An Observation-Based Assessment of Nonlinear Feedback Processes Associated with the Indian Ocean Dipole

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ABSTRACT

A well-known feature of the Indian Ocean dipole (IOD) is its positive skewness, with cold sea surface temperature (SST) anomalies over the east pole (IODE) exhibiting a larger amplitude than warm SST anomalies. Several mechanisms have been proposed for this asymmetry, but because of a lack of observations the role of various processes remains contentious. Using Argo profiles and other newly available data, the authors provide an observation-based assessment of the IOD skewness. First, the role of a nonlinear dynamical heating process is reaffirmed, which reinforces IODE cold anomalies but damps IODE warm anomalies. This reinforcing effect is greater than the damping effect, further contributing to the skewness. Second, the existence of a thermocline–temperature feedback asymmetry, whereby IODE cold anomalies induced by a shoaling thermocline are greater than warm anomalies associated with a deepening thermocline, is the primary forcing of the IOD skewness. This thermocline–temperature feedback asymmetry is a part of the nonlinear Bjerknes-like positive feedback loop involving winds, SST, and the thermocline, all displaying a consistent asymmetry with a stronger response when IODE SST is anomalously cold. The asymmetry is enhanced by a nonlinear barrier layer response, with a greater thinning associated with IODE cold anomalies than a thickening associated with IODE warm anomalies. Finally, in response to IODE cool anomalies, rainfall and evaporative heat loss diminish and incoming shortwave radiation increases, which results in damping the cool SST anomalies. The damping increases with IODE cold anomalies. Thus, the IOD skewness is generated in spite of a greater damping effect of the SST–cloud–radiation feedback process.

1. Introduction

The recent unprecedented high frequency of positive Indian Ocean dipole (pIOD) occurrences, including the three consecutive pIOD episodes during 2006–08 (Luo et al. 2008; Cai et al. 2009c), continues to highlight the associated impacts of tropical Indian Ocean variability, ranging from drought in Indonesia and Australia (Ashok et al. 2003; Cai et al. 2005, 2009b; Ummenhofer et al. 2009), conditioning of the southeast region of Australia for major summer wildfires (Cai et al. 2009a), and floods in eastern Africa (Black et al. 2003; Lareef et al. 2003). Through its coherence with the El Niño–Southern Oscillation (ENSO) in the austral spring season, the Indian Ocean dipole (IOD) is found to be a major pathway for delivering ENSO’s impact (Cai et al. 2011). ENSO reinforces tropical Indian Ocean SST anomalies and, in addition to a Gill-type response in the tropics, the associated adiabatic heating anomalies trigger equivalent barotropic Rossby wave trains (Hoskins and Karoly 1981), transmitting the IOD’s influence into midlatitude regions such as Australia. Furthermore, the impact associated with a pIOD is greater than that associated with a negative IOD (nIOD), underlined by the IOD’s skewness (Cai et al. 2012).

However, the cause of the IOD skewness remains a contentious issue. Possible mechanisms have previously been postulated using model outputs and reanalyses. These include a nonlinear dynamic heating (NDH) process (Hong et al. 2008), an asymmetric SST–cloud–radiation feedback (Hong and Li 2010), and an asymmetric thermocline–temperature feedback (Zheng et al. 2008).
During pIOD events, nonlinear zonal and vertical advection reinforces IODE cold SST anomalies but damps IODE warm SST anomalies. Despite a consensus on the role of the NDH feedback, the role from the other two feedbacks remains controversial.

The mechanism of a breakdown of the SST–cloud–radiation feedback is based on the idea that cold SST anomalies lead to a reduction in cloud cover that causes an increase in net downward shortwave radiation, damping the cold SST anomalies (Ramanathan and Collins 1991; Li et al. 2000). This feedback (damping) operates in an unhindered manner, with warm IODE SST anomalies leading to an increase in convection and cloud amount, which causes a decrease in the shortwave radiation reaching the surface. However, cold IODE SST anomalies, upon reaching a threshold value, may completely suppress the mean convection resulting in cloud-free conditions, at which point further cooling is not subject to any damping (Hong et al. 2008). Thereafter, cool SST anomalies are able to grow “freely” because of the lack of thermal damping, and the IOD skewness is generated.

Using observations of outgoing longwave radiation (OLR) as a measure of convective activity, it is recently shown that the SST–cloud–radiation feedback actually strengthens as cold IODE SST anomalies develop, with a greater response of OLR (Cai et al. 2012). The results suggest that the strong cold SST anomalies develop, not because of the stronger damping, but in spite of it.

The asymmetric thermocline–temperature feedback operates in the following manner: as the climatological mean thermocline in the IODE region is deep, a shoaling thermocline is more effective in inducing a surface cooling than a deepening thermocline is in inducing a surface warming (Zheng et al. 2010). For the same one unit of change of the thermocline, the size of surface cooling is far greater when the thermocline is shoaling than the size of warming induced by a deepening thermocline. Zheng et al. (2010) suggest that the asymmetric thermocline–SST feedback is a key process responsible for the negative SST skewness in the IODE. However, Hong and Li (2010) argue that the asymmetric thermocline–SST feedback can be accounted for by the breakdown of the SST–cloud–radiation feedback. Using a reanalysis temperature product, Hong and Li (2010) show that there is little asymmetry in the thermocline–subsurface temperature feedback or in the response of equatorial easterly winds to the thermocline. Cai et al. (2012) show that these features reported Hong and Li (2010) are a consequence of biases in the reanalysis product used in the their study.

With an unprecedented collection of temperature profiles from Argo floats and other concurrent observations such as winds and oceanic currents over the past decade, it is now possible to provide an observational description of these feedbacks and to assess their relative importance. This is the focus of the present study. Already, Argo profiles have been used to describe the evolution of the three consecutive 2006–08 pIOD occurrences (Cai et al. 2009e) and interannual variability of the barrier layer (BL) over the IODE (Qiu et al. 2012). The BL has been found to covary with the IOD, with a thinning during pIOD events enhancing the positive feedback (Masson et al. 2004). The extent to which the covarying BL contributes to the feedback asymmetry has not been explored by previous studies and will also be examined here. We confirm the role of the NDH, describe the asymmetry of the thermocline–temperature feedback, and assess the role of the SST–cloud–radiation feedback.

2. Data and method

Monthly 1° × 1° gridded temperature anomalies, with a vertical resolution ranging from 7.5 to 10 m for the upper 150 m and derived from quality controlled Argo profiles (Roemmich and Gilson 2009), are used to generate thermocline depth, defined as the depth of 20°C isotherm (Z20). Over the eastern Indian Ocean, while the initial Argo float coverage was not as dense as in the Pacific (e.g., during 2000–04), since 2004 the density of the float coverage and the number of the bin counts of the temperature and salinity profiles have been at least at the global average, if not above. By mid-2005 it had reached approximately the 75% level of the planned density of coverage at an average of 3° latitude–longitude spacing, which was achieved by 2007. It is for this reason that Roemmich and Gilson (2009) constructed the dataset from 2004 onward. We use the dipole mode index (DMI) as defined by Saji et al. (1999). During this Argo period 2004–11, there are pIOD and nIOD events, but because the number of events is limited we use monthly data, from May to November, to increase our sample size.

We use surface wind data from QuikSCAT (Liu et al. 2000) covering the period of 2004–09 to examine the Bjerknes feedback, and oceanic surface currents from monthly Ocean Surface Current Analysis—Real Time (OSCAR) currents (Bonjean and Lagerloef 2002) to describe the NDH process. To calculate gridpoint NDH, vertical velocity is required and is inferred from the continuity equation using the OSCAR current data. Precipitation data from the Tropical Rainfall Measuring Mission (TRMM) 3B43-V6 product (Huffman et al. 2007) and individual heat flux components from the Objective Analysis Flux (OAFlux) (Yu and Weller 2007) are also used. Monthly anomalies are computed by subtracting the corresponding monthly climatology over the
The basic analysis technique used in this study is linear regression (or, equivalently, correlation) using all samples and regression coefficients using samples containing only positive or negative index values, separately, to assess possible asymmetry.

As in Qiu et al. (2012), the BL thickness is defined as the difference between the depths of the isothermal layer (IL) and the mixed layer (ML). Depths of the ML and IL are calculated from individual profiles and then interpolated onto a regular grid of $1^\circ$ latitude $\times 1^\circ$ longitude for each month using the kriging algorithm (Oliver and Webster 1990). Following Bosc et al. (2009), the IL depth is defined as the depth where temperature is $0.2^\circ$C lower than that at the 10-m depth. The ML depth is defined in terms of a depth with a density equal to that at the 10-m depth plus an increment in density equivalent to a $0.2^\circ$C cooling.

We point out that the monthly samples are not independent, but given that we use various observed datasets that cover a varying length, we highlight contours of correlation coefficients that would be statistically significant above the 95% confidence level, if independent samples are assumed. This facilitates a comparison of coherence on an equal basis when different observed fields with a varying length are used.

### 3. The asymmetric role of NDH

Over the IODE region the SST skewness is highly negative, reaching a value of $-1^\circ$C, confirming the presence of skewness over the period. The intrinsic asymmetric nature of the NDH has been shown to contribute to the skewness of ENSO (Jin et al. 2003; An and Jin 2004) and the IOD (Li et al. 2003; Hong et al. 2008). The effect may be described as

$$NDH = (-U) \frac{\partial T'}{\partial x} + (-W) \frac{\partial T'}{\partial z}.$$
where $T'$, $U'$, and $W'$ represent the anomalous temperature, zonal current, and vertical current anomalies (positive eastward and upward), respectively. During a pIOD or nIOD event, an anomalous zonal and vertical temperature gradient leads to anomalous cold advection in the IODE region, reinforcing the cool SST anomalies that occur during a pIOD and damping the warm SST anomalies during nIOD events. Our

![Diagram](image1)

**Fig. 2.** Scatter diagram of average (a) zonal current and (b) zonal temperature gradient anomalies in the region of 90°–100°E, 7.5°–2.5°S vs IODE SST, normalized by their respective standard deviations.

![Diagram](image2)

**Fig. 3.** (a) Point-to-point correlation between SST and $Z20$ anomalies using all samples assuming linearity. Coefficients in areas confined by the black contours are statistically significant above the 95% confidence level. Scatter diagram of anomalies of (b) SST and (c) temperature at 75-m depth ($T75$) vs $Z20$, all averaged over the region of 90°–100°E, 7.5°–2.5°S, normalized by their respective standard deviations.
assessment focuses on the upper 50 m, a depth within which oceanic currents are taken as represented by the near ocean current data.

The correlation of gridpoint NDH with the DMI is generally weak across the tropical Indian Ocean (Fig. 1a) because of the known nonlinearity. A scatterplot of NDH values versus SST anomalies averaged within the IODE region (Fig. 1b) or NDH values against the DMI (Fig. 1c) shows that, for cold IODE SST anomalies (i.e., a positive DMI value), the NDH tends to be negative (although some weak positive values are present), reinforcing the cold anomalies. Examining the relative contribution of the vertical and zonal advection, we find that it is mostly dominated by the zonal component (not shown). During nIOD, most NDH values are negative, representing damping, although the damping increases only weakly with SST anomalies. Thus, the reinforcing effect during pIOD is far greater than the damping effect during nIOD, as indicated by a far greater slope during pIOD. Despite this, the central point is that there is a strong asymmetry, which will enhance the skewness.

The larger amplitude of the stronger reinforcing effect is directly linked to an asymmetric response of dynamical fields to SST anomalies. For example, zonal current anomalies within the IODE region are far greater and more responsive to SST (i.e., with a greater slope) when IODE SST anomalies are negative than when IODE SST anomalies are positive (Fig. 2a). This feature is also seen in the associated zonal SST gradient (Fig. 2b). These asymmetric features are a part of the asymmetric processes of the Bjerknes-like feedback, which we will discuss below.

4. Asymmetry in the Bjerknes positive feedback loop

a. Thermocline–temperature feedback

We first examine centers of known feedback processes in terms of a map of correlation between gridpoint anomalies and a key index using the monthly samples. An identification of centers of action is necessary because the center of IOD feedback action during the
short period with a limited number of events could be
different from that identified using data over a longer
period. It turns out the identified centers are located
mostly within the IODE region, although it encompasses
a smaller area. We use the region of highest correlation to
illustrate the process. This correlation analysis is not to
establish whether a process is at work, because these pro-
cesses have been established by numerous previous stud-
ies. Here, our focus is on the asymmetry of these processes.

Figure 3a plots point-to-point correlations of Z20 with
SST anomalies using all samples under a linear assump-
tion. A thermocline–temperature feedback operates,
particularly in the IODE region, where a shoaling ther-
mcline is associated with cold SST anomalies, and vice
versa. A scatterplot of SST anomalies versus Z20 anom-
alies within the IODE region brings out the asymmetry of
the feedback: a shoaling thermocline is more effective in
inducing a surface cooling than a deepening thermocline
that causes a surface warming. For the same one unit of
change in the thermocline, the size of surface cooling is
far greater when the thermocline is shoaling than the size
of warming when the thermocline is deepening. The slope
ratio is 0.6:0.03 (Fig. 3b).

The importance of the thermocline feedback in gen-
erating the skewness is highlighted by its greater
influence on temperature at depth than at the surface.
Variations of Z20 are more coherent with temperatures
at the 75-m depth, regardless of whether the thermocline
is deepening or shallowing (cf. Figs. 3b,c). Two features
emerge: First, the influence of a deepening thermocline
on SST is not statistically significant, in contrast to the in-
fluence on temperature at the 75-m depth. Second, when
the thermocline is shoaling, the influence is statistically
significant and is greater at depth than at the surface. These
features highlight that the source of the surface asymmetry
originates from the subsurface, that is, the thermocline.
This is a feature that is absent in the reanalysis used by
Hong and Li (2010).

b. Response of wind to SST and thermocline
to wind

There is a consistent asymmetry in the primary element
of the Bjerknes feedback, the response of anomalous
winds to underlying SST anomalies. Using all samples,
the maximum response lies in the equatorial eastern Indian
Ocean, not coinciding with the IODE region (Fig. 4a). This
reflects the fact that IODE Z20 anomalies, despite their
localized impact on SST, are induced by wind-driven
equatorial Kelvin waves that propagate into the region
(Fig. 4c). For the same one unit of change in IODE SST
anomalies, the size of anomalous easterlies is far greater when the IODE region is cooling than the size of anomalous westerlies when the region is warming (Fig. 4b). This asymmetry is also manifested when replacing IODE SST with DMI (not shown). Further, because less energy is required to lift the thermocline when it is shallow, for the same one unit of change of zonal wind, the uplift induced by easterlies is greater than the deepening induced by westerlies (Fig. 4d).

Thus, an asymmetry operates in the IODE region to generate a greater cooling anomaly than a warming anomaly in every element of the Bjerknes feedback loop, involving the response of wind to SST, SST to thermocline, and thermocline to wind. This positive feedback is primarily responsible for the observed IOD amplitude asymmetry.

5. Bjerknes feedback asymmetry and a covarying BL

The quasi-permanent presence of a BL near the Sumatra–Java region and its covariance with the IOD may further reinforce the asymmetry. As a blockadoe to turbulent entrainment of the cold thermocline water into the ML (Lukas and Lindstrom 1991), variability of the BL strongly influences SST (Masson et al. 2004). Using Argo profiles, Qiu et al. (2012) find that during the 2006 pIOD event, equatorial easterly induced upwelling Kelvin waves raised the IL off the Sumatra coast. A salinity-stratified ML also shoals because of a reduced near-surface eastward salty water transport by a weaker Wyrtki jet (Wyrtki 1971), but this ML shoaling is in part offset by a reduced freshwater flux associated with a decrease in rainfall during a pIOD. Consequently, thinning of the BL is dominated by thinning of the IL. During the 2010 nIOD event, a similar process operated, but in the opposite direction.

A more general relationship of the IL with the IOD is constructed in Fig. 5, showing that a thinning in the IODE region during a pIOD enhances the thermocline–ML coupling, and hence, the Bjerknes feedback loop. In
contrast, a thickening of the BL enhances the decoupling of the ML and the thermocline. Even if the covariance of the BL with pIOD and nIOD is symmetric, this is an additional effect to the purely thermal consideration without taking into account the salinity effect and will reinforce the asymmetry of the Bjerknes feedback.

It turns out that the covariance of the BL with the IODE SST is in itself asymmetric. For the same one unit of change of the IODE SST, the size of BL thinning is far greater when the IODE region is anomalously cold than the size of thickening when the IODE region is anomalously warm (Fig. 5b). This is not surprising given that the variance is dominated by the IL, that is, the depth of the top of the thermocline, which has an asymmetric response (Fig. 5d). There is little response in the IODE ML to the IOD, as discussed in Qiu et al. (2012), who also showed that BL variability is driven by SST variability. Below we show that SST anomalies, positive or negative, are not forced by heat flux.

6. SST–cloud–radiation feedback

On the basis of data from several reanalyses, an asymmetry in the SST–cloud–radiation feedback is thought to be responsible for the IOD skewness (Hong and Li 2010). The idea is that after cold IODE SST anomalies reach a threshold value, the mean convection is completely suppressed, resulting in cloud-free conditions, upon which point further cooling will not generate a response in rainfall or radiation. Therefore, cold SST anomalies are not subject to damping and are able to grow freely, generating the IOD skewness. Cai et al. (2012) show that such cloud-free conditions are rarely seen in the satellite-era rainfall data; therefore, a breakdown of this feedback is rare. Furthermore, Cai et al. (2012) find that the damping when IODE SST anomalies are negative is far greater than that when the IODE SST anomalies are positive.

There are several lines of evidence from available independent datasets supporting that such a breakdown of the SST–cloud–radiation feedback does not operate. Rainfall data from the Global Precipitation Climatology Project (GPCP) (Adler et al. 2003) shows a negative skewness near the IODE region (Fig. 6a), indicating that negative rainfall anomalies grow larger than positive rainfall anomalies. This is accompanied by a skewness in several flux components (Figs. 6b–f): a positive skewness of shortwave radiation (positive into the ocean, Fig. 6e),
a negative skewness of evaporative heat loss (positive out of the ocean, Fig. 6d), and a positive net heat flux skewness (positive into the ocean, Fig. 6c). Other components show no well-defined skewness direction for reasons unknown. These features suggest a greater damping over the IODE region when the IODE SST anomalies are negative.

This is indeed the case. Over the IODE region, the response of negative rainfall anomalies to cold IODE anomalies is far greater than the response of positive rainfall anomalies to warm anomalies (Fig. 7). Consistently, there is a greater damping for negative SST anomalies (Fig. 7d) than for positive SST anomalies. In fact, there is a breakdown in the damping of positive IODE SST anomalies in that the regression using positive IODE SST samples is not statistically significant, completely opposite to what Hong and Li (2010) have proposed. Thus, heat flux is not a forcing but a damping for negative SST anomalies; heat flux is not a forcing for positive SST anomalies either, because no relationship between them exists.

The greater damping for negative SST anomalies is contributed to by shortwave radiation, as expected, as a result of a decrease in cloud cover, rainfall, and by latent heat flux (Fig. 8). A smaller latent heat loss (in response to negative SST anomalies and, hence, a positive correlation) local to the IODE region occurs despite an enhanced heat loss to the west, which occurs as a consequence of easterly anomalies superimposed on climatological easterlies. Overall, there is no evidence to support a decreased damping for IODE cold anomalies. Instead, the larger IODE cold anomalies develop despite a larger damping.

7. Conclusions

Using Argo profiles and other newly available data, we provide an observation-based assessment of various feedback processes that lead to the IOD skewness. Several contributions are made to our understanding of the processes associated with the IOD.

First, these observations confirm the role of the NDH process, which reinforces negative but damps positive IODE SST anomalies (Li et al. 2003; Hong et al. 2008). Additionally, our analysis reveals a greater reinforcing effect than the damping counterpart, providing an additional factor that will further enhance the skewness.

Second, we provide observational evidence that the thermocline–temperature feedback asymmetry is the source of the IOD skewness, as suggested by previous modeling studies (e.g., Zheng et al. 2010; Ogata et al. 2013), in which IODE cold anomalies induced by a shoaling thermocline are greater than warm anomalies induced
by a deepening thermocline. We do so by showing that the feedback and its asymmetry are stronger at depth than at the surface. The absence of this feature has been used as a line of evidence supporting the notion that the skewness is not forced by the thermocline–temperature feedback (e.g., Hong and Li 2010). This thermocline–temperature feedback asymmetry is a part of the nonlinear Bjerknes-like positive feedback loop involving winds, SST, and the thermocline. We also show that the thermocline–temperature feedback asymmetry is enhanced by a nonlinear response of the BL, showing a greater thinning associated with an anomalous IODE cooling than a thickening associated with an anomalous IODE warming.

Finally, in response to an IODE cold anomaly, the reduction in rainfall, evaporative heat loss, and the increase in shortwave radiation are greater than the corresponding changes in response to an IODE warm anomaly. The higher sensitivity when the IODE SST anomalies are negative means that there is no evidence of a breakdown in the SST–cloud–radiation feedback. This is contrary to that suggested by Hong and Li (2010). The lack of a breakdown is further supported by a negative skewness in rainfall and outgoing evaporative heat loss and a positive skewness in the incoming shortwave radiation flux and in the net heat flux into the ocean.

We conclude that there is no observational evidence to support a breakdown of the SST–cloud–radiation feedback when the IODE negative SST anomalies are large; instead, the associated damping is greater than that when the IODE SST anomalies are positive. The negative skewness of the IODE SST anomalies occurs in spite of a stronger damping of negative SST anomalies, rather than by a breakdown of the damping. The skewness is therefore caused by the asymmetry of the thermocline–temperature feedback.

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