The role of the New Guinea cross-equatorial flow in the interannual variability of the western North Pacific summer monsoon

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Abstract
The western North Pacific (WNP) monsoon trough from 1958 to 2001 shows a binary-like feature in August and September, with more than half being either an imposing presence or a total absence. One of the major moisture sources maintaining the WNP monsoon trough is the low-level moisture advection laterally driven by the low-level cross-equatorial flow that originates from the Banda Sea and Solomon Sea. By decomposing contributions to the cross-equatorial flow based on the method proposed by Back and Bretherton in 2009, the boundary-layer pressure gradient in the Maritime Continent plays a major role. This pressure gradient is further found to be associated with the densely packed sea surface temperature (SST) gradient near the equator around New Guinea, which is well correlated with the SST anomalies in the equatorial eastern Pacific, a concurrent El Niño/Southern Oscillation (ENSO) condition.

Keywords: monsoon trough, El Nino/Southern Oscillation, cross-equatorial flow, western North Pacific monsoon, air–sea interaction

1. Introduction
A monsoon trough is one of well-known characteristics associated with the western North Pacific (WNP) summer monsoon (LinHo and Wang 2002). The monsoon trough, which shows pronounced interannual variability (figure 1), provides a favorable condition for tropical convective systems, such as tropical cyclones, to develop (Molinari and Volland 2013). A peculiar property of the WNP summer monsoon interannual variability is a binary-like structure of monsoon troughs, a fact that can be easily missed if only viewing anomalous fields. This remarkable contrast can be seen in figure 1. The 850-hPa streamlines show that the westerlies invade the lower left part of the blue box in strong monsoon trough years (the left column), whereas the westerlies are completely absent within the box in no monsoon trough years (the right column). The westerlies are closely linked to the cross-equatorial flow characterized by the equatorial ridge in strong monsoon trough years. Although Tao and Chen (1987) described the linkage of the WNP summer monsoon to the Australian High and the cross-equatorial flow, the role of the New Guinea cross-equatorial flow has not been extensively studied. Previous studies (e.g., Chou et al 2003, Wang et al 2003) have shown that the WNP summer monsoon is affected by El Niño/Southern Oscillation (ENSO). Hannachi and Turner (2013) showed that the positive and negative ENSO forcings influence the frequency of the different Asian summer monsoon intraseasonal modes. However, none of them discussed possible mechanisms of ENSO via the monsoon trough and the cross-equatorial flow.

Analyzing the monthly re-analysis product, ERA-40, of the European Centre for Medium-range Weather Forecasts (ECMWF) (Uppala et al 2005), and the Hadley Centre Global sea ice and SST dataset (HadISST) (Rayner et al 2003) in 1958–2001, we first investigate the linkage between the
monsoon trough and the cross-equatorial flow, based on a composite analysis, in section 2. In order to identify the momentum balance of the surface meridional wind, a semi-empirical mixed-layer model (MLM) (Stevens et al 2002) is used to evaluate possible processes that can affect the intensity of the cross-equatorial flow in section 3. Discussion and conclusions are shown in section 4.

2. Role of the low-level cross-equatorial flow

To quantify the binary-like feature of the WNP summer monsoon, a circulation index, the Monsoon Trough Index (MTI), is defined as the 850-hPa vorticity averaged over the region of 5°N–20°N, 130°E–175°E (see the blue box in figure 1) in August–September (Aug–Sep). 12 years with the MTI smaller than −0.7 standard deviation are classified as no monsoon trough years; they are 1966, 1970, 1971, 1973, 1974, 1975, 1983, 1984, 1988, 1996, 1998, and 1999. Similarly, 12 years with the MTI greater than 0.7 standard deviation are classified as strong monsoon trough years. More than half of the studied period, 1958–2001, are sorted into the binary feature of the monsoon trough associated with the WNP summer monsoon.

Figures 2(a) and (b) depict the composite maps of strong monsoon trough and no monsoon trough years, respectively.
The strong monsoon trough composite is characterized by a local Hadley cell straddling over the equator. The monsoon trough is distinctly found driven laterally by two branches of low-level cross-equatorial flows coming from two sides of New Guinea: the Banda Sea and the Solomon Sea. This implies that the moisture sustaining the monsoon trough in the WNP is mainly supplied by the tropical Southern Hemisphere. In contrast, during the no monsoon trough years the low-level moisture fluxes are confined to south of the equator, where the corresponding moisture divergence is also weakened and shrunken (figure 2(b)). It suggests that the intensity and, to a lesser degree, the location of the descending leg of the local Hadley cell, control the binary switch for the existence of the monsoon trough. In addition, the composite results are shown in figures 2(a) and (b) for which the local null hypothesis of equal means can be rejected with the Student $t$ test at the 10% significance level.

The 925-hPa meridional winds across the equator between 120°E and 160°E (over the 5°S–5°N) and the New Guinea Cross-Equatorial Flow (abbreviated as CEF) are used as a supplemental measurement to trace the origin of the monsoon trough. Figure 2(c) shows the scattering chart between the MTI and the CEF. This diagram demonstrates that the MTI and the CEF are closely correlated, with a correlation coefficient of $CC = 0.77$ that is statistically significant at the 1% significance level. Figure 2(c) further implies a critical value of the lateral forcing, i.e., the CEF, for inducing the monsoon trough. The normalized Aug–Sep SST averaged over the Niño3.4 region (5°S–5°N, 170°W–120°W) is used as a concurrent ENSO index. The CEF is found more susceptible to concurrent ENSO ($CC = 0.8$), whereas concurrent El Niño and La Niña are neatly divided by $CEF = 2.1 \text{ m s}^{-1}$. The warm phase of concurrent ENSO drives a more intensified CEF and spans a wider range of the CEF than its cold phase counterpart. The relatively small positive values of the climatological means for both MTI and CEF (figure 2(c)) indicate a slow cross-equatorial flow and a subdued monsoon trough, a result expected from a binary-like feature. The concurrent SST anomalies over the equatorial eastern Pacific, the Niño3.4 regions, show a strong correlation with the CEF and the monsoon trough, both with $CC = 0.8$, greater than critical values for the 1% significance level test. This implies that the concurrent SST anomalies over the Niño3.4 region can either directly affect the monsoon trough or affect the monsoon trough via the CEF. We will provide evidence in next section for the second route, which seems most likely the case.
Figure 3. The differences of the meridional component of the August–September surface winds between the strong and no monsoon trough years obtained from the MLM for (a) all contributions, (b) $P_i$, the free-tropospheric contribution, (c) $\Delta P_{BL}$, the boundary-layer contribution, and (d) the downward momentum mixing by 850-hPa winds. The contours intervals are 25 Pa in (a) and (b), 12.5 Pa in (c), and 1.5 m s$^{-1}$ in (d) for $P_s$, $P_i$, $\Delta P_{BL}$, and 850-hPa zonal wind anomalies, respectively. The thick gray contour denotes the zero value, and the dashed (solid) line is for the negative (positive) values. Composite differences of SST anomalies with a rejection of the null hypothesis at the 10% significance level are shaded in (c). Black (green) vectors denote the southerly (northerly) winds greater than 1 m s$^{-1}$.

3. Regulate the cross-equatorial flow

In the previous section, we demonstrated that the CEF is closely related to the monsoon trough on the interannual time scale. Next, we would like to study the interannual variability of the CEF through the momentum balance of the meridional component of surface winds calculated in a steady-state MLM (Stevens et al. 2002, Back and Bretherton 2009). The semi-empirical MLM assumes a boundary-layer momentum balance between the influences of boundary layer and free-tropospheric processes to obtain the surface winds. The top of atmospheric boundary layer is set at 850 hPa. The meridional surface winds can then be decomposed as

$$ V = V_T \varepsilon_i \varepsilon_x - U_T f \varepsilon_x + \rho_{air}^{-1} \left( f \frac{\partial P_s}{\partial x} - \varepsilon_i \frac{\partial P_s}{\partial y} \right) \varepsilon_i^2 + f^2, $$

(1)

where $f$ is the Coriolis parameter, $V$ is the meridional component of the horizontal bulk boundary-layer winds, $\rho_{air}$ is a constant reference density, and $P_s$ is surface pressure. $U_T$ and $V_T$ are the zonal and meridional components of 850-hPa winds, respectively, $\varepsilon_x$ and $\varepsilon_i$ are adopted constant values, $2 \times 10^{-3}$ $S^{-1}$ and $3.5 \times 10^{-3}$ $S^{-1}$, from Stevens et al. (2002). The topography effect strongly influences surface winds, so we focus only on tropical oceans to avoid the complication. The first two terms on the right-hand side of (1) are the 850-hPa wind downward mixing contribution, and the last two terms denote the surface pressure gradient forcing. Based on (1), the meridional surface winds can be calculated with the ERA–40 sea level pressure and 850-hPa horizontal winds as inputs. The MLM can well reproduce the meridional component of the surface winds (figure 3(a)), with a spatial correlation coefficient of 0.88 with the composite differences of the ERA-40 Aug–Sep mean 925-hPa meridional winds in the tropical western Pacific region ($20^\circ$S–$20^\circ$N, 120$^\circ$E–180$^\circ$E) between strong monsoon trough and no monsoon trough years (not shown).

According to Back and Bretherton (2009), the boundary-layer winds are related to surface convergence in two ways. One is associated with the free atmospheric pressure gradient, which can be regarded as a response to latent heat release by deep convection in the tropics (Matsumo 1966, Gill 1980, Matsuno–Gill hereafter), whereas deep convection inclines to appear over oceans where the corresponding SST surpasses a certain threshold (Gadgil et al. 1984). The other driver of boundary-layer winds comes from the SST gradient, a mechanism proposed by Lindzen and Nigam (1987) (Lindzen–Nigam hereafter). Here, we apply the Lindzen–Nigam’s back
pressure formulation (Back and Bretherton 2009) that can linearly decompose the surface pressure $P_s$ as following:

\[ P_s = P_i + \Delta P_{BL}, \quad \text{and} \]

\[ P_i = 850 \text{ hPa} + \rho_{850}(\Phi_{850} - \bar{\Phi}_{850}). \]

where $P_i$ denotes the free-tropospheric contribution to the surface pressure and can be deduced hydrostatically from the perturbed height at 850 hPa, the top of atmospheric boundary layer. $\rho_{850}$ is 850-hPa air density (1.15 kg m$^{-3}$), $\Phi_{850}$ denotes the 850-hPa geopotential height, and $\bar{\Phi}_{850}$ is taken as the unperturbed 850-hPa geopotential height averaged over the tropical belt of 20°S–20°N. The residual surface pressure $\Delta P_{BL}$ is closely correlated to the mean temperature between the surface and 850 hPa, which, in turn, is proportional to the underlying SST. Based on (2), the surface pressure gradient terms in (1) can be roughly divided into the $P_i$ and $\Delta P_{BL}$ contributions. Thus, the meridional surface winds can be calculated by using equations (1)–(3) and the Aug–Sep averages from ERA-40, including geopotential height, sea level pressure, and 850-hPa winds.

Examining the differences of anomalous surface pressure between the strong and no monsoon trough composites (figure 3(a)), regional high pressure anomalies appear to occur in the Maritime Continent and low pressure anomalies are found in the WNP, a meridionally dipole-like pattern west of 165°E. Compared to figures 3(b) and (c), the two pressure components, $P_i$ and $\Delta P_{BL}$, and their corresponding contributions to the meridional component of the surface winds are very different. A clear meridional gradient of $P_i$ is found in the WNP (figures 3(b) and 4(a)). Negative anomalies of $P_i$, which are induced by enhanced latent heating (the Matsuno–Gill mechanism), only take place over the WNP in the North Hemisphere where the monsoon trough develops. Positive anomalies of $P_i$, on the other hand, occur over the Maritime Continent in the tropical Southern Hemisphere. The anomalous southerlies associated with $P_i$ indeed appear in the WNP, particularly north of 5°N. This implies that the intensified monsoon trough with active convection drives flows northward to 15°N.

Different from that associated with $P_i$, the main meridional gradient of $\Delta P_{BL}$ is a zonal belt located along the southern rim of the warm pool (figure 3(c)), where the New Guinea CEF is located. Positive anomalies of $\Delta P_{BL}$ are mainly found south of the equator, with cold SST anomalies also south of the equator, especially the Coral Sea (the northeast coast of Australia). This implies a possible contribution of the SST gradient to the boundary-layer flow (the Lindzen–Nigam mechanism). This is consistent with the finding of Chiang et al (2001), in which a strong meridional gradient of surface temperature dominates the meridional winds in regions near the equator.

The third contribution to the meridional component of the surface winds is associated with the downward momentum mixing from the direct entrainment of 850-hPa winds into the mixed layer, which includes the downward mixing of the 850-hPa meridional winds (the first term on the right of (1)) and the contribution from the 850-hPa zonal winds forced by the Coriolis forcing (the second term on the right of (1)). The downward momentum mixing displays noticeable northerlies in the WNP (figure 3(d)), which are mainly associated with the Coriolis forcing onto local westerly winds, especially north of 5°N. The 850-hPa downward momentum mixing only induces a small portion of equatorial southerly anomalies around the Solomon Sea. Because of the strong cancellation between the southerlies in figure 3(b) and the northerlies in figure 3(d), the equatorial meridional surface wind anomalies are then dominated by the boundary-layer pressure gradient, implying that the SST anomalies around New Guinea regulate the interannual variability of the CEF.

This Lindzen–Nigam mechanism has been overlooked in previous studies, in which the deep convection anomalies in the WNP associated with the Matsuno–Gill mechanism are generally considered as the dominant forcing for the interannual variability of the local Hadley circulation. Here we provide evidence that the Lindzen–Nigam mechanism could also play an important role for inducing the meridional component of the surface winds in the tropical Southern Hemisphere, particularly around the Solomon Sea.

Figure 4(a) further displays the meridional variations of the difference of the meridional component of the Aug–Sep surface winds between the strong and no monsoon trough years averaged over 120°E–160°E from the three major contributions, $P_i$, $\Delta P_{BL}$, and the downward momentum mixing from 850-hPa winds. The meridional component of the surface wind anomalies peaks around 2.5°S. The southerlies from the $P_i$ contribution mainly occur north of the equator, the regime of the Catsuno–Gill mechanism. The southerlies from the $\Delta P_{BL}$ contribution dominate the tropical Southern Hemisphere, the regime of the Lindzen–Nigam mechanism. The contribution of the 850-hPa winds induces anomalous southerly winds only in the tropical Southern Hemisphere (figure 3(d)), which are relatively weaker than the one contribution from $\Delta P_{BL}$. More importantly, it shows a damping effect, i.e., anomalous northerlies, on anomalous southerlies north of the equator.

Next, we would like to establish the link between the CEF and ENSO. First, figure 4(b) clearly indicates a dry/moist dipole located at 7.5°S/12.5°N, which coincides well with the subsidence/convection associated with an enhanced local Hadley circulation pattern (not shown). The SST underneath mostly shows cold anomalies with the coldest just south of 5°S, which enhances the meridional SST gradient over areas that have already been dominated by a strong meridional gradient of mean SST (figure 4(c)). The background Aug–Sep mean SST north of the equator, on the other hand, is mainly associated with the warm pool (SST $> 29^\circ$C), which has a very weak meridional gradient of mean SST (figure 4(c)), due to a self-regulation mechanism (Waliser and Graham 1993). This implies that the anomalous moisture convergence associated with the monsoon trough over the warm pool is not directly driven by local SST anomalies. Instead, it is related to the moisture convergence via the New Guinea CEF induced by the enhanced SST gradient, so the Lindzen–Nigam mechanism-dominated regime becomes crucial for the interannual variability of the New Guinea CEF. Based on
Figure 4. The meridional variations of the August–September differences between the strong and no monsoon trough years averaged over 120°E–160°E: (a) the meridional component of surface winds (m s$^{-1}$) obtained from the MLM for all contributions (black line) and three contributions, $P_i$ (box), $P_{BL}$ (bar), and downward momentum mixing from 850-hPa winds (blue dotted line); (b) the vertically-integrated (surface–200 hPa) moisture convergence (green line, unit: 10$^{-6}$ kg m$^{-2}$ s$^{-1}$) and SST anomalies (black dashed line, unit: °C). The error bars of the SST anomalies and moisture convergence in (b) are represented by gray shaded and green bars, respectively. (c) The corresponding climatological SST (solid black line) and the difference of the meridional SST gradient (unit: °C deg$^{-1}$) between concurrent El Niño and La Niña (red dotted line), along with the error bar (pink shaded).

previous studies (Gordon and Susanto 2001, Kida and Richards 2009), SST around New Guinea is under the influence of southeastern monsoon via a cooling effect associated with surface latent heat flux, which is controlled by the concurrent ENSO state. In fact, the SST anomalies regressed with the concurrent ENSO in Aug–Sep (not shown) are quite similar to figure 3(c), and the coldest SST anomalies appear around New Guinea. Figure 4(c) shows a stronger meridional gradient of SST south of the equator during concurrent El Niño than the La Niña, which induces greater CEF (figure 2(c)). Overall, a possible chain is that concurrent El Niño induces SST cooling around New Guinea, creates the greater SST gradient at the southern rim of the warm pool, enhances the CEF, and then induces the monsoon trough to set up the WNP summer monsoon.

4. Conclusions and discussions

The less renowned cross-equatorial flows on two sides of New Guinea, which can reach up to 1/3 of the intensity of the Somali Jet, play a major role in regulating the monsoon trough north of the equator. In this study, we found that the WNP summer monsoon exhibits a binary-like variation in the monsoon trough, and the New Guinea CEF controls the switch. Observations and model results show the binary switch of the monsoon trough is installed on the southern rim of the warm pool, where the corresponding SST gradient is densely packed. The concurrent ENSO state affects this SST gradient via the cooling/warming of the Banda Sea and Solomon Sea. The SST gradient further induces low-level moisture convergence in the WNP via the New Guinea CEF, and then influences the monsoon trough.

The Lindzen–Nigam mechanism affects the New Guinea CEF through the New Guinea SST anomalies, which are closely related to concurrent ENSO. However, such causality should not be regarded as a nailed fact even though concurrent ENSO is supposed to be the ‘external’ forcing for the monsoon trough variability. The CEF can accelerate the growth of ENSO by enhancing the anomalous westerlies in the WNP, which is another feature of an enhanced monsoon trough.
Thus, a positive feedback between the local Hadley cell in the tropical western Pacific and the ENSO source areas could come into play. If so, this binary feature can contribute to the Niño/Niña asymmetry via a non-Bjerknes feedback, a subject we currently are exploring.

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