# Estimating the Meridional Energy Transports in the Atmosphere and Ocean

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ABSTRACT

The poleward energy transports in the atmosphere–ocean system are estimated for the annual mean and the four seasons based on satellite measurements of the net radiation balance at the top of the atmosphere, atmospheric transport data, and atmospheric and oceanic storage data. To eliminate the occurrence of spurious nonzero transports of energy at the north or south poles various types of corrections had to be made, so that the global balances are maintained. This also enabled us to estimate the uncertainties in the procedures used. The uncertainties found are similar to those reported by Hastenrath based on different satellite data sets but using the same correction method.

Finally, oceanic heat transports are computed for the annual mean and seasons. One of the crucial terms in the heat budget, the interseasonal storage of energy in the oceans, is estimated for three different layers, 0–112, 0–275 and 0–550 m, enabling a further error estimate in the inferred oceanic heat transports.

The present results confirm the presence of a strong annual cycle in the transport of energy by the oceans.

## 1. Introduction

Satellite measurements of the radiation flux have made possible global estimates of the energy budget for the earth. These estimates show latitudinal variations requiring large energy transports by the atmosphere–ocean system.

The energy budget for a unit-width latitude band that includes both the atmosphere and oceans can be written as

$$S_A + S_O = F_{TA} - \partial(T_A + T_O)/a\partial\theta, \qquad (1)$$

where $S_A = \partial E_A/\partial t$ is the rate of storage of energy in the atmospheric part of the band, $S_O = \partial E_O/\partial t$ the rate of storage in the oceanic part, $F_{TA}$ the net downward flux of radiation at the top of the atmosphere, $T_A$ and $T_O$ the transports of energy across a latitudinal wall in the atmosphere and oceans, respectively, $a$ the mean radius of the earth, and $\theta$ the latitude. In Eq. (1) we have neglected the storage of energy in the land, snow and ice (see Oort and Vonder Haar, 1976, for further discussion). Integration of (1) over a polar cap (see Fig. 1) yields

$$T_A(\theta) + T_O(\theta) = -\langle F_{TA}\rangle + \langle S_A\rangle + \langle S_O\rangle, \qquad (2)$$

where $\langle(\ \ )\rangle = \int_\theta^{\pi/2} (\ \ )a d\theta'$.

The primary purpose of this paper is to present the results for transports $T_A$ and $T_O$ for the year and the four seasons. Our calculations are based on radiation data from Campbell and Vonder Haar (1980), atmospheric energy storage and flux data from Oort and Peixoto (1983) and Oort (1983), and oceanic heat storage data from Levitus (1982, 1984).

The presently used values are improved over those used by Oort and Vonder Haar (1976) by including more recent measurements and better analysis techniques. For example, in the radiation sample two years of Nimbus 6 data were added and some earlier months with incomplete satellite coverage were deleted. A total of 48 months of satellite data are contained in the present data set. The radiation data were taken from polar orbiting, sun-synchronous satellites which may be affected by a diurnal sampling bias. Campbell and Vonder Haar (1980) made an extensive error analysis. They concluded that for the satellite measurements the present sampling was adequate to determine the relative magnitude of an average beyond a month but that large systematic errors were still possible because of the lack of absolute calibration of the instruments. Their error estimates for net radiation combined the errors estimates for infrared radiation and planetary albedo measurements in ten-degree latitude zones for each

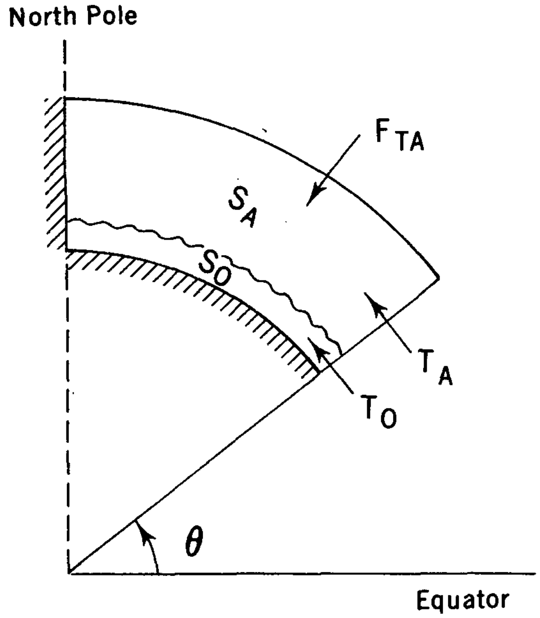


FIG. 1. Principle of the integration over a polar cap (see text).

of the twelve months. They found error estimates ranging from ±2.6 to ±15.6 W $m^{-2}$ for the net radiation with an average value of ±8 W $m^{-2}$. Future satellite experiments are expected to reduce the uncertainty in these measurements.

In earlier studies for the Northern Hemisphere only (Vonder Haar and Oort, 1973; Oort and Vonder Haar, 1976) it was recognized that the accumulation of systematic errors may lead to important uncertainties in the transports computed in the tropics and subtropics. Now that global data sets are available the different terms in the energy balance equation (1) may be computed from pole to pole and the problems become more evident (Oort and Peixoto, 1983).

For example, if we start integration of (1) with zero transport at the north pole, any systematic errors in the measurements will accumulate and eventually produce a spurious meridional transport of energy at the south pole (see Fig. 2, curve A). In order to obtain more reliable transport values we will attempt to remove this error by correcting the flux data so that the transports at both poles vanish. We will use various methods of correction in order to determine the uncertainties involved in this procedure. Similar methods of correction were used before by Hastenrath (1980).

Obviously there are other errors that do not produce a spurious energy transport at the south pole and which will not be corrected by the methods presented here. As a measure for those additional uncertainties we will consider the difference between independent satellite estimates that are in global balance, that is, for which there is no spurious transport computed at the south pole. In the case of the annual radiation budget we will take as independent estimates the transport curves computed by Hastenrath (1982) for various satellite samples but using the same systematic correction. For a further discussion of possible interannual variations in the poleward energy transports see a recent paper by Hastenrath (1984).

The annual case for which $S_A \approx 0$, and probably also $S_O \approx 0$, is presented in Section 2. Usually global balance has been achieved by correcting the net flux of radiation at the top of the atmosphere by the same constant value (in most cases less than 10 W $m^{-2}$) at all latitudes (Campbell and Vonder Haar, 1980; Oort and Peixoto, 1983). We will refer to this approach as the "standard" method.

In the seasonal case, we have the additional problem of cumulative errors in oceanic heat storage (the storage in the atmosphere is an order of magnitude smaller than that in the oceans); this is particularly true over the Southern Hemisphere where data are sparse. This problem will be examined further in Section 3.

To obtain separate estimates of the atmospheric and oceanic transports one can follow at least two approaches. The first is to use oceanic transport data, obtained by surface heat budget calculations (Hastenrath, 1980 and 1982; Talley, 1984) or from hydro-

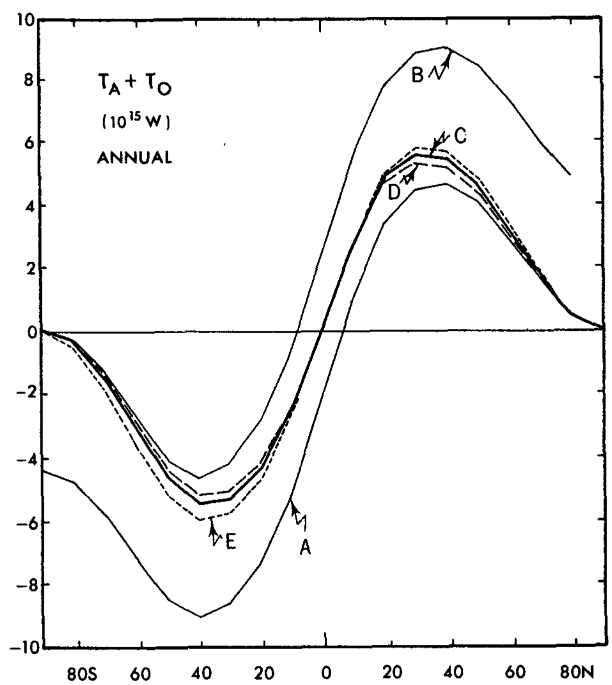


FIG. 2. Total annual transport of energy ($10^{15}$ W); curve A: integration starting at north pole based on uncorrected data, curve B: integration starting at south pole based on uncorrected data, curve C: standard correction [see Eq. (3)], curve D: latitude-dependent correction [see Eq. (4)], curve E: average of curves A and B [see Eq. (5)].

graphic cross sections (Bryden and Hall, 1980) and to compute the atmospheric transport as a residual. However this first approach does not seem very attractive because the oceanic transports are the least reliable since they have to be inferred indirectly using various approximations, whereas the atmospheric transports can be obtained directly from *in situ* observations. The second approach is to compute the oceanic transports as a residual using the measured atmospheric transport values. We will adopt the latter method in the following sections.

## 2. The annual transport

### *a. Methods of correction*

In the annual case, we will simplify Eq. (1) by assuming that $S_A$ and $S_O$ are zero when taken over a "normal" year or, in other words, that interannual variability can be neglected. This is probably a reasonable assumption since in the present case the radiative fluxes and other energy parameters represent composites of measurements made over several years. However, interannual variations, especially in oceanic heat storage, may still affect our results because the years included in the various data sets are different.

As mentioned before, the simplest approach, the standard method, is to apply a constant correction to the original radiation values. In what follows we will denote the uncorrected estimates by starred quantities. A subscript NP for a transport term will indicate that the integration was carried out starting at the north pole, and a subscript SP that it was carried out from the south pole.

If $T_1 \equiv T_{NP}(-\pi/2)$ represents the accumulated total northward transport $T_A + T_O$ at the south pole, the flux correction per unit area is $T_1/(4\pi a^2)$, and the corrected transport at a latitude $\theta$ is given by

$$T(\theta) = T^*_{NP}(\theta) - \int_\theta^{\pi/2} \frac{T_1}{4\pi a^2} 2\pi a^2 \cos\theta d\theta$$

$$= T^*_{NP}(\theta) - \frac{1}{2} T_1(1 - \sin\theta). \quad (3)$$

The result is shown in Fig. 2 as curve C.

Another approach is to consider a correction that varies with latitude, and takes into account the uncertainties in the reflected radiation associated with the diurnal cycle of the albedo, which is most important in the tropics (Saunders *et al.*, 1983). To accomplish this we have chosen a flux correction that affects mainly the low latitudes and is of the form:

$$\frac{T_1 \cos\theta}{-\int_{-\pi/2}^{\pi/2} 2\pi a^2 \cos^2\theta d\theta} = \frac{-T_1 \cos\theta}{\pi^2 a^2}.$$

The corrected transport is now given by

$$T(\theta) = T^*_{NP}(\theta) - \int_\theta^{\pi/2} \frac{T_1}{\pi^2 a^2} 2\pi a^2 \cos^2\theta d\theta$$

$$= T^*_{NP}(\theta) - T_1\left(\frac{1}{2} - \frac{\theta}{\pi} - \frac{\sin 2\theta}{2\pi}\right). \quad (4)$$

The result is shown in Fig. 2 as the dashed curve D. Comparing expressions (3) and (4) we note that

$$\frac{1}{2}(1 - \sin\theta) \geq \left(\frac{1}{2} - \frac{\theta}{\pi} - \frac{\sin 2\theta}{2\pi}\right) \quad \text{for} \quad 0 \leq \theta \leq \frac{\pi}{2},$$

$$\frac{1}{2}(1 - \sin\theta) \leq \left(\frac{1}{2} - \frac{\theta}{\pi} - \frac{\sin 2\theta}{2\pi}\right) \quad \text{for} \quad -\frac{\pi}{2} \leq \theta \leq 0.$$

Since there is a global excess of net incoming radiation in the uncorrected data (i.e., $T_1 < 0$) it is clear that the poleward transports have to be smaller for curve D than for curve C in both hemispheres.

Finally we show in Fig. 2, as a dotted curve E, the weighted average of the two uncorrected (starred) transport curves A and B which were obtained by starting the integration at the north and south poles, respectively.

$$T(\theta) = \left(\frac{1}{2} + \frac{\theta}{\pi}\right)T^*_{NP}(\theta) + \left(\frac{1}{2} - \frac{\theta}{\pi}\right)T^*_{SP}(\theta) \quad (5)$$

This expression reduces at the north pole to

$$T\left(\frac{\pi}{2}\right) = T^*_{NP}\left(\frac{\pi}{2}\right) = 0,$$

at the equator to

$$T(0) = \frac{1}{2} T^*_{NP}(0) + \frac{1}{2} T^*_{SP}(0),$$

and at the south pole to

$$T(-\pi/2) = T^*_{SP}(\pi/2) = 0.$$

It is of interest to note that the midlatitude extremes are the largest for curve E, implying that the corresponding flux corrections are larger in midlatitudes than in the tropics. However, this result depends on the particular shape of the uncorrected curves and cannot be expected to occur in general.

Beside the three methods discussed here there are, of course, many other ways of correcting the data but the present methods appear to be the most reasonable ones.

### *b. Results*

There is excellent agreement in Fig. 2 between the three curves C, D and E at low and high latitudes but there are considerable differences (on the order of $0.5 \times 10^{15}$ W) in midlatitudes. These differences, which represent uncertainties resulting from the corrections for global imbalance, are comparable with the spread in the total transport curves obtained by Hastenrath (1982), that are reproduced in Fig. 3. Hastenrath's curves are based on a variety of satellite data sets but were corrected for global imbalance using the same standard method as was used in our curve C. We excluded the values of Jacobowitz *et al.* (1979) which are very different from all other estimates. As mentioned in the introduction we consider the spread between Hastenrath's curves as a measure of the additional uncertainty independent from the corrections for global imbalance. Therefore the total uncertainty in the transport of energy by the atmosphere and the ocean amounts to about $1 \times 10^{15}$ W in midlatitudes and somewhat less in the tropics and high latitudes.

Using published estimates of the atmospheric transports (Oort and Peixoto, 1983) we are now able to deduce the oceanic transports $T_O$, corresponding with the three total transport curves obtained before. The results for $T_O$ are presented in Fig. 4c, while the

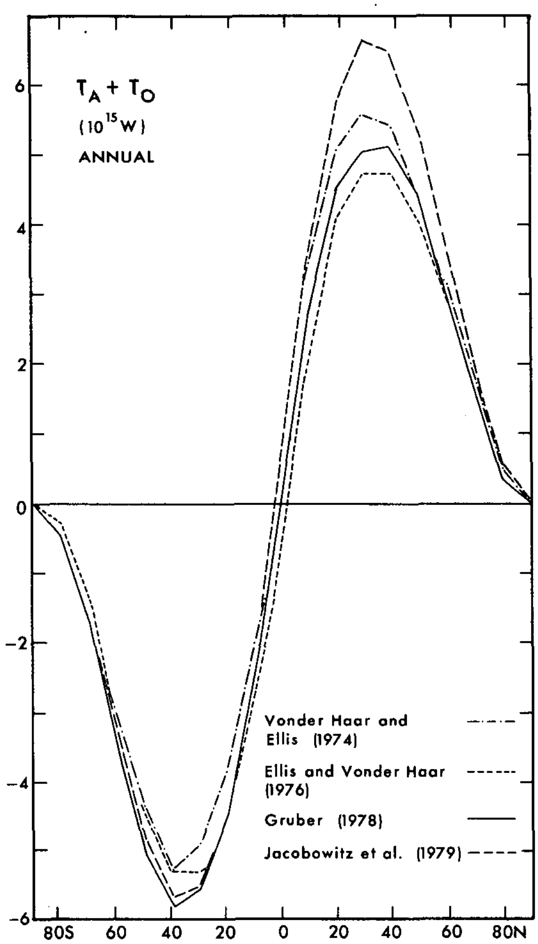


FIG. 3. Various estimates of the total annual transport of energy based on satellite data using a constant correction (from Hastenrath, 1982).

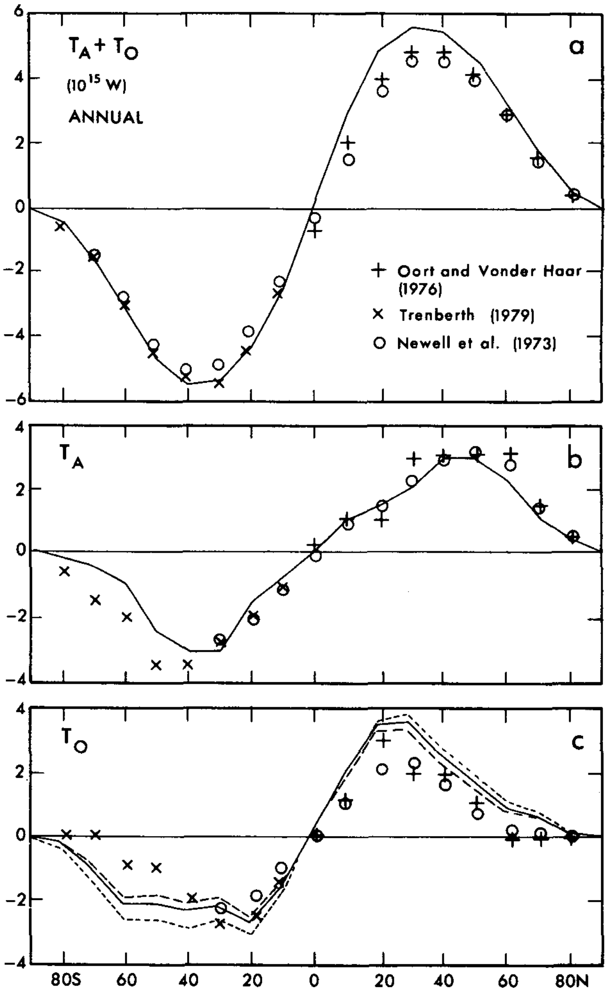


FIG. 4. Annual transport of energy for (a) total transport (as in Fig. 2, curve C), (b) atmospheric transport and (c) oceanic transport computed as a residual. In (c) the three curves for $T_O$ are based on different corrections for $T_A + T_O$ as shown in Fig. 2 by curves C (solid), D (dashed) and E (dotted). Results obtained by some previous investigators are added for comparison.

standard total transport, $T_A + T_O$, is given in Fig. 4a and the atmospheric transport ($T_A$) in Fig. 4b. For comparison, earlier estimates of the various transport terms by Newell *et al.* (1972), Oort and Vonder Haar (1976) and Trenberth (1979) are also presented in Fig. 4. Based on the differences for $T_A$ of about 0.5 $\times 10^{15}$ W shown in Fig. 4b together with our earlier estimate of an uncertainty in $T_A + T_O$ of $1 \times 10^{15}$ W, we derive a value of $1.5 \times 10^{15}$ W for the uncertainty in the inferred oceanic heat transport in midlatitudes.

In Fig. 4 we see that the total transport curve $T_A + T_O$ is approximately symmetric with respect to the equator reflecting the symmetry that is also found in the radiation budget. We notice a weak northward transport of energy across the equator.

The poleward transport in the atmosphere between 55 and 70° latitude in our data is stronger in the Northern than in the Southern Hemisphere. For example at 60° a difference of $1.3 \times 10^{15}$ W is found. This may be associated with a lack of standing eddy transports in the Southern Hemisphere. We will see later that most of this difference occurs during the Northern Hemisphere winter season. Near 30° latitude the poleward transport in the atmosphere is weaker in the Northern Hemisphere by about $1.0 \times 10^{15}$ W, leading to an apparent shift in the maximum of $T_A$ from 40–50° latitude in the Northern Hemisphere to 30–40° latitude in the Southern Hemisphere.

The oceanic transport has a pronounced maximum of $\sim 3.5 \times 10^{15}$ W between 20 and 30°N and a very broad but less intense maximum of $\sim 2.2 \times 10^{15}$ W between 20 and 60°S. Consequently we find convergence of heat due to ocean heat transport in the Northern Hemisphere between 30 and 80°N and in the Southern Hemisphere only between 60 and 80°S. To show the differences between the Northern and Southern Hemispheres better, we present in Fig. 5a, as a function of latitude, the poleward heat transport curves for the two hemispheres superposed. Qualitatively similar features are found in the numerical model experiments of Bryan and Lewis (1979) and Meehl *et al.* (1982), shown in Figs. 5b and c. On the other hand, Hastenrath's (1982) results in Fig. 5d show smaller differences between the two hemispheres.

We confirm Trenberth's (1979) finding of an appreciable poleward transport of heat in the region of the Antarctic Circumpolar Current near 60°S. At these latitudes the two numerical experiments cited above suggest that the poleward energy transport due to transient eddies and mixing dominates the equatorward heat flux associated with Ekman transport caused by the westerly winds.

Finally we must emphasize that it is difficult to explain why our ocean values are much larger than other independent estimates. For example at 25°N our oceanic heat transport was estimated to be $3.5 \times 10^{15}$ W whereas Bryan and Lewis found a value of $1 \times 10^{15}$ W, Meehl *et al.* (1982) $1.8 \times 10^{15}$ W and Hastenrath (1982) $2.2 \times 10^{15}$ W.

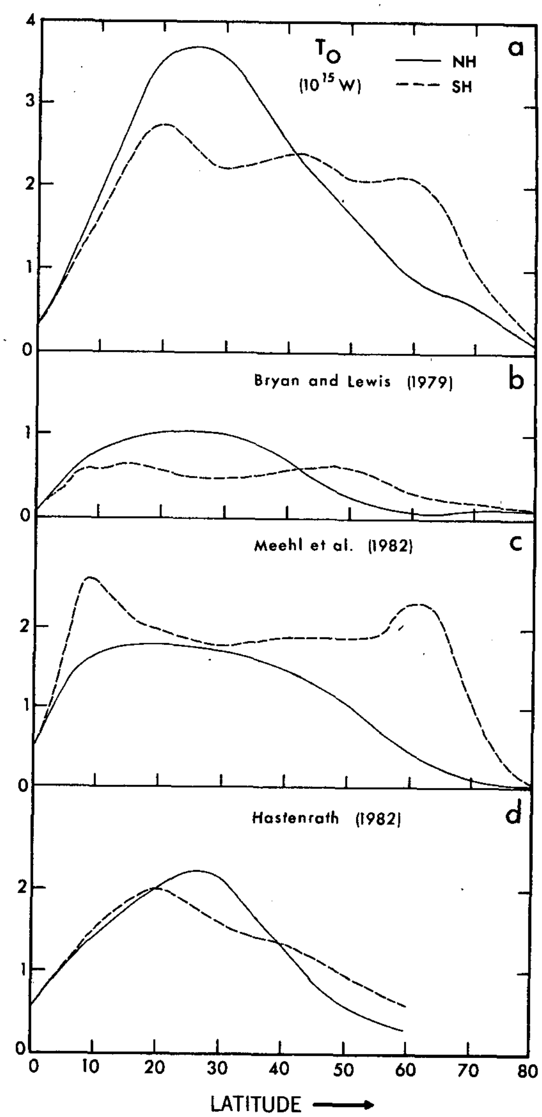


FIG. 5. Oceanic annual poleward transport of heat ($10^{15}$ W): (a) present study, (b) after Bryan and Lewis (1979), (c) after Meehl *et al.* (1982) and (d) after Hastenrath (1982).

## 3. The seasonal transports

### *a. Method of correction*

Uncertainties now come from the radiation flux ($F_{TA}$) the oceanic rate of heat storage ($S_O$) and to a much smaller extent from the atmospheric rate of storage ($S_A$). Because of the large uncertainties in the oceanic storage in the Southern Hemisphere we will start the integration at the north pole.

In order to be consistent with the earlier annual computations of $T_A + T_O$ and also to ensure global annual radiation balance we have first corrected all seasonal net radiation flux values for annual mean imbalance using the standard method discussed above. After applying this correction we still compute a spurious, non-zero value for the transport at the south pole because 1) the annual correction is not entirely appropriate to use for the seasonal flux values and 2) there is an additional imbalance due to the oceanic storage term, $S_O$ ($S_A$ being usually much smaller). We have decided to correct the data for the remaining imbalance only in the Southern Hemisphere since the oceanic and atmospheric measurements are especially sparse in that hemisphere . This procedure also has the advantage of avoiding contamination of the Northern Hemisphere values by unreliable Southern Hemisphere data.

In mathematical terms, the procedure we have used can be expressed as follows:

$$T(\theta) = T^*_{\rm NP}(\theta) - \frac{1}{2}T_1(1 - \sin\theta) \quad \text{for} \quad 0 \leq \theta \leq \frac{\pi}{2}, \tag{6a}$$

$$T(\theta) = T^*_{\rm NP}(\theta) - \frac{1}{2}T_1(1 - \sin\theta) + T_2 \sin\theta \quad \text{for} \quad -\frac{\pi}{2} \leq \theta \leq 0, \tag{6b}$$

where

$$\left.\begin{aligned} T_1 &= T^{*{\rm ann}}_{\rm NP}\left(-\frac{\pi}{2}\right) \\ T_2 &= T^*_{\rm NP}\left(-\frac{\pi}{2}\right) - T_1 \end{aligned}\right\}$$

and the asterisks again indicate that the transports are computed from the raw, uncorrected data. A superscript "ann" indicates annual-mean values. We note in (6b) that at the south pole the corrected transports reduce to zero, as they should:

$$T\left(-\frac{\pi}{2}\right) = T^*_{\rm NP}\left(-\frac{\pi}{2}\right) - T_1 - T_2 = 0.$$

There is an additional uncertainty regarding to what depth one chooses to integrate the oceanic storage values. If we choose a too shallow layer we may miss an important fraction of the seasonal signal but, on the other hand, for a deep layer the measurements become less reliable and we may thereby introduce even larger errors (Levitus, 1984). To show the dependence on depth we will present the results for $S_O$ for three layers between the surface and the depths 112,225 and 550 m (a value of 275 m was used before by Oort and Vonder Haar, 1976, Ellis *et al.*, 1978, and Levitus, 1984).

*b. Results*

Figure 6 shows the total atmospheric plus oceanic transports for the northern winter (December–February, denoted by DJF) and for the northern summer (June–August, denoted by JJA) seasons obtained directly from the raw data without applying any corrections. Obviously there is a large imbalance at the South Pole. Figures 7a, 8a, 9a and 10a show, at a different scale, the total transport values, $T_A + T_O$, for the four seasons after correction using the method (6) indicated above.

As before in the annual case, we have used previously published seasonal data for $T_A$ (Oort and Peixoto, 1983; Oort, 1983) shown in Figs. 7b, 8b, 9b and 10b to compute the oceanic transports ($T_O$) as a residual. The results are presented in Figs. 7c, 8c, 9c and 10c.

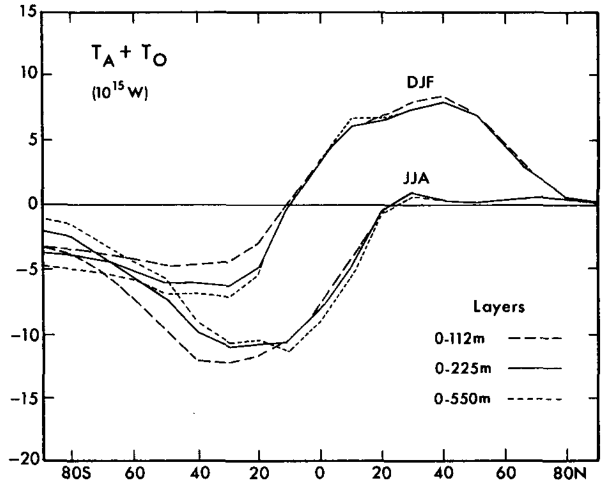


FIG. 6. Uncorrected total seasonal transports for December–February (DJF) and June–August (JJA), clearly showing the effects of the global imbalance at the south pole. The dashed, solid and dotted lines show the transports computed using oceanic storage values for the layers 0–112, 0–225 and 0–550 m, respectively.

We, first of all, notice in Figs. 7c through 10c that the oceanic transport in the 112 m deep layer is very different from the transports in the two other layers indicating that the 112 m layer is probably too shallow. The other two transport curves, for the 225 and 550 m deep layers, are in close agreement with each other.

The oceanic transport exhibits a strong similarity between the two hemispheres in the corresponding seasons, especially during winter. The transport is maximum, on the order of $6 \times 10^{15}$ W, at 25–30° latitude in the winter hemisphere, and is fairly weak in the summer hemisphere except in the inner tropics between 10°S and 10°N, where we find a very strong transport of heat toward the winter pole. There is a strong convergence of heat poleward of 35° in the winter hemisphere, and a strong divergence of heat in the tropics between 20°S and 20°N throughout the year.

We find at the equator that $T_O$ is subject to large seasonal variations (approximately $4 \times 10^{15}$ W into the winter hemisphere), twice as large as in the atmosphere (compare Figs. 7c and 9c with Figs. 7b and 9b). According to the numerical model results of Bryan and Lewis (1979) the primary mechanism for this cross-equatorial flow of heat appears to be Ekman transport as further discussed below. It is of interest to pursue the comparisons with the numerical ocean model of Bryan and Lewis (1979) and Bryan (1982a,b).

To emphasize the seasonal variations Bryan and Lewis (1979) computed the January–July difference in ocean heat transport. We have superposed their results for the two hemispheres in the form of winter–summer differences of the poleward heat transport in

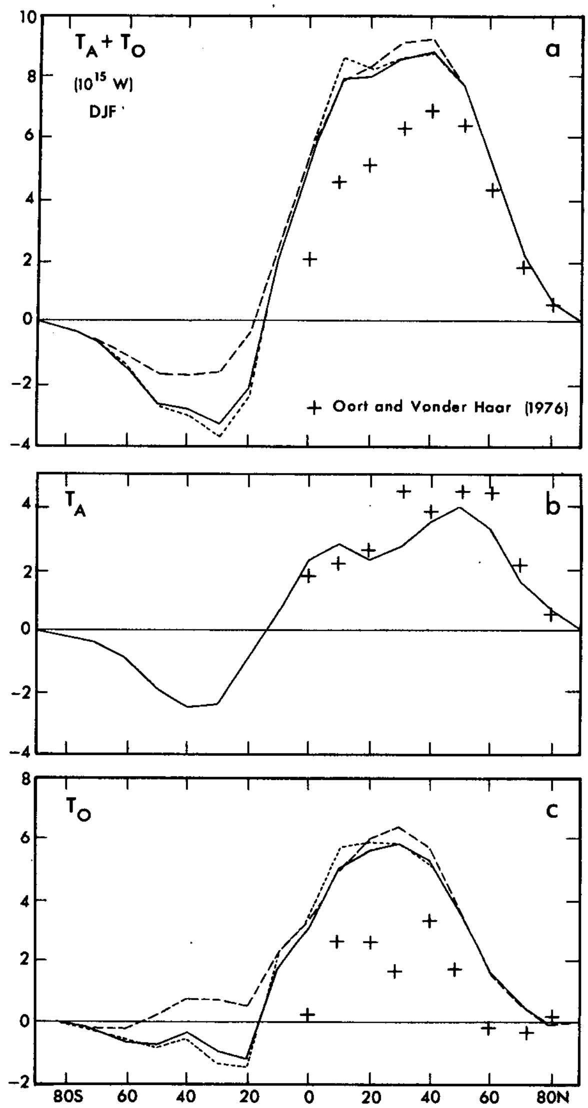


FIG. 7. Energy transports for December–February: (a) total transport, (b) atmospheric transport (from Oort and Peixoto, 1983) and (c) residual oceanic transport. In (a) and (c) the dashed, solid and dotted lines show the transports computed using oceanic storage values for the layers 0–112, 0–225, 0–550 m, respectively. Values obtained by Oort and Vonder Haar (1976) are plotted for comparison.

Fig. 11b in order to contrast the Northern and Southern Hemispheres. These curves can be compared with similar curves based on our present values in Fig. 11a.

In the model, almost the entire seasonal variation is due to changes in the meridional overturning associated with variations in the Ekman transport. This is attested by the similarity between heat transport and zonal wind stress divided by the Coriolis parameter, shown in Bryan (1982a).

Our results (Fig. 11a) are quite different from the model results, first, quantitatively by roughly a factor of two, and in midlatitudes also qualitatively because the poleward transport there is greater during winter than summer. It is an interesting unresolved issue why these differences occur. A definitive answer will probably require better observations, better numerical simulations and more detailed comparisons. Here we will only mention Bryan's (1982a) conclusion that the strength of the meridional overturning in his model was only 50% of the observed one. Furthermore, Bryan and Lewis (1979) found that their model results systematically led the observations by 1.5 months. This may be an important factor in view of the large seasonal variation in oceanic heat transport. To show this we have added in Fig. 11a, as dashed curves, the computed differences between the observed March–May (MAM) and September–November (SON) transport which are drastically different from the DJF–JJA differences. There are nevertheless some similarities between our data and the model integra-

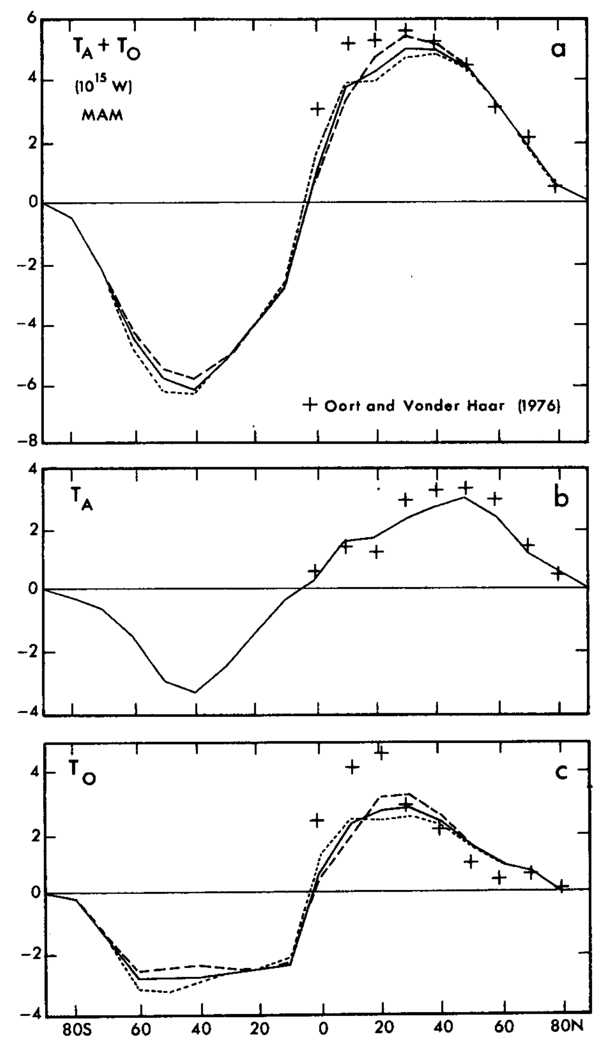


FIG. 8. As in Fig. 7 but for March–May.

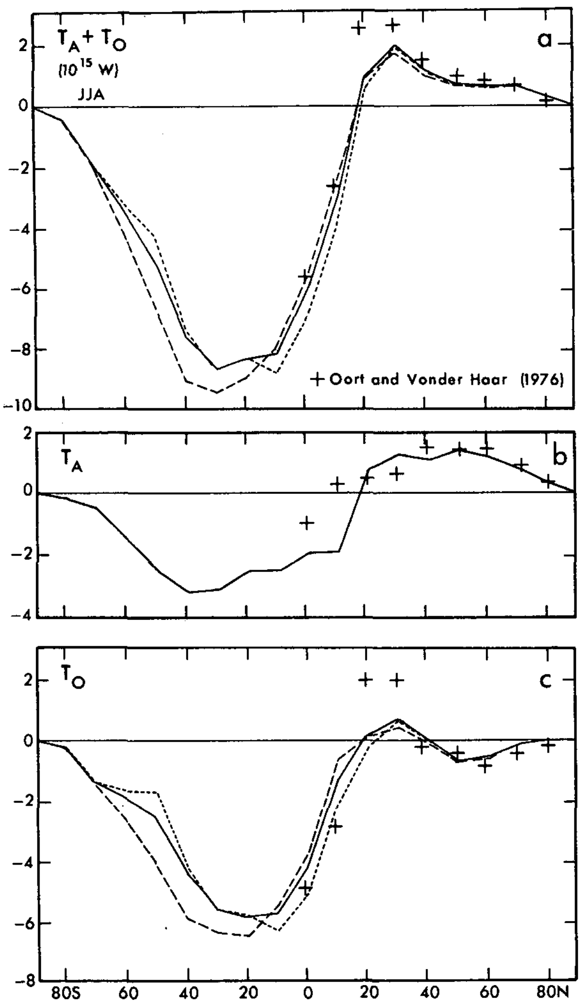


FIG. 9. As in Fig. 7 but for June–August.

tions regarding the winter-summer differences in the tropics. For example in both results the seasonal differences in the tropics appear to be larger in the Southern than in the Northern Hemisphere.

Coming back to the seasonal transport curves in Figs. 7 through 10, one notices some significant differences, especially during winter, between the present results and those reported earlier in 1976 for the Northern Hemisphere as shown by the plus symbols. When writing the previous paper we did not realize how important it was to apply consistent global corrections to the raw data. One of the reasons was that we only considered the Northern Hemisphere, and therefore did not encounter the problem of nonvanishing transports at the south pole. The differences with the earlier analyses are due mainly to differences in correcting for the observed (but spurious) annual mean excess of net radiation (see curve A in Fig. 2), and for the global heat storage in the atmosphere, oceans, snow and ice (see Fig. 6 of Ellis *et al.,* 1978; Levitus, 1984). However, we believe that the present estimates are more consistent and realistic than the previous ones. As can be seen from the uncorrected curves in Fig. 6 larger corrections had to be applied to the DJF than to the JJA curves, perhaps explaining the larger discrepancies between the new and the old results during the DJF season. In summary, we find a stronger annual cycle in $T_A + T_O$ than in the 1976 study as shown Figs. 7 through 10 (see also numerical values in Table 1).

When comparing the atmospheric transport values from Oort and Peixoto (1983) and Oort and Vonder Haar (1976) we find substantial differences near 30 and 60°N in the annual mean (Fig. 4b) and the DJF (Fig. 7b) and SON (Fig. 10b) values, the older values being larger by about $2 \times 10^{15}$ W. These differences are due to the use of more recent 1963–73 versus 1958–63 data, the use of indirectly computed mean meridional velocities (see Oort and Peixoto, 1983,

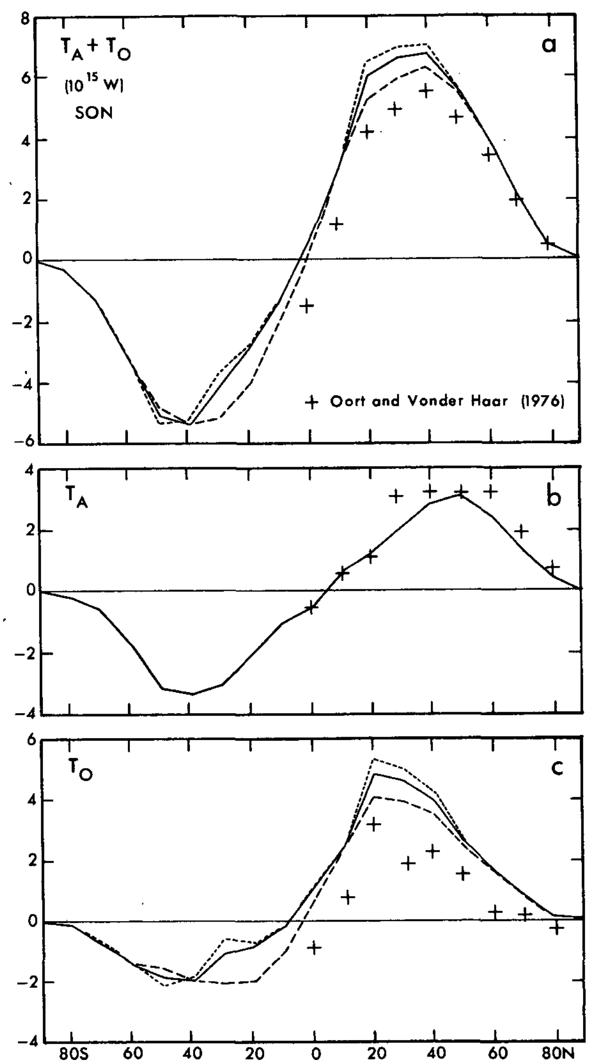


FIG. 10. As in Fig. 7 but for September–November.

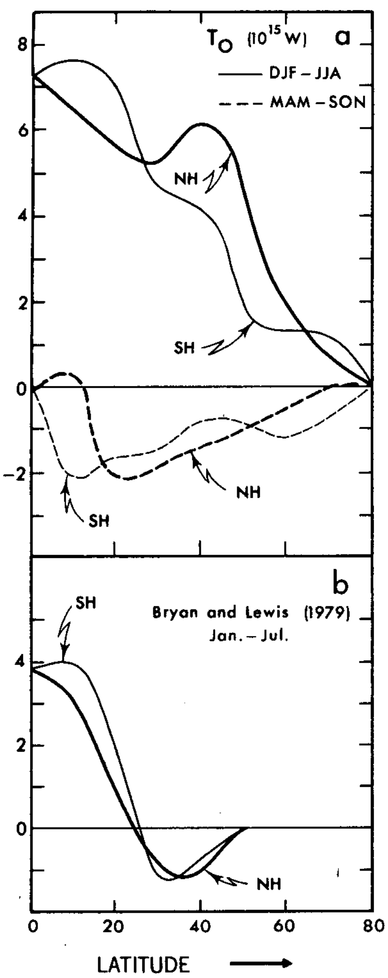


FIG. 11. Seasonal differences in poleward oceanic energy transport ($10^{15}$ W): (a) full curves, DJF–JJA; dashed curves, MAM–SON; and (b) full curves, January–July from Bryan and Lewis (1979).

Appendix), and some modifications in the analysis techniques.

Of course, the differences in the total transport and in the atmospheric transport curves are accumulated and more apparent in the residual, oceanic transport curves ($T_O$) in Figs. 7c through 10c. In spite of the large uncertainties in these oceanic transport values, it is clear that our data show the need for a very large seasonal variation in the oceanic heat transports.

Finally, we confirm that there exist large and unexplained discrepancies between the oceanic heat transports computed using the planetary-energy-balance method (Oort and Vonder Haar, 1976) and those based on oceanographic considerations (Hall and Bryden, 1982).

## 4. Conclusions

This study presents estimates of the meridional transport of energy by the atmosphere and the ocean from pole to pole, for the four seasons and for the year. These estimates were made using the planetary energy-balance method, similar to the method used by Oort and Vonder Haar (1976) for the Northern Hemisphere only, but with data based on more recent measurements.

Our main conclusions are:

1) The annual mean oceanic transport shows a substantial asymmetry between the two hemispheres. This asymmetry must be related to the differences in physiography of the more land-covered Northern and ocean-covered Southern Hemispheres. The heating due to convergence of oceanic heat transport in midlatitudes is much larger in the Northern than in the Southern Hemisphere, whereas the reverse holds at high latitudes. Qualitatively similar effects are also found in numerical simulations of the world ocean.

2) We find large seasonal variations in the oceanic heat transport. In the tropics the poleward transports are larger in winter than in summer for both observations and numerical models. However, in midlatitudes, the present observations show a stronger poleward heat transport and the numerical experiments a weaker one. This last model result is due to the increased equatorward Ekman transport associated with stronger westerlies in winter.

Previously Oort and Vonder Haar (1976) included in their error estimates the accumulation of systematic errors which lead to large uncertainties in the tropics. In a global study, systematic errors may accumulate and produce unphysical energy transports at the poles which must be corrected. Here we have used different methods of correction to evaluate the uncertainties.

We estimate the uncertainty in the total transport ($T_A + T_O$) for the annual case to reach a maximum value of $1 \times 10^{15}$ W in midlatitudes, taking into account the uncertainties both due to the correction of the systematic error and due to additional errors that do not affect the global balance.

For the atmospheric transport ($T_A$) we can use the spread between various independent estimates as a measure of uncertainty. This leads to an uncertainty in $T_A$ of $0.5 \times 10^{15}$ W in midlatitudes. The uncertainty in the residual oceanic transport ($T_O$) is therefore on the order of $1.5 \times 10^{15}$ W in midlatitudes and somewhat less in the tropics.

In the seasonal computations we make use of the oceanic heat storage which is poorly known. A tentative estimate of the uncertainty in the oceanic heat transport is 1.5–2 ($\times 10^{15}$) W based on our analyses using storage values for different layer depths.

Finally we must repeat that our analysis relies heavily on the assumption that the interannual variability is small compared with the seasonal variations. In doing so we are considering our data as being representative of a long-term climatology. The validity of this hypothesis clearly requires further testing which in turn requires more accuracy and better

TABLE 1. Northward transports of energy in the total atmosphere–ocean system ($T_A + T_O$), and in the atmosphere ($T_A$) and oceans ($T_O$) computed in the present study for the four seasons and the year in units of $10^{15}$ W.

| | DJF | | | MAM | | | JJA | | | SON | | | Year | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Latitude | $T_A + T_O$ | $T_A$ | $T_O$ | $T_A + T_O$ | $T_A$ | $T_O$ | $T_A + T_O$ | $T_A$ | $T_O$ | $T_A + T_O$ | $T_A$ | $T_O$ | $T_A + T_O$ | $T_A$ | $T_O$ |
| North | | | | | | | | | | | | | | | |
| 80 | 0.6 | 0.7 | −0.1 | 0.5 | 0.5 | 0.0 | 0.3 | 0.3 | 0.0 | 0.4 | 0.4 | 0.0 | 0.5 | 0.4 | 0.1 |
| 70 | 2.2 | 1.6 | 0.6 | 1.8 | 1.1 | 0.7 | 0.7 | 0.8 | −0.1 | 2.0 | 1.3 | 0.7 | 1.7 | 1.1 | 0.6 |
| 60 | 4.9 | 3.3 | 1.6 | 3.2 | 2.3 | 0.9 | 0.7 | 1.2 | −0.5 | 4.0 | 2.4 | 1.6 | 3.2 | 2.3 | 0.9 |
| 50 | 7.7 | 4.0 | 3.7 | 4.4 | 2.9 | 1.5 | 0.8 | 1.4 | −0.6 | 5.6 | 3.1 | 2.5 | 4.6 | 2.9 | 1.7 |
| 40 | 8.8 | 3.5 | 5.3 | 5.0 | 2.6 | 2.4 | 1.2 | 1.1 | 0.1 | 6.7 | 2.8 | 3.9 | 5.4 | 2.9 | 2.5 |
| 30 | 8.5 | 2.7 | 5.8 | 5.0 | 2.2 | 2.8 | 2.0 | 1.3 | 0.7 | 6.6 | 2.0 | 4.6 | 5.6 | 2.0 | 3.6 |
| 20 | 7.9 | 2.3 | 5.6 | 4.3 | 1.6 | 2.7 | 1.0 | 0.8 | 0.2 | 6.0 | 1.2 | 4.8 | 4.9 | 1.4 | 3.5 |
| 10 | 7.9 | 2.8 | 5.1 | 3.8 | 1.5 | 2.3 | −3.2 | −1.9 | −1.3 | 3.1 | 0.6 | 2.5 | 2.9 | 1.0 | 1.9 |
| 0 | 5.4 | 2.3 | 3.1 | 1.0 | 0.2 | 0.8 | −6.1 | −1.9 | −4.2 | 0.5 | −0.6 | 1.1 | 0.3 | 0.0 | 0.3 |
| South | | | | | | | | | | | | | | | |
| 10 | 2.4 | 0.6 | 1.8 | −2.8 | −0.4 | −2.4 | −8.2 | −2.5 | −5.7 | −1.3 | −1.1 | −0.2 | −2.4 | −0.8 | −1.6 |
| 20 | −2.1 | −0.9 | −1.2 | −4.0 | −1.5 | −2.5 | −8.3 | −2.5 | −5.8 | −2.9 | −2.1 | −0.8 | −4.3 | −1.6 | −2.7 |
| 30 | −3.3 | −2.4 | −0.9 | −5.2 | −2.6 | −2.7 | −8.6 | −3.1 | −5.5 | −4.2 | −3.1 | −1.1 | −5.3 | −3.1 | −2.2 |
| 40 | −2.8 | −2.5 | −0.3 | −6.2 | −3.4 | −2.8 | −7.6 | −3.2 | −4.4 | −5.4 | −3.4 | −2.0 | −5.5 | −3.1 | −2.4 |
| 50 | −2.7 | −1.9 | −0.8 | −5.8 | −3.0 | −2.8 | −5.1 | −2.5 | −2.6 | −5.1 | −3.2 | −1.9 | −4.6 | −2.5 | −2.1 |
| 60 | −1.6 | −0.9 | −0.7 | −4.4 | −1.6 | −2.8 | −3.4 | −1.5 | −1.9 | −3.1 | −1.7 | −1.4 | −3.1 | −1.0 | −2.1 |
| 70 | −0.6 | −0.4 | −0.2 | −2.2 | −0.7 | −1.5 | −1.9 | −0.5 | −1.4 | −1.3 | −0.6 | −0.7 | −1.5 | −0.5 | −1.0 |
| 80 | −0.2 | −0.2 | 0.0 | −0.5 | −0.3 | −0.2 | −0.5 | −0.2 | −0.3 | −0.3 | −0.2 | −0.1 | −0.4 | −0.2 | −0.2 |

spatial and temporal coverage in the measurements. In our future research we intend to study the radiational forcing and the atmospheric transports for individual years in order also to learn more about possible year-to-year fluctuations in the oceanic heat transport, which must be of great importance for the climate.

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