Understanding the Anomalously Cold European Winter of 2005/06 Using Relaxation Experiments

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ABSTRACT

Experiments with the atmospheric component of the ECMWF Integrated Forecasting System (IFS) have been carried out to study the origin of the atmospheric circulation anomalies that led to the unusually cold European winter of 2005/06. Experiments with prescribed sea surface temperature (SST) and sea ice fields fail to reproduce the observed atmospheric circulation anomalies suggesting that the role of SST and sea ice was either not very important or the atmospheric response to SST and sea ice was not very well captured by the ECMWF model. Additional experiments are carried out in which certain regions of the atmosphere are relaxed toward analysis data thereby artificially suppressing the development of forecast error. The relaxation experiments suggest that both tropospheric circulation anomalies in the Euro–Atlantic region and the anomalously weak stratospheric polar vortex can be explained by tropical circulation anomalies. Separate relaxation experiments for the tropical stratosphere and tropical troposphere highlight the role of the easterly phase of quasi-biennial oscillation (QBO) and, most importantly, tropospheric circulation anomalies, especially over South America and the tropical Atlantic. From the results presented in this study, it is argued that the relaxation technique is a powerful diagnostic tool to understand possible remote origins of seasonal-mean anomalies.

1. Introduction

It is well known that persistent large-scale extratropical circulation anomalies such as the North Atlantic Oscillation (NAO) have a profound impact on the climate of populated areas such as Europe and North America (e.g., van Loon and Rogers 1978; Hurrell 1995). Attempts have therefore been made to understand the mechanisms that drive extratropical atmospheric circulation anomalies. It is now widely accepted that a large part of the extratropical variability in the North Atlantic region is governed by internal atmospheric processes (e.g., Kushnir et al. 2002; Rowell 1998), especially on seasonal and interannual time scales. This suggests that predictability of such anomalies is limited to a few weeks. There is observational and modeling evidence, however, that the atmosphere in the North Atlantic region is also affected (i) locally by North Atlantic sea surface temperature (SST) anomalies (e.g., Czaja and Frankignoul 1999; Rodwell and Folland 2002; Rodwell et al. 1999; Latif et al. 2000) and (ii) remotely by tropical SST anomalies via atmospheric teleconnections (e.g., Fraedrich 1994; Greatbatch and Jung 2007). Furthermore, it has been suggested that the Northern Hemisphere stratosphere may provide some additional memory that could result in some useful monthly and seasonal forecast skill (e.g., Baldwin et al. 2003; Scaife and Knight 2008). However, the relative impact of the North Atlantic, the tropics, and the extratropical stratosphere has yet to be assessed.

In this study, which can be seen as an extension of the paper by Jung et al. (2008), a diagnostic technique is introduced—the so-called relaxation or nudging technique—which has the potential to help understand possible “remote” influences in the generation of extratropical atmospheric circulation anomalies. The relaxation technique has been widely used by the atmospheric science community on relatively shorter “weather” time scales (Kalnay 2003; Bauer et al. 2008). Here it will be illustrated as a diagnostic tool to understand processes on longer seasonal and climatic time scales.

To this end the anomalously cold European winter of 2005/06 makes an interesting case study for various reasons. First, it was the coldest winter in Europe in about a decade (Scaife and Knight 2008), which was brought
about by an increased frequency of occurrence of Euro–Atlantic blocking events (Croci-Maspoli and Davies 2009). This increase becomes apparent in the form of a seasonal-mean anticyclonic anomaly in geopotential height fields at the 500-hPa level (hereafter Z500, Fig. 1a) over the eastern North Atlantic extending to the Barents and Kara Seas. Second, most seasonal forecasting systems showed some skill in predicting the anomalously cold temperatures several months in advance (Graham et al. 2006; Folland et al. 2006) suggesting that some external forcing (or slowly varying internal dynamics) might have played a role. Finally, the winter of 2005/06 was marked by the presence of a number of seasonal anomalies, in both the Northern Hemisphere extratropics and in the tropics, which might explain the anomalous atmospheric circulation giving rise to the cold European winter of 2005/06.

Observed anomalies for the winter of 2005/06 and various different parameters are shown in Fig. 1. Evidently, in a seasonal-mean sense, the tropospheric circulation anomaly in the Euro–Atlantic region is accompanied by an anomalously weak stratospheric vortex (Fig. 1b). This, along with results from numerical experiments in which a stratospheric forcing is applied that imposes the observed sudden stratospheric warming in January 2006, which lasted for several weeks and explains most of the stratospheric seasonal-mean anomalies, led Scaife and Knight (2008) to conclude that the stratospheric warming has contributed to the cold European winter of 2005/06.

As previously mentioned, numerous studies have argued that North Atlantic SST anomalies were crucial. Folland et al. (2006), for example, point out that the statistical prediction scheme of Rodwell and Folland (2002) was successful in predicting the anomalously cold European winter and the physical basis of the Rodwell and Folland (2002) scheme includes North Atlantic SST anomalies affecting the atmospheric circulation as one key component. Although the Rodwell and Folland (2002) scheme employs SST in both the tropical and extratropical part of the North Atlantic, usually the role of the extratropics is highlighted though the so-called reemergence mechanism (Namias and Born 1970; Alexander et al. 1999). Synoptic-dynamical diagnosis of the 2005/06 winter by Croci-Maspoli and Davies (2009) also points to the importance of North Atlantic SST anomalies (and surface temperatures over North America). Croci-Maspoli and Davies (2009) argue that the Euro–Atlantic region is sensitive to upstream cloud–diabatic processes, which in turn is sensitive to near-surface temperature.

Closer inspection of Fig. 1 also reveals anomalies in the tropics. The winter of 2005/06 was marked, for example, by a La Niña event of moderate (Fig. 1c) strength that had a marked impact on outgoing longwave radiation (OLR; Fig. 1d) and the velocity potential at 200 hPa (χ200, Fig. 1e). The so-called “canonical” link between La Niña and the atmospheric circulation in the North Atlantic region (Fraedrich 1994; Gouriard and Moron 2003; Greatbatch and Jung 2007) predicts a positive phase of the North Atlantic Oscillation (NAO), that is, the opposite of what was observed (i.e., −0.42 standard deviations below average; Folland et al. 2006). Relatively strong negative SST anomalies are also found in the Indian Ocean. The fact that cloudiness in the area of this negative SST anomaly was reduced during the 2005/06 winter (Arguez et al. 2007, see their Fig. 2.12) suggests that the atmospheric anomalies are the response to the anomalously cold SST. The modeling study of Bader and Latif (2003) finds that the warming of the Indian Ocean in recent decades leads to an increased NAO via the jet stream waveguide, suggesting that Indian Ocean SST anomalies can have an influence on the atmospheric circulation in the Euro–Atlantic region (see also Hoerling et al. 2001). Furthermore, strong tropospheric anomalies were also evident over South America and over North Africa, both of which, potentially, may have triggered a Rossby wave response over the North Atlantic (Hoskins and Ambrizzi 1993).

Finally, the winter of 2005/06 was marked by the negative phase of the quasi-biennial oscillation (QBO, see Fig. 1f). According to Holton and Tan (1980), the negative phase of the QBO leads to a weakening of the Northern Hemisphere stratospheric polar vortex, which in turn may lead to the negative phase of the NAO through “downward propagation” of stratospheric anomalies (Baldwin and Dunkerton 1999). In fact, in a more recent study, Boer and Hamilton (2008) find a significant link between the phase of the QBO and the tropospheric circulation, especially in the North Atlantic region (see also Marshall and Scaife 2009).

The paper is organized as follows. In the next section the relaxation technique and its use in the present study will be described in some detail. The results section starts with a discussion of seasonal-mean anomalies. In this context, the influence from the tropics will be studied in considerable detail. The influence from the tropics will then be compared with the role played by the Northern Hemisphere stratosphere. This is followed by a short discussion of possible extratropical–tropical interactions. The section on seasonal-mean circulation anomalies finishes with an investigation into the sensitivity of results to details of the relaxation formulation. In the second part of the results section, the intraseasonal evolution during the 2005/06 winter will be discussed. The paper closes with a discussion of the results.
FIG. 1. Observed mean anomalies for the period 1 Dec 2005–28 Feb 2006: (a) 500-hPa geopotential height (contour interval is 20 m), (b) 50-hPa geopotential height (contour interval is 50 m), (c) SST (in K), (d) OLR (in W m\(^{-2}\)), (e) velocity potential at 200 hPa (contour interval is 0.5 m\(^2\) s\(^{-1}\)), and (f) zonal wind at 50 hPa (contour interval is 3 m s\(^{-1}\)). Negative (positive) values in (a), (b), (e), and (f) are dashed (solid). Notice that in (c) and (d) positive values are contoured (as well as shaded). All results are based on data from ERA-Interim (Simmons et al. 2007), but (d), which is based on estimates of OLR from NOAA satellites (Liebmann and Smith 1996).
2. Methodology

a. Experimental setup

The numerical experimentation, carried out in this study, is based on a recent version of the European Centre for Medium-Range Weather Forecasts (ECMWF) atmosphere model (cycle 32R1 used operationally from 5 June to 5 November 2007). All forecast experiments employ a horizontal resolution of $T_{L95}$ (linear Gaussian grid $\approx 1.85^\circ \times 1.85^\circ$) with 60 levels in the vertical. About half of the levels are located above the tropopause (Untch and Simmons 1999) extending up to 0.1 hPa. All experiments were carried out using initial conditions and lower boundary conditions (daily SST and sea ice) from the ECMWF Re-Analysis (ERA-Interim; Simmons et al. 2007). Aspects of the model’s performance are discussed elsewhere (Jung 2005; Jung et al. 2010b).

b. Relaxation formulation

To understand the origin of the anomalously cold winter of 2005/06, a large number of seasonal forecast experiments with and without relaxation have been carried out. The experiment without relaxation constitutes the control integration (CNT). The control integration is used to understand the possible influence of SST and sea ice anomalies. In the relaxation experiments the model is drawn toward ERA-Interim reanalysis data (Simmons et al. 2007) in a specific region during the course of the integration; this is achieved by adding an extra term of the following form to the ECMWF model:

$$-\lambda(x - x_{\text{ref}}).$$

The model state vector is represented by $x$ and the reference field toward which the model is drawn by $x_{\text{ref}}$. The strength of the relaxation is determined by $\lambda = a\lambda_0$, where $a$ defines the geographic region and model levels where the relaxation is applied. Here $\lambda_0$ defines the time scale of the relaxation. Unless stated otherwise $\lambda_0 = 0.1 \text{ h}^{-1}$ is used throughout the study. For a time step of 1 h used here a value of 0.1 h$^{-1}$ indicates that at each time step the model tendency is “corrected” using 10% of the departure of $x$ from $x_{\text{ref}}$. In this study the parameters being relaxed include $u$, $v$, $T$, and $\ln p_s$. Notice, that $\ln p_s$ is not relaxed for stratospheric relaxation experiments; $x_{\text{ref}}$ is deduced from the 6-hourly, $T_{L255}$ ERA-Interim reanalysis linearly interpolated in time to 1-hourly values and spatially interpolated to $T_{L95}$ by applying a sophisticated horizontal interpolation package [used routinely within the ECMWF Integrated Forecasting System (IFS)].

Table 1. Summary of the main seasonal forecast experiments used in this study. Unless mentioned otherwise, $\lambda_0 = 0.1 \text{ h}^{-1}$ is used throughout.

<table>
<thead>
<tr>
<th>Expt</th>
<th>Relaxation region</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNT</td>
<td>No relaxation</td>
</tr>
<tr>
<td>TROP</td>
<td>20°S–20°N, 0°–360°</td>
</tr>
<tr>
<td>TROP-T</td>
<td>Troposphere + stratosphere</td>
</tr>
<tr>
<td>TROP-S</td>
<td>Troposphere*</td>
</tr>
<tr>
<td>TROP-T/30°–90°E</td>
<td>Stratosphere**</td>
</tr>
<tr>
<td>TROP-T/150°E–120°W</td>
<td>Troposphere*</td>
</tr>
<tr>
<td>TROP-T/90°W–0°</td>
<td>Troposphere*</td>
</tr>
<tr>
<td>NH</td>
<td>30°–90°N, 0°–360°</td>
</tr>
<tr>
<td>NH-S</td>
<td>Troposphere + stratosphere**</td>
</tr>
</tbody>
</table>

* Actual strength of the relaxation at 500, 200, 50, and 20 hPa is approximately $0.99\lambda_0$, $0.018\lambda_0$, $8.3 \times 10^{-3}\lambda_0$, and $1.5 \times 10^{-2}\lambda_0$ h$^{-1}$, respectively.

** Actual strength of the relaxation at 500, 200, 50, and 20 hPa is approximately $1.1 \times 10^{-2}\lambda_0$, $2.3 \times 10^{-4}\lambda_0$, $0.018\lambda_0$, and $0.5\lambda_0$ h$^{-1}$, respectively.

To allow for an effective localization, the relaxation was carried out in gridpoint space. When applying masks to localize the relaxation, care has to be taken in order to reduce adverse effects close to the relaxation boundaries. Here the transition from relaxed to nonrelaxed regions in the horizontal is smoothed using the hyperbolic tangent. The smoothing is such that the relaxation coefficient $\lambda$ goes from $\lambda_0$ to 0 within a 20° belt, both in longitude and latitude. Boundaries stated in the text refer to the center of the respective 20° belt. To reduce the generation of spurious potential vorticity features, changes of $\lambda$ are also smoothed in the vertical. Here, the relaxation coefficient effectively goes from $\lambda_0$ to 0 in a vertical layer encompassing about 13 model levels (see Table 1 for actual values of $\lambda$ at various heights).

c. Forecast experiments

For the winter of 2005/06 a set of seasonal ensemble forecasts with and without relaxation was carried out. The ensembles were generated by starting forecasts in 6-hourly intervals from 1200 UTC 16 November to 1200 UTC 20 November 2005 giving a total of 17 ensemble members. A summary of all seasonal forecast experiments along with their abbreviations is given in Table 1.

Throughout this paper “anomalies” refer to departures of the ensemble mean or individual ensemble members from the climate of the model. The climate of the model was obtained from calibration runs. These runs are single integrations (i.e., one ensemble member) covering winters of the period 1990/91 to 2005/06. Forecasts were started at 1200 UTC 15 November. Notice that
the calibration runs for the relaxation experiments were carried out with the same relaxation as for the respective winter 2005/06 ensemble experiment. By doing this, the anomalies reflect the anomalous conditions during the 2005/05 winter rather than the remote influence of reduced systematic errors in the relaxation regions.

d. Statistical significance testing

Throughout this study the statistical significance of anomalies, defined in section 2c, is tested using a two-sample, two-sided Student’s $t$-test taking serial correlation into account.

3. Results

a. Seasonal-mean diagnostics

1) TROPICAL VERSUS STRATOSPHERIC INFLUENCES

Observed Z500 anomalies for the 2005/06 winter are shown in Fig. 2 alongside corresponding anomalies for the control experiment with observed SST/sea ice (CNT), the tropical relaxation experiment (TROP), and the experiment with relaxation of the Northern Hemisphere stratosphere (NH-S). Figure 2b shows that prescribing...
the observed SST/sea ice fields is not sufficient to reproduce the observed circulation anomalies in an ensemble-mean sense, especially over North America, the North Atlantic, and Europe. The Z500 response produced by TROP is highly significant and resembles the negative phase of the Arctic Oscillation–North Atlantic Oscillation (AO–NAO; Thompson and Wallace 1998; Walker 1924). Especially over North America, the North Atlantic, and Europe, the ensemble-mean response to tropical relaxation closely resembles the observed anomalies. The influence of the Northern Hemisphere stratosphere, NH-S, on Northern Hemisphere Z500 anomalies is weaker and different in terms of its spatial structure compared to that from the tropics. The Northern Hemisphere Z500 response for NH-S shows a significant anticyclonic circulation anomaly in the eastern North Atlantic, which shows little resemblance to the AO–NAO-like response expected to arise from the “downward propagation” of polar vortex anomalies (e.g., Baldwin and Dunkerton 1999; Ambaum and Hoskins 2002; Jung and Barkmeijer 2006). That the ensemble-mean responses are very different for TROP and NH-S can be inferred from the fact that spatial pattern correlation coefficients between the two fields for both the Northern Hemisphere ($r = 0.1$) and the Euro–Atlantic region ($r = 0.3$) are very small.

Although tropical relaxation does produce significant atmospheric circulation anomalies that reduce the westerly flow into middle and northern Europe, there are differences between the ensemble-mean response and the observed anomalies (Figs. 2a,c). Care has to be taken, however, when interpreting these differences. This is because a single realization (the observed trajectory) is compared with an average over 17 realizations (the ensemble mean). Figure 3 shows the response of each of the 17 ensemble members to the tropical relaxation. Evidently, there is considerable spread about the ensemble mean. Spatial correlation for Z500 anomalies between the ensemble mean and individual ensemble members range from $r = 0.92$ ($r = 0.96$) for the best member to $r = 0.47$ ($r = 0.36$) for the worst member over the Northern Hemisphere (Euro–Atlantic region). The spatial correlation between the ensemble mean and the observations is $0.57$ ($0.71$). By this measure the observed anomalies are closer to the ensemble-mean response than the “worst” ensemble members. These result show that considerable differences of the ensemble-mean response from observed anomalies can be expected for single-season case studies, making it difficult to decide whether the simulated response is realistic or not. The extent to which this is a problem depends on the ratio of signal (ensemble-mean anomaly) to noise (ensemble spread).

So far, the results suggest that primarily the tropical anomalies and secondarily the anomalously weak stratospheric polar vortex contributed to the tropospheric circulation anomalies observed during the 2005/06 winter. Figure 4 shows observed 50-hPa geopotential height (Z500) anomalies; also shown are ensemble-mean anomalies for CNT and TROP. Not surprisingly, the Z500 anomalies produced by NH-S are very similar to the observations and therefore not shown. CNT shows weak and nonsignificant Z500 anomalies suggesting that the observed SST and sea ice anomalies contributed little to the anomalously weak stratospheric polar vortex. The ensemble mean for TROP, on the other hand, reproduces the weakened stratospheric polar vortex, with the observations lying within the ensemble spread. These results suggest that the origin of anomalously weak stratospheric vortex during the 2005/06 winter lies in the tropics.

2) FURTHER EXPLORING THE TROPICAL INFLUENCE

Velocity potential anomalies at the 200-hPa level (i.e., $\chi_{200}$ anomaly) are shown in Fig. 5 for ERA-Interim, CNT, and TROP. The control integration with observed SST and sea ice distribution captures the anomalous convergent flow (positive $\chi_{200}$ anomaly) in the central tropical Pacific associated with the La Niña conditions both in terms of the structure and size of the anomaly. In other parts of the tropics, however, CNT fails to reproduce the observed $\chi_{200}$ anomalies. The fact that CNT fails to reproduce the observed Z500 anomalies over the Northern Hemisphere (Fig. 2b) suggests that La Niña was not responsible for the extratropical response; consequently the origin of the anomalous European 2005/06 winter seems to lie outside the central tropical Pacific region. The fact that TROP, which shows a stronger and more realistic extratropical response, captures the observed $\chi_{200}$ anomalies very well, shows that the tropical relaxation is efficient in imposing the observed tropical anomalies.

As mentioned in the introduction, the 2005/06 winter was marked by the easterly phase of the QBO. Consistent with the observational study by Holton and Tan (1980) the negative phase of the QBO during the winter 2005/06 is associated with an anomalously weak stratospheric polar vortex (Fig. 6a). Interestingly, CNT is able to reproduce the easterly phase of the QBO; CNT fails, however, to produce the observed weakening of the stratospheric polar vortex (Fig. 6b). A more detailed investigation reveals that CNT simulates the observed QBO structure by persisting the anomalous initial conditions throughout the whole winter (not shown). Persistence of QBO anomalies has been found in relatively low-resolution versions of the ECMWF model before
FIG. 3. Geopotential height anomalies at the 500-hPa level (contour interval is 20 m) for the period 1 Dec 2005–28 Feb 2006: (a) ERA-Interim, (b) ensemble mean, and (c)–(s) individual ensemble members for TROP. Values in parenthesis give spatial correlation coefficients of the respective anomaly with that of the ensemble mean for the Northern Hemisphere (north of 30°N)/Euro–Atlantic region (30°–70°N, 80°W–40°E).
At the first glance the results for CNT suggest that the Holton–Tan mechanism was not crucial during the 2005/06 winter. However, it is worth pointing out that (i) the QBO anomaly in CNT weakens throughout the 3-month period leaving it rather weak by the end of the winter and that (ii) CNT does not capture the downward propagation of zonal wind anomalies leading to relatively large errors during the second part of the winter when the stratospheric warming occurred. Alternatively, it might be argued that the Holton–Tan mechanism does work in CNT but that it is weak and, hence, that other signals dominate (see below for further discussion).

After having presented evidence for a link (possibly causal) between the tropical and extratropical anomalies the question arises as to which region of the tropical atmosphere is most strongly involved. First, the forcing associated with the tropical troposphere is separated from that associated with the tropical stratosphere. Such an approach seems physically reasonable given that different processes are likely to be crucial for explaining the observed anomalies in these two parts of the tropical atmosphere. This notion is further supported by the fact that relaxation of the tropical troposphere only (TROP-T) has a negligible impact on the tropical stratosphere (in terms of zonal-mean zonal wind anomalies, not shown); similarly, relaxation of the tropical stratosphere (TROP-S) has a very small impact on the tropical troposphere (in terms of $x_{200}$ anomalies, not shown). Figure 7 shows the extratropical response for TROP-T and TROP-S in terms of Northern Hemisphere Z500 anomalies. In the Euro–Atlantic region, the tropical tropospheric influence is larger than that of the tropical stratosphere. Over the northwest North Pacific, on the other hand, tropospheric and stratospheric tropical influences seem to be comparable. Whereas the QBO seems crucial for explaining the role of the tropical stratosphere, the presence of multiple anomalies in the tropical troposphere makes it more difficult to identify the relevant tropospheric region(s). In the following an attempt is made to pinpoint the origin of the extratropical circulation anomaly regionally by relaxing different regions of the tropical troposphere. Here, the focus will be on three regions (cf. Fig. 1): (i) the Indian Ocean (30°–90°E) and its associated anomaly (TROP-T/30°–90°E); (ii) the tropical Pacific (150°E–120°W) capturing the circulation anomaly associated with the La Niña (TROP-T/150°E–120°W); and (iii) South America, the tropical Atlantic, and western parts of tropical Africa (TROP-T/90°W–0°). Figure 8 shows Northern Hemisphere Z500 anomalies for the three relaxation experiments. Relaxing the tropical atmosphere over the Indian Ocean clearly fails to explain the extratropical anomalies (Branković et al. 1994).
Z500 anomalies in TROP-T (cf. Figs. 7a and 8a). Over the Euro–Atlantic region (30°–70°N, 80°W–40°E) the pattern correlation between these two responses amounts to \( r = 0.23 \). For the experiment with relaxation in the tropical Pacific only, this correlation increases to \( r = 0.60 \). It seems necessary, however, to relax the tropical troposphere between 90°W and the Greenwich Meridian in order to best reproduce the Z500 response in TROP-T for the Euro–Atlantic region (\( r = 0.81 \), Figs. 7a and 8c). Similar conclusions are obtained if the local relaxation experiments are compared with the response for the full tropical relaxation experiments.

It is worth pointing out that the experiments with spatial localization in the tropical troposphere have to
be interpreted carefully. This is because our interpretation is based on the assumption of linearity. Furthermore, relaxation in a certain region of the tropical troposphere is likely to have an indirect effect on other remote tropical regions in which no relaxation has applied at all. This is particularly true for seasonal integrations in which the atmosphere has time to adjust to the forcing implied by the relaxation term.

3) EXTRATROPICAL FORCING OF TROPICAL ANOMALIES

So far, the focus has been on tropical–extratropical interactions. To correctly assess cause and effect it is crucial to study possible extratropical–tropical interactions as well. In fact, it is well known that the tropics do respond to extratropical forcing (e.g., Kiladis and Weickmann 1992; Hoskins and Yang 2000; Jung et al. 2010a). The experiment in which the whole Northern Hemisphere north of 30°N, NH, is relaxed toward reanalysis data is designed to study a possible extratropical forcing of the observed tropical anomalies. The NH produces tropical anomalies both in the troposphere and stratosphere that are very similar to those found in CNT (not shown). Given that NH also sees realistic SST in the tropics, this suggests that the extratropical forcing of tropical anomalies during the 2005/06 winter, if existent, was relatively weak compared with tropical–extratropical interactions and hence the sense of the causality is more likely from the tropics to the extratropics.

4) SENSITIVITY EXPERIMENTS

As mentioned in the methods section, the tropical relaxation experiment, TROP, has its northern relaxation boundary at 20°N (with a transition zone covering 10° on either side). Synoptic studies of the sudden stratospheric warming (SSW) in January 2006 show that tropospheric precursor waves in the North Atlantic extended partly into the subtropics (Coy et al. 2009; Nishii et al. 2009). To ascertain that the origin of the anomalous circulation in the Euro–Atlantic region is truly tropical an additional tropical relaxation experiment has been carried out in which the relaxation boundaries have been moved equatorward to 10°S and 10°N, respectively. The same latitudinal smoothing is applied.

Fig. 6. Average zonal-mean zonal wind anomalies (shading in m s⁻¹) for the period 1 Dec 2005–28 Feb 2006: (a) ERA-Interim, (b) CNT, and (c) TROP. (d) The difference between TROP and CNT. In (a) climatological average zonal-mean zonal wind anomalies from ERA-Interim (contour interval is 5 m s⁻¹, negative values are dashed) are superimposed. Statistically significant (b), (c) anomalies and (d) differences (at the 95% confidence level) are hatched. The same contour shading is used in (a)–(d).
as in the other experiments. Restricting the tropical relaxation to the tropical belt 10°S–10°N yields a very similar Z500 response to TROP over the Northern Hemisphere (cf. Figs. 2c and 9a) confirming the importance of the tropics.

Fig. 7. Geopotential height anomalies at the 500-hPa level (contour interval is 20 m) for the period 1 Dec 2005–28 Feb 2006: (a) TROP-T and (b) TROP-S. Results are based on ensemble-mean data. Statistically significant anomalies (at the 95% confidence level) are hatched.

Fig. 8. Geopotential height anomalies at the 500-hPa level (contour interval is 20 m) for the period 1 Dec 2005–28 Feb 2006: (a) TROP-T/30°–90°E, (b) TROP-T/150°E–120°W, and (c) TROP-T/90°W–0°. Results are based on ensemble-mean data. Statistically significant anomalies (at the 95% confidence level) are hatched.
The choice of the relaxation coefficient ($\lambda_0 = 0.1$ h$^{-1}$) in this study is somewhat arbitrary. Therefore, it is important to test whether the conclusions of this study are sensitive to $\lambda_0$. Figure 9b shows Z500 anomalies for a tropical relaxation experiment (20°S–20°N) with $\lambda_0 = 0.01$ h$^{-1}$. Evidently, the spatial structure of the Z500 response is not strongly affected by the exact choice of $\lambda_0$; however, reducing the strength of the relaxation reduces the magnitude of the Z500 response, especially in the subtropics and midlatitudes (cf. Figs. 2c and 9b).

One might argue that some of the results presented in this study may depend on the model used to carry out the experiments. To address this issue the two experiments CNT and TROP have been repeated using the more recent model version 33R1. This model cycle was used operationally at ECMWF from 3 June to 29 September 2008. Compared with model version 32R1, which is the main model version employed in this study, model version 33R1 comprises substantial changes to almost every part of the ECMWF physics package (Bechtold et al. 2008; Jung et al. 2010b). The climate of the tropics (Bechtold et al. 2008) and the atmospheric circulation of the Northern Hemisphere extratropics (Jung et al. 2010b) are much improved in version 33R1. Figure 10 shows Northern Hemisphere Z500 anomalies for CNT and TROP based on ECMWF model version 33R1. The control integration with the newer model version appears to perform better in the Northern Hemisphere high latitudes. For the relaxation experiments, the results are very similar to the ones obtained with the older model version (Figs. 2b,c) suggesting that the conclusions of this study are not overly sensitive to model formulation and systematic model error and that they are statistically robust (reproducible).

b. Intraseasonal evolution

So far, the focus has been on seasonal-mean fields for the whole winter. However, the 2005/06 winter was marked by large intraseasonal changes, especially in the Northern Hemisphere stratosphere. In fact, Nishii et al. (2009) point out that the zonal-mean polar night jet weakened gradually from late December and then rapidly became easterly at the end of January. Therefore, taking the seasonal evolution of circulation anomalies into account rather than focusing solely on seasonal-mean anomalies should help to improve our understanding about what happened during the 2005/06 winter.

The observed temporal evolution of the strength of the stratospheric polar vortex during the 2005/06 winter can be inferred from Fig. 11 (solid black). Here the strength of the stratospheric polar vortex is defined by area-averaged Z50 anomalies north of 70°N; the larger the Z50 anomaly, the weaker the vortex. The control integration, CNT, shows some interesting intraseasonal variability: the first half of the winter is marked by an anomalously weak vortex in agreement with the observations; during the second half, however, the stratospheric polar vortex in CNT gradually intensifies rather than weakens as shown by the observations. The experiment TROP shows a gradual decrease of the strength of
the polar vortex throughout the winter. Inspection of the individual ensemble members for TROP reveals that the observations lie well within the ensemble throughout the whole winter (Fig. 12b). This is in contrast to the CNT ensemble, which only marginally captures the observation toward the end of the winter (Fig. 12a).

As mentioned above, the 2005/06 winter was marked by strong circulation anomalies in both the tropical troposphere (e.g., La Niña) and the tropical stratosphere (easterly phase of the QBO). In an attempt to disentangle the influence of processes in the tropical troposphere from those in the tropical stratosphere, the temporal evolution of the strength of the stratospheric polar vortex is considered separately for TROP-T and TROP-S. During the first half of the 2005/06 winter both TROP-T and TROP-S show a moderately weak stratospheric polar vortex (Fig. 11). In an ensemble-mean sense, differences between TROP-T and TROP-S start to develop during the second half of the winter: whereas the stratospheric polar vortex remains moderately weak for TROP-S, it further weakens for TROP-T during the second half of the winter because of an increased frequency of occurrence of relatively large-amplitude stratospheric warmings (Fig. 12c). While both ensembles, TROP-T and TROP-S, do capture the observations during the second half of the 2005/06 winter (Figs. 12c,d), it should be pointed out that the observations for TROP-S (TROP-T) are located at the margin (center) of the implied probability density function (PDF). In summary, therefore, the results suggest that the negative phase of the QBO led to a moderate weakening of the stratospheric polar vortex; however, observed anomalies in the tropical troposphere are most likely to be responsible for the occurrence of the large-amplitude stratospheric warming during the second half of the 2005/06 winter.

After having shown that it is useful to consider the intraseasonal evolution to shed further light on the mechanisms underlying the stratospheric conditions during...
the 2005/06 winter, in the following the low-frequency
intraseasonal evolution of the tropospheric circulation is
considered. To this end, the winter of 2005/06 is divided
into two parts, one representing early winter (1 December
2005–15 January 2006) and the other late winter (16
January–28 February 2006). A comparison of the ob-
served Z500 anomalies over the Northern Hemisphere
between early and late winter shows large intraseasonal
changes in the North Pacific; in other parts of the North-
eren Hemisphere, such as over Europe, tropospheric cir-
culation anomalies were more persistent throughout the
winter (Figs. 13a and 14a). The reversal of the anomalies
in the North Pacific is captured by all experiments in-
cluding CNT. To get a realistic representation of the
strong persistence of the Z500 anomalies in the Euro–
Atlantic region, on the other hand, it seems necessary to
impose the correct tropical forcing (Figs. 13 and 14). The atmospheric
La Niña response is especially clear for CNT and NH-S.
For TROP the atmospheric La Niña response seems to be
somewhat obscured by the presence of other circulation
anomalies over the Northern Hemisphere. The negative
PNA response in late winter in CNT could explain why
the strength of the stratospheric polar vortex increases in
CNT during the second half of the 2005/06 winter. In
fact, from diagnosis of observational data it has been
found that La Niña–type conditions are associated with
a strengthened stratospheric polar vortex (see Fig. 17 in
Brönnimann 2007). Therefore, one possible explanation
is that the negative phase of the PNA leads to a re-
duction of the stationary planetary wave amplitude and,
therefore, to a reduced slow down of the stratospheric
polar vortex through the reduced breaking of stationary
planetary waves of tropospheric origin (Taguchi and
Hartmann 2006; Ineson and Scaife 2008).

To test the idea of a negative correlation between
ENSO-related SST anomalies in the tropical Pacific and

**Fig. 12.** As in Fig. 11, but for all ensemble members (thin gray lines): (a) CNT, (b) TROP, (c) TROP-T, and
(d) TROP-S. Observed values from ERA-Interim are also shown (thick solid curves).
the strength of the stratospheric polar vortex more specifically for the ECMWF model, seasonal forecast experiments with the ECMWF model described in Greatbatch and Jung (2007) were diagnosed in more detail. Indeed, these experiments confirm that La Niña–type diabatic forcing leads to a strengthening of the stratospheric polar vortex in the ECMWF model (not shown).

4. Discussion

Numerical experiments with the ECMWF model have been carried out in order to understand the origin of the atmospheric circulation anomaly that led to the anomalously cold European winter of 2005/06. In contrast with most other previous studies, which explain observed atmospheric circulation anomalies primarily in terms of SST anomalies in the extratropical North Atlantic (Graham et al. 2006; Folland et al. 2006; Scaife and Knight 2008; Croci-Maspoli and Davies 2009), the relaxation experiments presented in this study indicate an important role for the tropical atmosphere. Further experimentation suggests that the largest forcing came from the tropical troposphere over South America, the Atlantic, and Africa. The results of this study further suggest that the easterly phase of the QBO also contributed to the observed circulation anomalies, especially in the Northern Hemisphere stratosphere. Scaife and Knight (2008) argue that the January 2006 sudden stratospheric warming is likely
to have contributed to the colder 2005/06 winter. While it cannot be excluded that the extratropical stratosphere might have increased the persistence of the cold spell, the results of this study suggest that the origin of the sudden stratospheric warming in January 2006 lies in the tropics.

Dynamical and statistical seasonal forecasts for the 2005/06 winter were relatively skilful. Previous studies have explained this relatively high level of skill in terms of North Atlantic SST anomalies (Graham et al. 2006; Folland et al. 2006; Scaife and Knight 2008). The results of this study provide an alternative perspective: good tropical forecasts, both for the stratosphere and especially the troposphere, were needed for accurate seasonal predictions.

Given the important role of the tropical atmosphere in forcing the extratropical circulation anomaly in the winter of 2005/06 suggested by this study along with widespread belief that low-frequency variations of the tropical atmosphere are largely driven by tropical SST anomalies, it seems surprising the control experiment with prescribed SST does not reproduce the cold European winter of 2005/06. The control integration cannot provide an accurate forecast because it fails to produce realistic tropospheric anomalies in parts of the tropics that seem to have mattered (i.e., South America, tropical Atlantic, and West Africa; Fig. 5). It is possible that the ECMWF model fails to respond realistically to the imposed SST anomalies. This might be either due to

FIG. 14. As in Fig. 2, but for the subperiod 16 Jan–28 Feb 2006.
model error or due to the fact that fixing SST can lead to an incorrect representation of air–sea interactions particularly in the Indian Ocean (e.g., Douville 2005; Copsey et al. 2006). It is also possible that land rather than sea surface anomalies mattered—and the land surface conditions were not prescribed in the control integration. In fact, the largest $\gamma_{200}$ anomalies in the tropical area 90°W–0° are found over tropical land areas (Fig. 1e). This conjecture could be tested in a future study by carrying out experiments with relaxation of land surface parameters (e.g., Douville 2003). It is encouraging to see, however, that the newer model version of the ECMWF model (33R1) with substantially revised physics, does produce better extratropical forecasts when forced with observed SST/sea ice fields, which suggests that it is beginning to reproduce what the tropical relaxation experiment “predicts.”

From the results presented in this study we argue that the relaxation method is an important diagnostic technique which may help to understand possible “remote origins” of seasonal-mean anomalies. Unlike in prescribed SST experiments, where the atmospheric response has to be simulated by the atmospheric model, the relaxation technique captures any possible SST-forced atmospheric response explicitly. The same technique has been applied to other prominent climate anomalies such as the European heat wave in the summer of 2003, the results of which will be reported in forthcoming publications.

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