Stratospheric Communication of El Niño Teleconnections to European Winter

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ABSTRACT

The stratospheric role in the European winter surface climate response to El Niño–Southern Oscillation sea surface temperature forcing is investigated using an intermediate general circulation model with a well-resolved stratosphere. Under El Niño conditions, both the modeled tropospheric and stratospheric mean-state circulation changes correspond well to the observed “canonical” responses of a late winter negative North Atlantic Oscillation and a strongly weakened polar vortex, respectively. The variability of the polar vortex is modulated by an increase in frequency of stratospheric sudden warming events throughout all winter months. The potential role of this stratospheric response in the tropical Pacific–European teleconnection is investigated by sensitivity experiments in which the mean state and variability of the stratosphere are degraded. As a result, the observed stratospheric response to El Niño is suppressed and the mean sea level pressure response fails to resemble the temporal and spatial evolution of the observations. The results suggest that the stratosphere plays an active role in the European response to El Niño. A saturation mechanism whereby for the strongest El Niño events tropospheric forcing dominates the European response is suggested. This is examined by means of a sensitivity test and it is shown that under large El Niño forcing the European response is insensitive to stratospheric representation.

1. Introduction

In observations of the Northern Hemisphere (NH) extratropics, the mean sea level pressure (MSLP) response to the El Niño–Southern Oscillation (ENSO) appears largest in wintertime, primarily as an enhancement of well-known tropospheric teleconnection patterns, such as the Pacific–North American (PNA) pattern and the tropical–Northern Hemispheric (TNH) pattern (Wallace and Gutzler 1981; Mo and Livezey 1986). Over the Euro-Atlantic sector, a robust surface signal has appeared in composite studies of long data- sets (Mann et al. 2000; Brönnimann et al. 2007), appearing as a projection onto a negative phase of the North Atlantic Oscillation (NAO). The signal is manifest as a zonally orientated dipole over Europe with anomalously high MSLP over northern Europe, and low MSLP over southern Europe and the Mediterranean (van Loon and Madden 1981; Fraedrich et al. 1992; Gouirand and Moron 2003). In general, the NH extratropical winter response takes on a structure that is highly reminiscent of the surface northern annular mode (NAM; Thompson and Wallace 2000) in its negative phase, with a high-latitude center of action that is dominated by positive anomalies over the pole and a band of negative anomalies in the midlatitudes.

In Europe this observed signal follows approximately two-thirds of the El Niño events (Toniazzo and Scaife 2006), tending to maximize in late winter (January–March) and lagging the central Pacific sea surface temperature (SST) maximum anomalies by a few months (Moron and Gouirand 2003). However, within the literature there is no consensus on a mechanism for this late-winter teleconnection (Brönnimann 2007), and authors have found a range of responses in the observations, some of which are nonlinear, nonstationary, or not statistically significant (Wu and Hsieh 2004; Greatbatch et al. 2004, Pozo-Vázquez et al. 2001).

In the stratosphere, observations and general circulation modeling (GCM) studies illustrate a more robust
response to El Niño, that of a climatologically weaker and warmer polar vortex (van Loon and Labitzke 1987; Hamilton 1993). An increase in the frequency of stratospheric sudden warming (SSW) events between the opposite (warm–cold) phases of ENSO was noted by Taguchi and Hartmann (2006) in a perpetual January simulation with the Whole Atmosphere Community Climate Model (WACCM), suggesting an influence of El Niño not only on the mean state of the stratosphere, but also on its variability. The weakened vortex is accompanied by a significant warming signal that begins in early winter (November) in the upper stratosphere and propagates downward. Upon reaching the lower stratosphere in midwinter the signal persists until spring. These dynamical responses have been supported in more recent studies (Brönnimann et al. 2004; Sassi et al. 2006). The temporal coincidence between polar lower-stratospheric anomalies and the MSLP signal over Europe in late winter following El Niño events may suggest a mechanism involving stratospheric communication of the tropical Pacific teleconnection to Europe. In recent years there has been growing evidence advocating for a significant two-way dynamical coupling between the stratosphere and troposphere (Shaw and Shepherd 2008), on time scales ranging from the seasonal to climatic (Baldwin and Dunkerton 2001; Gillett and Thompson 2003). In both the observational and modeling studies, weak stratospheric westerly regimes tend to be a precursor to negative NAO phases in the winter troposphere (Baldwin and Dunkerton 2001; Scaife et al. 2005) and, as such, a stratospheric role in the El Niño teleconnection to Europe has gained credence (Bronnimann 2007).

However, despite recent interest there have been relatively few studies investigating such a possibility. Unable to reproduce the observed NAO response to El Niño forcing using the third Hadley Centre Atmosphere Model (HadAM3), Toniazzo and Scaife (2006) suggested a possible inhibition of any stratospheric pathway in the teleconnection to Europe resulting from their low 10-hPa model top. A high/low model-top comparison was performed by Cagnazzo and Manzini (2009) who found a more realistic European surface response in the model with a fully resolved stratosphere. Recently, it has also been demonstrated that the frequency modulation of SSW events plays an important role in the late winter MSLP signal, such that the modeled canonical response is only simulated in winters with a dynamically active stratosphere (Ineson and Scaife 2009).

In this paper we directly investigate the role of the stratosphere with experiments that act to remove any stratospheric pathway in the teleconnection. Our methodology differs to that of high/low model-top experiments in that we make use of a damping of the stratospheric mean state and variability, such that any pathway via the stratosphere is suppressed. The results suggest that the canonical European response is modulated by the influence of the stratosphere and that in order to capture the observed MSLP signal, an accurate representation of stratospheric processes is required.

In a sensitivity experiment we also investigate the linearity of the Euro-Atlantic surface response in terms of El Niño amplitude. Observations demonstrate a clear non-linearity in the MSLP signal. For the strongest El Niño events, the zonal NAO response over Europe gives way to a high pressure signal, evident as a wave train originating in the tropospheric tropical Atlantic (Toniazzo and Scaife 2006). We investigate this nonlinearity by examining the European response to larger-amplitude El Niño events, and its sensitivity to stratospheric representation.

The paper is laid out as follows: section 2 describes the model and experimental setup. The wintertime stratospheric and surface responses to El Niño are examined in section 3. The role of stratospheric feedback onto the tropospheric response, including its sensitivity to El Niño amplitude, is investigated and described in section 4. The results are summarized and discussed in section 5.

2. Model and experimental setup

a. The Reading IGCM

The Reading Intermediate General Circulation Model (IGCM) is a fast general circulation model of intermediate complexity, based on the spectral dynamical core of Hoskins and Simmons (1975) with a range of moist physical processes, such as surface, boundary layer, convective, and radiative schemes (e.g., see Forster et al. 2000). The model runs are carried out with a fully evolving seasonal cycle at T31 horizontal resolution which has 48 × 96 grid points in latitude and longitude. It extends from 1000 to 0.1 hPa with 26 levels and has a vertical resolution that varies from ~1 km in the upper troposphere/lower stratosphere to ~3 km in the upper stratosphere. The decelerating effect of breaking gravity waves on the mean flow in the upper stratosphere and mesosphere is parameterized by a Rayleigh drag term that takes the form of an inverse time scale, given by \( r = \frac{1}{\tau} \) [1 − (\( \ell - 1 \)/3)]/\( 2\pi \). The drag is applied only over the top three model levels \( \ell \), above 1 hPa. This simple parameterization means that the quasi-biennial oscillation (QBO) is not simulated. At the lower boundary the model is forced with monthly mean climatological SSTs from the period of 1979–2000. Dynamical representation
of both the stratospheric mean state and its variability at T31 resolution marks a significant improvement to that of T21, used in previous studies (Forster et al. 2000; Joshi et al. 2003; Taylor and Bourqui 2005). Figure 1 (top row) shows the January–February climatology in zonal-mean temperature and zonal wind from the T31 model. Improved horizontal resolution has alleviated the significant wintertime zonal-mean temperature anomalies reported in Taylor and Bourqui (2005) by ~10 K. The improved stratospheric mean-state temperature structure at T31 is most likely due to a better simulation of planetary wave forcing on the polar vortex at T31, and as a result has a more realistic mean meridional circulation (MMC) in the stratosphere.

**FIG. 1.** (top) T31 control run climatology for the zonally averaged (left) temperature and (right) zonal wind for January–February. Contour intervals are 10 K and 10 m s$^{-1}$, respectively. (middle) The January–February eddy heat flux between 40$^\circ$ and 80$^\circ$N at 100 hPa against the February–March 50-hPa temperature over the polar cap shown for the (left) T21 and (right) T31 model runs. Thick line is the approximate NCEP average. (bottom) Time series of daily 10-hPa zonal-mean zonal wind at 60$^\circ$N for the (left) T21 and (right) T31 model runs. Thin gray line denotes the daily climatological mean.
Previous studies have indicated a strong relationship between meridional eddy heat flux and polar cap temperature anomalies (Newman et al. 2001; Charlton et al. 2007). The meridional eddy heat flux at 100 hPa can be used as a proxy for the vertical component of the Eliassen-Palm flux (Andrews et al. 1987; Polvani and Waugh 2004) and therefore as a diagnostic tool for stratospheric planetary wave forcing. To illustrate this, Fig. 1 (middle) shows scatterplots of the 50-hPa temperature (February–March) averaged over the polar cap, against the eddy heat flux at 100 hPa (January–February) averaged over the vortex edge (40°–80°N), at both T21 (left) and T31 (right) resolution. Although the T31 model shows a slight cold bias in comparison to that of the National Centers for Environmental Precipitation (NCEP), possibly due to the lack of a gravity wave scheme, the linear relationship is captured well, with higher stratospheric temperatures associated with increased planetary wave forcing. The corresponding plot is shown for T21 and, as noted in Taylor and Bourqui (2005), an underestimation of the eddy heat flux entering the stratosphere at 100 hPa results in a large cold bias in the NH winter stratosphere.

Not only is the modeled mean state improved at T31, but so too is the simulation of interannual and intraseasonal variability and, in particular, the SSW events. Here we use the World Meteorological Organization (WMO) definition of SSW (i.e., a reversal of zonal-mean zonal winds to easterly at 10 hPa and 60°N) to count the number of SSW events within the December–March (DJFM) period. We make use of an algorithm similar to that described in Charlton et al. (2007), such that once a warming is identified, no day within the following 20 days can be counted as SSW. Similarly, we discount final warmings by excluding cases in which the zonal-mean zonal wind becomes easterly and does not return to westerly for at least 10 consecutive days before the end of April. The time series of daily zonal-mean zonal wind at 10 hPa and 60°N is shown in Fig. 1 (bottom). It shows that, at T31 resolution, the model provides a realistic SSW climatology, simulating 0.42 events per year and falling within the range determined by Charlton et al. (2007). The T21 model, however, does a poor job at representing the correct stratospheric variability, with virtually zero SSW events simulated.

### b. Experimental setup

To investigate the influence of El Niño on the model climate, two 40-yr IGCM simulations were carried out; a control experiment (CTRL) in which the SSTs at the lower boundary were imposed as climatological averages, and an El Niño (ELNO) experiment, which differed from the control only in the tropical Pacific SST field. A monthly averaged composite El Niño anomaly field was derived from the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) data (online at http://bader.nerc.ac.uk/data/hadisst), for the El Niño years of 1982/83, 1986/87, 1991/92, 1997/98 in the region located between 5°N–5°S and 140°E–100°W. Any anomalies outside this region are masked out. The composite tropical Pacific SST field was then added to the climatological mean for each month, thus introducing an El Niño forcing into the annual cycle. Control and El Niño runs of 40 yr each were performed, and the responses (ELNO − CTRL) are presented. Statistical significance is determined using a Student’s t test and a confidence level of 95% is used throughout the paper.

### 3. IGCM response to El Niño

#### a. Stratospheric zonal-mean response

The zonal-mean January–February latitude–height response in temperature and zonal wind is shown in Figs. 2a,b. An influence on the zonal-mean flow is evident at virtually all latitudes and heights; however, the largest January–February response to El Niño occurs in the stratosphere. A significant zonal wind deceleration extends throughout the depth of the NH stratosphere, reaching −14 m s⁻¹ at the stratopause. The associated zonal-mean temperature anomaly reaches −6 K over the pole at 10 hPa. The stratospheric temperature dipole response, with warming at the polar latitudes and cooling in the tropics, is consistent with an enhancement in the mean meridional circulation and agrees well with observations (García-Herrera et al. 2006).

Using a diagnostic previously employed by Manzini et al. (2006), the time–height evolution of the zonal-mean temperature response at 60°–90°N and zonal wind response at 60°N are shown in Fig. 3. They reveal a clear zonal-mean response to El Niño by way of a downward-propagating signal in the stratosphere. Anomalies in both fields start in early NH winter in the upper stratosphere and reach the tropopause by midwinter. Warm temperature anomalies (Fig. 3, top) of 1–2 K appear at the stratopause in November, which slowly propagate downward and reach the lower stratosphere in January. Anomalies of ~2 K persist above the tropopause (possibly due to the long radiative time scales there) until late winter/early spring. With the variability of the stratosphere dominated by that in mid- to upper levels (associated with SSW events) the response in the lower stratosphere is statistically significant during February, March, and April.

In thermal wind balance with the temperature response the modeled zonal-mean zonal wind anomalies also show a similar downward propagation (Fig. 3, bottom). Appearing during November in the mesosphere,
anomalies descend to the lower stratosphere by January and persist until early spring.

The modeled IGCM mean-state response to El Niño therefore shows a warming and weakening of the polar vortex, consistent with that found in other modeling studies (Sassi et al. 2004; Manzini et al. 2006) and are in good agreement with reanalysis data (Manzini et al. 2006; Garcia-Herrera et al. 2006).

b. Stratospheric variability

To give an indication of changes in planetary wave forcing and its associated effect on the wintertime polar vortex, Fig. 4 shows a scatterplot of the January–February–averaged meridional eddy heat flux anomaly at 100 hPa ($40^\circ$–$80^\circ$N), against the corresponding February–March temperature anomaly over the NH polar cap ($60^\circ$–$90^\circ$N) at 50 hPa. Anomalies are taken from the CTRL seasonal cycle.

In the CTRL experiment (crosses) a strong degree of correlation (0.67) between eddy heat flux and polar cap temperature suggests that higher temperatures in the midstratosphere are preceded by changes in eddy heat flux entering the lower stratosphere during midwinter. In general, the anomalies associated with recent El Niño events taken from the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) ReAnalysis (ERA-40; triangles) show a bias toward the top-right quadrant of the figure, indicating a warmer stratosphere in response to stronger wave driving. This relationship is captured in the IGCM with anomalies taken from the ELNO experiment (gray spots), showing a similar bias toward the top-right quadrant. In comparison with the absolute T31 heat flux values (Fig. 1), the mean change in heat flux of $\sim 3$ K m s$^{-1}$ represents a 25% increase in the planetary wave flux during El Niño events. This propagation of the El Niño signal into the stratosphere has been shown as being robust in observations (Garcia-Herrera et al. 2006), and a strengthening of the IGCM mean meridional circulation is confirmed by an increase in tropical upwelling of $5 \times 10^9$ kg s$^{-1}$ ($\sim 10\%$) at 100 hPa in the ELNO experiment compared with that in the CTRL run.

The large change in flux of vertically propagating planetary waves affects not only the stratospheric mean state, but also the variability of its circulation. To illustrate the change in variability more directly, Fig. 5 shows a bar chart of the monthly climatology of SSW frequency. The CTRL run compares favorably with the reference climatology set out by Charlton et al. (2007a) using the NCEP data (0.4 events per year compared to 0.6 events per year), although it underestimates the observed frequency of SSW in January. Immediately obvious is the increase in SSW events in all winter months for the ELNO experiment, with the overall winter frequency for DJFM increasing from 0.42 to 0.55 events per year. The largest increase is in January, consistent with the largest stratospheric zonal-mean zonal wind and temperature anomalies.

c. Canonical tropospheric winter response

The January–March extratropical MSLP response is shown in Fig. 6a and displays a structure that is highly reminiscent of the surface NAM pattern in its negative phase (Thompson and Wallace 2000), with significant height anomalies of opposite sign between the poles and midlatitudes. Over the PNA region the signature is one
that has been well established as a response to tropical SST forcing (Van Loon and Madden 1981; Hoerling et al. 1997; Lee et al. 2002). A deepening of the Aleutian low, and a shift in the center of action toward the western coast of North America gives rise to a wave train structure resembling the tropical–Northern Hemisphere pattern (Mo and Livezey 1986).

In the Euro-Atlantic sector the canonical late winter El Niño signature of a negative NAO-like pattern is captured well by the model. It features a dipole in MSLP, with centers of action over high-latitude Greenland and the Azores, giving a strongly zonal response over the European sector. The spatial structure is in good agreement with observations, although the node line is shifted southward. In terms of circulation changes, the result is a significant equatorward shift in the zonal-mean zonal winds that can be seen extending throughout the depth of the troposphere (Fig. 2b).

To quantify changes in circulation further, the NAO index for both the CTRL and the ELNO experiments is computed using a principal component–based method (Hurrell 1995). We define the NAO pattern as the leading EOF of daily zonal-mean MSLP anomalies from the control experiment within the region (20°–80°N, 70°W–40°E) from December to March. The first mode explains 44% of the total variance within the region and consists of a dipole in pressure with centers of action at 40° and 65°N. Indices for the CTRL and ELNO experiments are constructed by removing the daily seasonal cycle of the CTRL experiment from both MSLP time series and regressing anomalies onto the spatial pattern of the CTRL experiment.

Histograms of the daily January–March average NAO index for the CTRL and ELNO experiments are compared in Fig. 6b. A pronounced shift in the distribution toward negative NAO values is apparent for the ELNO
case, with a greater probability of late winter NAO indices less than −1.0 in response to tropical SST forcing compared to the CTRL.

The observed seasonal modulation in the MSLP signal over Europe (Moron and Gouirand 2003) is also captured by the IGCM with a shift in sea level pressure between November–December and January–March. To illustrate this, the time evolution of modeled MSLP anomalies in the Euro-Atlantic sector (zonally averaged between 30°E and 30°W) is shown in Fig. 7a. In November the response displays a positive NAO-like anomaly pattern with low pressure to the north and high pressure to the south. Moving into January, however, the relationship is reversed and the pressure signal projects negatively onto the NAO. Thus, the late winter
canonical signal, beginning in January, is captured and persists well into late spring.

4. Test of stratospheric influence

Comparison between the timing of anomalies in the NH polar cap (Fig. 3) and those in the European sector (Fig. 7a), suggests that the midwinter NAO surface response coincides with the arrival of positive temperature anomalies in the lower stratosphere. Following El Niño events, significant anomalies persist above the tropopause throughout late winter, which may contribute to the development and persistence of the canonical surface European MSLP signal. To test this possibility, a second IGCM experiment has been carried out following a method employed by Norton (2003). Both the CTRL and ELNO experiments are repeated, but with the lower limit of the applied Rayleigh drag (see section 2) moved from 1 down to 10 hPa. Two 40-yr runs are performed each with a “degraded stratosphere.” Responses (ELNO_D – CTRL_D) are investigated (where subscript D denotes degraded stratosphere) and compared to those from the full-stratosphere experiment (ELNO – CTRL).

Regarding the degradation method, the effect on the stratospheric mean state and variability is shown in Fig. 8 for the two control experiments—one having a normal Rayleigh drag profile (CTRL) and one with a lowered Rayleigh drag profile (CTRL_D). As expected, the most prominent effect on the January–February-averaged zonal-mean zonal wind is a large deceleration of the stratospheric polar vortex (right). In the mid- and lower stratosphere the jet is weakened by up to 30 m s\(^{-1}\) compared to that with the standard Rayleigh drag profile. The weakened jet is shifted from the high to midlatitudes throughout the depth of the stratosphere.

In terms of variability, the evolution of North Pole temperatures at 50 hPa for 40 yr is shown for both CTRL and CTRL_D experiments in Fig. 9. In the CTRL experiment (left), as the vortex intensifies from October onward, polar temperatures undergo a cooling. However, abrupt departures from the mean state also occur during winter, associated with SSW events. In the CTRL_D experiment the polar temperatures remain relatively constant and warm throughout winter, and the zonal-mean flow does not exhibit any SSW events, with a much reduced variability.

This demonstrates that the degraded polar vortex remains in a weak westerly state, with large anomalies (SSW) strongly damped. Thus, any stratospheric response to El Niño in the degraded-stratosphere experiments will be strongly suppressed and the surface response can be interpreted as effectively tropospheric in origin. A comparison between the full-stratosphere and degraded-stratosphere experiments will therefore
help indicate the importance of the inclusion of stratospheric processes in the teleconnection.

a. Wintertime tropospheric response

In comparison with the canonical winter tropospheric response of the fully resolved stratosphere experiments (section 3), Fig. 10 (left) shows the modeled response taken from the experiments with degraded stratospheric flow. The extratropical MSLP pressure response over the North Pacific sector takes on a structure very similar to that shown in Fig. 6a, both in terms of magnitude and spatial structure. The Aleutian low is deepened, although not as strongly, and shifted to the North American coast.

At high latitudes however the responses differ greatly between that of the full and degraded-stratosphere experiments. The MSLP signal in the degraded-stratosphere experiment is much less annular, with the positive center of action being weaker and shifted from the polar cap into northwest Europe. Although there appears to be a NAO-like dipole over the North Atlantic, in mainland Europe the MSLP signal takes on a much less zonal NAO response than that shown in the observations and reproduced in the full-stratosphere experiments. Significant
anomalies of ~2 hPa are centered over the United Kingdom and Ireland and reach southeastward toward the Mediterranean, with the region dominated by anomalous high pressure.

As a test of the Rayleigh drag approach, an alternative method was also employed for comparison. The CTRL and ELNO experiments were both repeated, except that a simple linear relaxation of the zonal-mean zonal winds above 10 hPa toward a daily control climatology was employed. This has the advantage of damping the variability of the vortex, and hence the stratospheric response to El Niño forcing, but without strongly affecting the mean state. The same relaxation time scale as the Rayleigh damping was used. The MSLP response using this approach is shown in Fig. 10 (right), and the results are extremely similar to that of the degraded-stratosphere experiments.

The time evolution of the European MSLP differences between the CTRL\(_D\) and ELNO\(_D\) simulations can be compared to the corresponding diagnostic from the full-stratosphere experiment, and is shown in Fig. 7b. In November a negative pressure anomaly is centered at 50°N. This gives way to the large positive anomaly over western Europe in December, reaching ~3 hPa in January and lasting until March.

The degraded-stratosphere MSLP signal therefore not only takes on a contrasting spatial structure in comparison to that of the full-stratosphere experiment, but also a different temporal evolution, which has less persistence into spring. Only in the experiments with full-stratospheric variability included does the response over Europe resemble observations, suggesting an active stratospheric role in the teleconnection.

b. Origins of European winter response

Interestingly, in the modeling study of Ineson and Scaife (2009), for El Niño winters without SSW events, the response over Europe is characterized by a large area of high pressure to the west of Europe. A very similar structure was shown by Toniazzo and Scaife (2006) to follow only the strongest observed El Niño events. The authors demonstrated a clear nonlinearity in the MSLP response over Europe with El Niño SST amplitude, and through ray-tracing diagnostics attributed the observed high pressure signal associated with strong El Niño events to wave trains following an entirely tropospheric pathway from the tropical Atlantic. Their results suggest strongly that, in observations of the largest-amplitude El Niño events, a tropospheric pathway dominates the European response over any stratospheric influence.

Figure 11 (top) illustrates the observed European responses to moderate (left) and strong (right) El Niño events, reproduced from Toniazzo and Scaife (2006) using the second Hadley Centre mean sea level pressure (HadSLP2) dataset (online at http://hadobs.metoffice.com/gmslp/hadslp2/index.html). The observed moderate events take on the canonical NAO response, and compare well to the modeled full-stratosphere response (middle left). The strongest observed events are dominated by the above-mentioned positive center of action to the west of Europe. If the tropospheric pathway does indeed dominate the European response to strong El Niño events, then we would expect a similar signal to be found in the degraded-stratosphere experiments, which by construction permit a European response that can only be tropospheric in origin. The similarity to
that of the observed strong response is shown clearly in Fig. 11 (middle right).

We therefore propose that the observed European response to El Niño consists of a superposition between both tropospheric and stratospheric influences. The SST forcing employed in our experiments falls into a “moderate” category, and we have shown successfully that a modeled stratospheric response is required in order to reproduce the canonical European NAO signal. However, by its nature any stratospheric influence on the extratropical tropospheric circulation is subtle and intermittent, largely being restricted to periods in which the vortex is in a strongly anomalous flow regime, for example, SSW events. Thus, for stronger El Niño events a mechanism may be in place whereby, for large enough SST forcing, the tropospheric pathway dominates the surface response and any stratospheric influence on the surface becomes “saturated.” By saturation we imply that, following a SSW event, the polar vortex has a natural limit to its dynamical warming, irrespective of the forcing strength. The frequency of SSW events may increase during periods of stronger El Niño forcing; however, following the breakdown of the vortex, any MSLP response at the surface will also have a natural limit.

To examine this hypothesis, we have performed a simple sensitivity test that aims to show that, for strong
tropical Pacific SST forcing, the European response is insensitive to stratospheric representation. We repeat the full and degraded-stratosphere experiments with the imposed composite SST anomalies doubled in magnitude to represent very strong El Niño cases. If the proposed saturation mechanism exists then the European responses should remain similar in structure, and independent of any stratospheric response.

First, the full-stratosphere experiments discussed in section 3 are repeated with the stronger SST forcing. The European sector MSLP late winter response is shown in Fig. 11 (bottom left). The figure shows that even with the inclusion of a fully resolved stratospheric response, the surface MSLP field is dominated over northwest Europe by a high pressure center of action. This is similar to the strong El Niño response found in observations, demonstrating the predominance of a tropospheric pathway over any stratospheric influence.

Similarly, by repeating the strong El Niño experiment in the degraded-stratosphere version of the model, the European response should remain similar, consistent with a purely tropospheric pathway. Figure 11 (bottom, right) shows the resulting MSLP signal when any stratospheric influence is removed. The response is characterized by the same high pressure dominating mainland Europe. For the strongest El Niño cases the European surface response therefore appears insensitive to any influence from the stratosphere and provides evidence for a saturation mechanism in the teleconnection.

5. Summary and discussion

Mechanisms to explain the spatial structure and temporal evolution of the European surface response to El Niño remain uncertain. However, given the coincidence in timing between high-latitude lower-stratospheric anomalies and the observed canonical NAO-like surface response, it is not unreasonable to suggest that one possible pathway for the European response to El Niño is via the stratosphere. The potential of a stratospheric pathway has received recent attention (Brönnimann 2007); however, few studies have explicitly investigated any mechanism involving the stratosphere. This study directly examines such a possibility by performing sensitivity studies using a GCM with a well-resolved stratosphere.

The canonical tropospheric and stratospheric responses to El Niño have been reproduced in the IGCM using a composite tropical Pacific SST forcing. The warming and weakening of the polar vortex reported in other modeling and observational studies are well captured and appear to be facilitated by an increase in frequency of SSW events throughout NH winter. At the surface the model fully reproduces the observed extratropical canonical signal of a negative surface NAM pattern, and over Europe the zonally orientated NAO response, with its correct seasonal modulation, is well captured.

While these stratospheric and tropospheric canonical responses have been reported in other studies, the results of section 3 aim to illustrate the capability of the IGCM in simulating not only the observed spatial and temporal responses, but also the results achieved with more complex GCMs (Manzini et al. 2006). The model therefore serves as a useful tool in examining the role of the stratosphere in the teleconnection.

Any mechanism for the spatial and temporal modulation of the canonical surface response over Europe to El Niño currently remains unexplained. However, the development of a significant negative NAO dipole during midwinter (Fig. 7a) coincides strongly with the arrival and persistence of anomalies in the lower stratosphere (Fig. 3). Therefore, to directly examine any stratospheric influence in the European surface response we repeat the earlier experiments using a version of the model in which the mean state and variability of the stratosphere are strongly degraded. The lower limit of the Rayleigh drag profile is moved down from 1 to 10 hPa in order to artificially damp the propagation of planetary waves into the stratosphere, and suppress any influence of the stratospheric response to El Niño on the underlying troposphere. The approach taken here is novel in that our methodology relies on a degradation of the stratospheric flow, rather than a high/low model-top comparison.

The key finding of the degraded-stratosphere experiments is that, unlike the response found in the full-stratosphere experiments, the model is unable to reproduce the canonical negative NAO-like signal. The observed zonally orientated dipole in MSLP over Europe is not captured, and instead a positive pressure anomaly persists over northwest Europe. The results of our experiments strongly suggest a sensitivity in the modeled El Niño teleconnection to the inclusion of any response in the stratosphere.

Recent modeling studies investigating the stratospheric role in the El Niño teleconnection have provided results that support our findings. Both Cagnazzo and Manzini (2009) and Ineson and Scaife (2009) conclude that the frequency modulation of SSW events plays a significant role in establishing the late winter canonical NAO signal, thereby reinforcing the stratospheric role in the teleconnection.

We have also investigated the nature of the observed European response by examining the nonlinearity found in observations between moderate and strong El Niño cases. We find evidence to suggest a balance between
the tropospherically forced response and the influence of any stratospheric response on the surface climate, by means of a saturation mechanism. For moderate El Niño events the stratosphere plays an active role in influencing the observed canonical MSLP signal. However, for the strongest El Niño events, the European response appears insensitive (in our model) to any stratospheric pathway in the teleconnection, appearing as a positive MSLP anomaly over mainland Europe. The lack of any stratospheric component to the teleconnection in low-top GCM studies may explain their inability in attaining the canonical response (e.g., Toniazzo and Scaife 2006).

Finally, our modeling study addresses the broader question concerning the need for adequate stratospheric representation within general circulation models. In terms of climate, the majority of modeling studies, such as those used in the recent Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4), which examine important tropospheric processes such as El Niño, typically have model upper boundaries near 10 hPa (Cordero and Forster 2006) and poorly resolve stratospheric processes. Our results have implications not only for climate modeling but for seasonal forecasting whereby realistic inclusion of the stratospheric response to El Niño may afford a source of potential predictability during European winter. Also, it appears that similar mechanisms for the communication of low-frequency boundary forcing involving the stratosphere may not be restricted to that of El Niño, and could be present in other seasonally varying forcings, such as Eurasian snow cover (Fletcher et al. 2007).

In summary, we have demonstrated that the stratosphere plays an active role in the El Niño–European teleconnection, and that without an adequate representation of the stratospheric pathway, the tropospheric response to El Niño, which is the largest source of interannual variability in the atmosphere, may not be accurately reproduced.

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