A Statistical–Dynamical Estimate of Winter ENSO Teleconnections in a Future Climate

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ABSTRACT

Changes in the atmospheric response to SST variability in the decade 2065–75 are estimated from time-slice-like experiments using the NCAR Community Atmosphere Model, version 3 (CAM3) AGCM forced by specified SST and external forcing. The current climate is simulated using observed monthly SST and external forcing for 1951–2000. The change in mean SST for the future is represented by the difference between the 2065–75 and 1965–75 decadal mean SST climatologies from coupled model twentieth-century/future climate simulations of the response to external forcing. The change in external forcing is similarly specified as the change of the external forcing concurrent with the SST change. These seasonally varying changes in SST and external forcing are added to the 50-year sequence of 1951–2000 observed SST and external forcings to produce the specified future climate forcings for the AGCM.

Changes in the December through February mean ENSO teleconnections are evaluated from the difference between ensemble means from future and current climate time slice simulations. The ENSO teleconnections are strengthened and displaced westward in the time slice simulations, which is not in agreement with CGCM projections. These changes are associated with increased precipitation/atmospheric heating anomalies due to the warmer tropical SST. The quasigeostrophic stationary wave activity flux indicates that the dominant cause of the changes is a southward shift in a midlatitude central Pacific wave activity source rather than changes in the basic-state stationary wave dispersion properties.

1. Introduction

Future climate change may involve not only the climatological mean and annual cycle but also the characteristics and predictability of the climate variability. Changes in sea surface temperature variability in the tropical Pacific associated with ENSO, and their remote impacts, are of special interest since ENSO is a dominant mode of interannual variability that affects the global climate. Meehl et al. (2006a) found that El Niño teleconnections in the North Pacific and North America become weaker and are shifted southward and eastward in a warmer future climate in the Community Climate System Model, version 3 (CCSM3) and Parallel Climate Model (PCM) coupled general circulation models (CGCMs). They associated the weaker teleconnections with weaker future El Niño SST anomalies in the tropical Pacific projected by these models. Müller and Roeckner (2008) found similar changes in ENSO teleconnections in the ECHAM5/Max Planck Institute Ocean Model (MPI-OM) CGCM from twenty-second-century and 4 × CO₂ scenarios, although that model simulated future strengthening of ENSO SST anomalies. A multimodel comparison (Meehl and Teng 2007), which included the six coupled
atmosphere–ocean models—Canadian Centre for Climate Modelling and Analysis (CCCma) Coupled General Circulation Model, version 3 (CGCM3); CCSM3; ECHAM and the global Hamburg Ocean Primitive Equation (ECHO-G); Geophysical Fluid Dynamics Laboratory Climate Model version 2.0 (GFDL CM2.0); Goddard Institute for Space Studies Atmosphere–Ocean Model (GISS-AOM); GISS Model E-H (GISS-EH); GISS Model E-R (GISS-ER); Institute of Atmospheric Physics (IAP) Flexible Global Ocean–Atmosphere–Land System Model gridpoint version 1.0 (FGOALS-g1.0); L’Institut Pierre-Simon Laplace Coupled Model, version 4 (IPSL CM4); Model for Interdisciplinary Research on Climate 3.2, high-resolution version (MIROC3.2.hires); MPI ECHAM5; MRI CGCM2.3.2a; NCAR CCSM3.0; NCAR PCM1; Third climate configuration of the Met Office Unified Model (UKMO HadCM3) (cf. Table 1)—found that El Niño events are associated with weakening and eastward shifts of the Aleutian low, for models with both increasing and decreasing amplitude of tropical Pacific ENSO SST anomalies. Since changes in the climatological background state were similar for the six models, Meehl and Teng (2007) attributed the changes in the teleconnections to the effects of these background state changes on wave propagation from the tropical Pacific.

Projections of changes in the characteristics and influences of ENSO variability, however, are suspect not only because there is, as yet, no verification of the model projections but also because the models have large biases in the simulation of relevant aspects of the current climate (Bader et al. 2008) and produce a large range of tropical Pacific ENSO SST changes in response to similar external forcing scenarios (e.g., Guilyardi 2006). CGCM simulations of the “current climate” have biases in the mean tropical SST and the annual cycle of equatorial SST that are larger than the observed SST anomalies associated with ENSO. These SST biases are associated with biases in the atmospheric circulation such as the “double ITCZ” syndrome (two intertropical convergence zones in the eastern Pacific, typically with one in each hemisphere, rather than the observed single ITCZ north of the equator) symptomatic of errors in the atmospheric heating distribution, which probably cause biases in the climatological atmospheric circulation both within and outside of the tropics. Additionally, the amplitude, spatial scales, and time scales of the SST variability associated with current climate simulations of ENSO also have substantial biases, such that the properties of the simulated ENSO variability are outside the range of observed ENSO variability. While CGCM climatologies and ENSO variabilities have improved over the past 20 years, there are still substantial remaining biases (e.g., AchutaRao and Sperber 2006; Randall et al. 2007; Bader et al. 2008).

Given these problems, it is reasonable to assume that the properties of the observed variability of the current climate are closer to those that will be found in a (near) future climate than those projected by the models. This is not to say that the changes in the model-projected variability are unrealistic. However, there is no way to determine which model, if any, is projecting the correct changes. Similarly, it is reasonable to assume that some properties of the climatological SST that will occur in the future are better represented by the climatology of SST as observed in the current climate than by CGCM projections. These properties include, for example, the structures of the tropical SST gradients that determine the position of the ITCZ, SST in regions where ocean

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**Table 1. Models used to construct 2070 minus 1970 decadal mean change in SST.** The number of simulations from each model that were used to find the model average decadal change is given. A table with definitions and relevant references for all these models may be found on the PCMDI Web site at http://www-pcmdi.llnl.gov/projects/cmip/Table.php.

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dynamics are important, and the annual cycle of tropical SST. On the other hand, given the history and projections of increasing greenhouse gases (GHG) in the atmosphere for the coming decades, it is likely that the surface temperatures will warm; since the mechanisms for this warming are reasonably well understood, the models represent the relevant physics, and they agree on the general characteristics of this warming. Our approach to estimating the atmospheric response to SST variability in the future is then to represent the future climate by time-slice-like experiments with an AGCM forced by the observed SST, GHG, and aerosol evolution from the latter half of the twentieth century, with the projected change in climatological SST and GHG from coupled model projections superimposed.

2. Experimental design

A series of simulations of the current climate (1950–2000) have previously been carried out using the NCAR Community Atmosphere Model, version 3 (CAM3) AGCM with specified SST and external forcing and analyzed by Deser and Phillips (2009). To represent future climate, our simulations superimpose an annually repeating change in SST and external forcing on the same sequence of current climate forcing used by Deser and Phillips (2009). We also use the same AGCM. The annually repeating SST change is taken from CGCM transient simulations of current climate/projections of future climate, and the changes in the external forcing are taken from the A1b future climate scenario, which was used to force the CGCMs.

a. Model and verification

The experiments were performed with the stand-alone version of CAM3 (Collins et al. 2006) at T42 (triangular spectral truncation at total wavenumber 42) horizontal resolution, forced by specified SST and sea ice distribution. The simulations of some features of the climatology and interannual variability of CAM3 forced by observed SST are described by Hurrell et al. (2006) and Deser et al. (2006). The simulation of precipitation is more realistic in CAM3 (with specified SST) than in the CGCM CCSM3 (which uses CAM3 as the AGCM component), which has a prominent double ITCZ. This is due to the constrained SST in CAM3. However, there are still noticeable remaining biases in CAM3. Relevant to the analysis done in this study, which is restricted to December–February (DJF) means, there is a tendency toward a split ITCZ in the western Pacific. Northern Hemisphere stationary waves are shifted westward. ENSO-related precipitation anomalies in the tropical Pacific penetrate too far westward relative to the observed, with those at the equator changing sign at 110°E rather than 150°E, and with a bias of similar scale in the tropical Indian Ocean. In contrast, ENSO warm minus cold event composite extratropical Northern Hemisphere geopotential anomalies do not appear to have serious biases.

b. Forcing data

The 1950–2000 SST and sea ice dataset described by Hurrell et al. (2008) was used. The atmospheric forcing fields for 1950–2000, referred to as the twentieth-century climate simulation (20C3M) forcing, include time histories of GHG, aerosols (anthropogenic and volcanic), ozone, and solar variation. These are a subset of the forcings used by Meehl et al. (2006b) in a CCSM3 simulation of the twentieth century.

The SST change for the future climate time slice simulations is taken as the 2070 minus 1970 difference of the climatological SST from CGCM simulations forced by the reconstructed twentieth-century forcing and the Intergovernmental Panel on Climate Change (IPCC) Special Report on Emissions Scenarios (SRES) (Nakicenovic and Swart 2000) A1b future GHG and aerosol scenario. The CGCM simulations were carried out as part of the IPCC Fourth Assessment (AR4). An overview of the AR4 simulations is given in Randall et al. (2007, twentieth century) and Meehl et al. (2007, future climate scenarios). The SST from these simulations was obtained from the data archive maintained at the Program for Climate Model Diagnosis and Intercomparison (PCMDI).

The 2070 climatological SST is defined as the mean SST (as a function of month) averaged from 2065 through 2075. Similarly, the 1970 climatological SST is taken as the mean from 1965 through 1975. The change in the atmospheric forcing from 1970 to 2070 was constructed similarly, using the GHG forcing fields from Meehl et al. (2006b) for 1970 and the A1b scenario for 2070.

Two versions of the SST change were constructed, one from CCSM3 (also called SST1) and the other from multiple models (also called SST2). The CCSM3 version was averaged over 8-member twentieth-century and 6-member A1b simulations, the decadal means were computed, and the difference was taken. The multi-model version used ensemble members available from 15 CGCMs. The models, number of twentieth-century ensemble members, and number of twenty-first-century ensemble members are listed in Table 1. Each model was weighted equally for the 1970 and 2070 means before taking the difference.

The time evolution of GHG is shown in Fig. 1. This figure illustrates some of the forcings from the AR4 CGCM simulations, including the 20C3M forcing and the continuation into the A1b scenario. The concentrations and changes in the concentrations of these gases
The annual mean SST change between the 1970 and 2070 decades is shown in Fig. 2 for both CCSM3 and the multimodel ensemble. The two SST changes are broadly similar. However, the CCSM3 response has larger east–west variations in high latitudes, particularly in the Northern Hemisphere near 45°N where there is strong warming in the North Pacific and both strong warming and slight cooling in the North Atlantic. In the equatorial Pacific, the CCSM3 response decreases slightly from west to east (La Niña–like change), while the multimodel ensemble SST response suggests increasing SST from west to east (El Niño–like change). The CCSM3 response appears to be more similar to the observed 1979–2005 trend (Trenberth et al. 2007, Fig. 3.9), which has a La Niña–like structure. The multimodel SST response averages over a variety of behaviors, from the

La Niña–like response of CCSM3 to an El Niño–like response of ECHAM5/MPI-OM.

For all the experiments in the current study, sea ice was taken to be that used in the Deser and Phillips (2009) simulations.

c. Simulations

Ensembles of simulations of the effects of the 1970–2070 SST and GHG forcing changes were conducted using CAM3 by superimposing these changes on the 1951–2000 SST and atmospheric forcing evolution. Each ensemble has five members, and each member is a 50-year simulation. The five-member 1950–2000 T42 CAM3 ensemble of Deser and Phillips (2009), which includes the 1951–2000 SST and atmospheric forcing applied in our simulations, is taken as the control (denoted CONTROL). The CONTROL data used in the following is from 1951 to 2000. The effects of superimposing the SST and GHG forcing changes, together (denoted chgSST+GHG) as well as separately (denoted chgSST and chgGHG), were calculated. The changed SST and GHG simulations used the 1 January 1951 states of the Deser and Phillips (2009) ensemble members for initial states. Each case with changed SST was done with both the CCSM3 (SST1) and multimodel changes (SST2). Results from using SST1 and SST2 were similar, and these were combined into 10-member ensembles to improve confidence. The experiments are summarized in Table 2. These simulations comprise 1500 years of simulation, 1250 of which we performed to
3. Results

The object of the time slice experiments is to simulate and understand the changes in the ENSO teleconnections due to the projected SST and GHG forcing changes. Since the imposed future minus current climate SST changes have no variability other than an annual cycle and since these changes are superposed on the observed 1951–2000 SST evolution, the monthly SST anomalies in the CONTROL, chgSST + GHG, chgSST, and chgGHG simulations are all identical. They are also the same as those that occurred in reality and follow the observed time evolution. Therefore, the ENSO SST events in the future climate chgSST + GHG, chgSST, and chgGHG simulations are the same as those in CONTROL, and the effect of the imposed changes of SST and GHG on the ENSO teleconnections can be isolated, in principle, for each event.

Since our experiments consist of relatively small ensembles of simulations, we have chosen to use a composite analysis to improve the statistical confidence in the signals that are being calculated. A warm event, El Niño, composite evolution was constructed with respect to Decembers of mature events in 1957, 1965, 1972, 1977, 1982, 1987, 1991, 1992, 1994, and 1997. A cold event, La Niña composite, was similarly constructed with respect to Decembers of mature events in 1955, 1956, 1970, 1971, 1973, 1974, 1975, 1988, and 1998. The years chosen for the composites were taken from a consensus of categorizations using different indices (http://ggweather.com/enso/years.htm). The analysis is restricted to DJF to represent the season of largest impact of ENSO teleconnections in Northern Hemisphere midlatitudes. The results described below have been tested for significance using the t test.

The effects of the GHG changes alone on ENSO teleconnections were calculated both from chgSST + GHG minus chgSST and from GHG minus CONTROL. These two estimates did not agree and additionally were not significantly different from zero in the extratropics at the 5% level. Therefore, we concentrate on the effects of SST. The future climate chgSST + GHG and chgSST simulations are pooled into a single 20-member ensemble of simulations, called FUTURE, to estimate the changes in the ENSO response.

The field significance of the differences in the bottom panels of Figs. 3–6 was tested by computing the percentage area covered by the areas significant at the 5% level and comparing to the percentage area expected by chance (5%). The differences are significant by this test (reported in the figure captions). Spatial correlations are taken into account by estimating the number of degrees of freedom (DOF) of the field below which the field significance found would be attainable by chance with frequency greater than 5% (Livesey and Chen 1983, Fig. 3).

a. Changes in ENSO teleconnections

The warm minus cold event composite SLP (sea level pressure) for CONTROL, FUTURE, and FUTURE minus CONTROL are shown in Fig. 3. The low in the Aleutian region is intensified and extended slightly westward in FUTURE. Additionally, the high in the Newfoundland region is oriented more east–west rather than northwest–southeast and extends farther east across the North Atlantic. These changes are associated with a difference field with a low in the Aleutian region and a high extending from southern Greenland into central Europe. There is also a low over the eastern United States extending into the North Atlantic.
The corresponding behavior for the 300-hPa geopotential is shown in Fig. 4. The low in the Aleutians is intensified and extended westward in FUTURE. The high over northeast Canada is shifted northwestward (the change only marginally significant and not robust), and the low over the southern United States is intensified and extended eastward. A high is found near 0°E at 50°–70°N that appears to be related to the North Atlantic Oscillation (NAO) and east Atlantic patterns. In the tropics, the highs flanking the equator in the eastern Pacific are weakened, which is an indication of a weaker atmospheric ENSO response. In the Southern Hemisphere extratropics, the change in the ENSO response is primarily also a weakening of the signal.

The vertical structure of the extratropical geopotential for the simulated ENSO composite anomalies is equivalent barotropic (small vertical tilt), a robust property found in observations/analyses as well as simulations. The extratropical changes in Fig. 4 contain the same features noted with respect to the changes in the SLP teleconnections, consistent with an equivalent barotropic structure in the anomalous response.

The 2-m air temperature composites are shown in Fig. 5. There are warming differences extending east–west from Alaska to western Greenland. These differences intensify the ENSO warming response over Alaska and reduce the cooling response to the west of Greenland. Examination of the SLP responses indicates that these surface temperature responses are consistent with changes in the dynamical responses (e.g., the increase in the response of the southerly geostrophic wind between Alaska and western Greenland). In low latitudes, there is a warming response in isolated regions in East Africa, south-central Asia, and northern South America, and cooling in land areas near 30°S. The lower-latitude warming response differences in East Africa and northern South America are associated with negative precipitation responses (Fig. 6) and can be attributed to the combined effects of

Fig. 3. Warm minus cold ENSO event composite sea level pressure (hPa), (top) Current climate CONTROL, (middle) ensemble FUTURE of all simulations with future climate SST, and (bottom) difference [(middle) minus (top)]. Shaded areas are significantly different from zero at the 5% level. Contours levels are color bar boundaries, negative contours dashed. 28.5% of the area in the bottom panel is significant at the 5% level (DOF < 15).

Fig. 4. As in Fig. 3 but for 300-hPa height. In the bottom panel 36.3% of the area is significant at the 5% level (DOF < 15).
the implied decreased evaporation and cloudiness on the surface heat budget. The opposite situation applies to the cooling response differences.

The ENSO composite precipitation responses in CONTROL and FUTURE (Fig. 6, top and middle panels) both show the characteristic El Niño–like precipitation anomalies in the central Pacific representing an eastward shift of the region of strongest precipitation. The FUTURE minus CONTROL climatological precipitation also shows an El Niño–like response in the central Pacific. The most prominent precipitation differences in FUTURE relative to CONTROL (Fig. 6, bottom) are an intensification and an eastward/southward expansion in the warm minus cold event response in the equatorial central Pacific and a band of negative difference extending eastward from West Africa across Indonesia into the western Pacific. There are also bands of positive precipitation difference in the North Atlantic and North Pacific near 45°N, which may be indicative of changes in the storm track response. The tropical precipitation changes are associated with an approximately 30% increase of the warm minus cold event area-averaged precipitation difference in the region of largest anomalies: 10°S–10°N, 180°–230°W. This change has a substantial projection on the zonal-mean precipitation difference, which increases by 30% from +1.3 mm day$^{-1}$ between 5°S and 5°N in CONTROL to 1.7 mm day$^{-1}$ in FUTURE. The increase in the zonal mean difference is due to a stronger increase in warm event precipitation than that for cold events in FUTURE relative to CONTROL. The changes in precipitation would then suggest stronger forcing (heating) by tropical convection of both the ENSO composite stationary waves and the Hadley circulation. However, as is well known, the dynamical response to tropical heating will be approximately inversely proportional to the increase in tropical static stability. The latter increases in FUTURE.

The 300-hPa height difference (Fig. 4, bottom) response is opposite to that which might be expected from the above arguments regarding the difference in precipitation. A strengthening of the ENSO tropical heating
forcing would be expected to strengthen the anticyclones flanking the equator in the eastern tropical Pacific. Instead, the difference shows that the FUTURE anticyclones are weaker than those in CONTROL, which would suggest a weakening of the forcing of the ENSO teleconnections.

We have also examined the change in the nonlinearity or asymmetry in the DJF ENSO behavior. The nonlinearity is defined following Hoerling et al. (1997) as the sum of the composite El Niño and La Niña anomalies with respect to the climatologies of the various cases. We included all years, including ENSO years, in the climatology, and the nonlinearity was calculated for FUTURE and CONTROL. Although the years chosen for compositing were not the same as those in Hoerling et al. (1997), and their climatology did not include ENSO years, our results for CONTROL SST and precipitation are very similar to theirs for observations/analysis. The nonlinearity in the CONTROL SST anomaly is a weak El Niño–like pattern, as is the nonlinearity in the precipitation. Our 300-hPa height nonlinearity has a similar structure to their 500-hPa pattern and similar magnitudes in high latitudes. However, their nonlinearity is several times stronger in middle latitudes. In our experiments, the nonlinearity in precipitation increases in FUTURE relative to CONTROL. However, the nonlinearity in the 300-hPa height field becomes much weaker. This is found to result primarily from a change in the FUTURE La Niña pattern. We are not yet able to explain the decrease in the nonlinearity of the teleconnections in FUTURE.

We point out that an alternative view to a nonlinear ENSO response is that the climatologies of the ENSO and non-ENSO years differ. Referenced to the climatology of the ENSO years, the ENSO El Niño and La Niña composite anomalies are antisymmetric, the nonlinearity as defined above is zero in all fields, and the dynamics of the anomalies is mathematically linear (Schneider 1988). The question of why the ENSO response is nonlinear is then equivalent to asking why the climatology of ENSO events differs from that of non-ENSO events.

The choice of climatology can have a substantial influence on the anomalous behavior and its interpretation (Wu et al. 2008). However, while the nonlinearity depends on the definition of the climatology, the response (difference between El Niño and La Niña) is independent of this definition.

b. Attribution

By applying wave–mean flow interaction theory, the changes in the midlatitude ENSO teleconnections can be attributed to changes in forcing and to changes in the wave dispersion properties of the climatological mean flow. We examine both of these influences.

1) Changes in Tropical Forcing

Tropical upper-tropospheric divergence anomalies, essentially proportional to precipitation anomalies associated with deep convection, provide a possible forcing for Rossby waves to produce the extratropical ENSO teleconnections. The distribution of the ENSO composite tropical divergence anomalies at 200 hPa in the numerical experiments reflects the ENSO composite precipitation anomalies in the tropical Pacific to a good approximation, as shown in Fig. 7 for CONTROL and FUTURE. However, interpretation of the differences of FUTURE minus CONTROL is not so clear. There are many regions in the difference field where divergence anomalies not associated with precipitation anomalies are comparable in magnitude to those associated with the ENSO composite precipitation anomaly differences. Thus the divergence forcing for the FUTURE minus CONTROL differences is not clearly distinguishable from the response to that forcing. This is at least not at odds with the inference from the 300-hPa height that the tropical forcing of FUTURE ENSO is, if anything, weaker than in CONTROL.

2) Changes in Background State

Changes in the background state influences on wave propagation and dispersion have been suggested as important to explain the differences between current climate and projected future climate ENSO teleconnections. To diagnose these changes, Meehl et al. (2006a) and Meehl and Teng (2007) show differences between the simulated current and future climate 300-hPa streamfunction, while Müller and Roeckner (2008) shows the difference in the 300-hPa zonal wind.

The 300-hPa climatological zonal winds in CONTROL and FUTURE in our simulations are compared in Fig. 8. The difference (Fig. 8, bottom) is similar in distribution and magnitude to the DJF difference found by Müller and Roeckner (2008, Fig. 2b) for stabilized A1b minus current climate simulations. However, the extremes in the tropical/subtropical Pacific in their simulations are of a factor 2 more intense and are shifted eastward by about 40° longitude, probably due in part to corresponding eastward shifts of the tropical Pacific SST and heating anomalies. Both CONTROL and FUTURE have regions of westerlies between the date line and the west coast of South America more or less throughout the tropics (Fig. 8, top and middle). Since stationary free Rossby waves can propagate only in regions of westerlies (neglecting basic-state meridional velocity effects), near-equatorial forcing in this longitude band has
the possibility of producing a Rossby wave response that propagates into midlatitudes (Webster and Holton 1982). There is strong upper-tropospheric divergence in the ENSO composites in the equatorial Pacific to the east of the date line, with stronger forcing in FUTURE than CONTROL (Fig. 7), so a stronger extratropical response to the ENSO forcing might be expected in FUTURE. Changes in the region of easterlies suggest that the region of potential propagation might be shifted eastward in FUTURE relative to CONTROL since the zero wind line at the equator near the date line shifts about 10° longitude to the east in FUTURE and a region of easterlies near 10°N, 90°W found in CONTROL disappears in FUTURE.

An additional feature of interest in the structure of the zonal winds at 300 hPa is the occurrence of regions of reversal of the meridional gradient of absolute vorticity. These regions flank the subtropical jets, enclosing the zero wind line in the western tropical Pacific and the extratropical North Pacific (FUTURE), or regions of weak westerlies in the North Pacific (CONTROL) and the North Atlantic. This association probably has dynamical significance, in that it could result from mixing by stationary wave absorption/breaking at the critical lines (Andrews and McIntyre 1978; McIntyre and Palmer 1983), which could then lead to reflecting critical lines (Schneider 1990; Walker and Magnusdottir 2003).

The 300-hPa heights and changes for DJF climatology are shown shaded in Fig. 9. The changes in the climatology (shown shaded in Fig. 9, bottom, and as superimposed contours in Fig. 9, middle) show an eastward shift in the Northern Hemisphere stationary wave pattern. The changes in heights (proportional to the geostrophic streamfunction) are similar in structure to those found in the 300-hPa extratropical streamfunction by Meehl et al. (2006a) in CCSM3, and Meehl and Teng (2007) in a multimodel ensemble of coupled CGCMs, for the differences of the DJF climatologies of stabilized A1b minus twentieth-century stabilization simulations. The future climate changes in climatological 300-hPa height have structures similar to the trends in 500-hPa
heights in the twentieth-century CAM3 simulations but with a westward shift (comparing the shaded results in the bottom panel of Fig. 9 with the Deser and Phillips Fig. 2b). The composite ENSO response is mostly in quadrature with the climatological stationary waves with an eastward shift (e.g., Fig. 9, top, for CONTROL). The change in the ENSO teleconnections and the climatological stationary wave changes also tend to be in quadrature in the Northern Hemisphere extratropics (Fig. 9, bottom).

The effect of changes in the climate on barotropic Rossby wave propagation are illustrated in Fig. 10, which shows the stationary wavenumber for propagating Rossby waves at 300 hPa for CONTROL, FUTURE, and the difference FUTURE minus CONTROL, calculated using Eq. (2.13) of Hoskins and Ambrizzi (1993). This formula, a generalization of \( K_s^2 = a^2 (\beta - U y) / U \) using Mercator coordinates (\( K_s \) is the stationary horizontal wavenumber, \( a \) is the radius of the earth, \( \beta \) is meridional gradient of planetary vorticity, \( U \) is zonal velocity, and \( y \) is meridional distance), takes into account the zonal and meridional variation of the zonal velocity in a WKB sense (background state length scales large compared to anomaly length scales) but does not include the meridional velocity, which Hoskins and Ambrizzi claim is of secondary importance. The stationary wavenumber is imaginary in the blanked out regions owing to easterly zonal winds (the zero zonal wind contour is drawn) or negative meridional gradient of the absolute vorticity (cf. top two panels of Fig. 8). The most prominent stationary Rossby waves will tend to have wavenumbers near the local stationary Rossby wavenumber. The inviscid response to stationary forcing is resonant at the stationary wavenumber (Charney and Eliassen 1949), but Hoskins and Karoly (1981) show that the stationary wavenumber is also the scale of the response when sufficient damping is included so that resonance plays no role. Rossby waves are kinematically refracted toward higher stationary wavenumber; consequently, they will be trapped in tongues of maximum
stationary wavenumber, which will then act as waveguides. They will also be attracted toward critical lines and presumably absorbed, provided the absolute vorticity gradient remains positive and finite in the vicinity of the critical line. Surfaces where the absolute vorticity gradient is small/zero away from critical lines will tend to repel the stationary waves, while the behavior near critical lines where the absolute vorticity gradient also goes to zero depends on the ratio of two small terms.

In the Northern Hemisphere, Fig. 10 shows that in both CONTROL and FUTURE there is a waveguide (i.e., area of maximum in stationary wavenumber) that extends northeastward from the tropical Pacific near the date line across Mexico and the southeastern United States into Europe and eventually northern Asia (“Western Hemisphere waveguide”). There is another waveguide (“Eastern Hemisphere waveguide”) extending eastward from northern Africa to the central Pacific, where it bends northeast into western North America. There is also an equatorial waveguide in the eastern Pacific extending from the date line to South America. Similar features can be seen in the analysis of the observed 300-hPa DJF climatological flow in Hoskins and Ambrizzi (1993). The FUTURE minus CONTROL differences between the structures of the Northern Hemisphere waveguides are small. One change is a contraction of the North American extension of the Eastern Hemisphere waveguide in FUTURE compared to CONTROL.

It is clear from a visual comparison of Fig. 4 (top) and Fig. 10 (top) that the Northern Hemisphere midlatitude anomalies do not “fit” in the waveguides, which indicates that the WKB assumption of a slowly varying mean state is violated. Thus, it is difficult to interpret how the waveguide structures or the changes in the waveguides between FUTURE and CONTROL affect the propagation of disturbances forced by the ENSO SST anomalies. However, results presented in the next subsection support the view that the propagation structures shown by the waveguide diagnosis are physically meaningful.

3) WAVE ACTIVITY DIAGNOSIS

A diagnostic of stationary wave activity flux (Takaya and Nakamura 2001) is applied to examine the forcing and propagation of the ENSO composite anomalies. The diagnostic extends the three-dimensional version (Plumb 1985) of the Eliassen–Palm flux diagnostic (Edmon et al. 1980) to include nonzero meridional velocities in the basic flow. This diagnostic, which we denote as $\mathbf{F}$, is a three-dimensional vector that can be interpreted in the context of linear quasigeostrophic dynamics. The horizontal projection of $\mathbf{F}$ is parallel to and in the direction of the local stationary Rossby wave group velocity (perpendicular to wave crests) in the WKB limit of a slowly varying background flow. Nonzero $\mathbf{V} \cdot \mathbf{F}$ indicates sources (positive divergence) and sinks (negative divergence) of a scalar quantity $A$ related to wave activity. The sources and sinks occur owing to wave–mean flow interaction, nonlinearity, transient forcing, and nonconservative physics such as heating and friction. The spherical coordinate version of $\mathbf{F}$ in log pressure coordinates, given as Eq. (38) in Takaya and Nakamura (2001), but with the obvious correction of using a latitudinally varying Coriolis parameter instead of a constant one, is used here. The formula we applied is given in the appendix.

The horizontal projection of $\mathbf{F}$ at 300 hPa, $\mathbf{F}_{300}$, which is related to the horizontal stationary wave momentum flux, and $\mathbf{V} \cdot \mathbf{F}$ at this level, are shown in Fig. 11. The data used for the analysis was gridded pressure level model output at 200, 300, and 500 hPa. The basic-state horizontal velocity used was the average of the 300-hPa flows during the winters chosen for the ENSO composite events. The quasigeostrophic analysis breaks down in low latitudes, leading to very large amplitude and meaningless results for $\mathbf{F}$. Values of both quantities outside of the region shown (excepting the excluded region within 7° of the equator) are small compared to those shown in Fig. 11.

The stationary waves associated with ENSO in both CONTROL and FUTURE have similar gross characteristics. They appear to originate in an extensive source region in the central North Pacific between 30° and 45°N, indicated by strong divergence of $\mathbf{F}$. Since the simulated stationary waves are equivalent barotropic in midlatitudes with maximum amplitude near 300 hPa, the vertical component of $\mathbf{F}$, related to the horizontal heat flux of the stationary waves, makes a secondary contribution in the evaluation of the divergence. That is, $\mathbf{V} \cdot \mathbf{F}_{300} \approx \mathbf{V}_h \cdot \mathbf{F}_{300}$, where $\mathbf{V}_h$ is the horizontal divergence. Flux vectors also appear to originate in low latitudes in the central Pacific north of the strong ENSO precipitation anomalies. However, due to the obvious breakdown of the quasigeostrophic diagnosis near the equator for $\mathbf{F}$, the implied equatorial source could be spurious (note that the consequences of this breakdown are not so obvious for the stationary wavenumber diagnosis in Fig. 10). The vector $\mathbf{F}_{300}$ originating in the North Pacific takes two distinct paths: a northern branch arcing to the northeast across northwest North America, then turning southward, and a southern branch starting to the southeast into the tropics, then curving to the northeast to cross the southern United States and northern Mexico. The flux from the deep tropics appears to join the southern branch. The flux weakens following each branch eastward, and the northern and southern branches merge over the central/southern United States.
FIG. 11. Horizontal wave activity flux (arrows, m$^2$ s$^{-2}$) and divergence of 3D wave activity flux (shaded, contour interval 10$^{-6}$ m s$^{-2}$) at 300 mb for ENSO composite in (top) CONTROL, (middle) FUTURE, and (bottom) FUTURE minus CONTROL. Arrows are not plotted if they are shorter than 0.1 m$^2$ s$^{-2}$ or originate from latitudes south of 7°N.

FIG. 12. Horizontal wave activity flux (arrows, m$^2$ s$^{-2}$) for the ENSO composite, as in Fig. 11, and stationary wavenumber (shaded) for the winds averaged over the years in the ENSO composites, both at 300 hPa, in (top) CONTROL, (middle) FUTURE, and (bottom) FUTURE minus CONTROL (fluxes), FUTURE (wavenumber).
The difference FUTURE minus CONTROL shows that the directions of $\mathbf{F}_{300}$ are similar in both, so the arrows in the bottom panel are approximately parallel to those in the top two panels. The picture is one of a strengthening of $\mathbf{F}_{300}$ in FUTURE in the northern branch and a weakening in the southern branch. Thus the $\mathbf{F}_{300}$ difference arrows have characteristics of a clockwise closed circulation (i.e., a nondivergent flux) but with strengthening and weakening along the path associated with sources and sinks. The change over North America is large but not robust. One speculative interpretation is that the origin of the circulation is the FUTURE increase in the flux from the tropics, and this stimulates an increase in the midlatitude source. A secondary feature is a FUTURE weakening of a northeast component of $\mathbf{F}_{300}$ between the Great Lakes and Greenland. This change is large; however, it is not robust.

The wave activity flux diagnosis suggests that the ENSO response is forced primarily in middle latitudes, and the response is split into two wave trains: one starting toward the northeast and then curving back to toward lower latitudes and the other starting southeast. These properties are similar to the results of Hoskins and Karoly (1981) of the linear stationary wave response to idealized subtropical and midlatitude sources. Their diagnosis provides a dynamical explanation of some of the above features, using ray tracing and Wentzel–Kramers–Brillouin–Jeffries (WKBJ) analysis (i.e., following the group velocity in a slowly varying basic state so that their results should then be more or less equivalent to the wave activity flux diagnosis). The low stationary wavenumber part of the response propagates initially poleward and follows great circles (northern branch), appearing to curve back toward lower latitudes in a Mercator projection and remaining trapped on the poleward side of the jet stream. The high stationary wavenumber part of the response starts equatorward (southern branch). However, in their simulations, the southern branch turns toward the equator, whereas in our simulations the southern branch turns poleward, possibly merging with the northern branch.

Figure 12 superposes $\mathbf{F}_{300}$, as in Fig. 11, on the stationary wavenumber, as in Fig. 10 (but computed using the mean of the winters used in the ENSO composites). We see that in both FUTURE and CONTROL, the ENSO composite $\mathbf{F}_{300}$ follows the northern (low wavenumber) and southern (high wavenumber) waveguides. The major differences from the Hoskins and Karoly (1981) picture are due to the zonally varying basic state. The southern waveguide is located between the North American jet to the north and a region of small/reversed meridional absolute vorticity gradient to the south, which gives the southern waveguide a southwest–northeast tilt.

The main FUTURE minus CONTROL differences in Fig. 12 look like an intensification in the northern branch of the ENSO response, possibly connected to an increase in the flux from the tropical central Pacific, with little change in the southern branch. The increased flux in the northern branch follows the same path in all three panels. It is not clear to us, however, why the flux in the southern branch does not intensify, while that in the northern branch does.

4. Conclusions

The potential effects of climate change on ENSO teleconnections have been investigated using a hybrid AGCM simulation approach. An annual cycle climate change SST signal superposed on the observed monthly time varying SST from the latter half of the twentieth century was used to represent future climate SST and its variability. Similarly, a time-independent change of CO$_2$ and other “externally specified” greenhouse gases was superposed on the observed time varying greenhouse gases from the latter half of the twentieth century to represent future greenhouse gases. AGCM simulations forced by the specified changed climate SST and greenhouse gases were compared to current climate simulation. The specified changes in greenhouse gases were taken as 2070 minus 1970 decadal averages from the A1b scenario. The specified changes in SST were 2070 minus 1970 decadal averages from CGCMs forced by the 20C3M/A1b scenario.

The motivation for this approach is that the ENSO SST variability in the future climate simulations is identical to that observed for the climate of the latter half of the twentieth century. Therefore the differences in the ENSO teleconnections result only from changes in the climate caused only by the constant specified SST and GHG changes. Changes in the properties of the ENSO SST variability or, for that matter, changes in other types of SST variability are eliminated from consideration. Additionally, the structures, amplitudes, and evolution of the ENSO SST anomalies (as well as all other SST anomalies) are as observed, and hence are realistic, at least for the current climate. This contrasts with coupled-model-simulated ENSO variability (as well as the simulated SST climatology and other SST anomalies), which has substantial biases compared to current climate. Given these model biases, coupled model simulations of future ENSO variability and its changes from current variability are probably also less realistic than in our approach. Thus, it is reasonable to expect that current observed ENSO SST variability is a better estimate of future variability than that obtained by coupled model projections.
In contrast to the results from coupled GCMs (Meehl et al. 2006a; Meehl and Teng 2007; Müller and Roeckner 2008), which simulate a weakening and eastward shift in the Aleutian low, the change in the simulated future ENSO teleconnections over the North Pacific and North America is an intensification and westward extension in the pattern relative to simulations of the present. This change in the pattern is associated with intensification of the ENSO surface temperature anomalies over North America—warmer in the north and cooler in the south. The ENSO SST anomalies produce a stronger precipitation response in the future simulations, with upper-tropospheric divergence anomalies in the deep tropics stronger in the central Pacific and convergence stronger near Indonesia. However, the relationship between the changes in precipitation and changes in tropical upper-tropospheric divergence is not strong, obscuring the difference between “forcing” and “response.”

Some changes in the structure of the climatology between the future and present climate simulations were noted. The wave propagation properties of the climatology, including location of zero zonal wind lines, regions of mixed vorticity, and stationary wavenumber, do not, however, show much change between present and future.

A wave activity flux diagnosis indicates instead that the changes in the teleconnections are due to changes in the forcing distribution in the subtropical/midlatitude central North Pacific. The wave activity diagnosis also suggests a tropical or subtropical source connected to this forcing change, but the diagnosis breaks down close to the equator, exactly the region of most interest, because of the breakdown of the quasigeostrophic assumption there, so it is not clear if this result is reliable. The wave activity paths are not changed much, in agreement with the lack of change in the waveguide structure. Our diagnosis of the cause of the changes in the ENSO teleconnections, therefore, contrasts with the conjecture of Meehl and Teng (2007) and Müller and Roeckner (2008) that the differences can be attributed to differences in the background state.

We have not identified the cause of the different changes in ENSO teleconnections between our results and those obtained from CGCM projections, although this would be interesting to pursue. While we have not yet identified the source of the change in the wave activity flux forcing, it seem likely that this change is associated with a change in the location and/or intensity of transient eddy activity (the storm track). Another issue, which we have not addressed, is how the statistics and predictability of the ENSO teleconnections might change in a future climate. While beyond the scope of the present study, the experimental design and dataset developed here are appropriate for exploring these issues.

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APPENDIX

Wave Activity Flux

The stationary wave activity flux diagnostic, $F$, in spherical pressure coordinates after Takaya and Nakamura [2001, Eq. (38)] is

$$F = \frac{p \cos \phi}{2 |U|} \left\{ \frac{U}{a^2 \cos^2 \phi} \left[ \left( \frac{\partial \psi'}{\partial \lambda} \right)^2 - \psi' \frac{\partial^2 \psi'}{\partial \lambda^2} \right] + \frac{V}{a^2 \cos^2 \phi} \left( \frac{\partial \psi'}{\partial \phi} \frac{\partial \psi'}{\partial \phi} - \psi' \frac{\partial^2 \psi'}{\partial \phi^2} \right) \right\}.
$$

(A1)

In this formula, $\lambda$ is longitude, $\phi$ is latitude, and $p$ is pressure; $z = -h \ln(p/p_0)$, where $h = RT_0/g$, $p_0$ and $T_0$ are a reference pressure and temperature, $R$ is the gas constant, and $g$ is gravity. Other parameters are the radius of the earth $a$; the Coriolis parameter $f = 2 \Omega \sin \phi$, where $\Omega$ is the earth’s angular speed of rotation; and the buoyancy frequency $N$. The basic-state horizontal velocity is $U = (U, V)$. The geostrophic perturbation
streamfunction is $\psi' = \Phi f$, where $\Phi$ is the geopotential. In applying Eq. (A1), the zonal mean of $\psi'$ is set to zero to control the amplitude of $F$ near the equator.

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