Meridional Heat Transport in the Pacific Ocean

L. D. TALLEY[1]

School of Oceanography, Oregon State University, Corvallis, OR 97331

(Manuscript received 9 June 1983, in final form 29 September 1983)

ABSTRACT

The heat transported meridionally in the Pacific Ocean is calculated from the surface heat budgets of Clark and Weare and others; both budgets were based on Bunker's method with different radiation formulas. The meridional heat transport is also calculated from the surface heat budget of Esbensen and Kushnir, who used Budyko's method. The heat transport is southward at most latitudes if the numbers of Clark and of Weare are used. It is northward in the North Pacific and southward in the South Pacific if Esbensen and Kushnir's numbers are used. Systematic errors in both calculations appear to be so large that confident determination of even the sign of the heat transport in the North Pacific is not possible. The amount of heat transported poleward by all oceans is obtained from the Pacific Ocean calculation and transports in the Atlantic and Indian Oceans based on Bunker's surface heat fluxes.

1. Introduction

The role of the ocean in transporting heat meridionally has been examined by many investigators using a variety of methods to estimate its importance relative to atmospheric heat transport. The usual reason given for determining the meridional heat transport is that it is a vital component of the climatic balance, the smoother of inequitable incoming solar radiation. If the ocean heat transport is well-determined, it can be used as a test of the validity of numerical, global climate models. The transport of heat by the oceans, both meridionally and zonally, is also important for problems of more direct oceanographic interest—it is related to the general circulation in the ocean basin and is important in water mass transformation and maintenance of the *T–S* relation (Stommel and Csanady, 1979). For oceanographic purposes, heat transport estimates for each ocean basin are more useful than estimates averaged over all oceans.

Hastenrath (1982) and Hall and Bryden (1982), among others, have reviewed the various methods for estimating meridional heat transport. The present calculation uses the net surface heating, determined from ship observations, to compute the meridional heat transport in the Pacific Ocean from 60°N to 40°S. Previous estimates of Pacific heat transport using the same method were made by Bryan (1962), using Sverdrup's (1957), Albrecht's (1960) and Budyko's (1956) surface heat fluxes, by Emig (1967), by Hastenrath (1980), and by Wyrtki (1965). Both Emig and Hastenrath used Budyko et al.'s (1962) surface heat fluxes. Wyrtki made a separate surface heat flux calculation and reported meridional heat transport at 6°N and at the equator: meridional heat transports at all latitudes were calculated by Wyrtki (private communication, 1983) and are presented here. Direct estimates of Pacific Ocean heat transport using hydrographic sections have been made by Bryan (1962) at 32°N, by Wunsch *et al.* (1983) and Bennett (1978) at 28°S and 43°S and by Georgi and Toole (1982) for the zonal transport in the Antarctic circumpolar region.

The present indirect estimate of heat transport was inspired by the success of a similar calculation in the North Atlantic, using Bunker and Worthington's (1976) surface heat fluxes (Hall and Bryden, 1982). As stated by Bunker (1976), the existence of a much larger set of ship observations and more up-to-date bulk formulas for computing the surface heat flux compared with those used by Budyko *et al.* (1962), makes a new calculation of meridional heat transport of interest. Clark (private communication, 1983), and Weare *et al.* (1981) computed the surface heat flux in the Pacific Ocean from 20°N to 60°N and 40°S to 30°N respectively, basically using Bunker's (1976) method with different radiation formulas. Both groups used individual ship observations obtained from the Fleet Numerical Weather Central and Southwest Fisheries Center. Weare *et al.* (1981) computed surface fluxes for 1957–76 while Clark computed the fluxes for 1947–79.

Esbensen and Kushnir (1981) computed surface heat fluxes with Budyko's (1974) method, using climatological means of measured variables to calculate heat fluxes rather than individual observations. Because they used updated climatological data from the National Climatic Center, the basic data set is more similar to the data used by Clark and Weare *et al.* than the data

[1] Present affiliation: Scripps Institution of Oceanography, La Jolla, CA 92093.

used by Budyko *et al.* (1962) and Budyko (1974). Meridional heat transports calculated from Esbensen and Kushnir's surface heat fluxes may thus provide a better reference point in comparing the transports based on Bunker-type calculations with those based on Budyko-type calculations. All three groups, Clark, Weare *et al.*, and Esbensen and Kushnir, were kind enough to supply their results for the present calculation of meridional heat transport.

2. Surface heat budget in the Pacific

The net heat flux from the atmosphere into the ocean is

$$Q = Q_I - Q_B - Q_L - Q_S,$$

where Q_I is the net incoming radiation, Q_B the net back radiation, Q_L the latent heat flux and Q_S the sensible heat flux. In the absence of direct measurements, all terms are computed from commonly-measured quantities using bulk parameterizations which are based on comparisons of direct heat flux measurements with the commonly-measured variables. There is not yet agreement on which parameterizations are the best; estimates of errors in the bulk formulas have been made by various authors and are summarized in this paper.

The method used by Weare *et al.* (1981) and Clark for calculating the surface heat flux is described by the latter. The formulas they (hereafter referred to as CW) used are listed in Table 1. CW used Seckel and Beaudry's (1973) formula for insolation, based on List's (1958) observations, with Reed's (1977) cloudiness correction and Payne's (1972) albedos. For net back radiation, CW used Berliand and Berliand's formula (Budyko, 1974). Clark used a constant cloudiness correction of $(1 - 0.8C)$, where C is the cloudiness in tenths, as suggested by Simpson and Paulson (1979) for midlatitudes. Weare *et al.* corrected for cloudiness depending on cloud type, following Laevastu (1967). If cloud type was not reported, they used a cloudiness correction of $(1 - 0.75C)$. CW computed latent and sensible heat fluxes using Bunker's (1976) transfer coefficients, which depend on wind speed and atmospheric stability. Clark computed the surface specific humidity using vapor pressures over salt water while Weare *et al.* used vapor pressures for fresh water. This means that Clark's latent heat fluxes are about 5% less than those of Weare *et al.* (Weare, personal communication, 1983). Lastly, it is important to note that CW follow Bunker (1976) in computing the heat fluxes for each observation before averaging in time and space. They averaged the individual fluxes by month and 5° square. Then climatological means were found for each month and finally, the twelve monthly values were summed to yield the net heat exchange.

Esbensen and Kushnir (1981), hereafter referred to as EK, followed Budyko's (1974) method with two minor differences. Their formulas are also listed in Table 1. They used Berliand's formula, described by Budyko (1974), for incoming solar radiation. The effect of clouds is included in the formula and Payne's (1972) tables were used for albedos. EK used Berliand and Berliand's formula for net back radiation with a correction for cloudiness suggested by Budyko [(1974), Eq. (2.13)]. EK's calculation differs from CW's in ex-

TABLE 1. Flux formulas.*

	Clark (1982); Weare *et al.* (1981)	Esbensen and Kushnir (1981)
Insolation	$A_0 + A_1 \cos\phi + B_1 \sin\phi + A_2 \cos 2\phi + B_2 \sin 2\phi$ (Seckel and Beaudry, 1973; Reed, 1977)	Budyko (1974, Table 3)
Cloud correction	$1 - 0.62C + 0.0019\alpha$ (Reed, 1977)	$1 - (a + 0.38C)C$ a depends on latitude (Budyko, 1974)
Albedo	Payne (1972)	Payne (1972)
Back radiation	$\epsilon\sigma T_s^4 (0.39 - 0.05e^{1/2}) + 4\epsilon\sigma T_s^3 (T_s - T_a)$	$\epsilon\sigma T_a^4 (0.39 - 0.05e^{1/2})(1 - dC^2) + 4\epsilon\sigma T_a^3(T_s - T_a)$ d depends on latitude (Berliand and Berliand, 1952; Budyko, 1974)
Cloud correction	$(1 - 0.8C)$ for 20°N to 60°N (Simpson and Paulson, 1979) $(1 - bC)$ for 40°S to 20°N b depends on cloud type (Laevastu, 1967)	already included
Latent heat flux	$L\rho_a C_E(q_s - q_a)\lvert\vec{u}\rvert$	$L\rho_a C_L(q_s - q_a)\lvert\vec{u}\rvert$
Sensible heat flux	$c_p\rho_a C_E(T_s - T_a)\lvert\vec{u}\rvert$ C_E from Bunker (1976)	$c_p\rho_a C_E(T_s - T_a)\lvert\vec{u}\rvert$ C_L and C_E from Liu *et al.* (1979)

* C cloudiness in tenths; T_s sea surface temperature; ρ_a air density; c_p specific heat of air; α noon solar altitude; T_a air temperature; q_s saturation specific humidity; $\phi = (t - 21)(360/365)$ where t is the time of year in days; ϵ emissivity of water; e vapor pressure; q_a specific humidity at 10 m; σ Stefan–Boltzmann constant; L latent heat of vaporization; $\lvert\vec{u}\rvert$ wind speed at 10 m.

changing T_a and T_s, the air and sea temperatures. Latent and sensible heat fluxes were calculated using transfer coefficients of Liu *et al.* (1979) which are lower than Bunker's (1976) coefficients at all wind speeds. EK followed Budyko's method, using monthly averages of temperature, humidity, wind speed and cloudiness to calculate heat fluxes.

Net surface heating for the Pacific Ocean, as computed by Weare *et al.* (1981) and Clark is shown in Fig. 1. Because results of Clark and Weare *et al.* for net heating differ greatly in the overlap band from 20°–30°N, the two sets are contoured separately. Clark's ocean heat gain is about 30 W m^{-2} higher than that of Weare *et al.* throughout the band of overlap, largely due to systematic discrepancies in insolation and latent heat flux. These differences are discussed in Section 4. The net result is that the ocean gains heat from 20°–30°N according to Clark's calculation and loses heat according to Weare *et al.*

Weare *et al.* (1981) compare the net heat fluxes from 40°S to 20°N (Fig. 1) with those of Budyko (1963), Wyrtki (1965) and Hastenrath and Lamb (1978), so I will not discuss this region here. North of 20°N, using Clark's calculation, there is large heat loss in the Kuroshio gyre and large heat gain along the eastern boundary, as found in all previous calculations of net heating. Clark's calculation shows a broad region of small heat loss in the central subtropical Pacific, as do the calculations of Weare *et al.*, Budyko (1974), Wyrtki (1965) and Esbensen and Kushnir (1981). Clark shows small, locally insignificant heat gain elsewhere in midocean. Wyrtki's distribution is similar, but Budyko and Esbensen and Kushnir (EK) find small heat loss throughout the midocean. The hatched regions of Fig. 1, the Sea of Okhotsk and the Bering Sea, were excluded by Wyrtki and EK because of lack of data; Clark shows large heat gain in these regions. This is highly suspect and is probably due to a seasonal bias in the obser-

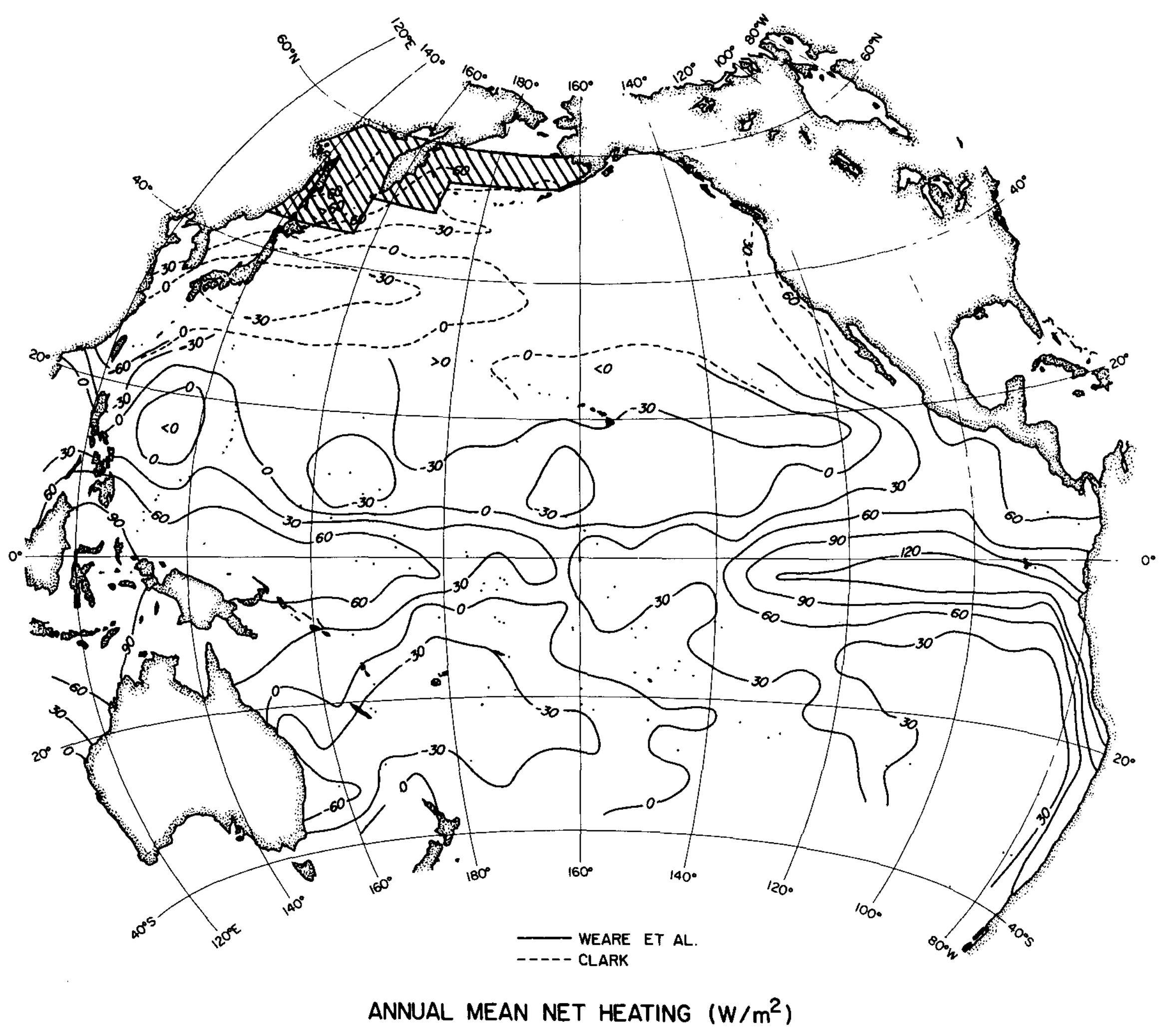

FIG. 1. Annual mean net surface heating (W m^{-2}). Data are from N. Clark for 20–60°N and from Weare *et al.* (1981) for 40°S–30°N. The hatched region is the region excluded by Wyrtki (1965) because of sparse data.

vations: it is difficult to imagine that there is net annual heat gain in these regions.

3. Meridional heat transport

a. Pacific Ocean

The net heat fluxes calculated by CW and EK were integrated from north to south to produce the meridional heat transports graphed in Fig. 2a. There are two curves for CW: the first includes the large heat gains in the Sea of Okhotsk and the Bering Sea and the second excludes all data in those areas excluded by Wyrtki (1965, shaded in Fig. 1). Values of net heating from Weare *et al.* (1981) are used in the band 20°–30°N. Meridional boundaries for the integrations were 120°E and the eastern boundary, in accordance with the divisions made by Hastenrath (1980). Emig (1967) used a variable western boundary in order to include the Indonesian archipelago in the Pacific Ocean.

Assumptions made in integrating CW's and EK's heat fluxes were 1) the heat transport across 60°N is negligible and 2) the heat transport through the Indonesian archipelago is negligible. The first assumption is good: Aagaard and Greisman (1975) estimated a northward transport of 0.2×10^{13} W through Bering Strait at about 65°N. It is unlikely that the heat transport across 60°N is much larger since the deep basin of the Bering Sea terminates just north of 60°N, so the strong northward current on the shelf break turns westward and then southward along the western boundary just north of 60°N (Hughes *et al.*, 1972): the net heat transported by these currents is probably small. An error of even 5×10^{13} W at 60°N would not be very important here since, as we will see, systematic errors in heat flux parameterization lead to much larger errors in heat transport.

Heat transport due to flow through the Indonesian archipelago may not be so small. Godfrey and Golding (1981) estimate that the volume transport $\bar{V}$ westward through the archipelago is about 10 Sv (1 Sv = 10^6 m^3 s^{-1}), mostly confined to the upper 400 m between 9° and 14°S. From the *Oceanographic Atlas of the Indian Ocean Expedition* (Wyrtki *et al.*, 1971), the average temperature $\bar{T}$ of the upper 400 m is about 18°C. An upper bound on the heat transport due to flow of this water into the Indian Ocean is obtained if it is assumed that the water returns to the Pacific in the Antarctic Circumpolar Current at a temperature of about 3°C (Georgi and Toole, 1982). With a temperature difference of 15°C, the heat transport across 110°E is $C_p\rho\bar{V}\Delta\bar{T} = 0.6 \times 10^{15}$ W, a significant fraction of the heat transport across 10° and 20°S. Of course, the returning flow may actually be much warmer: if it returns south of Australia but north of the Antarctic Circumpolar Current and in the upper part of the water column, its temperature could be between 10° and 15°C, with corresponding heat transports of 0.3 and 0.1×10^{15} W. The transport through the Indonesian archipelago may actually be much smaller or distributed below 400 m (Wyrtki, personal communication, 1983). Because this component of the heat transport is not well-defined, it will not be taken into account.

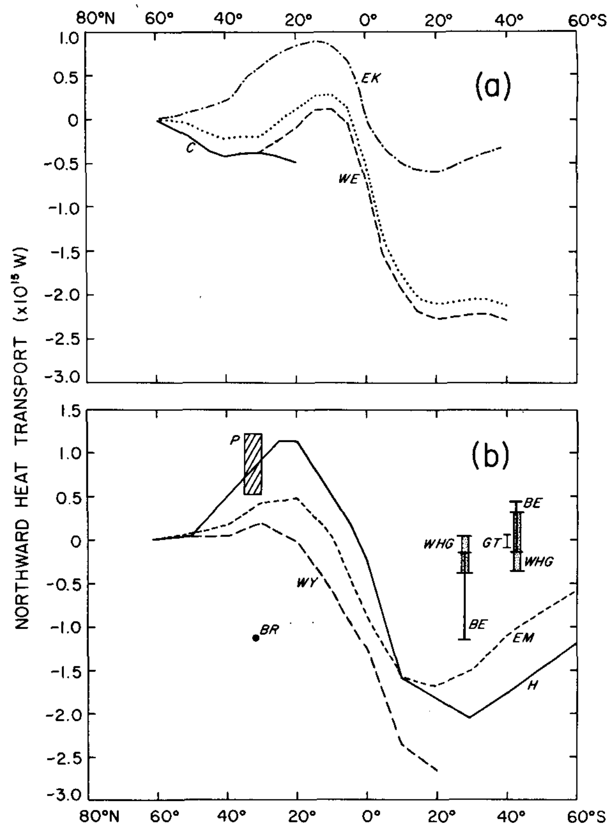

FIG. 2. Annual mean meridional heat transport in the Pacific Ocean. (a) Computed from surface heat fluxes of Clark and Weare *et al.* (solid, dash and dot) and Esbensen and Kushnir (dot-dash). The heat transport at 60°N is assumed to be zero. The solid curve is based on Clark's heat fluxes. The long dashes are based on heat fluxes of Weare *et al.*: integration south of 30°N is based on Weare *et al.* The middle dotted curve deletes the heat fluxes in the hatched region of Fig. 1 (see text). (b) Meridional heat transports based on surface heat fluxes of Emig (EM) Hastenrath (H) and Wyrtki (WY). These heat transports are not corrected for heat transport into the Indian Ocean. Also shown are heat transports from Georgi and Toole (GT) and Pullen, and three direct estimates of heat transport: Bryan (BR), Bennett (BE) and Wunsch *et al.* (WHG).

The main features of the heat transport based on calculations of Clark and Weare *et al.* are:

1) southward heat transport at almost all latitudes;

2) a plateau in heat transport from 25°N to 40°N, resulting from near-cancellation of large heat losses in the Kuroshio gyre and heat gains in the eastern, coastal-upwelling areas;

3) large southward heat transport in the Southern Hemisphere, due to large heat gains in the eastern Tropical Pacific and along the coast of South America.

Meridional heat transports calculated by other researchers are shown in Fig. 2b for comparison with the present calculations. The initial impression is of great scatter, although some order is discernible if indirect and direct estimates are separated. Indirect estimates based on surface heat fluxes have been computed by Emig (1967), Hastenrath (1980) and Wyrtki (personal communication, 1983). Pullen (1977) and Georgi and Toole (1982) used a combination of direct and indirect methods. Direct methods were used by Bryan (1962), Bennett (1978) and Wunsch *et al.* (1983). The direct estimates vary in magnitude and some in sign from the indirect estimates.

Emig's (1967) calculation was based on her own chart of net heat flux obtained from the component heat flux charts of Budyko *et al.* (1962). I have used her tabulated values of heat flux and area to construct the curve shown in Fig. 2b. Hastenrath (1980) used net surface heat fluxes from Budyko (1963, 1974) for latitudes 30–66°N and 20–60°S and from Wyrtki (1965) and Hastenrath and Lamb (1978) for 30°N–20°S. Wyrtki used the surface heat fluxes shown in Wyrtki (1965) in which transport values at 6°N and the equator were given. While the general shape of all curves but CW is the same due to net gain in the tropics and loss at higher latitudes, there is wide variation in extreme values and in the latitude of zero-crossing. The equatorial transport was negative for all calculations but EK.

Pullen's (1977) study focused on variations in meridional heat transport in the band 30–35°N; hence the wide range of values shown in Fig. 2b. The mean value of heat transport which can be calculated from her tables (1.02×10^{15} W) is strongly dependent on an initial indirect estimate of the heat transport which was derived from Budyko *et al.* (1962) and Vonder Haar and Oort (1973). If instead, Wyrtki's estimate of 1.96×10^{14} W had been used, the mean temperature of the return flow would have been 15°C and the mean heat transport would have been 1.97×10^{14} W.

Bryan's (1962) direct estimate of heat transport across 32°N was based on a hydrographic section and the Sverdrup transport estimated from wind-stress curl and is very different from transports calculated from surface heat budgets. He found 1.17×10^{15} W of southward transport, which might have been as small as 0.82×10^{15} W if errors due to the Sea of Japan had been included. There may also be errors due to unknown, eddy-driven, barotropic circulations, eddy heat fluxes and seasonal bias (the section was made in August). The latter two would decrease the southward transport. Since the section is at the edge of an eddy-rich region, since the wind stress data available at the time was not comparable to what is presently available and since the systematic and random errors in the indirect estimates at 32°N are not very large (Section 4), the indirect estimates are probably more reliable.

This is not the case in the Southern Hemisphere where Bennett (1978), Wunsch *et al.* (1983) and Georgi and Toole (1982) have estimated heat transport at 28, 43 and 40°S. Bennett produced rather wide ranges of heat transports, dependent on the assumed widths of the western boundary current. He also estimated errors due to eddy fluxes which were not resolved by the section: his error estimates can be interpreted to be 0.25×10^{15} W if ill-defined barotropic eddy fluxes are ignored and 1.03×10^{15} W if a rough estimate of their magnitude is included. Wunsch *et al.* claim that the errors are relatively small, 0.22×10^{15} W at 28°S and 0.35×10^{15} W at 43°S. Georgi and Toole (1982) determined the meridional heat transport across 40°S from direct estimates of heat transported into the Pacific by the Antarctic Circumpolar Current and surface heat fluxes south of 40°S. Their 40°S heat transport is very small; its sign depends on assumptions about the barotropic component of the Antarctic Circumpolar Current. Since the indirect portion of their estimate depends on northward integration of surface heat fluxes over a relatively small region, from Antarctica to 40°S, the error is much smaller than error in the completely indirect estimate which was integrated southward from the northern North Pacific.

All three groups estimate much lower values of heat transport than the indirect methods and may even show northward heat transport at 43°S. Eddy heat flux may be the cause of the differences. However, the cumulative effect of random and, especially, systematic errors in the integration of surface heat fluxes from 60°N to 30 and 40°S is likely to be the real culprit. These errors are discussed in Section 4 where it is shown that the direct estimates are within the error bounds of the indirect estimates.

b. Global

The meridional heat transport estimated from CW's surface fluxes in the Pacific Ocean is added to heat transports in the Atlantic and Indian Oceans to yield the total amount of heat transported meridionally by the ocean. Hall and Bryden (1982) computed the meridional heat transport in the Atlantic Ocean based on Bunker and Worthington's (1976) and Bunker's (1980) surface heat fluxes for the North and South Atlantic Oceans, respectively. Bryden (personal communication, 1983) provided the heat transports for the Indian Ocean based on Bunker's surface heat fluxes, which are available at the Woods Hole Oceanographic Institution. The meridional heat transports in the Pacific, Atlantic and Indian Oceans and in all oceans are listed in Table 2 and shown in Fig. 3. (In the Pacific, the transport calculation which excludes the Sea of Okhotsk and the Bering Sea was used here.) The total heat transport can be compared with Oort and Vonder Haar's (1976) and Trenberth's (1979) estimates of ocean heat transport based on measurements of net radiation at the top of the atmosphere and atmospheric heat transport: their results are also shown. Hastenrath (1982) discusses difficulties with the satellites radiation

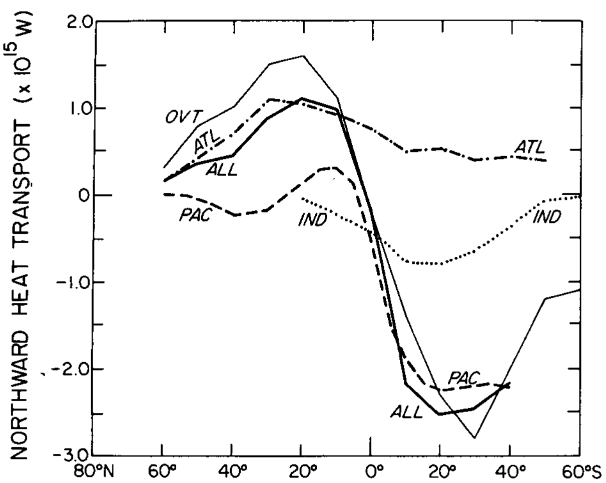

FIG. 3. Annual mean meridional heat transported by all oceans, all based on surface heat fluxes calculated using Bunker's method. The Atlantic Ocean heat fluxes were calculated by Hall and Bryden (1982). Oort and Vonder Haar's (1976) and Trenberth's (1979) ocean heat transports for the Northern and Southern Hemisphere, respectively, are shown for comparison (OVT).

measurements and uses a composite of four data sets for a calculation of atmospheric transport; such a composite or the Stephens *et al.* (1981) data could also be used for computing ocean heat transport.

The net ocean heat transport calculated from the surface heat budgets of Clark, Weare *et al.* and Bunker is fairly asymmetric with a maximum poleward heat transport of 1.1×10^{15} W in the Northern Hemisphere and a maximum poleward heat transport of 2.5×10^{15} W in the Southern Hemisphere. The estimates from Oort and Vonder Haar (1976) and Trenberth (1979) are slightly more symmetric: the maximum poleward heat transport in the Northern Hemisphere is 1.6 $\times 10^{15}$ W and in the Southern Hemisphere, 2.8×10^{15} W. Thus there is actually fair agreement between the ocean transport calculated from the surface heat budget and the transport calculated from the radiation balance and atmospheric transport. The maximum magnitude of the ocean transport calculated from the surface budgets is smaller in both hemispheres, so that, say, a 1×10^{15} W increase in northward transport at 20°N in the Pacific would actually degrade the agreement between the two estimates of transport. However, since the various calculations of heat transported by the North Pacific disagree in both sign and magnitude, it is necessary to examine the accuracy of the estimates, as done in Section 4.

The atmospheric heat transport can be estimated as the difference between ocean heat transport and net heat transport determined from satellite radiation measurements. Hastenrath (1982) used an average of four satellite data sets which were each uniformly corrected to yield zero heat transport at the poles. The Stephens *et al.* (1981) data are very similar to Hastenrath's composite and are used here: their data has an average imbalance of 9 W m^{-2} which leads to a 4.6×10^{15} W imbalance in heat transport when integrated over the globe. An adjustment of 9 W m^{-2} is applied here at all latitudes. The resulting heat transports are listed in Table 2 and shown in Fig. 4. The ocean heat transport is then subtracted from the net to yield the atmospheric heat transport, also in Table 2 and Fig. 4.

TABLE 2. Net northward heat transport in the oceans, atmosphere and total ($\times 10^{15}$ W). In the Pacific, Weare *et al.* (1981) and Clark's surface heat budgets were used for 40°S–30°N and 30°N–60°N, respectively, omitting the Sea of Okhotsk and the Bering Sea. The Atlantic Ocean transports were computed by Hall and Bryden (1982). The Indian Ocean transports were calculated from surface heat fluxes from Bunker's files at the Woods Hole Oceanographic Institution (Bryden, personal communication, 1983). In the column labeled OVHT are the ocean heat transports estimated by Oort and Vonder Haar (1976) and Trenberth (1979) for the Northern and Southern Hemispheres, respectively. Net transports were calculated from net heat fluxes of Stephens *et al.* (1981). Atmospheric transports were calculated from the net heat transport and total ocean heat transport.

	Oceans					Atmosphere	Net
Latitude	Pacific	Indian	Atlantic	Total	OVHT		
80°N	—	—	—	—	—	—	0.5
70°N	—	—	—	—	—	—	1.7
60°N	—	—	0.16	0.16	0.3	3.1	3.2
50°N	−0.05	—	0.42	0.37	0.8	4.3	4.6
40°N	−0.23	—	0.69	0.46	1.0	5.0	5.5
30°N	−0.19	—	1.07	0.88	1.5	4.7	5.6
20°N	0.13	−0.05	1.03	1.11	1.6	3.8	5.0
10°N	0.31	−0.24	0.92	0.99	1.1	2.0	3.0
0°	−0.53	−0.43	0.75	−0.21	−0.2	0.4	0.2
10°S	−1.88	−0.78	0.49	−2.17	−1.4	−0.4	−2.5
20°S	−2.26	−0.80	0.52	−2.54	−2.3	−1.8	−4.4
30°S	−2.20	−0.65	0.39	−2.46	−2.8	−2.9	−5.3
40°S	−2.22	−0.36	0.43	−2.15	−2.0	−3.3	−5.4
50°S	—	−0.06	0.40	—	−1.2	—	−4.7
60°S	—	—	—	—	—	—	−3.1
70°S	—	—	—	—	—	—	−1.5
80°S	—	—	—	—	—	—	−0.4

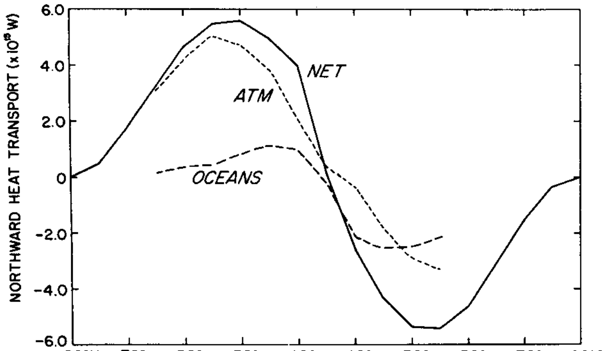

FIG. 4. Northward heat transported by the oceans, the atmosphere and the combined ocean-atmosphere. Sources were Fig. 3 for the oceans and the Stephens *et al.* (1981) zonal mean net flux from satellite measurements.

4. Errors

It is unsettling that even the sign of the meridional heat transport in the Pacific is in doubt. Unfortunately, systematic errors in net heat flux for each 5° square can be no larger than about 5 W m^{-2} if the minimum acceptable error in heat transport at 20°N is, say, 0.3 $\times$ 10^{15} W. To even determine the sign of the heat transport at 20°N, the systematic error would have to be less than 10 W m^{-2}. This contrasts with the situation in the North Atlantic Ocean, where systematic errors must be less than 15 W m^{-2}, in order to produce acceptable heat transport numbers at 20°N, emphasizing the great difference in size of the Pacific and Atlantic Oceans and the more vigorous heat exchange at high latitudes (subarctic region) in the North Atlantic. Of course, random errors can be, and are, much larger than systematic errors. Weare *et al.* (1981) estimate a random error of about 50 W m^{-2} for net heating over a 5° square. The maximum error in heat transport at each latitude due to this random error is $\epsilon A N^{1/2}$, where ϵ = 50 W m^{-2}, A is the area of a 5° square and N is the total number of 5° squares contributing to the transport. This results in an error of $\pm$0.10 $\times$ 10^{15} W at 25°N, $\pm$0.17 $\times$ 10^{15} W at the equator and $\pm$0.25 $\times$ 10^{15} W at 40°S. If all errors were random and of this order, the heat transport calculation would be acceptable.

However, the error is not all random. The systematic errors calculated by various observers for each flux term are summarized in Table 3 for both the CW and the EK calculations. Let us discuss each term briefly.

Weare *et al.* and Clark used Seckel and Beaudry's (1973) clear-sky insolation formula with Reed's (1977) cloud correction formula and Payne's (1972) albedo. Reed (1977, 1982) found that these insolation and cloud correction formulas underestimate observations by 2% and 2.6%, respectively, thus about 5% in total. On the other hand, Simpson and Paulson (1979) found that the formulas overestimated incoming radiation by 6%. EK used Berliand's formula for insolation and cloudiness correction (Budyko, 1974). Reed (1977) found that Berliand's clear-sky insolation formula was 3 to 18% too high (in summer and winter, respectively). The cloudiness factor, Q_I/Q_0, where Q_0 is the clear-sky insolation, also has errors. We can roughly estimate from Reed's (1977) figures that Berliand's cloudiness factor is 20 to 25% too low. Simpson and Paulson found that Seckel and Beaudry's clear-sky insolation, together with Berliand's cloudiness factor, underestimated Q_I by 17%: using Reed's finding of 2% error in clear-sky insolation, one notes that this implies that Berliand's cloudiness factor is 19% too low. The total error in Budyko's Q_I using Reed's figures of 3–18% and 25% is an 11–23% underestimate. The total error in Budyko's Q_I using Reed's 3–18% and Simpson and Paulson's 19% is a 6–17% underestimate. Smith and Dobson (1983) found that Budyko's clear-sky insolation agreed with observations but that the cloud formula underestimates observations by 29%. Reed (1977) and Simpson and Paulson (1979) agree that an insolation and cloud correction formulation due to Lumb (1964) is the most reliable; however, the necessary information on cloud type is often not available in routine observations. Smith and Dobson (1984) have proposed new formulas for incoming radiation: their estimates of mean flux at Ocean Weather Stations Bravo and Papa are about 29% higher than estimates based on Budyko's (1974) formulas.

The back radiation formulas used by CW and EK are nearly identical except that EK use air temperature rather than sea-surface temperature and apply the cloud correction to the first term only. The latter affects the result but little since the second term is small.

TABLE 3. Average values of heat flux components (W m^{-2}) for 20–30°N and percent systematic underestimate of observations.

	Clark	Weare *et al.*	EK
Q_I	201	239	184
Reed (1976)	5%	5%	11 to 23%
Simpson and Paulson (1979)	−6%	−6%	(6 to 17%)*
Smith and Dobson (1983)			29%
Q_B	37	43	56
Reed and Halpern (1975)	−16%	−16%	−19%
Simpson and Paulson (1979)	1%	1%	(−2%)
Q_L	138	199	129
Large and Pond (1982) } Smith and Dobson (1983) }	−16%	−21%	−8%
Q_S	16	19	10
Large and Pond (1982)	−21%	−21%	(7%)
Smith and Dobson (1983)	−32%	−32%	0%
Q_{total}	9	−22	−11

* Values in parentheses were derived from other results listed in the table and were not taken from the source papers: see text.

Simpson and Paulson (1979) found that the formula used by CW for clear-sky longwave radiation underestimated observations by an average of 8% or 1% if a weighted average was used. They did not list systematic errors for cloudy conditions. The use of air temperature rather than sea-surface temperature in the formula used by EK means that they overestimate clear-sky longwave radiation by about 2%. Reed and Halpern (1975), in contrast to Simpson and Paulson, found that the Berliand formulation overestimated the clear-sky back radiation by about 16%, and is even worse when a nonlinear cloud factor is used. Because of the contradictory results of the studies, it is difficult to decide whether CW's and EK's back-radiation estimates are low or high. Clearly further work is needed to estimate errors in the back radiation formulas to determine which, if any, of the presently used parameterizations, using the type of data recorded by ships of opportunity, is acceptable.

To calculate latent heat flux, CW used the usual bulk parameterization with Bunker's (1976) transfer coefficients which were empirically derived from the then-available observations. EK used transfer coefficients derived by Liu *et al.* (1979) which have both observational and theoretical bases. Bunker's values at neutral stability increase at high wind speeds; in addition, he increased all coefficients by 10% for use with ship observations because of possible, systematic differences between latent heat fluxes based on typical observations and those made in more controlled circumstances. Values of Liu *et al.* decrease at high wind speed and do not incorporate any correction for ship observations. Recent work by Large and Pond (1982), Smith and Dobson (1984) and Anderson and Smith (1980) indicates that there may be an increase in transfer coefficient with increased wind speed. The latent heat flux coefficient also does not have great dependence on stability, although the sensible heat flux coefficient does. Smith and Dobson's coefficient at neutral stability is about 1.13×10^{-3}, in excellent agreement with Large and Pond's value of 1.15×10^{-3}. Smith and Dobson find that Bunker's adjusted transfer coefficients are 13 to 40% higher than theirs for wind speeds of 2 to 27 m s^{-1}, respectively. At 5 m s^{-1}, the difference was 16%. Neither Large and Pond nor Smith and Dobson agreed with Bunker's upward adjustment of 10%. Liu *et al.* determine an average transfer coefficient of 1.2–1.3 ($\times 10^{-13}$) over all wind speeds at neutral stability. This is somewhat higher than Large and Pond's and Smith and Dobson's values, but well within the range given by many others (cf. Anderson and Smith).

There are a number of additional problems with latent heat flux calculations. According to Large and Pond (1982), the same transfer coefficients do not apply to all situations, implying that some physics has been left out. Smith and Dobson (1984) note that anemometers are not placed at uniform height and are generally considerably higher than 10 m; if coefficients appropriate for 10 m winds are used, the flux will be overestimated. They take this into account in determining the coefficients. The use of mixing ratio and specific humidity is not clearly differentiated in the literature and could lead to an error of a few percent. The vapor pressures used in calculating specific humidity differ over salt and fresh-water: one difference between calculations of Weare *et al.* (1981) and Clark is that the latter used fresh-water vapor pressure so their latent heat fluxes should be 5 to 20% high, using typical values of saturation humidity and humidity differences from Large and Pond.

The sensible heat flux is a small part of the total heat budget on average, but error in its calculation can also be estimated. CW followed Bunker and used the same transfer coefficients as for latent heat flux, which means that their values are too large since the transfer coefficients for sensible heat are lower by a factor of 1.1 (Large and Pond, 1982) or 1.2 (Smith and Dobson, 1984) and since they incorporate Bunker's 10% adjustment.

Sampling biases, editing procedures and averaging methods also affect the net heat flux. Weare and Strub (1981) discussed sampling biases. Procedures for editing the large ship-of-opportunity data base differ and may be partially responsible for differences between the results of Weare *et al.* (1981) and Clark discussed below. Averaging methods also differ: Bunker (1976) piloted the more accurate method of determining best fluxes from each individual observation before averaging. CW followed this method while EK calculated heat fluxes from averaged ship observations; since the heat fluxes are nonlinear in temperature, cloudiness and windspeed, this type of averaging may affect the results. However, Esbensen and Reynolds (1981) found that latent heat fluxes calculated from averaged data using transfer coefficients of Liu *et al.* were at most 5% higher than fluxes calculated from individual observations and the same transfer coefficients. (There is an overestimate rather than an underestimate because there is a small decrease in the transfer coefficients of the former with wind speed.) They did not consider the effects of averaging on the radiation terms.

An additional source of difficulty in making the net meridional heat transport calculation for the North Pacific may be a large seasonal cycle: Bryan (1982) showed that the seasonal heat transports in the North Pacific in a general circulation model were much larger than the net heat transport and that the meridional heat transport changed sign from winter to summer in the tropics. The annual meridional heat transport was nearly zero. In some ways this is in agreement with the two calculations presented here, particularly that based on CW where the transport in the North Pacific is of rather small magnitude.

The total systematic error for any calculation of net heat flux is difficult to determine since there are many different estimates of error and since it is undoubtedly spatially-dependent. Accordingly, three error estimates for Clark, Weare *et al.* and EK are made based on the maximum and minimum net fluxes allowed by the errors listed in Table 3 and based on a personally chosen set of errors. Of course, future observations may yield different errors and perhaps even wider ranges of possible heat fluxes. Average values of each component in the latitude band 20–30°N as computed by each of the three groups are also listed in Table 3. Clark's net flux is 9 W m^{-2}; the minimum and maximum fluxes calculated from the possible errors listed are -1 W m^{-2} and 49 W m^{-2}. The average net heat flux from Weare *et al.* is -22 W m^{-2}; the minimum and maximum possible fluxes are -36 W m^{-2} and 37 W m^{-2}. Similarly, EK's net flux is -11 W m^{-2} with minimum and maximum fluxes of -11 W m^{-2} and 103 W m^{-2}. (The last, large range of possible error is due to the large Smith and Dobson correction to Q_I.) One could choose a given set of errors, based on the amount of data used to estimate the error, and recalculate the error. A personal choice would be Reed's (1977) error for Q_I since a large number of stations at many different latitudes were used, Simpson and Paulson's weighted-average error for Q_B, the listed error for Q_L and Large and Pond's error for Q_S. The resulting net fluxes for Clark, Weare *et al.* and EK would be 43 W m^{-2}, 29 W m^{-2} and 22 W m^{-2}. Thus the systematic errors in total heat flux would be 34 W m^{-2}, 51 W m^{-2} and 33 W m^{-2}; all are positive and would tend to make the meridional heat transport even more southward.

The average differences in each heat flux component between Clark and EK are not very different from those calculated from the systematic errors, except for back radiation (whose error is poorly determined). However, there are large systematic differences in calculations of Q_I and Q_L of Clark's and Weare *et al.* which cannot be easily explained. The differences in calculations were in vapor pressure used for the latent heat flux (5–10% difference) and the cloudiness correction to back radiation. Two remaining possibilities are 1) different averaging periods (Clark used 1947–79 while Weare *et al.* used 1957–76) and 2) different editing procedures. We can dismiss the first possibility since there was no large climatological shift between 1947 and 1977. In Fig. 5 the annual net heat flux from Clark's calculations is plotted for a representative 5° square. The average heat flux is 4.6 W m^{-2} for the entire period and -8.7 W m^{-2} for 1957–76, while Weare *et al.* obtained -40.4 W m^{-2} for 1957–76. Thus, editing procedures remain as the sole possibility, but it is difficult to imagine that such large systematic differences could result. Without repeating the calculations, it is not possible to explain the differences at this point.

The differences between the individual terms in the calculations of Weare *et al.* and EK are the largest of all, but the net heat fluxes are in closest agreement.

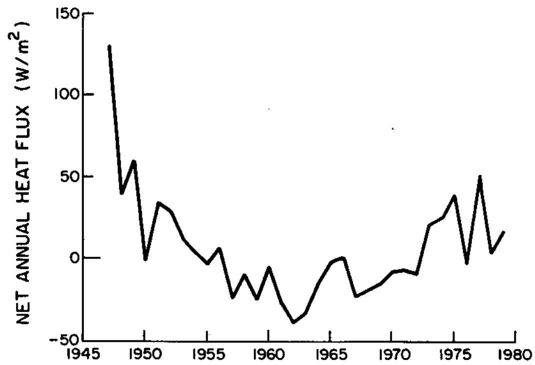

FIG. 5. Annual net heat flux (W m^{-2}) for the area 175°E–180°, 20–25°N, as calculated by Clark. The mean net heat flux is 4.6 W m^{-2} for the entire period and -8.7 W m^{-2} for 1957–76. Weare *et al.* (1981) obtain -40.4 W m^{-2} for 1957–76.

Note that the average differences in almost every term are larger than 5 W m^{-2}, which is the approximate accuracy needed for the heat transport in the North Pacific. The total differences are larger than the accuracy required for heat transport.

The foregoing exercises really serve only to point out the possible magnitude of systematic error in the net heat flux: it is probably safe to say that meridional heat transports calculated for the Pacific Ocean from surface heat fluxes are invalid since the systematic errors are, at the smallest, on the order of 10 W m^{-2}. The cumulative error in Pacific heat transport due to systematic errors of 5, 10 and 20 W m^{-2} when integrated from north to south is plotted in Fig. 6a. At 40°S the error is 1.5×10^{15} W if the systematic error is 10 W m^{-2}, a good fraction of the total signal. In Fig. 2b, CW's heat flux from Fig. 3a is replotted with ± 10 W m^{-2} error. The resulting envelope easily encompasses the direct estimates shown in Fig. 2b.

5. Summary

The meridional heat transport in the Pacific Ocean was calculated using net heat fluxes from Clark (private communications, 1983), Weare *et al.* (1981) and Esbensen and Kushnir (1981). Heat transport based on Weare *et al.* and Clark is southward almost everywhere in the Pacific. Major heat transport in the North Pacific is zonal rather than meridional: from 25° to 40°N, there is a region of large heat loss in the Kuroshio gyre and a region of high heat gain along the eastern boundary. The net heat gain in this latitude band is nearly zero, indicating a large westward transport of heat by the ocean. Clark found net heat gain from 20–30° and 40–60°N and small net heat loss from 30–40°N. Weare *et al.* (1981), on the other hand, calculated net heat loss in the band 20–30°N. In contrast, the heat transport based on Esbensen and Kushnir's (1982) net heat fluxes was northward in the North Pacific and southward in the South Pacific.

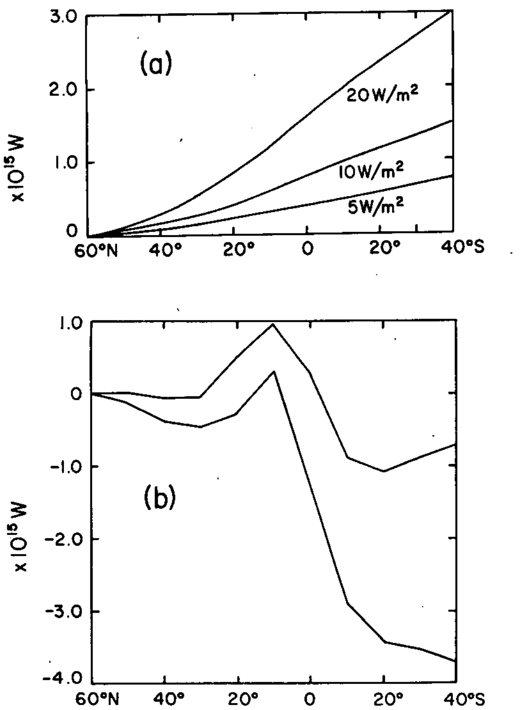

FIG. 6. (a) Cumulative error in meridional heat transport when integrated from north to south for three values of systematic error. (b) Meridional heat transport of Fig. 3b, with a correction of ± 10 W m^{-2}.

It was seen that systematic errors in the bulk parameterizations of heat fluxes were too large to inspire confidence in heat transport calculations in the Pacific Ocean where the meridional heat transport is small and the area large compared with the Atlantic. Confident determination of the sign of meridional heat transport in the North Pacific requires a maximum error on the order of 5 W m^{-2}; none of the bulk formulas on their own are this accurate. Poleward heat transport in the South Pacific, however, is large enough to be recognized even with the present errors in bulk parameterizations although its magnitude may be very different from the predictions of indirect methods.

Direct methods used for the South Pacific by Wunsch *et al.* (1983) and Bennett (1978), as well as a combined method used by Georgi and Toole (1982), provide more believable estimates of South Pacific heat transport than the present method in which a potentially large systematic error is integrated from north to south. While the accumulated systematic error is not as large in the North Pacific, there is considerable disagreement between the various indirect estimates of heat transport because they are dominated by small, mid-ocean heat flux, of uncertain sign.

There is reasonable agreement between the total ocean heat transport summed from Bunker-type calculations in all oceans and the estimates of Oort and Vonder Haar (1976) and Trenberth (1979), based on net radiation and atmospheric transport. This may mean that, although systematic error affects the sign of the Pacific heat transport, the North Pacific plays a small role in the global heat balance: the heat transport in the North Pacific is much lower than in the North Atlantic, in agreement with observations of the lack of deep water renewal in the North Pacific (Warren, 1983). There are, however, reasons for attempting to estimate the North Pacific heat transport more carefully: if the heat transport were poleward, there would be even better agreement between Oort and Vonder Haar's ocean heat transport and the net transport based on surface heat budgets. If the heat transport were equatorward, it would reflect a large-scale meridional cell in which upwelling occurs throughout the North Pacific in response to net downwelling in the Antarctic and North Atlantic-Arctic, as discussed by Wyrtki (1965). Direct measurement of the heat transport at several latitudes is necessary if we are even to be certain of its sign.

Acknowledgments. The author thanks N. Clark, B. Weare and S. Esbensen for making their heat flux calculations available, H. Bryden for supplying heat transports for the Atlantic and Indian Oceans and K. Wyrtki for providing heat transports for the Pacific Ocean. Discussion with R. de Szoeke, C. Paulson, J. Richman, B. Weare, S. Esbensen and H. Bryden were helpful as were comments by the referees. Nathan Clark's computations at Scripps Institution of Oceanography were supported by the Office of Naval Research under Contract N00014-79-C-0587. This work was supported by the Office of Naval Research under Contract N000174-79-C-0004, Project NR 083-102 and by the National Science Foundation under Grant OCE-8117694.

REFERENCES

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

Albrecht, F., 1960: Jahreskarten des Wärme-und Wasserhaushaltes der Ozeane. *Ber. Dtsch. Wetterdienstes*, **66**, Bd. 9, 19 pp.

Anderson, R. J., and S. D. Smith, 1981: Evaporation coefficient for the sea surface from eddy flux measurements. *J. Geophys. Res.*, **86**, 449–456.

Bennett, A. F., 1978: Poleward heat fluxes in Southern Hemisphere oceans. *J. Phys. Oceanogr.*, **8**, 785–798.

Bryan, K., 1962: Measurements of meridional and heat transport by ocean currents. *J. Geophys. Res.*, **67**, 3403–3413.

——, 1982: Seasonal variation in meridional overturning and poleward heat transport in the Atlantic and Pacific Oceans: A model study. *J. Mar. Res.*, **40**(Suppl), 39–53.

Budyko, M. I., 1956: *The Heat Balance of the Earth's Surface*, translated by N. A. Stepanova, 1958, U.S. Department of Commerce, Washington, DC, 259 pp.

——, 1963: *Atlas of the Heat Balance of the Earth.* Kartfabrika Gosgeoltehizdata, Leningrad, 75 pp.

——, 1974: *Climate and Life*, Academic Press, 508 pp.

——, N. A. Epinova, L. I. Zubenok and L. A. Strokina, 1962: The heat balance of the earth's surface. *Izv. Akad. Nauk. SSSR, Ser. Geograf.*, No. 1, 6–16.

Bunker, A. F., 1976: Computations of surface energy flux and annual air-sea interaction cycles of the North Atlantic Ocean. *Mon. Wea. Rev.*, **104**, 1122–1140.

——, 1980: Trends of variables and energy fluxes over the Atlantic Ocean. *Mon. Wea. Rev.*, **108**, 720–732.

——, and L. V. Worthington, 1976: Energy exchange charts of the North Atlantic Ocean. *Bull. Amer. Meteor. Soc.*, **57**, 670–678.

Emig, M., 1967: Heat transport by ocean currents. *J. Geophys. Res.*, **72**, 2519–2529.

Esbensen, S. K., and V. Kushnir, 1981: *The Heat Budget of the Global Ocean: An Atlas Based on Estimates from Surface Marine Observations.* Climatic Research Institute, Rep. No. 29. Oregon State University, 27 pp., 188 figs.

——, and R. W. Reynolds, 1981: Estimating monthly averaged air–sea transfers of heat and momentum using the bulk aerodynamic method. *J. Phys. Oceanogr.*, **11**, 457–465.

Georgi, D. T., and J. M. Toole, 1982: The Antarctic Circumpolar Current and the oceanic heat and freshwater budgets. *J. Mar. Res.*, **40**(Suppl), 183–197.

Godfrey, J. S., and T. J. Golding, 1981: The Sverdrup relation in the Indian Ocean, and the effect of Pacific–Indian Ocean throughflow on Indian Ocean circulation and on the East Australian Current. *J. Phys. Oceanogr.*, **11**, 771–779.

Hall, M. M., and H. L. Bryden, 1982: Direct estimates and mechanisms of ocean heat transport. *Deep-Sea Res.*, **29**, 339–359.

Hastenrath, S., 1980: Heat budget of tropical ocean and atmosphere. *J. Phys. Oceanogr.*, **10**, 159–170.

——, 1982: On meridional heat transports in the world ocean. *J. Phys. Oceanogr.*, **12**, 922–927.

——, and P. Lamb, 1978: *Heat Budget Atlas of the Tropical Atlantic and Eastern Pacific Oceans.* University of Wisconsin Press, 90 pp.

Hughes, F. W., L. K. Coachman and K. Aagaard, 1972: Circulation, transport and water exchange in the western Bering Sea. *Oceanography of the Bering Sea.* D. W. Hood and E. J. Kelley, Eds., Institute of Marine Science, University of Alaska, 59–98.

Laevastu, T., 1967: Cloud factor in longwave radiation formulas. *J. Geophys. Res.*, **72**, 4277.

Large, W. G., and S. Pond, 1982: Sensible and latent heat flux measurements over the ocean. *J. Phys. Oceanogr.*, **12**, 464–482.

List, R. J., 1958: *Smithsonian Meteorological Tables,* 6th ed., Smithsonian Institution Press, Washington, DC, 527 pp.

Liu, W. T., K. B. Katsaros and J. A. Businger, 1979: Bulk parameterization of air–sea exchanges of heat and water vapor including molecular constraints at the interface. *J. Atmos. Sci.*, **36**, 1722–1735.

Lumb, F. E., 1964: The influence of cloud on hourly amounts of total solar radiation at the sea surface. *Quart. J. Roy. Meteor. Soc.*, **90**, 43–56.

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**, 781–800.

Payne, R. E., 1972: Albedo of the sea surface. *J. Atmos. Sci.*, **29**, 959–970.

Pullen, P., 1977: Ocean heat flux variations related to the Kuroshio (1956–1964). *J. Geophys. Res.*, **82**, 4915–4920.

Reed, R. K., 1976: On estimation of net longwave radiation from the oceans. *J. Geophys. Res.*, **81**, 5793–5794.

——, 1977: On estimating insolation over the ocean. *J. Phys. Oceanogr.*, **7**, 482–485.

——, 1982: Comparison of measured and estimated insolation over the eastern Pacific Ocean. *J. App. Meteor.*, **21**, 339–341.

——, and D. Halpern, 1975: Insolation and net longwave radiation off the Oregon Coast. *J. Geophys. Res.*, **80**, 839–844.

Seckel, G. R., and F. H. Beaudry, 1973: The radiation from sun and sky over the North Pacific Ocean (abstract). *Trans. Amer. Geophys. Union,* **54**, 1114.

Simpson, J. J., and C. A. Paulson, 1979: Mid-ocean observations of atmospheric radiation. *Quart. J. Roy. Meteor. Soc.*, **105**, 487–502.

Smith, S. D., and F. W. Dobson, 1984: The heat budget at Ocean Weather Ship Bravo. *Atmos.-Ocean,* (in press).

Stephens, G. L., G. G. Campbell and T. H. Vonder Haar, 1981: Earth radiation budgets. *J. Geophys. Res.*, **86**, 9739–9760.

Stommel, H. M., and G. T. Csanady, 1980: A relation between the *T-S* curve and global heat and atmospheric water transports. *J. Geophys. Res.*, **85**, 495–501.

Sverdrup, H. U., 1957: Oceanography. *Handbuch der Physik,* **48**, Springer-Verlag, 608–670.

Trenberth, K. E., 1979: Mean annual poleward energy transports by the ocean in the Southern Hemisphere. *Dyn. Atmos. Oceans,* **4**, 57–64.

Vonder Haar, T. H., and A. H. Oort, 1973: New estimate of annual poleward energy transport by northern hemisphere oceans. *J. Phys. Oceanogr.*, **3**, 169–172.

Warren, B. A., 1983: Why is no deep water formed in the North Pacific? *J. Mar. Res.*, **41**, 327–347.

Weare, B. C., and P. T. Strub, 1981: The significance of sampling biases on calculated monthly mean oceanic surface heat fluxes. *Tellus,* **33**, 211–224.

——, P. T. Strub and M. D. Samuel, 1981: Annual mean surface heat fluxes in the tropical Pacific Ocean. *J. Phys. Oceanogr.*, **11**, 705–717.

Wunsch, C., D. Hu and B. Grant, 1983: Mass, heat, salt and nutrient fluxes in the South Pacific Ocean. *J. Phys. Oceanogr.*, **13**, 725–753.

Wyrtki, K., 1965: The average annual heat balance of the North Pacific Ocean and its relation to ocean circulation. *J. Geophys. Res.*, **70**, 4547–4559.

——, E. B. Bennett and D. J. Rochford, 1971: *Oceanographic Atlas of the International Indian Ocean Expedition.* National Science Foundation NSF-IOE-1, 531 pp.