Estimates of Oceanic Horizontal Heat Transport in the Tropical Pacific

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Meridional heat transport in the tropical Pacific is estimated using a linear numerical model with realistic boundaries and forced by 18 years of observed wind, covering the period from January 1962 to December 1979. The long-term mean heat transport estimated in this study is similar to the estimates based on heat balance and radiation considerations and on complex numerical models that account for thermodynamics as well. This points to the dominant role played by the adiabatic process, the only heat transport mechanism present in this study, in the heat balance for the equatorial Pacific. The combined Ekman and geostrophic heat transport can account for the net meridional heat transport, except near the equator, where continuity requirements dictate. The Ekman and geostrophic transport oppose each other, and their small difference in magnitude gives rise to the net meridional heat transport, resulting in transport away from the equator for the southern hemisphere and north of 6°N, while for the band between 6°N and the equator an equatorward transport is present. Seasonal and interannual variations are found to be as large as, or even larger than, the long-term mean. Seasonal variations in meridional heat transport are in accordance with seasonal variations in zonal winds via Ekman transport, while the geostrophic transport remains more or less constant on this time scale. The results are a net poleward heat transport in the winter hemisphere and an equatorward transport in the summer hemisphere. At the interannual time scale, variabilities in both Ekman and geostrophic components contribute to the interannual variability in heat transport. Major features of the interannual variations in meridional heat transport appear to be associated with El Nino events. It is interesting to note that the interannual variations associated with El Nino events are not restricted to the near-equatorial region. Phase locking between the interannual variations and the annual cycle is evident in the data set. Major findings in this study, based on an adiabatic model, are expected to carry over to more realistic nonlinear numerical models.

1. INTRODUCTION

The earth receives shortwave radiation from the sun predominantly in the tropics, but longwave radiation back into space is relatively uniform over the globe. This results in a heat gain at low latitudes and a heat loss at high latitudes. Because of this differential heating, a net poleward heat transport by the earth's fluid envelope is required. Over continents, redistribution of energy takes place exclusively in the atmosphere, while over the oceans, both the ocean and the atmosphere intervene. The relative importance of the atmosphere and ocean in effecting this transport has not yet been established. Recent results suggest, however, that the ocean heat transport is not only a relevant mechanism but, in some areas, provides the dominant transport [Vonder Haar and Oort, 1973]. Using radiation budget calculations, Oort and Vonder Haar [1976] found that the oceanic contribution to the total net poleward heat transport at 20°N is about 50%. This percentage seems to increase equatorward, while at middle and high latitudes the atmospheric contribution seems dominant.

Heat exchange through the ocean-atmosphere interface is a primary driving force for the atmospheric circulation; wind stresses at the ocean surface, in turn, drive the major ocean currents. An important component of this feedback mechanism must be the redistribution of heat by the ocean. A better understanding of the oceanic heat transport (OHT) mechanisms seems necessary in order to effectively model and predict climate.

Oort and Vonder Haar [1976] showed evidence of a large annual cycle of OHT at low latitudes with heat flowing from the summer to the winter hemisphere. They found that a cross-equatorial, seasonally varying heat flux of the order of 1 PW (1 PW = 10^{15} W) is needed to maintain the heat balance.

This large, necessary heat flux shows the primary role played by the tropical ocean in the earth's climate. Since the pioneering work of Oort and Vonder Haar, widespread attention has been given to the subject. Recent reviews of the progress in calculating OHT were presented by Bryan [1982, 1983].

Three basic methods have been used to estimate OHT [Bryan, 1983]. The first one is the heat balance method. Implementation of this method has followed two different approaches: heat balance considerations at the ocean surface and a global heat balance. In the former, empirical formulas are used to compute heat fluxes through the ocean surface, while in the latter a global heat balance of the entire ocean-atmosphere-earth system is considered. In both cases, transport of heat by the ocean is estimated as a residual term from the requirement of heat balance. The advantage of this method is that it permits an estimation of the heat balance components for large areas of the ocean by using standard meteorological-oceanographic observations. A disadvantage of this method is that the uncertainties in the parameterization of surface heat fluxes create unknown uncertainties in the estimated transport. This is the "traditional method" [Hall and Bryden, 1982], and it has been used by a number of investigators [e.g., Budyko, 1963; Wyrtki, 1965; Oort and Vonder Haar, 1976; Hastingh and Lamb, 1977].

A second method is the direct method. It uses oceanographic observations of velocity and temperature to calculate direct heat fluxes. This method permits investigation into the mechanisms by which heat is transported [Bryan, 1983]. Although the direct method seems to be most appropriate, the lack of detailed hydrographic or direct current measurements across an entire ocean makes its use very limited. Nevertheless, its value lies in the fact that it gives "direct observations" of heat transport against which other indirect calculations or models can be compared. Reviews on the direct estimates and mechanisms of OHT were presented by Hall and Bryden [1982].

A third possible way of estimating OHT is through the use...
of numerical models. The problem of constructing a simple model for OHT is basically the same as finding a simple model of the ocean circulation [Bryan, 1982b]. As with most models, models of heat transport can be classified into two overlapping categories: those that try to isolate basic physical mechanisms [e.g., Schopf, 1980; Cane and Sarachik, 1983] and those that try to reproduce observed data [e.g., Bryan et al., 1975; Bryan and Lewis, 1979]. In this study an intermediate approach is taken: A simple, linear, reduced-gravity model, but with realistic geometry and forced by real winds, is used to estimate heat transports. The objective of this study is to understand better how the tropical Pacific Ocean transports heat away from the equator.

Cane and Sarachik [1981] solved the linear shallow-water equation forced by latitudinally varying (but zonally averaged) seasonal wind. Estimating heat transports with this model, they found [Cane and Sarachik, 1983] that the meridional transport by the western boundary current is of the same magnitude as, but out of phase with, the seasonally varying interior transport, and both tend to cancel each other. The role of Ekman flow and planetary waves on the OHT has been investigated by Schopf [1980]. Using a numerical model, Schopf found that basic Ekman pumping and drift could account for most of the net cross-equatorial heat flux. He also found that additions of planetary and gravity waves do not alter the basic pattern. The values of OHT were found to be comparable to those estimated by Oort and Vonder Haar [1976] [Schopf, 1980]. Schopf points out that the adiabatic advection mechanism for heat transport implicit in his model would not be appropriate for higher latitudes (≥ 20°) where other physics should be included (i.e., surface heat fluxes, mixing processes, gyral effect, etc.).

The present study is somewhat similar to that of Schopf [1980] and Cane and Sarachik [1983]: estimation of oceanic heat fluxes are drawn from a simple numerical model. Our study differs from theirs, however, in that our model includes the realistic boundaries of the equatorial Pacific and is forced by real winds.

The same model has been used before to study the seasonal [Busalacchi and O'Brien, 1980] and interannual variability of the equatorial Pacific for the 1960's [Busalacchi and O'Brien, 1981] and 1970's [Busalacchi et al., 1983].

A very good agreement with observations of the thermocline topography and its seasonal variations was obtained when the model was forced by mean monthly winds [Busalacchi and O'Brien, 1980]. Using 18 years of observed winds, covering the periods 1961–1978, the variability of sea level at the Galapagos Island was found to be similar to the variability of the model pycnocline. El Nino events of the 1960's and 1970's are characterized in the model by a persistently deep pycnocline. The results also indicate that the variability at the equator is related to the excitation and propagation of Kelvin and Rossby waves [Busalacchi and O'Brien, 1981].

For the present study the same wind data, extended to 1979, is used to force the model. A description of the procedures used to estimate OHT is presented in section 2. In section 3 the results are presented and discussed. Comparison with other authors' results is also given in this section. Finally, in section 4, conclusions and criticisms are considered.

2 PROCEDURES

2a. Net Heat Transport

In this study the model developed by Busalacchi and O'Brien [1980, 1981] was used to compute heat transport. The model is a one-layer, reduced-gravity, linear transport model on an equatorial β-plane. The model covers a domain extending from 126°E to 79°W and from 18°N to 12°S, with idealized boundaries to represent the tropical Pacific Ocean (Figure 1). The northern and southern boundaries are treated as open boundaries.

Through the equation of state, the difference in density between the two layers is related to a difference in temperature and heat content. For the upper layer of thickness h and temperature T the heat content per unit area, s, can be expressed as

\[ s = \rho \hat{c}_p (T - T_0) h \]  

where \( \rho \) is the density of the upper layer, \( \hat{c}_p \) is the heat capacity, and \( T_0 \) is the reference temperature (temperature of the lower layer). Using the equation of state in the form

\[ \rho = \rho_o (1 - \alpha (T - T_0)) \]  

where \( \rho_o \) is the reference density and \( \alpha \) is the coefficient of thermal expansion, \((T - T_0)\) can be eliminated from (1) to give an expression for heat content in terms of the density jump \( \Delta \rho = \rho_o - \rho \) and the layer thickness \( h \). After integration across a latitudinal band extending from the western \( (x_w) \) to the eastern \( (x_e) \) boundary, the rate of change of heat in the volume is given by

\[ \frac{\partial s}{\partial t} = \gamma \int_{x_w}^{x_e} \frac{\partial h}{\partial t} dy dx \]  

where \( \gamma = c_p (\Delta \rho / \alpha) \) is taken to be a constant in this study. A list of symbols and values used for the constants may be found in the appendix.

Following Busalacchi and O'Brien [1980], the ocean model equations are

\[ \frac{\partial V}{\partial t} = -\beta \gamma U c^2 \frac{\partial h}{\partial y} + \frac{\partial^2 V}{\partial y^2} + A V^2 \frac{\partial V}{\partial x} \]  

\[ \frac{\partial U}{\partial t} = \beta \gamma V c^2 \frac{\partial h}{\partial x} + \frac{\partial^2 U}{\partial x^2} + A V^2 U \]  

\[ \frac{\partial h}{\partial t} + \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0 \]  

where \( V = U H \) and \( U = U_H \) are the transports in the north-south and east-west directions, respectively; \( H \) is the undisturbed depth of the layer; \( c \) is the baroclinic phase speed, taken to be 2.45 m s \(^{-1}\); \( \tau \) is the wind stress applied as a body force, and \( A \) is the horizontal eddy viscosity coefficient to make the horizontal velocity along the closed boundaries vanish.

Integration of the continuity equation over a latitudinal band gives, after multiplication by \( \gamma \) and using (3),

\[ \frac{\partial s}{\partial t} = B(y_e - \Delta y) - B(y_w + \Delta y) \]  

where

\[ B(y) = \gamma \int_{x_w}^{x_e} V(x, y) dx \]  

The first and second terms on the right side of (5a) can be identified as the northward heat transports at the southern and northern boundaries of the band, respectively.

2b. Components of oceanic heat transport

For a basin closed at one end, two important modes of meridional heat transport can be identified [Bryan, 1982b]:
(1) overturning in the meridional plane, where poleward heat transport can take place because of the poleward movement of warm water at the surface compensated by the equatorward movement of colder deep water;

(2) gyre effect, where correlations between velocities and temperature in the horizontal plane contribute to a meridional heat transport. At low latitudes the temperature contrast between surface and deep water is much greater than the temperature contrast in a typical east-west section. Meridional overturning is therefore more efficient in transporting heat than a gyre circulation of similar strength. Earlier calculations [e.g., Bryan and Lewis, 1979; Meehl et al., 1982] seem to corroborate this idea.

Meridional overturning is the only mechanism included in a one-layer, reduced-gravity model. In this model, net meridional transport in the upper layer is compensated by an equal transport in the opposite direction of higher-density (colder) water in the lower layer. Away from the equator the overturning circulation can be thought of as being composed of two components. One is related to the Ekman transport, and the other is a geostrophically balanced component.

For low-frequency forcing, away from the equator and boundaries, the system of equations (4a) and (4b) can be approximated as [Busalacchi and O'Brien, 1980]

\[ \beta y U = -c^2 \frac{\partial h}{\partial y} + \tau^* \frac{\rho}{\rho} \]  
\[ -\beta y V = -c^2 \frac{\partial h}{\partial x} + \tau^* \frac{\rho}{\rho} \]

For this linear model one can think of the transports \( U \) and \( V \) as being composed of two components satisfying the geostrophic and Ekman relations. Meridional heat transport contributions resulting from Ekman (\( B_e \)) and geostrophic (\( B_g \)) components are defined as

\[ B_g(y, t) = \gamma \int_{-\infty}^{\infty} V_e \, dx \]  
\[ B_e(y, t) = \gamma \int_{-\infty}^{\infty} V_e \, dx \]

where \( V_e \) and \( V_g \) refer to the transport caused by Ekman and geostrophic components, respectively, and are defined as

\[ V_e = -\frac{\tau^*}{\beta y \rho} \]  
\[ V_g = \frac{c^2 \frac{\partial h}{\partial x}}{\beta y} \]

Away from the equator these two contributions must nearly sum up to the net transport. The integrated geostrophic transport depends on large-scale features that adjust in much longer time than the adjustment time for the Ekman transport. It can be expected that most of the seasonal and shorter time scales variability in the net meridional transport is related to the Ekman transport changes. On the other hand, for the interannual variability the geostrophic component is more likely to be important.

At the equator the vanishing of the Coriolis force enables one to write an equation for \( V \) in the form [Busalacchi and O'Brien, 1980]

\[ V_{int} = -\frac{\beta^2 y^2}{c^2} \]  
\[ V_e + \beta V_e = \kappa \left( \nabla \tau \frac{\tau}{\rho} \right) \]  
\[ + \frac{\beta y}{c^2} \left( \frac{\tau^*}{\rho} \right) + \frac{1}{c^2} \left( \frac{\tau^*}{\rho} \right) \]
Solutions of this equation can be represented in terms of dispersive and nondispersive baroclinic Rossby waves [McCreeary, 1976]. A further wavelike solution of (4a) through (4c) is the equatorially trapped Kelvin wave [Moore and Philander, 1977]. This solution can be derived more easily by constraining the meridional velocity to be zero in the homogeneous ($f = 0$) part of (4a) and (4b) (also with $A = 0$). The resulting Kelvin-wave solution is a nondispersive, eastward propagating wave with phase speed $c$ [Moore and Philander, 1977]. Since this latter wave has zero north-south transport, it cannot alter the meridional heat transport pattern. However, it might be indirectly important through excitation of meridionally propagating, coastally trapped Kelvin waves. Equatorial Kelvin waves have been shown to propagate part of their energy in the form of boundary-trapped Kelvin waves [Moore, 1968].

3. RESULTS AND DISCUSSION

3a. Net Meridional Transport

In the preceding section we discussed a relation that enables one to estimate meridional heat fluxes from the interface depth of our model. To interpret the results, an analysis of the heat budget components is in order. For a zonally integrated ocean, if one defines $Q_s$ as the storage of heat in the ocean (rate of gain or loss of heat), $Q_a$ as the horizontal advection of heat by the wind-driven circulation, $Q_t$ as the horizontal advection of heat by the thermohaline circulation, and $Q_{ac}$ as the net input or loss of heat through the ocean surface, the heat budget for a particular body of water can be stated symbolically by the equation

$$Q_t = Q_s + Q_a + Q_{ac}$$

(12)

For a long-term average, since the ocean maintains a constant temperature, the storage term $Q_s$ in (12) can safely be neglected. At low latitudes, as noted before, the ocean receives a net input of heat through its surface, and hence for a balance the advection terms $Q_a$ and $Q_t$ can be expected to be poleward. The net heat transport estimated from the model via (5) is thought to represent the adiabatic advection of heat by the wind-driven circulation ($Q_a$). Only $Q_s$ is estimated in this study. The model does not include the transport caused by the thermodynamic circulation or the surface heat fluxes. Unless otherwise indicated, a reference to the net heat transport in the following sections is a reference to $Q_t$ not to the "total" horizontal transport ($Q_w + Q_s$).

The 18-year average heat transport, computed by using (5b), is shown in Figure 2 as a function of latitude. North of 6°N, the net transport is northward, while south of this latitude the transport is to the south. The cross-equatorial heat transport is from the northern to the southern hemisphere with a magnitude of about 0.18 PW.

Three regions can be identified in Figure 2: (1) From the southern boundary of the model (12°S) to around 5°N the heat transport increases linearly from $-0.35$ PW at 12°S to $-0.15$ PW at around 5°N. (2) A turning latitude band from 5°N to about 8°N, where the transport changes from southward to northward. (3) To the north of 8°N the transport remains more or less constant at about 0.25 PW until 12°N, where it decreases slowly to the north.

Except for the "abnormal" latitudinal band north of the equator, our estimates imply a "normal" ocean in the sense introduced by Stommel [1980]: one in which the flow of heat is poleward. Wyrtki [1965] states that for heat balance calculations the boundary between the north and south Pacific Ocean should not be at the geographic equator but at the thermal equator, which is situated near 6°N. If the thermal equator, instead of the geographic equator, is considered in Stommel's definition of a normal ocean, our "abnormal" band would be removed and our estimates would imply a completely normal Pacific Ocean.

Our estimates compare well with those calculated indirectly by Hastenrath [1980] from heat budget and radiation considerations. In both cases there is a southward cross-equatorial heat transport of similar strength (~0.20 PW) and a turning latitude at about the same latitude: "somewhat north of the equator" [Hastenrath, 1980], 6°N in our case. Estimates with a numerical model presented by Meehl et al. [1982] forced by monthly mean winds and atmospheric temperatures also show very similar values for the cross-equatorial transport and turning latitude. The magnitudes of the transport at higher latitudes, however, are somewhat smaller in our study. Bryan [1982a] used the January-July average as an approximation to the annual mean heat transport. In Figure 3 our January-July average is compared to Bryan's [1982a] results. The similarity of both calculations is striking if one considers the very different characteristics of the models. The model used by Bryan [1982a] is a fully nonlinear model with thermodynamic effects included and much greater vertical resolution than the one used in this study. The conclusion drawn from this result is that the heat storage and lateral heat advection play a dominant role in the overall heat budget in the equatorial Pacific over the seasonal time scale.

![Figure 2](image1.png)

Fig. 2. Mean net meridional heat transport by the ocean model. Northward transport positive. Units are in PW (1 PW = 10^{15} W).

![Figure 3](image2.png)

Fig. 3. January-July average heat transport by the ocean model compared to the January-July estimates from Bryan [1982a]. Units are in PW.
or less constant.

...Ekman transport remains more evident in the combined Ekman and geostrophic parts, as with the net transport (Figure 2). One observes that, except for both contributions is 10 PW, one order of magnitude smaller than those computed by Hastenrath [1980] and Meehl et al. [1982]. The characteristics of the heat transport for the Pacific and Atlantic oceans are, consequently, very different. While the Pacific, as mentioned before, transports heat away from the equator (normal-sign ocean), the transport in the Atlantic is northward at all latitudes (abnormal sign in the southern hemisphere) [Hastenrath, 1980].

The difference has been explained in terms of the thermohaline circulation in the Atlantic [Stommel, 1980]. The Atlantic loses sufficient heat in the North Atlantic that it causes the heat flux from the South Atlantic to be reversed in sign [Stommel, 1980]. Since the Pacific does not extend to such high latitudes in the northern hemisphere, the thermohaline circulation is weaker and does not constitute a dominant mechanism for transporting heat. It is possible, however, that a reason for our values being smaller than those computed by Hastenrath [1980] and Meehl et al. [1982] is the absence of the thermohaline circulation in our model. The effect of the thermohaline circulation in the Pacific trade wind regions would be to reinforce the Ekman drift, producing a stronger poleward transport of surface water [Bryan et al., 1975], hence a stronger poleward transport of heat.

Transport of heat in a horizontal plane (i.e., gyral effect) and the contribution to the ocean heat transport by synoptic-scale eddies have been shown to be negligible in the equatorial regions [Bryan and Lewis, 1979].

3b. Seasonal Variation

Oort and Vonder Haar [1976] showed evidence of a large seasonal variation of the OHT. They found the amplitude of the annual variation in the tropical ocean to be of the order of...
variation of the meridional heat transport in the Atlantic and Pacific oceans. With the January minus July pattern as a representative range of the annual cycle, he found the vertical overturning associated with the Ekman transport to be the dominant term in the overall equatorial meridional transport. Bryan [1982a] estimated the northward equatorial transport to vary from about +1 PW in January to about -1 PW in July.

The effect of long planetary waves on the seasonal heat transport was studied by Schopf [1980]. Using a simple numerical model, Schopf demonstrated that, although these waves strongly affect the local behavior, they do not significantly modify the zonally integrated transport. The equatorial and extra-equatorial dynamics in our model (4a) through (4c), as noted before, include long planetary waves. Specifically, close to the equator our model response includes dispersive and nondispersive Rossby waves as well as the equatorially trapped Kelvin wave. The results of Schopf [1980] suggest that, at the seasonal time scale at least, and for the zonally integrated transport, all these waves are not relevant, and the

Using a numerical model of the world ocean [Bryan, 1979; Bryan and Lewis, 1979], Bryan [1982a] described the seasonal variation of the meridional heat transport in the Atlantic and Pacific oceans. With the January minus July pattern as a representative range of the annual cycle, he found the vertical overturning associated with the Ekman transport to be the dominant term in the overall equatorial meridional transport. Bryan [1982a] estimated the northward equatorial transport to vary from about +1 PW in January to about -1 PW in July.

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solution across the equator is simply determined by continuity requirements.

Following is a brief description of the characteristics of the wind used to force the model to provide background for the discussion on the seasonal variation of the estimated heat transport. A more extensive description of the wind in the area was given by Wyrtki and Meyers [1975] and by Goldenberg and O'Brien [1981].

Seasonal variation of the zonally averaged east-west wind pseudo stress is shown in Figure 6a as a function of latitude and month. Figure 6b shows a similar plot for the wind stress anomaly, calculated by subtracting from the values in Figure 6a the annual average wind at each latitude.

As seen in Figure 6a, the zonal component of the wind is toward the west throughout the year at all latitudes. An annual oscillation is present with maximum easterlies (−58 m² s⁻²) at 10°N and at 12°S (−42 m² s⁻²) in February and August, respectively, and weaker easterlies in September at around 8°N (−3 m² s⁻²). From Figure 6b the zonal component of the wind shows a seasonal variation markedly anti-symmetric across the equator, with stronger than the average easterlies in the winter-spring hemisphere and weaker than the average in the summer-fall hemisphere. The magnitude of the annual variation appears strongest around 8-10°N and weakest at the equator.

Figure 7 shows the seasonal variation of heat transport as a function of latitude and time. There is a strong seasonal variation in the OHT, with a maximum northward heat transport (NHT) at 8°N in February and a maximum southward heat transport (SHT) in September at around 5°N. The band from 5°N to 8°N is the band of maximum annual variation going from 1.4 PW in February to −1.1 PW in September. The magnitude of the heat transport decreases to the north and south of these latitudes, becoming less positive (northward) away from around 8°N in the northern hemisphere winter-spring and less negative (southward) in the northern hemisphere fall-summer. In the southern hemisphere the transport is mostly to the south, with a relative maximum SHT in July.

The cross-equatorial transport is from the summer to the winter hemisphere. From December to April there is a northward cross-equatorial heat transport, while from May to November a southward cross-equatorial heat transport is present.

The Ekman and geostrophic components of the heat transport are shown in Figures 8 and 9, respectively. As expected, they oppose each other, with Ekman transport poleward and geostrophic transport equatorward in both hemispheres throughout the year.

The geostrophic transport (Figure 9) is nearly anti-symmetric across the equator, it increases in magnitude toward the equator in a way dictated by the reduction of the Coriolis parameter (equation (10)). The transport at a given latitude remains almost constant throughout the year, except for a slight increase on the equatorward transport in August-September. The Ekman transport (Figure 8), on the other hand, presents a marked seasonal variation. The transport is more or less symmetric across the equator, with weak poleward heat transport (PWHT) in August-September and strong PWHT in February in the northern hemisphere and vice versa for the southern hemisphere.

The importance of the Ekman transport on the seasonal variation of net heat transport becomes evident by comparing Figure 6 and 8. It can be seen from those figures that, away from the equator, almost all the seasonal variation in the net transport is given by a similar seasonal variation in the Ekman transport associated with the seasonal variation in zonal wind. The stronger easterlies in the winter hemisphere (Figure 6b) drive a stronger poleward Ekman transport that can overcome the equatorward geostrophic transport, resulting in a net poleward transport. In the summer hemisphere the reduced easterlies result in a weaker poleward Ekman transport, and the equatorward geostrophic transport dominates.

3c. Interannual Variation

In the preceding section it was shown that the seasonal variation of the winds in the tropical Pacific is responsible for
the seasonal variation of the meridional heat transport and that its effect is very strong compared to the mean transport.

In the tropical Pacific the interannual variations in the wind stress can be as strong as the seasonal variations. For example, in their study of the wind stress variability, Goldenberg and O'Brien [1981] found an area of high interannual variability centered just north of the equator in the central Pacific. It seems only reasonable to expect the meridional heat transport to respond to such large-scale interannual variations in the wind pattern as well as to the seasonal variations. In this section a description of the interannual variability of the estimated heat transport is made.

During recent years, the recognition of the El Nino phenomenon as a part of the very large scale anomalies in both the atmosphere and the ocean has produced a dramatic increase in the interest in the phenomenon, as can be judged by the large amount of literature published on the subject. In this study it is not our intention to analyze in detail the El Nino events in our estimates of heat transport but only to describe the interannual variability of the estimated heat transport and to recognize that it must be a part of the large-scale variability in the ocean-atmosphere system associated with El Nino events.

Historically, the term El Nino refers to an annual occurrence of warm water off the coast of Peru and Ecuador. At irregular intervals a massive version of El Nino occurs, where
the warm water accumulation is excessive, leading to a natural catastrophe. This irregular and more dramatic event is what recently has been referred to as the El Nino phenomenon. The interannual variability of the model response (including El Nino events) in the 1960's and 1970's have been discussed by Busalacchi and O'Brien [1981] and Busalacchi et al. [1983]. The model response was found to correlate very well with the observed sea level records.

Probably because of historical reasons, what have been referred to as El Nino characteristics are mostly related to eastern boundary responses (i.e., warming of the sea surface, higher sea level, deepening of the thermocline, etc.). The response of the model (and presumably of the real ocean), however, is characterized by longitudinal variation [Busalacchi et al., 1983]. A zonally integrated meridional heat transport is the response of the whole basin, and it is difficult to recognize the dynamical mechanisms that have been usually linked to the El Nino phenomenon (i.e., equatorial and coastally trapped Kelvin waves).

Intuitively, one would expect a decrease in the poleward heat transport during the onset and development of El Nino event, when the ocean at the eastern boundary is warming and the pycnocline is deepening. During the retreat of the warm water, the final stages of the phenomenon, an increase in the poleward transport, can be expected in order to bring the pycnocline depth and heat transport back to its "normal" averaged position.

The interannual variability of the estimated heat transport is obtained by applying a 12-month running mean filter to the computed time series. The result is shown in Figure 10a as a
latitude-time plot. Figure 10b shows the heat transport anomalies as a function of latitude and time. A positive anomaly means stronger northward transport or weaker southward transport than the long-term average. A negative anomaly means a weaker northward transport or a stronger southward transport compared to the long-term mean. In Figures 10a and 10b the years that have been classified as El Nino years (1963, 1965, 1969, 1972, 1976) are marked to facilitate the description.

Inspection of Figures 10a and 10b shows an important result of this section: the interannual variability in the estimated meridional transport is not at all negligible and is not restricted to the equator. The actual magnitude of the variability is bigger at some latitudes than the mean and almost as large as the seasonal variation. This result itself implies that estimated values of heat transport based on the direct method [e.g., Hall and Bryden, 1982] do not necessarily represent the “true” long-term average. Long time series are called for to define a meaningful average annual transport.

From Figure 10a the region of maximum interannual variability is found to be around 8°N. This latitude corresponds approximately to the latitude of maximum seasonal variability, which seems to reinforce the idea of the interannual variability being linked to the seasonal cycle. To corroborate this idea, complex demodulation [see Bloomfield, 1976] of the heat transport time series at the annual frequency was performed (not shown here). The complex demodulation analysis showed that the interannual variation is mainly due to changes in the amplitude of the annual cycle, which explains the link between the two time scales. Interpretation of interannual variability in terms of an anomalous amplification of the annual cycle is not new. It has been discussed by Horel [1982] for atmospheric parameters and by Meyers [1982] for oceanic parameters.

From Figure 10b it can be seen that the anomalies are not restricted to the equator. Bands of positive and negative anomalies extend throughout the whole latitudinal extent of the model. Maximum positive anomalies north of the equator occur at the end (beginning of next year) of 1963, 1969, 1972, 1976, all of these years having been classified as El Nino years. Other positive anomaly maxima occur north of the equator at the beginning of 1975 and south of the equator at the end of 1964 (beginning of 1965). The year 1975 has been classified as an aborted El Nino year and 1965 as a major El Nino year [Rasmussen and Carpenter, 1982].

Although the latitudinal focus, strength, and duration of the maximum positive anomalies differ for each event, Figure 10b shows a clear correlation between the occurrences of El Nino and the heat transport anomalies. For 1963, 1965, and 1969 the pattern seems to be a stronger-than-average poleward heat transport, positive anomalies in the north, and negative anomalies in the south. The strongest latitudinal variation on the heat transport anomalies occurs at around 5°N in 1972, a year that has been recognized as a major El Nino [Rasmussen and Carpenter, 1982]. For this year a region of convergence anomaly, negative anomaly to the north of a positive anomaly, also occurs to the north of 8°N.

4. Summary and Conclusions

Net meridional heat transport in the equatorial Pacific has been estimated with a simple numerical model. The model used is that of Bualacchi and O’Brien [1980, 1981]: a linear, reduced-gravity model in an equatorial β-plane forced by 18 years of monthly surface winds (from January 1962 to December 1979). The coastline geometry of the model basin is an approximation of the tropical Pacific. With a linear equation of state the density difference between the dynamically active upper layer and the motionless lower layer was related to a difference in temperature and heat content.

The estimated long-term mean heat transport by the wind-driven circulation was first analyzed. Our estimated heat transport was found to be southward to the south of 6°N and northward north of this latitude. Qualitatively, the latitudinal shape and magnitude of the transport compare well with the results presented by Hastenrath [1980] and Meethal [1982], which shows that the lateral advection is a major term in the total heat budget. However, the different characteristics of the models previous to the model used in this study (e.g., realistic geometry, real winds) and the lack of observations that are dense enough (in space and time) make it impossible to carry out a detailed comparison of our results with previous studies of heat transport.

The physical mechanisms involved in heat transport were investigated by decomposing the net meridional heat transport into two parts: Ekman and geostrophic. These two components oppose each other, with poleward and equatorward transport for the Ekman and geostrophic components, respectively. Both Ekman and geostrophic components were found to be an order of magnitude larger than their sum. Except near the equator, the net heat transport is almost exclusively due to the combined effect of the Ekman and geostrophic parts. The mechanism suggested by these results is that of an equatorward geostrophic transport of water maintained by a west-to-east pressure gradient balanced by the easterly trade winds and opposed by a direct poleward Ekman drift. The resulting net meridional transport of water in the upper (warmer) layer is compensated by an equal transport of lower-layer (colder) water, producing a net meridional heat flux.

Seasonal and interannual variations are found to be large compared to the long-term mean. At the seasonal time scale the geostrophic transport remains more or less constant, and large seasonal variations in the estimated heat transport were found to be directly associated with the seasonal variation of the zonal wind via Ekman transport. The stronger-than-average easterlies in the winter hemisphere drive a stronger poleward Ekman transport that can overcome the equatorward geostrophic transport, resulting in a net poleward heat transport. In the summer hemisphere the reduced easterlies result in a weaker poleward Ekman transport, and the equatorward geostrophic transport dominates.

Maximum interannual variability was found to be north of the equator at a latitude that approximately coincides with the latitude of maximum seasonal variability. Interpretation of the interannual variability as an anomalous intensification of the annual cycle is thought to explain this coincidence. Correlation between the occurrence of El Nino phenomenon and the interannual anomalies in the estimated heat transport is evident. The most conspicuous being the presence of large positive anomalies at the end of all the years classified as El Nino years. It is interesting to note that the interannual variations associated with El Nino events are not restricted to the near-equatorial region.

It is recognized that the model used in this study is a very simple one. The present model only provides information on heat storage and lateral heat advection, two important but not exclusive terms on the total heat budget. A complete thermodynamic model could only shed information on the remaining terms of the heat budget. Inclusion of surface heat fluxes and
other thermodynamics would certainly modify the magnitude, and probably the phase, of our estimated heat transport. However, comparison between our results and previous estimates, based on observations and numerical models that incorporate thermodynamics, suggests that the adiabatic dynamics, with near cancellation of the Ekman and geostrophic transports, is a dominant process in the total heat budget of the equatorial Pacific. Of course it would be of interest to test this idea with a more complete model.

**Notation**

- $A$: horizontal eddy viscosity coefficient, $10^2$ m$^2$ s$^{-1}$
- $B_{E}$: Ekman and geostrophic component of heat transport, respectively.
- $c$: baroclinic phase speed, $(g'H)^{1/2}$, 245 ms$^{-1}$
- $c_p$: heat capacity of water at constant pressure, 4.02 $\times 10^3$ J kg$^{-1}$ K$^{-1}$
- $g$: acceleration resulting from gravity, 9.8 m s$^{-2}$
- $g'$: reduced gravity, $g(p_0 - p)/p_0$
- $h$: upper-layer thickness.
- $h$: initial upper-layer thickness, 300 m.
- $k$: $z$-directed unit vector.
- $s$: heat content.
- $t$: time.
- $T$: temperature of upper and lower layer, respectively.
- $x$, $y$, $z$: tangent plane cartesian coordinates: $x$ positive eastward, $y$ positive northward, and $z$ positive upward.
- $x_0$, $x_0$': meridional boundaries of the model basin.
- $\alpha$: coefficient of thermal expansion, 2.4 $\times 10^{-4}$ deg$^{-1}$.
- $\beta$: meridional derivative of Coriolis parameter, 2.25 $\times 10^{-11}$ m$^{-1}$ s$^{-1}$.
- $\gamma$: constant coefficient $\Delta x p x^{-1}$.
- $\rho$, $\rho_0$: density of seawater, upper and lower layer, respectively.
- $\Delta \rho$: density difference between upper and lower layer, 2.0 kg m$^{-3}$.
- $r^*$, $r^*$: zonal and meridional components of wind stress, respectively.

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**References**


