Sudden Stratospheric Warmings and their Effect on the General Circulation of the Atmosphere

A case study of 2013

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Abstract

The general circulation and the motion of air in the atmosphere can be influenced by many reasons and mechanisms and an important one is derived from the propagation of planetary waves from the lower levels of the troposphere into the stratospheric levels. The zonal forcing applied on the circulation pattern over the northern hemisphere, expressed by the momentum and heat fluxes, can be highly affected leading to a possible breakup of the polar vortex or a distortion of it. This phenomenon sometimes occurs during the winter months in the northern hemisphere and is called a Sudden Stratospheric Warming (SSW) event and can be observed in the upper parts of the stratosphere. SSWs are often accompanied with tropospheric blocking events which are defined on mid-tropospheric levels and they can be identified by an anti-cyclonic flow in the northern hemisphere persisting for a long period of time during the winter months. The analysis and study of these two phenomena requires the understanding of the Eliassen-Palm (E-P) theory because the equations characterizing the E-P field constitute a clear representation of the eddy fluxes, moreover the applied forcing on the general circulation is included in the divergence of the E-P field. In the winter of 2013 there was a SSW event observed which did lead to a breakup of the polar vortex and this is the main event considered in this study. Using ECMWF re-analysis data from 1st of November 2012 until 31st of March 2013 on 25 different pressure levels for the northern hemisphere and across all longitudes, the E-P theory will be put in practice trying to identify the effects from the propagation of the planetary waves into the stratosphere and connect this event with any tropospheric blockings found prior or after the warming event.

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Summary

Almost every year during the winter months and especially in the northern hemisphere a large scale phenomenon called Sudden Stratospheric Warming (SSW) is occurring. This phenomenon usually lasts a few days to one month and affects the polar vortex by either distorting it or breaking it up into two smaller vortices. Moreover it affects the circulation of the polar vortex by reversing it. In order for these events to occur there must be an active forcing on the general circulation which will eventually lead to a SSW event. This forcing is expressed by the propagation of planetary waves from the troposphere into the stratosphere through the eddy fluxes, more specifically the momentum and heat flux. This is why the SSW events constitute a clear indication of the existence of a dynamic coupling between the troposphere and stratosphere. The way these fluxes affect the circulation of the polar vortex is done with the study of the evolution of the Eliassen-Palm vector field and its divergence. There are different kinds (major, minor, final, Canadian) of SSWs with different attributes and therefore there is a standard way of identifying and classifying them. Tropospheric blockings (TB) are other phenomena that usually develop right before or after the SSW events but more often and therefore their connection cannot be neglected. Higher pressure systems (anticyclones in the northern hemisphere) usually persist for a specific period of time leading regions beneath them to experience certain weather systems for longer than normal.

The purpose of this study is to examine all these attributes of the warming event and put the theory into practice by applying it in a real case. More importantly we try to identify the tropospheric impact of a SSW event by using the combination of them. The results show that the winter of 2012/2013 can be characterized as a normal one with one SSW event and three TB events and during those events the propagation of Rossby waves showed a normal distribution as long as the temporal evolution of the eddy fluxes indicate how they affected the atmosphere and the Eliassen-Palm vector field.

1. Introduction

The evolution of weather systems always appealed to humans and the necessity for further understanding of these phenomena got essential. With the development of technology, mankind acquired higher knowledge for better weather forecast methods which improved people's everyday lives. To reach that point many years had passed and a lot of effort had been put into establishing a mathematical framework on the interactions between atmosphere and environment and between the interactions in the atmosphere itself.

The creation of fronts and the dynamical behavior of large scale weather systems bring forward the importance of studying atmosphere on a synoptic scale and this is the reason why the upper parts of the atmosphere are of great interest beside the lower parts. Apart from weather systems though, many other phenomena can be found occurring at these heights. One of them is the creation and evolution of a sudden warming in the upper part of the stratosphere with many subsequent events. This phenomenon is called a Sudden Stratospheric Warming (SSW) and the study of these events underlies in both satellite data as long as observations of certain variables from the surface of the Earth such as temperature. These events are triggered by the alteration of the way planetary waves propagate into the stratosphere and their magnitude. This change is caused by the enhancement or diminishing in the amplitude of the eddy fluxes that govern the general circulation in the atmosphere. They are not included in the normal stratospheric circulation as it seems that SSW events do not occur every year but in some years there were more than one event observed. The evolution of SSWs can be identified through the study of the Eliassen-Palm flux (hereafter E-P flux) and its divergence while both include the temporal evolution of the eddy fluxes which characterize the vertical propagation of planetary waves into the stratosphere. SSWs are usually dynamically coupled with another phenomenon that is highly important due to its effects on the actual weather we experience on the surface of the Earth and is called a Tropospheric Blocking event (TB). These events occurring in the troposphere are characterized by a persistent blocked area in the atmosphere over a specific region which implies the existence of an anticyclone over that area.

The information included in the previous paragraph gave me the motivation to study more in depth these phenomena and in order to achieve that the next research questions were formulated:

How does the E-P flux evolve during a SSW event in the winter northern hemisphere?

Is the SSW event dynamically coupled with the TB events of 2013?

Throughout the report a clear study of the SSW's will be provided. After this brief introduction, in the second chapter a theoretical background is given to all aspects of this research topic. The third chapter will deal with the implementation of the model in use and how the outcome was produced. The actual analysis will be presented and supported with the mathematical framework. The fourth chapter will be about the presentation of the results and answering the research questions as well as with the graphical solutions whilst the final two chapters will include the conclusions of the report in which all the striking points will be mentioned and the recommendations for further study on the topic together with the limitations of this particular study.

2. Theoretical Background

This chapter will provide the necessary knowledge in order to further understand the analysis of the topic. The theory underlies in the understanding of the basic equations that describe the interactions but also their practical influence on the changes we observe. A vital step towards the analysis of a SSW event is to firstly assess the attributes of the atmospheric compartments that interact. Afterwards the analysis of the terminology used, like the E-P flux and the wave propagation will be explained and their importance will be shown.

2.1 Atmosphere

The atmosphere can be separated in different parts and layers that will help us understand the interactions in its one separately but also the interactions between these parts. The atmosphere is generally divided in four compartments namely the troposphere, the stratosphere, the mesosphere and the thermosphere. A vertical profile of the temperature in the atmosphere can be seen in the Figure 2.1.

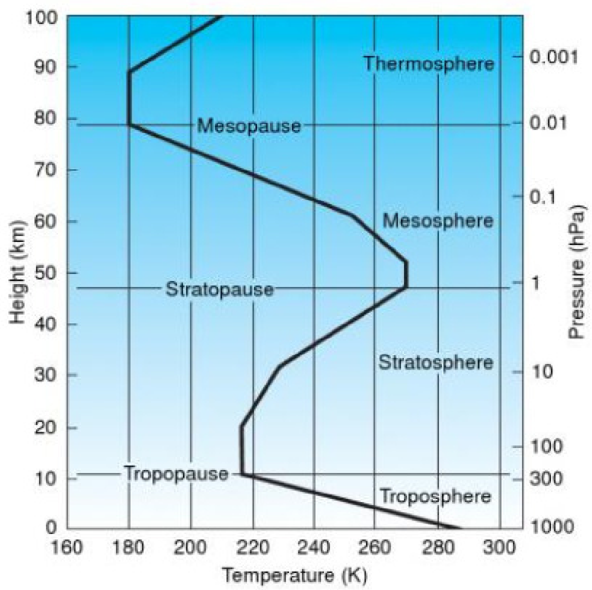

Figure 2.1: Vertical cross-section of the atmosphere with the temperature profile (Wallace and Hobbs, 2006).

Some characteristics of each layer like the temperature, density of air, chemical composition and others differ as the elevation increases. The vertical profile presented in Figure 2.1 shows the height changing distribution of temperature (black thick line) between the different compartments of the atmosphere. In contrast, the air density follows a uniform distribution and decreases with height. This correlates with the drop in pressure from the hydrostatic equation:

$$dp = -\rho * g * dz \tag{2.1}$$

where p is pressure, ρ is the density, g is the acceleration due to gravity and z is the height (Wallace and Hobbs, 2006).

Because the topic of this analysis is concentrated mainly in the stratospheric part of the atmosphere and it is also dynamically coupled with the troposphere no further details on other variables and compartments will be provided here. Nevertheless a full description of the most important layers will be given before actually moving forward to the dynamic analysis.

2.1.1 Troposphere

The lower part of the atmosphere namely the troposphere extends from the surface of the Earth till the height of the tropopause approximately at 10 km. The most crucial attribute of the troposphere is the decrease of temperature as height increases. At the top of the troposphere we find the so-called tropopause, the level at which the temperature starts to increase again. Dynamically speaking, the troposphere is not a stable layer. Due to the fact that the lower parts of the atmosphere heat up from the surface of the Earth, the air is forced to mix in these parts of the troposphere and is transported to the upper parts. In that way the created circulation will mix the troposphere even more to a certain degree and subsequently the tropospheric flow will be forced to break through the tropopause (Mak, 2011). The ascending air will need to be replaced by colder air from the layers that lie above the tropopause. So a circulation pattern will be created which will assist in the exchange of air parcels between troposphere and stratosphere.

2.1.2 Stratosphere

Above the tropopause we find the stratosphere and it extends till the height of approximately 50 km where the stratopause is present. Up to this level the temperature is increasing due to radiation absorption by ozone but a SSW event can lead to further temperature increase of 40 K in a time span of few weeks (Mak, 2011) as we will see later. The stratosphere is characterized by its stable stratification and the general stability that dominates this layer of the atmosphere. The potential temperature θ is a conserved variable for adiabatic motions and increases with height along with the actual temperature for non-adiabatic motions (Haynes, 2005). It is therefore an important aspect of the stratospheric dynamics which strongly affects the E-P flux as we will see later.

2.2 Dynamic Meteorology

An essential part to meteorology is the dynamical part which deals with the motion of air mass in the atmosphere and the formation of different weather systems (Holton and Hakim, 2013). Dynamic meteorology studies the general circulation of the atmosphere in an integrated way and because of the existence of this circulation many phenomena may develop.

2.2.1 Eddy Fluxes

One of the main attributes of the northern hemisphere (hereafter NH) during winter time is the existence of planetary waves or Rossby waves which propagate both vertically and horizontally and are characterized as quasi-stationary. The existence of these waves is due to the dependency of the Coriolis parameter on latitude and they can be considered as departures from the mean zonal flow values before and after the warming event (Holton and Hakim, 2013). Rossby waves can highly affect the circulation of the polar vortex and distract it so areas of high and low pressure may develop as we can see in Figure 2.2. This kind of forcing on the polar vortex is done through the effect of the eddy fluxes, the momentum and heat flux, which alter the available amounts of energy in the polar vortex.

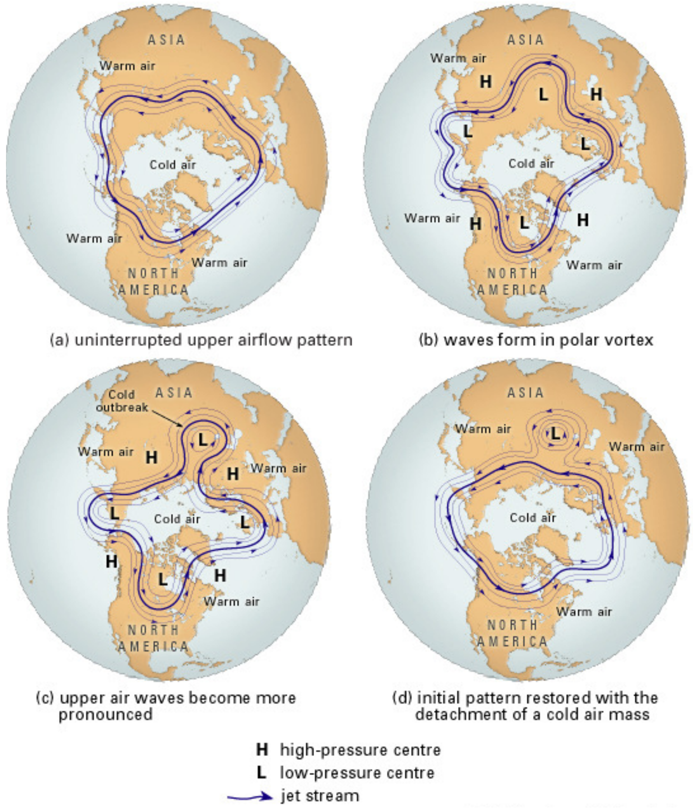

Figure 2.2: Rossby waves on the polar vortex and the creation of high and low pressure systems. Source: Encyclopedia Britannica 2007.

In their composite study of many warming events Charlton and Polvani (2007) found that the heat and momentum flux showed a rather anomalous activity before and after the onset of any SSW event type. These anomalies were amplified even more as the onset of the SSW was approaching and after the warming the observations showed negative anomalies. When the momentum flux is taken into consideration and therefore the meridional component of the E-P flux, the findings clearly stated that before the SSWs negative anomalies were present which lead to a decrease in the zonal westerly flow and the same anomalies were recorded after the SSWs. This pattern was found for both splitting and displacement events with the difference that in vortex displacement events the momentum flux anomalies were smaller.

2.2.2 Mean Meridional Circulation

The mathematical framework (Mak, 2011) contains the primitive equations where a successful description of the mean meridional circulation (MMC) of the atmosphere is based on. According to his analysis the eddy fluxes are an attribute of the vertical propagating waves from the troposphere into the stratosphere and vice versa. This is a fact that enhances the dynamical coupling between stratosphere and troposphere and through the MMC the eddy fluxes will affect the E-P flux under the SSW influence.

The air masses in the atmosphere are in an endless motion creating the general circulation of the atmosphere. Above the equator warm air is rising and transferred poleward until almost 30^{o} N where it sinks and moved towards the equator again aloft the surface of the Earth. This loop is called the Hadley cell and through this loop warm air is transferred at higher latitudes. Above the subtropics another loop is found, called the Ferrel cell, which follows the opposite circulation pattern. Through this cell, colder air from high latitudes is transferred to the subtropics where is sinks and increase its temperature before being transferred towards northern latitudes (Palmén and Newton, 1969). The whole atmospheric circulation can be seen in Figure 2.3 and from this figure we can see the existence of the westerlies in the region between 30° N and 60° N. Above this latitude the polar vortex is formulated and circulates under the influence of the westerly flow.

The existence of the meridional circulation, beside the zonal mean flow and the general circulation of the atmosphere is equally important in the atmosphere and well known since the study of George Hadley in 1735. In this sense the vertical propagation of planetary waves create the MMC, meaning the circulation along a meridian or to the north-south direction. The importance of the MMC was highlighted from the study of Lorentz in 1967 who showed that the eddy fluxes apart from accelerating the zonal mean flow could also trigger the MMC by bringing more energy to higher levels in the atmosphere (Wen and Ronghui, 2002). The tropospheric flow, as part of the MMC, is carrying the appropriate amount of energy needed to initialize a large-scale warming to higher altitudes. With this amount of energy transported, a SSW event may develop and alter the way the polar vortex circulates (Trenberth, 1973).

During the atmospheric circulation the main interaction that we are interested in is the one between the eddies and the mean flow. Therefore it is necessary to use the transformed Eulerian mean (TEM) residual circulation and this is essential also because the Lagrangian motion cannot be directly measured. A clearer view on the effect of the eddy forcing on the whole circulation can be more easily interpreted through the TEM as in that way the combined effect of the eddy fluxes on the atmospheric circulation can be studied (Holton and Hakim, 2013). The meridional circulation can be subdivided to smaller motions in order to complete the bigger picture of it. The rising flow in the tropics will follow the poleward direction and in the high latitudes will begin its descent till it crosses the tropopause again in order to complete the circulation (Cevolani et al., 1990).

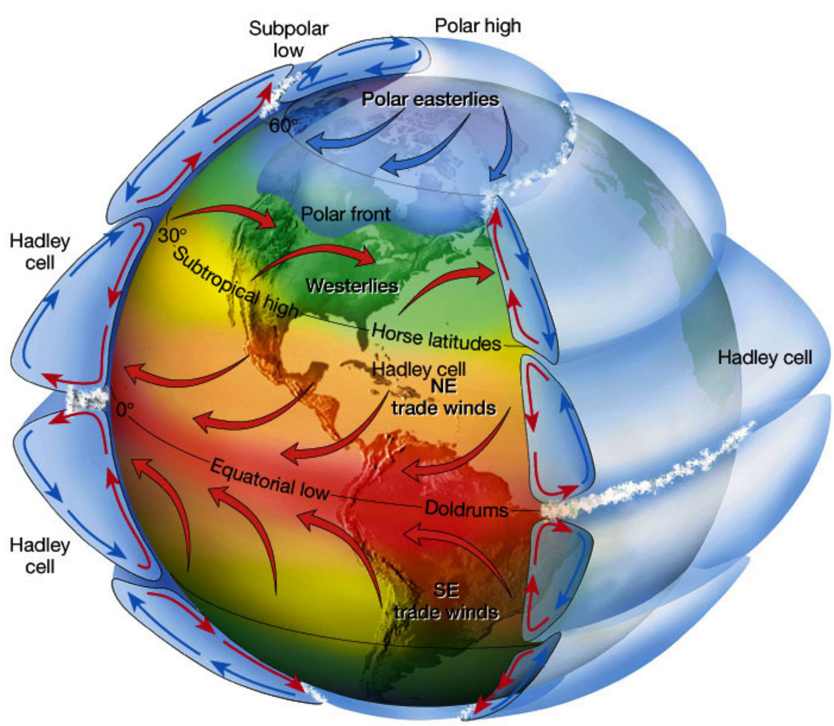

Figure 2.3: General circulation in the atmosphere. Visible are the Hadley and Ferrel cells as long as the wind patterns.

2.2.3 Eliassen Palm Theory

Since the discovery of the SSWs by Scherhag in 1952, it became clear that the coupling between troposphere and stratosphere had to be further studied and the SSWs were a proof for the existence of this coupling. One way to analyze this coupling between different compartments of the atmosphere is to use the E-P flux which can be a profound representation of the wave activity in the troposphere-stratosphere system. The Rossby waves number 1 or 2 affect the structure of the SSW events and according to their amplitude they affect the tropospheric flow which propagates into the stratosphere (Cevolani et al., 1990).

According to the analysis of Mak (2011) using the TEM equations, the derivation of the equations for the E-P vector and its divergence are calculated in spherical-isobaric (ϕ,p) coordinates. These equations are presented below with the first one calculating the E-P vectors and the second one the divergence of the E-P flux. These two equations will be the starting point of the analysis when it comes to study a SSW event. They contain all the variables needed to quantify the influence of the eddy fluxes on the polar vortex and the combination of the effects of each flux can be addressed through the divergence.

$$\hat{\vec{E}} \equiv \left(\widehat{E_1}, \widehat{E_2}\right) = \left(-a[u^*v^*]\cos\varphi, \frac{fa\cos\varphi}{\Theta_p}[v^*\theta^*]\right) \tag{2.2}$$

$$\nabla \cdot \hat{\vec{E}} = \frac{\partial}{\partial\varphi}\left(-[u^*v^*]\cos\varphi\right) + [u^*v^*]\sin\varphi + \frac{\partial}{\partial p}\left(\frac{f}{\Theta_p}[v^*\theta^*]\right) \tag{2.3}$$

In the previous equations, a is the mean radius of the Earth, ϕ is the latitude transformed in radians according to the formula of conversion: $rad = \frac{\varphi\pi}{180}$, f is the Coriolis parameter:

$$f = 2\Omega\sin\varphi \tag{2.4}$$

with Ω the angular speed of rotation of the Earth, $\Theta_p = \frac{\partial\theta}{\partial p}$ is the background part of the potential temperature field, u is the zonal component of the velocity field, v the meridional one and p is the pressure (a full list of the constants can be found on Table 1). The square brackets denote zonal mean values and the asterisk denotes the deviation from the zonal average values of the variables. By first calculating the zonal mean and using the expression $X = [X] + X^*$ one can easily determine the deviation of a specific variable X. Because the warming events occur in the upper part of the stratosphere the eddy fluxes can be considered to be attributes of the propagating waves from the troposphere (Mak, 2011). This implies that we can have a clear view on which latitudes the waves are amplified.

The E-P components $\widehat{E_1}, \widehat{E_2}$ according to the equation 2.2 represent the meridional momentum and heat fluxes respectively. The momentum flux acts as the meridional component of the E-P flux and the heat flux as the vertical one (Peixoto and Oort, 1992). This discrimination is useful when it comes to examine a vertical cross-section of the E-P flux and its divergence. In such a cross-section these vectors act as a clear representation of the upward propagating planetary waves and the divergence of the E-P flux acts as a forcing on the mean-state flow. The momentum flux is included in the first two terms of the equation 2.3 and more in details the first term explores the spatial evolution of the momentum flux across all latitudes. These two terms are derived from the differentiation of the E_1 component across all latitudes. Moreover, the third term represents the heat flux of the E-P flux equation and it describes the spatial evolution of the heat flux across all levels.

At this point it is necessary to point out that the main reason we are interested in the divergence of the E-P flux is the fact that the eddy fluxes included in the calculation will not alter or influence the zonal-mean circulation on their own but only their combination through the divergence of the E-P flux can eventually cause changes to the circulation (Holton and Hakim, 2013). If both momentum and heat fluxes are taken into consideration and combined they can create the zonal forcing necessary to affect the meridional circulation. This zonal forcing is expressed by the divergence of the E-P flux, as presented in the equation 2.3, on the zonal mean winds and therefore to the jetstream (Cevolani et al., 1990). Moreover another connection can be made between the potential vorticity and the divergence of the E-P flux. The transportation of the potential vorticity towards higher latitudes is directly proportional to the divergence of the E-P flux (Edmon Jr et al., 1980). More importantly it needs to be clarified that the contours representing the divergence of the E-P flux express the zones in which the eddies will be affecting the zonal flow. Depending on the fact that the E-P flux is divergent or convergent, the forcing on the zonal flow may lead to an acceleration or deceleration of the zonal winds (Robinson, 1986). Depending on whether the divergence of the E-P flux is positive or negative we can reach the conclusion that there should be a region where enhanced wave activity is present. This conclusion is derived from the fact that planetary waves which are produced in the troposphere and propagate into the stratosphere will eventually be subjects to dissipation. Therefore the existence of a region with positive divergence leads to the conclusion that the wave activity is declining.

2.3 Sudden Stratospheric Warming

As already discussed, the stratosphere is a much more stable compartment of the atmosphere compared to the troposphere. The systems that evolve at these heights may last for a long time before they fade away and also deviations from average situations will take time until they return to normal. The coupling of troposphere and stratosphere can lead to the creation of SSW every one or two years during the winter months in the NH (Holton and Hakim, 2013).

2.3.1 Description

During winter time in the NH, the polar vortex circulates in a counter-clockwise way meaning that the winds are westerly in the stratosphere. This pattern though can be influenced by the propagation of planetary waves from the troposphere. These waves are created mainly due to differences in temperature between land and ocean. Due to the fact that the NH contains a bigger amount of land compared to the southern hemisphere we can conclude that the planetary waves are more enhanced and more abundant on the NH. Bearing that in mind it can be justified why the frequency of SSW events in the NH is larger compared to the southern hemisphere (Mak, 2011).

An important feature of a SSW event is the starting day of the event. This day is called central date and is defined as the first day when the polar vortex changes its circulation and the zonal mean zonal winds become easterly at the 10 hPa level and at 60^{o} N latitude. An event like a SSW can be categorized into two different types namely a vortex displacement and a vortex split. The difference between these two types is visible from the first day that the event occurs and it has to do with the breaking up of the vortex or not (Charlton and Polvani, 2007). The difference in the structure of each type of SSW is clear as can be seen in the Figure 2.4 from previous cases. In Figure 2.4 (a) we can see a vortex displacement event, noted by the distortion of the polar vortex and in Figure 2.4 (b) we can see a vortex split event, noted by the breaking up of the polar vortex into two different smaller vortices. In other words as it will be shown later SSW events cuase the appearance of a positive zonal mean temperature gradient between 60° N and 90° N latitudes at the level of 10 hPa (Krüger et al., 2005).

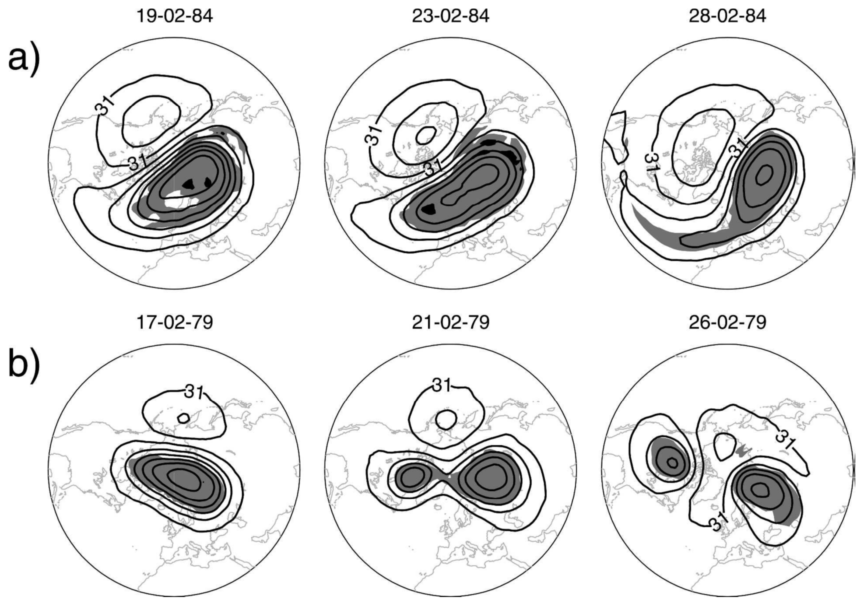

Figure 2.4: Geopotential height (contours) on the 10 hPa pressure surface in a polar stereographic plot. Shaded areas show potential vorticity greater than $4*10^{-6}$ $K*kg^{-1}*m^{2}*s^{-1}$. (a) Vortex displacement warming in 1984 and (b) vortex split warming in 1979 from Charlton and Polvani (2007).

The atmosphere can be characterized to be hydrostatically balanced and therefore the warming of the polar vortex will lead to an increase in the thickness of a column of air between two isobaric levels. In case the warming is sufficient enough, the whole process way conclude to a reversal of the mean meridional pressure gradient which consequently will lead to a change in the polar vortex circulation from cyclonic flow to anti-cyclonic flow (Holton, 1980). Such a situation can be observed when a major SSW event is developed. Apart from major warmings some studies refer also to minor, final and Canadian warmings which have to do with the reversal or not of the zonal mean flow and the time of occurrence. Minor warmings will not reverse the zonal mean flow even though the increase in temperature is still of the same magnitude as in a major warming event. Final warmings have the characteristic that when they occur the reversal of the vortex circulation cannot be restored to its normal westerly flow until the next winter and finally the so-called Canadian warmings occur in the early winter (mainly in November or December) (Charlton and Polvani, 2007).

SSWs can be also categorized according to their wavenumber into vortex displacement or vortex split events. Wavenumber 1 events characterize a vortex displacement event and on the other hand wavenumber 2 events characterize a vortex split event. Generally SSWs are observed during the NH winter time and more specifically in January and February even though some are detected as early as December and some as late as March (Charlton and Polvani, 2007). The fact that SSW events have these characteristics can generalize our view upon them in the way that when a SSW occurs we can observe changes in amplitude of certain variables as temperature and wind, in the timing of occurrence as in early or late winter and in the structure of the SSW as a vortex displacement or vortex split event (Coughlin and Gray, 2009).

During the winter of 2012/2013 a SSW event was detected and as it will be proved later it can be categorized as a major event with its central date on 7th of January 2013. Thus this winter can be characterized, dynamically speaking, as a normal winter for the NH.

2.3.2 Evolution

The observed general circulation in the stratosphere during the NH winter has some basic characteristics which combined with observations of how this circulation is altered before a warming event is the main purpose for the study conducted by Limpasuvan et al. (2004). Their findings gave an insight on what should someone expect when monitoring the evolution of a SSW event. First of all, for an event to occur the polar vortex needs to meet some requirements. When the zonal flow is not placed in its regular position but it seems to be compressed above the pole and therefore displaced, it is easier for the upward propagating planetary waves to penetrate the flow. Just before the warming event is initialized, the observations show a very weak zonal flow from 60^o N towards the equator and the opposite situation towards the North Pole. This is the reason in most of the cases why the polar vortex is displaced and the propagation of the waves can break it down or displace it even more. More generally speaking, the occurrence of a warming event can be highly dependent on the magnitude of the westerly winds in the lower parts of the stratosphere and the amplitude of the vertically propagating planetary waves from the troposphere (Wen and Ronghui, 2002). In that way the energy carried by these waves can be transported to the higher levels of the atmosphere and therefore a SSW event may develop.

As the phenomenon evolves all the changes in the E-P fluxes and the effects on the stratosphere (reversal of the winds, increase in polar temperature gradient) move downward towards the tropopause (Limpasuvan et al., 2004). Through the evolution of the heat and momentum fluxes we can perform the so-called downward control which has to do with the effects of SSWs in the troposphere. The E-P vectors point towards the higher levels of the stratosphere in the extratropical latitudes and as we move upwards they turn southwards (Mak, 2011). More specifically it was shown by Vial et al. (2013) that before the onset of a warming event the higher parts of the stratosphere experience strong westerly winds which were reversed in most of the cases (their analysis was based on the composite of 480 SSW events) and afterwards they slowly recover to the normal flow. During the mature stage of the warmings though the zonal-mean zonal winds propagating downwards slowed down and met the easterly flow that will dominate the troposphere.

2.4 Tropospheric Blockings

Tropospheric blockings can be defined as the reversal of the geopotential height gradient at the 500 hPa level (Vial et al., 2013). They usually precede the SSW events and can be observed some days or a couple of weeks prior to the central date of the warming. The reason why tropospheric blockings are important to the dynamical coupling of the troposphere with the stratosphere, beside their time of occurrence, is that their geographical distribution affects the type of the SSW event. In the study of Martius et al. (2009) this dependency was examined in a 1000 years data simulation using CMIP5 coupled model. It was well proven that tropospheric blockings located in the Atlantic region were usually followed by vortex displacement events and blockings located in the eastern Pacific region or at the Atlantic and Pacific region at the same time were usually followed by vortex split events.

One striking difference though between tropospheric blockings and SSW events, apart from the fact that they are identified in different levels, is that blockings can occur several times per year and they are located in different locations (Woollings et al., 2010). SSW events are identified by zonal averaging all variables and therefore we can never really identify the exact location that they occur. Taking into consideration only this fact, makes it obvious that the coupling between the two phenomena is not an easy procedure. In addition to that, observations of blockings occurring after the mature stage of SSWs make it unclear whether blockings are those who cause the onset of SSWs or it is the other way around. Nevertheless, there is strong evidence that blocking and warming events are coupled because both of them are affected by the propagation of planetary waves between the stratosphere and troposphere (Castanheira and Barriopedro, 2010).

More generally speaking blocking events can be described as those events which force the westerly winds to be blocked by an anticyclonic flow and this will be a persistent phenomenon as it can last for several weeks (Woollings et al., 2010). It is one of the major reasons why certain areas, whose climate highly depends on maritime air flow, experience severe weather for a long time during winter as the high pressure system (anticyclone in the NH) persists and the weather systems stay blocked above the regions for a long time.

In order for an event to be classified as a blocking one, it should meet the criterion that it covers at least 15^{o} of longitude and lasts at least for 5 days before or after the SSW event (Vial et al., 2013). In that way it can be classified as a large-scale event that could alter the upward or downward propagation of planetary waves depending on the time of occurrence. Beside this classification criterion, other studies such as Lejenäs (1983) and Tibaldi and Molteni (1990) suggest a different approach on the blocking event identification procedure. The reason behind the existence of many different ways to identify these events is that smaller patterns of blocking episodes may be observed in the atmosphere but they cannot be characterized as blocking episodes coupled with the SSW events. Therefore by taking into account the possibility that these patterns are present also in this case study we follow the identification method suggested by Barriopedro et al. (2006) which includes three criteria in total. Before setting these criteria though two gradients should be calculated and for certain cases. These gradients, which are northward and southward of the reference latitude of 60^{o} N respectively, are calculated according to the following equations:

$$GHGN = \frac{Z(\lambda,\varphi_N)-Z(\lambda,\varphi_o)}{\varphi_N-\varphi_o} \tag{2.5}$$

$$GHGS = \frac{Z(\lambda,\varphi_o)-Z(\lambda,\varphi_s)}{\varphi_o-\varphi_s} \tag{2.6}$$

The two gradients presented in the equations (2.5) and (2.6) refer to the 500 hPa geopotential height gradient across all longitudes but for certain latitudes and with a small deviation from the original values Barriopedro et al. (2006) used. Because their study was based on a 2.5^{o} latitude by 2.5^{o} longitude grid, the selection of latitudes was different compared to the one used in this research as they were able to use an interval of 0.5 degrees. On the other hand the latitudes used in this study had an interval of 1 degree as the grid under study was 1^{o} latitude by 1^{o} longitude.

Therefore the values used in their study are rounded up to meet our grid resolution. For that reason the value of 77.5^{o}+Δ for ϕ_N had become 78^{o}+Δ but the rest remained the same as in the original study meaning 60^{o}+Δ and 40^{o}+Δ for ϕ_o and ϕ_s respectively. In addition, the deviation Δ is rounded up as well from the original values of -5^{o}, -2.5^{o}, 0^{o}, 2.5^{o} and 5^{o} to the values of -4^{o}, -2^{o}, 0^{o}, 2^{o} and 4^{o} respectively. These will produce five different cases of selection criteria which will provide the identification procedure with more credibility and a better spatial resolution.

After the definition of these gradients the criteria that an event should meet in order to be classified as a blocking one are the following. First of all the difference between the 500 hPa geopotential height at 60^{o} N and the 500 hPa zonal average geopotential height at 60^{o} N should be positive. Secondly the geopotential height gradient GHGN should be less than -10 geopotential meters per latitude and thirdly the geopotential height gradient GHGS should be positive. If these criteria are met for any event for more than 5 consecutive days and persist for at least three degrees in longitude at the same time then the event can be identified as a blocking event.

3. Analysis

This section will provide the relevant information on the data which were used, the reasoning behind this selection and the tools used to manipulate those data. Finally, using the Grid Analysis and Display System (GrADS) the relevant graphs will be produced in order to support the mathematical analysis and visualize the results.

3.1 Data Description

For a complete study on the evolution of the SSW and the E-P flux, the values of certain variables and fluxes were collected and calculated as mentioned before. The data used in the analysis were obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF) database and consisted of the three components of the wind (u-wind, v-wind and w-wind) in m/s, the actual temperature in K, the vorticity and the potential vorticity in PVU[1] and the Geopotential: $\Phi = \mathrm{Z} * g_o$ in $\frac{m^2}{s^2}$. The data consisted of daily average values from the 1st of November 2012 until the 31st of March 2013 meaning 151 time steps. For all these variables, 25 different pressure levels were selected with no stable interval starting from the surface of the Earth (1000 hPa) until the stratopause (1 hPa). The data referred to all longitudes (0^o – 360^o) and for the NH only (Eq. – 90^o N) as the case under study has evolved in the NH. The main interest is on the u and v component of the wind, the potential vorticity and the temperature from which the potential temperature can be calculated because these variables are included in the calculation of the fluxes. Therefore some of the other variables will be used only for validation reasons as we will see in the next section and studying the tropospheric blocking events using the geopotential height data.

According to a satellite observation, in the winter of 2012/2013 a major SSW event over the NH was observed[2]. The breakdown of the polar vortex can be easily recognized when we look at the distribution of the Ertel's potential vorticity in a stereographic view of the NH in Figure 3.1. Due to the time of occurrence of the warming event, the data chosen to study this event had this specific spatial and temporal distribution as presented before. The breakdown of the polar vortex can be seen in Figure 3.1 which is a clear evidence of the warming event.

[1] 1 PVU = 10^{-6}K kg^{-1} m^2 s^{-1}

[2] http://gmao.gsfc.nasa.gov/researchhighlights/SSW/

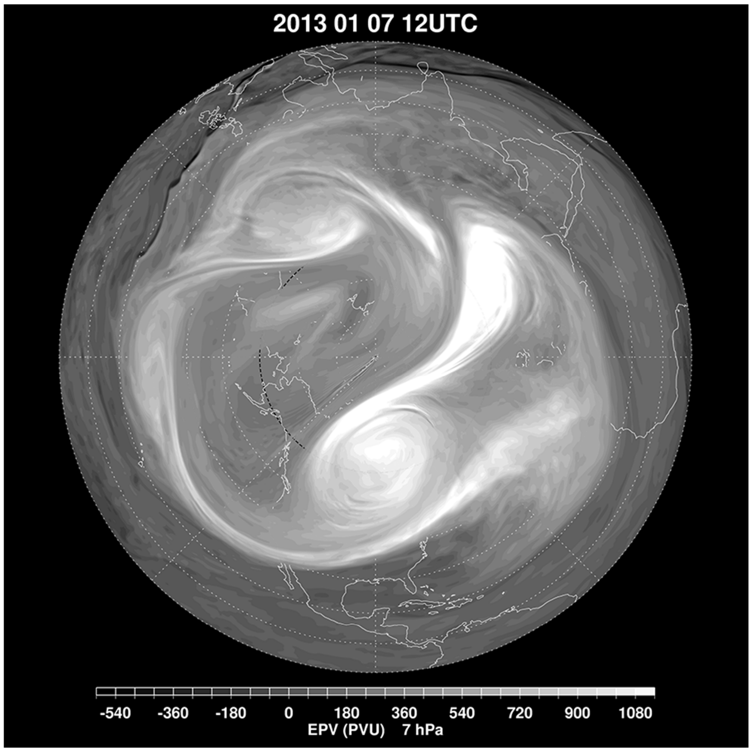

Figure 3.1: Distribution of EPV at 7 hPa on January 7, 2013, central date of the SSW event. The EPV field is from the GEOS-5 data assimilation system, and shows EPV in potential vorticity units (PVU, 10^{-6}K kg^{-1} m^{2} s^{-1}). Source: NASA.

A full description of the available files, data and the derived files used for each script used can be found in the Appendix I, Ia, Ib, Ic and Id. In these Appendices all the available means of analysis are documented and the way these files were manipulated.

3.2 Methodology

The steps followed on the analysis of the case 2013 will be in logical sequence and the first point that needed to be addressed will be the classification of the SSW event. By using the horizontal component of the wind we can identify by its sign when the wind is becoming easterly (negative values denote the transition to easterly winds) taking into account that the positive values characterize the westerly winds. The first day that the u-wind is negative is the central date of the warming event and until the recovery of the wind to positive values again the warming is developing. The normal state of the stratospheric flow is restored when the wind turns back westerly and therefore it reaches positive values.

Secondly a cross section of the E-P flux and its divergence will assess towards the identification of how the forcing is applied on the stratospheric flow and how the E-P flux evolves in the stratosphere, towards which direction and at which levels. With this cross section it can be checked not only the evolution of the fluxes before the onset of the warming but also after the mature stage of the event. Hence it could be easier to assess the tropospheric impact of the SSW and connected to the tropospheric blocking event.

In addition tropospheric blocking events present before or after the warming event will be addressed through the geopotential height and its anomalies. This method will produce an additional reasoning on whether the SSW should be classified as a vortex displacement or split event. Because warming and blocking events occur at different periods of time and levels we need to address their analysis accordingly and assess their overall impact at the end.

The connection between TB events and the SSW will be done by the use of the temperature profiles at different levels and the identification of possible areas where the temperature is changing according to these phenomena.

3.3 Re-analysis

Before proceeding to the analysis a validation of the files is needed. Therefore for an indication on whether the acquired files contain the data in the appropriate form a very simple checking will be performed. By plotting the vertical profile of the temperature we will be able to assess the validity of the included data. Moreover, it will be shown whether or not an increase in the polar temperature is observed which characterizes a SSW event.

Moving forward to the raw data, some calculations were necessary in order to obtain an insight of how the SSW event and the E-P flux evolved and eventually connected them with the tropospheric blocking event. Those calculations were required because of the form of the E-P flux equation. This is the basic equation on which the whole analysis stood on and therefore a correct manipulation of it, according to our needs, was vital.

The complexity of the equation can be seen in the section 2.2.3 where it was presented. Each variable and flux was calculated separately in order to avoid confusion and miscalculations. Prior to the final calculation of the E-P flux, the potential temperature, θ, needed to be introduced into the scripts. Following the equation of the potential temperature:

$$\theta = T\left(\frac{p_s}{p}\right)^{\frac{R}{c_p}} \tag{2.7}$$

and using the data from the actual temperature and pressure levels, the values of θ were calculated. In equation 2.7 R is the gas constant for dry air, p_s is the standard constant pressure level (1000 hPa), usually the surface of the Earth and c_p is the specific heat of dry air at constant pressure (see Table 1 for more details on the constants). Following this calculation, the zonal mean values of the potential temperature were calculated and for the heat and momentum fluxes the zonal mean values of the wind components were computed.

First of all the calculations of the zonal mean values of both wind components and their deviations were performed. These calculations were made using all longitudes, latitudes, levels and timesteps. By acquiring these values, the temporal evolution of the wind components could be plotted and therefore find the central date of the warming event as long as the date where the polar vortex restored its normal circulation.

Secondly with the manipulation of the temperature data, the potential temperature had been calculated and afterwards the zonal mean values and their deviations. These calculations were made once again using all longitudes, latitudes, levels and timesteps.

In contrast with the previous calculations those required to examine the E-P vector field had some restrictions. This was necessary because of the form of the E-P flux equation and the terms included. More in details it is essential to mention that even though for the calculation of the eddy fluxes and their zonal average values all longitudes, latitudes, levels and timesteps could be used, the same does not apply for the calculation of the background potential temperature. For a clear representation of the background potential temperature profile the area average values were computed which were used for the calculation of the zonal mean values. In that way it was assured that the whole area of the NH was included in the calculation including the North Pole as long as the equator. This correction in the calculation was vital because of the integration way when calculating the background potential temperature profile.

Moving on to the vectors of the E-P flux the only restriction made is that the values of these vectors were calculated using the whole domain of interest excluding the top and bottom pressure level.

Table 1: Constants used in the calculations with their symbols and values

Constant	Symbol	Value
Angular speed of rotation of the Earth	Ω	7.292×10^{-5} rad/s
Gravity at sea level	g_o	9.81 m/s^2
Mean radius of the Earth	a	6.37×10^{6} m
Specific heat of dry air at constant pressure	c_p	1004 J/K*kg
Gas constant of dry air	R	287.05 J/K*kg
π	pi	3.1415

For the tropospheric blocking event identification we will use the criteria mentioned in section 2.4 and for that reason the geopotential height data will be used to calculate all gradients for all different cases. The regions in the atmosphere who meet these criteria will be classified as blocked. Finally to check on possible effects on the weather on the surface of the earth, the areas that will be classified as blocked will be examined in terms of the evolution of the geopotential height in order to get an insight on the creation of anticyclonic flows.

4. Results

Following the analysis, in this chapter the results will be presented and commented. One of the most crucial features of this chapter is the connection that will be made between different outcomes and the two different phenomena under study.

4.1 Actual case and stratosphere

Before moving to the results the validation of the files will be done by choosing to plot the vertical profile of the time mean zonal mean temperature. This procedure will provide the appropriate validation of the correct interpretation of the data. The results can be seen in Figure 4.1.

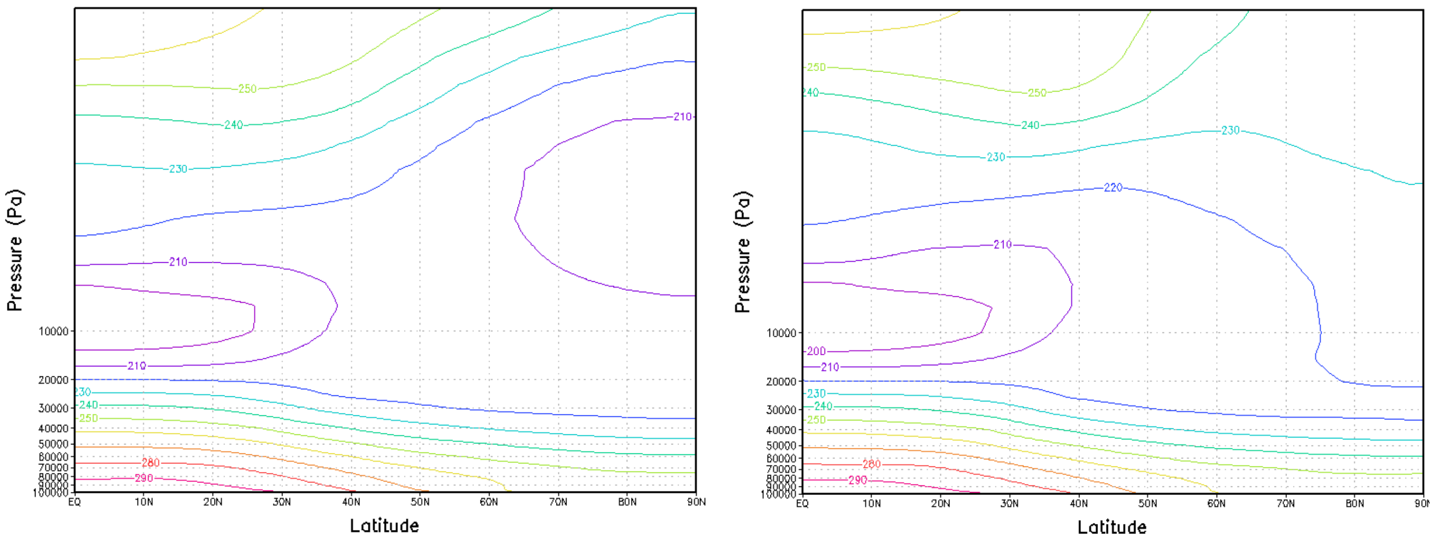

Figure 4.1: Vertical profile of the zonal mean time mean temperature in K (contour lines) during the whole period and during only the SSW event (left and right respectively).

The profile of the temperature seems to follow the normal evolution. Higher temperatures are observed towards the equator and the lower levels close to the surface. As we move northward and to higher levels the temperature is decreasing reaching a minimum value of 200 K at 100 hPa. From that level and above, the temperature increases again clearly indicating the beginning of the stratospheric compartment. The atmosphere is in a normal state and during the SSW event the temperature profile shows an increase in the polar temperature compared to the one observed over the equatorial latitudes. This is something normal according to the theory presented and the characteristics of a warming event.

After this small validation comes the first step of the analysis which was to identify the SSW event and to determine the central date using the horizontal component of the wind. By defining the west to east direction as the positive one it follows that westerly winds will have positive values and the transition to easterly ones will lead to a decrease in the value of the wind component which eventually will become negative. The first day that the value of the wind at 60^{o} N and 10 hPa is negative can be therefore determined as the central date of the warming event according to the definition presented before. The outcome can be seen in Figure 4.2 where the zonal mean zonal wind values are presented. Every point in this figure denotes the daily average zonal mean value which means that the central date is identified as the 7th of January 2013. Consequently it is easy to see that the polar vortex restores its circulation on the 28th of January 2013 when the zonal mean wind becomes again westerly and with positive values.

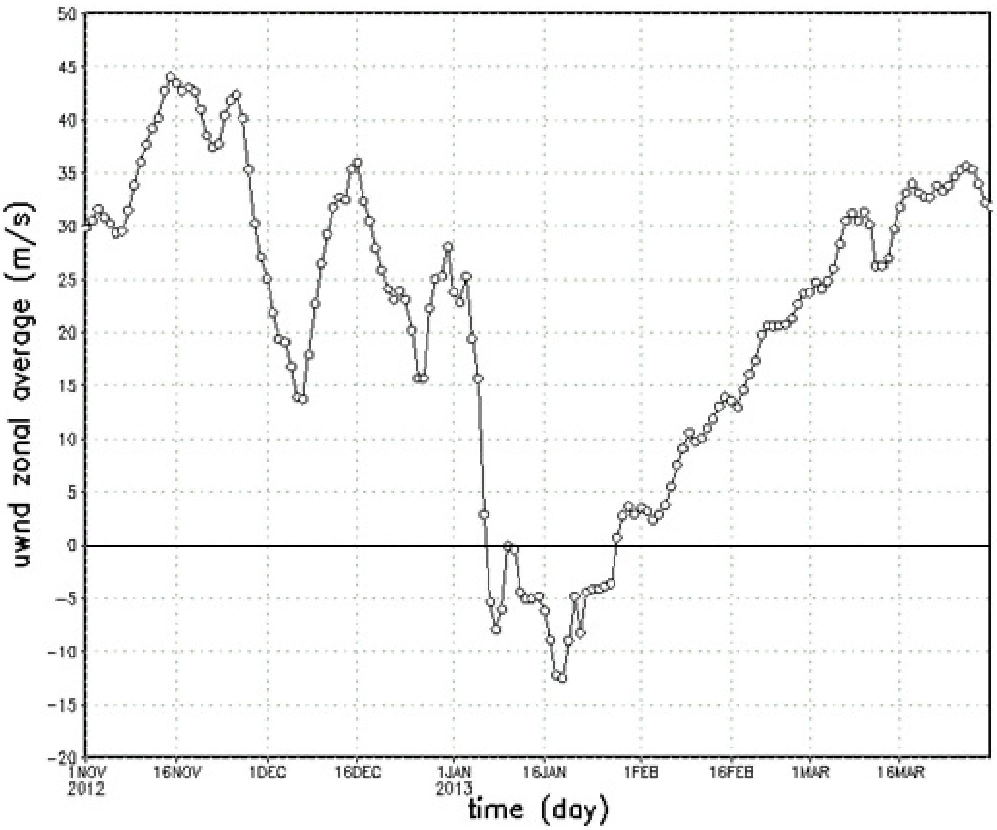

Figure 4.2: Evolution of the zonal mean zonal wind (m/s) at 10 hPa and 60^o N.

A not straightforward but important result is the observation of the temporal evolution of the wind throughout the dates under study. Before the onset of the warming, the horizontal component of the wind fluctuates until the 1st of January 2013 when a major and sudden decrease occurs, indicating the warming event. This pattern is not present after the restoring of the circulation. A continuous increase in the value of the wind is observed up to almost 35 m/s.

Beside the reversal of the zonal wind direction, another main characteristic of a SSW event is the sudden increase in the polar temperature. Therefore one way to look at it is to plot the temperature difference between the central date of the warming event and an arbitrary one before the onset of the warming. This can be seen in Figure 4.3 where it is presented and for a clearer view the y-axis is in a logarithmic scale. In the stratospheric levels and over the polar region the temperature increased almost 45 K between the 7th of January 2012 and the final day of the warming. This increase can be characterized also as a rather sudden one as it occurs in a time span of 20 days.

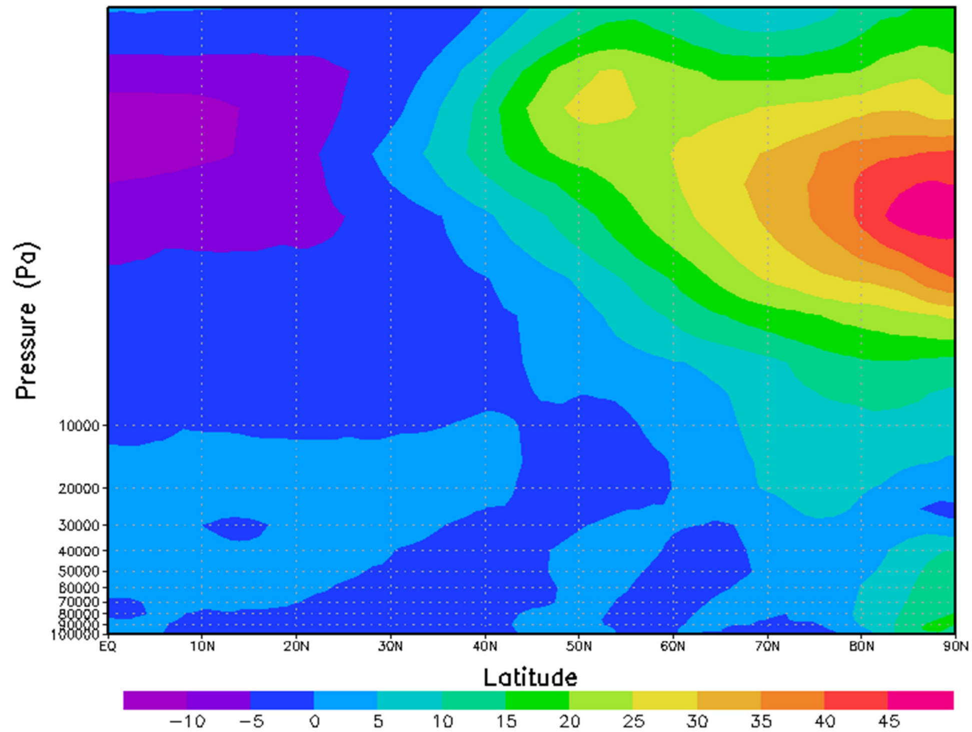

Figure 4.3: Temperature difference (K) between 7th of January 2013 and 16th of December 2012.

After the identification of the event the next step is to classify the event as either a vortex displacement or a vortex split event. This can be achieved by plotting the temporal profile of the potential vorticity in a polar stereographic plot or by plotting the temporal distribution of the geopotential height at 10 hPa again in a stereographic plot. Both procedures should lead to the same outcome and to a vortex split event as seen from the satellite observation in the section 3.1. By comparing the temporal evolution of the geopotential height with the temporal evolution of the potential vorticity we can confirm the original observations. The polar vortex breaks up into three smaller vortices as can be seen in Figure 4.4 and Figure 4.5. In these figures three dates have been selected to be plotted, one before the central date, the central date and one after the central date of the warming event respectively. Therefore it is fair to classify the warming event as a vortex split event.

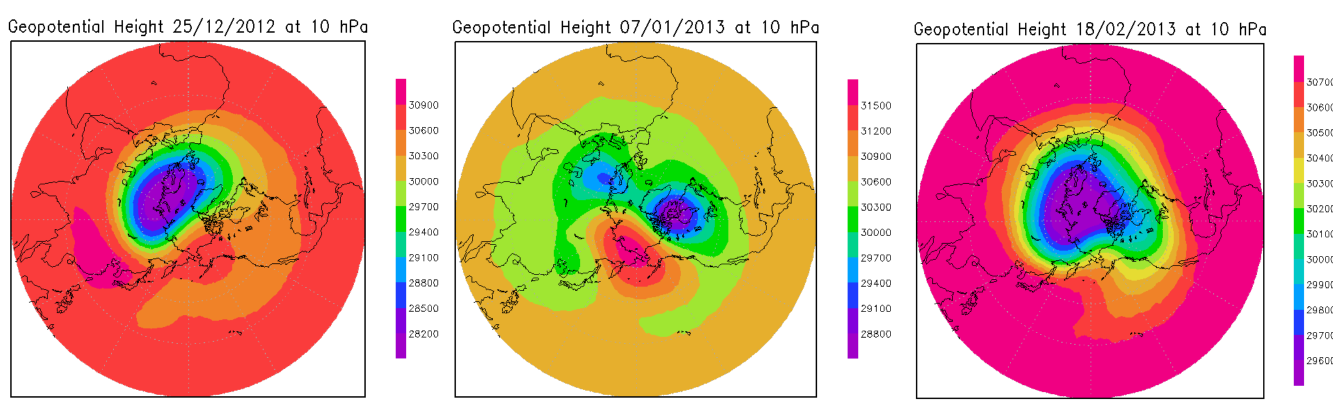

Figure 4.4: Temporal Evolution of geopotential height at 10 hPa. From left to right the dates represent the state before the central day, the central day and after the central day of the warming.

The temporal evolution of both the geopotential height and potential vorticity are in agreement with the satellite observations and with each other. In both sequences of events we can see the polar vortex to break up and later to restore to its circulation. The plots of the temporal evolution of the geopotential height show the three vortices during the central date of the warming. When the polar vortex is split into the three smaller vortices we observe lower values of geopotential height for the two vortices and higher values for one of the vortices compared to the situation before the central date.

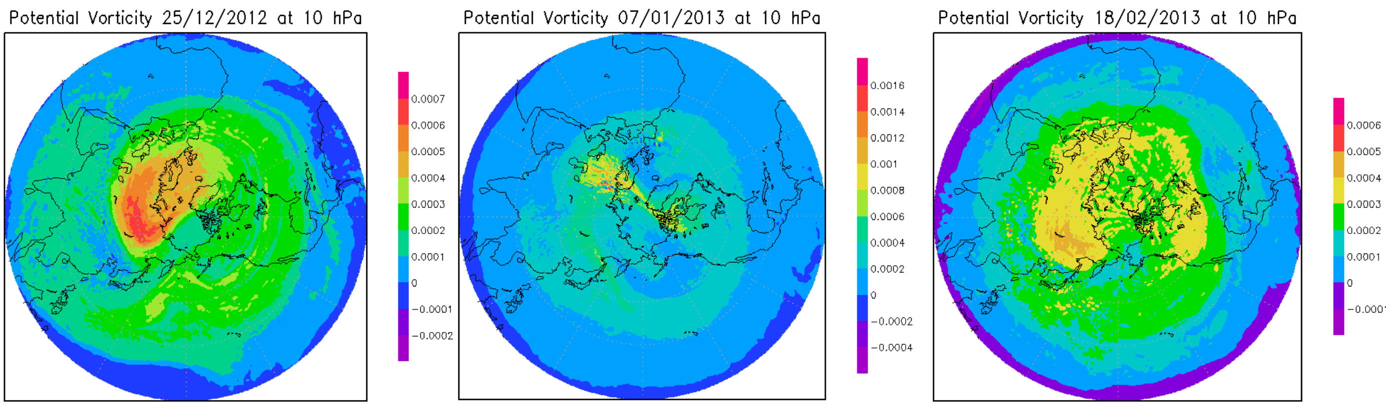

Figure 4.5: Temporal Evolution of potential vorticity at 10 hPa in s^{-1}. From left to right the dates represent the state before the central day, the central day and after the central day of the warming.

In addition to the previous figures, the plots of the potential vorticity come to justify our findings. The clear existence of the polar vortex before the onset of the SSW denoted by the higher values of the potential vorticity is disrupted and on the central date two smaller areas of even higher values of potential vorticity appear. These are the small vortices seen through the geopotential height plots which later converge into one vortex with normal values of the potential vorticity as before the onset of the event. One needs to pay attention on the different scaling between the three figures of the temporal evolution of the potential vorticity as during the central date the smaller vortices appear to have values of one order of magnitude larger than the values before the onset of the warming event.

Following the identification of the warming event and the visualization of it, the E-P theory will be applied on the case in order to determine how the eddy fluxes evolve and how do they affect the E-P vector field. The whole study underlies in the equation 2.2 and 2.3 but before actually plotting the vector field and its divergence it would be preferable to study how the two components of this field evolve. In Figure 4.6 the time-mean distribution of the two components is presented and the outcome is multiplied by a factor of 10^6 for a better understanding of the values. The horizontal component E_1 shows a minimum area centred at the level of 200 hPa and at latitude 30^0 N accompanied with a maximum area centred at 300 hPa and 60^0 N whereas the vertical component shows an increasing distribution towards higher levels at the latitude of 50^0 N.

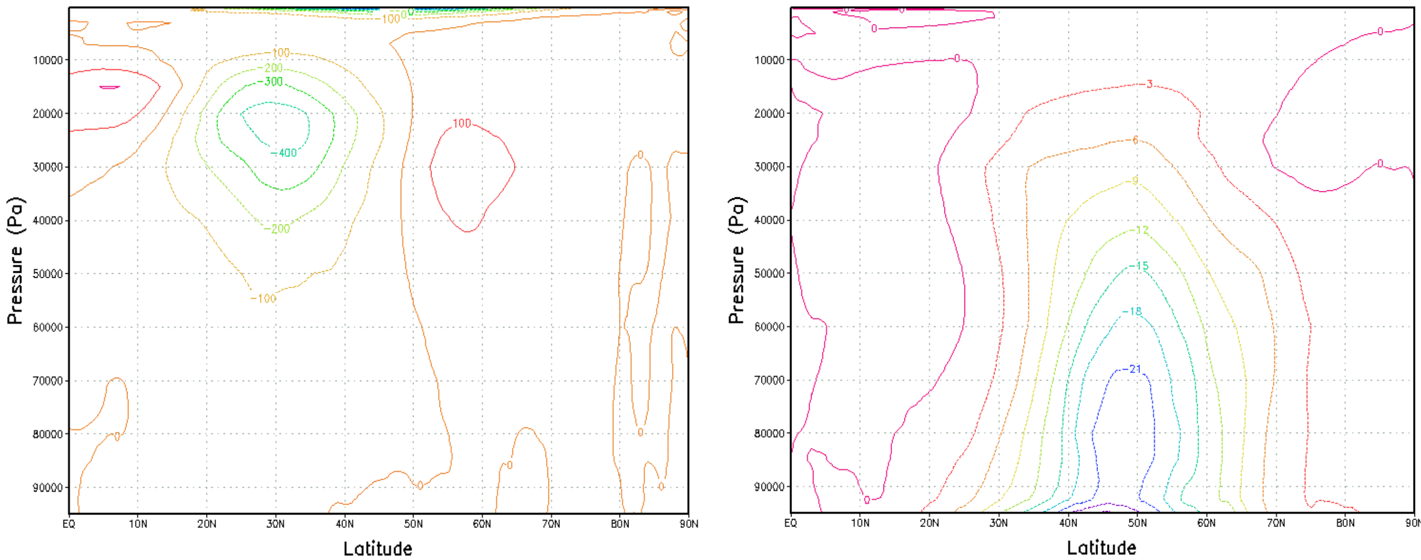

Figure 4.6: Time-mean distribution of the E_1 and E_2 components (left and right respectively) multiplied by 10^6.

The vertical component of the vector field is found to have the larger values at the lower levels and therefore becomes dominant. The negative values can be justified by the selection of the positive direction as mentioned before. On the other hand the horizontal component is dominant at higher levels and we expect to observe an influence on the whole vector field at these levels.

More in details when we take a closer look on the temporal evolution of the E_1 component of the E-P vector field we can observe stronger negative values of this component as the elevation increases before the onset of the warming event. The horizontal component of the vector field is a negative quantity meaning that an increase of the momentum flux will lead to further decrease of the value of the E_1 component. In Figure 4.7 we can see this temporal evolution at 2 hPa, 10 hPa and 20 hPa and in all levels there is clear activity before the warming event around 60^0 N. More specifically and especially at the level of 10 hPa we observe large negative values of the E_1 component before the central date of the SSW event which become positive as the warming event is initialized. This means that the momentum flux will become negative and therefore a northward flux of momentum will be observed.

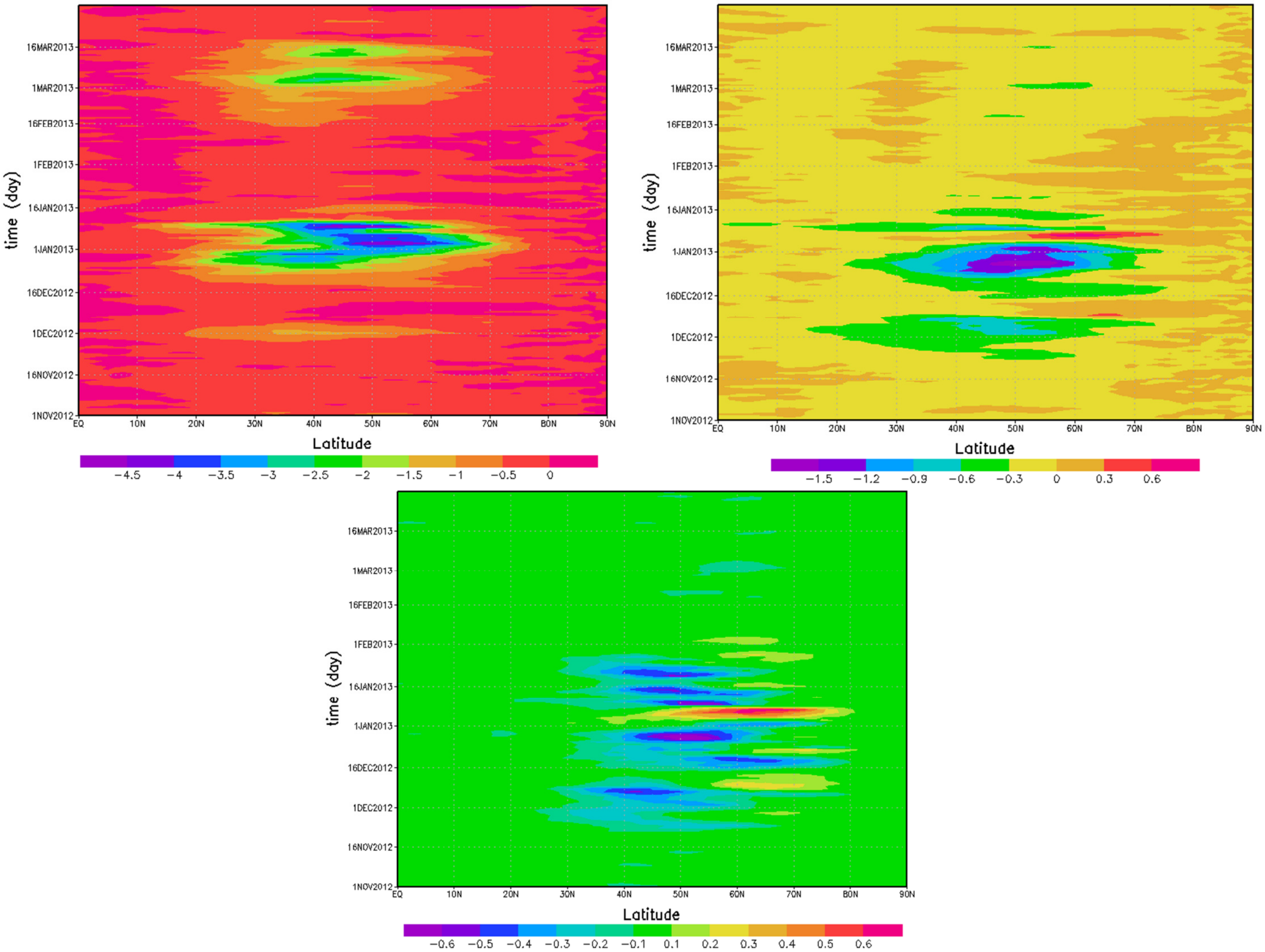

Figure 4.7: Temporal evolution of the E_1 component multiplied by 10^{-9} at 200 Pa, 1000 Pa (upper figures left and right respectively) and 2000 Pa (bottom figure).

When applying the same way of analysis for the E_2 component of the E-P vector field we can observe the exact opposite result. Therefore there is an increase in the value of the component as we move towards higher levels in the atmosphere. This can be seen in Figure 4.8. The minimum value around the central date of the SSW event is found to occur around 60^o N in all three different levels. Even though the difference between the values of the top level and the bottom one chosen to be plotted here is not big we can clearly understand the trend of the E_2 component of the vector field. There is no big activity going on before the warming event but in the level of 10 hPa the strong negative values denote an increase in the heat flux propagating towards higher levels, keeping in mind that that areal average of the potential temperature is a negative quantity when moving towards lower pressure levels.

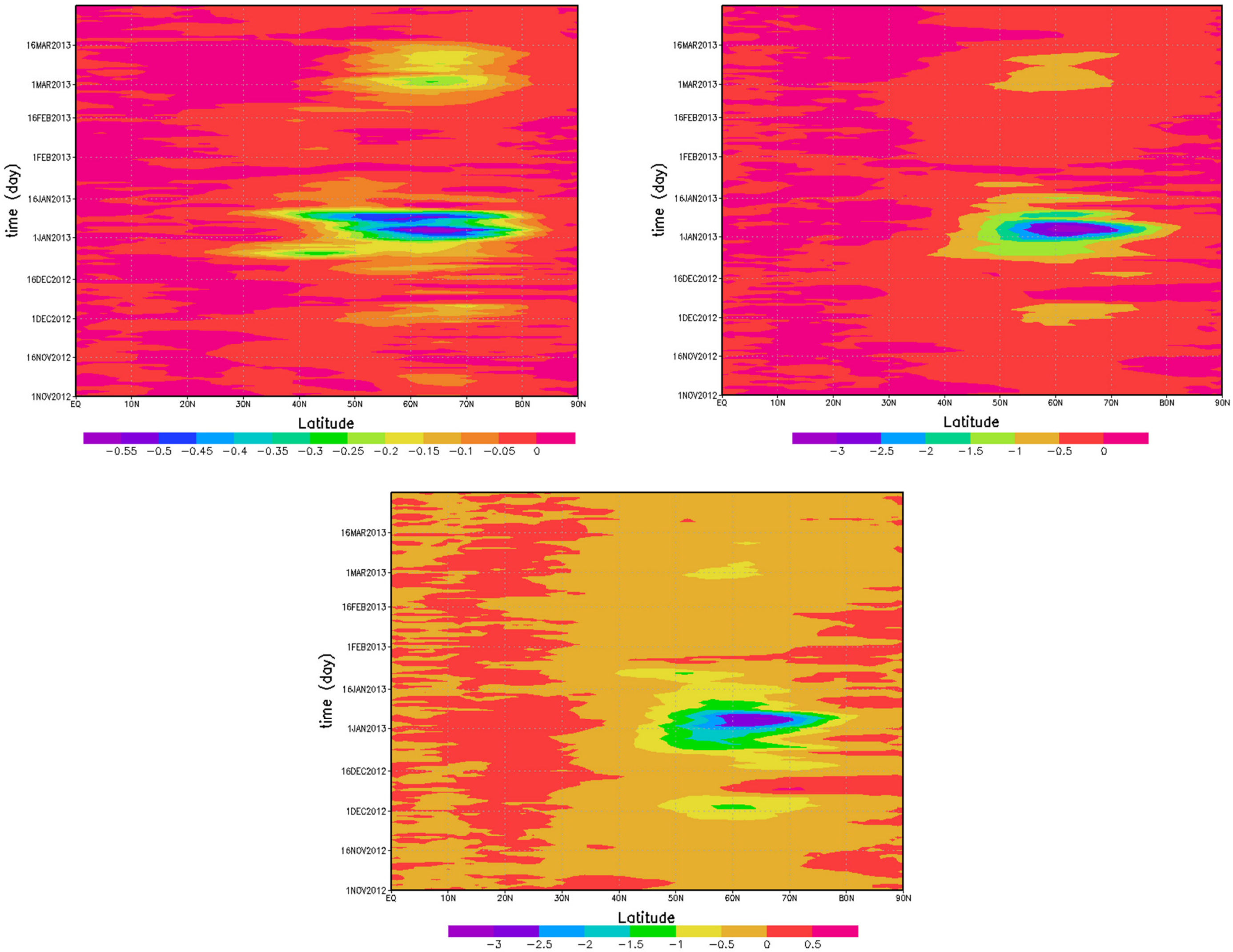

Figure 4.8: Temporal evolution of the E_2 component multiplied by 10^{-6} at 200 Pa, 1000 Pa (upper figures left and right respectively) and 2000 Pa (bottom figure).

Both components of the E-P vector field show a rather uniform distribution as we examine them in lower levels. The higher anomalies are observed a few days before the onset of the warming event and more specifically the horizontal component shows big anomalies at the level of 10 hPa almost 20 days before the central date of the warming event. The same applies for the vertical component of the vector field where we can observe anomalies around the central date of the warming. These anomalies persist after the central date which is a typical attribute of a vortex split event according to the composite analysis performed by Charlton and Polvani (2007).

One of the most important features of the whole E-P theory though is the fluxes included in the calculations of the vectors. These fluxes, namely the momentum and heat flux, have a different form and they can determine the evolution of the E-P field in the atmosphere. The fact that these fluxes are the core of the E-P field clearly shows their importance as in that way one can realize how a SSW event is developed. According to the theory of Mak (2011) the distribution of the climatological zonal mean eddy fluxes should be the same for every case. His analysis was based on climatological annual mean average values. The results in this case study come to almost perfect agreement with the theory. The momentum flux distribution can be seen in Figure 4.9 where a maximum value of more than 70 m^2/s^2 is found at latitude 30° N whereas the minimum of -40 m^2/s^2 is found at 60° N. Both minimum and maximum values are located between levels of 200 and 300 hPa. In this case study the time of interest is only the winter months thus the values of the momentum flux have a broader range compared to the values found by Mak (2011) as the summer months are not taken into consideration. The whole distribution though follows the same pattern and the same characteristics for both eddy fluxes.

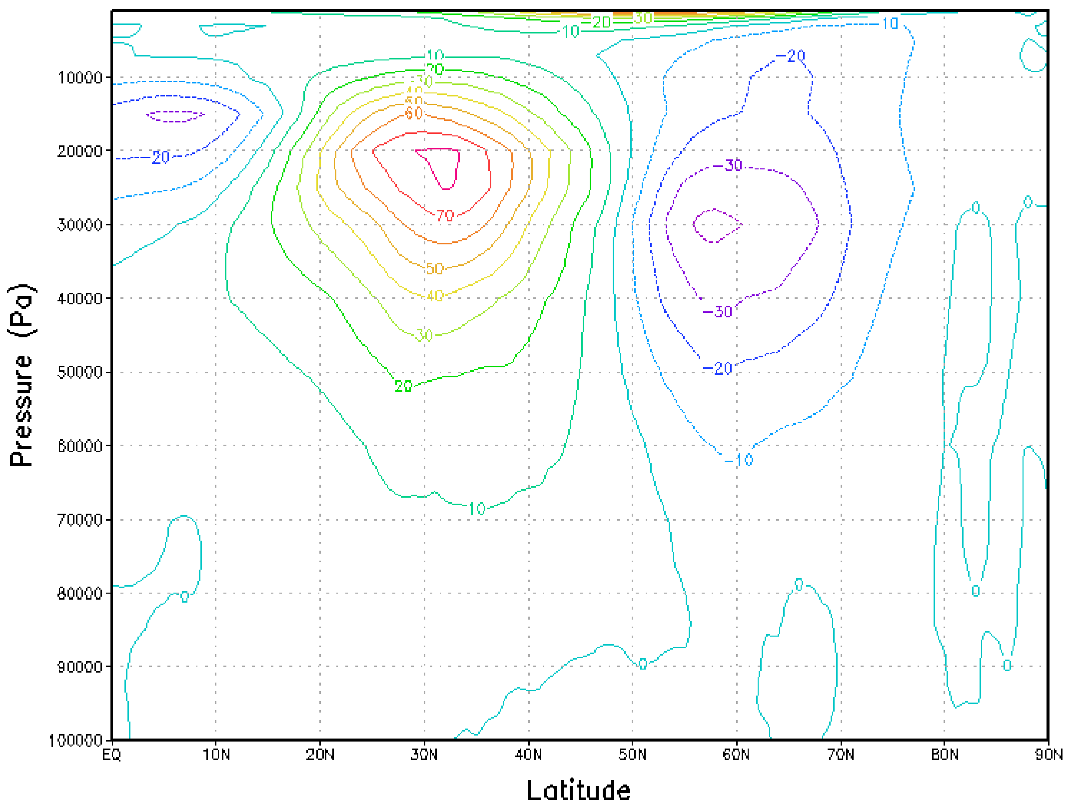

Figure 4.9: Time-mean zonal mean distribution of the momentum flux in m^2/s^2.

Following the same procedure for the distribution of the heat flux we find in Figure 4.10 where we can observe two areas of maximum values. The first maximum is found on the level of 200 hPa and the second one approximately on the level of 800 hPa. Both areas are centred on latitude 50° N which coincides with the literature. It needs to be mentioned though that due to the fact that the potential temperature increases very rapidly when entering the stratosphere, for a correct interpretation of the heat flux the upper part of the troposphere should be cut off. In that way all the high values are excluded and therefore the values and contours of the heat flux in the lower levels are clearer.

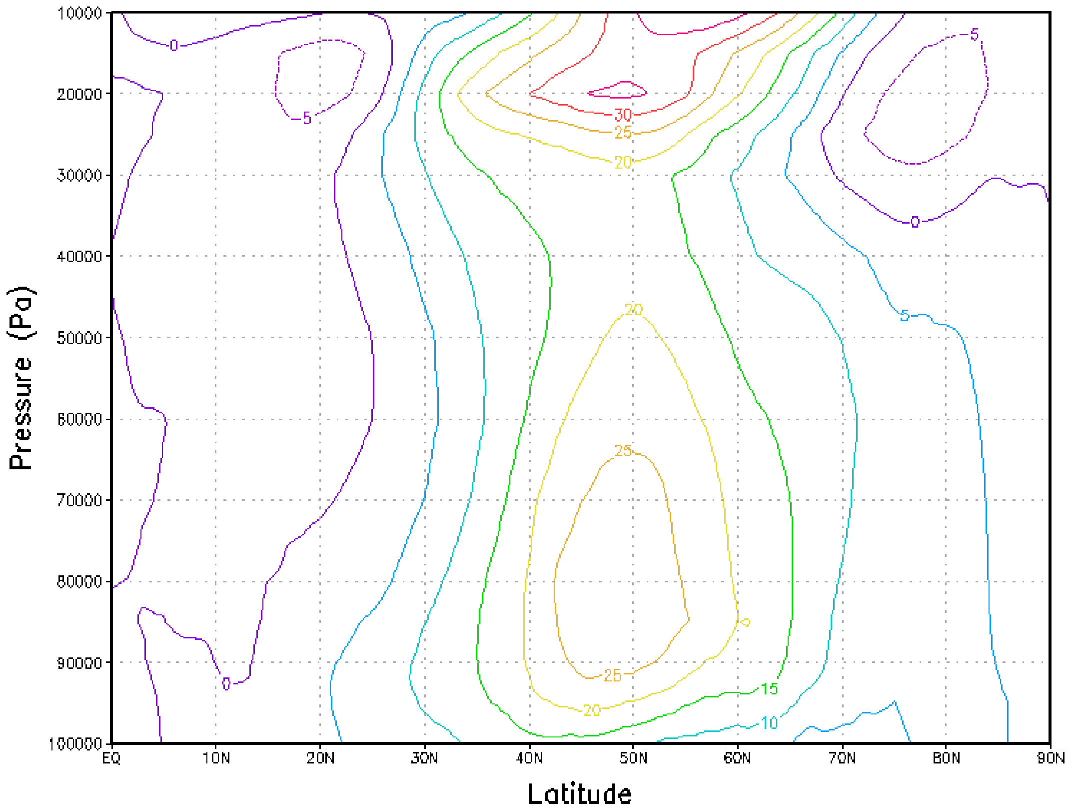

Figure 4.10: Time-mean zonal mean distribution of the heat flux in m*K/s.

The maximum values of the heat flux are approximately 30 m*K/s and they are a bit higher than those found using the climatological annual mean by Charlton and Polvani (2007) and the reason is once again the fact that our case study contains only the winter months in the calculations. Thus the forcing on the polar vortex is found to be enhanced which is the reason why the SSW event occurs during January in the NH for the year 2013.

More generally speaking the two eddy fluxes in combination will affect the E-P vector field through the two components E_1 and E_2. Therefore the distribution of the eddy fluxes will have a clear impact on the vector orientation of the whole E-P field. On the lower levels the heat flux is dominant and as we move towards higher levels the momentum flux is becoming dominant whereas the heat flux is diminishing. The form of the time averaged distribution of the E-P vector field across all levels and all latitudes for the months under study accompanied with its divergence can be seen in Figure 4.11 keeping in mind that negative values of the vertical component are plotted towards higher levels and therefore lower pressure areas according to Mak (2011). Compared to the results of Mak (2011) we can observe that the values of the vector field are large between the zone of 20° N and 75° N and centred around 50° N. Moreover the vector field is decreasing in magnitude with increasing height almost until the pressure level of 200 hPa.

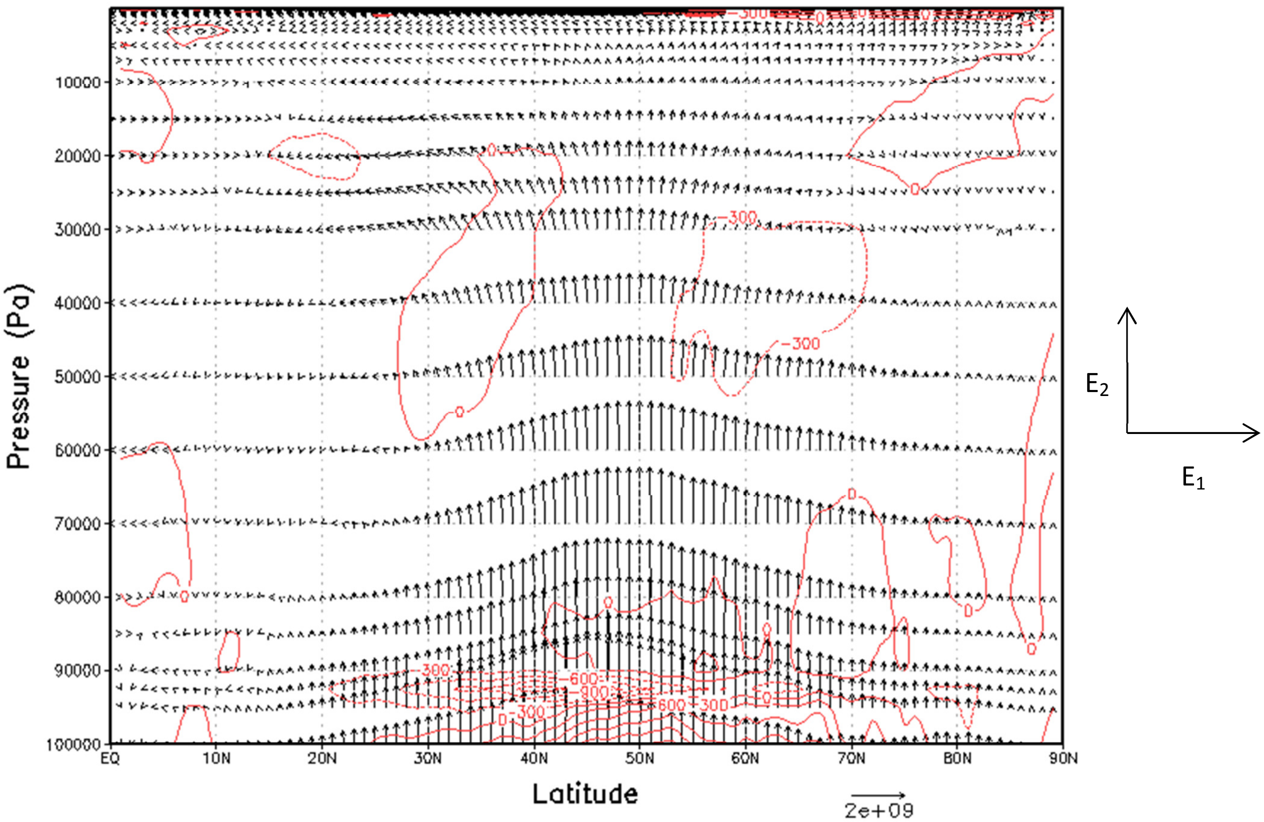

Figure 4.11: Climatological time-mean distribution of the E-P vector field. The vertical component of the E-P field is multiplied by 100. Vertical coordinate is Pressure in Pa and the horizontal is Latitude.

The vectors are pointing towards higher levels in the lower parts of the troposphere as described before due to the domination of the heat flux. Due to the fact that as the elevation increases the momentum flux is becoming more dominant thus the horizontal component of the vector field, we can observe the vectors to start being tilted towards the equator. The orientation of the vectors indicate the propagation of the Rossby waves through the influence of the eddy fluxes to the upper parts of the atmosphere whereas the divergence of the field the total zonal forcing on the polar vortex. Consequently the southward tilting of the vectors at the level of 300 hPa coincides with the maximum values observed on the time-mean distribution of the momentum flux.

The combination of both eddy flux terms contribute to the forcing on the zonal mean flow leading to the conclusion that on higher altitudes the forcing is decreasing. This can be identified through the divergence of the vector field. In Figure 4.11 the red contours represent the divergence of the vector field showing a large convergence zone in the upper parts of the troposphere indicating a deceleration of the zonal winds under the enhanced wave activity. In contrast in the lower parts of the troposphere the field is divergent and the wave activity not that strong.

Beside the time-mean distribution, the temporal evolution of the vector field's divergence can provide with the information about the evolution of the forcing on the polar vortex and what was the situation in the atmosphere before the warming event in different levels. The evolution of the divergence, shown in Figure 4.12, shows how the forcing on the polar vortex evolved and periods when the atmosphere experienced an enhanced forcing are characterized by a convergent vector field. This can be observed when the warming event occurred and especially at the levels of 10 and

20 hPa. That is exactly when the wave activity was enhanced which could initialize the deceleration of the zonal winds and eventually the reversal of the polar vortex circulation.

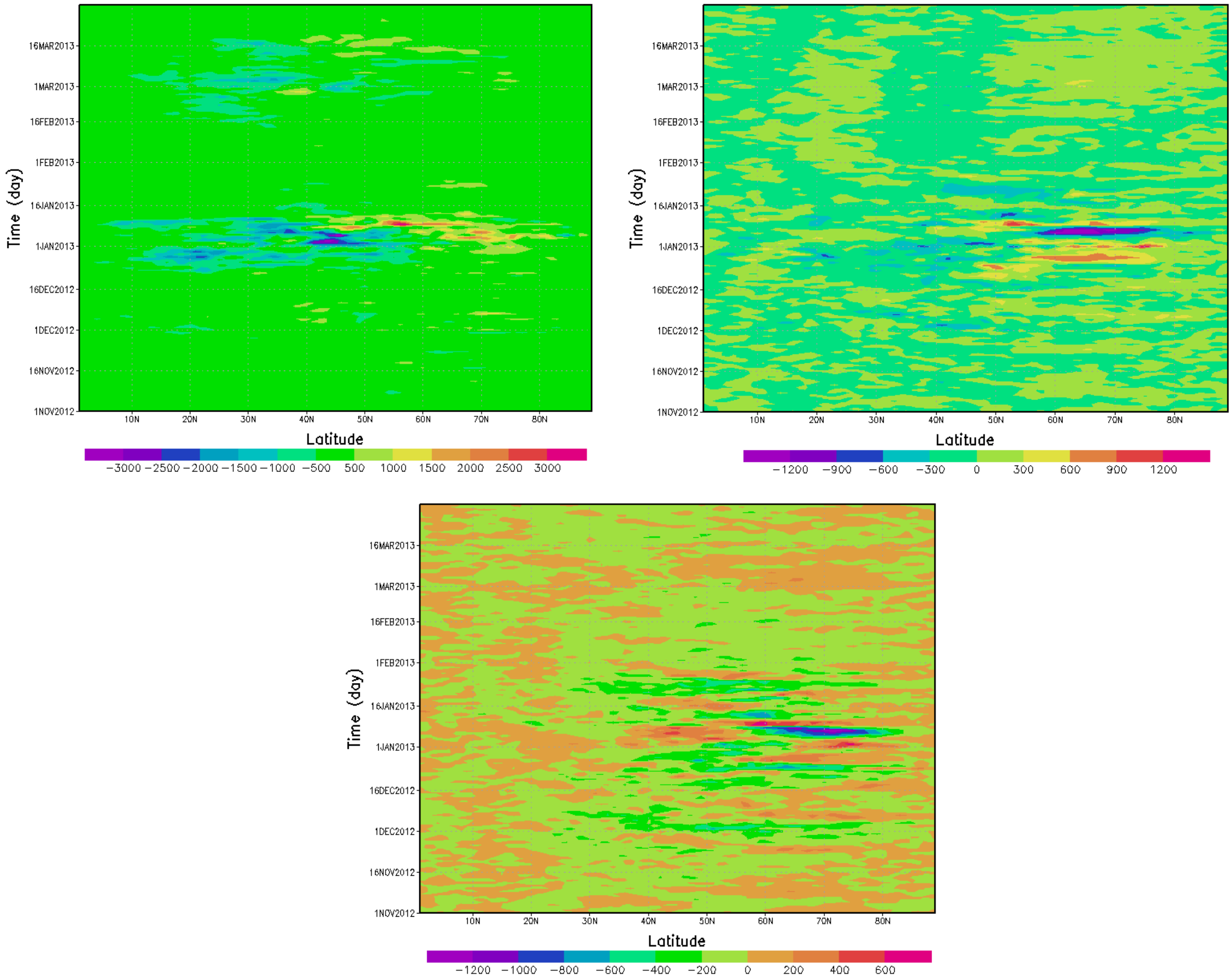

Figure 4.12: Temporal evolution of the divergence of the vector field at 200 Pa, 1000 Pa (upper figures left and right respectively) and 2000 Pa (bottom figure).

Another way to look at the evolution of the SSW event is to study the temporal evolution of the temperature at different levels. An analysis like this will assist on the understanding how the temperature evolved in the atmosphere and how close to the central date of the warming event these changes occurred. At the level of 10 hPa the temporal evolution of the temperature can be seen in Figure 4.13. The increase in temperature over the North Pole is really sudden when looking at the period of time around 7th of January 2013. More importantly we observe another but yet smaller increase in temperature at the beginning of January suggesting that there was an event affecting the value of the temperature at that time.

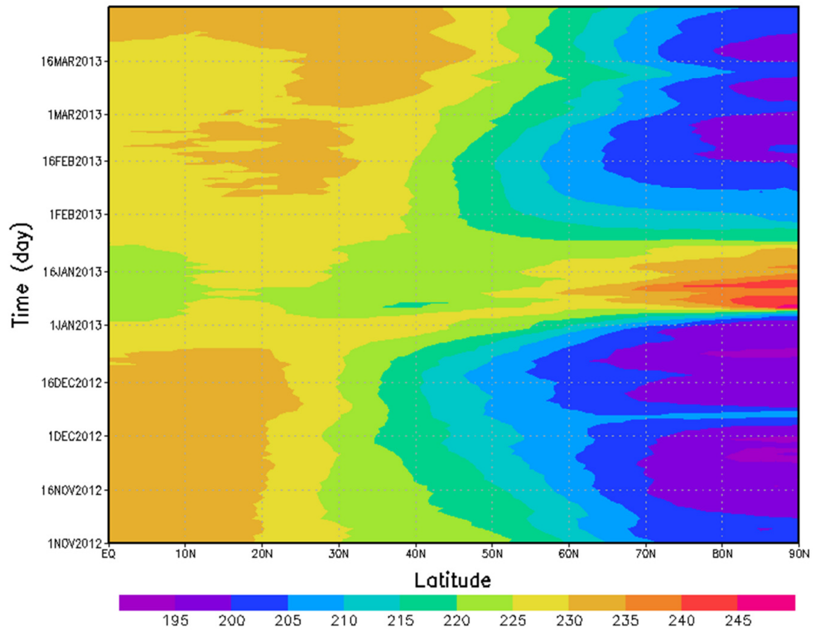

Figure 4.13: Temporal evolution of the zonal averaged temperature (K) at 10 hPa.

Even though the increase in temperature at the upper part of the stratosphere just before the warming event occurs within a period of very few days the same situation doesn't apply when we take a look at a lower level. Figure 4.14 shows the temporal evolution of temperature at 50 hPa. The increase in temperature before the onset of the warming event is less sudden and the same applies for the early December case. At this level the time period when the stratosphere experiences an increase in temperature compared to the normal values is larger. The big difference between these two levels is observed on the time needed for the polar cap to restore its normal temperature value. It seems that after the warming event the stratosphere at the level of 50 hPa needs more time until the temperature decreases again.

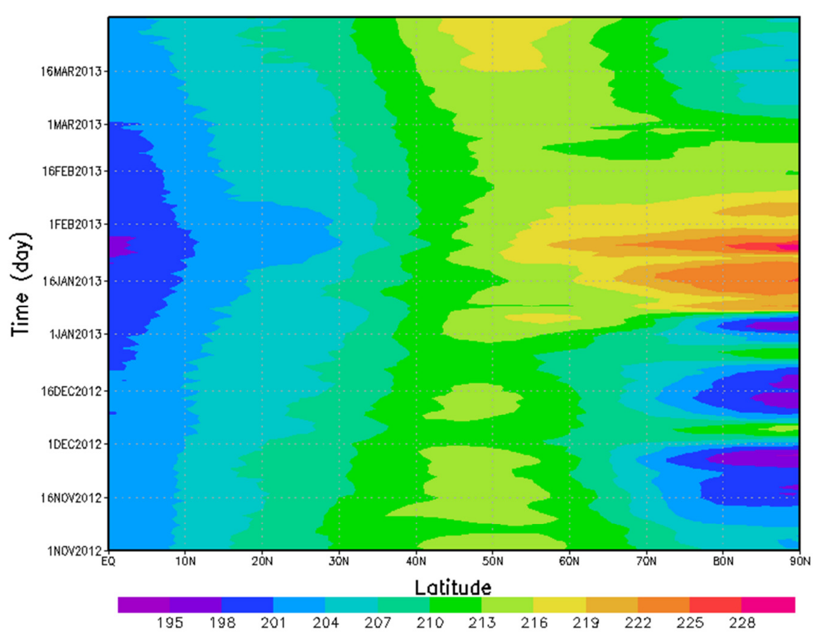

Figure 4.14: Temporal evolution of the zonal averaged temperature (K) at 50 hPa.

4.2 Tropospheric impact and blocking events

The regions that can be characterized as blocked in the troposphere will be those that meet all three criteria presented in section 2.4 and can be seen in Figure 4.15. According to that figure there are three regions that experience a tropospheric blocking event. The characteristics such as duration and location of these regions can be found on Table 2.

Table 2: Characteristics of the blocking events

Blocking event	Given name	Duration	Location
1	TB1	12/12/2012 – 18/12/2012	32^{o} E – 81^{o} E
2	TB2	17/01/2013 – 25/01/2013	9^{o} W – 22^{o} E
3	TB3	18/02/2013 – 06/03/2013	82^{o} W – 28^{o} E

A closer look to these regions can be found in the Appendix II where Figure 4.15 is zoomed in over them. In that way the validity of the results can be proven. The first TB event occurs before the onset of the SSW event, the second TB event occurs during the SSW event at the late phase of it and the third TB event is found after the SSW event. Even though the theory suggested by Martius et al. (2009) indicates that the SSW event of 2013 should be preceded by a TB event over both the Pacific and the Atlantic basin, here in this study the findings do not come in agreement with this hypothesis. The only TB found before the onset of the warming event is over Eurasia. On the other hand the results showed a rather big in terms of areal coverage and longtime lasting TB event after the restoration of the polar vortex circulation over the north Atlantic.

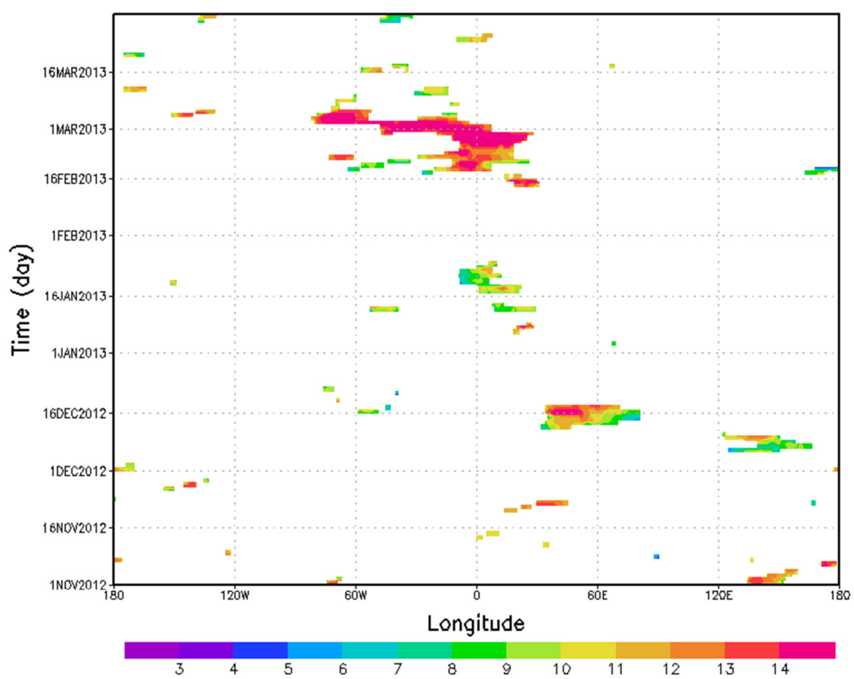

Figure 4.15: Longitudes that meet tropospheric blocking identification criteria.

Furthermore in Figure 4.15 one can also understand the magnitude of the TB event according to the numbers of criteria that each region meets. This can be done by the scoring values each region achieves and this scoring can be of maximum 15 points. This is due to the fact that the criteria that should be met were examined for three different latitudes and five values of Δ were used for all latitudes. In that case a specific region will score one point for each latitude and each value of Δ that the criteria are met for leading to the maximum of 15 points if a specific region meets all criteria for all latitudes and all Δ values. By understanding this point system the color bar of Figure 4.15 makes sense and provides with assistance when it comes to evaluating the magnitude of each TB event.

The areas that can be characterized as blocked have a clear effect on the troposphere and the actual weather on the surface of the Earth. All three TB events show one of their basic characteristics clearly over the areas they occurred. The results can be seen in the next figures which show the geopotential height at 500 hPa level for an arbitrary day during each TB event. More specifically for each TB event we can see in Figure 4.16 the creation of a blocked are over Eurasia at the 500 hPa level which is the TB1 event.

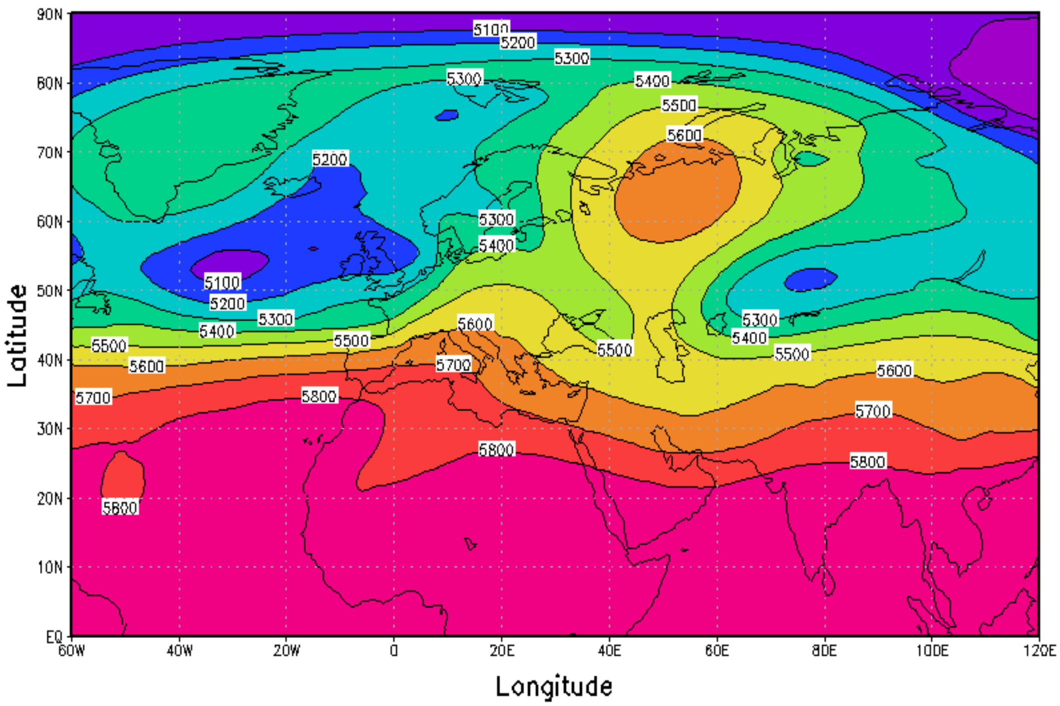

Figure 4.16: Geopotential height at 500 hPa on 15/12/2012.

This region experiences large geopotential height values compared to the normal ones and from the shape of the contours it is clear that the area beneath can be characterized as blocked. The geopotential height values are up to 5800 m whereas the geopotential height of regions at different longitudes and same latitude is approximately 5300 m. These larger values of geopotential height compared to the normal ones are an indication of an anticyclonic flow at this area.

The TB2 event can be seen in the Figure 4.17 and it occurs simultaneously with the SSW event. This TB event in contrast with the TB1 event has a smaller difference between the normal and the observed geopotential values over the region it occurs. The maximum value of 5400 m is centered between Iceland and Norway and is less than the values observed in TB1 event. The same applies though for the regions surrounding the blocked one as the minimum value of the TB2 case in geopotential height is 5000 m.

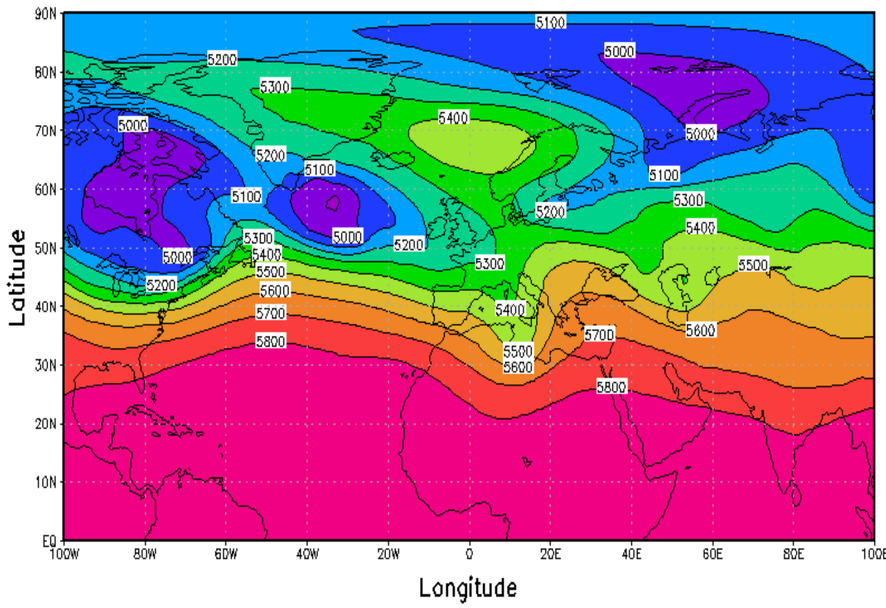

Figure 4.17: Geopotential height at 500 hPa on 21/01/2013.

Once again the anticyclonic flow is blocked between two low pressure systems which are characterized by their low geopotential height values. Because TB2 event scored the less compared to the other two TB events it is logical not to characterize it as an important one and to question its origin or even if the effect on the surface would be significant.

The same does not apply for the last TB event which is the most significant one. TB3 event can be seen in Figure 4.18 where the anticyclonic flow is characterized by geopotential value of 5700 m whereas the normal one is approximately 4800 m. This big difference makes the TB3 event a really important one which persists in the area for the longest compared to the rest and highly affects the weather on the surface. The significance of this TB event can also be understood by the fact that it is located over a highly populated area, the UK and north part of Europe.

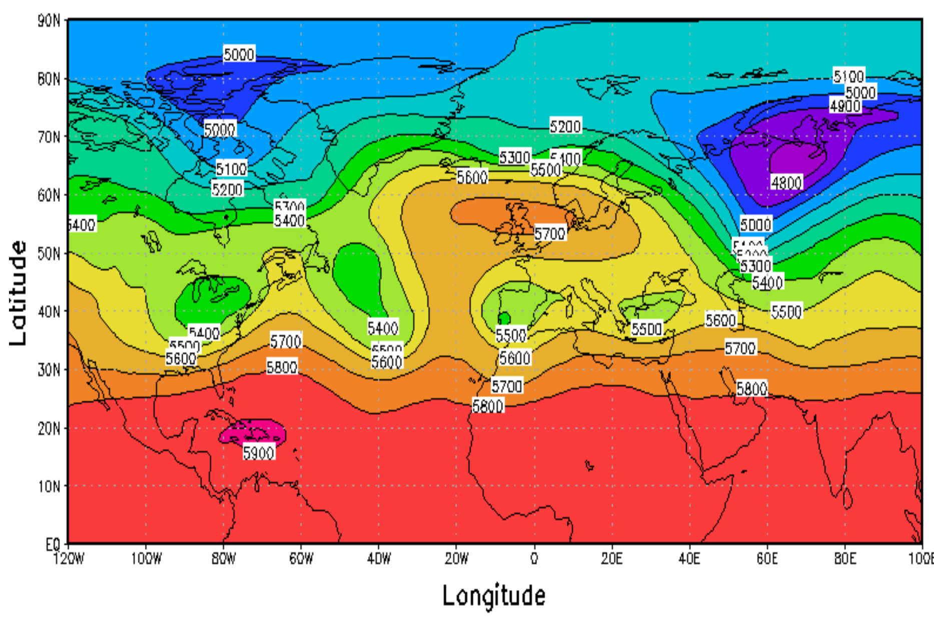

Figure 4.18: Geopotential height at 500 hPa on 28/02/2013.

Beside the effect on the actual weather, TB events are assumed according to the theory to be dynamically coupled with SSW events and consequently the troposphere to be influencing the stratosphere and vice versa via the upward and downward propagation of planetary waves. In order to examine such a hypothesis we are obliged to study the interactions between these compartments of the atmosphere and try to find any possible connections. When looking at Figure 4.2 and the zonal mean zonal wind we observe that on the dates the TB events occurred there is an increase on the zonal wind value. Moreover, it is no coincidence that the anticyclonic flow is observed almost at 60^0 N almost in every case. With this information and keeping in mind that any changes at one level will need time until they have an impact on another level, it is fair to say that TB events are connected to the sudden increase in the zonal wind at higher levels. This is the case even for the TB2 event where we observe an increase in the value of the wind during the SSW event. This is more visible in Figure 4.19 which informs us about the evolution of the zonal mean zonal wind at 60^0 N and 10 hPa level but for a shorter period of time and not the whole time series as Figure 4.2.

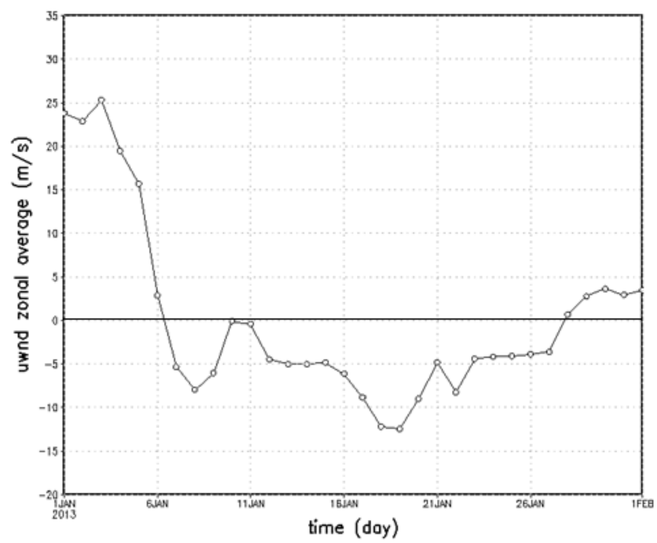

Figure 4.19: Evolution of the zonal mean zonal wind (m/s) at 10 hPa and 60^o N.

During the TB1 event the peak in the value of the zonal mean zonal wind is clear and is preceded by a huge decrease in the value of the wind. After reaching its maximum value, the zonal wind starts decreasing again moving towards the SSW event. On the other hand during the TB3 event the zonal mean zonal wind increases its value after a short period of couple of days when the value was stabilized.

More importantly the connection between stratosphere and troposphere can be seen through the temporal evolution of temperature in the stratospheric levels. The anticyclonic flows created during the TB events lead to an increase in temperature over the polar cap and that is shown in Figure 4.13 and Figure 4.14. Especially the impact of TB1 and TB2 events is clear on the temporal evolution of the temperature at the levels of 10 and 50 hPa respectively. The temporal evolution of temperature at these two levels has already been described but the connection with the TB events is a straightforward observation. Early in December of 2012 there is an increase in temperature over the polar region which coincides with the TB1 event. As a matter of fact the temperature starts to increase earlier than the initialization date of the TB1 event in both levels. The same applies for the TB2 event whereas during the TB3 event there is not a big significant increase in temperature observed in the stratosphere.

5. Discussion

This purpose of this study was to examine in detail the theory on the SSW events and their connection with the TB events. The potential connection between these phenomena would be the proof that stratosphere and troposphere are dynamically coupled. The results from the analysis of the case study of 2013 prove this hypothesis. In the winter of 2012/2013 a major SSW occurred which was due to upward propagation of planetary waves from the troposphere. Through the general circulation of the atmosphere the Rossby waves brought energy to higher levels in the atmosphere which eventually affected the circulation of the polar vortex. This can be seen by the study of the E-P vector field which is a representation of how the eddy fluxes affected the atmosphere. The whole E-P vector field of this winter shows the normal distribution as described in the studies mentioned in the theoretical part. At the lower levels the meridional momentum flux dominates the atmosphere increasing the forcing towards the higher levels and as the elevation increased the heat flux became dominant leading to the tilting of the vector field in the southward direction. In addition to the distribution of the E-P vector field, the evolution of the eddy fluxes shows the levels and regions in the atmosphere where the atmospheric circulation is influenced. The TB events observed in the winter of 2012/2013 come to support the findings. Large anticyclones where formulated over north Atlantic and Europe which meant that these areas experienced persisting high pressure systems leading to clear sky and light winds on the surface of the earth. The fact that the TB events were once again present in the troposphere before and after the SSW event gives us the necessary reasoning for connecting the troposphere and the stratosphere in a dynamic way. The fact that all TB events had an influence on the stratosphere suggests that the propagation of the Rossby waves into the stratosphere is the reason for the existence of the SSW event. On the other hand the appearance of a TB event after the restoration of the circulation of the polar vortex questions this assumption. Because of the fact that the atmospheric circulation not only brings air masses to higher levels but also to the lower levels through the different loops included, it means that phenomena occurring in the stratosphere indeed will affect the tropospheric circulation. This is one of the reasons why there is a TB event even after the SSW one. The findings of several composite studies mentioned in the theory could not guarantee robust evidence on the debate of which phenomenon trigger the initialization of the other. The analysis of this case study could not of course provide with more evidence on this topic but helped on the understanding how different compartments of the atmosphere interact and to understand the role of the atmospheric circulation on the creation of large scale phenomena at different levels. The fact though that this specific winter according to the dynamical analysis can be characterized as a normal one does not come to an agreement when one tries to understand why several areas of Europe experienced such a severe winter in terms of weather systems and the prolonged period in which those weather systems were in action. This has to do with the fact that meteorological systems in general are chaotic and due to the huge amount of interactions it is really difficult to correctly interpret the findings when studying a topic.

6. Conclusions

Concluding this study it needs to be mentioned that even though SSWs are not fully studied yet it is proven that they are phenomena which play a key role on the stratospheric structure. Since their discovery, almost every year in the winter time of the NH there is at least one SSW event observed. Apart from SSWs, in the troposphere there are other phenomena that occur and in fact more often than the warming events. These phenomena are strongly connected but neither the theory nor the analysis of this case study could prove which ones triggers the other one if there is such a case. There were some winters in the past when the TB events were not followed by SSWs. What is for sure is that because they are large scale phenomena they may be affected by other factors which were not taken into account in this study. The theory used in this study was put in practice through the analysis of a specific case study. The role of the eddy fluxes was examined in the general circulation of the atmosphere and even though the propagation of Rossby waves is pretty clear till now, the creation of warming events up in the stratosphere is something that needs to be examined further in details. The stratosphere is a different compartment compared to the troposphere in many ways and the studies that need to be done in order to fully understand the creation, evolution and impact of SSW on the rest of the atmosphere need to take into account the interactions with the other compartments of the atmosphere. A first step would be to firstly try to find out what triggers a TB event when this occurs early in the winter before the SSW event initialization. If such a study produces clear and robust evidence on how a TB event is initialized then it would easier to assess the characteristics of a SSW event as it would be clearer how this phenomenon is triggered. Because the stratosphere is a more stable compartment than the troposphere, large scale systems at these levels can assist on the large scale weather forecast and this the reason why understanding SSW events could eventually help us improve the forecast methods.

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Appendix I

<u>Description of the available files for this research.</u>

For this research the next data were available which were included in **.nc files**, mentioned in the brackets:

1. u-wind (CM_u.nc)
2. v-wind (CM_v.nc)
3. w-wind (CM_w.nc)
4. temperature (CM_T.nc)
5. potential vorticity (CM_PV.nc)
6. vorticity (CM_vort.nc)
7. Phi (geopotential) (CM_Phi.nc)

To manipulate the raw data from the .nc files the following **.f90 files** were created.

1. u.f90
2. v.f90
3. w.f90
4. temp.f90
5. No file was created
6. vorticity.f90
7. Phi.f90

Each one of the .f90 files has as an input the .nc file that corresponds to the same variable and from the .f90 files certain calculations were made. The outcome of these calculations is shown below for each .f90 file respectively and written in .dat files.

1. uwnd_st.dat, uwnd_za.dat, uwnd_in.dat
2. vwnd_st.dat, vwnd_za.dat
3. wwind2013_out.dat
4. theta.dat, theta_za.dat, theta_st.dat
5. No calculations made
6. Vort2013_out.dat
7. Phi2013_out.dat

From the file labelled as 1, the output files contain the u^*,$[u]$ and u.

From the file labelled as 2, the output files contain the u^* and $[u]$.

From the file labelled as 3, the output file contains the w.

From the file labelled as 4, the output files contain the θ, $[\theta]$ and θ^*.

From the file labelled as 5, no output files were created.

From the file labelled as 6, the output file contains the Vorticity.

From the file labelled as 7, the output file contains the Φ.

In the previous statements an asterisk (*) denotes the departure from the zonal mean and the square brackets denote the zonal mean of each variable. The u, v and w output files denote the actual values of u, v and w component of the wind and θ is the potential temperature.

In order to calculate the Eliassen-Palm flux the following were calculated:

[u*v*], [v*θ*] namely ustvst and vsttst in a .f90 file called readall_version**.f90 (different versions available) as long as other variables needed to be used for the divergence of the Eliassen-Palm vector field and also for the Eliassen-Palm vectors E_1 and E_2.

Finally this .f90 file is producing output files named as uv.dat, vth.dat and ep.dat. These files are respectively the [u*v*], [v*θ*] and Eliassen-Palm flux values.

It also produces output files readable by grads which are e1n.dat, e2n.dat and dive.dat which are the Eliassen-Palm vectors E_1 and E_2 and the divergence of the Eliassen-Palm vector field.

In the .nc files that no output files were created, the manipulation of these data occurred straight away in GrADS

Appendix Ia

program u_wind

```
use netcdf
implicit none

! This is the name of the data file we will read.
character (len = *), parameter :: FILE_NAME = "CM_u.nc"
integer :: ncid

! We are reading 4D data, a 25 x 91 x 360 lvl-lat-lon grid, with 151
! timesteps of data.
integer, parameter :: NDIMS = 4, NRECS = 151
integer, parameter :: NLVLS = 25, NLATS = 91, NLONS = 360
character (len = *), parameter :: LVL_NAME = "lev"
character (len = *), parameter :: LAT_NAME = "lat"
character (len = *), parameter :: LON_NAME = "lon"
character (len = *), parameter :: REC_NAME = "time"

! The start and count arrays will tell the netCDF library where to
! read our data.
integer :: start(NDIMS), count(NDIMS)

! In addition to the latitude and longitude dimensions, we will also
! create latitude and longitude variables which will hold the actual
! latitudes and longitudes. Since they hold data about the
! coordinate system, the netCDF term for these is: "coordinate
! variables."
real :: lats(NLATS), lons(NLONS), recs(NRECS), lvls(NLVLS)
integer :: lon_varid, lat_varid, rec_varid, lvl_varid
```

```
! We will read wind fields only. In netCDF
! terminology these are called "variables."
character (len = *), parameter :: uwnd="u"
integer :: uwnd_varid

! Loop indices
integer :: lvl, lat, lon, rec, irec, time
real :: sumu

! We recommend that each variable carry a "units" attribute.
character (len = *), parameter :: UNITS = "units"
character (len = *), parameter :: uwnd_UNITS = "m/s"
character (len = *), parameter :: LAT_UNITS = "degrees_north"
character (len = *), parameter :: LON_UNITS = "degrees_east"

! Program variables to hold the data we will read in. We will only
! need enough space to hold one timestep of data; one record.
real :: uwnd_za(NLATS, NLVLS, NRECS)
real :: uwnd_st(NLONS, NLATS, NLVLS, NRECS)
real :: uwnd_in(NLONS, NLATS, NLVLS, NRECS)

! Open the file.
call check(nf90_open(FILE_NAME, nf90_nowrite, ncid))

! Also open an output file
open (unit=20,file="uwnd_st.dat",status="replace",action="write")
open (unit=21,file="uwnd_za.dat",status="replace",action="write")
open (unit=22,file="uwnd_in.dat",status="replace",action="write", &
      form="unformatted",access="direct",recl=NLATS*NLONS*4)

! Get the varids of the latitude and longitude coordinate variables.
call check(nf90_inq_varid(ncid, LAT_NAME, lat_varid))
call check(nf90_inq_varid(ncid, LON_NAME, lon_varid))

! Read the latitude and longitude data.
call check(nf90_get_var(ncid, lat_varid, lats))
call check(nf90_get_var(ncid, lon_varid, lons))

! Get the varids of the rec variable.
call check(nf90_inq_varid(ncid, REC_NAME, rec_varid))

! Read the rec data.
call check(nf90_get_var(ncid, REC_varid, recs))

! Get the varids of level variable.
call check(nf90_inq_varid(ncid, LVL_NAME, lvl_varid))

! Read the level data.
call check(nf90_get_var(ncid, lvl_varid, lvls))

! Get the varids of the uwnd netCDF variables.
call check(nf90_inq_varid(ncid, uwnd, uwnd_varid))
```

```
  ! Read all records of NLVLS*NLATS*NLONS values, starting at the beginning
  ! of the record (the (1, 1, 1, rec) element in the netCDF file).
  count = (/ NLONS, NLATS, NLVLS, NRECS /)
  start = (/ 1, 1, 1, 1 /)

  ! Read the uwnd data from the file, one record at a time.
  do rec = 1, NRECS
    call check( nf90_get_var(ncid, uwnd_varid, uwnd_in, start, count))
  end do

  ! Close the file. This frees up any internal netCDF resources
  ! associated with the file.
  call check(nf90_close(ncid))

  do rec=1,NRECS
   write (20,900) rec
   write (21,900) rec
   do lvl=1,NLVLS
    write (20,900) lvl
    write (21,900) lvl
    do lat=1,NLATS
     sumu=0
     do lon=1,NLONS
          sumu=sumu+uwnd_in(lon,lat,lvl,rec)
     end do
     uwnd_za(lat,lvl,rec)=sumu/NLONS
     write (21,902) lat,uwnd_za(lat,lvl,rec)
     do lon=1,NLONS
          uwnd_st(lon,lat,lvl,rec)=uwnd_in(lon,lat,lvl,rec)-uwnd_za(lat,lvl,rec)
     end do
     write (20,901) lat,(uwnd_st(lon,lat,lvl,rec),lon=1,NLONS)
    end do
   end do
  end do

  irec=1
  do time=1,NRECS
   do lvl=1,NLVLS
    write (22,rec=irec) ((uwnd_in(lon,lat,lvl,time),lon=1,NLONS),lat=1,NLATS)
    irec=irec+1
    print *, irec
   end do
  end do

900 Format(I4)
901 Format(I4, 360F10.4)
902 Format(I4, F10.4)

!------------------------------------------------------------------------------------------
contains
  subroutine check(status)
   integer, intent (in) :: status
```

```
    if(status /= nf90_noerr) then
      print *, trim(nf90_strerror(status))
      stop "Stopped"
    end if
  end subroutine check
end program u_wind
```

Appendix Ib

program v_wind

```
  use netcdf
  implicit none

  ! This is the name of the data file we will read.
  character (len = *), parameter :: FILE_NAME = "CM_v.nc"
  integer :: ncid

  ! We are reading 4D data, a 25 x 91 x 360 lvl-lat-lon grid, with 151
  ! timesteps of data.
  integer, parameter :: NDIMS = 4, NRECS = 151
  integer, parameter :: NLVLS = 25, NLATS = 91, NLONS = 360
  character (len = *), parameter :: LVL_NAME = "lev"
  character (len = *), parameter :: LAT_NAME = "lat"
  character (len = *), parameter :: LON_NAME = "lon"
  character (len = *), parameter :: REC_NAME = "time"

  ! The start and count arrays will tell the netCDF library where to
  ! read our data.
  integer :: start(NDIMS), count(NDIMS)

  ! In addition to the latitude and longitude dimensions, we will also
  ! create latitude and longitude variables which will hold the actual
  ! latitudes and longitudes. Since they hold data about the
  ! coordinate system, the netCDF term for these is: "coordinate
  ! variables."
  real :: lats(NLATS), lons(NLONS), recs(NRECS), lvls(NLVLS)
  integer :: lon_varid, lat_varid, rec_varid, lvl_varid

  ! We will read wind fields only. In netCDF
  ! terminology these are called "variables."
  character (len = *), parameter :: vwnd="v"
  integer :: vwnd_varid

  ! Loop indices
  integer :: lvl, lat, lon, rec
  real :: sumv

  ! We recommend that each variable carry a "units" attribute.
  character (len = *), parameter :: UNITS = "units"
  character (len = *), parameter :: vwnd_UNITS = "m/s"
  character (len = *), parameter :: LAT_UNITS = "degrees_north"
```

```
character (len = *), parameter :: LON_UNITS = "degrees_east"

! Program variables to hold the data we will read in. We will only
! need enough space to hold one timestep of data; one record.
real :: vwnd_za(NLATS, NLVLS, NRECS)
real :: vwnd_st(NLONS, NLATS, NLVLS, NRECS)
real :: vwnd_in(NLONS, NLATS, NLVLS, NRECS)

! Open the file.
call check(nf90_open(FILE_NAME, nf90_nowrite, ncid))

! Also open an output file
open (unit=20,file="vwnd_st.dat",status="replace",action="write")
open (unit=21,file="vwnd_za.dat",status="replace",action="write")

! Get the varids of the latitude and longitude coordinate variables.
call check(nf90_inq_varid(ncid, LAT_NAME, lat_varid))
call check(nf90_inq_varid(ncid, LON_NAME, lon_varid))

! Read the latitude and longitude data.
call check(nf90_get_var(ncid, lat_varid, lats))
call check(nf90_get_var(ncid, lon_varid, lons))

! Get the varids of the rec variable.
call check(nf90_inq_varid(ncid, REC_NAME, rec_varid))

! Read the rec data.
call check(nf90_get_var(ncid, REC_varid, recs))

! Get the varids of level variable.
call check(nf90_inq_varid(ncid, LVL_NAME, lvl_varid))

! Read the level data.
call check(nf90_get_var(ncid, lvl_varid, lvls))

! Get the varids of the uwnd netCDF variables.
call check(nf90_inq_varid(ncid, vwnd, vwnd_varid))

! Read all records of NLVLS*NLATS*NLONS values, starting at the beginning
! of the record (the (1, 1, 1, rec) element in the netCDF file).
count = (/ NLONS, NLATS, NLVLS, NRECS /)
start = (/ 1, 1, 1, 1 /)

! Read the vorticity data from the file, one record at a time.
do rec = 1, NRECS
  call check( nf90_get_var(ncid, vwnd_varid, vwnd_in, start, count))
end do

! Close the file. This frees up any internal netCDF resources
! associated with the file.
call check(nf90_close(ncid))

do rec=1,NRECS
```

```
    write (20,900) rec
    write (21,900) rec
    do lvl=1,NLVLS
      write (20,900) lvl
      write (21,900) lvl
      do lat=1,NLATS
        sumv=0
        do lon=1,NLONS
                sumv=sumv+vwnd_in(lon,lat,lvl,rec)
        end do
        vwnd_za(lat,lvl,rec)=sumv/NLONS
        write (21,902) lat,vwnd_za(lat,lvl,rec)
        do lon=1,NLONS
                vwnd_st(lon,lat,lvl,rec)=vwnd_in(lon,lat,lvl,rec)-vwnd_za(lat,lvl,rec)
        end do
        write (20,901) lat,(vwnd_st(lon,lat,lvl,rec),lon=1,NLONS)
      end do
    end do
  end do

900 Format(I4)
901 Format(I4, 360F10.4)
902 Format(I4, F10.4)

!------------------------------------------------------------------------------------------
contains
  subroutine check(status)
    integer, intent (in) :: status

    if(status /= nf90_noerr) then
      print *, trim(nf90_strerror(status))
      stop "Stopped"
    end if
  end subroutine check
end program v_wind
```

Appendix Ic

program temperature

```
  use netcdf
  implicit none

  ! This is the name of the data file we will read.
  character (len = *), parameter :: FILE_NAME = "CM_T.nc"
  real, parameter  :: rdry = 287.05, cp = 1004.0, pstand = 100000.0
  integer :: ncid

  ! We are reading 4D data, a 25 x 91 x 360 lvl-lat-lon grid, with 151
  ! timesteps of data.
  integer, parameter :: NDIMS = 4, NRECS = 151
  integer, parameter :: NLVLS = 25, NLATS = 91, NLONS = 360
```

```
character (len = *), parameter :: LVL_NAME = "lev"
character (len = *), parameter :: LAT_NAME = "lat"
character (len = *), parameter :: LON_NAME = "lon"
character (len = *), parameter :: REC_NAME = "time"

! The start and count arrays will tell the netCDF library where to
! read our data.
integer :: start(NDIMS), count(NDIMS)

! In addition to the latitude and longitude dimensions, we will also
! create latitude and longitude variables which will hold the actual
! latitudes and longitudes. Since they hold data about the
! coordinate system, the netCDF term for these is: "coordinate
! variables."
real :: lats(NLATS), lons(NLONS), recs(NRECS), lvls(NLVLS)
integer :: lon_varid, lat_varid, rec_varid, lvl_varid

! We will read temperature fields only. In netCDF
! terminology these are called "variables."
character (len = *), parameter :: temp="T"
integer :: temp_varid

! Loop indices
integer :: lvl, lat, lon, rec
real :: sumtheta

! We recommend that each variable carry a "units" attribute.
character (len = *), parameter :: UNITS = "units"
character (len = *), parameter :: Temp_UNITS = "K"
character (len = *), parameter :: LAT_UNITS = "degrees_north"
character (len = *), parameter :: LON_UNITS = "degrees_east"

! Program variables to hold the data we will read in. We will only
! need enough space to hold one timestep of data; one record.
real :: temp_in(NLONS, NLATS, NLVLS, NRECS)
real :: theta_in(NLONS, NLATS, NLVLS, NRECS)
real :: theta_za(NLATS, NLVLS, NRECS)
real :: theta_st(NLONS, NLATS, NLVLS, NRECS)

! Open the file.
call check(nf90_open(FILE_NAME, nf90_nowrite, ncid))

! Also open an output file
open (unit=20,file="theta.dat",status="replace",action="write")
open (unit=21,file="theta_za.dat",status="replace",action="write")
open (unit=22,file="theta_st.dat",status="replace",action="write")

! Get the varids of the latitude and longitude coordinate variables.
call check(nf90_inq_varid(ncid, LAT_NAME, lat_varid))
call check(nf90_inq_varid(ncid, LON_NAME, lon_varid))

! Read the latitude and longitude data.
call check(nf90_get_var(ncid, lat_varid, lats))
```

```
call check(nf90_get_var(ncid, lon_varid, lons))

! Get the varids of the rec variable.
call check(nf90_inq_varid(ncid, REC_NAME, rec_varid))

! Read the rec data.
call check(nf90_get_var(ncid, REC_varid, recs))

! Get the varids of level variable.
call check(nf90_inq_varid(ncid, LVL_NAME, lvl_varid))

! Read the level data.
call check(nf90_get_var(ncid, lvl_varid, lvls))

! Get the varids of the uwnd netCDF variables.
call check(nf90_inq_varid(ncid, temp, temp_varid))

! Read all records of NLVLS*NLATS*NLONS values, starting at the beginning
! of the record (the (1, 1, 1, rec) element in the netCDF file).
count = (/ NLONS, NLATS, NLVLS, NRECS /)
start = (/ 1, 1, 1, 1 /)

! Read the vorticity data from the file, one record at a time.
do rec = 1, NRECS
  call check( nf90_get_var(ncid, temp_varid, temp_in, start, count))
end do

! Close the file. This frees up any internal netCDF resources
! associated with the file.
call check(nf90_close(ncid))

! Calculate the potential temperature (theta)
do rec=1,NRECS
  do lvl=1,NLVLS
        do lat=1,NLATS
      do lon=1, NLONS
            theta_in(lon,lat,lvl,rec)=temp_in(lon,lat,lvl,rec)*((pstand/lvls(lvl))**(rdry/cp))
      end do
        end do
  end do
end do

do rec=1,NRECS
  do lvl=1,NLVLS
        do lat=1,NLATS
          write (20,900) lvl,rec
          write (20,901) lat, (theta_in(lon,lat,lvl,rec), lon=1,NLONS)
        end do
  end do
end do

do rec=1,NRECS
 write (21,900) rec
```

```
   write (22,900) rec
   do lvl=1,NLVLS
     write (21,900) lvl
     write (22,900) lvl
     do lat=1,NLATS
       sumtheta=0
       do lon=1,NLONS
            sumtheta=sumtheta+theta_in(lon,lat,lvl,rec)
       end do
       theta_za(lat,lvl,rec)=sumtheta/NLONS
       write (21,902) lat,theta_za(lat,lvl,rec)
       do lon=1,NLONS
            theta_st(lon,lat,lvl,rec)=theta_in(lon,lat,lvl,rec)-theta_za(lat,lvl,rec)
       end do
       write (22,901) lat,(theta_st(lon,lat,lvl,rec),lon=1,NLONS)
     end do
   end do
 end do

900 Format(I4)
901 Format(I4, 360F10.4)
902 Format(I4, F10.4)

!---------------------------------------------------------------------------------------------
contains
 subroutine check(status)
   integer, intent (in) :: status

   if(status /= nf90_noerr) then
     print *, trim(nf90_strerror(status))
     stop "Stopped"
   end if
 end subroutine check
end program temperature
```

Appendix Id

program read_all

```
 implicit none

 ! This is the name of the data file we will read
 character (len = *), parameter :: FILE_NAME1 = "uwnd_st.dat"
 character (len = *), parameter :: FILE_NAME2 = "vwnd_st.dat"
 character (len = *), parameter :: FILE_NAME3 = "theta_st.dat"
 character (len = *), parameter :: FILE_NAME4 = "theta_za.dat"

 ! We are reading 4D data, a 25 x 91 x 360 lvl-lat-lon grid, with 151 timesteps of data
 integer, parameter :: NRECS = 151, NLVLS = 25, NLATS = 91, NLONS = 360
 real, parameter :: a=6371000, omega=0.00007292, pi=3.1415
 real, parameter :: deltaphi=2.0*pi/180.0
```

```
! Loop indices
integer :: lvl, lat, lon, rec, i, j, k, time, irec
! Miscellaneous reals
real :: sum1, sum2, sum3, sum4, ccos
real :: area, north, south, dw, up, dp

! Program variables to hold the data we will read in 4D arrays
real, dimension(:,:,:,:), allocatable :: uwnd_st
real, dimension(:,:,:,:), allocatable :: vwnd_st
real, dimension(:,:,:,:), allocatable :: theta_st
real :: ustvst(NLONS, NLATS, NLVLS, NRECS)
real :: vsttst(NLONS, NLATS, NLVLS, NRECS)

! 3D arrays
real, dimension(:,:,:), allocatable :: theta_za
real :: uv_za(NLATS, NLVLS, NRECS), vth_za(NLATS, NLVLS, NRECS)
real :: epflux(NLATS, NLVLS, NRECS), e1(NLATS, NLVLS, NRECS), e2(NLATS, NLVLS, NRECS)
real :: termA(NLATS,NLVLS,NRECS), termB(NLATS,NLVLS,NRECS), termC(NLATS,NLVLS,NRECS)

! 2D arrays
real :: dthetadp(NLVLS,NRECS), theta_aa(NLVLS,NRECS)

! 1D arrays
real :: phi(NLATS+1), f(NLATS+1), plvl(NLVLS)

! Open input files
open (unit=20,file=FILE_NAME1,status="old",action="read")
open (unit=21,file=FILE_NAME2,status="old",action="read")
open (unit=22,file=FILE_NAME3,status="old",action="read")
open (unit=23,file=FILE_NAME4,status="old",action="read")

! Open output files
open (unit=30,file="uv.dat",status="replace",action="write")
open (unit=31,file="vth.dat",status="replace",action="write")
open (unit=32,file="ep.dat",status="replace",action="write")

! Open GrADS output files
open (unit=41,file="e1n.dat",status="replace",action="write", &
      form="unformatted",access="direct",recl=NLATS*4)
open (unit=42,file="e2n.dat",status="replace",action="write", &
      form="unformatted",access="direct",recl=NLATS*4)
open (unit=43,file="diven.dat",status="replace",action="write", &
      form="unformatted",access="direct",recl=(NLATS-2)*4)
open (unit=51,file="ustvst.dat",status="replace",action="write", &
      form="unformatted",access="direct",recl=NLATS*NLONS*4)
open (unit=52,file="vsttst.dat",status="replace",action="write", &
      form="unformatted",access="direct",recl=NLATS*NLONS*4)

! Calculation of Coriolis parameter and real latitudes
do lat=1,NLATS+1
  phi(lat)=91.0-lat
  f(lat)=2.0*omega*sin(phi(lat)*pi/180.0)
end do
```

```
! Definitions of real pressure levels
plvl(1)=100.0
plvl(2)=200.0
plvl(3)=300.0
plvl(4)=500.0
plvl(5)=700.0
plvl(6)=1000.0
plvl(7)=2000.0
plvl(8)=3000.0
plvl(9)=5000.0
plvl(10)=7000.0
plvl(11)=10000.0
plvl(12)=15000.0
plvl(13)=20000.0
plvl(14)=25000.0
plvl(15)=30000.0
plvl(16)=40000.0
plvl(17)=50000.0
plvl(18)=60000.0
plvl(19)=70000.0
plvl(20)=80000.0
plvl(21)=85000.0
plvl(22)=90000.0
plvl(23)=92500.0
plvl(24)=95000.0
plvl(25)=100000.0

! Read in data
allocate (uwnd_st(NLONS,NLATS,NLVLS,NRECS))
allocate (vwnd_st(NLONS,NLATS,NLVLS,NRECS))
allocate (theta_st(NLONS,NLATS,NLVLS,NRECS))

print *,"Reading u_st"

do i=1,NRECS
  read (20,900) rec
  do j=1,NLVLS
    read (20,900) lvl
    do k=1,NLATS
      read (20,*) lat,(uwnd_st(lon,k,j,i),lon=1,NLONS)
    end do
  end do
end do

print *,"Reading v_st"
do i=1,NRECS
  read (21,900) rec
  do j=1,NLVLS
    read (21,900) lvl
    do k=1,NLATS
      read (21,*) lat,(vwnd_st(lon,k,j,i),lon=1,NLONS)
    end do
```

```
  end do
end do

print *,"Reading theta_st"
do i=1,NRECS
  read (22,900) rec
  do j=1,NLVLS
    read (22,900) lvl
    do k=1,NLATS
      read (22,*) lat,(theta_st(lon,k,j,i),lon=1,NLONS)
    end do
  end do
end do

print *,"Calculating & writing u*v* and v*th*"
! Calculation of u*v* and v*th*
do rec=1,NRECS
  write (30,900) rec
  write (31,900) rec
  do lvl=1,NLVLS
    write (30,900) lvl
    write (31,900) lvl
    do lat=1,NLATS
      do lon=1,NLONS
        ustvst(lon,lat,lvl,rec)=uwnd_st(lon,lat,lvl,rec)*vwnd_st(lon,lat,lvl,rec)
        vsttst(lon,lat,lvl,rec)=vwnd_st(lon,lat,lvl,rec)*theta_st(lon,lat,lvl,rec)
      end do
        write (30,901) lat,(ustvst(lon,lat,lvl,rec),lon=1,NLONS)
        write (31,901) lat,(vsttst(lon,lat,lvl,rec),lon=1,NLONS)
    end do
  end do
end do

CLOSE(unit=30)
CLOSE(unit=31)

! Write GrADS ustvst and vsttst files (4D values)
irec=1
do time=1,NRECS
  do lvl=1,NLVLS
    write (51,rec=irec) ((ustvst(lon,NLATS+1-lat,NLVLS+1-lvl,time),lon=1,NLONS),lat=1,NLATS)
    write (52,rec=irec) ((vsttst(lon,NLATS+1-lat,NLVLS+1-lvl,time),lon=1,NLONS),lat=1,NLATS)
    irec=irec+1
  end do
end do

deallocate (uwnd_st)
deallocate (vwnd_st)
deallocate (theta_st)

print *,"Calculation of u*v* zonal average"
! Calculation of u*v* zonal average
do rec=1,NRECS
```

```
  do lvl=1,NLVLS
    do lat=1,NLATS
      sum1=0.0
      do lon=1,NLONS
        sum1=sum1+ustvst(lon,lat,lvl,rec)
      end do
      uv_za(lat,lvl,rec)=sum1/NLONS
    end do
  end do
end do

print *,"Calculation of v*th* zonal average"
! Calculation of v*th* zonal average
do rec=1,NRECS
  do lvl=1,NLVLS
    do lat=1,NLATS
      sum2=0.0
      do lon=1,NLONS
        sum2=sum2+vsttst(lon,lat,lvl,rec)
      end do
      vth_za(lat,lvl,rec)=sum2/NLONS
    end do
  end do
end do

print *,"Reading theta_za"
! Read in data for theta zonal average
allocate (theta_za(NLATS,NLVLS,NRECS))

do i=1,NRECS
  read (23,900) rec
  do j=1,NLVLS
    read (23,900) lvl
    do k=1,NLATS
      read (23,*) lat,theta_za(k,j,i)
    end do
  end do
end do

print *,"Calculation of theta areal average"
! Calculation of [THETA]
do rec=1,NRECS
  do lvl=1,NLVLS
    sum3=0.0
    sum4=0.0
    do lat=2,NLATS
      area=(a**2)*2*pi*(sin(phi(lat-1)*pi/180.0)-sin(phi(lat+1)*pi/180.0))
      sum3=sum3+area
      sum4=sum4+area*theta_za(lat,lvl,rec)
    end do
    theta_aa(lvl,rec)=sum4/sum3
  end do
end do
```

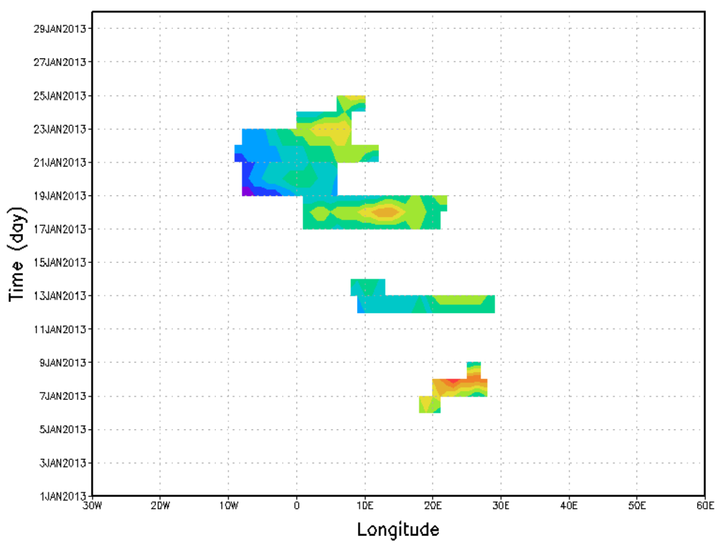
```
deallocate (theta_za)

print *,"Calculation of dtheta/dp"
! Calculation of dthetadp
! Centered differences for lvl 2..24, non-centered differences for lvl 1 and 25
do rec=1,NRECS
 do lvl=2,NLVLS-1
   dp=plvl(lvl+1)-plvl(lvl-1)
   dthetadp(lvl,rec)=(theta_aa(lvl+1,rec)-theta_aa(lvl-1,rec))/dp
 end do
 dthetadp(1,rec)=(theta_aa(2,rec)-theta_aa(1,rec))/(plvl(2)-plvl(1))
 dthetadp(25,rec)=(theta_aa(25,rec)-theta_aa(24,rec))/(plvl(25)-plvl(24))
end do

print *,"Calculation of (E1,E2)"
! Calculation of E1, E2
do rec=1,NRECS
 do lvl=1,NLVLS
   do lat=1,NLATS
    ccos=cos(phi(lat)*pi/180.0)
    e1(lat,lvl,rec)=(-1.0)*a*uv_za(lat,lvl,rec)*ccos
    e2(lat,lvl,rec)=a*f(lat)*vth_za(lat,lvl,rec)*ccos/dthetadp(lvl,rec)
   end do
 end do
end do

print *,"Write E1 and E2"
irec=1
do time=1,NRECS
 do lvl=1,NLVLS
   write (41,rec=irec) (e1(NLATS+1-lat,NLVLS+1-lvl,time),lat=1,NLATS)
   write (42,rec=irec) (e2(NLATS+1-lat,NLVLS+1-lvl,time),lat=1,NLATS)
   irec=irec+1
 end do
end do

print *,uv_za(2,2,2)
print *,cos(phi(2)*pi/180)
print *,f(2)
print *,vth_za(2,2,2)
print *,plvl(2)
print *,dthetadp(2,2)

!.........................................................................

! Calculate the divergence of (e1,e2)
! Term A
do rec=1,NRECS
 do lvl=1,NLVLS
   do lat=2,NLATS-1
     north=uv_za(lat-1,lvl,rec)*cos(phi(lat-1)*pi/180.0)
     south=uv_za(lat+1,lvl,rec)*cos(phi(lat+1)*pi/180.0)
```

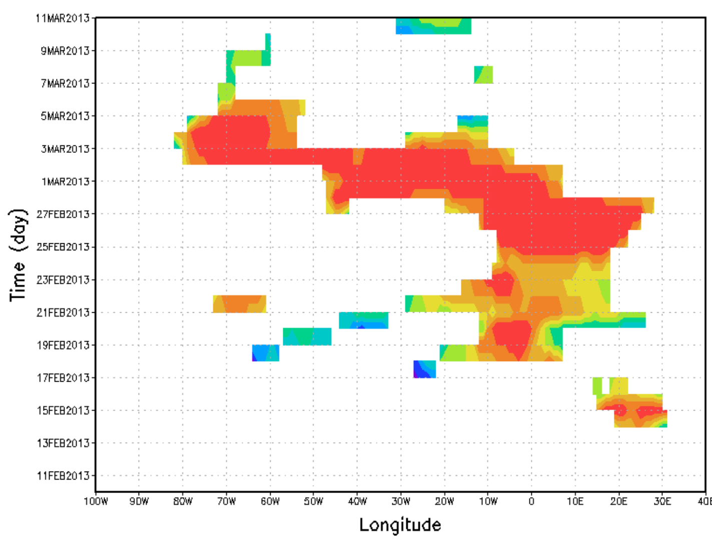
```
      termA(lat,lvl,rec)=(-1.0)*(north-south)/deltaphi
    end do
  end do
 end do

 ! Term B
 do rec=1,NRECS
  do lvl=1,NLVLS
   do lat=2,NLATS-1
    termB(lat,lvl,rec)=uv_za(lat,lvl,rec)*(sin(phi(lat)*pi/180.0))
   end do
  end do
 end do

! Term C
 do rec=1,NRECS
  do lvl=2,NLVLS-1
   do lat=2,NLATS-1
    dw=vth_za(lat,lvl+1,rec)/dthetadp(lvl+1,rec)
    up=vth_za(lat,lvl-1,rec)/dthetadp(lvl-1,rec)
    dp=plvl(lvl+1)-plvl(lvl-1)
        termC(lat,lvl,rec)=a*cos(phi(lat)*pi/180.0)*f(lat)*(dw-up)/dp
   end do
  end do

! And also for the top (lvl=1) and bottom (lvl=25) layers:
   do lat=2,NLATS-1
    dw=vth_za(lat,2,rec)/dthetadp(2,rec)
    up=vth_za(lat,1,rec)/dthetadp(1,rec)
    dp=plvl(2)-plvl(1)
        termC(lat,1,rec)=a*cos(phi(lat)*pi/180.0)*f(lat)*(dw-up)/dp
    dw=vth_za(lat,25,rec)/dthetadp(25,rec)
    up=vth_za(lat,24,rec)/dthetadp(24,rec)
    dp=plvl(25)-plvl(24)
        termC(lat,25,rec)=a*cos(phi(lat)*pi/180.0)*f(lat)*(dw-up)/dp
   end do
 end do

 ! Calculation of the divergence of E
 do rec=1,NRECS
  write (32,900) rec
  do lvl=1,NLVLS
   write (32,900) lvl
   do lat=2,NLATS-1
     epflux(lat,lvl,rec)=termA(lat,lvl,rec)+termB(lat,lvl,rec)+termC(lat,lvl,rec)
     write (32,905) lat,phi(lat),termA(lat,lvl,rec),termB(lat,lvl,rec),termC(lat,lvl,rec),epflux(lat,lvl,rec)
   end do
  end do
 end do

 ! Write GrADS epflux files
 ! Default GrADS reads from south to north, hence the reversal of "lat".
 ! To open these files in GrADS we need special .CTL files. These can be created with eg Notepad.
```

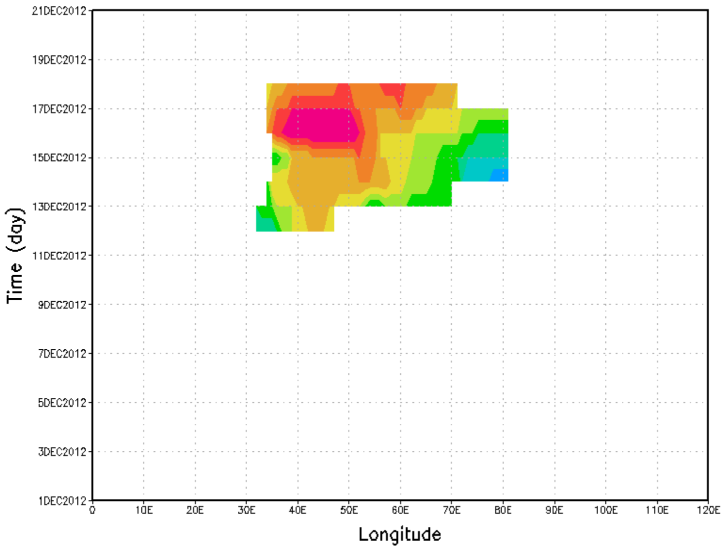
```
 irec=1
 do time=1,NRECS
  do lvl=1,NLVLS
   write (43,rec=irec) (epflux(NLATS+1-lat,NLVLS+1-lvl,time),lat=2,NLATS-1)
   irec=irec+1
  end do
 end do

900 Format(I4)
901 Format(I4, 360F10.4)
902 Format(I4, I4, F9.4)
903 Format(I4, I4, I4, I4, F9.4)
904 Format(I4, F10.4)
905 Format(I4, 5F10.4)

end program read_all
```

Appendix II

Figure 4.15 zoomed in for the TB1, TB2 and TB3 events respectively.






21DEC2012
19DEC2012
17DEC2012
15DEC2012
13DEC2012
11DEC2012
9DEC2012
7DEC2012
5DEC2012
3DEC2012
1DEC2012
Time (day)
0
10E
20E
30E
40E
50E
60E
70E
80E
90E
100E
110E
120E
Longitude