Estimating vertical velocity on the Equator

Equatorial undercurrent
Pacific circulation
Upwelling
Vertical velocity
Cross-isotherm flux

Sous-courant équatorial
Circulation du Pacifique
Upwelling
Vitesse verticale
Flux trans-isotherme

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ABSTRACT

Vertical velocity on the Equator in the Central Pacific is estimated in four ways using mean CTD measurements assuming the following : 1) that the Equatorial Undercurrent core is a trajectory of flow ; 2) that the flow is along isotherms ; 3) that mass is conserved in the meridional-vertical plane ; and 4) that mass is conserved three-dimensionally. It is remarkable that by using such simple and independent methods, all results agree so well, exhibiting an average vertical velocity at 90 m of 1.9×10^{-3} cm s^{-1}, with a standard deviation of only 0.3×10^{-3} cm s^{-1}. The results reported here demonstrate the ease with which reliable estimates of vertical velocity may be obtained on the Equator, which may be useful in studies of the effects of upwelling on productivity and chemical nutrient renewal in the equatorial surface layer.

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RÉSUMÉ

Estimations de la vitesse verticale à l'équateur

Quatre techniques pour estimer la vitesse verticale à l'équateur dans la partie centrale du Pacifique sont présentées ; elles utilisent toutes des données hydrologiques moyennes et reposent sur les hypothèses suivantes, respectivement : 1) le noyau du sous-courant équatorial constitue une trajectoire de l'écoulement ; 2) l'écoulement se fait le long des isothermes ; 3) la masse est conservée dans le plan vertical méridien ; 4) la masse est conservée dans les trois dimensions. Les estimations issues de ces méthodes indépendantes pourtant très simples présentent un accord remarquable : la moyenne des valeurs obtenues pour la vitesse verticale à 90 m est de $1,9 \times 10^{-3}$ cm s^{-1}, et l'écart quadratique moyen associé de $0,3 \times 10^{-3}$ cm s^{-1} seulement. Les résultats décrits dans cet article montrent qu'il est facile d'obtenir une bonne estimation de la vitesse verticale à l'équateur ; cela peut être utile pour étudier les effets de l'upwelling sur la production et le renouvellement des sels nutritifs dans la couche de surface équatoriale.

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INTRODUCTION

Unlike the situation at mid-latitudes, vertical velocity on the Equator is easy to estimate for low-frequency large-scale motions. This is because equatorial upwelling, manifested in a cold tongue of sea surface temperature extending along the Equator westward from the South American coast, is a first order consequence of equatorial dynamics and a direct result of divergent Ekman flow, caused by the prevailing easterly winds and compensated within the shallow thermocline about equally by both geostrophic meridional inflow due to the eastward pressure gradient and zonal convergence due to the decelerating upward and eastward flowing equatorial undercurrent (Bryden, Brady, 1985). Four ways of obtaining reliable estimates of vertical velocity, useful for application to large scale, low-frequency problems, are demonstrated here.

It is necessary to render a distinction between the vertical velocity, which is the velocity descriptive of the actual vertical motion a parcel undergoes, and the diapycnal or cross-isotherm velocity, which is the component of velocity normal to isopycnals or isotherms, because each is relevant to separate problems (Bennett, 1986). Upward vertical velocity represents a transport of nutrients from the deeper ocean into the euphotic zone, resulting in enhanced biological productivity along the Equator. Cross-isotherm velocity, even in the absence of vertical motion, is the central quantity in the study of water mass conversion, thus important in the global heat budget. Both are estimated and discussed in fuller detail here, though in practice the cross-isotherm velocity is more difficult to obtain than an estimate of the vertical velocity.

METHODS TO ESTIMATE VERTICAL VELOCITY

The following four methods may be easily applied at any given location along the Equator to obtain a reasonable estimate of vertical velocity associated with low-frequency large scale motions.

Undercurrent core trajectory as streamline

This method is based on the crudest assumption allowable : that the trajectory of the undercurrent velocity " core " is an approximate streamline of flow. The picture which obtains is that of a coherent jet flowing upward from west to east within the ambient fluid. Implicitly assumed is the idea that the rate at which fluid flows upward and eastward away from the core of the jet is much less than the rate at which fluid flows upward and eastward along the core of the jet. Hence the vertical velocity $w_{(1)}$, at the depth of the core, equal to the vertical component of velocity along the trajectory, is approximated as $w_{(1)} \simeq u_{core} \tan \varphi$, where u_{core} is the zonal component of velocity at the depth of the core and $\tan \varphi$ is the slope of the core trajectory.

Flow along isotherms

Assuming that the undercurrent flows parallel to isotherms, the vertical motion at depths of the undercurrent results mostly from fluid flowing along the thermocline which slopes upwards to the east and less from fluid being warmed from above and effectively crossing isotherms. Vertical velocity can be then estimated at particular isotherms under quasi-steady conditions as $w_{(2)} \simeq u(\partial z/\partial x)_\theta$, where $\partial z/\partial x$ is the slope of the isotherm at constant potential temperature θ. For slowly varying time-dependent motions, a better estimate for vertical velocity may result if an estimate for the local vertical displacement motion of the isotherm is added to the estimate above, though for annual and longer timescales, it will be shown that this addition is negligible.

Simple mass conservation

With only mean wind stress estimates at the surface and integrated pressure gradients (from the dynamic height field determined from mean CTD data), there are two ways to make an estimate of vertical velocity at the Equator with simple ideas of two dimensional mass conservation in the vertical-meridional plane. This method, most successfully exploited by Wyrtki (1981), is based on the assumption that the upwelling at the base of the surface layer is due to the combined effects of the divergent meridional Ekman flow, a result of the easterly wind stress, and the convergent geostrophic flow due to the eastward pressure gradient set up by the wind stress. The first estimate $w_{(3\,a)}$, is determined assuming that the equatorial wind stress is exactly balanced by the vertically integrated pressure gradient as predicted by steady linear theory. It is assumed that the equatorial wind stress and the pressure gradient set up by the wind stress are pervasive throughout the entire equatorial upwelling zone. Thus, off the Equator at the edge of the upwelling zone the integrated meridional velocity field, Ekman and geostrophic, must be zero. The Ekman transport over a suitable depth, easily calculated from the equatorial wind stress value at the Equator, must be returned by the geostrophic flow distributed throughout a column of greater depth, also suitably chosen. Vertical velocity at the base of the surface layer throughout the upwelling zone is determined from the net divergence in the upper surface layer. A second estimate, $w_{(3\,b)}$, is made using the wind stress estimates off the Equator, but calculating the geostrophic flow directly from the measured integrated pressure gradients to balance mass. The uncompensated Ekman transport is placed as a zonal convergence in the surface layer.

Three-dimensional mass conserving model

Here, the three dimensionality of the flow field is taken into account. Upwelling at a particular depth is a result not only of meridional geostrophic convergence and Ekman divergence but also of zonal divergence and convergence associated with the undercurrent sloping up and decelerating to the East — effects which are quite important to the overall mass balance in the equatorial zone. The procedure for determining a steady state vertical velocity profile is as follows. The relative horizontal flow normal to the sides of a large "box" is determined both geostrophically and by estimating the meridional Ekman transport. Absolute velocity is obtained by adjusting the horizontal velocity field at the chosen reference level so that mass is conserved, subject to the condition that vertical velocity is zero at both the sea surface and bottom of the box. The continuity equation is integrated meridionally from the extreme northern (southern) edge of the box down to the northern (southern) edge of a defined upwelling zone along the Equator, assuming the flow is horizontally non-divergent in the extra-equatorial region, in order to obtain vertical profiles of meridional velocity at the northern and southern edges of the upwelling zone. A vertical profile of vertical velocity for the equatorial zone is then obtained after vertically integrating the three-dimensional continuity equation from the surface to the bottom of the box.

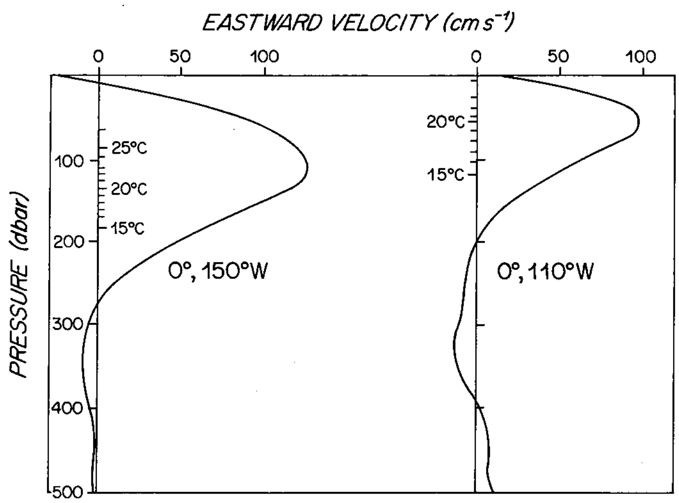

Figure 1

Vertical profile of eastward velocity on the Equator at 150°W and 110°W. Also shown is the depth at which selected mean isotherms are located within the high velocity core of the undercurrent.

DATA/PROCEDURE/RESULTS

In order to calculate vertical velocity employing the various methods described above, mean meridional CTD sections from the Norpax Hawaii to Tahiti Shuttle Experiment along 150°W and the EPOCS cruises along 110°W are used. The temperature and salinity data extend from 5°N to 5°S at 1° latitude increments and from the surface to 500 dbar at 5 dbar increments. The mean zonal wind stress estimates needed for methods (3) and (4) are provided by Wyrtki and Meyers (1976). The horizontal velocity field is taken from the three-dimensional diagnostic model of Bryden and Brady (1985).

Using methods (1) through (4), described in the previous section, and the data described above, the following estimates for vertical velocity result :

Method (1)

The zonal velocity profiles, Figure 1, show an undercurrent maximum at 110°W of 98.6 cm s^{-1} at 60 dbar, and at 150°W of 127.3 cm s^{-1} at 120 dbar. From these profiles an average undercurrent core velocity of 113.0 cm s^{-1} and a core slope of 1.35×10^{-5} are computed yielding an estimate for vertical velocity, $w_{(1)}$ at a single depth of 90 dbar of 1.52×10^{-3} cm s^{-1}.

Method (2)

From the equatorial profiles of temperature at both longitudes, the zonal slope of isotherms, typically 1.8×10^{-5} is calculated (Tab. 1, column 4). Multiplying the isotherm slope by the zonally averaged zonal velocity (Tab. 1, column 3), gives an estimate of vertical velocity, $w_{(2)}$ (Tab. 1, column 5). A typical value at a depth corresponding to the undercurrent core is 1.9×10^{-3} cm s^{-1} at 93 dbar. Note that this value is much larger than the vertical motion due to the local vertical displacement of an isotherm on an annual timescale estimated to be about $0.4 \pm 0.2 \times 10^{-3}$ cm s^{-1}, given that a typical equatorial thermocline displacement amplitude in the central Pacific is about 20 ± 10 m (Meyers, 1979). The advantage of this method over method (1) is that a vertical profile of vertical velocity is obtained, with a maximum of 2.0×10^{-3} cm s^{-1} at 86 dbar.

Method (3)

Using an equatorial wind stress value of −.478 dynes cm^{-2} gives a total depth integrated meridional Ekman flow, combined over both north and south edges of a 4° latitude width upwelling zone spanning the Equator, of about 18.7×10^{4} cm^2 s^{-1}.

Table 1

Vertical velocity estimates.

θ(°C)	depth (dbar)	U(cm s^{-1})	$(\partial z/\partial x)_\theta$ (10^{-5})	$w_{(2)}$	$w_{(4)}$ (10^{-3} cm s^{-1})	w_c
24	51.4	74.2	2.06	1.53	2.48	.95
23	66.0	98.9	1.88	1.86	2.86	1.00
22	76.7	107.5	1.81	1.95	2.62	.67
21	85.8	111.0	1.78	1.98	2.42	.44
20	93.4	109.9	1.77	1.94	2.24	.30
19	102.2	105.6	1.80	1.90	2.01	.11
18	112.8	98.6	1.67	1.65	1.71	.06
17	123.5	86.7	1.58	1.37	1.39	.02
16	134.1	74.6	1.52	1.13	1.08	−.05

For $w_{(3\,a)}$, this flow is assumed to be exactly compensated by a 200 m deep meridional flow because the pressure gradient vanishes beyond that depth (Bryden, Brady, 1985). The difference in Ekman outflow and meridional inflow yields a net meridional flow out of the upper 50 m deep surface layer of $14.0 \times 10^4\ cm^2\ s^{-1}$ which results in a vertical velocity at 50 m, $w_{(3\,a)}$, of $3.16 \times 10^{-3}\ cm\ s^{-1}$. Using the actual wind stress values at 2°N and 2°S of $-.437$ and $-.555$ dynes cm^{-2} respectively instead for $w_{(3\,b)}$, a total Ekman outflow in the upper 50 m of $19.4 \times 10^4\ cm^2\ s^{-1}$ is obtained. For $w_{(3\,b)}$, instead of calculating the meridional geostrophic inflow as necessary to balance the Ekman outflow, the actual vertically integrated pressure gradients of $-.325$ and $-.425$ dynes cm^{-2} at 2°N and 2°S respectively, are used to calculate a total meridional geostrophic inflow of $14.7 \times 10^4\ cm^2\ s^{-1}$. Thus, there is a net imbalance of $4.7 \times 10^4\ cm^2\ s^{-1}$ between the Ekman outflow and the geostrophic inflow. If this imbalance is placed in the upper 50 m as a zonal convergence, a minimum estimate for vertical velocity, $w_{(3\,b)}$, of $2.48 \times 10^{-3}\ cm\ s^{-1}$ is obtained.

Method (4)

This method using the same data is described by Bryden and Brady (1984). The results are cited here for comparison with the other methods. The upwelling profile which obtains is shown in Figure 2. A maximum value of $2.9 \times 10^{-3}\ cm\ s^{-1}$ is attained at a depth of 62 dbar. Below 180 dbar, downwelling achieving a maximum at 322 dbar of $-0.9 \times 10^{-3}\ cm\ s^{-1}$ is observed. The gross features of this profile are similarly observed in the profile obtained with current meter measurements by Halpern and Freitag (1987). Values of vertical velocity using this method, $w_{(4)}$ at

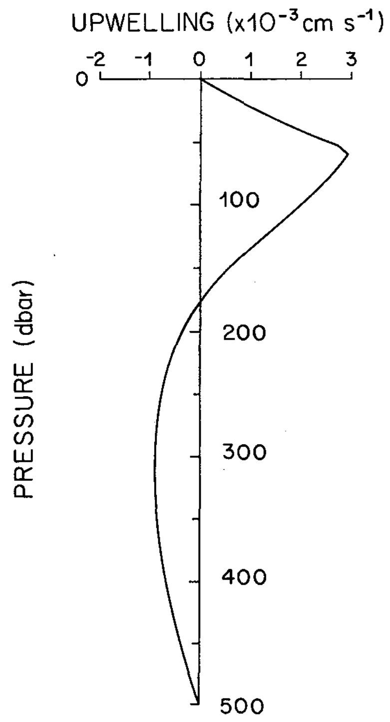

Figure 2

Vertical profile of vertical velocity for the equatorial region, 150°W to 110°W, 0.75°N to 0.75°S (from Bryden, Brady, 1985).

depths corresponding to particular isotherms are listed in Table 1 (column 6), for comparison with $w_{(2)}$. The difference between the two values is equivalent to an effective cross-isotherm velocity, w_c (column 7).

DISCUSSION

It is apparent from Table 2, a tabulation of the vertical velocity estimates from all methods, that there is little scatter in the estimates made at a particular depth, chosen as either 50 or 90 dbar. Treating the results from each method as a separate estimate, a mean and standard deviation is obtained

Table 2

A comparison of the vertical velocity estimates.

	$w_{(1)}$	$w_{(2)}$	$w_{(3\,a)}$	$w_{(3\,b)}$	$w_{(4)}$
		($\times 10^{-3}$ cm s^{-1})			
50 dbar :		1.5	3.2	2.5	2.5
90 dbar :	1.5	1.9			2.2

for the results as a group at each depth. For the group of 4 estimates made at 50 dbar depth, the mean obtained is $2.4 \times 10^{-3}\ cm\ s^{-1}$ with a standard deviation of $0.6 \times 10^{-3}\ cm\ s^{-1}$. At 90 dbar depth for the group of 3 estimates, a mean of $1.9 \times 10^{-3}\ cm\ s^{-1}$ and a standard deviation of $0.3 \times 10^{-3}\ cm\ s^{-1}$ is obtained. Quite clearly, better than " order of magnitude " estimates are attainable by any of the 4 methods cited.

Taking the most sophisticated estimate $w_{(4)}$, determined from three-dimensional mass conservation, as a standard by which to compare the other estimates not only reveals certain dynamical and thermodynamical implications, but also tests the assumptions made in methods (1) and (2). For example, using the undercurrent core trajectory as a streamline produces an underestimated result $w_{(1)}$, compared not only to $w_{(4)}$ but to $w_{(2)}$ as well. This underestimation implies that parcels of water in and above the undercurrent velocity core flow up and away from the core while replaced by water from just below the core (where the flow is still upward as evident in Figure 2 to a depth of 180 dbars) and from the meridional and zonal convergence at core depth. Since the estimates determined from both methods differ in a relative sense by at most 30 %, it appears that the actual parcel trajectories only slightly diverge away from the path of the undercurrent core, validating the crucial assumption in method (1).

Comparing the vertical velocity $w_{(2)}$, associated with the vertical motion of water parcels flowing along isotherms, to $w_{(4)}$ is revealing of how much vertical mixing of heat exists on the Equator. The difference in the two velocities, approximately w_c, equal to the volume flux per unit area across isotherms, is listed in column 7 of Table 2. This cross-isotherm velocity is upward in water warmer than 17°C and everywhere much smaller than $w_{(4)}$. In the region corresponding to the velocity core of the undercurrent associated with water warmer than 17°C but colder than 22°C,

w_c is less than 40 % of the magnitude of $w_{(4)}$. Hence the assumption of method (2), that water parcels moving with the undercurrent core flow mostly upward along isotherms, is also validated.

An upward cross-isotherm flux represents a water mass conversion and as such must be balanced by a net heating accomplished through the vertical mixing of heat downwards. Inspecting the size of w_c compared to $w_{(4)}$, it appears that significant vertical mixing probably occurs only in the warmest water classes but is present though negligible down to the velocity maximum of the Untercurrent. Assuming that this thermal mixing or heating vanishes at some level z_0, the depth where the cross-isotherm flux vanishes, the term $w_c\ \partial\theta/\partial z$, representing vertical mixing of heat in the equation for the conservation of potential temperature, can be integrated from z_0 to the surface to obtain an estimate for the net surface heat flux required to balance the flux across isotherms. Taking z_0 as 93 dbar, since below that depth, the depth of the 20°C isotherm and just deeper than the velocity maximum, the estimate for w_c is indistinguishable from zero within expected errors, a net surface heat flux of about $160\ W\ m^{-2}$ results. This estimate exceeds the best estimates of net air-sea heat flux by about a factor of two but can be reduced by that much if $w_{(4)}$ is merely over-estimated by 15 % — certainly within the errors acceptable in the method employed.

In summary, these findings demonstrate that vertical velocity under quasi-steady or geostrophic conditions, can be easily estimated with confidence. The methods outlined here are well-suited for use in studies where an independently obtained upwelling rate is necessary to determine vertical fluxes of nutrients and chemicals into the surface layer eliminating the need to employ over-simplified, one-dimensional advection-diffusion models which use *ad hoc* diffusion parameters. For such purposes where a vertical transport must be estimated, the simplest method giving the best results is method (2), especially at depths near the undercurrent core where mixing is negligible. All that is required is a mean undercurrent velocity, obtained from either suitably averaged direct measurements or geostrophically from the dynamic height field, and sufficient knowledge of the zonal and vertical thermal gradients.

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REFERENCES

Bennett S. L., 1986. The relationship between vertical, diapycnal and isopycnal velocity and mixing in the ocean general circulation, *J. Phys. Oceanogr.*, **16**, 167-174.

Bryden H. L., Brady E. C., 1985. Diagnostic model of the three-dimensional circulation in the upper equatorial Pacific Ocean, *J. Phys. Oceanogr.*, **15**, 1255-1273.

Halpern D., Freitag, H. P., 1987. Vertical motion in the upper ocean of the equatorial Eastern Pacific, *Proc. Inter. Symp. on equatorial vertical motion, Oceanol. Acta*, this vol., 19-26.

Meyers G., 1979. Annual variation in the slope of the 14°C isotherm along the equator in the Pacific Ocean, *J. Phys. Oceanogr.*, **9**, 885-891.

Wyrtki K., 1981. An estimate of equatorial upwelling in the Pacific, *J. Phys. Oceanogr.*, **11**, 1205-1214.

Wyrtki K., Meyers G., 1976. The trade wind field over the Pacific Ocean, *J. Appl. Meteorol.*, **15**, 689-704.