On Oceanic Heat and Freshwater Fluxes at 30°S*

ALBERTO R. PIOLA

Servicio de Hidrografía Naval, Buenos Aires, Argentina

ARNOLD L. GORDON

Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964

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ABSTRACT

A simple box model based on mass, heat and salinity conservation combined with existing estimates of ocean-atmosphere heat and freshwater exchanges is used to calculate the oceanic mean meridional volume fluxes of three water masses at 30°S. Model results lead to relatively large volume fluxes and questionable flow directions in the South Pacific Ocean. It is shown that although solutions are sensitive to changes in the mean temperature and salinity of each water mass, changes of these properties within realistic limits cannot lead to large changes in the mass fluxes or flow reversals.

The effect of changes in the ocean-atmosphere fluxes north of 30°S on the oceanic mass transports is evaluated. In the South Atlantic Ocean, reduction of the excess evaporation would significantly reduce the required volume fluxes whereas in the South Pacific Ocean, reduction of the volume fluxes and reversing the flow direction in the Antarctic Intermediate Water requires reduction of both heat and freshwater fluxes into the ocean. The role of a possible flow of Pacific Ocean waters into the Indian Ocean at equatorial latitudes, through the Southeast Asian Seas, and its effects on the Pacific and Indian Oceans' mass, heat and freshwater budgets are evaluated. Such a flow would strongly relax the requirement of southward freshwater flux at 30°S in the South Pacific Ocean.

1. Introduction

Using simple mass, heat and salinity conservation statements for three water masses with specified temperatures and salinities, Stommel (1980) calculated the oceanic meridional volume transports within each water mass across 30°S required to balance the heat and freshwater exchanges with the atmosphere north of that latitude. The so-called 3-point model is based on the following system of equations:

$$\left.\begin{aligned} m_1+m_2+m_3+F&=0\\ m_1T_1+m_2T_2+m_3T_3+H&=0\\ m_1S_1+m_2S_2+m_3S_3&=0\end{aligned}\right\} \qquad (1)$$

where m_i are the unknown net volume transports across 30°S within each water mass of temperature T_i and salinity S_i. The volume transport m_1 refers to mid-thermocline waters, m_2 to the intermediate salinity minimum water, and m_3 to the deep water. These water masses include the primary features of the temperature-salinity structure at 30°S. The heat and freshwater fluxes from the atmosphere to the ocean north of 30°S are taken from Hastenrath (1980, 1982) and Baumgartner and Reichel (1975), respectively. The rather large volume transports obtained by Stommel, particularly those in the Pacific Ocean deepest layer [m_3 of well over 100 Sv (100×10^6 m^3 s^{-1}) where 1 Sv $\equiv 10^6$ m^3 s^{-1}] and intermediate layer (m_2 of 80 Sv), are unrealistic. For example, from geochemical tracers, the residence time for the Pacific Ocean is estimated as approximately 510 years (Stuvier et al., 1983). However, an inflow of 130 Sv would reduce the residence time in this ocean to 130 years. Stommel (1980) suggests that a simple way of reducing the large volume fluxes, particularly in the South Pacific, is by increasing precipitation in the Atlantic and reducing it in the Pacific Ocean, as would be the case for a more globally uniform equatorial precipitation.

Gordon and Piola (1983) have used the Baumgartner and Reichel estimates to force a box model for the upper layer (water less dense than sigma-theta of 27.6) of the Atlantic Ocean. In the model, sinking of water in the northern North Atlantic associated with the formation and subsequent export of North Atlantic Deep Water is compensated by northward flow of upper-layer water. The observed salinity difference between upper-layer inflow at 30°S and its value in the northern North Atlantic, combined with the freshwater flux estimates, leads to a volume transport of 20 Sv. Since this formation rate of North Atlantic Deep Water is similar to values based on other methods, this result suggests that freshwater flux estimates over the Atlantic

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not be unreliable (see Gordon and Piola, 1983, for sensitivity study). In the Indian and Pacific Oceans however, similar box models lead to unreasonably large volume transports. These transports are significantly reduced if water within the upper layer is allowed to flow from the Pacific to the Indian Ocean through the Southeast Asian Seas (Piola and Gordon, 1984).

There are large discrepancies in the magnitude of the climatological meridional heat flux near 30°S (see Hsiung, 1985, Table 1). Because of nonavailability of time averaged, eddy resolving hydrographic sections along the southern boundaries of each ocean, comparison between direct estimates of meridional heat flux (Bennett, 1978; Fu, 1981, 1986; Toole and Raymer, 1985; Wunsch et al., 1983) and the climatological mean surface fluxes is difficult to interpret. The purpose of this paper is to introduce realistic modifications to Stommel's (1980) 3-point model which have the effect of reducing the magnitude of meridional volume transports across 30°S to more "acceptable" values. In section 2 the modifications of the 3-point model are given. The model results and sensitivity are presented in section 3. The effect of changing the H and F exchanges between the atmosphere and ocean are presented in section 4.

2. The model

The 3-point model was first presented by Stommel (1980) using temperature and salinity values typical of the Central, Intermediate and Deep waters at 30°S (Table 1). These values, combined with the estimated heat and freshwater exchanges with the atmosphere within each ocean north of 30°S, were then used to solve Eqs. (1) to determine the volume transports. In Table 2 the results arrived at by Stommel (1980) are summarized.

In order to investigate possible ways of reducing the oceanic volume transports, the following changes are introduced:

1) The water mass temperature and salinity values used by Stommel (1980) are replaced by mean temperature and salinity values averaged over potential density intervals characteristic of each water mass core within each ocean. These mean values (Table 3) are calculated from the gridded temperature and salinity data of Gordon and Baker (1982), between 30° and 35°S. Within this zone, temperature and salinity were averaged over potential density ranges characteristic of the Central, Intermediate and Deep waters.

TABLE 1. Temperature and salinity for each water mass (Stommel, 1980).

Layer	T (°C)	S (‰)
m_1	10	35
m_2	4	34
m_3	2	34.7

2) A flow of 1.5 Sv with temperature of 0°C and salinity of 33‰ is imposed to flow northward through the Bering Strait and into the Atlantic Ocean (Gordon and Piola, 1983).

3) An updated value of H in the Atlantic north of 30°S is used (Hastenrath, 1982).

4) It is noted that Hastenrath (1980, 1982) used 120°E as boundary between the Pacific and Indian Oceans while Baumgartner and Reichel (1975) give F in the Southeast Asian Seas (except for the Timor and Arafura Seas) as part of the Pacific Ocean. In order to obtain H over the same area represented by F, the value of H in the Southeast Asian seas west of 120°E (Wyrtki 1961, Table 2) was added to the Pacific and subtracted from the Indian Ocean.

3. Model results

The results obtained after solving Eqs. (1) for each ocean are given in Table 4.

What is the impact of specific modifications, as described in the previous section? First, the results of using the updated value of H in the Atlantic and the adjustment of F and H areas for the Indian and Pacific Oceans are inspected. In the Atlantic Ocean, decreasing the heat loss to the atmosphere from -275×10^6 °C m^3 s^{-1} (Hastenrath, 1980) to -165×10^6 °C m^3 s^{-1} (Hastenrath, 1982) acts to decrease the oceanic volume transports at 30°S by about 10 Sv in the upper and intermediate layers and about 20 Sv in the deep layer. The modification of the heat gained by the ocean, from 117 to 86 ($\times 10^6$ °C m^3 s^{-1}) in the Indian Ocean and from 458 to 489 ($\times 10^6$ °C m^3 s^{-1}) in the Pacific Ocean, produces smaller changes in the volume transports (see Table 4).

The flow from the Pacific into the Atlantic Ocean through the Bering Strait produces volume transport changes smaller than 5 Sv, more typically 1 Sv.

Introduction of temperature and salinity values derived from observations in the 30° to 35°S zone (Table 3) leads to a substantial decrease in the calculated volume transports (see Table 4; Figs. 1 and 2). The greatest range of temperature and salinity occurs in the upper layer. In the Atlantic, if the temperature T_1 is increased to 15°C, the m_1, m_2, m_3 volume transports are not greatly altered. A decrease of T_1 to 10°C leads to a small increase approximately (2 Sv) in m_1 and m_2 while m_3 increases by 8 Sv. Changes of the Intermediate Water temperature (T_2) produce larger changes in the solutions. Only a 1°C increase in T_2 causes an 8 Sv decrease in m_3. Because of the large temperature contrast between inflow (m_1, m_2) and outflow (m_3), the volume transports of all water masses decrease when T_2 increases. Temperature changes of 0.5°C in T_3 produce volume transport changes of less than 5 Sv.

Salinity changes have a stronger effect on the Atlantic solution. Figure 1b reveals that m_1 is not strongly affected by salinity changes. However, treating the At-

TABLE 2. Heat flux (H, relative to 0°C) and freshwater flux (F) through the sea surface north of 30°S and volume transports calculated by Stommel (1980). Values given in parentheses are calculated after introducing the updated value of H from Hastenrath (1982) for the Atlantic and after correction of H for different boundaries for the Pacific and Indian Oceans. (1 Sv ≡ 10^6 m^3 s^{-1}.)

Ocean	H (°C × 10^6 m^3 s^{-1})	F (Sv)	Volume transport (Sv) m_1	m_2	m_3
Atlantic	−275	−0.552	14	78	−92
	(−165)	(−0.552)	(3.6)	(67.5)	(−70.5)
Indian	117	−0.510	−24	36	−12
	(86)	(−0.510)	(−20.5)	(38.5)	(−17.5)
Pacific	458	0.500	−34	−92	125
	(489)	(0.500)	(−37.3)	(−94.6)	(131.5)

lantic as an inverted (or negative) estuary (Gordon and Piola, 1983), m_2 and m_3 are decreased markedly with decreasing the salinity of inflow water (S_1, S_2) and salinity increase of outflow water (S_3). An increase of 0.1‰ in S_3 produces a decrease of 15 Sv in m_2 and 10 Sv in m_3. A combined decrease in the salinity of the intermediate waters (S_2) to 34.2‰ with an increase in the salinity of the deep water (S_3) to 34.9‰ lead to a solution: 12, 30, −44 Sv, for m_1, m_2 and m_3, respectively.

In the Pacific Ocean the 3-point model solutions are somewhat more sensitive to changes in the T and S values (Fig. 2a and 2b). However, volume transport changes greater than 10 to 15 Sv require rather large and unrealistic changes in the T and S properties. Further reduction of the volume transports and reversing the flow direction in the Intermediate Water in the South Pacific requires modification of the surface fluxes H or F.

4. Surface fluxes

The uncertainties in the magnitude of the heat and freshwater fluxes through the sea surface are large (see Hsiung, 1985, for heat flux comparisons). In this section, the effect of changes in the surface fluxes of heat and freshwater north of 30°S is quantified and discussed. This effect is best displayed by giving the volume fluxes (m_i) in each ocean as a function of the net heat and freshwater fluxes through the sea surface north of 30°S (Fig. 3a–c). In Fig. 3 we have included the heat flux estimates of Hsiung (1985) based on surface marine observations for the period 1946–79. Because Hsiung used 120°E as a western boundary in the North Pacific, we have also revised her estimate to include the Southeast Asian seas in the Pacific Ocean budget.

In the Atlantic Ocean (Fig. 3a) reduction of the excess evaporation (along a constant H Line) will lead to a reduction in m_2 and m_3 with a small increase in m_1. This situation is perhaps more acceptable in terms of water mass flux. Thus accepting Hastenrath's H value, one suspects the net excess of evaporation over precipitation for the Atlantic north of 30°S given by Baumgartner and Reichel (1975) is an overestimate, as suggested by Stommel (1980).

In the Indian Ocean (Fig. 3b) solution volume transports for the present estimates of H and F are not excessively large (see Table 4). In order to balance the rather large excess evaporation, there is a northward flow in the intermediate water (salinity minimum) and outflow in the upper and deep waters. The heat flux into the Indian Ocean north of 30°S estimated by Hsiung (1985) is about three times larger than Hastenrath's. Using Hsiung's value of H (~309 × 10^6 °C m^3 s^{-1}) leads to an inflow of 20 Sv of deep water and outflow in the upper layer (Fig. 3b).

In the Pacific Ocean (Fig. 3c), a significant reduction of volume transports in the Deep Water to values smaller than 50 Sv would require reversing the sign of F or reducing H to about one-half of its value. Note that, after correction for different boundaries, Hsiung (1985) has estimated that the net heat flux into the Pacific Ocean north of 30°S is 87 × 10^6 °C m^3 s^{-1} or about 18% of Hastenrath's (1982) estimate. One particularly discouraging aspect of the Pacific solution is

TABLE 3. Mean temperatures (°C) and salinities (‰) for the 3-point model.

Ocean	m_1 T	m_1 S	m_2 T	m_2 S	m_3 T	m_3 S
Atlantic	13.27	35.29	4.07	34.32	2.83	34.86
Indian	13.38	35.32	4.73	34.43	1.78	34.76
Pacific	11.20	34.78	5.20	34.35	1.54	34.70
Bering Strait	0	33.00				
Pacific-Indian	22	33.60				

TABLE 4. Heat (H, relative to 0°C) and freshwater fluxes (F) through the sea surface north of 30°S and calculated mass transports through 30°S. Values given in parentheses are calculated with flow through the Bering Strait. Values in brackets are calculated after imposing a flow of 14 Sv, 22°C and 33.6‰ from the Pacific to the Indian Ocean.

			Volume transport (Sv)		
Ocean	H (°C × 10^6 m^3 s^{-1})	F (Sv)	m_1	m_2	m_3
Atlantic	−165	−0.552	10.4	43.9	−53.8
			(11.4)	(39.5)	(−51.8)
Indian	86	−0.510	−14.8	28.6	−13.4
			[−23.1]	[−34.7]	[44.2]
Pacific	489	0.500	−29.4	−55.8	84.7
			(−32.2)	(−49.1)	(82.3)
			[−20.2]	[−2.4]	[37.6]

the requirement of a strong southward flow in the intermediate layer (m_2 of −50 Sv). This flow is a consequence of the large excess precipitation in the Pacific. Because the Intermediate Water temperature (T_2) in the Pacific is close to the average temperature at 30°S, its volume flux is relatively insensitive to changes in the heat exchange to the atmosphere. Therefore, in order to reduce the flow in the deep water and reverse the flow in the intermediate water, the alternative of reversing the sign of F appears to be the better choice. It should be noted, however, that this sign reversal will lead to substantial increase in the southward flow of the upper layer, from $m_1 = -32$ Sv (Table 4) to $m_1 > 50$ Sv. Thus, "correction" of the Pacific values cannot be easily handled with alteration of T, S, H or F. Another possibility needs to be explored, as now follows.

5. Discussion

Simple budget models, as those presented here, have obvious shortcomings. What is the contribution of eddy fluxes and horizontal recirculations on the heat and freshwater fluxes? Answers to these questions can be provided in part by studies such as Bennett's (1978) for the heat flux or inverse calculations such as those of Wunsch et al. (1983) and Fu (1986). Part of the heat gain and freshwater loss to the atmosphere north of 30°S in the Indian Ocean could be balanced within the upper layer by southward flow of warm, salty waters of subtropical origin and northward flow of relatively cold and fresh subantarctic waters. Similarly, eddies could transfer heat and freshwater downgradient (poleward), thus contributing to reduce the flows in the 3-point model. However, in an ocean with excess pre-

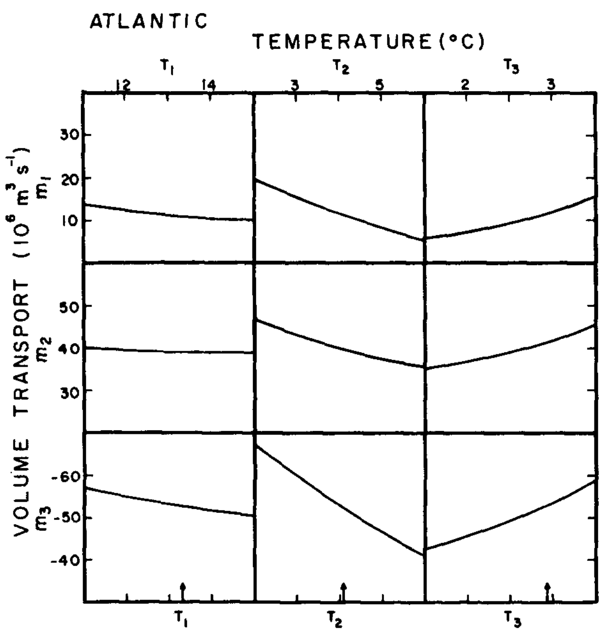

FIG. 1a. Sensitivity of the 3-point model solutions to changes in the temperatures in the Atlantic Ocean. The arrows at the base indicate the averaged T values calculated from Gordon and Baker (1982) grid data and used to obtain the solutions given in Table 4.

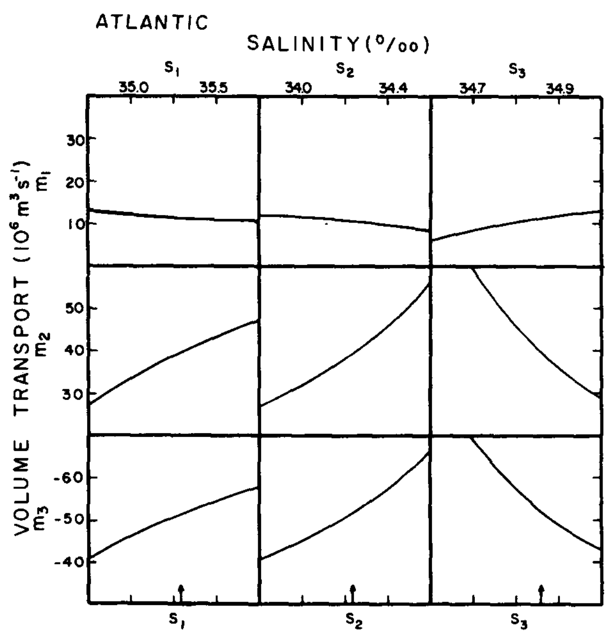

FIG. 1b. As in Fig. 1a but for changes in the salinity in the Atlantic Ocean.

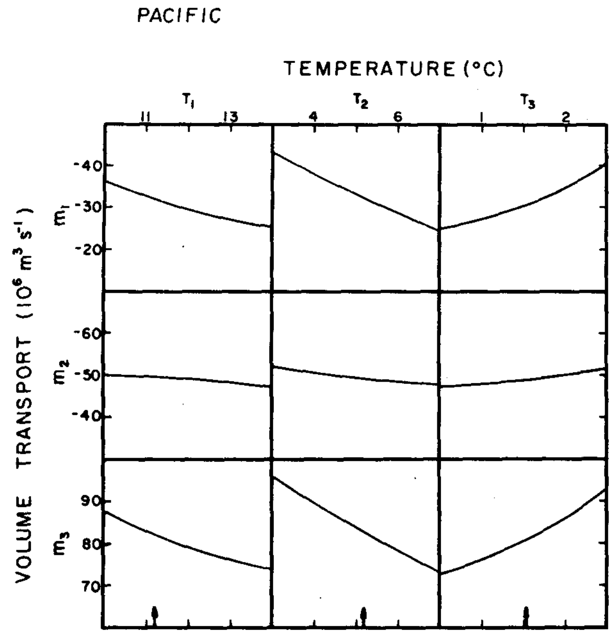

FIG. 2a. As in Fig. 1a but for changes in the temperature in the Pacific Ocean.

cipitation north of 30°S, such as the Pacific, outflow of relatively low salinity and inflow of relatively high salinity water is required, no matter what mechanisms are responsible. Horizontal recirculations within the upper layer, where a sufficiently large salinity contrast

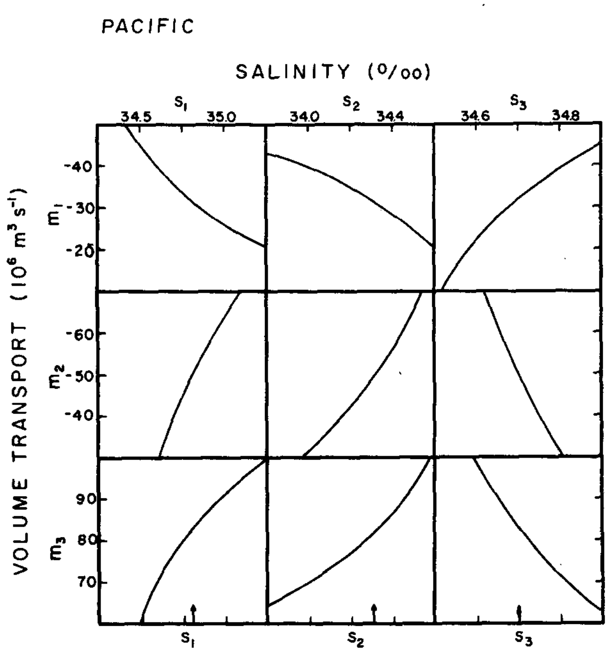

FIG. 2b. As in Fig. 1a but for changes in the salinity in the Pacific Ocean.

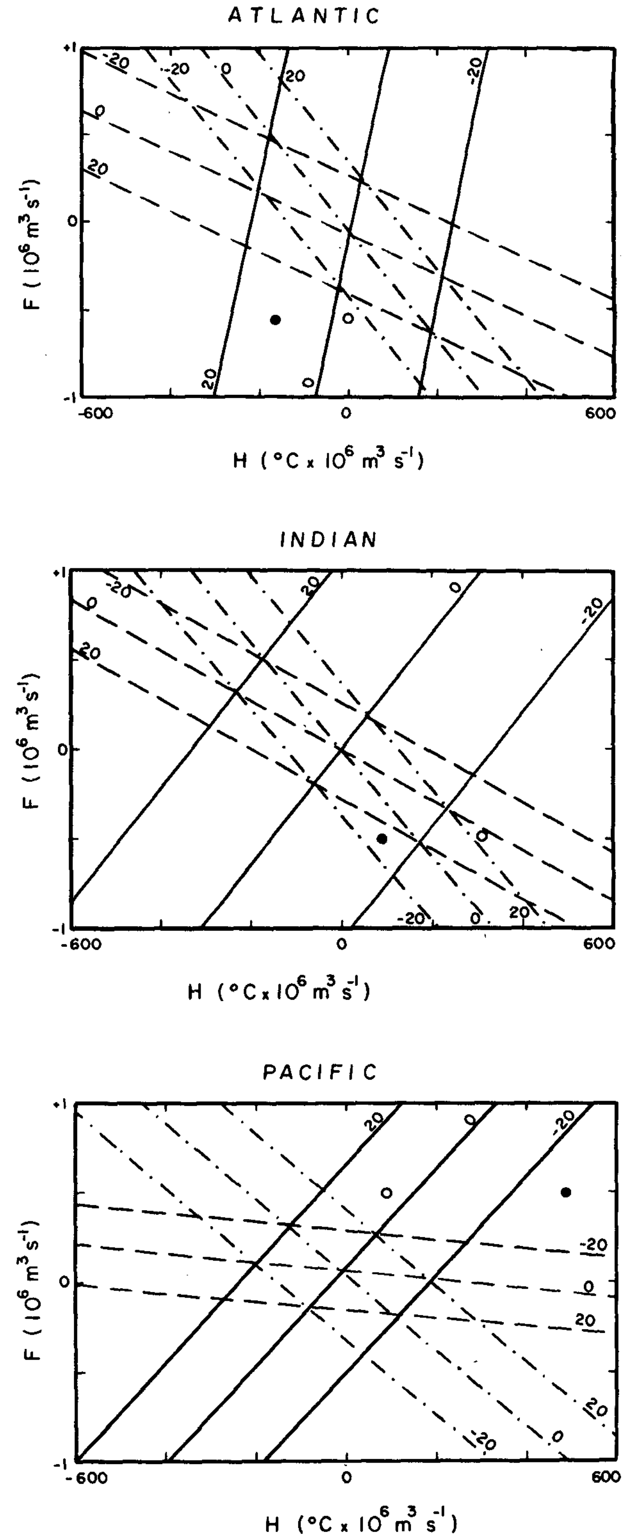

FIG. 3. Volume transport solutions as a function of the heat (H) and freshwater (F) fluxes through the sea surface: (a) for the Atlantic, (b) for the Indian and (c) for the Pacific Oceans. The solid line indicates transports for m_1, the dashed line for m_2 and the dashed-dotted line for m_3. The solid dot indicates the value of H from Hastenrath (1982) and F from Baumgartner and Reichel (1975). The open dot indicates H from Hsiung (1985) and F from Baumgartner and Reichel (1975).

exist, are unlikely to achieve the freshwater balance in the South Pacific since high salinity water is originated at lower latitudes and low salinity water is originated at higher latitude.

Piola and Gordon (1984) have suggested that a flow from the Pacific to the Indian Ocean at tropical latitudes through the Southeast Asian Seas might be a way of redistributing the freshwater surplus of the equatorial Pacific into the Indian Ocean. Flow from the Pacific into the Indian Ocean in this region has been recently estimated at rates of 10 Sv (Godfrey and Golding, 1981) and 6.6 Sv (Fu, 1986). Such a flow would strongly relax the transport dilemma in the Pacific. Based on mass and freshwater conservation, Piola and Gordon estimated a flow of 14 Sv, 33.6‰ from the Pacific into the Indian Ocean. The effect of such a flow in the 3-point model is studied by imposing a flow of this characteristic and 22°C in the Pacific and Indian Ocean models. Because this flow would occur at low latitudes, where temperature is higher and salinity lower than at 30°S, it will introduce heat into the Indian Ocean and remove freshwater from the Pacific Ocean. Calculated values of volume transport for each ocean and each water mass are given in Table 4. The Pacific–Indian exchange acts to substantially decrease the volume transport in the deep layer in the Pacific from 80 to less than 40 Sv and in the intermediate layer from −50 to −2 Sv. In the Indian Ocean the Pacific–Indian equatorial link acts to reverse and increase the volume transport in the deep layer and reverse the flow direction from northward to southward in the intermediate layer.

The hypothetical flow through Southeast Asian Seas would occur at substantially higher temperature (>20°C) and lower salinity (33.5‰) than the averaged T and S values at 30°S; it is expected to be significant in interocean redistribution of heat and freshwater. It is therefore useful to evaluate the contributions of an interoceanic exchange to H and F in each ocean. From temperature and mass conservations, these contributions are

$$F_I = m(S_I - S_{30})/S_{30}$$

$$H_I = m(T_{30} - T_I) \qquad (2)$$

where T_{30} and S_{30} are the average temperature and salinity of the flow at 30°S, m, T_I and S_I are the volume transport, temperature and salinity of the exchange through the Southeast Asian Seas. F_I and H_I are the freshwater and heat fluxes required to convert Southeast Asian Sea exchange water to 30°S water.

Inspection of the 3-point model solution for a throughflow of 14 Sv (Table 4) suggests that the net inflow into the Pacific Ocean occurs in the deep layer, whereas outflow at 30°S in the Indian Ocean occurs in the upper and intermediate layers. If the averaged temperature and salinity at 30°S in Eqs. (2) are replaced by the temperature and salinity of the deep water in the Pacific and the temperature and salinity of the upper layer of the Indian Ocean, we can estimate for each ocean the amount of heat (H_I) and freshwater (F_I) required to convert 30°S water to throughflow water. Toole and Raymer (1985) have presented graphically the results of a similar calculation. Their Fig. 3b suggests that if the salinity difference between throughflow waters and 30°S waters is of order 0.5‰ (probably a lower bound for the mean salinity difference), even if the transoceanic flow is reduced to 2 Sv (a value close to that estimated by Wyrtki, 1961) it will cause a freshwater flux into the Indian Ocean greater than 0.2 Sv or about 30% of the estimated net freshwater loss to the atmosphere north of 30°S in that ocean.

Estimates of the meridional mass, heat and freshwater transports at 32°S in the Indian Ocean based on the inverse method (Fu, 1986) indicate that the heat and freshwater budgets are relatively insensitive to the flow through the Southeast Asian Seas. Given the low salinity characteristic of the upper layer in the Southeast Asian Seas (33.6‰), our calculations based on Eq. 2 lead to values close to those estimated by Toole and Raymer (1985).

6. Summary and conclusions

Meridional oceanic volume transports at 30°S required to balance the net heat and freshwater fluxes through the sea surface have been calculated. Present estimates of the net heat and freshwater fluxes through the sea surface in the Atlantic and Indian Oceans lead to meridional transports of about 50 Sv and 30 Sv, respectively. Although somewhat large, given the uncertainties in the surface fluxes, these transports and the associated flow directions of each water mass are acceptable. In the Pacific Ocean, however, the solution is unrealistic, with transports greater than 80 Sv for the deep water and southward flow (opposite to the flow direction inferred from property distributions) at the intermediate salinity minimum. The difficulty in the South Pacific arises from the large excess precipitation estimated in this ocean (Baumgartner and Reichel, 1975). This excess precipitation requires the inflow of relatively high salinity water and outflow of relatively low salinity water.

The sensitivity of the 3-point model solution to changes in the temperature and salinity of the water masses and to changes in the surface fluxes was examined. Changes in T and S within realistic limits in the Pacific could somewhat reduce the deep water inflow but are unlikely to produce a flow reversal in the Antarctic Intermediate Water. It is apparent that a substantial reduction of the freshwater flux, from excess precipitation to excess evaporation, is required in order to reverse the flow direction in the intermediate water. Figure 3c also suggests that in order to further decrease the deep water inflow, a reduction in the heat flux into the ocean is required. Given the uncertainties in the magnitude and sign of the net heat flux into the Pacific Ocean north of 30°S (see Hsiung, 1985) this possibility cannot be ruled out. Introduction of Hsiung's estimate of heat flux at the sea surface in the Pacific leads to a substantial reduction of deep water inflow (to less than 40 Sv), which is balanced primarily by outflow of intermediate water. The transport of the latter is relatively insensitive to changes in the heat flux at the sea surface and its reduction or reversal requires a substantial reduction of F allowing for interocean exchange north of 30°S.

The role of a possible equatorial link between the Pacific and Indian Oceans through the Southeast Asian Seas was examined. This flow would act to strongly relax the required freshwater flux through 30°S in the South Pacific and would introduce heat and freshwater into the Indian Ocean, which would lead to a flow reversal in the deep and intermediate water masses in this ocean. It is recognized, however, that in an ocean with excess heat and evaporation north of 30°S, other mechanisms, such as horizontal recirculations and eddy fluxes, can contribute to balance the fluxes through the sea surface.

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