retrievals from ERBE (Feb. 1985 to Apr. 1989, Barkstrom and Hall 1982; FT07) and CERES (Mar. 2000 to May 2004, Wielicki et al. 1996; FT07) are used. While the ERBE retrievals offer multi-satellite sampling, the CERES data show lower noise, improved ties to ground calibration, and smaller fields of view than
ERBE (Loeb et al. 2007). CERES instrument calibration stability on Terra is claimed to be typically better than 0.2%, and calibration consistency from ground to space is better than 0.25%. As both the ERBE and CERES estimates are known to contain spurious imbalances however (Trenberth 1997, Wielicki et al. 2006), adjustments are required, as described in FT07, such that estimates of the global imbalance during the ERBE and CERES periods (Hansen et al. 2005; Willis et al. 2004; Huang 2006; Levitus et al. 2005) are matched. Associated uncertainty estimates, equal to two sample standard deviations of interannual variability ($\pm 2\sigma_i$), are reported here in order to quantify the uncertainty associated with the limited temporal span of the ERBE and CERES time periods. For all analysis here these adjusted ERBE and CERES fields are thus used and in no instances are the raw fields used.

c) Reanalysis datasets

In order to solve for $F_S$ from (2) estimates of the atmospheric tendency and divergence are required, except for the global average where divergence by definition is zero. In constructing estimates of these terms we use only fields strongly influenced by observations, such as surface pressure and atmospheric temperature and humidity from NRA and ERA reanalyses. Estimates of the monthly mean vertically integrated storage, transports and divergences of energy within the atmosphere for ERA and NRA were computed and evaluated as in Trenberth et al. (2001) and for NRA have been updated through 2006 (http://www.cgd.ucar.edu/cas/catalog/newbudgets, see Trenberth and Stepaniak 2003a, b, FT07). However, as ERA fields are not available beyond 2001, the ERBE period is the primary focus of the present study.

d) Ocean surface fluxes and storage

The computation of $F_S$ as a residual is superior to both model-based and Comprehensive Ocean Atmosphere Data Set (COADS) surface fluxes in terms of biases, as the latter both suffer
from systematic biases and fail to satisfy global constraints (Trenberth et al. 2001). Grist and Josey (2003) wrestled with how to best adjust their COADS-based estimates to satisfy energy transport constraints, suggesting adjustments to the fluxes and the need for further refinements associated with clouds. Trenberth and Caron (2001) used the long-term annual means of $F_S$ during the ERBE period to compute the implied meridional ocean energy transports. Results agreed quite well with independent estimates from direct ocean measurements within the error bars of each in numerous sections. Moreover, they are reasonably compatible with estimates from state-of-the-art coupled climate models. However, the annual mean $F_S$ (Trenberth et al. 2001; Trenberth and Stepaniak 2004) likely has various problems, especially over the southern oceans (Trenberth and Caron 2001).

The ocean datasets used here to diagnose ocean heat include the WOA 2005, JMA, and the recently corrected (6 Feb 2006) GODAS. While some insight into the likely biases of $O_E$ from these data can be gained from their degree of closure with $F_S$ (FT07), uncertainty in the observations, and particularly the decomposition of error into its systematic and random components, has been hampered by a lack of observations, particularly at depth. Here we also use departures from this mean for the ERBE period and thus gain the advantage of subtracting out most systematic errors or biases. Estimates of $F_S$ over ocean are based on (2) whose accuracy is of order 20 W m$^{-2}$ over 1000 km scales while satisfying closure among $F_S$, $\nabla \cdot F_A$, and $\delta A_E/\delta t$, and $R_T$ (Trenberth et al. 2001). As errors in $\nabla \cdot F_A$ are largely random and since divergence is zero globally by definition, a cancellation of error occurs over larger scales. Uncertainty in $F_S$ is governed mainly by the uncertainties in $R_T$ and $\nabla \cdot F_A$. By exploiting the constraint that $F_S$ and $\partial O_E/\partial t$ must balance globally, FT07 identify an excessive annual cycle of $O_E$ in JMA and WOA relative to that which can be explained by either a broad range of $F_S$ estimates or GODAS fields.
In estimating ocean energy from (3), we essentially follow the calculations of Antonov et al. (2004). However, there the density of seawater (\( \rho \)) was assumed to be 1020 kg m\(^{-3}\) and the specific heat (\( C_w \)) was assumed to be 4187 J kg\(^{-1}\) K\(^{-1}\) whereas for the typical salinity of the ocean of \( \sim 35 \) PSU, \( \rho \) is \( \sim 1025 \) to 1028 kg m\(^{-3}\) and \( C_w \) is 3985 to 3995 J kg\(^{-1}\) K\(^{-1}\) for temperatures from 2 to 20ºC. The product \( \rho C_w \) is more nearly constant than either of the two components but the Antonov et al. value is 4.4% too high, leading to an overestimate of \( O_E \) and its annual cycle. Therefore we have adjusted these constants and performed our own integration, using ocean temperatures provided at multiple levels in the WOA where the depth of each layer is assumed to extend between the midpoints of each level. In the case of the surface, the layer is assumed to begin at 0 m and at 250 m depth the layer is assumed to terminate at 275 m – half way to the next WOA layer at 300 m. We assign the density of ocean water to 1026.5 kg m\(^{-3}\) and specific heat of 3990 J kg\(^{-1}\) K\(^{-1}\), although it is their product that determines \( O_E \) per (3).

We further edit out the obvious spurious values south of 20ºS, by accepting only the monthly departures from the annual mean that are within two standard deviations of the zonal ocean mean to take advantage of the lack of land over the southern oceans. Globally, the data are also filtered temporally by retaining the first three harmonics of the annual cycle. Oceanic energy tendencies are then computed by reassembling the Fourier series and differencing \( O_E \) between the start and end days of each month. Other small systematic differences may arise from methods of compiling the vertical integral and numerical aspects in computing \( O_E \) (which are not described by Antonov et al. (2004)), however the magnitude of the annual cycle is not significantly affected by these refinements. In general the patterns of \( O_E \) derived here are quite similar to the actual \( O_E \) of Antonov et al. (2004) outside of the Tropics except the latter are \( \sim 10\text{-}20\% \) larger. One source of discrepancy can be the depth of integration, and Levitus and Antonov (1997) show that depths below 150 m are often somewhat out of phase with near surface values. Deser et al. (1999) show
how decadal changes in the Kuroshio extension are associated with temperature changes exceeding 1°C in the main thermocline from 300 to 1000 m depth and such changes can occur in association with subduction and ventilation of the thermocline. However, integrating to greater depth brings in regions of fewer data and thus increases errors and noise and we have therefore opted to limit our depth of integration to 275 m.

e) Regridding and standard deviations

To provide a consistent delineation of land-sea boundaries among the datasets used here, all fields are transformed to a grid containing 192 evenly spaced longitudinal grid points and 96 Gaussian-spaced latitudinal grid points using bilinear interpolation (i.e., to a T63 grid). Spatial integrals are calculated using Gaussian weights over the T63 grid and a common land-sea mask is applied. Total energy is expressed in units of petawatts ($10^{15}$ W or PW) and monthly mean values are used for all calculations. In quantifying seasonality, the estimated population standard deviation of monthly values is used.

f) Secondary terms

Within the ocean, formation and melting of sea ice can influence $O_E$. An estimate of this effect from the latent heat of fusion can be obtained from estimates of the annual cycle in ice volume from an ice model by Köberle and Gerdes (2003) for the Arctic, driven by NRA atmospheric fields, of about 1.5 to $3.2\times10^4$ km$^3$, which is also broadly consistent with energy budget estimates of Serreze et al. (2006, 2007). These studies thus suggest a heating amplitude of 0.5 PW for the annual cycle first harmonic. Moreover, transport of water vapor to land and storage as either snow or water affects the mass of the ocean and sea level (Minster et al.1999) and is associated with an energy flux estimated near 0.1 PW (FT07). For global budgets, it is expected that some degree of compensation between sea ice formation in the Antarctic balances that in the
Arctic, however reliable estimates of the net global budget remain unavailable. Nevertheless, this initial scale comparison suggests that overall the global imbalance at TOA should be reflected primarily in $O_E$ (Levitus et al. 2005).

3. Results

a. The annual mean $F_S$

The annual mean $F_S$ based on ERBE-NRA residuals is shown in Fig. 1a with stippled and hatched regions highlighting regions of difference ERA-based estimates exceeding $\pm 10$ W m$^{-2}$. For both estimates, the Tropics exist as a region of strong energy flux into the ocean ($F_S < 0$), particularly in the eastern Pacific Ocean. In contrast, the subtropics, where evaporative cooling is strong (Trenberth and Stepaniak 2004), and high latitudes are characterized by strong ocean cooling in the annual mean ($F_S > 0$). Zonal gradients in $F_S$ are large in the midlatitudes with particularly strong ocean cooling in the western boundary currents (e.g., Gulf Stream, Kuroshio Current) and generally modest ocean warming in the eastern basins. The values in the Southern Hemisphere (SH) mid-latitudes are smaller and more zonally symmetric than in the Northern Hemisphere (NH). Key differences between the ERA- and NRA-based estimates (Fig. 1a) are found in the Tropics and SH midlatitudes, where $\nabla \cdot F_A$, and thus $F_S$, in ERA exceed NRA, and through much of the subtropics, where ERA estimates of $\nabla \cdot F_A$, and $F_S$ are less than those of NRA.

b. The annual cycle

Figure 2 presents the annual cycle of the two of the dominant contributors to $F_S$, namely $R_T$, and $\nabla \cdot F_A$ ($\delta A_E/\delta t$ is relatively small). The zonal mean is averaged over the oceans only and both the total and the departure from the annual mean are presented. A distinct annual cycle is evident in $\nabla \cdot F_A$ in the NH as energy is transported in winter from ocean to land, where it is mostly
radiated to space (FT07). The reverse occurs in summer, though to a lesser extent than during winter, when high temperatures from mid-latitude land contribute to the divergence of energy to ocean regions. However, in the SH the lack of significant extratropical land extent results in a steady pattern of poleward energy transport with few zonal asymmetries and the annual cycle of $\nabla \cdot F_A$ is very small (Fig. 2d).

The zonal means in Fig. 2 integrate over large domains and thus mask the strong zonal structure of $F_S$; see Figs. 1b,c for June-July-August (JJA) and December-January-February (DJF) $F_S$ departures from the annual mean with stippling and hatching for differences with ERBE-ERA estimates of $\pm 30$ W m$^{-2}$. The solstitial season $F_S$ fields are dominated by the change in net surface solar radiative flux, and hence the large negative values in summer designate energy flux into the ocean and contrast with the positive values that are pervasive in the winter hemisphere. Some finer scale structures correspond to components of the ocean circulation. In NH winter (Fig. 1c) the largest values of annual mean departures of $F_S$ off the east coasts of the major continents exceed 240 W m$^{-2}$ and arise as cold dry air is advected off the continents over the warm Gulf Stream and Kuroshio Current, resulting in large sensible and latent heat fluxes into the atmosphere and corresponding maxima in atmospheric energy divergence (Fig. 2d). The lack of analogous features in $F_S$ in summer highlights the skewed influence of the boundary currents toward winter and contributes to their rectified impact on the annual mean fluxes, which exceed 140 W m$^{-2}$ in these western boundary regions (Fig. 1a), so that the total winter fluxes exceed 360 W m$^{-2}$ in these regions. In contrast, both the annual mean and seasonal extreme fluxes are much weaker in the eastern North Pacific, North Atlantic, and SH oceans. In the Tropics, seasonal contrasts are less than in midlatitudes and $F_S$ is positive in the primary upwelling zones for both solstitial seasons.

The mean annual cycle of $F_S$ zonally-averaged over the oceans (Fig. 3a), which is approximately the difference between Figs. 2b,d, features peak values at time of solstice in both
hemispheres, with maximum values exceeding 0.3 PW deg\(^{-1}\). [Note that zonal means are expressed as integrals rather than averages, to better depict the meridional structure. Rather than use per meter as the remaining dimension, we have chosen to use 1 degree latitude, as 111.3 km. Thus the units are given PW per deg.] From approximately 5\(^\circ\)S to 10\(^\circ\)N a semiannual oscillation peaks positively in May through July and November through December, and negatively in January through March and September through October. Uncertainty in \(F_S\) associated with differences in ERA and NRA fields is greatest in the Tropics and high latitudes, where estimates of \(\nabla \cdot F_O\) and \(F_S\) in ERA are larger than in NRA, and in the subtropics, where ERA estimates are less than those of NRA.

Integration of \(F_S\) in time provides an estimate of the annual cycle of equivalent \(O_E\) (Fig. 3b) in \(10^{20}\) Joules per degree latitude for comparison with direct calculations of \(O_E\) (Antonov et al. 2004). Upper ocean energy content for the Pacific, Atlantic and Indian Oceans (Fig. 4) can be similarly compared. In general, the patterns of equivalent \(O_E\) are quite similar to the values of Antonov et al. (2004) outside of the Tropics, except the latter are \(~10-20\%\) larger, even with the corrected physical constants. Differences in equivalent \(O_E\) values derived from \(F_S\) based on ERA and NRA fields are associated primarily with differences in the meridional structure of \(\nabla \cdot F_A\) (Fig. 2d). For the seasonal cycle, these differences translate to smaller (larger) equivalent \(O_E\) in the north (south) equatorial ocean from March to June and larger (smaller) equivalent \(O_E\) from October through December in NRA-based estimates relative to ERA-based estimates; differences that are evident in each of the three ocean basins (Fig. 4).

The zonal mean \(\delta O_E/\delta t\) estimated from GODAS, its differences with WOA, and the implied \(\nabla \cdot F_O\) based on (2) are given in Fig. 5. The general pattern of change in \(O_E\) is quite similar to that of \(F_S\) (Fig. 3a) in the extratropics, with peak values occurring at the solstices. However, \(\delta O_E/\delta t\) exceeds \(F_S\) in the midlatitudes by order 0.01 PW deg\(^{-1}\) in both hemispheres. The implication
then is that, if real, this difference is related to the contribution of $\nabla \cdot F_O$ (Fig. 5b) and interaction between the middle and lower latitudes, as divergence and its areal extent at high latitudes is relatively small. Strong meridional structure in the seasonal cycle of $\nabla \cdot F_O$ (Fig. 5b) consists of divergence maxima in the Tropics and summer subtropics that exceed those at mid-latitudes, and convergence maxima in subtropical winter subtropics and summer mid-latitudes. The complexity of the annual cycle in $\delta O_E / \delta t$ between 15°S and 20°N is not reflected in $F_S$ and thus the key role of tropical ocean current variations and their effects on $\nabla \cdot F_O$ in the mixed layer energy budget is indicated. We will show later that the exchanges are likely related mainly to Ekman transports. Discrepancies between GODAS and WOA estimates are widespread in SH winter and have been associated with the lack of observations in austral winter (Fasullo and Trenberth 2007b).

Uncertainty in $\nabla \cdot F_O$ that stems mainly from uncertainty in $R_T$, the atmospheric energy budget, and $\delta O_E / \delta t$ is quantified later (Fig. 10). The principal features of Fig. 5b are robust in the context of these uncertainties.

b. Extended EOF analysis

To identify first order aspects of variability in the spatially and temporally complex fields of $\nabla \cdot F_O$, a joint EOF analysis is performed that combines the $F_S$ and $\delta O_E / \delta t$ fields and extracts common modes of variability. The first extended joint EOF (Fig. 6) gives the dominant annual cycle and accounts for 63% of the combined variance. The second EOF (not shown), which accounts for 12% of the variance, has a semiannual time series with peaks in April and October-November, and minima in July and January-February. It is dominated by $O_E$ changes in the Tropics, particularly the Indian Ocean, and is not primarily related to $F_S$. We therefore focus here only on the first EOF. In Fig. 6 we rescale both the EOF pattern and the principle component time series so that the latter has
a maximum value approaching unity, and thus the EOF values can be interpreted reasonably well in terms of W m$^{-2}$. It can be seen (Fig. 6) that the patterns in the top two panels are somewhat similar in the extratropics with $\delta O_E/\delta t$ containing an amplitude larger by order 50 W m$^{-2}$. The associated $\mathbf{V} \cdot \mathbf{F}_O$ (Fig. 6c) is tied particularly tightly with $F_S$ and $\delta O_E/\delta t$ at mid- and high-latitudes. In the Tropics, the mode strongly resembles the spatial structure and seasonal coherence of the Inter-Tropical Convergence Zone.

The zonal mean of EOF1 of $\mathbf{V} \cdot \mathbf{F}_O$ (Fig. 7a), simplifies the zonal structure from Figs. 6c,d and the seasonal cycle of $\mathbf{V} \cdot \mathbf{F}_O$ can be described as strong divergence in summer subtropics, and convergence in winter, that are balanced by variability in the Tropics and midlatitudes, and by cross equatorial transports. The annual cycle in midlatitude $O_E$ is thus enhanced by $\mathbf{V} \cdot \mathbf{F}_O$ over that expected from $F_S$ alone by over 0.09 PW $^\circ$lat$^{-1}$ in the NH and 0.06 PW $^\circ$lat$^{-1}$ in the SH. Moreover, the seasonal phase is quite uniform between hemispheres in time, and its spatial structure differs from that of $F_S$, with maxima in the Tropics and subtropics, particularly north of the equator, and in the fine meridional structure near the equator. The primary coherent contributions in EOF1 arise from spatial coherence across the subtropics (8 to 30$^\circ$N and 5 to 29$^\circ$S), with midlatitude and tropical peaks of opposite sign but weaker magnitude, that together compensate for the subtropical variability. The temporal characteristics of spatially coherent domains and the degree of closure between the regions suggested in Fig. 7 are investigated further in Section 4.

c. Principal balances in EOF 1

By integrating Fig. 7a over regions of coherent variability, the basic energy exchanges associated with $\mathbf{V} \cdot \mathbf{F}_O$ between the subtropics and lower and higher latitudes that characterize the first mode can be quantified. As the integral from 65$^\circ$N to 65$^\circ$N yields cancellation during all months (Fig. 7c), and thus on its own satisfies the fundamental global constraint of zero net
divergence, the linear approach is adopted. In the NH (Fig. 7a,b), the principal balance in $\nabla \cdot \mathbf{F_O}$ is between the northern Tropics and subtropics ($8^\circ$N to $30^\circ$N) and the nearly equal contributions from a portion of the deep Tropics ($2^\circ$S to $8^\circ$N) and mid-latitudes ($30^\circ$N to $65^\circ$N). The net magnitude of the implied exchange is $\sim \pm 2.2$ PW. A residual NH energy transport of $\sim \pm 0.6$ PW occurs in phase with the subtropical domain.

In the SH, compensation between the southern Tropics and subtropics ($2^\circ$S to $29^\circ$S) is largely balanced by variability in mid-latitudes that is $\sim 70\%$ stronger than in the NH with the subtropical contribution exhibiting a magnitude of $\pm 3.7$ PW. An equal and opposite residual transport is also evident from the $5^\circ$S to $65^\circ$S integral that is in phase with the southern subtropical contribution and balances that from the NH, and thus results in negligible net divergence from $65^\circ$S to $65^\circ$N.

**d) Implied meridional transports**

From the derived fields of $\nabla \cdot \mathbf{F_O}$, we estimate the ocean’s integrated meridional transport (Fig. 8) and its associated uncertainty, both for the zonal mean annual cycle (Fig. 8a) and the zonal annual mean, both globally and for individual ocean basins (Fig. 8b). There are nine estimates considered in Fig. 8, associated with each permutation of the three estimates of $F_S$ and $\delta O_E/\delta t$. Also shown in Fig. 8b are the mean values obtained from numerous ocean sections. The annual cycle of transport is significant in the Tropics and subtropics, and transports are strongest in the winter hemisphere.

In winter, poleward transports exceed 4 PW from November through April in the NH, and 3 PW from June through October in the SH. Thus the seasonal minimum identified by Zhang et al. (2002) at $24^\circ$N in February is an exception for NH winter, where northward transports are otherwise strong and positive. There is often a broad coherent meridional scale for the estimated
transports. For example, in southern winter, southward transports extend from 15ºN to 60ºS with a substantial cross equatorial flux that exceeds 4 PW in September. Perhaps more surprising is the broad meridional scale of northward transports during December, at which time global ocean transports are northward throughout most of the SH (to 50ºS) and cross equatorial transports again exceed 4 PW. From April to September, the poleward transports of heat in the SH coincide with reductions in ocean heat content (from May to October; Fig. 5), and hence mostly help feed surface energy fluxes into the atmosphere, especially in the tropics and subtropics.

A key to understanding the relationship between divergence and transports is consideration of the summer tropical to subtropical divergence maxima, which contribute substantially to the winter poleward transport and cross-equatorial transports in some months. The importance of higher order modes is underscored, as a large portion of the transport is not accounted for by EOF 1 (Fig. 7). Uncertainty in annual cycle of transport, estimated by two standard deviations (2σ) of the estimates of transport included in Fig. 8, exceeds 0.5 PW in the Tropics and subtropics during much of the year and is greater than 1.0 PW from Dec through Feb in the Tropics and SH subtropics and from Apr through Jul in the Tropics and subtropics (not shown).

The annual zonal mean structure of transport is characterized by a northward maximum at 15ºN of 1.7±0.3 PW, and a southward maximum at 11ºS of 1.2±0.5 PW where the uncertainty is ±2σ. A steep meridional gradient in transport is present in the deep Tropics and is reflective of the strong net energy flux into the ocean on the equator (Fig. 1). Despite the pronounced seasonal variability in cross equatorial transports (Fig. 8a), the annual mean cross equatorial transport is negligible (<0.1 PW) according to the median estimate, with an upper bound transport estimated from the ±2σ range of 0.6 PW.

The structure of global annual mean transports is generally supported by observed oceanographic sections (Fig. 8), with close agreement at 47ºN and 30ºS and agreement within the
range of uncertainty at 24ºN, 8ºN, and 20ºS. At 36ºN, the observed estimate of Talley (2003) exceeds the values derived here by about 0.6 PW, which lies beyond the 20% (0.3 PW) error range provided by Talley. Our NH northward transports are systematically less than the ocean estimates between 0 to 40ºN and this bias occurs in the North Atlantic.

Mean transport in the Atlantic Ocean is northward north of 40ºS, while in the Indian Ocean, it is southward at all latitudes. In the Pacific Ocean, transport is positive in the NH and negative in the SH. As in Trenberth and Caron (2001), the magnitude of mean transports is comparable among the basins, with a southward peak of 0.8 PW at 12ºS in the Indian Ocean, a northward peak of 0.8 PW at 40ºN in the Atlantic Ocean, and for the Pacific Ocean, northward and southward peaks of 0.9 PW at 13ºN and 0.6 PW at 10ºS, respectively.

In the Atlantic, agreement with some direct ocean estimates is good, including those of B97 at 47ºN, S96 at 36ºN, and T03 at 8ºN and 18ºS. In other instances, disagreement between the estimates is large, such as for T03 at 47ºN, M98 at 8ºN, and S96 and M98 at 8ºS, although in several of these cases there are also disagreements between the direct ocean estimates that are outside the estimated error bars. Agreement with observed sections in the Pacific is good for T03 at 47ºN and numerous estimates at 30ºN, and for both the Indian and Pacific basins at 30ºS. At 10ºN in the Pacific the two direct estimates of M98 and T03 are at odds and our value is in between. Uncertainty in transports, based on the range of estimates derived herein, is largest in the Indian Ocean, where sampling of the upper ocean is sparse (Locarini et al. 2006), variability in $F_S$ is large, and the heat budget of the upper ocean is complex (Loschnigg and Webster 2000).

4. Discussion

While agreement between estimates of transport derived herein and those from ocean sections is reassuring, it is not definitive evidence that the transports are correct, as both the values
derived herein and the values reported from ocean sections are subject to considerably uncertainty (e.g., Bryden et al. 2005) and temporal sampling is an issue (Koltermann et al. 1999) as indicated for instance by variability of Florida Strait cable estimates of current transports (Baringer and Larson 2001). Similarly, instances in which disagreement between the estimates are large does not preclude the validity of our analysis. Moreover, in some regions, such as in the tropical Pacific and northern Atlantic Oceans, it is clear that the observations themselves are mutually inconsistent. Indeed, in many instances where the observations are at odds, the transports provided herein represent a reasonable compromise among the observed values. Hence the current analysis thus represents an important additional basis for assessing ocean energy transports.

a. Consistency of results: global constraints

Here we further briefly explore the robustness of our findings based on global constraints (Fig. 9), a topic dealt with in more detail by FT07. TOA solar energy coming into the Earth system is a maximum January 3rd at the time of perihelion, with annual cycle amplitude of 3.4%. From ERBE, the net globally integrated $R_T$ peaks in February of 4.5 PW, as the displacement from the January peak in solar irradiance results in February from lower albedo, related to changes in cloud and other factors, and the annual minimum of outgoing longwave radiation. The minimum in $R_T$ in June and July is ~5 PW. This imbalance in $R_T$ should be reflected primarily in both $F_S$ and $O_E$, as energy storage in the atmosphere and on land is small. The larger ocean area in the SH dominates the global ocean annual cycle, which in turn balances $R_T$. However, in the NH there is a major transport of energy from ocean to land in winter by atmospheric winds (Fig. 2 and FT07), so that the TOA radiation is not the only driver of $F_S$ over the oceans. Net evaporation over the oceans and transport of latent energy to land areas results in annual cycle variations in water (or
snow and ice) on land and sea level, as observed by altimetry (Minster et al. 1999) with amplitude equivalent to 9.5 mm of sea level, peaking in September.

Globally, the difference between the TOA radiation and $F_S$ over ocean (Fig. 9) arises primarily from the small contributions from land storage tendency (FT07). While some energy storage also occurs in sea ice as a deficit built up in winter that is released in summer, $F_S$ over ocean appears principally as a change in $O_E$; and presumably the energy tendency associated with sea ice largely cancels between the two hemispheres when integrated globally. The ocean temperature datasets imply a larger annual cycle of $O_E$ than do $F_S$ estimates, and tendencies are outside the error bars in southern winter (too low) and in October-November (too high) (Fig. 9), and correspond to $O_E$ values that are too high in March-April and too low in August to October. In FT07 and Fasullo and Trenberth (2007b) it is shown that the errors most likely arise from $O_E$ south of 40ºS, where WOA and JMA values are further astray from GODAS.

b. Ekman transports

As given in Fig. 5b, the biggest differences between the results in Figs. 3 and 5 and those of Antonov and Levitus (2004) are in the Tropics, notably from ~5 to 15ºN, and especially in the Pacific where there is strong evidence for the annual cycle being dominated by ocean dynamics. The zero wind stress curl line over the North Pacific migrates from 11ºN in March to 20ºN in September and induces upper layer thickness anomalies across the Pacific Basin resulting in major seasonal changes in the North Equatorial Current (NEC) and where it bifurcates along the western boundary. The NEC is farthest north in October and farthest south in February (Qiu and Lukas 1996). Furthermore, large seasonal changes in Ekman transports lead to a substantial annual cycle in northward energy transports throughout the Tropics (Jayne and Marotzke 2001) and at 24ºN (Zhang et al. 2002) range from about zero in winter to maxima in July and November of 1 PW.
In the Atlantic, the NEC peaks in boreal summer and weakens during spring and fall (Arnault, 1987). Boning and Herman (1994) present results for a North Atlantic Ocean model simulation and the annual cycles of surface fluxes and $O_E$ show similar results to those presented here. In the Atlantic at 8ºN they also find a large annual cycle in northward energy transports in which the strong NEC-induced changes in $O_E$ have little to do with $F_S$, but depend rather on changes in surface stress. Kobayashi and Imasato (1998) determine that substantial annual cycles exist in meridional energy transports in the Atlantic and Pacific, with largest values around 10ºN. Seasonal variability in energy transport of 100% is also suggested for the Indonesian Throughflow, and in the Indian Ocean the annual cycle just south of the equator is suggested to be +1.4 PW in December through February and −1.8 PW in June through September (Loschnigg and Webster 2000). The regional results appear to be consistent with the inferred fields presented here.

For the total meridional transport at 24ºN, Zhang et al. (2002) estimate that the annual cycle ranges from 1.1 PW in February to 2.8 PW in August, with a mean of 2.1±0.4 PW. The inference is that in both the Atlantic and the Pacific, transports enhance the annual cycle of $O_E$ from what it would be based on $F_S$ alone by about 10%, as was shown for the Atlantic in a model by Böning and Herrmann (1994). While the transports at 24ºN from Zhang et al. (2002) are consistent with the fields derived here, albeit somewhat stronger, the behavior at 24ºN is not representative generally of variability at other latitudes (Fig. 8), owing to the steep gradients in transport that exist in the subtropics seasonally and their complex relationship to cross equatorial flow and midlatitude variability, which is relatively weak. Indeed, in model simulations, Jayne and Marotzke (2001) demonstrate the dominant role of the annual cycle in Ekman transports associated with broad overturning circulations in the Tropics and subtropics that contribute to strong energy transport and underlie the deduced changes in $O_E$. Based on the surface forcing they use, Ekman transports reverse sign sharply at about 25º latitude in each hemisphere through the course of the
annual cycle. Hence the dipole structures seen in Figs. 6 and 7 near the Tropics and midlatitudes in both hemispheres are qualitatively consistent with established changes due to Ekman transports. Moreover, the suitability of assessing these transports near the node of the overturning circulations, where the gradient in transport is strong, rather than between the nodes, where sensitivity of estimates to the circulation’s precise meridional location is less, is called into question.

c. Factors contributing to uncertainty

Terms that contribute to uncertainty in $\nabla \cdot \mathbf{F}_O$ include contributions from: 1) $R_T$ (Fig. 10a) – estimated from ERBE and CERES period FM1 and FM2 fluxes, 2) the atmospheric budget (Fig. 10b) – estimated from NRA and ERA during ERBE and NRA during CERES, and 3) $\delta O_E/\delta t$ (Fig. 10c) – estimated from the WOA climatology and JMA and GODAS fields during the 1990s. The estimated population standard deviation of zonal means for each term (Fig. 10a-c) and, the zonal mean structure of uncertainty in monthly means (Fig. 10d) are shown. The estimates are independent (e.g., ERBE and CERES fluxes are not used in the compilation of the reanalysis or ocean datasets) and there is no expectation that cancellation of uncertainty will occur when the fields are combined to infer $\nabla \cdot \mathbf{F}_O$. Also, as some of the differences between the fields in Fig. 10 represent real differences between the ERBE, CERES, and WOA time periods, the estimates (Fig. 10) somewhat overstate the uncertainty associated with analysis methods. At TOA, uncertainty associated with $R_T$ is largest in the SH but it is still small relative to other terms (less than 0.01 PW deg$^{-1}$). Uncertainty in $\nabla \cdot \mathbf{F}_A$, particularly in the Tropics, is a substantial contributor to the uncertainty in $\nabla \cdot \mathbf{F}_O$, with values exceeding 0.05 PW deg$^{-1}$ from March through November. Uncertainty associated with both the atmospheric budget and ocean tendencies is large in the subtropics but near the node of EOF 1 (Fig. 7), uncertainty returns to a relative minimum. In midlatitudes, large uncertainty is associated primarily with disagreements in estimates in the ocean.
tendencies, however it decreases towards the poles. In the extratropics, where the moisture holding capacity of the atmosphere is greatly reduced from the Tropics, agreement between ERA and NRA estimates of $\nabla \cdot \mathbf{F}_A$ is better and uncertainty is reduced. For the ocean, uncertainty associated with $\delta O_E / \delta t$ is greatest in SH midlatitudes, and is likely to be associated with the lack of observations, and therefore the disproportionate reliance on infilling techniques used in constructing ocean analyses in these regions. As the overall trend in ocean temperatures is small across the 1990s, the annual-zonal mean of $\delta O_E / \delta t$ is small compared to other terms (Fasullo and Trenberth 2007b).

Together, the various contributions to uncertainty in $\nabla \cdot \mathbf{F}_O$ represent a spatially complex pattern of diverse contributions with distinct underlying causes. Despite uncertainty, the principle features of Fig. 6 are larger (order 0.2 PW deg$^{-1}$) than the combined uncertainty of terms in Fig. 10 (order 0.05 PW deg$^{-1}$) and thus it is likely that its primary features are robust to data shortcomings. A caveat to these conclusions would be if there is a substantial systematic error across all the estimates for a particular field in Fig. 10. For instance, based on analysis of data from ARGO floats, it has recently been suggested that WOA overstates the annual cycle in the North Atlantic Ocean (Ivchenko et al. 2006) and perhaps globally; see also Willis et al. (2007). The potential presence of such biases is particularly likely for ocean analyses, which share both similar analytical bases and observational shortcomings. Revisions of ocean datasets that have been suggested (Ivchenko et al. 2006) would alter the values presented here.

5. Concluding remarks

A test of how well we understand the global energy budget of the Earth system and the oceans is to examine the annual cycle in detail and the degree to which closure can be reached among a variety of independent data. Significant advances in the observation and analysis of TOA,
atmospheric, and ocean budget have been made in recent decades such that the errors are now relatively modest and allow for an initial estimate of ocean energy divergence.

In midlatitudes, the seasonal ocean energy tendency is dominated by surface fluxes, with divergence of energy from ocean currents enhancing magnitudes by order 50 W m$^{-2}$. This finding is consistent with model results and explained by Ekman transports driven by the annual cycle of surface winds. In contrast, in the Tropics and subtropics, divergence is found to play the dominant role in the upper ocean energy budget with secondary but important contributions from surface fluxes.

The first mode of the annual cycle of divergence and the dominant aspects of spatial coherence in this mode are identified and the total transports are calculated, globally and by basin. An additional important implication is that the annual cycle of energy transports is primarily a response to surface winds, rather than instabilities associated with temperature gradients, and it increases seasonal energy extremes in the upper ocean and the amplitude of the annual cycle. This contrasts with the thermohaline circulation, which feeds on density (and thus temperature) gradients.

Inferred annual mean ocean heat transports are somewhat lower than direct ocean estimates in the North Atlantic and thus for the zonal average ocean. Although there are uncertainties in the atmospheric energy transports, there is not much scope for the ocean transports to be increased much as their sum is quite strongly constrained. The fields used to infer ocean energy divergence here are indirect and are sensitive to the accumulation of error through the depth of the atmosphere and oceans. While retrievals of radiative fluxes of TOA are among the most accurate fields to be observed globally, uncertainty in atmospheric divergence and ocean energy tendency can be substantial, particularly on regional scales. With further refinements of these fields and more complete temporal sampling in the ocean, a more accurate diagnosis of ocean energy divergence
will be possible for comparison with model assimilated fields from analysis of new ocean data. With sufficient data and continued improvements, it will also be possible to extend these kinds of analysis to examine interannual variability. This task has already been realized for the tropical Pacific to examine changes in energy content with ENSO (Trenberth et al. 2002), although improvements in data quality continue to be desirable. With new and improved TOA radiation from the CERES and much better spatial and temporal from global ARGO float measurements, the prospects for doing this diagnosis routinely should become realistic. Accordingly, this approach has the potential to provide a more holistic view of the climate system, help validate ocean models, and better assess the role of ocean energy uptake, release and transport in climate variability.

Acknowledgments. This research is partially sponsored by the NOAA CLIVAR and CCDD programs under grant NA17GP1376 and NA04OAR4310073. NCAR is sponsored by the National Science Foundation. We thank Bill Large, Frank Bryan and Clara Deser for comments and Dave Stepaniak for help with computations.
References


Figure captions

Figure 1: a) Annual mean $F_S$ as computed by residual from $R_T$ and NRA atmospheric energy budgets for the ERBE period in W m$^{-2}$. Departure from the annual mean of the b) JJA and c) DJF surface flux out of the ocean in W m$^{-2}$ is also shown. Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by more than 10 W m$^{-2}$ in a) and by more than 30 W m$^{-2}$ in b) and c).

Figure 2: Zonal mean of the total (left) and departure from the annual mean (right) for the incoming radiation at TOA (top) and column integrated atmospheric energy divergence (bottom) over the oceans in W m$^{-2}$ for the ERBE period. Stippling (hatching) denote areas where ERBE and NRA based estimates exceed (fall below) those of CERES and ERA by more than 0.02 PW deg$^{-1}$.

Figure 3: Zonal mean over the world oceans of the departure from the annual mean of the a) net surface flux out of the ocean in PW deg$^{-1}$. Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by more than 0.02 PW deg$^{-1}$; b) vertically integrated equivalent $O_E$ corresponding to Fig. 3a. Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by $5 \times 10^{19}$ J deg$^{-1}$.

Figure 4: Zonal mean of the annual cycle of equivalent $O_E$ for the a) Pacific, b) Atlantic and c) Indian Oceans. Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by 2.5 $\times 10^{19}$ J deg$^{-1}$.

Figure 5: Zonal integral over the world oceans of a) $O_E'$, b) $\delta O_E/\delta t$ and c) inferred $\nabla \cdot F_O$ in PW deg$^{-1}$. Hatching (stippling) in a) corresponds to instances in which WOA values of monthly values of $\delta O_E/\delta t$ exceed (fall below) those of WOA by 0.03 PW deg$^{-1}$, or $10^{20}$ J o'lat$^{-1}$ for $O_E'$.

Figure 6: First EOF of the combined extended EOF analysis for $F_S$ and $O_E$, plus their principle component time series. a) $F_S$, b) $\delta O_E/\delta t$, c) the inferred $\nabla \cdot F_O$ from (2). The principal component time series (d) is scaled to have extremes of ±1, so that the EOFs approximately depict units of W m$^{-2}$.

Figure 7: a) The zonal mean of the implied $\nabla \cdot F_O$ from panel c) of Fig. 6 is shown for the annual cycle, in PW deg$^{-1}$. Regional integrals of $\nabla \cdot F_O$ EOF 1 (Figs. 6c, 7a) for zones as labeled in b) and c). Zones are chosen to isolate the main regions of coherence from Fig. 7a and to illustrate fundamental balances.

Figure 8: Integrated meridional ocean heat transports and their uncertainty are shown for the a) zonal mean annual cycle, b) zonal annual global mean, and zonal ocean basin means. Stippling in a) represents regions and times of year in which two standard deviations of monthly mean values among estimates, some of which include the CERES period (see text), exceeds 0.5 PW. Transport in b) is the median annual mean transport by latitude from all estimates and the associated ±2σ range (shaded). Observational estimates include Bacon 1997 (B97), Bryden et al. 1991 (B91), Ganachaud and Wunsch 2000 (GW00), Holfort and Siedler 2001 (HS01), Klein et al 1995 (K95), Lavin et al. 1998 (L98), Macdonald 1998 (M98), Macdonald and Wunsch 1996 (MW96), Robbins and Toole 1997 (RT97), Roemmich et al. 2001 (R01), Sato and
Rossby 2000 (SR00), Saunders and King 1995 (SK95), Speer et al. 1996 (S96) and Talley 2003 (T03).

Figure 9: Global mean annual cycles for the ERBE period of $R_T$ (solid/blue), $F_S$ over ocean (dot/purple), and GODAS $\delta O_E/\delta t$ (dash/grey) in PW. Shading represents $\sigma$ of monthly means for each field.

Figure 10: Uncertainty analysis of terms contributing to estimation of $\nabla \cdot F_O$ including a) $R_T$, b) $\nabla \cdot F_O + \delta A_E/\delta t$, c) $\delta O_E/\delta t$, and d) the zonal means of the uncertainty in monthly means of each term. Estimates of uncertainty are based on the standard deviation among three estimates of each field including: 1) $R_T$ derived from ERBE-period and CERES-period FM1 and FM2 fluxes, b) estimates of $\nabla \cdot F_O$ and $\delta A_E/\delta t$ taken from NRA and ERA during ERBE, and NRA during CERES, and c) estimates of $\delta O_E/\delta t$ from WOA fields, and from JMA and GODAS fields during the 1990s.
Figure 1: a) Annual mean $F_S$ as computed by residual from $R_T$ and NRA atmospheric energy budgets for the ERBE period in W m$^{-2}$. Departure from the annual mean of the b) JJA and c) DJF surface flux out of the ocean in W m$^{-2}$ is also shown. Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by more than 10 W m$^{-2}$ in a) and b more than 30 W m$^{-2}$ in b) and c).
Figure 2: Zonal mean of the total (left) and departure from the annual mean (right) for the incoming radiation at TOA (top) and column integrated atmospheric energy divergence (bottom) over the oceans in W m$^2$ for the ERBE period. Stippling (hatching) denote areas where ERBE and NRA based estimates exceed (fall below) those of CERES and ERA by more than 0.02 PW deg$^{-1}$. 
Figure 3: Zonal mean over the world oceans of the departure from the annual mean of a) the net surface flux out of the ocean in PW deg\(^{-1}\). Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by more than 0.02 PW deg\(^{-1}\); b) vertically integrated equivalent \(O_E\) corresponding to Fig. 3a. Stippling (hatching) denotes areas where NRA based estimates exceed (fall below) those of ERA by \(5 \times 10^{19}\) J deg\(^{-1}\).
Figure 4: Zonal mean of the annual cycle of equivalent $O_E$ for the a) Pacific, b) Atlantic and c) Indian Oceans. Stippling (hatching) denote areas where NRA based estimates exceed (fall below) those of ERA by $2.5 \times 10^{19} \text{ J deg}^{-1}$.
Figure 5: Zonal integral over the world oceans of a) $O_E'$, b) $\partial O_E/\partial t$ and c) inferred $\nabla \cdot \mathbf{F}_O$ in PW deg$^{-1}$. Hatching (stippling) in a) corresponds to instances in which WOA values of monthly values of $\partial O_E/\partial t$ exceed (fall below) those of WOA by 0.03 PW deg$^{-1}$, or $10^{20}$ J o'lat$^{-1}$ for $O_E'$. 
Figure 6: First EOF of the combined extended EOF analysis for $F_S$ and $O_E$, plus their principle component time series. a) $F_S$, b) $\delta O_E/\delta t$, c) the inferred $\nabla \cdot F_O$ from (2). The principal component time series (d) is scaled to have extremes of $\pm 1$, so that the EOFs approximately depict units of $W \text{ m}^{-2}$. 
Figure 7: a) The zonal mean of the implied $\nabla \cdot F_O$ from panel c) of Fig. 6 is shown for the annual cycle, in PW deg$^{-1}$. Regional integrals of $\nabla \cdot F_O$ EOF 1 (Figs. 6c, 7a) for zones as labeled in b) and c). Zones are chosen to isolate the main regions of coherence from Fig. 7a and to illustrate fundamental balances.
Figure 8: Integrated meridional ocean heat transports and their uncertainty are shown for the a) zonal mean annual cycle, b) zonal annual global mean, and zonal ocean basin means. Stippling in a) represents regions and times of year in which two standard deviations of monthly mean values among estimates, some of which include the CERES period (see text), exceeds 0.5 PW. Transport in b) is the median annual mean transport by latitude from all estimates and the associated ±2σ range (shaded). Observational estimates include Bacon 1997 (B97), Bryden et al. 1991 (B91), Ganachaud and Wunsch 2000 (GW00), Holfort and Siedler 2001 (HS01), Klein et al 1995 (K95), Lavin et al. 1998 (L98), Macdonald 1998 (M98), Macdonald and Wunsch 1996 (MW96), Robbins and Toole 1997 (RT97), Roemmich et al. 2001 (R01), Sato and Rossby 2000 (SR00), Saunders and King 1995 (SK95), Speer et al. 1996 (S96) and Talley 2003 (T03).
Figure 9: Global mean annual cycles for the ERBE period of $R_T$ (solid/blue), $F_S$ over ocean (dot/purple), and GODAS $\partial O_E/\partial t$ (dash/grey) in PW. Shading represents $\sigma$ of monthly means for each field.
Figure 10: Uncertainty analysis of terms contributing to estimation of $\nabla \cdot \mathbf{F}_O$ including a) $R_T$, b) $\nabla \cdot \mathbf{F}_O + \delta A_E/\delta t$, c) $\delta O_E/\delta t$, and d) the zonal means of the uncertainty in monthly means of each term. Estimates of uncertainty are based on the standard deviation among three estimates of each field including: 1) $R_T$ derived from ERBE-period and CERES-period FM1 and FM2 fluxes, b) estimates of $\nabla \cdot \mathbf{F}_O$ and $\delta A_E/\delta t$ taken from NRA and ERA during ERBE, and NRA during CERES, and c) estimates of $\delta O_E/\delta t$ from WOA fields, and from JMA and GODAS fields during the 1990s.