On the Tropical Atmospheric Signature of El Niño

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ABSTRACT

The linear atmospheric signature of ENSO, obtained by regressing fields of geopotential height $Z$, wind, vertical velocity, and rainfall upon the Niño-3.4 sea surface temperature (SST) index, is partitioned into zonally symmetric and eddy components. The zonally symmetric component is thermally forced by the narrowing and intensification of the zonally averaged equatorial rain belt during El Niño and mechanically forced by the weakening of the upper-tropospheric equatorial stationary waves and their associated flux of wave activity. The eddy component of the ENSO signature is decomposed into barotropic (BT) and baroclinic (BC) contributions, the latter into first and second modal structures BC1 and BC2, separable functions of space $(x, y)$, and pressure $p$, using eigenvector analysis. BC1 exhibits a nearly equatorially symmetric planetary wave structure comprising three dumbbell-shaped features suggestive of equatorial Rossby waves, with out-of-phase wind and geopotential height perturbations in the upper and lower troposphere. BC1 and BT exhibit coincident centers of action. In regions of the tropics where the flow in the climatological-mean stationary waves is cyclonic, BT reinforces BC1, and vice versa, in accordance with vorticity balance considerations. BC1 and BT dominate the eddy ENSO signature in the free atmosphere. Most of the residual is captured by BC2, which exhibits a shallow, convergent boundary layer signature forced by the weakening of the equatorial cold tongue in SST. The anomalous boundary layer convergence drives a deep convection signature whose upper-tropospheric outflow is an integral part of the BC1 contribution to the ENSO signature.

1. Introduction

The atmospheric signature of ENSO is clearly defined, robust, and multifaceted and, as such, offers an opportunity to hone our understanding of the processes that shape the tropical general circulation. ENSO’s extratropical signature has been the subject of numerous studies, but its tropical signature has not been investigated as intensively and has yet to be treated in a holistic way, taking into account the dynamical relationships between different variables at different levels. In tutorials for lay audiences, the atmospheric circulation during El Niño is often depicted as a weakening of a deep overturning circulation cell in the equatorial plane, referred to by Bjerknes (1969) as the “Walker circulation.” However, in reality the tropical atmospheric signature of ENSO is fully three-dimensional, and it involves regional features, equatorial planetary waves, and the zonally symmetric circulation.

The sea level pressure (SLP) signature of ENSO, known as the Southern Oscillation (SO; Walker 1924; Walker and Bliss 1932), consists of an east–west, planetary-scale dipole with its western pole in the Indo-Pacific warm pool region and its eastern pole in the southeast Pacific trades (Trup 1965; Trenberth 1976; van Loon and Madden 1981; Trenberth and Shea 1987; Trenberth and Caron 2000). The upper-tropospheric geopotential height signature of ENSO in the tropics is dominated by an equatorially symmetric dumbbell pattern centered near 140°W, with anticyclonic gyres during El Niño (Horel and Wallace 1981; Yulaeva and Wallace 1994, hereinafter YW; Calvo Fernandez et al. 2004). The upper- and lower-tropospheric equatorial zonal wind signatures of ENSO are contrasted in plate 7 (P7) of Wallace et al. (1998).
An important departure from the Walker cell paradigm is the presence of perturbations in the meridional component of the surface wind signature, strongest in the central Pacific but extending along much of the length of the equatorial Pacific SST cold tongue, first noted by Rasmusson and Carpenter (1982) and clearly evident in P7. When the cold tongue weakens during El Niño, the anomalous meridional flow is equatorward, producing boundary layer convergence.

The upper-tropospheric planetary wave signature of ENSO is positioned such that El Niño is attended by an anomalous weakening of the equatorial stationary waves in the upper troposphere (Dima and Wallace 2007; Grise and Thompson 2012, hereinafter GT). A weakening of the waves implies a reduction of the poleward flux of wave activity out of the equatorial belt. It follows from the Eliassen–Palm flux formalism (Edmon et al. 1980) that El Niño should be marked by anomalous poleward fluxes of westerly momentum out of the equatorial belt. GT showed observational evidence that these relationships apply, not only to ENSO, but also to the MJO, and more generally to any weakening or strengthening of the equatorial stationary waves. They further showed that, just as the climatological-mean poleward eddy fluxes of westerly momentum at 30°N/S drive the thermally direct Hadley cells on their equatorward flanks and the thermally indirect Ferrel cells on their poleward flanks, the anomalous poleward eddy fluxes associated with a weakening of the equatorial stationary waves, which are centered ~15°N/S, drive thermally direct anomalous circulation cells in the inner tropics and thermally indirect cells in the outer tropics. The Coriolis torque associated with the anomalous meridional flow in the upper branches of the cells balances the anomalous eddy flux divergence.

Other zonally symmetric features of the ENSO signature observed during El Niño include the anomalous warmth of the tropical troposphere, documented by YW, Seager et al. (2003), and Randel et al. (2009), a strengthening and equatorward displacement of the tropospheric jet streams (Pack and Huang 2012; Seager et al. 2003; L’Heureux and Thompson 2006; GT), resulting in a narrowing of the tropics, and positive anomalies in atmospheric angular momentum (Rosen et al. 1984). Another feature that has received less attention in the literature but may play an important role in the zonally symmetric dynamics is the intensification and narrowing of the band of (zonally averaged) equatorial rainfall that marks the rising branch of the Hadley cells.

Here we undertake a systematic examination of the three-dimensional structure of the atmospheric signature of El Niño, with emphasis on the interpretation of the geopotential height, temperature, wind, and rainfall anomalies. The anomaly fields associated with the ENSO signature are generated by linear regression, using the Niño-3.4 SST index as the reference time series. The analysis protocol is summarized in outline form in Fig. 1. First, we partition the El Niño signature into zonally symmetric and eddy components. We show that the zonally symmetric component is consistent with response to the redistribution of zonally averaged diabatic heating, plus the changes in the eddy fluxes of zonal momentum associated with the weakening of the equatorial stationary waves and the deeper penetration of baroclinic waves into the subtropics during El Niño. We then decompose the eddy component into baroclinic and barotropic contributions and the former into modal contributions. The first baroclinic mode is suggestive of a deep planetary wave response to the ENSO-related diabatic heating anomalies. The second is a shallow overturning circulation with height and wind anomalies of opposing polarity in the boundary layer and free troposphere and a horizontal pattern suggestive of a local response to the underlying SST anomalies. We show evidence that the anomalous deep convection associated with this boundary-forced second baroclinic mode drives the first baroclinic mode and its attendant barotropic signature.

In the next section we list the datasets used in the paper and describe the analysis methodology. In section 3 we present a brief overview of the ENSO signature, and in section 4 we examine its zonally symmetric component. In section 5 we examine the eddy component. In the first part of this section we examine the first baroclinic mode and the barotropic mode, with which it is closely associated. These patterns dominate the
eddy signature of ENSO in the free atmosphere, but they leave unexplained much of the boundary layer signature. In the second part of section 5, we examine the second baroclinic mode, a shallow overturning circulation related to the strengthening/weakening of the equatorial cold tongue. The meridional wind component of this second mode accounts for most of the boundary layer convergence in the ENSO signature. Results are summarized and discussed in section 6.

2. Data and analysis techniques

The geopotential height, temperature, horizontal wind, and vertical velocity fields analyzed in this study are based on the ERA-Interim for the period of record 1979–2013, obtained from ECMWF (http://apps.ecmwf.int/datasets/data/interim-full-moda/levtype=pl/). The data are monthly means of daily means, with a horizontal resolution of 1.5° × 1.5°, and they include all pressure levels from 50 to 1000 hPa at intervals of 50 hPa. Rainfall data are based on the Global Precipitation Climatology Project (GPCP; Adler et al. 2003; Huffman et al. 2009) downloaded from NOAA’s Earth System Research Laboratory (ESRL; http://www.esrl.noaa.gov/psd/data/grid/d8/data/gpcp.html). Monthly values of the Niño-3.4 index were downloaded from the NOAA’s Climate Prediction Center website.

The fields analyzed in this study are anomalies (departures from the mean for each calendar month). The three-dimensional signature of ENSO is constructed by linearly regressing the fields of geopotential height and other variables on the widely used ENSO index known as Niño-3.4, the time series of monthly mean SST averaged over the region 5°N–5°S, 170°E–120°W. The regressions are based on all 12 calendar months for the period of record 1979–2013. Similar results are obtained using data from the extended boreal winter months of November–March, when ENSO activity is strongest. We use the standardized version of the index, created by dividing it by its own standard deviation (0.8°C). Positive anomalies are indicative of conditions observed during El Niño years and negative anomalies of La Niña years. The ENSO-related anomalies in each field are decomposed into zonally symmetric and eddy (i.e., departure from the zonal mean) components at each latitude and on each pressure level.

The eddy component of the ENSO signature in the geopotential height field is decomposed into a barotropic contribution, the mass-weighted vertical average, and a baroclinic contribution, the departure from the mass-weighted vertical average. The baroclinic contribution is decomposed into separable functions of space (x, y) and pressure p, using two variants of EOF analysis. In the first it is performed on the temperature signature, and the corresponding geopotential height pattern is obtained indirectly, making use of the hypsometric equation. In the second the geopotential height signature is obtained directly from EOF analysis and the temperature signature indirectly from the hypsometric equation. Results for the first mode are virtually identical, and results for the second mode are broadly similar. Here we show results based on EOF analysis of temperature, which yields a more robust second mode. Wind signatures are obtained by projecting the ENSO wind signature upon the principal components (i.e., vertical profiles) of the respective modes. The EOF analysis is performed on the two-dimensional data matrix consisting of the regression coefficients derived from Niño-3.4, with each row containing values for a single pressure level, defined within the tropical domain 30°N–30°S, and the columns consisting of the 20 pressure levels ranging from 50 to 1000 hPa. The analysis is pressure weighted by multiplying each row by the increment of pressure that it represents. We weight by pressure rather than square root of pressure so that in the temperature EOFs a 1°C anomaly represents the same amount of enthalpy in each layer and in the geopotential height EOFs a 1-m anomaly implies the same amount of momentum of the geostrophic wind. However, we found that results are not sensitive to the choice of weighting. The data are also area weighted by the square root of cosine of latitude. The procedure is similar to that described schematically in Fig. 4 of Adames and Wallace (2014), but we use EOF analysis to determine the vertical structure of the temperature and height fields instead of maximal covariance analysis.

When performing EOF analysis in the time domain, the time mean is removed so that the EOFs consist of anomaly fields, the polarities of which are assigned by the analyst. In the analysis performed in this paper, neither the row nor the column means of the regression coefficients in the data matrix are removed. Hence, each EOF and its associated principal component (PC) possess an intrinsic polarity.

The regression analysis presented here recovers features very similar to the ones that have appeared in Wallace et al. (1998) and prior studies dating back to the early 1980s. By virtue of the longer period of record, the regression coefficients shown here possess about twice as many statistical degrees of freedom as those in Wallace et al. (1998).

3. An overview

Figure 2 provides a summary of the anomalies observed during El Niño (i.e., when Niño-3.4 is positive), as
inferred from the regression coefficients. All amplitudes in this and in most subsequent figures are in dimensional units, as indicated, per standard deviation of the Niño-3.4 index. The top panel shows SST, with anomalies concentrated within the region of the equatorial Pacific cold tongue; the middle panel shows the characteristic Southern Oscillation SLP signature accompanied by 1000-hPa wind anomalies directed down the anomalous pressure gradient and converging into the belt of anomalously warm SST in the central and eastern equatorial Pacific. The patterns are reminiscent of the “warm SST episode” composites of Rasmusson and Carpenter (1982) and P7. The middle panel also shows GPCP rainfall anomalies that exhibit a strong maximum in the region of convergence in the 1000-hPa wind field. The anomalous 400-hPa vertical velocity field shown in the bottom panel mirrors the rainfall anomalies. Also shown in the bottom panel are the 150-hPa height and wind patterns, which resemble those shown in Fig. 22a of YW.

Figure 3 contrasts the 400-hPa vertical velocity field and the upper-tropospheric height and wind fields during El Niño and the contrasting, so-called La Niña polarity of the ENSO cycle. They are composites based on days when the standardized Niño-3.4 index exceeded a magnitude of 1.5 and −1.5, respectively. It is evident that the equatorial stationary waves are much stronger...
during La Niña than during El Niño, in agreement with the results of Dima and Wallace (2007) and GT.

4. The zonally symmetric component

Figure 4 shows the zonally symmetric component of the ENSO signature. The westerly wind anomalies at ~25°N/S in all three panels, with amplitudes up to 2.5 m s⁻¹, are indicative of a strengthening and equatorward displacement of the climatological-mean subtropical jet streams during El Niño and, hence, a narrowing of the tropics. In accordance with geostrophic balance, the geopotential height gradient across the subtropics is anomalously strong, as shown in Fig. 4a. Temperature, shown in Fig. 4b, is above normal throughout the tropical troposphere up to the 150-hPa level, with anomalies in excess of 0.4°C in the middle and upper troposphere. The anomalous tropospheric warmth is hydrostatically consistent with the positive upper-tropospheric geopotential height anomalies in Fig. 4a and in agreement with prior results of YW, Seager et al. (2003), and Randel et al. (2009). The vertical profile of the anomalous warmth, with maximum amplitude in the 400–200-hPa layer is consistent with the notion that the lapse rate remains moist adiabatic as the tropical troposphere warms and cools in response to ENSO-related variability.

The zonally symmetric component of the ENSO signature in the mean meridional circulation, indicated by the arrows in Fig. 4c, is characterized by a pronounced strengthening of the Hadley cell in the inner tropics and a more subtle weakening in the outer tropics, consistent with results of Quan et al. (2004). The increased ascent over the equator and descent at ~15°N/S is consistent with the meridional profiles of tropical rainfall shown in Fig. 5, from which it is evident that the equatorial rain belt is narrower and more intense during El Niño, and the zonally averaged ITCZ is less pronounced. During La Niña, the equatorial rain area weakens and broadens.

The anomalous diabatic heating associated with the meridional redistribution of zonally averaged tropical rainfall shown in Fig. 5 is not the only forcing that drives the anomalous mean meridional circulation in Fig. 4c. The anomalous poleward fluxes of westerly momentum by the equatorial stationary waves shown in Fig. 4c also contribute. In agreement with results presented in Fig. 4 of GT, the anomalous fluxes are strongest ~15°N/S. The peak amplitudes of up to 5 m² s⁻² are stronger than those shown in GT. As explained in the introduction, these...
fluxes are a response to the weakening of the stationary waves during El Niño. GT showed that they drive a dipole pattern of mechanically forced cells in each hemisphere, with thermally direct anomalous cells in the inner tropics and thermally indirect cells in the outer tropics. The mean meridional circulation cell signature in Fig. 4c projects upon this idealized distribution, but the outer cells (relative to the equator) are weaker than the inner ones, and they extend 10°–15° of latitude farther poleward than the boundaries of the tropics.

Eddy fluxes can move angular momentum around in the meridional plane, but they cannot create or destroy it. Hence, they cannot be responsible for the overall increase in atmospheric angular momentum (AAM) during El Niño relative to La Niña. To account for the observed increase, it is necessary to invoke the thermally direct inner cells in Fig. 4c, which reflect the greater concentration of the tropical rainfall in the equatorial belt during El Niño. Focusing the ascending branches of the Hadley cells more tightly about the equator ventilates the entire tropical upper troposphere with air with greater angular momentum by virtue of its larger moment arm about Earth’s axis \( \text{AAM} = R_E \cos \phi (\Omega R_E \cos \phi + \omega) \), where \( \phi \) is latitude, \( \Omega \) is Earth’s angular frequency, and \( R_E \) is Earth’s radius. The resulting increase in vertical (westerly) wind shear throughout the tropics requires that the meridional temperature gradient increases throughout the tropics, and this can happen only if the tropics as a whole warms relative to the extratropics, as observed in Fig. 4b. Inferring the anomalous zonally averaged temperature distribution from the momentum balance, it is implicitly assumed that free-tropospheric Kelvin waves of the kind described by Chiang and Sobel (2002) redistribute the anomalous latent heat release associated with the ENSO signature as needed to maintain thermal wind balance.

The ENSO-related zonal wind anomalies shown in Fig. 4 influence the propagation characteristics of extratropical transient eddies. Seager et al. (2003, 2010) and Harnik et al. (2010) showed how the resulting changes in the poleward fluxes of westerly momentum by the transient eddies induce a strengthening and equatorward displacement of the subtropical jet streams during El Niño, as observed, accompanied by an anomalous thermally indirect mean meridional
circulation with midlatitude ascent and tropical descent, consistent with Fig. 4c. Wave–mean flow interactions involving the extratropical transient eddies can thus be viewed as a positive feedback on the tropically induced, zonally symmetric ENSO signature.

5. The eddy component

The eddy components of the geopotential height and horizontal velocity signatures are compared with the total signatures in Fig. 6. At the 1000-hPa level the eddy component exhibits the same tropical Southern Oscillation signature as shown in Fig. 2, but the 150-hPa height pattern is noticeably different. Removing the zonal mean at that level reveals the existence of three Rossby wave dumbbells of alternating sign. For reference, we will assign letter A to the one over the Maritime Continent, B to the one centered at ~140°W over the eastern Pacific, and C to the one over the Atlantic sector. Dumbbell B is the strongest; A is ~2/3 as strong as B, and C is ~1/3 as strong as B. The pattern is similar to the one shown in the bottom-right panel of Fig. 8 of GT, which, in turn, resembles the pattern observed in association with variations in the amplitude of the equatorial planetary waves, shown in that same figure. At the 1000-hPa level, only dumbbells A and B are clearly discernible: they are of roughly comparable amplitude, defining an east–west seesaw in the Pacific sector.

a. The first baroclinic mode and its attendant barotropic contribution

The eddy component of the ENSO signature in the geopotential height field is decomposed into a barotropic contribution (BT), the mass-weighted vertical average, and a baroclinic contribution (BC), the departure from the mass-weighted vertical average. The baroclinic contribution is further decomposed into separable functions of \((x, y)\) and \(p\), using EOF analysis, performed on the temperature signature, as described in section 2. The corresponding geopotential height pattern is obtained indirectly, making use of the hypsometric equation.

The first baroclinic mode (BC1) accounts for 77% of the spatial variance of the eddy geopotential height signature. Vertical profiles of its amplitude in the temperature and geopotential height fields, shown in Fig. 7, are typical of tropical planetary waves (Gill 1980). The node at 400 hPa in the geopotential height profile corresponds to the midtropospheric maximum in diabatic heating and vertical velocity in the ENSO-related vertical velocity anomalies (not shown) and in the equatorial stationary waves (Dima and Wallace 2007, their

![Fig. 6. The ENSO signature in the (a),(c) 150-hPa height and wind fields and (b),(d) 1000-hPa height and wind fields; (left) total fields and (right) eddy fields. The contour interval is 3 m for 150-hPa height and 1 m for 1000-hPa height. The zero contours are thick and black. The longest arrows correspond to 4 m s\(^{-1}\) at 150 hPa and 1 m s\(^{-1}\) at 1000 hPa.](image-url)
Fig. 7c). The amplitudes of the induced geopotential height and wind anomalies are roughly proportional to the anomalous divergence, which is of opposing sign above and below the node and nearly zero in the 1000–100-hPa vertical integral. BC1 is dominant in the tropics, and BT is dominant in the extratropics, as shown in Fig. 8. The first baroclinic and barotropic components of the MJO exhibit an analogous latitudinal dependence (Adames and Wallace 2014, their Fig. 7). This latitudinal dependence of the first baroclinic and barotropic modes is related to the distribution of horizontal and vertical shear in the climatological-mean zonal flow, as discussed by Wang and Xie (1996) and Ji et al. (2016).

Figure 9 provides a more detailed comparison of BC1 and BT in the tropics. Rossby wave dumbbells B and C are apparent in BT with the same polarity as they appear in the upper troposphere in BC1. Because of this reinforcement by BT, dumbbells B and C are more top-heavy than dumbbell A (i.e., the geopotential height anomalies in their upper branches are accentuated relative to those in their lower branches). In agreement with the results of Ji et al. (2015), the northern center of dumbbell B in BT is more prominent than the southern center. In contrast to BT, BC1 is more equatorially symmetric (i.e., the Northern and Southern Hemisphere centers of the three dumbbells are of comparable amplitude). It follows that at lower tropospheric levels, where the baroclinic and barotropic contributions are of opposing sign, the southern center of dumbbell B is stronger than the northern center, consistent with the Southern Hemisphere bias of the SLP signature of ENSO, as reflected in the term “Southern Oscillation.”

That the profiles of planetary wave amplitude are more top-heavy in dumbbells B and C than in dumbbell A is consistent with vorticity considerations. The planetary wave response to an imposed mass source/sink should be proportional to the ambient absolute vorticity (Sardeshmukh and Hoskins 1988). Dumbbells B and C are in the cold sector of the tropics, where the vertical shear and the upper-tropospheric flow are cyclonic and the ambient absolute vorticity consequently is larger than in dumbbell A, which is situated over the warm pool (Dima and Wallace 2007, their Fig. 2; GT, their Fig. 1). The equatorial asymmetry of BT with larger amplitude of the Northern Hemisphere center of dumbbell B, may derive from the fact that climatological-mean SST is lower near its Northern Hemisphere center than at its Southern Hemisphere center, and, accordingly, the upper-tropospheric flow is more cyclonic.

BC1 alone captures the salient features of the eddy signature of ENSO in the upper-tropospheric geopotential height field, as evidenced by the similarity between the patterns in Figs. 6c and 9a. In combination with BT, it captures most of the 800-hPa eddy structure as well, as shown in Fig. 10. However, the sum of BC1 and BT at the 1000-hPa level, which is nearly identical to those in the 800-hPa level (not shown), does not fully explain the 1000-hPa Z signature shown in Fig. 6d.

To compare the contributions of BC1 and BT to the boundary layer wind field with the 1000-hPa eddy winds in Fig. 6d, the effect of boundary layer drag must be taken into account. To do this, we partition the (BC1 and BT) eddy wind and divergence signatures into “free atmosphere” and “frictionally induced” components, as shown in Figs. 11a and 11b. The former is the sum of the wind fields in BC1 and BT above the BL (850 hPa), and the latter is estimated applying Eq. (2) of Back and Bretherton (2009) to the free atmosphere signature, as detailed in the appendix. The free atmosphere signature is characterized by strong convergence along the cold tongue region and divergence over the Maritime Continent. Frictional drag weakens the zonal wind anomalies in the equatorial belt and induces an anomalous poleward Ekman drift over the central and western Pacific.
frictionally induced component partially cancels the free atmosphere component so that the net BC1 + BT boundary layer convergence in the equatorial belt, shown in Fig. 11c, does not exhibit a strong divergent signature.

b. The second baroclinic mode

These unexplained boundary layer features of the ENSO signature are closely related to the second mode.
of the baroclinic contribution, BC2, which accounts for 15% of the temperature variance of the eddy component of the ENSO signature; 2/3 of the variance remaining after the removal of the first baroclinic mode. Its vertical profile of temperature, shown in Fig. 12, is characterized by positive anomalies extending from the surface up to 700 hPa, with weaker anomalies of opposing polarity above that level. The corresponding geopotential height anomalies are of opposing polarity in the boundary layer and the free troposphere, reminiscent of sea-breeze circulations and the shallow, boundary-forced circulations in the slab model of Lindzen and Nigam (1987).

The 1000-hPa height pattern of BC2, shown in Fig. 13a, resembles the residual signature [i.e., the total minus the BC1 + BT signatures (Fig. 13b)], and it also resembles the underlying pattern of SST anomalies Fig. 2b. The negative $Z_{1000}$ anomalies reflect the weakening of the climatological-mean ridge of high pressure that overlies the equatorial cold tongue in SST during El Niño. Upon close inspection, it is evident in the ENSO signature in both the total and $Z_{1000}$ field and the eddy $Z_{1000}$ field in (Fig. 6).

The wind vectors in all three panels of Figs. 13 depict the first residual boundary layer wind signature $R_1$, defined as the total eddy component minus the sum of BC1 and BT, including the frictional contribution in Fig. 11b. The flow is directed down the gradient of the $Z_{1000}$ signature of BC2 (Fig. 13a). Hence, it is evident that BC2 also dominates the ENSO signatures in the boundary layer mass flux and divergence (Fig. 13c).

c. Further interpretation of the boundary layer signature

In Fig. 14 the ENSO signature in boundary layer mass flux and divergence is partitioned into zonal ($\partial u/\partial x$) and meridional ($\partial v/\partial y$) components. It is evident that the meridional component dominates: without it, the rainfall anomalies in the central Pacific would not be as strong and would not extend as far to the east as observed. The observed eastward shift of equatorial Pacific rainfall observed during El Niño is qualitatively consistent with a weakening and eastward displacement of the Walker circulation in the equatorial plane, but that interpretation cannot account for the strength

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![Figure 11](image1.png)

**Fig. 11.** Boundary layer (1000–850-hPa vertically averaged) wind and divergence signatures during El Niño. (a) Free-atmosphere contribution obtained by summing BC1 and BT. (b) The frictional contribution obtained using Eq. (A1) in the appendix. (c) The net wind and divergence fields (a) plus (b). The longest arrows correspond to 1 m s$^{-1}$.

![Figure 12](image2.png)

**Fig. 12.** Vertical profiles of (left) temperature and (right) geopotential height in BC2, as inferred from the second EOF of the regression field of temperature (solid) and geopotential height (dashed).
and broad longitudinal extent of the rainfall anomalies. The paradigm of a perturbed equatorial circulation cell is too simple to explain the ENSO signature in wind and rainfall.

Figure 15 offers a three-dimensional perspective on the boundary layer wind and divergence signatures, relating them to the corresponding fields in the 300–100-hPa layer and implicitly to the midtropospheric vertical velocity. The boundary layer convergence over the weakened equatorial cold tongue during El Niño feeds a deep plume of ascent. The divergent outflow from the top of the plume corresponds to a band of diffluence in the equatorial Rossby wave signature in that layer. The distinctions between the boundary layer and upper-tropospheric wind patterns are more clearly revealed by Fig. 16, which shows the residual (R1) boundary layer wind and divergence fields together with (BC1 + BT) in the 300–100-hPa layer. The position of the plume of ascent is determined by the residual boundary layer wind field, which is dominated by BC2, and the upper-tropospheric outflow from the plume is an integral part of the (BC1 + BT) response to the diabatic heating in the plume.

Much of the structure in the residual signature is associated with a shallow meridional overturning circulation that extends along the northern flank of the equatorial cold tongue, as shown in Fig. 17. The node in the wind and pressure fields is located just above 900 hPa, and the upper branch is centered at ~700 hPa. There is a hint of the upper branch in the zonally averaged mean section shown in Fig. 4c. This shallow circulation is reminiscent of the ITCZ-related shallow meridional circulations identified by Zhang et al. (2004, 2008) and shown in Fig. 1b of Nolan et al. (2010). The layer of boundary layer divergence extending from 6° to 10°N along the northern flank of the ITCZ shows up clearly in the maps of boundary layer divergence field such as Fig. 15b, and it is reflected in rainfall statistics (Ropelewski and Halpert 1987). In contrast to the low-level features in Fig. 17, the upper-tropospheric divergence (colored shading) bears little relation to the v or $\partial \bar{v} / \partial y$ signatures in this cross section.

6. Summary and discussion

The outline in Fig. 1 is helpful in putting the results of this study into perspective. With respect to the zonally symmetric component of the ENSO signature, we confirm results of previous studies relating to the anomalous

- increased atmospheric angular momentum,
- warmth of the entire tropical upper troposphere,
• intensification and equatorward displacement of the subtropical jet streams,
• poleward fluxes of westerly momentum by the equatorial stationary waves \( \sim 15^\circ \text{N/S} \),
• intensification of the Hadley cells in the inner tropics, and
• thermally indirect mean meridional circulation cells centered at \( \sim 25^\circ \text{N} \) and \( 30^\circ \text{S} \)
during El Niño. We have proposed that several of these features are a consequence of the narrowing and intensification of the equatorial rain belt during El Niño, which ventilates the entire tropical upper troposphere with air with anomalously high angular momentum. It follows from thermal wind balance and equatorial symmetry that anomalously high angular momentum implies anomalous warmth. We have argued that the anomalous

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**FIG. 14.** Boundary layer (1000–850 hPa) winds and divergence: (a) total, (b) the zonal component, and (c) the meridional component. The longest arrows correspond to 1 m s\(^{-1}\).

**FIG. 15.** (a) Upper-troposphere (300–100 hPa) and (b) boundary layer (1000–850 hPa) vertically averaged wind and divergence signatures. The longest arrows correspond to 4 m s\(^{-1}\) in (a) and 1 m s\(^{-1}\) in (b).
poleward fluxes of westerly momentum around 15°N/S are a consequence of the weakening of the equatorial planetary waves. The intensification of the Hadley cell in the inner tropics is consistent with the rainfall anomalies and the anomalous, thermally indirect circulation cells centered in the subtropics are consistent with the response to the enhanced poleward fluxes of westerly momentum by the stationary waves. Changes in the structure of the transient eddies dispersing into the subtropics from higher latitudes also contribute to the driving of these anomalous outer cells (Seager et al. 2003, 2010; Harnik et al. 2010).

We have decomposed the eddy component of the ENSO signature into the barotropic mode (BT) and the first and second baroclinic modes (BC1 and BC2). BC1 assumes the form of an equatorially symmetric planetary wave, reminiscent of the response to an equatorial heat source (Matsuno 1966; Webster 1972; Gill 1980). BT is closely related to BC1. Their dumbbell centers of action are nearly coincident: in phase wherever the flow in the climatological-mean stationary waves is cyclonic and out of phase where it is anticyclonic. Hence, the combined contribution (BC1 + BT) tends to be proportional to the absolute vorticity of the background flow, in accordance with the baroclinic term \((\zeta + f)\text{Div}\) in the vorticity equation, which is presumably the source term for BC1. Together, BC1 and BT account for most of the eddy signature of ENSO in the free troposphere, but they leave unexplained most of the wind and divergence signature in the boundary layer, because the frictionally induced flow cancels much of the (BC1 + BT) boundary layer wind and divergence fields. The unexplained (R1) features project strongly upon BC2, which appears to be a shallow response to the underlying SST anomalies of the kind hypothesized by Lindzen and Nigam (1987). The boundary layer convergence associated with BC2 is dominated by the meridional wind component. This result is in accord with the diagnostic study of Chiang et al. (2001), who compared the contributions of elevated heating by cumulus convection and SST gradients to forcing the surface winds over the tropical oceans using a linear primitive equation model with idealized parameterizations. They found that, while elevated heating dominates the surface zonal wind response and contributes significantly to the meridional wind response, the surface temperature gradients dominate the meridional wind forcing in the equatorial Pacific cold tongue region.
The striking similarity between the thermally forced boundary layer divergence signature in BC2 and the upper-tropospheric (BC1 + BT) signature, as shown in Fig. 16, suggests that the shallow thermal response to the SST anomalies is the primary driver of the distinctive deep convective signature centered over the western flank of the equatorial cold tongue during El Niño. The convection, in turn, drives the (BC1 + BT) signature that dominates the upper-tropospheric geopotential height, wind, and divergence regression patterns. The forced planetary wave response (BC1 + BT) feeds back positively upon the boundary layer convergence signature, enhancing the low-level convergence over the equatorial central Pacific during El Niño. In the absence of friction, this positive feedback would be strong enough to roughly double the amplitude of the convergence signature over the equatorial cold tongue, but, in fact, boundary layer friction substantially weakens the feedback, as shown in Fig. 11. Figure 18 shows a three-dimensional rendering of the eddy wind and divergence fields in Fig. 15, augmented with a schematic of the equatorial vertical velocity $\omega$. In contrast to the familiar two-dimensional representation of the overturning circulation in the equatorial plane, which emphasizes the weakening and eastward shift of the so-called Walker circulation cell, this schematic incorporates the meridional confluence into the equatorial Pacific in the boundary layer flow and the diffluence in the Rossby wave gyres in the upper troposphere.

We are not the first to offer an interpretation of the barotropic component of the ENSO signature. Ji et al. (2015) performed a similar decomposition of the baroclinic and barotropic components, but they applied it to the total ENSO signature rather than just the eddy component. The patterns of the barotropic contribution as determined in the two studies are nonetheless quite similar. Ji et al. (2015) suggested that the barotropic contribution derives from nonlinear interactions between the baroclinic contribution and the climatological-mean basic state. Our proposed explanation is similar to theirs in that regard, but whereas they envisioned the vertical shear of the zonally averaged subtropical jets as playing the primary role in these nonlinear interactions, we believe that, in order to explain the distinctive spatial pattern of the barotropic contribution, it is necessary to invoke the longitudinally dependent features of the background flow.

There are several directions in which this study could be extended:
Our interpretation of the zonally symmetric ENSO signature is admittedly speculative and needs to be tested by numerical experiments of the type performed by Tandon et al. (2013). Rather than imposing an equatorial heat source, as they did in their “Phi5” experiment, it would be illuminating to apply a heat source that mimics the narrowing of the equatorial convection during El Niño.

It would be interesting to see whether the baroclinic contribution to the ENSO signature could be replicated in a simple (e.g., steady-state shallow-water wave equation) model with a realistic tropical mean state, driven by a patch of surface heating or boundary layer convergence along the equator that resembles BC2.

We have not formally evaluated the terms that contribute to the barotropic contribution. It would be interesting to repeat the analysis of Ji et al. (2015), focusing on the eddy component of the ENSO signature and taking into account the zonal dependence of the background vertical wind shear.

Trenberth and Smith (2009) have examined the atmospheric signature of ENSO with emphasis on seasonality, lagged relationships, and the “flavors” of El Niño and La Niña episodes. Rather than deriving and diagnosing a single set of three-dimensional regression fields, as in our study, their analysis is based on the leading EOFs of the tropical temperature field. It is conceivable that combining the zonal versus eddy decomposition and the dynamically motivated modal separation technique illustrated in our study with EOF analysis in the time domain, as in their study, could yield new insights into the time-dependent behavior of ENSO.

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APPENDIX

The BC1 + BT Contribution to the Boundary Layer Divergence Signature

In this study we decomposed the boundary layer winds into the contributions from the first baroclinic and barotropic modes (BC1 + BT) and what we termed the first residual contribution (R1), as depicted in Fig. 1. However, the interpretation of the boundary layer wind is complicated by the effects of frictional drag and moment fluxes from low-level clouds, which cause the boundary layer wind field to depart substantially from geostrophic balance. To take these processes into account, Stevens et al. (2002) developed a mixed-layer model to estimate the boundary layer winds in the tropics. Back and Bretherton (2009) expanded on this model by writing an explicit equation for the boundary layer (BL; 850–1000 hPa) winds, which has the following form in isobaric coordinates

\[
\begin{align*}
\mathbf{u}_{\text{BL}} &= \frac{\epsilon_f \mathbf{u}_T + f \epsilon_e \mathbf{v}_T - f \partial_x \Phi_{\text{BL}} - \epsilon_e \partial_y \Phi_{\text{BL}}}{f^2 + \epsilon_e}, \\
\mathbf{v}_{\text{BL}} &= \frac{\epsilon_f \mathbf{v}_T + f \epsilon_e \mathbf{u}_T + f \partial_y \Phi_{\text{BL}} - \epsilon_e \partial_x \Phi_{\text{BL}}}{f^2 + \epsilon_e},
\end{align*}
\]

where \( \mathbf{u}_T \) and \( \mathbf{v}_T \) are the 850-hPa winds, \( f \) is the planetary vorticity, and \( \epsilon_f = 3.5 \times 10^{-5} \text{s}^{-1} \) and \( \epsilon_e = 2 \times 10^{-5} \text{s}^{-1} \) are drag coefficients.

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