TECHNICAL MEMORANDUM

604

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

T. Jung, T.N. Palmer, M.J. Rodwell and S. Serrar

Research Department

ECMWF, Shinfield Park, Reading RG2 9AX, United Kingdom

Submitted to Monthly Weather Review

October 2009

European Centre for Medium-Range Weather Forecasts
Europäisches Zentrum für mittelfristige Wettervorhersage
Centre européen pour les prévisions météorologiques à moyen terme

Series: ECMWF Technical Memoranda

A full list of ECMWF Publications can be found on our web site under:
http://www.ecmwf.int/publications/

Contact: library@ecmwf.int

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 towards analysis data thereby artificially suppressing the development of forecast error. It is shown 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, diabatic heating anomalies over South America and the tropical Atlantic. From these results it is argued that the relaxation technique is a very powerful diagnostic tool to understand 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, 1996). This suggests that predictability of such anomalies is limited to a few weeks. There is observational and modelling evidence, however, that the atmosphere in the North Atlantic region is affected locally by sea surface temperature (SST) anomalies in the North Atlantic (e.g Czaja and Frankignoul, 1999; Rodwell and Folland, 2002; Rodwell et al., 1999; Latif et al., 2000) and 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 which might increase 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 practice this is very difficult to diagnose.

In this study, which can be seen as an extension of the paper by Jung et al. (2008), a diagnostic technique 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. Here, the relaxation technique, which has been widely used by the atmospheric science community on relatively shorter 'weather' time scales (Kalnay, 2003; Bauer et al., 2008), will be illustrated for the special case of the cold European winter of 2005/06.

The anomalously cold European winter of 2005/06 makes an interesting case study for various reasons. Firstly, 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. This increase manifested itself in, for example, the form of an anti-cyclonic anomaly in geopotential height fields at the 500 hPa level (hereafter Z500, Fig. 1a). Secondly, most seasonal forecasting system 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 rather internal atmospheric dynamics might have played a role. Thirdly, the anomalous atmospheric circulation giving rise to the cold European winter has been studied in some detail (e.g. Folland et al., 2006; Scaife and Knight, 2008; Croci-Maspoli and Davies, 2009). Finally, the winter of 2005/06 was marked by the presence of a number of climate anomalies, both in the Northern Hemisphere extratropics and in the

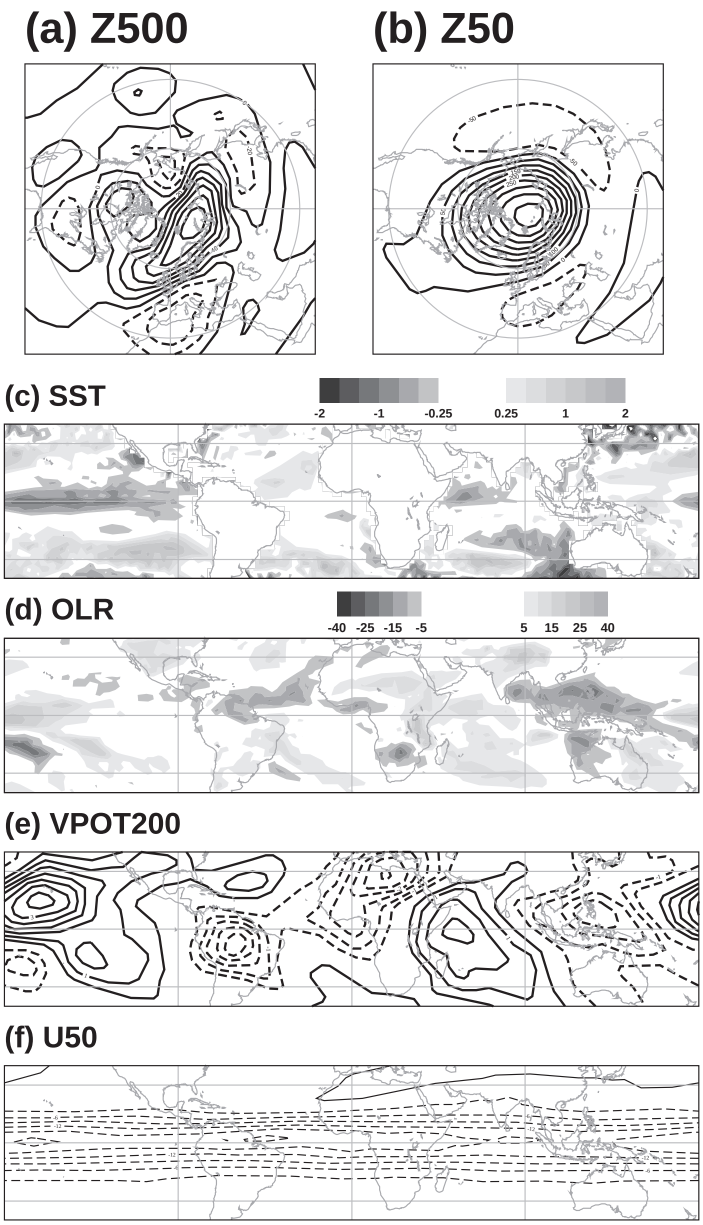

Figure 1: Observed mean anomalies for the period 1 December 2005 to 28 February 2006: (a) 500 hPa geopotential height (contour interval is 20 m), (b) 50 hPa geopotential height (contour interval is 50 m), (c) sea surface temperature (in K), (d) outgoing longwave radiation (in $\mathrm{W\,m^{-2}}$), (e) velocity potential at 200 hPa (contour interval is 0.5 $\mathrm{m^2\,s^{-1}}$) and (f) zonal wind at 50 hPa (contour interval is 3 $\mathrm{m\,s^{-1}}$). Negative (positive) values in (a), (b), (e) and (f) are dashed (solid). All results are based on data from ERA Interim (Simmons et al., 2007), except for (d) which is based on estimates of outgoing longwave radiation from NOAA satellites (Liebmann and Smith, 1996).

tropics, which might explain the observed circulation

Observed anomalies for the winter of 2005/06 and various different parameters are shown in Figure 1. Evidently, the anomalously tropospheric circulation anomaly in the Euro-Atlantic region is accompanied by an anomalously weak vortex. This, along with results from numerical experiments in which a stratospheric perturbation is applied that mimics the observed stratospheric warming, led Scaife and Knight (2008) to conclude that the sudden stratospheric warming has contributed to the cold European winter of 2005/06.

As mentioned above, 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 ciruclation as one key component. Although the Rodwell and Folland (2002) scheme employs SSTs 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. 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 cloud-diabatic processes upstream which in turn is sensitive to near-surface temperature.

Closer inspection of Figure 1 reveals strong seasonal anomalies in the tropics. The winter of 2005/06 was marked, for example, by a La Niña event of moderate strength which had a marked impact on the outgoing longwave radiation (OLR) and the velocity potential at 200 hPa (χ_{200}). The so-called 'canonical' link between La Niña and the atmospheric circulation in the North Atlantic region (Fraedrich, 1994; Greatbatch and Jung, 2007) predicts a positive phase of the North Atlantic Oscillation (NAO), that is, the opposite of what was observed. Relatively strong negative SST anomalies are also to be found in the Indian ocean. These anomalies can explain the strong local atmospheric anomalies (i.e., reduced cloudiness). The modelling 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 wave guide, suggesting that Indian ocean SST anomalies can have an influence on the atmospheric circulation in the Euro-Atlantic region. Finally, strong tropospheric anomalies were also evident over South America, the tropical Atlantic and over North Africa, all 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 (2008) find a significant link between the phase of the QBO and the tropospheric circulation, especially in the North Atlantic region.

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 interaction. The section on seasonal-mean circulation anomalies finishes with an investigation into the senstivity 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 dicussed. In this context, the origin of the sudden stratospheric warming in January 2006 will be discussed in some detail. The paper closes with a discussion of results.

2 Methodology

The numerical experimentation carried out in this study is based on a recent version of the ECMWF atmosphere model (cycle 32R1 used operationally from 5 June to 5 November 2007). All forecast experiments use a horizontal resolution of T_L95 (linear Gaussian grid $\approx 1.85^o \times 1.85^o$) and employ 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 observed lower boundary conditions (SST and sea ice). Aspects of the model's performance are discussed elsewhere (Jung, 2005; Jung et al., 2009).

In order 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 hereafter). The control integration is used to understand the rolf of SST and sea ice anomalies. In the relaxation experiments the model is drawn towards ERA-Interim reanalysis data (Simmons et al., 2007) during the course of the integration; this is achieved by adding an extra term of the following form to the ECMWF model:

$$-\lambda(\mathbf{x}-\mathbf{x}_{ref}). \tag{1}$$

The model state vector is represented by $\mathbf{x}$ and the reference field towards which the model is drawn (ERA-Interim reanalysis data) by $\mathbf{x}_{ref}$. The strength of the relaxation is determined by $\lambda = a \cdot \lambda_0$, where m is a function of longitude, latitude, height and the parameter being considered and λ_0 is a constant. The units of λ are in (time step)$^{-1}$. Unless stated otherwise $\lambda_0 = 0.1\,hrs^{-1}$ is used throughout the study. For a time step of one hour used here a value of $0.1\,hrs^{-1}$ indicates that at each time step the model is 'corrected' using 10% of the departure of $\mathbf{x}$ from $\mathbf{x}_{ref}$. 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. The reference fields ($\mathbf{x}_{ref}$) were obtained from the ERA-Interim reanalysis and have been interpolated from their native resolution of T_L255 to T_L95 using a sophisticated horizontal interpolation package used routinely within the ECMWF Integrated Forecasting System.

In order to allow for an effective localization, the relaxation was carried out in grid point 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 non-relaxed regions in the horizontal is smoothed using the hyperbolic tangent. The smoothing is such that the relaxation coefficient λ goes from λ_0 to zero within a 20^o belt, both in longitude and latitude. Boundaries stated in the text refer to the centre of the respective 20^o belt. In order to reduce the generation of spurious potential vorticity features, changes of λ are also smoothed in the vertical. Here, the relaxation coefficient effectively goes from λ_0 to zero in a vertical layer encompassing about 13 model levels (see Tab. 1 for actual values of λ at various heights).

For each control and relaxation experiment a separate calibration run covering winters of the period 1990/91 to

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

Experiment	Relaxation Region	
CNT	no relaxation	—
TROP	20^oS–20^oN, 0^o–360^oE	troposphere+stratosphere
TROP-T	20^oS–20^oN, 0^o–360^oE	troposphere*
TROP-S	20^oS–20^oN, 0^o–360^oE	stratosphere†
TROP-T/30–90E	20^oS–20^oN, 0^o–90^oE	troposphere*
TROP-T/150E–120W	20^oS–20^oN, 150^oE–120^oW	troposphere*
TROP-T/90W–0	20^oS–20^oN, 90^oW–0^o	troposphere*
NH	30^oN–90^oN, 0^o–360^oE	troposphere+stratosphere
NH-S	20^oN–90^oN, 0^o–360^oE	stratosphere†

* Actual strength of the relaxation at 500, 200, 50 and 20 hPa is approximately $\lambda_0 \cdot 0.999, \lambda_0 \cdot 1.8 \cdot 10^{-2}, \lambda_0 \cdot 8.3 \cdot 10^{-7}$ and $\lambda_0 \cdot 1.5 \cdot 10^{-8}$ hrs^{-1}, respectively.

† Actual strength of the relaxation at 500, 200, 50 and 20 hPa is approximately $\lambda_0 \cdot 1.1 \cdot 10^{-7}, \lambda_0 \cdot 2.3 \cdot 10^{-6}, \lambda_0 \cdot 1.8 \cdot 10^{-2}$ and $\lambda_0 \cdot 0.5$ hrs^{-1}, respectively.

2005/06 was carried out in order to obtain the model's climatology. These integrations were started at 12UTC on 15 November. For the winter of 2005/06 a set of seasonal ensemble forecasts with and without relaxation was carried out using a lagged approach. The ensembles were generated by starting forecasts in 6-hourly intervals from 12 UTC on 16 November to 12 UTC on 20 November 2005 giving a total of 17 ensemble members. Throughout this paper 'anomalies' refer to departures of the ensemble mean or ensemble members from the climate of the model obtained from the calibration run. A summary of all seasonal forecast experiments along with their abbreviations is given in Table 1.

3 Results

3.1 Seasonal-mean diagnostics

3.1.1 Tropical versus stratospheric influences

Observed Z500 anomalies for the 2005/06 winter are shown in Figure 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 esemble 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). 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 only weak resemblance with 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). Rather, NH-S locally leads to a significant anti-cyclonic circulation anomaly in the eastern North Atlantic.

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 3 shows observed 50 hPa geopotential height (Z50) anomalies; also shown are ensemble mean anomalies for CNT and TROP. The Z50 anomalies produced by NH-S are very similar to the observations (not shown). CNT shows weak and non-significant Z50 anomalies suggesting that the observed SST and sea ice anomalies have contributed little to the anomalously weak stratospheric polar vortex. The ensemble mean for TROP, on the other hand, produces a weakened stratospheric polar vortex, with an anomaly which is stronger than observed. As mentioned above, inspection of the individual ensemble members (not shown) suggests that the stratospheric response to a tropical forcing is consistent with the observations. These results highlight that the anomalously weak stratospheric vortex during the 2005/06 winter might have actually been forced from the tropics

3.1.2 Further exploring the tropical influence

Velocity potential anomalies at the 200 hPa level (hereafter χ_{200} anomaly) are shown in Figure 4 for ERA-Interim, CNT and TROP. The control integration with observed SST and sea ice distribution captures the anomalous convergent flow (positive χ_{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

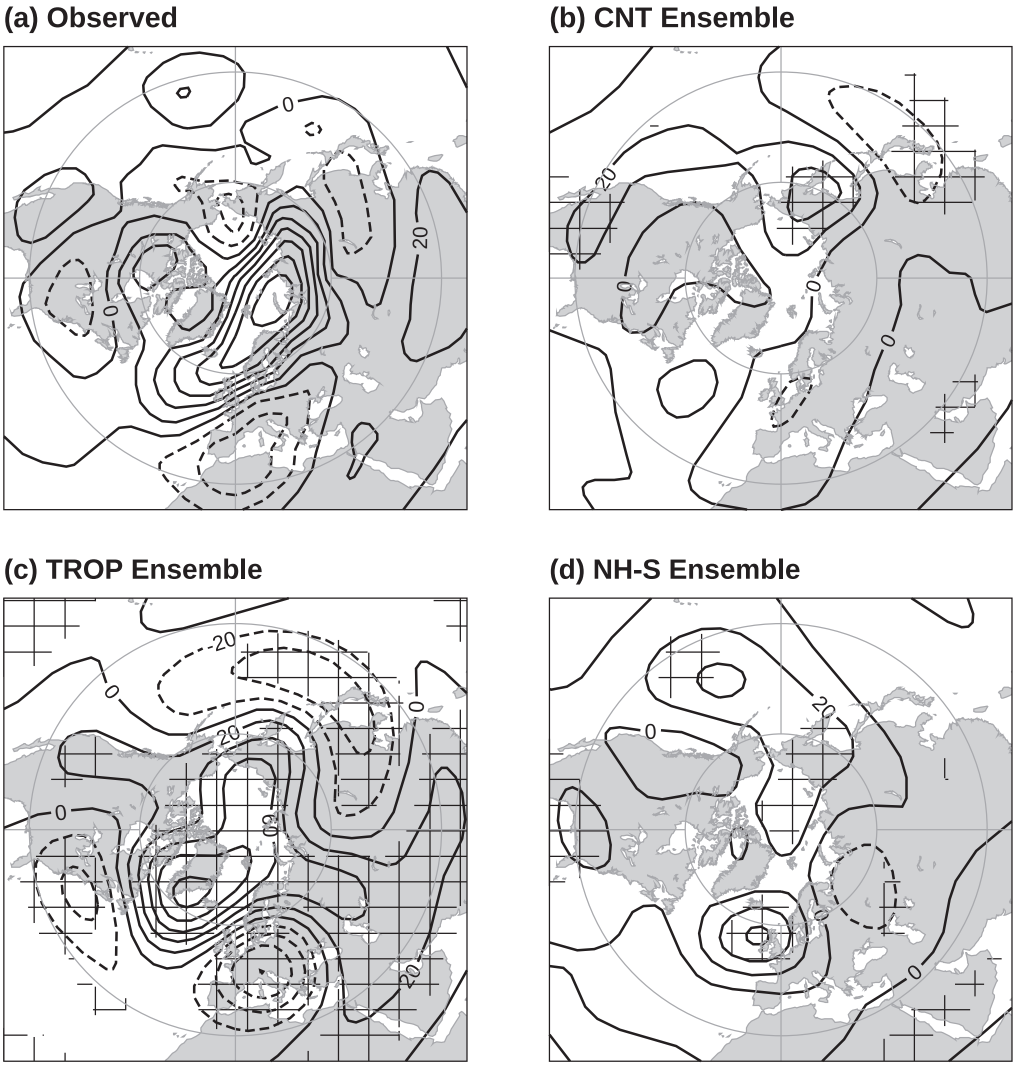

Figure 2: Geopotential height anomalies at the 500 hPa level (contour interval is 20 m) for the period 1 December 2005 to 28 February 2006: (a) ERA Interim, (b) CNT ensemble, (c) TROP ensemble and (d) NH-S ensemble. Results in (b)–(d) are based on ensemble mean data. Statistically significant differences (at the 95% confidence level) in (b)–(d) are hatched.

fails to reproduce the observed χ_{200} anomalies. Given that CNT fails to reproduce the observed Z500 anomalies over the Northern Hemisphere (Fig. 2b), it can be concluded that La Niña was not responsible for the extratropical response suggesting that the cause of the anomalous European 2005/06 winter lies outside the central tropical Pacific region. The fact that TROP, which shows a stronger and more realistic extratropical response, captures the observed χ_{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. 5a). Interestingly, CNT is able to simulate the easterly phase of the QBO; CNT fails, however, to produce the observed weakening of the stratospheric polar vortex (Fig. 5b). 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

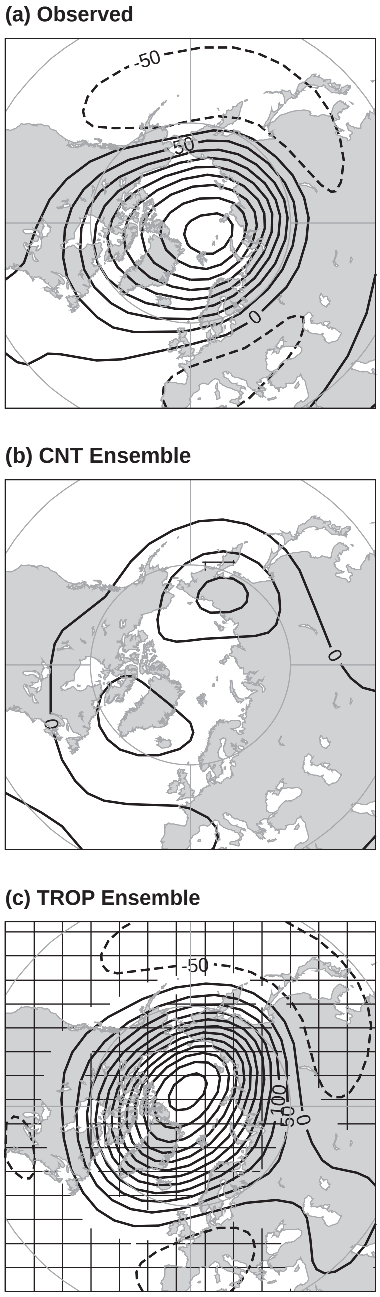

Figure 3: Geopotential height anomalies at the 50 hPa level (contour interval is 20 m) for the period 1 December 2005 to 28 February 2006: (a) ERA Interim, (b) CNT ensemble and (c) TROP ensemble. Results in (b) and (c) are based on ensemble mean data. Statistically significant differences (at the 95% confidence level) in (b) and (c) are hatched.

(Branković et al., 1994). 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 (i) that the QBO in CNT weakens throughout the 3-month period leaving it rather week by the end of the winter when it might have mattered most and (ii) that the error associated with the missing downward propagation matters, especially by the end of the winter 2005/06 (not shown). An alternative explanation is that the Holton-Tan mechanism does work in CNT but is obscured by other signals of tropospheric origin (see below).

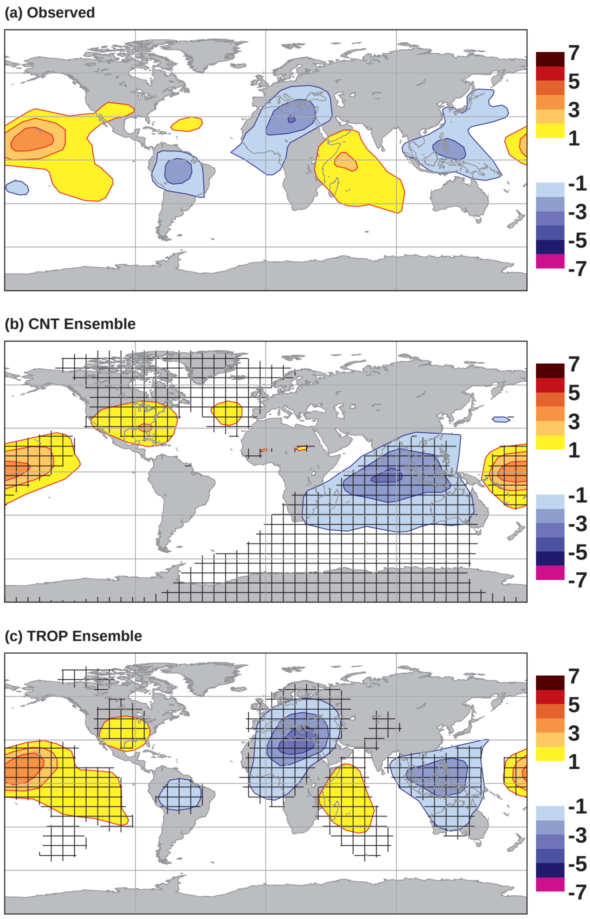

Figure 4: Velocity potential anomalies ($m^2\,s^{-1}$) at the 200hPa level for the period December–February 2005/06: (a) ERA Interim, (b) CNT and (c) TROP. Differences significant at the 95% confidence level are hatched (b and c only).

After having established the crucial role of the tropics for explaining the anomalous atmospheric circulation over large parts of the Northern Hemisphere during the 2005/06 winter, the question arises which region of the tropical atmosphere contributed to the extratropical forcing. 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 χ_{200} anomalies, not shown). Figure 6 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 north-west North Pacific, on the other hand, tropospheric and stratospheric influences seem 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 indentify the relevant tropospheric physical processes. 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 (compare Fig. 1): (i) the Indian ocean (30^o–90^oE) and its associated anomaly (TROP-T/30–90E, hereafter), (ii) the tropical Pacific (150^oE–120^oW) capturing the circulation anomaly associated with the La Niña (TROP-T/150E–

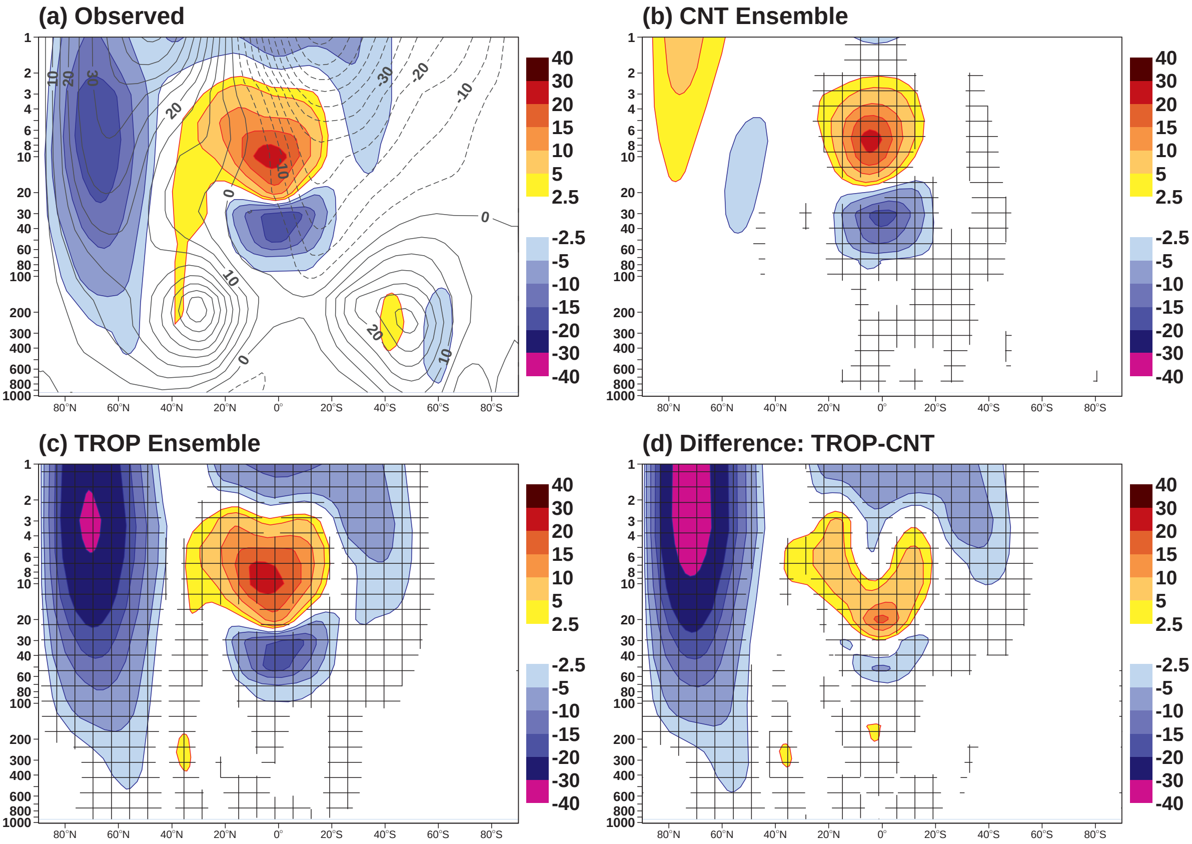

Figure 5: Average zonal mean zonal wind anomalies (shading in m/s) for the period 1 December 2005 to 28 February 2006: (a) ERA Interim, (b) CNT, (c) TROP. Also shown is (d) the difference between TROP and CNT. In (a) climatological average zonal mean zonal wind anomalies from ERA-Interim (contour interval is 5 m/s, negative values are dashed) are superimposed. Statistically significant differences (at the 95% confidence level) in (b)–(d) are hatched.

120W) and (iii) South America, the tropical Atlantic and western parts of tropical Africa (TROP-T/90W–0). Figure 7 shows Northern Hemisphere Z500 anomalies for the three relaxation experiments. Relaxing the tropical atmosphere over the Indian ocean clearly fails to explain the extratropical Z500 anomalies (compare Figs. 6a and 7a). Relaxing the troposphere over the tropical Pacific captures some of the anomalies produced by TROP-T, especially in the North Pacific region. It is necessary, however, to relax the tropical troposphere between 90^oW and the Greenwich Meridian in order to reproduce a Z500 response in the Euro-Atlantic region which similar to that foounf for TROP-T (Figs. 6a and 7c).

It is worth pointing out that the experiments with spatial localization in the tropical troposphere have to be interpreted carefully. This is because relaxation in a certain region of the tropical troposphere is likely to have an indirect effect on other tropical regions as well. This is particularly true for seasonal integration in which the atmosphere has time to adjust to the forcing applied by carrying out the relaxation.

3.1.3 Extratropical forcing of tropical anomalies

So far, the focus has been on tropical-to-extratropical interactions. In order to correctly assess cause and effect it is crucial to study possible extratropical-to-tropical interactions as well. In fact, it is well-known that the the tropics respond to an extratropical forcing (e.g. Kiladis and Weickmann, 1992; Hoskins and Yang, 2000;

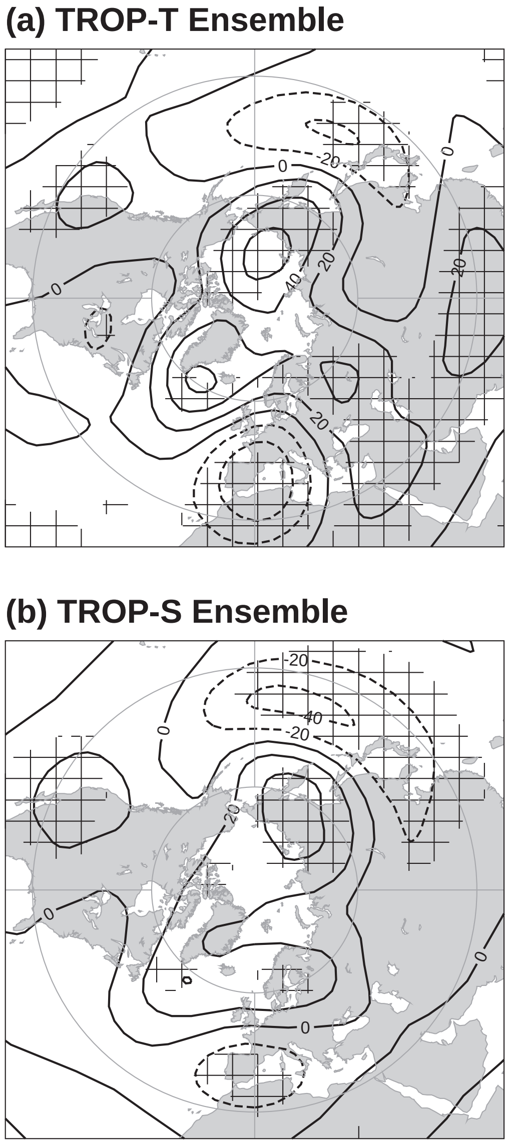

Figure 6: Geopotential height anomalies at the 500 hPa level (contour interval is 20 m) for the period 1 December 2005 to 28 February 2006: (a) TROP-T and (b) TROP-S. Results are based on ensemble mean data. Statistically significant differences (at the 95% confidence level) in (b)–(d) are hatched.

Jung and Palmer, 2009). The experiment in which the whole Northern Hemisphere north of 30^{o}N, NH, is relaxed towards reanalysis data is designed to study a possible extratropical forcing of the observed tropical anonmalies. NH produces tropical anomalies both in the troposphere and stratosphere that are very similar to those found in CNT (not shown). This suggests that the extratropical forcing of tropical anomalies during the 2005/06 winter, if existent, was relatively weak compared with tropical-to-extratropical interactions.

3.1.4 Sensitivity Experiments

As mentioned in the Methods section, the tropical relaxation experiment, TROP, has its northern relaxation boundary at 20^o (with a transition zone covering 10^o on either side). Synoptic studies of the sudden stratospheric

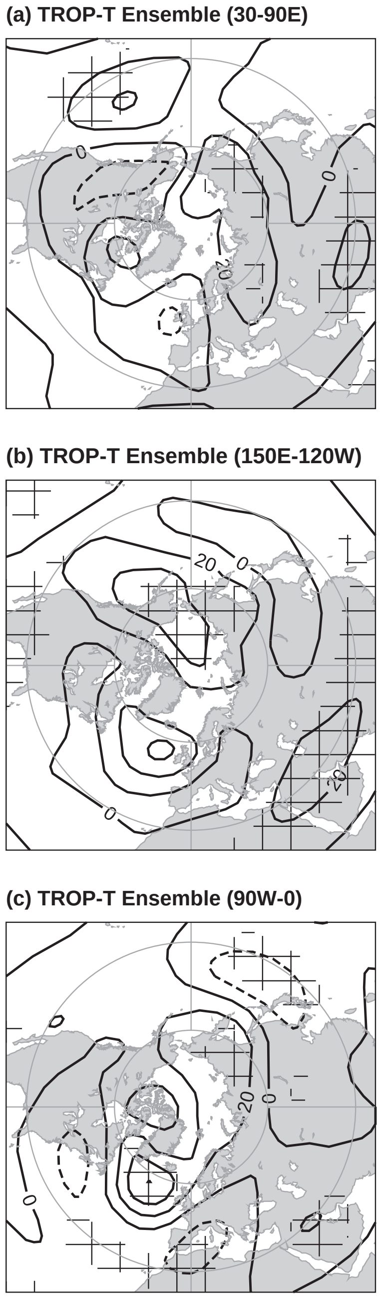

Figure 7: Geopotential height anomalies at the 500 hPa level (contour interval is 20 m) for the period 1 December 2005 to 28 February 2006: (a) TROP-T/30–90E, (b) TROP-T/150E–120W and (c) TROP-T/90W-0. Results are based on ensemble mean data. Statistically significant differences (at the 95% confidence level) in (b)–(d) are hatched.

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). In order to ascertain that the origin of the anomalous circulation in the Euro-Atlantic region is truly of *tropical* origin another tropical relaxation experiment has been carried out in which the relaxation boundaries have been moved equatorward to 10^oS and 10^oN, respectively. The same latitudinal smoothing is applied as in the other experiments. Restricting the tropical relaxation to the tropical belt 10^oS–10^oN yields a very similar Z500 response to TROP over the Northern Hemisphere (compare Figs. 2c and 8a) confirming the importance of the tropics.

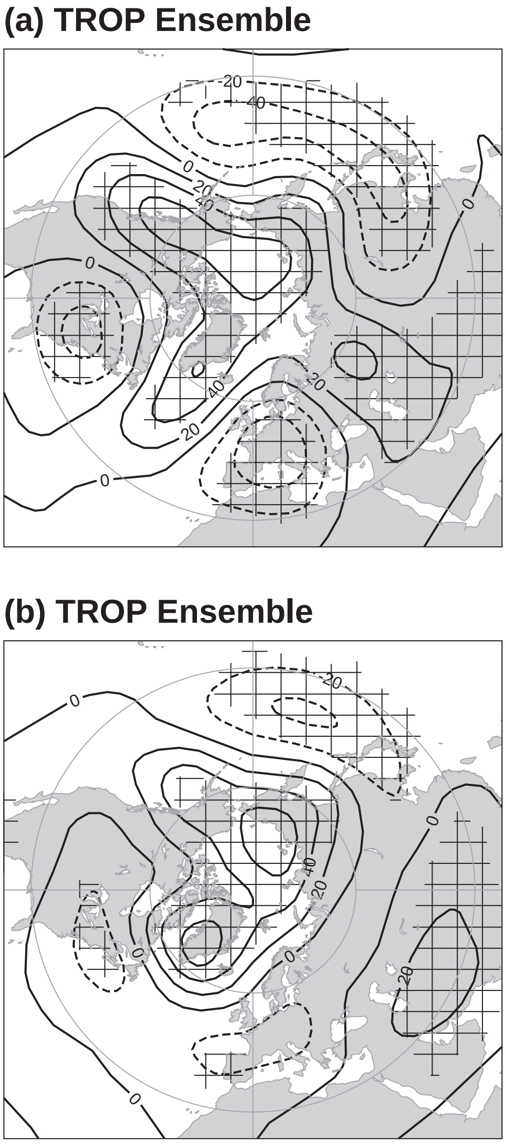

Figure 8: As in Fig. 2, but for (a) tropical relaxation in the belt 10°S–10°N ($\lambda = 0.1$ hrs^{-1}) and (b) tropical relaxation with $\lambda = 0.01$ hrs^{-1} (20°S–20°N).

The choice of the relaxation coefficient ($\lambda_0 = 0.1\,hrs^{-1}$ in this study) is somewhat arbitrary. Therefore, it is important the test whether the conclusions of this study depend on the exact choice of λ. Figure 8b shows Z500 anomalies for a tropical relaxation experiment (20^{o}S–20^{o}N) with $\lambda = 0.01\,hrs^{-1}$. Evidently, neither the spatial structure nor the magnitude of the Z500 response is strongly affected by the exact choice of λ_0 (compare Figs. 2c and 8b).

One might argue that some of the results presented in this study may depend on the model used to carry out the experiments. In order to address this issue the two experiments CNT and TROP have been repeated using the more recent model version 33R1, which has been used operationally at ECMWF from 3 June to 29 September 2008. Compared with model version 32R1, on which most of the experimentation presented in this study is based, model version 33R1 comprises substantial changes to almost every part of the ECMWF physics package (Bechtold et al., 2008; Jung et al., 2009). Figure 9 shows Northern Hemisphere Z500 anomalies for CNT and

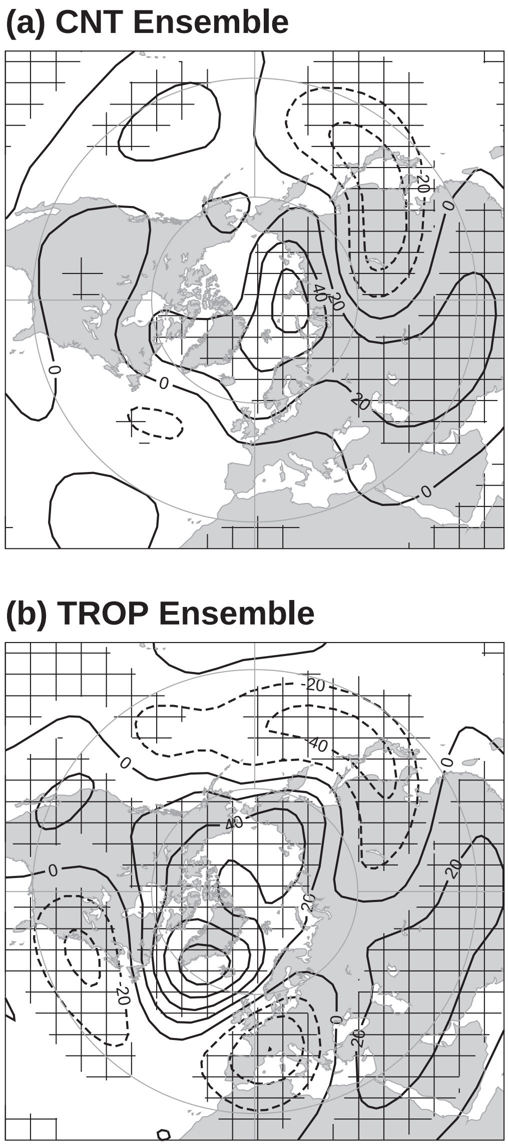

Figure 9: As in Fig. 2, but for the more recent ECMWF model version 33R1.

TROP based on ECMWF model version 33R1. The results are very similar to the ones obtained with the older model version (Fig. 2b,c) showing that the conclusions of this study are not overly sensitive to the model formulation employed.

3.2 Intraseasonal evolution

So far, the focus has been on seasonal-mean fields for the whole winter. The 2005/06 winter was marked, however, by large intraseasonal changes particularly in the Northern Hemisphere stratosphere. Nishii et al. (2009), for example, pointed 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 focussing solely on seasonal-mean anomalies should help to illuminate what happened during the 2005/06 winter.

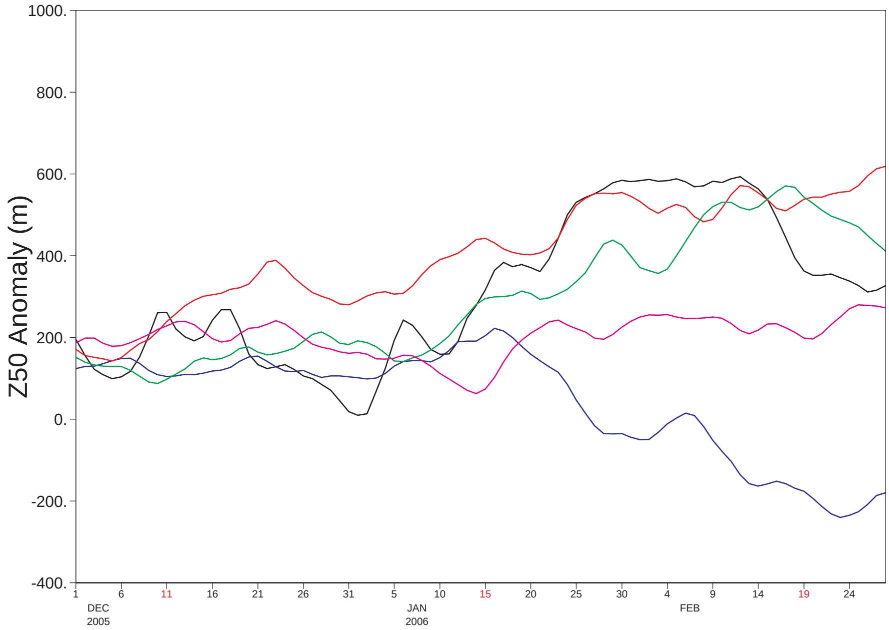

Figure 10: Time series of area averaged (70–90^{o}N, 0–360^{o}W) 50 hPa geopotential height anomalies for the period 1 December 2005 to 28 February 2006: ERA Interim (black), CNT (blue), TROP (red), TROP-T (green) and TROP-S (purple). Results for the forecast experiments are based on ensemble mean data.

The observed temporal evolution of the strength of the stratospheric polar vortex during the 2005/06 winter can be inferred from Figure 10 (black). Here the strength of the stratospheric polar vortex is defined by area-averaged Z50 anomalies north of 70^{o}N. 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 increase of the strength of the polar vortex throughout the winter. Inspection of the individual ensemble members (not shown) for TROP (also TROP-T and TROP-S) reveals that the observations lie within the ensebmle throughout the whole winter; in contrast, the CNT ensemble clearly fails to capture the observation towards the end of the winter (not shown).

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 order to disentangle the response to tropical tropospheric forcing from tropical stratospheric forcings the temporal evolution of the stength 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 (Figure 10). For TROP-S the stratospheric polar vortex remains moderately weak throughout the second half of the winter; for TROP-T, on the other hand, the stratospheric polar vortex further weakens during the second half of the winter due to an increased frequency of occurrence of stratospheric warmings (not shown). In summary, observed anomalies in the tropical *stratosphere* led to a more or less constant weakening of moderate strength through the winter. Observed anomalies in the tropical *troposphere*, however, have to be considered in order to explain the stratospheric warming that occurred in January 2006.

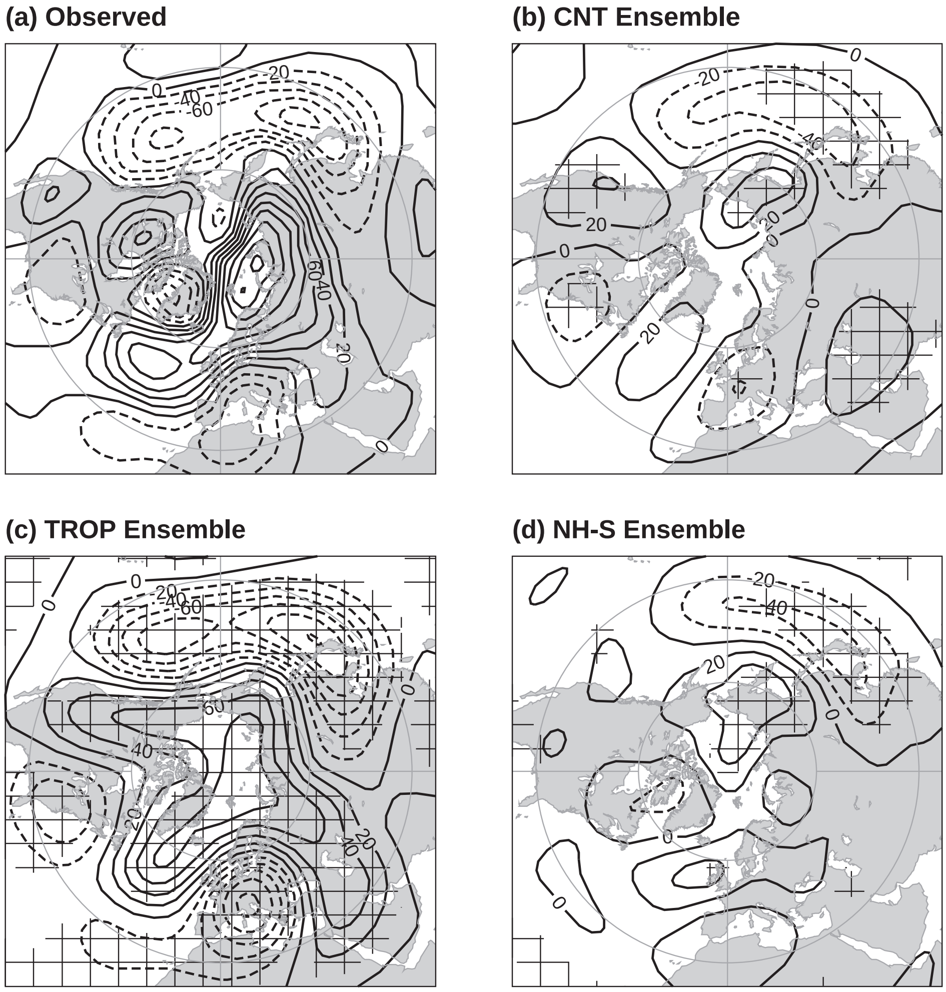

Figure 11: Same as in Fig. 2, but for the sub-period 1 December 2005 to 15 January 2006.

From the temporal evolution of the strength of the stratospheric polar vortex shown in Figure 10 it seems reasonable, for diagnostic purposes, to divide the 2005/06 winter into two parts, one representing early winter (1 December 2005 to 15 January 2006) and the other late winter (16 January to 28 February 2006). A comparison of the observed Z500 anomalies over the Northern Hemisphere between early and late winter shows marked differences between the North Pacific and the rest of the Northern Hemisphere (Figs. 11a and 12a). For the former the observed Z500 anomalies changed sign from early to late winter; and for the latter the tropospheric circulation was rather persistent. Interestingly, the reversal of the anomalies in the North Pacific are captured by all experiments including CNT. In order to get a realistic representation of the strong persistence of the Z500 anomalies in the Euro-Atlantic region, on the other hand, it is necessary to relax the tropical atmosphere (Figs. 11 and 12). Perhaps not too surprisingly, the tropospheric response in the Euro-Atlantic region in NH-S is most pronounced during late winter when the stratospheric circulation was most anomalous.

As mentioned in the Introduction, La Niña conditions prevailed in the eastern tropical Pacific during the 2005/06 winter. The typical extratropical atmospheric reponse to a cold SST anomlies in the central and eastern tropical Pacific, resembling the negative phase of the Pacific North America (PNA) pattern, can be found only during late winter (Fig. 12). The atmospheric La Niña response is especially clear for CNT and NH-S. For TROP the atmospheric La Niña response is somewhat obscured by the presence of other circulation anomalies over the

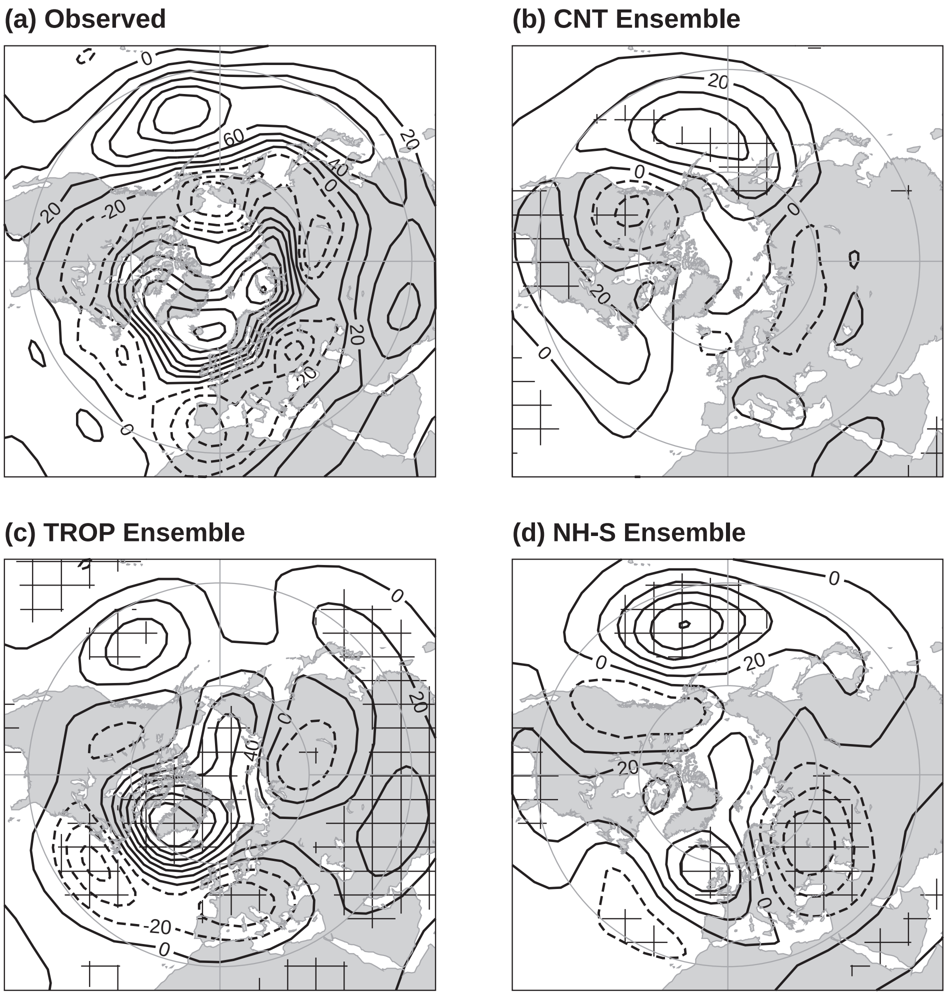

Figure 12: Same as in Fig. 2, but for the sub-period 16 January to 28 February 2006.

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). One possible explanation is that the negative phase of the PNA leads to a reduction 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).

To test the idea of a negative correlation between ENSO-related SST anomalies in the tropical Pacific and the strength of the stratospheric polar vortex more specifically for the ECMWF model, seasonal forecast experiments with the ECMWF model and La Niña-type diabatic forcing the experiments described in Greatbatch and Jung (2007) were diagnosed in more detail. Indeed it is found that a La-Niña-type diabatic forcing applied to the ECMWF model leads to a strengthening of the stratospheric polar vortex (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 clearly highlight the important role of the tropical atmosphere. Further experimentation suggests that the largest forcing came from the tropical troposphere in the region 90^{o}W–30^{o}E. Interestingly, this area does *not* cover the apparently most prominent tropical anomalies in the Indian ocean and the central Pacific. The easterly phase of the QBO also contributed to the observed circulation anomalies, especially in the Northern Hemisphere stratosphere. Scaife and Knight (2008) suggest 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 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 lies in the tropics; hence the stratosphere was not a primary cause for the cold winter.

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.

As mentioned in the Introduction, most of the observed anomalies of the tropical troposphere can be explained by the underlying SST anomalies (Fig. 1): Positive (negative) SST anomalies are accompanied by reduced (increased) OLR and anomalously strong (weak) divergent outflow at the upper troposphere. If the tropical troposphere was crucial in explaining the anomalously cold European winter, and the tropical tropospheric circulation anomalies were the response to SST anomalies, then the question arises as to why the control integration, with specified observed SST anomalies, fails to produce the observed extratropical response. The control integration does not provide an accurate forecast because it fails to show a realistic tropospheric response in that part of the tropics that mattered (South America, tropical Atlantic and West Africa, Fig. 4). One possible explanation is that the ECMWF model fails to respond realistically to the imposed SST anomalies. An alternative explanation is that *land* rather than sea surface anomalies mattered—and the land surface conditions were not prescribed in the control integration. In fact, the largest χ_{200} anomalies in the tropical area 90^{o}W–30^{o}E are found over land (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). Diagnosing the realism of the response of the ECMWF model to SST anomalies is another important challenge that is left for future study.

For the results presented in this study we argue that the relaxation method is an important diagnostic technique which helps 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 and, hence, is prone to model uncertainty, the relaxation technique captures any SST-forced atmospheric response explicitly. The same technique has been applied to other prominent climate anomalies such as the European heat wave in summer 2003, the results of which will be reported in a forthcoming paper.

Acknowledgements Dr. Hervé Douville provided useful comments on an earlier version of the manuscript.

References

Ambaum, M. H. P. and B. J. Hoskins, 2002: The NAO troposphere-stratosphere connection. *J. Climate*, **15**, 1969–1978.

Bader, J. and M. Latif, 2003: The impact of decadal-scale Iindian Ocean sea surface temperature anomalies on Sahelian rainfall and the North Atlantic Oscillation. *Geophys. Res. Lett.*, **30**, 2169, doi:10.1029/2003GL018426.

Baldwin, M. P. and T. J. Dunkerton, 1999: Propagation of the Arctic Oscillation from the stratosphere to the troposphere. *J. Geophys. Res.*, **104**, 30937–30946.

Baldwin, M. P., D. B. Stephenson, D. W. J. Thompson, T. J. Dunkerton, A. J. Charlton, and A. O'Neill, 2003: Stratospheric memory and skill of extended-range weather forecasts. *Science*, **301**, 636–640.

Bauer, H.-S., V. Wulfmeyer, and L. Bengtsson, 2008: The representation of synoptic-scale weather system in a thermodynamically adjusted version of the ECHAM4 general ciruclation model. *Meteorol. Atmos. Phys.*, **99**, 129–153.

Bechtold, P., M. Köhler, T. Jung, F. Doblas-Reyes, M. Leutbecher, M. Rodwell, F. Vitart, and G. Balsamo, 2008: Advances in simulating atmospheric variability with the ECMWF model: From synoptic to decadal time-scales. *Quart. J. Roy. Meteor. Soc.*, **134**, 1337–1351.

Boer, G., 2008: Qbo influence on extratropical predictive skill. *Clim. Dyn.*, **31**, 987–1000.

Branković, C., T. N. Palmer, and L. Ferranti, 1994: Predictability of seasonal atmospheric variations. *J. Climate*, **7**, 217–237.

Brönnimann, S., 2007: Impact of El Niño-Southern Oscillation on European climate. *Rev. Geophys.*, **45**, 1–28.

Coy, L., S. Eckermann, and C. Hoppel, 2009: Planetary wave breaking and tropospheric forcing seen in the stratospheric warming of 2006. *J. Atmos. Sci.*, **66**, 495–507.

Croci-Maspoli, M. and H. C. Davies, 2009: Key dynamical features of the 2005/06 European winter. *Mon. Wea. Rev.*, **137**, 664–.

Czaja, A. and Frankignoul, 1999: Influence of the North Atlantic SST on the atmospheric circulation. *Geophys. Res. Lett.*, **26**, 2969–2972.

Douville, H., 2003: Assessing the influence of soil moisture on seasonal climate variability with AGCMs. *J. Hydromet.*, **4**, 1044–1066.

Folland, C. K., D. E. Parker, A. A. Scaife, J. J. Kennedy, A. W. Colman, A. Brookshaw, S. Cusack, and M. R. Huddleston, 2006: The 2005/06 winter in Europe and the United Kingdom: Part II—prediction techniques and their assessment against observations. *Weather*, **61**, 337–346.

Fraedrich, K., 1994: ENSO impact on Europe? *Tellus*, **46A**, 541–552.

Graham, R. J., C. Gordon, M. R. Huddleston, M. Davey, W. Norton, A. Colman, A. A. Scaife, A. Brookshaw, S. Cusack, E. McCallum, W. Elliott, K. Groves, D. Cotgrove, and D. Robinson, 2006: The 2005/06 winter in Europe and the United Kingdom: Part I—How the Met Office forecast was produced and communicated. *Weather*, **61**, 327–336.

Greatbatch, R. J. and T. Jung, 2007: Local versus tropical diabatic heating and the winter North Atlantic Oscillation. *J. Climate*, **20**(10), 2058–2075.

Holton, J. R. and C.-H. Tan, 1980: The influence of the equatorial Quasi-Biennial Oscillation on the global circulation at 50mb. *J. Atmos. Sci.*, **37**, 2200–2208.

Hoskins, B. J. and T. Ambrizzi, 1993: Rossby wave propagation on a realistic longitudinally varying flow. *J. Atmos. Sci.*, **50**, 1661–1671.

Hoskins, B. J. and G. Y. Yang, 2000: The equatorial response to higher-latitude forcing. *J. Atmos. Sci.*, **57**, 1197–1213.

Hurrell, J. W., 1995: Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science*, **269**, 676–679.

Jung, T., 2005: Systematic errors of the atmospheric circulation in the ECMWF forecasting system. *Quart. J. Roy. Meteor. Soc.*, **131**, 1045–1073.

Jung, T., G. Balsamo, P. Bechtold, A. Beljaars, M. Köhler, M. Miller, J.-J. Morcrette, A. Orr, M. J. Rodwell, and A. M. Tompkins, 2009: The ECMWF model climate: Recent progress through improved physical parametrizations. *Quart. J. Roy. Meteor. Soc.*, p. submitted.

Jung, T. and J. Barkmeijer, 2006: Sensitivity of the tropospheric circulation to changes in the strength of the stratospheric polar vortex. *Mon. Wea. Rev.*, **134**, 2191–2207.

Jung, T. and Palmer, 2009: Diagnosising extended-range forecast error. *Mon. Wea. Rev.*, p. submitted.

Jung, T., T. N. Palmer, M. J. Rodwell, and S. Serrar, 2008: Diagnosing forecast error using relaxation experiments. ECMWF Newsletter 82, ECMWF, Shinfield Park, Reading, Berkshire RG2 9AX, UK.

Kalnay, E., 2003: *Atmospheric Modelling, Data Assimilation and Predictability*. Cambridge University Press.

Kiladis, G. N. and K. M. Weickmann, 1992: Extratropical forcing of tropical Pacific convection during northern winter. *Mon. Wea. Rev.*, **120**, 1924–1939.

Kushnir, Y., W. A. Robinson, I. Bladé, N. M. J. Hall, S. Peng, and R. Sutton, 2002: Atmospheric GCM response to extratropical SST anomalies: Synthesis and evaluation. *J. Climate*, **15**, 2233–2256.

Latif, M., K. Arpe, and E. Roeckner, 2000: Oceanic control of decadal North Atlantic sea level pressure variability in winter. *Geophys. Res. Lett.*, **27**, 727–730.

Liebmann, B. and C. A. Smith, 1996: Description of a complete (interpolated) outgoing longwave radiation dataset. *Bull. Amer. Meteor. Soc.*, **77**, 1275–1277.

Nishii, K., N. H., and T. Miyasaka, 2009: Modulations in the planetary wave field by upward-propagating Rossby wave packets prior to a stratospheric warming event in January 2006. *Quart. J. Roy. Meteor. Soc.*, **135**, 39–52.

Rodwell, M., D. P. Rowell, and C. K. Folland, 1999: Oceanic forcing of the wintertime North Atlantic Oscillation and European climate. *Nature*, **398**, 320–323.

Rodwell, M. J. and C. K. Folland, 2002: Atlantic air-sea interaction and seasonal predictability. *Quart. J. Roy. Meteor. Soc.*, **128**, 1413–1443.

Rowell, D. P., 1996: Assessing potential seasonal predictability with an ensemble of multidecadal GCM simulations. *J. Climate*, **11**, 109–120.

Scaife, A. A. and J. R. Knight, 2008: Ensemble simulations of the cold European winter of 2005–2006. *Quart. J. Roy. Meteor. Soc.*, **134**, 1647–1659.

Simmons, A. J., S. Uppala, D. Dee, and S. Kobayashi, 2007: ERA-Interim: New ECMWF reanalysis products from 1989 onwards. ECMWF Newsletter 110, ECMWF, Shinfield Park, Reading, Berkshire RG2 9AX, UK.

Taguchi, M. and D. L. Hartmann, 2006: Increased occurence of stratospheric sudden warmings during El Niño as simulated by WACCM. *J. Climate*, **19**, 324–332.

Thompson, D. W. J. and J. M. Wallace, 1998: The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophys. Res. Lett.*, **25**, 1297–1300.

Untch, A. and A. J. Simmons, 1999: Increased stratospheric resolution. ECMWF Newsletter 82, ECMWF, Shinfield Park, Reading, Berkshire RG2 9AX, UK.

van Loon, H. and J. C. Rogers, 1978: The seesaw in winter temperatures between Greenland and Northern Europe. Part I: General description. *Mon. Wea. Rev.*, **106**, 296–310.

Walker, G. T., 1924: Correlation in seasonal variation of weather, IX. *Mem. Indian. Meteor. Dep.*, **24**(9), 275–332.