Nonlinearity of the combined warm ENSO and QBO effects on the Northern Hemisphere polar vortex in MAECHAM5 simulations

Natalia Calvo,[1,2] Marco A. Giorgetta,[3] Ricardo Garcia-Herrera,[1] and Elisa Manzini[4,5]

Received 10 November 2008; revised 6 February 2009; accepted 5 March 2009; published 14 July 2009.

[1] The influence of the quasi-biennial oscillation (QBO) on the Northern Hemisphere (NH) polar vortex response to warm El Niño–Southern Oscillation (ENSO) events and the impact of the warm ENSO events on the QBO signal in the NH polar stratosphere have been analyzed using the Middle Atmosphere ECHAM5 model. The experiment setup was designed to include simulations of extended NH winter seasons for either strong easterly or strong westerly phases of the tropical QBO, forced with either sea surface temperatures (SSTs) from the strong ENSO event that occurred in 1997/1998 or with climatological SSTs. It has been found that the weakening and warming of the polar vortex associated with a warm ENSO are intensified at the end of the winter during both QBO phases. In addition, the westerly QBO phase delays the onset of the warm ENSO signal, while the easterly QBO phase advances it. Warm ENSO events also impact the extratropical signal of the QBO by intensifying (weakening) the QBO effects in early (late) winter. Therefore, it appears that during warm ENSO events the duration of QBO signal in the northern extratropics is shortened while its downward propagation accelerated. Our dynamical analysis has revealed that these results are due to changes in the background flow caused by the QBO combined with changes in the anomalous propagation and dissipation of extratropical waves generated by warm ENSO. In both cases, a nonlinear behavior in the response of the polar vortex is observed when both warm ENSO and the easterly phase of the QBO operate together. These results suggest that the Arctic polar vortex response to combined forcing factors, in our case warm ENSO and the QBO phenomena, is expected to be nonlinear also for other coexistent forcing factors able to affect the variability of the vortex in the stratosphere.

Citation: Calvo, N., M. A. Giorgetta, R. Garcia-Herrera, and E. Manzini (2009), Nonlinearity of the combined warm ENSO and QBO effects on the Northern Hemisphere polar vortex in MAECHAM5 simulations, *J. Geophys. Res.*, *114*, D13109, doi:10.1029/2008JD011445.

1. Introduction

[2] The first evidences of the quasi-biennial oscillation (QBO) signal in the extratropical northern stratosphere were presented by *Holton and Tan* [1980, 1982]. Eventually, other works confirmed these findings with both models and observations [e.g., *Naito and Hirota*, 1997; *Calvo et al.*, 2007] and showed that a stronger (weaker) polar vortex accompanied by a colder (warmer) polar stratosphere appear during the westerly (easterly) phase of the QBO. The alternation of easterly and westerly winds in the tropics and subtropics due to the QBO seems to modify the effective channels where the extratropical waves propagate and the regions where they dissipate, changing the background flow and affecting the polar vortex.

[3] Although El Niño–Southern Oscillation (ENSO) is mainly a tropospheric phenomenon, it also has an effect on the polar winter stratosphere. The first studies regarding the ENSO signal in the stratosphere were based on observations and were largely inconclusive [*Wallace and Chang*, 1982; *van Loon and Labitzke*, 1987; *Hamilton*, 1993; *Baldwin and O'Sullivan*, 1995]. This was mainly due to the difficulty in isolating the ENSO phenomena from other signals, especially from the QBO, as their phases tend to coincide (warm ENSO events are observed to coincide with easterly QBO phases). In addition, the major volcanic eruptions of Mt. Agung in 1963, El Chichon in 1982, and Mt. Pinatubo in 1991 coincided with warm ENSO events. Therefore, general circulation model (GCM) experiments, capable of isolating the ENSO signal from those generated by other sources of variability became one of the few tools available to study the ENSO effect in the stratosphere. These works showed that strong, warm ENSO events generate a weaker and warmer polar vortex through the anomalous upward propagation and dissipation of Rossby waves which accelerate the residual

[1]Departamento Fisica de la Tierra II, Universidad Complutense de Madrid, Madrid, Spain.
[2]National Center for Atmospheric Research, Boulder, Colorado, USA.
[3]Max Planck Institute for Meteorology, Hamburg, Germany.
[4]Instituto Nazionale di Geofisica e Vulcanologia, Bologna, Italy.
[5]Centro Euro-Mediterraneo per i Cambiamenti Climatici, Bologna, Italy.

0148-0227/09/2008JD011445

circulation in the stratosphere [*Sassi et al.*, 2004; *Manzini et al.*, 2006; *Garcia-Herrera et al.*, 2006]. However, a significant signal in the polar stratosphere has not always been detected for cold ENSO events [*Manzini et al.*, 2006]. Recently, *Camp and Tung* [2007] did establish the statistical significance of the ENSO perturbation at high latitudes using observations without having to stratify the data according to QBO phase and their results agree with previous analysis from model outputs.

[4] While the majority of the works mentioned before focused on disentangling the QBO and ENSO influences on the polar stratosphere, not many studies have analyzed the combined effect of both phenomena when operating together or the possible influence of one on the other. Very recently, *Garfinkel and Hartmann* [2007] used ERA-40 data composites for ENSO months and showed that the effect of the QBO or ENSO in the Arctic region seemed weaker when coinciding with warm ENSO or east QBO phase events, respectively; although their results were extremely sensitive to the months included in the composites. *Wei et al.* [2007] obtained a similar conclusion when looking exclusively at the extratropical QBO signal during warm and cold ENSOs through a regression analysis and suggested that the anomalous propagation of planetary waves at high latitudes are responsible for this behavior. Dynamical models, however, are still needed to go deeper into the actual combined effect in a comprehensive way and investigate the dynamical mechanisms involved.

[5] In this paper, the GCM MAECHAM5 has been used to design an experiment that considers strong westerly and easterly QBO phases during one single extreme warm ENSO event. We analyze the QBO impact on the warm ENSO signals in the polar stratosphere and the warm ENSO influence on the NH extratropical QBO effects. We focus on the NH during boreal winter months when previous studies have characterized the ENSO and QBO effects when any of these phenomena operated independently [*Garcia-Herrera et al.*, 2006; *Manzini et al.*, 2006; *Calvo et al.*, 2007]. Section 2 presents the model and experimental design, the main results are described in section 3 and section 4 summarizes.

2. Experiments

2.1. MAECHAM5

[6] The ECHAM5 general circulation model [*Roeckner et al.*, 2003, 2006] is used in this study in the Middle Atmosphere configuration MAECHAM5 [*Manzini et al.*, 2006] that resolves the atmosphere vertically from the surface to 0.01 hPa (or 80 km) in the mesosphere. The spectral transform method is employed for the horizontal dynamics, using a triangular truncation in spectral wave number space. Nonlinear dynamical terms, as well as transport and physical parameterizations are computed on an associated Gaussian longitude latitude grid. For this study the horizontal dynamics is truncated to wave number 42 and the longitude latitude grid has a resolution of 128×64 points, or about 2.8°. The vertical resolution chosen here has 90 vertical levels with approximately 700 m vertical resolution between the middle troposphere and 42 km height and better than 1 km resolution up to the stratopause [*Giorgetta et al.*, 2006, Figure 1]. At this vertical resolution, the model is able to internally simulate a realistic QBO in the tropical stratosphere by both resolved and parameterized wave–mean flow interaction [*Giorgetta et al.*, 2002, 2006]. The model also reproduces the expected Holton and Tan relationship in the Northern Hemisphere high latitudes [*Calvo et al.*, 2007]. For the simulation used here, the designated control simulation (CT simulation), sea surface temperature (SST) and sea ice distribution are prescribed following the monthly mean climatology of the period 1979–1996, thus excluding any direct effect from ENSO. Ozone concentrations are prescribed as climatology and any external interannual forcing as the 11-year solar cycle or the volcanic eruptions are excluded. A total of 100 years has been run in this simulation corresponding to 42 complete QBO cycles.

2.2. Experimental Design

[7] On the basis of the 100-year simulation with climatological SST boundary conditions described above (CT simulation), two ensembles have been defined consisting of periods starting in July and ending in September of the following year with their QBO phase being in either westerly (QBO/W) or easterly (QBO/E) phase during the Northern Hemisphere winter months. The criterion used considers westerly (easterly) winds at 30 hPa, where the QBO has the largest impact in the extratropical boreal winter [*Calvo et al.*, 2007], above 10 m s^{-1} (below -10 m s^{-1}) from December to February. This procedure yields a total of 12 realizations for westerly QBO phases and 14 for easterly QBO phases. The ensembles from this control simulation will be referred to as control ensembles (CTW and CTE ensembles). Figure 1 shows the temporal evolution from July 1997 to August 1998 of the ensemble members (thin gray lines) together with the ensemble means (thick black lines) for each QBO phase. Between November and April all members of the westerly and easterly ensembles are westerly and easterly, respectively, and strong westerlies and easterlies prevail from December to March.

[8] The initial states of the selected periods were then used to run experiments with the prescribed SST of the period July 1997 to September 1998 which features one of the strongest warm ENSO events in the last 50 years and it is known to have had a strong effect on the polar stratosphere [*Sassi et al.*, 2004; *Manzini et al.*, 2006; *Garcia-Herrera et al.*, 2006]. These new simulations (EN simulations) result in two additional ensembles, designated as ENW and ENE, which combine the strong warm ENSO SST anomalies that peaked in November 1997 with the selected westerly and easterly QBO phase, respectively, inherited from the initial conditions of the CTW and CTE ensembles.

[9] The four ensembles CTW, CTE, ENW, and ENE resulting from this experimental design are compared below to study the influence of warm ENSO events on the extratropical QBO signal and the effect of the QBO on the warm ENSO signal in the NH polar vortex during the boreal winter months.

3. Results

3.1. QBO Impact on the Warm ENSO Response in the Polar Stratosphere

[10] Considering our experimental setup, the effect of warm ENSO on the extratropical stratosphere without stratifying with respect to a particular QBO phase can be

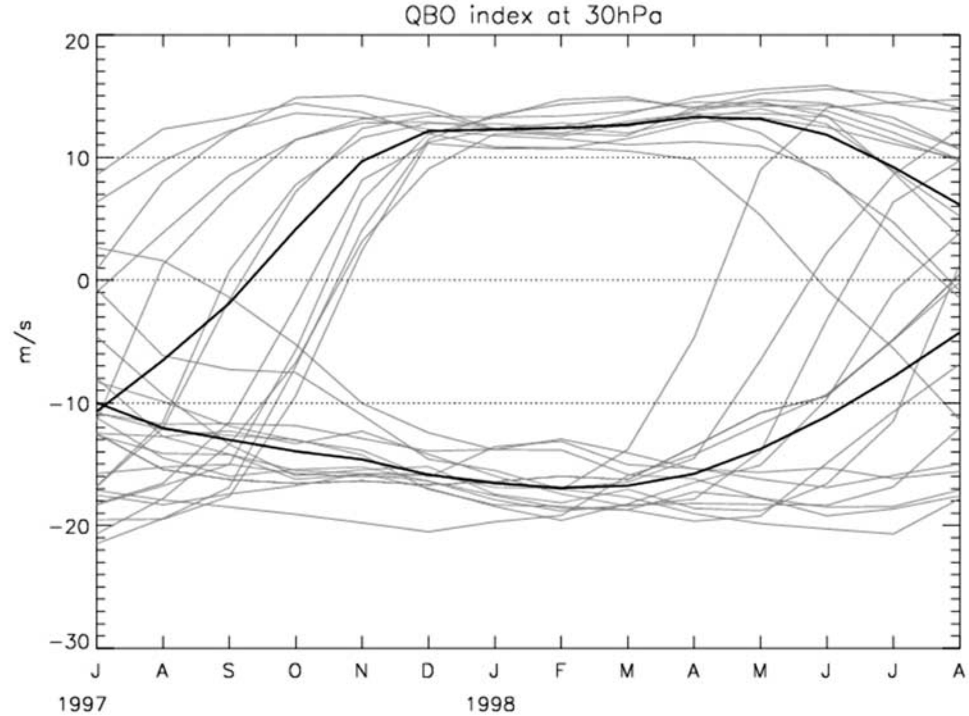

Figure 1. Temporal evolution of the quasi-biennial oscillation (QBO) index computed as the zonal mean zonal wind at the equator at 30 hPa from July 1997 to August 1998 for all the ensemble members (thin lines) of the control simulation (CT) for the westerly and easterly QBO phases. The thick lines denote the ensemble means for each QBO phase.

computed as the mean value (ENW+ENE)/2 minus the 100-year mean from the CT simulation (hereafter, this case is referred to as 'EN−CT'). In addition, the effect of warm ENSO stratified according to the QBO phase can be obtained by computing differences between the ensemble means with and without ENSO for the westerly and easterly phase (ENW−CTW and ENE−CTE). Figure 2 shows the temporal evolution, from November to April, of the zonal mean temperature at 80°N and zonal mean zonal wind at 60°N for these three ensemble mean differences, EN−CT (Figure 2, top), ENW−CTW (Figure 2, middle) and ENE−CTE (Figure 2, bottom). The shadowing denotes those areas where the differences are 95% (dark gray) and 90% (light gray) significant according to a mean difference *t* test. A Monte Carlo test is not suitable here since no temporal series are available for the simulations with observed SST (only ENW and ENE cases were computed). Thus, the computation of random composites is not possible.

[11] When no particular phase of the QBO is selected (EN−CT; Figure 2 (top)), the ENSO that occurred in 1997/1998 generates a warmer and weaker polar vortex in the stratosphere from December to March with the largest anomalies in January. The signal propagates downward with time from the upper stratosphere (December) toward the lower stratosphere below 20 km (March) as a result of the wave−mean flow interaction. This pattern is consistent with previous results obtained from general circulation models with no QBO [*Sassi et al.*, 2004; *Garcia-Herrera et al.*, 2006; *Manzini et al.*, 2006] and reanalysis data where no particular QBO phase was selected [*Garcia-Herrera et al.*, 2006]. In particular, our results agree very well with Figure 4 from *Manzini et al.* [2006], who analyzed the ENSO signal in a different set of experiments from MAECHAM5, run from 1980 to 1999 with observed SSTs and no QBO (neither prescribed nor internally generated because of lower vertical resolution). Larger anomalies are observed in our Figure 2 (top) probably because only the strong ENSO event of 1997/1998 has been considered here while *Manzini et al.* [2006] used a total of 36 warm ENSO members from 9 different simulations to compute the ensemble average which smooths the signal. In fact, the anomalous values in our EN−CT case agree better with *Garcia-Herrera et al.* [2006], who used a single four-member ensemble in their composites.

[12] When stratifying according to the QBO phase (Figure 2, middle and bottom), a warm ENSO still warms the polar stratosphere and weakens the polar vortex in both QBO phases but several differences are observed when comparing with the EN−CT case (Figure 2, top). During the westerly phase of the QBO, the warm ENSO signal starts later in the winter season (especially noticeable in the zonal mean zonal wind) with the largest anomalies also delayed compared with the EN−CT case. Significant maxima anomalies around 7 K in temperature and 14 m s^{-1} in zonal wind are observed here in February versus January in the EN−CT case. On the contrary, during the easterly QBO phase, the signal of warm ENSO advances and shows the largest values at first time in January as when no stratification with respect to the QBO phase was done (EN−CT case). Further, during the west QBO phase, a monotonous downward propagation of the warm ENSO signal is observed from early to late winter as in the EN−CT case; however, in the easterly QBO phase, the descending of the warm ENSO

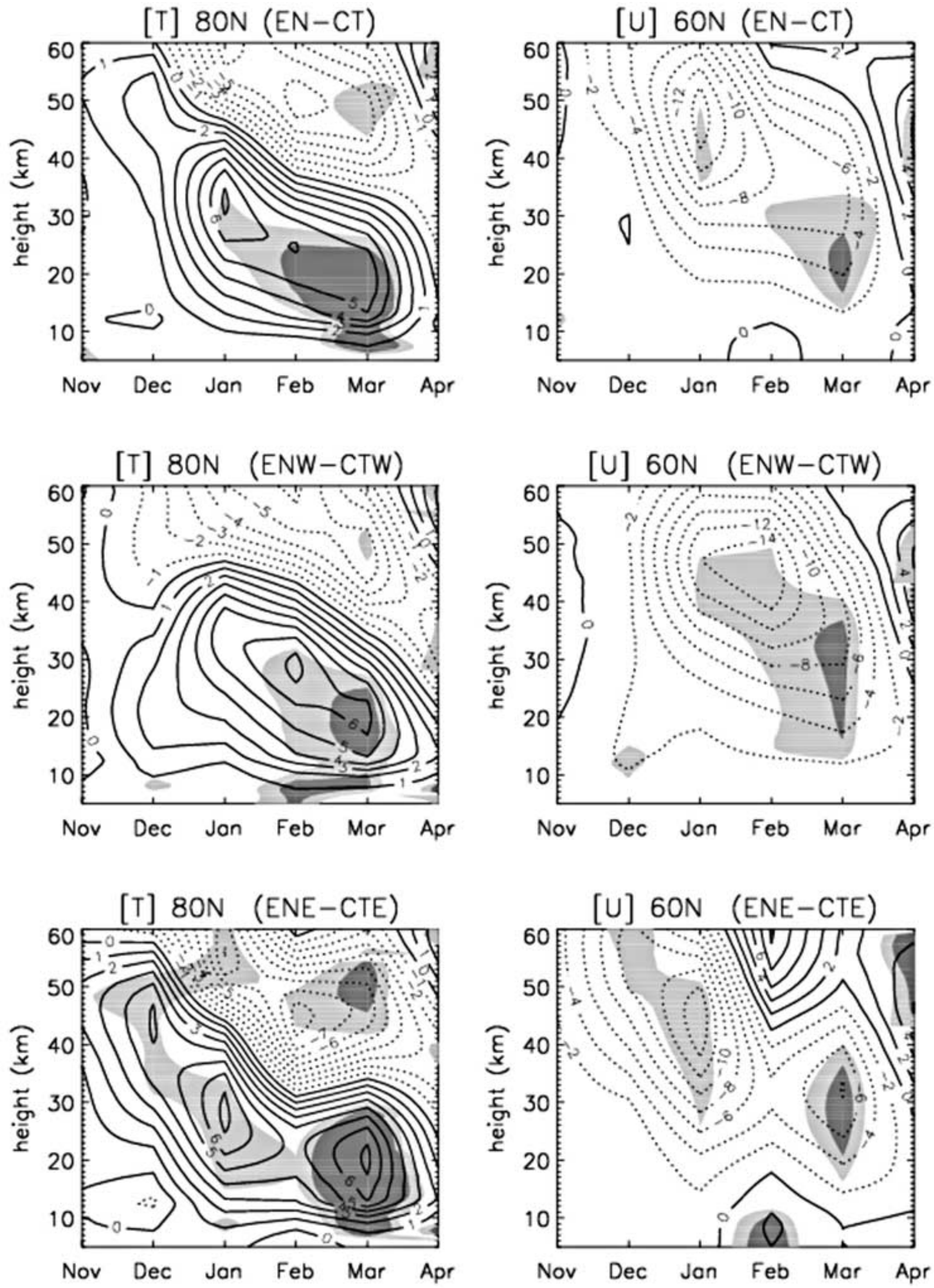

Figure 2. Temporal evolution of the zonal mean temperature at 80°N (left) and zonal mean zonal wind at 60°N (right) for the (top) EN−CT case where no particular QBO phase was selected (more details in the text), (middle) ENW−CTW, and (bottom) ENE−CTE ensemble differences. Solid (dashed) contours for positive (negative) values. Contour intervals are 1 K for temperature and 2 m s^{-1} for zonal winds. Shaded areas denote 90% (light gray) and 95% (dark gray) significant regions according to a different mean *t* test.

signal stops quite abruptly in February when the anomalous values are very low and not significant neither in zonal mean temperature nor in zonal wind. In March, the effect of a warm ENSO intensifies again but does not descend any longer. This behavior is especially well illustrated in the zonal mean zonal wind pattern (Figure 2, bottom right) where the zonal mean zonal wind anomalies with respect to the 100-year climatology drop from -10 m s^{-1} in January to -4 m s^{-1} in February and recover again in March reaching up to -8 m s^{-1}.

[13] To summarize, it seems that the westerly phase of the QBO delays the onset of warm ENSO signal and in particular its largest values toward the end of the winter. The easterly QBO phase, on the other hand, advances the beginning of the warm ENSO signal to November elongating its length from November to March but minimizes largely its effects in February, being nonsignificant. Therefore, the ENSO 1997/1998 event hardly generates any response in the polar vortex in February when it coincides with the easterly phase of the QBO.

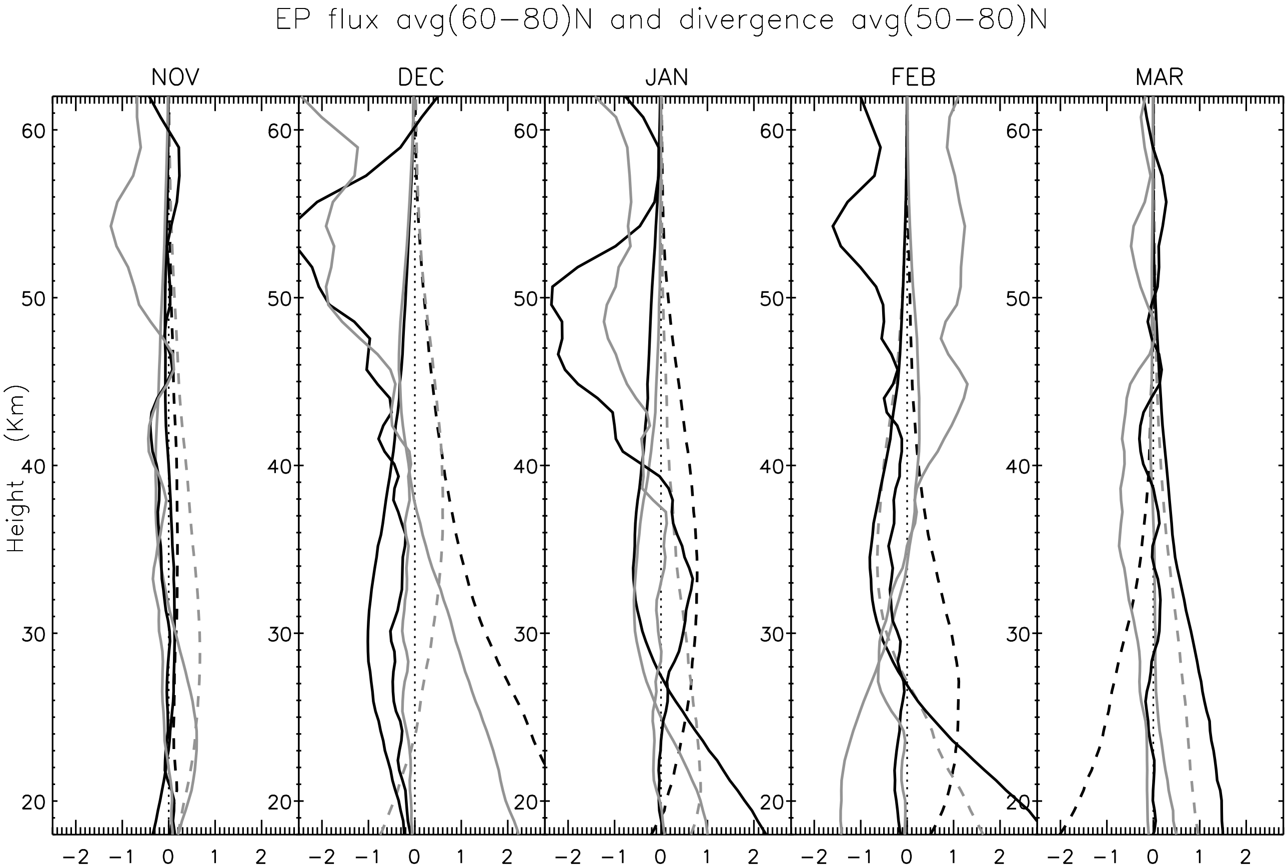

Figure 3. Vertical profiles for the ENW−CTW (black) and ENE−CTE (gray) ensemble differences of the vertical (dashed) and meridional (dashed-dotted) components of the EP flux averaged from 60°N to 80°N and the Eliassen Palm (EP) flux divergence (solid) averaged from 50°N to 80°N. Values plotted are 10^4Fy (Nm^{-1}), 10^6Fz (Nm^{-1}), and 0.5 × div(EP) in m $s^{-1}d^{-1}$.

[14] Previous works dealing with the influence of ENSO on the stratosphere have shown that ENSO affects the polar region through the anomalous generation and dissipation of extratropical waves at middle latitudes which, in turn, changes the background flow [*Sassi et al.*, 2004; *Manzini et al.*, 2006; *Garcia-Herrera et al.*, 2006]. In addition, the QBO, by modulating the zonal mean zonal winds in the tropics, also gives rise to a response of the polar vortex as it modifies the wave−mean flow interaction. Thus, both the generation and dissipation of planetary waves and the zonal mean flow interaction must be taken into account together to understand the effect of the QBO on the warm ENSO signal in the extratropical stratosphere. A simple way to visualize these changes is analyzing the Eliassen Palm (EP) flux and its divergence. The EP flux can be considered as a measure of the vertical wave propagation [*Edmon et al.*, 1980], while its divergence measures the changes that the dissipating waves originate in the background flow [*Andrews et al.*, 1987]. In MAECHAM5 the EP flux and EP flux divergence, respectively, account for propagation and dissipation of planetary waves resolved by the model. The EP flux for CT and EN ensembles in both QBO phases (not shown) indicates upward wave propagation during boreal winter as expected since westerly winds occur in the NH extratropics and allow vertical wave propagation [*Charney and Drazin*, 1961]. Figure 3 shows the vertical profile in winter months for the ENW−CTW (black) and ENE−CTE (gray) ensemble mean differences of the Fy (dashed-dotted) and Fz (dashed) components of the EP flux averaged between 60°N and 80°N. This latitudinal average maximizes the signals and clarifies the differences between easterly and westerly QBO phases. Figure 3 also shows the differences in the EP flux divergence (solid) averaged from 50°N to 80°N. In this case, the 50°N−80°N latitude range have been chosen instead of 60°N−80°N because the largest dissipation areas tend to occur slightly equatorward from the latitudes with the largest upward propagation (waves propagate upward, bend equatorward and then dissipate, not shown here) and thus, an extended region toward the subtropics maximizes the signal in this case.

[15] In November, the larger anomalous upward wave propagation (dashed gray line) and dissipation (solid gray line) generated by warm ENSO during the easterly QBO phase agrees with the fact that the warm ENSO signal on zonal mean temperature and zonal wind appears earlier in the polar stratosphere than during the westerly QBO phase (see Figure 2). As was mentioned before, these differences follow changes in the background flow due to the QBO: a strong polar vortex favored during the westerly QBO phase makes its perturbation by warm ENSOs more difficult while the easterly QBO phase, whose effects on the polar vortex are in the same direction as those from warm ENSOs, favors

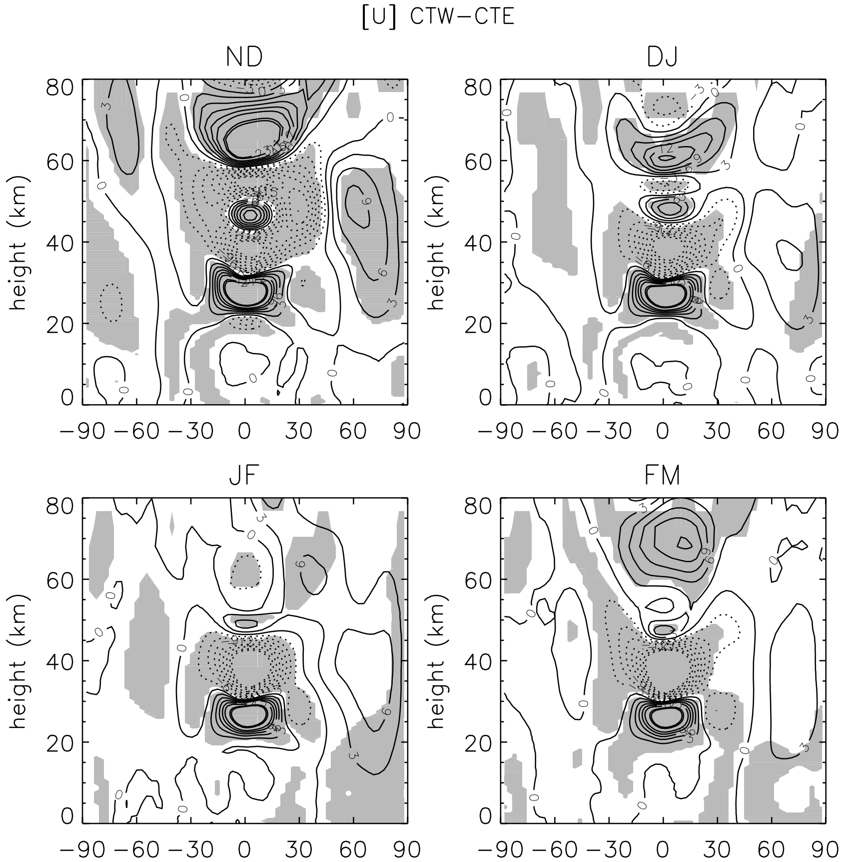

Figure 4. Westerly minus easterly QBO phase differences of the zonal mean zonal wind for the ensemble means of the CT simulation (CTW−CTE) for (top left) November–December average, (top right) December–January, (bottom left) January–February, and (bottom right) February–March. Contours are drawn every 3 m s^{-1}; solid (dashed) lines for positive (negative) anomalies. Shaded regions denote those 95% significant following a different mean *t* test.

its weakening. In December and January, the upward propagation and dissipation of waves due to warm ENSO intensifies in both QBO phases compared with November which increases the warm ENSO signal in both zonal wind and temperature fields. In February, the anomalous upward propagation generated by a warm ENSO intensifies toward the North Pole during the westerly QBO phase (dashed and dashed-dotted black lines). However, in the easterly QBO phase, below 30 km, the wave propagation is intensified upward toward the subtropics (dashed and dashed-dotted gray lines) instead of toward high latitudes as would be expected if a warm ENSO or the easterly phase of the QBO operated independently [e.g., *Garcia-Herrera et al.*, 2006; *Calvo et al.*, 2007]. This is indicated by the dashed-dotted gray line in February which shows equatorward anomalies (negative values) in the meridional component of the EP flux (Fy) below 30 km while the positive anomalies in Fz (dashed gray) decrease with height. As a result of this strong refraction toward the subtropics, above 30 km, the upward propagation ENE−CTE difference becomes negative, indicating less intense upward propagation in this region when including warm ENSO and less intense wave dissipation in the middle and upper stratosphere (positive differences in the EP flux divergence, solid gray line). This anomalous behavior in the wave activity in the entire extratropical stratosphere when a warm ENSO coincides with the easterly phase of the QBO originates the response of the polar stratosphere observed in Figure 2. Finally, in March, the upward wave propagation becomes negative in the westerly QBO phase (which means weaker upward propagation in EN simulations than in CT) and the dissipation gets weaker. However, in the easterly phase, as the wave propagation gets weaker in February, the vortex might recover again and thus, be affected by warm ENSO in March (as shown by the positive upward EP flux anomalies and negative EP flux divergence anomalies) which generates warm temperature anomalies and weaker winds as shown in Figure 2.

[16] Therefore, the analysis carried out here indicate that at some point, the polar vortex already perturbed by the easterly QBO phase cannot be more perturbed by the additional forcing that a strong warm ENSO event might generate. This agrees with some of the results in the work by *Garfinkel and Hartmann* [2007], who composited ERA-40 data for warm ENSO winter months (NDJF) and obtained a weak and nonsignificant ENSO effect on the polar region at 10 hPa when the QBO was in its easterly phase.

3.2. Warm ENSO Impact on the QBO Effect in the Polar Stratosphere

[17] As mentioned in the Introduction, the QBO affects the polar stratosphere originating a weaker (stronger) and warmer (colder) polar vortex during the easterly (westerly)

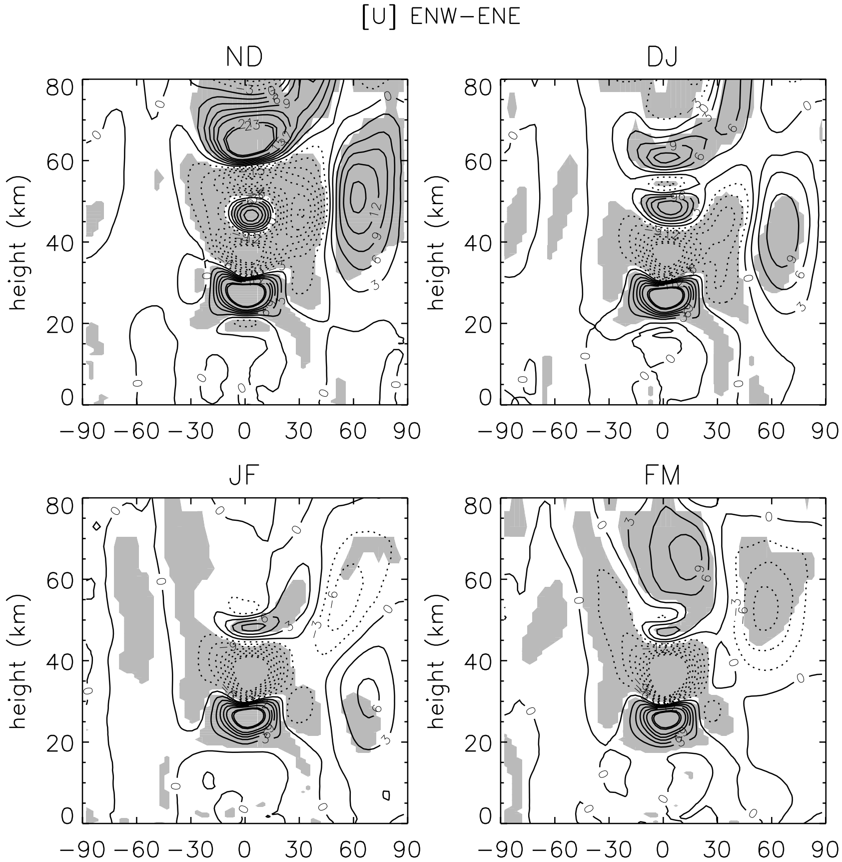

Figure 5. As Figure 4 for the ensemble means of EN simulations.

phase of the QBO [e.g., *Holton and Tan*, 1982; *Calvo et al.*, 2007]. Traditionally the QBO signal has been studied through the analysis of composite differences between QBO phases (typically westerly minus easterly QBO phase) defined with respect to the equatorial wind at near 30 hPa. The impacts, if any, of ENSO on the extratropical QBO signal are analyzed here by computing and comparing the westerly minus easterly QBO phase ensemble differences for CT and EN simulations: CTW–CTE and ENW–ENE. Figure 4 shows latitude-height cross sections of the westerly minus easterly QBO phase difference of zonal mean zonal wind for the ensemble means of the control simulation (CTW–CTE) where the SSTs are climatological and thus, no direct ENSO signal is included. The plots are averaged from early to late winter every 2 months applying the same methodology followed by *Calvo et al.* [2007]. A mean difference *t* test has been performed computing the ensemble mean and standard deviation of each group of 2 months.

[18] Similarly, Figure 5 shows the westerly minus easterly QBO phase ensemble differences for the EN simulations (ENW–ENE) which represents the QBO signal during a strong warm ENSO event. In both cases, CT and EN simulations, the largest QBO signal in U occurs in early winter, mainly during November–December. Thereafter the positive U anomalies descend and weaken. The comparison between Figures 4 and 5 indicates that the ENSO 1997/1998 intensifies the QBO effect in zonal mean zonal wind in early winter (November/December (ND) and December/January (DJ)) in the polar upper stratosphere and lower mesosphere (anomalies reach up to 15 m s^{-1}), and accelerates its downward propagation shortening the total length of the extratropical QBO signal in winter. The positive QBO signal in U is last observed in January/February (JF) in the EN simulations versus February/March (FM) in CT. In the EN simulations, a negative QBO signal in U is observed at 60°N in JF in the upper stratosphere and lower mesosphere being stronger in FM. The analysis of the QBO signal in zonal mean temperature shows consistent results and is not shown here.

[19] For the further discussion it is helpful to split the W–E signal into separate phases with respect to the 100-year CT climatology: W–E = (W-clim.)–(E-clim.). This gives some indication on when the effect of the westerly or easterly QBO phase with respect to climatology may dominate the total W-E QBO signal although new plots are not included here for brevity. For example, the QBO effect in U in the CT simulation is mainly due to the intensification of the polar vortex during the westerly QBO phase in ND and DJ while in JF the QBO signal is mainly due to the weakening of the zonal winds during the easterly QBO phase. In the EN simulations, both phases contribute equally to the QBO signature in early winter, while in middle and late winter the influence of the easterly QBO phase becomes larger. The impact of a warm ENSO on the extratropical QBO signal thus results from the effect of warm ENSO on the westerly and easterly QBO phases which might be different, including the possibility that warm ENSO modifies only one of the QBO phases at some time in the winter.

[20] This decomposition of the W–E QBO signal is especially useful for the interpretation of the wave–mean flow interaction related to the impact of warm ENSO on the

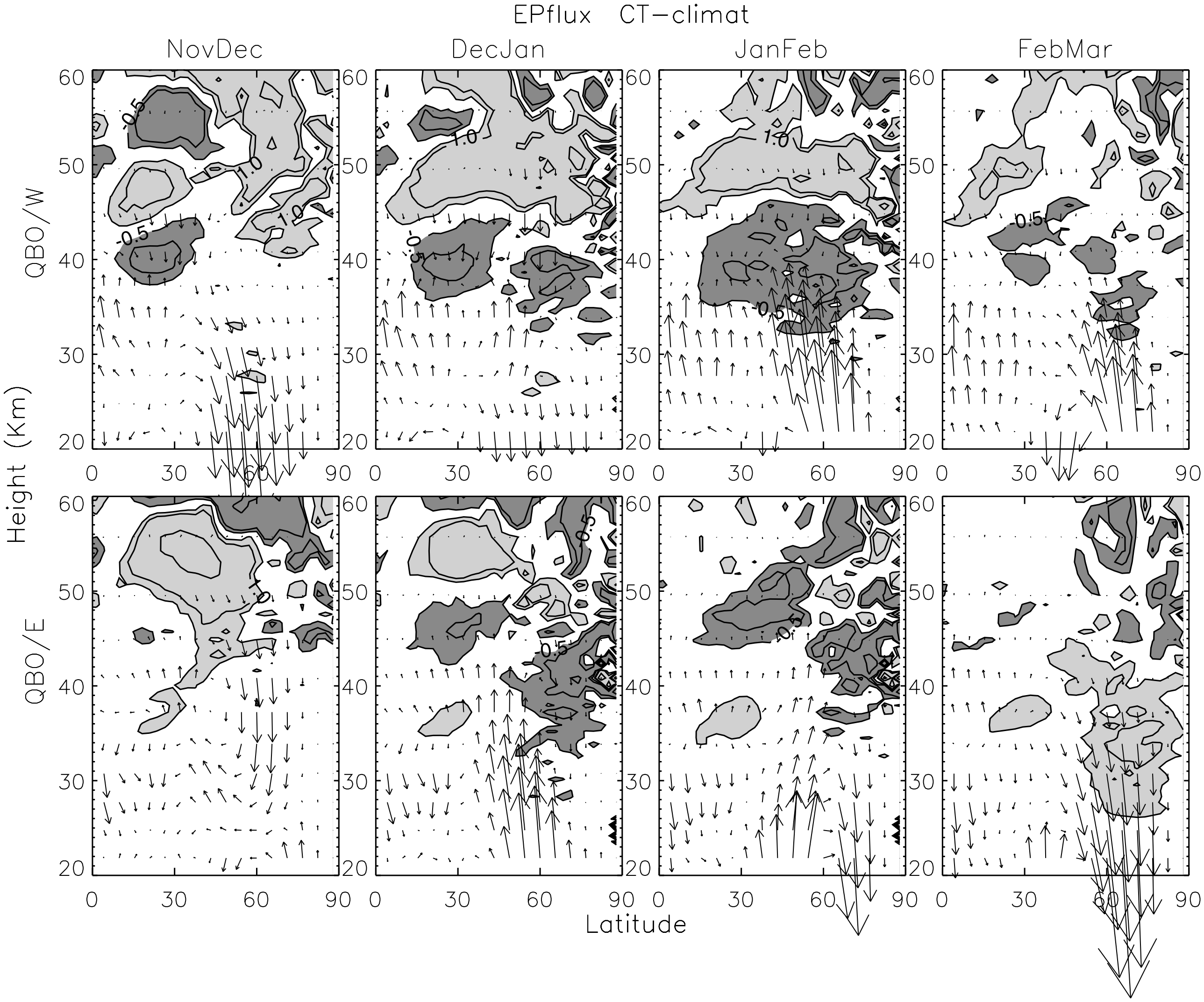

Figure 6. Latitude-height cross sections of the EP flux (arrows) and EP flux divergence (contours) anomalies with respect to the 100-year CT climatology for the (top) CTW and (bottom) CTE ensembles from November to December (leftmost) to February–March (rightmost). ($Fz/F\varphi = 300$.) Contours are drawn at ±0.5, 1, 3, 5 m s^{-1} d^{-1}. Light (dark) shaded for positive (negative) divergence larger than ±0.5 m s^{-1} d^{-1}.

W–E QBO signal shown in U (Figures 4 and 5). It is well known that during the westerly phase of the QBO waves refract more toward the subtropics and as a result, the polar vortex intensifies compared to easterly QBO phase conditions [e.g., *Calvo et al.*, 2007]. This behavior is also shown in Figure 6 where anomalies of the EP flux (arrows) and its divergence (contours) with respect to the 100-year CT climatology are displayed in the NH for the CTW and CTE ensembles. During the westerly QBO phase, the EP flux divergence anomalies in ND are positive and result in acceleration of the zonal mean zonal wind U at high latitudes. Acceleration occurs also in DJ and JF at about 50 km altitude. Deceleration at high latitudes occurs in DJ and JF between 35 and 45 km altitude, but not poleward than 75°N. In FM, the EP flux–related tendencies in U are small. During the easterly QBO phase, the anomalies in the EP flux divergence result in small tendencies at high latitudes in U in ND, and deceleration in DJ and JF. Only during FM we find acceleration between about 25 and 40 km altitude.

[21] On the other hand, the effect of a strong warm ENSO generally intensifies the upward EP flux and the dissipation at higher heights and latitudes with respect to the climatology and as a result, the polar vortex weakens compared to neutral or cold ENSO conditions [*Garcia-Herrera et al.*, 2006]. Thus, it may be expected that during warm ENSO, when upward EP fluxes at high latitudes are stronger than the average and their refraction still depends on the QBO, opposite phases of the QBO can generate stronger wave–mean flow interaction differences and hence a stronger QBO signal in the polar vortex than the average. However, the observed effect of the QBO on the timing of the warm ENSO signals, as reported in section 3.1, may add complications to this picture, and linearity may not apply. Figure 7 shows the differences in the EP flux and its divergence between the EN simulations and the climatology for ENW and ENE ensembles. From a quick comparison with Figure 6 where climatological SSTs were used, it is immediately evident that the warm ENSO 1997/1998 intensifies the upward wave

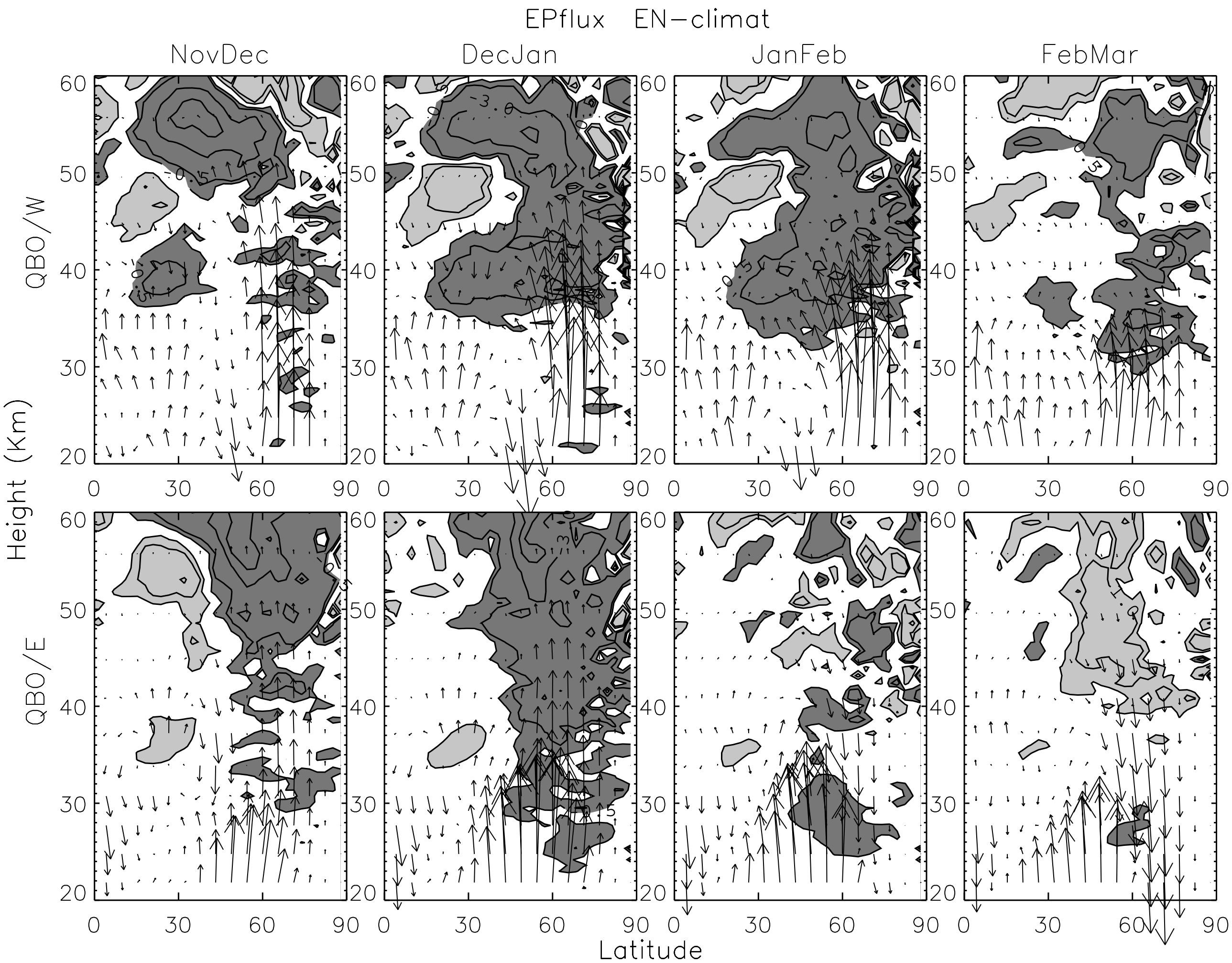

Figure 7. As Figure 6 for the ENW and ENE ensembles.

propagation and dissipation at high latitudes in both QBO phases. A detailed analysis of Figures 6 and 7 reveals several features that explain the effect of a warm ENSO on the extratropical QBO signal: its intensification in early winter and the earlier dissipation in late winter.

[22] In ND, despite the enhanced propagation in both QBO phases, large anomalies in the EP flux divergence are observed at high latitudes only during the east phase of the QBO, giving rise eventually to negative anomalies in U during this QBO phase (U anomalies with respect to climatology in CTW, CTE, ENW, ENE ensembles are not shown here but will be discussed when necessary). During the westerly QBO phase, however, the anomalous dissipation occurs at middle latitudes and does not have an effect on the polar vortex. Thus, the intensification of the QBO signature in U observed in early winter (Figure 5) is mainly due to the effect of ENSO on the easterly QBO phase signal (ENE). In this case, the polar vortex is weakened even more than in the CTE ensemble mean owing to an enhancement of the propagation and dissipation of the waves. This result agrees with section 3.1 where it was shown that the ENSO effect on the polar region appeared earlier during the easterly phase of the QBO than during the westerly phase.

[23] In DJ the wave propagation and dissipation is still enhanced at polar latitudes during both phases of the QBO in the EN simulations. In JF and FM, the upward propagation and dissipation at high latitudes is still large in the westerly phase (ENW) compared with the CTW ensemble (Figure 6), which originates strong negative anomalies in U during this phase (not shown). However, in the easterly QBO phase, the differences between EN and CT simulations are not relevant at latitudes higher than 60°N. For both SST boundary conditions, similar downward anomalies occur in the EP flux in JF compared to climatology, and acceleration in the mid or upper stratosphere. As was mentioned before, this is the result of the nonlinear behavior already showed in section 3.1 when both warm ENSO and the easterly phase of the QBO operate together.

[24] These results explain the earlier vanishing of the QBO signature in U observed in the EN simulations (Figure 5). The large effect during the westerly QBO phase in late winter generate negative anomalies in U while the small effect of warm ENSO during the easterly QBO phase hardly changes the negative anomalies already observed in ND and DJ in the CT simulation. This results in negative anomalies of about the same magnitude in both phases which cancel the total QBO signature (W−E QBO) in the lower/middle stratosphere in late winter. In the upper stratosphere, the negative anomalies in U observed in FM in Figure 5 are the result of the ENSO effect during the easterly QBO phase with

less upward propagation and dissipation than the climatology (positive EP flux divergence above 40 km) and thus, anomalous strengthening of the polar vortex.

4. Summary and Conclusions

[25] We have studied the combined effect of warm ENSO events and extreme phases of the QBO on the NH polar stratosphere in the most recent version of MAECHAM5 GCM, which is able to internally generate a realistic QBO [*Giorgetta et al.*, 2002, 2006]. Experiments were performed in which winters with strong easterly or westerly QBO phases have been forced by either SSTs of the strong ENSO 1997/1998 or climatological SSTs obtaining a total of 14 realizations for the easterly QBO phase and 12 for the westerly phase for each of the two SST forcings.

[26] The well-known effect of warm ENSO on the polar stratosphere observed when it operates independently (a warmer polar stratosphere and a weaker polar vortex [e.g., *Garcia-Herrera et al.*, 2006]) is obtained here when no particular QBO phase is selected. However, our analysis has shown that the phase of the QBO does influence the signal of warm ENSO in the polar stratosphere. The warm ENSO signal is intensified by the QBO at the end of the winter (March), whether the QBO is in westerly or easterly phase. The westerly QBO phase delays the onset of the signal of the warm ENSO while the easterly QBO phase advances it. Further, in February the effect of warm ENSO diminishes considerably when it coincides with the easterly phase of the QBO, being hardly noticeable and not significant, in agreement with results from *Garfinkel and Hartmann* [2007], who analyzed the warm ENSO response for ERA-40 data during winter months.

[27] We have demonstrated that the changes in the background flow associated with the QBO affect the propagation and dissipation of extratropical waves generated by ENSO and explain the impact of the QBO phase on the polar stratospheric response to a warm ENSO. The larger anomalous wave propagation toward the subtropics, as made possible by the westerly QBO phase, delays the onset of the warm ENSO signal while the easterly QBO phase favors the poleward wave propagation and advances the onset of the warm ENSO effect. However, the strong propagation toward the North Pole strongly weakens at the end of the winter, and in February the propagation and dissipation observed in the NH extratropical stratosphere is much weaker when a warm ENSO event coincides with the easterly phase of the QBO. Thus, despite both warm ENSO and easterly QBO originate a polar response in the same direction when they operate independently; their effects are not additive when both phenomena coincide.

[28] Warm ENSO events also have an impact on the extratropical QBO signal intensifying the effect of the QBO at the beginning of the winter and advancing its signal so that it disappears earlier in late winter, shortening its length and accelerating its downward propagation. These changes are also the result of anomalies in the wave–mean flow interaction due to both ENSO and QBO operating together. The earlier effect of warm ENSO on the polar stratosphere during the easterly phase of the QBO compared with the westerly phase intensifies the weakening of the polar vortex in the easterly phase but hardly changes the values in the westerly phase. This intensifies the total westerly minus easterly QBO phase signature. In middle and late winter, the signal of the westerly phase of the QBO in the polar stratosphere reverses signs due to the effect of warm ENSO and an anomalously weaker polar vortex is observed. However, during the east phase of the QBO, the effect of warm ENSO is not additive and it does not intensify the signal of the easterly QBO phase (despite both easterly QBO phase and warm ENSO independently would act to warm the polar stratosphere and weaken the polar vortex) but even debilitates. Thus, in late winter both westerly and easterly QBO phases combined with ENSO originate anomalous weaker winds in the polar region and thus, the total QBO signature computed as westerly minus easterly phases minimizes. Then, this yields to a faster downward propagation of the extratropical QBO signal from early winter to late winter, mainly because the QBO signal disappears earlier in late winter. This partially agrees with results from *Wei et al.* [2007], who analyzed the ENSO modulation of the extratropical QBO in ERA-40 data and found a much weaker Holton and Tan relationship during warm ENSO events than during cold ENSOs in the DJF average. We obtained that a warm ENSO weakens the Holton and Tan relationship in late winter but intensifies it in early winter.

[29] Therefore, both the QBO impact on the warm ENSO signal in the NH polar stratosphere and the effect of a warm ENSO on the extratropical QBO response point to the same conclusion: a nonlinear behavior of the extratropical stratosphere to external forcings. *Camp and Tung* [2007] also found a similar nonlinear response between the solar cycle and the QBO with a weaker Holton and Tan relationship during solar maxima, which highlights the importance of isolating the phenomena we are studying from other forcings in the stratosphere. In this sense our results show for the first time, the ENSO/QBO interaction without the possible influence of other sources or variability included in observational data as the effects of solar cycle or volcanic eruptions. Furthermore, the model experiment has been designed to choose a strong warm ENSO event coinciding with strong phases of the QBO for a number of realizations, which adds confidence to the results. It also allows dynamical analysis in terms of wave activity and wave driving which facilitates the interpretation and understanding of the mechanisms involved. These results suggest that the nonlinearity found in this work for the combined effects of the QBO and ENSO has to be expected also in the analysis of combined effects including other phenomena that influence the variability of the stratosphere, such as the solar cycle, volcanic eruptions, and climate change.

[30] **Acknowledgments.** N. Calvo was supported by the visitor program of the Max Planck Institute for Meteorology in Hamburg, the Spanish Ministry of Education and Science, and the Fulbright Commission in Spain. The simulations described here were carried out on the NEC-SX6 computer at the German Climate Computing Center (DKRZ). We thank Rolando R. Garcia and Daniel Marsh for their helpful comments and suggestions.

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N. Calvo and R. Garcia-Herrera, Departamento Fisica de la Tierra II, Universidad Complutense de Madrid, Madrid E-28040, Spain. (calvo@ucar.edu)

M. A. Giorgetta, Max Planck Institute for Meteorology, Bundesstrasse 53, Hamburg D-20146, Germany.

E. Manzini, Instituto Nazionale di Geofisica e Vulcanologia, Via Donato Creti 12, Bologna I-40128, Italy.