Influence of Low-Frequency Indonesian Throughflow Transport on Temperatures in the Indian Ocean in a Coupled Model*

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ABSTRACT

The relationship between 3- and 10-yr variability in Indian Ocean temperatures and Indonesian throughflow (ITF) volume transport is examined using results from a 300-yr integration of the coupled NCAR Parallel Climate Model (PCM). Correlation and regression analyses are used with physical reasoning to estimate the relative contributions of changes in ITF volume transport and Indian Ocean surface atmospheric forcing in determining low-frequency temperature variations in the Indian Ocean. In the PCM, low-frequency variations in ITF transport are small, 2 Sv (1 Sv = 10⁶ m³ s⁻¹), and have a minimal impact on sea surface temperatures (SSTs). Most of the low-frequency variance in Indian Ocean temperature (rms > 0.5°C) occurs in the upper thermocline (75–100 m). These variations largely reflect concurrent atmospheric forcing; ITF-induced temperature variability at this depth is limited to the outflow region between Java and Australia extending westward along a band between 10° and 15°S.

1. Introduction

The Indian Ocean receives heat and mass from the Pacific at a low latitude via the Indonesian throughflow (ITF; see Godfrey 1996 for a review). A potential consequence is that variations in Indian Ocean temperature may not be only a result of atmospheric forcing over the Indian Ocean, but also may be influenced by changes in the ITF. An important question, and the focus of this study, is to what degree low-frequency changes in upper-ocean temperatures in the Indian Ocean are related to low-frequency changes in ITF transport as opposed to reflecting low-frequency changes in Indian Ocean atmospheric forcing.

One might infer the impact of low-frequency variations in ITF transport from the fate of ITF water in the Indian Ocean. These ITF waters enter the southeastern Indian Ocean between 10° and 15°S, continue across the Indian Ocean as part of the westward South Equatorial Current (SEC) within a limited meridional range between 5° and 20°S, and reach the east coast of Africa after about 10 yr (Song et al. 2004). Observations and numerical models suggest that one-third of the ITF water then exits directly into the Madagascar Current and Agulhas Current while the remainder first recirculates through the northern Indian Ocean (Song et al. 2004). The heat that is carried in this flow is then released in either the southeast Indian Ocean, south of 30°S (Vranes et al. 2002), or in the interior Indian Ocean, or it is exported to the Atlantic via the Antarctic Circumpolar Current (Hughes et al. 1992) as part of the warm water route of the global thermohaline circulation (Gordon 1986).

Consistent with these hypotheses, the magnitude of observed decadal changes in southern Indian Ocean SSTs has been linked to a modulation of the ITF forced by changes in Pacific wind stress (Allan and Haylock 1993; Allan et al. 1995; Reason et al. 1996a,b). However, these studies did not consider the competing impact of forcing these anomalies by local air–sea interaction.

Numerical experiments that contrast ocean simulations with open and closed ITF have been used to explore the ramifications of the existence of the ITF on ocean circulation and on the climate system. These experiments suggest that the ITF affects subsurface temperature throughout the Pacific and Indian Oceans in...
accordance with Sverdrup theory and baroclinic wave propagation. Changes in surface temperatures are found in specific areas, such as the Agulhas Current and the Leeuwin Current, where perturbations of the thermocline affect the surface (Hirst and Godfrey 1993, 1994, hereafter HG93 and HG94). This impact appears to be frequency dependent: low-frequency ITF variations lead to changes in the southern Indian Ocean, while the impact of higher-frequency variations is limited to the western Indian Ocean (Veschell et al. 1995).

Coupled model experiments contrasting open and closed ITF passages suggest that the existence of the ITF affects Pacific and Indian Ocean surface temperature and winds (Schneider 1998; Wajsowicz and Schneider 2001; Wajsowicz 2002). Details of climate sensitivities differ among these models, however a recurring feature is a warmer low-latitude Indian Ocean when the ITF passageways are open.

Here, we seek to understand the impact of a time-varying ITF on the Indian Ocean, and comparing model runs with open and closed ITF passages is of limited value. For example, with continuous land from Sumatra to Australia, a zero throughflow transport would be associated with a pressure difference across this land bridge (e.g., Schneider 1998). In contrast, a reduction of the ITF would be associated with atmospheric and oceanic states that reduce the pressure difference between the Pacific and Indian Oceans (Wyrtki 1987; Potemra and Lukas 1999). Thus, covariations of the ITF and Indian Ocean temperatures could either imply a genuine effect of the ITF, or can reflect atmospheric forcing that affects both.

The present paper seeks to separate these hypotheses. This question requires knowledge of Indian Ocean climate not available from direct observations. Instead, we analyze output from a 300-yr integration of the National Center for Atmospheric Research (NCAR) Parallel Climate Model (PCM) to separate ITF and direct atmospheric influences on Indian Ocean temperatures. Our focus in this study is on time scales longer than ENSO (24- to 30-month periods in the PCM run), since the ENSO variability in the ITF is being studied elsewhere. Due to the length of the coupled model run, we further limit our study to time scales of less than 10 yr, to preserve significant degrees of freedom in our analysis.

Our approach is to combine correlation and regression analysis with physical reasoning to identify the areas of influence of the ITF in the presence of ubiquitous forcing of the Indian Ocean by the atmosphere. In the next section, the model is described. Low-frequency variations in the model Indian Ocean temperature field and ITF transport are then diagnosed. In the final section the relationship between these two is examined.

2. Coupled model

The key model quantities for our study are long time series of ITF transport and its continuation into the Indian Ocean, ocean temperature, and air–sea fluxes in the Indian Ocean. Since this combination is unavailable from observations, this study relies on the dynamical consistency of a coupled model, rather than absolute accuracy of either component, so that relationships between ITF transport and heat content variability may be obtained for a fully coupled environment.

The model used in this study is the PCM version 1 (Washington et al. 2000). The atmospheric component is the NCAR Community Climate Model version 3 (CCM3) on a T42 grid (an approximate spacing of 2.8° × 2.8°) and 18 layers in the vertical. The ocean model, based on the Parallel Ocean Program (POP) has a mean horizontal resolution of 7/ degree near the equator and 32 vertical levels. These components are coupled without flux correction and simulate a stable climate that is similar to observations (Washington et al. 2000). The model has been used to investigate processes affecting El Niño amplitudes (Meehl et al. 2001), twentieth-century climate (e.g., Dai et al. 2001; Meehl et al. 2004), climate change scenarios (Meehl et al. 2005), and the physics of megadroughts (Meehl and Hu 2006). We used 299 yr of control integration of the model with constant greenhouse gas concentrations.

Overall, the Indian Ocean climate simulated by the PCM is realistic. In the following, we discuss the simulation of the strong seasonal cycle of the surface flows and variations of ENSO and then proceed to the presentation of the mean and variations of the ITF and Indian Ocean temperature.

a. PCM climatology

Certain upper-ocean flows in the Indian Ocean change direction due to the influence of the monsoons. This is evident in observations, for example, near the equator where the Southwest Monsoon Current (SMC) flows toward the east in July/August and the Northeast Monsoon Current (NMC) flows toward the west in January/February. The South Equatorial Counter Current (SECC) also appears in observations in January/February as a return flow toward the east between the westward South Equatorial Current (SEC) and the NMC.

The PCM most clearly reproduces the SEC, which appears as a zonal jet between 10° and 15°S (Fig. 1). The model has a seasonally reversing NMC and SMC,
but also has near-equatorial westward flow in June through August and in the annual mean (interannual variations of the latter could be an important contributor to variations in SST for the 3–10-yr time scale). There are also indications of eastward flow in the SECC in December–February in the model, but they are farther south than observations at about 7°S.

Seasonal changes in SST are also realistic in the model (Fig. 1). The equatorial region remains above 28°C, with cooler temperatures in the east during the northeast monsoon (northern summer months).

ITF transport was computed as the depth and cross-strait integral of the model zonal velocity field along a meridional section from Australia to Java. Similar to the results of Potemra et al. (2003), PCM ITF transport occurs mainly in three layers: a surface layer that responds to local and Indian Ocean winds, a subsurface, thermocline layer that responds to lower-frequency forcing in the Pacific, and a deeper layer below the thermocline with very weak flows. Following an EOF analysis of vertical variations of PCM zonal velocity along a section from Australia to Indonesia, transport from the PCM was divided into three layers: upper-layer transport (ULT) from 0 to 132 m (the top five model levels), middepth transport (MLT) from 132 to 559 m (nine model levels), and 559 to 1844 m (five

Fig. 1. Sea surface temperatures (shading in °C) and mean upper-ocean (top 132 m) velocities (streamlines) of the PCM integration: (a) annual mean, and seasonal averages of (b) Dec–Jan–Feb, (c) Mar–Apr–May, (d) Jun–Jul–Aug, and (e) Sep–Oct–Nov.
model levels extending to the bottom of the section). Positive values of ITF transport variations indicate a decrease in flow from the Pacific to the Indian Ocean.

The mean ITF transport from the PCM is 12 Sv (1 Sv = 10^6 m^3 s^-1), which is higher than recent observed estimates of 6 to 10 Sv (e.g., Hautala et al. 2001). Similar to observations, most of the transport occurs in the surface layer (8.2 Sv), with a much smaller contribution at middepth (2.8 Sv) and very low flow at depth (1 Sv; not shown). The annual cycle is realistic, with peak transport of 20 Sv in August during the northern summer (Fig. 3a) when local winds are south-easterly. During boreal winter ITF transport is 6 Sv (January).

b. PCM variability

Perhaps the most well-documented, low-frequency variability in the ocean is that related to the El Niño–Southern Oscillation (ENSO) phenomena. The ENSO amplitude in the PCM is close to observations (Meehl et al. 2001). The ENSO time scale, measured from the power spectrum of the first EOF of equatorial Pacific SST, is between 26 and 36 months. Like most coupled models, the PCM suffers from having an ENSO that is too regular (temporally) and extends too far west.

ENSO affects the Indian Ocean SST primarily via an atmospheric bridge (Latif and Barnett 1995), while associated changes of the ITF are uncertain. We examine the impact of ENSO on ITF variability in more detail in a subsequent study and here remove the ENSO signal by applying a 3- to 10-yr bandpass filter to the model results. The low-pass cutoff ensures a large number of degrees of freedom for our subsequent correlation and regression analyses.

1) INDIAN OCEAN TEMPERATURE VARIABILITY

Bandpass-filtered (36 to 120 months) PCM temperatures in the Indian Ocean do not show much variance in the upper level except for a region near the coast of Sumatra (rms > 0.2°C; Fig. 2a). This occurs where the mean eastward equatorial flow downwells at the coast. The temperature variance is similar to those due to interannual winds associated with ENSO (Annamalai et al. 2005) that cause changes in equatorial flow and coastal upwelling.

Temperature variance in the model’s second level
(25 to 50 m) reflects changes in the Indian Ocean’s southern tropical gyre, between the equator and 15°S in the western half of the basin. It is only in the nearthermocline depth (77 to 104 m) where temperature variability seems to coincide with ITF pathways (Fig. 2c). There is also relatively high variance in the subtropical gyre east of north Madagascar. At deeper depths, the variance in temperature is small and confined to higher latitudes.

2) THROUGHFLOW VARIABILITY

Apart from peaks at the semiannual and annual periods the frequency spectrum of ITF transport (Fig. 3b) is overall flat, with variation of less than an order of magnitude for periods longer than 1 yr. Within this range, the spectrum shows enhanced energy at ENSO periods, between 2 and 3 yr in this model. For ULT and total ITF transport, these raised levels extend to lower frequencies corresponding to periods of 4 to 5 yr with a slight drop in variance to periods of 10 yr. MLT shows no comparable energy at these low frequencies. Most of the presented results, therefore, will focus on ULT variations in the frequency band beyond ENSO to the decadal time scales.

Figure 3c shows the 36- to 120-month bandpass-filtered variability of the PCM ITF transport. The variations, about ±1.0 Sv, are associated with anomalies in heat flux on order 0.1 PW (assuming a reference
temperature of 0°C) and could have a significant impact in the Indian Ocean. Two key points are that the low-frequency variations in ULT and MLT are not significantly correlated, suggesting different forcing mechanisms, and that the variations of MLT are only slightly smaller those for ULT.

3. Covariations in ITF transport and ocean temperatures

The correlation of low-frequency changes in transport (upper and middepth) and ocean temperature (at various depths) occurs in specific regions (Fig. 4) for different depths. Almost all of the significant correlation occurs with temperatures in the surface layer (0 to 25 m) and in the upper thermocline (77 to 104 m). There are three key regions of high correlation: in the ITF outflow region of the eastern Indian Ocean near Sumatra, in the eastern equatorial Indian Ocean, and in the southern tropical gyre. Positive ITF transport variations mean a reduction in the usual Pacific to Indian Ocean flow, so a positive correlation between temperatures and ITF transport occurs when low (high) throughput is coincident with warmer (colder) temperatures. Correlations above 0.3 are significant at the 98% level for these filtered time series.

There are several things to note about these correlations. First, in the direct outflow region the correlations of ULT and MLT to temperatures in the upper and thermocline levels, respectively, are limited to the eastern side of the basin, similar to the correlations of depth-integrated pressure (Fig. 5), and do not show a continuous band across the Indian Ocean as one might expect from Sverdrup dynamics (e.g., HG93). In both cases the negative correlations indicate that low transport is associated with cooler temperatures, as is to be expected for temperature changes driven by advection. The correlation of ULT and surface temperature drops at 110°E (Fig. 4a), while the correlation of MLT subsurface temperatures extends to about 80°E (Fig. 4d). This suggests that interactions with the atmosphere dominate the variance of temperature in the surface
Fig. 5. Depth-integrated ocean pressure from the PCM was computed off southern Java ($P_{\text{java}}$) and northwestern Western Australia ($P_{\text{aust}}$). (a)–(d) The correlations of these signals to model pressure in the Indian Ocean [monthly mean in (a) and (c) and 36- to 120-month filtered in (b) and (d)]. Depth-integrated pressure was also correlated (e), (f) to the cross-ITF pressure difference ($P_{\text{java}} - P_{\text{aust}}$), and (g), (h) to model MLT.
layer and thus diminish the role of the ITF there; below the surface, the interactions with the atmosphere are less prominent and lead to a larger area of correlation between temperature and throughflow.

The correlation of ULT and thermocline depth temperature (Fig. 4c) shows high values off the coasts of Sumatra and Java and drops toward the south. This suggests a geostrophic response through which a stronger ULT deepens the thermocline (Fig. 4a) and lowers temperatures to the north of the ITF path (positive correlation to the north) and has weak negative correlation, corresponding to higher temperatures to the south.

Low-frequency variations in ULT correlate to surface temperatures near the outflow region (stronger ULT corresponds to warmer SSTs) and near the equator off Sumatra (weaker ULT corresponds to warmer SSTs; Fig. 4a). In the latter case, anomalous strengthening of the mean westerlies during monsoon transitions over the equatorial Indian Ocean causes downwelling at the coast of Sumatra, and subsequent Kelvin waves reduce the ITF, so warm SSTs correspond to a weaker ULT by a common surface forcing. The region of correlation is small due the effect of surface heat fluxes. The correlation with subsurface temperatures in this region (Fig. 4c) is more extensive.

There are also negative correlations in the area off Madagascar (Fig. 4c), along the path of the throughflow. These negative correlations correspond to weaker ULT occurring with colder temperatures at depth. This could either be due to reduced advection from the ITF, or from a spinup or shift of the southern tropical gyre. The southern tropical gyre is a cyclonic gyre; if the gyre strength increases, upwelling in the interior of the gyre is also increased. The anomalous westerlies at the equator would account for such an increase. There is no clear link back to the ITF outflow, and so changes in the gyre are the most likely cause for correlations in this region. This is examined in a latter section by comparison to the model wind field.

Outside the direct outflow region, the correlations of ULT to temperature (Figs. 4a,c) are of the opposite sign to those between MLT and temperature (Figs. 4b,d).

4. Influences on Indian Ocean temperatures

The ITF can be viewed along with surface atmospheric forcing as a time-varying boundary condition to the Indian Ocean, supplying both heat and mass. Anomalous heat supplied by the ITF, through advection, can lead to changes in Indian Ocean temperatures. On the other hand, changes in ITF mass transport can influence Indian Ocean temperatures by adjusting the oceanic pressure field along the Indian Ocean’s eastern boundary. These ITF-induced pressure changes will propagate into the Indian Ocean along Rossby wave trajectories and adjust the thermocline throughout the basin.

In addition to computing depth-integrated transport, temperature-weighted transport (in each depth level) was also computed from the model. The former gives an indication of boundary-induced pressure changes while the latter can be used to study changes to the supplied temperature at the boundary. Relative variations in the volume transport are significantly higher than for the transport-weighted temperature; variance in ULT is 36.8 Sv$^2$ with a mean of 8.2 Sv, while variance in transport-weighted temperature is 0.02 (GJ m$^{-3}$)$^2$ with a mean of 0.45 GJ m$^{-3}$. Therefore, only volume transport variations were used in this analysis.

The covariability of ITF and Indian Ocean temperature can be attributed either to changes in ITF transport, or to surface forcing in the Indian Ocean that alters both ITF and temperatures. To untangle this relationship, we consider next the relationship of the ITF with upper-ocean pressure throughout the Pacific. This shows that a large part of the correlation of SST and ITF shown in Fig. 4a is consistent with the wind stresses both laying down SST anomaly patterns and forcing the ITF.

We then consider the combined forcing of ITF and the Indian Ocean wind stress to determine their relative roles.

a. Pacific vs Indian Ocean forcing of the ITF

Following HG93 and HG94, the depth-integrated pressure signal is used to examine the relative contributions of Pacific and Indian Ocean forcing on the ITF. The ITF is determined by the pressure difference between the coasts of Australia and Indonesia. Neglecting local, alongshore wind forcing, the pressure signal off northwestern Western Australia, $P_{\text{aust}}$, reflects Pacific Ocean forcing (Clarke and Liu 1994), while pressure near Indonesia, $P_{\text{java}}$, reflects Indian Ocean forcing. This is indeed the case in the PCM. Figures 5a–d show the correlation of upper-ocean, depth-integrated pressure throughout the Indian Ocean with $P_{\text{java}}$ and $P_{\text{aust}}$.

Here, $P_{\text{java}}$ correlates to depth-integrated pressure in the equatorial Indian Ocean (Fig. 5a). This is consistent with large-scale waves, driven by equatorial winds, that act to raise and lower the thermocline off Java (Wyrtki 1987). At low frequencies (Fig. 5b), the area of negative correlation in the western side of the basin extends to much higher latitudes, indicative of gyre-scale adjustments or of a delayed response to the anomalies of $P_{\text{java}}$. 
Pacific Ocean forcing is apparent for $P_{\text{aust}}$ (Fig. 5c), with maximum correlation along the eastern boundary in the southern Indian Ocean. The correlations become larger at lower frequencies, and extend farther throughout the basin (Fig. 5d).

Finally, the correlation of $P_{\text{java}} - P_{\text{aust}}$, which gives an estimate of ULT, to pressure in the Indian Ocean (Fig. 5e) matches the correlation between ULT and $P$ (not shown). At monthly mean time scales, the correlation is similar to $P_{\text{java}}$, supporting the hypothesis that Indian Ocean forcing controls the higher-frequency changes in ULT. At lower frequencies, however, high (negative) correlations are found along $15^\circ$S. At higher frequencies, using monthly mean model output, lagged correlations show the Rossby wave propagation, but correlations drop below significant values near $80^\circ$E (not shown). Note that there is not a continuous region of high correlation from the ITF region into the Indian Ocean, but that Indian Ocean processes (e.g., surface wind stress curl) erode the signal.

This correlation analysis suggests that low-frequency changes in surface forcing in the equatorial and tropical Indian Ocean impact the variability introduced by the ITF. In other words, further analysis is required to understand what the direct impact of the ITF is. In the subsequent sections we attempt to quantify the effect of Indian Ocean surface forcing as well as ITF-induced changes in Indian Ocean temperatures by using a linear regression model. This method cannot address nonlinear interactions. Furthermore, by construction we will only address changes in throughflow volume transport.

b. Low-frequency variations in surface winds

The correlations between ITF transport and temperature outlined above do not necessarily show cause and effect. It could be the case that both ocean temperatures and ITF transport are responding to similar, low-frequency changes in atmospheric forcing. A similar correlation procedure was performed on atmospheric forcing, transport, and ocean temperature (results not shown). Low-frequency changes in Indian Ocean zonal wind stress are correlated to ULT in a region near the ocean’s eastern equatorial boundary. This is consistent with previous studies (albeit at shorter time scales) that showed the annual and interannual variations of the ITF were controlled in large part by winds in the Pacific (Potemra 1999). Intraseasonal variability in ITF transport is mainly controlled by equatorial winds [MJO or intraseasonal oscillations (ISOs)] in the Indian Ocean. Low-frequency variability in the ISOs, therefore, may be reflected in low-frequency variations in ULT. There is also evidence of correlation between wind stress curl and level-4 temperatures in the model in the tropical and subtropical gyres in the southern Indian Ocean.

c. The competition of throughflow and Indian Ocean atmospheric forcing

In order to quantify the relative impacts of ITF we consider the hypothesis that Indian Ocean temperature fluctuations result from the combined forcing of the Indian Ocean atmosphere, as represented by the wind stress and its curl, and by the ITF. Since the focus here is on time scales longer than the adjustment time of the Indian Ocean by Rossby waves, and longer than typical damping times of surface temperature anomalies by air–sea fluxes (3/4 yr; Barnett et al. 1991), Indian Ocean temperature variations are related to contemporaneous time series of the ITF and the Indian Ocean atmosphere using a multilinear regression (MLR) model. At each point $(x, y, z, t)$ anomalies of temperature $T(x, y, z, t)$ are fit using the following linear model:

$$T(x, y, z, t) = \sum_i \alpha_i (x, y, z) F_i(t) + \epsilon,$$

(1)

where $F_i(t)$ are low-frequency time series of ULT, MLT, and the atmospheric indices all normalized to unit standard deviation. The regression coefficients $\alpha_i$ are obtained by least squares fitting that minimizes the variance of the residual $\epsilon$ (e.g., Schneider and Cornuelle 2005). The skill of the model is evaluated by the correlation of the left-hand side and the reconstruction from the $F_i(t)$ time series.

For atmospheric forcing indices we use the leading EOFs of the 36- to 120-month bandpass-filtered zonal wind stress and wind stress curl. It should be noted that the first 70 yr were removed from all time series, to eliminate model spinup. The input time series were also orthogonalized (to zonal wind stress EOF-1), although this orthogonalization does not alter the spatial pattern of the resulting $\alpha_i$ values. Figure 6 shows the projection of zonal and meridional wind stress as well as wind stress curl onto the leading EOF of zonal wind stress as an example of one index of forcing. This mode, and variations associated with it, are characterized by changes in equatorial zonal winds and alongshore winds off Sumatra. Both are important factors for adjusting the thermocline off Sumatra as well as throughflow transport and, for the signs shown in Fig. 6, tend to deepen the thermocline in the eastern equatorial Indian Ocean, and decelerate the ITF. South of the equator in the region of the subtropical gyre, veering of winds from southeasterlies to southweste]
The MLR (1) was first constructed using bandpass-filtered ULT, MLT, the first three EOFs of zonal wind stress, and the first three EOFs of wind stress curl (8-function fit). Most low-frequency variance in temperature occurs in model level 4 (77 to 104 m; see Fig. 2c), so the fit was made using temperatures at this model level as well as the surface level. Correlations between the MLR fit and model temperatures do not significantly increase in the regions of high temperature variance by including MLT, as suggested by the spectra in Fig. 3b, and higher EOFs of wind stress and curl, so the discussion here is limited to the fit of ULT, the first two EOFs of zonal wind stress, and the first EOF of wind stress curl (4-function MLR fit).

Figure 7 shows the correlation of both the 8- and 4-function MLR fits described above to model temperature in level 1 (0 to 25 m) and level 4 (77 to 104 m). The skill of the MLR fit (measured by a correlation between the MLR fit and the original temperature field) is high in the region off Sumatra and in the tropical and northern subtropical gyres.

Regions of high standard deviation in temperature (contours in Fig. 7) are reproduced well by the 4-function MLR. Most of the variability at these frequencies (36 to 120 months) is found in model level 4, and the skill of the MLR fit is highest off Sumatra and in the tropical gyre. The subtropical gyre is better reproduced when more functions, specifically the three EOFs of wind stress curl (rather than just the first EOF), are used. Nevertheless, the MLR fit of the zonal wind
stress, wind stress curl, and ITF transport is able to reproduce 36- to 120-month variations in temperature in the PCM.

The regression coefficients, \( \alpha_i \), of the MLR fit (1) represent the response of temperature to a unit amplitude variation of the respective forcing. While the forcing time series may share significant correlations, their unique regression patterns are fixed by where they are independent, and the regression coefficients are an estimate of the importance of each forcing index.

For surface temperature, the regression coefficients show the dominance of Indian Ocean wind stress, in particular EOF-1 of the zonal wind stress. The zonal stress in the equatorial region (Fig. 8b) creates downwelling off Sumatra and is associated with warm SST in the eastern Indian Ocean and cool SST in the west, which takes a form similar to the Indian Ocean dipole (Saji et al. 1999). The impact of ULT (Fig. 8a), in contrast, is very weak, and at best is limited to the entry region between Java and Australia. Overall, however, the Indonesian throughflow accounts for a small fraction Indian Ocean temperature variance in this model. The second EOF of the wind stress and index of the wind stress curl have some expression in the throughflow entry region but show most of their impact on the far South Indian Ocean (Figs. 8c,d).

In contrast to the very limited impact of the ULT on SST, the 100-m temperatures show stronger sensitivity to the ITF. The regression coefficient for ULT (Fig. 9a) shows that an increase of ULT (negative anomaly) warms the thermocline temperature in the outflow region and between 10° and 15°S in the central Pacific. In addition, in the South Indian Ocean, large-scale but weak, cool temperature anomalies occur.

However, even for ULT, atmospheric forcing clearly dominates the temperature response. The regression pattern of zonal wind stress EOF-1 (Fig. 9b) shows a large dipole in the equatorial and southern tropical Indian Ocean. Comparison of this pattern with wind stress and wind stress curl associated with this EOF (Fig. 6) shows that the equatorial warming in the east and cooling in the west are a response to the equatorial zonal wind stress, where the slight shift off the equator to 5°S in the west is consistent with the \( \beta \) effect. In the west, thermocline depth warming results from the combination of downwelling due to poleward alongshore
wind stress and offshore Ekman downwelling. Farther to the west this warming is then reduced by the Sverdrupian response to Ekman pumping at 90°E. The secondary maximum of the regression pattern between 10° and 15°S east of Madagascar is clearly a result of the Sverdrup transport induced by central Indian Ocean wind stress curl at these latitudes. The higher EOF of wind stress and the leading EOF of its curl (Figs. 9c,d) add to the variance of thermocline temperature between 10° and 15°S and in the South Pacific. Overall, we conclude that the Indonesian throughflow plays at best a minor role in determining the low-frequency variance of SST. Its impact on thermocline depth in the Indian Ocean is limited to the direct outflow region and its extension across the basin. The widespread correlation between the throughflow and Indian Ocean temperature reflects their respective responses to the common atmospheric forcing.

5. Summary and conclusions

The PCM, a global, coupled model was used to investigate the impact of low-frequency changes in throughflow transport on upper-ocean temperatures in the Indian Ocean. We find that low-frequency increases in the upper layer (ULT) of the ITF lead to warmer temperatures in the thermocline in the western portion of the SEC near Madagascar. For the surface temperatures, the ITF does not account for a significant fraction of the variance at low frequency. Other low-frequency changes in Indian Ocean temperatures are correlated with low-frequency atmospheric variability. Most notable of these are temperatures within the southern gyres and temperatures along the coast of Sumatra. In this coupled model, therefore, throughflow effects on low-frequency Indian Ocean temperatures are confined to the direct SEC path, at the depth of the thermocline. Surface ITF transport (transport above the thermocline) is the main contributor. Changes in subsurface ITF transport, while coincident with temperature changes in the Indian Ocean subtropical gyre, are caused by similar atmospheric forcing. Low-frequency variations in surface temperature are small, and in this model are controlled by variations in wind stress and wind stress curl.

The implications of this model study are therefore

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**Fig. 8.** Regression patterns of the multilinear regression model (°C) for unit standard deviation anomalies of the forcing time series, to PCM level 1 (0 to 25 m); the color shading gives the regression coefficients of (a) the ULT throughflow transport, leading EOFs (b) 1 and (c) 2 of the zonal wind stress, and (d) leading EOF of the wind stress curl. The title over each panel indicates the particular function, and the contours show the standard deviation of the low-frequency temperatures (see Fig. 2).
that the impact of low-frequency variations in Indonesian throughflow transport, while appearing large in model simulations with open/closed Indonesian straits, is in fact modest. The atmospheric states that affect the strength of the throughflow also impact, and in fact dominate, the low-frequency variance of upper Indian Ocean temperature. It remains, however, a possibility that the anomalous atmospheric forcing of the Indian Ocean is in part a response to changes of the upper ocean in the outflow region between Java and Australia. The correlations of the throughflow and the atmospheric forcing suggest that this explains only a portion of the atmospheric forcing and suggests future studies of this possible atmospheric interaction. In addition, the results presented here are based on a single coupled model only, with very small low-frequency ITF variability, and are therefore subject to the limitations inherent to this model, including the strong numerical diffusion present in coarse-resolution models. However, the dynamic consistency provides a plausible hypothesis for interannual to decadal variations in the Indian Ocean.

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Fig. 9. Same as in Fig. 8, but for PCM level 4 (77 to 104 m).


