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Application of Munk's abyssal recipes to tracer distributions in the deep waters of the South Banda basin

South Banda basin
Tracers
Abyssal recipes
Diffusion coefficient
Source terms

Bassin sud de Banda
Traceurs
Formules abyssales
Coefficient de diffusion
Termes de source

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ABSTRACT

The vertical structures of distribution of potential temperature, oxygen, orthophosphate, dissolved silica and ^{14}C in the South Banda basin (East Indonesia) have been studied quantitatively by means of a simple one-dimensional model as reviewed by Munk (1966). The effective source and sink terms for several parameters, as well as the effective diffusion coefficient, have been estimated. The relatively large vertical diffusion coefficient ($1.3 \cdot 10^{-3}$ m^2/s) is probably mainly due to tidally driven boundary mixing along the side walls of the basin. The budgets of dissolved oxygen, phosphate and silicate can be modelled adequately with a constant source or sink term. While oxygen consumption mainly takes place in the water column, the source for dissolved silica can be explained completely from sediment-water exchange. Because of the small flushing time of the basin and the large O_2:P Redfield ratio, dissolved phosphate behaves as a conservative tracer.

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RÉSUMÉ

Application des formules abyssales de Munk sur les répartitions de traceurs dans les eaux profondes du bassin sud de Banda

A l'aide de modèles simples unidimensionnels discutés par Munk (1966), nous avons étudié la répartition verticale de la température potentielle, de l'oxygène, de l'orthophosphate, du silicate dissous et du ^{14}C. Les termes source et puits pour différents paramètres ainsi que le coefficient de diffusion ont pu être estimés. La valeur relativement importante du coefficient de diffusion verticale ($1{,}3 \cdot 10^{-3}$ m^2/s) est probablement due au mélange sur les bords du bassin. Dans les modèles, les bilans sont repoduits de façon satisfaisante avec des termes source et puits constants. Alors que l'oxygène est principalement consommé dans la colonne d'eau, le terme source pour le silicate dissous peut être entièrement expliqué par les échanges eau-sédiment. A cause du renouvellement rapide des eaux du bassin et de la valeur élevée du rapport de Redfield, O_2:P, le phosphate dissous se comporte comme un traceur conservatif.

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INTRODUCTION

Tracers, either conservative or non-conservative, are used in oceanography to determine circulation and mixing in the ocean. With Wüst's well-known core-layer method the direction of the flow, or better the direction of the tracer transport, is inferred from the observed large-scale distribution of tracers. Recently, much attention has been given to the quantitative use of tracers to determine velocities and mixing coefficients by inverse modelling (*e.g.* Fiadeiro and Veronis, 1984; Wunsch, 1985; 1987; Tziperman, 1988; Charnock *et al.*, 1988).

If non-conservative tracers are to be used for this purpose, knowledge of the value of source or sink terms is a necessity. As long ago as 1966, Munk reviewed "abyssal recipes" for modelling the one-dimensional vertical distribution of oceanographic variables, assuming that the vertical structure of the tracer distribution in the ocean is solely determined by advection due to large-scale or global upwelling, turbulent diffusion and a simple source or sink term. From the vertical distribution of conservative tracers a scale height of the advection-diffusion balance is estimated. If the upwelling velocity is known from independent measurements, it is possible to infer the value of the turbulent mixing coefficient as well as source/sink terms from the observed non-conservative tracer distributions.

In this paper we shall discuss the assumptions underlying Munk's "abyssal recipes". These recipes or models are used for the quantitative interpretation of tracer observations from the South Banda basin, carried out during the cooperative Indonesian-Dutch Snellius II expedition, in terms of diffusion coefficients and source and sink terms. The Banda basin, situated in the central part of the Banda Sea (Fig. 1), is divided into a northern and a southern basin, separated by a ridge with a sill depth of about 4 000 m. Since the deep waters in the South Banda basin form a small-scale semi-enclosed system, the condition of a one-dimensional system is better met there than in the Pacific where Munk has applied his recipes. In the Banda Sea, the water below a depth of 1 500 m comes from a single source, the Lifamatola sill (*see* below), whereas in the larger ocean basins the intermediate, deep and bottom waters may come

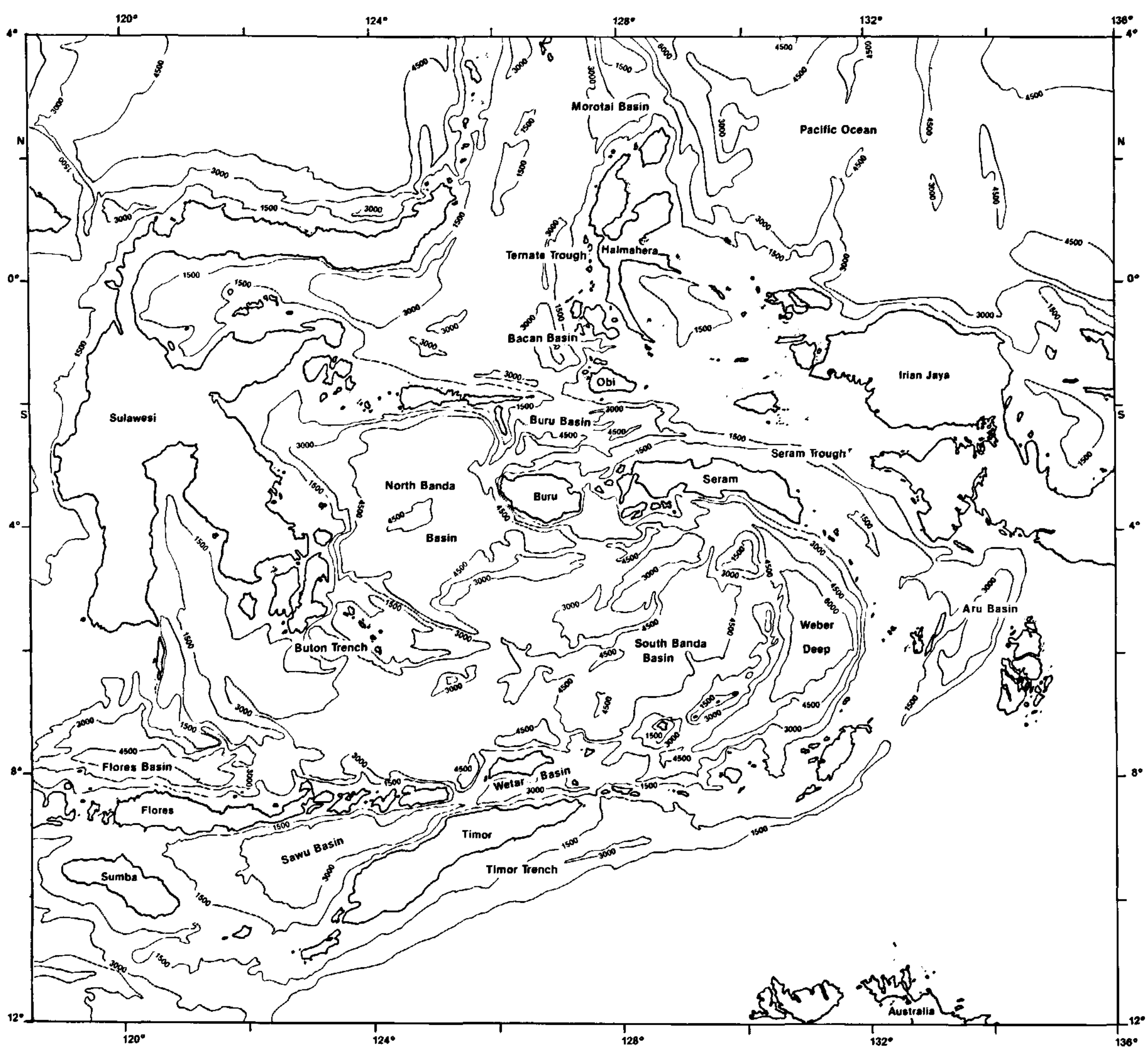


Figure 1

Depth contours (metres) of the east Indonesian basins. The main inflow of Pacific deep water into the Banda Sea occurs through Lifamatola strait, west of the island Obi. From there this water flows via the Buru basin and the North Banda basin towards the South Banda basin.

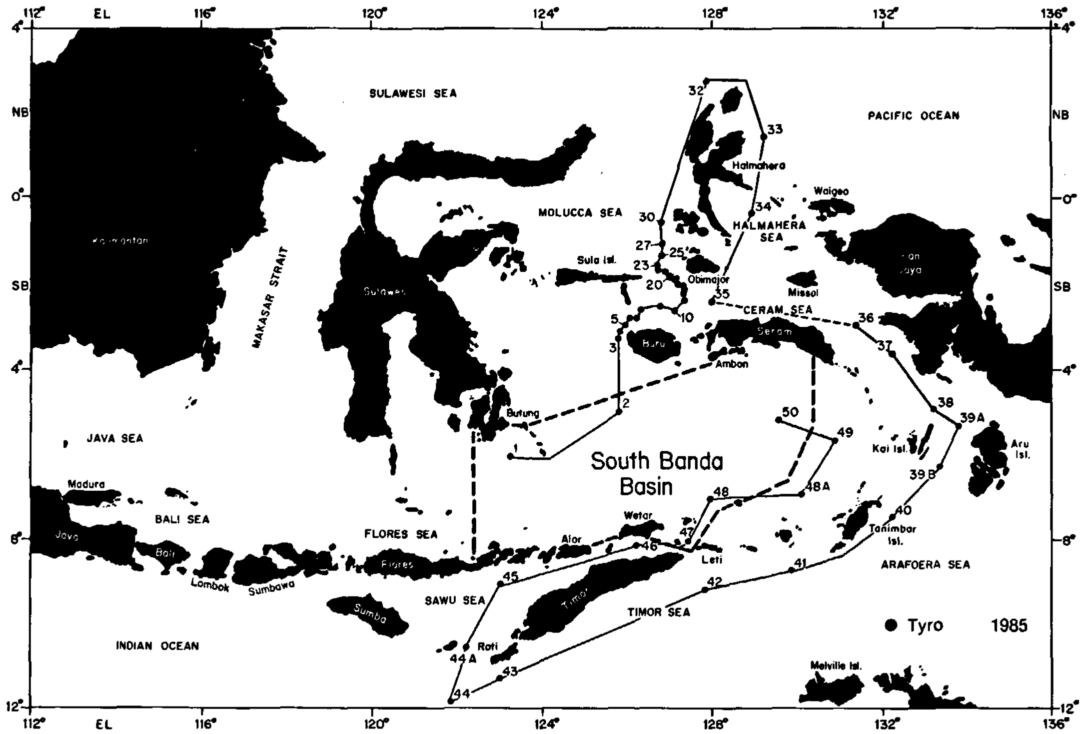


Figure 2

Map of East Indonesia. The stations occupied by R/V Tyro *are indicated with black dots. The dashed line delimits the South Banda basin as used in this study. The main inflow sill in Lifamatola strait is near station 20.*

from a range of geographically distributed sources, each generating its own typical water mass properties. Edmond *et al.* (1979) even stated that in the large ocean basins the vertical structure of tracer distributions may be mainly determined by horizontal advection.

The flushing of the deep water in the South Banda basin appears to occur so rapidly that the decay of ^{14}C does not cause an observable deviation in the ^{14}C profile. Therefore no scale height for the decay of ^{14}C or subsequent estimate of the upwelling velocity can be established from observations. Estimates of the flushing time of the Banda Sea from other sources are used instead.

THE FLUSHING OF THE SOUTH BANDA BASIN

In January and February 1985, R/V *Tyro* carried out a hydrographic survey in East Indonesian waters (Fig. 1). At 47 different positions (Fig. 2) CTD-casts were recorded and water samples were taken at several different levels. The positions were chosen to follow the flow of deep water from the Pacific and Indian Ocean through the deep East Indonesian basins. The observed hydrography confirmed earlier reports (van Riel, 1934; Postma, 1958; Wyrtki,

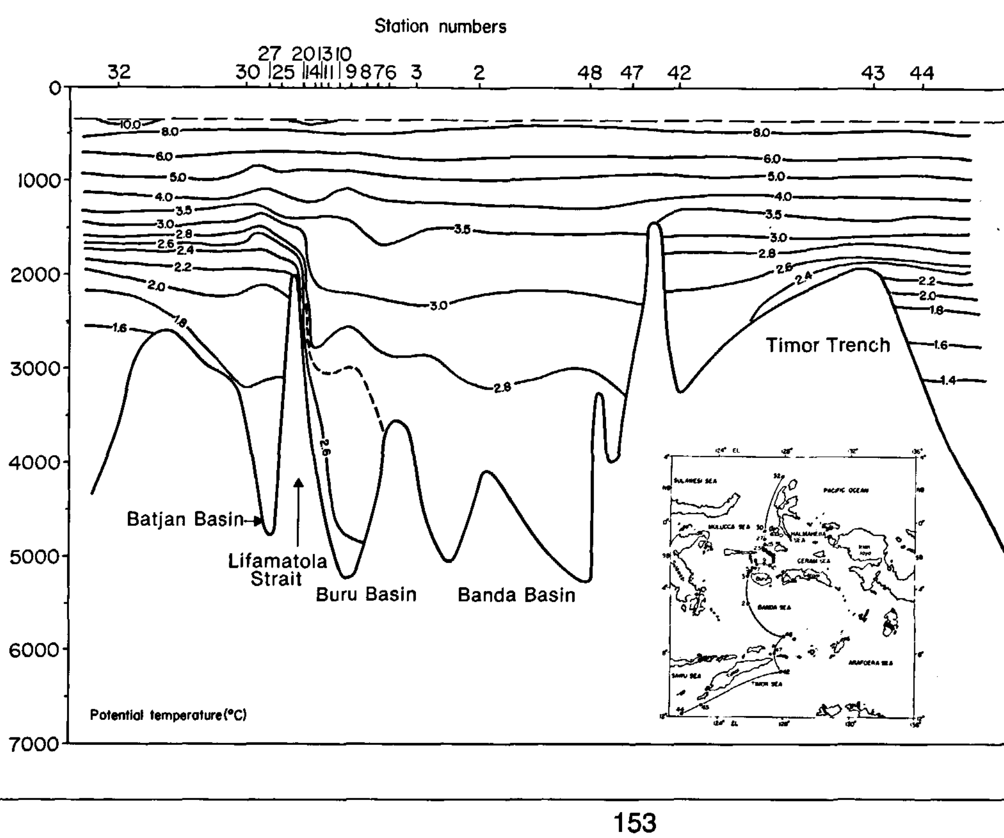


Figure 3

North-south section of the potential temperature (°C) from the Morotai basin (station 32) to the Indian Ocean (station 44)

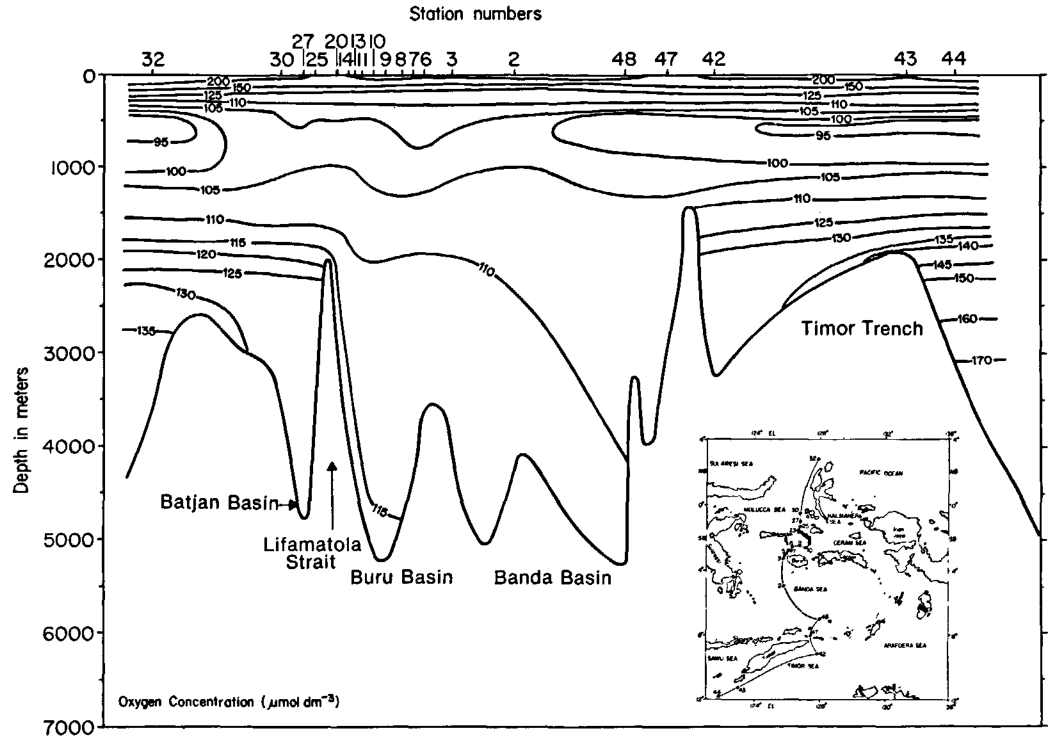
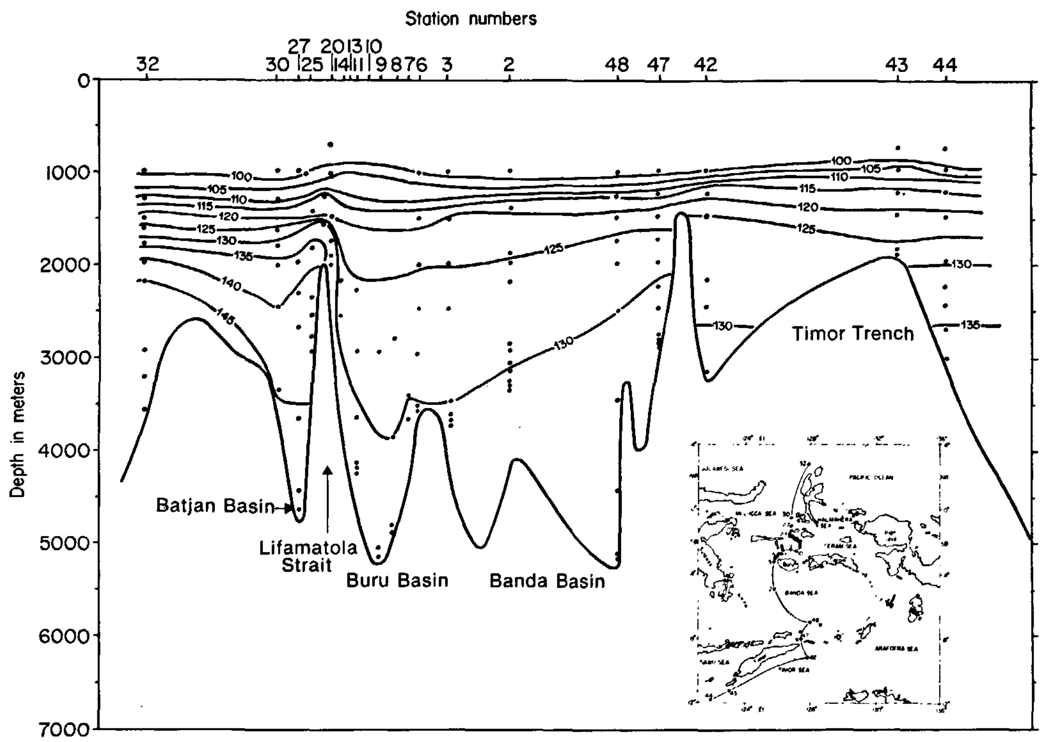


Figure 4

North-south section of the dissolved oxygen concentration ($\mu mol/dm^3$) from the Morotai basin (station 32) to the Indian Ocean (station 44).

Figure 5

North-south section of the dissolved silica concentration ($\mu mol/dm^3$) from the Morotai basin (station 32) to the Indian Ocean (station 44).



1961; Broecker *et al.*, 1986) and has been extensively discussed elsewhere (van Aken *et al.*, 1988; van Bennekom, 1988; Rommets, 1988; Postma and Mook, 1988).

The distribution of the potential temperature (Fig. 3) shows that at depths between about 1 500 and 2 000 m Pacific deep water ($\theta \approx 2.3$ °C, $S \approx 34.62$ at the inflow level) flows from the Batjan basin across the sill in Lifamatola strait into the Buru basin. Downstream from the sill the steep gradient of the isotherms below 1 500 m shows that the overflow water, while entraining and mixing with overlying water, reaches the bottom layer of the Buru basin quite rapidly. The individual θ and S profiles published by van Aken *et al.* (1986) show that the water descending from the Lifamatola sill flows in a quasi- homogeneous bottom layer with a thickness of about 500 m to depths certainly greater than 3 000 m. From the Buru basin, water with a potential temperature slightly above 2.7 °C enters the North Banda basin across a sill at a depth of 3 500 m, whence it flows across a sill at about 4 000 m into the South Banda basin. In the Seram and Banda Seas the lowest potential temperature was always encountered near the bottom. The bottom water in the Banda basins ultimately moves upwards to a depth of some 1 250 m in order to leave the basin, probably in the direction of the Timor trench (van Bennekom, 1988; van Aken *et al.*, 1988). During this upwelling the water is heated as a result of turbulent mixing with the overlying warmer water. The nearly horizontal position of the isotherms (Fig. 3) in the Banda basins indicates that the southward transport proceeds relatively rapidly and that the deep upwelling is a basin-wide phenomenon.

The distribution of dissolved O_2 (Fig. 4) confirms this flushing scheme. However the isolines of O_2 in the deep parts of both Banda basins tend to descend southwards due to O_2 consumption. The rate of O_2 consumption in the deep waters is however not known quantitatively. The distribution of dissolved silica, Si (Fig. 5), shows southward ascending isolines, due to the introduction of dissolved silica into the southward flowing water (van Bennekom, 1988). The dissolution of diatom frustules in the water column is not well known quantitatively, but is assumed to play a role predominantly in the upper 1000 m. Edmond *et al.* (1979) and Broecker (1981) conclude from qualitative arguments that *in situ* dissolution can be neglected as a source term for the Si budget of deep watermasses. The main agent for introduction of Si into the watermasses is assumed to be diffusion from the pore water of the sediment. Quantitative estimates of the dissolution of Si in the deep water column will improve the understanding of the different terms in the silica budget.

With the exception of the Weber Deep (van Bennekom, 1988), no indication for the lateral spreading of inflowing water at intermediate depths below the sill is observed in any of the Indonesian basins. Therefore the flushing of deep basins like the South Banda basin may be summarized schematically, as in Figure 6. The inflow of water from the inflow sill into the bottom layer takes place in a fast-moving and narrow gravity current. Thereupon a convergent bottom current feeds basin-wide deep upwelling which is ultimately incorporated in the lower thermocline. Here it can leave the basin across an outflow sill. In a stationary state, advection of conservative properties due to upwelling is balanced by vertical diffusion, while for non conservative properties source or sink terms have to be added to this balance.

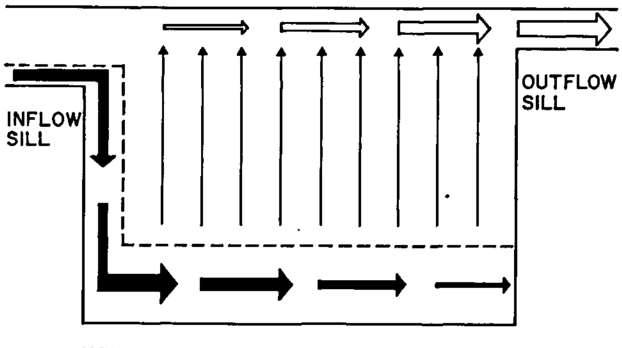


Figure 6

Schematic picture of the flushing of the South Banda basin. The thick black arrows indicate the inflow and the convergent southward transport in the bottom layer while the thick open arrows indicate the divergent outflow in the deep thermocline. The thin vertical arrows show the basin-wide upwelling.

Table

Estimates of flushing times, τ, and vertical velocities, w, of the Banda Sea.

Source	τ (a)	w (m/a)
van Aken *et al.* (1988)	29	75
Broecker *et al.* (1986)	88	25
van Bennekom (1988)	20	110
Postma and Mook (1988)	25	88

Early estimates of the flushing rate of the Banda Sea have resulted in a flushing time of the deep water of about 300 a (Postma, 1958; 1 a is the time unit of one year, according to ISO recommendations). However, Broecker *et al.* (1986) found evidence for a flushing time well below 100 a. From current measurements in Lifamatola strait, carried out in February-March 1985, van Aken *et al.* (1988) have estimated an inflow below 1 500 m of $1.5 \cdot 10^6$ m³/s. With a mean depth of about 3 700 m and a volume of the whole Seram and Banda Sea below a depth of 1 500 m amounting to about $1.4 \cdot 10^{15}$ m³, this implies a flushing time for the water below 1 500 m in the Seram and Banda Seas of 29 a and a mean upwelling velocity of w = 75 m/a. The current velocities reported by Broecker *et al.* (1986) for August 1976 are, however, about three times lower than those reported by van Aken *et al.* (1988), suggesting a flushing time of 88 a and an upwelling velocity w = 25 m/a. Van Bennekom (1988) has estimated the flushing time of the Banda Sea from the dissolved silica excess and the flux of dissolved silica out of the sediment to be 20 a, which implies an upwelling velocity w = 110 m/a. Postma and Mook (1988) reported the flushing time of the Banda basins, based on ^{14}C - O_2 correlations, to be of the order of 25 a. This estimate results in an upwelling velocity w = 88 m/a. All these recent estimates confirm that the Banda basins are flushed relatively rapidly. The distribution of the isolines (Fig.s 3, 4 and 5) shows the upwelling to be distributed evenly over the basins. Here we take as a typical value for the upwelling velocity in the South Banda basin the average of the estimates of w, listed in the Table. This mean value amounts to 75 m/a. The standard deviation of the individual estimates relative to the mean is 36 m/a, so that the estimated standard deviation of the mean value is $36/\sqrt{4} = 18$ m/a.

In order to study quantitatively the advection-diffusion balance in the South Banda basin, data from CTD-stations 1, 2, 47, 48 and 50 (Fig. 2) have been selected. These stations cover almost the entire South Banda basin. Stations 48A and 49 are situated over the Weber Deep, which is separated from the South Banda basin by a sill at about 4 000 m depth. These two stations have not been included in the selection, although the θ and S profiles hardly differ from the more westerly stations. However there are indications of non-stationarity in the Weber Deep (van Bennekom, 1988), while at levels below 3 000 m the distributions of Si, O_2 and ^{14}C seem to be influenced by relatively slow flushing of the deeper parts of the Weber Deep (van Bennekom, 1988; Postma and Mook, 1988). Data from the

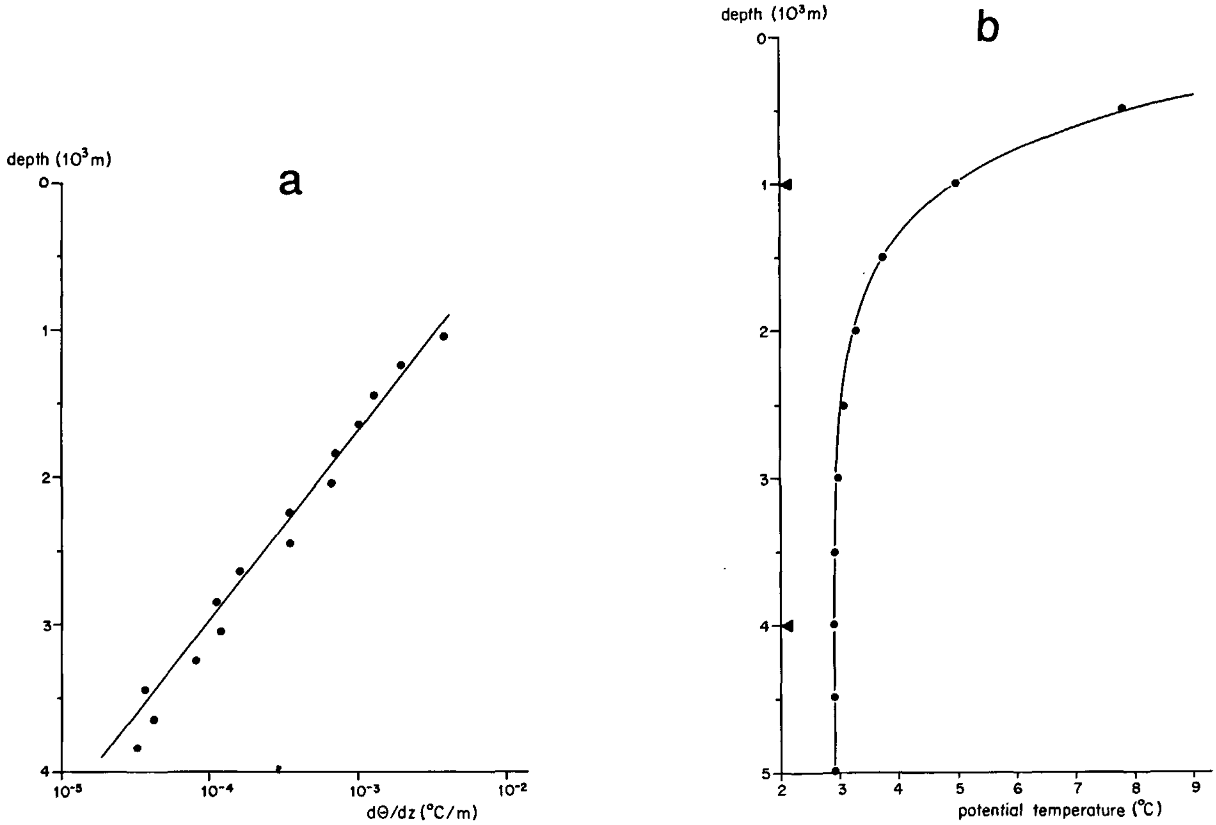


Figure 7

Profiles of (a) the mean vertical gradient of the potential temperature and (b) the mean potential temperature in the South Banda basin. Dots are observed values and lines are model results according to equation 4. The advective-diffusive scale height L is estimated from (a) to be 560 m.

selected stations have been averaged in order to obtain profiles of the mean vertical structure of tracer distributions in the South Banda basin, beyond the inflow region. The typical standard deviation of the potential temperature profiles between 1 500 and 4 000 m is less than 0.03 °C. In this depth range the potential temperature in the Weber Deep differs by less than 0.03 °C from the calculated mean profile. Comparison of these profiles with data published elsewhere (van Riel *et al.*, 1957; Postma, 1958; Broecker *et al.*, 1986) reveal that our mean profiles are representative for the whole South Banda basin, although they are based on only a small number of hydrographic stations. Mean profiles of potential temperature gradient (Fig. 7 *a*), potential temperature (Fig. 7 *b*), O_2 concentration (Fig. 9), orthophosphate (PO_4-P) concentration (Fig. 10) and Si concentration (Fig. 11) have been determined. The mean θ-S relation for the South Banda basin is shown in Figure 8. The ^{14}C data are scarce and unevenly distributed over the selected stations, so that it is not possible to construct a reliable mean profile of ^{14}C. Therefore only the individual data points have been plotted (Fig. 12).

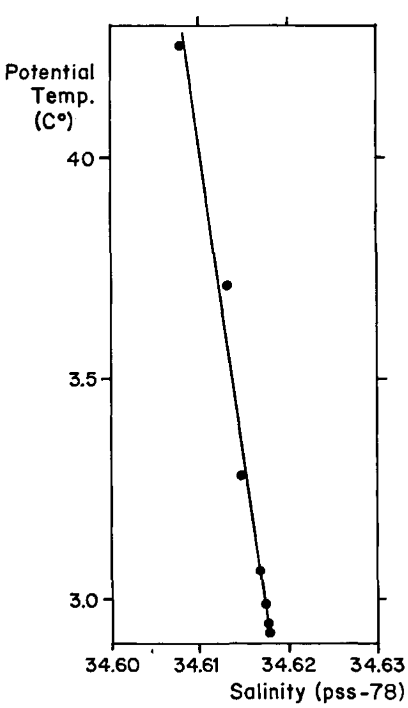


Figure 8

Potential temperature vs salinity diagram for the mean profiles in the South Banda basin. Dots are observed values below 1000 m. The expected linear relation according to equation 4 is given by the straight line.

THE MODEL

In order to clarify the assumptions used by Munk (1966) and to facilitate a subsequent interpretation of the results, the derivation of a one-dimensional advection-diffusion model from the full three-dimensional equations is discussed in this chapter.

The budget of a tracer concentration, C, in an incompressible fluid can be described by:

$$\partial C/\partial t + \nabla_h.U_hC + \partial wC/\partial z = \nabla_h.D_l\nabla_hC + \partial/\partial z\,(D_v.\partial C/\partial z) + Q \qquad (1)$$

In this equation t and z are the time and the vertical coordinate (positive upwards). The index h indicates horizontal components only, U is the velocity vector while w is the vertical component of the velocity. D_l and D_v are the lateral and vertical coefficients of turbulent diffusion, while the source/sink term Q takes care of non-conservative processes. By assuming stationarity and averaging over a horizontal surface A, with boundary s, equation 1 can be reduced to a one-dimensional equation which reads:

$$1/A \int U_nC\,ds + \partial/\partial z <wC> = 1/A \int D_l\,\nabla_n\,C\,ds + \partial/\partial z <D_v.\,\partial C/\partial z> + <Q> \qquad (2)$$

Horizontal averages are indicated with brackets <.>, while the index n indicates the horizontal component, perpendicular to the boundary.

In order to come from equation 2 to the equation as used by Munk (1966), some additional assumptions have to be made. Firstly, horizontal advection across the boundary s is assumed to be negligible. This implies that A has to be chosen outside the layers where the inflowing water feeds the convergent horizontal transport in the bottom layer or where the outflowing thermocline water is fed by the basin-wide upwelling. That is, A has to be positioned well above the bottom layer, but below the level of the outflow sill. And at the levels of A, the vertical variation of w is assumed to be negligible. Secondly, it is assumed that w and C are not correlated so that their averaged product can be written as the product of the averages. This shows that it is important to choose A well within the upwelling area and completely outside the descending inflow. Otherwise the resulting covariance of w and C will act as a kind of diffusion term, thereby reducing the effective vertical diffusion. Thirdly, if the lateral diffusion across the boundary s cannot be neglected, this term will act as a spurious source/sink term, even for conservative tracers. The one-dimensional model to be derived from equation 2 cannot discriminate between sediment-water interaction at the walls of a basin and processes in the water column. Additional information is needed to discriminate these processes. Finally we assume that D_v is constant, not depending on the depth within the basin, and not correlated with $\partial C/\partial z$. Note here that no assumption is made on horizontal homogeneity of C; C may vary within A, as long as the four assumptions are fulfilled satisfactorily. With these assumptions, we can reduce equation 2, dropping the use of brackets for averages, to the form used by Munk (1966), that is:

$$w.\partial C/\partial z - D_v.\,\partial^2 C/\partial z^2 = Q \qquad (3)$$

The profile of a conservative tracer (Q = 0) with known values C_1 and C_2 at levels z_1 and z_2 can be derived from equation 3 to be

$$C\,(z) = C_1 + (C_2 - C_1)\,\frac{\exp\,[(z-z_1)/L] - 1}{\exp\,(\Delta z/L) - 1} \qquad (4)$$

with $\Delta z = z_2 - z_1$. The scale height $L = D_v/w$ is typical for the advective diffusive balance controlling the vertical structure of conservative tracers.

For non-conservative tracers whereby the source/sink term in first approximation can be modelled as a zeroth-order reaction, that is Q = constant, the solution of equation 3 is a linearly modified version of equation 4,

$$C\,(z) = C_1 + (C_2-C_1)\,\frac{\exp\,[(z-z_1)/L] - 1}{\exp\,(\Delta z/L) - 1}\,(1 - \beta) + \beta.\,(z - z_1)/\Delta z \qquad (5)$$

the dimensionless parameter $\beta = (Q\Delta z)/w(C_2-C_1)$ is the ratio of the constant source term and the mean advective convergence between z_1 and z_2.

If the source/sink term can be modelled as a first-order reaction, that is $Q = - C/\tau$ as in radioactive decay, C can be written as:

$$C(z) = C^+ \exp\,[(z - z_1)\alpha^+/2L] - C^- \exp\,[(z-z_1)\alpha^-/2L] \qquad (6)$$

with

$$C^{\pm} = \frac{C_2 - C_1 \exp\,(\Delta z.\alpha^{\pm}/2L)}{\exp\,(\Delta z.\alpha^+/2L) - \exp\,(\Delta z\ \alpha\text{-}/2)}$$

and

$$\alpha^{\pm} = 1/2\,[1 \pm \sqrt{(1 + 4L/w\tau)}]$$

The dimensionless parameters $\alpha^{\pm}$ are determined by the ratio of the advective-diffusive scale height L and a scale height typical for the decay, $w\tau$. This latter scale height is the vertical distance a water parcel, with velocity w, will travel during the e-folding period τ of the decay process. If this ratio becomes very small, that is if $L << w\tau$, solution (6) naturally converges to equation 4, the solution for conservative tracers.

DISCUSSION

According to equation 4, the vertical gradient of conservative tracers varies exponentially with depth. The logarithmic plot in Figure 7 *a* shows that the potential temperature is indeed stratified according to equation 4. From these data the scale height L is estimated to be L = 560 m. Since salinity is also a conservative property which is expected to have a vertical structure according to equation 4 with the same scale height L, the θ-S relation will be linear. This is confirmed for the mean θ and S profiles (Fig. 8). With boundary conditions at 1 000 and 4 000 m, the mean profile of the potential temperature in the South Banda basin has been calculated with equation 4. Comparison of the theoretical curve with the data (Fig. 7 *b*) shows no systematic effects of spurious source/sink terms at intermediate depths due to lateral advection and diffusion from the inflow area or from the Flores basin or Weber deep. The assumptions underlying equation 4 as discussed in the previous section appear to be met satisfactorily for the temperature and salinity fields in the depth range between 1 000 and 4 000 m. But the horizontal area enclosed by the 4 000 m depth contour in the South Banda basin is about one-half of the area enclosed by the 1 000 m contour. Therefore one can expect w to decrease upwards. The diffusion coefficient D_v, however, is also expected to decrease upwards due to increasing stability in that direction (Sarmiento *et al.*, 1976). Both effects appear to counteract each other so that L is nearly constant over the whole depth range as shown in Figure 7 *a*. The quantitative estimates presented below are only a representative average for the parameter values in the depth range where the model is applied.

With the estimate of L, the value of D_v can be computed if a representative estimate of the upwelling velocity, w, is known. With the recent estimates of w, discussed in the second section, that is w = 75 (± 18) m/a, a diffusion coefficient $D_v = 1.3\ (\pm\ 0.3)\cdot 10^{-3}$ m²/s is obtained. This value seems to be high in comparison with most other studies which give, for the ocean interior, values of the order of 10^{-4} m²/s (Garrett, 1979), but are of the same order of magnitude as the values Sarmiento *et al.* (1976) obtained above the benthic boundary layer by using ^{228}Ra profiles. Minster (1989) also derived such large diffusivities from TTO observations in the western boundary of the Atlantic Ocean. Garrett (1979) attributes such high diffusivities to boundary mixing driven by internal tides at the side walls of the basins, and subsequent lateral redistribution along isopycnals. Berger *et al.* (1988) have estimated the diffusion coefficient in the benthic boundary layer of the deep East Indonesian basins outside the sill areas from ^{222}Rn distributions to be $5.5\ (\pm\ 2.2)\cdot 10^{-3}$ m²/s, while over strongly sloping topography D_v values of the order of 10^{-1} m²/s have been observed occasionally in the lowest 200 m.

The power density P, needed to maintain the turbulent mass flux between depth levels z_1 and z_2, can be defined to be

$$P = gD_v\,(\rho_1 - \rho_2), \qquad (7)$$

where g is the gravitational acceleration and $\rho_1 - \rho_2$ is the density difference between z_1 and z_2, corrected for adiabatic effects. With the mean profiles determined for the South Banda basin, P between 4 000 and 1 500 m proves to be $1.6\cdot 10^{-3}$ W/m². With a mixing efficiency of 25 % (Thorpe, 1973), this requires a turbulent energy input of at least $6.4\cdot 10^{-3}$W/m² or for the whole Banda Sea a total of $4\cdot 10^{+9}$ W. This energy may be derived from the tides.

In restricted basins like the Banda Sea, internal tides can be generated by interaction of the barotropic tide with the sill topography in a stratified sea. Stigebrandt (1976) derived with a simple two-layer model the energy input into the internal tides at the sill to be:

$$W = 1/2\ \rho_0\,\alpha^2\,(h_0^2/h_1 + h_0)BC_g. \qquad (8)$$

In this expression ρ_0 is a reference density for seawater; a is the velocity amplitude of the barotropic tide; h_0 and h_1 are the thickncss of the lower and the upper layer over the sill; B is the width of the sill and C_g is the group velocity of the internal tides. For the situation near the sill in Lifamatola strait (van Aken *et al.*, 1988) equation 8 leads to an estimate of $W = 2 - 8\cdot 10^{+9}$ W, depending on the choice of h_0 and h_1. Therefore Lifamatola strait alone can already supply a considerable part of the energy needed to maintain the deep turbulent mixing in the Banda Sea. Since there are several similar, although shallower sills between the islands surrounding the Banda Sea with comparable tidal amplitudes (Lek, 1938), the energy input into the internal tides is probably large enough to support the turbulent mixing in a large part of the deep Banda Sea. Therefore tidally driven boundary mixing is a probable explanation for the high D_v values derived for the South Banda basin.

It has to be noted here that for a certain observed value of L, w and D_v are proportional. Therefore any value of w can be obtained by an appropriate choice of D_v. In this way

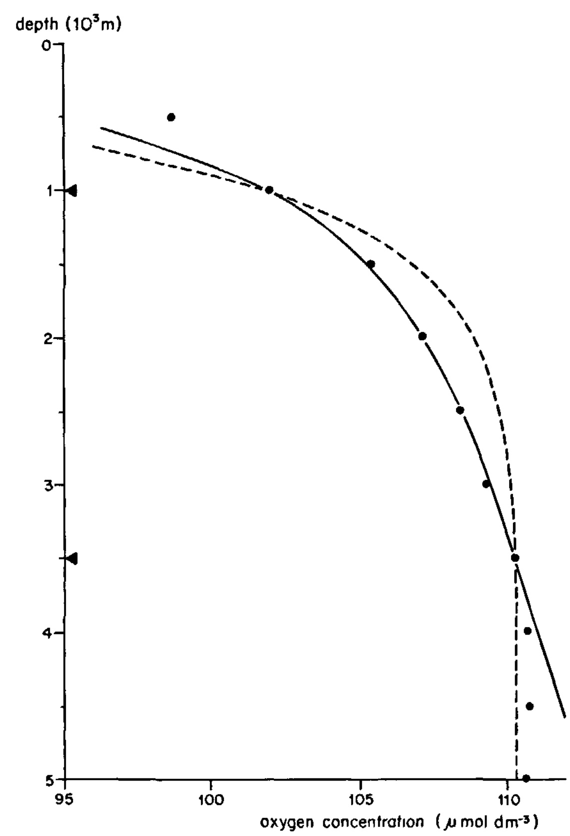


Figure 9

Profile of the mean oxygen concentration in the South Banda basin. Dots are based on observations. The dashed line has been constructed according to equation 4 with L = 560 m. The full line is based on equation 5 with L = 560 m and β = 0.5. Boundary conditions have been applied at the levels of the triangles.

Wyrtki (1961) obtained confirmation of an earlier estimate of the flushing time of the Banda Sea to be 300 a by an appropriate but arbitrary choice of D_v.

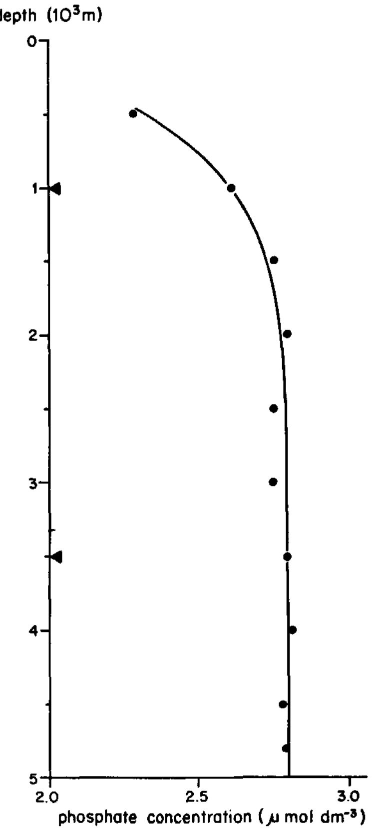
Comparison of the mean O_2 profile with a curve according to equation 4 with L = 560 m (the dashed line in Fig. 9) shows that the vertical distribution of O_2 cannot be explained without a sink term. The good fit of a curve according to equation 5 with $\beta = 0.5$ (the full line in Fig. 9) indicates that the O_2 consumption between 1000 and 3500 m depth can be quite satisfactorily modelled with a zeroth-order process, independent of depth. With $\beta = 0.5$ and w = 75 (± 18) m/a the O_2 consumption appears to be -Q = 0.12 (± 0.03) mmol/(m^3 a), which is of the same order of magnitude as the O_2 consumption estimated independently by Postma and Mook (1988) from O_2-$\Delta^{14}C$ correlations to be 0.08 mmol/(m^3 a). It should be noted here that if the lateral diffusion in equation 2 cannot be ignored, any O_2 flux into the bottom at levels between z_1 and z_2 will be incorporated as a spurious sink term. Since 45 % of that part of the South Banda basin which has depths greater than 1 000 m is in the 1 000 to 3 500 m depth range, the maximum possible O_2 flux into the sediment of the side walls of the basin is 2 500 . 0.12/0.45 = 670 mmol/(m^2 a). This value is at least 2.5 time the mean value determined by Berelson *et al.* (1987) for Californian borderland basins and up to seven times the values determined by Reimers *et al.* (1984) for the deep tropical central Pacific, both by means of benthic measurements. The vertical O_2 gradients in most East Indonesian basins, with the exception of the Aru and Sawu basin, were negligible in the benthic boundary layer (van Aken *et al.*, 1988), in contrast with the vertical Si gradients. This suggests that probably a large part of the O_2 consumption in the deep South Banda basin occurs in the water column, while the O_2 flux into the sediment is of minor importance for the O_2 budget.

If O_2 is used for the mineralization of organic matter, nutrients are released with a more or less constant ratio, the Redfield ratio. With an atomic Redfield ratio $-O_2 : PO_4\text{-}P$ = 172 : 1 (Postma, 1964; Takahashi *et al.*, 1985) the source term for PO_4-P can be calculated from the already estimated O_2 consumption. Substitution of this source term into equation 5, with L = 560 m, w = 75 m^2/a and application of boundary conditions at 1000 and 3500 m gives a model curve for the PO_4-P concentration. Comparison of this model curve with the observational data (Fig. 10) shows that such a zeroth-order source term with the already derived w and D_v values matches the observations within the accuracy of the PO_4-P determinations (0.05 mmol/m^3; van Bennekom and Muchtar, 1988). However, a conservative profile according to equation 4 will not deviate more than 0.01 mmol/m^3 from the drawn line in Figure 10, so no firm conclusions on the existence of a source term can be made.

Comparison of a curve according to equation 4 with L = 560 m with the Si data from the South Banda basin shows a small but systematic mismatch (Fig. 11, dashed line). Apparently the Si budget needs a source term. Curve-fitting of equation 5 with the already derived values for w

Figure 10

Profile of the mean orthophosphate concentration in the South Banda basin. Dots are based on observations. The line has been constructed according to equation 5 with L = 560 m and β =-0.14. This β-value has been determined from the oxygen sink term by means of a Redfield ratio. The conservative profile according to equation 4 deviates less than 0.01 μmol/dm^3 from the non-conservative line and is therefore not drawn. Levels where boundary conditions have been applied are indicated with triangles.



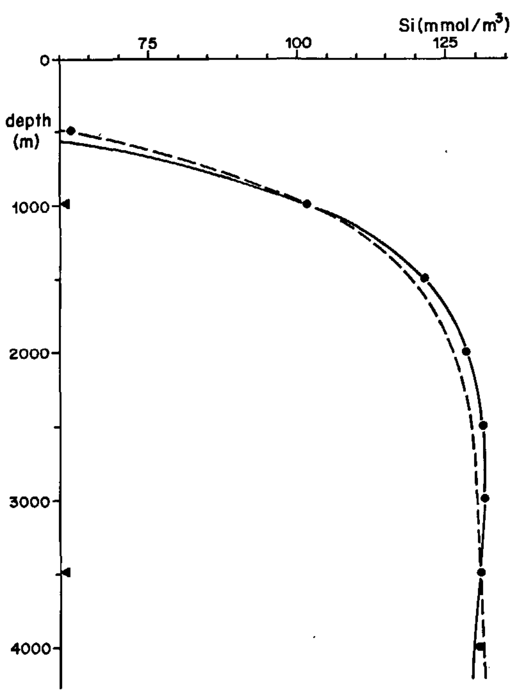


Figure 11

Profile of the mean dissolved silica concentration in the South Banda basin. Dots are based on observations, whereby the lowest 200 m of each station, where considerable gradients due to sediment-water interaction have been observed, have been left out of the analysis. The dashed line indicates the conservative profile according to equation 4 with L = 560 m, while the full line is based on the non-conservative profile according to equation 6 with L = 560 m and β =-0.18. Levels where boundary conditions have been applied are indicated with triangles.

and D_v shows that the data between 1 000 and 3 500 m can be modelled quite well with a zeroth-order source term Q = 0.17 ± 0.04 mmol/(m^3 a) (Fig. 11, full line). Integration of Q between z = 1 000 m and z = 3 500 m results in a total Si input into this layer of 400 ± 100 mmol/(m^2 a). Van Bennekom (1988) has determined the mean Si flux from the sediment into the water of the South Banda basin from pore water gradients to be 900 mmol/(m^2 a). This Si flux supports a small but significant vertical Si gradient in the benthic boundary layer of the South Banda basin (van Bennekom, 1988). As stated above, of the part of the South Banda basin with a depth greater than 1 000 m, 45% is in the range 1 000 to 3 500 m. With a mean Si flux from the bottom of 900 mmol/(m^2 a), the sediment-water interaction will contribute a spurious source term to Q of 0.45·900 = 405 mmol/(m^2 a^{-1}) which happens to be nearly equal to the total estimated Q. It can be concluded that a large part, if not all, of the apparent Si input into the deep South Banda basin originates from the sediment. Dissolution of sinking diatom frustules in the water column below 1 000 m depth can be ignored as a term in the Si budget, relative to the effects of sediment-water interaction.

Ideally the sink term for the decay of the radio-isotope ^{14}C in the dissolved components of the CO_2 system is known to be Q = -C/τ, where τ = 8 268 a is the decay time (τ = half life/ln 2). Complications may arise from the input of CO_2 either by mineralization of organic matter or by dissolution of carbonate. Both processes are known to occur in the East Indonesian basins (Rommets, 1988). If the CO_2 introduced in this way is relatively young, this process will reduce the sink term in equation 3 and even may turn it into a source term. If the introduced CO_2 happens to be relatively old, the sink term will increase. We assume here with Postma and Mook (1988) that both effects are relatively small, so that (6) can be used for the modelling of the ^{14}C profile. With w = 75 m/a, the decay scale height wτ = 620 km. This gives values for the =-parameters of α^+ = 1.0009 and α^- = 0.0009. Therefore, due to the fast flushing of the deep water, the ^{14}C profile is expected hardly to differ from that of a conservative tracer. This prevents us from obtaining an independent estimate of w. Comparison of the curve according to equation 6, based on the previously estimated L and w values, with the individual data points (Fig. 12) does not show any systematic deviation. A spurious source or sink term due to the introduction of either new or old CO_2 appears not to be needed to model the ^{14}C distribution on the time scale of the flushing of the South Banda basin.

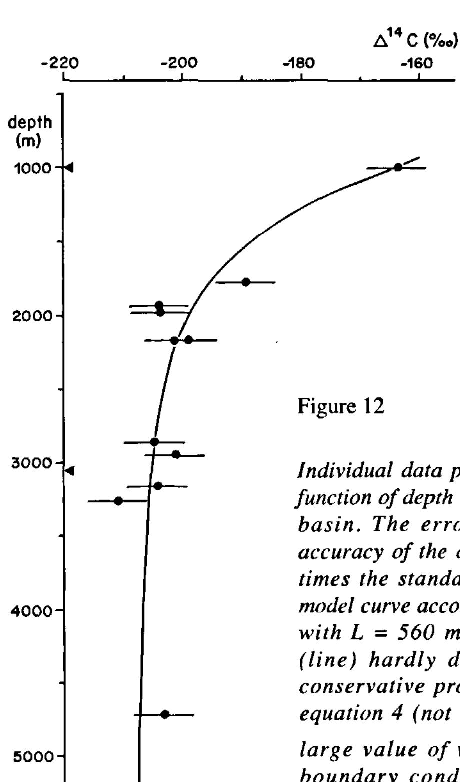


Figure 12

Individual data points of $\Delta^{14}C$ as a function of depth in the South Banda basin. The error bars show the accuracy of the determination (two times the standard deviation. The model curve according to equation 6 with L = 560 m and w = 75 m/a (line) hardly deviates from the conservative profile according to equation 4 (not drawn) due to the large value of wτ. Levels where boundary conditions have been applied are indicated with triangles.

SUMMARY AND CONCLUSIONS

The South Banda basin is a relatively small but deep semi-enclosed basin. The deep water in this basin is flushed in about 30 a with Pacific deep water which flows across the sill in Lifamatola strait, through the Buru basin and the North Banda basin into the South Banda basin. The mean vertical profiles for several tracers between 1 000 and 3 500 m depth can be modelled satisfactorily with the simple, one-dimensional models reviewed by Munk (1966) although the horizontal area of this basin almost doubles between 3 500 and 1 000 m depth.

The vertical gradient of the potential temperature has the expected exponential form with an advective-diffusive scale height of 560 m while the θ-S relation has the expected linear form. The residence time of the deep water is so small, O (30 a), that the flushing time cannot be estimated independently from the vertical ^{14}C distribution. Therefore the upwelling velocity, w, has been estimated from a compilation of recent, mutually independent estimates of the flushing time to be w = 75 (± 18) m/a. From these estimates the effective vertical diffusion coefficient has been computed to be D_v = 1.3 (± 0.3)·10^{-3} m^2/s. This value is an order of magnitude larger than most other estimates for the deep ocean, probably due to the relatively large influence of boundary mixing, driven by internal tides. The observed vertical distribution of ^{14}C is consistent with these estimates.

The vertical distributions of the biogeochemical tracers O_2 and PO_4-P appear to be consistent with constant, zeroth-order source/sink terms whereby the ratio of the O_2 consumption and PO_4-P production is prescribed to be the Redfield ratio: -O_2 : PO_4-P = 172 : 1. However, a source term for PO_4-P is not needed to fit the observations. By curve fitting an oxygen consumption -Q = 0.12 (± 0.03) mmol/(m^3 a) has been estimated. The upper limit of the O_2 flux into the sediment between a depth of 1 000 and 3 500 m, based on this Q estimate amounts to 670 mmol/(m^2 a), which value is much larger than the range of estimates for

California borderland basins and for the deep tropical Pacific. No significant O_2 gradients have been observed in the benthic boundary layer. So it is probable that a large part of the O_2 loss in the deep South Banda basin is due to processes in the water column.

The observed mean Si profile is also consistent with a constant zeroth-order source term. The total Si input in the depth interval between 1 000 and 3 500 m has been estimated by curve fitting to be 400 ± 100 mmol/(m^2 a). This value appears to be equal to the total input due to sediment-water interaction in that depth interval, calculated from independently determined Si fluxes from the sediment. Dissolution of diatom frustules in the water column below 1 000 m depth can therefore be ignored as a source of Si in the South Banda basin.

In conclusion, we may note that Munk's simple recipes can be used successfully to model the mean vertical distributions of tracers as observed between 1 000 and 3 500 m in the South Banda basin. However, only effective bulk parameters can be obtained from one dimensional modelling. Additional information from other sources strongly suggests that some parameters which have been estimated quantitatively are mainly determined by processes in the benthic boundary layer and in the sediment.

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