

## Vertical Mixing in the Indonesian Thermocline\*

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(Manuscript received 16 January 1991, in final form 10 June 1991)

### ABSTRACT

Western Pacific central and tropical waters characterized by a subsurface salinity maximum spread into the Indonesian seas as part of the Indonesian throughflow. Within the Indonesian seas this salinity maximum is attenuated and, in some places, completely removed. A simple advection-diffusion model verifies the importance of vertical mixing in the transformation of Western Pacific waters to Indonesian thermocline water. The profiles indicate a predominant North Pacific presence in most of the seas, although some South Pacific water is present in the eastern seas of Halmahera, Seram, and Banda. The main interocean route is through the western seas of Sulawesi, Makassar, and Flores, while the flow pathway in the eastern seas is less certain. The Banda Sea can be renewed from either the northern passages (Halmahera and Maluku) or from the south via the Flores Sea. Using representative basin property profiles derived from the archived data allows determination of a range of vertical diffusivities and residence times that best reproduce the transformation of Pacific waters into Indonesian water. In the Makassar thermocline a lower limit of  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  for vertical diffusivity is inferred from the model results with reasonable throughflow and precipitation values. This estimate is roughly an order of magnitude greater than those deduced for the interior oceanic thermocline in an environment not conducive to salt fingers. In the Banda Sea a  $K_z$  of  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  implies a predominant North Pacific source. If  $K_z$  is higher, then a larger South Pacific presence is possible.

### 1. Introduction

Western Pacific central and tropical waters, called subtropical lower waters, are characterized by a shallow salinity maximum (Wyrtki 1961). Some of this Pacific water advects into the Indonesian seas as part of the Indonesian throughflow, where the salinity maximum is attenuated and, in some places, completely removed (Gordon 1986, Fig. 1). To parameterize the effective cross-isopycnal salinity flux necessary to modify the western Pacific salinity gradients to the Banda Sea gradients, Gordon (1986) estimates a vertical diffusivity ( $K_z$ ) of  $3 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  within the Indonesian thermocline. In a region not conducive to salt fingers, this estimate is more than an order of magnitude greater than those typically deduced for the interior oceanic thermocline by different means: for example, microstructure measurements, process-inspired parameterizations, or property distributions (Gregg 1987). However, dissipation of tidal energy along topographical boundaries may provide additional energy for mixing

in the Indonesian seas. In the deep east Indonesian basins, Van Aken et al. (1988) estimated  $K_z = 9 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  from an advection-diffusion model, while Berger et al. (1988) inferred  $K_z$  in excess of  $5 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$  on slopes and sills. These vertical diffusivities are much greater than those typically estimated for the deep interior ocean (Gargett 1984; Ledwell and Watson 1991). In addition, recent observations along an eastern sill in the Sulawesi Sea (Fig. 2) support such high  $K_z$  values in the Indonesian seas. These observations include isothermal layers between 300 and 1100 m, an obliterated oxygen maximum near  $8^\circ\text{C}$ , and a strongly modified salinity maximum between  $8^\circ$  and  $9^\circ\text{C}$  (Lukas et al. 1991).

Enhanced vertical mixing in the Indonesian seas implies an influence on the regional climate system: heat and freshwater are driven down into the oceanic thermocline, in turn affecting the radiative-convective equilibrium in the atmosphere. Enhanced vertical mixing may also be important for biological interactions and the carbon cycle, because nutrients are brought to the surface, fueling productivity. Thus, identifying and characterizing such diffusive heat and salt flux extremes in the oceanic thermocline is essential for understanding the link between ocean circulation and climate.

The aim of this paper is to quantify the vertical mixing in the thermocline of the Indonesian seas by means of a simple advection-diffusion model and archived

\* Lamont-Doherty Geological Observatory Contribution Number 4859.

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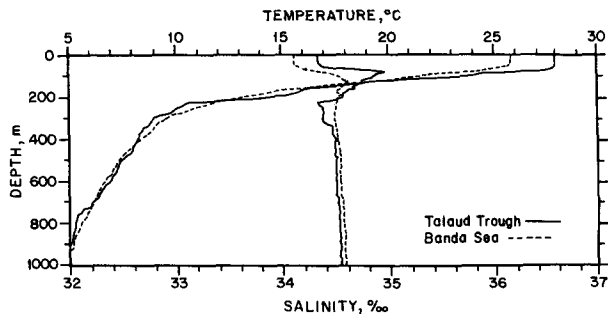


FIG. 1. Temperature and salinity profiles from two CTD stations (Indopac 1978) in the Talaud Trough (northern Maluku Sea; station 3: 3°45.0'N, 126°14.6'E) and the Banda Sea (station 10: 6°27.0'S, 126°0.2'E). A salinity maximum (Western North Pacific Central Water, 34.9‰) and a salinity minimum (North Pacific Intermediate Water, 34.42‰) are observed in the Talaud Trough profile, but not in the Banda Sea profile.

Pacific water entering the Indonesian seas. Using observed profiles and vertical diffusivity estimates, we calculate the residence time necessary to account for observed changes in salinity gradients. The results provide evidence for enhanced vertical mixing in the thermocline.

2. The archived data

The archived data from the National Oceanic Data Center (NODC) was used for this study. As of May 1988, the NODC dataset included 14 356 bottle and low-resolution CTD (conductivity, temperature, depth) stations over the region 20°N to 20°S and 100° to 150°E. A data subset was selected by requiring that each retained station have oxygen data, a minimum of five observations, and measurements deeper than 100 meters. This procedure produces a subset of 4733 stations with approximately 1000 stations within the Indonesian seas. Most of the Indonesian stations fall between the years 1929 and 1981 and were acquired by a host of different countries: Australia, Indonesia,

data. The dataset—which includes profiles of salinity, temperature, and oxygen—also is used to determine the relative contributions of North Pacific and South

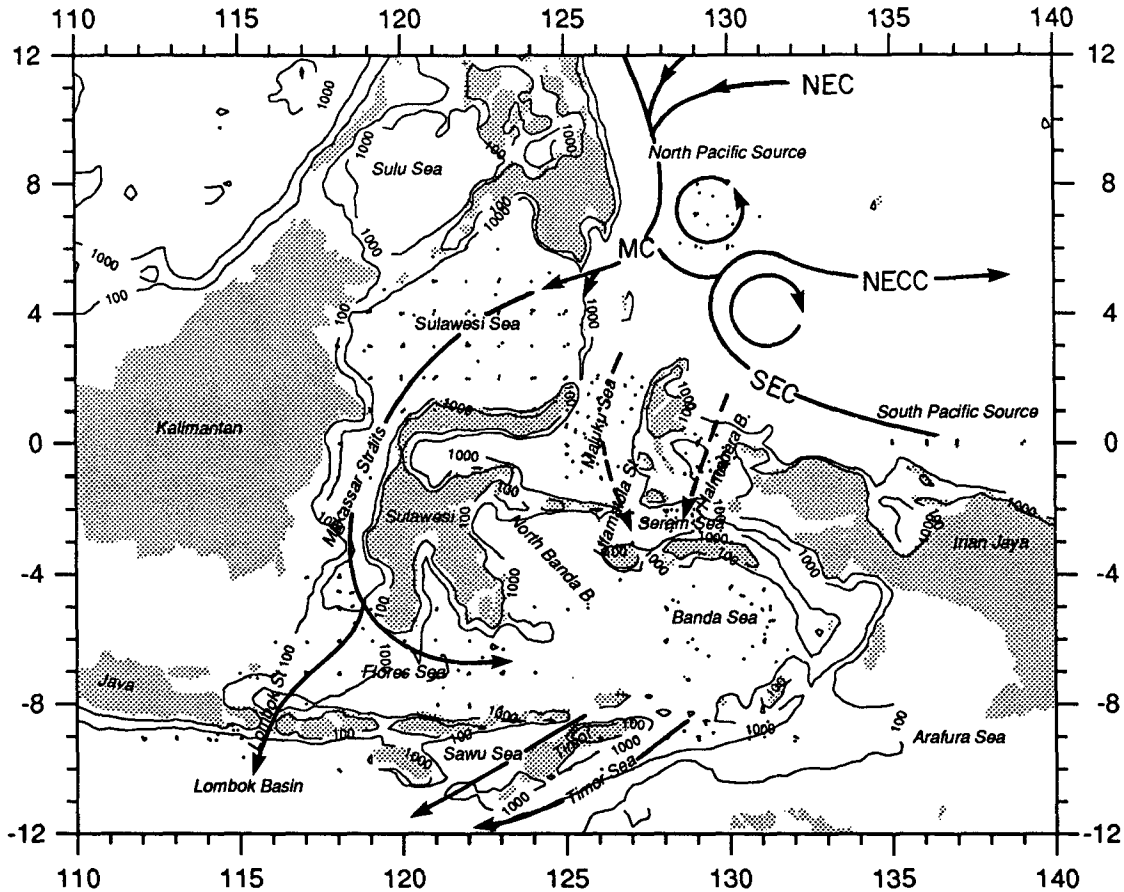


FIG. 2. Map of the Indonesian region with bathymetry contoured at 100 and 1000 m. The North Equatorial Current (NEC), Mindanao Current (MC), South Equatorial Current (SEC), and North Equatorial Counter Current (NECC), derived from the Wepocs III cruise, are shown by solid arrows (Hacker et al. 1989, Fig. 1). The main throughflow is confined to the Makassar Strait (solid arrow) with secondary throughflows in the eastern seas (dashed arrows). The stations selected from the archived dataset for the representative basin profiles are shown as dots. Note that the stations used to represent the NP are east of the MC; however, their water properties are the same as those within the MC, and more data is available there.

United States, Netherlands, Japan, USSR, Denmark, and the Philippines.

Subset groups of stations were constructed for the major seas of the Indonesian archipelago: Sulawesi, Makassar, Flores, Maluku, Halmahera, Seram, and Banda (Fig. 2). The far western seas (Sulu, Cina Selatan, Jawa) and the far eastern seas (Arafura) are not considered here because their sills are less than 100 m deep. The western Pacific central and tropical waters forming the possible sources of the Indonesian throughflow may be separated objectively by their thermocline properties into two classes: North Pacific (NP) and South Pacific (SP). Stations thought to capture the most representative characteristics of the NP and SP water types were selected; the NP is best described by the Mindanao Current water around 7°N, 130°E, and the SP is exemplified by the South Equatorial Current water around 137°E at the equator. Stations with suspect data were deleted from the station groupings. The remaining data was averaged within standard level bins to form representative basin profiles. Each grouping was found to have a unique salinity and oxygen profile, distinct from adjoining basins (Fig. 3). However, the standard deviations of the basin profiles are fairly high, and the component data making up the averages is not well distributed in time or space. Therefore, the average basin profiles can only be considered approximate representations of the Indonesian seas.

### 3. North Pacific versus South Pacific inflow

Before presenting the model, it is worthwhile to first inspect the thermocline stratification in a more qualitative sense for the probable sources and paths of the throughflow. The NP and SP carry different properties, and they are easily distinguished by their salinity or oxygen profiles. The NP salinity maximum is 34.75‰ at 100 m, whereas the SP salinity maximum is 35.41‰ at 150 m (Figs. 3 and 4a). With the oxygen curves, the NP values gradually decrease from the surface to a minimum around 10°C [which Sverdrup et al. (1942) called the North Pacific Intermediate Water], while the SP oxygen values are nearly constant from 25° to 10°C (Figs. 3 and 4b). The amount of vertical mixing needed to transform the Pacific stratification into Indonesian thermocline water depends on the source. For example, the larger gradients associated with the SP salinity maximum require considerably more mixing (and freshwater input) than the NP to produce the same result. Because the potential temperature–salinity ( $\theta$ - $S$ ) and potential temperature–oxygen ( $\theta$ - $O_2$ ) characteristics of the NP and SP are very distinct, their signatures can be mapped to trace the sources of Indonesian throughflow.

#### a. Western seas

An NP source is suspected for the western seas of Sulawesi, Makassar, and Flores, because there is only

minor modification from the original NP  $\theta$ - $S$  and  $\theta$ - $O_2$  curves (Fig. 3). Tritium evidence supports an overall NP source for the Indonesian throughflow, because the high tritium values of the NP source are apparently required to account for high values found in the Indian Ocean (Fine 1985). The only discernible difference between the properties in these seas and the original NP profile is that the gradients of the salinity maximum are slightly reduced and the surface layers are fresher, especially in the Flores Sea. These relatively slight changes imply a short travel time, a conclusion consistent with evidence from surface currents (Wyrski 1961), drifters (Hacker et al. 1989), and models (Kindle et al. 1989). They all show a portion of the Mindanao Current entering the Sulawesi Sea, then traveling through the Makassar Strait. In addition, the NP salinity maximum can be traced through the western seas and into the Banda with gradual freshening and lower oxygen values (Fig. 4). From the Flores Sea, a portion of the throughflow leaves the Indonesian seas through the Lombok Strait. There, current meter measurements revealed an annual mass transport of  $1.7 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  with a maximum flow of  $4 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  during August 1985 (Murray and Arief 1988). Within the Flores Sea NP water may also pass to the Indian Ocean by way of the Banda Sea (discussed in the next section).

#### b. Eastern seas

In the eastern seas (Halmahera, Seram, Maluku, and Banda) the origin and circulation of the thermocline water is less certain, despite the comprehensive oceanography studies by the *Sneillius I* expedition (Lek 1938), Wyrski (1961), and the *Sneillius II* expedition (Van Aken et al. 1988). Above 500 m, there are three possible throughflow entrances of Pacific water to the Banda Sea: direct pathways from the Pacific via the Halmahera and the Maluku seas, and a “back door” pathway from the western route via the Flores Sea, which implies an NP origin of Banda Sea water (arrows on Fig. 2 delineate the possible routes). An Indian Ocean origin can be ruled out, since there is no evidence of Indian Ocean water in the upper thermocline in the *Sneillius II* salinity sections (Van Aken et al. 1988). The Banda Sea  $\theta$ - $S$  curve in the thermocline is isohaline, with the Pacific salinity maximum completely removed (Figs. 1 and 3). Remnants of the low salinity thermocline similar to that of the Banda Sea can be found to stretch across the Indian Ocean from the Timor Sea (Gordon 1986).

By tracing the low oxygen values of the SP salinity maximum, Wyrski (1961, p. 112) argues that SP water spreads through the Halmahera Sea into the Banda Sea between 100 and 200 meters (this can be seen in Fig. 4b). The archived oxygen data does not necessarily support this suggestion, since the oxygen profiles can also be explained by vertical mixing within an NP source. Note that while at the salinity maximum the

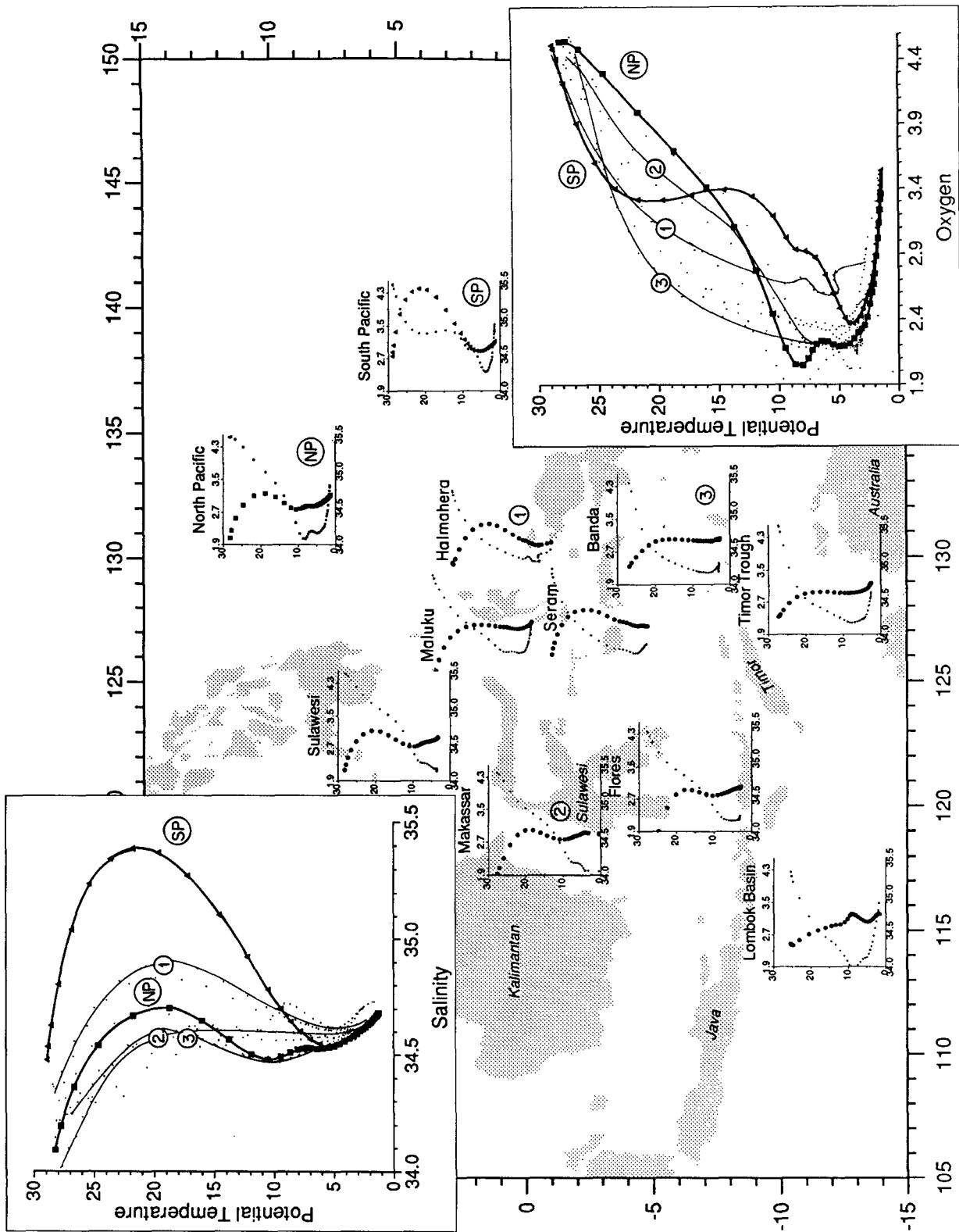


FIG. 3. Potential temperature-salinity ( $\theta-S$ ) and potential temperature-oxygen ( $\theta-O_2$ ) curves derived from the archived data are shown at the approximate location of the component data, and in a group  $\theta-S$  and  $\theta-O_2$  plot. The North Pacific (NP), South Pacific (SP), Halmahera (1), Makassar (2), and Banda (3) curves are labeled on the group plots.

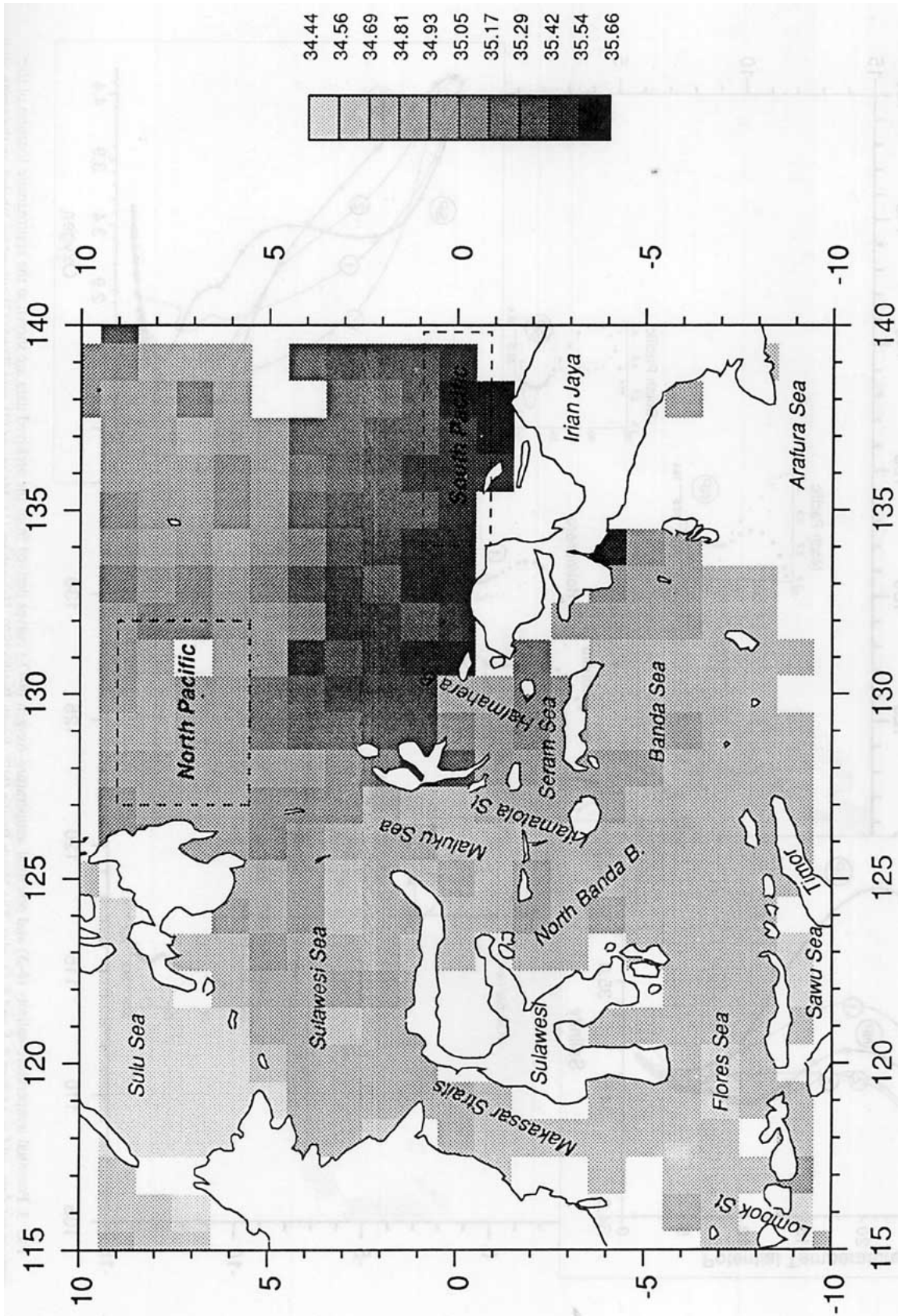


FIG. 4. (a) Salinity values (‰) of the upper salinity maximum from the full archived dataset. High salinity values designated by darker shading are found only north of Irian Jaya in the region of South Pacific thermocline water. Gradual freshening of North Pacific thermocline water can be traced through the Sulawesi, Makassar, and Flores seas. (b) Oxygen values (ml l<sup>-1</sup>) of the upper salinity maximum from the full archived dataset. High oxygen values designated by darker shading are found only in the western seas in the region of North Pacific thermocline water.

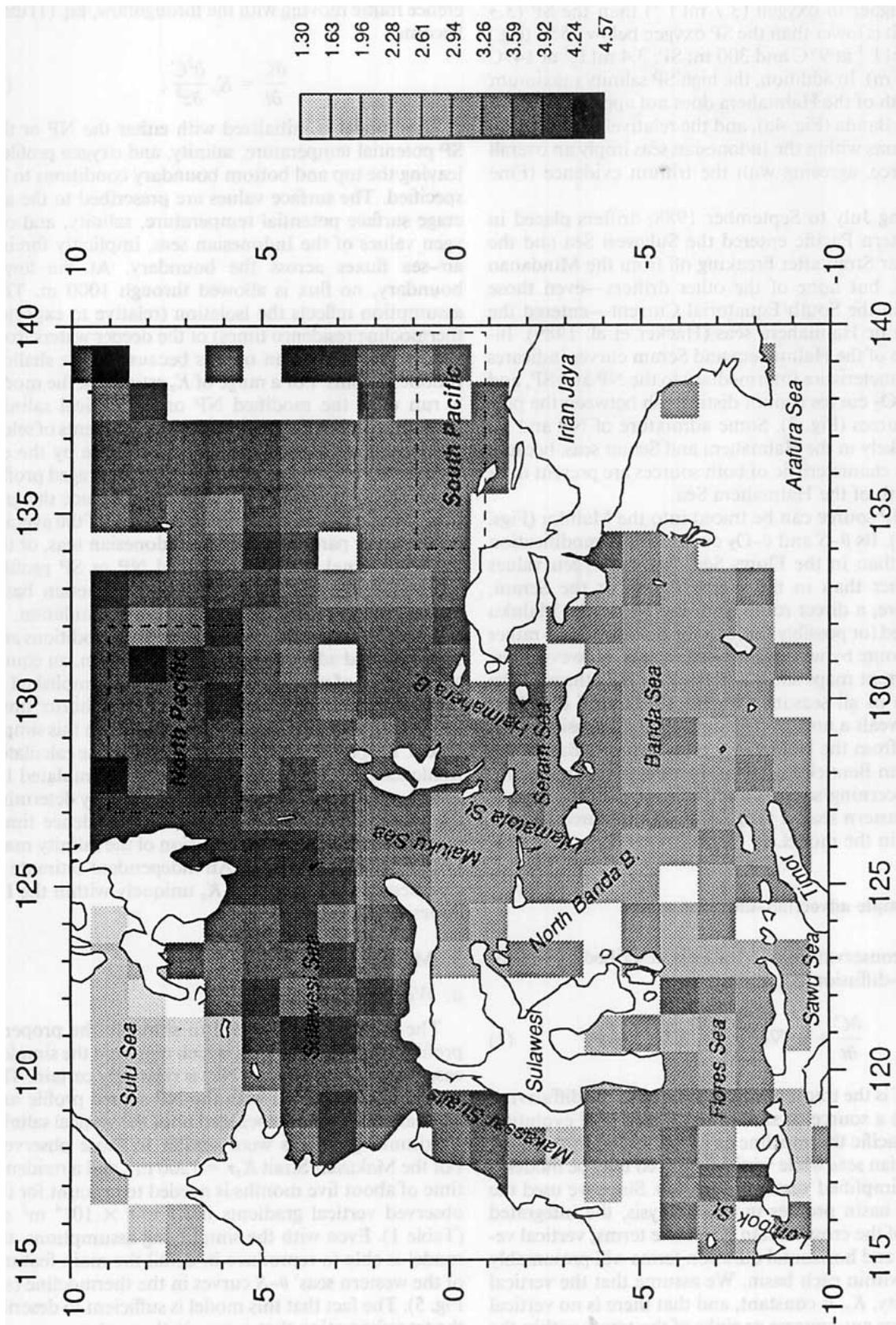


FIG. 4. (Continued)



NP is higher in oxygen ( $3.7 \text{ ml l}^{-1}$ ) than the SP ( $3.3 \text{ ml l}^{-1}$ ), it is lower than the SP oxygen below  $15^\circ\text{C}$  (e.g., NP:  $2 \text{ ml l}^{-1}$  at  $9^\circ\text{C}$  and  $300 \text{ m}$ ; SP:  $3.4 \text{ ml l}^{-1}$  at  $14^\circ\text{C}$  and  $300 \text{ m}$ ). In addition, the high SP salinity maximum just north of the Halmahera does not appear to spread into the Banda (Fig. 4a), and the relatively low salinity maximums within the Indonesian seas imply an overall NP source, agreeing with the tritium evidence (Fine 1985).

During July to September 1988, drifters placed in the western Pacific entered the Sulawesi Sea and the Makassar Strait after breaking off from the Mindanao Current, but none of the other drifters—even those placed in the South Equatorial Current—entered the Maluku or Halmahera seas (Hacker et al. 1989). Inspection of the Halmahera and Seram curves indicates  $\theta$ - $S$  characteristics intermediary to the NP and SP, and their  $\theta$ - $\text{O}_2$  curves cannot distinguish between the possible sources (Fig. 3). Some admixture of NP and SP seems likely in the Halmahera and Seram seas, because features characteristic of both sources are present in or just north of the Halmahera Sea.

An NP source can be traced into the Maluku (Figs. 3 and 4). Its  $\theta$ - $S$  and  $\theta$ - $\text{O}_2$  curves imply modification greater than in the Flores Sea, but its oxygen values are higher than in the North Banda or the Seram. Therefore, a direct route from the NP to the Maluku is implied (or possibly through the Sulawesi first), rather than a route by way of the western seas. However, surface current maps show a northward flow through the Maluku in all seasons (Wyrki 1961), and dissolved silica reveals a northward flow at  $500 \text{ m}$ , possibly originating from the Halmahera or as part of a Banda Sea gyre (Van Bennekom 1988). In view of the uncertainties concerning sources and paths of the throughflow in the eastern seas, a range of possible sources is considered in the model.

#### 4. A simple advection-diffusion model

The conservation of a tracer is described by the advection-diffusion equation,

$$\frac{\partial C}{\partial t} + \mathbf{u} \cdot \nabla C = \nabla \cdot (K \nabla C) + F \quad (1)$$

where  $C$  is the tracer concentration,  $K$  is the diffusivity, and  $F$  is a source or sink of the tracer. The evolution of the Pacific thermocline as it is advected through the Indonesian seas while vertically mixed may be modeled with a simplified version of Eq. (1). Since we used the average basin profiles in this analysis, the integrated effects of the cross-stream advective terms, vertical velocities, and horizontal diffusion terms will presumably cancel within each basin. We assume that the vertical diffusivity,  $K_z$ , is constant, and that there is no vertical shear, nor any sources or sinks of the tracer within the water column. With these assumptions and in a ref-

erence frame moving with the throughflow, Eq. (1) then becomes

$$\frac{\partial C}{\partial t} = K_z \frac{\partial^2 C}{\partial z^2}. \quad (2)$$

This model is initialized with either the NP or the SP potential temperature, salinity, and oxygen profiles, leaving the top and bottom boundary conditions to be specified. The surface values are prescribed to the average surface potential temperature, salinity, and oxygen values of the Indonesian seas, implicitly forcing air-sea fluxes across the boundary. At the lower boundary, no flux is allowed through  $1000 \text{ m}$ . This assumption reflects the isolation (relative to expected thermocline residence times) of the deeper waters from the Pacific and Indian oceans because of the shallow Indonesian sills. For a range of  $K_z$  estimates, the model is run until the modified NP or SP vertical salinity maximum gradients are similar to the gradients of select Indonesian seas. "Similar" is defined here by the estimated accuracy of the model and the averaged profile data. The simulated time required to reduce the gradients is noted (this is the model estimate of the average time a water parcel stays in the Indonesian seas, or the residence time), and the modified NP or SP profiles are then compared visually to the Indonesian basin profiles to determine the quality of the simulation.

With the time-independent boundary conditions and the simplified advection-diffusion equation, an equivalent degree of vertical mixing can be accomplished in two ways: by either using a larger  $K_z$  for a shorter time, or by using a smaller  $K_z$  for a longer time. In this simple model, the product of the  $K_z$  estimate and the calculated residence time,  $\tau$ , is a constant for each simulated Indonesian sea. Therefore, the model can only determine the range of vertical diffusivities and residence times that are consistent with the erosion of the salinity maximum by vertical mixing. An independent estimate of  $\tau$  is then needed to specify  $K_z$  uniquely within the Indonesian seas.

#### 5. Model results

##### a. Western seas

The model was first used to simulate the property profiles in the western seas, which represent the simplest cases because the source (NP) is relatively certain. The model was initialized with the NP source profile and vertical mixing then proceeded until the vertical salinity maximum gradients were similar to those observed. For the Makassar Strait  $K_z \tau = 1200 \text{ m}^2$ , and a residence time of about five months is needed to account for the observed vertical gradients if  $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  (Table 1). Even with the simplifying assumptions, the model is able to reproduce in detail the main features of the western seas'  $\theta$ - $S$  curves in the thermocline (see Fig. 5). The fact that this model is sufficient to describe the transformation that occurs in the western seas supports the assumption that the  $\theta$ - $S$  stratification is dom-

TABLE 1. The advection-diffusion model results. The constant,  $K_z\tau$ , is characteristic of the amount of vertical mixing necessary to transform an NP or SP profile into an Indonesian basin profile. A larger vertical diffusivity,  $K_z$ , can be compensated by a shorter residence time,  $\tau$ . The mixing depth is the square root of  $K_z\tau$ .

Indonesia Sea	Pacific source	$K_z\tau$ (m <sup>2</sup> )	Mixing depth (m)	Residence time if $K_z = 1 \times 10^{-4}$ m <sup>2</sup> s <sup>-1</sup>
<i>Western Seas</i>				
Sulawesi	NP	600	24	2 months
Makassar	NP	1200	35	5 months
Flores	NP	1400	37	5 months
<i>Eastern Seas</i>				
Halmahera	50% NP-50% SP	1700	41	6 months
Maluku	NP	3100	56	1.0 year
Seram	50% NP-50% SP	3400	58	1.1 years
Banda	NP	5300	73	1.7 years
Banda	50% NP-50% SP	27 600	166	8.7 years
Banda	SP	53 600	232	17.0 years

inated by vertical mixing. However, it is apparent from the model  $\theta$ -O<sub>2</sub> curves that, as expected, an oxygen consumption term is necessary. The model  $\theta$ -O<sub>2</sub> curve is not very different from the initial NP curve, in contrast to the observed  $\theta$ -O<sub>2</sub> curve, where lower O<sub>2</sub> values

in the 15°-25°C stratum provide evidence that oxygen is consumed within the thermocline.

These results are not sensitive to the surface flux formulation (described in section 4), because in these seas the time necessary to reduce the vertical salinity gradients for a given  $K_z$  is less than the time it would take a surface signal to penetrate to the salinity maximum. With a smaller  $K_z$  the residence time would be larger, but the surface flux would still reach the same depth (the square root of  $K_z\tau$  is characteristic of this depth). The range of  $K_z$  and residence times that can duplicate the western curves is illustrated in Fig. 6a and catalogued in Table 1. Suggested by increasing residence times for a given  $K_z$  value, the NP water is spreading along the expected path: Sulawesi, Makassar, Flores.

*b. Eastern seas*

Allowing for the uncertainties in the source for the eastern seas, the model was run with various combinations of NP and SP profiles (Fig. 6b and Table 1). The Halmahera, Maluku, and Seram simulations reproduce the observed  $\theta$ -S curves well, but an oxygen consumption term is needed to simulate the  $\theta$ -O<sub>2</sub> curves. In this model, none of the Banda Sea simulations can completely reproduce the observed Banda Sea  $\theta$ -S or  $\theta$ -O<sub>2</sub> curves.

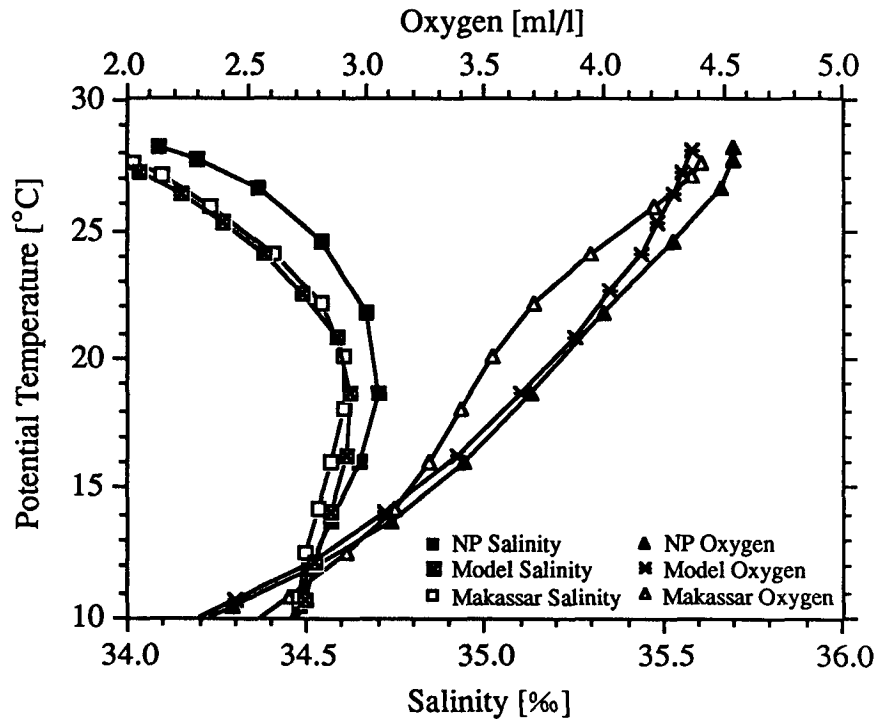


FIG. 5. The  $\theta$ -S and  $\theta$ -O<sub>2</sub> curves for the North Pacific, Makassar, and model. The model results are for a North Pacific source ( $K_z = 1 \times 10^{-4}$  m<sup>2</sup> s<sup>-1</sup> and a residence time of 5 months).



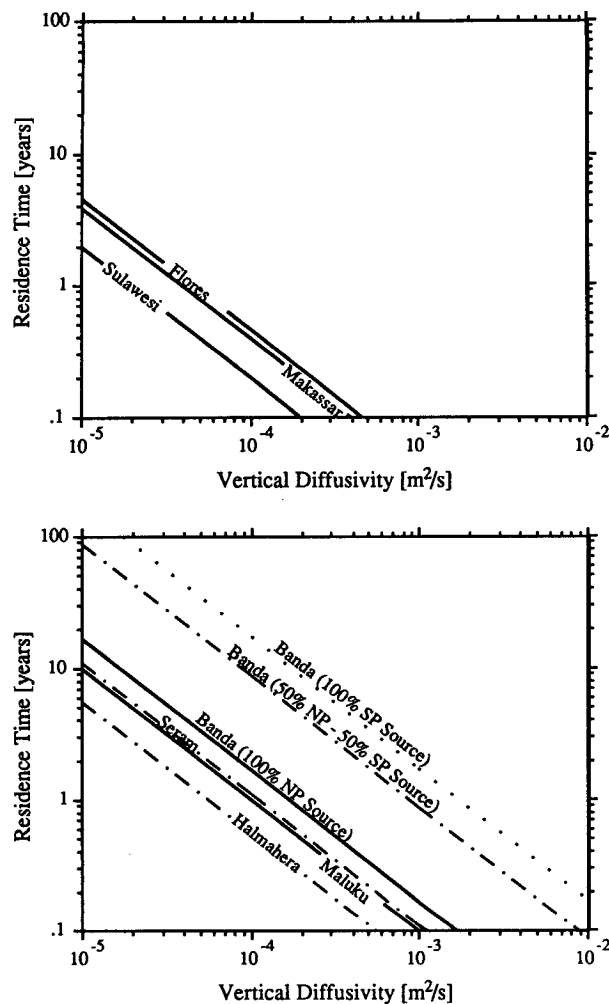


FIG. 6. (a) The range of  $K_z$  and residence times that can duplicate the  $\theta$ - $S$  curves of the western Indonesian seas (Sulawesi, Makassar, and Flores) when the model is initialized with the NP. The path of the throughflow is easily seen, as the basins farthest from the original source require longer residence times for equivalent values of vertical diffusivity. (b) The range of  $K_z$  and residence times that can duplicate the  $\theta$ - $S$  curves of the eastern Indonesian seas (Halmahera, Maluku, Seram, and Banda). The model is initialized either with the NP, SP, or a mixture. A solid line represents an NP source, a dotted line an SP source, and a dashed-dotted line a 50% NP-50% SP source. In the Banda Sea, a greater residence time or vertical diffusivity is necessary with increasing amounts of South Pacific source water.

In the Banda Sea  $K_z\tau = 5300 \text{ m}^2$  when the model is initialized with an NP source. Thus, when the model is run with  $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ , for example,  $\tau$  must be about two years to account for the observed changes in the Banda Sea vertical salinity maximum gradient. The  $\theta$ - $S$  structure of the model thermocline, in this case, is similar to the nearly isohaline Banda curve, but the salinity values are uniformly too low (Fig. 7). A source of salt is needed within the thermocline and the SP thermocline is the likely candidate. With exclusively an SP source,  $K_z\tau = 53 \text{ 600 m}^2$ , and a residence time of about 17 years is needed for the model to simulate

the Banda Sea salinity maximum gradients if  $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ . However, for a residence time of this length the model shows that the downward flux of low salinity surface water reaches well into the salinity maximum, which is not observed. With a smaller  $K_z$ , the residence time would be larger, and the surface flux would still penetrate to the same depth. A mix of 50% NP-50% SP (along isopycnal surfaces) requires a residence time of about nine years if  $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  and  $K_z\tau = 27 \text{ 600 m}^2$ . If the model profiles for this case were shown, they would be found intermediary to the NP and SP extremes in Fig. 7. For both an NP and SP source, the vertical diffusion necessary to reduce the salinity gradients for the Banda Sea also destroys the extremes of the oxygen curves (Fig. 8). Therefore, the oxygen curves cannot be used to distinguish between sources, and, in any case, oxygen consumption would be necessary to replicate the observed curves. While the simple model does not simulate the shape of the Banda Sea profiles well in the upper thermocline, the  $K_z\tau$  estimates can still be used to indicate the different degrees of vertical mixing necessary for transforming various source profiles.

Suggested by increasing residence times for a given  $K_z$  value, the model results are consistent with the throughflow entering the Banda Sea either from the Flores, Maluku, or Halmahera, or from a combination of these entrances (remembering that some SP water is needed in the Banda Sea to satisfy salinity requirements and that too much SP water requires too long a residence time).

### c. A Nonconstant $K_z$ and vertical shear

The sensitivity of these results to a depth-dependent  $K_z$  and vertical shear must be considered. To test the

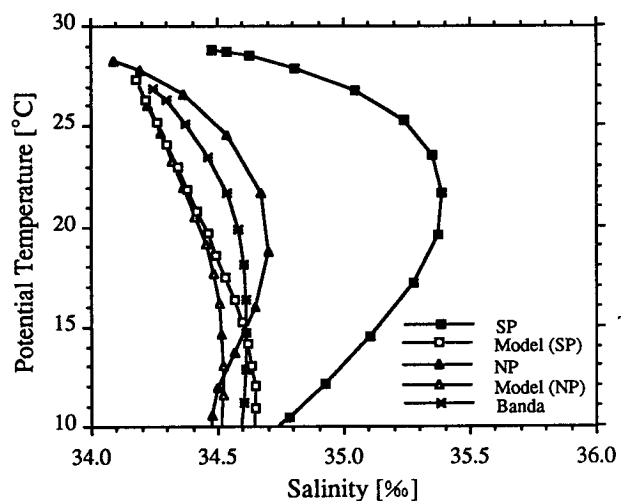


FIG. 7. The  $\theta$ - $S$  curves for the North Pacific, South Pacific, Banda, and model. The model results are for a North Pacific source ( $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  and a residence time of 1.7 years) and a South Pacific source ( $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  and a residence time of 17 years).

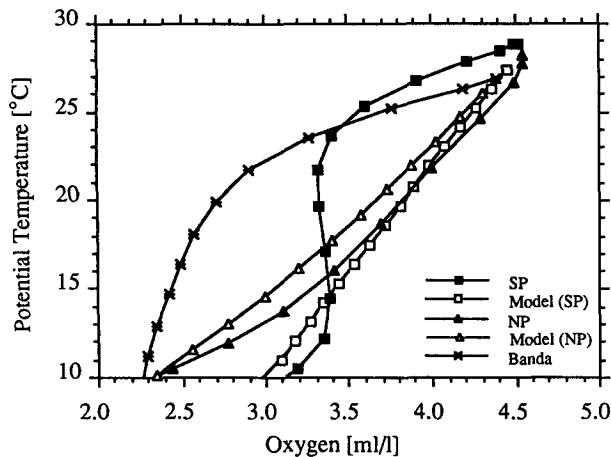


FIG. 8. The  $\theta$ - $O_2$  curves for the North Pacific, South Pacific, Banda, and model. The model results are for a North Pacific source ( $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  and a residence time of 1.7 years) and a South Pacific source ( $K_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  and a residence time of 17 years).

model sensitivity to a vertically varying diffusivity profile, a stratification-dependent parameterization was used (Garrett 1984). With this parameterization  $K_z$  is low where the water column is highly stratified and high where the water column is unstratified. Equation (1) is therefore simplified to

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left( K_z \frac{\partial C}{\partial z} \right) \quad (3)$$

where  $K_z$  is proportional to inverse buoyancy frequency. The results are nearly the same as with the constant diffusion case. Examination of the curves reveals that in the deep where the vertical diffusivity is high (because the stability is low) the water is already well mixed, and a high vertical diffusivity has no apparent effect. At the salinity maximum where  $K_z$  is low (because the stability is high) the model must run until the salinity maximum is eroded, producing the same  $K_z\tau$  estimate as in the constant  $K_z$  case.

To evaluate the sensitivity of the results to vertical shear, the shape of the velocity profile was inferred from the structure of the pressure gradient driving the Indonesian throughflow (Wyrki 1987). Because most of the flow is in the upper 150 to 200 m (Murray et al. 1989; Wyrki 1987), the velocity in the upper thermocline was assumed to be roughly an order of magnitude greater than in the lower thermocline. Equation (1) is again modified, and in a stationary reference frame, the throughflow velocity is explicitly modeled by  $u(z)$ ,

$$u(z) \frac{\partial C}{\partial x} = K_z \frac{\partial^2 C}{\partial z^2} \quad (4)$$

Again, the results did not significantly vary with this

sensitivity test. Inspection of the curves indicates that at the salinity maximum—where the velocity is high and therefore the residence time short—there is little time for diffusion to work on the high gradients. On the other hand, in the lower thermocline—where the velocity is low and hence the residence time is long—previously mixed water just becomes more thoroughly mixed. Therefore, the model continues until the residence time at the salinity maximum is long enough for the vertical salinity maximum gradients to be eroded, producing the same  $K_z\tau$  estimate as in the case without vertical shear. The lack of sensitivity to vertical shear and stratification-dependent diffusivity suggests that the  $K_z$  and  $\tau$  estimates may be constraining only at the salinity maximum.

### 6. Discussion

The results provide a constraint for circulation and budget models by specifying the allowed range of vertical diffusivities and residence times consistent with the erosion of the salinity maximum by vertical mixing. Although the relationship of  $K_z$  to  $\tau$  has been estimated (Fig. 6), an independent estimate of the residence time is necessary to specify  $K_z$  in the Indonesian seas. Such an estimate can be made by extracting the residence times inherent in throughflow and freshwater flux estimates and drifter velocity data.

Because most of the throughflow is thought to be in the upper 200 m, the residence time at the salinity maximum (around 100 to 200 m) may be a good approximation of the average throughflow residence time in the upper 200 m. Therefore, the residence time axis in Fig. 6 can be converted into units of throughflow

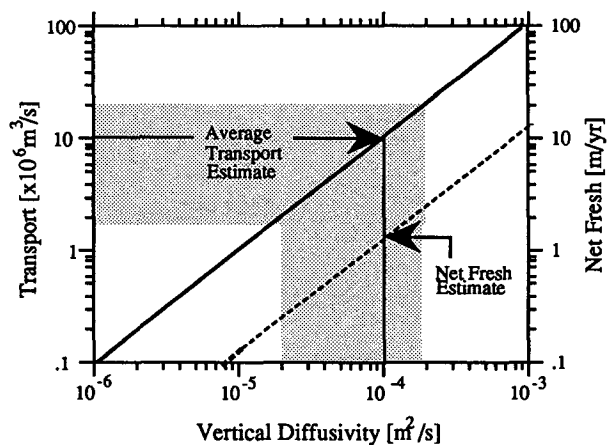


FIG. 9. The relationship of  $K_z$  to throughflow and to freshwater flux in the Makassar Strait. Both a  $10 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  throughflow estimate and a  $1.2 \text{ yr}^{-1}$  net freshwater flux estimate suggest a vertical diffusivity around  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ . The 1.7 to 20 ( $\times 10^6 \text{ m}^3 \text{ s}^{-1}$ ) throughflow range for the entire Indonesian seas (Gordon 1986) is shaded.

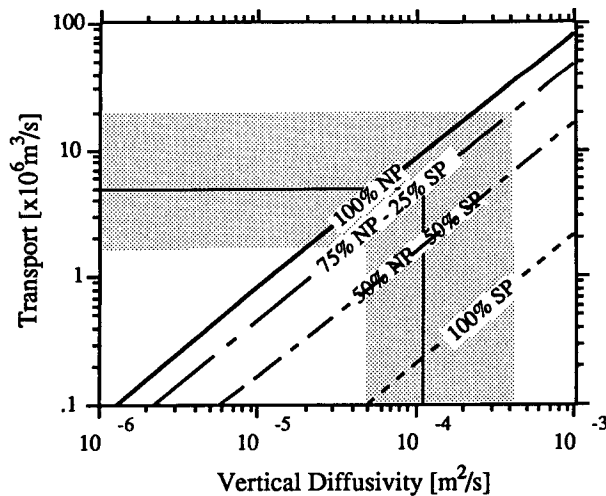


FIG. 10. The relationship of transport to vertical diffusivity in the Banda Sea for a 100% North Pacific source, a 75% North Pacific–25% South Pacific isopycnal mix, a 50% North Pacific–50% South Pacific isopycnal mix, and a 100% SP source. Reasonable throughflow estimates ( $5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ ) and Pacific source combinations (75% NP–25% SP) imply a vertical diffusivity around  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ . The  $1.7$  to  $20$  ( $\times 10^6 \text{ m}^3 \text{ s}^{-1}$ ) throughflow range for the entire Indonesian seas (Gordon 1986) is shaded.

by dividing the appropriate upper 200-m water volume by  $\tau$  (Figs. 9 and 10). For the Makassar Strait an additional constraint is obtained from estimates of the atmospheric and riverine freshwater flux and drifter velocity data. Thus, the residence time axis in Fig. 6a also can be scaled to freshwater flux by dividing the freshwater needed to convert the NP profile into the Makassar profile (0.5 m, as calculated from the average salinity of the profiles) by  $\tau$  (Fig. 9).

There have been many indirect estimates of the Indonesian throughflow ranging from  $1.7 \times 10^6$  to  $20 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  (Gordon 1986). For the Makassar Strait, both an average transport estimate of  $10 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  and a freshwater flux estimate of  $1.2 \text{ m yr}^{-1}$  (Oberhuber 1988) suggest that  $K_z \approx 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  in order to account for the observed reduction in vertical salinity gradients (Fig. 9). A higher throughflow or freshwater flux estimate demands more vigorous mixing, and a smaller throughflow or freshwater flux estimate demands less vigorous mixing. Lukas et al. (1991) report near-surface speeds over  $1 \text{ m s}^{-1}$  in the Mindanao Current and  $0.8 \text{ m s}^{-1}$  in the Makassar Strait from July to September. Drifters traveling that distance at those speeds imply a residence time on the order of 1 month. From Fig. 6a, this suggests that  $K_z$  is about  $3 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ .

In the Banda Sea an average transport estimate of  $10 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  implies that  $K_z \approx 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  for an NP source,  $6 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$  for a 50% NP–50% SP isopycnal mix, and  $4 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}$  for a pure SP source in order to account for the observed reduction in ver-

tical salinity gradients (Fig. 10). For a  $K_z$  of  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ , the throughflow range restricts the source to one that is predominantly NP. A large proportion of SP water—greater than 50%—requires that either  $K_z$  is two orders of magnitude greater in the Banda Sea than in the interior ocean or there is little or no throughflow in the eastern Indonesian seas (because of the long residence time required to reduce the vertical salinity maximum gradients). Nevertheless, the model shows that some SP presence is needed to balance the salt budget. A Banda throughflow of  $5 \times 10^6 \text{ m}^3 \text{ s}^{-1}$  with a blend of 75% NP–25% SP requires  $K_z \approx 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ .

In general, the throughflow route suggested by our modeling results indicates NP water flowing through the Sulawesi, Makassar, and Flores in less than a year (for an average water parcel in the upper 200 m) while vertically mixed by a diffusivity of at least  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ . As most of the mixing is expected to occur at the boundaries or sills rather than in the interior, these basin-scale vertical mixing estimates may be a measure of the integrated effect of vertical mixing. An SP presence in the Banda Sea is required by salinity profiles. However, a pure SP source implies a long residence time and a minor throughflow. There is little evidence to suggest substantial NP thermocline water entering the eastern seas through the Maluku, and SP water necessarily implies minimum transports. Therefore, most of the throughflow must originate from the NP and pass through the Makassar Strait.

## 7. Conclusions

The vertical mixing within the Indonesian seas can be quantified by simulating the degree of modification of the subsurface salinity maximum with a simple advection–diffusion model. The results provide a possible constraint for circulation and budget models by specifying the allowed range of vertical diffusivities and residence times consistent with the erosion of the salinity maximum by vertical mixing. The results also suggest that a high proportion of the throughflow is of North Pacific origin flowing through the Makassar Strait and into the Banda Sea before entering the Indian Ocean. A high vertical diffusivity—greater than  $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ —may be inferred in the Indonesian seas. Such a high value for thermocline vertical mixing indicates that this process may play an exceptional role in the region's dynamics and thermohaline and nutrient vertical fluxes. Further studies must be directed toward understanding the physical mechanisms in the Indonesian seas responsible for enhanced vertical mixing. A likely candidate is tidal energy dissipated along topographical boundaries.

*Acknowledgments.* The authors thank S. Zebiak, H. Ou, D. Martinson, J. Ledwell, M. Cane, and the

anonymous reviewers for their helpful comments and criticisms. This work was supported by the National Science Foundation Grant OCE-8614323 and the Office of Naval Research Contract N00014-90-J-1233.

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