NOTES AND CORRESPONDENCE

A Relationship between Interannual Variations in the South Pacific Wind Stress Curl, the Indonesian Throughflow, and the West Pacific Warm Water Pool

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ABSTRACT

Simple theory gives that the depth-integrated flow between the Pacific and Indian Oceans, on interannual timescales and longer, is driven by the integral of the wind stress along a line from the northern tip of Papua–New Guinea across the equatorial Pacific, south along the South American west coast, westward across the South Pacific at the latitude of the southern tip of Australia, and northward along the west coast of Australia. Evaluation of this integral using ECMWF/JMA 1000-mb wind data for the decade 1980–89 yields an interdecadal and interannual signal. The interannual signal peaks in 1981 and 1985, then decreases sharply through the ensuing ENSO events. The variations are chiefly attributable to variations in the integral of wind stress across the midlatitude line in the South Pacific. The presence of the shallow sills within the Indonesian seas will partially block the midlatitude contribution, but local baroclinic adjustment over the sills will reduce the blocking effect and produce a corresponding interannual variation in upper-layer transport through the seas. Two mechanisms by which variations in throughflow magnitude could contribute to warm water pileup in the west Pacific are proposed.

1. Introduction

Interest in the magnitude and variability of the flow between the Pacific and Indian Oceans through the Indonesian archipelago has arisen because of its possible effect on the evolution of the warm water pool in the west Pacific. Observations in the region are limited, and global numerical general circulation models, for example, Semtner and Chervin (1988), give annual mean values of 15–20 Sv (Sv = 1 × 10⁶ m³ s⁻¹), in agreement with Godfrey’s (1989) Island Rule estimate of 16 ± 4 Sv using Hellerman and Rosenstein’s (1983) climatological annual mean wind stresses. Wajswowicz (1993) rederived the Island Rule taking into account topographic effects and frictional effects along the oceans’ eastern boundaries, and argued that it could be used to describe variations in the magnitude of the depth-integrated Indonesian Throughflow on interannual timescales and longer. Knowledge of variations in the depth-integrated transport is not particularly useful when considering the heat budget of the warm water pool. However, the very shallow sills within the Indonesian seas will induce a significant first baroclinic mode adjustment in response to variations in the depth-integrated transport, and so a corresponding variation in upper-layer transport on interannual timescales.

The depth-integrated throughflow magnitude is set by the integral of the wind stress along a path around the South Pacific, and modulated by the dynamical processes within the Indonesian seas or frictional effects along the oceans’ eastern boundaries. Little is known at this stage about the size of the modulation; Wajswowicz (1993) gives some estimates, but the wind stress component can be calculated from available wind stress data. In section 2, the relevant extension of Godfrey’s Island Rule, derived in Wajswowicz (1993), is recapped and discussed. Results from evaluating the theoretical throughflow magnitude, that is, the line integral of the wind stress, using European Centre for Medium-Range Weather Forecasts/Japanese Meteorological Agency (ECWMF/JMA) 1000-mb winds for 1980–89, and Comprehensive Ocean–Atmosphere Data Set (COADS) psuedo–wind stresses for 1970–79, are presented in section 3. A discussion of how variations in the theoretical magnitude could contribute to warm water pileup in the west Pacific is given in section 4.

2. The Island Rule for interannual variations

Before presenting the results, it is worth recapitulating the derivation of the Island Rule in terms of the barotropic streamfunction ψ given by

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\[ \psi_x = \int_{-H}^{0} vdz, \quad \psi_y = -\int_{-H}^{0} udz, \]

where \( z = -H(x, y) \) is the ocean floor. The existence of \( \psi \) may be deduced by vertically integrating the continuity equation and making the rigid-lid approximation \((w = 0 \text{ at the surface})\). The barotropic vorticity equation, assuming linear dynamics, is

\[
\nabla_b \cdot \left( \frac{1}{H} \nabla \psi \right) = \int \frac{f}{H} \cdot \psi \\
= J[\mathcal{P}, H] + \text{curl} \frac{\tau}{\rho_0 H} + \text{curl} \frac{F}{H}, \quad (2.1)
\]

where

\[ J[a, b] = a_x b_y - a_y b_x, \quad \text{and} \]

\[ \mathcal{P} = \frac{g}{\rho_0 H^2} \int_{-H}^{0} \{ H_p - \int_{z}^{0} \rho dz \} \cdot dz, \]

and where \( \nabla_b \) is the horizontal gradient operator. The explicit time derivative in (2.1) can be neglected in comparison with the planetary vorticity term on timescales greater than the barotropic topographic adjustment scale, which is \( O(\text{months}) \). Frictional effects are represented by the term \( F \).

If the ocean depth is assumed uniform over the Pacific, that is, \( H = H_0, \text{const} \), then integrating (2.1) over the area \( AA'BB'B' \), bounded by the line \( \partial D \), see Fig. 1, and applying Stokes theorem, gives the streamfunction on Australia–Papua–New Guinea (PNG), \( \psi_A \), namely

\[
-(f_N - f_S) \psi_A = H_0 \int_{\partial D} \frac{\tau}{\rho_0 H} \cdot dl + H_0 \int_{\partial D} F \cdot dl, \quad (2.2)
\]

where \( f_N, f_S \) are the values of the Coriolis parameter at the northern tip of Irian-Jaya and southern tip of Australia, respectively, and \( dl \) is the line element tangential to the curve \( \partial D \). For brevity in this derivation, New Zealand has been neglected. Its presence can be readily incorporated by using the Multiple Island Rule derived in Wajszcz (1993).

Integrating the horizontal momentum equations along Australia–PNG’s boundary, \( \partial l \), yields the torque equation

\[
-\int_{\partial l} \frac{1}{H} \nabla_h \times (\psi_k) \cdot dl \\
= \int_{\partial l} \left( \frac{\tau}{\rho_0 H} + \mathcal{P} \nabla_h H + \frac{F}{H} \right) \cdot dl. \quad (2.3)
\]

The pressure is assumed single valued, and the Coriolis contribution has been eliminated as the no normal flow boundary condition implies \( \nabla_h \psi \cdot dl = 0 \); variable topography has been retained. The explicit time derivative in (2.3) can be neglected on timescales greater than that needed for friction in the western boundary layer to dissipate the angular momentum imparted on the island by the wind stress or bottom pressure torques, which is typically of \( O(10 \text{ days}) \).

Suppose friction and topographic variations are negligible along the Australia–PNG western boundary except through the Indonesian seas, which will be denoted by a segment \( \mathcal{E} \) of \( \partial D \). If friction is assumed negligible along the segment \( BB'AA' \) of \( \partial D \), then the friction integral along the Australia–PNG east coast can be eliminated between (2.2) and (2.3) to give

\[
-(f_N - f_S) \psi_A = H_0 \int_{\partial D} \frac{\tau}{\rho_0 H} \cdot dl \\
+ H_0 \int_{\partial D} \left( \mathcal{P} \nabla_h H + \frac{F}{H} \right) \cdot dl, \quad (2.4)
\]

where \( \mathcal{E} = \partial D + \partial l \). In Godfrey’s (1989) original Sverdrup model derivation, a simple western boundary

![Fig. 1](https://example.com/figure1.png)

**Fig. 1.** The geometry used in deriving Godfrey’s Island Rule. The path \( \partial D: AA'BB' \) follows the arrows (→). The path \( \partial l \) is around Australia–PNG following the arrows (⇒). The path \( \mathcal{E} \) is \( \partial D \) plus \( \partial l \), that is, \( AA'BB' \), but along the west coast of Australia–PNG.
layer closure scheme was adopted in which vorticity was dissipated at the latitude of creation in the western boundary layer. This choice was necessary for constructing the streamfunction field for the global oceans. In deriving the Island Rule, we just require that the frictional integral along the path $B'BAA'$ is negligible compared with that along $A'B'$.

The expression (2.4) is valid on seasonal timescales and longer, but only in a diagnostic sense, as $P$ contains baroclinic variations due to varying wind stress, as well as those due to explicit buoyancy forcing. A more useful diagnostic is given by expressing the integral along $\delta'$ as a difference in depth-integrated pressure. Applying the operator $(n \cdot \partial_t + 1 \cdot f)$ to the depth-integrated momentum equations yields

$$\nabla_h P_t \cdot n + f \nabla_h P \cdot l = (n \cdot \partial_t + 1 \cdot f) \left( P \nabla_h H + \frac{\tau}{\rho_0 H} + \frac{F}{H} \right),$$

where $P = (\rho_0 H)^{-1} \int_0^H \rho d\zeta$, and $n$ is the unit normal to the boundary such that $l = k \times n$. This equation expresses the propagation of coastal Kelvin waves along boundaries. Therefore,

$$\int_S \left( P \nabla_h H + \frac{F}{H} \right) \cdot dl$$

$$\approx \int_S \left( \nabla_h P - \frac{\tau}{\rho_0 H} \right) \cdot dl = \Delta P_{\text{EXCESS}}$$

on timescales for which it takes the depth-integrated pressure to equilibrate around Australia–PNG. As Australia–PNG intersects the equatorial waveguide, to first order this is the time taken for a first baroclinic mode coastal Kelvin wave to propagate around Australia–PNG plus the time for an equatorial Kelvin wave to propagate across the Pacific and the reflected long Rossby wave to return, which is $O(1 \text{ year})$. Therefore, on interannual timescales and longer,

$$\psi_A \approx -\frac{H_0}{(f_N - f_S)} \int_{y_0} \frac{\tau}{\rho_0 H} \cdot dl + \frac{H_0 \Delta P_{\text{EXCESS}}}{(f_N - f_S)}.$$

If the oceans were homogeneous and the sills high, for example, blocking three-quarters of the fluid depth, then the second term in (2.5) would be of similar magnitude to the first, and $\psi_A$ would be much reduced from the value in the absence of sills. The flow would not be wholly blocked, as a fraction would cross on and off the sill through frictional boundary layers on the Asian coast on the northward slope and the Australian coast on the southern slope. Therefore, in the presence of sills, frictional effects along $\delta'$ cannot be neglected. They are necessary for a quasi-steady state to be achieved. If the oceans are stratified and a steady wind stress is switched on, then the full depth-integrated throughflow in the absence of sills is achieved after baroclinic planetary waves and coastal Kelvin waves generated by the wind stress have propagated around the basin and set up a level of no motion above the sill top. The problem is similar to that of the effect of the mid-Atlantic ridge on the magnitude of the Gulf Stream described by Anderson et al. (1979). The "sheltering" of the upper-layer flow from the underlying topography is actually a more complex process in a continuously stratified fluid. As the barotropic flow crosses the ridge, it will generate upwelling and downwelling, which produce horizontal temperature gradients yielding geostrophic flow directed to cancel the barotropic flow. These signals will in turn generate very slow bottom-trapped waves, which if the wind-generated planetary waves had not switched off the flow at depth on a much faster timescale, would eventually switch off the flow below the ridge top. Figure 16 of Anderson et al. shows temperature perturbations over the ridge as well as those associated with the long planetary waves directly generated by the wind stress curl. The "sheltering" process was discussed in a slightly different context by Wajszowicz (1991), who described in detail the processes occurring due to the interaction of a coastal Kelvin wave with a sill. Similar processes will occur in the Indonesian seas, but on a much faster timescale, as the sills are high and situated near the equator, so the up- and downwelling generated will project largely onto the fast, first baroclinic mode. Also, the sills are narrow and connect small seas, so mixing processes will smooth out gradients rapidly. Therefore, although long, first baroclinic mode planetary waves take $O(15 \text{ years})$ to cross the South Pacific at midlatitudes, the interannual variations in the barotropic flow through the Indonesian seas driven by the midlatitude wind stress curl will only be partially blocked. A further important point is that if the oceans were flat bottomed, the upper-layer transport through the Indonesian seas would not respond on interannual timescales to changes in the South Pacific midlatitudes, and the expression (2.5) for the depth-integrated transport would not be particularly useful when considering the heat budget of the west Pacific warm water pool. However, the presence of shallow sills in the archipelago implies that there has to be a significant first baroclinic mode adjustment in response to a varying depth-integrated transport. Detailed modeling studies are needed to determine the magnitude of the resulting upper-layer adjustment and whether it is chiefly confined to just the Indonesian seas, or whether there is a significant amount of energy scattered into equatorial and coastal Kelvin waves. Studies using an idealized geometry and simplified forcing, described in Wajszowicz (1994, unpublished manuscript) confirm that through topographic interaction, changes in the wind stress at midlatitudes in the South Pacific can affect the upper-layer transport through the seas on interannual timescales. Further confirmation is provided by results from a GCM forced by ECMWF wind stress anomalies reported in Wajszowicz (1994, unpublished manuscript).

No information is available at present to estimate the variability of the second term in (2.5). The shal-
lowness of the sills in the Indonesian seas indicate it could be important. It would be through the $\mathcal{P}V_s H$ integral that interannual variations in the wind stress at latitudes other than that of the northern tip of PNG and the southern tip of Australia would produce interannual variations in the depth-integrated throughflow. From Godfrey's (1989) analysis of depth-integrated steric height, using the Levitus (1982) climatological hydrographic data, the mean value of $\Delta P_{EXCESS}$ may be as small as 2 Sv or as large as 10 Sv; better data is needed to narrow the range. In the following section, the first term is investigated using available wind stress data.

3. Evaluation of the wind stress integral

Let us define the component of the depth-integrated throughflow driven by the South Pacific wind stress as

$$\psi_{IT} = \frac{1}{(f_N - f_s)} \int_{\rho_0} \tau \cdot d\tau > 0.$$  \hspace{1cm} (3.1)

The difference arising from taking $H = H_0$ in the archipelago in the wind stress integral is negligible for interannual timescales. Quality wind stress data for the midlatitude South Pacific are very sparse. As a first effort compromise, ECMWF 1000-mb winds for 1980–86 and JMA 1000-mb winds for 1987–89 have been used to calculate the wind stress integral. A monthly average of the daily winds is taken, and the wind stress is calculated using the simple formula

$$\tau = \rho_0 C_d |u| u,$$

$$C_d = \begin{cases} 0.8 \times 10^{-3}, & |u| < 6.7 \text{ m s}^{-1} \\ 2.6 \times 10^{-3}, & \text{otherwise}, \end{cases}$$  \hspace{1cm} (3.2)

where $\rho_0 = 1.2 \times 10^{-3}$ g cm$^{-3}$. A slightly more tentative calculation is made for the 1970–79 period using COADS psuedo–wind stresses with $C_d = 1.5 \times 10^{-3}$.


The path chosen for calculating (3.1) is based on the $1^\circ \times 1^\circ$ coastal outline from a typical Levitus (1982) dataset. It was found easier to calculate (3.1) as an equivalent area integral than a line integral along a jagged coastline line. The presence of New Zealand is taken into account through the Multiple Island Rule derived in Wajszowicz (1993). The ECMWF/JMA wind data has a resolution of 5° zonally and 2.5° meridionally. It was linearly interpolated onto the $1^\circ \times 1^\circ$ geometry grid. The calculation for the 1980–89 period shows an interdecadal trend as well as an interannual one (see Fig. 2). The magnitude of the predicted throughflow decreases over the decade. This is possibly enhanced by the switch to JMA winds in 1987.
The annual means derived from ECMWF winds, and using (3.2), are in the range 16 ± 4 Sv determined by Godfrey (1989) from climatological wind stress data. Those derived from JMA winds are about 5 Sv weaker. Superimposed on the decrease is an interannual variation, which shows a sharp decrease in magnitude around the onset of the 1982–83 and 1986–87 ENSO events with a gradual increase in between.

Further calculations show that the wind stress integral is quite insensitive to whether New Zealand is taken into account, or whether the integral is along actual coastlines or the 200-m isodepth, say; cf. Fig. 2 and 3a. For the remainder of the calculations, New Zealand is ignored and the 200-m isodepth chosen. The contribution to the wind stress integral from different parts of the path C was calculated and is shown in Fig. 3a. The zonal paths are readily calculated, and those along the coastlines are calculated as the difference between the equivalent area integral and the integrals along lines of latitude. The mean magnitude of the depth-integrated throughflow is set by the contribution from the South Pacific, that is, along B'B in Fig. 1. It typically accounts for about 70%. The variability about the decadal mean is shown in Fig. 3b. Surprisingly, it is the contribution from the South Pacific that accounts for both interdecadal and interannual variability. The contribution from the equatorial wind stress modulates the phase slightly. Of course, there is the question of how much of the contribution from midlatitudes will translate into upper-layer variations. If topography blocks 50% of the midlatitude variability, then the surfaced-trapped equatorial contribution will dominate. However, the result is still interesting. It may have been supposed that the throughflow would be weaker during an ENSO event, because of the weakening of the easterly trade winds. The contribution from the weakening of the westerlies in the southern midlatitudes emphasizes the Southern Oscillation aspect of the El Niño phenomenon. Meteorologists have known for many years that changes in sea level pressure in the South Pacific midlatitudes precede an El Niño event by almost a year; see, for example, Trenberth (1976) and van Loon (1984). An atmospheric link via wave propagation, for example, has not been forthcoming. The above arguments provide a quite simple and plausible oceanic link, which is explored further in the next section.

The only information from modeling studies available to date is that of Kindle et al. (1989). Using a global reduced gravity model forced by ECMWF winds, reduced by 20% and $C_v = 1.5 \times 10^{-3}$, they obtained a mean upper-layer transport of only 4.5 Sv. The interannual variability is different from the predicted depth-integrated transport in Fig. 2. Their model had peak transports in 1982 and 1984/85, and minimum in 1980/81 and 1983/84. Also, they did not find an interdecadal signal, but an increase from 1980 to 1984 followed by a decrease. No variation associated with the 1986–87 ENSO event was found. The difference between the Kindle et al. results and Fig. 2 can be explained by the fact that only the first baroclinic mode timescale is present in the Kindle et al. model. As stated earlier, interannual variations in the wind stress over the South Pacific midlatitudes cannot produce corresponding variations in the upper-layer transport through the Indonesian seas in such a model with an O(1 year) phase lag. In a model with both baroclinic and barotropic modes, the depth-integrated throughflow will respond to variations in the wind stress at midlatitudes rapidly. Interaction with the shallow sills in the Indonesian archipelago will produce variations in upper-layer transport on interannual timescales.

Wyrtki (1987) found little interannual variability in the sea level difference between Davao, Philippines, and Darwin, Australia, but as noted by Wajisowicz (1993), sea level difference between any location at the northern and southern entrances of the Indonesian archipelago is unlikely to be a good measure of throughflow magnitude or variability because of the small proportionality constant, which is very sensitive to the choice of latitude of the sea level measurements. The above also suggests that Clarke (1991) and Du Penhoat and Cane (1991) may not have taken into account the full interannual variability of the throughflow in their assumption that it was just that due to the transmission of low-frequency equatorial planetary waves through the gappy western boundary of the Pacific.

b. Variability during 1970–79

The COADS dataset, see Woodruff et al. (1987), is typically too sparse over the relevant latitudes in the Pacific to calculate (3.1). An attempt has been made for the decade 1970–79, as data are more plentiful. Even still, the data had to be averaged over four 2° × 2° boxes spanning 44°S to get a good monthly coverage, and only data for 86°E of longitude east of Tasmania were considered. As in 1980–89, there appears to be an interdecadal as well as interannual variation. The decrease over the decade is due to a decrease from the southern midlatitude contribution. Once again, the interannual variation is such that the implied throughflow decreases around the onset of an ENSO event, but now the variability is determined by the equatorial contribution. The midlatitude contribution is out-of-phase with the equatorial contribution for the 1972–73 event, and shows no clear interannual variation for the 1976–1977 event. The decadal mean of the 44°S signal shown in Fig. 4 is only about 3 Sv, whereas that of the climatological mean contribution from 44°S is over 11 Sv, which indicates that the signal may not be representative of the total contribution from B'B.

4. Implications for the west Pacific warm water pool

Wyrtki (1989) defines the warm water pool in the west Pacific Ocean as the area enclosed by the 28°
FIG. 3. The wind stress integral (12-month running mean) for the path $0^\circ: AA'B'B'A$, denoted by solid line, is plotted against time for the decade (1980–1989) in (a) from ECMWF/JMA data. The contributions to the wind stress integral (12-month running means) from the paths $AA'$ along the equator (dashed line), $B'B$ along $44^\circ$S (dot-dashed line), and the residual along the west Australian–PNG and South American coasts, i.e., that along $A'B'$ and $BA$ (double dot-dashed line) are also plotted against time in (a). The variability about the decadal mean for each contribution is plotted against time in (b). The 12-month running mean filter has been applied twice to the data in (b). The decadal means are 2.33 Sv, 7.55 Sv, and 0.65 Sv for the equatorial, $44^\circ$S, and residual contributions, respectively, giving a total of 10.53 Sv. These compare with climatological annual mean contributions, calculated from Hellerman and Rosenstein's (1983) wind stress data, of 5.83 Sv, 11.40 Sv, and $-0.56$ Sv, giving an annual mean total of 16.76 Sv.
isotherm. Its extent undergoes large annual variations, but there is a persistent region centered between 10°N and 10°S and around 170°E. The pool is fed by waters from the subtropical gyres, and warm water exits the pool through the North Equatorial Countercurrent and Indonesian Throughflow. The residence time of water in the pool is only about 15 months; therefore the pool will be sensitive to interannual variations in the heat input and heat advection. Variations in the Indonesian Throughflow are particularly significant, as heat is lost from the Pacific system in this flow. Equation (2.5) and the results from ECMWF data suggest two mechanisms by which throughflow variations may contribute to the pileup of warm water in the west Pacific pool.

a. Decrease in the midlatitude South Pacific westerlies

Suppose the depth-integrated throughflow decreases due to a decrease in the westerly winds in the midlatitude South Pacific. There will be a corresponding decrease in upper-layer transport because of the adjustment due to the shallow sills in the Indonesian seas. If the easterly trades driving the main flow in and out of the pool are unchanged or their decrease lags that in midlatitudes, then from Fig. 2, there is O(4 Sv) circulating in the pool, which cannot transport its heat out of the Pacific system into the Indian Ocean.

b. Limitation on throughflow magnitude due to frictional or nonlinear effects

Alternatively, the interannual variation in Fig. 2 could be interpreted as peaking prior to an El Niño event yielding a large theoretical throughflow. However, if the dynamics within the Indonesian seas are such that not more than 8 Sv, or 60% of the theoretical value, say, can pass through the narrow, shallow channels, that is, the second term in (2.5) is important, then once again there will be an excess of warm water circulating in the pool. Wajszczuk (1993) discussed and gave estimates of the reduction in throughflow due to frictional processes, and also showed evidence that the flow could be critical in sections of the straits, and therefore that some form of hydraulic control is possible. It is interesting to note that if the second term in (2.5) does provide a cutoff for throughflow magnitude, then observations and models would not show any evidence of interannual variations in the upper-layer transport.

5. Discussion

The above calculations emphasize the importance of obtaining better wind stress data for the southern midlatitude Pacific Ocean. Data from numerical weather prediction models for the 1980s show that there could be an interesting link between variations in the Indonesian Throughflow and evolution of the
west Pacific warm water pool. Also, the Island Rule provides a connection between midlatitude and equatorial events.

Many centers are now building and developing GCMs of the Pacific and Indian Oceans, so it will be interesting to see what variability they find in the throughflow. If the variability is not similar to that shown in Fig. 2, then the second term in (2.5) must play an important role and needs further investigation.

REFERENCES


