Meteorologische Zeitschrift, Vol. **20**, No. 3, 259-266 (June 2011)
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Article

Influence of sudden stratospheric warmings on tropospheric winds

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(Manuscript received September 23, 2010; in revised form December 21, 2010; accepted December 22, 2010)

Abstract

The influence of changes in the zonal mean stratospheric potential vorticity, associated with sudden stratospheric warmings, on the zonal mean zonal wind in the troposphere is investigated by piecewise potential vorticity inversion. The focus is on the major sudden stratospheric warming that occurred in January 2009. It is shown that changes in the potential vorticity distribution associated with sudden stratospheric warmings may result in substantially less westerly or more easterly winds in the mid-troposphere in midlatitudes. The potential vorticity changes accompanying the sudden stratospheric warming persist for several months in the lower stratosphere. A comparison of years with and years without a sudden stratospheric warming leads to the conclusion that the change in the zonal mean tropospheric windspeed due to the sudden stratospheric warming can be as large as -5 m s^{-1} and that this change can easily persist until the end of the winter. Information on the stratospheric potential vorticity distribution may therefore be valuable for winter weather forecasts on a monthly timescale.

Zusammenfassung

Mit plötzlichen stratosphärischen Erwärmungen zusammenhängende Einflüsse von Änderungen der zonal gemittelten stratosphärischen potentiellen Vorticity werden durch stückweise Inversion der potentiellen Vorticity untersucht. Das Schwergewicht liegt auf dem starken Erwärmungsereignis vom Januar 2009. Es wird gezeigt, dass Änderungen in der Verteilung der potentiellen Vorticity, welche mit plötzlichen stratosphärischen Erwärmungen verbunden sind, zu deutlich verringerten westlichen oder mehr östlichen Winden in der mittleren Troposphäre der mittleren Breiten führen können. Die Änderungen der potentiellen Vorticity, die die plötzlichen stratosphärischen Erwärmungen begleiten, halten sich für mehrere Monate in der unteren Stratosphäre. Ein Vergleich von Jahren mit und ohne plötzlicher stratosphärischer Erwärmung zeigt, dass die Änderung der zonal gemittelten troposphärischen Windgeschwindigkeit bis zu -5 m s^{-1} betragen kann, und dass diese Änderung problemlos bis zum Ende des Winter erhalten bleiben kann. Informationen über die Verteilung der stratosphärischen potentiellen Vorticity können daher wertvoll für Winterwettervorhersagen auf der monatlichen Zeitskala sein.

1 Introduction

The winter of 2008/2009 in the northern hemisphere was relatively cold in western Europe and parts of North America. This was due to a persistent negative phase of the tropospheric annular mode (a measure of the pressure difference between the pole and the midlatitudes), which is associated with less advection of relatively warm air from the Atlantic Ocean over Western Europe. From observational data (BALDWIN and DUNKERTON, 1999, 2001) and from model data (GERBER and POLVANI, 2009) it is noted that the tropospheric annular mode remains in the negative phase for an extended period of time after a sudden stratospheric warming (SSW). During an SSW the polar stratosphere is substantially warmed and the polar stratospheric vortex is severely weakened or even broken down (LIMPASUVAN et al., 2004; CHARLTON and POLVANI, 2007; MATTHEWMAN et al., 2009). THOMPSON et al. (2002) noted that a weakening of the wintertime stratospheric polar vortex is often followed by cold conditions over north-east America, northern Europe and eastern Asia, that can persist for about 2 months. An SSW is classified as "major" if the zonal average zonal wind at 10 hPa and 60°N and the zonal average temperature gradient between 60°N and the North Pole change sign.

In the winter of 2008/2009 this condition was satisfied first on 24 January (LABITZKE and KUNZE, 2009). The persistence of the coldness throughout the winter of 2008/2009 might be linked to this SSW. Here we investigate to what extent an SSW influences the troposphere, by examining how the zonal average winds in the troposphere are affected by the changes in the zonal mean state of the stratospheric potential vorticity (PV) that accompany an SSW, focusing on the major SSW of January 2009.

An SSW is accompanied by strong and widespread changes in the polar cap stratospheric PV distribution. From the PV invertibility principle (KLEINSCHMIDT, 1950; HOSKINS et al., 1985) we know that this must

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0941-2948/2011/0503 $ 3.60
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DOI 10.1127/0941-2948/2011/0503

be accompanied by changes in the wind distribution that may also penetrate into the troposphere (e.g., HARTLEY et al., 1998; BLACK, 2002; BLACK and MCDANIEL, 2004; HINSSEN et al., 2010). Here we use PV inversion to examine the effect on the zonal tropospheric winds of daily changes in the zonal mean stratospheric PV distribution during the period running from November 2008 to March 2009. We compare this winter with so-called "undisturbed winters" that were devoid of significant SSWs. This comparison brings out very clearly the long-lasting (in the order of months) influence of the major SSW on the tropospheric zonal flow in the winter of 2009. Moreover, an examination of two other winters with significant SSWs reiterates this conclusion and suggests a possibility for winter weather forecasts on a monthly time scale.

Our study is related to that of BLACK and MCDANIEL (2004), who used quasi-geostrophic PV inversion to show that PV anomalies in the stratosphere associated with so-called "northern annular mode events" (based on the 10 hPa flow structure) affect the tropospheric circulation. They show that a change in stratospheric PV over a nine day period around a stratospheric event is related to a geostrophic wind change in the troposphere. Here we relax the quasi-geostrophic approximation and perform PV inversion in isentropic coordinates.

Section 2 presents an overview of the data used and our definition of splitting the PV into a reference state and an anomaly. Also a short description of the PV inversion method used is given in this section. The results of inverting the stratospheric part of the PV distribution is shown in section 3. Finally, section 4 presents a summary and some conclusions.

2 Data and inversion method

The European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-interim reanalysis data on 37 pressure levels between 1000 hPa and 1 hPa is used for the analysis. This is the most recently produced reanalysis dataset, with an improved representation of the stratosphere compared to the previous ERA-40 reanalysis (FUEGLISTALER et al., 2009). The daily data (from the analysis of the atmospheric model at 12:00 UTC) of the zonal wind and the temperature for November to March are interpolated from isobaric to isentropic levels by the method described by EDOUARD et al. (1997) and zonally averaged for the winters 1993/1994, 1995/1996, 1996/1997, 2003/2004, 2005/2006, and 2008/2009.

Due to the large increase of θ with height in the stratosphere, a higher vertical resolution in terms of θ is desired in the lower stratosphere compared to the upper stratosphere. In order to make optimal use of the available ERA-interim data, a stretched grid in the vertical direction is therefore employed:

$$\theta = \theta_s e^{z^*/H} \tag{2.1}$$

with θ_s = 250 K and H a vertical scale height ($H = R\theta/g$, with R the gas constant for dry air and g the gravitational acceleration). Levels are placed at equal distances in z^* (50 m), implying that the vertical distance between two levels in terms of θ increases with height.

The isentropic potential vorticity Z_θ (HOSKINS et al., 1985) is then calculated from:

$$Z_\theta = \frac{\zeta_\theta + f}{\sigma} \tag{2.2}$$

Here ζ_θ is the isentropic relative vorticity, f is the Coriolis parameter and σ is the isentropic density:

$$\zeta_\theta = -\frac{1}{a}\frac{\partial u}{\partial \phi} + \frac{u \tan\phi}{a} \tag{2.3}$$

$$\sigma = -\frac{1}{g}\frac{\partial p}{\partial \theta} \tag{2.4}$$

Here u is the zonal wind, θ is the potential temperature, a is the radius of the Earth, ϕ is the latitude, p is the pressure and g is the gravitational acceleration. This gives a dataset of zonal mean PV, zonal wind and pressure on isentropic levels. The domain of this dataset ranges from 10.5°N to 90°N in the horizontal, with a resolution of 1.5°, and from the lower troposphere to about 1250 K in the vertical, with a resolution varying from 2 K in the troposphere to about 10 K in the upper layers. 10°N is chosen as the equatorward boundary of the domain, since the PV changes sign near the equator, and the PV inversion equation is not solvable for negative PV values. The lower boundary of the domain is determined by the available data and for each latitude this is defined as the first isentropic level for which data is available along the full latitude circle, meaning that the pressure is below 1000 hPa at all longitudes along this latitude circle. We will refer to this lower boundary as the 'surface', but it should be noted that this does not correspond to the actual surface of the Earth, but is located in the low troposphere. This 'surface-isentropic level' varies with latitude and from day to day, from around 260 K at 90°N to about 310 K near the equator.

To study the relation between the potential vorticity and the zonal wind we use a PV inversion equation. We assume axi-symmetry around the North Pole. The PV inversion equation is derived from the definition of the isentropic potential vorticity (Eq. (2.2)). The zonal mean zonal wind is assumed to be in hydrostatic balance and in gradient wind balance with the zonal mean temperature and pressure. The PV inversion equation in isentropic coordinates reads (see also HINSSEN et al., 2010):

$$\frac{Z_\theta}{g}\frac{\partial}{\partial \theta}\left(\rho\theta f_{loc}\frac{\partial u}{\partial \theta}\right) + \frac{\partial^2 u}{\partial r^2} + \frac{\tan\phi}{a}\frac{\partial u}{\partial r} - \frac{u}{a^2\cos^2\phi} = \sigma\frac{\partial Z_\theta}{\partial r} - \frac{\partial f}{\partial r} \tag{2.5}$$

with

$$f_{loc} = f(r) + \frac{2\tan\phi}{a}u; r = a\left(\frac{\pi}{2} - \phi\right); \\ f(r) = 2\Omega\sin\phi = 2\Omega\cos\left(\frac{r}{a}\right) \quad (2.6)$$

Here, r is the distance from the pole measured along the surface of the Earth, ρ is the density and Ω is the rotation rate of the Earth.

The PV inversion equation (Equation 82.5)) can be seen as the formulation of thermal wind balance in terms of potential vorticity. It describes the flow pattern that is associated with a specific pattern of the potential vorticity in a balanced axi-symmetric vortex centred at the pole.

The boundary condition at the pole ($r = 0$) is simply that $u = 0$, because the vortex is assumed to be axi-symmetric and centred at the pole. At the outer boundary at 10.5°N we prescribe the wind according to the circulation theorem (HOSKINS et al., 1985, p 897). At the upper boundary we impose the ERA-interim wind. At the lower boundary thermal wind balance is imposed, which is approximated in isentropic coordinates by:

$$\frac{\partial u}{\partial \theta} = \frac{c_p}{f\theta}\left(\frac{\partial T}{\partial r}\right)_\theta \quad (2.7)$$

with c_p the heat capacity of dry air at constant pressure and T the temperature averaged over the lowest layer between two isentropic levels. The temperature is derived from the zonal mean pressure on isentropic levels.

On the basis of the PV inversion equation, the zonal mean PV and the isentropic density are split into a reference state and an anomaly as follows:

$$Z_\theta \equiv Z_{\theta,ref} + Z'_\theta; \sigma \equiv \sigma_{ref} + \sigma' \quad (2.8)$$

with

$$Z_{\theta,ref} = \frac{f}{\sigma_{ref}} \quad (2.9)$$

and

$$\sigma_{ref} = \frac{\int \sigma \cos(\phi) d\phi}{\int \cos(\phi) d\phi} \quad (2.10)$$

In other words, the reference isentropic density σ_{ref} is the area-weighted average of σ over the domain in question. In our case we choose the area poleward of 10°N. Therefore, σ_{ref} depends only on θ. The reference PV is associated with the solution $u = 0$ and $\sigma = \sigma_{ref}$, indicating that the PV anomaly represents that part of the PV field that induces a wind field, according to the PV inversion equation. The total PV is used for the inversion, but the PV anomaly will be discussed in relation to the resulting wind field, since variations in the PV anomaly are associated with variations in the wind field. Both the seasonal cycle of the stratospheric PV distribution and the impact of SSWs are very well visible in the PV anomaly, as is illustrated in the next section.

3 PV inversions

Equation (2.5) is solved by successive relaxation. The wind field that is obtained from inversion of the daily PV in the full domain is reasonably close to the ERA-interim wind field during November 2008 to March 2009. There are some differences between the inverted and ERA-interim wind field, especially in the subtropics to midlatitudes near the lower boundary, where the inverted wind speed is on average a few m s^{-1} lower than the ERA-interim wind speed. We, however, believe that the inversion of the stratospheric PV between 400 K and 1250 K will give a good indication of the balanced influence of the stratospheric PV on the tropospheric wind. In the stratospheric inversions, the reference PV is used as a background distribution in the whole domain, and zero thermal wind is applied as the lower boundary condition, while the other boundary conditions remain the same as described in the previous section. By applying zero thermal wind at the lower boundary, we assume that the influence of the stratospheric PV on the surface thermal wind is negligible. This is an assumption, but it is believed that the surface temperature distribution is mainly related to the tropopause PV anomaly, since this PV anomaly is located much closer to the lower boundary than the stratospheric polar cap PV anomaly.

The daily stratospheric inversions are discussed in section 3.1, where comparison between 2008/2009 and 1996/1997 shows that an influence of the January 2009 SSW is still present in late winter. The winters of 1993/1994, 1995/1996, 2003/2004 and 2005/2006 are also discussed, to indicate that the 2009 SSW event was quite extreme, but that the tropospheric winds are also affected by weaker SSWs. Inversion results for 25 day averages before and after the 2009 SSW are discussed in section 3.2. Comparison with the ERA-interim wind allows us to examine the importance of the stratosphere in explaining the observed wind difference that is associated with the SSW.

3.1 Daily influence of the stratosphere on the troposphere

Figure 1 shows the polar cap PV anomaly as function of time and potential temperature for 6 different years. The polar cap PV anomaly is defined as the area weighted average PV anomaly poleward of 70°N. The left panels correspond to winters with SSWs, while the right panels correspond to relatively undisturbed winters. Negative values of the polar cap PV anomaly are characteristic for the summer, except near the tropopause, where we observe a persistent positive anomaly, which is referred to

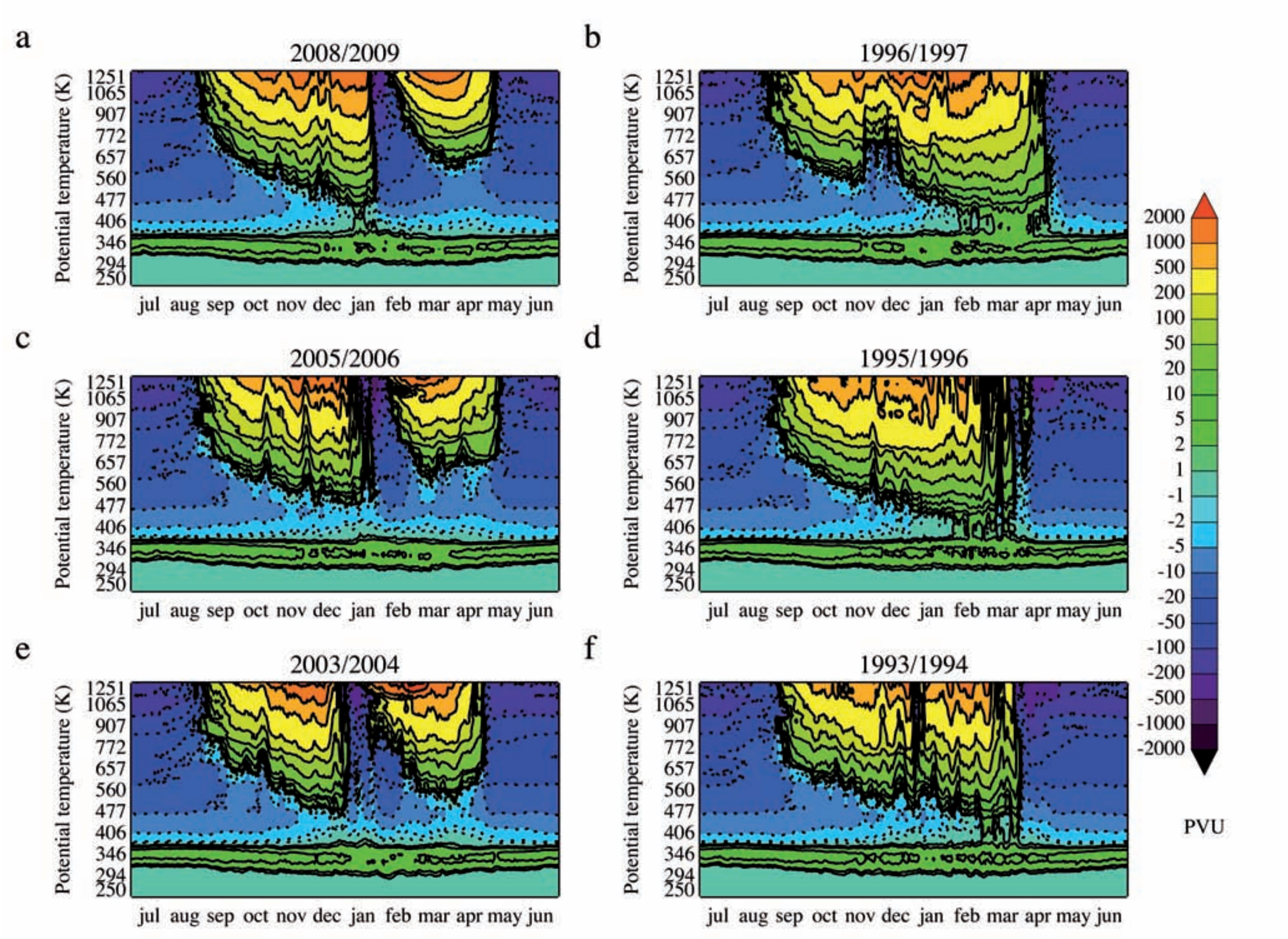

Figure 1: Daily polar cap PV anomaly (area weighted average PV anomaly poleward of 70°N, in PVU) for (a) July 2008 to June 2009, (b) July 1996 to June 1997, (c) July 2005 to June 2006, (d) July 1995 to June 1996, (e) July 2003 to June 2004, (f) July 1993 to June 1994, as a function of potential temperature (K) and time. Contours at ± 1, 2, 5, 10, 20, 50, 100, 200, 500, 1000, 2000, 3000 PVU and negative values represented by dashed lines.

as the Upper Troposphere Lower Stratosphere (UTLS) PV anomaly. A positive stratospheric PV anomaly, and an associated cyclonic stratospheric vortex, begins to develop at high levels in September and subsequently expands downwards. However, a negative PV anomaly persists at levels around 450 K until at least December. If there are no significant SSWs during the winter, such as in the winters of 1993/1994, 1995/1996 and 1996/1997, the negative PV anomaly ultimately vanishes completely in February. The positive stratospheric PV anomaly then makes a connection with the UTLS PV anomaly. The SSWs are clearly visible by the very abrupt restoration of the summer PV anomaly distribution, i.e. the PV anomaly becomes negative at all levels above 500 K. The polar UTLS PV anomaly does not seem to be affected by the SSW. After the SSW event a new positive stratospheric PV anomaly forms at upper levels and expands downwards over a period of many weeks, analogous to the developments in September and October. This recovery of the positive stratospheric PV anomaly is, however, not completed because of the the final stratospheric warming and the associated vortex break-up in March or April.

Figure 2 shows the zonal average zonal wind speed, obtained from inverting the daily stratospheric (400 K–1250 K) PV anomaly, as a function of time from November 1 to March 31, at 304 K and 50°N, at 298 K and 60°N and at 292 K and 70°N for the same six winters as shown in Figure 1. These isentropic levels are located in the mid-troposphere approximately half way between the average tropopause level and the average lower boundary for each latitude (averaged over all 151 days). The tropopause is defined as the 2 PVU isopleths (1 PVU = 10^{-6} K m^2 kg^{-1} s^{-1}). The results do not depend strongly on the choice of the isentropic level. Figure 2a demonstrates that the PV changes due to the 2009 SSW have a profound influence on the tropospheric winds. This influence is easterly in early winter

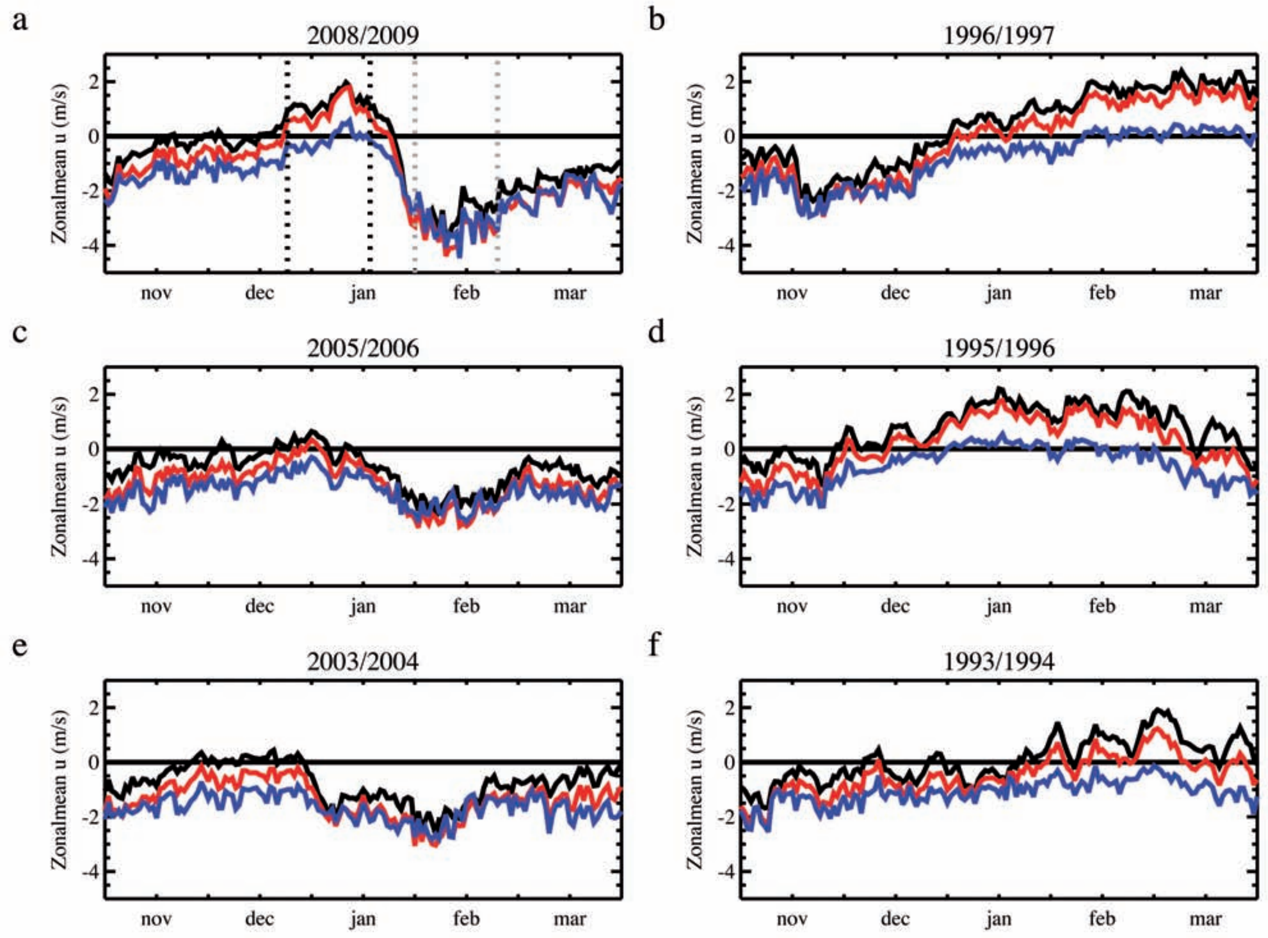

Figure 2: Daily zonal mean zonal wind (m s^{-1}), obtained from inverting the stratospheric PV anomaly between 400 K and 1250 K, at 70°N and 292 K (black), at 60°N and 298 K (red) and at 50°N and 304 K (blue), as a function of time for (a) 1 November 2008 to 31 March 2009, (b) 1 November 1996 to 31 March 1997, (c) 1 November 2005 to 31 March 2006, (d) 1 November 1995 to 31 March 1996, (e) 1 November 2003 to 31 March 2004, and (f) 1 November 1993 to 31 March 1994. The tickmark accompanying the name of a month corresponds to approximately the middle of the month. In (a), the vertical black dashed lines indicate the 25 day period 25 December 2008 to 18 January 2009 before the SSW, and the vertical grey dashed lines indicate the 25 day period 31 January to 24 February 2009 after the SSW.

(–1 to –2 m s^{-1}), decreasing in magnitude and becoming westerly in January (1 to 2 m s^{-1}), until the occurrence of the SSW in mid-January, after which the stratospheric influence becomes easterly (–3 to –4 m s^{-1}) again. The largest easterly influence is observed in early to mid-February, but the stratospheric influence remains easterly at least up to the end of March.

This behaviour constrasts with the relatively undisturbed winter of for instance 1996/1997. Totally undisturbed winters in the northern hemisphere are not observed between 1989 and 2009. A distortion of the vortex is even observed in mid-November to mid-December 1996 below 800 K. The associated decrease of the stratospheric PV anomaly is observed to have an easterly influence on the tropospheric winds in mid- to high latitudes (figure 2b). Nevertheless, in this winter the westerly influence of the stratosphere on the tropospheric winds increases further in the second half of the winter, especially at latitudes north of 60°N.

A comparison of Figures 2a and 2b makes clear that the 2009 SSW influences the tropospheric winds for several months. The difference in the stratospheric influence on the tropospheric winds in the mid- to high latitudes in March between the undisturbed winter of 1996/1997 and the highly disturbed winter of 2008/2009 is of the order of 5 m s^{-1} in the high latitude troposphere. This is consistent with the findings of BALDWIN and DUNKERTON (2001), who also find that the anomalies in the lower stratosphere can persist for more than two months, while anomalies in the middle to upper stratosphere are less persistent. The total tropospheric wind difference in March between 2009 and 1997 is not dominated by the stratospheric influence, indicating that tropospheric processes also play a role.

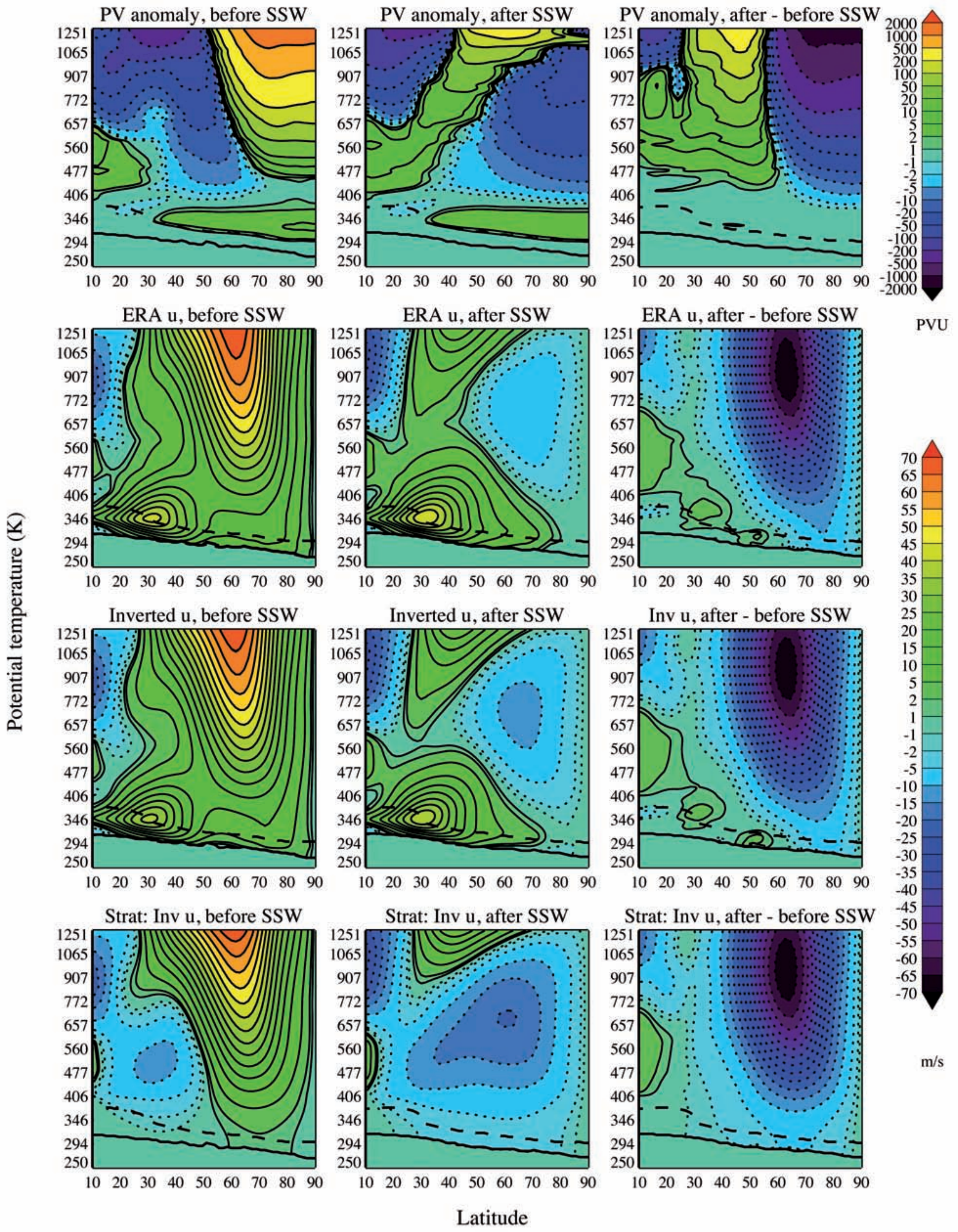

Figure 3: Daily zonal mean PV anomaly (in PVU, first row), ERA-interim zonal wind (in m s^{-1}, second row), wind obtained from inverting the total PV (in m s^{-1}, third row) and wind obtained from inverting the stratospheric PV between 400 K and 1250 K (in m s^{-1}, fourth row), averaged over a 25 day period before the SSW (25 December 2008 to 18 January 2009, left column), averaged over a 25 day period after the SSW (31 January to 24 February 2009, middle column), and the difference between the period after the SSW minus the period before the SSW (right column), as a function of potential temperature (K) and latitude (°N). The position of the tropopause is indicated by the thick dashed line, while the thick solid line near the bottom of each figure represents the lower boundary of the inversion domain. Contours for the PV anomaly as in Figure 2, and contours for the wind field are every 5 m s^{-1} with the ± 1 and 2 contours added, with negative values represented by dashed lines and zero contours omitted.

During the winters of 2005/2006 (Figure 1c) and 2003/2004 (Figure 1e) the polar cap PV anomaly was also strongly disturbed by an SSW. In Figure 2 it can be seen that these events had a similar influence on the midlatitude tropospheric zonal wind as the 2009 SSW. The effect of an SSW on the tropospheric zonal winds is felt for quite some time after the event. The magnitude of the effect depends on the strength of the SSW, i.e. whether it is a vortex split (SSW of 2009, primary contribution by wavenumber 2) or a vortex displacement (SSWs of 2003/2004 and 2005/2006, primary contribution by wavenumber 1) (HARADA et al., 2010).

3.2 Wind and PV changes due to the SSW

To further examine the importance of the stratosphere in explaining the change in the zonal wind due to the SSW we focus on the major SSW of January 2009. Two 25 day periods are considered: 25 December 2008 to 18 January 2009 (before the SSW) and 31 January to 24 February (after the SSW, see Figure 2a). The daily zonal mean zonal wind and pressure are averaged over these 25 day periods, after which the PV and PV anomaly are calculated for each period (using Eq. 82.2) to 82.4) and Eq. (2.8) to (2.10)). The stratospheric PV anomaly above 400 K is then inverted for both periods.

Figure 3 presents the PV anomaly (first row), the ERA-interim zonal mean zonal wind field (second row), the inverted zonal wind field from the inversion of the total PV (third row) and the inverted zonal wind from the inversion of the stratospheric PV above 400 K (fourth row), for the period before the SSW (left column), the period after the SSW (middle column), and for the difference between the period after minus the period before the SSW (right column). The stratospheric PV anomaly over the pole is strong and positive before the SSW, while it is negative after the SSW. This is clearly visible in the difference plot, which shows a strong decrease of the PV anomaly over the pole after the SSW compared to before. A strong decrease is also seen in the ERA-interim wind field, showing strong westerly winds in the stratosphere before the SSW and weak easterlies afterwards. The wind difference shown in the right panel of the second row shows that the change in wind is not restricted to the stratosphere, but that also the high latitude troposphere experiences an easterly "forcing" of about 5 m s^{-1}. The wind speeds from inverting the total PV are somewhat lower than the ERA-interim wind speeds, mainly in the subtropics to midlatitudes in the lower part of the domain, but the inverted wind change due to the SSW (Figure 3 third row, right panel) is practically the same as the ERA-interim wind change (Figure 3 second row, right panel). The wind fields obtained from inverting the stratospheric PV (fourth row of Figure 3) show that there is a weak westerly influence (1 to 2 m s^{-1}) of the stratosphere above 400 K on the troposphere before the SSW in the high latitudes, and a stronger easterly influence (-2 to -4 m s^{-1}) poleward of 20° after the SSW. The difference (lower right panel of Figure 3) is not restricted to the stratosphere, but attains values of the order of 5 m s^{-1} in the mid to high latitude troposphere.

Both the wind change related to the stratospheric PV change (Figure 3 lower right panel) and the total observed and inverted wind change (Figure 3 right column middle panels) show clear tropospheric changes in the wind, indicating that the stratosphere played an important role in the 2008/2009 winter. The maximum stratospheric influence on the troposphere is found between 60°N and 70°N, while the maximum wind difference seen in the ERA-interim data and also in the inverted wind from the total PV is between 70°N and 80°N, similar to the results BLACK and MCDANIEL (2004) find for the negative stratospheric northern annular mode event during March 1989 (their figure 5). Internal tropospheric variability can mask part of the stratospheric signal (BLACK and MCDANIEL, 2004; GERBER et al., 2009), so tropospheric processes can adjust the wind change that would result from the stratosphere alone. However, it is clear from Figure 3 that the stratosphere imposes a strong easterly forcing on the tropospheric wind due to the SSW, and that without the stratospheric changes the tropospheric winds would have been substantially more westerly.

4 Conclusions

We investigated the influence of the zonal mean stratospheric PV on the zonal mean zonal wind in the troposphere for the period November 2008 to March 2009, with special focus on the SSW that occurred during January 2009. Piecewise potential vorticity inversion was used to invert the stratospheric PV between 400 K and 1250 K and study its influence on the tropospheric wind field.

The 2009 SSW had a strong influence on the stratospheric PV anomaly over the pole, with positive values before the SSW, and negative values after the SSW, especially in the lower to middle stratosphere. Accordingly, the influence of the stratospheric PV on the tropospheric winds was westerly before the SSW, while it was easterly afterwards. This resulted in zonal mean zonal wind changes up to 5 m s^{-1} within a few weeks, which are also visible in the difference between a 25 day average before the SSW and a 25 day average after the SSW. These changes were found to be a substantial contribution to the wind difference observed in the ERA-interim data, indicating that the stratosphere influenced the tropospheric conditions, leading to a weakening of the westerlies or even a switch to easterlies after the occurrence of the January 2009 SSW.

These conclusions are consistent with the conclusions of BLACK and MCDANIEL (2004) and GERBER et al. (2009). The results of BLACK and MCDANIEL (2004) are based on quasi-geostrophic PV inversion. We used isentropic PV inversion, where σ is part of the solution. The similarity between our results and those of BLACK and MCDANIEL (2004) shows that σ in the troposphere is hardly affected by the stratospheric PV above 400 K, an assumption already made in advance in the quasi-geostrophic theory. Our results thus confirm that the quasi-geostrophic theory is suitable to examine the influence of the stratospheric PV on the wind field in the troposphere.

Comparison of the winter of 2008/2009 with a relatively undisturbed winter (1996/1997) indicated that the effect of the SSW in the troposphere and lower stratosphere is felt for months after the occurrence of the SSW. In March 2009, an easterly influence of about 2 m s^{-1} was found, while a westerly influence of the same order of magnitude was found in absence of an SSW in March 1997. Qualitatively similar results are found for two other recent winters during which major SSWs occurred (2003/2004 and 2005/2006), but the 2009 SSW was a more extreme event. The quantitative results for 2009 are therefore not representative for all winters with an SSW, but examination of this strong, clear event indicates that the stratospheric changes that accompany an SSW can affect the tropospheric winds.

An SSW, initially forced by waves from the troposphere, affects the stratospheric PV distribution for a long time. This presents a possibility for extended range winter weather forecasts.

Acknowledgements

The ECMWF ERA-interim data used in this study have been provided by ECMWF. We thank Anna VON DER HEYDT for the German translation of the abstract.

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