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Recipe for Banda Sea water

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ABSTRACT

Water from the western Pacific flows through the Indonesian Seas following different pathways and is modified by various processes to form the uniquely characterized isohaline Banda Sea Water. The processes contributing to the isohaline structure are studied using data from three ARLINDO cruises in 1993, 1994, and 1996. An inverse-model analysis using salinity and CFC-11 data is applied to a vertical section along the main path of flow, from the Makassar Strait to the Flores Sea and Banda Sea. The model reproduces the seasonal and interannual variability of the throughflow and shows reversals of flow in the vertical structure. The model solutions suggest strong baroclinic flows during the southeast monsoon of 1993 and 1996 and a small, more barotropic flow during the northwest monsoon of 1994. The isohaline structure can be accounted for by isopycnal mixing of different source waters and by vertical exchanges, which are significant in this region. A downward flux equivalent to a downwelling velocity of $5 \times 10^{-7}$ m/s is estimated for the section. The total balance also suggests that seasonally and possibly interannually variable backflushing of water from the Banda Sea into the Straits contribute to the isohaline structure of Banda Sea water.

1. Introduction

The Indonesian Seas are the tropical link between the Pacific Ocean and the Indian Ocean and are part of the global thermohaline circulation (Gordon, 1986). The circulation and transport in the Indonesian Seas have a large annual variation due to the strong monsoonal change in the wind pattern in this area. The throughflow has the maximum transport during the south monsoon (northern summer), and the flow reaches its minimum in the north monsoon (northern winter) (e.g., Wyrtki, 1987; Murray and Arief, 1988; Meyers et al., 1996; Molcard et al., 1996). The change in the circulation pattern and volume of the throughflow must affect the different processes that occur in the Indonesian Seas. Dynamics of interaction with coastlines and shelves invigorate processes like upwelling, downwelling, and internal wave action.

Changes in the flow pattern and transport variability in turn produce changes in water masses and properties. During different monsoon seasons, waters from different sources flow to the Indonesian Seas, and the result can be seen in the variation of salinity, temperature, and other tracers from the surface down to the lower thermocline. Local

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processes such as precipitation, tidal mixing and coastal upwelling also contribute to these changes. Historical and recent data show the seasonal as well as interannual changes in the water properties in the Indonesian Seas (e.g., Wyrtki, 1961; Boely et al., 1990; Zijlstra et al., 1990; Gordon and Fine, 1996; Ilahude and Gordon, 1996). The source of the throughflow waters depends on the nonlinear properties of the Mindanao Current and South Equatorial Current retroflection at the Pacific entrance to the Indonesian Seas and the tropical Pacific wind fields (Nof, 1996; Wajsowicz, 2000) (Fig. 1). There are two main pathways of the throughflow in the Indonesian Seas, the western route and the eastern route. The primary is water that flows through the western route which is mostly from the North Pacific. It flows through Makassar Strait, Flores Sea, and Banda Sea before exiting to the Indian Ocean via Timor Sea and Ombai Strait (Fine, 1985; Ffield and Gordon, 1992; Fieux et al., 1994, 1996; Gordon et al., 1994; Bingham and Lukas, 1995; Ilahude and Gordon, 1996; Gordon and Fine, 1996). Part of this water also exits to the Indian Ocean through Lombok Strait (Murray and Arief, 1988). Prior to entering the Banda Sea, water that takes the eastern route (east of Sulawesi Island) is mostly from the South Pacific (Van Aken et al., 1988; Gordon, 1995; Ilahude and Gordon, 1996; Gordon and Fine, 1996; Hautala et al., 1996). It flows via Halmahera and Seram Sea to the Banda Sea and
Timor Sea. There is probably also a mixture of water coming from the North Pacific through the eastern route via the Maluku Sea. Most studies neglect the throughflow from the western Pacific through the South China Sea, Karimata Strait and Java Sea because of the shallowness of the shelf (mean depth of Java Sea is around 40 m). There is also some evidence of flow from the Indian Ocean into the Banda Sea through the Straits of Lesser Sunda Islands. This flow from the Indian Ocean is mostly confined to the surface layer during boreal winter (Michida and Yoritaka, 1996).

As the water flows through the Indonesian sills and straits, the extremely complex topography and the combination of active monsoonal changes and higher frequency forcing in the low latitude region contribute to momentum transfer and water mass transformation throughout the water column. Different studies show a large range of vertical diffusivity estimates for the different basins in Indonesia. Gordon (1986) uses a high value of vertical diffusivity ($K_z > 3 \times 10^{-4} \text{ m/s}$) within the Indonesian thermocline to account for the modification of the North Pacific thermocline temperature-salinity structure. Ffield and Gordon (1992), using a simple advection-diffusion model for average basin profiles, estimate a lower limit of vertical diffusivity, $K_z$ at $1 \times 10^{-4} \text{ m/s}$, for transformation of Pacific waters into Indonesian water. These vertical diffusivities are greater than those typically estimated for the interior ocean (e.g., Gargett, 1984; Ledwell and Watson, 1991; Toole et al., 1994).

Hautala et al. (1996) examine isopycnal advection and mixing between North and South Pacific low latitude western boundary current sources to form Banda Sea water. Three regimes are apparent in their analysis. The surface and upper thermocline layers, down to $\sigma_0 = 25.8$, required vertical mixing of surface precipitation and runoff. Vertical mixing is also apparent below $\sigma_0 = 27.0$. In the lower thermocline purely isopycnal spreading gives a simple variation in the ratio of sources toward a larger South Pacific contribution with depth.

Besides a large range of transport and diffusivity estimates, little is known about vertical distribution of flow and mixing in the Indonesian Seas. Changes in the direction of the currents suggest that the flow is highly baroclinic. Change in direction and strength of the wind force the change in the surface Ekman flows, moving surface water toward opposite directions. The build up of water along the boundaries change the pressure gradient across the shallow sills surrounding this area and presumably force the baroclinic flow throughout the whole region. Vertical motion could also play an important role in the water distribution and transformation in this region. Lack of sufficient direct and long-term measurements make it difficult to get a complete picture of circulation and processes in the region, especially the vertical distribution of the circulation.

In this paper we will discuss some new observations from the Arlindo Mixing cruises in 1993 and 1994 (AM93, AM94) onboard KAL Baruna Jaya I and Arlindo Circulation 1996 cruise (AC96) onboard KAL Baruna Jaya IV. We will describe an inverse analysis for quantifying the mixing and advection parameters along the pathways of the throughflow
that form Banda Sea water. This analysis makes use of Arlindo temperature, salinity and CFC-11 data.

As more tracer data become available, different techniques are developed to quantify the velocity field and mixing parameters from the tracer distribution. One of the first attempts to obtain velocities from tracer distributions was done by Fiadeiro and Veronis (1984). Lee and Veronis (1989) show that velocities and mixing coefficients can be obtained from an inversion of perfect tracer data, but inversion of imperfect data leads to substantial errors. Hogg (1987) obtains the flow fields and diffusivities on two isopycnal surfaces from temperature, oxygen and potential vorticity distribution of Levitus Atlas data (Levitus, 1982).

One way to solve this problem is by using different tracer distributions, with at least one transient tracer. Wunch (1987, 1988) studies the transient tracer inverse as a problem in control theory and the regularization aspect of using transient tracers. Jenkins (1998) combines $^3$H-$^3$He age with salinity, oxygen, and geostrophy to compute isopycnal diffusivities, oxygen consumption rates, and absolute velocities for the North Atlantic subduction area. McIntosh and Veronis (1993) show that an underdetermined tracer inverse problem can be solved by spatial smoothing and cross validation. Here an inverse method similar to that of McIntosh and Veronis (1993) is used to quantify the advective and diffusive processes in the eastern Indonesian Seas.

2. Observations
   a. Data

   A total of 103 hydrographic stations were occupied during AM93 cruise from 6 August to 12 September 1993 in the Southeast Monsoon (SEM). The AM94 cruise from 26 January to 28 February 1994 during the Northwest Monsoon (NWM) has a similar station distribution (station 104–209). Cruise AC96, including 37 hydrographic stations into the eastern Banda Sea were occupied from 18 November to 15 December 1996 (Fig. 1).

   The CTD data were collected by Lamont Doherty Earth Observatory (LDEO) group and results from AM93 and AM94 have been reported before (Gordon and Fine, 1996; Ilahude and Gordon, 1996; Mele et al., 1996). Chlorofluorocarbon-11 (CFC-11) was measured during the three cruises by the University of Miami. The CFC analyses were done via a custom built extraction system and gas chromatograph similar to Bullister and Weiss (1988). The average analytical blanks for CFC-11 for the AM93, AM94, and AC96 cruises are 0.001, 0.029, and 0.001 pmol/kg, respectively. The high blank levels especially at the beginning of AM94 are due to high CFC concentrations in the laboratory air where the samples were analyzed. The average bottle blanks for CFC-11 for the AM93, AM94, and AC96 cruises are 0.009, 0.060, and 0.002 pmol/kg, respectively. Improvements in the analytical system and lower concentrations of CFC in the laboratory air during AC96 cruise increased the quality of CFC measurements significantly. The precision of the CFC analyses was estimated by drawing duplicate samples from the same Niskin bottle. For
measurement of CFC-11 greater than 0.1 pmol/kg, the averaged concentration-difference between duplicates in the eight to twenty replicates are 3.48%, 3.32%, and 0.32% for the AM93, AM94, and AC96 cruises, respectively.

b. Observed properties and circulation based on AM93, AM94, and AC96 data

The Banda Sea water is unique in its almost isohaline water column at around 34.5 from the thermocline down to the bottom (see for example Gordon, 1986; Field and Gordon, 1992; Gordon et al., 1994; Gordon and Fine, 1996; Ilahude and Gordon, 1996; Fieux et al., 1996). Observations of water masses in the Banda Sea during the three Arlindo cruises show the same thermocline isohaline structure in the Banda Sea (Fig. 2). Lower salinity water is observed at the Banda surface especially during AM93 in the SEM. At the intermediate depths the temperature-salinity (T/S) characteristics match the Antarctic Intermediate Water and so must come from the South Pacific.
i. **Upper layer.** All around Indonesia, the upper layer tracer distribution changes seasonally and interannually (Figs. 2, 3 and 5). During AM93 (SEM), North Pacific Subtropical Water (NPSW) salinity maximum of 34.75 is observed at about 150 m depth in the Makassar Strait ($\sigma_0 = 24.5$). This maximum is observed into the Flores Sea with diminished intensity. It is not apparent in the Banda Sea. During AM94 (NWM), very low surface salinity is found in the southern Makassar Strait. The low salinity signal penetrates to over 100 m depth. At this time, a very weak salinity maximum (34.6) is observed in Flores Sea at 150 m.

The T/S curves in the Makassar Strait between $\sigma_0 = 23$ and $\sigma_0 = 25.5$ during NWM lack the NPSW salinity maxima and are similar to the Banda Sea T/S curves during SEM. There are four possible explanations for this lack of salinity maxima. (1) Qiu *et al.* (2000) state that the salinity maxima input to the Indonesian Seas displays strong variability at intraseasonal time scales, resulting from eddy formations at the Mindanao retroreflection. Gordon and Fine (1996) suggest that relaxation of the throughflow during NWM allows more attenuation of the salinity maximum signal in the Makassar Strait. (2) Ilahude and Gordon (1996) suspect a southward drift of Sulu Sea water wipes out the NPSW salinity maxima causing a localized salinity minimum. (3) Upwelling in the frontal boundary between the outflow of Makassar Strait and current from the Flores Sea can contribute to the properties of water in the south Makassar Strait as well (Ilahude, 1970; Hughes, 1982). (4) Here a backflushing of modified Banda Sea Water into the Makassar Strait is offered as another possible explanation. The term backflushing as used here does not distinguish between horizontal advection and diffusion processes.

The backflushing possibility is best addressed by using a transient tracer that can differentiate temporal changes. In 1993 (SEM), a subsurface CFC-11 maximum is observed just above the salinity maximum core at about 100 m ($\sigma_0 = 24.5$) (Fig. 5). CFC-11 concentration is up to 1.8 pmol/kg at this depth, north of Dewakang sill in the Makassar Strait and decreases to about 1.4 pmol/kg in the Banda Sea at the same level. During NWM, the CFC-11 distribution follows the trend of salinity in the upper 150 m, the subsurface maxima in the Makassar Strait is not observed; however, a weak CFC-11 maximum with concentration of 1.7 pmol/kg is observed in the Flores Sea at 100 m depth. By examining the salinity distribution alone, one may conclude that the salinity maxima in the thermocline is destroyed by vertical mixing of fresh water at the surface. Localized CFC mimima above the Flores maximum suggest that it cannot be a result of mixing down of surface water that is high in CFC. The strong stratification suppresses such mixing.

The salinity distributions during AC96 are similar to AM93 with the exception of a lower surface salinity in the beginning of NWM, and a deeper thermocline by about 50 m compared with AM93 and AM94 (Fig. 5). The latter may be a clue to interannual variations (Ffield *et al.*, 2000). In 1996, a more continuous subsurface CFC maximum is observed from the northernmost station in the Makassar Strait to Flores Sea at about 150 m depth. The concentration is 1.9 pmol/kg in Makassar Strait and decreases to 1.7 pmol/kg in the western Banda Sea. Higher CFC-11 concentrations compared to the previous cruises are
expected in the subsurface waters because of the increase in the atmospheric input function. CFC-11 concentrations in the atmosphere increased until year 1993 and decrease slightly after 1993 (Walker et al., 2000). Water below 50 m in this region, during AM93, AM94, and AC96, has been ventilated prior to 1993 and, therefore, increases in CFC-11 concentration with time. Changes in the circulation pattern and ventilation rate due to seasonal and interannual variability can also alter the CFC concentration. Later, the inverse analysis results will show that part of the increase is due to the change in the circulation.

Fresh water influx at the surface is an important contributor to the T/S structure. Fresh water increases the stability of the upper layers. The major sources of fresh water are river runoff from islands plus precipitation over the ocean. Maximum fresh water inflow occurs during the NWM, when the wind brings the warm moist air from the South China Sea. In the NWM, there is a sea-surface salinity minimum in the Southern Makassar Strait, possibly advected from the Java Sea. The surface salinity minimum in Banda Sea during the month of AM93 cruise is suspected to be advected to the east from the previous NWM season. Other possible sources of fresh water in the upper Banda Sea are river outflows from Irian Jaya (New Guinea; Wyrtki, 1961; Zijlstra et al., 1990). In the SEM, the surface temperature of the Banda Sea is about 4°C cooler than in the NWM. The surface cooling is expected to be the result of upwelling and local evaporative cooling processes when the wind is blowing from the southeast bringing dry air.

**ii. Mid and lower thermocline.** The mid and lower thermocline layers have a less variable tracer distribution. The North Pacific Intermediate Water (NPIW) salinity minimum is clearly observed in the Makassar Strait at \( \sigma_0 = 26.5 \) with salinity 34.45 during all three cruises (Figs. 2 and 5). The NPIW salinity minimum around 300 m is a permanent feature in both monsoon seasons. This suggests that the mid-depth circulation field is more constant. The salinity minimum can be traced downstream to Flores Sea and all the way to the Banda Sea with diminished intensity.

The increase in salinity at the NPIW level along the path from Makassar Strait to the Flores and Banda Seas can alternatively occur from vertical mixing or as a product of isopycnal mixing between the low salinity NPIW water with a saltier water of South Pacific origin. The latter conclusion is drawn from the Hautala et al. (1996) mixing analysis, where purely isopycnal mixing between North and South Pacific thermocline water can produce the observed Banda Sea water. Here CFC’s are used to constrain the problem.

In the upper layer, CFC-11 partial pressure age is younger in the northern Makassar, ages increase in the Flores and Banda Seas (Fig. 3). Along the section in the middle of the Banda Sea, above 500 m, CFC-11 ages increase from south to north during AM93. CFC ages in the Southern Banda Sea suggest that it is more recently ventilated and that the central Banda Sea main thermocline is ventilated by flow from the Flores Sea. However, the trend changes below 400 m, where younger ages of CFC-11 are observed in the Seram Sea and northern Banda Sea, as compared with the Flores Sea and Makassar Strait. Figure 4 shows
the CFC-11 profiles in the northern Makassar Strait, Banda Sea, and Seram Sea during AC96. Below the subsurface maximum, the CFC-11 concentrations in the northern Makassar Strait decrease rapidly to near blank levels. In the Banda Sea the CFC-11 profile has no subsurface maxima, and the concentration decreases less with depth.

Banda Sea lower thermocline and intermediate water is influenced by South Pacific water entering above the sills at the eastern entrance to the Banda Sea. The deepest sill connecting the Pacific and the Banda Sea is the Lifamatola sill (1950 m) at 126° 57’ E, 1° 49’ S, between Sulu Archipelago and Obi Island (Fig. 1). The Halmahera Sea sill depth is around 500 m. Seram Sea water is characterized by salty lower thermocline water that can only be derived from the South Pacific. In 1993 the salinity maximum is 34.7 at 250 m depth (σ₀ = 26.0) at station 41. A more pronounced intrusion of South Pacific Subtropical Water is found during the AM94 in the Halmahera and Seram Seas. The intrusion is characterized by a salinity maximum located on σ₀ = 26.0, with salinity up to 34.9 (Ilahude and Gordon, 1996). CFC-11 profiles in the Banda Sea are similar to the profiles in Seram Sea. The scatter in the data at the low concentrations makes it difficult to determine the trend of the CFC-11 age along the eastern route (Fig. 3). It is possible that some recirculation occurs between the Seram Sea and the Banda Sea at this level.

Figure 3. CFC-11 derived ages on σ₀ surfaces along different sections in the Indonesian Seas during AM93, AM94, and AC96.
iii. Intermediate water. Below the Dewakang sill depth (550 m), at the end of Makassar Strait (118° 48' E, 5° 24' S), the CFC-11 concentrations are low, but still show a clear trend between basins. In Makassar Strait, north of the sill, CFC-11 concentrations are near blank level at 1000 m and below. The presence of this old water suggests a blocked flow. At depths around 1000 m, the Banda Sea has definitely higher concentrations than the Makassar water at the same depth. At this depth the Seram Sea has slightly higher CFC-11
concentrations as compared to the Banda Sea (Fig. 4). At these intermediate depths the concentrations decrease from Banda Sea to the west (Flores Sea). The relatively high CFC-11 below 500 m in the Flores Sea in the AM94 data set have been discounted here due to blank problems at the beginning of the cruise.

Both T/S and CFC profiles suggest the possibility of some backflushing of water from the Banda Sea to Seram Sea, Flores Sea and Makassar Strait. There is also exchange with the Java Sea for the top 50 m. These observations described in this section agree with the pattern of circulation illustrated by Wyrtki (1961) and Van Bennekom (1988).

3. An inverse model for the passage to the Banda Sea

The tracer distribution at a point is a result of the advection and mixing history prior to the observation. The result is also subject to the different boundary conditions for each individual tracer. The observed tracer distributions reflect the dynamics integrated over time. If a system is stationary and there are enough linearly independent tracers, it is possible to obtain an optimal solution for the circulation field and mixing parameters that can produce the observed tracer field. In most cases, there are not enough observations, or the information in the tracer fields are not linearly independent, leading to an underdetermined problem. The Arlindo program provides sufficient data for a multi-tracer inverse analysis for the region.

The domain of interest is a two-dimensional vertical section along the flow path from Makassar Strait to Flores Sea and Banda Sea (Fig. 5), for 1993, 1994, and 1996. It is assumed that these Straits and Seas act as a channel and the flow and tracer fields are uniform across the channel. The 2-D channel assumption does not ignore the input for the Banda Sea from the eastern channel. In fact, the model solutions show that there is flow from the east in the lower thermocline during AM93 and AC96. The sections are gridded into 200 km bins in horizontal direction and by 50 m in vertical direction. By excluding the top 50 m of the water column, the Java Sea influence and the surface exchange is minimized. The region below 650 m is not evaluated because the tracer field is almost uniform there. When tracer gradients are parallel in a region, or gradients disappear, the problem becomes degenerate and solutions are indeterminate (McIntosh and Veronis, 1993). The model uses open boundary conditions. The top, bottom, northern and southern boundary conditions allow varying streamfunctions, so that exchanges with regions outside the domain are determined by the solutions of the model only. A quasi-steady state is assumed for the tracer distributions for each cruise in turn. The tracers used in the calculations are salinity and CFC-11 age. The information contained in the potential temperatures is used in the CFC-11 age calculation.

The steady advective-diffusive equation for conservative tracer $C$ can be written as

$$-v \cdot \nabla C + \nabla \cdot (K \nabla C) = 0.$$ (1)
A steady advective-diffusive equation for an ideal tracer age $T$ becomes

$$-\mathbf{v} \cdot \nabla T + \nabla \cdot (K \nabla T) + 1 = 0$$

(2)

where factor one comes from a linear increase in age with time. Here the calculation is posed in terms of the two-dimensional streamfunction $\varphi$, where $u = -\partial \varphi / \partial z$ and $w = \partial \varphi / \partial x$. Vertical diffusivity $K_z$ is assumed uniform throughout the channel. There is no horizontal diffusivity in the model. The terms chosen to balance the advective-diffusive equation are similar to the ones used by Ffield and Gordon (1992), where advection is balanced by vertical diffusion. Ffield and Gordon (1992) use a moving reference frame, so the advection term appears as the local rate of change in tracer concentration.

For an $m \times n$ grid (see Fig. 6), there are $(m + 1)(n + 1)$ unknowns for $\varphi$. For $k$ tracers there will be $k \times m \times n$ equations, in this case $k = 2$ (salinity and CFC-11 age). For the case where vertical diffusivity is equal to zero, the inhomogeneous terms needed to solve the inverse problems come from the transient tracer, in this case the CFC-11 age.

In each grid, tracer concentrations are evaluated at the center of the grid and streamfunction $\varphi$ at the corners. The discrete form of the advective-diffusive equations in the inverse...
The calculation is:

\[
\left[ \frac{1}{2} (C_{i+1,j} - C_{i,j+1}) \Psi_{ij} + (C_{i,j-1} - C_{i+1,j}) \Psi_{i,j-1} \\
(C_{i,j+1} - C_{i-1,j}) \Psi_{i-1,j} + (C_{i-1,j} - C_{i,j-1}) \Psi_{i,j-1} \right]
\]

\[
= (C_{i,j+1} + C_{i,j-1} - 2C_{i,j}) K^* + t_{ij}
\]

where

\[\Psi = \varphi / (\Delta x \Delta z)\]
\[K^* = K_c / (\Delta z)^2\]

and \(t_{ij}\) equals zero for conservative tracers and minus one for age equations. Over the entire grid, the difference equation can be written in matrix form as

\[Ax = b\]

where \(A\) is a \((mnk + 1)\) by \((m + 1)(n + 1)\) coefficient matrix, \(x\) is the unknown vector of \(\Psi\) and \(b\) is a vector of zeros and minus ones plus the diffusive terms. The one additional equation augmented on the set of the tracer equations is for the arbitrary constant for the streamfunction. This system of equations contains two kinds of errors, errors in the coefficients that come from measurement error, and errors in the model.

To solve this system of linear equations, the technique similar to that of McIntosh and Veronis (1993) is used. This inverse method seeks the optimal solution for the advective-diffusive tracer problem by spatial smoothing and cross validation.
a. Smoothing

This technique solves the system of equations $Ax = b$ by balancing the solution smoothness, based on spatial derivatives of the solution, and the residual in the advective-diffusive equation. A penalty function, $P$, is formed as a weighted linear combination of roughness and the residual equation $e = Ax - b$.

$$P = x^T W_x x + e^T W_e e.$$  \hspace{1cm} (5)

$W_x$ is a roughness matrix containing semi-norm operators and weights for the streamfunction.

$$W_x = \mu W_\Psi.$$  \hspace{1cm} (6)

$W_\Psi$ is a derivative operator. When operated on the solution field it gives the measure of roughness. A smooth solution is obtained by minimizing the second derivatives (second order seminorm) for streamfunction. $W_e$ is the error covariance matrix. In the calculation, a noncorrelated, uniformly weighted covariance matrix is used. It is a diagonal matrix with diagonal elements that are inversely proportional to the relative measurement error for salinity ($w_s$) and CFC-11 age ($w_a$). The numerical values of $w_s$ and $w_a$ are unimportant, it is only the size of $W_e$ relative to $W_\Psi$ that is important, and the relative size is controlled by the smoothing parameter $\mu$. The CFC measurements during AC96 are improved significantly compared to the AM93 and AM94. In the calculation, different error weights, $w_a$, are applied. For AC96 $w_a = 1$ year, and $w_a = 10$ years for AC93 and AC94. For salinity $w_s = 0.001$ for all three observations. These weights are relative weights to the error in the streamfunctions. The optimal solution (smooth solution consistent with observation) is obtained by minimizing $P$ as a function of $\mu$ parameters.

b. Cross validation

To estimate $\mu$, a parameter estimation technique known as the generalized cross-validation technique is used. The idea is to withhold one equation, solve the $M^{-1}$ problem, and compare the withheld datum with the value predicted from the reduced problem (McIntosh and Veronis, 1993). A cross validation function $V$ is formed.

$$V(\mu) = \frac{M^{-1}e^TW_e e}{[M^{-1}\text{trace}(I-R_b)]^2}.$$  \hspace{1cm} (7)

where $R_b = AA^{-g}$ and

$$A^{-g} = [A^TW_e A + W_x]^{-1}A^TW_e$$

$R_b$ is called the data resolution matrix. $A^{-g}$ is the generalized matrix inverse and $x = A^{-g}b$.

The entire minimization process is done by minimizing this single function $V$. The minimization procedure used is the simplex minimization routine $fmins$ in MATLAB.
c. Model results

The inverse calculations were carried out over a range of vertical diffusivity coefficients, \( K_z \). Figure 7 shows the ratio between the horizontal velocity, \( u \), at distance = 100 km and depth = 300 m to the input vertical diffusivity. Although this figure only shows the value from one point in the section, the pattern is the same throughout the entire grid. A linear relation between velocity and vertical diffusivity is obtained when \( K_z \) is larger than \( 10^{-2} \) m\(^2\)/s. When \( K_z \) is smaller than \( 10^{-3} \) m\(^2\)/s, \( u/K_z \) is inversely proportional to \( K_z \). This behavior can be explained as follows. When \( K_z \) is large, the factor one in the age equation becomes negligible. The balance between the terms in the advective-diffusive equation is then between the advective term and the vertical diffusion term. When \( K_z \) is small, the balance is between the horizontal advection and the vertical advection. A velocity-diffusivity ratio \( (u/K_z) \) that is inversely proportional to \( K_z \), shows that \( u \) stays constant even though \( K_z \) decreases. This gives the limit of advection in the absence of diffusion. In other words, the solution for \( u \) remains the same for any diffusivity below \( K_z = 10^{-5} \) m\(^2\)/s.

Figures 8 and 9 show the solutions of the inverse model for two extreme cases. The first one is for a no diffusivity case \( (K_z = 0) \), and the second figure shows the solution for a large vertical diffusivity case \( (K_z = 10^{-3} \) m\(^2\)/s). Table 1 summarizes the parameters used and the smoothing parameter of the optimal solution for each case. The no diffusivity case gives an unrealistic small advection and is shown mainly for comparison. The vertical diffusivity \( 10^{-3} \) m\(^2\)/s is chosen to give the solution in the range of observed horizontal velocities. The surface currents based on ship drift data in the Makassar Strait are of the order of a few tens cm/s (Mariano et al., 1995). Other diffusivities larger than \( 10^{-5} \) m\(^2\)/s will give very similar velocity profiles, scaled with the mixing coefficient used, as explained above. The dashed lines on the velocity profiles are the residuals between the optimal solution and the observed tracer coefficient multiplied by the generalized inverse matrix. The solution \( x = A^{-1}b \) is equal to a vector of streamfunction on every grid. The streamfunction residual is
calculated as \( x_e = A^{-1}e \) where \( e = Ax - b \) is the residual. Horizontal and vertical velocity residuals are then obtained from \( x_e \). Except for the no diffusion case in AC96, all the residuals are small compared with the optimal solution.

The solution of the inverse calculation is limited somewhat because of the nature of the minimization process. The model minimizes the residuals, and the residuals are minimum when \( K_z \) is small. Therefore the transport tends to be small for a given value of diffusivity. In a different experiment where \( K_z \) is taken as an unknown variable, the optimal solution gives unrealistically small \( u \) and \( K_z \).

4. Discussion

Results from the inverse calculation are limited by the accuracy in the data and the assumptions of the model. The model assumes steady state conditions, while the data are taken from three synoptic cruises. This condition may apply assuming that the circulation adjustment time to monsoon change is short, which is generally the case in monsoon
regions in the world (Ray, 1984; Savidge et al., 1990; Liu et al., 1992). The flushing time of the Makassar Strait depends on the advective velocity. Examples of advective velocities in the Makassar Strait are given by surface drifters. The surface drifter data are compiled by the Drifting Buoy Data Assembly Center (DAC), Atlantic Oceanographic and Meteorology Laboratory (AOML), NOAA, Miami. Four out of five surface drifters found in the Makassar Strait entered from the Sulawesi Sea between February and May, but none of them entered during the Arlindo cruises. These drifter velocities reach up to 1 m/s. At these velocities, it takes only 16 days to carry water and tracers as far as 1400 km. The velocity is expected to be lower in the intermediate water. At velocities on the order of a few tens of centimeters per second, the flushing time can increase to on order two to three months. Because the flushing time is less than the semi annual period, the quasi steady state assumption is still, however, applicable.

The model does not allow $K_z$ to vary spatially. The impacts of this assumption are lessened because the model domain does not include the top mixed layer or the bottom friction layer. This suppresses the effect of varying $K_z$ due to turbulence in the surface and
bottom layer. Vertical mixing above rough bottoms and sills indeed can be larger (e.g., Polzin et al., 1996, 1997; Van Bennekom et al., 1988). Adding these spatially varying $K_z$ terms in the model will change the solutions, but having more unknown terms in the system of equations will make the system more underdetermined, and a solution might not be obtained at all.

Some dynamical terms like horizontal diffusion are not represented in the model. Since horizontal advection is the only horizontal process represented in the model, it has to be interpreted as a combination of all processes that carry or distribute tracers horizontally. Likewise, in the case where $K_z = 0$, vertical advection is the only way to distribute the tracers vertically. Assuming a channel-like geometry means that the velocity field in the model solutions also has to compensate for flows across channel that might occur in reality, such as the Lombok outflow and Java Sea exchange.

The errors in the data are accounted for in the error covariance matrix that contains the weights for measurement errors for CFC-11 and salinity. With these weights, the size of the residual equation turned out to be relatively small compared with the optimal solution. This shows that, despite the assumptions in the model, the simple dynamics chosen can reproduce the observed tracer fields with relatively small error.

The optimal solutions of the inverse calculation suggest that strong baroclinic flows occur during AM93 and AC96. On the other hand, a small and more barotropic flow is the optimal solution obtained for the tracer distribution in AM94. The general flow is southward and eastward, and sometimes accompanied by a reversal in flow. These flow patterns in the thermocline have different directions as compared with the surface currents that usually follow the local wind direction (Figs. 2 and 10). During SEM, the surface current in the Flores Sea is westward, but the solutions for AM93 and AC96 show a strong eastward flow in the upper thermocline. During NWM, the surface current is eastward, while the AM94 model gives a weak eastward flow in the upper thermocline.

There are two possible explanations for the difference between the observed surface circulation and the solutions given by the model for subsurface flows. First, tracer

Table 1. Input parameters and smoothing parameters $\mu = e^m$ for no diffusion and large diffusion case. $\mu$ is obtained by minimizing the generalized cross validation function $V$. Shown also are the mean optimal $u$ and $w$ for each case. $ws$ and $wa$ are the weights for error in salinity and age coefficients, respectively.

<table>
<thead>
<tr>
<th>$K_v$ (m$^2$/s)</th>
<th>$ws$</th>
<th>$wa$</th>
<th>$m^*$</th>
<th>$V(\mu)$</th>
<th>Mean $u$ (10$^{-2}$ m/s)</th>
<th>Mean $w$ (10$^{-7}$ m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>AM93 0</td>
<td>10$^{-3}$</td>
<td>10</td>
<td>1.929</td>
<td>1.144</td>
<td>0.046</td>
<td>-3.55</td>
</tr>
<tr>
<td>AM94 0</td>
<td>10$^{-3}$</td>
<td>10</td>
<td>2.615</td>
<td>1.297</td>
<td>-0.019</td>
<td>-2.90</td>
</tr>
<tr>
<td>AC96 0</td>
<td>10$^{-3}$</td>
<td>1</td>
<td>2.304</td>
<td>4.699</td>
<td>-0.006</td>
<td>-5.28</td>
</tr>
<tr>
<td>AM93 10$^{-3}$</td>
<td>10$^{-3}$</td>
<td>10</td>
<td>-0.350</td>
<td>5709</td>
<td>7.91</td>
<td>57.9</td>
</tr>
<tr>
<td>AM94 10$^{-3}$</td>
<td>10$^{-3}$</td>
<td>10</td>
<td>2.480</td>
<td>10841</td>
<td>1.25</td>
<td>74.4</td>
</tr>
<tr>
<td>AC96 10$^{-3}$</td>
<td>10$^{-3}$</td>
<td>1</td>
<td>0.037</td>
<td>23536</td>
<td>6.62</td>
<td>32.5</td>
</tr>
</tbody>
</table>

*(\mu = e^m)*
distributions reflect the circulation prior to observation. The model result may reflect the circulation during the opposite season of the measurement. This could explain the opposite flow in the thermocline as compared with the surface current and wind direction.

A more likely explanation for the difference between observed surface circulation and model subsurface flows is that the flows in the deeper thermocline are not directly driven by the local wind. During SEM, upwelling occurs in the Arafura and the eastern Banda Sea. The westward-flowing surface current induces coastal upwelling along the eastern boundary. The upwelled water is supplied by the thermocline, and so strengthens the eastward flow in the thermocline, as observed in the solutions for AM93 and AC96. During NWM, the surface circulation in the eastern Banda Sea is downwelling favorable, pushing the water westward in the lower thermocline. A westward flow from the Banda Sea to the Flores Sea is required for the AM94 solution (Fig. 10). Recent current meter data in the Makassar Strait indicates that for the boreal winter the surface layer is moving opposite to the thermocline flow (Gordon et al., 1999). It is interesting to note that the solutions for the NWM vertical velocities show weakest downwelling in the no diffusivity case or strongest upwelling in the large diffusivity case (Table 1), supporting the type of circulation described above.

Flow to the west is also part of the solution for AC96 in the lower thermocline and is more pronounced when the vertical mixing is small. A secondary downwelling cell may occur below the upwelling cell during SEM. The idea that the flow to the east in the upper thermocline is driving a secondary cell with a reversal in flow in the lower thermocline is suggested in Gordon et al. (1994). Here it is shown that the reverse flow can extend to the upper thermocline, like in the AM94 case, and vary seasonally and interannually.

The backflushing of water from the Banda Sea to the Makassar Strait through the Flores Sea could contribute to the similarity observed in the T/S curve in the Makassar Strait during AM94 as compared to the Banda Sea. The salinity minimum in the Flores Sea is clearly coming from the North Pacific Intermediate Water, which flows through the Makassar Strait. The water below NPIW could have more variable sources. Lower thermocline water in the Flores Sea is affected by sources from the east, through the passages east of Sulawesi. Higher CFC-11 concentrations in the deep Banda Sea than what are found at the comparable depths in the Makassar Strait suggest a South Pacific component. The exact route and ratio of North and South Pacific sources is undetermined. Considering the longer residence time of water in the Banda Sea (Ffield and Gordon, 1992), a modified Banda Sea water, that is a mixture of North Pacific, South Pacific and Banda Sea water from previous seasons, appears to be flushed back into the Makassar Strait from time to time. Recent observations of current meter data in the Makassar Strait from December 1996 to July 1998 show some evidence of reversal of the flow in the current meter at 350 m and deeper (Gordon et al., 1998; Gordon and Susanto, 1999). The current meter at 240 m does not show a flow reversal. A shorter record of ADCP data (1 December 1997–9 March 1997) from the same location shows a maximum southward flow near 110 m depth, and a variable northward and southward flow in the top 60 m (Gordon et
Thus, although from the property distributions and model an advective or diffusive backflushing cannot be determined, the current meter data are consistent with an advective backflushing process.

The changes in the flow character are caused by the changes in the pressure field along the channel. The pressure changes because of the change in the local wind direction and intensity, and the changes in the remote pressure field. The solutions suggest that the changes vary with depth. The exact mechanism of how and why these changes occur cannot be determined by an inverse model. A diagnostic or process study is needed to infer
the forcing and their responses. The seasonal flow variability given by the model solution supports the observation that the throughflow is minimum during the northwest monsoon season. In the large diffusivity case, the horizontal velocity for the AM94 (NWM) is about 15% to 20% of the velocity in the AM93 (SEM) and AC96 (end of SEM). Interannual variability may play an important role in the variability of the flow. The three observations were in different ENSO phases, the 1993 and 1994 cruises occur during a prolonged El Niño period, and the 1996 cruise was in a La Niña period. The solutions show a relatively stronger flow to the east in the upper thermocline during AC96 as compared with AM93. This supports the argument that during a La Niña phase, the build up of the warm pool in the western Pacific will strengthen the throughflow (Gordon et al., 1999). While the pattern of the circulation in the Indonesian Seas is strongly dependent on the local forcing and the distribution of boundaries, the strength of the throughflow seems to be more closely related to the pressure difference between the remote Pacific and the Indian Ocean.

The model solutions show that the upper thermocline eastward flow in 1996 is also accompanied by a reverse flow to the west in the lower thermocline. The strengthening of the pressure gradient in the lower thermocline between the Pacific Ocean and the Indonesian Seas during a La Niña phase could increase the flow in the deeper thermocline through the eastern passages. Since the sill depths in the eastern passages are much deeper than the Makassar Strait sill depth, an increase in the pressure gradient in the lower thermocline will increase the throughflow from the eastern passages.

Mean vertical velocities of the solutions range from $-5 \times 10^{-7}$ m/s to $74 \times 10^{-7}$ m/s. Although the values are not unrealistic, the pattern depends on the size of the vertical diffusivity chosen. In the case where vertical diffusivity is small, the downward vertical advection is needed to satisfy the advective-diffusive balance, and when $K_z$ is large, the optimal vertical velocity is mostly upward in the upper thermocline. Whichever process actually occurs, whether it is large vertical mixing or downward advection, the model results suggest a generally downward flux in this area. Vertical exchange is an important process that carries the heat and tracers downward. Downward movement in one place must be accompanied by upwelling somewhere else. Upwelling brings cooler and nutrient-rich water to the surface. A mean downward velocity of $0.5 \times 10^{-6}$ m/s will carry about 42 W/m² of heat over a 20°C temperature gradient from the surface to the lower thermocline.

This study shows a marked change in the throughflow volume and character seasonally and interannually. The inverse model is also able to estimate the horizontal and vertical exchange processes that produce the observed water masses and are important to the local weather patterns and productivity.

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