

# The Spatiotemporal Structure of Twentieth-Century Climate Variations in Observations and Reanalyses. Part I: Long-Term Trend

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## ABSTRACT

The dominant interannual El Niño–Southern Oscillation (ENSO) phenomenon and the short length of climate observation records make it difficult to study long-term climate variations in the spatiotemporal domain. Based on the fact that the ENSO signal spreads to remote regions and induces delayed climate variation through atmospheric teleconnections, an ENSO-removal method is developed through which the ENSO signal can be approximately removed at the grid box level from the spatiotemporal field of a climate parameter. After this signal is removed, long-term climate variations are isolated at mid- and low latitudes in the climate parameter fields from observed and reanalysis datasets. This paper addresses the long-term global warming trend (GW); a companion paper concentrates on Pacific pan-decadal variability (PDV).

The warming that occurs in the Pacific basin (approximately 0.4 K in the twentieth century) is much weaker than in surrounding regions and the other two ocean basins (approximately 0.8 K). The modest warming in the Pacific basin is likely due to its dynamic nature on the interannual and decadal time scales and/or the leakage of upper ocean water through the Indonesian Throughflow.

Based on the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40), a comprehensive atmospheric structure associated with the GW trend is given. Significant discrepancies exist between the two datasets, especially in the tightly coupled dynamics and water vapor fields. The dynamics fields based on NCEP–NCAR, which show a change in the Walker Circulation, are consistent with the GW change in the surface temperature field. However, intensification in the Hadley Circulation is associated with GW trend in ERA-40 instead.

## 1. Introduction

The global warming (GW) trend of the past one hundred years or so has been of concern and debated by climate scientists since the late 1930s (Callendar 1938) or even earlier (Weart 2003). In recent decades, based on carefully calibrated global surface temperature (ST) records, a consensus has been gradually reached that the global mean temperature increased by about 0.8 K

during the twentieth century (Ellsaesser et al. 1986; Hansen and Lebedeff 1987; Jones et al. 1999; Hansen et al. 1999, and other references therein), which is unprecedented in the most recent millennium of the earth's history (Mann et al. 1999). These studies indicate a relatively uniform warming over the earth's surface, except that the warming over land and higher latitudes is stronger than over ocean and lower latitudes. The spatial distribution of the trends is roughly consistent with general circulation model simulations forced by anthropogenic greenhouse gas and aerosol increases.

Schneider and Held (2001) use discriminant analysis to decompose the long-term ST field. The GW trend

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appears to be the first discriminant in their analysis. Based on its spatial pattern, they suggest that this trend results from an increase in anthropogenic radiative forcing. Because they acquire the long-term trend discriminant by maximizing the ratio of interdecadal-to-intrdecadal variance, it is difficult to use this method to detect long-term climate change in other datasets with shorter time periods.

Accompanying the global surface warming trend, there must be systematic changes in the atmospheric structure, circulation, and hydrologic cycle. Global atmospheric profile data come from radiosonde network and satellite sounding data in more recent decades. Temperature trends derived from these data have disagreed with each other and with global climate model predictions, but the derived trends have begun to converge because the understanding of discontinuities and biases in sonde records and satellite data has increased (Karl et al. 2006). Suffering little from spatial and temporal subsampling, reanalysis products that assimilate observations into a global weather prediction model (Kalnay et al. 1996; Kistler et al. 2001; Uppala et al. 2005) can potentially be used to diagnose aspects of long-term changes in the atmosphere (Santer et al. 2004; Trenberth and Smith 2005; Trenberth et al. 2005; Mitas and Clement 2005, 2006).

Although the reanalyses are produced with a fixed model and analysis system, the input observations are not homogeneous because of historical changes in the observing system, especially when new kinds of observations are introduced. After 1948, which marked the beginning of the longest National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis, two major changes occurred in the global observing system: the global radiosonde network was established in 1958 and the global operational satellite observation system was introduced in the 1970s.

Several studies have indicated limitations of the reanalysis products for global climate change studies. Kistler et al. (2001) show that the strongest impact of the introduction of satellite observations in 1979 on the NCEP–NCAR reanalysis is above 200 hPa and south of 60°S. The discontinuities in the lower stratosphere and upper troposphere around 1979 are caused by biases from both the NCEP model and the National Environmental Satellite, Data, and Information Service (NESDIS) retrievals (Santer et al. 1999). Sturaro (2003) finds that the spurious shifts in the temperature fields in 1979 are broader than earlier estimates: though the biggest changes still occurred around the 100-hPa level, the affected regions include the tropics between 700 and 50 hPa and the Southern Ocean region between

500 and 250 hPa. For the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40), Bengtsson et al. (2004a) argue that the changes in the observing system have had huge impacts on the global mean linear trends of temperature, integrated water vapor, and kinetic energy.

The time periods covered by reanalysis datasets are on the order of decades, which is too short to derive reliable regional GW trends. For example, the linear trend in equatorial eastern Pacific sea surface temperature (SST) is strongly positive during the second half of the twentieth century, because of the 1976 climate regime shift in the Pacific basin, but it is very weak over the entire century. The 1976 shift is one of several off-setting phase changes associated with natural pan-decadal variability (PDV) in the Pacific basin over the past century (see, e.g., Zhang et al. 1997).

Another complicating factor is the dominant interannual El Niño–Southern Oscillation (ENSO) phenomenon. When a dataset is long enough, a low-pass filter can be applied to remove the ENSO signal before attempting to isolate other modes of variability using techniques such as empirical orthogonal function (EOF) analysis (Zhang et al. 1997). If, however, the record is too short to apply a low-pass filter, then ENSO not only dominates the first EOF but also contaminates several following modes, as shown in appendix A. The reason is that the ENSO signal appears in remote regions away from the tropical Pacific with a time delay of a few months due to atmospheric teleconnections or ocean dynamics. The contamination of the following modes prevents the isolation of other aspects of climate variability.

To extract long-term climate information from limited and imperfect data, we describe in this paper an ENSO-removal method that is based on the observation that regional ENSO effects can be approximately described by a lead–lag relationship between the Niño-3.4 ENSO index and the local anomaly time series (Trenberth et al. 2002). After ENSO is removed, the GW trend and PDV are isolated as the leading modes in EOF analyses of ST datasets; however, the results are unstable, if based on individual parameters from the reanalyses. The reason is unclear but appears to be related to the unevenly distributed discontinuities in the reanalyses.

A systematic investigation is therefore performed on the discontinuity issue in the NCEP–NCAR and ERA-40 reanalyses (appendix B). Temporal discontinuities in the reanalyses are not uniformly present in different parameters or coherently distributed in different regions and pressure levels. On the other hand, the real climate signal should be distributed across multiple pa-

rameters in a more coherent fashion. Thus, the effects of artificial discontinuities can be reduced if a large number of parameter fields are included in a combined EOF analysis. We perform such an analysis for both reanalyses, in an attempt to isolate the GW trend and PDV signals in atmospheric thermodynamic structure and circulation fields. This paper concentrates on the GW trend, while a companion paper (Chen et al. 2008) focuses on the PDV mode.

## 2. Data and analysis procedures

### a. Datasets

#### 1) SURFACE TEMPERATURE

In this study, two long-term monthly global ST datasets are used. The first is an SST dataset, the extended reconstruction of global SST, version 2 (ERSST.V2; Smith and Reynolds 2003; Smith and Reynolds 2004), which is based on SST data in the International Comprehensive Ocean–Atmosphere Data Set (ICOADS; Slutz et al. 1985; Worley et al. 2005) from 1854 to the present. ICOADS is a global dataset based on in situ marine observations; ERSST.V2 is an improved version of ICOADS SST with updated quality control procedures and advanced reconstruction methods. The data are reliable after 1880, and the uncertainty decreases to a minimum after 1950. The other ST dataset is the National Aeronautics and Space Administration (NASA) Goddard Institute for Space Studies (GISS) surface temperature (GISTEMP) analysis (Hansen et al. 1999), which is a combination of land surface air temperature (SAT) observed by meteorological stations and SST observed by ships. In GISTEMP, land SAT is based on the Global Historical Climatology Network dataset, and SST is based on the Met Office Hadley Centre’s sea ice and SST dataset (HadISST1; Rayner et al. 2003) from 1880 to November 1981 and the National Oceanic and Atmospheric Administration (NOAA) Optimum Interpolation SST, version 2, dataset (NOAA OI SST V2; Reynolds et al. 2002) from December 1981 to present. ERSST.V2 is the latest version of the SST dataset available; thus, we use it in the calculation of the ENSO index and in an ocean-only analysis. For a global analysis covering both land and ocean, we use GISTEMP instead.

#### 2) ATMOSPHERIC PARAMETER FIELDS FROM REANALYSES

In addition to the two ST datasets, five atmospheric parameters at eight pressure levels (1000, 925, 850, 700,

600, 500, 400, and 300 hPa) from the NCEP–NCAR and ERA-40 reanalyses are analyzed. These parameters—air temperature ( $T_{\text{air}}$ ), zonal wind speed ( $U$ ), meridional wind speed ( $V$ ), vertical velocity ( $\omega$ ), and specific humidity ( $q$ )—are in the higher-reliability classes of the NCEP–NCAR reanalysis products. The  $T_{\text{air}}$ ,  $U$ , and  $V$  are ranked in class A, and  $\omega$  and  $q$  are ranked in class B, indicating that these parameters depend more on observations than on the NCEP model, especially for class-A parameters (Kalnay et al. 1996). The same should be true for ERA-40 (Uppala et al. 2005), which assimilates similar and additional radiosonde and satellite sounding data, though no ranking system is provided.

### b. Data processing and analysis procedures

For each parameter at each grid point, the long-term mean annual cycle was subtracted from the original monthly time series to obtain the anomaly time series. The long-term mean annual cycle for the STs is based on the period 1982–2000, following the definition of GISTEMP. For the reanalyses, we use 1968–96, consistent with that used by the NCEP–NCAR reanalysis. The difference in the definition of annual cycle has negligible effects on the analysis results.

Next, the ENSO-removal method (see appendix A) was applied to the anomaly time series of each parameter at each grid point. Then, the ENSO-removed ST data were analyzed using EOF analysis for 60°S–60°N from 1900 to 2003. The GISTEMP and ERSST.V2 analyses were performed based on the covariance matrix of the parameter field, and the components of the matrix were weighted by the area of the corresponding grid. Before EOF analysis, to emphasize the long-term variations, the ENSO-removed monthly time series at each grid is smoothed with an annual average filter (with weights of 1., 1.5, 3., 4.5, 5.5, 6.5, 7., 6.5, 5.5, 4.5, 3., 1.5, and 1) and then adjusted to a zero mean by subtracting its mean value from each point of the time series.

The ENSO-removed reanalysis monthly atmospheric parameter fields were interpolated to  $5^\circ \times 5^\circ$  resolution, then were smoothed with the annual average filter and adjusted to a zero mean at each grid point. Each parameter field was normalized by dividing the standard deviation of the whole spatiotemporal field of this parameter, thus providing similar weights. Then, the rescaled parameter spatial fields at each month were put together to construct a combined field, which includes the normalized spatial patterns of the 40 parameters. After the EOF analysis was applied to the time series of the combined field, the resulting EOF of each

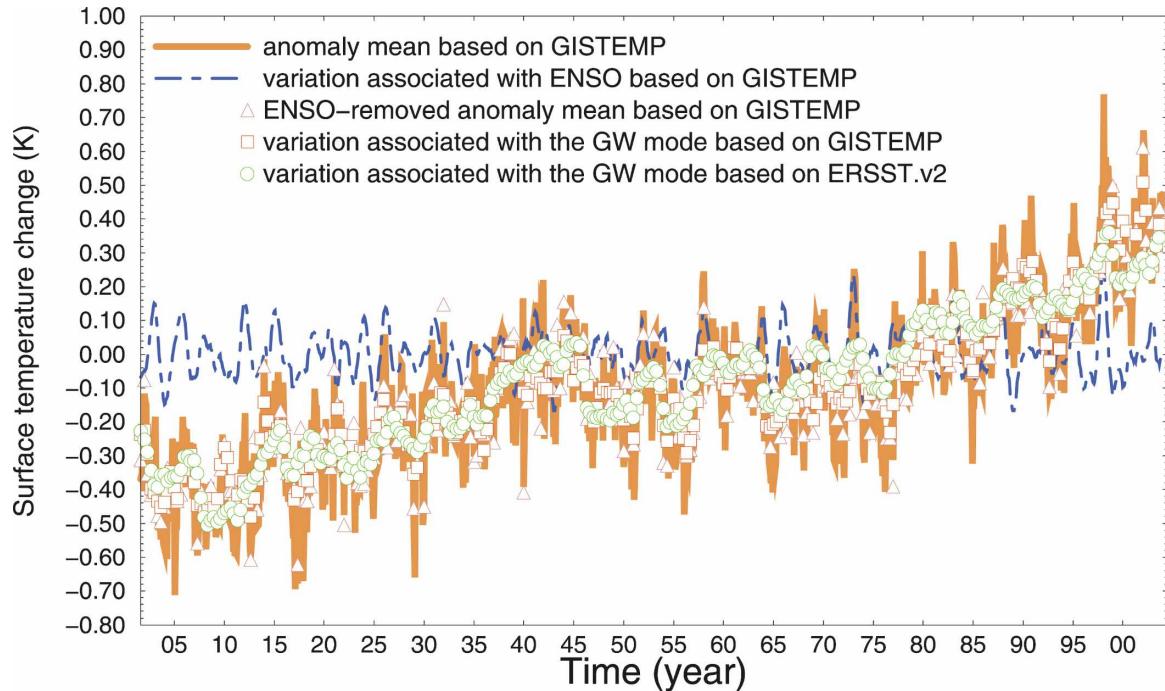


FIG. 1. The time series of the GW trend in the 60°S–60°N latitude band. Because the 60°S–60°N zonal band covers 86.6% of the earth's surface and a reliable long-term observation record is lacking in the polar regions, the average temperature for this zonal band is assumed to roughly represent the known average temperature of the entire area of the earth. Orange solid line: original anomaly mean based on GISTEMP; blue dashed–dotted line: ENSO-related anomaly mean; brown triangle: ENSO-removed anomaly mean; red square: first-mode PC of GISTEMP ENSO-removed EOF, scaled to represent the mean temperature change associated with this mode; green circle: first-mode PC of ERSST.V2 ENSO-removed EOF, scaled to represent the mean temperature change associated with this mode.

mode was separated back into 40 individual spatial patterns, one for each parameter. The 40 patterns share the principal component (PC) time series; that is, a given mode depicts related variations in all the parameters. As in the ST EOF analyses, the spatial domain of the combined EOF analyses is also over the 60°S–60°N zonal band. The time period is from 1970 to 2003 (NCEP–NCAR) and from 1970 to 2002 (ERA-40). The starting point of 1970 was chosen for the combined EOF analyses to include the 1976 climate regime shift and to exclude the relatively low-quality reanalysis data before the 1970s (as shown in appendix B). The combined EOF analyses based on reanalysis datasets are abbreviated as the NCEP1–40p and ERA40–40p analyses, respectively.

### 3. The global warming trend mode in the ST field

The mean anomaly ST time series (Fig. 1, orange solid line), based on GISTEMP data, shows no obvious trend from 1900 to 1910. There was an increase (about 0.3 K) in global ST from the mid- to late 1910s through the early 1940s, a slight cooling from the 1940s to the

1970s, and then the global ST increased again by about 0.5 K from the 1970s through the late 1990s. The ENSO-related deviation (Fig. 1, blue dashed–dotted line), defined as the global mean ENSO signal based on the ENSO-removal method in appendix A, is mostly around  $\pm 0.1$  K, with only a few exceptional El Niño events reaching the level of 0.2 K. The ENSO-removed ST time series (Fig. 1, brown triangle) retains most of the long-term variability in the original ST time series.

We can also see from Fig. 1 that the principal component (Fig. 1, red square, which is scaled to represent the global mean ST change associated with the EOF mode) of the first mode (27.3% variance explained) of the GISTEMP analysis matches well with the original global average ST time series, in both its low-frequency characteristics and the magnitude of the long-term increase. (This mode will hereafter be called the GW trend mode.) The GW trend mode also appears to be the dominant mode of the ERSST.V2 analysis (Fig. 1, green circle). Both PC time series share the same long-term trend with few differences and are consistent with the global average temperature trend.

The EOF spatial patterns of the GW trend mode

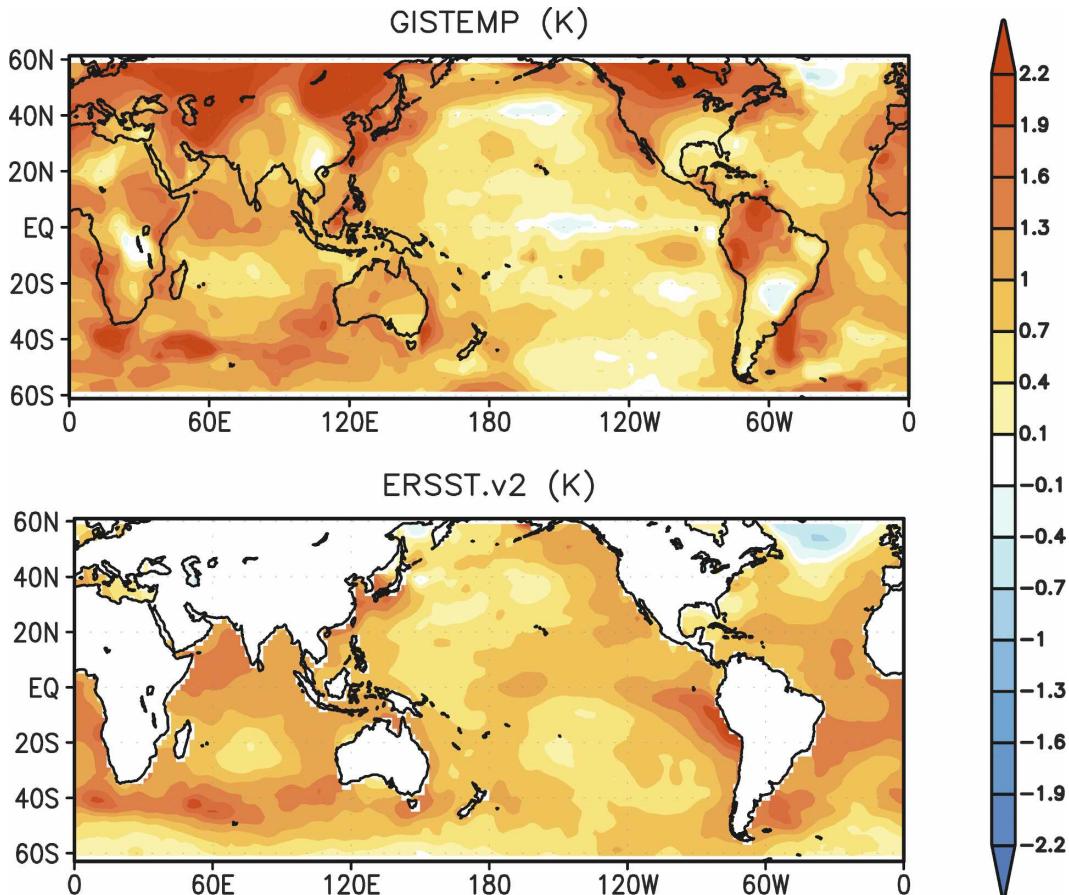


FIG. 2. The spatial patterns of the trend (first) modes of the 1900–2003 ENSO-removed EOF analyses based on GISTEMP and ERSST.V2. This mode explains 27.3% and 31.1% of the total variance in the ENSO-removed annually smoothed anomaly data of GISTEMP and ERSST.V2, respectively. (top) The EOF spatial pattern from GISTEMP. (bottom) The EOF spatial pattern from ERSST.V2. Each grid of the EOF spatial patterns is scaled to represent the range of the local temperature change associated with the PC time series.

(Fig. 2) show that most of the earth's surface warmed in the twentieth century. The GW spatial patterns suggest a response to globally imposed radiative forcing rather than a long-term internal oscillation of the ocean circulation, which usually appears as a polarized structure in the SST field (Schneider and Held 2001). The locations of the cooling regions over land approximately coincide with the regions with negative radiative forcing from sulfate aerosol, organic aerosol, soil dust, and land-use change (Hansen et al. 1998; Haywood et al. 1997; Penner et al. 1998; Tegen et al. 1996). The connection between aerosol loading and the regional cooling trend over land has been noted in previous studies (Hansen et al. 1998; Karl et al. 1995; Mitchell and Johns 1997). Generally, warming over land is substantially stronger than over the ocean because of the large thermal inertia of the ocean compared to the negligible heat-storage capability of the land and the smaller evaporative cooling over dry land (Trenberth et al. 2007). The warming

over high latitudes is stronger than over low latitudes because of the positive temperature-albedo feedback caused by snow cover and the absence of convective heat removal at high latitudes.

The ocean-only spatial pattern of the ERSST.V2 trend mode (Fig. 2, lower) is not identical to the ocean part of the GISTEMP trend mode pattern (Fig. 2, upper; correlation coefficient: 0.43). An ocean-only analysis applied to GISTEMP (not shown) has a nearly identical trend mode to that from the original GISTEMP analysis: the correlation coefficients are 0.97 and 0.99 for the PC time series and the EOF spatial pattern, respectively. Thus, the difference in spatial patterns stems from the disagreement between the two datasets, which are different in historical bias corrections, input data, and analysis procedures (Smith and Reynolds 2003, 2004; Rayner et al. 2003). There is almost no cooling in the Pacific anywhere in ERSST.V2. The warming along the west coast of South America and over the

Central Pacific region is stronger in ERSST.V2, while warming off the west coast of North America is larger in GISTEMP. Furthermore, there are differences in the detailed features of the two patterns over the Indian and Atlantic Oceans. Otherwise, the patterns from GISTEMP and ERSST.V2 are generally consistent in their global-scale ocean-warming patterns.

The average warming over the Pacific is much weaker than over the Atlantic and Indian Oceans in both datasets. This feature has not been emphasized by previous observational studies, though it can be seen in local ST linear trend maps: for example, in Fig. 3.9 of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4; Trenberth et al. 2007), in which it was attributed to Pacific decadal oscillation effects. There is no sign of this feature in the multimodel ensemble annual mean ST change based on different scenarios in Fig. 9.6 of the AR4 report (Hegerl et al. 2007). The weakened warming in the Pacific basin is consistent with the ocean heat content change in the analysis of Levitus et al. (2000). About half of the increased heat is stored in the upper 300 m of the Atlantic and Indian Oceans. In the Pacific Ocean, however, the heat content in the upper 300 m seems to have stronger decadal variability and a small increasing trend, which is consistent with our weaker Pacific SST increase.

It is natural to ask why the warming in the Pacific Ocean is weaker than in other ocean basins. First, the largest interannual and decadal variability in mid- and low-latitude ocean dynamics, due to ENSO and PDV, is located in the Pacific basin. ENSO involves large heat storage and release on interannual time scales (Trenberth et al. 2002). The same is probably also true for the decadal oscillation in the Pacific. Thus, radiative heating anomalies may mix more easily to the deeper ocean and/or be released to the atmosphere, instead of being stored in the upper ocean, thus diminishing surface warming in the Pacific basin.

Another possible influence on the Pacific is the Indonesian Throughflow (ITF). Warming induced by radiative forcing is concentrated in the upper levels of the ocean. In the Pacific, surface mixed-layer and thermocline waters are continuously exported to the Indian Ocean through the ITF; the lost upper-level water is replaced by cold lower-level water from the Southern Ocean. Through this renewal process, at least part of the heat from radiative forcing is transported from the Pacific Ocean to the Indian Ocean. The amplitude of this cooling effect can be roughly estimated through the following assumptions: (i) the temperature increase occurs uniformly in the upper 300 m; (ii) the temperature increase rate in the Pacific Ocean would equal the global mean ST increase rate if there were no leakage

through the ITF; (iii) the ITF transport is 5 Sv ( $\text{Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$ ) in the upper 300 m (about half of the total ITF transport that occurs down to 700 m; Song et al. 2004; Susanto and Gordon 2005); (iv) the ITF water is only from the upper 300 m of the Pacific Ocean; and (v) the temperature of the Southern Ocean water that replaces the ITF water is stable. The change in temperature in the upper 300 m of the Pacific Ocean can be expressed as

$$dT(t) = Rdt - \text{ITF} \left[ \frac{T(t) - T(0)}{V} \right] dt, \quad (1)$$

where  $T$  denotes temperature,  $t$  refers to time ( $t = 0$  is assumed here to be 1900),  $R$  is the global mean ST increase rate (0.8 K per century), ITF is the ITF flux (approximately 5 Sv, treated as constant), and  $V$  is the volume of the water mass in the upper 300 m ( $5.43 \times 10^{16} \text{ m}^3$ ) of the Pacific. The solution of (1) is

$$\Delta T = \frac{RV}{\text{ITF}} \left( 1 - e^{-\frac{\text{ITF}}{V} \Delta t} \right). \quad (2)$$

Thus, in the last century ( $\Delta t = 100 \text{ yr}$ ), the temperature change in the Pacific Ocean estimated from (2) is  $\Delta T \sim 0.7 \text{ K}$ . The difference between the global mean temperature increase and  $\Delta T$  is  $0.8 - 0.7 \text{ K} = 0.1 \text{ K}$ . This implies a 0.15-K difference between the Pacific and the rest of world, given the relative ocean areas. ITF could be a factor in the weakened warming of the Pacific Ocean.

The warming pattern in the Indian Ocean is also modified by the ITF. Although the average warming in the Indian Ocean is much stronger than in the Pacific Ocean, the relatively weaker warming east of Madagascar (Fig. 2) may be caused by the upwelling of the relatively cooler ITF water that is transported from the Indonesian Sea to the West Indian Ocean in the thermocline (Song et al. 2004). Along the west coast of the Indian Ocean, the Agulhas Current transports warm water from low latitudes to southern midlatitudes. The strong warming in the South Indian Ocean (around  $40^\circ\text{S}$ ) probably is caused by the retroflexion of the Agulhas Current (Gordon 2001) near the south end of Africa. Part of the Agulhas water enters the South Atlantic, thus contributing to the strong warming in the southeast Atlantic region.

The northern North Atlantic Ocean appears to be weakly cooling (Fig. 2). This feature has been simulated and attributed to a reduction of poleward heat transport resulting from the weakening of thermohaline circulation, initiated by an increase in the freshwater input at high latitudes (Cubasch et al. 2001). Another possible reason is the strong positive phase of the North Atlantic Oscillation in the 1980s and 1990s (Scaife et al.

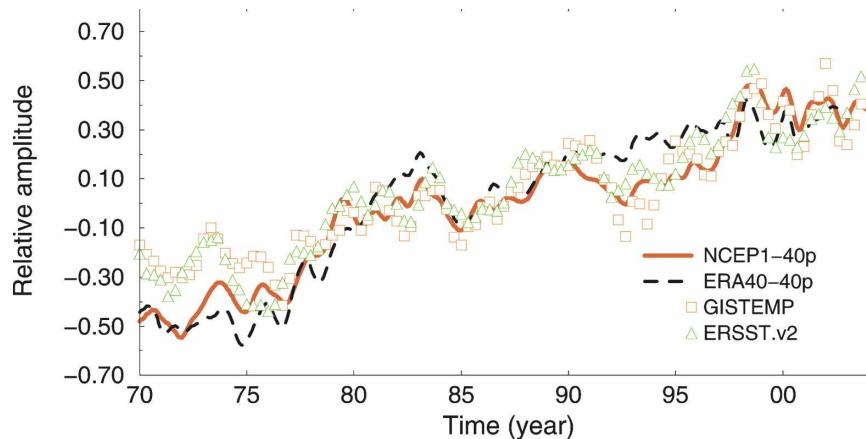


FIG. 3. The PCs of the GW trend (first) mode of the ENSO-removed combined EOF analyses based on reanalyses data. Red solid: PC from 1970–2003 NCEP1–40p analysis, based on NCEP–NCAR data; this accounts for 16.6% of the total variance. Black dashed line: PC from 1970–2002 ERA40–40p analysis, based on ERA-40 data; this accounts for 15.1% of the total variance. For reference, the 1970–2003 segments of the 1900–2003 ENSO-removed EOF analyses based on GISTEMP (orange square) and ERSST.V2 (green triangle) are also shown. The correlation coefficients for every pair of the time series in this plot are all above 0.9.

2005), which causes southward Ekman drift of cold polar water and advection of cold air from Labrador.

#### 4. The GW trend mode manifested in the atmospheric fields of reanalyses

Figure 3 shows the dominant-mode PCs from both the NCEP1–40p (red solid line) and the ERA40–40p (black dashed line) analyses and the 1970–2003 segment of the GW trend mode PCs of the GISTEMP (orange square) and ERSST.V2 (green triangle) analyses. The NCEP1–40p and ERA40–40p PCs share a similar long-term increasing trend with the GISTEMP and ERSST.V2 PCs, though they are not identical. During the early 1970s, the NCEP1–40p and ERA40–40p PCs are more negative than the GISTEMP and ERSST.V2 PCs. This difference is unlikely to be caused by the satellite discontinuity effect on the reanalysis data, as all the curves agree after 1975. The timing may instead suggest gaps in sampling of the 1976 climate shift by the radionsonde network. The early 1990s minimum in the GISTEMP PC, due to the 1991 Mount Pinatubo eruption, is not so obvious in the ERSST.V2 and NCEP1–40p PCs and is absent in the ERA40–40p PC. This difference is plausible for NCEP1–40p and ERA40–40p, because volcanic aerosol forcing is not explicitly included in the reanalyses. In the ocean-only GISTEMP analysis (not shown), the volcanic signal is weaker than in the original GISTEMP analysis but still larger than in ERSST.V2. Therefore, the diminished

volcano signal in the ERSST.V2 analysis is caused not only by the resistance of the ocean to the transient volcano aerosol forcing but also by the difference between the two datasets. Otherwise, the NCEP1–40p and ERA40–40p PCs match the GISTEMP PC and the ERSST.V2 PC well, considering that they are based on different parameter fields and time ranges.

The spatial patterns of the atmospheric temperature changes associated with the trend mode are shown in Fig. 4 (NCEP1–40p in the left column; ERA40–40p in the right column). Warming is prevalent at all tropospheric levels for both NCEP1–40p and ERA40–40p. This is consistent with global radiative forcing and with the increase of tropospheric temperature projected by climate models forced by natural and anthropogenic radiative forcings (Hansen et al. 2002; Karl et al. 2006).

The near-surface  $T_{\text{air}}$  patterns from the NCEP1–40p (Fig. 4, bottom left) and ERA40–40p (Fig. 4, bottom right) analyses differ in some ways from the ST pattern from the 1900 to 2003 GISTEMP and ERSST.V2 analyses, especially outside the tropics (Fig. 2). This discrepancy mainly results from the different time ranges. The correlation coefficient between the NCEP1–40p 1000-hPa pattern (Fig. 4-12 in Chen 2005) and the GISTEMP pattern becomes much greater (rising from 0.46 to 0.71) when the GISTEMP 1900–2003 spatial pattern (Fig. 2, top) is replaced by the GISTEMP 1970–2003 spatial pattern (Fig. 4-13 in Chen 2005).

In NCEP1–40p, the temperature increase at the 850-hPa level is larger than at the levels below (925 hPa)

# Air temperature (K)

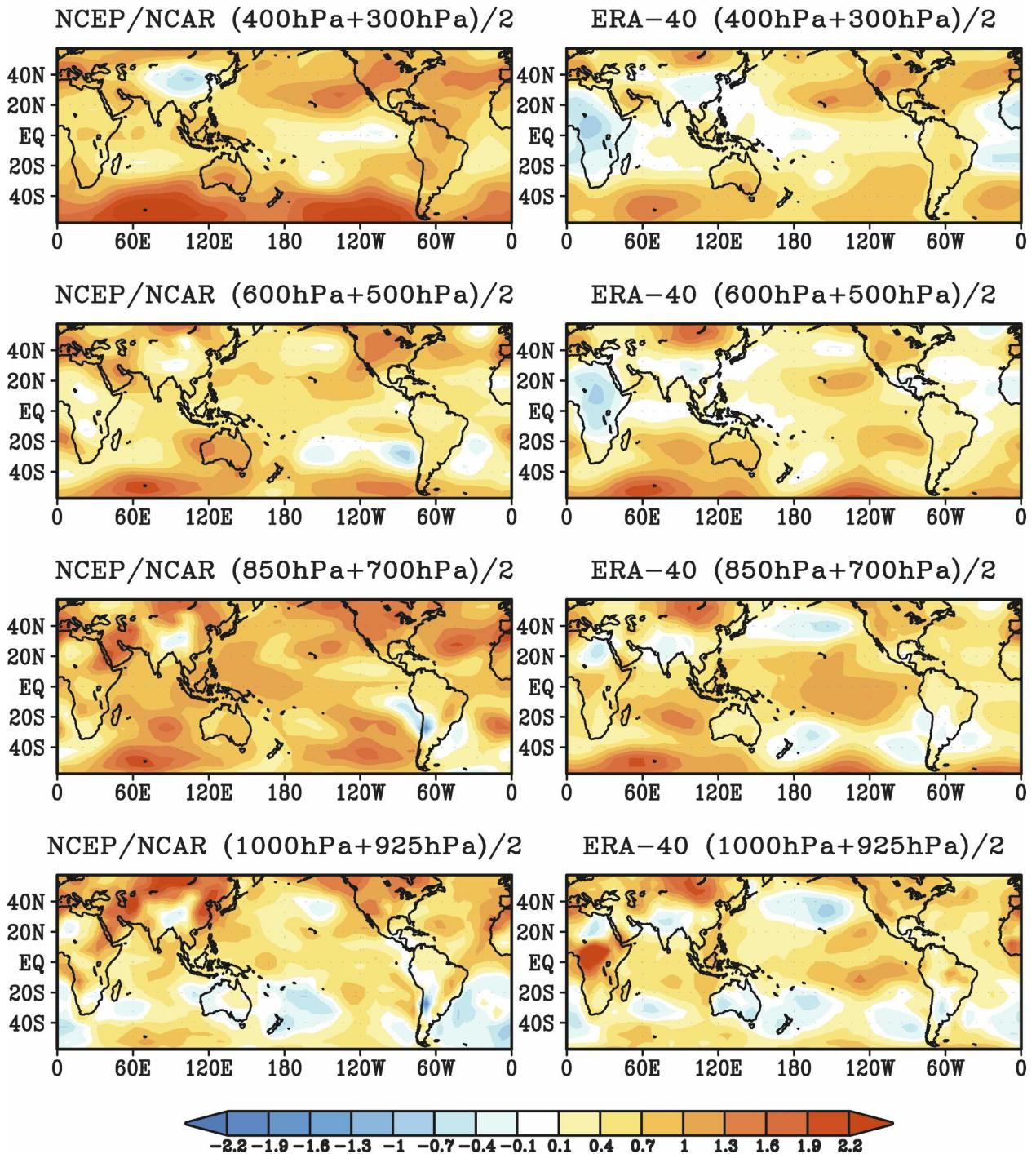


FIG. 4. The spatial patterns of  $T_{\text{air}}$  variation from 1000 to 300 hPa associated with the GW trend mode. (left) From NCEP1-40p analysis, based on NCEP-NCAR reanalysis data. (right) From ERA40-40p analysis, based on ERA-40 reanalysis data.

TABLE 1. The correlation coefficients between NCEP–NCAR reanalysis and ERA-40 reanalysis in GW spatial patterns.

	1000 hPa	925 hPa	850 hPa	700 hPa	600 hPa	500 hPa	400 hPa	300 hPa
T	0.65	0.49	0.11	0.27	0.50	0.61	0.68	0.52
U	0.59	0.63	0.53	0.39	0.48	0.47	0.46	0.53
V	0.43	0.42	0.45	0.52	0.56	0.58	0.59	0.60
$\omega$	0.17	0.22	0.25	0.23	0.29	0.34	0.37	0.41
q	0.45	0.32	0.30	0.09	0.01	0.18	0.18	0.01

and above (700 hPa), over the South Indian Ocean and Southeast Pacific regions (Fig. 4 left column; shown more clearly in Fig. 4-12 of Chen 2005). In the midtroposphere of ERA40–40p and the upper troposphere of NCEP1–40p, warming in the southern midlatitudes is suspiciously high. The temperature discontinuity patterns of NCEP–NCAR and ERA-40 (appendix B; fig. B2) suggest that these might be artifacts of satellite discontinuity in the 1970s.

Although  $T_{\text{air}}$  patterns (Fig. 4) are contaminated by the above artifacts, moderate to high correlations in  $T_{\text{air}}$  patterns between the two reanalyses (Table 1) imply useful information content, considering they are based on different models, input observations, and assimilation systems. Broad warming patterns indicate similar trend directions with observations (Lanzante et al. 2006) and GCMs (Santer et al. 2006). At the same time, the contamination from discontinuities makes it hard to answer a more subtle question based on reanalyses: whether the troposphere has warmed more than the surface. The answer is positive based on GCM simulations (Santer et al. 2006), but still not in consensus in different observation datasets (Lanzante et al. 2006). The proposed new reanalyses (Folland et al. 2006), for example, Modern Era Retrospective-analysis for Research and Applications (MERRA; <http://gmao.gsfc.nasa.gov/research/merra/>), which focus more on hydrological cycle and climate homogeneity, will help to solve this problem.

In Fig. 5, atmospheric dynamics changes at 850, 500, and 300 hPa associated with the trend mode of NCEP1–40p (left column) and ERA40–40p (right column) are shown. In general, the dynamical fields significantly differ between the two reanalyses. For NCEP1–40p, a strong 850-hPa divergence trend (Fig. 5, bottom left) appears in the equatorial central Pacific, while convergence trends exist in the Amazon, east Pacific, west Pacific, and equatorial Indian Ocean. In the midtroposphere (Fig. 5, middle left), these features appear as corresponding 500-hPa descent/ascent trends but on somewhat broader spatial scales. The ascending anomalies imply that convective activity over the Maritime Continent and Amazon basin should have increased in association with the GW trend. In the central

Pacific, the strong subsidence feature extends to the northern and southern Pacific at weaker intensity. Coherent with the midtroposphere vertical velocity field, the upper-troposphere horizontal wind field (Fig. 5, top left) diverges in the 500-hPa ascending regions and converges in the descending regions. The three-dimensional wind field defines an anomalous overturning circulation, which has one descending branch over the central Pacific and two ascending branches over the western and eastern Pacific coastal regions. This anomalous overturning circulation is consistent with the large-scale trend in the zonal SST gradient over the Pacific basin and its surrounding regions, caused by the weakened warming in the Pacific basin (Fig. 2).

For ERA40–40p (Fig. 5, bottom right), a chain of 850-hPa convergence anomalies circles most of the earth roughly along the equator, and divergence appears over the subtropical Pacific. Consistent with this, midtroposphere ascending motion (negative  $\omega$ ) (Fig. 5, middle right) dominates in the vicinity of the equator, except over Africa; descending motion is prevalent in the subtropics. In the upper troposphere (Fig. 5, top right), strong divergence (convergence) appears more in the equatorial (subtropical) regions. This describes a coherent three-dimensional dynamic change in the GW mode of ERA40–40p: more convective activity occurs in the equatorial region and more subsidence motion in the subtropical regions, indicating an intensification of the Hadley Circulation (Mitas and Clement 2005).

The long-term mean of the dynamical fields and the changes in these associated with the GW trend mode in the two reanalyses are summarized in the 10°S–10°N meridional mean zonal cross section and zonal mean meridional cross section wind fields shown in Fig. 6. The two reanalysis datasets are similar in the climatological mean overturning circulations (Fig. 6, green vectors). The zonal mean Hadley Circulation (top) has climatological ascending motion from 10°S–10°N, with a maximum at  $\sim 7.5^\circ\text{N}$ ; descending motion occurs at 15°–40° in both hemispheres. The climatological 10°S–10°N meridional mean Walker Circulation (Fig. 6 bottom, green vectors) has a dominant convection center over the Maritime Continent and west Pacific warm pool (80°–180°E) region, and two secondary ascending re-



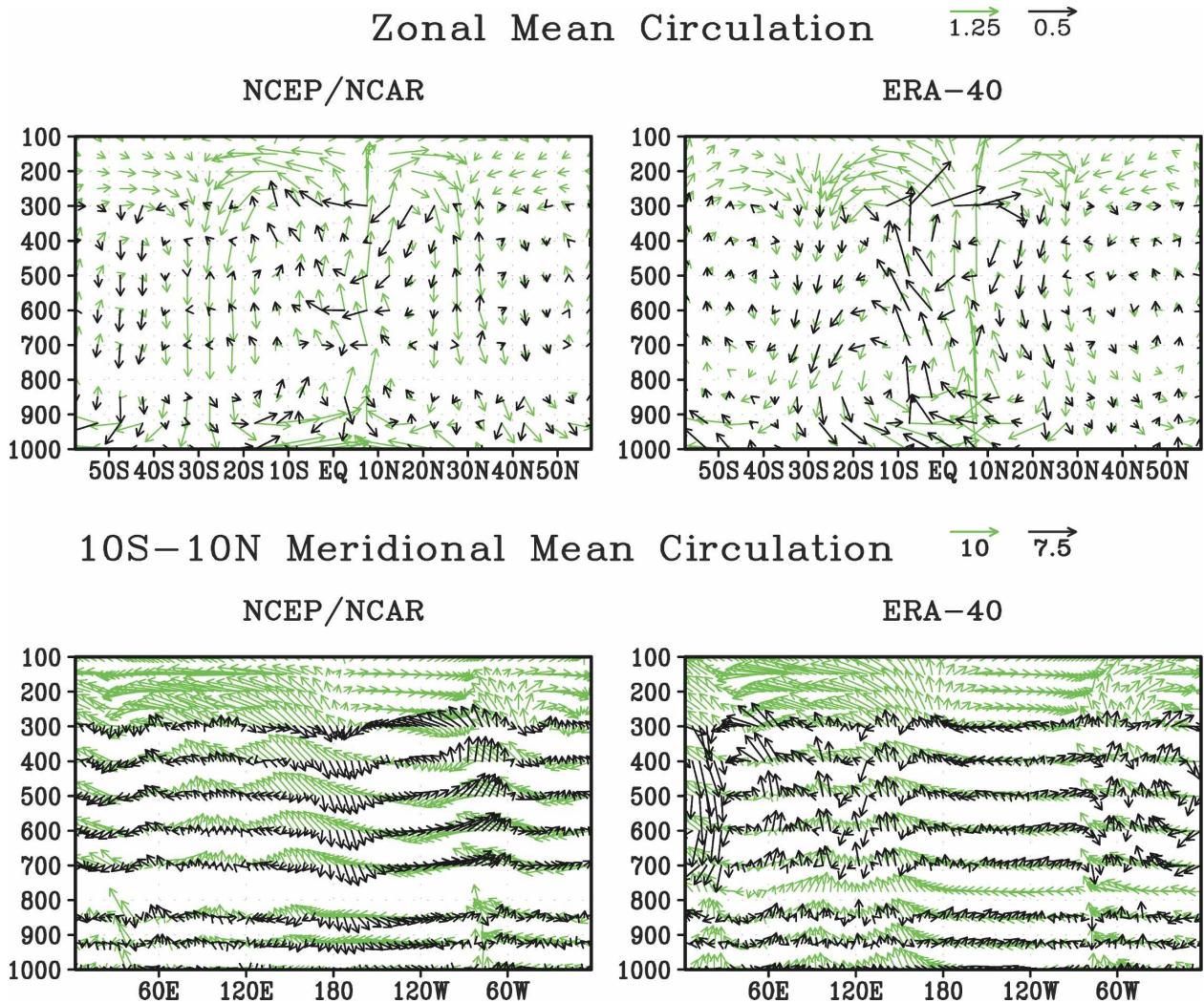


FIG. 6. The circulation changes (black vectors) associated with the (top row) GW trend mode in the zonal mean meridional vertical cross section and the (bottom row)  $10^{\circ}\text{S}$ – $10^{\circ}\text{N}$  meridional mean zonal-vertical cross section. (left column) From NCEP1–40p analysis, based on NCEP–NCAR reanalysis data. (right column) From ERA40–40p analysis, based on ERA-40 reanalysis data. The climatology is drawn in green vectors. In the top plots, the horizontal component of the vectors is meridional wind with unit  $1(\text{m s}^{-1})$ , and the vertical component of the vectors is negative  $\omega$  with unit  $-1/60(\text{hPa s}^{-1})$ . In the lower plots, the horizontal component of the vectors is zonal wind with unit  $1(\text{m s}^{-1})$ , and the vertical component of the vectors is negative  $\omega$  with unit  $-1/120(\text{hPa s}^{-1})$ . The arrow lengths of the vectors are scaled as shown on the top of each row.

dramatically in the two reanalyses. In NCEP1–40p, consistent with the broad surface warming (Fig. 2),  $q$  broadly increases near the surface 1000-hPa and 925-hPa levels (the latter is not shown). In the mid- to upper troposphere, the water vapor concentration seems tightly coupled with the dynamics (Trenberth et al. 2005). Strong drying in the midtroposphere (Fig. 7, two middle left plots) over the central Pacific is consistent with the strong descending branch of the anomalous zonal overturning circulation (Fig. 5, middle left; Fig. 6, bottom left); at the same time, strong moistening (Fig. 7, two middle left plots) occurs in both the western and

eastern Pacific ascending branches (Fig. 5, middle left; Fig. 6, bottom left). In the upper troposphere, moistening is strong over the east Pacific and Amazon regions, but drying now occurs over the Maritime Continent region. This asymmetric change might be attributed to the stronger anomalous ascending motion over the east equatorial Pacific and Amazon regions (Fig. 5, middle left; Fig. 6, lower left), thus allowing more water vapor to be transported to the upper troposphere, in comparison with the situation over the Maritime Continent region.

In ERA40–40p, the water vapor change is closely

# Specific Humidity (g/kg)

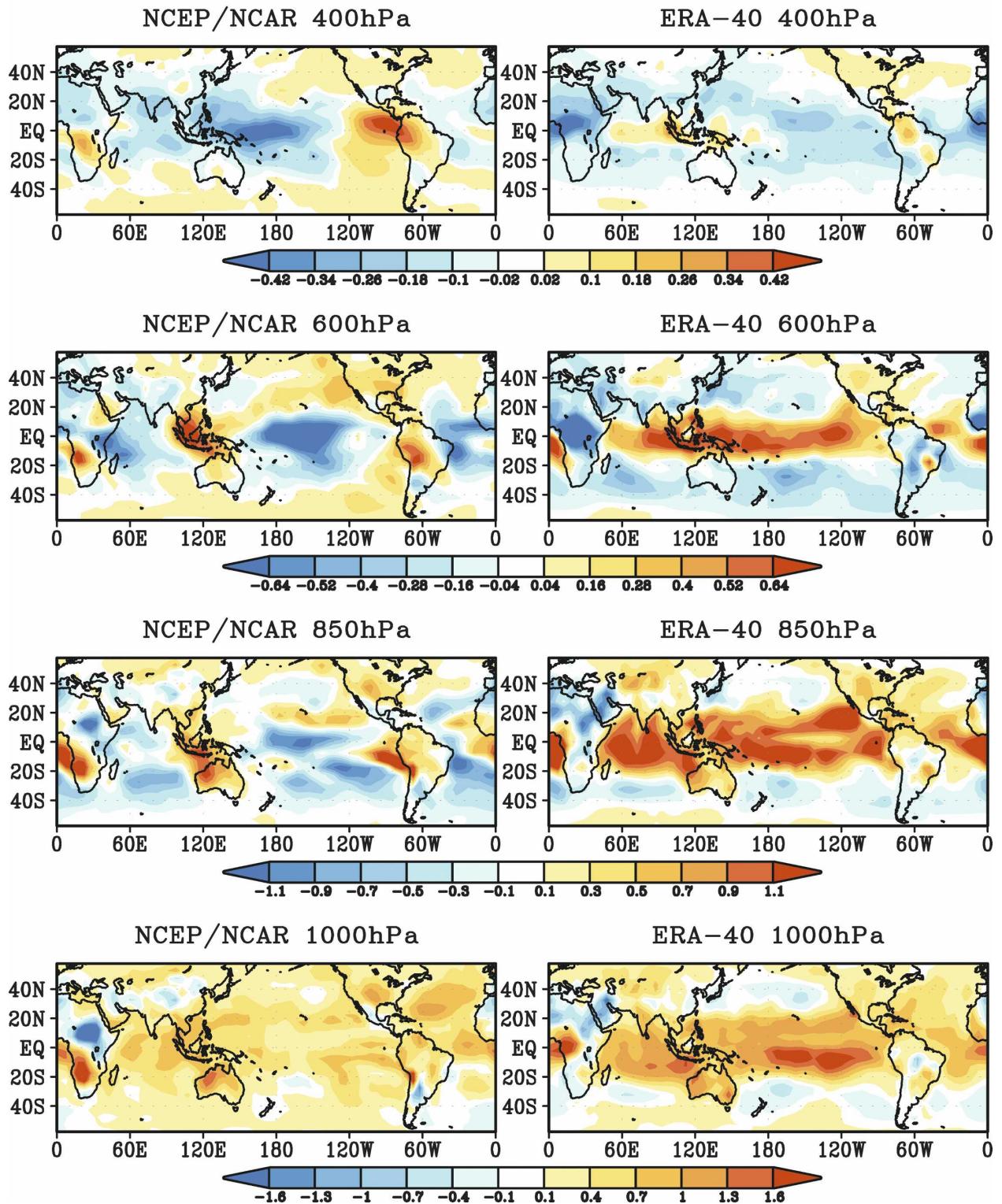


FIG. 7. The  $q$  variation at selected pressure levels from 1000 to 300 hPa associated with the GW trend mode. (left column) From NCEP1-40p analysis, based on NCEP-NCAR reanalysis data. (right column) From ERA40-40p analysis, based on ERA-40 reanalysis data.

associated with the anomalous intensification of the Hadley Circulation in the mid- to lower troposphere. Strong moistening happens in the  $\pm 20^\circ$  band, where the ascending branch of the Hadley Circulation is located; drying appears in the subtropical and midlatitudes, where the descending branches of the Hadley Circulation occur. From the 600-hPa to the upper-pressure levels, the tropical moistening feature weakens, and subtropical drying strengthens and expands into the tropics, except for the Indonesian and Amazon regions.

## 5. Conclusions and discussion

In this study, the GW trend in the twentieth century emerged as the dominant mode in the ENSO-removed EOF analyses based on ST datasets. The relatively uniform warming implies that the global long-term trend should be attributed to relatively evenly distributed radiative forcing. The amplification of that warming in the midhigh-latitude land regions is consistent with previous linear trend analyses (Ellsaesser et al. 1986; Hansen and Lebedeff 1987; Jones et al. 1999; Hansen et al. 1999) and anthropogenic greenhouse gas model simulations (Hegerl et al. 2007). Several embedded cooling spots that may be related to negative radiative forcing from sulfate aerosol, organic aerosol, soil dust, and land-use change (Hansen et al. 1998; Karl et al. 1995; Mitchell and Johns 1997) are also detected in our analysis.

Beyond these previously documented aspects of the twentieth-century long-term trend, we reveal another important characteristic of the GW trend: warming in the Pacific basin is much weaker than the warming at the east and west boundaries of the Pacific Ocean and in the other ocean basins. The modest warming in the Pacific basin is consistent with the long-term ocean heat content change (Levitus et al. 2000). We hypothesize that the weakened warming originates from the more dynamic nature of the Pacific Ocean on interannual and interdecadal time scales and/or the leakage of the upper ocean water through the Indonesian Throughflow. The failure of climate models to simulate these phenomena could explain why the apparent weakened warming in the Pacific basin has not appeared in the multimodel ensemble annual mean ST change based on different scenarios (Hegerl et al. 2007).

Changes in the atmospheric parameter fields associated with the GW trend have also been isolated from the NCEP-NCAR and ERA-40 reanalyses. The broad warming in both reanalyses at different pressure levels of the troposphere is consistent with an evenly distributed radiative forcing, and the increasing trend PC time

series are generally consistent with the PC time series based on century-time scale ST analyses.

GW-related dynamics and water vapor changes are coherent within each reanalysis, but significant discrepancies exist between the two reanalyses. The main atmospheric circulation change in NCEP1-40p is a modification in the Walker Circulation over the Pacific basin and its surrounding regions, with anomalous ascending moist branches over the Maritime Continent and east Pacific-Amazon regions and a descending dry branch over the central Pacific. This anomalous zonal overturning circulation is consistent with the large-scale zonal ST contrast caused by the weakened surface warming in the Pacific basin and the strong warming in surrounding regions. On the other hand, an anomalous intensification of the Hadley Circulation appears to be the major dynamical feature of the GW trend mode in the ERA40-40p, corresponding to strong moistening in the tropics and drying in the subtropics.

Besides the difference in the underlying models, which is not expected to have a large effect on long-term trends (Cai and Kalnay 2005), significant differences exist between NCEP-NCAR and ERA-40 in what satellite data are assimilated and how, and these probably cause the discrepancies in dynamics and water vapor fields. In the NCEP-NCAR reanalysis, only retrieved atmospheric temperature profiles are assimilated. In the ERA-40, the radiances from satellites are directly assimilated, including both temperature-sensitive and water vapor-sensitive channels. After 1987, the column-integrated precipitable water retrieved from Special Sensor Microwave Imager (SSM/I) microwave channels is also assimilated in ERA-40. That is, more water vapor data are included in the ERA-40 than in the NCEP-NCAR reanalysis, where the major water vapor data input is from radiosonde observations. Because the satellite data from different spacecraft are introduced at different times, the ERA-40 has a more severe discontinuity issue in humidity fields than the NCEP-NCAR reanalysis (appendix B; Figs. B1, B2). At the same time, the water vapor increment from the satellite data can artificially change the large-scale circulation when additional latent heat is released through precipitation (Bengtsson et al. 2004b).

The signal-noise ratio of long-term climate variations can be enhanced (Santer et al. 2001) by removing the known signals of short-term climate phenomena, especially ENSO. Compared to the previous ENSO-isolation attempts discussed in appendix A, our ENSO-isolation method keeps the simplicity of linear regression yet considers the lag relationship between the 6-yr

high-pass Niño-3.4 index (N34h) index and local time series. More sophisticated methods may be developed to address the nonlinear and asymmetrical aspects of ENSO. The GW trend time series obtained in this study is close to the trend time series in Kelly and Jones (1996). Although the trend time series in Penland and Matrosova (2006) is significantly different. The cause of this discrepancy is not clear, yet we notice that their ENSO signal includes long-term variation in the interdecadal time scale, traditionally not attributed to ENSO.

Combining multiple parameters in the EOF analysis helps to isolate the real climate signal in imperfect datasets, when the artifacts are not systematically distributed among parameters and across different regions. Aspects of the surface and atmospheric trend patterns that are common to both reanalyses and not associated with data discontinuities can be compared with model simulation results to confirm whether the GW trend is caused by anthropogenic forcing or natural variability and/or to test and improve the model capability to simulate climate change on regional scales. However, this study is based on annually smoothed anomaly data; thus, no seasonal information is included in the results. It is possible that long-term climate changes confined to specific seasons are not detected by our approach.

The weakened warming in the Pacific basin appears to be an important and consistent feature of the GW trend. We propose two possible mechanisms to explain this weakened warming in the Pacific basin. These hypotheses must be explored using model simulations emphasizing the long-term heat budget of the ocean and including realistic basin topography and proper simulation of the interannual and decadal variability. At the same time, other mechanisms may exist, for example, century-time scale random variabilities, which could be verified based on the output of ensemble runs of the coupled models with the capability to realistically simulate long-term climate variations. Until this weakened warming in the Pacific basin is understood and incorporated in coupled model simulations, predictions of twenty-first-century regional climate changes in surrounding continental regions should be viewed circumspectly.

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Reconstructed SST data, version 2 (ERSST.V2) were provided by the NOAA/CIRES Climate Diagnostics Center, Boulder, Colorado, from their Web site (<http://www.cdc.noaa.gov/>; NCEP-NCAR reanalysis data were downloaded from <http://www.cdc.noaa.gov/cdc/reanalysis/>). The ERA-40 monthly mean pressure-level analysis data were downloaded from the European Centre for Medium-Range Weather Forecasts Web site (<http://www.ecmwf.int/products/data/archive/descriptions/e4/index.html>).

## APPENDIX A

### ENSO Removal

#### *a. Introduction*

There have been several attempts to remove ENSO from the ST field (Kelly and Jones 1996). Jones (1988) regresses the high-pass-filtered hemispheric and global mean ST time series on the Southern Oscillation index (SOI) with SOI leading by 6 months, and then removes the regression result from the ST time series. Privalsky and Jensen (1995) use a bivariate autoregressive (AR) model to regress the global average air temperature on SOI. They find that more variance (about 30%) can be attributed to ENSO with the bivariate AR model than with a linear regression method (about 10%), which implies that ENSO affects surface air temperature with different time lags in different regions; thus, the global average temperature has a multilag relationship with SOI. Robock and Mao (1995) linearly regress the ST to SOI with no lag over the ocean and one season lag over land. They note that spurious small-scale patterns can be induced if a specific lag is applied to every grid box. Kelly and Jones (1996) identify the first and third modes in an EOF analysis of global ST as related to ENSO. They remove the ENSO signal based on the cross correlations between the SOI and the principal component of these two EOF modes. Santer et al. (2001) remove ENSO and volcanic signals in the global mean surface and lower-tropospheric temperature using an iterative procedure to account for the collinear relation between the ENSO and volcanic signals; they find more significant trends and better agreement between the observations and model simulations when the effects of ENSO are removed. Penland and Matrosova (2006) use a linear inverse model to isolate the ENSO signal in a broad spectral range, from interannual to interdecadal.

Recent advances in understanding the mechanisms of ENSO and its remote effects (Neelin et al. 1998; Trenberth et al. 1998; Wallace et al. 1998; Harrison and

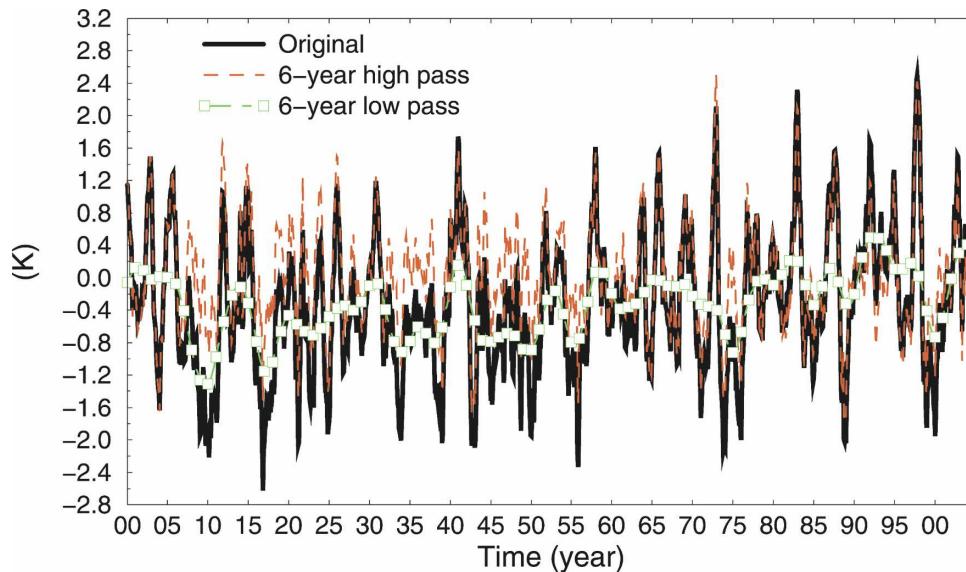


FIG. A1. Black solid: The time series of the Niño-3.4 index (N34h), which is defined as the averaged SST anomaly in the equatorial eastern Pacific region of  $5^{\circ}\text{N}$ – $5^{\circ}\text{S}$  and  $170^{\circ}$ – $120^{\circ}\text{W}$ . It is calculated from ERSST.V2. Red dashed: 6-yr high-pass-filtered N34h, used the same filter as in Zhang et al. (1997). Green dashed–dotted square: 6-yr low-pass-filtered N34h.

Larkin 1998; Klein et al. 1999; Trenberth et al. 2002; Alexander et al. 2002; Carleton 2003) lead us to suggest a more reliable ENSO-removal method at the grid box level, which we describe below.

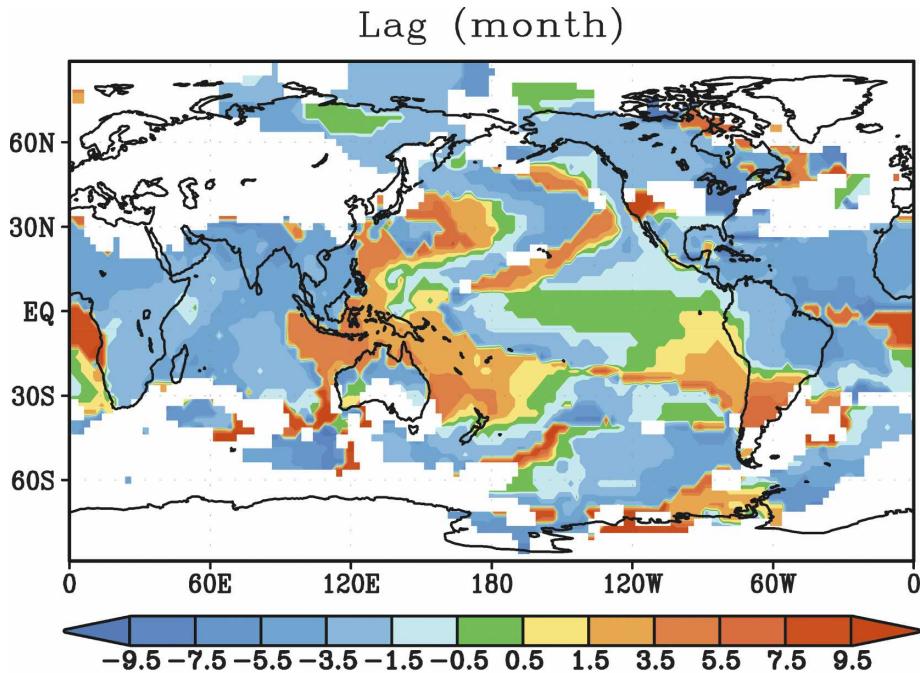
#### *b. The ENSO index and how ENSO affects remote regions*

The primary surface manifestation of ENSO is a seesaw between the west and east Pacific; thus, its phase and strength can be approximately depicted with an index. Hanley et al. (2003) comprehensively evaluate various indices and find strong similarities in the Niño-3, Niño-3.4, and Japan Meteorological Agency (JMA) indices. Trenberth et al. (2002) suggest that Niño-3.4 is useful for systematically exploring lead and lag relationships in the evolution of ENSO. We therefore calculate the time series of the Niño-3.4 index, defined as the averaged SST anomaly in the eastern equatorial Pacific region of  $5^{\circ}\text{N}$ – $5^{\circ}\text{S}$  and  $170^{\circ}$ – $120^{\circ}\text{W}$ , based on the SST field from ERSST.V2. Because ENSO is an interannual phenomenon, we remove the low-frequency (decadal and longer) variability (Fig. A1, green dashed–dotted square) with a 6-yr low-pass filter (Zhang et al. 1997) from the original Niño-3.4 index time series (Fig. A1, black solid line). Hereafter, we will use the resulting 6-yr high-pass Niño-3.4 index (referred to as N34h; Fig. A1, red dashed line) to represent ENSO temporal variability.

Previous studies (Bjerknes 1969; Opsteegh and Van

den Dool 1980; Hoskins and Karoly 1981; Webster 1981; Horel and Wallace 1981; Lau 1997; Trenberth et al. 1998; Wallace et al. 1998) address the fact that the tropical Pacific ENSO signal spreads to remote regions around the world through atmospheric teleconnection mechanisms. In the zonal direction, the anomalous deep convective activities in the central Pacific depress the regional circulation over the tropical Atlantic and Indian Oceans, resulting in anomalous heating and a delayed ST change (Lanzante 1996; Enfield and Mestas-Nuñez 1999; Klein et al. 1999; Alexander et al. 2002; Chiang and Sobel 2002). In the meridional direction associated with enhanced upper-troposphere divergence in the central-eastern Pacific, strong upper-troposphere convergence in the subtropics acts as a source of Rossby waves and thus induces teleconnection patterns, namely, the northeastward Pacific–North America (PNA) pattern and southeastward Pacific–South America (PSA) pattern (Carleton 2003). In these mechanisms, the atmosphere acts as a bridge to link the tropical Pacific to remote regions (Lau and Nath 1996; Klein et al. 1999). Associated water vapor (Klein et al. 1999), cloud (Fu et al. 1996), and wind (Enfield and Mayer 1997) variations affect the surface energy flux, adjusting the ST in remote regions over a few months. In the Pacific, ENSO-generated coastally trapped ocean waves (mainly Kelvin waves) propagate SST anomalies to high latitudes (Alexander et al. 2002).

From the top of Fig. A2, we see that ST is in phase



### Extreme Crosscorrelations at the Least Lag

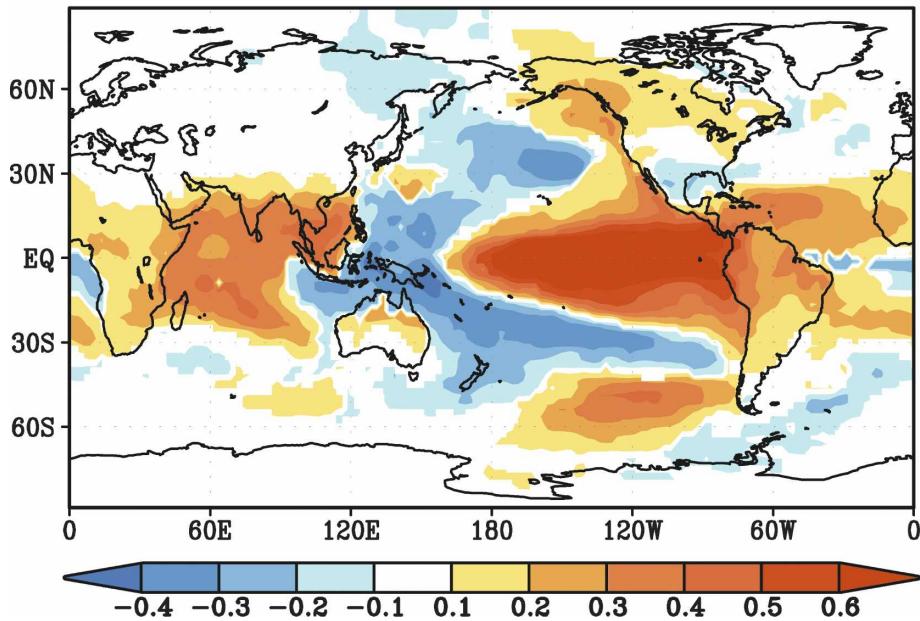


FIG. A2. The map of ECLL of the N34h index with ST. (top) The map of lags at which ECLL is found in  $\pm 12$ -month range. Negative value means the N34h index leads temperature. (bottom) The values of ECLL. In these plots, only the values in the grids for which the ECLL confidence level exceeds 99% are shown.

with the N34h index in the central-eastern tropical Pacific, where the strongest extreme cross correlations with the least lag (ECLL) appear (Fig. A2, bottom). In much of the extratropical Pacific and North America, ST lags N34h by 1–6 months and appears in patterns

that alternate positive and negative ECLL with N34h and deflect eastward from low latitude to high latitude (Fig. A2, bottom). These features are consistent with the PNA- and PSA-pattern responses to enhanced equatorial convection (see Fig. 3-3 of Chen 2005). In

the wave train, ascending motion and cooling occur in low-pressure cyclonic regions, and the opposite in high-pressure regions. The SST change is delayed relative to the atmosphere change by the larger thermal inertia of the ocean but has the same sign (Alexander et al. 2002), because Ekman transport associated with the cyclonic flow results in upwelling and a shoaling thermocline, and increased storminess intensifies ocean mixing and heat flux to the atmosphere (Deser et al. 1996).

In the northeast Pacific, there is a narrow region in which the ST appears to lead N34h, but the ECLL is small here and has little effect on the ENSO-removal technique described below. Near the west coast of Chile and Peru, ST also leads N34h, and here ECLL is high. This precursor feature has been mentioned by Rasmusson and Carpenter (1982). Enfield and Mestas-Nuñez (1999) attribute it to a surface heat flux change due to weaker trade winds off the Chilean coast, which precede the main Pacific trade wind weakening. Centered on the Maritime Continent, three regions in which ST leads N34h extend to the north, west, and southeast. These have also been observed by Trenberth et al. (2002).

ST over the Indian Ocean and South China Sea highly correlate with N34h at about 2–6 months' lag (Lanzante 1996; Enfield and Mestas-Nuñez 1999). El Niño causes anomalous subsidence there that reduces cloud amount; thus, more solar radiation reaches the surface and warms the ocean but with a delay due to its large thermal inertia. The smaller lag in the Indian Ocean is attributed to the fact that cloud changes in the Indian Ocean lead the ENSO index by 3 months because of the altered summer monsoons (Webster and Yang 1992; Lau and Yang 1996). In the South China Sea, the change in cloud amount coincides with the ENSO index (Klein et al. 1999).

In the tropical Atlantic Ocean, ST lags N34h by about 5–6 months (Lanzante 1996). The local Hadley cell weakens during El Niño, manifesting in a negative (positive) anomalous change near the intertropical convergence zone (in the subtropics) in water vapor, cloud amount, and reflected shortwave flux, although the surface latent heat flux change due to weakened trade winds seems more important in the tropical North Atlantic (Enfield and Mayer 1997; Klein et al. 1999; Alexander et al. 2002). In part of the Gulf of Guinea, ST anticorrelates with N34h, leading it by about 6 months. This feature is also consistent with Trenberth et al. (2002), although the reasons have not been addressed.

### c. ENSO-removal method

Because most of the lag and ECLL features in Fig. A2 can be explained by known ENSO and teleconnec-

tion mechanisms, the N34h index and the ECLL relationship between the N34h index and a given climate parameter provide a good approximate description of the ENSO effect on that parameter. Based on this, we can roughly remove the ENSO signal at the grid box level from the spatiotemporal field of that parameter. We first calculate the ECLL between N34h and the anomalies of climate parameter  $T$  at each grid box. For example, at grid box  $i$ , we obtain  $ECLL_i$  at lag  $j$ . Negative lag means that the N34h index leads the climate parameter for  $|j|$  months. We calculate the standard deviation of N34h, denoted as  $\sigma_{N34h}$ , and the standard deviation of  $T_i$ , denoted as  $\sigma_{T_i}$ . The residual value of  $T$  with ENSO-removed at grid box  $i$  and time  $m$ , denoted as  $Tres_{i,m}$ , is then calculated as follows:

$$Tres_{i,m} = T_{i,m} - N34h_{m+j} \times ECLL_i \times \sigma_{T_i} / \sigma_{N34h}$$

In the above equation, the second term on the right-hand side is the regression of  $T$  onto the N34h index for the least lag at which the extreme cross-correlation magnitude between  $T$  and N34h occurs. The removal procedure is applied only to grid boxes where  $ECLL_i$  is statistically significant above the 99% confidence level.

This ENSO-removal method is tested with the cross correlations for a  $\pm 12$ -month lag range between the N34h index and the principal component time series of the EOF analyses, based on GISTEMP and ERSST.V2 with or without the ENSO-removal procedure. In the EOF analyses of the original data (Fig. 3-6 in Chen 2005), the principal components of the first few modes significantly correlate with N34h at different lags. However, none of the principal components from the ENSO-removed data significantly correlates with the N34h index at any lag. In both the GISTEMP and ERSST.V2 ENSO-removed annually smoothed EOF analyses, the first two modes clearly represent global warming and the pan-decadal variability in the Pacific, respectively. The fact that these long-term climate variations emerge after the ENSO signal is removed suggests that the removal method is successful. The technique has also successfully been applied to extract known ENSO features in satellite precipitation datasets (Chen et al. 2007).

The ENSO-removal method is based on the broadly accepted assumptions that the ENSO phenomenon can be represented by the N34h index and that stable global linear lag relationships exist between N34h and climate parameters. However, we note that global lag relationships between the N34h index and the climate parameters can be modified when the climate background is shifted. For example, in Fig. 3-7 of Chen (2005), the lag relationships are compared before and after the 1976

climate shift. In the period 1958–75, the equatorial east Pacific leads the central Pacific; from 1979 to 1996, the equatorial east Pacific lags the central Pacific. Trenberth et al. (2002) hypothesize that a shift in the balance of terms within the ocean associated with the 1976 climate shift in the equatorial Pacific may have caused the change, but Wang and An (2002) attribute it to equatorial winds and an associated upwelling change after 1976. In addition, although ENSO and its effect on remote regions can be explained mostly based on linear dynamics, nonlinearity also plays a role (Trenberth et al. 1998), and asymmetry between warm and cold events needs to be considered (Larkin and Harrison 2002) in the future.

## APPENDIX B

### Effects of Data Discontinuities on Reanalyses

The reliability of reanalysis products for estimating long-term climate trends is considered here. Although there are deficiencies in the reanalysis models (Bony et al. 1997), the fixed-model strategy and constraints from the assimilated observations cause parameters with relatively more observations to be less affected by model problems. But because the model has inherent biases and the system is forced to fit the assimilated observations, changes in the observation system can introduce artificial climate change. The global observation system has experienced two major changes: the global radiosonde network established in 1958 and the global operational satellite observation system established in the 1970s (Kistler et al. 2001).

Several studies indicate weaknesses of the reanalysis products for global climate change studies. Kistler et al. (2001) address the fact that the NCEP–NCAR data pre-1958 have poorer quality and that the effect of the introduction of the global satellite observation system in 1979 is largest at or above 200 hPa and south of 60°S. Discrepancies in the lower stratosphere and upper troposphere before and after 1979 are caused by biases from both the NCEP model and NESDIS retrievals (Santer et al. 1999). Sturaro (2003) finds that the 1979 spurious shifts in the temperature fields are broader than earlier estimates: the largest changes occurred near 100 hPa, but the affected regions include the tropics between 700 and 50 hPa and the Southern Ocean region between 500 and 250 hPa. In Kinter et al. (2004), discrepancies in precipitation time series between observations and reanalysis in several tropical regions are attributed to the problematic divergent circulation in the reanalysis.

Through EOF analysis applied to individual or com-

bined parameter fields in different time ranges, we find that suspicious climate variations occur in earlier reanalysis data, especially before the 1970s. We also find that the effect of the 1979 discontinuity can be reduced if we add a large number of parameter fields to the combined EOF analysis. This is understandable because the essence of the EOF analysis is to draw the greatest possible commonality from the parameter fields sequentially under the orthogonal constraint. The impact of the 1979 discontinuity differs among parameters, geographical locations, and pressure levels (Sturaro 2003). Thus if we exclude the most affected regions, above 300 hPa and high latitudes, then the commonality of the data fields (i.e., the real climate variation) should emerge as the dominant mode of the combined EOF analysis.

A quantitative analysis of the discontinuities in the reanalysis data determines how successful this approach is in suppressing them. The discontinuity testing algorithm is based on the work of Rodionov (2004). This method is used to sequentially test the validity of a null hypothesis (in this case, the existence of discontinuity) for each value in a time series. If a value in the time series exceeds a criterion range, which is decided based on the  $l$  values earlier than the test value with the Student's  $t$  test method, then a regime shift index (RSI, which is a cumulative sum of the normalized anomalies relative to the range limit of the  $l$  values after the test point) at this test point is calculated. If the RSI becomes negative during the calculation, then the hypothesis is rejected, and the RSI is set back to zero. If the RSI remains positive, then the discontinuity at the test point is significant at the probability level set with the Student's  $t$  test. With this algorithm, we can obtain a discontinuity time series corresponding to a given time series. The discontinuity time series has the same time length as the original time series. In the discontinuity time series, a value of zero indicates that no discontinuity has occurred at that point, and a positive value indicates that a discontinuity has occurred. The value itself represents the strength of the discontinuity.

To obtain a general picture of the discontinuities in the reanalyses, we calculate the discontinuity time series for each grid box of the five parameters at each pressure level in the troposphere (1000–100 hPa) from 1948 (NCEP–NCAR) and 1958 (ERA-40). Figure B1 shows the mean discontinuity time series for each parameter of the two reanalyses. For NCEP–NCAR, discontinuities of  $T_{\text{air}}$  have almost double the range of strength compared with those of the other four parameters. Discontinuities for all parameters are relatively large before 1958. For  $U$ ,  $V$ ,  $\omega$ , and  $q$ , discontinuities are relatively small after 1970 except during the 1974 to

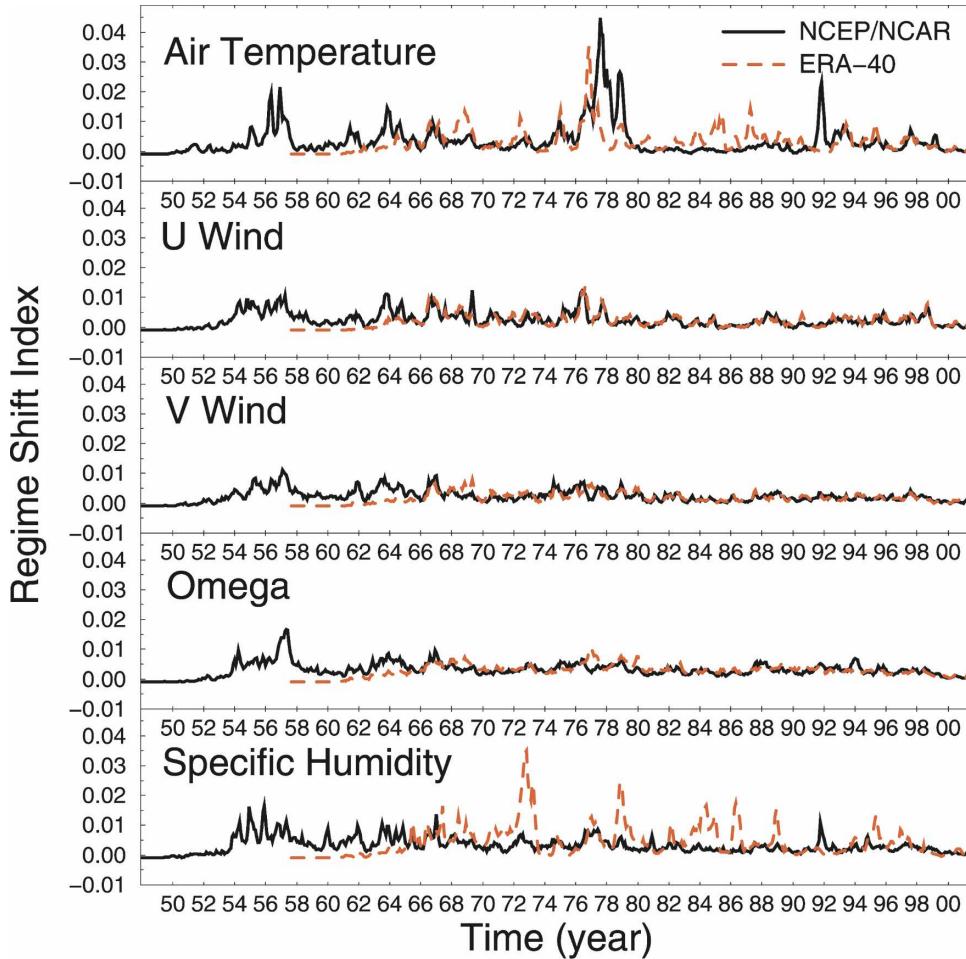


FIG. B1. The global, 1000–100-hPa averaged discontinuity (RSI index defined in Rodionov 2004) index time series for five atmospheric parameters ( $T_{\text{air}}$ ,  $U$ ,  $V$ ,  $\omega$ , and  $q$ ) from NCEP–NCAR (black solid) and ERA-40 (red dashed) reanalyses.

1980 period. Discontinuities in the ERA-40 dynamical fields are comparable to that of NCEP–NCAR, and the ERA-40  $q(T_{\text{air}})$  has more (fewer) high discontinuity values than its counterpart in NCEP–NCAR.

These discontinuities, especially around January 1979, are considered to be an obstacle to using reanalysis data in climate research (Chelliah and Ropelewski 2000; Sturaro 2003) due to the impact from the introduction of Television and Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) data. Figure B2 shows the zonal mean of the averaged discontinuity index from January 1978 to December 1979 for the five parameters in the vertical meridional plane. NCEP–NCAR  $T_{\text{air}}$  shows more than double the discontinuity value of the other four parameters in NCEP–NCAR. For all parameters except  $q$ , the discontinuity maximum occurs at 100 hPa in the tropics. The maximum discontinuity in  $T_{\text{air}}$  at 100 hPa extends to

200 hPa in the tropics; the maximum is around 300 hPa in the Southern Hemisphere; and there is also a weak local tropical maximum at 700 and 850 hPa.

In the three NCEP–NCAR dynamical parameter plots, discontinuities are relatively small compared to the  $T_{\text{air}}$ , but there are extreme regions around 60°S throughout the troposphere for  $V$  and  $\omega$ . Discontinuities in  $q$  have the same strength as those in the dynamical parameters, and the maximum occurs in the tropics at 300 hPa. This is the highest level for this parameter, but there are extremes near the surface around the equator and at 60°S. From the above plots, it is clear that the 1979 discontinuity is unevenly distributed. Excluding data above the 300-hPa level eliminates much of the discontinuity problem. The mid- to high latitudes in the Southern Hemisphere have a stronger discontinuity signal than other latitudes. To weaken the effect of the 1979 discontinuity, the analysis latitude range is

## Zonal Mean Discontinuity around 1979

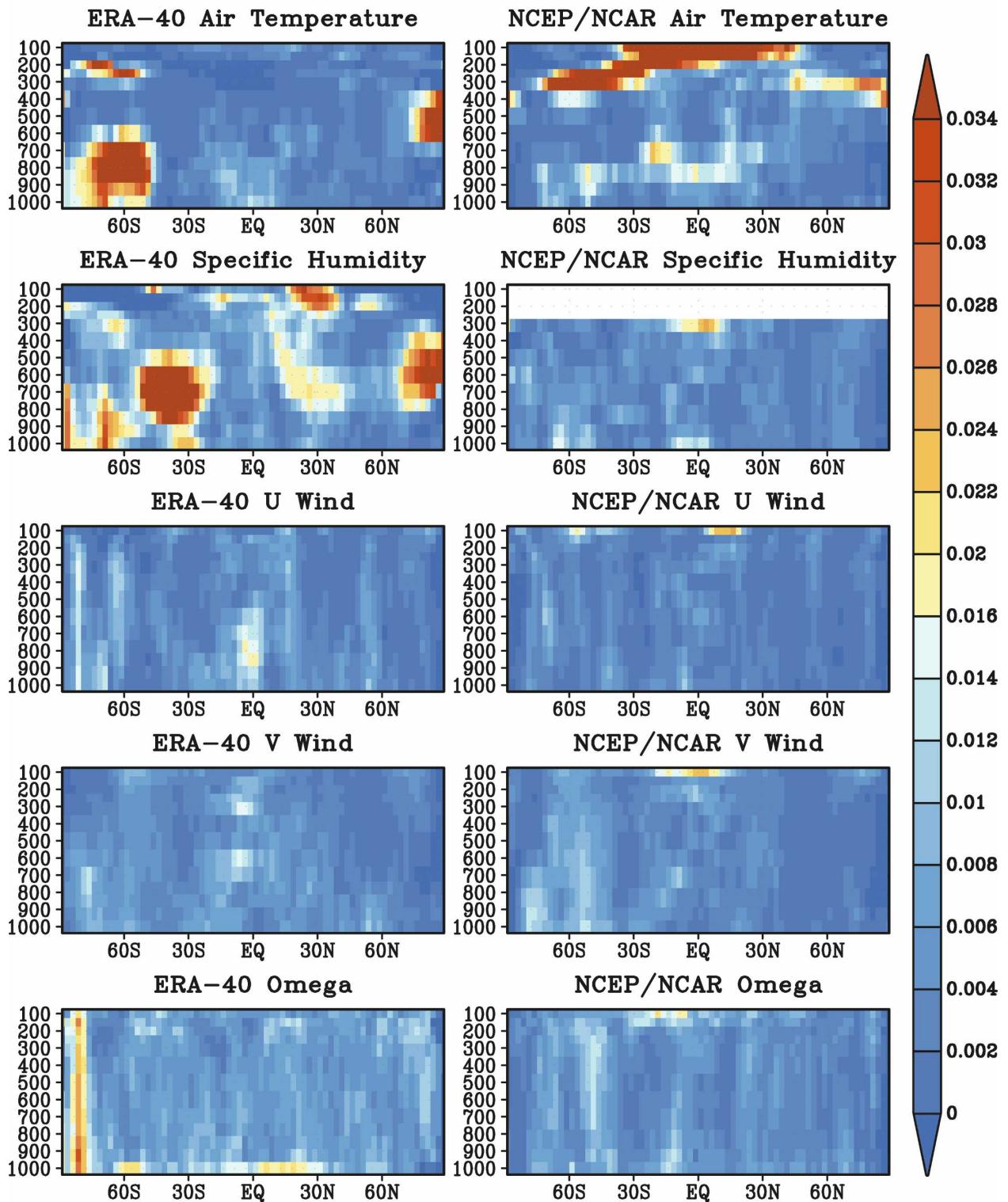


FIG. B2. The zonal mean discontinuity index averaged around 1979 (from January 1978 to December 1979) for five parameters from (right column) NCEP-NCAR and (left column) ERA-40. The parameters shown from top to bottom:  $T_{\text{air}}$ ,  $q$ ,  $U$ ,  $V$ ,  $\omega$ . The color bar shows the strength of the discontinuity; the horizontal axes are latitude; and the vertical axes are pressure levels.

confined to 60°S–60°N. Other significant discontinuity features located at lower-altitude (latitude) regions cannot be easily avoided and must be considered when interpreting results in the main text.

The ERA-40 has different discontinuity patterns, but these discontinuities are also not evenly distributed in space and among different parameters. Shown in the left column of Fig. B2, strong discontinuities are located at a high latitude in  $T_{\text{air}}$ ,  $q$ , and  $\omega$ . In the midlatitudes of the Southern Hemisphere, there is a strong discontinuity in  $q$ .

To determine how the introduction of satellite data around 1979 affects EOF analysis, we conducted an experiment based on the NCEP–NCAR reanalysis. First, EOF analysis was performed individually on each of the five parameters at each of the eight levels from 1000 to 300 hPa for 60°S–60°N from 1948 to 2001. We then calculated the discontinuity time series corresponding to the first 10 principal component time series for each EOF analysis. The mean discontinuity values from January 1978 to December 1979 corresponding to the PCs are shown in Fig. 4–4 of Chen (2005). For  $T_{\text{air}}$ , except at 1000 hPa, the 1979 discontinuity appears in all first-mode PCs. For  $q$ , the 1979 discontinuity appears in the first two modes at 400, 500, 600, and 850 hPa. For  $U$ , the 1979 discontinuity appears in the first two modes at 500, 600, and 850 hPa. For  $V$ , there is no sign of the 1979 discontinuity in the first two modes. For  $\omega$ , the 1979 discontinuity appears in the first two modes at 300, 400, and 700 hPa.

The 1979 discontinuity problem is largely overcome by a single combined EOF analysis for all five parameters at all eight pressure levels. In the combined EOF analysis for 1948 to 2001, a weak discontinuity appears in the second-mode PC around 1979. Besides the 1979 discontinuity, strong suspicious variability appears in the second and third PCs early in the time period. In the analysis based on the 1970 to 2001 period, no discontinuity appears in the first four-mode PCs around 1979. Thus, the combined EOF analysis can indeed suppress the 1979 discontinuity, because this discontinuity is not coherently distributed in reanalysis parameters.

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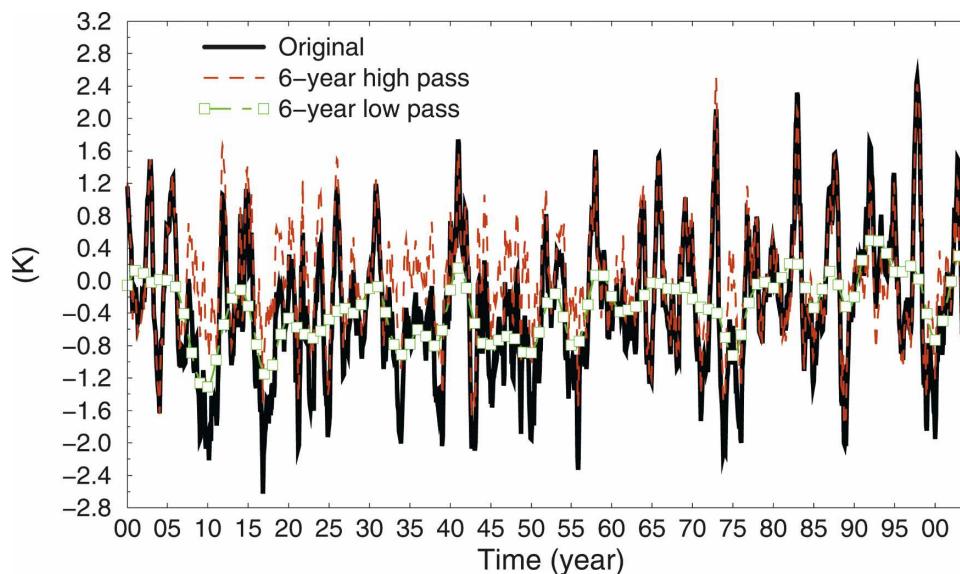


FIG. A1. Black solid: The time series of the Niño-3.4 index, which is defined as the averaged SST anomaly in the equatorial eastern Pacific region of  $5^{\circ}\text{N}$ – $5^{\circ}\text{S}$  and  $170^{\circ}$ – $120^{\circ}\text{W}$ . It is calculated from ERSST.V2. Red dashed: 6-yr high-pass-filtered Niño-3.4 index (N34h), using the same filter as in Zhang et al. (1997). Green dashed–dotted square: 6-yr low-pass-filtered Niño-3.4 index.

In Fig. A1, there was a typographical error in the figure caption. The correct versions of both of these figures appear here.

The staff of the *Journal of Climate* regret any inconvenience these errors may have caused.