Meridional Ekman Heat Transport: Estimates from Satellite Data

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ABSTRACT

Analyses of satellite-derived SSM/I winds and AVHRR sea surface temperatures are used to compute weekly estimates of global meridional ocean Ekman heat transport for the 4-year period 1987–1991. The heat transport is consistently poleward throughout the year over the Atlantic and much of the Pacific between 30°S and 30°N and equatorward at higher latitudes. The zonally integrated Ekman heat transport in the Pacific was weak and equatorward at 10°N in September 1989 and 1990, whereas in other years it is poleward throughout the year. In the Indian Ocean, equatorward heat transport was strongest in Northern Hemisphere summer 1990. The weekly time series provides better temporal resolution than previous studies that at best used monthly averages. The higher-frequency variations are explored through rotated empirical orthogonal functions (REOFs) of nonseasonal heat transport anomalies. The REOFs show large-scale coherence across the tropical and subtropical Pacific and Indian Oceans. The first REOF has a strong spectral peak at periods of 50–60 days and is dominated by the variability in the Southern Hemisphere Indian Ocean. The second REOF is dominated by the variance in the Northern Hemisphere Indian Ocean and western tropical North Pacific. The first four REOFs, which explain 22.5% of the nonseasonal variance, have spectral peaks at periods consistent with the 30–60 day atmospheric Madden–Julian oscillation. These periods were not resolved in the monthly averaged data from COADS. Singular spectrum analysis has been used as a filter to show the long timescale variations. The early phase of the 1988–89 La Niña has a strong influence on the heat transport, indicating enhanced poleward heat transport in the eastern Indian Ocean, tropical North Pacific, and tropical and subtropical South Pacific and enhanced equatorward heat transport in the midlatitude North Pacific and western tropical Indian Ocean.

1. Introduction

Heat transport due to surface Ekman dynamics is an important component of the total oceanic heat transport. At the tropic circles, Kraus and Levitus (1986) found that about one-half of the total heat transport in the Atlantic and nearly all of the heat transport in the Pacific was due to Ekman effects. The model simulation studies of Bönig and Hermann (1994) show that this component dominates the contributions from the horizontal gyre circulation equatorward of the 35° parallel. Those studies confirm the theoretical analysis of Gill and Niiler (1973) that, except for the equatorial zone where the structure of the meridional circulation is baroclinic, the response of the ocean to the seasonal and shorter timescale variability in the surface layer is primarily barotropic. In the Tropics, surface heat fluxes and the balance between heat transport due to Ekman dynamics and that due to other baroclinic processes are responsible for the large seasonal variations in heat storage (e.g., Gill and Niiler 1973; Pares-Sierra et al. 1985; Sarmiento 1986). Elsewhere, the transport divergence has a minor impact on the local heat budget, and surface heat flux variations alone are primarily responsible for seasonal variations in heat storage (Gill and Niiler 1973; Sarmiento 1986). For longer timescales, as shown by Gill and Niiler, Ekman dynamics may have more influence on heat storage and large-scale sea surface temperature variations.

The meridional Ekman heat transport is the only component of the total global heat transport budget for which time variations can be estimated directly from existing observations. Levitus (1987) estimated the annual cycle of zonally integrated meridional heat transport using monthly mean wind stress climatology from Helleman and Rosenstien (1983) and monthly mean sea surface temperature (SST) climatology from Levitus (1982). The annual cycle of heat transport is similar in the Pacific and Atlantic Oceans, although the zonally integrated heat transport in the Pacific is much

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larger. Zonally integrated heat transport in the tropical and subtropical Indian Ocean is large (greater than 3.0 pW) and southward from April through October, with the most intense heat transport during Northern Hemisphere summer due to the Indian monsoon.

Temporal and spatial variations in meridional Ekman heat transport were estimated by Adamec et al. (1993; henceforth referred to as ARV) using monthly means of wind stress and SST from the Comprehensive Ocean-Atmosphere Data Set (COADS). Poleward heat transport of up to $6 \times 10^8$ W m$^{-1}$ was estimated in the tropical Atlantic and Pacific Oceans throughout the year. Empirical orthogonal functions (EOFs) of the Ekman heat transport were used to investigate interannual variability and to highlight the various phases of Ekman heat transport during Los Niños. The first EOF indicated increased equatorward Ekman heat transport in the central and western tropical Pacific in the initial phase of Los Niños, followed by increased poleward heat transport in the eastern tropical Pacific.

ARV used monthly averaged winds to calculate the monthly mean wind stresses. The magnitudes of monthly mean wind stress estimated in such a manner can differ substantially (by up to 50% in the midlatitudes of the North Pacific) from those calculated from daily wind stress (Rienecker et al. 1994). Kraus and Levitus (1986; henceforth referred to as KL) suggest that the Ekman heat transport calculated from monthly averaged quantities may underestimate the monthly averaged heat transports by up to 50% compared to estimates calculated from instantaneous quantities.

Previous estimates of meridional Ekman heat transport (e.g., Levitus 1987; ARV) have been based on in situ observations. Satellite data provides better spatio-temporal resolution than in situ measurements over the entire globe and especially over the southern ocean. Implicit in the use of satellite data (or other surface data) is the assumption that the Ekman layer depth is less than that of the mixed layer. Both KL and ARV discuss the assumption in detail and argue that useful estimates of the Ekman heat transport can still be made even for the few instances when this assumption is violated. When Ekman depths exceed the mixed layer depth, most of the transport within the Ekman layer would take place within the mixed layer since the Ekman velocity profile decays exponentially with depth.

In this paper, time-varying meridional Ekman heat transport is calculated for the World Ocean for the 4-year period of July 1987 through June 1991 using satellite-derived weekly mean values of SST and wind stress on a 2° grid. The magnitude and temporal variations of those weekly estimates are compared with previous estimates of the Ekman heat transport. The higher-frequency signals contained in the weekly time series are also explored.

The data used to compute the meridional Ekman heat transport are described in section 2. Procedures, as well as the sensitivity of the Ekman heat transport calculation to different assumptions for the depth range of the return flow and to different averaging schemes, are detailed in section 3. Analyses of global maps, zonally integrated heat transport, and nonseasonal variability are presented in section 4, and the findings are summarized in section 5.

2. Data

a. Sea surface temperatures

The SSTs are daytime weekly mean data produced by the University of Miami/Rosenstiel School of Marine and Atmospheric Science from NOAA’s Advanced Very High Resolution Radiometer (AVHRR). The SST data used in this study are averaged from a global 18-km grid to a 2° grid. The SST product uses Laplacian interpolation to provide SST estimates when AVHRR data is not available. The AVHRR SST data density distribution (Fig. 1), that is, the percentage of time for each grid point for which AVHRR data are available, shows that the best coverage is between 30°S and 30°N. In the higher latitudes of each hemisphere, the tropical Western Pacific, and near 20°N in the Eastern Atlantic the data density is lower. These are areas of high cloudiness. In spite of cloud-affected areas, observations are, in general, present over 80% of the time between 30°N and 30°S. Early years, AVHRR data were not recovered poleward of 50° in the winter hemisphere. This cutoff was part of the original satellite data-retrieval system and contributes to the low data density in the high latitudes. In the Tropics, the data distribution is better than that of COADS (e.g., ARV), which is primarily limited to widely spaced ship tracks.

Subjective quality control was performed on the SSTs by investigation of seasonal means, variances, zonal averages, and comparisons with climatology. These analyses identified areas and times when the SSTs were suspect, particularly east of Canada, north of Japan, in the North Sea, and south of Australia. These regions were clearly affected by the incorporation of the ice boundary into the original Laplacian interpolation scheme. When possible, erroneous SSTs were replaced using temporal interpolation. Comparison with an independent climatology also revealed anomalous high SSTs south of 40°S during austral winter in all years except 1989. During the entire 4-yr period, data were not available for four separate weeks, none of which were contiguous. Due to the low data density poleward of 50° latitude, the Ekman heat transport calculation is limited to 50°S to 50°N.

The 4-yr monthly averaged SST for January and July are shown in Fig. 2. The strong meridional temperature gradients of the Gulf Stream and the Kuroshio are well resolved. Equatorial upwelling can be seen in the Eastern Pacific in both months shown. The SST distributions are in general agreement with the climatology derived by Bottomley et al. (1990) and that of ARV,
except that the eastern equatorial Pacific is colder than climatology by over 2°C in January.

b. Wind stress

The wind stress is calculated using the Atlas SSM/I wind analyses, which are based on data from the Special Sensor Microwave Imager (SSM/I) on the Defense Meteorological Satellite Program (DMSP) platform. A variational analysis method is used to produce surface wind vector analyses from SSM/I-derived wind speeds, ship and buoy observations, and 1000-mb winds from the European Centre for Medium-Range Weather Forecasts (ECMWF). The constraints and weights used in the analysis may be found in Atlas et al. (1993).

The Atlas SSM/I winds are available at 6-h intervals on a 2° latitude by 2.5° longitude grid and span the period from July 1987 to June 1991. For this study, the winds have been linearly interpolated to a 2° by 2° grid subsampled to a 12-h interval and reduced from a height of 19.5 m to 10 m using a neutrally stable logarithmic boundary-layer algorithm. Occasional (e.g., December 1987, January 1988, March 1989) satellite problems led to data gaps in the SSM/I time series. These gaps were filled with ECMWF analyses at 10 m. The merged dataset was then used to produce weekly averaged wind stress from 12-hourly wind stress data. It should be kept in mind that Rienecker et al. (1994) found that in the North Pacific the Atlas SSM/I wind stresses are stronger than those from ECMWF—by about 0.01 N m⁻² on average and by up to 0.05 N m⁻² in some midlatitude areas during winter. The differences are usually less than the underlying model variances.

The zonal wind stress, \( \tau_x \), was calculated using the bulk formula:

\[
\tau_x = \rho_a c_D |u| u,
\]

where \( \rho_a \) is the density of air (1.3 kg m⁻³), \( |u| \) is the wind speed, \( u \) is the zonal wind component of the surface wind, and \( c_D \) is the drag coefficient calculated according to the Large and Pond (1981) formulation:

\[
c_D = \begin{cases} 
1.14 \times 10^{-3} & \text{for } |u| < 10 \text{ m s}^{-1} \\
(0.49 + 0.065|u|) \times 10^{-3} & \text{for } |u| > 10 \text{ m s}^{-1}.
\end{cases}
\]

The zonal component of the 4-year monthly averaged wind stress for January and July is shown in Fig. 3. In January easterlies prevail in the Tropics, except for an area in the Indian Ocean, the western equatorial Pacific, and small areas in the eastern Pacific and Atlantic Oceans. The strongest easterlies occur in the Northern Hemisphere. Westerlies prevail at the mid- and high latitudes. The wind stresses at the higher latitudes of each hemisphere are stronger in hemispheric winter than in hemispheric summer due to a higher frequency and intensity of storms during winter.

3. Procedures

The meridional ocean Ekman heat transport is calculated as in Levitus (1987):

\[
F = -c_p \tau_y [\text{SST} - \bar{\theta}] / f, \tag{1}
\]

where \( c_p \) is the specific heat of seawater (4000 J kg⁻¹ K⁻¹), \( \bar{\theta} \) is the vertically averaged monthly mean potential temperature, and \( f \) is the Coriolis parameter. The studies of Levitus (1987), KL, ARV, and Bryden et al.
(1991) use slightly higher values for $c_p$, resulting in Ekman heat transport estimates about 4.5% higher than would have been obtained from the lower $c_p$ value (e.g., Gill 1982) used in this study. Positive (negative) values of $F$ represent poleward (equatorward) heat transport.

For mass conservation, a return flow is needed to balance the surface Ekman flow. It is assumed that the heat associated with the return flow can be approximated according to a depth-averaged potential temperature, $\theta$, which is evaluated from the Levitus (1982) climatology. An assumption must be made concerning the depth range of the return flow and so the region over which $\theta$ is to be averaged. Levitus (1987) and KL used $\theta$ averaged throughout the water column. Bryden et al. (1991), using about 2.5 months of CTD station data along 24°N in the Pacific, estimated that the return flow occurs within the upper 700 m of the water column. However, Böning and Herrmann (1994), using a high-resolution model, found that the return flow extends to at least 3000 m in the extratropics and that formula (1) with $\bar{\theta}$ an average over the entire water column provides a good estimate of the meridional Ekman heat transport for the global ocean. The sensitivity of (1) to different depth range assumptions is explored below. The Ekman heat transport was calculated for the global ocean outside regions masked according to bathymetry of 1000 m or less. The latitudinal band from 4°S to 4°N is excluded from this study to avoid the singularity in the Ekman heat transport at the equator.

**Sensitivity studies**

The global Ekman heat transport is calculated according to (1) with $\bar{\theta}$ determined over three different depth ranges:

(i) the entire water column, $\bar{\theta}_o$,
(ii) 100–3000 m, $\bar{\theta}_r$, and
(iii) 100–1000 m, $\bar{\theta}_c$.

The sensitivity of the calculation to the depth range for $\bar{\theta}$ is summarized in the annual mean zonally integrated heat transport (Fig. 4). For the World Ocean, the heat transport calculated using $\bar{\theta}_o$ is smaller than that calculated using $\bar{\theta}_r$ by up to 5%–10% in the Tropics and 10%–15% in the midlatitudes. The heat transport calculated using $\bar{\theta}_c$ is smaller than that calculated using $\bar{\theta}_r$ by up to 20%–25% in the Tropics and 40%–50% in the midlatitudes. These results also hold in the separate ocean basins (not shown). Using a transect along 24°N in the Pacific, Bryden et al. (1991) estimated the Ekman heat transport based on climatological wind stress to be 0.36 pW if the return flow is con-
fined to the upper 700 m and 0.92 pW if the return flow occurred over the entire water column, the latter being in close agreement with the annual mean value of 0.86 pW estimated in this study. The heat transport based on $\bar{\rho}$, as used in the remainder of this study, serves as an upper bound for heat transport estimates and is supported by the model calculations of Böning and Hermann (1994).

As mentioned above, the use of monthly averaged components in a nonlinear calculation can grossly underestimate the desired monthly averaged fields. The impact of different averaging schemes on global monthly averaged meridional Ekman heat transports is assessed by a comparison of

- $F_A$ using weekly averaged $\tau_x$ and SST, that is, $\bar{\tau_x}(u)$ SST,
- $F_B$ using monthly averaged $\tau_x$ and SST, that is, $\bar{\tau_x}(u)$ SST,
- $F_C$ using monthly averaged $\tau_x$, derived from monthly averaged winds $\bar{u}$, and monthly averaged SST, that is, $\bar{\tau_x}(\bar{u})$ SST,

where the overbar denotes monthly average.

Global maps of Ekman heat transport for January and July (not shown) reveal that $F_B$ is within ±10% of $F_A$ in the Tropics and ±33% of $F_A$ in the midlatitudes. In the Tropics, $F_C$ is always less than $F_A$, sometimes by as much as 47%, but usually by less than 20%; $F_C$ is always less, by between 33% and 75%, than $F_A$ in the midlatitudes. These results reflect the higher variability of zonal wind stress at higher latitudes. The annual mean zonally integrated Ekman heat transport from $F_C$ is smaller than $F_A$ everywhere (Fig. 5). The curve for $F_B$ is not discernible in Fig. 5 because it is nearly identical to $F_A$. For the global ocean, the maximum from $F_C$ is smaller than $F_A$ by up to 4.6% in the Tropics, by up to 44.6% in the midlatitude Southern Hemisphere, and by up to 62.4% in the midlatitude Northern Hemisphere. The choice of averaging scheme has a marked effect on the magnitude of the heat transport results. However, the crucial element appears to be a good estimate of the monthly mean wind stress (from data with fine temporal resolution) — with such an estimate, there is little sensitivity in the monthly averaged heat transport calculation. This finding is supported by additional calculations (not shown) using daily AVHRR data and daily-averaged wind stresses. For the remainder of this study, only $F_A$ is discussed.
4. Analyses

a. Global meridional Ekman heat transport patterns

The annual cycle of monthly averaged meridional Ekman heat transport for the World Ocean is shown in Fig. 6. As mentioned previously, anomalous high SSTs were found at high southern latitudes during winter. These suspect data were not included in the calculation of the annual cycle so that, for example, the mean for July south of 40°S represents July 1989 data only.

Heat is transported poleward through the Ekman layer in the Tropics of the Atlantic and most of the Pacific in all months. The poleward heat transport is largest in the winter–spring hemisphere and weakest in the summer–fall hemisphere. Due to larger seasonal variations in the zonal wind stress in the Northern Hemisphere, the annual range is much higher in the Northern Hemisphere, about $4.3 \times 10^8$ W m$^{-1}$, than in the Southern Hemisphere, about $1.3 \times 10^8$ W m$^{-1}$. In January, the maximum heat transport is higher in the tropical North Atlantic ($5.9 \times 10^8$ W m$^{-1}$) than in the tropical North Pacific ($4.0 \times 10^8$ W m$^{-1}$). In July, the maximum values in the Tropics are diminished to $1.5 \times 10^8$ W m$^{-1}$ in the North Atlantic and $2.4 \times 10^8$ W m$^{-1}$ in the North Pacific. Areas of weak equatorward heat transport in the eastern and western tropical Pacific and in the eastern Atlantic are evident in both July and October. This is due to the movement of the ITCZ away from this region and a resulting strengthening of the westerly zonal wind stress in these localized areas. In October, the maximum heat transport is higher in the tropical South Atlantic ($4.6 \times 10^8$ W m$^{-1}$) than in the tropical South Pacific ($3.5 \times 10^8$ W m$^{-1}$). Equatorward heat transport in the midlatitudes is generally less than $1 \times 10^8$ W m$^{-1}$; however, it is higher than this in January in two areas in the midlatitude Northern Hemisphere.

The annual cycle in the Indian Ocean has a different character from that in the Atlantic and Pacific Oceans. The Ekman heat transport is northward near the Equator during January and southward north of 30°S in July and October. A maximum of $7.4 \times 10^8$ W m$^{-1}$ is found off Somalia in July due to strong monsoonal forcing. The maximum poleward heat transport in the southern Tropics reaches $4.0 \times 10^8$ W m$^{-1}$ in October and diminishes to $0.5 \times 10^8$ W m$^{-1}$ in January.

These results agree qualitatively with the calculations of ARV using monthly mean winds from COADS. The present study shows larger poleward heat transport in the tropical Atlantic in January, April, and October, whereas the COADS study found the larger values in the tropical Pacific. The equatorward heat transport is stronger in the midlatitude Pacific and Atlantic in January (greater than $1 \times 10^8$ W m$^{-1}$) than in COADS (less than $1 \times 10^8$ W m$^{-1}$). In April, the area of poleward heat transport in both hemispheres extends about 10° farther poleward in the western Pacific than in the COADS study. The present study indicates a maximum poleward heat transport of $3.9 \times 10^8$ W m$^{-1}$ in the tropical Southern Indian Ocean in July compared with a maximum of less than $3 \times 10^8$ W m$^{-1}$ in the COADS study. The poleward heat transport east of Japan and in the South Pacific in July is not seen in the COADS study.
Fig. 6. Monthly averaged meridional Ekman heat transport $F_\text{e}$ for (a) January 1988–1991, (b) April 1988–1991, (c) July 1987–1990 and (d) October 1987–1990. The contour interval is $0.5 \times 10^8$ W m$^{-1}$; negative contours are dashed; the $2 \times 10^8$ W m$^{-1}$ contour is dotted.
The 4-yr Ekman heat transport variance (Fig. 7a) is highest in the Indian Ocean off Somalia due to the Indian monsoon. The highest variance in the tropical North Pacific is located 60° east of that in the COADS study, whereas the maximum variance in the South Pacific is located about 50° west of that in the COADS study. The 4-yr nonseasonal Ekman heat transport variance (calculated from weekly averaged transports, Fig. 7b) over the tropical Indian Ocean is about twice that found in the COADS study. The nonseasonal variance in the Pacific is less since the longer time series of the COADS study encompasses more ENSO cycles, including the intense 1982/83 event, than this study. In contrast, the Indian Ocean evidently is subject to much stronger high-frequency variability such as the Madden–Julian oscillation, as discussed below, explaining the higher nonseasonal variance there. The region of largest nonseasonal variability in the tropical North Pacific is found in the extreme west of the basin compared with COADS where the largest variability is at the date line.

b. Zonally integrated meridional Ekman heat transport patterns

The global Ekman heat transports have been zonally-integrated by basin and for the World Ocean (Fig. 8). No adjustments for suspect wintertime SST data (south of 40°S in all basins and north of 40°N in the Atlantic) have been made to the monthly time series.

In the Pacific Ocean, the heat transport is weak and equatorward at 10°N in September 1989 and 1990, whereas in the other years, it is weak and poleward. Weak westerly wind stress at 10°N in the eastern and western Pacific in 1989 and 1990 is sufficient to change the sign; in September 1987 and 1988, these are areas, primarily, of weak easterlies. There is a weakening of the poleward heat transport in March 1989 in the Trop-
ics of the Northern Hemisphere. In March 1989 the Atlas SSM/I winds had data gaps that were filled with ECMWF wind analyses. Since the ECMWF winds tend to be weaker than the Atlas SSM/I winds (Rienecker et al. 1994), the anomalous weak poleward heat transport in March 1989 is probably an artifact of the data merger. (Similarly, data gaps in the December 1987 and January 1988 winds are likely contributing to the anomalous weak equatorward heat transport found in the Southern Hemisphere Indian Ocean in those months). There is anomalous high heat transport in the North Pacific midlatitudes during winter 1990/91. Equatorward heat transport in the Atlantic is largest in the Tropics in July 1988 and 1989 associated with strong westerlies off west Africa. In the Indian Ocean, the strongest equatorward heat transport is found in the Northern Hemisphere in summer 1990.

The annual cycle based on 4-year monthly averages is shown in Fig. 9. The Pacific and Atlantic basins have poleward heat transport in the Tropics for all months, except September in the Pacific basin and August in the Atlantic basin when there is weak equatorward heat transport of about 0.5 pW at 10°N. Using climatological data, Levitus (1987) indicates a minimum poleward heat transport at this time. The integrated heat transport in the Atlantic is much smaller than in the Pacific because of the width differential, with a maximum of 1.6 pW in the Atlantic compared to a maximum of 5.1 pW in the Pacific. In general, for all basins and both hemispheres, the maxima presented in this study are less than those of Levitus (1987), usually by about 0.5 pW, but up to 2 pW (or nearly 40%) in the tropical North Pacific. This discrepancy is consistent with the Atlas SSM/I wind product being weaker in the Tropics than the climatological winds used by Levitus (Rienecker et al. 1994). As was found by Levitus, the signal from the Indian Ocean dominates the World Oceans zonally integrated heat transport during boreal summer.

c. Nonseasonal variability

EOFs are calculated from the time series of nonseasonal anomalies formed by subtracting the annual cycle of monthly mean heat transport from the weekly time series. Rotation of the EOFs according to the varimax technique (e.g., Harmon 1976) localizes the EOFs into a small number of components, yielding modes that have simpler structures than the EOFs. As discussed by Richman (1986), such rotations potentially provide more physically realistic spatial relationships and reduce the dependence of the EOF analysis on the domain shape. They also suffer less from subdomain in-
stability and sampling problems than conventional EOFs. The varimax rotation retains the orthogonality of the modes.

The first two EOFs especially highlight the covariability between the Indian and tropical Pacific Oceans (Fig. 10). In EOF1, which explains 6.53% of the nonseasonal variance, enhanced poleward heat transport in the eastern tropical North Pacific and western tropical South Pacific is coherent with equatorward heat transport in the tropical southern Indian Ocean. In EOF2, which explains 5.75% of the total nonseasonal variance, the variability in the western tropical Pacific is coherent with variability in the Northern Hemisphere Indian Ocean. The patterns of these first two EOFs in the Indian Ocean are each very close to the first two unrotated EOFs determined for the Indian Ocean alone. The first Indian Ocean EOF (not shown) explains 21.25% of the total nonseasonal variance and up to 68% of the local nonseasonal variance in the southern tropical Indian Ocean. The second EOF explains 17.02% of the total nonseasonal variance and up to 73% of the local nonseasonal variance in the northern Indian Ocean. The variability in the Atlantic in EOF1 is similar to the first Atlantic basin unrotated EOF (although the latter has a larger signal in the tropical South Atlantic), the latter explaining 15.65% of the total nonseasonal variance in the Atlantic and up to 50% of the local variance in both hemispheres of the tropical Atlantic. The variability in the Pacific in EOF2 is close to the second unrotated mode determined from the Pacific basin alone. The latter explains 8.20% of the total nonseasonal variance in the Pacific, up to 20% of the local nonseasonal variance in the western Pacific. The global EOF3 and EOF4 are North and South Pacific basin modes, explaining 5.31% and 4.86%, respectively, of the total nonseasonal variance. EOF3 is essentially the same as the first unrotated EOF determined for the Pacific basin alone. The latter explains 9.10% of the total Pacific nonseasonal variance and up to 47% of the local nonseasonal variance in the tropical North Pacific. The global EOF shows coherence across the entire tropical to subtropical North Pacific and covariability with the central tropical South Pacific. These signals are out of phase with variations in the tropical Indian Ocean. EOF4 is close to the fourth Pacific basin EOF, which explains 5.81% of the total Pacific nonseasonal variance and up to 33% of the local nonseasonal variance in the western tropical South Pacific. In the global EOF, variations in the tropical to subtropical South Pacific are out of phase with variations in the eastern tropical North Pacific and the Atlantic.

The time series of the principal components show large variations on many timescales, including inter-
annual variations that are not well resolved in a spectrum analysis (see Fig. 12). The ensemble spectra, produced by averaging four individual spectra calculated from yearly time series, do not show much energy in the short timescales (Fig. 11); however, they show several peaks at periods from 30 to 60 days, indicative of the influence of the atmospheric Madden–Julian oscillation (MJO, e.g., Lau et al. 1994) on the Ekman heat transport. One marked feature is the strong peak at 50 to 60 days in the spectrum from the principal component of REOF1. REOF2 has more energy than REOF1 at periods longer than 100 days.

The multiple timescales in the time series have been investigated through singular spectrum analysis (SSA), which is used to discriminate dominant frequencies in the time series of the EOF amplitudes. SSA finds the eigenfunctions (which can be shown to be either symmetric or antisymmetric) of the covariance matrix calculated from the time series of an EOF amplitude. These eigenfunctions form an orthonormal ba-
The dominant mode of variation for REOF1 (Fig. 12a) is at a period of 40 to 50 days (RC-1) and shows up as one of a pair of near-degenerate eigenfunctions, 90° out of phase. The oscillation has a distinct seasonal modulation of its amplitude with larger amplitudes in the boreal spring. The amplitude is reduced during the 1988–89 La Niña. The longer timescales in REOF1 do not show up until the third eigenfunction. The signal indicates enhanced poleward heat transport in the Indian Ocean and equatorward heat transport in the eastern tropical Pacific Ocean during the first half of the 1988–89 La Niña, followed by the reverse of this during the second half of this event (e.g., Delcroix 1993).

The dominant signal (RC-1) in the time series of REOF2 amplitudes has positive peaks in December 1987, July 1989, and July 1990 (Fig. 12b), contributing to enhanced southward heat transport in the Northern Hemisphere Indian Ocean and western tropical North Pacific and poleward heat transport in the central tropical South Pacific. REOF3 is dominated by an anomaly during boreal summer 1988 (Fig. 12c), associated with enhanced poleward heat transport across the entire tropical North Pacific. With the enhanced equatorward heat transport in the midlatitude Pacific, the anomalies in the equatorial Pacific would lead to enhanced convergence of heat transport in the subtropical Pacific at that time. REOF4 is largely influenced by variations with periods of 40 to 50 days. This signal has reduced amplitude during 1988 and a modulation of about 1.5 years (Fig. 12d). The long timescale signal (RC-3) shows a negative anomaly in early 1989, indicating enhanced equatorward heat transport in the tropical South Pacific and the western equatorial North Pacific, with enhanced poleward transport in the eastern tropical North Pacific.

The influence of the 1988 event on the Ekman heat transport, as summarized in the longer timescale signals in the first four REOFs, is shown as the average of these signals over June–August 1988 in Fig. 13. (This does not include the modulation of the higher-frequency signals during this event.) The major influence is the enhanced (by up to $1 \times 10^8$ W m$^{-1}$) poleward heat transport in the tropical North Pacific and eastern Indian Ocean. The enhanced poleward heat transport in areas of the tropical and subtropical South Pacific during this period is evident in the greater poleward extent of high values in the time series of zonally integrated heat transport in Fig. 8. The reduced poleward (enhanced equatorward) heat transport in the subtropical North Pacific is also evident near 25°N in Fig. 8. The zonally integrated anomaly in the Tropics is sufficient to change the equatorward heat transport seen near 10°N in the other summers to a weak poleward heat transport in boreal summer 1988. In the Indian Ocean, the equatorward heat transport in the northern basin is enhanced in the western Tropics but reduced in the northeast; the poleward heat transport in the southern basin is enhanced in the eastern Tropics.

5. Summary

Weekly global meridional Ekman heat transports on a 2° grid have been estimated from satellite-derived data products, namely, AVHRR SSTs and Atlas SSM/I wind analyses. For comparison with previous estimates that used in situ observations and monthly averaged or climatological winds (e.g.,
Fig. 12. Time series of the REOF amplitudes and chosen reconstructed components in their SSA decomposition (in units of $10^6$ W m$^{-1}$) with the variance explained by each eigenfunction. (a) REOF1, (b) REOF2, (c) REOF3, and (d) REOF4.

Fig. 13. The average of the nonseasonal anomaly for June–August 1988 calculated from the long timescale SSA reconstructed components of REOFs 1–4. The contour interval is $0.2 \times 10^6$ W m$^{-1}$; zero and negative contours are dashed.
ARV; Levitus 1987), monthly averaged heat transport was calculated. Zonally integrated, monthly averaged heat transport estimated from monthly averaged wind components and monthly averaged SST were consistently lower than the monthly averaged estimates from weekly averaged wind stress and SST—by only up to 5% in the Tropics but by up to 62% in the midlatitudes.

Global maps show that the meridional Ekman heat transport is poleward over the tropical Atlantic and over most of the tropical Pacific throughout the year, with maxima in the winter–spring hemisphere and minima in the summer–fall hemisphere. In January, April, and October, poleward heat transport is higher in the tropical Atlantic than in the tropical Pacific, in contrast with results based on 30 years of COADS data presented by ARV. In January, the equatorward heat transport in the Northern Hemisphere midlatitudes is larger than $1 \times 10^8 \text{ W m}^{-1}$.

The zonally integrated, monthly averaged Ekman heat transport was weak and equatorward at $10^\circ\text{N}$ in the Pacific in September 1989 and 1990. Weak westerly wind stress at $10^\circ\text{N}$ in the eastern and western basin was sufficient to change the sign of the weak integrated heat transport. At all other times, the zonally integrated heat transport is poleward at this latitude. In the Indian Ocean, equatorward heat transport is strongest in the Northern Hemisphere in summer 1990.

The weekly time series provides better temporal resolution than previous studies. The higher-frequency variations are explored through rotated empirical orthogonal functions (REOFs) of nonseasonal heat transport anomalies. The REOFs show large-scale coherence across the tropical and subtropical Pacific and Indian Oceans and are close to the unrotated EOFs determined from each basin separately. The first REOF has a strong spectral peak at periods of 50 to 60 days and is dominated by the variability in the Southern Hemisphere Indian Ocean. The second REOF is dominated by the variance in the Northern Hemisphere Indian Ocean and western tropical North Pacific. The first four REOFs, which explain 22.5% of the nonseasonal variance, have spectral peaks at periods consistent with the 30–60 day atmospheric Madden–Julian oscillation. This signal was not resolved in the monthly averaged data from COADS. Singular spectrum analysis has been used to show the amplitude modulations of these signals. The early phase of the 1988–1989 La Niña has a strong influence on the heat transport, indicating enhanced poleward heat transport in the eastern Indian Ocean, tropical North Pacific, and tropical and subtropical South Pacific and enhanced equatorward heat transport in the midlatitude North Pacific and western tropical Indian Ocean.

The fine temporal and spatial resolution of satellite-derived data has provided interesting additional information on the global meridional Ekman heat transport variability that is not available in conventional in situ observations. This study, and preliminary calculations using the finer temporal resolution and increased coverage that will be available from the NASA/NOAA AVHRR Pathfinder data, indicates that good estimates of the monthly averaged wind stress are essential for good estimates of global heat fluxes and the transports necessary to maintain the earth’s thermal equilibrium.

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REFERENCES


