SUDDEN STRATOSPHERIC WARMING

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One of the most dynamically active processes in the stratosphere results in abnormal warming in the region above the pole and it is called the sudden stratospheric warming (SSW). Although relatively precise monitoring of such events is done, there is still some confusion referring SSWs, because various different methods and criteria are being used to describe a SSW which leads to differences in the number of observed events. In this paper, the mechanisms, leading to the SSW are reviewed. In addition, by using ERA5 reanalysis data, it is illustrated, how the upward wave propagation interacts with the mean zonal flow and causes the SSW in winters 2009 and 2018.

NENADNO SEGREVANJE STRATOSFERE

Nenadno segrevanje stratosfere sodi med najbolj dinamično aktivne procese, ki se dogajajo v tem delu ozračja. Pri tem nad polarnimi predeli v stratosferi temperatura bistveno naraste v relativno kratkem časovnem intervalu. Čeprav so ti dogodki opazovani že nekaj desetletij, pa je klasifikacija samega dogodka zaradi mnogih definicij in kriterijev neenotna, kar se odraža v različni pogostoti pojavljanja teh dogodkov. Članek obravnava obnašanje vertikalnega širjenja planetarnih valov, kar je dominanten razlog za nastanek nenadnega segrevanja stratosfere. Članek zajema tudi vpliv planetarnih valov na časovni razvoj tega dogodka. Za izračune so bili uporabljeni podatki reanalize ERA5 in sicer za leti 2009 in 2018.

1. Introduction

The stratosphere is a layer in the Earth's atmosphere that lies just above the troposphere. It is highly stratified and as a result, the horizontal motions dominates there. The stratosphere extends up to the tropopause, which is approximately fifty kilometers above the Earth's surface, while the lower boundary of the stratosphere varies from six to eighteen kilometers with respect to the latitude. This layer contains fifteen percent of the atmospheric mass and one of its most important characteristics is ozone content, which protects life on Earth from harmful ultraviolet radiation. The ozone absorbs ultraviolet radiation and this process warms up the stratosphere. Stratospheric circulation is most variable during winter when the polar vortex plays a dominant role in stratospheric dynamics. This variability can sometimes be large enough to trigger the sudden stratospheric warming. Although this event happens in the upper stratosphere, it can sometimes influence the tropospheric weather. If the anomalies are large enough, they can reach the troposphere, where the tropospheric polar vortex is affected. This can result in cold air outbreaks that can reach mid-latitudes thus bringing unusually low temperatures to that regions [1].

2. Stratospheric polar vortex

When the polar night occurs, the radiation income decreases in the polar region, so more radiation is emitted than received, which leads to the temperature drop. The temperature gradient between that polar region and extratropics increases, which leads to strong westerly winds. These winds capture the cold air and prevent it from outbreaking from the polar region. So, the polar vortex is a planetary-scale mid-to-high latitude circumpolar circulation. The extent of the polar vortex is defined by several definitions, the most frequently used are by the region of low geopotential height or by the region of high potential vorticity. Potential vorticity is a quantity, proportional to the product of vorticity and stratification. Vorticity is a measure of the rate of rotation of air parcels. Polar vortices can be observed in both the troposphere and the stratosphere, but the two have different characteristics and are not directly connected to each other. The stratospheric polar vortex tends to be smaller in size, but stronger and well organized, while its tropospheric counterpart tends to be greater in size and it usually contains slower winds and its shape is more disrupted. The reasons for its disrupted shape are mainly the Earth's topography and the land-sea contrasts. The stratospheric polar vortex occurs in both northern and southern hemisphere. The necessary condition for the existence of the stratospheric polar vortex is the polar night, while the tropospheric polar vortex can be seen during all seasons [2]. During the polar night, the winter's hemisphere polar region is isolated from getting any of the Sun's radiation and consequently the temperature there drops. This temperature drop increases the temperature gradient between the equatorial and polar region and this leads to strong westerly winds in mid latitudes. The ozone layer in the stratosphere absorbs the ultra-violent radiation that comes from the Sun and during this process, the stratosphere warms up. But during the polar night, the isolated area does not receive much radiation and this is why the polar region on the winter's hemisphere gets cooler.

The edge of the stratospheric polar vortex is defined by the strong circumpolar westerlies that maximize at the latitude of around $60°$, from just above the tropopause (around $100\,\mathrm{hPa}$) into the mesosphere (above $1\,\mathrm{hPa}$) [2]. During the winter months the polar vortex varies in strength and it is disrupted by planetary-scale Rossby waves which originate in the troposphere and transport the momentum upward into the stratosphere. These waves deposit momentum into the stratosphere, affecting the dynamics of the stratospheric polar vortex and leading to the variability of stratospheric circulation. The stratosphere is modified by upward propagating planetary waves, which leads to the downward propagation of the stratospheric circulation anomalies and in some cases this anomalies can reach the troposphere. These anomalies are a result of wave-breaking events that occur in the stratosphere. When a sudden stratospheric warming occurs, the waves break on the lower boundary of the affected area and these events creates an illusion of downward propagation of anomalies, that represent a sudden stratospheric warming. These variations are well characterized by temperature anomalies, zonal wind anomalies or annular modes (AMs). AMs are the hemispheric-scale patterns, characterized by synchronous fluctuations in pressure of one sign above the poles and of opposite sign at mid-latitudes [3].

Stratosphere and troposphere are dynamically coupled but not all stratospheric anomalies can reach the troposphere. Only the strongest anomalies can travel through the tropopause, which is the border layer between the stratosphere and the troposphere, and then affect the tropospheric dynamics. When there is a lot of wave forcing present, this results in a weak, disorganized polar vortex but when there is little wave forcing, the polar vortex tends to be stronger. Stronger polar vortex results in storm track that is displaced farther north. In the case of strong polar vortex, the jet stream is stronger which results in more intense surface winds and more intense cyclones. Another consequence of an unusually strong polar vortex is the formation of the ozone hole above the polar region. On the other hand, the weak state of the polar vortex can result in a sudden stratospheric warming which can affect the tropospheric weather regimes and trigger cold air outbreaks. Such events are shown to be more frequent when the quasi-biennial oscillation is in the easterly phase [1].

3. Sudden stratospheric warming

Waves propagate from the troposphere to the stratosphere where they break and these wave-breaking events decelerate the zonal flow in the polar vortex region in the stratosphere, sometimes to the point, where the zonal winds reverse from westerlies to easterlies. The amplitude of these waves increases with decreasing density and the waves eventually break. They break where the speed of the surrounding air equals to the the phase speed of these waves. Zonal wind reversal is accompanied by the temperature increase of tens of degrees Celsius over a period of a few days above the pole. This warming happens when the meridional advection in the polar region increases, so the air exchange between the polar vortex and surrounding air becomes significant. This leads to the outbreak of cold air to the mid-latitudes. Cold air above the polar region is then replaced with the much warmer air from mid-latitudes which results in warming that happens abnormally fast in the polar region. This event is called the sudden stratospheric warming and it occurs, on average, approximately six times per decade in the northern hemisphere, but only one event has ever been observed in the southern hemisphere in the year of 2002 [1].

The polar vortex in the southern hemisphere is far stronger than the one in the northern hemisphere. The main reason for that is the difference in the wave forcing on each hemisphere. While the northern hemisphere has diverse orography and a great land-sea contrast, the mid-latitudes area in the southern hemisphere has much more uniform properties, which leads to a smaller upward wave propagation and as a consequence, the polar vortex tends to be stronger in the southern hemisphere as it is harder to disturb [1].

In general, we divide the sudden stratospheric warming event into two categories: the displacement event and the split event. The displacement event represents an event, at which the whole polar vortex is displaced from the pole. The split event occurs when the polar vortex gets split into two or more smaller vortices. Another important term is the term final warming, which occurs every winter and it announces the end of the polar vortex for that season (the polar vortex does not form back from the final warming). Final warmings do not contribute to the total frequency of the sudden stratospheric warmings. Sudden stratospheric warmings can be further divided into two groups - the minor and the major events. In general, we are mostly interested in the major events, that have a higher possibility of reaching the troposphere. Unfortunately, there is no uniform criteria for a sudden stratospheric event in meteorology, so that different definitions of such event results in different frequency of the sudden stratospheric warming event.

The comparison of SSW event frequencies for different criteria of the event is shown in Table 1. There are actually many other definitions that characterizes sudden stratospheric warming, but only a few of them are shown here. Although the definitions are very similar, the frequency of the sudden stratospheric warming event differs among definitions [4].

Table 1. The average number of major SSW events per year (SSW/yr) for some different SSW definition for 57 winters in ERA-40 (Jan 1958-1989)/ERA-interim (Mar 1989-Apr 2014), adapted from [4].

Definition	Description	SSW/yr
Zonal wind and temperature gradient reversal	Zonal-mean zonal winds at 10 hPa and 60°N fall below $0\,\mathrm{m\,s^{-1}}$ and meridional temperature gradient reverses. Winds must return to westerly for at least 20 days between the events and they must remain westerly for at least 10 days (to exclude final warmings).	0.65
Zonal wind reversal.	Zonal-mean zonal winds at 10 hPa at 60°N fall below $0\,\mathrm{m\,s^{-1}}$ from November to March. Events must return to westerly for at least 20 days between events and they must remain westerly for at least 10 days.	0.65
Zonal wind reversal.	Zonal-mean zonal winds at 10 hPa at 65°N fall below $0\,\mathrm{m\,s^{-1}}$ from November to March. Events must return to westerly for at least 20 days between events and they must remain westerly for at least 10 days.	0.84
Averaged zonal winds reversal.	Zonal-mean zonal winds at 10 hPa, cosine weighed and averaged from 60°N to 90°N fall below $0\,\mathrm{m\,s^{-1}}$ from November to March.	0.91

Sudden stratospheric warming can be easily detected by observing time series of zonal-mean zonal winds $\overline{u}(t)$ or the zonal-mean temperature $\overline{T}(t)$ at the 10 hPa pressure level. Figure 1 shows these timeseries at three different latitudes. The wind direction reversal and the temperature rise can be observed between January 17 and January 26. The chosen latitudes are often used to determine the sudden stratospheric warming event and, as it is seen from the figure, these latitudes give similar results for the wind reversal and for the temperature rise. The red line in the Figure 1a represents the threshold for declaring a sudden stratospheric warming event.

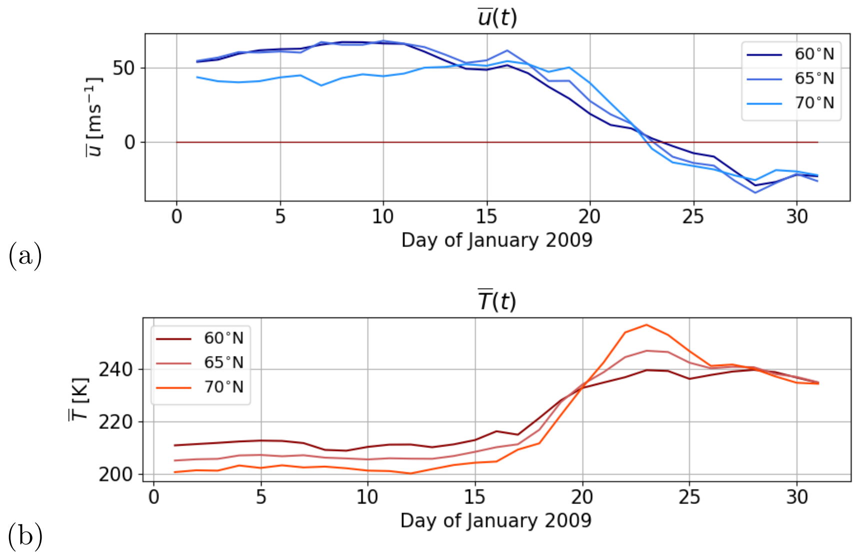

Figure 1. Time series of zonal-mean zonal wind $\overline{u}$ (a) and zonal-mean temperature $\overline{T}$ (b) at the 10 hPa pressure level at three different latitudes for January 2009 during the sudden stratospheric warming. The red line on the Figure 1a represents the zero wind line.

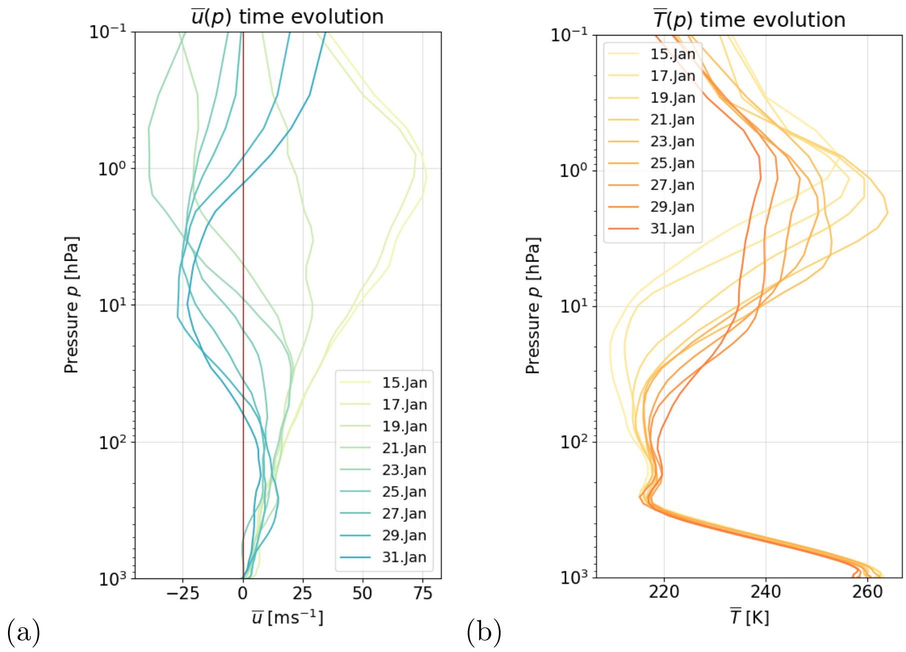

Figure 2. Vertical profile of zonal-mean zonal wind $\overline{u}$ (a) and zonal-mean temperature $\overline{T}$ (b) for specific days on January 2009 at latitude of 60°N during the sudden stratospheric warming. The brighter colors represent earlier days (days before sudden stratospheric warming event) and the stronger colors represent days during and after the event. The red line on the Figure 2a represents the zero wind line.

There are some differences in intensity of the change, but the dates of the event occurrence on different latitudes overlap. The changes in wind intensity or the changes in temperature are significant which proposes that the sudden stratospheric event happened. The vertical structures of zonal-mean zonal wind and zonal-mean temperature during the studied event are shown in Figure 2. The vertical profile for several different dates is plotted on Figure 2. The plots were produced using ERA5 reanalysis data. The reversal of zonal-mean zonal wind can be seen on the left graph and it is shown how variable and non-uniform wind in the stratosphere and in the lower mesosphere actually is. The sudden stratospheric warming has different influence at different pressure levels. For example, the graph on Figure 1a is only compatible with the data on 10 hPa pressure level on the Figure 2a and differs a lot from values on different pressure levels. The same applies to the vertical profile of zonal-mean temperature.

4. Wave-mean flow interaction

The circulation from equatorial regions to the winter's polar region in the stratosphere is primarily driven by upward propagating waves from the troposphere and this large-scale dynamical process is called the Brewer-Dobson circulation. It can be studied through the transformed Eulerian mean (TEM) equation. This equation is derived from the momentum equation

$$\frac{\mathrm{D}u}{\mathrm{D}t} - fv + \frac{\partial \Phi}{\partial x} = X, \tag{1}$$

where u is the zonal component of wind velocity, t is time, f is the Coriolis parameter, v is the meridional component of wind velocity, Φ is the geopotential and X is the zonal component of drag due to small scale eddies. We are interested into zonal mean circulation, so this equation is then zonally averaged. Zonally averaged circulation is represented by the longitudinally averaged flow and the longitudinally varying disturbances. Such average that is calculated at fixed latitude, height, and time is called the Eulerian mean. By applying the Eulerian zonal mean on the momentum equation, we can obtain the conventional Eulerian mean (CEM) momentum equation

$$\frac{\partial \overline{u}}{\partial t} - f\overline{v} = -\frac{\partial(\overline{u'v'})}{\partial y} + \overline{X}. \tag{2}$$

The material derivative is now described by partial derivative and flux form. Primes represent the departure from the zonal mean, overbar denotes the zonal mean, and $\overline{X}$ is the zonally averaged turbulent drag which represents unresolved waves. In this expression, the advection term is neglected by the ageostrophic mean meridional circulation and vertical eddy flux divergence because these terms are small for quasi-geostrophic scales, compared to the other ones in equation 2. The CEM momentum equation can be transformed into the TEM equation, which provides a more direct view of transport processes in the meridional plane [5]. Here the residual circulation is defined by residual velocity components in spherical coordinates

$$\overline{v}^* = \overline{v} - \frac{1}{\rho_0}\frac{\partial(\rho_0 \overline{v'\theta'}/\frac{\partial\overline{\theta}}{\partial z})}{\partial z}, \tag{3}$$

$$\overline{w}^* = \overline{w} + \frac{1}{a\cos(\varphi)}\frac{\partial(\cos(\varphi)\overline{v'\theta'}/\frac{\partial\overline{\theta}}{\partial z})}{\partial \varphi}, \tag{4}$$

where w is the vertical velocity, ρ_0 is the standard density, θ is the potential temperature, a is the radius of the Earth and φ is the latitude. Residual circulation is associated with diabatic processes that are directly related to the mean meridional mass flow. Now the TEM equation can be obtained by substituting mean velocity components with residual mean velocities. This yields the TEM equation

$$\frac{\partial \overline{u}}{\partial t} = -\left[\frac{\frac{\partial(\overline{u}\mathrm{cos}(\varphi))}{\partial\varphi}}{a\mathrm{cos}(\varphi)} - f\right]\overline{v}^* - \bar{w}^*\frac{\partial \overline{u}}{\partial z} + \frac{\nabla\cdot F}{\rho_0 a\mathrm{cos}(\varphi)} + \overline{X}. \tag{5}$$

Here we examine the atmospheric dynamics using residual mean wind velocities that are used in the TEM equation to describe the zonal-mean meridional circulation. It describes the changes in zonal-mean zonal wind including contributions from small-scale gravity waves, that are not resolved by models (the $\overline{X}$ term), the mean westerly momentum transported across latitudes, and height (first two terms on the RHS) that helps to conserve angular momentum and mass continuity. The most important part of the TEM equation in this case is the Eliassen–Palm (E-P) flux divergence term (the third term on the RHS) which contributes the most to changes in zonal-mean zonal winds. Wave flux term F consists of meridional and vertical components and these components can be calculated as

$$F^{(\varphi)} = \rho_0 a\mathrm{cos}(\varphi)\left(\frac{\overline{v'\theta'}}{\frac{\partial\theta}{\partial z}}\frac{\partial\overline{u}}{\partial z} - \overline{v'u'}\right) \quad \text{and} \tag{6}$$

$$F^{(z)} = \rho_0 a\mathrm{cos}(\varphi)\left(\left[f - \frac{\frac{\partial(\overline{u}\mathrm{cos}(\varphi))}{\partial\varphi}}{a\mathrm{cos}(\varphi)}\right]\frac{\overline{v'\theta'}}{\frac{\partial\theta}{\partial z}} - \overline{w'u'}\right), \tag{7}$$

If the E-P flux divergence term is positive, then the wave forcing is accelerating the zonal winds thus strengthening the polar vortex, but if this term is negative, then the convergence of wave forcing is present which decelerates the jet stream and weakens the polar vortex [6]. Figure 3 represents wave forcing vectors, E-P flux divergence and the zonal-mean zonal winds at different months of 2018.

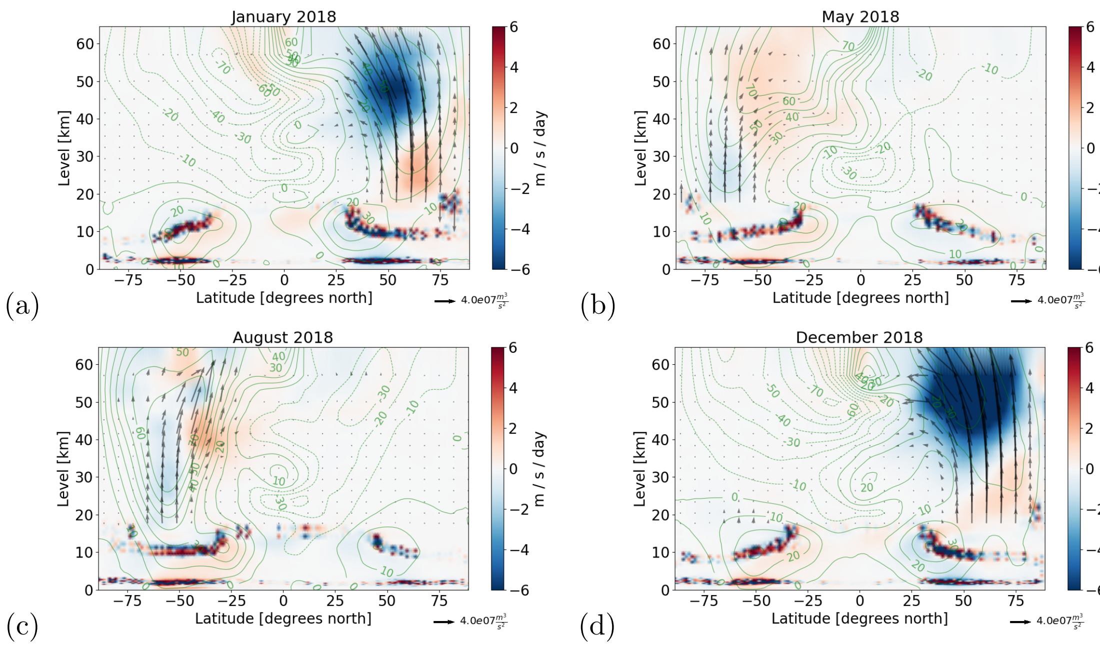

Figure 3. The average behaviour of wave forcing (arrows), E-P flux divergence (colors) and zonal-mean zonal winds (contours) for months in a year of 2018. The plots were produced using ERA5 reanalysis data.

The red areas on the plots show the divergence of E-P flux and the blue areas represent its convergence which can be understood as decelerating rate of the zonal winds in a velocity change units per day. These areas of convergence weaken the polar vortex and this increases the possibility of sudden stratospheric warming event. Areas of convergence/divergence are a result of strong wave forcing from the troposphere that is also seen on these plots. Another noticeable thing is the intensity of both the wave forcing and the E-P flux divergence on the southern hemisphere compared to the northern hemisphere. As discussed in the previous section, the amount of wave forcing on the southern hemisphere is way smaller than the amount on the northern hemisphere and consequently, the E-P flux divergence is not significant there, leading to a stronger and more organized polar vortex on the southern hemisphere. The wave forcing occurs only in the winter's hemisphere, where the stratosphere is dominated by strong westerly winds in which vertical propagation of planetary waves is possible. On the other hand, summer's hemisphere in the region of the stratosphere is dominated by easterly winds in which the planetary waves upward propagation is not possible. This is known as a Charney-Drazin criterion which implies that upward wave propagation in the atmosphere is only possible when the flow is westerly and not too strong [7]. This is well visible in the plotted subfigures of Figure 3, where the wave forcing only occurs when the winds are eastward and the wave forcing is significantly more intense in the northern hemisphere.

5. Examples of past SSW events

The frequency of sudden stratospheric warming events varies between six and nine per decade using different criteria to describe the event, but most of the criteria result in six to seven major events per decade [6]. Two major sudden stratospheric warmings happened in the years 2009 and 2018. Both years have been mentioned in previous sections of this paper, but now we will examine the behavior of these two events using the TEM equation (equation 5) and its impact on zonal winds.

Figure 4 represents a set of subfigures, plotted for days in January of 2009 when the sudden stratospheric warming event happened thus showing the time evolution of the event in an altitude-latitude diagram. Before the SSW occurred (Fig. 4a, Fig. 4b), the winds in mid-latitudes and above the north pole are strictly westerly, which enables strong upward wave propagation from the

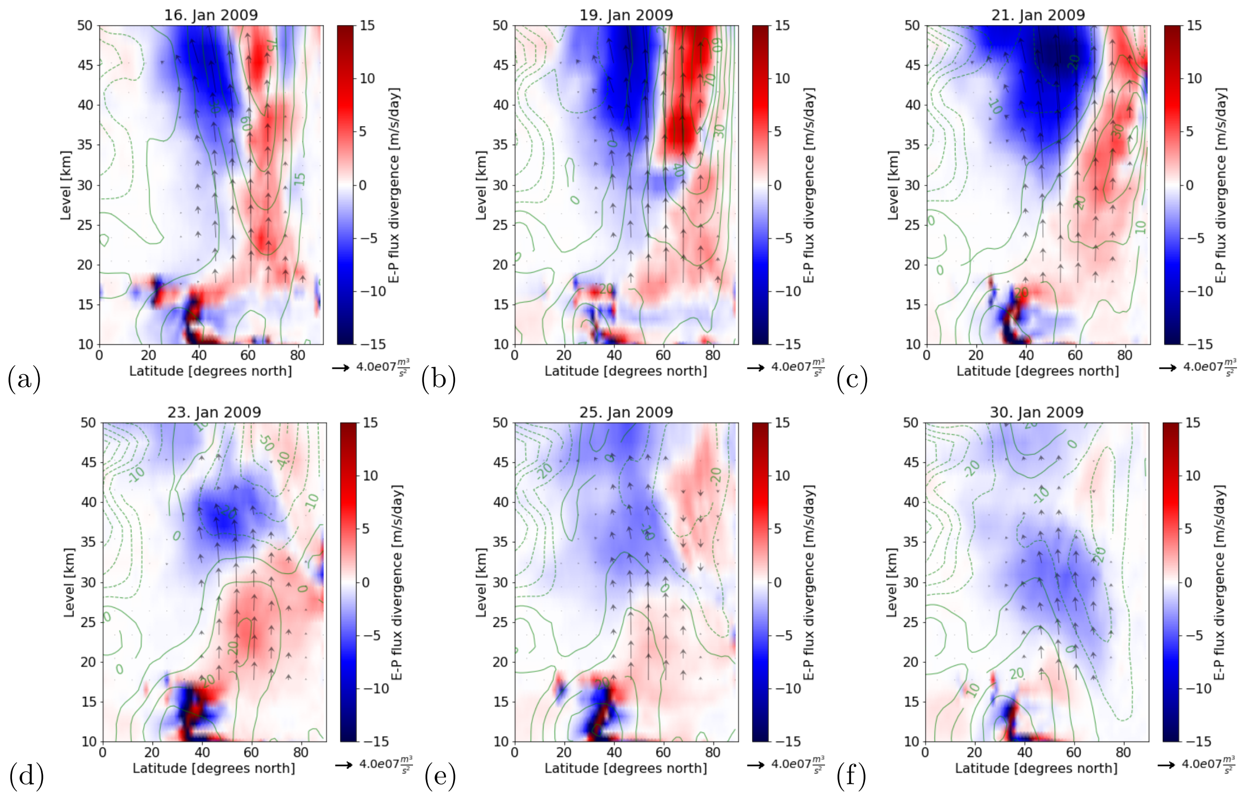

Figure 4. The behaviour of the same quantities as in Figure 3 for the SSW event that occurred in January of 2009.

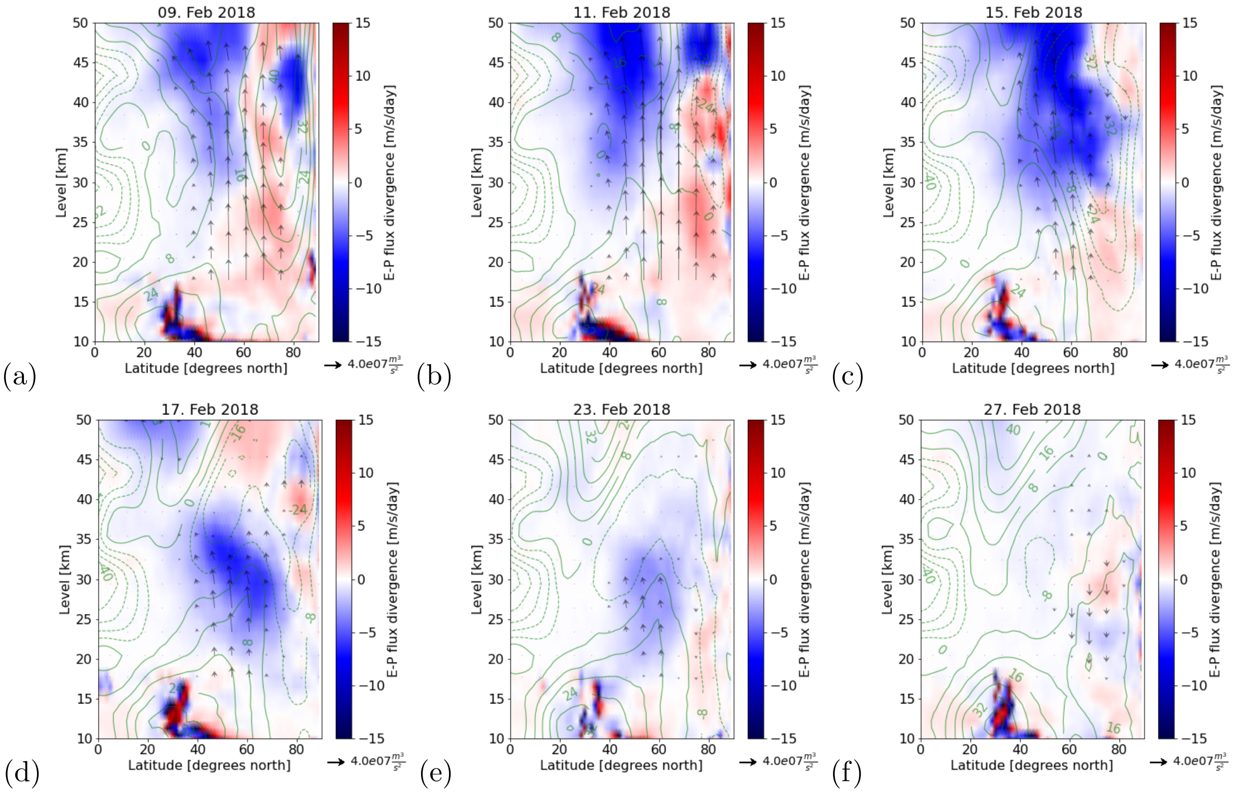

Figure 5. The same as Figure 4, only for the SSW event that occurred in February of 2018.

troposphere (characterized with arrows). Around January 19 and January 21 (Fig. 4b, Fig. 4c), the early stage of the SSW is visible in the upper stratosphere, where the winds reverse to easterly in mid-latitudes, while the winds above the polar region remain westerly at that point. Wind reversal happens as a consequence of strongly negative values of E-P flux divergence (marked with blue colors) that decelerate the westerly flow. The twenty-third of January (Fig. 4d) shows the SSW event in the upper half of the stratosphere. Since the winds there are now easterly, the waves can not propagate above the region, affected by the SSW event. Therefore, the upward-propagating waves break on the edge of that region, resulting in negative values of E-P flux divergence, which decelerate the westerly winds there and this gives the illusion of downward-propagating zonal wind anomalies which is seen in the following days (Fig. 4e, Fig. 4f).

Figure 5 shows another time evolution of another typical SSW event behavior in the northern hemisphere that happened in February of 2018. This event shares the same characteristics as the one that happened in the year 2009. On the ninth of February of 2018 (Fig. 5a), there are convergence areas of wave forcing in the upper stratosphere which two days later result in wind reversal in middle to upper stratosphere above the polar region (Fig. 5b). On February fifteenth (Fig. 5c) another large region of convergence is seen that enhances the intensity of the sudden stratospheric event. On this day, the zonal wind reversal anomaly reaches the lower part of the stratosphere. During the next days (Fig. 5e, Fig. 5f), the waves are not able to propagate to the upper stratosphere so they break and interact with the flow at lower altitudes.

6. Conclusion

The E-P flux divergence term of the TEM equation is key for understanding the tropospheric influence on the stratosphere. Stratospheric circulation is modified by upward propagating planetary waves, which can trigger the sudden stratospheric warming event which can then influence the troposphere by downward propagation of zonal wind anomalies. Even though the sudden stratospheric warming brings a rapid zonal wind reversal with the synchronized temperature rise above the poles, the timescale of influencing the troposphere can be longer. This means that the sudden stratospheric warming effect on the tropospheric weather could be predicted and included in meteorological weather forecast models, but the coupling mechanism of troposphere and stratosphere is not completely known and understood yet. On the other hand, not all of the sudden stratospheric warming events occur due to the wave forcing. This means that predicting sudden stratospheric warmings only by calculating and predicting the Eliassen-Palm flux is not satisfactory.

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