Genesis of Indian Ocean Mixed Layer Temperature Anomalies: A Heat Budget Analysis

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(Manuscript received 3 February 2009, in final form 15 May 2010)

ABSTRACT

The genesis of mixed layer temperature anomalies across the Indian Ocean are analyzed in terms of the underlying heat budget components. Observational data, for which a seasonal budget can be computed, and a climate model output, which provides improved spatial and temporal coverage for longer time scales, are examined. The seasonal climatology of the model heat budget is broadly consistent with the observational reconstruction, thus providing certain confidence in extending the model analysis to interannual time scales. To identify the dominant heat budget components, covariance analysis is applied based on the heat budget equation. In addition, the role of the heat budget terms on the generation and decay of temperature anomalies is revealed via a novel temperature variance budget approach. The seasonal evolution of the mixed layer temperature is found to be largely controlled by air–sea heat fluxes, except in the tropics where advection and entrainment are important. A distinct shift in the importance and role of certain heat budget components is shown to be apparent in moving from seasonal to interannual time scales. On these longer time scales, advection gains importance in generating and sustaining anomalies over extensive regions, including the trade wind and midlatitude wind regimes. On the other hand, air–sea heat fluxes tend to drive the evolution of thermal anomalies over subtropical regions including off northwestern Australia. In the tropics, however, they limit the growth of anomalies. Entrainment plays a role in the generation and maintenance of interannual anomalies over localized regions, particularly off Sumatra and over the Seychelles–Chagos Thermocline Ridge. It is further shown that the spatial distribution of the role and importance of these terms is related to oceanographic features of the Indian Ocean. Mixed layer depth effects and the influence of model biases are discussed.

1. Introduction

The Indian Ocean (IO) has a significant influence on regional climate variability (see a review by Schott et al. 2009). Modes of IO SST variability (e.g., Saji et al. 1999; Reason 1999; Behera and Yamagata 2001; England et al. 2006) are found to impact rainfall over various regions such as Africa (Black 2003; Ummenhofer et al. 2009b), India (Ashok et al. 2001), Sri Lanka (Zubair et al. 2003), and Australia (Ansell et al. 2000; Ashok et al. 2003; Ummenhofer et al. 2008; Ummenhofer et al. 2009a). It has also been shown that summer rainfall over South Africa is linked to temperature variations in the Agulhas Current (Walker 1990; Jury et al. 1993). Understanding how SST anomalies evolve in the IO is therefore of great importance for improving long-term rainfall prediction in these regions. SST variability is essentially set by a combination of local and remote processes affecting the whole of the surface mixed layer via air–sea heat fluxes, advection, entrainment, mixing, and diffusion. It is thus important to study the heat budget evolution of the mixed layer, which essentially governs the genesis of modes of climate variability.

A number of studies have recently been undertaken focusing on the mixed layer heat budget of the IO on seasonal time scales, for which observational data are sufficient (e.g., Rao and Sivakumar 2000). Studies on interannual time scales are, in contrast, relatively few in number, as oceanographic observations tend to be sporadic with few time series of sufficient duration and often sparse spatial coverage. Modeling studies exist but mainly use OGCMs forced by prescribed air–sea heat fluxes (e.g., de Boyer Montégut et al. 2007b) or coupled to a mixed-layer atmospheric model (e.g., Murtugudde and Busalacchi 1999). It has been suggested, however,
that ocean–atmosphere coupling is important over certain regions of the IO (Murtugudde and Busalacchi 1999; Saji et al. 1999), implying the need for a fully coupled model to resolve air–sea feedbacks for simulating ocean variability. An ocean-only forced model would omit these coupled feedback processes while relying on prescribed fluxes derived from reanalysis products whose reliability is questionable back before the satellite era. Moreover, most of these studies focus on limited regions of the IO. In this study, we analyze output from a fully coupled global circulation model and conduct a heat budget analysis of the mixed layer for the entire IO Basin covering seasonal-to-interannual time scales. In this way, we aim to obtain a broad understanding of the mixed layer thermal variations in terms of the governing heat flux components and their role in the genesis of temperature anomalies. Given the large domain and long time series, we present and implement two complementary heat budget analysis methods for revealing the dominance and role of the different heat budget terms, and demonstrate how these differ on seasonal and interannual time scales.

The rest of this paper is organized as follows. Section 2 details the heat budget equation of the mixed layer and describes and validates the climate model. The heat budget climatology for both the climate model and available reanalysis data are also compared. The seasonal cycle of the heat budget components at selected regions is discussed in a more detail in section 3. The heat budget analyses are presented and implemented in section 4. Section 5 provides further discussions on the interannual time-scale thermal variations. Finally, the results are summarized in section 6.

2. Methodology

a. Heat budget of the mixed layer

The heat budget in the mixed layer can be expressed as

\[ \theta_1 = Q_{\text{net}} - \mathbf{u} \cdot \nabla \theta - \frac{w_{\text{ent}} (\theta - \theta_d)}{h} + \text{Res} \]  

(1)

(e.g., Qiu 2000; Qu 2003; Du et al. 2005), where \( \theta \) is the potential temperature averaged over the mixed layer, the subscript \( d \) denotes time differential operator, and \( h \) is the mixed layer depth (MLD). Here, \( Q_{\text{net}} \) represents the effective net surface heat flux retained within the mixed layer, specified by

\[ Q_{\text{net}} = \frac{Q}{\rho_0 C_p h} \]  

(2)

where \( \rho_0 \) is a reference density (1026 kg m\(^{-3}\)), \( C_p \) is the specific heat capacity of seawater (3986 J kg\(^{-1}\) K\(^{-1}\)), and \( Q \) is the net air–sea heat flux absorbed within the mixed layer. Here, \( Q \) can be further decomposed into its components: \( Q = SW + LW + LH + SH - q_d \), where \( SW, LW, LH, \) and \( SH \) are shortwave, net longwave, latent heat, and sensible heat fluxes at the ocean surface, respectively, and \( q_d \) is the portion of the shortwave radiation that escapes through the base of the mixed layer. Following Paulson and Simpson (1977), \( q_d \) is estimated as

\[ q_d = SW[Re^{(-h/\gamma_1)} + (1 - R)e^{(-h/\gamma_2)}], \]  

(3)

where \( R, \gamma_1, \) and \( \gamma_2 \) are coefficients that depend on water turbidity as classified by Jerlov (1968). For example, the Southern Ocean approximately falls into a water type with \( R = 0.67, \gamma_1 = 1, \) and \( \gamma_2 = 17 \) (i.e., Jerlov’s water type IB; see also Dong et al. 2007).

The second term on the rhs of (1) expresses the oceanic advection of heat. The horizontal current velocity \( \mathbf{u} \) is an average over the MLDO and can be decomposed into the Ekman \( [\mathbf{u}_e = ((\rho_0 f)/h)^{-1}(\tau^x, -\tau^y)] \) and non-Ekman components, where \( f \) is the Coriolis parameter and \( (\tau^x, \tau^y) \) are the zonal and meridional components of sea surface wind stress.

The third term on the rhs of (1) expresses the heat flux due to entrainment. Ignoring diffusion, the entrainment rate \( (w_{\text{ent}}) \) can be diagnosed as

\[ w_{\text{ent}} = h_1 + \nabla \cdot (h \mathbf{u}) \quad \text{if this term} > 0 \]  

\[ w_{\text{ent}} = 0 \quad \text{otherwise}, \]  

(4)

after Qiu and Kelly (1993), where \( h_1 \) is the rate of change of the MLD, and \( \nabla \cdot (h \mathbf{u}) \) represents the divergence of mass in the mixed layer, which includes vertical advection at the base of the mixed layer. Here, \( \theta_d \) is the temperature of entrained fluid immediately below the mixed layer (taken at 5 m below MLD; Du et al. 2005).

The residual term (Res) represents all remaining unresolved processes, such as lateral diffusion.

b. The climate model

We analyze output from the 1990 control integration of the National Center for Atmospheric Research (NCAR) Community Climate System Model, version 2 (CCSM2). CCSM2 is a fully coupled atmosphere–land–ocean–sea ice GCM. The atmospheric model employs a spectral T42 horizontal resolution and 26 vertical levels with a prognostic scheme for cloud condensate (Rasch and Kristjannsson 1998), a cloud water vapor emissivity–absorptivity scheme (Collins et al. 2002), and a cloud overlap scheme (Collins 2001). The ocean model employs the Parallel Ocean Program (POP) code with a uniform zonal resolution of \( \approx 1^\circ \), a meridional resolution varying
from 0.27° at the equator to about 0.5° at midlatitudes, and with 40 levels in the vertical. There are 14 levels in the top 200 m with increasing thickness from 10 to 30 m. The ocean model implements the Gent and McWilliams (1990) parameterization for mesoscale eddies, the K-profile parameterization (KPP) scheme of Large et al. (1994) for vertical mixing, and spatially variable anisotropic horizontal viscosity coefficients (Smith and McWilliams 2003). The model allows for variations in sea surface height but with a fixed ocean volume. The model implements Jerlov’s (1968) water type IB (see section 2a) for computing $q_d$ and determines MLD within the KPP scheme (Large et al. 1994). The model was run without flux adjustment. Further model details and general performance can be found in Kiehl and Gent (2004) and Sen Gupta and England (2006). As we show later, the model simulates the overall features of the IO climate reasonably well, such as the monsoon circulation and the southwest tropical IO Thermocline Ridge [5°–12°S, 55°–90°E; also termed the Seychelles–Chagos Thermocline Ridge (SCTR); Hermes and Reason 2008], as well as overall IO climate variability.

Figures 1a and 1b present the zonally averaged monthly zonal wind stress in the model and the National Centers for Environmental Prediction (NCEP) reanalysis. The seasonality of the monsoon and midlatitude circulations are well captured by the model, including the semiannual cycle of the equatorial winds. There are, however, some discrepancies in the magnitude (Fig. 1c) that can help explain biases in the thermocline depth. The model’s annually averaged thermocline depth is plotted along with the Commonwealth Scientific and Industrial Research Organisation (CSIRO) Atlas of Regional Seas 2006 (CARS06) gridded dataset (Ridgway et al. 2002) in Figs. 1d and 1e. The SCTR is simulated by the model, although it is shallower and extends farther east than “observed.” The thermocline in the eastern tropical IO (ETIO) is also too shallow. These biases can be linked to the stronger than observed trade winds (see also Wajsowicz 2004; Hermes and Reason 2009). The eastern tropical thermocline bias is most apparent in austral winter to spring (Fig. 1f), when the overly strong trade winds are located around the equator, shoaling the eastern thermocline. In addition, the weaker simulated equatorial westerlies in October (Figs. 1a–c) result in weaker Wyrtki jets and thus weaker eastward downwelling Kelvin waves. The bias in the SCTR region is most apparent over the first half of the year (Fig. 1g), when the trade winds and the westerlies to the north are too strong (Fig. 1c), implying a stronger upwelling favorable wind stress curl. The thermocline bias in the northern parts of the basin can be linked to biases in the monsoon circulation (Fig. 1c).

Figures 1h–j compare the model annual-mean MLD against the observations (de Boyer Montégut et al. 2007a). In agreement with observations, the deepest mixed layers occur at midlatitudes, where strong westerlies and winter cooling lead to vigorous mixing. The largest biases are also found in this region, where the MLD is too deep along the Subantarctic Front and too shallow farther south. A number of factors can contribute to these biases, including wind biases, particularly as the model midlatitude westerlies are too strong and situated too far north (Figs. 1a–c; see also Sen Gupta and England 2006, their Fig. 3). This can cause changes in turbulent mixing, heat fluxes, and the position and strength of Ekman pumping. Another possible contributing factor is the larger simulated precipitation at high latitudes (Figs. 1k–m). Excessive precipitation can spuriously enhance stratification, leaving the MLD too shallow and giving rise to a thicker barrier layer compared to the observed (figure not shown; see also, e.g., de Boyer Montégut et al. 2004). This results in a small temperature gradient across the base of the mixed layer, thus underestimating entrainment. Thick barrier layers are also known to occur off Sumatra (Qu and Meyers 2005; de Boyer Montégut et al. 2007a). The lack of precipitation in the CCSM2 in this region (Fig. 1m), along with the deeper than observed MLD (Fig. 1j), is an indication that the barrier layer thickness there is underestimated. This would potentially enhance the effect of entrainment on the mixed layer. Despite these biases, the seasonal evolution of the MLD at various regions is in fairly good agreement with the observed (see Fig. 2 for model and observed MLD averaged over regions indicated in Fig. 1h). Apparent biases in the Bay of Bengal, western Arabian Sea, off Sumatra, and off northwestern Australia (Figs. 2a,b,d,f) may also be linked to discrepancies in the wind strength and precipitation (Figs. 1c,m). Overall, the model captures many of the pertinent features of the IO, as well as many of the Intergovernmental Panel on Climate Change (IPCC)-class climate models we are familiar with (see Sen Gupta et al. 2009).

The depth of the thermocline plays an important role in the strength of the coupling between ocean and atmosphere. The spurious shoaling of the eastern tropical thermocline enhances the Bjerknes feedback in the model, forcing tropical variability that is too strong. This can be seen in Figs. 3a and 3b, which presents the standard deviation of SST in the model and observed [Hadley Centre Sea Ice and Sea Surface Temperature (HadISST)]. Not surprisingly, the model then simulates higher tropical rainfall variability than in observations [based on Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) rainfall; Figs. 3c,d]. However, the large-scale IO SST and rainfall variability patterns are
comparable with the observed. For example, the model captures the major modes of IO variability as demonstrated by a standard EOF analysis applied to the monthly IO SST north of 20°S. EOF1 captures an IO basin mode (not shown), accounting for 20% of total SST variance, which is known as a delayed response to the El Niño–Southern Oscillation (ENSO; e.g., Klein et al. 1999; Du et al. 2009). By correlating the model Niño-3.4 index
against gridpoint SSTs at a 6-month lag, a basin-wide warming pattern is obtained (Fig. 4a; see also Saji et al. 2006, their Fig. 4). There is, however, some reminiscence of an Indian Ocean Dipole (IOD; Saji et al. 1999) signal, likely due to the thermocline bias mentioned earlier. A subtropical dipole pattern (Behera and Yamagata 2001; England et al. 2006) is also evident, supporting its possible connection to ENSO (Hermes and Reason 2005). EOF2 reveals an IOD-like pattern (17% of total variance), with peak variability in September–November (SON) as observed (not shown). Linear regression of wind stress and thermocline depth patterns against the EOF2 principal component (Fig. 4b) depict a state of positive IOD marked by easterly wind anomalies and an east–west sloped thermocline. These patterns are in qualitative agreement with those inferred from reanalysis products (Saji et al. 2006; see their Fig. 9).

Existing climate models simulate IO climate variability and teleconnections to varying degrees of realism (e.g., Saji et al. 2006; Cai et al. 2009), with many exhibiting biases, such as those described previously. The model used here simulates IO mean climate and its variability reasonably well, capturing most aspects of the real system. It is encouraging that the seasonality simulated by the model is in good agreement with the observed. Coupled with the model’s ability to simulate the general features of the IO, this implies a decent representation of the atmosphere–ocean heat fluxes—given the model was run without flux corrections. We will show in this study that on seasonal time scales there is a close correspondence between the simulated heat budget terms and those reconstructed from observational products. This provides a certain confidence in extending our analysis out to interannual time scales. Nevertheless, the model biases must be borne in mind in interpreting our results.

c. Heat budget climatology

The model’s heat budget terms are calculated as in Eq. (1), using 100 yr of monthly output. Long-term monthly means of these terms are shown in Fig. 5. For comparison, we construct a monthly climatology of the heat budget terms (Fig. 6) utilizing reanalysis data and the de Boyer Montégut et al. (2007a) 2° × 2° MLD climatology (Figs. 1i, 2). The MLD climatology was reconstructed using temperature and salinity profiles collected between 1967 and 2006 as provided by the National Oceanographic Data Center (NODC), World Ocean Circulation Experiment (WOCE) database, and Argo Global Data Centers (GDAC). We utilize the NCEP Reanalysis 1 for surface heat and momentum flux data from 1970–2006 and the Archiving, Validation, and Interpretation of Satellite Oceanographic data (AVISO) [TOPEX/Poseidon, Jason-1 + European...
Remote Sensing Satellite (ERS), Envisat, and the Geosat Follow-On (GFO)]. For ocean temperature, we utilize the CARS06 gridded datasets that are on a 0.5° × 0.5° horizontal grid with 79 depth levels. Prior to computing the heat budget terms, all the reanalysis variables are mapped onto a uniform 1° × 1° grid. A 6° × 6° triangle filter is applied horizontally on the temperature and geostrophic velocities to exclude the effect of mesoscale processes that are not resolved by the coarse NCEP surface heat flux and wind stress fields.

The monthly evolution of the terms in the model is in broad agreement with the observations. The discrepancies that exist can be partly attributed to the disparity in MLD (Figs. 1j, 2). It should be noted, however, that the observed MLD estimate may itself suffer from uncertainties as a result of the estimation method. Uncertainties in the NCEP heat flux product (Yu et al. 2007) are also likely to contribute to the model–observed differences. For example, omitting $q_d$ in the model would cause an overestimate of the warming rate by about 2°C month$^{-1}$ off northwestern Australia in January when the mixed layer there is only about 15 m deep (Fig. 2f). As noted earlier, the Res term also includes lateral and vertical mixing and diffusion. Thus, the larger observed residual might also be due to an underestimation of diffusive-scale processes represented in the model. Numerical errors in the offline estimation of the model heat budget are also included in Res.

For the rest of the paper, we focus our analysis on the model heat budget terms, particularly the air–sea heat flux, advection, and entrainment terms. Figures 5 and 6 reveal a broadly consistent picture between the heat budget seasonal cycle in the model and reconstructed observed climatology. We will again observe this consistency as we review the seasonal cycle in section 3. This motivates us to then extend our analysis to the interannual variability in the model. The reconstruction of observed IO MLD, in particular at a monthly resolution for an extended period, makes such an analysis impractical for the observational data.

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**Fig. 3.** (top) Standard deviation of annual-mean SST in (a) the model and (b) the HadISST reanalysis (1950–2008). Values $>$0.3°C are shaded in gray. Contour interval is 0.05°C. The model analysis uses 59 yr of data picked arbitrarily to match the length of the time series in the reanalysis. (bottom) Standard deviation of annual-mean rainfall in (c) the model and (d) CMAP reanalysis (1979–2005). Values $>$0.5 mm day$^{-1}$ are shaded in gray. Contour interval is 0.25 mm day$^{-1}$. The model analysis uses 27 yr of data to match the reanalysis.
On a hemispheric scale, the seasonal pattern of net air–sea heat flux $Q_{\text{net}}$ roughly matches that of the heat storage rate $u_t$, indicating the dominance of the air–sea heat flux on these time scales (Figs. 5, 6). The influence of insolation is apparent in the annual cycle of $u_t$ in the extratropical Southern Hemisphere, which is in a warming (cooling) phase in austral spring–summer (autumn–winter). In the northern IO, there is a strong semiannual cycle in $u_t$ with a cooling phase during boreal winter and summer and a warming phase during the premonsoon seasons. This also matches the evolution of $Q_{\text{net}}$. As illustrated later, however, there are regions where the seasonal cycle of $u_t$ shows significant discrepancies with $Q_{\text{net}}$. In these regions, advection and entrainment play an important role. In this section, we provide an overview of the seasonal cycle in the mixed layer heat budget for selected regions and the associated large-scale atmosphere–ocean processes driving these changes. To aid in the discussion, we present the simulated monthly heat budget components in Figs. 7 and 8, averaged over the regions indicated. The simulated seasonal wind stress, MLD, ocean currents, and mixed layer temperature are presented in Fig. 9. The seasonal evolution of the regional heat budget is generally insensitive to the spatial extent of the box averaging. Processes of interest that occur outside the box will be noted. For comparison, the corresponding observational estimate of $u_t$ is also shown in the left panels of Figs. 7 and 8. For most regions, there is a reasonable agreement between the model and observed $u_t$. Slight discrepancies can be seen in boreal spring over the northern Indian Ocean regions, which appear to be linked to the MLD and atmospheric biases noted in section 2b.

### a. Bay of Bengal

In the Bay of Bengal, there is a warming tendency in boreal spring and autumn and a cooling tendency in winter and summer (Fig. 7a) driven by air–sea heat fluxes, with limited contribution from ocean processes during certain months. Cooling by the net air–sea heat flux ($Q_{\text{net}}$) is strongest in winter [November–January (NDJ)], mostly because of latent heat loss enhanced by the cold and dry northeasterly winds that deepen the mixed layer (Fig. 9a). The spring [March–May (MAM)] is marked by weaker winds (Fig. 9b) and peaked shortwave radiation (Fig. 7a, middle) because of clear-sky conditions (e.g., Rao and Sivakumar 2000), resulting in the strongest heating of the year. During this time, the intertropical convergence zone (ITCZ), a region of intense convection and high cloud cover, is still farther south straddling the equator (Waliser and Gautier 1993). The weak boreal summer–autumn $Q_{\text{net}}$ is due to low shortwave radiation, an indication of increased cloud cover as the ITCZ is expanding northward (see also de Boyer Montégut et al. 2007b). Enhanced latent heat loss in summer by southwesterly winds (Fig. 9c) leads to further cooling of the summer mixed layer.

The area-averaged advection term is small, as it is limited to the southern entrance of the bay (Figs. 5, 6). The effect of the entrainment term is also localized, with a cooling off southeast India in boreal summer–autumn balancing the warming due to advection. A wind-driven upwelling has been previously reported for this region during this time of the year (Shetye et al. 1991). The dominance of the air–sea heat flux over ocean advection was also detected by Shenoi et al. (2002) using observational products for the upper 50 m.
b. Western Arabian Sea

In this region, there is again a distinct semiannual cycle to the temperature tendency. This is driven by competing influences of the air–sea heat flux and entrainment (Fig. 7b). Entrainment starts in late boreal spring, peaks in June–August (JJA), and terminates in October, linked to the evolution of the summer monsoon southwesterlies that cause coastal upwelling (Figs. 9c,h; e.g., Shenoi et al. 2002). On the other hand, $Q_{\text{net}}$ warming starts earlier (Fig. 7b, left) with the demise of the monsoon northwesterlies (Fig. 9b). Then, $Q_{\text{net}}$ gradually builds up and peaks in August (Fig. 7b, left) despite little change in the shortwave radiation (Fig. 7b, middle). This enhanced air–sea heat flux is instead driven by weaker outgoing longwave and suppressed evaporation because of the presence of cold winter upwelling waters (Rao and Sivakumar 2000; de Boyer Montégut et al. 2007b). The slight increase in MLD in JJA (Fig. 2b) also contributes, to a lesser extent, by reducing the amount of shortwave radiation escaping from the base of the mixed layer. As a result, the positive boreal spring and autumn heat storage rates are set by $Q_{\text{net}}$ with a damping effect from entrainment, whereas the early summer cooling is dominated by entrainment. On the other hand, the winter cooling is solely due to $Q_{\text{net}}$ as a result of wind-driven latent heat loss and reduced insolation (see also de Boyer Montégut...
et al. 2007b). It is interesting to note that $Q_{\text{net}}$ appears more amplified in boreal summer (Fig. 7b, left), despite the comparable magnitudes of summer and winter atmospheric heat flux ($Q$; Fig. 7b, middle). This is due to the modulating effect of MLD variation that is shallow in summer ($\approx 25$ m) and deep in winter ($\approx 70$ m; Fig. 2b).

c. Equatorial regime

Compared to higher latitudes, the magnitude of $Q_{\text{net}}$ variations within the equatorial trough is relatively weak (Figs. 5, 6), with weak seasonal insolation, extensive cloud cover, and consistently high humidity. It is thus not surprising that ocean processes contribute more substantially to the evolution of the heat balance. In the western tropical Indian Ocean (WTIO), off Somalia, there is an interplay between the air–sea and ocean heat fluxes, which produces a semiannual signal in $\theta_t$ (Fig. 7c, left). The heat flux variations are essentially linked to the reversal of the monsoon winds and Somali Current (Fig. 9). During the southwest monsoon season, the southwesternly winds force the northward Somali Current, which deflects eastward at about $5^\circ$N. This creates a region of coastal upwelling, entraining colder water into the mixed layer. During the northeast monsoon, the winds reverse and the Somali Current flows southward to meet the East African Coastal Current before joining the South Equatorial Counter Current, creating a coastal upwelling just south of the equator (Schott and McCreary 2001).
As shown in Fig. 7c, entrainment dominates the heat budget in May–July and November–January. The seasonal variations in ocean currents drive the evolution of the advection term that is in phase with $u_t$. The alongshore monsoon winds also enhance latent heat loss and deepen the mixed layer in DJF and JJA (Figs. 2c, 5a,c). Thus, $Q_{\text{net}}$ dominates the heat budget during monsoon transitional months, particularly in September–October, when latent heat loss and ocean heat fluxes are weakest and the mixed layer is shallowest, resulting in a warming of the mixed layer.

Off the west coast of Sumatra, there is substantial contribution by entrainment from July to November (Fig. 7d). Ekman upwelling is known to occur in June–October because of the monsoon southeasterlies (Susanto et al. 2001). However, using a high-resolution OGCM, Du et al. (2005) found that the entrainment cooling there is small because of the existence of a thick barrier layer. They also found a large negative residual term (see also Figs. 5, 6), implying the importance of unresolved mesoscale processes. The entrainment in our model is likely to be too strong, mainly in association with the shallow thermocline bias during austral spring (section 2b). At this time, the entrainment is balanced by weakened evaporation and enhanced shortwave radiation, which may in turn be preconditioned by the cold entrainment-induced SST and clear skies. These factors contribute to the discrepancy between the model and observed $\theta$, from August to December (Fig. 7d, left).

d. South tropical Indian Ocean

The magnitude of advection and entrainment is relatively large between the equator and 15°S, that is, under the influence of the trade winds (Fig. 9c). The entrainment term is concentrated over the SCTR, with increasing magnitude from austral summer to autumn diminishing in spring (Fig. 8a, left; see also Figs. 5, 6). The autumn peak coincides with the seasonal minimum depth of the SCTR (Fig. 1g). The SCTR and its variations are set by the interaction between the southeast trades and the Indian monsoon, via Ekman pumping as well as by Rossby waves from the east (Hermes and Reason 2008; Yokoi et al. 2008; see also a review by Vialard et al. 2009). When the southeast trades reach their strongest and most equatorward position in JJA (Fig. 9c), along with the northward shift of the ITCZ, the mixed layer is
cooled via latent heat flux (Fig. 8a, middle). At the same time, the winds drive warm meridional Ekman advection, moderating the air–sea cooling (Fig. 8a, right). This Ekman advection dominates the heat budget in SON when entrainment becomes small, coinciding with the annually deepest thermocline (Fig. 1g). This is also around the time when the Indonesian Throughflow (ITF) is at a maximum, contributing to the warming primarily to the southeast of the region (Figs. 5, 6). Reduced latent heat loss and increased solar radiation then drive mixed layer warming during DJF, amplified by the mixed layer shoaling (Fig. 2e). As noted by Yokoi et al. (2008), the SST evolution is dominated by the annual harmonic. As described here, the annual signal is set by interactions of the air–sea and ocean heat fluxes.

In contrast, for the region off northwestern Australia, $\theta_r$ is almost entirely determined by $Q_{\text{net}}$ (Fig. 8b). Entrainment and advection have only a minor effect. This is consistent with the modeling study by Du et al. (2005). The evolution of $Q_{\text{net}}$ is driven by changes in both the shortwave and latent heat fluxes, with an amplification in DJF due to the shallow mixed layer (Fig. 2f). Advection has a weak warming effect during austral winter because of increased meridional temperature gradients and southward Ekman transport (Fig. 8b, right; Fig. 9h).

The geographical configuration of the northwestern shelf under the southeastlies allows a southwestward alongshore Ekman advection. This, coupled with a deep thermocline ($\approx 150$ m; Fig. 1d), means that there is little large-scale Ekman upwelling. Nevertheless, some entrainment does occur during austral summer at around 20°S (Fig. 5) when the vertical temperature gradient is larger with weak northwesterlies (Fig. 9a), causing coastal upwelling at the southern edge of the Northwestern Shelf.

e. Subtropics (15°–35°S)

The seasonal evolution of the mixed layer in the subtropics is largely dictated by insolation (Figs. 5, 6). Here, there is a year-round warming tendency from the southwestward advection concentrated at 15°S. This is due to the South Equatorial Current (SEC), which is strongest during austral summer–autumn. This advection weakens farther south because of the weaker zonal currents. Particularly weak currents occur during austral winter as the SEC shifts northward, contributing to the intensification of the subtropical high. The advection is only on the order of 0.2°C month$^{-1}$, compared to the 1.5°C month$^{-1}$ warming–cooling from $Q_{\text{net}}$.

In the Agulhas region, the advection term is associated with the southward flowing Agulhas Current, with
FIG. 9. (a)–(d) Seasonal mixed layer depth and wind stress vectors in the model. (f)–(i) Seasonal mixed layer temperature and current vectors. The corresponding annual-mean fields are shown in (e) and (j). The boxes shown in (e) indicate the locations referred to in Figs. 7 and 8.
a year-round warming effect over the region exhibiting little seasonal variation (Fig. 8c, right; Figs. 9f–j). On the other hand, $Q_{\text{net}}$ cools the region for most of the year (Fig. 8c, left) because of persistent latent, sensible, and longwave radiative cooling (Fig. 8c, middle) that result from the anomalously warm water in that region. The seasonal cycle of $Q_{\text{net}}$ and subsequently $\theta$, however, is almost entirely driven by changes in the shortwave radiation.

### f. Midlatitudes (35°–50°S)

In the midlatitudes, the annual cycle of the mixed layer temperature is also largely controlled by $Q_{\text{net}}$ (Fig. 8d), consistent with observational studies by Sallée et al. (2006) and Dong et al. (2007). The air–sea heat flux variation follows the cycle of seasonal insolation (Fig. 8d). Latent heat loss enhances the net atmospheric cooling primarily in austral winter. The MLD exhibits large variations ranging from 50 m in austral summer to more than 200 m in late winter (Fig. 2h), driven by strong atmospheric cooling coupled with wind stirring (Figs. 9a–d). The deep winter MLD damps the atmospheric cooling, whereas the shallow summer MLD amplifies it. Advection cools the region year round because of the persistent northward Ekman transport of subpolar waters by the midlatitude westerlies. As noted by Dong et al. (2007), entrainment is generally small between 50° and 40°S and has a cooling effect. The model, on the other hand, exhibits a weak entrainment warming ($<0.1^\circ$C month$^{-1}$) during austral winter because of temperature inversions below the mixed layer that are stronger than observed (figure not shown). This error is likely associated with the atmospheric biases described in section 2b. However, this does not appear to significantly affect the simulated heat budget.

### g. Summary

The air–sea heat flux is the dominant component driving the seasonal evolution of mixed layer temperatures over the majority of the Indian Ocean. Advection and entrainment have more localized effects, becoming more important during certain seasons. These terms are particularly important in the tropics, where the seasonality in insolation is relatively weak.

### 4. Analysis of heat flux contributions to temperature variations

#### a. Detecting dominant heat budget terms

Section 3 examined how the seasonal cycle of the IO mixed layer is controlled by air–sea heat fluxes, advection, and entrainment. We now turn our attention to the importance of these components in controlling the rates of warming and cooling on interannual time scales. In this case, the focus is now on the background variations that cause the seasonal cycle to vary from year-to-year. For this purpose, the 100-yr monthly time series of all the model variables making up the budget terms are bandpass filtered in the frequency domain to retain signals with periods of 2–9 yr. The heat budget terms are then reconstructed as in section 2.

The importance of each heat flux term is assessed using a covariance analysis based on the heat budget equation (see appendix). The analysis is applied to each model grid point revealing that the magnitude and spatial–temporal structure of $\theta$, variations are “shaped” by the contribution of each of the covariance terms. The interannual covariance terms are presented in Fig. 10, along with the total covariance terms calculated from the raw data. Positive covariances indicate where the corresponding heat flux terms tend to evolve in phase with $\theta$. Negative values indicate a damping influence by the individual components. The dominance of the terms is considered when their covariance magnitudes are greater than 0.5. Regions where the covariance magnitude of a heat flux component is larger than the covariance magnitude of all other components are indicated in Fig. 10.1

Because the raw data are dominated by seasonal variations, we first demonstrate below that the analysis of the total covariance is consistent with the seasonal overview in section 3.

1. **1) Seasonal time scales**

The total covariance for $Q_{\text{net}}$ is above 0.5 in all regions over the IO Basin, except in the tropical region (Fig. 10b). This is consistent with the seasonal results presented in section 3. In the tropics, variations in both advection and entrainment are required to explain the evolution of $\theta$. In WTIO, as discussed earlier, there is a strong influence of the seasonally varying trade winds and Somali Current. As a result, the covariance is large and positive for advection (Fig. 10c), reflecting the fact that the advection term is in phase with $\theta$ (Fig. 7c). This indicates that advection plays a major role in driving mixed layer thermal variations in the WTIO. A damping effect by advection is evident in various regions where the covariance is negative. This is clear along the band of the trade winds where the year-round heating by Ekman advection peaks in winter when the mixed layer is cooling (Fig. 8a). In the midlatitudes, the negative covariance for advection is to be expected, as the

---

1 For succinctness of the paper, the analysis of the residual term, which may also contain nonphysical estimation errors, is not shown. Nonetheless, it may be noted in Fig. 10 that, where none of the three budget terms dominate, the residual term must dominate.
FIG. 10. Covariance analysis of the heat budget terms (see appendix and section 4a for details) in (left) the raw data and (right) bandpass-filtered data focusing on interannual signals. The covariance for (b),(f) $Q_{\text{net}}$, (c),(g) advection, and (d),(h) entrainment have been normalized by the variance of $u_t$. (a),(e) The standard deviation of $u_t$. The dominance of the terms is considered when the covariance magnitude is $>0.5$. Regions where the covariance magnitude of a heat flux component is larger than the covariance magnitude of the other components (including the residual term) are contoured.
seasonal cycle of advection is out of phase with $\theta$ variations (Fig. 8d).

The substantial entrainment contribution over the SCTR, off Sumatra, and along the coasts of Oman and Somalia is also identified by this analysis (see also Figs. 7b–d, 8a). Note that the entrainment dominance off Sumatra may be exaggerated, as a result of the model bias described in section 2b. The importance of entrainment in the SCTR region might also be located slightly farther east than in the real system. Nonetheless, these results are consistent with, and complement, the overview presented in section 3. The covariance analysis can thus provide a useful comparison of the importance of budget terms, particularly given a large spatial domain and long time series.

2) INTERANNUAL TIME SCALES

On interannual time scales, advection dominates across larger portions of the basin, most notably in WTIO, along the band of the trade winds, in the midlatitudes, and in the Agulhas and Leeuwin Current regions (Fig. 10g). Note that the contribution from advection does not extend to the region off northwestern Australia. Here, the air–sea heat flux term dominates (Fig. 10f). The air–sea heat flux term is also dominant over almost the entire northern IO and over the subtropical region off Madagascar. Contributions from entrainment are significant off Sumatra and over the SCTR (Fig. 10h). Moreover, there is now a significant entrainment contribution along the equator, although this constitutes a damping effect. These features will be further explored in section 5. In the following section, we assess the role of the heat budget components in the genesis of temperature anomalies.

b. Genesis of temperature anomalies

The covariance analysis presented thus far is useful for identifying which terms dominate in controlling the rate of change of the mixed layer temperature. It does not, however, give a direct indication of whether the terms drive the growth of temperature anomalies. A positive temperature tendency, for example, can either correspond to the decay of a cold anomaly or the growth of a warm anomaly (or vice versa for a negative temperature tendency; see illustration in Figs. 11a, b). It is thus necessary to distinguish between these two alternatives to understand how temperature anomalies grow and decay. This may be achieved by employing a temperature variance budget analysis (TVB; see appendix), based on an equation derived in a similar fashion as the density variance budget in Arzel et al. (2007). The TVB equation is derived by multiplying both sides of Eq. (1) by the temperature anomaly ($\theta$) to yield

$$\frac{1}{2} \left( \theta^2 \right) = \frac{\theta \mathcal{O}_{\text{net}}}{\rho_0 C_p h} - \theta \mathbf{u} \cdot \nabla \theta - \frac{\theta w_{\text{entr}}(\theta - \bar{\theta})}{h} + \theta \text{Res},$$

where the overbars denote temporal averaging. Note that the heat budget components are also in the form of anomalies relative to the long-term averages. The lhs is termed temperature variance tendency. A positive
Arabian Sea, it is Q_{net} that tends to generate and sustain anomalies (Figs. 12b,f) while entrainment contributes to the decay of anomalies (Fig. 12h) via summertime upwelling (section 3).

The dominance of entrainment and advection in the WTIO as detected by the covariance analysis is accompanied by positive TVB terms across growth phases (Figs. 12c,d), indicating that advection and entrainment generate temperature anomalies in this region. On the other hand, advection over the south IO, within the band of the trade winds, appears to oppose the growth of temperature anomalies (Fig. 12c). This damping role would be related to the warming by Ekman advection and ITF transport during austral winter cooling (Fig. 8a). The negative TVB term for advection during decay phases in this region (Fig. 12g) reflects the action of warm advection that lasts until spring, dissipating the cold winter anomaly. Entrainment, which was previously shown to dominate the heat budget in this region (Fig. 10d), actually dissipates rather than generating temperature anomalies (Fig. 12h). This is primarily associated with the decay of warm summertime anomalies by cold entrainment that peaks in autumn (Fig. 8a).

The advection term contributes to the growth and persistence of anomalies in the subtropics (Figs. 12c,g). However, this contribution is relatively small, which is expected given that the advection term constitutes only a small component of the heat budget in this region as determined by the covariance analysis (Fig. 10c; see also Fig. A2).

2) Interannual Time Scales

The picture changes considerably on interannual time scales. Now, the role of advection in generating temperature anomalies covers extensive areas of the IO Basin (Fig. 13c). The similarity between the TVB and covariance patterns for advection (Figs. 13c, 10g) suggests that where ocean advection is important, it also tends to drive the growth of the temperature anomalies. Advection also tends to increase the persistence of anomalies as indicated by the positive TVB term during the decay phases (Fig. 13g). On the other hand, the control of Q_{net} on the growth and decay of thermal anomalies becomes localized over the northern IO regions and southern subtropical regions, including off northwestern Australia (Figs. 13b,f). Over the rest of the domain, particularly in the tropics, Q_{net} acts primarily as a damping and decay term (Figs. 10f, 13b,f). The contribution of entrainment to the growth and persistence of anomalies off Sumatra extends to interannual time scales (Figs. 13d,h). These roles also extend over the SCTR, where entrainment was previously found to dissipate seasonal anomalies (Fig. 12h).

5. Characteristics of the interannual heat budget

Based on the analyses presented in section 4, we can identify certain regions where particular heat flux terms are crucial for driving interannual variability of the mixed layer temperature (Fig. 14). These regions are defined based on both 1) a dominant covariance term (i.e., a
FIG. 12. TVB terms in the raw data for (left) temperature anomaly growth phase: composites over instances when the variance tendencies are positive indicating growth of anomalies (region G in Fig. 11); and (right) decay phase: composites based on negative variance tendencies (region D in Fig. 11). The terms presented are (a),(e) temperature variance tendency, TVB terms of (b),(f) $Q_{\text{net}}$, (c),(g) advection, and (d),(h) entrainment [see Eq. (5) and text for details]. Gridpoint terms that are not significantly different from zero, as determined by a two-tailed $t$ test at the 99% level, are shown in white.
Fig. 13. As in Fig. 12, but for the bandpass-filtered data focusing on interannual signals.
particular heat budget component explains most of the variance in the temperature evolution; see section 4a) and 2) a positive TVB term during anomaly growth phases (i.e., a particular heat budget component contributes to the growth of temperature anomalies; see section 4b). The white areas in Fig. 14 indicate where two or more heat flux components are of similar magnitude, and thus depict regions where complex interactions between multiple components are driving temperature variations. Regions where the residual term appears important may indicate the control by diffusive processes, or regions where numerical errors, associated with the computation of the heat budget terms, are large. We limit our discussion to those regions dominated solely by air–sea heat fluxes, advection, or entrainment. We demonstrate later that these regions can be categorized in terms of certain basic physical characteristics.

Most of the locations where entrainment dominates occur within 10°S–10°N (Fig. 14). As noted in previous studies (e.g., Rao and Behera 2005), this latitude band is known for significant subsurface influence on SST, and it is associated with a relatively shallow thermocline [see Fig. 15a for the model 20°C isotherm depth (D20)]. Figure 15b compares the distribution of D20 at locations where entrainment is important and where $Q_{\text{net}}$ dominates. The two are significantly different above the 95% confidence level, demonstrating that entrainment clearly tends to be dominant in regions of shallow thermocline.

On the other hand, where the thermocline is deep (e.g., off northwestern Australia), the air–sea heat flux tends to be dominant. This relationship is further supported by significant positive correlations between D20 and SST in the entrainment regions, compared to little systematic correlations in $Q_{\text{net}}$ regions (Fig. 15c).
One striking feature from the covariance and TVB analyses is the shift in the importance of advection in moving from seasonal-to-interannual time scales (Figs. 10c,g, 12c, 13c). The significance of advection is most apparent in regions associated with strong wind stress (i.e., in the trade wind region and the midlatitudes) and boundary currents (the Leeuwin Current, WTIO, and the Agulhas regions). In these regions, \( Q_{\text{net}} \) acts to damp the growth of advection-driven temperature anomalies (Fig. 13b) via changes to the latent heat flux. Indeed, there are significant negative correlations between SST and evaporation in advection regions, that is, surface warming corresponds to increased evaporation. These negative correlations imply that evaporation is driven by SST. If it were the converse, we would expect a positive correlation—that is, surface warming corresponding to suppressed evaporation. Figure 16a shows that the distribution of 0-lag correlations between SST and latent heat flux averaged over 1–6-month lag times with latent heat leading SST.

\[ \frac{Q}{\rho_0 C_p h} = \frac{\overline{Q}}{\rho_0 C_p h} + \frac{\delta Q}{\rho_0 C_p h}. \]  

(6)

The first term on the rhs represents the effect of MLD variations acting on the long-term-averaged net air–sea heat flux. The second term accounts for the \( Q \) anomalies that are also modulated by the MLD variations. The contribution of the first and second terms to \( \theta_t \) variability can again be assessed using the covariance analysis (Figs. 17b,a, respectively). Note that the two covariance terms represent the decomposition of the total \( Q_{\text{net}} \) covariance previously shown in Fig. 10f. By comparing Fig. 10f and Figs. 17a and 17b, it is apparent that the negative covariance term in the Agulhas region (Fig. 10f), for instance, is largely due to MLD variability. Other regions where MLD variations can be important include coastal regions of the northern IO, the area affected by the trade winds, and off southwestern Australia. It is worth noting that in general the MLD effect tends to be substantial where the mixed layer is shallow and the magnitude of the net air–sea heat flux is relatively large (Fig. 18). In other regions, the second term of Eq. (6) dominates, which is largely because of the \( Q \) anomalies, with only negligible modulation by the MLD. This can be demonstrated by computing the covariance of the second term with MLD held fixed at its long-term mean value (Fig. 17a, contoured)—there are minimal differences from the original covariance. Furthermore, over most parts of the IO, it is the latent heat component that dominates the effect of air–sea heat fluxes on \( \theta_t \) (Fig. 17c).

Dong et al. (2007) showed that MLD variations are important over the midlatitudes on seasonal time scales. In particular, they noted that the seasonal variations in the advection term disappear when the temporal

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**Fig. 16.** (a) Distribution of SST and latent heat flux correlation at 0-lag separated over advection-dominated regions (gray-shaded histogram) and \( Q_{\text{net}} \)-dominated regions (black-contoured histogram) as defined in Fig. 14. (b) As in (a), but for the correlation of SST and latent heat flux averaged over 1–6-month lag times with latent heat leading SST.
variations in MLD were neglected in the calculation. They note that this is because the Southern Ocean has strong seasonal variations in MLD but weak wind stress seasonality (see also Fig. 9). We conducted a similar calculation with time-averaged MLD. On seasonal time scales, the difference in the covariance for advection with a fixed MLD from that using a time-varying MLD (Fig. 19a) is most pronounced in the midlatitudes. Holding MLD constant indeed removes the advection variability (not shown), which otherwise plays a role in damping $\theta_t$ (Fig. 10c). On interannual time scales, the most pronounced differences occur along the band of the trade winds, rather than in the midlatitudes (Fig. 19b). This is because the Southern Ocean advection variation is more strongly controlled by the interannual variability in current velocities and temperature gradients. This analysis suggests that MLD variations are more important over the trade wind regime on interannual time scales and are an important consideration when conducting interannual surface heat budget analyses.

Another interesting point to note is the dominance of non-Ekman advection in the midlatitudes on interannual time scales (Fig. 19d) despite its relatively small mean value. For example, the mean value of non-Ekman advection heat budget terms averaged over the region $37^\circ$–$50^\circ$S, $53^\circ$–$109^\circ$E is $-0.3 \times 10^{-7}$C s$^{-1}$, which is an order of magnitude smaller than the averaged Ekman advection terms of $-2.0 \times 10^{-7}$C s$^{-1}$. As noted by Dong et al. (2007), the small magnitude is because the geostrophic velocities follow the contours of constant sea surface height, which are generally parallel to the contours of constant SST. As a result, there is the tendency for the zonal and meridional components to cancel each other out. However, we find that the variability in the advection components is comparable in magnitude. Averaged over the same region defined earlier, the standard deviation of the non-Ekman advection is $1.0 \times 10^{-8}$C s$^{-1}$, which is comparable to the Ekman term standard deviation of $0.8 \times 10^{-8}$C s$^{-1}$. The non-Ekman advection is particularly large, however, in the Agulhas Current and

![Figure 17](image.png)

Fig. 17. The decomposition of the covariance of $Q_{\text{net}}$ with $\theta_t$ into (a) covariance calculated with $Q$ expressed as anomalies and (b) covariance with long-term averaged $Q$ accounting for purely MLD variations [see Eq. (6)]. The contours overlaid in (a) indicate a similar covariance analysis but with MLD held to its long-term mean. (c) Covariance of latent heat flux anomalies with $\theta_t$, showing a close resemblance to (a). (d) Scatterplot of the covariance shown in (a) against covariances for its $Q$ components, shown only at $Q_{\text{net}}$-dominated regions (as defined in Fig. 14). The lines in (d) are best-fit lines by the least squares method on the corresponding scatter points. The dashed blue line indicates the line of $y = x$. Note that the latent heat flux component is larger than the other components, indicating a dominant air–sea heat flux component in controlling $\theta_t$ variations.
Retroflection, both in terms of the mean and variability. Considering the Agulhas Retroflection carries a significant amount of heat into the Southern Ocean (Sun and Watts 2002), interannual variability sourced from geostrophic advection is also likely to be important in this region. Ekman advection does, however, control the thermal variations along the band of the trade winds (Fig. 19c).

The contribution of the heat budget components described earlier sets the overall pattern of IO SST variability (Fig. 3a). If the model were to be driven by ocean dynamics alone—for example, by fixing $Q_{\text{net}}$ to climatological values—then the structure of the interannual variability would essentially match the combined pattern of the advection and entrainment contribution (i.e., Figs. 10g,h). By forcing an OGCM with solar radiation set to climatology and fixed latent heat loss but allowing wind stress to vary interannually, Murtugudde and Busalacchi (1999) found an SST standard deviation pattern stretching from the WTIO to the southeast IO, which is similar to our advection covariance pattern (Fig. 10g; see Murtugudde and Busalacchi 1999, their Figs. 11 and 16). Obvious reductions in their SST standard deviation occur in regions where we have identified $Q_{\text{net}}$ to be dominant—namely, in the Arabian Sea, the Bay of Bengal, and off northwestern Australia.

Finally, note that the roles and dominance of the heat flux terms in the equatorial regions (Figs. 13, 14) govern the genesis of the IOD. Although these may be somewhat enhanced in the model, because of the spurious thermocline shoaling that leads to the overly strong

![Fig. 18. Scatterplot of MLD against $Q$ magnitude at locations where the MLD effect of Eq. (6) is substantial (dark markers) vs where the MLD effect is small (light markers), as determined by the covariance magnitude in Fig. 17b. The large and small threshold is defined as covariance magnitude $>0.5$ and $<0.5$, respectively. The mean values for the case of substantial and nonsubstantial MLD effect are indicated by the corresponding circled dots.](image1)

![Fig. 19. The difference between the covariance for advection calculated using long-term averaged MLD and that with varying MLD in (a) the raw data and (b) the bandpass-filtered data retaining interannual signals. Covariance analysis of (c) Ekman advection and (d) non-Ekman advection in the filtered data.](image2)
tropical variability (section 2b), they are generally consistent with previous studies. For example, using reanalysis products Du et al. (2008) found that the generation of SST anomalies off Sumatra during IOD events is mainly via entrainment when southeasterly wind anomalies drive anomalous upwelling. Net air–sea heat flux was found to be the primary damping term (see also Tokinaga and Tanimoto 2004). In the WTIO, ocean dynamics are known to be the dominant mechanism driving SST anomalies, whereas net air–sea heat flux again plays a damping role (Webster et al. 1999; Li et al. 2002; Tokinaga and Tanimoto 2004). The role of advection on the persistence of SST anomalies in this region, and air–sea heat fluxes on the decay of anomalies in the east, provide a mechanism for an IOD event to evolve into a basin-wide mode (Tokinaga and Tanimoto 2004). Changes in atmospheric circulation associated with the IOD and/or ENSO force westward-propagating off-equatorial Rossby waves (Xie et al. 2002; Feng and Meyers 2003; Rao and Behera 2005). These also affect SCTR, where SST anomalies are generated via anomalous local Ekman pumping and Rossby waves. As shown in Fig. 10h, entrainment is dominant in this region. The ENSO influence on IO is largely via air–sea heat fluxes, except over the SCTR (e.g., Klein et al. 1999; Du et al. 2009). This coincides with regions indicated by the air–sea heat flux dominance in Fig. 14.

6. Summary

We have determined the relative importance and role of air–sea heat fluxes, advection, and entrainment on the genesis of mixed layer temperature anomalies across IO from seasonal–to-interannual time scales using a coupled climate model. The use of a model is still essential to achieve sufficient spatial and temporal resolution and to extract a suitably long time series to investigate interannual variability over the IO. The seasonal climatology of the model heat budget components is presented and is in good qualitative agreement with a reconstructed climatology based on available reanalysis and gridded datasets. Despite biases in both the model and the observational products, the similarity between the seasonal cycle of the heat budget terms over many regions in the model and observations, and consistency with previous studies, suggest that the model analysis can be usefully extended to interannual time scales.

To detect the dominant heat budget terms that drive the variations of the heat balance, a covariance analysis is applied on the heat budget equation [Eq. (1)]. To reveal their role in the generation and decay of temperature anomalies, a TVB analysis is used (see section 4b and the appendix). Based on these two analyses, regions where certain heat budget terms are identified to be important for the generation of interannual temperature anomalies are summarized in Fig. 14. Model biases do, however, exist and may distort the spatial extent or location of the identified regions. An interesting future extension of this work would be to apply our analysis to a whole suite of multicentury runs of climate models to further examine how model biases could influence the role and dominance of the heat fluxes.

The primary results from our study can be summarized as follows:

1) The air–sea heat flux controls the seasonal thermal variations over most regions of the IO except in the tropics. On interannual time scales, the growth and decay of temperature anomalies in the Arabian Sea, the Bay of Bengal, and southern subtropical regions including off northwestern Australia are primarily forced by air–sea heat flux variations. Over the rest of the domain, particularly in the tropics, air–sea heat fluxes act as a damping and decay term. Over most of the domain, it is the latent heat flux component that is found to dominate the air–sea heat flux for interannual variations. In the Agulhas region and off southwestern Australia, however, the modulating effect of MLD variations appears to be substantial.

2) On seasonal time scales, ocean advection is dominant only in WTIO, where it can generate temperature anomalies. On interannual time scales, advection becomes dominant over many regions. These include WTIO, the trade winds region, in the Leeuwin Current, the Agulhas Current and its retroflection, and in the midlatitudes. Non-Ekman advection is found to be important in the midlatitudes and the boundary current regions. On the other hand, Ekman advection is more important in areas associated with the trade winds. In these regions, advection generates temperature anomalies and also tends to increase their persistence. Furthermore, the SST variations were shown to drive variations in the latent heat flux, demonstrating the potential for ocean heat fluxes to drive atmospheric variability in these regions on interannual time scales. It is found that MLD variations are particularly important in modulating advection in the midlatitudes and on seasonal time scales, whereas they are more important for the trade winds region on interannual time scales.

3) Driven by the monsoon winds, the seasonal evolution of entrainment is found to be important over the Seychelles–Chagos Thermocline Ridge in the south tropical IO, off Sumatra, and along the coasts of Somalia and Oman. The importance of entrainment in these regions extends to interannual time scales.
contributing to the generation and persistence of thermal anomalies there. Entrainment tends to be important where the thermocline depth is relatively shallow, whereas regions with a deeper thermocline tend to be dominated by air–sea heat fluxes.

Acknowledgments. The authors gratefully thank Olivier Arzel for suggesting the implementation of the temperature variance budget equation. Comments and suggestions by three anonymous reviewers helped improve the paper. The NCEP–NCAR reanalysis datasets are provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, available from their Web site at http://www.cdc.noaa.gov/. The mixed layer depth climatology was provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado. The altimeter products were produced by Ssalto/Altimeter home page at http://www.marine.csiro.au/~dunn/cars2006/. The CARS06 dataset is made available at http://www.marine.csiro.au/~dunn/cars2006/.

APPENDIX

Heat Budget Analysis

a. Covariance analysis

Given a \( \theta_t \) time series, and by the covariance property, the variance of \( \theta_t \) can be expressed in terms of the covariance \( \gamma \) of \( \theta_t \) with each budget term:

\[
\sigma^2(\theta_t) = \gamma(\theta_t, Q) + \gamma(\theta_t, \text{Adv}) + \gamma(\theta_t, \text{Ent}) + \gamma(\theta_t, \text{Res}).
\]  
(A1)

Because the correlation of \( x \) and \( y \), \( \rho(x, y) = \gamma(x, y)/\sqrt{\sigma(x)\sigma(y)} \), (A1) can be written as

\[
\sigma(\theta_t) = \rho(\theta_t, Q)\sigma(Q) + \rho(\theta_t, \text{Adv})\sigma(\text{Adv}) + \rho(\theta_t, \text{Ent})\sigma(\text{Ent}) + \rho(\theta_t, \text{Res})\sigma(\text{Res}).
\]  
(A2)

Equation (A2) reveals how each heat flux term contributes to the variability of \( \theta_t \). In our analysis, the terms on the rhs of (A1) and (A2) are normalized such that their sum is 1. Each of these normalized covariances provides a measure of the relative importance of the term in the heat budget. It may be noted that these normalized covariances are equivalent to regression coefficients obtained from regressing the heat flux term onto \( \theta_t \). For brevity, and to highlight the previously defined relations, we refer to the normalized covariance simply as the covariance.

b. TVB analysis

When the sum of the heat budget terms is nonzero, the temperature anomaly \( \dot{\theta} \) will grow or decay. This is illustrated in Fig. A1, which shows a portion of the interannual time series for all the relevant variables for the WTI0 region. Here, \( \dot{\theta} \) and \( \theta \) are by definition in quadrature: they are precisely in phase and out of phase during growth and decay phases of \( \dot{\theta} \), respectively (Figs. A1a,b; see also Figs. 11a,b). Thus, the growth or decay phase of \( \dot{\theta} \) at a given time is marked by the sign of the covariability between \( \theta \) and its time derivative \( \dot{\theta} \) at that instant (i.e., a growth phase when \( \dot{\theta} \theta > 0 \) and a decay phase when \( \dot{\theta} \theta < 0 \); see Figs. A1a,b). This implies that the heat budget terms can have a different impact on the growth and decay of \( \theta \), depending on whether their anomalies are in phase or out of phase with \( \dot{\theta} \), during periods when \( \theta \) is either growing or undergoing decay (Fig. A1b). However, \( \theta \dot{\theta} \) is equivalent to the temperature variance tendency \( 0.5(\dot{\theta}^2) \); lhs of the TVB equation, Eq. (5)]. Therefore, the role of each budget term on the growth and decay of \( \theta \) can be equivalently inferred by the sign of their corresponding covariability with \( \dot{\theta} \) [rhs of Eq. (5); see Figs. A1b,c]. The TVB equation as derived from Eq. (1) provides the physical framework for conducting further statistical analyses as discussed later.

To obtain a generalization over a given time series, the budget terms are often regressed onto \( \dot{\theta} \). This actually involves taking the time average in Eq. (5) over all phases of \( \dot{\theta} \) growth and decay. In this study, however, the time averaging is applied over growth and decay phases separately. This approach allows one to determine whether a certain budget term participates in the actual generation of \( \dot{\theta} \) or in the increase of its persistence, and similarly, in the decay of \( \dot{\theta} \) or in limiting its growth (see also section 4b). Clearly, these physically distinct roles are “averaged out” in the conventional regression approach.

This becomes problematic when the controlling heat budget term is nonlinearly related to \( \theta \). One clear example is the seasonal heat budget over the subtropics, where \( \theta_t \) is entirely due to \( Q_{\text{net}} \) as illustrated in Fig. A2 (see also Fig. 10b). Because \( \dot{\theta} \) and its time derivative are in quadrature, the regression of \( Q_{\text{net}} \) on \( \dot{\theta} \) is zero (i.e., the covariability term cancels out when averaging across all phases; Fig. A2c). This leads to an erroneous conclusion. Alternatively, by considering the growth and decay phases, separately, the underlying physical process is revealed—that is, the growth and decay of \( \dot{\theta} \) is driven by the seasonal variation of \( Q_{\text{net}} \) (see Figs. 12b,f, A2c). Another example of such a nonlinear case would be the control of \( Q_{\text{net}} \) on the interannual heat budget over the northern IO region (Figs. 10f, 13b,f).
It should be emphasized that the TVB analysis described earlier is based on the temperature time series governed deterministically by the respective heat budget terms via Eq. (1). Thus, the covariability, $\theta_t \theta$, instantaneously marks the actual growth and decay phases of $\theta$ as described previously, allowing examination of the impact of the heat budget terms on the genesis of $\theta$. However, statistical significance should still be computed for the covariability of the budget terms to generalize their role over the whole time series.
Figure A3 illustrates how the heat budget covariance analysis described previously is related to the TVB analysis. Figure A3 shows covariance and correlations plotted in the TVB space, based on a periodic $\theta_t$ and a periodic budget term ($Q$) with varying phase and standard deviation. Note how the correlations between $Q$ and $\theta_t$, and between $Q$ and $\bar{\theta}$, are in quadrature in the TVB space (Figs. A3b,c). As determined by the correlation between $Q$ and $\theta_t$, the covariance between $Q$ and $\theta_t$ increases in magnitude toward quadrant I and III (Fig. A3a). It is in these quadrants that care should be taken when interpreting results using the conventional regression approach, as noted previously. Finally, note that the covariance magnitude in Fig. A3a increases with the variability magnitude of the budget term. This allows a heat budget term to be meaningfully detected as important via its large variability magnitude, despite a small correlation with $\theta_t$ [an implication from Eq. (A2)]. For example, advection in WTIO is an important term because of its large standard deviation (Fig. A1b). Its correlation with $\theta_t$ is rather weak, as it is generally in phase and often out of phase with $\theta_t$ over the growth and decay phases, respectively. However, its correlation with $\bar{\theta}$ is highly positive, as it is mostly in phase with $\bar{\theta}$ over the decay phases (Fig. A1c). Thus, its effect is expected to lie in quadrant II, where its TVB terms are positive.

**FIG. A2.** As in Fig. A1, but for seasonal anomalies in the subtropical region (32°–15°S, 53°–92°E). The seasonal heat budget in this region is almost exclusively driven by $Q_{net}$ variations. The correlation between $Q_{net}$ and $\theta_t$ is close to unity. Thus, $Q_{net}$ is in quadrature with $\bar{\theta}$, resulting in positive and negative covariability during growth and decay phases, respectively. This indicates that the growth and decay of $\bar{\theta}$ are driven by variations in $Q_{net}$.
for both growth and decay phases (as in Figs. A3c, A1c). This can be seen in Fig. A4, which shows the covariance for advection at each grid point in the WTIO region presented in the TVB space. Other examples are also shown in Fig. A4.

It is noted that dominant covariance terms do not always translate to dominant TVB terms of either sign (figure not shown). Thus, in this study, the two analyses are used in combination to meaningfully extract the drivers of thermal anomalies in the surface mixed layer.

REFERENCES


