Estimating Meridional Energy Transports by the Atmospheric and Oceanic General Circulations Using Boundary Fluxes

Y.-C. ZHANG

Department of Applied Physics, Columbia University, New York, New York

W. B. ROSSOW

NASA/Goddard Institute for Space Studies, New York, New York

(Manuscript received 1 July 1996, in final form 30 January 1997)

ABSTRACT

The annual-mean meridional energy transport in the atmosphere–ocean system (total transport) is estimated using 4-yr mean net radiative fluxes at the top of the atmosphere (TOA) calculated from the International Satellite Cloud Climatology Project cloud datasets. In addition, the net atmospheric and surface radiative fluxes are calculated. When supplemented by a climatology of the surface latent and sensible heat fluxes, these radiative fluxes are used to derive the separate atmospheric and oceanic energy transports using a surface and planetary energy-balance method. Most previous results are based on direct calculations of the atmospheric energy transport from in situ measurements of horizontal wind velocity, temperature, and humidity in the atmosphere and on inference of oceanic heat transports as the difference between the atmospheric transports and the total energy transport (the planetary energy-balance method). Total, atmospheric, and oceanic energy transports from this study are in good agreement with more recent results (within mutual uncertainties). A detailed assessment is made of the uncertainties in the atmospheric and ocean energy transports that arise from uncertainties in the TOA and surface energy fluxes: the largest uncertainties are associated with the surface radiative and latent heat fluxes. Since the errors in the present method are from different sources and have different geographic distributions, the results of this study complement previous estimates of the atmospheric and oceanic energy transports. Assessment of error sources also suggests that improvement of this type of result is more likely in the near future than for the other methods. Because the radiative fluxes are calculated from physical quantities, the authors can characterize the mean effects of clouds on the atmospheric and oceanic energy transports: 1) cloud effects on the TOA radiation budget reduce hemispheric differences introduced by hemispheric differences of surface properties, 2) the cloud effects on the atmospheric and surface radiation budgets induce hemispheric differences in the heating/cooling of the atmosphere and ocean that require cross-equatorial transports in opposite directions by the atmosphere and ocean, and 3) all other factors held constant, clouds tend to reduce oceanic energy transports and increase atmospheric energy transports.

1. Introduction

Energy transports by the general circulations of the earth's atmosphere and oceans are the fundamental components of the climate system and hence have drawn the interest of many authors since Hadley's time (e.g., Lorenz 1967). The most common method for estimating the energy transports is to calculate directly the atmospheric energy transport based on in situ measurements of the three-dimensional distributions and time variations of atmospheric temperature, humidity, and wind velocity and to derive the total annual mean transport by the whole atmosphere–ocean system from the net radiative flux at the top of atmosphere (TOA). Then the oceanic energy transport is estimated from the difference between total transport and the atmospheric transport (e.g., Oort and Vonder Haar 1976; Peixoto and Oort 1992; Trenberth and Solomon 1994; see other references in section 5a). This method is called the planetary energy-balance method by Carissimo et al. (1985). The reason to use such a hybrid method that combines direct calculations (for the atmospheric storage and transports) and indirect inference (for the surface energy flux or the oceanic storage and transports) is that there are no global and synoptic measurements of current velocity, temperature, and salinity in the oceans. Although oceanic energy transports at particular latitudes in some ocean basins have been estimated from measurements (e.g., Bryden 1991; MacDonald 1993), there is generally not enough data to calculate the mean energy transports for the global ocean. Even for the atmosphere, there are still many difficulties and significant uncertainties in avail-

Corresponding author address: Dr. Yuanchong Zhang, Department of Applied Physics, Columbia University, 2880 Broadway, New York, NY 10025.
E-mail: clyxz@giss.nasa.gov

able measurements, especially poor coverage of the Southern Hemisphere (Trenberth and Solomon 1994; Keith 1995).

For estimating multiyear annual-mean energy transports, there is an alternative method, which we call the surface and planetary energy balance (SPEB) method, that employs only the energy fluxes at the TOA and surface boundaries of the atmosphere to infer the balancing transports within the atmosphere and ocean, rather than direct measurement and calculation of the dynamic transports [e.g., Hastenrath (1982) and Hsuing (1985) used this boundary flux method to estimate the ocean heat transport]. Recently, we developed a very detailed method for calculating radiative fluxes at TOA, the surface, and in the atmosphere (as the difference of the first two) from the International Satellite Cloud Climatology Project (ISCCP) cloud datasets using the NASA/Goddard Institute for Space Studies (GISS) radiative transfer model (Zhang et al. 1995; Rossow and Zhang 1995). We have produced a global radiative flux dataset (called the ISCCP–FC dataset) covering 1985–88 with which it is possible to estimate the energy transports in the atmosphere and oceans using the SPEB method if the radiative fluxes are combined with estimates of the latent and sensible heat fluxes at the surface.

This paper presents our estimates of the mean meridional energy transports by the atmosphere and ocean in comparison with previous results to highlight three points. First, the errors in the estimated transports from our alternative method, though currently comparable in magnitude with those of the other approaches, have a very different character with regard to their geographic and time distribution so that this alternative provides a complementary constraint on the transport uncertainties. The current uncertainties in the radiative, latent, and sensible heat fluxes are discussed in some detail. Of particular note is that the largest disagreements with previous results occur in the Southern Hemisphere, where the satellite-based results have much better sampling than the conventional observations. Because of the even greater geographic inhomogeneity of the quality of the conventional datasets, we focus on zonal mean fluxes and the mean meridional energy transports at this stage of our analysis. Second, looking at the quantities that are used in all methods and their current observational uncertainties identifies changes in our current observing system that are needed to improve our understanding of the energy cycle of the climate system. Finally, we exploit the unique properties of our results to characterize, for the first time, the mean cloud-radiative effects on the atmospheric and oceanic energy transports.

2. Method

As required by energy conservation, the equation for the rate of change of all forms of energy for the atmosphere–earth system is

$$\frac{\partial E}{\partial t} = S_a + S_o + S_l + S_{is} = F_t - \nabla \cdot T_a - \nabla \cdot T_o, \quad (1)$$

where E is total energy of the atmosphere–earth system; S_a, S_o, S_l, and S_{is} are rates of energy storage in the atmosphere, oceans, land and ice/snow, respectively; F_t is net energy flux at TOA (i.e., net radiative flux since there is no other energy flux at TOA); and T_a and T_o are energy transports in the atmosphere and oceans, respectively. Positive and negative divergences of T_a and T_o indicate the energy source and sink regions, respectively, that maintain the general circulation of the atmosphere–ocean system. Separation of Eq. (1) gives the following two equations for the atmospheric and oceanic circulations, respectively:

$$S_a = F_a - \nabla \cdot T_a \quad (2)$$

and

$$S_o + S_l + S_{is} = F_s - \nabla \cdot T_o, \quad (3)$$

where F_a and F_s are net energy fluxes into the atmosphere and into the surface of the earth, respectively, and $F_t = F_a + F_s$. If we average over vertical columns and latitude zones, then Eqs. (1), (2), and (3) are reduced to one-dimensional equations for meridional energy transport. For multiyear averages, all the rates of storage of energy can be neglected; that is, we assume that their interannual variations average out over many years as required for a long-term thermal balance of the climate system. In fact, these storage terms might differ from zero by a few W m^{-2} in any given year or decade. After integration, Eqs. (1), (2), and (3) reduce to

$$T_a(\theta) + T_o(\theta) = \int_{-\pi/2}^{\theta} 2\pi R^2 \cos\theta F_t \, d\theta \quad (4)$$

$$T_a(\theta) = \int_{-\pi/2}^{\theta} 2\pi R^2 \cos\theta F_a \, d\theta \quad (5)$$

$$T_o(\theta) = \int_{-\pi/2}^{\theta} 2\pi R^2 \cos\theta F_s \, d\theta, \quad (6)$$

where R is the radius of the earth and θ is the latitude. Here, $T_a(\theta)$ and $T_o(\theta)$ are now meridional energy transports northward across a latitude θ in the atmosphere and oceans, respectively, and the values of F_t, F_a, and F_s are now understood to be the zonally averaged fluxes. If F_t (TOA net radiative flux) is known, Eq. (4) gives the energy transport by the combined atmosphere–ocean system.

The net energy flux at the surface is the net radiative flux at the surface (F_{sr}) less the latent heat flux (evaporation, LH) and sensible heat flux (SH):

$$F_s = F_{sr} - \mathrm{LH} - \mathrm{SH}, \quad (7)$$

where minus sign means that the direction of *LH* and *SH* is from the surface of the earth into the atmosphere.

Correspondingly, the net energy flux into the atmosphere is

$$F_a = F_{ar} + \mathrm{LH} + \mathrm{SH}, \tag{8}$$

where F_{ar} is the net radiative flux into the atmosphere. Therefore, the net radiative fluxes at the surface and into the atmosphere, together with surface latent and sensible heat fluxes, give the zonally averaged net energy fluxes F_a and F_s, which can be used in Eqs. (5) and (6) to evaluate the energy transports in the atmosphere and oceans, respectively. Alternatively, we can evaluate either Eq. (5) or (6) and obtain the other transport by calculating the difference with the total transport from Eq. (4). These methods are totally indirect since only the surface and the planetary (TOA) boundary energy fluxes are involved. These boundary energy fluxes must balance the energy transports in the long-term annual mean. In the planetary energy-balance method, as defined by Carissimo et al. (1985), only the net radiative flux at TOA is satellite derived and the surface energy flux is inferred as the difference between it and the (directly calculated) divergence of the atmospheric energy transport and its rate of energy storage (if any).

In Eq. (6), the transport is determined from the zonal mean of F_s, which includes both land and ocean surface fluxes. If we neglect the horizontal energy transport by water runoff from land to ocean and the storage of energy in land ice, then the net energy flux into the land surface should be zero for a multiyear annual mean, so the land areas do not contribute to the meridional energy transport. Thus, Eq. (6) gives the ocean transport, even though the zonal mean value of F_s is used. If only the net energy flux at the ocean surface, F_o, is known, we still can derive T_o from a revised version of (6):

$$T_o(\theta) = \int_{-\pi/2}^{\theta} 2\pi R^2 \left(\frac{Z_0}{Z_g}\right) \cos\theta F_o \, d\theta, \tag{9}$$

where (Z_o/Z_g) is the fraction of each zone covered by ocean and F_o is the longitudinal average of the net surface energy flux over the ocean-covered portion of latitude θ. The "global" adjustment for balance then applies only to the ocean area. Equation (9) is used to derive the oceanic poleward energy transport for ocean-only surface flux datasets. In such calculations, our surface radiative flux data is also limited to ocean areas.

3. Data

The ISCCP–FC radiative flux dataset (henceforth FC data) is calculated from the ISCCP C1 and C2 datasets (Rossow and Schiffer 1991; Rossow et al. 1991), augmented by other information about the surface and cloud layering (see Zhang et al. 1995 for details), and includes all the radiative flux components: upward and downward; total, cloudy, and clear; and shortwave (SW) and longwave (LW) fluxes at TOA, the surface, and into the atmosphere. This version of FC data covers the whole globe for 4 yr (every third month from April 1985 to January 1989) with a temporal resolution of 3 h and a spatial resolution of about 280 km. Our validation studies (Rossow and Zhang 1995) show that the overall uncertainties of the monthly mean SW and LW fluxes at TOA and the surface for 280-km regions are about 5–15 and 10–25 W m^{-2}, respectively, by comparison with Earth Radiation Budget Experiment (ERBE) results at TOA and with various surface radiation measurements. The 4-yr zonal averages of the net radiative fluxes at TOA, the surface, and into the atmosphere are used in this study; these multiyear and zonal averages likely have somewhat smaller uncertainties. Basing our annual mean fluxes on the average of monthly means from every third month introduces little error: comparing the annual mean TOA fluxes obtained from ERBE using every third month of the year and all 12 months shows differences <1 W m^{-2}. The error at the surface may be slightly larger because of the time lag of the ocean surface temperatures, but the contribution of the surface net LW flux to the total surface energy flux is relatively small.

We also examine radiative, latent, and sensible heat fluxes (all based on 12-month annual averages) from several other sources, some summarized in Peixoto and Oort (1992). Their TOA radiative fluxes are obtained from Campbell and Vonder Haar (1980). Another estimate of TOA radiative fluxes is obtained from ERBE (e.g., Harrison et al. 1990). Climatological zonal means (land and ocean) of evaporation rate, precipitation rate, and sensible heat flux are taken from Sellers (1965), some of whose data are compiled from Budyko (1963). Another estimate of surface radiative, sensible, and latent heat fluxes over oceans is obtained from the Comprehensive Ocean–Atmosphere Data Set (COADS) compilation (da Silva et al. 1994). Our main results come from the combination of our FC radiative fluxes with the Sellers (1965) climatology of surface latent and sensible heat fluxes (henceforth Sellers data), but we consider some other combinations in section 4.

4. Results

a. Energy transport by the whole atmosphere–ocean system

Figure 1 shows the 4-yr annual and zonal average of the total net radiative flux at TOA from the FC and ERBE results. Both results have been adjusted by a globally uniform amount to obtain a global mean net flux of zero so that the global integral of the horizontal energy transports is zero as required by total energy balance (cf. Carissimo et al. 1985): for the FC data the adjustment is $+4$ W m^{-2} and for the ERBE results the adjustment is -5 W m^{-2}. Such adjustments[1] are com-

[1] The adjustments have the opposite sign of the global mean values.

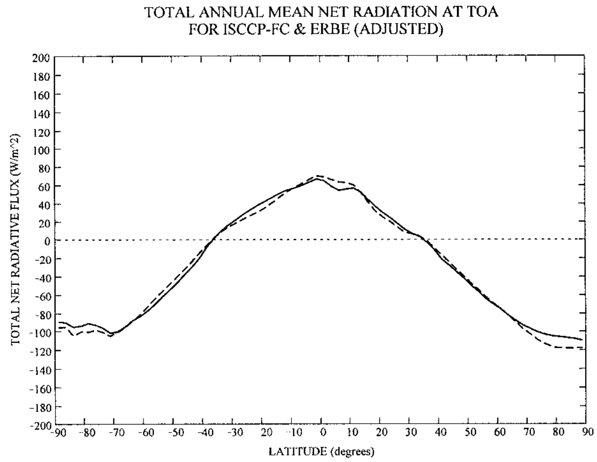

FIG. 1. Zonal annual mean net radiative fluxes at top of atmosphere (W m^{-2}) from our FC data adjusted by $+4$ W m^{-2} (solid line) and from ERBE adjusted by -5.1 W m^{-2} (dashed line). Both datasets are averages of every third month from April 1985 through January 1989.

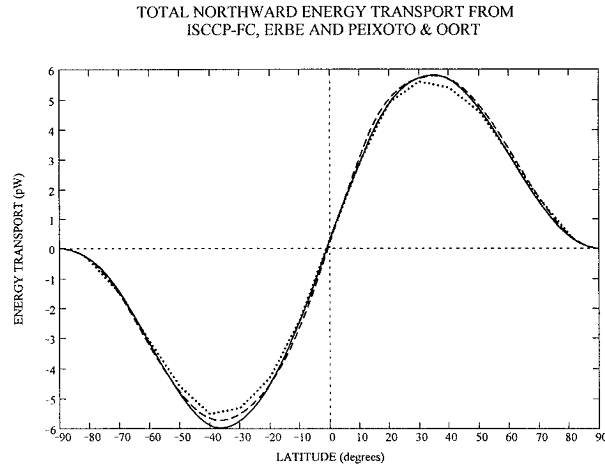

FIG. 2. Mean northward total energy transport by the atmosphere–ocean system (PW = 10^{15} W) inferred from the zonal annual mean net TOA radiative fluxes using Eq. (4). Results are shown from our FC (solid line), ERBE (dashed line), and Peixoto and Oort (1992) (dotted line).

monly required because of errors in the radiative flux estimates and/or true interannual variations in the energy balance (e.g., Peixoto and Oort 1992, PO92 in the following). Note that after adjustment, the FC and ERBE TOA net flux variations with latitude are nearly identical: rms differences of zonal mean net fluxes are 5.4 W m^{-2}. As expected, the zones equatorward of about 35° in the Northern and Southern Hemispheres (tropical and subtropical zones) are net source regions of radiative energy, balanced by positive divergences of (T_a + T_o), and poleward latitudes are net sink regions of radiative energy, balanced by negative divergences of (T_a + T_o). This latitudinal variation of the net radiation is sometimes said to be the forcing for the general circulations of the atmosphere and ocean; but actually the net radiation, particularly the LW component, is also part of the climate system response. The forcing for the general circulations is the net SW heating that would exist without any feedbacks, such as by clouds, which cannot be determined directly from observations.

By evaluating Eq. (4), we obtain the total meridional (northward) energy transport by the atmosphere–ocean system (Fig. 2). Figure 2 also shows for comparison the result obtained using the ERBE net fluxes and the result reported by PO92. The PO92 result is based on 48 consecutive months of TOA net radiation from satellite measurements made in the 1960s to 1970s (Campbell and Vonder Haar 1980) with uncertainties estimated to range from 2.6 to 15.6 W m^{-2} with an average of 8 W m^{-2} (Carissimo et al. 1985), and a global adjustment of about -9 W m^{-2} required (PO92). The ERBE zonal annual mean net flux uncertainties are estimated to be about 5–10 W m^{-2} (see discussion and references in Trenberth and Solomon 1994 and Rossow and Zhang 1995), whereas our zonal mean net flux uncertainties are estimated to be less than about 10 W m^{-2} (Rossow and Zhang 1995). Figure 2 shows excellent quantitative agreement among the three estimates of the total transport, despite significant differences in data sources and time periods covered (Table 1 shows some other results). The largest differences, ≈0.5 petaWatt (PW = 10^{15} Watts), occur near the peak transports at 30°–40° in both hemispheres. The peak transports in both the ERBE[2] and FC results are larger than PO92's values, but by an amount within PO92's estimated uncertainty of 1 PW (Carissimo et al. 1985). The meridional transports obtained from our calculated radiative fluxes differ from those of ERBE by only 0.13 PW rms.

b. Energy transports in the atmosphere and oceans

Using Eqs. (5) and (8), we obtain the annual mean atmospheric energy transport and the oceanic energy transport as the difference between the total transport (i.e., for the whole atmosphere–ocean system) and the atmospheric transport. Alternatively, we could use Eqs. (6) and (7) to obtain the oceanic energy transport and

[2] The ERBE results cannot be obtained directly from the ERBE datasets because there are significant areas of missing data at high latitudes in any given month. Our version of the ERBE net fluxes is formed by using the astronomical solar elevation to determine all zero values of the net SW flux, averaging each month over 4 yr (1985–88) in a 2.5° resolution map (a result is reported if at least one value is available); averaging over January, April, July, and October to obtain an annual mean map (all four values must be available); and averaging over latitude, and then interpolating to fill missing zones (only one was missing). The implied total energy transports shown in Fig. 2 are within 0.2 PW of the values derived by Trenberth and Solomon (1994) from 1 yr (1988) of ERBE data. On the other hand, Keith (1995) shows total energy transports derived from a 3-yr average of ERBE data that are nearly identical in the Northern Hemisphere but are about 1 pW lower in the Southern Hemisphere.

TABLE 1. Comparison of reported values of the annual mean northward energy transport (in petaWatts = 10^{-15} = PW) by the atmosphere–ocean system at the indicated latitudes (see text for references). Values in parentheses were adopted from another work without an independent assessment. Values from S88, M88, MD91, TS94, K95, H82, and H85 are estimated from published figures and are uncertain by ±0.1 PW.

	65°N	45°N	24°N	0°	30°S	45°S	65°S
PO92	2.4	5.0	5.3	0.3	−5.3	−5.1	−2.8
S88	(2.3)	(5.2)	(5.3)	(0)	(−5.6)	(−5.6)	(−2.0)
M88	2.0	5.5	5.5	0.3	−5.6	−5.6	−2.3
MD91	2.1	5.0	5.5	0.5	−5.7	−5.3	−2.4
TS94	2.4	5.1	5.5	0.7	−5.5	−5.6	−2.8
K95	2.4	5.3	5.4	0.5	−5.5	−4.6	−2.2
H82	2.3	4.8	5.3	−0.1	−5.4	−5.1	−2.0
H85							
B91							
M93							
daS94							
COADS							
ERBE	2.5	5.4	5.4	0.2	−5.5	−5.3	−2.4
FC + Sellers	2.4	5.3	5.3	0.3	−5.7	−5.5	−2.3

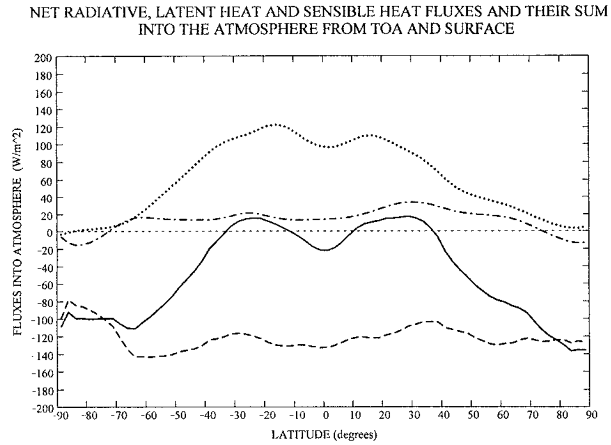

FIG. 3. Zonal annual mean values (W m^{-2}) of 1) net radiative cooling of the atmosphere (dashed line) from FC data, 2) latent heat flux (evaporation) at the surface into the atmosphere (dotted line) from Sellers (1965), 3) sensible heat flux at the surface into the atmosphere (dash–dotted line) from Sellers (1965), and 4) net energy flux into the atmosphere (solid line), which is the sum of 1–3.

subtract that from the total transport to get the atmospheric transport. The results are mathematically identical,[3] so we show only results from the first method. Figure 3 shows the annual zonal mean fluxes of surface net radiation (from FC), the surface latent and sensible heat fluxes (from Sellers), and their sum—that is, the zonal mean net energy flux, F_a—*into the atmosphere.* Here, F_a is reproduced in Fig. 4 with a globally uniform adjustment of +26 W m^{-2} (i.e., our total surface flux underestimates the heating of the atmosphere and overestimates the heating of the surface) to eliminate spurious net transport at the north pole caused by the average imbalance. From Fig. 4, we see that the energy sources for the atmospheric circulation are in the subtropical regions, showing a bimodal distribution about the equator, and that the energy sinks are at higher latitudes. Integration of Eq. (5) gives the atmospheric energy transport (Fig. 5), where PO92's result (based on Oort and Peixoto 1983, OP83 in following) is also shown. The difference between total and atmospheric transports gives the oceanic transport (Fig. 6). The alternate procedure gives the same transports (not shown here), but the global adjustment of the total energy flux *into the surface is* −22 W m^{-2} (the difference in the required adjustments is just the global imbalance of the net radiative fluxes at TOA).

Figures 5 and 6 compare our atmospheric and oceanic energy transports with those reported by PO92. They can also be compared with more recent results (Tables 1, 2, and 3) obtained by analyzing atmospheric observations as in PO92 (Savijärvi 1988, S88 in the following) or by using the results of a global data assimilation model to estimate the atmospheric transports (Masuda 1988; Michaud and Derome 1991; Trenberth and Solomon 1994; Keith 1995; henceforth M88, MD91, TS94, and K95, respectively). Our atmospheric transports are generally larger than PO92's, by more than 1 PW at 40°N and by more than 2 PW at 40°S, and show more cross-equatorial transport. Both results show that the oceanic transport dominates at lower latitudes, whereas the atmospheric transport is more important at middle and high latitudes.

PO92's atmospheric energy transports are based on a 10-yr (May 1963 to April 1973) upper-air dataset from the global rawinsonde network and surface ship reports (OP83) with uncertainties estimated to be 0.5 PW (Carissimo et al. 1985). S88 also analyzed this dataset, but

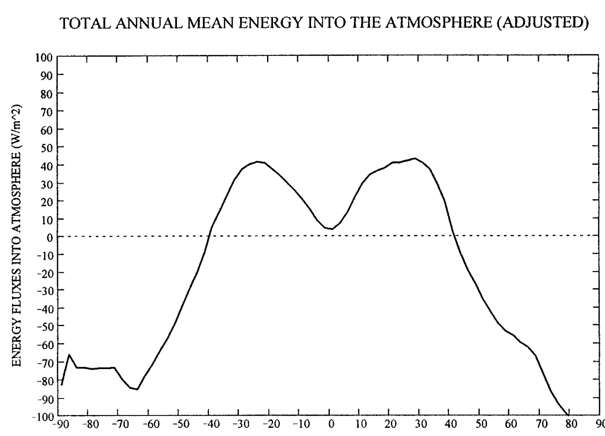

FIG. 4. Zonal annual mean net energy flux into the atmosphere (W m^{-2}) redrawn from Fig. 3 after being adjusted by +26 W m^{-2} so that the global mean value is zero.

[3] We calculate the total transports required to balance zonal mean net fluxes, which neglects small differences in the average fluxes over land and ocean.

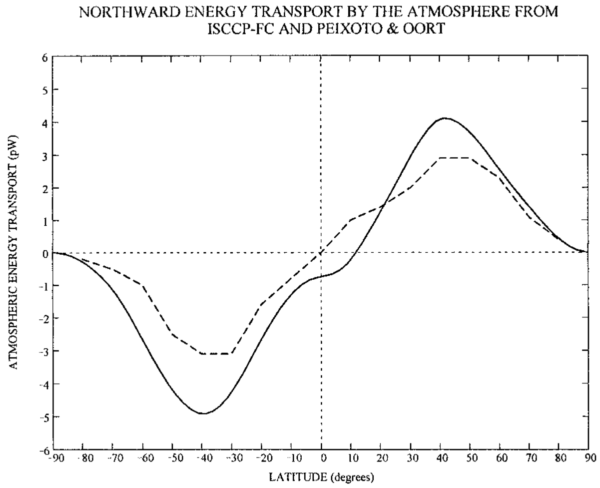

FIG. 5. Zonal annual mean northward total energy transport by the atmosphere (PW) determined from Eq. (5) in this work (FC + Sellers data, solid line) and reported by Peixoto and Oort (1992) (dashed line).

TABLE 2. Comparison of reported values of the annual mean northward energy transport (PW) by the atmosphere at the indicated latitudes (see text for references). Values in parentheses were adopted from another work without an independent assessment. Values from S88, M88, MD91, TS94, K95, H82, and H85 are estimated from published figures and are uncertain by ±0.1 PW.

	65°N	45°N	24°N	0°	30°S	45°S	65°S
PO92	1.7	2.9	1.6	0	−3.1	−2.9	−0.8
S88	2.0	4.2	2.7	0.6	−3.5	−4.5	−1.0
M88	1.9	4.0	2.9	−0.8	−4.0	−3.5	−1.3
MD91	1.8	3.7	2.5	−0.3	−3.8	−3.7	−1.7
TS94	1.9	3.8	3.0	−0.1	−3.8	−3.8	−1.9
K95	2.1	4.2	2.9	0	−4.3	−3.4	−1.7
H82	2.3	3.9	3.0	0	−3.8	−4.0	—
H85							
B91							
M93							
daS94							
COADS							
ERBE							
FC + Sellers	2.0	4.0	1.9	−0.7	−4.2	−4.7	−1.8
Fcm + Sellers	2.2	4.6	2.6	−0.6	−4.8	−5.2	−2.1

added a dynamical constraint on the divergent wind component, and obtained results more like ours with peak midlatitude atmospheric transports of about 4.0–4.5 PW. Notably, the reanalysis with the dynamic constraint also increased the cross-equatorial transport relative to PO92's result but produced the opposite sign from our result (Table 2). Use of data assimilation models applied to the FGGE dataset (M88), applied to 6 yr (1981–86) of European Centre for Medium-Range Weather Forecasts (ECMWF) results (MD91) or applied to 1 yr (1988) of ECMWF results (TS94) to estimate the atmospheric energy transports, all produce results very similar to those of S88 and ours (Table 2), showing peak atmospheric transports, near 4 PW at ±40° latitude—differences are generally ≈0.5 PW in the Northern Hemisphere but are larger, almost 1 PW, in the Southern Hemisphere where the atmospheric data sources are very sparse. All of these results show southward transport at the equator similar to our result.

Our ocean energy transports are similar in magnitude to those of PO92 in the Northern Hemisphere, though the peak is shifted equatorward, but smaller by more than 1 PW than those of PO92 in the Southern Hemisphere. The uncertainties of PO92's oceanic energy transports are estimated to be 1.5 PW (Carissimo et al. 1985). The ocean energy transports estimated in the

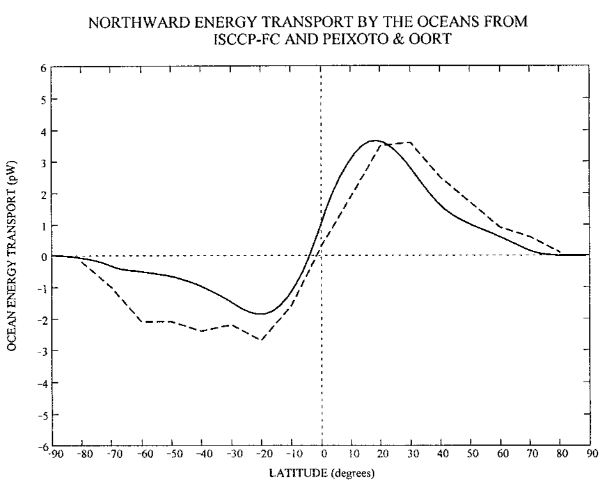

FIG. 6. Zonal annual mean northward energy transport by the oceans (PW) determined from the difference between the total and atmospheric transports (Figs. 2 and 5) from this work (FC + Sellers data, solid line) and from Peixoto and Oort (1992) (dashed line).

TABLE 3. Comparison of reported values of the annual mean northward energy transport (PW) by the ocean at the indicated latitudes (see text for references). Values in parentheses were adopted from another work without an independent assessment. Values from S88, M88, MD91, TS94, K95, H82, and H85 are estimated from published figures and are uncertain by ±0.1 PW. Results from COADS use flux values directly; results attributed to da Silva (1994) are based on "constrained" fluxes and differ from their published results by no more than ±0.1 PW.

	65°N	45°N	24°N	0°	30°S	45°S	65°S
PO92	0.7	2.1	3.7	0.3	−2.2	−2.2	−1.4
S88	0.3	1.0	2.6	−0.6	−1.8	−1.1	−1.0
M88	0.1	1.5	2.6	1.1	−1.6	−2.1	−1.0
MD91	0.3	1.3	3.0	0.8	−1.9	−1.6	−0.7
TS94	0.5	1.3	2.5	0.8	−1.7	−1.8	−0.9
K95	0.3	1.1	2.5	0.5	−1.2	−1.2	−0.5
H82	0	0.9	2.3	−0.1	−1.6	−1.1	—
H85	0.2	0.5	1.6	0.1	−1.6	—	—
B91	—	—	2.0	—	—	—	—
M93	—	—	—	—	−0.7	—	—
daS94	0.5	1.1	2.3	0.6	−0.3	0.4	0
COADS	0.9	1.7	2.7	0.7	−1.2	−0.5	−0.2
ERBE							
FC + Sellers	0.4	1.2	3.4	1.0	−1.5	−0.8	−0.5
Fcm + Sellers	0.1	0.7	2.8	0.9	−0.9	−0.3	−0.2
FC + COADS	0.9	1.6	3.1	1.5	−0.7	−0.3	−0.3
FCm + COADS	0.7	1.2	2.7	1.4	−0.3	0.1	−0.2

TABLE 4. Peak values, their latitudes, and the equatorial values for the northward oceanic and atmospheric energy transports.

Northward transport						
Oceanic (O) or atmospheric (A)	Derived from (data/authors)	Equatorial value	N. Hemispheric peak		S. Hemispheric peak	
			Value (PW)	Latitude	Value (PW)	Latitude
O	FC + Sellers	1.0	3.7	18.0°N	−1.9	20.0°S
O	Modified FC + Sellers	0.9	3.0	17.0°N	−1.5	18.5°S
O	FC-Oceanic + COADS	1.5	3.2	21.0°N	−0.8	23.0°S
O	Modified FC-Oceanic + COADS	1.4	2.8	19.5°N	−0.4	21.5°S
O	COADS	0.7	2.7	24.0°N	−1.4	20.5°S
O	Constrained COADS	0.9	2.3	21.0°N	−0.7	16.0°S
A	FC + Sellers	−0.7	4.1	42.0°N	−4.9	39.5°S
A	Modified FC + Sellers	−0.6	4.7	41.5°N	−5.5	39.5°S

other "atmospheric" studies (S88, M88, MD91, TS94) show somewhat larger disagreements in their oceanic energy transports than they did in their atmospheric energy transports, especially in the Southern Hemisphere (Table 3). The results of S88 show a peak Southern Hemisphere transport of about 3 PW, similar to PO92, but about 1 PW larger than all the other results (M88, MD91, TS94), including ours. At 30°S (Table 3) our result is 0.8 PW larger than found from direct ocean measurements by MacDonald (1993) but very similar to the transport inferred by Hastenrath (1982, henceforth H82) and Hsuing (1985, henceforth H85). The analysis of ocean heat transport by da Silva et al. (1994), based on the COADS data compilation and constrained by the direct estimates at particular latitudes (and therefore, not entirely independent of the direct transport estimates), obtains a Southern Hemisphere peak transport of about 0.7 PW, more than 1 PW smaller than our value (Table 4). The results of TS94 show a peak Northern Hemisphere ocean transport of about 2 PW that is smaller by about 0.5–1.0 PW than all the other results (S88, M88, MD91), including ours. At 24°N (Table 3) our result is 1.4 PW larger than found from direct ocean measurements by Bryden (1991) and 1–2 PW larger than the values inferred by H82 and H85. The COADS analysis by da Silva et al. (1994) obtains a Northern Hemisphere peak transport of about 2.3 PW, more than 1 PW smaller than our value (Table 4).

c. Latent heat transport in the atmosphere

If we replace LH (or evaporation, E) in Eq. (8) by the precipitation (P), the atmospheric energy transport derived by the same procedure is the dry static energy transport that excludes the transport of latent heat (water vapor) in the atmosphere. Figure 7 shows the zonal annual mean E and P fluxes from Sellers (1965), and Fig. 8 compares the zonal annual mean total energy fluxes into the atmosphere obtained using either E or P to represent the latent heat contribution. The difference between the total (dry static plus latent) and dry static atmospheric energy transports, both shown in Fig. 9, is produced by $E - P$ and gives the latent heat transport in the atmosphere (Fig. 10). Comparison of Sellers' values for $E - P$ with the values given by Baumgartner

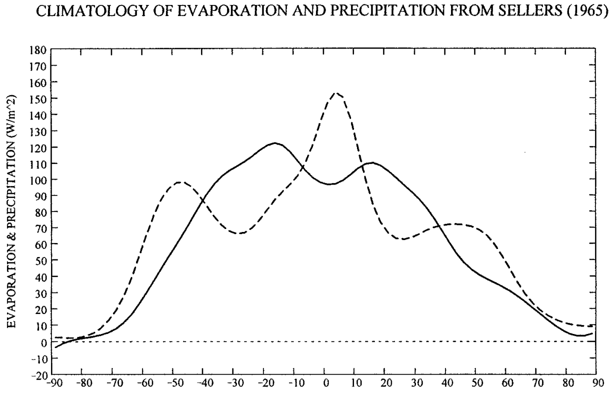

FIG. 7. Zonal annual mean latent energy fluxes (W m^{-2}) into the atmosphere implied by surface evaporation (solid line) and by precipitation (dashed line) from Sellers (1965). The actual heating of the atmosphere is caused by precipitation, but the transport of water vapor, implied by the difference between evaporation and precipitation, is included in the total energy transport when evaporation is used to represent the latent heat contribution.

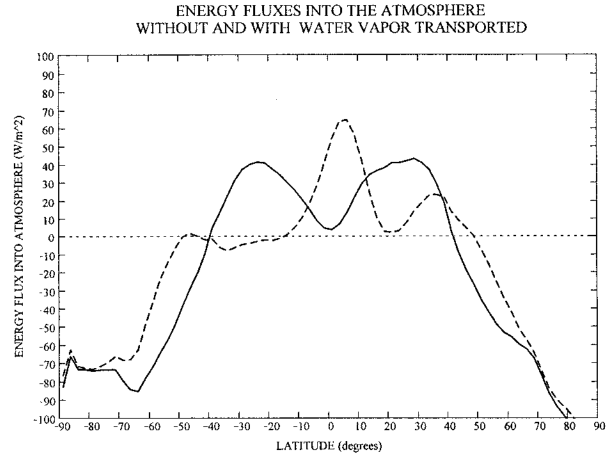

FIG. 8. Zonal annual mean total energy fluxes (W m^{-2}) into the atmosphere when using surface evaporation (solid line) and precipitation (dashed line) from Sellers (1965) to represent the latent heat contribution.

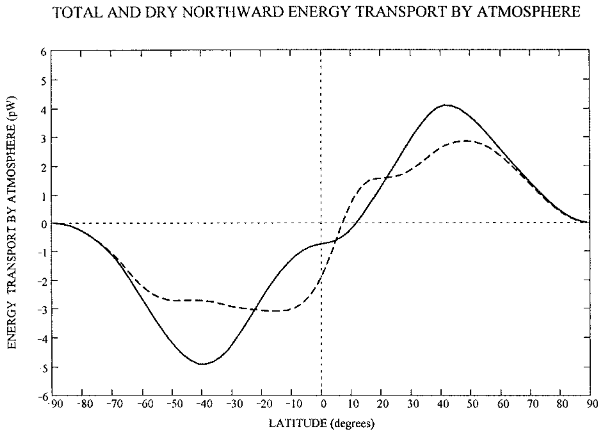

FIG. 9. Zonal annual mean northward dry static energy (dashed line) and total energy (solid line) transport (PW = 10^{15} W) by the atmosphere. The former is obtained from Eq. (5) when precipitation is used to represent latent heating of the atmosphere and excludes the transport of water vapor. The latter includes water vapor transport.

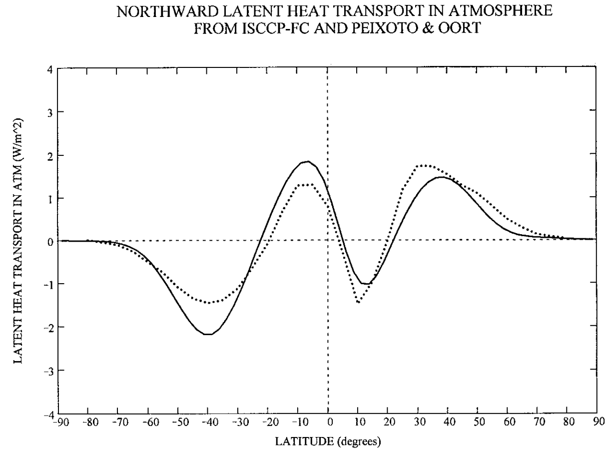

FIG. 10. Zonal annual mean water vapor transport (solid line) in W m^{-2} calculated from the difference of the total and dry static energy transports (shown in Fig. 9) compared with the corresponding result from Peixoto and Oort (1992) (dots).

and Reichel (1975, BR75 in following) and OP83 shows rms differences of 6–11 W m^{-2} in the annual zonal means; the agreement among these older $E - P$ climatologies is misleading, however, since many of these studies use essentially the same sources of information and the same bulk formulas. Comparison of Seller's values with COADS values of $E - P$ show rms differences of $\approx$40 W m^{-2}, caused mostly by large (46 W m^{-2} rms) differences in precipitation, particularly at higher latitudes. OP83 also directly calculate the latent heat transport and get values similar to those implied by their $E - P$ values, but the larger latent heat transports implied by the $E - P$ values of Sellers (1965) or BR75 agree better with more recent direct calculations reported by S88, M88, MD91, and K95.

d. Sensitivity study of the radiative fluxes

The main source uncertainty in the atmospheric and oceanic heat transports comes from the surface fluxes, since as Fig. 1 shows there is little difference in the latitudinal gradients of the FC and ERBE net radiative fluxes at TOA. Figure 11 shows all of the surface fluxes used to estimate the transports in this work. From comparisons of our calculated surface radiative fluxes with other measurements, the net radiative fluxes into the atmosphere are uncertain by about 10–25 W m^{-2} overall (Rossow and Zhang 1995). The Weare (1989) and Gleckler and Weare (1997) studies suggest overall uncertainties of the latent and sensible heat fluxes of about 30 and 10 W m^{-2}, respectively. Formally then, the total uncertainty of the surface energy flux is about 35–40 W m^{-2}, but some cancellation must occur because the global adjustment in the total surface energy fluxes, required to eliminate spurious transports at the north pole, is only about 25 W m^{-2}. More importantly, we must consider systematic errors in the meridional variation of the surface fluxes that would affect the calculated energy transport. The comparisons of the FC surface radiative fluxes with other measurements imply a possible systematic latitudinal variation of our SW flux errors from a high bias of about 20 W m^{-2} at the equator to almost no bias at middle latitudes [uncertainties in polar surface fluxes may be somewhat larger, cf. Curry et al. (1996)], which we believe is caused mostly by an underestimate of water vapor abundances in the particular dataset used (Rossow et al. 1991) and the aerosol optical thicknesses assumed (cf. Bishop et al. 1997). There may also be a much smaller ($\sim$5 W m^{-2} from equator to pole) meridional bias associated with the

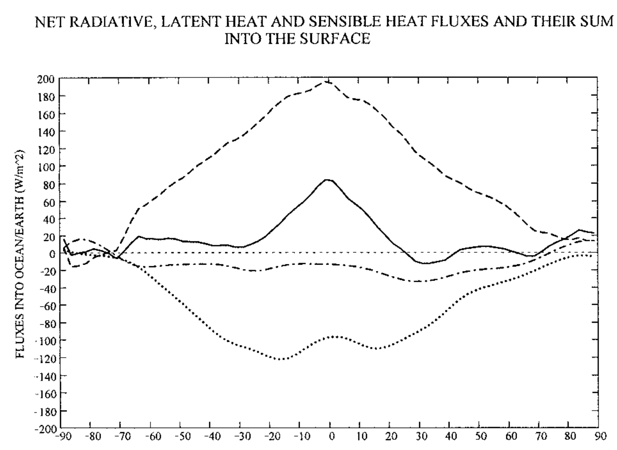

FIG. 11. Zonal annual mean values (W m^{-2}) of 1) net radiative heating of the surface (dashed line) from FC data, 2) latent cooling (evaporation) of the surface (dotted line) from Sellers (1965), 3) sensible cooling of the surface (dash–dotted line) from Sellers (1965), and 4) net energy into the surface (solid line), which is the sum of 1–3. The global mean net energy into the surface is adjusted by −22 W m^{-2} to determine ocean energy transport.

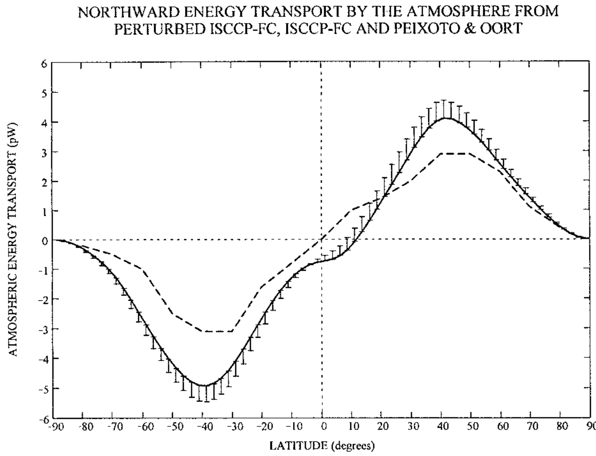

FIG. 12. Same as Fig. 5 but the bars on our (solid line) mean northward atmospheric energy transport (PW) show the change caused by reducing the net radiative heating of the surface by 20 W m^{-2} at the equator and nearly zero at the poles, giving a global mean reduction of 12 W m^{-2} (called FCm data). The results of Peixoto and Oort (1992) are shown as a dashed line.

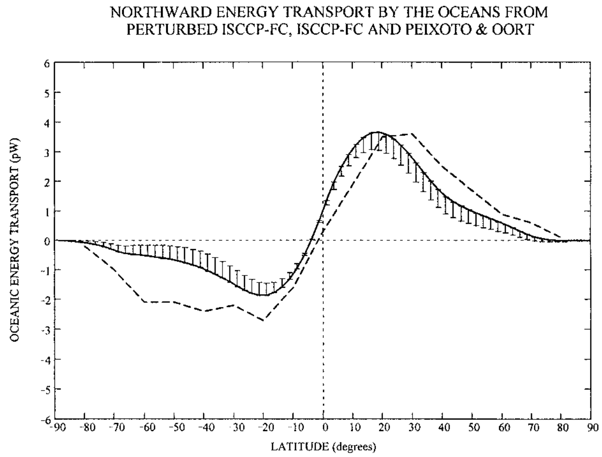

FIG. 13. Same as Fig. 6 but the bars on our (solid line) mean northward oceanic energy transport (PW) show the change caused by reducing the net radiative heating of the surface as in Fig. 12. The results of Peixoto and Oort (1992) are shown as a dashed line.

treatment of surface skin temperatures and emissivities (Rossow and Zhang 1995).

We explore the sensitivity of our results to possible errors in the meridional gradient of the net surface radiation by applying a multiplicative factor to reduce our net surface radiative fluxes by 20 W m^{-2} at the equator, but only by <1 W m^{-2} at the poles. The required global adjustment is reduced to $+14$ W m^{-2} from $+26$ W m^{-2}, which could easily be accounted for by systematic errors in (LH + SH) (we investigate errors in the surface latent and sensible heat fluxes in the next section). Figures 12 and 13 show the changes in the atmospheric and oceanic energy transports caused by this reduction of the surface net radiative flux in the form of bars on the original values from Figs. 5 and 6, respectively: the solid line at one end of the bars is the original result and the new result is at the other end of the bars. The global mean decrease in the surface net radiative flux of 12 W m^{-2} corresponds to an overall change in the northward energy transports of about 0.5 PW, an increase in atmospheric transport and a decrease in oceanic transport. This sort of an error in the surface radiative flux does not narrow the differences between our results and the older results of PO92 and it degrades the agreement with the newer atmospheric analyses, making our atmospheric transports about 0.5 PW larger (Table 2). On the other hand the revised oceanic transports agree better with the direct determinations in both hemispheres (Table 3).

e. Sensitivity study of the latent and sensible heat fluxes

Comparison of Sellers (1965) results with some of the other commonly used climatologies (e.g., OP83 and BR75) shows rms differences in annual zonal mean values of LH about 10–15 W m^{-2}; however, these estimates all use similar sources of information and similar bulk formulas. Weare (1989) examined the sources of error in using the common ''bulk'' formulas to estimate mean values of LH and SH and found global uncertainties to be about 30 and 10 W m^{-2}, respectively. Included in this estimate was a possible error in the meridional variation of LH that could be $\approx$10 W m^{-2} from equator to pole. Kent and Taylor (1995) compared several different calculations for the same dataset in the North Atlantic and found that all of the analyses tended to overestimate the annual mean latent and sensible heat fluxes in a range from about 10–30 W m^{-2}, mostly because of an overestimate of the wintertime fluxes, but that the Esbensen and Kushnir (1981) values were closest to their best estimate. Gleckler and Weare (1997) find similar magnitude uncertainties for the flux climatology of Oberhuber (1988). Comparison of the Sellers' fluxes with those from the COADS compilation over the oceans shows rms differences in annual longitudinal means of LH and SH of 16 and 5 W m^{-2}, respectively. [Comparison of Sellers' values of LH and SH with values reported in the recently released National Centers for Environmental Prediction (NCEP) reanalysis (Kalnay et al. 1996) shows rms differences in annual zonal means of 9 and 12 W m^{-2}, respectively.] Thus, we estimate the total uncertainty of the annual longitudinal mean value of (LH + SH) was to be at least $>$20 W m^{-2}.

We explore the sensitivity of the calculated energy transports to uncertainties in LH and SH by repeating the ocean transport calculations with the COADS values of LH and SH in place of the Sellers's values. The global adjustment to the net surface flux *into* the ocean from (FC + COADS), required to produce zero global mean transport, is -40 W m^{-2}, as compared with an adjustment of -22 W m^{-2} for (FC + Sellers). The main

TABLE 5. Differences (A − B) of oceanic radiative, latent, and sensible heat fluxes (W m^{-2}).

Compared quantity	A	B	Global mean A	Global mean B	Global mean diff	Global rms	Zonal rms
Oceanic net SW	FC	COADS	173.4	172.3	1.0	6.8	9.0
Oceanic net LW	FC	COADS	−39.8	−49.8	10.0	12.7	16.2
Oceanic total net radiation	FC	COADS	133.6	122.5	11.0	12.9	12.4
Ocean latent heat	Sellers	COADS	105.5	90.1	15.4	16.6	16.1
Ocean sensible heat	Sellers	COADS	10.4	9.4	1.1	3.5	4.5

systematic difference (Table 5) is that the COADS values of LH are about 16 W m^{-2} smaller in magnitude on average than the Sellers's values (the global mean SH is the same to within 1 W m^{-2}, but rms differences in zonal mean values are about 4 W m^{-2}; that is, the overestimate of surface heating that we had with (FC + Sellers) is increased because the latent cooling flux from COADS is smaller. (If we combine the modified FC values with COADS, the global adjustment is about −27 W m^{-2}; if we use the COADS surface net radiative fluxes in place of FC, the global adjustment is still −29 W m^{-2}.) Most importantly, the difference between the COADS and Sellers's values of (LH + SH) is larger in the Tropics than in midlatitudes. Figure 14 shows the resulting oceanic energy transport.[4] Since the COADS values of (LH + SH) reduce the meridional gradient of the surface net flux relative to the Sellers's values, less ocean energy transport is required. The magnitude of the reduction is similar to that produced by reducing the surface net radiative fluxes, about 0.5 PW in the Northern Hemisphere, but about 1 PW in the Southern Hemisphere (Table 4). The reduced transports improve agreement with direct ocean energy transport estimates in both hemispheres, but the Northern Hemisphere transport is still about 0.5–1.0 PW higher (Table 3).

5. Discussion

a. Uncertainties

Based in part on the range of values reported in recent literature, we summarize the uncertainties in the three components of the energy budget as follows[5]: 1) the range of TOA net radiative fluxes implies uncertainties in the total energy transports of 0.5–1.0 PW, about 10%–20% of peak values; 2) direct estimates of atmospheric energy transports have uncertainties of 0.5–1.5 PW, about 15%–40% of peak values; 3) direct estimates of oceanic energy transports at a few locations and estimates based on surface fluxes have uncertainties of 0.5–1.0 PW, about 30%–75% of peak values. The uncertainty of our values of the atmospheric and oceanic transports caused by uncertainties in the radiative fluxes is about 0.5–1.0 PW; but because of the additional and larger uncertainty in the surface latent heat flux, our total uncertainty is about 1.0–1.5 PW. A comparison of various results at specific latitudes in each hemisphere is shown in Tables 1, 2, and 3.

The uncertainty of the total energy transport implied by the TOA net radiation is only as large as 1 PW because of the difference between newer and older estimates. Even larger differences exist between the satellite-based estimates and presatellite values. The newer estimates from ERBE and the fluxes calculated from the ISCCP datasets (Zhang et al. 1995), both of which account for diurnal variations better than earlier results, agree to within 0.13 PW. Since the ERBE and ISCCP results are completely independent results, their agreement suggests that, of the three energy budget components, the total energy transport estimated from the TOA net radiation is now the most certain with an un-

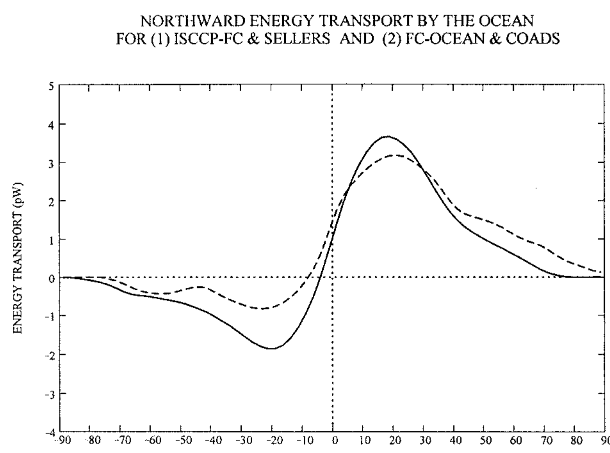

FIG. 14. Zonal annual mean northward energy transport (PW) by the oceans calculated by combining our net surface radiative fluxes with the latent and sensible heat fluxes from Sellers (1965) (solid line) and from COADS (dashed line). In the first case, zonal mean fluxes are used (covering both land and ocean), whereas in the second case the longitudinal averages cover only ocean areas. To obtain zero global means, the total surface fluxes are adjusted by −22 W m^{-2} and −40 W m^{-2}, respectively.

[4] Since the COADS fluxes are given only for the ocean surface, we use the FC fluxes only over ocean areas in this calculation.

[5] There is no direct relation between flux and transport errors; however, based on our sensitivity tests, we crudely estimate energy transport uncertainties from flux uncertainties by assuming that about two-thirds of the flux error appears as an error in the latitudinal gradients that alters the calculated transport, so that a 7 W m^{-2} flux error corresponds to a 0.5-PW transport error (cf. Keith 1995). Thus, these estimates are not precise. Fractional errors are determined relative to the peak transports, since the integral equations constrain all analyses toward zero transport at the poles.

certainty no more than 0.2 PW, certainly <0.5 PW. Also, of the three components, the satellite-based net radiation has the most complete, uniform, and dense sampling over the globe and over time. In particular, the total transport is known equally well in both hemispheres.

Our estimate of the atmospheric transport uncertainty is larger than claimed by most authors because it is based on the range of results reported in the literature, particularly for the Southern Hemisphere, and on cited comparisons of the net surface fluxes inferred from the calculated atmospheric transports (cf. TS94; Gleckler and Weare 1997). As with TOA net radiation, more recent estimates exhibit better agreement (to within about 0.5–1.0 PW) and show generally larger atmospheric transports than older results. PO92 comment that the deficiency in the geographic coverage of the upper-air data, particularly over oceans and especially in the Southern Hemisphere, would tend to underestimate the atmospheric transport (cf. K95). This conclusion is supported by S88, who obtained significantly larger atmospheric transports from the same dataset analyzed by PO92 (actually OP83) by applying dynamic constraints to the divergent wind component. These constraints are similar to those applied by the model assimilation techniques used in all of the more recent studies, but the model-based results continue to be sensitive to how other physical processes are represented in particular models, especially tropical connection (see M88; TS94). The recent values of the atmospheric energy transport are all generally within about 0.5 PW of the values reported by S88 in the Northern Hemisphere, but the range of disagreement in the Southern Hemisphere is still about 1.0 PW, suggesting that it is the poor quality of Southern Hemisphere wind observations that still controls the uncertainties there.

The upper limit of our estimated uncertainty range of 1.5 PW arises from the large surface flux errors, $\approx$30 W m^{-2} (e.g., TS94; Gleckler and Weare 1997), implied by the atmospheric analyses. The fact that these ''residual'' errors are similar in magnitude to the uncertainties of direct estimates of the surface fluxes (Weare 1989) may result partly from the use of similar ''bulk formulations'' to calculate surface fluxes: Weare (1989) estimates formulation errors to be at least 20 W m^{-2}. However, there is also some evidence that such large errors in surface fluxes may be due, in part, to poor representation of clouds and radiation in the assimilation models (Gleckler et al. 1995). Such large surface flux errors could correspond to errors in the atmospheric energy transports of as much as 1.5 PW (cf. K95).

Although some uncertainty in direct estimates of atmospheric energy transport is contributed by errors in the measurements of atmospheric temperature and humidity, the dominant cause of uncertainty is likely to be errors in the wind measurements (cf. Holopainen and Fortelius 1986; Trenberth and Solomon 1994; Baker et al. 1995). This conclusion is supported by the magnitude of the changes caused in the results when use is made of a dynamic constraint on the winds (S88) or when different models are used (M88; K95) and by the persistent large disagreements in the Southern Hemisphere.

Our estimate of the oceanic transport uncertainty is larger than cited by most authors (but consistent with PO92's estimate) because the range of values reported in the literature is larger than the claimed uncertainties (cf. the range reported by MacDonald 1993) and because the so-called direct determinations, in fact, include calculations of some components, in particular the Ekman transport, which can be about one-third to one-half the total transport. These calculations use surface wind climatologies that are likely to be particularly inaccurate in the Southern Hemisphere (cf. Atlas et al. 1996) and may underestimate the Ekman transport by averaging out the contributions from shorter-term circulation variations, particularly seasonal changes (cf. Liu et al. 1993; Atlas et al. 1996). In addition, the ocean transport estimates are derived from what are effectively single samples of the ocean state at particular locations and in particular seasons. Consequently, neglect of the storage terms in the energy budget may introduce more error. The complete latitudinal distribution of the ocean heat transport has not been determined directly from ocean measurements. Instead, such latitude distributions are derived indirectly from estimates of the surface energy fluxes (e.g., Hastenrath 1982; Hsuing 1985; da Silva et al. 1994), as we do; however, these results are subject to the uncertainties discussed by Weare (1989) and Gleckler and Weare (1997) with somewhat larger uncertainties in the surface net radiation than we have.

The uncertainty in our indirect results is controlled mostly by the uncertainties of the surface energy fluxes. As discussed above, uncertainty in the TOA net radiation corresponds to uncertainties of the total energy transport of no more than 0.2 PW, but the uncertainty in the surface fluxes, based on our sensitivity studies, corresponds to transport uncertainties of at least 1 PW. Although the radiative flux uncertainties are significant, the bigger problem is still the latent heat flux, LH. Direct comparisons of several datasets (Sellers 1965, KR75, OP83, COADS) suggest uncertainties in zonal annual mean values of only $\approx$15 W m^{-2}, but these estimates are not entirely independent since they use similar formulas and ship observations (see Kent and Taylor 1995). Weare's (1989) estimate of the LH uncertainty is 30 W m^{-2}, but this estimate does not indicate how much the errors distort the large-scale pattern that alters the transports. The large uncertainties can be illustrated by two analyses of the COADS dataset: using their radiative,[6]

[6] The annual zonal mean SW fluxes in COADS agree with our FC values to within about 7 W m^{-2}, with almost no bias (Table 5); however, a map of the differences shows regional differences of both signs that are >10 W m^{-2} that are clearly related to large-scale variations in cloud properties. The COADS annual zonal mean LW flux is systematically larger than the FC value by 10 W m^{-2} (Table 4),

latent, and sensible fluxes in our calculations produces ocean energy transports that differ by almost 1 PW from the analysis of the same data by da Silva et al. (1994), particularly in the Southern Hemisphere (Tables 3 and 4). In addition to the large adjustments in the surface wind speed and drag coefficients employed, da Silva et al. (1994) also strongly constrain their transport estimates using the few direct estimates from ocean measurements. Thus, we estimate that the ocean transport uncertainty is still near 1 PW, possibly larger.

What are the prospects for improving accuracy of the estimates of the three global energy budget components? Further refinements of the TOA net radiation budget are being worked on (cf. discussions in Rossow and Zhang 1995; Wielicki et al. 1995, Wielicki et al. 1996), but this component is already the most accurately determined one. Moreover, since this component is determined from satellite observations, its space–time sampling is much superior, being much denser and more complete and uniform in its coverage than any of the other datasets used in these analyses. Combining the *Nimbus-7* Earth Radiation Budget data, ERBE data, and the ISCCP-based flux calculations, we will soon have a dataset covering almost 20 yr; new satellite missions will extend the record to at least 35 yr. With work already in progress and refinements planned, this component should eventually be known to at least 0.1 PW.

Improving the direct determination of the ocean heat transport requires much better, time-resolved surface wind datasets to calculate the Ekman component more accurately and more complete measurements with better space–time sampling of the three-dimensional currents, temperature, and salinity to determine the transport by the mean thermohaline circulation. The global ocean surface winds are already being improved by analysis of passive microwave measurements from satellites and should be improved further by forthcoming scatterometer measurements from satellites (Atlas et al. 1996). More complete information on the deep circulation will come from the continuing series of satellite altimeter missions. The World Ocean Circulation Experiment (World Climate Program 1986) is conducting a one-time survey of the state and circulation of the world's deep oceans, but this difficult and expensive set of measurements may not be repeated soon. Assimilation of these data by ocean circulation models may improve the dynamic consistency of the results; however, more complete observations are needed to verify the models. Thus, although there should be significant improvements in the direct estimates of ocean energy transport in the coming years, the accuracy will still be significantly limited by the poor time sampling of the observing system.

Improving the direct determination of the atmospheric energy transport requires much better measurements of the synoptic variations and three-dimensional distribution of, in order of priority, winds, humidity, and temperature. These measurements need to be globally complete. Current satellite measurements have already improved the global coverage of temperature and humidity measurements and planned improvements of operational satellite systems will usefully increase both space–time resolutions and measurement accuracy. However, these improvements will have little direct effect on the energy transport calculations because the uncertainties are dominated by the uncertainties in the winds (cf. Holopainen and Fortelius 1986; Trenberth and Solomon 1994; Baker et al. 1995). Although there are some operational determinations of winds in certain regions at some altitudes by tracking cloud motions (see Baker et al. 1995), there are currently no plans to improve direct wind measurements by obtaining global coverage and increasing vertical resolution. The availability of better ocean surface winds from satellite scatterometers may improve Southern Hemisphere results somewhat. A more important improvement may come from direct assimilation of satellite-measured radiances that are sensitive to atmospheric trace constituents, particularly water vapor; such an assimilation approach should more strongly constrain the divergent component of the circulation that transports energy (Andersson et al. 1994; McNally and Vesperini 1996). Nevertheless, without improvements in the direct measurements of the 3D wind field, further reduction of the uncertainties in direct estimates of atmospheric energy transports, particularly in the Southern Hemisphere, is likely to be limited in the near future.

A major advantage of the SPEB method is realized if all the surface fluxes can be determined from satellite observations: satellite-based results would have very dense space and time sampling that is globally complete and uniform. Improving the accuracy of the indirect determination of the atmospheric and oceanic energy transports requires reducing the systematic uncertainties in the surface energy fluxes. Uncertainties in the surface net radiation can be reduced by 1) improved treatment of the effects of cloud inhomogeneities (e.g., Barker and Davies 1992a,b; Cahalan et al. 1994) and ice clouds (e.g., Minnis et al. 1993; Mishchenko et al. 1996) on SW fluxes; 2) improved information on the location of cloud bases that affects the LW fluxes (e.g., Zhang et al. 1995; Wang and Rossow 1995); and 3) improved aerosol information (e.g., Bishop et al. 1997). Improved atmospheric temperature and humidity retrievals will also be important (see Zhang et al. 1995). All of these items are being worked on now, so the uncertainty of the energy transports related to surface net radiation should soon be reduced below 0.5 PW. Reducing un-

---

but exhibits even stronger regional differences >20 W m^{-2} associated with variations in cloud properties. The pattern of these radiative flux differences indicates that the bulk formulas commonly used to calculate surface radiative fluxes, which depend only on cloud cover, fail to capture all of the cloud-induced variations in surface radiative fluxes (see Fung et al. 1984; Seager and Blumenthal 1994).

certainties in LH and SH requires improved measurements of surface winds, near-surface humidities, and surface and near-surface temperatures. Improved determinations of these quantities over land presents significant challenges (e.g., Hall et al. 1992), but tests of satellite-based techniques for estimating LH and SH over oceans show significant promise (Liu and Gautier 1990; Miller and Katsaros 1992; Liu et al. 1994; Chou et al. 1995; Clayson et al. 1996). In particular these studies show that the surface flux uncertainty is reduced with higher time resolution surface wind measurements, which can be obtained from satellite-based measurements. Thus, expected improvements in satellite-based determinations of surface energy fluxes may reduce the atmosphere and ocean energy transport uncertainties below 1.0 PW in the near future.

The different sources and geographic distributions of the uncertainties in the surface fluxes obtained from the SPEB method compared with in situ atmospheric and oceanic observations provide a useful complement to the other methods of estimating the atmospheric and oceanic energy transports. Moreover, attempts to determine the surface energy fluxes will also be useful as independent checks of the atmospheric and oceanic assimilation models used to determine transports. For the time being, indirect determination of the ocean energy transports from the difference between the TOA radiation budget and calculated atmospheric transports or from net surface energy fluxes continues to be more accurate than direct measurements and provides more complete coverage.

b. Cloud effects on energy transports

The main features of our results are the same as in previous studies (Table 4); namely, the total energy transport is dominated by ocean transport in a zone from about 15°S to 30°N and by atmospheric transport at higher latitudes (Fig. 15). Total energy transport peaks at about 6 PW near 35° lat in both hemispheres, atmospheric transport peaks at 4–5 PW near 40° lat, and oceanic transport peaks at 2–3 PW near 20° lat in both hemispheres (Fig. 15). Total and atmospheric transports are larger in the Southern than in the Northern Hemisphere, but oceanic transport is larger in the Northern than in the Southern Hemisphere. Peak atmospheric transport is about three times larger than peak oceanic transport in the Southern Hemisphere, but in the Northern Hemisphere the two transports have nearly equal peak values. The atmospheric transport of dry static energy is offset by more than 50% by latent transport near the equator (±20°) and reinforced by about 40% by latent transport in midlatitudes (Fig. 9). Most studies exhibit small cross-equatorial energy transports, southward by the atmosphere and northward by the ocean, resulting in a small northward total transport (Tables 1, 2, and 3); exceptions are the results of H82, where all the transports are nearly zero, and S88, where the atmospheric transport is northward and the oceanic transport is southward (H82, PO92, and K95 actually have zero cross-equatorial transport by the atmosphere). The hemispheric symmetry of both the ERBE and FC TOA net radiative fluxes (as well as earlier estimates) requires negligible (<5%) interhemispheric transport by the total atmosphere–ocean system; however, the hemispheric asymmetry of the Hadley circulation (e.g., Lorenz 1967, PO92) and estimates of ocean transports (Wunsch 1984, see also da Silva 1994) both suggest that there are significant hemispheric differences in both atmospheric and oceanic transports.

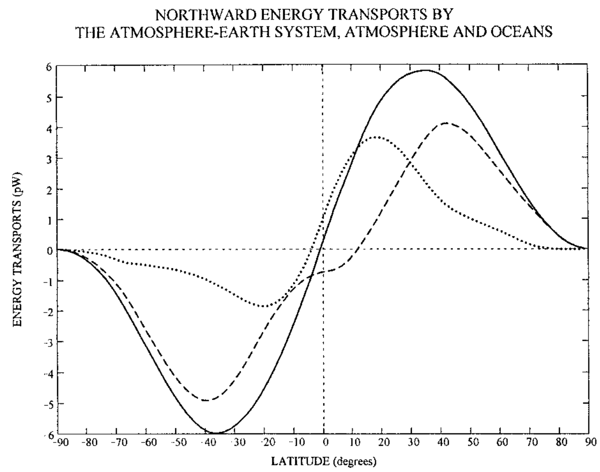

FIG. 15. Zonal annual mean northward total energy transports (PW $= 10^{15}$ W) in the atmosphere–ocean system as required by the top-of-atmosphere radiation budget (solid line), by the atmosphere (dashed line), and by the ocean (dotted line) from this work (FC + Sellers data).

The hemispheric symmetry of the net TOA radiation is, in fact, remarkable because there is a significant hemispheric difference of the surface albedos: the surface albedo is significantly lower in the Southern Hemisphere, especially in midlatitudes (Table 6). If there were no clouds, the net SW absorbed by the Southern Hemisphere would exceed that in the Northern Hemisphere by about 7 W m^{-2} (Table 6); if the difference between the albedos of the two polar regions were removed, this difference would be about 15 W m^{-2}. Although this net SW heating difference might simply be balanced by higher temperatures in the Southern Hemisphere, producing correspondingly more LW cooling, this difference might also produce a cross-equatorial energy transport. In fact, despite the albedo difference, the annual average surface temperature of the Southern Hemisphere is 2 K *lower* than that of the Northern Hemisphere and the LW cooling at TOA is 3 W m^{-2} lower (Table 6). Hemispheric differences in cloud properties nearly cancel the hemispheric differences in surface albedo producing a net TOA SW radiative flux that is nearly identical in both hemispheres. At midlatitudes, smaller cloud amounts associated with more land actually reverse the situation so that the net SW absorbed

TABLE 6. Annual, zonal, and hemispheric mean radiation budget quantities from Rossow and Zhang (1995), where α_s and T_s are surface albedo and temperature, $S_t\downarrow$ is the solar insulation of top of atmosphere (TOA), NS_t and NL_t are the net shortwave and longwave fluxes at TOA averaged over all conditions, and CLR indicates clear conditions.

	Mean values for SH/NH zones						
Latitude range	α_s	T_s	$S_t\downarrow$	NS_t	NS_t − CLR	NL_t	NL_t − CLR
0°–20°	0.097/0.113	299.2/300.5	409.9/409.8	302.5/298.5	356.9/352.6	−251.4/−249.8	−275.6/−276.3
20°–40°	0.099/0.150	293.6/295.5	366.4/366.2	260.9/255.1	317.1/302.9	−247.0/−248.9	−268.1/−268.8
40°–60°	0.092/0.180	277.8/278.9	287.8/287.5	164.4/169.8	246.3/228.2	−216.6/−218.3	−239.1/−239.7
60°–90°	0.573/0.405	253.1/263.4	202.1/201.7	82.8/102.7	112.5/130.9	−180.0/−199.3	−191.9/−210.5
0°–90°	0.135/0.157	287.8/289.8	341.7/341.5	229.8/230.5	287.5/280.2	−232.8/−235.7	−254.0/−257.1

in the 40°–60° zone is slightly larger in the Northern instead of the Southern Hemisphere (Table 6). Although the clouds significantly reduce the LW cooling, they have little effect on the hemispheric difference in LW cooling.

The consequence of the near-symmetry of the TOA net radiation produced by *offsetting asymmetries of surface and cloud properties* is that the ocean and atmosphere energy balances in the two hemispheres must differ because the majority of the TOA net SW flux appears as *surface* heating and the majority of the TOA net LW flux appears as *atmospheric* cooling, so that the near-symmetry of the separate components requires interhemispheric energy transports in both the atmosphere and ocean. In our results there is net energy transport from north to south in the atmosphere and from south to north in the ocean of about 1 PW magnitude (Fig. 15). The analysis of COADS data by da Silva et al. (1994) is consistent with Wunsch's (1984) estimate of about 0.6 PW northward transport by the oceans. The atmospheric transport of dry static energy is larger still because it is opposed by a net latent transport from south to north (Fig. 9), consistent with estimates of more evaporation in the Southern Hemisphere and more precipitation in the Northern Hemisphere, constituting a net water vapor transport from ocean to land in the atmosphere (PO92, da Silva et al. 1994). The north to south transport in the atmosphere also produces a peak atmospheric energy transport that is larger in the Southern Hemisphere than in the northern by almost 1 PW. The ocean exhibits south to north transport, which is consistent with more net SW flux there, and a peak transport that is 1.5 PW larger in the Northern Hemisphere than in the Southern Hemisphere (Fig. 15). The larger peak transport in the north is consistent with direct determinations that show northward transport in both the Atlantic and Pacific basins, whereas the smaller transport in the south results from southward transports in the Indian and Pacific basins offset by northward transport in the Atlantic (MacDonald 1993).

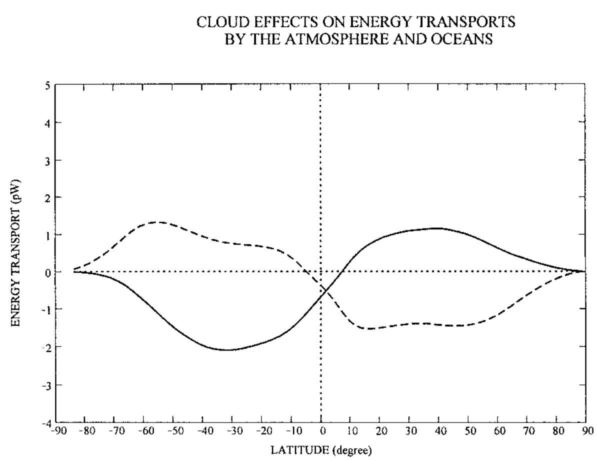

FIG. 16. Qualitative indication of cloud-radiative effects on the zonal annual mean northward total energy transports (PW) by the atmosphere (solid line) and the ocean (dashed line) given by the differences in the inferred transports using total radiative fluxes (cloudy plus clear) and clear-sky radiative fluxes. The changes in the transports shown result from adding clouds, all other factors being held constant.

To obtain a *qualitative* indication of the cloud effects on these energy transports, the entire analysis was repeated using clear-sky radiative fluxes, which are calculated from the same surface and atmospheric properties with all clouds removed (Zhang et al. 1995), in place of the total fluxes. The ''clear'' energy transports are calculated by the same procedure, including adjusting the global mean TOA and surface fluxes to zero (the adjustments are −28 W m^{-2} and +21 W m^{-2}, respectively, where the latter is the adjustment to the surface fluxes *into* the atmosphere[7]). The values of LH and SH remain the same. These ''clear'' transports represent only the amount of energy that must be moved from equator to pole to balance differences in the net radiation and are not the real transports that would occur in the atmosphere and ocean if clouds were actually removed. In particular, such a change in the net radiation would also trigger significant changes in temperature and LW cooling as well as many feedbacks, especially in the latent heat exchanges. Figure 16 shows the differences between the actual atmospheric and oceanic transports and their ''clear'' values. The change in net radiative cooling of the atmosphere produced by *adding clouds* enhances the atmospheric energy transports by about

[7] If we calculate the ocean transport directly from the clear-sky surface energy fluxes, then the global adjustment is −49 W m^{-2}, the sum of the global mean TOA and surface fluxes.

25% in the Northern Hemisphere and by almost 40% in the Southern Hemisphere (Fig. 16). This effect is the consequence of the fact that clouds, all other factors held constant, decrease the radiative cooling of the tropical atmosphere (an effective heating) and increase the radiative cooling of the atmosphere at higher latitudes (Rossow and Zhang 1995). One can think of this effect either as a strengthening of the mean circulation by the cloud-radiative effects, thereby increasing energy transport, or as an addition of energy at the source and a removal of energy at the sink, which requires more transport. Notice that the cloud effect reaches its maximum on the equatorward side of the peak atmospheric transport, serving to shift the peak atmospheric transport equatorward.

The decrease of the net radiative heating of the surface produced by *adding clouds* reduces the oceanic transports by more than 30% in the Northern Hemisphere subtropics and by more than 50% in the Southern Hemisphere subtropics (Fig. 16). At higher latitudes, *removing clouds* without other changes would actually require a change in the direction of oceanic energy transport as the consequent SW radiative heating would require removal of significant energy from midlatitude oceans. The reduced surface solar heating in the presence of clouds also serves to shift the peak oceanic transport equatorward.

Because our radiative fluxes are calculated from physical quantities, we can for the first time identify how cloud-induced changes in the radiation budget affect the atmospheric and oceanic energy transports. Cloud-induced reductions in SW radiative heating mostly perturb the surface radiation budget and the balancing oceanic energy transports, whereas cloud-induced reductions in LW radiative cooling mostly perturb the atmospheric radiation budget and the balancing atmospheric energy transports. In the real case, changes in the clouds would affect both the atmosphere and the ocean, but they would also interact so that the actual change in transports cannot be estimated directly. Clouds also cancel the effect of the lower Southern Hemisphere surface albedo, producing a near-perfect hemispheric symmetry of the TOA SW heating; consequently the poleward energy transport required of the ocean tends to be smaller in the Southern than in the Northern Hemisphere, all other things being constant. Clouds also enhance the poleward energy transport required of the atmosphere more in the Southern Hemisphere, all other things being constant. These hemispheric imbalances are indicated in the net energy transport from south to north by the ocean and from north to south by the atmosphere. All other factors held constant, adding clouds tends to require more energy transport by the atmosphere and less by the ocean.

Acknowledgments. This research is supported by the NASA Global Radiation and Data Analysis program (Dr. Robert A. Schiffer and Dr. Robert J. Curran). We were helped significantly by two thorough reviews. We also acknowledge helpful conversations with Dr. T. H. Vonder Haar.

REFERENCES

Aagaard, K., and P. Greismann, 1975: Toward new mass and heat budgets for the Arctic Ocean. *J. Geophys. Res.,* **80,** 3821–3827.

Andersson, E., J. Pailleux, J.-N. Thépaut, J. R. Eyre, A. P. McNally, G. A. Kelly, and P. Coutier, 1994: Use of cloud-cleared radiances in three/four-dimensional variational data assimilation. *Quart. J. Roy. Meteor. Soc.,* **120,** 627–653.

Atlas, R., R. N. Hoffman, S. C. Bloom, J. C. Jusem, and J. Ardizzone, 1996: A multiyear global surface wind velocity dataset using SSM/I wind observations. *Bull. Amer. Meteor. Soc.,* **77,** 869–882.

Baker, W. E., and Coauthors, 1995: Lidar-measured winds from space: A key component for weather and climate prediction. *Bull. Amer. Meteor. Soc.,* **76,** 869–888.

Barker, H. W., and J. A. Davies, 1992a: Solar radiative fluxes for broken cloud fields above reflecting surfaces. *J. Atmos. Sci.,* **49,** 749–761.

——, and ——, 1992b: Solar radiative fluxes for stochastic, scale-invariant broken cloud fields above reflecting surfaces. *J. Atmos. Sci.,* **49,** 1115–1126.

Baumgartner, A., and E. Reichel, 1975: *The World Water Balance.* Elsevier, 179 pp.

Bishop, J. K. B., W. B. Rossow, and E. G. Dutton, 1997: Surface solar irradiances from the International Satellite Cloud Climatology Project 1983–1991.. *J. Geophys. Res.,* **102,** 6883–6910.

Bryden, H. L., D. H. Roemmich, and J. A. Church, 1991: Ocean heat transport across 24°N in the Pacific. *Deep-Sea Res.,* **38,** 297–324.

Budyko, M. I., 1963: *Atlas Teplovogo Balansa.* Gidrometeorologicheskoe Izdateel'skoe, 69 pp.

Cahalan, R. F., W. Ridgway, W. J. Wiscombe, S. Gollner, and Harshvardhan, 1994: Independent pixel and Monte Carlo estimates of stratocumulus albedo. *J. Atmos. Sci.,* **51,** 3776–3790.

Campbell, G. G., and T. H. Vonder Haar, 1980: Climatology of radiation budget measurements from satellites. Paper 323, Dept. of Atmospheric Sciences, Colorado State University, Fort Collins, CO, 74 pp. [Available from Dept. of Atmospheric Science, Colorado State University, Fort Collins, CO 80523.]

Carissimo, B. C., A. H. Oort, and T. H. Vonder Haar, 1985: Estimating the meridional energy transports in the atmosphere and ocean. *J. Phys. Oceanogr.,* **15,** 82–91.

Chou, S., R. M. Atlas, C. Shie, and J. Ardizzone, 1995: Estimates of surface humidity and latent heat fluxes over oceans from SSM/I data. *Mon. Wea. Rev.,* **123,** 2405–2425.

Clayson, C. A., and J. A. Curry, 1996: Determination of surface turbulent fluxes for TOGA COARE: Comparison of satellite retrievals and in situ measurements. *J. Geophys. Res.,* **101,** 28 503–28 513.

——, ——, and C. W. Fairall, 1996: Determination of surface turbulent fluxes for TOGA COARE: Comparison of in situ measurements and satellite retrievals. *J. Geophys. Res.,* in press.

Curry, J. A., W. B. Rossow, D. Randall, and J. L. Schramm, 1996: Overview of Arctic cloud and radiation. *J. Climate,* **9,** 1731–1764.

da Silva, A. M., C. C. Young, and S. Levitus, 1994: *Atlas of Surface Marine Data 1994.* Vol. 1, *Algorithms and Procedures,* NOAA Atlas NESDIS 6, U.S. Dept. of Commerce, 83 pp.

Esbensen, S. K., and Y. Kushnir, 1981: The heat budget of the global ocean: an atlas based on estimates from surface marine observations. Climatic Research Institute Rep. 29, Oregon State University, 27 pp. [Available from College of Oceanographic and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331.]

Fung, I. Y., D. E. Harrison, and A. A. Lacis, 1984: On the variability

of the net longwave radiation at the ocean surface. *Rev. Geophys. Space Phys.,* **22,** 177–193.
Gleckler, P. J., and B. C. Weare, 1997: Uncertainties in global ocean surface heat flux climatologies derived from ship observations. *J. Climate,* in press.
——, and Coauthors, 1995: Interpretation of ocean energy transports implied by atmospheric general circulation models. *Geophys. Res. Lett.,* **22,** 791–794.
Hall, F. G., K. F. Huemmrich, S. J. Goetz, P. J. Sellers, and J. E. Nickeson, 1992: Satellite remote sensing of surface energy balance: Success, failures and unresolved issues in FIFE. *J. Geophys. Res.,* **97,** 19 061–19 089.
Harrison, E. F., P. Minnis, B. R. Barkstrom, V. Ramanathan, R. D. Cess, and G. G. Gibson, 1990: Seasonal variation of cloud radiative forcing derived from the Earth Radiation Budget Experiment. *J. Geophys. Res.,* **95,** 18 687–18 703.
Hastenrath, S., 1982: On meridional heat transports in the world ocean. *J. Phys. Oceanogr.,* **12,** 922–927.
Holopainen, E., and C. Fortelius, 1986: Accuracy of estimates of atmospheric large-scale energy flux divergences. *Mon. Wea. Rev.,* **114,** 1910–1921.
Hsuing, J., 1985: Estimates of global oceanic meridional heat transport. *J. Phys. Oceanogr.,* **15,** 1405–1413.
Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteor. Soc.,* **77,** 437–471.
Keith, D. W., 1995: Meridional energy transport: Uncertainty in zonal means. *Tellus,* **47A,** 30–44.
Kent, E. C., and P. K. Taylor, 1995: A comparison of sensible and latent heat flux estimates for the North Atlantic Ocean. *J. Phys. Oceanogr.,* **25,** 1530–1549.
Liu, W. T., and C. Gautier, 1990: Thermal forcing on the tropical Pacific from satellite data. *J. Geophys. Res.,* **95,** 13 209–13 217.
——, W. Tang, and R. Atlas, 1993: Sea surface temperature exhibited by an ocean general circulation model in response to wind forcing derived from satellite data. *Remote Sensing of the Oceanic Environment,* Y. Sugimori and R. W. Stewart, Eds., Seibutsu Kenkyusha, 350–355.
——, A. Zhang, and J. K. B. Bishop, 1994: Evaporation and solar irradiance as regulators of sea surface temperature in annual and interannual changes. *J. Geophys. Res.,* **99,** 12 623–12 637.
Lorentz, E. N., 1967: The nature and theory of the general circulation of the atmosphere. WMO-No. 218, TP-115, World Meteorological Organization, Geneva, Switzerland, 161 pp. [Available from World Meteorological Organization, Avenue Guiseppe-Motta, 1211 Geneva, Switzerland.]
MacDonald, A., 1993: Property fluxes at 30°S and their implications for the Pacific–Indian throughflow and the global heat budget. *J. Geophys. Res.,* **98,** 6851–6868.
Masuda, K., 1988: Meridional heat transport by the atmosphere and the ocean: Analysis of FGGE data. *Tellus,* **40A,** 285–302.
McNally, A. P., and M. Vesperini, 1996: Variational analysis of humidity information from TOVS radiances. *Quart. J. Roy. Meteor. Soc.,* **122,** 1521–1544.
Michaud, R., and J. Derome, 1991: On the mean meridional transport of energy in the atmosphere and oceans as derived from six years of ECMWF analyses. *Tellus,* **43A,** 1–14.
Miller, D. K., and K. B. Katsaros, 1992: Satellite-derived surface latent heat fluxes in a rapidly intensifying marine cyclone. *Mon. Wea. Rev.,* **120,** 1093–1107.
Minnis, P., P. W. Heck, and D. F. Young, 1993: Inference of cirrus cloud properties using satellite-observed visible and infrared radiances. Part II: Verification of theoretical cirrus radiative properties. *J. Atmos. Sci.,* **50,** 1305–1322.
Mishchenko, M. I., W. B. Rossow, A. Macke, and A. A. Lacis, 1996: Sensitivity of cirrus cloud albedo, bidirectional reflectance, and optical thickness retrieval accuracy to ice-particle shape. *J. Geophys. Res.,* **101,** 16 973–16 985.
Oberhuber, J. M., 1988: An atlas based on the coads data set: the budgets of heat, buoyancy and turbulent kinetic energy at the surface of the global ocean. Max-Planck-Insitut für Meteorologie Rep. 15, 20 pp. + figures.
Oort, A. H., and T. H. Vonder Haar, 1976: On the observed annual cycle in the ocean–atmosphere heat balance over the northern hemisphere. *J. Phys. Oceanogr.,* **6,** 781–799.
——, and J. P. Peixoto, 1983: Global angular momentum and energy balance requirements from observations. *Advances in Geophysics,* Vol. 25, Academic Press, 355–490.
Peixoto, J. P., and A. H. Oort, 1992: *Physics of Climate,* AIP, 520 pp.
Rossow, W. B., and R. A. Schiffer, 1991: ISCCP cloud data products. *Bull. Amer. Meteor. Soc.,* **72,** 2–20.
——, and Y.-C. Zhang, 1995: Calculation of surface and top of atmosphere radiative fluxes from physical quantities based on ISCCP data sets, 2. Validation and first results. *J. Geophys. Res.,* **100,** 1167–1197.
——, L. C. Garder, P.-J. Lu, and A. Walker, 1991: International Satellite Cloud Climatology Project (ISCCP) documentation of cloud data. WMO/TD-No. 266 (revised), World Climate Research Programme (ICSU and WMO), Geneva, Switzerland, 76 pp. plus three appendices. [Available from World Meteorological Organization, Avenue Guiseppe-Motta, 1211 Geneva, Switzerland.]
Savijärvi, H. I., 1988: Global energy and moisture budgets from rawinsonde data. *Mon. Wea. Rev.,* **116,** 417–430.
Seager, R., and M. B. Blumenthal, 1994: Modeling tropical Pacific sea surface temperature with satellite-derived solar radiation forcing. *J. Climate,* **7,** 1943–1957.
Sellers, W. D., 1965: *Physical Climatology.* University of Chicago Press, 272 pp.
Trenberth, K. E., and A. Solomon, 1994: The global heat balance: Heat transports in the atmosphere and ocean. *Climate Dyn.,* **10,** 107–134.
Wang, J., and W. B. Rossow, 1995: Determination of cloud vertical structure from upper air observations. *J. Appl. Meteor.,* **34,** 2243–2258.
Weare, B. C., 1989: Uncertainties in estimates of surface heat fluxes derived from marine reports over the tropical and subtropical oceans. *Tellus,* **41A,** 357–370.
Wielicki, B. A., R. D. Cess, M. D. King, D. A. Randall, and E. F. Harrison, 1995: Mission to Planet Earth: Role of clouds and radiation in climate. *Bull. Amer. Meteor. Soc.,* **76,** 2125–2153.
——, B. R. Barkstrom, E. F. Harrison, R. B. Lee, G. L. Smith, and J. R. Cooper, 1996: Clouds and the Earth's Radiant Energy System (CERES): An earth observing system experiment. *Bull. Amer. Meteor. Soc.,* **77,** 853–868.
World Climate Program, 1986: Scientific plan for the World Ocean Circulation Experiment. WMO/TD-No. 122, World Meteorological Organization, Geneva, Switzerland, 83 pp. [Available from World Meteorological Organization, Avenue Guiseppe-Motta, 1211 Geneva, Switzerland.]
Wunsch, C., 1984: An eclectic Atlantic Ocean circulation model. Part I: The meridional flux of heat. *J. Phys. Oceanogr.,* **14,** 1712–1733.
Zhang, Y.-C., W. B. Rossow, and A. A. Lacis, 1995: Calculation of surface and top of atmosphere radiative fluxes from physical quantities based on ISCCP data sets, 1. Method and sensitivity to input data uncertainties. *J. Geophys. Res.,* **100,** 1149–1165.