Model Sensitivities and Stratosphere - Troposphere Interactions

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In recent years it has been shown that the troposphere is affected by changes in the climate of the stratosphere as well as vice versa. Investigating the downward influence has implications for understanding not only past climate but also for predicting future climate. A simplified, Newtonian-forced general circulation model is used to investigate the impact of changes in the stratosphere on the tropospheric circulation.

First the sensitivity of tropospheric climate, tropospheric climate variability and response to stratospheric forcing to the surface temperature relaxation timescale are investigated. Changes to this parameter are shown to have a significant impact on the model’s climatology, influencing both the thermal structure of the lower troposphere and the position of the eddy driven mid-latitude jet. The change in mean tropospheric climate influences the annular variability, including its timescale. A strong relationship between this timescale and the magnitude of response to forcing is found, which is consistent with the fluctuation - dissipation theorem.

The tropospheric response for both the surface parameter experiments and stratospheric temperature forcings is shown to be remarkably similar. This indicates that the same dynamical feedbacks are triggered, and thus resulting in the same annular mode-like response.

Further, the impact of an improved representation of the stratosphere (in a model of greater vertical extent) and its effect on the response to a range of stratospheric heating perturbations is investigated. The extent of the heating perturbations, both in latitude and altitude, are shown to have a significant impact on the tropospheric response. These experiments further reveal an influence of the tropospheric eddy response back onto the stratospheric forcing region.
which modifies the direct stratospheric response to the heating. This suggests a strong two way coupling between the lower stratosphere and the tropospheric jets and storm track eddies.
Declaration of Originality

The work presented in this thesis is that of the author, and any information, figures or data taken from other sources have been referenced accordingly.

Alice Margareta Flint, 27th June 20112
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List of Symbols

\( a = 6371 \cdot 10^3 \text{m}, \) radius of Earth

\( c_p = 1005 \text{JK}^{-1} \text{kg}^{-1}, \) specific heat capacity of dry air at constant pressure

\( f = 2\Omega \sin \phi, \) Coriolis parameter

\( g = 9.8 \text{m/s}^2, \) acceleration due to gravity

\( H = 8000 \text{m}, \) density scale height

\( J, \) rate of heating per unit mass due to radiation, conduction and latent heat release

\( k, \) dimensionless zonal wavenumber

\( n^2, \) refractive index squared

\( N, \) buoyancy frequency

\( p, \) pressure (hPa)

\( p_s, \) surface pressure

\( Q, \) diabatic heating

\( q, \) potential vorticity

\( R = 287 \text{JK}^{-1} \text{K}^{-1}, \) gas constant

\( t, \) time

\( u, \) zonal wind (m/s)

\( v, \) meridional wind (m/s)

\( \bar{v^*}, \) TEM meridional wind

\( \omega, \) vertical velocity (m/s)

\( \bar{\omega^*}, \) TEM vertical velocity

\( z = -H \ln(p/p_s), \) height

\( \theta, \) potential temperature

\( \kappa = R/c_p \approx 2/7 \)

\( \lambda, \) longitude (°)

\( \rho, \) density

\( \sigma = p/p_s \)

\( \phi, \) latitude

\( \Phi, \) geopotential height
\[ \Omega \approx 7.29 \cdot 10^{-5} \text{1/s, angular velocity of the Earth} \]
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Results for the experiments run with the L26 IGCM2.2 for Chapter 5. The errors indicated are one standard error for the jet positions and the jet shifts. The errors for the decorrelation times $\tau$ were calculated following the method described in Chapter 3. The jet shifts are always given relative the respective control run (i.e. L26complete, B1L26 or B3L26). .................................................. 225
Results for the experiments run with the L15 IGCM2.2 for Chapter 5. The errors indicated are one standard error for the jet positions and the jet shifts. The errors for the decorrelation times $\tau$ were calculated following the method described in Chapter 3. The jet shifts are always given relative the the respective control run (i.e. L15control or B1L15).
1 Introduction

1.1 The Troposphere and the Stratosphere

The troposphere and stratosphere are the two lowermost layers of the Earth’s atmosphere. The troposphere extends from the surface to an altitude of between 8 km (at polar latitudes) and 16 km (at equatorial latitudes). In this layer the temperature generally decreases with altitude (with a global mean lapse rate of about 6.5 K/km). The troposphere contains about 80% of the atmosphere’s mass and most weather systems occur here. The circulation in the troposphere is characterised by large overturning cells, the Hadley, Ferrel and polar cells (see Fig. 1.1).

![Schematic representation of Hadley-, Ferrel- and Polar-Cell and jet streams](image)

Figure 1.1: Schematic representation of Hadley-, Ferrel- and Polar-Cell and jet streams (from NOAA).

As a straightforward consequence of the geometry of the Earth and its orbit, the annually averaged incoming solar radiation per unit surface at the equator is much larger than at the poles. In the mean the net incoming solar energy absorbed by the Earth and atmosphere must be balanced by the infrared energy that is emitted back to space. However, while the annually averaged solar heating is strongly latitude dependent, the emitted infrared radiation exhibits only a weak dependence on latitude. This means there is a net radiation surplus at the equator and a net deficit at the poles leading to the equatorial region being warmer than the polar regions. This equator-to-pole temperature difference
drives a thermally direct overturning circulation in the tropics and subtropical latitudes. Warm air rises at the equator, moves poleward and cools at higher altitudes before sinking back towards the surface. Such a convection cell, the Hadley cell, dominates tropical and sub-tropical climates and is confined to latitudes of about $\pm 30^\circ$.

As air moves away from the equator to higher latitudes the Coriolis parameter becomes larger and, in the Northern Hemisphere, turns the wind to the right thus introducing a westerly component to the flow. This strong westerly flow, seen at the poleward upper edge of the Hadley cell (see Fig. 1.1) is one of the most prominent features of the global atmospheric general circulation are the tropospheric jet streams. These are bands of strong westerly winds ($>30 \, m/s$) meandering around the globe in the upper troposphere at about $30-35^\circ$ latitude in both the southern and northern hemisphere.

Outside the tropics the influence of the Coriolis effect is much larger and impedes the meridional flow. Thus the mean meridional circulation at these latitudes is relatively weak and cannot be responsible for the required poleward heat transport. In the extratropics large scale eddies are responsible for the bulk transport of heat, mass and angular momentum. The observed weak thermally indirect cell in the mid-latitudes, the Ferrel cell, is caused by the combined eddy heat flux and eddy momentum flux distributions in each hemisphere which drive a mean meridional cell with sinking motion equatorward of about $45^\circ$ latitude and rising motions poleward of this latitude.

The polar cell (about $60^\circ$ to $90^\circ$ latitude) is again a thermally direct cell. Cold dense air descends over the poles, as the air sinks, it spreads southward. The southward moving air is deflected sharply to the right (in the Northern Hemisphere) by the Coriolis effect (creating surface polar easterlies). The cold air moves along the surface to lower latitudes. At around $\pm 60^\circ$ latitude, the air rises upwards again.

The boundary between the troposphere and stratosphere is the tropopause. It is characterised by an abrupt change of atmospheric lapse rate.
The World Meteorological Organisation (WMO) defines the tropopause as the lowest atmospheric level at which the lapse rate decreases to 2 K/km or less, and the average lapse rate from this level to any level within the next 2 km does not exceed this lapse rate.

Above the troposphere the temperature increases with altitude due to ozone heating. This atmospheric layer is called the stratosphere and has an upper boundary (the stratopause) at about 50 km above the Earth’s surface. The stratosphere is characterised by a stable stratification which makes it dynamically very different to the troposphere. The abrupt increase in stratification acts as a filter allowing only longer waves to propagate out of the troposphere and trapping smaller scale disturbances in the troposphere. The chemical composition of the stratosphere is very different from the troposphere with much lower water vapor concentrations and much higher ozone concentrations.

![Schematic of the tropical Hadley cell and the stratospheric Brewer-Dobson circulation at solstice](image)

Figure 1.2: Schematic view of the tropical Hadley cell and the stratospheric Brewer-Dobson circulation at solstice (from WMO 1985).

The meridional circulation in the stratosphere is characterised by the Brewer-Dobson circulation, a hemispheric-scale cell with rising air in the tropics and sinking air over the winter pole. Figure 1.2 shows a schematic of the Brewer-
Dobson circulation for solstice conditions. The arrows indicate the main effects on tracers with upward and poleward transport of atmospheric constituents such as water vapor and ozone above the tropical tropopause and a sinking motion poleward of the mid-latitudes. This circulation is wave-driven (also called the ‘wave-driven pump’). This meridional overturning circulation in the stratosphere is mainly restricted to the winter season when the westerly flow allows for upward propagation of waves.

Figure 1.3 shows a vertical profile of mid-latitude temperature based on the U.S. standard atmosphere (1976), indicating the different atmospheric layers and their boundaries.

Figure 1.3: Mid-latitude temperature profile based on the U.S. standard atmosphere (Holton, 1994).

1.2 Stratosphere - Troposphere Interactions

It has been long known that the stratosphere is influenced by changes in the troposphere. However, only in the last decades has it become apparent that the
troposphere is also susceptible to changes in the stratosphere. Many studies, involving both observations and modeling, have confirmed a two-way dynamical coupling (Haynes, 2005) between stratosphere and troposphere. Figure 1.4 gives an overview of various key drivers for stratosphere - troposphere interactions. In order to understand the complete troposphere-stratosphere dynamics both directions of the interaction have to be considered. The most important of these interactions will be discussed in this chapter.

Figure 1.4: Schematic diagram of the some key drivers for stratosphere - troposphere interactions.

1.2.1 Observations of Stratosphere - Troposphere Interactions

Changes in the state of the stratosphere arise due to various causes, both natural (e.g. solar heating, volcanism) and anthropogenic (e.g. cooling effects from greenhouse gas, GHG, emissions). Investigating the tropospheric response to such perturbations is crucial for understanding not only past but also fu-
ture climate. Over the past decades an increasing number of observations have shown that stratospheric processes influence the troposphere over a wide range of timescales. Key examples will be presented in more detail in the following sections.

**Intraseasonal timescales**

On these timescales (10–100 days) stratosphere - troposphere coupling is mostly observed in the Northern Hemisphere winter. Baldwin et al. (1995) used daily height fields from 1964 to 1993 to study correlations between tropospheric and stratospheric circulation anomalies. They were able to show that variability in the stratospheric polar vortex has a significant impact on the tropospheric circulation. Further studies (e.g. Perlwitz and Graf, 1995; Kitoh et al., 1996; Thompson and Wallace 2000; Baldwin and Dunkerton, 1999 and 2001) confirmed that month-to-month variations in the Northern Hemisphere stratospheric polar vortex are coupled with a pattern of anomalies in the tropospheric circulation centered over the North Atlantic sector. The stratospheric changes often precede the tropospheric changes, e.g. an equatorward shift of the tropospheric jet is often observed after a weakening and warming of the stratospheric polar vortex (i.e a so-called sudden stratospheric warming, SSW, event).

**Interannual timescales**

On interannual timescales several examples of stratosphere - troposphere coupling are observed. One example for stratospheric processes that influence the tropospheric climate on these timescales is the Quasi-Biennial Oscillation, QBO. This is a periodic (timescale of ≈ 26 months) descent of alternating westerly and easterly winds in the equatorial stratosphere between about 10 hPa and 70 hPa. Coughlin and Tung (2001) found a QBO signal at the extratropical surface using detrended daily zonal winds from 1953 to 1997. Further, Thompson et al. (2002) analysed 42 year reanalysis (NCEP-NCAR) data and found an impact of the mid-winter QBO on the Northern Hemisphere weather, where the easterly
phase QBO was shown to favour an increased occurrence of extreme cold events.

**Timescales of several years**

On timescales of several years, volcanic eruptions offer another opportunity to study stratosphere - troposphere interactions (e.g. Robock and Mao, 1992; Kodera and Yamazaki, 1994; Stenchikov et al., 1998). Robock and Mao (1992) examined the changes in Northern Hemisphere winter surface temperature patterns after massive volcanic eruptions during the past century. As well as local effects, this research indicated large volcanic eruptions can affect the global climate. They found a warming over Eurasia and North America and a cooling over the Middle East and northern Africa which was independent of the location of the volcanoes. Aerosols injected into the lower stratosphere by volcanic eruptions (mainly sulfate aerosols) are heated by the sunlight preferentially at equatorial latitudes which leads to an increased meridional temperature gradient as well as a strengthening of the stratospheric polar vortex. Modeling studies suggest that the observed winter warming is caused by a positive phase of the Arctic Oscillation/North Atlantic Oscillation (AO/NAO) (e.g. Graf et al., 1994; Stenchikov et al., 2006). As the aerosol concentrations are highest in the tropical stratosphere, the heating for volcanic aerosol is highly localised in this region and due to differences in the radiative balance can lead to cooling in the extra-tropics. It is thought that the increase of the stratospheric meridional temperature gradient due to this radiative heating in the aerosol cloud over the tropics induces this positive phase of the AO/NAO (Kodera, 1994a; Kirschner et al., 1999). Later research confirmed that during volcanic eruptions huge amounts of aerosols can be injected into the lower stratosphere, exerting a large influence on the radiative budget on a timescale of about 2 years (Foster et al. 2007), causing stratospheric warming and tropospheric cooling at the same time. Further, strongly enhanced scattering of incoming solar radiation causes global annual cooling at the surface at similar timescales (2 - 3 years) (Robock and Mao, 1995; Stenchikov 2006, Fischer 2007). While the volcanic aerosol heating in the lower stratosphere is mainly
due to long-wave absorption, the surface cooling is associated with short-wave reflection of the volcanic sulfate aerosols. Modeling studies of volcanic aerosols reproduced qualitatively the tropospheric and stratospheric temperature changes in response to volcanic forcings. Most models are however unable to exactly simulate the observed response; either over- or underestimating the magnitudes of the local tropospheric temperature changes and showing uncertainty towards the exact locations of these changes (Stenchikov et al. 2006).

Another example of how the tropospheric circulation may be influenced through a stratospheric pathway is the irradiance change that is associated with the 11-year solar cycle (e.g. Labitzke and van Loon, 1997; Labitzke et al., 2002). A positive correlation between the globally averaged surface temperature and the solar activity was found by several studies (e.g. Douglass and Clader, 2002; Coughlin and Tung, 2004; Tung and Camp, 2008). Common to all these studies is that the observed temperature signal is larger than would be expected from direct radiative forcing alone. Further studies show consistent changes in the temperature throughout the troposphere and the tropospheric circulation over the solar cycle (e.g. Haigh, 2003; Haigh et al., 2005; Crooks and Gray, 2005). Such tropospheric changes have also been found in modeling studies. Stratosphere - troposphere coupling is found to play an important role in both transmitting the solar signal down into the troposphere and in enhancing the tropospheric response. Early numerical studies (e.g. Kodera, 1990) in which the solar ultraviolet heating rate in a general circulation model was varied leading to stratospheric temperature variations forcing changes in the upper stratospheric mean zonal wind, showed that this leads to changes in the tropospheric circulation. Many studies have been done since. Haigh (1996 and 1999) and Haigh et al. (2005 and 2006), for example, investigated solar influences on the dynamical coupling between the stratosphere and troposphere in a simple general circulation model, GCM. They show that heating perturbations imposed on the lower stratosphere lead to a latitudinal shift of the subtropical jets and a weakening of the jets as well as the tropospheric mean meridional circulation. This is in agreement with observa-
tions of tropospheric behavior associated with solar forcing.

**Decadal timescales**

One of the main changes in the stratosphere over the last decades was the ozone depletion in the second half of the last century and subsequent ozone recovery in recent years. The cooling of the polar stratosphere that is associated with ozone depletion was found to lead to a poleward shift of the tropospheric jet (Thompson and Solomon 2002 and 2009; Chen and Held 2007). Further it was found that stronger ozone depletion in late spring generally leads not only to a stronger poleward displacement and intensification of the jet but also to a stronger expansion of the Southern Hemisphere Hadley cell in the summer (Seidel et al., 2008; Johanson and Fu, 2009; Son et al. 2010). In the global mean stratospheric ozone depletion has been shown to have a small cooling effect on the troposphere whereas a warming at the surface was seen caused by increases in stratospheric water vapor (Foster et al. 2007). However the influence of ozone depletion on the tropospheric climate is not only global, but can be found on regional scales. Modeling studies have suggested that a previously observed cooling near the surface during early austral summer in the Antarctic could be forced by ozone depletion in the Antarctic (Lal et al. 1987; Grise et al. 2009).

Of particular importance for the future climate is the impact of changes in the greenhouse gas (GHG) concentrations on the stratosphere - troposphere system. As GHG concentrations rise the temperature will increase in the troposphere whilst the stratosphere cools due to increased longwave emission by CO₂ (Manabe and Wetherald, 1967, 1975 and 1980). Such a cooling trend in the stratosphere has been found in satellite observations over the last few decades (Ramaswamy et al. 2001). This cooling leads to a reduction of static stability in the tropopause region and thus to an increase in its height. This response to increased CO₂ levels has been both observed (Santer et al. 2003) and simulated with climate models (Kushner et al. 2001). Combined tropospheric warming
and stratospheric cooling leads to an increased meridional temperature gradient at the level of the tropopause due to the tropopause sloping downward at the poles. Simulations of the 21st century found a consistent strengthening and poleward shift of the subtropical jets (Kushner et al. 2001; Yin 2005; Lorenz and DeWeaver, 2007; Kidston and Gerber, 2010; Kushner, 2010).

Even though most climate change studies found a general cooling of the stratosphere, some model studies also found a warming of the Arctic stratosphere in mid-winter (Mahfouf et al. 1994, Rind et al. 1998, Signal et al. 2003). Mahfouf et al. (1994) found that the temperature increases in the polar lower stratosphere of the Northern Hemisphere are caused by an intensification in the frequency of Sudden Stratospheric Warmings in their double CO₂ simulation. Signal et al. (2003) attributed the small warming in the arctic lower stratosphere to increased downward motions associated with an increased residual circulation that is mainly due to the tropospheric CO₂ doubling in their studies. Together these observations provide evidence for the existence of a dynamical stratosphere-troposphere coupling which enables signals to propagate downward from the stratosphere into the troposphere. The main tropospheric response to any stratospheric forcing seems to be a latitudinal shift of the mid-latitude jets and this provided a major driver for the work described in this thesis.

1.2.2 Annular Modes

Many of the stratospheric changes discussed in the previous section influence the troposphere through the annular modes of variability. For example, a shift to the higher index state of the annular modes that is predicted with increased greenhouse gas concentrations might be associated with a cooling of the stratosphere (Lorenz and DeWeaver, 2007; Kidston and Gerber, 2010; Butler et al. 2010). Another example is the tropospheric response to stratospheric ozone depletion and recovery which seems to occur through the annular modes (e.g. Thompson and Solomon, 2002 and 2009; Son et al., 2008 and 2010).

At middle and high latitudes in both the Northern and Southern Hemisphere,
the annular modes are the dominant pattern of climate variability (Thompson and Wallace, 2000). The annular modes are characterised by geopotential height anomalies of opposite signs in the midlatitude and polar regions and corresponding anomalies in the mean zonal wind equator- and poleward of about 45° latitude, see Fig. 1.5.

Figure 1.5: Zonal-mean geostrophic wind (top) and lower-tropospheric geopotential height (bottom) regressed on the standardized indices of the annular modes based upon monthly data, Jan 1958-Dec 1997. The SH is shown in the left panels and the NH in the right panels. Units are m/s with contour intervals of 0.5 m/s (top) and m/std dev with contour intervals of 10 m (bottom). Figure from Thompson and Wallace (2000).

The annular modes can be viewed as both a north-south shift in atmospheric mass between the polar regions and the mid-latitudes and as a north-south shift of the extratropical zonal wind with centres of action located at about 55-60 degrees and about 30-35 degrees (Thompson, 2007). Both the Northern (NAM)
and Southern (SAM) annular modes exist year round in the troposphere. During winter they are strongest and extend well into the stratosphere. The annular modes explain about $\approx 20 - 30\%$ of the total variability in the zonal wind fields and geopotential height in the corresponding latitudes in both hemispheres. Since their influence is strongest at high latitudes, the annular modes are sometimes referred to as Arctic and Antarctic Oscillations. The NAM (or AO - Arctic Oscillation) is an annular mode. The NAO (North Atlantic Oscillation) may be part of this but is distinct with a particular $\Delta p$ pattern in the North Atlantic (Ambaum et al., 2001).

There seems to be a strong link between the annular modes and the configuration of the extratropical storm tracks and jet streams. It has been shown that changes in the phase of the annular modes can occur as a result of eddy - mean flow interactions, and that external forcings are not necessary to sustain them (De Weaver and Nigam, 2000; Limpasuvan and Hartmann, 2000). Lorenz and Hartmann (2001) discussed the importance of eddy feedbacks in contributing to SAM persistence. They show that synoptic scale eddy feedbacks are very important in contributing to the annular mode persistence, resulting in annular mode variability emerging as the dominant mode of variability. They estimated that internal positive eddy feedback on zonal wind anomalies is responsible for half of the low frequency variance of the annular mode pattern.

A useful tool for climate diagnostics are the indices of the annular modes. The spatial structure of the annular modes can be expressed through the leading empirical orthogonal function (EOF) of near-surface, monthly-mean geopotential anomalies poleward of $20^\circ$ (Karoly, 1990; Thompson and Wallace 2000; Limpasuvan and Hartmann, 2000). Alternatively, the leading EOFs of extratropical zonal-mean zonal wind may be used (Kidson, 1988; Lorenz and Hartmann, 2001 and 2003). The corresponding leading Principal Component (PC) time series is often used to define the temporal variability in the annular modes.

Since the annular modes have significant impacts on climate throughout much of their respective hemispheres, studying and understanding them is very im-
important. The NAM, for example, is associated with large anomalies in surface temperatures and precipitation in North America and Eurasia (Marshall and Plumb, 2001) while the SAM is linked to variations in the sea-surface temperature in the Southern Ocean (Kwok and Comiso, 2002). Both modes are sensitive to variability in the stratospheric flow and show long term trends that might be linked to stratospheric ozone and greenhouse gas concentrations (Rind et. al, 2005; Butler et. al, 2007).

Thompson and Wallace (2000) found what seems to be a strong correlation between annular mode variability in the troposphere and that in the stratosphere. During seasons of strong stratosphere/troposphere coupling the annular modes show stronger variability on monthly time-scales (Thompson, 2007). Further it was shown that the vertical structure of NAM variation in the northern hemisphere winter exhibits a downward propagation from the middle stratosphere to the troposphere (Baldwin and Dunkerton, 2001). Shifts in the probability distributions of the location of storm tracks and the local likelihood of mid-latitude storms seem to be preceded by large variations in the strength of the stratospheric circulation. However this does not necessarily imply a downward propagation of information, but can be caused by forcing from the troposphere (Plumb and Semenik, 2003). Understanding how variations in the stratospheric annular modes influence the surface climate might lead to an improved long-term weather prediction (Baldwin et al., 2003).

Several studies have shown that annular modes similar to those observed in the Earth's atmosphere are produced by models, e.g. in barotropic models (Vallis et al. 2004) or in dynamical core GCMs (Polvani and Kushner, 2002; Kushner and Polvani, 2004, Kushner 2010). Ring and Plumb (2007) showed that changes in eddy fluxes are needed to produce both the full the strength and the structure of the annular mode responses. Even though the above research has shown that the stratospheric and tropospheric components of the annular modes are coupled in both hemispheres, the full cause of coupling is not yet understood (Baldwin et al. 2010). While General Circulation Models generally capture annular modes and
their timescales, it is found that the predicted tropospheric timescales are generally too long in the SH summer compared to observations (Gerber et al. 2010). Simpson et al. (2011b) used the Canadian Middle Atmosphere Model to investigate the influence of stratospheric variability on the tropospheric annular mode timescales. They showed that part of the reason for the unrealistically long timescales is attributable to a late breakdown of the stratospheric polar vortex. This allows the tropospheric influence of stratospheric variability to extend into early summer. They further identified an enhanced summer-time persistence of the models’ SAM that is thought to be of tropospheric origin as it is shown to be unrelated to stratospheric variability. While the effect of stratospheric variability in enhancing the annular mode timescale in the troposphere is found in both hemispheres, the tropospheric feature found for the SAM in this model was not seen in the Northern Hemisphere. Another factor causing the long annular mode timescales in models could be the tropospheric jet latitude. Many GCMs exhibit mid-latitude jets that are too far equatorward (Fyfe and Seanko, 2006) and many studies have shown that the timescales of variability increase when the jet is further equatorward (e.g. Gerber et al. 2008; Chan et al. 2009, Simpson et al. 2010; Kidston and Gerber, 2010 - this is discussed further in section 1.2.5).

1.2.3 Mechanisms of Stratosphere - Troposphere Interactions

Several mechanisms aiming to explain how changes in the state of the stratosphere might influence the troposphere have been proposed during the last decade. These can be grouped into direct and indirect mechanisms.

- Direct: mechanisms where stratospheric changes influence the tropospheric circulation directly.

- Indirect: mechanisms involve wave-mean flow interactions between tropospheric waves and the stratospheric flow.

One example for a direct mechanism involves changes in the potential vorticity (PV) distribution in the lower stratosphere (Hartley et al., 1998; Ambaum and
Hoskins, 2002; Black 2002). Quantities such as velocity or temperature can be determined from the potential vorticity (PV) distribution via the non-local inversion operator \( \ell \). Under the assumption of quasi-geostrophy PV anomalies \( (q') \) are linearly related to geopotential height anomalies \( (\Phi') \) as \( q' = \ell(\Phi') \) (Hartley et al. 1998). Any change in the potential vorticity distribution in the lower stratosphere gives rise to changes in wind and temperature in the troposphere. Hoskins and Ambaum (2002) show that this PV inversion mechanism can be used to explain part of the observed relationships between surface pressure variations, tropopause height, and the strength of the stratospheric vortex. In addition to the PV inversion mechanism they suggest that a coupling between a stratospheric NAO component to its tropospheric component occurs via wave propagation and geostrophic and hydrostatic adjustment. A schematic of the connections is shown in Figure 1.6.

![Schematic of the connections between modulations in the NAO, the height of the tropopause, and the strength of the stratospheric jet](image)

Figure 1.6: Schematic of the connections between modulations in the NAO, the height of the tropopause, and the strength of the stratospheric jet (Ambaum and Hoskins, 2002).
If the North-Atlantic Oscillation (NAO) index increases this leads to an enhanced cyclonic circulation over Iceland (IC) and the tropopause lowers with associated positive potential vorticity anomaly (+). Upward propagating Rossby waves are now refracted more toward the equator and break less in the stratospheric jet. As a consequence the stratospheric jet enhances with associated positive potential vorticity anomaly (+). This leads to a rise of the tropopause below the anomaly and a stretching of the tropospheric column (↑). This prompts an enhanced cyclonic circulation over the North Pole.

The PV inversion mechanism can be refined when longer timescales are considered. The meridional circulation tends to be deeper and narrower on longer timescales (with radiative damping included). This might allow for an enhanced response of the troposphere to stratospheric wave forcings (Haynes et al. 1991, Holton et al. 1995).

One particular case is the downward control principle (Haynes et al., 1991) which applies to zonal mean fields. This is also an example of a direct mechanism. Here the PV inversion mechanism is equivalent to the statement that a wave-induced force applied to the stratosphere will give rise to an acceleration in the troposphere. These accelerations are due to instantaneously induced meridional circulations. Haynes et al. (1991) show that a zonal force produces a response that is predominantly downward.

They further show that this principle of downward control is only relevant to regions of the atmosphere where the angular momentum contours span the atmosphere in the vertical (e.g. mid-latitudes) and fails in other regions (e.g. the tropics). Moreover they demonstrate that if timescales considered are longer than the radiative timescale, the downward control principle still applies for non steady-state (i.e. time dependent forcings). The response to the forcing is then however not exclusively downward. Still, due to the much larger mass contained in the troposphere compared to the stratosphere, this principle is unlikely account for the full tropospheric response to stratospheric perturbations.
While most of the proposed mechanisms of stratosphere - troposphere interaction are dynamical, Grise et al. (2009) investigated radiative processes. They studied the impact radiative processes due to altered stratospheric temperatures and ozone concentrations have on the troposphere. Explicitly neglecting changes in the tropospheric dynamics, they were able to isolate the component of tropospheric temperature response that is driven only by radiative effects. They were able to show that the radiative effect of a stratospheric cooling can explain a significant proportion of the temperature trends in the upper and middle troposphere. However, the lower tropospheric temperature response cannot be accounted for by radiative effects. Hence, while direct radiative effects contribute to the tropospheric response, a dynamical mechanism is necessary to explain the full response.

Other mechanisms are based on communication between troposphere and stratosphere via propagation of large-scale waves (Hartman et al., 2000; Limpasuvan and Hartman, 2000). Such mechanisms are examples of an indirect mechanism. One proposed mechanism involves wave reflection; Rossby waves propagating out of the troposphere might be sensitive to changes in the refractive state of the lower stratospheric flow and are reflected downward back into the troposphere from higher in the stratosphere (Perlitz and Harnik, 2003 and 2004; Ortland and Dunkerton, 2004). This means that an altered stratospheric circulation, for example due to radiative forcings, could change the way planetary waves are reflected and thereby alter the tropospheric circulation. Observational evidence for such planetary wave reflection influencing the troposphere has been found during NH winter (Kodera et al., 2008).

Another mechanism that could be important is wave breaking in the stratosphere; Large-scale planetary waves propagating up into the stratosphere can be absorbed and deposit their momentum there. When the stratosphere is not in a reflective state the waves are not reflected downward back into the troposphere. If the stratospheric flow is changed, this can alter the propagation of the large-
scale waves and hence change where their momentum is deposited. This could then lead to descent of zonal wind anomalies. Such alteration of wave-mean flow interactions was suggested by Boville (1984). Reichler et al. (2005) used an idealised model to study the dynamics of such zonal-mean flow anomalies propagating downward from the stratosphere into the troposphere. They generated a pulse of lower-tropospheric wave activity propagating upwards in their model atmosphere and investigated the subsequent evolution in the stratosphere and troposphere. Figure 1.7 shows a schematic illustration of such a planetary wave breaking event in the stratosphere; Planetary waves form over an obstacle (1) during the time $\Delta t$ and propagate up from the troposphere into the stratosphere (2) where they dissipate and break. This results in mixing potential vorticity in the stratosphere and generating a mean flow and residual circulation response there (3). Zonal flow anomalies propagate downwards (4) which lead to a tropospheric response at time lag of $\tau$ (5).

![Figure 1.7: Schematic illustration of planetary wave breaking in stratosphere (Reichler et al., 05).](image)

It was found that even though the initial pulse was always in the troposphere, the evolution of the perturbation was influenced by the state of the stratosphere-
troposphere system at the time of pulse generation. The initial anomalies in the zonal mean circulation and zonal mean wave drag then control how long the signal takes to propagate downward from the stratosphere (Reichler et al. 2005). Several studies (e.g. Kodera et al., 1990; Scott and Polvani, 2004) have shown interactions between planetary waves and the mean flow to be responsible for downward propagation of zonal wind anomalies in the stratosphere. When the stratosphere is in the state where wave-mean flow interactions are strong, interaction between planetary waves and the mean flow can reach down into the troposphere (Perlwitz and Harnik, 2004). Song and Robinson (2004) confirm the importance of Rossby waves by showing that the full tropospheric response cannot be obtained when the Rossby waves in the stratosphere are artificially damped.

Large-scale waves (i.e. wavenumber 1 and 2) are created by air-flow over large obstacles such as mountain ranges or by land-sea temperature contrasts. Therefore, such waves are far more prominent in the Northern Hemisphere than in the Southern Hemisphere. Further, stationary planetary waves are only able to propagate up into the stratosphere when the flow is westerly which is the case only during winter. A mechanism based on the vertical propagation of such waves would therefore be far more effective in the Northern Hemisphere winter than at any other place or time. Therefore, for stratosphere-troposphere coupling observed in the SH or during NH summer a different mechanism is necessary.

Song and Robinson (2004) conducted numerical experiments to investigate the effects of imposed stratospheric perturbations on the troposphere. They found that the effect of stratospheric wave forcing is communicated to the troposphere through a downward penetrating response in the mean circulation. Further they showed that this response is amplified by synoptic scale eddy feedbacks. They called this mechanism 'downward control with eddy feedback'. Eddies are present in both hemisphere as they are created by baroclinic instability which occurs in the strong meridional temperature gradient in mid-latitudes. Mechanisms involve-
ing eddies are therefore valid both in the NH and SH. Eddies are essential for heat and momentum transfer and for maintaining the mid-latitude westerly jet in the upper troposphere. Many modeling studies have shown that tropospheric eddies are influenced by perturbations (mostly studied through temperature perturbations) to the lower stratosphere and that they are crucial for generating the full tropospheric response (Polvani and Kushner, 2002; Song and Robinson, 2004; Kushner and Polvani, 2004 and 2006; Haigh et al., 2005; Williams, 2006; Lorenz and DeWeaver, 2007; Simpson et al. 2009).

Simpson (2009) and Simpson et al. (2009) extended the studies by Haigh et al. (2005) to explain the role of eddies in driving the tropospheric response to stratospheric heating perturbations. An overview of this mechanism is presented in Fig. 1.8.

They show that changes in wave refractivity are important in driving changes in the tropospheric circulation. Furthermore they demonstrate that the vertical temperature gradient around the tropopause and its latitudinal extent are important in determining the direction of the shift of the tropospheric jet associated with a stratospheric heating.

Applying a heating perturbation to the stratosphere causes a change in the vertical as well as horizontal temperature gradients around the tropopause. Changes in the vertical temperature gradient have a direct effect on the meridional potential vorticity (PV) gradient, eqn. 1.1

\[
\overline{q}_\phi = 2\Omega \cos \phi - \left[ \frac{\nabla \cos \phi}{a \cos \phi} \right]_\phi + \frac{af^2}{R_d} \left( \frac{p \theta \overline{v}_\phi}{T \theta_p} \right)_p
\]  

(1.1)

where \( \overline{q}_\phi \) is the potential vorticity. The second term on the right hand side, \( \left[ \frac{\nabla \cos \phi}{a \cos \phi} \right]_\phi \), is a measure of meridional curvature of the zonal wind and the third term, \( \frac{af^2}{R_d} \left( \frac{p \theta \overline{v}_\phi}{T \theta_p} \right)_p \), is a measure of the changes of vertical shear and curvature of zonal wind and potential temperature. All symbols have the usual meaning as given in the list of symbols.

Changes in the meridional temperature gradient effect the shear and curvature of the zonal wind due to the thermal wind balance and these changes again impact the meridional PV gradient around the tropopause. The change of the
Figure 1.8: Eddy feedback mechanism by which heating perturbations in the stratosphere influence the tropospheric circulation (adapted from Simpson et al., 09).

PV gradient in turn influences the refractive index, eqn. 1.2, and hence the direction of the eddy propagation. Thus creating horizontal eddy momentum flux anomalies as well as eddy heat flux anomalies.

\[
n^2 = \left\{ \frac{q_0}{a|\Pi - c|} - \left( \frac{k}{a \cos \phi} \right)^2 - \left( \frac{f}{2NH} \right)^2 \right\} a^2
\]

where \( n^2 \) is zonal mean quasi-geostrophic refractive index (Matsuno, 1970) and \( c \) is the zonal phase speed. All other symbols have the usual meaning as given in the list of symbols.

The anomalous eddy momentum fluxes locally drive changes in the zonal wind around the tropopause and upper troposphere, see momentum balance equation...
(eqn. 1.3). The changes in both the eddy momentum and the eddy heat fluxes induce an anomalous meridional circulation which leads to temperature and zonal wind changes throughout the troposphere.

\[
\frac{\partial \pi}{\partial t} = f\pi - \frac{1}{a \cos^2 \phi} \frac{\partial \bar{u} \bar{v} \cos^2 \phi}{\partial \phi} - k\pi + \text{ageostrophic terms}
\]  

(1.3)

where \( k \) is the boundary layer frictional damping coefficient. All other symbols have the usual meaning as given in the list of symbols.

Two types of positive eddy feedback enhance the zonal wind and temperature changes throughout the troposphere.

- (i) Reductions in the zonal wind lead to an increase of ambient positive refractive index, see eqn. 1.2. From Karoly and Hoskins (1982) it follows that the wave activity is refracted towards latitudes of increased positive refractive index. In case of an easterly anomaly this leads to an increasing EP-flux convergence which drives a further easterly acceleration (and vice versa for a westerly anomaly), thus creating a positive feedback.

- (ii) A second positive feedback loop is created by the spreading of zonal wind anomalies throughout the troposphere: An easterly anomaly with easterly vertical wind shear causes a reduction in local baroclinicity. This leads to a weakening of the eddy flux source and thus reduces the eddy forcing of westerly flow at that latitude, again providing a positive feedback on displacements of the midlatitude jet.

Recent studies suggested alternative eddy feedback mechanisms (e.g. Chen et al., 2007; Chen and Held, 2007; Son et al. 2008). Chen et al. (2007) and Chen and Held (2007) propose a mechanism based on the displacement of critical lines. Thereby changes in vertical wind shear that arise in response to altered meridional temperature gradients lead to an increase of the tropospheric eddy phase speed and shift the latitude of eddy breaking, and hence the jet, poleward.

All of the mechanisms discussed in this section seem to be significant and are
backed up by observational and modeling evidence. None of mechanisms can on its own account for the full tropospheric response. It is therefore likely that a combination of several mechanisms is needed to fully explain how changes in the state of the stratosphere can influence the tropospheric climate. Depending on the nature of the stratospheric forcing and where it takes place different mechanisms could be working together. Whilst research on this has greatly improved our knowledge of stratosphere - troposphere mechanisms, our understanding of this far from complete.

1.2.4 Timescales of internal variability

Chan and Plumb (2009) and Simpson et al. (2010) investigated the dependence of the tropospheric response to stratospheric forcing on the state of the troposphere. This work was motivated by the results from Polvani and Kushner (2002) who found unrealistically long time-scales for the persistence of one particular phase of their model's leading mode of variability. According to the Fluctuation-Dissipation Theorem (FDT) this could lead to unrealistically high magnitudes of responses to external forcings. The FDT was originally devised by Nyquist (1928) in the context of thermal agitation of electric charge in conductors. Later, Leith (1975) used the FDT to calculate the climate response to small external forcing or other parameter changes by estimating suitable statistics in the present climate. The FDT states that the linear response to an imposed forcing is proportional to the projection of that forcing onto the system's natural (unforced) modes of variability and to the decorrelation time associated with that mode (Leith et al., 1975; Chan and Plumb, 2009):

\[(\text{Response } \cdot \text{mode}) \propto \text{decorrelation time } \tau \times (\text{forcing } \cdot \text{mode}) \quad (1.4)\]

Despite this relation not being quantitatively precise in a simple GCM, several studies confirmed that its qualitative nature is accurate (e.g. Ring and Plumb, 2007 and 2008; Gerber et al. 2008; Chan and Plumb 2009). Understanding the atmosphere's natural annular variability could hence be helpful for improving
the understanding of the jet displacements in response to various stratospheric forcings.

Chan and Plumb (2009) found that for the experiment configuration as used by Polvani and Kushner (2002), the tropospheric jet exhibits a bimodality. The jet occurs in two distinct regions, the subtropics and the mid-latitudes and persists in one region for several thousand days before suddenly switching to the other region. This gives rise to the unrealistically long timescales associated with the model’s intrinsic variability. Using a simple GCM to conduct experiments similar to Polvani and Kushner’s experiments and systematically varying the tropospheric equilibrium temperature profiles, Chan and Plumb (2009) found that the bimodality of the jet’s spatial distribution does only occur in the particular configuration used by Polvani and Kushner (2002). It does not occur when modest changes to the tropospheric state are applied. Hence the timescales associated with the model’s internal mode were significantly reduced in the other set-ups. They concluded that the decorrelation timescales associated with the model’s internal variability are just as important as how the external forcings project onto the mode. The same conclusion was reached by Gerber et al. (2008) who tested the annular mode autocorrelation timescale in simple AGCMs. They pointed out that the response of a model to external perturbations, e.g. increased greenhouse gas forcing or ozone recovery, may be abnormal if the low frequency variability is not simulated well. The Fluctuation-Dissipation Theorem was previously used to predict climatological changes in response to tropospheric forcings (Ring and Plumb, 2007 and 2008). Chan and Plumb (2009) demonstrated that the FDT can also be used predict the qualitative response in a stratosphere-troposphere system where stratospheric forcings are applied.

Sparrow et al. (2009) focused on the natural variability of the tropospheric jet and how it can be characterised by examining the combined fluctuations of the two leading modes of annular variability. Their aim was to understand the processes involved in controlling the natural variability of the tropospheric flow and hence help explaining how the stratospheric forcings lead to the tropospheric re-
response in Haigh et al. (2005). They used the same simplified, Newtonian forced, GCM to study the evolution of the first two modes of annular variability at high and low frequencies (short and long time-scales). The behaviour of the annular variability and eddy forcing was shown to depend on the timescale:

- at low frequencies the zonal flow and baroclinic eddies are in quasi-equilibrium and a poleward propagation of the anomalies is observed.

- at high frequencies the flow is strongly evolving and anomalies propagate equatorward.

In the first case (low frequencies) the eddies seem primarily to reinforce the anomalous state and are closely balanced by the linear damping whereas for the high frequency case eddies are shown to drive the evolution strongly. The study was then extended to investigate the evolution on a more refined range of timescales using empirical mode decomposition (S. Sparrow, personal communication). This is currently focusing on the low and mid-frequency range aiming to explain the dynamics associated with the different behaviours in a set of idealised model runs with varying tropospheric states (same set of experiments used by Simpson et al. 2010).

1.2.5 Importance of jet position

Many studies with simplified GCMs consistently showed that the decorrelation timescale $\tau$ of the annular mode variability depends on the position of the mid-latitude jet (e.g. Son and Lee, 2005; Gerber and Vallis, 2007; Gerber et al. 2008; Chan et al. 2009, Simpson et al. 2010; Barnes et al., 2010). A lower latitude jet results in a longer decorrelation timescale while model configurations with a higher latitude jet exhibit shorter annual model decorrelation times. As the fluctuation - dissipation theorem predicts a larger response to a forcing for configurations with long decorrelation time scales, the magnitude of the response is also related to the position of the jet. Simpson et al. (2010), for example, showed that the magnitude of the poleward jet shift in response to their stratospheric
heating depended on the latitude of the climatological jet. Lower latitude jets were found to shift further poleward and have a longer decorrelation timescale than higher latitude jets. Simpson et al. (2011a) investigated the response dependence on the jet latitude further and showed it is related to a difference in the strength of the feedback between eddies and mean-flow anomaly on the equatorward side of the jet. One of the feedbacks responsible for producing a poleward shift of the jet in response to equatorial stratospheric heating is an increase in the equatorward refraction of the eddies. An increase in momentum flux on the poleward side of the low latitude critical line results from this enhanced eddy refraction which produces a momentum flux convergence/divergence anomaly which acts to accelerate the zonal wind on the poleward side of the jet and to decelerate it on the equatorward side. Key to the difference in response magnitude between different jet structures is the coherence of the latitudinal location of the momentum flux anomalies of different phase speeds. Simpson et al. (2011) suggest that for mid-latitude jets located at high latitudes, a wider latitudinal range of critical latitudes exists which results in a lack of coherence between the behaviour of different phase speeds. In this situation low phase speeds can act to damp the effects of high phase speeds and thus provide a negative feedback onto the zonal flow anomaly. In case of a mid-latitude jet located further equatorward, the latitudinal range of critical latitudes is much smaller and the behaviour of the different phase speed eddies is more coherent. This then leads to all phase speeds together providing a strong positive feedback onto the zonal mean flow anomaly.

Other recent studies have suggested alternative possible reasons for the jet latitude dependence on the jet shift. Lee et al. (2007) and Son et al. (2009) also proposed a difference in the strength of the feedback between mean-flow and eddies for the different latitude jets. They proposed that the presence or absence of a low latitude reflecting surface is responsible for the dependence of the timescale of variability on the jet position. This was, however, found to be inconsistent with the behaviour of the natural variability of the different jet structures investi-
tigated by Simpson et al. (2009) (Simpson et al., 2011a). Another possibility is the existence of a 'high latitude limit' that has been suggested by Kidston and Gerber (2010), where low latitude jets can move further towards the pole than high latitude jets before they reach this limit. The nature and reason for this 'high latitude limit' however could not be explained. Barnes et al. (2010) investigated the asymmetry in the persistence of the eddy-driven jet. Their study shows that in a simple barotropic model equatorward-shifted (low-phase) jets are more persistent than poleward-shifted (high-phase) jets. They propose that the asymmetry in phase persistence is due to a turning-latitude near the pole that prevents waves from breaking there and thus inhibits the positive eddy-mean flow feedback for high latitude jets. It is still subject to further research if any of the proposed mechanisms (or a combination thereof) provides a full explanation for the importance of the mid-latitude jet position.

1.3 Thesis questions and structure

The purpose of this research is to improve the understanding of stratosphere - troposphere interactions. For this a simplified general circulation model, the IGCM2.2, is used. As it is crucial to understand the model behaviour in order to correctly interpret any simulation results, the model sensitivities to parameter changes have also been studied as part of this thesis. The main research aims and questions are outlined below.

Investigation of the sensitivity of tropospheric climate, tropospheric climate variability and response to stratospheric forcing to boundary layer parameters.

This research is split into several questions:

- How does a change in boundary layer parameters affect the (unrealistic) cold surface layer in the model climatology?
- Is it possible to reduce the mid-latitude jet variability by reducing the
surface relaxation temperature timescale?

- How are the model responses to stratospheric forcings and to changes in the surface relaxation temperature timescale linked?

Answering these questions will improve the understanding of the IGCM2 and will allow assessment of the sensitivity of previous model results regarding the response to various stratospheric forcings. Even though the IGCM2 is a simplified model, it has been previously shown to be very useful in identifying and understanding stratosphere-troposphere interactions. The IGCM2 is ideally suited to investigate eddy feedback processes which were shown to be crucial for generating a tropospheric response to stratospheric forcings (e.g. Simpson et al. 2009). While not all of the proposed mechanisms of stratosphere-troposphere interactions (see section 1.2.3) can be studied with the IGCM2, the answers gained from simple/intermediate modeling studies offer valuable insights on the underlying dynamics. Answering the above questions will hence not only improve the understanding of this particular model, but also contribute to understanding real-world stratosphere-troposphere interactions.

**Investigation of the impact of an improved representation of the stratosphere (in a model of greater vertical extent) and its effect on the response to a range of stratospheric heating perturbations.**

Several questions are addressed for this research:

- How does a vertical extension of the IGCM2.2 including a better representation of the stratosphere, influence the model climatology?

- What is the effect of retracting heating perturbations in altitude above the tropopause?

- What is the effect of restricting stratospheric heating perturbations in latitude?

- How are the components of the response and the mean state linked?
Answering these questions will again improve the understanding of the model dynamics and hence the interpretation of previous and future modeling studies. The investigation of the influence of the position and extent of stratospheric heating perturbations on the tropospheric response will be useful for understanding real-world forcings. Especially forcings such as solar forcings or temperature perturbations associated with water vapor which have a more complex heating/cooling pattern than previous experiments performed with this model (e.g. Haigh et al., 2005). Further these investigations could confirm the importance of eddy feedback mechanisms and improve the understanding of the magnitude and direction of tropospheric mid-latitude jet shift which occurs in response to stratospheric forcings. Thus this research will help to build up a more complete picture of real world stratosphere-troposphere interaction mechanisms.

Before the questions are addressed, Chapter 2 will give an overview of the IGCM2.2 model and outline the main diagnostics used for this research. In Chapter 3 the model sensitivities with respect to changes in the surface temperature relaxation timescale will be presented. The vertical extension of the model will be discussed in Chapter 4, followed by an investigation of their effects on the model climatology. In Chapter 5 the effects of changes in the position and shape of heating perturbations applied to the extended model will be expounded. Finally the main results are summarised and the general conclusions together with directions for future work are presented in Chapter 6.
2 The Model and Diagnostics

2.1 The Reading Intermediate General Circulation Model - the IGCM2.2

The Reading IGCM2.2 was used for all experiments conducted for this research. The IGCM2.2 is a model of the global atmospheric circulation that has been developed from the primitive-equations baroclinic model described by Hoskins and Simmons (1975). It is a spectral model using a dynamical spectral core to solve the primitive equations on a sphere. The IGCM2.2 is modified to include the angular-momentum conserving vertical discretisation of Simmons and Burridge (1981) while retaining the original sigma coordinate \((\sigma = p/p_s)\). The model uses the spectral transform method (Orszag, 1970) to solve the primitive equations in terms of vorticity and divergence and transforms between spherical harmonics wave number space and a latitude-longitude grid at every time step. Physical processes such as radiative heating and condensation are computed in physical grid space whereas differential dynamical quantities are calculated in spectral space (the linear and quadratic terms can be calculated exactly while for the cubic terms aliasing occurs).

All experiments use a T42 resolution, denoting a triangular truncation at wave-number 42. This is roughly equivalent to a horizontal resolution of about 475 km for the equivalent linear transform grid. The simulations for the investigation of the model sensitivity to changes in the surface relaxation timescale (Chapter 3) are using 15 vertical sigma levels between the surface and 18.5 hPa (about 30 km) which are chosen to increase resolution around the tropopause. The model levels are at: 18.5, 59.6, 106, 152, 197, 241, 287, 338, 400, 477, 569, 674, 784, 887 and 967 hPa (for a surface pressure of \(p_s = 1000 \text{ hPa}\)). Later the model is extended to 26 vertical sigma levels between the surface and 0.1 hPa (Chapter 4).

Though the representation of the large scale dynamics in this model is accurate, the physical processes are highly parameterised. Rather than using detailed ra-
diative, turbulence and moist parameterisation as included in full GCMs, the IGCM climate is maintained using the linear forcing and drag scheme as described by Held and Suarez (1994). The temperature is relaxation to a prescribed zonally symmetric equilibrium temperature field:

$$\frac{\partial T}{\partial t} = ... - k_T(\Phi, \sigma) \left[ T - T_{eq}(\Phi, p) \right]$$  \hspace{1cm} (2.1)

where ... represents the usual dynamical terms which act to change the temperature.

The relaxation timescale $k_T$ is given by

$$k_T = k_a + (k_s - k_a) \max \left( 0, \frac{\sigma - \sigma_b}{1 - \sigma_b} \right) \cdot \cos^4 \Phi$$  \hspace{1cm} (2.2)

where $k_a = 1/40 \text{ day}^{-1}$ and $k_s = 1/4 \text{ day}^{-1}$, such that the temperature is relaxed on a timescale of 40 days (representing radiation and deep moist processes) above the boundary layer, $\sigma < 0.7$ (about 700 hPa), reducing to 4 days at the equatorial surface (representing the planetary boundary layer).

Rayleigh damping of low-level winds, with a timescale of 1 day at the surface, is included to represent boundary-layer friction:

$$\frac{\partial (\overline{v}, \overline{u})}{\partial t} = ... - k_{(\overline{v}, \overline{u})}(\sigma)(\overline{v}, \overline{u})$$  \hspace{1cm} (2.3)

where ... represents the usual dynamical terms that act to accelerate the zonal wind ($\overline{v}$) or meridional wind ($\overline{u}$). The damping timescale $k_{(\overline{v}, \overline{u})}$ is given by

$$k_{(\overline{v}, \overline{u})} = k_f \max \left( 0, \frac{\sigma - \sigma_b}{1 - \sigma_b} \right)$$  \hspace{1cm} (2.4)

where $k_f = 1 \text{ day}^{-1}$ and $\sigma_b = 0.7$.

The zonally symmetric Newtonian relaxation temperature profile $T_{eq}$ is a function of pressure ($p$) and latitude ($\phi$) only.

$$T_{eq} = \max \left\{ (T_{teq} + \Delta T_{tp} \cos^2 \phi), \right. \left. \left[ T_0 - \Delta T_g \sin^2 \phi - \Delta \theta_s \cos^2 \phi \cdot \log \left( \frac{p}{p_0} \right) \right] \left( \frac{p}{p_0} \right)^\nu \right\}$$  \hspace{1cm} (2.5)
where $T_{tepq}$ is the equatorial tropopause temperature, $T_0$ is the equatorial surface temperature. $\Delta T_y$ is the difference between the polar and equatorial surface temperature, $\Delta T_{tp}$ is the difference between the polar and equatorial tropopause temperature. $(\Delta \theta)_z$ is the equatorial lapse rate factor (increase in potential temperature with an increase in altitude of one pressure scale height). $p_0$ is the reference surface pressure ($=1000\,\text{hPa}$) and $\kappa = R/c_p$.

For the T42L15 IGCM2.2 a $\nabla^8$ hyperdiffusion is used to parameterise the effects of sub-grid scale processes and the model top boundary condition is simply a reflective lid (i.e. $\sigma = 0$). For the T42L26 version of the model a $\nabla^2$ diffusion is added for the top three levels (see chapter 4). Further the model does not include any orography. The surface boundary conditions are a spherical surface with no zonal asymmetries or topography. Therefore planetary waves are weak and only eddy forced. This is a significant simplification especially for the Northern Hemisphere where in the real world large planetary waves are forced by airflow over mountains and land-sea temperature contrasts.

The reference state equilibrium temperature and the model control run climatology are shown in Figs. 2.1 and 2.2, respectively, with $T_{tepq} = 200\,\text{K}$, $T_0 = 315\,\text{K}$, $\Delta T_y = 60\,\text{K}$, $\Delta T_{tp} = 0\,\text{K}$ and $\Delta \theta = 10\,\text{K}$. These are the standard settings used by Held and Suarez (1994).

The reference equilibrium temperature profile (Fig. 2.1) is isothermal in the stratosphere, has a negative latitudinal temperature gradient in the troposphere and is symmetric about the equator (i.e. perpetual equinox). The occurrence of gravitational instability is minimised by some positive static stability in the tropics which decreases to zero at the poles. Baroclinic instability is initiated at the start of a model run by applying a white noise perturbation to the surface pressure field.

The results of the model for a $10000\,\text{day}$ control (an initial spin-up period of $1000\,\text{days}$ has been discarded) run is shown in Fig. 2.2. Figure 2.2 (a) shows the zonal mean temperature field. In the stratosphere the dynamical response of the model leads to a profile that is no longer isothermal. The equatorial re-
Figure 2.1: Newtonian relaxation temperature profile for the control run climatology

gion is now cooler and the polar region warmer than in $T_{eq}$ leading to a positive latitudinal temperature gradient. In the troposphere the model response to $T_{eq}$ is a cooling at the equatorial/low latitudes and a warming at mid-high latitudes resulting in a reduction of the negative latitudinal temperature gradient. The zonal mean zonal wind ($\mathbf{u}$) is shown in Fig. 2.2 (b). In thermal wind balance with the latitudinal temperature gradient, a mid-latitude westerly jet is present in both hemispheres. The jets are centered at about $\pm 45^\circ$ with maximum zonal wind speeds just below the tropopause ($\approx 30 \, ms^{-1}$).
The mean meridional circulation, Fig. 2.2 (c), in each hemisphere consists of a thermally direct cell between the equator and about $30^\circ$ (Hadley cell) and a thermally indirect cell extending polewards from the Hadley cell to about $60^\circ$ (Ferrel cell). There is also a weak (hence not visible in the plot) thermally direct polar cell at high latitudes.
Figure 2.2: (a) control run zonal mean temperature, (b) zonal mean zonal wind ($ms^{-1}$), (c) control run stream function of the mean meridional circulation ($10^{10} kgs^{-1}$)
Figure 2.3 shows ERA40 reanalysis plots of the zonal mean temperature (top) and zonal mean zonal wind (bottom) for March-May conditions (close to equinox conditions as used for the IGCM).

![Zonal Mean Temperature and Wind](image)

Figure 2.3: (top) ERA40 zonal mean temperatures (K), (bottom) ERA 40 zonal mean zonal wind (m s⁻¹)

Comparing these with the model results (Fig. 2.2 a and b), it is apparent that the overall structure of the climatology is well captured by the model. The model zonal mean temperatures are however somewhat (about 10 K) higher at the surface and lower (again by about 10 K) in the tropical stratosphere, i.e. show a stronger vertical temperature gradient. The mid-latitude jet shown in the mean zonal mean winds plot produced by the model are overall slightly faster (by about 5 m/s) than the ERA40 results. The main differences in the zonal mean zonal wind are in the tropical surface easterlies (about 7 m/s faster in the model) and
the easterlies in the upper tropical regions where the model produces winds of > 30 m/s while the ERA40 data show only weak winds. Overall despite the simple forcing and dissipation the model produces a climatology that is fairly close to that of the real world atmosphere.

The simplifications in the model allow for a fast computation of the results and make long integrations feasible. This makes the model an ideal tool for sensitivity studies where model responses to a variety of influences are investigated. Moreover, in full GCMs it is often very difficult to attribute specific model responses to any specific causes. Using a simplified model such as the IGCM2.2 it is possible to study basic dynamical mechanisms without them being obscured as can be the case in more complicated GCMs. For these qualities the IGCM2.2 was chosen for this study to investigate stratosphere - troposphere interactions and to conduct model sensitivity studies with the aim of improving the interpretation of the stratosphere - troposphere interaction experiments.

### 2.2 Diagnostics

The diagnostics used for analysing the model responses to both parameter changes and stratospheric temperature perturbations are based on the system of primitive equations which govern the global atmospheric flow. The primitive equations in log-pressure coordinates on a sphere are given as (e.g. Andrews et al. 1987):

- **momentum equations in zonal (eqn. 2.6) and meridional (eqn. 2.7) direction**

\[
\frac{Dv}{Dt} - \left( f + \frac{u \tan \phi}{a} \right) v + \frac{\Phi_\lambda}{a \cos \phi} = X
\]  

\[
\frac{Du}{Dt} + \left( f + \frac{u \tan \phi}{a} \right) u + \frac{\Phi_\phi}{a} = Y
\]  

- **hydrostatic balance in the vertical**

\[
\Phi_z = H^{-1}R\theta \exp^{-\kappa z/H}
\]
• continuity of mass

\[
\frac{u_\lambda + (v \cos \phi)_{\phi}}{a \cos \phi} + \frac{\rho_0 w_z}{\rho_0} = 0
\]  

(2.9)

• thermodynamic relation between diabatic heating and the material rate of change of potential temperature

\[
\frac{D\theta}{Dt} = Q
\]

(2.10)

where the vertical coordinate is the log-pressure coordinate \( z \equiv -H \ln(p/p_s) \) and the horizontal coordinates are \( \lambda = \) longitude and \( \phi = \) latitude. The velocity components are

\[
(u, v, w) = \left[ (a \cos \phi) \frac{D\lambda}{Dt}, a \frac{D\phi}{Dt}, \frac{Dz}{Dt} \right]
\]

(2.11)

where the time rate of change following the fluid motion \( D/Dt \) is

\[
\frac{D}{Dt} \equiv \frac{\partial}{\partial t} + \frac{u}{a \cos \phi} \frac{\partial}{\partial \lambda} + \frac{v}{a} \frac{\partial}{\partial \phi} + w \frac{\partial}{\partial z}
\]

(2.12)

The Coriolis parameter is \( f = 2\Omega \sin \phi \). \( X \) and \( Y \) in equations 2.6 and 2.7 are horizontal components of friction and other nonconservative mechanical forcing. The diabatic heating term in eqn. 2.10 is given by \( Q \equiv (J/c_p) \exp^{\alpha z} H \).

### 2.2.1 The zonally averaged circulation

As a first approximation for the theoretical analysis of synoptic-scale wave disturbances, it can be useful to look at deviations from a zonally average mean flow. In order to analyse the zonally averaged circulation, the interaction of eddies (deviations from the zonal mean) with the mean flow must be investigated. This can be done by using the Eulerian mean, that is expanding any variable \( A \) to \( A = \bar{A} + A' \). In this notation the eddy components are denoted by the primed variables and the longitudinally averaged flow components are denoted by overbars, i.e.

\[
\bar{A}(\phi, z, t) = \frac{1}{2\pi} \int_0^{2\pi} A(\lambda, \phi, z, t) d\lambda
\]

(2.13)

\[
A'(\lambda, \phi, z, t) \equiv A - \bar{A}
\]

(2.14)
In the Eulerian mean the primitive equations become (e.g. Andrews et al. 1987)

- momentum equations in zonal (eqn. 2.15) and meridional (eqn. 2.16) direction

\[
\frac{d\mathbf{u}}{dt} - f\mathbf{u} + \frac{\mathbf{v}}{a \cos \phi} \frac{\partial (\mathbf{v} \cos \phi)}{\partial \phi} + \mathbf{w} \frac{\partial \mathbf{u}}{\partial z} = -\frac{1}{a \cos^2 \phi} \frac{\partial (\mathbf{u} \mathbf{v}' \cos^2 \phi)}{\partial \phi} + \frac{1}{\rho_0} \frac{\partial (\rho_0 \mathbf{u} \mathbf{v}')}{\partial z} + \mathbf{X}
\]

\[
\frac{d\mathbf{v}}{dt} + f\mathbf{u} + \frac{\mathbf{v}}{a \partial \phi} + \mathbf{w} \frac{\partial \mathbf{v}}{\partial z} + \frac{\mathbf{w} \tan \phi}{a} + \frac{\partial \Phi}{a \partial \phi} = \mathbf{Y}
\]

- hydrostatic balance in the vertical

\[
\Phi_z - \frac{1}{H} \bar{R} \bar{\theta} \exp^{-cz/H} = 0
\]

- continuity of mass

\[
\frac{1}{a \cos \phi} \frac{\partial (\mathbf{v} \cos \phi)}{\partial \phi} + \frac{1}{\rho_0} \frac{\partial (\rho_0 \mathbf{w})}{\partial z} = 0
\]

- thermodynamic energy equation

\[
\frac{\partial \bar{\theta}}{\partial t} + \frac{\mathbf{v} \partial \bar{\theta}}{a \partial \phi} + \frac{\partial \bar{\theta}}{\partial z} = -\frac{1}{a \cos \phi} \frac{\partial (\mathbf{v} \mathbf{w}' \cos \phi)}{\partial \phi} - \frac{1}{\rho_0} \frac{\partial (\rho_0 \mathbf{w} \mathbf{v}')}{\partial z} + \mathbf{Q}
\]

When analysing extratropical synoptic-scale motions the quasi-geostrophic approximation is a useful tool. That is, for motions that are hydrostatic and nearly geostrophic (the Coriolis force balances the pressure gradient force), the 3D flow field is determined by isobaric distribution of geopotential \( \Phi(x, y, p, t) \). The horizontal velocity is expressed as the sum of geostrophic and ageostrophic components, \( \mathbf{V} = \mathbf{V}_g + \mathbf{V}_a \), where the geostrophic wind \( \mathbf{V}_g \) is defined as \( \mathbf{V}_g = f_0 \mathbf{k} \times \nabla \Phi \). Further considering quasi-geostrophic scales, advection by the ageostrophic mean
meridional circulation and vertical eddy flux divergences can be neglected as these terms are small compared to the other terms (e.g. Holton, 2004). Applying this reduces the zonal mean momentum balance (eqn. 2.15) and the thermodynamic energy equation (eqn. 2.20)

$$\frac{\partial \pi}{\partial t} - f\vec{v} = -\frac{1}{a \cos^2 \phi} \frac{\partial (\vec{v'} \cos^2 \phi)}{\partial \phi} + \overline{X} \quad (2.20)$$

where the terms on the left hand side are the zonal mean zonal wind acceleration and the Coriolis force on zonal mean meridional wind and the terms on the RHS the convergence of horizontal eddy momentum flux and friction.

$$\frac{\partial \theta}{\partial t} + \overline{w} \frac{\partial \overline{\theta}}{\partial z} = \overline{Q} - \frac{1}{a \cos \phi} \frac{\partial (\vec{v'} \overline{\theta} \cos \phi)}{\partial \phi} \quad (2.21)$$

where the terms on the LHS are the rate of change of potential temperature and the vertical advection and the terms on the RHS the diabatic heating and the convergence of eddy heat flux.

The meridional circulation \((\overline{\pi}, \overline{\pi})\) is required to maintain a state of thermal wind balance. Without this meridional circulation the divergences of eddy momentum flux and the eddy heat flux as well as friction and heating would act to separately change the mean zonal wind (in eqn. 2.20) and temperature fields (in eqn. 2.21). The streamfunction of the mean meridional circulation can be qualitatively expressed as (e.g. Holton, 2004)

$$\overline{\Psi} \propto -\frac{\partial}{\partial \phi} \text{(diabatic heating)} + \frac{\partial^2}{\partial \phi^2} \text{(large - scale eddy heat flux)} + \frac{\partial^2}{\partial \phi \partial z} \text{(large - scale eddy momentum flux)} + \frac{\partial}{\partial z} \text{(zonal drag force)} \quad (2.22)$$

The sign of the streamfunction \(\Psi\) has been chosen such that a thermally direct circulation (i.e. rising in low latitudes and sinking in high latitudes) corresponds to a positive maximum in \(\Psi\) in the NH.

### 2.2.2 Transformed Eulerian Mean and Eliassen-Palm flux

In the conventional *Eulerian mean* a strong cancellation occurs between the eddy heat flux convergence and the adiabatic cooling term. The diabatic heating
term, with which the residual meridional circulation is associated, is only a small residual. However, since the residual meridional circulation is directly related to the mean meridional mass flow, the conventional Eulerian mean is not very convenient for detailed analysis. A clearer diagnostic of eddy forcing is provided by the transformed Eulerian mean (TEM) which defines a residual circulation \((\tilde{u}, \tilde{w})\) that represents the part not cancelled by the eddy heat fluxes.

\[
\tilde{u} \equiv \bar{u} - \frac{1}{\rho_0} \cdot \left( \frac{\bar{v}' \theta'}{\bar{\theta}_z} \right)_z \tag{2.23}
\]

\[
\tilde{w} \equiv \bar{w} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \cdot \left( \cos \phi \frac{v' \theta'}{\bar{\theta}_z} \right)_\phi \tag{2.24}
\]

This formulation (Andrews and McIntyre, 1976) of the residual meridional circulation identifies the net effect of eddies on the mean flow. The residual vertical velocity \((\tilde{w})\) represents the part of the mean vertical velocity whose contribution to adiabatic temperature change is not cancelled by the eddy heat flux divergence. Using eqns. 2.23 and 2.24 to substitute for \(\tilde{u}\) and \(\tilde{w}\) in the zonal mean zonal wind acceleration (eqn. 2.20) and thermodynamic equation (eqn. 2.21) the respective quasi-geostrophic TEM equations are then defined as

\[
\frac{\partial \tilde{v}}{\partial t} - f \tilde{u} - \tilde{X} = \frac{1}{\rho_0 a \cos \phi} \Delta \tilde{F} \tag{2.25}
\]

\[
\frac{\partial \tilde{\theta}}{\partial t} + \tilde{w} \frac{\partial \tilde{\theta}}{\partial z} - \tilde{Q} = 0 \tag{2.26}
\]

With this formulation of the thermodynamic energy equation (eqn. 2.26) the dominant balance between the adiabatic heating (cooling) and the divergence (convergence) of horizontal eddy heat flux has been removed. A particularly useful diagnostic tool is the Eliassen-Palm flux, EP flux (Eliassen and Palm, 1961), the vector \(\tilde{F}\) in eqn. 2.25. The EP flux components, \(\tilde{F}(\phi, z)\), for large-scale quasi-geostrophic eddies in spherical coordinates are given by

\[
F_\phi = -a \cos \phi \bar{u} \bar{v}' \tag{2.27}
\]

\[
F_z = a f \cos \phi \frac{\bar{v}' \theta'}{\bar{\theta}_z} \tag{2.28}
\]
i.e. the EP-flux is a vector in the latitude-height plane. Its latitudinal component is proportional to the negative horizontal eddy momentum flux and its vertical component proportional to the horizontal eddy heat flux. The divergence of the EP flux gives the changes in the zonal mean circulation which are driven by a combination of both the eddy heat fluxes and the eddy momentum fluxes. It represents the net eddy forcing of the zonal mean state, taking account of the induced meridional circulation seen in the conventional Eulerian mean. From the TEM equation for the zonal mean zonal wind acceleration (eqn. 2.25) it is evident that the zonal mean zonal wind can be changed by both the horizontal eddy momentum flux and horizontal eddy heat flux contributions. This was not obvious from the conventional Eulerian mean formulation (eqn. 2.20). A convenient way to display the eddy forcing of the zonal mean flow is provided by mapping the EP flux and contouring the isolines of its divergence. The direction of the EP flux indicates the relative importance of eddy heat and momentum fluxes and can be useful in determining which properties of the waves are changing in order to drive an acceleration in the mean flow (Edmon et al., 1980). Moreover, Edmon et al. (1980) also show that, the direction of the EP flux can be related to the direction of wave propagation.

2.2.3 Principal Component Analysis

Principal Component Analysis - PCA is a multivariate statistical technique for analysing large data sets. It is also called Empirical orthogonal function (EOF) analysis. PCA can be used to reduce data sets containing a large number of variables to a data set with lower dimensionality while still retaining most of the original variability. The higher correlation of the variables in the original data set (i.e. the data set contains redundant information - as usually the case for data sets of atmospheric fields) the greater the reduction of variables achieved by PCA.

A rough outline of the most important steps of PCA is given below (adapted from Wilks, 2006):
• obtain original data set \( \bar{x} \) (in meteorology usually \( \bar{x}(\text{space}, \text{time}) \))

• subtract the mean (PCA is conducted on anomalies): \( \bar{x}' = \bar{x} - \bar{x}_{\text{mean}} \)

• calculate the correlation (or covariance) matrix of the centered data (\( \bar{x}' \))

• calculate the eigenvectors of the matrix

• calculate the new variables:

\[
u_m = e_m^T \bar{x}' = \sum_{k=1}^{K} e_{km} \bar{x}'_k
\]

with \( K \) the dimensions of the original data set and \( m = 1, ..., M \) dimension of the new data set. \( u_m \) is the \( m \)th Principal Component (projection of the data vector \( \bar{x}' \) onto the \( m \)th eigenvector, \( e_m \)). The new variables, \( u_m \), will account successively for the maximum amount of variability, i.e. the eigenvector with the highest eigenvalue is the 1st Principle Component of the data set (ordering the eigenvectors by eigenvalue, highest to lowest, gives the Principal Components in order of significance).

• view the new data set on a coordinate system defined by the eigenvectors.

These eigenvectors are called empirical orthogonal functions (EOFs).

In the context of this work PCA is used to analyse the internal variability of the model rather than as a tool for compression. Most of the model’s zonal mean state internal variability is contained in the first two Principle Components. Figure 2.4 shows the first two empirical orthogonal functions of zonal-mean zonal wind for a 10000 day control run. These first two EOFs together explain about 70 percent of the variance (50.75 percent from EOF1 and 19.09 percent from EOF2). The EOF patterns are shown in meters per second by projecting the anomaly data onto the normalised principal component (PC) time series (S. Sparrow, personal communication). The leading mode of variability is a latitudinal shift of the centre of the mid-latitude jet. The second mode of variability represents a weakening (strengthening) and broadening (narrowing) of the jet.
Figure 2.4: EOF1 (left panel) and EOF2 (right panel) of zonal-mean zonal wind for a 10000 day control run. The contour interval is 0.5 m s$^{-1}$ and the percentage of the variance explained by each EOF is given in the figure title.

Baldwin et al. (2003) define a decorrelation timescale $\tau$ of annular variability which quantifies the persistence of midlatitude jet anomalies. These timescales are obtained by finding the first EOF of the zonal mean zonal wind variability, projecting the zonal wind anomaly from the time mean onto that EOF and then calculating the mean e-folding timescale of the autocorrelation of that projection. This method is described in detail by e.g. Gerber et al. (2008).

2.2.4 Estimating the uncertainty in the decorrelation times

Gerber et al. (2008) describe a method for a rough estimation of the uncertainty in the decorrelation time $\tau$. Their method is based on the assumption that the annular mode index can be described as an order 1 autoregressive process (AR-1). The persistence of such an AR-1 process only depends on the decay timescale of its autocorrelation function $r(t)$. Gerber et al. (2008) estimate the true AR-1 autocorrelation function $r(t) = \exp(-t/\tau)$ of a time series of length $N$ (days) by computing the empirical autocorrelation function $r_N(t)$. Bartlett (1935 and 1946) shows that the variance of $r_N(t)$ about $r(t)$ can be described as a function of $N$, $\tau$, and lag $t$

$$\text{var}[r_N(\tau)] = \frac{1}{N} \left\{ \frac{(1 + e^{-\frac{\tau}{2}})(1 - e^{-\tau})}{1 - e^{-\frac{\tau}{2}}} - 2\tau e^{-\tau} \right\}$$  \hspace{1cm} (2.29)
Assuming that $\tau$ is large ($\tau \gg 2$), the expression $e^{\frac{-2}{\tau}}$ can be approximated by taking the first two terms of its Taylor expansion

$$e^{\frac{-2}{\tau}} \approx 1 - \frac{2}{\tau}$$  \hspace{1cm} (2.30)

Applying this to eqn. 2.29 yields

$$\text{var}[r_N(\tau)] \approx \frac{1}{N} \left\{ \frac{[1 + (1 - \frac{2}{\tau})[1 - e^{-2}]]}{1 - (1 - \frac{2}{\tau})} - 2\tau e^{-2} \right\}$$ \hspace{1cm} (2.31a)

$$\text{var}[r_N(\tau)] \approx \frac{1}{N} \left\{ \frac{[2 - \frac{2}{\tau}][1 - e^{-2}]}{-\frac{2}{\tau}} - 2\tau e^{-2} \right\}$$ \hspace{1cm} (2.31b)

$$\text{var}[r_N(\tau)] \approx \frac{1}{N} \left\{ \frac{[2 - \frac{2}{\tau}]}{2} \cdot (1 - e^{-2})\tau - 2\tau e^{-2} \right\}$$ \hspace{1cm} (2.31c)

as $\tau \gg 2$, $2/\tau \approx 0$, thus

$$\text{var}[r_N(\tau)] \approx \frac{1}{N} \{ (1 - e^{-2})\tau - 2\tau e^{-2} \}$$ \hspace{1cm} (2.31d)

$$\text{var}[r_N(\tau)] \approx \frac{\tau}{N}(1 - 3e^{-2})$$ \hspace{1cm} (2.31e)

The standard error in the height of $r_N(t)$ therefore is

$$\text{std}[r_N(\tau)] \approx \sqrt{\frac{\tau}{N}(1 - 3e^{-2})}$$ \hspace{1cm} (2.32)

The variability in the e-folding time of $r_N(t)$ is then used to determine the uncertainty in $\tau_N$. With the previous assumption that $r_N(t)$ can be expressed as an exponential, $r_N(t)$ is linearised about the point where it reaches $1/e$. Gerber et al (2008) then use the slope $(dr/dt)|_{t=\tau} = -e^{-1}/\tau$ to estimate the standard error in $\tau_N$. Therefore, as shown in Fig. 2.5

$$\left| -e^{-1}/\tau \right| \approx \left| \frac{\text{std}(r_N(\tau_N))}{\text{std}(\tau_N)} \right|$$ \hspace{1cm} (2.33)

$$e^{-1}/\tau \approx \frac{\sqrt{\frac{\tau}{N}(1 - 3e^{-2})}}{\text{std}(\tau_N)}$$ \hspace{1cm} (2.34)

Thus, the uncertainty in $\tau_N$ can be approximated as

$$\text{std}(\tau_N) \approx \sqrt{\frac{\tau}{N}(1 - 3e^{-2}) \cdot \tau \cdot e}$$ \hspace{1cm} (2.35a)

$$\text{std}(\tau_N) \approx e(1 - 3e^{-2})^{\frac{1}{2}} \cdot \tau^{\frac{3}{4}} \cdot N^{-\frac{1}{4}}$$ \hspace{1cm} (2.35b)
Figure 2.5: Schematic depicting the strategy for estimating the uncertainty in the empirical decorrelation time scale $\tau_N$, given the uncertainty in the autocorrelation function $r_N(t)$. First linearizing $r_N(t)$ around the point $(\tau, e^1)$ then estimating the derivative as that of a pure exponential (Gerber et al., 2008)

$$\text{std}(\tau_N) \approx \kappa \cdot \tau^{\frac{3}{2}} \cdot N^{-\frac{1}{2}} \quad (2.35c)$$

where $\kappa = e(1 - 3e^{-2})^{\frac{1}{2}} \approx 2$.

This is however only a rough estimate as the annular mode index is not entirely an AR-1 process and hence $r_N(t)$ is not perfectly exponential. This however is sufficient for the purpose of this work as we are rather concerned with the relative magnitudes of the decorrelation times and not the absolute values.

### 2.2.5 Significance testing

In order to test the statistical significance of the differences between the model control run and various experimental runs, a Student t-test is used. This is a statistical hypothesis test used for determining differences between two populations (e.g. data sets) and the significance of these differences. Once the t-value and the number of degrees of freedom are determined, a table of values of Student’s t-distribution can be used to find the corresponding p-value. Depending on the
chosen statistical significance level (for this work 0.05), the p-value determines if the null hypothesis (two data sets do not differ) is valid or not to a certain probability. The t-test used for this work is a paired t-test which uses lagged autocorrelations and assesses the effective degrees of freedom, \( N_{\text{eff}} \), following Trenberth (1984). The t-value of the difference between two data sets, e.g. \( c \) (control) and \( \text{exp} \) (experiment), is given by

\[
t = \frac{\overline{c} - \overline{\text{exp}}}{\left[ \frac{s_c^2}{N_{\text{eff}}} + \frac{s_{\text{exp}}^2}{N_{\text{eff}}} \right]^{1/2}},
\]

where \( s_c \) and \( s_{\text{exp}} \) are the variances of the two data sets

\[
s_x = \frac{1}{N_x - T_\alpha} \sum_{i=1}^{N_x} (x_i - \overline{x})^2
\]

and \( N_{\text{eff}}^c = N_c/T_\alpha \) and \( N_{\text{eff}}^\text{exp} = N_{\text{exp}}/T_{\text{exp}} \) are the effective number of degrees of freedom of the population of size \( N_c \) and \( N_{\text{exp}} \), respectively. The members of each population are not necessarily independent. This is accounted for by the factor \( T_\alpha \) which reduces the number of degrees of freedom by the amount that is dependent on the autocorrelation within the population (Trenberth, 1984)

\[
T_\alpha = 1 + 2 \sum_{L=1}^{N} \left( 1 - \frac{L}{N} \right) \cdot r_L
\]

where \( r_L \) is the autocorrelation at lag \( L \) and given by

\[
r_L = \frac{\sum_{i=1}^{N-1} (x_i - \overline{x})(x_{i+L} - \overline{x})}{\sum_{i=1}^{N} (x_i - \overline{x})^2}.
\]

The number of effective degrees of freedom (\( N_{\text{eff}} \)) used for the t-test was calculated by (Welch, 1947)

\[
N_{\text{eff}} = \frac{(s_c^2/n_c + s_{\text{exp}}^2/n_{\text{exp}})^2}{(s_c^2/n_c)^2/(n_c - 1) + (s_{\text{exp}}^2/n_{\text{exp}})^2/(n_{\text{exp}} - 1)}.
\]

It was noted in Trenberth (1984) the calculation of \( r_L \) becomes unreliable at large lags. The implications of this were investigated by Simpson (2009) who used the same model. The large amount of variability in the simulations performed with
the IGCM2.2 can lead to strong autocorrelations at large lags that are not a real 
correlation but rather are due to the variability. For runs of 5000 days length 
these autocorrelations can have a significant influence on the number of degrees 
of freedom. Simpson (2009) suggested to calculate the autocorrelation only until 
the lag at which it drops to zero. Trenberth (1984) proposed a more sophisticated 
method, fitting an autoregressive noise model to the autocorrelation, Simpson 
(2009) however found that their method is sufficient for the purpose of signifi-
cance testing used for their work. Despite the fact that the choice of cut-off lag 
does have an effect on the number of degrees of freedom, the significance of the 
difference between their experiments and control is very large in most regions 
so that this has only a slight influence around the edges of the regions of strong 
response. This also applies to the simulations run for this research, therefore, 
the same method as Simpson (2009) is used.
3 Model sensitivity to changes in the surface temperature relaxation timescale

In order to analyse correctly and to interpret IGCM2.2 results from the stratosphere - troposphere interaction studies it is crucial to have a good understanding of the fundamental behaviour of the model. An important aspect of this is its sensitivity to parameter changes. Parameterisations in simplified models such as the IGCM2.2 are based on a combination of current physical understanding of the respective processes and available observations. This provides constraints on the values of a parameter within a range of possible physically plausible values. In this chapter the influence of changing the surface temperature relaxation timescale on the model climatology is investigated. The motivation for choosing this parameter is twofold; Firstly, even though the model in its original formulation produces an overall climatology that is fairly close to that of the real world atmosphere (as shown in Fig. 2.2), an unrealistic cold layer develops near the surface. By clamping the model temperature more strongly to the equilibrium temperature profile (i.e. decreasing the temperature relaxation time) in the boundary layer, a more realistic temperature profile might be achieved. Secondly, as the magnitude of the latitudinal jet variability is of great importance (see Chapter 1.2), the possibility of reducing this variability by relaxing the low-level baroclinicity more strongly is explored. The experiments in this chapter further allow assessment of the sensitivity of previous model results regarding the response to various forcing, to the change of this parameter. First, an overview of the experiment set-up is given. Then, the effect of these changes on the mean climate as well as on the internal variability is discussed. Further, the effect on the response to stratospheric heating is examined. The magnitude of the response to the stratospheric heating perturbations is expected to be related to the timescale of annular variability. This prediction is tested in section 3.4.
3.1 The experiments

Experiments with four (three plus control) different surface temperature relaxation timescales $k_s$ are presented. The coefficient $k_T$ in eqn. 3.1 determines how quickly the temperature is relaxed back to the prescribed equilibrium temperature $T_{eq}$.

$$\frac{\partial T}{\partial t} = -k_T(\phi, \sigma) [T - T_{eq}(\Phi, p)]$$  \hspace{1cm} (3.1)

where $k_T = k_a + (k_s - k_a)\max\left(0, \frac{\sigma_b - \sigma}{1 - \sigma_b}\right) \cdot \cos^4 \Phi$.

It is defined such that above the boundary layer ($\sigma_b = 0.7 \approx 700\text{hPa}$) only weak radiative-convective damping exists. The relaxation time for this is about 40 days ($k_a = 1/40 \text{[day}^{-1}]$). In the boundary layer the surface damping is increased to account for boundary layer effects and to prevent the development of an unrealistic thin cold layer near the surface. Held and Suarez (1994) suggest the boundary layer relaxation time to be set to 4 days at the equatorial surface ($k_s = 1/4 \text{[day}^{-1}]$).

![control run zonal mean temperature](image)

Figure 3.1: 10000 day control run zonal mean temperature distribution.

Whilst the overall structure and magnitudes of the temperatures are in quite good agreement with a real long term mean zonal mean temperature distribution, the profile in the lower tropics (area highlighted in Fig. 3.1) is not realistic.
Here the zonal mean temperature distribution exhibits an unphysical structure - almost like an additional temperature inversion layer. It is clear that the surface relaxation time should be decreased in the boundary layer, however there is no physical justification for it to be exactly 4 days.

An overview of the experiments conducted for the first part of this investigation is given in Table 1. All experiments are run for 11000 days with the first 1000 days discarded as spin-up and with an integration time step of 30 minutes. Two of the experiments, B1 and B2, have shorter surface temperature relaxation times than the control (4 days) where as experiment, B3, has a longer relaxation time.

<table>
<thead>
<tr>
<th>experiment</th>
<th>$k_a$ [day$^{-1}$]</th>
<th>run length [days]</th>
<th>time step [minutes]</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1</td>
<td>1/2</td>
<td>10000</td>
<td>30</td>
</tr>
<tr>
<td>B2</td>
<td>1/3</td>
<td>10000</td>
<td>30</td>
</tr>
<tr>
<td>control</td>
<td>1/4</td>
<td>10000</td>
<td>30</td>
</tr>
<tr>
<td>B3</td>
<td>1/8</td>
<td>10000</td>
<td>30</td>
</tr>
</tbody>
</table>

Table 1: Overview of the configurations for each surface temperature relaxation timescale experiment.

These four set-ups are then used in conjunction with a heating perturbation applied to the lower stratosphere region. The temperature perturbation used for this is the preferentially equatorial heating, E5, previously used by Haigh et al. (2005), Simpson (2009) and Simpson et al. (2009). E5 denotes a maximum increase of 5 K in the equatorial stratosphere decreasing to 0 K at the poles. The difference in the relaxation temperature profile, $T_{eq}$, between this perturbation and the control run are shown in Fig 3.2. The stratospheric $T_{eq}$ is altered but the tropospheric $T_{eq}$ remains unchanged.

The heating perturbation is applied by simply changing the parameters that determine $T_{eq}$.
\[ T_{eq} = \max \left\{ \left( T_{\text{tpeq}} + \Delta T_{tp} \cos^2 \phi \right), \right. \]
\[
\left. \left[ T_0 - \Delta T_y \sin^2 \phi - \Delta \theta_z \cos^2 \phi \cdot \log \left( \frac{p}{p_0} \right) \right] \left( \frac{p}{p_0} \right)^n \right\} \tag{3.2} \]

where, for the E5 heating, \( T_{\text{tpeq}} = 200 \text{ K} \) and \( \Delta T_{tp} = 5 \text{ K} \). All other parameters are as described in section 2.1.

This heating perturbation was chosen as it is this forcing that was most intensively investigated (Haigh et al. 2005; Simpson, 2009 and Simpson et al., 2009). One focus of their research was the investigation of the mechanisms by which the tropospheric response to solar activity is produced. The E5 heating pattern is a crude idealisation of that seen over the solar cycle. The magnitude (5 K at equator) of the applied temperature perturbation is however, considerably larger than that observed (\( \approx 1.2 \text{ K} \)) (Haigh, 2003; Labitzke et al., 2002). The results of Haigh et al. (2005), however suggest that the mechanisms involved in the tropospheric response are independent of the heating magnitude. Therefore Simpson (2009) and Simpson et al. (2009) used this larger magnitude heating to obtain statistically significant results for short integration periods. Comparing their E5 results with the surface temperature relaxation time plus E5 experiments allows assessment of the sensitivity of these results to a change of temperature surface
relaxation $k_s$.

Again, all E5 experiments for this work have been run for 10000 days (plus the initial 1000 days discarded as spin-up). The model runs in Haigh et al. (2005) and Simpson (2009) and Simpson et al. (2009) are for 1000 days and for 5000 days, respectively. Both discarded additional 200 days as spin-up. For this work longer equilibrium runs have been chosen as, due to the variability in the model, even a 5000 day run may not be of sufficient length for accurate determination of the magnitudes of response (Simpson et al., 2010). For the control, the B1 and B3 experiments runs of 30000 day length have been performed to test the accuracy of the results. It has been found that the longer runs do not give significantly different magnitudes than the 10000 day experiments.

3.2 The effect on the mean climate

First the effect of the changes in the surface temperature relaxation time on the mean climate is discussed. Figure 3.3 which shows the zonal mean temperatures for all experiment runs and Figure 3.4 the zonal mean temperature and zonal mean zonal wind for the control ($k_s = 1/4 \text{[days}^{-1}]$) and the differences between each experiment run and the control. The fields are averaged over both hemispheres, thus only one hemisphere is shown. Regions where the difference between the experiments and control run is not statistically significant at the 95% level are shaded in grey (calculated by using the Student t-test).

It is expected that a change of the surface temperature relaxation time alters the static stability and the baroclinicity on the equatorward side of the mid-latitude jet in the boundary layer. These changes are expected to increase the growth rates of baroclinic waves on the equatorward flank of the jet.

It is clear from the differences in the zonal mean temperatures (LHS of Fig. 3.4) that changing $k_s$ influences the low level baroclinicity as would be expected. A faster surface temperature relaxation, i.e. a stronger relaxation of the temperatures to the equilibrium temperature profile $T_{eq}$ in the tropics, results in a warming (up to $\approx +4 \text{K}$) of the equatorial surface region (B1 and B2) while a
Figure 3.3: Zonal mean temperature fields for the $k_s$ experiments. The contour interval is 5 K.

Weakening of the relaxation leads to a cooling (up to $\approx -5$ K) of the same region. This is also apparent from Fig. 3.3; the unrealistic cooler surface layer previously observed in the control run disappears for B1 ($k_s = 1/2 \text{[days}^{-1}\text{]}$) and is weakened for B2 ($k_s = 1/3 \text{[days}^{-1}\text{]}$) while it is increased for B3 ($k_s = 1/8 \text{[days}^{-1}\text{]}$). The influence of $k_s$ on the temperatures above the boundary layer or poleward of about 50deg is small ($\leq 1.5$ K), but still significant.
Figure 3.4: Temperature (LHS) and mean zonal mean wind (RHS) of the control run (top panel) and the differences between the experiments and the control run (lower panels). The contour intervals for the control are 5 Kelvin and 1 m/s, respectively, and for the differences 1 K and 1 m/s, respectively. The shaded areas are deduced to be less than 95% significant using a Student t-test.
Associated with the changes in the surface temperature structure is a change of the latitudinal position of the mid-latitude jet. The changes in zonal wind and zonal wind shear (du/dz) throughout the troposphere, and related changes in surface pressure, are associated with the jet displacements. The differences in the zonal mean zonal winds (RHS of Fig. 3.4) show a strengthening of the zonal winds on the equatorward side of the jet maximum and a weakening on the poleward side for the B1 and B2 experiments, indicating an equatorward shift of the jet. The opposite is true for the B3 experiment. The changes in the jet positions are related to the changes in the low level baroclinicity via the thermal wind relation. An increase in the surface baroclinicity at a certain latitude region leads to a strengthening of the source of eddies at that region. As the eddies propagate away from that latitude, the momentum flux towards this region increases and therefore the eddy-driven jet tends to exist at the latitude of the maximum baroclinicity. For the B1 and B2 experiments the low-level baroclinicity is increased at lower latitudes and weakened at higher latitudes (compared to the control run) resulting in a shift of the eddy source region towards lower latitudes and hence an equatorward shift of the eddy-driven jet.

The weaker surface temperature relaxation in experiment B3 results in a low-level baroclinicity that is weaker at low latitudes and stronger at high latitudes (compared to the control run), resulting in a poleward shift of the jet. The zonal winds change throughout the troposphere in response to altering $k_s$ in the lower troposphere only. The change in the eddies induced by the altered low level baroclinicity is hence changing the winds throughout the troposphere. The zonal wind anomaly patterns (RHS of Fig. 3.4) are remarkably similar to those in e.g. Haigh et al. (2005) and Simpson et al. (2009). This implied that the same dynamical feedbacks are triggered in the $k_s$ experiments. The eddy feedback by baroclinic mechanism (see Fig. 1.6) seems to be a robust response independent of the nature of the forcing or the location where it is applied.

The change in zonal mean zonal winds is most pronounced for a halving of the surface temperature relaxation timescale. The position of the mid-latitude jet
versus the surface temperature relaxation timescale for all experiments as well as for the control run is plotted in Figure 3.5. It is apparent that a strong linear relation exists between $k_s$ and the jet position.

The jet latitudes given in Figure 3.5 are the latitudinal positions ($\phi_{jet}$) of the eddy driven jet for the control and experiment runs. They are calculated following the method of Kidston and Vallis (2010). $\phi_{jet}$ is taken as the latitude of the maximum zonal mean zonal winds at the level closest to the surface (i.e. 967hPa). The value of $\phi_{jet}$ is determined by fitting a quadratic to the zonal wind $u$ between the two latitudes either side of the maximum. This method is also used by Kidston and Gerber (2010).

Figure 3.5: Latitudinal positions of the eddy driven jets for the control and experiment runs. The error bars indicate one standard error.

The effects of $k_s$ on the mean climate suggest that a shorter relaxation timescale leads to both a more realistic temperature profile in the equatorial surface region as well as to a more equatorward position of the mid-latitude jet. While the former might be desirable, previous research (see section 1.2.4) suggests that the shift in the jet position is likely to affect the timescales of internal variability. This is investigated in the following section.
3.3 The effect on annular variability

One motivation for studying the model sensitivity to variations of the surface temperature relaxation timescale $k_s$ was to investigate if altering $k_s$ would lead to a reduced latitudinal jet variability. The idea behind this was that a stronger relaxation of the low-level baroclinicity might be a way to clamp the jet position. The main mode of variability for the mid-latitude jet is a fluctuation in latitude about its average position. Figure 3.6 shows the standard deviations from the mean of the 10000 day control and experiment runs as a measure of the latitudinal jet variability.

![Figure 3.6: Latitudinal jet variability: The plots show the standard deviations from the mean of the 10000 day runs (B1, B2, control and B3). The contour interval is 0.5 m/s.](image)

It is clear from this figure that reducing the surface relaxation time does not lead to a reduction in variability. In fact, the experiment with the shortest relaxation time (B1, $k_s = 1/2$ [days$^{-1}$]) exhibits the strongest variability while the variability is reduced for longer relaxation times (control, B2 and B3). This however does not suggest that the change in low-level baroclinicity which is brought about
by altering $k_s$ causes the observed changes in latitudinal jet variability. Rather it indicates that any direct influence $k_s$ might have on the jet variability is small compared to the effect the jet position in latitude has on its variability. As discussed in section 1.2.5. the timescale of variability has been found in previous studies to be strongly connected to the position of the mid-latitude jet.

To investigate the natural variability of the zonal mean flow, a Principal Component Analysis has been conducted. The leading mode of variability is found to be a latitudinal displacement of the jet while the second mode of variability is found to be a variation in jet strength and width. This is shown to be true for all runs. Figure 3.7 shows the first two empirical orthogonal functions (EOFs) of zonal-mean zonal wind. The colors indicate positive (red) and negative (blue) anomalies. Together the first two EOFs explain between 65% and 80% (44.59 - 65.77% from EOF1 and 15.05 - 20.21% from EOF2) of the variance. The EOF patterns are shown in metres per second by projecting the anomaly data onto the normalised principal component (PC) time series. For each of the experiments EOF1 (left hand side of Fig. 3.7) shows a deep structure throughout the troposphere and represents latitudinal displacement of the mid-latitude jet relative to its mean position for each run (mean jet positions are given in Table 2). A strong similarity in the EOF1 structure can be seen for all experiments. A secondary effect is the increased EOF1 variance percentage for more equatorward jets. EOF2 (right hand side of Fig. 3.7) represents both broadening/narrowing and weakening/strengthening of the mid-latitude jet.

The natural variability of the zonal mean flow can be characterised by the decorrelation timescale $\tau$ of the leading mode time series, PC1 (see sections 2.2.3 and 2.2.4). The results of the calculations of $\tau$ for the control and experiment runs are given in Table 2.
Figure 3.7: The first two EOFs of the zonal mean zonal wind $u$ for each run. The first EOF is shown on the LHS and the second EOF on the RHS. The contour interval is 0.5 m/s and the percentage variance explained by each EOF is given in the figure title. The red and blue coloring indicate positive and negative zonal wind anomalies, respectively.
It is apparent that the timescale of the annular variability is influenced systematically by the choice of $k_s$. This is also demonstrated in Fig. 3.8 where the decorrelation times $\tau$ for control and experiment runs are plotted against the surface relaxation temperature timescales $k_s$. It was found that the faster the relaxation timescale, the longer the timescale of annular variability. This is the opposite of what would be expected if the surface temperature relaxation timescale was the only factor influencing the timescale of annular mode variability.

Figure 3.8: Plot of zonal mean zonal wind PC1 decorrelation time $\tau$ (days) vs. surface relaxation timescale $k_s$ (day$^{-1}$) for the control run and experiments. The error bars for the decorrelation time show one standard deviation estimated as described in section 2.2.4.

It has to be noted that the decorrelation times for these experiments as well as for the control run are significantly longer than the decorrelation times that characterize annular modes in the real world atmosphere. These were observed to be about 10-20 days (e.g. Feldstein, 2000; Baldwin et al. 2003). This is common to many other studies using simplified GCMs as features that can reduce the
decorrelation times, such as orography (i.e. planetary waves) and seasonality, are absent in the models. This indicates that knowledge and understanding of the decorrelation times is also very important for studies using more complicated GCMs with the aim to draw conclusions about the magnitudes of response to forcings (e.g. Climate Change CO₂ studies).

<table>
<thead>
<tr>
<th>experiment</th>
<th>( k_a ) [day(^{-1})]</th>
<th>mean jet position [deg]</th>
<th>decorrelation time ( \tau ) [days]</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1</td>
<td>1/2</td>
<td>41.1 ± 0.4</td>
<td>165 ± 30</td>
</tr>
<tr>
<td>B2</td>
<td>1/3</td>
<td>43.9 ± 0.3</td>
<td>120 ± 19</td>
</tr>
<tr>
<td>control</td>
<td>1/4</td>
<td>45.4 ± 0.2</td>
<td>55 ± 6</td>
</tr>
<tr>
<td>B3</td>
<td>1/8</td>
<td>47.7 ± 0.1</td>
<td>30 ± 2</td>
</tr>
</tbody>
</table>

Table 2: Overview of the configurations and results for each surface temperature relaxation timescale experiment. The errors for the jet positions are one standard error and the errors for the decorrelation time standard deviation.

Further, a strong relation between the position of the mid-latitude jet and the decorrelation times is found (see Fig. 3.9). For this figure the decorrelation times \( \tau \) are plotted against the mean jet positions of the mid-latitude jet for each model run. It is apparent that the decorrelation times decrease approximately linearly with increasing jet latitude. This result is qualitatively consistent with previous studies using simplified GCMs (e.g. Son and Lee, 2005; Gerber and Vallis, 2007; Gerber at al. 2008; Simpson et al. 2010). All those consistently showed that the decorrelation timescale \( \tau \) of the annular mode variability depends on the position of the mid-latitude jet where configurations with a jet at lower latitudes exhibit longer decorrelation times.

In order to further investigate the difference in decorrelation times between the different runs, times series of the zonal wind anomalies at the level closest to the jet maximum (240 hPa) were obtained. These are given in Figure 3.10, where negative zonal wind anomalies are shown in blue and positive anomalies in red. The top panel shows the control run result while the lower panels are sorted by in-
creasing surface relaxation time from top to bottom (i.e 2 - 3 - 8 days; B1, B2 and B3). Two distinct types of behaviours can be identified; a propagating behaviour of lower amplitude and a quasi-stationary behaviour of higher amplitude. Typical examples of both behaviours can be seen e.g. in the first panel: day 1000-2000 for stationary and day 4000-5000 for quasi-propagating behaviour. The control run time series shows both propagating and quasi-stationary behaviour. The anomaly time series for the B1 experiment \( (k_s = 1/2 \, [\text{day}^{-1}]) \) exhibits mainly

![Plot of zonal mean zonal wind PC1 decorrelation time \( \tau \) (days) vs. jet position (deg) for the control run and experiments. The error bars for the decorrelation time show one standard deviation estimated as described in section 2.2.4. The error bars for the jet positions show one standard error.](image)

quasi-stationary behaviour while the B3 run \( (k_s = 1/8 \, [\text{day}^{-1}]) \) shows almost exclusively propagating behaviour and B2 \( (k_s = 1/3 \, [\text{day}^{-1}]) \) a more balanced mix of both behaviours. For the control run and B3 (and partly B2) the timescale of zonal wind variability is mostly short with the anomalies from the time mean being relatively small and poleward propagating. During some parts of the times series the jet appears to enter into an anomalous regime where there are large
Figure 3.10: Daily zonal mean zonal wind anomalies from the time mean on the 240 hPa level for control run (upper panel) and experiment runs (lower panels). Red indicates positive zonal wind anomalies and blue indicated negative anomalies.
zonal wind anomalies on either side of the time mean jet for extended periods of time. This is true especially for the B1 run where first a long time with a large negative zonal wind anomaly on the equatorward side of the time mean jet and a positive anomaly on the poleward side (e.g. days ≈1000 to ≈2000) and then vice versa (e.g. days ≈2500 to ≈4000) is observed (Fig. 3.10). These different types of variability, propagating and quasi-stationary, have previously been observed in models as well as in the real world atmosphere (e.g. Son and Lee, 2006; Codron, 2007; Simpson et al., 2010; Sparrow et al., 2009). It is however not clear why the jet goes into these anomalous states and understanding the different behaviours associated with the quasi-stationary/propagating zonal wind anomalies is a subject of ongoing research (e.g. S. Sparrow, personal communication).

Chan and Plumb (2009) used a simplified model to investigate the dependence of the decorrelation times on the state of the troposphere by varying the model’s equilibrium temperature profile (both altering the strength of the stratospheric polar vortex and the seasonal asymmetry). This was mainly motivated by the very long decorrelation times found by Polvani and Kushner (2002). Some of their experiments exhibited a bimodality in the tropospheric eddy-driven jet where the jet had a preference for two distinct regions, the subtropics and mid-latitudes. Model configurations (like the one used by Polvani and Kushner, 2002) that result in a bimodal jet have very long (> 200 days) decorrelation times with the jet persisting in one region for thousands of days before switching to another. Applying only small changes to the tropospheric state was shown to remove the bimodality in the zonal-mean zonal jet and hence to significantly reduce the decorrelation timescales (< 70 days).

While the sensitivity studies performed here show the existence of two different behaviours of variability (propagating and quasi-stationary), the decorrelation time has been shown to increase for lower jet latitudes and no such bimodality of the jet has been found.
The occurrence of long decorrelation times (> 100 days) for model configurations with short surface relaxation times suggests that a bimodality of the jet as found by Chan and Plumb (2009) is not the only factor that can lead to long decorrelation times.

3.4 The effect on the response to stratospheric heating

In this section the effect of changes to the surface temperature relaxation time scale on the response to stratospheric heating perturbations are explored. As described in section 3.1, 10000 day equilibrium runs have been performed for E5 experiments on the control run and each of the experiment runs (B1-3). It is expected that all these stratospheric heating experiments qualitatively follow the results obtained by Haigh et al. (2005) (as well as Simpson, 2009 and Simpson et al., 2009). The differences in latitudinal position of the jet, which is influenced by the choice of $k_s$, allow investigation of the relation between the magnitude of the response and the jet position of the unperturbed run. From theory (Fluctuation-Dissipation Theorem) a stronger response is expected for more equatorial positions of the eddy driven jet.

Figure 3.11 shows the differences in zonal mean temperature and zonal mean zonal wind between the perturbed and unperturbed runs. The results are shown in order of increasing surface temperature relaxation time (i.e. B1 E5 - B1, B2 E5 - B2, E5 - control and B3 E5 - B3). The areas shaded in grey show the regions where the difference between the equilibrium and control runs are not statistically significant at the 95% level (calculated using the Students t-test).

The response to the applied stratospheric heating shown in Figure 3.11 is qualitatively similar to the results from Haigh et al. (2005), Simpson (2009) and Simpson et al. (2009). For the control run the results are also quantitatively similar to the results from Simpson (2009) and Simpson et al. (2009). Haigh et al. (2005) performed run of 1000 days length while Simpson (2009) and Simpson et al. (2009) used 5000 day runs, both used the same model, the IGCM2.2, as was used for this work. Differences in the magnitude of the response arise from
Figure 3.11: Differences between perturbed (E5) and unperturbed experiments in zonal mean temperature (left panels) and zonal mean zonal wind (right panels) for all E5 experiments. The results are shown in order of increasing surface temperature relaxation time from top to bottom. The contour intervals are 0.5 K (left panels) and 0.5 m/s (right panels). Shaded areas are not statistically significant at the 95% level.
the internal variability and Simpson (2009) and Simpson et al. (2009) showed that the magnitude of response resulting from a 1000 days run are not representative. They even raised doubts as to whether 5000 day runs are sufficient to accurately determine this. The qualitative patterns of response however, have been found to be robust even in 1000 day runs. For this work tests have been performed with runs of 30000 day length for the control run as well as for the extremes of the experiments, B1 and B3. Both the magnitude of response (to \(k_s\) changes) and the decorrelation times were found to be consistent with the values obtained by the 10000 day runs. Therefore it was decided that runs of 10000 day length would be sufficient for this work. Especially, as the relative magnitudes of the differences between the control and various experiments are of interest for this work rather than obtaining absolute values.

All E5 experiments show a heating in the stratosphere preferentially at the equator as applied (left panels in Fig. 3.11). The heating response is however not exactly mirroring the E5 heating perturbation. There is, for example, a larger increase of temperature at the polar regions in the upper stratosphere than originally imposed (i.e. the imposed \(\Delta T\) is zero at the pole). In the lower stratospheric polar regions a negative temperature difference is found, i.e. a cooling. These changes are brought about by both mean circulation changes arising from the heating perturbation and eddy feedbacks (as discussed by Haigh et al., 2005 and Simpson (2009) and Simpson et al., 2009).

All of the E5 experiments also exhibit regions of statistically significant warming in the troposphere, with B1 (\(k_s = 1/2 \text{[days}^{-1}\)])) showing the largest warming. The tropospheric temperature response consists of a banded increase in temperature in the mid-latitudes (centered on the jet) with a decrease on either side. These temperature changes are in thermal wind balance with the zonal wind response. For each E5 experiment shown in Figure 3.11 it is apparent that a heating perturbation applied only to the stratosphere also results in a significant response in the tropospheric circulation (right panels). In each case the zonal
mean zonal wind response consists of a decreased westerly wind on the equatorward side of the mid-latitude jet and an increase on the poleward side resulting in a poleward shift in the mid-latitude jet. This is accompanied by a westerly wind anomaly in the subtropics. Again B1 shows the strongest response and B3 the weakest, i.e. the longer the surface temperature relaxation time the weaker the response. Haigh et al. (2005) found this pattern of response in zonal mean temperatures and zonal mean zonal winds to be similar to that reported to results from increased solar activity in NCEP–NCAR data (Haigh, 2003) as well as in full GCM studies (Haigh, 1996 and 1999; Larkin et al., 2000).

Simpson 2009 and Simpson et al. (2009) found that changing eddy momentum fluxes are important for maintaining the anomalous circulation and also crucial for creating it. Figure 3.12 shows the differences in horizontal eddy momentum fluxes, horizontal eddy heat fluxes and the streamfunction of the mean meridional circulation between the perturbed and unperturbed runs. The results are shown in order of increasing surface temperature relaxation time (i.e. B1 E5 - B1, B2 E5 - B2, E5 - control and B3 E5 - B3). The contour intervals are 2.0 m²/s² (upper panels), 0.5 Km/s (middle panels) and 10¹⁰ kg/s (lower panels). The areas shaded in grey show the regions where the difference between the equilibrium and control runs are not statistically significant at the 95% level (calculated using the Students t-test). The poleward eddy momentum flux anomalies (upper panels) show a dipole structure, with a weakening of the fluxes on the equatorward side of the time mean jet and a strengthening on its poleward side that stretches downward and equatorward so that in the troposphere the maximum $u'v'$ is around the jet centre. This is visible in all E5 experiments, though the relative magnitude of the weakening/strengthening varies. For short surface temperature relaxation times only a small weakening of the momentum fluxes on the equatorward side of the jet is observed, while the strengthening on the poleward side is very pronounced. This changes with increasing relaxation time, where the magnitudes in the B3 E5 experiment ($k_s = 1/8$ [day⁻¹])
are of similar size for both the weakening as well as the strengthening of the eddy momentum fluxes. Haigh et al (2005) found that the anomalous meridional circulation locally counteracts the anomalous eddy momentum forcings and its effects are balanced by the anomalous horizontal eddy heat fluxes that act to damp the temperature anomalies. This is also found to be true for each of the experiments performed here (Fig. 3.12 middle panels for horizontal eddy heat flux anomalies and mean meridional circulation anomalies shown in the lower panel).

Figure 3.12: Differences between perturbed (E5) and unperturbed experiments in horizontal eddy momentum fluxes (upper panels), horizontal eddy heat fluxes (middle panels) and streamfunction of the mean meridional circulation (lower panels) for all E5 experiments.
The strongest effect on both the horizontal eddy heat fluxes and the mean meridional circulation are again seen for the B1 E5 ($k_x = 1/2$ [day$^{-1}$]) experiment which is consistent with this experiment also showing the strongest jet shift.

![Graph showing jet shift vs. PC1 decorrelation time](image)

Figure 3.13: Zonal mean zonal wind PC1 decorrelation time $\tau$ (days) of the unperturbed runs vs. jet shift (deg) in response to heating.

The results show that there is a clear trend for a stronger response to the stratospheric heating perturbation for experiments with shorter surface temperature relaxation times. These experiments also exhibit more equatorward mid-latitude jets and longer decorrelation times (as shown in section 3.2 and 3.3.). From the Fluctuation-Dissipation Theorem it would be expected that runs with longer decorrelation times respond more strongly to external forcings like the applied heating perturbation (see section 1.2.4). This is also observed for the control and B heating experiments. Figure 3.13 shows a clear linear relation between the magnitude of the jet shift and the decorrelation time of the unperturbed run, i.e. this is in agreement with the expectations from Fluctuation-Dissipation theory.
3.5 Discussion and Conclusion

The objective of the experiments in this chapter was to investigate the influence of changing the surface temperature relaxation timescale \(k_s\) on the model climatology as well as to study its effect on the response to a stratospheric heating perturbation. The motivation for this was to improve the understanding of the sensitivities of the model as this is of great importance for the interpretation of stratosphere-troposphere coupling studies undertaken with this model. Specifically studying the effect of variations in the surface temperature relaxation timescale was chosen as this parameter influences the low level baroclinicity and was expected to have an effect on the latitudinal jet variability which is subject of many recent studies (e.g. Ring and Plumb, 2007 and 2008; Gerber et al. 2008; Chan and Plumb 2009, Chan et al. 2009, Barnes et al., 2010, Simpson et al. 2011).

A set of model configurations with varying \(k_s\) were run for 10 000 days each and analysed. Further, an E5 heating perturbation has been applied to all runs of this set to investigate the influence of \(k_s\) on the response to this perturbation.

In this chapter is was shown that changes to the model’s surface temperature relaxation timescale have a significant impact on several aspects of model behaviour:

- The model climatology:
  
  Both the thermal structure of the lower troposphere and the position of the mid-latitude jet are influenced by the choice of \(k_s\).
  
  It was expected originally that a shorter surface relaxation time might improve the temperature structure here as with the previously used configuration an unrealistic cold layer was observed. The results confirm that it is indeed possible to improve the temperature structure at the surface by using shorter relaxation times. Longer surface relaxation times (e.g. the B3 experiment with \(k_s = 1/8 \, \text{day}^{-1}\)) results in an intensifying of the cold layer.
The mid-latitude jet is found to be located at different latitudes for different choices of $k_s$. A strong temperature relaxation at the surface is shown to lead to a more equatorward jet while weaker relaxations result in more poleward jets. The changes in the jet positions are related to the changes in the low level baroclinicity and static stability via the thermal wind relation. Even though $k_s$ is altered in the lower troposphere only, the zonal winds change throughout the troposphere. The remarkable similarity of the zonal wind anomaly patterns with those in e.g. Haigh et al. (2005) and Simpson et al. (2009) implies that the same dynamical feedbacks are triggered in the $k_s$ experiments. The eddy feedback mechanism seems to be a robust response independent of the nature of the forcing or the location where it is applied.

- The natural variability of the mid-latitude jet:
  It was expected that the latitudinal jet variability could be reduced by relaxing the low-level baroclinicity more strongly. The model response in fact was found to be in the opposite sense. This can be explained by the change in jet position related to the change in $k_s$; While a stronger damping of the temperatures might still reduce the latitudinal jet variability, any such effect is small compared to the effect the mean jet position has on the variability. As changing the surface temperature relaxation timescale has a significant effect on the jet position, the variability was investigated with respect to this. It was found that model configurations with lower latitude jets exhibit longer timescales of annular variability than when the jet is located at higher latitudes. This is in agreement with previous research (e.g. Son and Lee, 2005; Gerber and Vallis, 2007; Gerber et al. 2008; Simpson et al. 2010). Further it was found that the mid-latitude jet exhibits a combination of two distinct behaviours, propagating and quasi-stationary. This occurrence of two distinct behaviours was also found by previous studies (e.g. Son and Lee, 2006; Codron, 2006 Simpson et al., 2009; Sparrow et al., 2009) and is subject to current research (e.g. S. Sparrow, personal commu-
nication). The low amplitude propagating behaviour is found to strongly dominate the zonal wind anomaly time series for the highest latitude jet (B3), while the lowest latitude jet (B1) exhibits mostly high amplitude quasi-stationary behaviour. The model runs (control and B2) with jets between these extremes show a more balanced mixture of both behaviours. The difference in zonal wind anomaly amplitudes explains the amplitude dependence in the zonal wind standard deviations (Fig. 3.6) which was opposite to the expectations.

In order to answer the original question whether the latitudinal jet variability could be reduced by relaxing the low-level baroclinicity more strongly, an altered experiment is proposed for future studies. A strengthening of the boundary layer relaxation without moving the jet, might be possible if not only $k_s$ but also both $\Delta T_y$ and $T_0$ (in eqn. 2.6) were changed to obtain the same polar $T_{eq}$ and a similar $T(\phi = 0, \sigma_{NL})$ as in the control experiment. However, as $k_s \propto \cos^4 \phi$ and $T_{eq} \propto \cos^2 \phi$, the design of such an experiment is not trivial.

- The magnitude of the response to a heating perturbation applied to the stratosphere:

The patterns of response were found to be remarkably similar to those observed by Haigh et al. (2005) and Simpson (2009) and Simpson et al. (2009). A stronger response is found when the longer timescale quasi-stationary type of variability dominates, i.e. where the jet is located more equatorward. A clear linear relation between the decorrelation times of the unperturbed run and the magnitude of poleward jet shift in response to the heating was found. This is in agreement with the Fluctuation- Dissipation Theorem. The importance of the jet latitude for the magnitude of response is subject of ongoing research (e.g. Simpson et al., 2011) to identify the mechanisms that lead to the differences in response.

The findings of the study of the model sensitivity to changes in the surface temperature relaxation timescale discussed in this chapter were shown to be impor-
tant for understanding and interpreting the results of stratosphere-troposphere interaction experiments performed with the IGCM2.2. Further it was found that, while changing $k_s$ can improve the low-level temperature structure, this advantage is clearly outweighed by the disadvantage of obtaining much more unrealistic decorrelation timescales associated with a more equatorward jet position.

A very important finding of this research is the remarkable similarity of the tropospheric response for both the studied $k_s$-experiments and the previously investigated stratospheric temperature forcings (e.g. Haigh et al., 2005; Simpson, 2009 and Simpson et al., 2009). Both perturbations to the basic state, whether from the equatorial stratospheric temperature (E5) or from low level sub-tropical temperature ($k_s$), result in tropospheric responses with close similarity to annular mode variability. This indicated that the same dynamical feedbacks are triggered (see Fig. 1.6). E5 triggers a refraction feedback in the upper troposphere, while the $k_s$ experiments trigger baroclinic feedback in the lower troposphere. The combined feedbacks must be identical, yielding the same annular mode like response.

It has further been shown that the magnitude of the tropospheric response to temperature perturbations applied to the lower stratosphere strongly depends on the position of the mid-latitude jet and that this is influenced in a systematic way by the choice of the surface temperature relaxation timescale. While the dependence of the magnitude of response on the jet latitude has been shown previously (e.g. Simpson et al., 2011), the influence of the surface relaxation timescale on the jet latitude and thereby on the magnitude of response has, to the author’s knowledge, not previously been investigated.

These key findings provide an important framework for the further research conducted for this thesis: knowledge of the mid-latitude jet position is crucial for correctly interpreting comparisons of experiments conducted with different versions of the IGCM2.2. This is important for both designing a vertically extended model (Chapter 4) and for directly comparing tropospheric responses to stratospheric forcings applied to the original IGCM2.2 and the extended version.
(Chapter 5). Further, using varying surface temperature relaxation timescales as a tool to shift the mid-latitude jet allows investigation of the tropospheric response in different model versions in more detail (Chapter 5).
4 Vertical extension of the IGCM2

A new 26 level version of the IGCM2.2 is introduced in this chapter. First a motivation for this vertical extension is given. This is followed by a presentation of each individual update to the model that was necessary to achieve stable model runs and its effects on the model results. Further, the full L26 model results and features are presented. The chapter finishes with a general discussion of the updated model and results and conclusions drawn from this.

4.1 Motivation

Using simplified models like the IGCM2.2 to investigate the tropospheric response to stratospheric forcings is of great importance for improving the understanding of stratosphere - troposphere interactions. However, due to the low vertical extent (only up to $\approx 30\text{km}$) of the previously used version of the IGCM2.2, only forcings very close to (and crossing) the tropopause could be investigated. While it is very important to understand how changes to the atmosphere at about this level are communicated to and throughout the troposphere, it is also of interest to investigate the effects of forcings applied higher up in the stratosphere. Extending the model to include the whole stratosphere and hence making it possible to retract forcings away from the tropopause helps to improve the understanding of the previous model results to temperature perturbations as well as to include perturbations that are closer to those observed in the real atmosphere. Figure 4.1 and 4.2 show two examples for such stratospheric perturbations.

Fig. 4.1 shows the difference between solar maximum and solar minimum of annually averaged zonal mean temperature (K) from the ERA-40 temperature analysis (Frame and Gray, 2010). A maximum temperature response of about 1.2 K occurs at about 1hPa (40 - 50 km) in the equatorial region between about 20 S and 20 N. A further maximum (\(\approx 1\text{K}\)) is seen in the lower stratosphere between 20 km and 30 km stretching slightly further poleward from about 45 S to
Figure 4.1: Annual averaged estimate solar maximum minus minimum signal (K) from the ERA-40 temperature analysis for the period 1979-2001. Light and dark shading indicates statistical significance levels at the 95% and 99% levels (Frame and Gray, 2010).

45°N. The original IGCM2 has 15 vertical levels going up to 18.5 hPa (about 30 km). Hence the vertical structure of the solar forcing cannot be fully resolved with this model version. An extended IGCM2 could however be used to investigate the effects of temperature perturbations similar to this pattern. Further the extended model could be used to study the tropospheric response to other stratospheric forcings, for example, a quasi-biennial oscillation (QBO)-like momentum forcing (Fig. 4.2). Such investigations are examples of the possible future applications of the extended model discussed in this chapter. The impact of variations in the position and shape of heating perturbations imposed on the stratosphere using the extended L26 IGCM2.2 version will be discussed in Chapter 5.

Further, the vertical extension of the IGCM2 makes it possible to compare the original “low-top” model results with the new “high-top” results both for unforced control runs and for perturbation experiments. This might help assessment of the importance of resolving the stratosphere (e.g. Shaw et al., 2009; Sigmoid et al. 2008; Hardimann et al., 2008). It is however important to note that due to
the limited stratospheric dynamics in the new L26 IGCM2, "low-top" - "high-top" comparisons using the results of this model are only relevant to cases with limited stratospheric dynamics (e.g. in the summer hemisphere).

Figure 4.2: QBO signal: (top) time-height section of the monthly-mean zonal mean wind (m/s) for 1964-1990, with seasonal cycle removed. The contour interval is 6 m/s, with the band between ± 3 unshaded. Red represents positive (westerly) winds and blue negative (easterly) winds. The bottom panel shows the same data after applying a 9-48 month band-pass filter. Figures from Baldwin et al. 2001.

4.2 Additions to the model

Several addition to the model were necessary in order to achieve stable runs and a reasonable representation of the stratosphere. It was found that a simple change in the vertical extent of the model resulted in it becoming unstable and failing to complete the runs. Two numerical improvements were made to increase the stability; a change in integration time step from 30 minutes to 15 minutes and an added $\nabla^2$-diffusion for the top three model levels. Further the model physics was adjusted to account for the presence of a stratosphere; both the temperature relaxation timescale and the equilibrium temperature profile for the stratosphere were altered accordingly. These updates will be outlined in the following sections.
This is then followed by an investigation into the effects each of these additions have on the model climatology and variability.

4.2.1 L15 to L26

The first change to the model was to exchange the previously used 15 vertical levels between the surface and 18.5 hPa (about 30 km) with 26 levels between the surface and 0.1 hPa (about 80 km). The model levels for L15 and the new L26 are at (see Fig. 4.3)

- L15: 18.5, 59.6, 106, 152, 197, 241, 287, 338, 400, 477, 569, 674, 784, 887 and 967 hPa
- L26: 0.1, 0.4, 1.1, 2.2, 4.2, 7.5, 12.7, 20.5, 31.3, 45.6, 63.5, 85.2, 110.8, 140.9, 175.9, 216.7, 263.9, 318.5, 381.7, 453.7, 534.1, 621.4, 712.4, 802.9, 888.3 and 964.6 hPa.

The vertical levels from L26 were previously used in the IGCM3 (e.g. Bell et al., 2009) which includes a more sophisticated radiation scheme, moist/clouds parameterisations as well as real-Earth surface conditions.

Figure 4.3: Pressure levels for L15 (red) and L26 (blue).
4.2.2 Change in time step

It was found that the model lost its numerical stability when simply changing from L15 to L26 and would blow-up before completing even the first 1000 days of the control run. An instability like this may be caused by a violation of the Courant–Friedrichs–Lewy (CFL) criterion (based on Courant et al., 1928). Whenever explicit time-marching schemes are used in models this criterion is a necessary condition for stability. The condition on the maximum time step given by this criterion is that the time step needs to be shorter than (or equal to) the time it takes for the fastest-traveling disturbance (i.e. wave) to reach from one grid point to the next.

To account for a possible violation of the CFL criterion, the time step in the model was reduced from originally 30 minutes (i.e. 48 time steps per day) to 20 minutes (72 steps per day). It was found that while this was sufficient to achieve stable control runs, the previously discussed B1 experiment would still blow-up before completing all 11000 days. Further, the results from the control run showed very fast easterly winds (> 100 m/s) in the equatorial stratosphere (for plots and discussion see section 4.2.7). Additional model updates to improve the representation of the stratosphere (sections 4.2.4 and 4.2.5) did not improve the stability sufficiently and some model runs, with 72 steps per day, including these updates would still blow-up. It was therefore decided that the time step in the final L26 version of the IGCM2.2 should be reduced further and was set to 15 minutes (96 steps per day).

4.2.3 Treatment of model top

The fulfillment of the CFL criterion is a necessary condition for stability, but not a sufficient condition. In order to further increase the numerical stability of the extended model, a change in the treatment of the model top was found to be necessary. The IGCM2.2 L15 version uses a simple reflective lid as boundary condition for the model top. This can lead to unwanted spurious downwards reflection of waves from the upper stratosphere. It was found that running the
L26 IGCM with reduced time step (96 steps per day) produced climatologies with a pronounced vertical structure in the equatorial stratosphere. Figure 4.4 shows a snapshot of the meridional winds during one day. A clear ‘pancake’-like structure where the meridional flow across the equator changes sign at each level is visible.

![Snapshot of meridional winds](image_url)

**Figure 4.4:** Snapshot of meridional winds at ±50° above 100 hPa. The contour interval is 0.25 m/s.

Such structures are characteristic of inertial instability and have been observed in modeling studies (e.g. Hunt, 1981; O’Sullivan and Hitchman, 1992) as well as in observations (e.g. Hitchman et al. 1987). This kind of instability arises in rotating fluids from an imbalance of pressure gradient force and centrifugal force (Rayleigh, 1916; Dunkerton, 1981).

This can be expressed as

\[
\frac{\partial M}{\partial y} = f \left( f - \frac{\partial u_g}{\partial y} \right) \begin{cases} 
> 0, & \text{stable} \\
= 0, & \text{neutral} \\
< 0, & \text{instable}
\end{cases}
\]  

(4.1)

where f is the Coriolis parameter, \( M = fy - u_g \) is the absolute momentum and \( \frac{\partial u_g}{\partial y} \) is the horizontal wind shear.
In an inertially unstable flow, perturbations are amplified by the imbalance of forces. In models this can lead to large errors causing the integration to fail. Several methods were tested to reduce the amount of vertical structure in the stratosphere.

- A vertical diffusion acting on the winds in the upper equatorial stratosphere:

\[
A_v(\phi, p) = \begin{cases} 
1, & \text{if } \phi \leq 15 \text{ deg and } p \leq 50 \text{ hPa} \\
 a(\phi, p), & \text{if } 15 \text{ deg} < \phi \leq 25 \text{ deg and } 50 \text{ hPa} < p \leq 250 \text{ hPa} \\
0, & \text{if } \phi > 25 \text{ deg or } p > 250 \text{ hPa} 
\end{cases}
\]  

(4.2) 

with

\[
a(\phi, p) = 1 \cdot \cos^2 \left( \frac{\phi - 15}{25 - 15} \right) \cdot \cos^2 \left( \frac{p - 50}{250 - 50} \right)
\]

(4.3) 

where \( A_v \) [\( m^2/s \)] is the vertical diffusion coefficient, \( \phi \) is the latitude and \( p \) the pressure. This gives a weak wind mixing within 50 - 250 hPa and 15-25 deg latitude falling off as a \( \cos^2 \)-function with increasing latitude and pressure and a constant diffusion with \( A_v = 1 \) [\( m^2/s \)] within 0 - 50 hPa and \( \pm 15 \text{ deg latitude} \). Further tests with a stronger vertical diffusion, \( A_v = 2 \) [\( m^2/s \)], were also conducted.

- Inclusion of a 'sponge' layer at the model top:

A sponge layer for the top three model levels (0.1, 0.4 and 1.1 hPa) was applied as a linear damping in spectral space acting on the divergence, vorticity and temperature and only damping the eddies. The damping coefficients were set to 1, 1/2 and 1/5 [1/days] for model level 1,2 and 3, respectively (i.e. damping time scales of 1, 2, and 5 days). Further, damping coefficients of 1/2, 1/4 and 1/10 [1/days] were tested.
- Added $\nabla^2$-diffusion for the model top:

Hyperdiffusion is used in atmospheric General Circulation Models to account for sub-grid scale processes. Effects of physical and dynamical processes that can be explicitly incorporated in GCMs are limited by the model resolution. Processes occurring on smaller spatial scales cannot be resolved. It is however necessary to account for such unresolved effects in order to correctly model the atmosphere as unresolved scales were found to significantly influence the large-scale flow (Burrows, 1976; Chen and Wiin-Nielsen, 1978). One commonly used approach is to parameterise the effects of the unresolved scales as a diffusion process acting on the resolved scales of motion. The hyperdiffusion $HD$ for a prognostic variable $F$ can be expressed as

$$HD_F = -(-1)^q \cdot k \cdot \nabla^{2q} \cdot F \quad (4.3)$$

where $k$ is the diffusion coefficient, $q$ is an integer and $\nabla^{2q} = \sum_{i=1}^{n} \hat{e}_i \frac{\partial^{2q}}{\partial x_i^{2q}}$. The IGCM2.2 uses a $\nabla^8$-hyperdiffusion throughout all levels. In order to reduce the amount of vertical structure near the equatorial lid which was found for the L26 runs, a stronger $\nabla^2$-diffusion was applied to the top three model levels. The diffusion coefficients were set to damp the smallest resolved waves at the truncation limit (i.e. $n=42$) with time scales of 1, 2 and 5 days for levels 1 (0.1 hPa), 2 (0.4 hPa) and 3 (1.1 hPa), respectively.

All three approaches were implemented and tested with the L26 model. All of them resulted in slightly more stable model results allowing the control run and B1 experiments to complete all 11000 days. The vertical diffusion and the added $\nabla^2$-diffusion yielded the best results (see section 4.2.7). It was decided to include the added $\nabla^2$-diffusion into the final version of the L26 IGCM2.2.

However, even with this change in the model top which successfully prevented the model from crashing, signatures of inertial instability and an unusual vertical zonal wind structure in the equatorial upper regions remained (see results in
section 4.2.7). The equatorial structure will be discussed in more detail in section 4.3.2.

4.2.4 Stratospheric temperature relaxation timescale

The temperature relaxation timescale $k_T$ determines how fast the temperatures are relaxed towards the prescribed equilibrium temperature. In Chapter 3 the impact of changing this parameter in the boundary layer (< 700 hPa) was investigated. The relaxation rate above the boundary layer was set to be constant at $k_T = 1/40 \text{ days}^{-1}$ for all model configurations. Here $k_T$ is representing the damping of thermal perturbations associated with planetary- and synoptic scale waves by radiative exchange (Newman and Rosenfield, 1997). For the 15 level version of the IGCM2.2 which only went up to about 30 km (18.5 hPa), $k_T = 1/40 \text{ days}^{-1}$ is a good representation for the long radiative timescales found in the lower stratosphere (Kiehl and Solomon, 1986; Newman and Rosenfield, 1997). In the upper stratosphere absorption of solar ultraviolet radiation (mainly by ozone) leads to warming, while longwave emission by CO$_2$ has a cooling effect. The stratospheric temperature increases with height and reaches a maximum at the stratopause (at about 1 hPa). Ramanathan et al. (1983) showed that the Newtonian cooling rates in the warmer upper stratosphere should become shorter. Several studies found the radiative relaxation timescales here to be about 5-10 days (Kiehl and Solomon, 1986; Gille and Lyjak, 1986; Newman and Rosenfield, 1997).

As the L26 model reaches up to 0.1 hPa it was decided to introduce a more realistic formulation for $k_T$ instead of keeping it at 40 days throughout the whole stratosphere. Charlton-Perez and O’Neill (2009) formulated $k_T$ such that it approximately followed the thermal relaxation rates in the real atmosphere that were calculated by Newman and Rosenfield (1997):

$$k_T(\phi, \sigma) = \begin{cases} k_{\text{tropo}}, & \text{for } \sigma \geq 0.1 \\ k_{\text{strat}}, & \text{for } 0.1 > \sigma \geq 0.01 \\ 5, & \text{for } \sigma < 0.01 \end{cases}$$  \quad (4.4)
where $\sigma = p/p_s$ and

$$k_{tropo} = k_a + (k_s - k_a)\max\left(0, \frac{\sigma - \sigma_b}{1 - \sigma_b}\right) \cdot \cos^4 \Phi \quad (4.5)$$

the formulation used for the L15 model version (with $k_a = 1/40$ [1/day], $k_s = 1/4$ [1/day] and $\sigma_b = 0.7$).

The temperature relaxation timescale for the stratospheric part ($> 100$ hPa) is given as (Charlton-Perez and O’Neill, 2009):

$$k_{strat} = \left(5 - \left\{ \frac{5 - 40}{4} \log_{10}(1000\sigma)^2 \right\} \right)^{-1} \quad (4.5)$$

Above 0.1 hPa (i.e. $\sigma < 0.01$) $k_T$ was set to be 5 days.

Figure 4.5 shows both the new temperature relaxation timescale $k_T$ (black) and the old version (red) as a function of pressure. This more realistic formulation of $k_T$ was included for the final L26 version of the IGCM2.2.

![Figure 4.5: Temperature relaxation timescale as a function of pressure. The red line indicates the previously used values (constant 40 days above 700 hPa) and the black line shows the values form the model update.](image-url)
4.2.5 **Stratospheric equilibrium temperature**

The equilibrium temperature field $T_{eq}$ used for the L15 model is isothermal (200 K) throughout the stratosphere. In order to include a better representation of the stratosphere in the L26 version, a more realistic stratospheric component $T_{eq}^S$ was added to the equilibrium temperature field.

The stratospheric equilibrium temperature profile was calculated to smoothly approximate the radiative equilibrium temperatures estimated from model (CMAM) data by Hitchcock et al. (2010) which they showed to agree well with the results from a time-marched computation of true radiative–photochemical equilibrium by Fels (1985), see Fig. 4.6.

![Figure 4.6](image)

**Figure 4.6:** (a) Linear estimate of the radiative equilibrium temperature for 15 January by Hitchcock et al. (2010). (b) Radiative–photochemical equilibrium temperatures on the same date computed by Fels (1985). Figures from Hitchcock et al. (2010).

The formulation of $T_{eq}^S$ is given by:

$$T_{eq}^S = T_{tpeq} + f(p) \cdot [T_1 + T_2 \cdot \sqrt{\cos(\phi)}]$$

where $T_{tpeq} = 200$ K (tropopause temperature), $T_1 = 10$ K, $T_2 = 60$ K and

$$f(p) = \frac{1}{2}c^2 \cdot (1 + c^2)$$

where

$$c = \cos \left( \frac{\ln(p) - \ln(p_u)}{\ln(p_d) - \ln(p_u)} \right)$$

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where \( p \) is pressure, \( p_u = 1 \text{ hPa} \) and \( p_d = 85 \text{ hPa} \).
This gives a temperature field with a stratopause temperature maximum of 270 K
at the equator, decreasing to 210 K at the poles. Figure 4.7 shows the new
temperature equilibrium field for equinox conditions now including this more
realistic representation of the stratosphere.

![Equilibrium temperature field](image)

Figure 4.7: Newtonian relaxation temperature profile for equinox conditions with
new formulation for the stratospheric part. The contour interval is 5 K.

### 4.2.6 Experiments

In order to assess the impact of the individual model updates several experiments
with combinations of these have been performed. Table 3 gives an overview of
the experiments used for this investigation.

In addition to the control runs with various update combinations, a set of B1
runs \((k_s = 1/2,\) as in chapter 3\) to test the different treatments of the model top
is discussed. The B1 run was chosen for this as this configuration was found to
be least stable and even with the short integration time step \((96 \text{ steps per day})\)
the B1 L26 run would fail without additional treatment of the model top.

All experiments were run for 10000 days (+1000 days discarded for spin-up),
unless instability caused the model to fail before completing full run (this is
the case for the runs L26t48 and B1L26t96, marked in italics in the table).
The notation 't48' and 't96' indicates the use of 48 and 96 time steps per day, respectively.

Further to the experiments described in Table 3 most of the L26 runs with $k_s = 1/4 \text{[1/day]}$ have been conducted with 72 time steps per day (i.e. a time step of 20 minutes). The results of these runs did not yield additional information and are therefore not discussed further. The B1L26 ($k_s = 1/2 \text{[1/day]}$) experiments failed to complete when run with 72 time steps per day (as mentioned in the previous section).
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Table 3: Overview of the configurations for each experiment. TSPD gives the time steps per day. $\nabla^2$, $k_{strat}$, $T_{eq}^S$ indicated whether the $\nabla^2$-diffusion, the adjusted stratospheric temperature relaxation timescale ($k_{strat}$) and/or the adjusted temperature equilibrium profile for the stratosphere ($T_{eq}^S$) were included. 'V-diff' and 'sponge' indicate whether a vertical diffusion or sponge layer was included as alternative to the $\nabla^2$-diffusion. Experiments in italics failed to complete the full run length.
4.2.7 Results

In this section the impact of the individual model updates on the model stability and climatology will be assessed.

Changing the number of time steps per day from 48 to 96:

In order to investigate the impact of the change in length of the time step, the L15t48 and L15t96 runs were compared. Figure 4.8 shows the differences zonal mean temperature (LHS) and zonal mean zonal wind (RHS) between these runs. The only difference between L15t48 and L15t96 is the change in

![Graph showing temperature and zonal wind differences](image-url)

Figure 4.8: As Fig 4.8, for L15t48 and L15t96.

integration time step. Fig. 4.8 shows that the difference in zonal mean temperatures and zonal mean zonal winds are less than 0.5 K and 0.5 m/s respectively and are not statistically significant throughout, thus reducing the length of the integration time step from 30 minutes to 15 minutes does not significantly alter the model climatology. As L26t48 was not sufficiently stable, an equivalent comparison for L26 could not be made.

Changing the vertical model extent from 18.5 hPa to 0.1 hPa:

As mentioned in section 4.2.2 a simple change from L15 to L26 led to a failure to complete the full run length which is believed to be caused by a violation of the CFL criterion. In order to investigate the impact of the vertical model extent change, from 18.5 hPa to 0.1 hPa, the L15 control run with a time step of 15 minutes (TSPD=96) and the L26 control run with the same time step have

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been chosen. Both runs successfully completed all 10000 days (plus 1000 days spin-up).

Figure 4.9: Differences between L15t96 and L26t96 zonal mean temperature (LHS) and zonal mean zonal wind (RHS). The contour intervals are 0.5 Kelvin and 0.5 m/s, respectively. The shaded areas are deduced to be less than 95% significant using a Student t-test.

Figure 4.9 shows the differences between L15t96 and L26t96 zonal mean temperature (LHS) and zonal mean zonal wind (RHS). The change from L15 to L26 has a significant impact on both. The zonal mean temperature difference about the tropopause level and above exhibits a dipole structure with cooling in the equatorial/tropical regions and a warming in the extra-tropics/polar regions. This coincides with a shift of the mid-latitude jet (RHS) towards the equator (by about 4°).

Giorgetta et al. (2007) found a similar pattern of differences in low-top versus high-top comparisons. Using an analysis of dynamical forcing terms they showed that a change in EP-flux divergence between 50 hPa and 10 hPa drives changes in the Brewer-Dobson circulation (strengthening for high-top) in the correct sense to explain the temperature differences in the lower stratosphere and thereafter the Hadley circulation changes.

The observed pattern for the L15 L25 comparison is also similar to the results of Shaw et al. (2009) who show a comparison between a 0.001 hPa (high) and 10 hPa (low) version of the Canadian Middle Atmosphere Model (CMAM). Different to the experiments here, a sponge layer has been applied to their 'high' version and the model levels below 10 hPa were identical.
In general a comparison with previous high-top versus low-top modeling studies was found to be difficult as most studies (e.g. Shaw et al., 2009; Sigmond et al. 2008; Hardiman et al., 2008) comparing high-top low-top model results use models that both are more complicated than the IGCM2 and include several differences (e.g. radiation details, sponge layers) other than the change in vertical extent between the high-top and low-top models.

![Graph](image)

Figure 4.10: Time series of the zonal winds about the equator (±4.19deg) for the L26t96 run. Shown are all 11000 days of the run, including the first 1000 days that were discarded for spin-up. Different colors indicate different pressure levels.

Even though the overall pattern (L26 - L15) is consistent with previous studies (Shaw et al., 2009), the zonal winds in the upper stratosphere equatorial region have not reached equilibrium even after a run time of 11000 days. Figure 4.10 illustrates this by showing the L26t96 time series of the zonal mean winds averaged about the equator (±4.19deg) for selected pressure levels in the stratosphere. This is a clear indication that a better treatment of the model top (as presented in section 4.2.3) is necessary.

**Treatment of model top:**

Three different treatments for the model top were tested to improve the model stability; a local vertical diffusion, a sponge layer and an additional $\nabla^2$- diffusion. These were applied to the B1L26t96 run ($k_s = 1/2$, as in chapter 3) as this configuration was found to be least stable and even with the short integration
time step (96 time steps per day) the B1 L26 run would fail without additional treatment of the model top.

Figure 4.11 shows the zonal wind time series for selected pressure levels (averaged within 4.2° deg of the equator). The results for B1L26t96 are displayed in the top panel (note this plot has a different y-axis range than the time series in the lower panels in order to show the full range of the zonal winds). This run gradually developed very large westerlies in the upper equatorial regions and failed after about 4300 days. As a first test the localised vertical diffusion was applied. The results for this are shown in the second panel. It is evident that the vertical diffusion has a strong effect on the zonal winds in the top three layers and acts to reduce the westerlies. The further development however shows that an equilibrium is not achieved and the zonal wind magnitudes (and direction) exhibit large fluctuations. While the inclusion of a vertical diffusion does improve the overall stability and allows the full 11000 days run length to complete, the results are clearly not satisfactory. Further tests with a stronger vertical diffusion coefficient (\(A_v = 2 \, m^2/s\)) did not significantly improve the results.

Next the sponge layer was tested. The results for this are shown in the third panel. The sponge layer also prevents the strong westerlies seen in B1L26t96 from developing. After the first 1000 days the zonal winds exhibit a stable behaviour for most of the levels shown. The zonal winds at the 7.5 hPa level however show an overall slight westerly trend throughout the run.

The results for applying the \(\nabla^2\)-diffusion update for B1L26t96 are shown in the bottom panel. This method is also successful in preventing the strong westerlies and allowing the run to complete the whole integration. The zonal winds in the top layers are slightly more reduced than for the sponge layer and do not exhibit strong trends throughout the run. Overall the zonal winds are however more variable (especially at the top level and at 7.5 hPa) than when a sponge layer is applied.

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Figure 4.11: As Fig. 4.9, for the B1 runs with different treatments of the model top. The upper panel shows the B1 run without any further treatment of the model top. The lower panels show the results for vertical diffusion, sponge layer and $\nabla^2$-diffusion, respectively.
All three methods of model top treatment succeed in sufficiently stabilising the B1L26t96 experiment to complete the full run length. Both the sponge layer and the $\nabla^2$-diffusion were shown to result in greater improvements than the vertical diffusion. A significant advantage for either the sponge layer or the $\nabla^2$-diffusion could not be determined. It was decided to used the $\nabla^2$-diffusion for the final L26 IG CM2.2 version.

This was then applied to the L26 control run (L26t96). Figure 4.12 shows equivalent plot for L26t96 with (lower panel) and without (upper panel) the $\nabla^2$-diffusion.

Figure 4.12: As Fig. 4.9, for the L26t96 and L26t96d. The upper panel shows the L26t96 run without $\nabla^2$-diffusion and the lower panel L26t96d, with $\nabla^2$-diffusion.

Again the westerlies in the top layers are significantly reduced (note the different vertical axes) and an equilibrium was reached after about 4000 days. The variability in zonal wind magnitude is reduced compared to the B1 experiment results. The reduction of the zonal wind magnitude in the upper equatorial regions is also visible in Fig. 4.13. Note that the zonal mean temperatures and zonal mean wind in Fig. 4.13 are only shown up to 1 hPa where the zonal winds are equilibrated (see Fig. 4.12). The top two levels have trends throughout the run for L26t96.
Figure 4.13: Differences between L26t96 and L26t96d (with $\nabla^2$-diffusion) zonal mean temperature (LHS) and zonal mean wind (RHS) with log-pressure scale. The contour intervals are 0.5 K and 2 m/s, respectively. The shaded areas are deduced to be less than 95% significant using a Student t-test.

Shown are the zonal mean temperature and zonal mean zonal wind differences between L26t96 and T26T96d. The shaded areas are deduced to be less than 95% significant using a Student t-test. It is also apparent that the $\nabla^2$-diffusion does not significantly influence the climatology in the troposphere, but only acts to alter the zonal winds and temperatures in the stratosphere near the top three layers where it is applied (down to about 20 hPa).

Combining both the change in integration time step and treating the model top with a $\nabla^2$-diffusion results in a greatly improved numerical stability. Further improvement is achieved by adjusting the model physics for the presence of a stratosphere as shown in the following sections.

**Adjusting the stratospheric temperature relaxation timescale:**
Introducing a more realistic formulation of the stratospheric temperature relaxation timescale is an important step towards a better representation of the stratosphere in the model. Figure 4.14 shows the effect of this (middle panel) on the zonal mean zonal wind time series as well as the effect of additionally including the $\nabla^2$-diffusion (bottom panel). The top panel shows the zonal mean time series of L26t96 for comparison.
Figure 4.14: As Fig. 4.9, for the L26t96, L26t96k and L26t96kd. The upper panel shows the L26t96 run without adjustment for $k_{strat}$ and the middle panel L26t96k, with new $k_{strat}$ and the lower panel L26t96kd, with new $k_{strat}$ and $\nabla^2$-diffusion.
These results again confirm that the $\nabla^2$-diffusion is necessary to achieve a stable run even when the relaxation timescale is adjusted for the presence of a stratosphere. Applying only $k_{\text{strat}}$ does slightly reduce the magnitude of the equatorial upper layer westerlies, but continues to exhibit zonal wind trends throughout the entire run time without coming to equilibrium.

The effect of the combined $\nabla^2$-diffusion and $k_{\text{strat}}$ is also shown in Fig. 4.15. Shown are the zonal mean temperature (RHS) and zonal mean zonal wind (LHS) differences between L26t96 and T26T96kd. The shaded areas are deduced to be less than 95% significant using a Student t-test. A significant impact on both the zonal mean temperatures and the zonal mean winds in mainly apparent in the stratosphere where the changes have been applied. The faster relaxation towards the equilibrium temperatures (which are isothermal at 200 K above the tropopause) leads to a slight warming of the equatorial/tropical regions in the stratosphere while a more pronounced cooling in the polar regions is visible.

![Figure 4.15](image-url)

Figure 4.15: Differences between L26t96 and L26t96kd (with $\nabla^2$-diffusion and $k_{\text{strat}}$) zonal mean temperature (LHS) and zonal mean wind (RHS) with log-pressure scale. The contour intervals are 1 Kelvin and 2 m/s, respectively. The shaded areas are deduced to be less than 95% significant using a Student t-test.

This decreases the meridional temperature gradients about the mid-latitudes at the model top region. This is consistent with the stronger temperature relaxation applied for the upper regions which acts to constrain the temperatures more closely to the equilibrium temperature $T_{eq}$ (which is a constant 200 K). Figure 4.16 show the zonal mean temperatures and zonal mean winds for L26t96
for comparison. Equivalent plots for L26t96kd can be found in the top panel of Fig. 4.21. This change in temperature at leads to a weakening of the strong easterlies in this region as seen in Fig. 4.15 (LHS) and Fig. 4.16 (LHS). Further a significant reduction of the equatorial easterlies between 10 hPa and 2 hPa and a reduction of the equatorial westerlies above 2 hPa can be seen. This decrease in zonal wind magnitude is consistent with the previously observed effects of the $\nabla^2$-diffusion. Note again that the zonal mean temperatures and zonal mean wind in Fig. 4.15 are only shown up to 1 hPa where the zonal winds are equilibrated (see Fig. 4.14).

Figure 4.16: Zonal mean temperatures and zonal mean winds for L26t96. The contour intervals are 5 Kelvin and 4 m/s, respectively.

**Adjusting the stratospheric equilibrium temperature:**

The last step for the full L26 version of IGCM2.2 is the change in the stratospheric equilibrium temperature from an isothermal to a more realistic profile. Figure 4.17 shows the effect of this (2nd panel) on the zonal mean zonal wind time series as well as the effect of additionally including the $\nabla^2$-diffusion (3rd panel) and finally the full L26complete result (which also includes $k_{strat}$). The top panel shows the zonal mean time series of L26t96 for comparison. While an adjusted $T_{eq}^S$ strongly reduces the amplitudes of the equatorial westerlies, an equilibrium even after 11000 days is not achieved. The zonal winds at 0.1hPa show strong variability with easterly wind speeds between -17 m/s to -82 m/s (minimum not shown in the plot range). The lower two panels show how the results are improved when the $\nabla^2$-diffusion and $k_{strat}$ are included as well.
Figure 4.17: As Fig. 4.9, for the L26t96, L26t96Ts, L26t96dTs and L26complete. The upper panel shows the L26t96 run without adjustment for $T_{eq}^S$ and the lower panels L26t96Ts, with new $T_{eq}^S$, L26t96dTs (with $\nabla^2$-diffusion and $T_{eq}^S$) and L26complete.
In the L26complete version zonal mean wind equilibrium is reached after about 1000 days.

The effect of the adjusted $T_{eq}^S$ is further shown in Figure 4.18. Shown are the zonal mean temperature (RHS) and zonal mean zonal wind (LHS) differences between L26t96 and T26T96Ts. The shaded areas are deduced to be less than 95% significant using a Student t-test.

![Figure 4.18: As Fig 4.15, but with contour intervals of 5 Kelvin for the temperature and 5 m/s for the zonal wind differences, for L26t96 and L26t96Ts.](image)

A significant impact on both the zonal mean temperatures and the zonal mean winds in apparent in the stratosphere where the equilibrium temperature change has been applied. The zonal mean temperature differences (LHS) show the effect of changing from an isothermal ($T_{eq}=200$ K) to the more realistic profile with a temperature increase toward higher altitudes in the stratosphere reaching a maximum at about the stratopause region (about 1 hPa). The zonal mean temperatures follow approximately the $T_{eq}^S$ that were prescribed while being modified by the dynamics in the stratosphere. The equatorial stratosphere is slightly cooler than prescribed and the polar region is slightly warmer as the Brewer-Dobson circulation transports air from the equatorial regions to the more polar regions. The strong meridional temperature gradients (T decreasing with latitude) in the polar regions in the stratosphere lead to the appearance of a polar jet (RHS) through the thermal wind relation.
4.3 Analysis of the extended model control run

The full L26 (L26complete) model results and features are presented in this section. First the main climatology is discussed, this is followed by a more detailed analysis of the equatorial stratosphere region. This region is of special interest due to the occurrence of anomalous winds/circulations in this area as well as the presence of inertial instability.

4.3.1 The mean climatology

The full L26 control run was run for 10000 days (plus 1000 days spin up time) and included all updates as discussed in section 4.2 (where the ∇^2-diffusion was chosen as treatment for the model top). The results of the L26complete run are presented in Figure 4.19. The zonal mean temperatures (top panel) and zonal mean zonal winds (bottom panel) are shown. The vertical axis is given in log-pressure for better visibility of the stratospheric part. The zonal mean temperature now exhibits a more realistic profile. The temperature decreases with height in the troposphere, reaching a temperature minimum of about 185 K at the equatorial tropopause before then increasing toward higher altitudes in the stratosphere and reaching a maximum at about the stratopause region (about 1 hPa, about 270 K). This zonal mean temperature profile follows approximately the $T_{eq}$ and $T^{S}_{eq}$ equilibrium profiles that were prescribed to the model but is modified by the dynamics. In addition to the mid-latitude zonal mean jets in the troposphere, seen in L15, polar jets with maximum westerly wind speed of about 60 m/s occur at about ± 55° in the stratosphere of the L26complete version (bottom panel). The pattern of easterlies observed in the equatorial stratosphere will be discussed in section 4.3.2.

The zonal mean temperature and zonal wind differences of the L26complete run from the L15t96 run are shown in Fig. 4.20. Note, this figure has a linear vertical axis for better comparison between L15 and L26. The shaded areas are deduced to be less than 95% significant using a Student t-test. A slight cooling of the equatorial/subtropical tropopause regions and a slight warming of the polar
Figure 4.19: a) Zonal mean temperature and b) zonal mean wind for the control run with vertical axis in log-pressure. The contour intervals are 5 Kelvin and 4 m/s, respectively.

regions in the stratosphere are observed for the L26complete with only small changes in the tropospheric temperatures (LHS). The difference in zonal mean zonal winds (RHS) shows a clear shift of the mid-latitude jet towards the equator (about 2 deg) which would be expected as response to the observed warming of the polar tropopause region (Haigh et al., 2005).

Gerber et al. (2008) caution that an increased vertical resolution might not always result in a better simulation of the dynamics. Specifically they found that doubling the vertical resolution (from L20 to L40) can lead to significantly longer (up to about 100%) annular mode decorrelation timescales (τ). This was also found to be the case for the IGCM2.2 when increasing the vertical resolution from L15 to L30, but the reverse was found for the change from L30 to L60.
Figure 4.20: As Fig 4.10, but with contour intervals of 0.5 Kelvin for the temperature and 0.5 m/s for the zonal wind differences, for L15t96 and L26complete. (Simpson et al., 2009). The change from L15 to L26 discussed for this chapter differs from the Gerber et al. (2008) and Simpson et al. (2009) experiments in that the vertical resolution in the troposphere and lower stratosphere was kept similar, but for the L26 version additional model levels at higher altitudes were included. A comparison of the decorrelation times between L26complete ($\tau = 72 \pm 9$) and L15t96 ($\tau = 63 \pm 7$) shows only a slight increase which is consistent with the more equatorward mid-latitude jet position found for L26complete.

Overall the new L26 version of the IGCM2.2 produces a realistic climatology within the constraints of the simplified GCM formulation.

4.3.2 The equatorial stratopause region

The equatorial region in the stratosphere exhibits some anomalous zonal wind patterns that are discussed in this section. This is of interest as any anomalous dynamics in this region might have an influence on the outcome of stratospheric forcing experiments conducted with the L26 version of the model. Especially for potential future experiments with QBO-like momentum forcings which would be applied in the lower equatorial stratosphere, a thorough understanding of the control run dynamics in this region would be needed.

The L15 IGCM2.2 control zonal wind in the equatorial regions consists of easterlies which increase in strength with height. While the L26t96 shows the same
pattern with the easterlies reaching very large (> 100 m/s) magnitudes in the upper stratosphere, any further addition to the model (aside from reducing the time step) leads to a pattern of zonal winds with minimum easterlies (or even westerlies) roughly at the stratopause level. Originally it was thought this equatorial structure might be linked to the inertial instability found in this region. The RMS (root mean squared) value of the zonally averaged meridional winds at the top six model levels was used as an indication of the strength of the inertial instability. The occurrence and strength of the equatorial zonal wind structure however could not be linked to the strength of inertial instability or the likelihood of the run failing to complete the integration. In order to investigate this further three experiments that exhibit this pattern with varying strength have been chosen; L26t96kd showing the strongest pattern, L26complete as control and L26complete with E5 heating (as in chapter 3) showing the faintest pattern. Figure 4.21 shows the zonal mean temperatures (LHS) as well as zonal mean zonal winds (RHS) for these three experiments, each run for 10000 + 1000 days. The temperature structure clearly reflects the absence of a modified equilibrium temperature in the stratosphere ($T_{eq}^S$) for L26t96kd. Further the effect of the prescribed E5 heating for the L26complete E5 temperatures are evident. The equatorial zonal wind pattern in the stratosphere, as mentioned above, can be seen in the zonal wind fields on the RHS of Fig. 4.21.

Plots of the streamfunction of the mean meridional circulation (Figure 4.22, LHS) for L26t96kd show a clear anomalous circulation confined between 10hPa - 1hPa and about ±10° on either side of the equator. This anomalous circulation is only rudimentary for L26complete and not visible at all for L26complete E5. The RHS of Figure 4.22 shows the EP-flux divergences for all three experiments. An increased EP-flux divergence can be seen at the position of the anomalous circulation. This is now also apparent for L26complete E5, though much reduced compared to the other two experiments. This positive EP-flux divergence locally reduces the poleward residual circulation.
Figure 4.21: Zonal mean temperature (LHS) and zonal mean zonal wind (RHS) for the L26t96kd (top panel), L26complete (middle panel) and L26complete E5 (bottom panel). The contour intervals are 5 Kelvin and 4 m/s, respectively.
Figure 4.22: Streamfunction of the meridional circulation (LHS) and total EP-flux divergence (RHS) for the L26t96kd (top panel), L26complete (middle panel) and L26complete E5 (bottom panel). The contour intervals are half-logarithmic at 1,2 and \(5 \times 10^m\) kg/s and \(m^3\), respectively.
This pattern of weak easterlies/westerlies is similar in structure and extent to the westerly phase of the semi-annual oscillation (SAO). As the IGCM2 experiments are in a fixed phase of the seasonal cycle (perpetual equinox), a fixed phase of the SAO might be expected. This would agree with the occurrence of maximum SAO westerlies in March and October (e.g. Reed, 1966). The main drivers of the SAO however arise from seasonality (Hamilton and Mahlman, 1988; Delisi and Dunkerton, 1988) which this model does not include. Further the forcing of the westerly acceleration of the SAO has been connected with dissipation of planetary scale waves (Mueller et al., 1997) which are strongly underrepresented in the IGCM2.2 as no orography is included. A possible solution for the occurrence of the westerlies in this model might be that different waves could have a similar effect. For further investigation of this a cospectral analysis might be useful, but is beyond the scope of this thesis.

The cause of the observed anomaly in the equatorial stratopause remains unclear. As shown earlier the occurrence and magnitude of this anomaly could not be linked with the numerical stability of the model nor with the magnitude of the equatorial inertial instability. The L26 complete mean climatology in the tropopause and lower stratosphere however is comparable with the L15 control simulation.

In order to investigate whether the tropospheric response to stratospheric heating perturbations is influenced by the differences in L15 and L26 stratosphere, various heating perturbation were applied to both model versions and compared. This will be presented and discussed in detail in Chapter 5. It will be shown that the tropospheric responses are qualitatively similar and the quantitative differences can be explained by other factors (e.g. the jet position difference between L15 and L26 control simulations). It was hence decided not to investigate the L26 equatorial stratopause anomaly further. It has to be noted, however, that for future investigations of the effects of momentum forcings applied to the stratosphere (e.g. QBO or SAO-like experiments), possible influences from this anomaly need to be considered.

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4.4 Discussion and Conclusions

A new 26 level version of the IGCM2.2, now including a stratosphere, has been introduced in this chapter. It was shown that a simple change of the model sigma levels resulted in the model becoming unstable and the runs failing to complete. Several model updates that were applied to both increase the numerical stability and to improve the representation of the stratosphere were presented and the results of each of them on the model climatology was discussed.

The updates included in the final L26 version of the model and the main effects of these on the model climatology are:

- Changing the vertical model extent from 18.5 hPa to 0.1 hPa
  The change from L15 to L26 has a significant impact on both the zonal mean temperature and the zonal mean winds. The zonal mean temperature difference about the tropopause level and above was found to exhibit a dipole structure with cooling in the equatorial/tropical regions and a warming in the extra-tropics/polar regions. This coincides with a shift of the mid-latitude jet towards the equator (by about 4°). Such a jet shift is in agreement with the tropospheric response that would be expected (e.g. Polvani and Kushnir, 2002; Haigh et al., 2005; Simpson et al., 2009) as result of a stratospheric temperature change like the equatorial cooling/polar warming observed here.

- Increasing the number of time steps per day from 48 to 96
  This update was included to prevent a possible violation of the CFL criterion. It was found to improve the model stability and stable control runs were achieved. However, the B1 experiment still blew-up before completing all 11000 days. It was further shown that a reduction in the length of the integration time step from 30 minutes (48 time steps per day) to 15 minutes (96 time steps per day) does not significantly alter the model control run climatology.
• Treatment of the model top
  Three different treatments for the model top were tested to further improve the model stability; a local vertical diffusion, a sponge layer and an additional $\nabla^2$-diffusion. It was shown that each of these further improved the model stability with the sponge layer and the additional $\nabla^2$-diffusion yielding the best results. As no significant difference in improvement between the sponge layer and the additional $\nabla^2$-diffusion could be determined, the additional $\nabla^2$-diffusion was chosen for the final L26 version. It was then shown that the $\nabla^2$-diffusion does not significantly influence the climatology in the troposphere, but only acts to alter the zonal winds and temperatures close to the top three layers where it is applied.

• Adjusting the stratospheric temperature relaxation timescale
  A more realistic formulation of the stratospheric temperature relaxation timescale was introduced to improve the representation of the stratosphere in the model. The relaxation time scales were calculated such that they approximately followed the thermal relaxation rates in the real atmosphere using the formulation by Charlton-Perez and O’Neill (2009). It was found that applying this update without additional treatment of the model top does not yield stable results. In combination with the $\nabla^2$-diffusion, the faster relaxation towards the equilibrium temperatures was shown to lead to a decrease the meridional temperature gradient which is consistent with with the stronger temperature relaxation applied for the upper regions which acts to constrain the temperatures more closely to the equilibrium temperature $T_{\text{eq}}$ (which is a constant 200 K).

• Adjusting the stratospheric equilibrium temperature
  A more realistic stratospheric component $T_{\text{eq}}^S$ was added to the equilibrium temperature field. This stratospheric equilibrium temperature profile was calculated to smoothly approximate the radiative equilibrium temperatures in the stratosphere estimated from previous studies by Hitchcock
et al. (2009) and Fels (1985). Such an improvement of the stratospheric equilibrium temperature was found to have a significant impact on both the zonal mean temperatures and the zonal mean winds mainly in the stratosphere where the equilibrium temperature change has been applied. The zonal mean temperatures were shown to now follow the more realistic profile which a temperature increase toward higher altitudes in the stratosphere reaching a maximum at about the stratopause region (about 1 hPa). The resulting strong meridional temperature gradients in the stratospheric polar regions were shown to lead to the appearance of a polar jet while the strong warming in the equatorial regions coincides with a reduction of the stratospheric westerlies.

Overall the new L26 version of the IGCM2.2 was shown to produce a realistic climatology within the constraints of the simplified GCM formulation. It has to be noted however that the model climatology does not include realistic stratospheric variability. The variability in the stratosphere is governed by tropospheric variability, mediated via planetary waves (e.g. Sudden Stratospheric Warming events that are forced from the troposphere by upward propagating planetary waves (e.g. Yoden et al., 1999) or the westerly phase of the Semi-Annual-Oscillation (Mueller et al., 1997)). The main drivers for the generation of such planetary waves are orography, land-sea contrasts and seasonality. These are all absent from the IGCM2.2 and hence large-scale waves and stratospheric variability are strongly underrepresented in this model.

Further, in the equatorial stratopause region an anomalous zonal wind signature was observed. While L15 IGCM2.2 control zonal wind in the equatorial regions consists of easterlies which increase in strength with height and the same hold true for the L26t96 experiment, any further addition to the model (aside from reducing the timestep) leads to a pattern of zonal winds with minimum easterlies (or even westerlies) roughly at the stratopause level. The cause of the observed anomaly in the equatorial stratopause could not be determined. It was shown that the occurrence and magnitude of this anomaly cannot be linked with the
numerical stability of the model nor with the magnitude of the equatorial inertial instability. Such a feature has to the author’s knowledge not been discussed previously. A more comprehensive investigation of this anomaly is beyond the scope of this thesis, especially as the occurrence of it does not alter the tropospheric response to heating perturbation applied to the stratosphere (as will be shown in Chapter 5). It has to be noted, however, that for future investigations of the effects of momentum forcings applied to the stratosphere (e.g. QBO or SAO-like experiments), possible influences from this anomaly need to be considered.
5 Impact of variations in the position and shape of heating perturbations imposed on the stratosphere

In this chapter the impact of variations in the shape and position of heating perturbations applied to the stratosphere is investigated. Firstly, the motivation for this research is given and then the different experiments used for this study are presented. The effects of retracting the heating perturbations in altitude are discussed in section 5.3. This is followed by an investigation of the effects of restricting the latitudinal extent of the heating (section 5.4). Further, this section includes a comparison between results gained using the L15 and the L26 versions of the IGCM2.2. Finally, the results are summarised and the main conclusions presented in section 5.5.

5.1 Motivation

The new L26 version of the IGCM2.2 allows further investigation of the tropospheric response to stratospheric forcings, both by altering the shape and position of heating perturbation as well as by enabling a comparison of the results obtained with this extended model and the original L15 version.

It is of great interest to investigate whether the same mechanisms that are important for creating the tropospheric response also apply to perturbations higher in the stratosphere. This as well as the relation between the magnitude of the tropospheric response and the distance of the perturbation from the tropopause are of relevance for understanding and interpreting observations from real world solar forcings (such as shown in Fig. 4.1.). The previous research of stratosphere-troposphere interactions with the IGCM2 (e.g. Haigh et al., 2005; Simpson (2009) and Simpson et al., 2009) used heating perturbations applied down to a reference tropopause which extends below the model’s tropopause. The original L15 version of the IGCM2 used for these experiments has only very few (5) lev-
els in the lower stratosphere region for solar regression and the top model level is well below the maximum temperature response that occurs at about 1hPa. Using the L26 model, it is possible to retract heating perturbations away from the tropopause and therefore to study perturbations that resemble the real world solar forcing more closely. This will also improve the understanding of the previous model results to temperature perturbations.

The investigation of the tropospheric response to the different retracted heating perturbations highlights the importance of the mid-latitude region. The magnitude of the heating perturbation at these latitudes is shown to have a great impact on the tropospheric response (section 5.3). As understanding this is of great importance for interpreting the previous results as well as the general understanding of stratosphere - troposphere interactions, it was decided to explore this further by designing a set of experiments with varying latitudinal extents of the stratospheric heating perturbation.

The use of the L26 IGCM2.2 version further allows the investigation of the stratospheric response to such heating perturbations. Previously this was severely compromised with IGCM2.2 due to the low vertical extent (≈30 km) of the L15 version. Understanding this may be important as a stratospheric response to the applied heating perturbation might alter the forcing region itself and thereby possibly influence the tropospheric response.

Another important aspect of the research presented in this chapter is the comparison between the L15 and L26 versions of the IGCM2.2. This allows investigation of the impact of the presence of a resolved stratosphere on the model response in both the lower stratosphere/ tropopause region and in the troposphere to heating perturbations applied in the lower stratosphere. This is not only important for understanding the role of the mid/upper stratosphere but also helping in the design of future experiments and in the choice of appropriate models. It is however again important to note that due to the limited stratospheric variability in the new L26 IGCM2, the results of this model are only relevant to cases with limited stratospheric dynamics (e.g. in the summer hemisphere). Another consideration
is computing time: it is important to have solid information on which model is sufficient for a given experiment. In the case of the IGCM.2.2 running the L26 version for 10000 days instead of the L15 version increases computing time by a factor of more than 3.

5.2 The experiments

Two types of heating perturbation experiments are presented. The first part of this chapter focuses on heating perturbations that are similar to the original E5 heating (see Chapter 3), but where the full heating is applied higher up in the stratosphere. In the second part the effects of varying the extent of the heating perturbation in latitude is investigated.

The heating perturbations are applied by changing the parameters that determine $T_{eq}^S$ (for E5) or adding a heating perturbation function to this equilibrium temperature. The formulation of $T_{eq}^S$ used in the L26 IGCM2.2, as introduced in Chapter 4, is given by:

$$T_{eq}^S = T_{tpeq} + f(p) \cdot [T_1 + T_2 \cdot \sqrt{\cos(\phi)}]$$  \hfill (5.1)

where $T_{tpeq} = 200$ K (tropopause temperature), $T_1 = 10$ K, $T_2 = 60$ K and

$$f(p) = \frac{1}{2} c^2 \cdot (1 + c^2)$$  \hfill (5.1a)

where

$$c = \cos\left(\frac{\ln(p) - \ln(p_u)}{\ln(p_d) - \ln(p_u)}\right)$$  \hfill (5.1b)

where $p$ is pressure, $p_u = 1$ hPa and $p_d = 85$ hPa.

This gives a temperature field with a stratosphere temperature maximum of 270 K at the equator, decreasing to 210 K at the poles.

The heating perturbations used for the experiments conducted for this research are labeled E5 - original heating perturbation, Aup/Bup/Cup/E5dp and E5dz for the heating perturbations that are retracted in altitude and SI - SV for the
varying latitudinally restricted perturbations. The experiments with the latitudinal restriction are also compared with a uniform 5 K heating, U5.

The functions applied for the different heating perturbations and short descriptions of the structures are given below and are displayed in Figures 5.1 and 5.3:

- E5, the same heating perturbation used in Chapter 3 (5 K at the equator, falling off with a $\cos^2 \phi$ towards the poles).

$$T_{eq}^{S(E5)} = T_{eq}^S + \Delta T_{tp} \cos^2(\phi)$$

(5.2)

where $\Delta T_{tp} = 5$ K.

- Three different vertically retracted heating perturbations are used:

  - Aup, this heating perturbation is equal to the E5 above 85 hPa. Below this it follows a linear decrease from 5 K at 85 hPa down to 0 K at the pressure value of the equatorial tropopause. Note, this lower level is equal for all latitudes, i.e. the heating does not reach the tropopause for latitudes poleward of the equator. Further, the perturbation also falls off with a $\cos^2 \phi$ towards the poles.

  - Bup, as Aup, but interpolated to zero at the local tropopause in $T_{eq}$ for all latitudes.

  - Cup, as Bup, but keeps the shape of the tropopause, i.e. the linear decrease does not start at 85 hPa for all latitudes, but at a value that keeps the same pressure distance from the local tropopause.

The vertically retracted heating perturbations are given by:

$$T_{eq}^{S(v)} = T_{eq}^S + v_{up} \cos^2(\phi)$$

(5.3a)

where

$$v_{up} = \Delta T_{tp} \cdot \frac{P - P_2}{P_1 - P_2}$$

(5.3b)

where $p_1$ and $p_2$ for the three different heating perturbations are:
- Aup: \( p_1 = 85\, \text{hPa} \) and \( p_2 = p_{\text{eq}} \) the pressure at the equatorial tropopause.

- Bup: \( p_1 = 85\, \text{hPa} \) and \( p_2 = p_t(\phi) \) the pressure at the local tropopause.

- Cup: \( p_1 = p_{\text{off}}(\phi) = p_t(\phi) - [p_{\text{eq}} - 85\, \text{hPa}] \) and \( p_2 = p_t(\phi) \) the pressure at the local tropopause.

- E5dp, like the E5 heating, but moved upwards by a constant pressure distance, \( \Delta p = p_{\text{eq}} - 85\, \text{hPa} \).

- E5dz, like the E5 heating, but moved upwards by a constant height distance, \( \Delta z \, [\text{m}] \) which is equivalent to the difference between the equatorial tropopause pressure and 85 hPa (about 4.8 km).

Figure 5.1. shows the shape of heating perturbations E5, Aup, Bup, Cup, E5dp and E5dz. Shown are the differences of the equilibrium temperature profiles with heating perturbation and without.
Figure 5.1: Shape of heating perturbations E5, Aup, Bup, Cup, E5dp and E5dz. Shown are the differences of the applied equilibrium temperature profiles with heating perturbation and without. The contour interval is 1 K.

The latitudinally restricted perturbations are described below and shown in Figs. 5.2 and 5.3.

- SI - SV, preferentially equatorial heating perturbations, with varying latitudinal extent. All S perturbations fall off with a \( \sin^2 \) function from 5 K to 0 K within a 30 deg latitude range. Equatorward of this the heating is kept at a constant 5 K and poleward of this at 0 K. The position of the heating gradient is shifted between 0-30 deg and 60-90 deg in 15 deg intervals. In all cases the full heating is applied immediately above the model tropopause.

\[
T_{eq}^{S(S)} = T_{eq}^S + S \cdot r
\]  

(5.6)
where
\[
    r = \begin{cases} 
        1, & \text{if } \phi < a \\
        r^*, & \text{if } a \leq \phi \leq (a + b) \\
        0, & \text{if } \phi > (a + b) 
    \end{cases} \tag{5.6a}
\]

where \( b = 30^\circ \) and
\[
    r^* = \cos^2 \left( \frac{|\phi| - a}{b} \right) \tag{5.6b}
\]

and
\[
    a = \begin{cases} 
        0, & \text{for } SI \\
        15, & \text{for } SII \\
        30, & \text{for } SIII \\
        45, & \text{for } SIV \\
        60, & \text{for } SV
    \end{cases} \tag{5.6c}
\]

- U5, uniform heating perturbation of 5 K everywhere applied above the reference tropopause.

Latitudinal profiles of the E5 and SI-SV heating perturbations above the model tropopause (shown at 50 hPa) are shown in Fig. 5.2.

![Heating perturbation profiles at 50hPa](image_url)

**Figure 5.2:** Profiles of the E5 and SI-SV heating perturbations.

Figure 5.3 shows the shape of heating perturbations SI - SV and U5. Shown are the differences of the equilibrium temperature profiles with heating perturbation and without. The contour interval is 1 K.
Figure 5.3: Shape of heating perturbations SI, SII, SIII, SIV, SV and U5 (see text).

The investigation of the effect of retracting the heating perturbation in altitude (Chapter 5.3) includes experiments using the L26 complete control ($k_s = 1/4 \ [\text{day}^{-1}]$) with applied heating as well as B1L26 ($k_s = 1/2 \ [\text{day}^{-1}]$) with heating perturbations.

Experiments using both the L26 and the L15 versions of the IGCM2.2 are used for the investigation of the effect of restricting the latitudinal extent of the heating. The heating perturbations with varying latitudinal extent were applied only to the control runs ($k_s = 1/4 \ [\text{day}^{-1}]$). For comparison B1L26 and B1L15 with E5 and U5 heating perturbations were also performed. In addition to this the E5, Cup and E5dz were also combined with a SII latitudinal restriction and applied to the L26 control run (L26E5-SII, L26Cup-SII and L26E5dz-SII).

All L26 experiments use the complete L26 set-up with all additions discussed...
in Chapter 4 (i.e. TSPD=96, added $\nabla^2$-diffusion, improved stratospheric temperature relaxation timescale and stratospheric equilibrium temperature). A complete overview of the experiments is given in Tables 4 (L15) and 5 (L26).

<table>
<thead>
<tr>
<th>experiment</th>
<th>$k_\sigma$ [1/days]</th>
<th>heating perturbation</th>
</tr>
</thead>
<tbody>
<tr>
<td>L15control</td>
<td>1/4</td>
<td>-</td>
</tr>
<tr>
<td>L15E5</td>
<td>1/4</td>
<td>E5</td>
</tr>
<tr>
<td>L15SI</td>
<td>1/4</td>
<td>SI</td>
</tr>
<tr>
<td>L15SII</td>
<td>1/4</td>
<td>SII</td>
</tr>
<tr>
<td>L15SIII</td>
<td>1/4</td>
<td>SIII</td>
</tr>
<tr>
<td>L15SIV</td>
<td>1/4</td>
<td>SIV</td>
</tr>
<tr>
<td>L15SV</td>
<td>1/4</td>
<td>SV</td>
</tr>
<tr>
<td>L15U5</td>
<td>1/4</td>
<td>U5</td>
</tr>
<tr>
<td>B1L15</td>
<td>1/2</td>
<td>-</td>
</tr>
<tr>
<td>B1L15E5</td>
<td>1/2</td>
<td>E5</td>
</tr>
<tr>
<td>B1L15U5</td>
<td>1/2</td>
<td>U5</td>
</tr>
</tbody>
</table>

Table 4: Overview of the configurations for each heating perturbation run performed with the original L15 IGCM2.2. Each experiment was run for 11000 days.
<table>
<thead>
<tr>
<th>experiment</th>
<th>$k_s$ [1/days]</th>
<th>heating perturbation</th>
</tr>
</thead>
<tbody>
<tr>
<td>L26complete</td>
<td>1/4</td>
<td>-</td>
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<tr>
<td>L26E5</td>
<td>1/4</td>
<td>E5</td>
</tr>
<tr>
<td>L26Aup</td>
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<td>Aup</td>
</tr>
<tr>
<td>L26Bup</td>
<td>1/4</td>
<td>Bup</td>
</tr>
<tr>
<td>L26Cup</td>
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<td>Cup</td>
</tr>
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</tr>
<tr>
<td>B1L26</td>
<td>1/2</td>
<td>-</td>
</tr>
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<td>Aup</td>
</tr>
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<td>Bup</td>
</tr>
<tr>
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</tr>
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<tr>
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</tr>
<tr>
<td>L26SII</td>
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</tr>
<tr>
<td>L26SIII</td>
<td>1/4</td>
<td>SIII</td>
</tr>
<tr>
<td>L26SIV</td>
<td>1/4</td>
<td>SIV</td>
</tr>
<tr>
<td>L26SV</td>
<td>1/4</td>
<td>SV</td>
</tr>
<tr>
<td>L26U5</td>
<td>1/4</td>
<td>U5</td>
</tr>
</tbody>
</table>

Table 5: Overview of the configurations for each heating perturbation run performed with the L26 IGCM2.2. Each experiment was run for 11000 days. Experiments performed with the L26 IGCM version use the L26complete set up discussed in Chapter 4.
5.3 Effect of retracting the heating perturbation in altitude

First the effects of retracting the heating perturbation in altitude are discussed. The main motivation for this is to investigate whether the same mechanisms that are important for creating the tropospheric response in previous studies (e.g. Haigh et al., 2005; Simpson, 2009 and Simpson et al., 2009) also apply to perturbations higher in the stratosphere. Further the relation between the magnitude of the tropospheric response and the distance of the perturbation from the tropopause are of relevance for understanding and interpreting observations from real world solar forcings (such as shown in Fig. 4.1.).

Three heating perturbations Aup, Bup and Cup are applied to the stratospheric equilibrium temperature and run for 10000 days (plus 1000 days) each. As described in the previous section, these heatings are similar to the E5 perturbation, but do not apply the full 5 K heating in full down to the model reference tropopause. Figure 5.4 shows the differences of zonal mean temperature (LHS) and zonal mean zonal wind (RHS) from the control run (L26complete) for the E5 and each of the vertically retracted experiments. The fields are averaged over both hemispheres, thus only one hemisphere is shown. Regions where the difference between the experiments and control run is not statistically significant at the 95% level are shaded in grey (calculated by using the Student t-test).

The temperature differences roughly reflect the respective heating perturbations applied for each experiment. Comparisons with the results from zonally symmetric runs show eddy feedbacks play an important role in causing the deviations of the temperature anomalies shown in Fig. 5.4 from the applied perturbations (Fig. 5.1). This will be discussed in detail for the heating experiments with varying latitudinal extent (section 5.4). The shape of the stratospheric temperature differences for the L26Cup (bottom panel) experiment resembles the original E5 result most closely. The corresponding L26Aup and L26Bup results show the full heating perturbations further away from the tropopause for in the extratropics. While these temperature differences closely match the results that would be
Figure 5.4: Differences between perturbed and unperturbed experiments in zonal mean temperature (left panels) and zonal mean zonal wind (right panels) for the E5, Aup, Bup and Cup experiments. The contour intervals are 0.5 K (left panels) and 0.5 m/s (right panels). Shaded areas are not statistically significant at the 95% level.
expected in response to the applied perturbations, the zonal mean zonal wind differences in the troposphere (RHS of Fig. 5.4) are less obvious to interpret. A strong poleward shift of the mid-latitude jet is visible in response to the E5 heating (top panel). The tropospheric responses for the L26Aup, L26Bup and L26Cup experiments also show a poleward shift of the jet, but considerably weaker then for L26E5. This would be expected as the full heating perturbation (5 K) is only applied higher up (see Fig. 5.1) in the atmosphere. While the Aup and Bup perturbations lead to similar magnitudes of response (Bup slightly stronger than Aup) in the troposphere, the Cup experiment shows a significantly weaker response. Comparing the temperature differences on the LHS of Fig. 5.4 out of all vertically retracted experiments, the Cup zonal mean zonal wind differences would have been expected to be stronger than for Aup and Bup as the temperature perturbation is the one that most closely follows the E5 result.

In order to investigate this further a closer comparison between the E5 and the vertically retracted temperature perturbations has been performed. On the LHS of Figure 5.5, the differences of the equilibrium temperature profiles with Aup (top), Bup (middle) and Cup (bottom) heating perturbations from the original equilibrium temperature with E5 perturbation are shown. The vertical blue lines indicate the jet position for the unperturbed control run (L26complete), the red line shows the 200 K model reference tropopause and the purple line the model mean tropopause of the control run as defined by the WMO (lapse rate of -2 K/km, equivalent to a static stability of $N^2 = 3.510^{-4} \text{1/s}^2$). The horizontal lines indicate the equatorial reference tropopause level (green) and the 85 hPa level (light blue) above which the full heating is applied for the Aup and Bup perturbations. The orange line in the bottom panel indicates the level above which the full heating is applied for the Cup perturbation. It is obvious that the E5 perturbation reaches well below the model mean tropopause as defined by the WMO (purple line) at all latitudes, as it is applied in full down to the 200 K level (reference tropopause - red line).
Figure 5.5: Difference of heating perturbations Aup, Bup and Cup from the original E5 perturbation. See text for further explanation.
For comparison, the differences of the zonal mean temperatures for the Aup (top), Bup (middle) and Cup (bottom) model runs from the original E5 results are shown on the RHS of Fig. 5.5. The colored lines are the same as Fig. 5.5 LHS. For the Cup perturbation the following two effects lead to an emphasis of the mid-latitude part of the heating perturbation at the tropopause relative to the equatorial part:

- The Cup perturbation raises the base of the full heating by a constant dp which implies a greater height (dz) in the tropics compared to the poles as visible in the bottom panel (comparing the red and the orange lines).

- For the E5 perturbation the distance between the reference tropopause and the model mean tropopause is small at the equatorial regions and largest at the poles. Compared to E5 this results in the full (5 K) heating perturbation in Cup being retracted to well above the tropical model mean tropopause, but to only slightly above it in mid-latitudes and below the model mean tropopause at polar latitudes (beyond \( \pm 60 {\text{deg N/S}} \)).

This emphasis on the mid-latitude / polar part of the heating perturbation at the tropopause is consistent with a weakening of the tropospheric response (weaker poleward mid-latitude jet shift) arising from preferentially equatorial heating. Previous studies (e.g. Haigh et al. 2005, Simpson et al. 2009) found an equatorward shift of the mid-latitude jet in response to both uniform as well as preferential polar warming. As the strongest heating at tropopause level occurs at mid-latitudes for the Cup perturbation, this is counteracting the effects of the equatorial part of the heating.

This is also evident when looking at eddy fluxes (Figure 5.6). Shown are the horizontal eddy momentum flux (left panels) and horizontal eddy heat flux (right panels) anomalies for the E5, Aup, Bup and Cup experiments. The vertical red line indicates the jet position for the unperturbed control run (L26complete).
Figure 5.6: Horizontal eddy momentum flux (left panels) and horizontal eddy heat flux (right panels) anomalies for the E5, Aup, Bup and Cup experiments. The contour intervals are $1.0 \, m^2/s^2$ (left panels) and $0.1 \, K\,m/s$ (right panels). The vertical red line indicates the jet position for the unperturbed control run (L26complete).
The poleward eddy momentum flux anomalies (left panels) show the previously seen (e.g. Chapter 3) dipole structure, with a weakening of the fluxes on the equatorward side of the time mean jet and a strengthening on its poleward side which stretches downward and equatorward so that in the troposphere the maximum $\overline{u'v'}$ is around the jet centre. This is visible in all heating perturbation experiments, though the relative magnitude of the weakening/strengthening varies. Especially for the Aup and Bup experiments only a small weakening (for Aup smaller than the contour interval) of the momentum fluxes on the equatorward side is observed, while the strengthening on the poleward side is more pronounced. The Cup results show a stronger weakening (about 60% of the E5 response) that extends further poleward than for the other experiments and only a very weak strengthening of the momentum fluxes. A weakening of the climatological $\overline{u'v'}$ at a greater latitudinal extent leads to the Cup response becoming more comparable to the uniform heating (U5) response previously studied by Haigh et al. (2005) and Simpson et al. (2009). This is consistent with the observed relatively weak tropospheric response. The eddy momentum flux anomaly results suggest that the magnitude of the weakening on the equatorward side of the jet is not directly related with the magnitude of the jet response. Using spin-up experiments with E5, U5 and P10 (preferentially polar) heating perturbations Simpson et al. (2009) found that it is the poleward momentum flux anomaly across the jet that constitutes the refraction feedback mechanism. It is this component of the response that scales approximately with the zonal wind response (as can also be seen for this work in Figs. 5.4 and 5.6). The negative eddy momentum flux response in the subtropics is part of the direct response not the feedback, but its extent is important to trigger the feedback. Further they found that the direction of the jet shift depends on the position of the momentum forcing and differences in the location of anomalous horizontal eddy momentum flux lead to varying latitudinal extents of the regions of anomalous momentum flux convergence/divergence. They found this to influence the initial zonal wind anomalies around the tropopause and the direct/indirect meridional circulation
changes. An anomalous meridional circulation in response to the E5, Aup, Bup and Cup perturbations is induced through the thermal wind imbalances created by changes in horizontal eddy momentum flux (Figure 5.9 LHS). This anomalous circulation locally counteracts the anomalous eddy momentum forcings and its effects are balanced by the anomalous horizontal eddy heat fluxes (RHS of Figure 5.6) that act to damp the temperature anomalies.

Figure 5.7 shows a schematic of the mechanism for the tropospheric response to a preferentially equatorial stratospheric heating as proposed in Simpson et al. (2011a).

![Figure 5.7: Schematic of the initial tropospheric response (LHS) and subsequent feedback (RHS) for stratospheric heating perturbations, such as E5, Aup and Bup (adapted from Simpson et al. 2011a).](image)

The initial response is shown on the LHS; The altered temperature gradient around the tropopause leads to a weakening of the upward flux of wave activity here. Associated with this is a reduction of eddy momentum fluxes (area in grey circles) around the tropopause on the equatorward side of the jet (vertical dark grey line indicates the jet position). This induces an anomalous mean meridional circulation which results in a decrease of mean zonal winds (red area) on the equatorward side of the jet. The subsequent feedback is shown on the RHS of Fig. 5.7; The initial zonal wind deceleration leads to an altered meridional wind shear which increases the equatorward refraction of the eddies (green arrow).
Associated with this is an increased poleward momentum flux across the jet centre (area in grey circles). This acts to enhance the initial deceleration of mean zonal winds on the equatorward side (red area) and to accelerate the mean zonal winds on the poleward side of the jet (blue area). A shift in baroclinicity and eddy source latitude (thick black arrow) results from the shift in the mean zonal winds. This again strengthens the equatorward propagation of the eddies, enhancing the poleward eddy momentum flux and zonal wind anomalies. The same mechanism is responsible for the jet shifts observed for the E5, Aup and Bup experiments. As shown above the poleward eddy momentum flux anomalies (Fig. 5.6 LHS) exhibit the weakening of the fluxes on the equatorward side of the jet which are part of the initial response and the $\overline{w^2}$ strengthening across the jet which is associated with the feedback response. The magnitude of the jet shift is clearly related to the strength of the poleward momentum flux anomaly across the jet (see Fig.5.4 RHS and Fig. 5.6 LHS). The same mechanism still holds true for the Cup experiment. However, due to the emphasis of the mid-latitude part of the heating perturbation at the tropopause relative to the equatorial part, a more latitudinally uniform decrease in the upward flux of wave activity leads to the initial $\overline{w^2}$ weakening extending much further poleward and even crossing the jet (Fig. 5.6, bottom panel on LHS). This is also shown in Fig. 5.8, which presents a schematic of the initial response to a Cup-like heating perturbation. The Cup poleward eddy momentum flux strengthening across the jet (associated with the feedback response) is much weaker than for the E5, Aup or Bup perturbations, but the feedback mechanism is the same. This is consistent with with the much weaker tropospheric response shown in Fig. 5.4. Overall the Cup experiment exhibits a response that is in between the responses to an E5 heating and a uniform (U5) perturbation. Simpson (2009) show that for uniform heating similar eddy feedbacks result in an equatorward shift of the jet; As the heating perturbation stretches all the way towards the pole (e.g. for U5) the eddy momentum flux anomalies are simply becoming a weakening of the control run eddy momentum flux around the tropopause. Hence the region of increased
Figure 5.8: Schematic of the initial tropospheric response for the Cup
perturbation.

divergence of eddy momentum flux that is responsible for the anomalous direct
circulation in the mid-latitudes has a much greater extent in latitude and the
resulting initial decrease in mean zonal wind stretches further poleward. For the
uniform heating perturbation these initial zonal wind anomalies are less effec-
tive at creating a strong meridional wind shear across the jet centre (as for E5)
and therefore are less effective initially at producing the tropospheric feedback.
However, Simpson (2009) show that some poleward eddy refraction can be seen
towards the poleward side of the jet in the troposphere associated with the de-
crease in zonal wind there. An increase in convergence of eddy momentum flux
(at about 40-45 deg) around where the zonal wind has decreased results in an
acceleration of the zonal wind in this region. They further propose that the tro-
pospheric eddy momentum flux anomalies act to migrate the increased westerly
wind on the equatorward side of the jet towards higher latitudes. A further pole-
ward migrated zonal wind anomaly is now more effective at producing a strong
meridional wind shear (and hence refractive index gradient) across the jet centre
resulting in the tropospheric eddy momentum flux anomalies becoming larger.
The eddy momentum flux feedbacks act to accelerate the zonal wind anomalies
further, leading to an overall equatorward shift of the jet.

The panels on the RHS in Fig. 5.9 show the anomalies (experiment - control) in static stability for the E5, Aup, Bup and Cup experiments. The vertical red line again indicates the jet position for the unperturbed control run (L26complete). Equatorward of the mid-latitude jet the retraced heating experiments show a similar structure of the static stability change as the E5 result, an increase of static stability, but less strong and at slightly higher altitude than for E5. The latitudinal structure of the static stability anomaly is very flat throughout all latitudes for Aup and going slightly lower with higher latitudes for Bup as would be expected from the respective change in the zonal mean temperature (Fig. 5.4). Similarly the Cup static stability anomaly follows the shape of the tropopause as does the anomaly for E5, but overall extends to higher altitudes.

Plots of the difference in static stability of the Aup/Bup and Cup experiments from E5 are shown in Figure 5.10. The static stability anomalies for Aup and Bup consist mainly of an upward shift of the E5 anomaly throughout all latitudes. This is again consistent with the weaker tropospheric response. Equatorward of the jet, the Cup static stability anomaly is weaker than for E5, while poleward of the jet very little difference is seen. This means that the static stability changes poleward of the jet are very similar to the E5 perturbation anomalies. Overall resulting in an increase of static stability close to the tropopause on the poleward side of the jet which is consistent with the emphasis on the mid-/higher latitudes seen for the applied Cup temperature perturbation.
Figure 5.9: Differences between perturbed and unperturbed experiments in streamfunction of the mean meridional circulation (left panels) and static stability (right panels) for the E5, Aup, Bup and Cup experiments. The contour intervals are $10^{10}$ kg/s (left panels) and $10^{-51}$ s$^{-2}$ (right panels). The vertical red line indicates the jet position for the unperturbed control run (L26complete).
Figure 5.10: Differences between Aup/Bup/Cup and E5 experiments in static stability. The contour intervals are $10^{-6}$/s². The vertical red line indicates the jet position for the unperturbed control run (L26complete) and the purple line gives the mean model tropopause (WMO definition of $3.5 \cdot 10^{-4}$/s²).

In addition to the experiments discussed above, two further heating perturbations, E5dp and E5dz, were applied to the control run. Both are E5-like heatings, but moved upwards by a constant pressure distance for E5dp and by a constant height distance for E5dz, i.e. no heating is applied right at the reference tropopause. Figure 5.11 shows the differences between perturbed and unperturbed runs for the E5dp (LHS) and E5dz (RHS) experiments. Shown are the zonal mean temperature (upper panels), zonal mean zonal wind (2nd panels), the horizontal eddy momentum flux (3rd panels) and the static stability (bottom panels) anomalies. Shaded areas in the zonal mean temperature and zonal mean zonal wind plots are not statistically significant at the 95% level. The vertical red line indicates the jet position for the unperturbed control run (L26complete).

In both cases no significant response was found in the troposphere. While a small weakening of the eddy momentum fluxes can be observed in the lower stratosphere equatorward and across the mid-latitude jet, no strengthening on the poleward side can be seen (3rd panels of Fig 5.11). This implies that a direct response (negative eddy momentum flux response) which stretches across the jet latitudes (not just equatorward of the jet as for the original E5 experiments).
Figure 5.11: Zonal mean temperature (upper panels), zonal mean zonal wind (2nd panels), the horizontal eddy momentum flux (3rd panels) and the static stability (bottom panels) anomalies for the E5dp and E5dz experiments. The contour intervals are 0.5 K, 0.5 m/s (right panels), 1.0 m²/s² and 1⁻⁶¹/s², respectively. Shaded areas in the zonal mean temperature and zonal mean zonal wind plots are not statistically significant at the 95% level. The vertical red line indicates the jet position for the unperturbed control run (L26complete).
does not trigger the refraction feedback. The static stability anomalies occur above the tropopause and are weaker than for the E5 experiment. For E5dp the anomaly follows the shape of the tropopause while it is flatter for E5dz consistent with the shapes and positions of the applied heating perturbations. Overall, especially for E5dp but also for E5dz, the results are similar to the Cup perturbation results as would be expected from comparing the applied heating perturbations. For both E5dp and E5dz perturbations the mid-latitude parts are emphasized over the equatorial regions, similar to the Cup heating. As the equatorial (and overall) heating perturbation for both E5dp and E5dz is further removed from the tropopause than for the Cup experiment, any tropospheric response to this heating would have been weak compared to the one caused by a Cup perturbation. The effects from the mid-latitude part of the heating perturbations therefore appear to be strong enough to counteract any effect from the equatorial perturbation, resulting in an insignificant tropospheric response. In order to confirm this it was decided to design further heating perturbations restricted in their latitudinal extent. It would be expected that a E5dp/E5dz/Cup-like heating that is constrained to the equatorward side of the mid-latitude jet, would result in a significant tropospheric response that is in the sense of the E5 heating (i.e. a strong poleward shift of the jet). Such an experiment (E5dz-SH) is presented in section 5.4.

A summary of the tropospheric response to the various retracted heating perturbations is given in Fig 5.12. Shown are the magnitudes of shift of the mid-latitude jet (positive indicating a poleward shift) with error bars showing one standard error on the horizontal axis. All black data points refer to experiments performed on the control run (L26complete). In addition to these the E5, Aup, Bup and Cup perturbations have also been applied to the B1L26 run. B1 again denotes a reduced temperature relaxation time at the surface ($k_s = 1/2 \, [1/\text{days}]$) as introduced in Chapter 3. As for the L15 IGCM experiments, a reduction of the temperature relaxation time for the L26 runs results in a more equatorward position of the mid-latitude jet, for L26 it is $38.18^\circ$ deg for B1 compared to $43.64^\circ$ deg
for the control \( (k_s = 1/4 \ [1/\text{days}]) \). The results for these runs are shown in blue. The vertical axis is used to offset the experiments from each other and has no physical meaning.

![Graph showing jet shift response to various heating perturbations]

Figure 5.12: Magnitude of jet shift (deg) in response to the various heating perturbations. Shown are the results for all experiments using L26complete and all experiments using B1L26. The error bars show one standard error.

As discussed previously the Cup, E5dp and E5dz perturbations applied to the L26complete do not result in a significant jet shift. The E5 perturbation has the largest effect while the Aup and Bup experiments show a weaker response. The corresponding B1L26 experiments (E5dp and E5dz not performed) follow the same pattern, but with stronger magnitudes of response for the E5, Aup and Bup perturbations. This is again consistent with the B1 mid-latitude jet located closer to the equator which, as seen in Chapter 3, is connected to a longer decorrelation time scale \( (\tau = 93\pm13 \ \text{days}) \) for B1L26 compared to \( \tau = 72\pm9 \ \text{for L26complete} \) and hence a stronger magnitude of response to stratospheric forcings. Overall the B1L26 results closely follow the L26complete results (see Appendix A for the equivalent plots of zonal mean temperature, zonal mean zonal wind, eddy
momentum flux and static stability anomalies for the B1 experiments).

5.4 Effect of restricting the latitudinal extent of the heating

Understanding the effect of the latitudinal extent of heating perturbations applied to the stratosphere can aid interpretation of the previous results as well as general understanding of stratosphere - troposphere interactions. The investigation of the tropospheric response to the different retracted heating perturbations in the previous section highlights the importance of the mid-latitude region. Understanding this is of great importance for interpreting the previous results as well as the general understanding of stratosphere - troposphere interactions. In this section a set of seven heating perturbations with varying latitudinal extent is investigated with both the L26 and the L15 versions of the IGCM2.2. The main objectives of this study are to:

- Confirm the role of the mid-latitude part of the Cup, E5dp and E5dz perturbations.

- Investigate the tropospheric response to the various perturbations.

- Study the impact of including a resolved stratosphere (L26 vs L15) on the tropospheric response: The comparison between the L15 and L26 versions of the IGCM2.2 allows investigation of the impact of the presence of a resolved stratosphere on the model response in both the lower stratosphere/tropopause region and in the troposphere to heating perturbations applied in the lower stratosphere. This is not only important for understanding the role of the mid/upper stratosphere but also helping in the design of future experiments and in the choice of appropriate models.

- Investigate the stratospheric response to the perturbations: The use of the L26 IGCM2.2 version allows the investigation of the stratospheric response to such heating perturbations. Previously this was severely compromised
with IGCM2.2 due to the low vertical extent (∼30 km) of the L15 version. Understanding this may be important as a stratospheric response to the applied heating perturbation might alter the forcing region itself and thereby possibly influence the tropospheric response.

5.4.1 Restricting the latitudinal extent for the retracted heating perturbations

In the previous section (5.3) it was shown that for heating perturbations retracted in altitude which emphasized the mid-latitude part of the heating relative to the equatorial part in the lower stratosphere/tropopause region, no significant (or just a very weak) tropospheric response was found. This was the case for the Cup, E5dp and E5dz perturbations. It was suggested that the effects from the mid-latitude part of the heating perturbations were strong enough to counteract any effect from the equatorial part of the perturbation. Previous research (Haigh et al., 2005; Simpson, 2009 and Simpson et al., 2009) showed that the mid-latitude jet shifts equatorward in response to uniform (U5) or preferentially polar heating perturbation applied to the lower stratosphere/tropopause region. In order to investigate whether the mid-latitude part of the Cup, E5dp and E5dz perturbations caused the overall tropospheric response to be reduced, the Cup and E5dz perturbations were restricted to latitudes equatorward of the mean control jet position (L26complete). The full Cup or E5dz heating was applied within 15° each side of the equator and then reduced to zero at and beyond 45° (as sin²(φ) function) - as in the SII perturbation. This sharper cut-off is on the equatorward side of the jet, therefore the heating does not fully cross the jet. For comparison the original E5 perturbation (not retracted in altitude) was also constrained in latitude in this manner.

Figure 5.13 shows the zonal mean temperature (LHS) and zonal mean wind (RHS) anomalies for the E5dz-SII perturbations. The contour intervals are 0.5 K (left panels) and 0.5 m/s (right panels). Shaded areas are not statistically significant at the 95% level. A significant tropospheric response can now be seen,
with the mid-latitude jet shifting polewards. The magnitude of the tropospheric response is about 75% of the full L26E5 response and about 35% of the full L26E5-SII response. A similar result is obtained for the Cup-SII perturbation, with a tropospheric response that is about 120% of the full L26E5 response and about 65% of the full L26E5-SII response.

![Figure 5.13](image_url)

Figure 5.13: Differences between perturbed and unperturbed experiments in zonal mean temperature (left panels) and zonal mean zonal wind (right panels) for the E5dzSII experiment. The contour intervals are 0.5 K (left panels) and 0.5 m/s (right panels). Shaded areas are not statistically significant at the 95% level.

An overview of the results is given in Table 6. These results confirm that the mid-latitude part of the previously used heating perturbations effectively counteracts the tropospheric response that would result from the equatorial part of the heating alone. For each experiment the SII latitude restriction gives a stronger tropospheric response. All retracted SII experiments (L25Cup SII and L26E5dz SII) result in a significant fraction of the original E5 response, unlike the retraction-only (L26Cup and L26E5dz) experiments. This indicates that a SII-like experiment would be better suited to investigate the tropospheric response to solar-forcing as such an experiment focuses more clearly on equatorial perturbations.

Further it is shown that the heating perturbations retracted in altitude still result in about 35% (E5dz-SII)/65% (Cup-SII) of the tropospheric response of the E5-SII (not retracted) perturbation. It is therefore not necessary to apply a
<table>
<thead>
<tr>
<th>experiment</th>
<th>jet position [deg]</th>
<th>jet shift [deg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>L26E5</td>
<td>45.46 ± 0.23</td>
<td>1.82 ± 0.33</td>
</tr>
<tr>
<td>L26Cup</td>
<td>44.04 ± 0.42</td>
<td>0.40 ± 0.48</td>
</tr>
<tr>
<td>L26E5dz</td>
<td>43.62 ± 0.47</td>
<td>-0.02 ± 0.52</td>
</tr>
<tr>
<td>L26E5-SII</td>
<td>47.14 ± 0.66</td>
<td>3.50 ± 0.70</td>
</tr>
<tr>
<td>L26Cup SII</td>
<td>45.79 ± 0.10</td>
<td>2.15 ± 0.25</td>
</tr>
<tr>
<td>L26E5dz SII</td>
<td>45.03 ± 0.21</td>
<td>1.39 ± 0.31</td>
</tr>
</tbody>
</table>

Table 6: Overview of results for the in altitude retracted SII experiments. The errors indicated are one standard error. The jet shifts are always given relative to the control run (i.e. L26complete).

heating perturbation directly onto the reference tropopause in order to achieve a significant tropospheric response. This is consistent with the results obtained by Tandon et al. (2011). As part of their investigation of the tropospheric response to thermal perturbations that mimic stratospheric water vapor changes, they also conducted experiments with a perturbation retracted in altitude. Varying the magnitude and sign of the heating in both the retracted and original perturbation, they found that about 40-45% of the full magnitude of the mid-latitude jet shift was reached with the retracted perturbation. The manner in which the perturbation was retracted in altitude is similar to E5dz. However, due to significant differences in the model physics (e.g. different tropopause shapes) and in the latitudinal structure of the heating perturbation (preferentially polar for Tandon et al. 2011) only a qualitative comparison is possible.

5.4.2 The tropospheric response

In order to investigate the effect of the latitudinal extent of the heating perturbation in a more systematic manner, seven heating perturbations with varying latitudinal extent were applied to the L26 control run (L26complete) and run for 10000 days. The perturbations are called E5, SI, SII, SIII, SIV, SV and U5, detailed descriptions can be found in section 5.2. SI to SV differ from each
other only in their latitudinal extent which increases by 15 deg in each hemisphere between consecutive perturbations. The following analysis is based on all seven experiments, but for clarity only the results for E5, SI, SIII, SV and U5 are plotted. An overview of jet position, decorrelation times and magnitude of jet shift in response to the respective perturbation for all seven experiments and the control run can be found in the Appendix D in Table 7.

Figure 5.14 shows the zonal mean temperature (LHS) and zonal mean wind (RHS) anomalies for the different heating perturbations. The temperature anomalies roughly follow the prescribed perturbations, but are somewhat modified. This modification is most pronounced for the SI experiments where the heating perturbation is spread out over a far greater latitudinal range than prescribed (1 K heating out to about 40 deg instead of ≈21 deg) and the magnitude of the heating at the equator is reduced by about 50% (2.5 K instead of 5 K). The SIII temperature anomaly is also spread out further than the applied perturbation, but instead of a reduction of the heating magnitude at the equator a increase by about 1 K is seen. Further a cooling at the polar regions at about 200hPa can be observed for the E5, SI and SIII experiments (also for SII, not shown). These differences between observed temperature anomalies and applied heating perturbations will be investigated further by comparisons with zonally symmetric runs later in this section. The zonal mean zonal wind anomalies (RHS of Fig. 5.14) show a significant tropospheric response for all experiments. E5, SI and SIII (as well as SII and SIV) show a poleward shift of the mid-latitude jet with SIII giving the largest magnitude and SI the smallest shift. For the SV and U5 perturbations an equatorward shift of the jet is seen, with U5 giving the strongest shift.

The tropospheric zonal wind response for SI is of similar magnitude as the original E5 response even though the magnitude of the temperature anomaly does not reach the full magnitude of the applied perturbation (2.5 K instead of 5 K).
Figure 5.14: As figure 5.11, but for the E5, SI, SIII, SV and U5 experiments.
This shows that if the heating perturbation is confined to the tropics, only a comparatively small magnitude of forcing is needed to obtain a similar magnitude of tropospheric response as in the original E5 experiment. The SII and SIII perturbations result in tropospheric responses that are of similar magnitude and the strongest overall. For SII this was expected as the heating is mostly restricted to the equatorward side of the jet and the temperature anomaly at the equator is about the magnitude of the applied perturbation (5 K). The applied temperature perturbation SIII allows for heating across the jet latitude (going from 5 K at 30 deg to 0 K at 60 deg). The strong poleward shift of the mid-latitude jet for this perturbations shows that the effect from the equatorial part of the heating (5 K for 0-30 deg) is still very strong compared to the mid-latitude part. The jet shift resulting from applying the SIV perturbation is about the same magnitude as for the E5 perturbation, i.e. significantly reduced from the SII/SIII shift (about 1.9 deg compared to about 3.7 deg). This again can be explained with the mid-latitude part of the SIV perturbation counteracting the poleward shift that would result from the equatorial part alone. For the SV and U5 perturbation the effects from the mid-latitude/polar heating become dominant over the equatorial part causing an overall equatorward shift of the jet.

Anomalies in the horizontal eddy momentum flux (hemf - $\overline{u'v'}$), as well as the static stability (SS), are shown for each experiment in Figure 5.15. The contour intervals are 1.0 m$^2$s$^{-2}$ ($\overline{u'v'}$ - left panels) and 10$^{-5}$ s$^{-2}$ (SS - right panels). The vertical red line indicates the jet position for the unperturbed control run (L26complete). The previously observed dipole structure in response to equatorial heating, with a weakening of the fluxes on the equatorward side of the time mean jet and a strengthening on its poleward side can be seen for E5, SI and SIII (also for SII and SIV). SV and U5 show only the weakening of the eddy momentum fluxes at mid-latitudes that was previously (Simpson et al, 2009) connected with equatorward jet shifts. The location of maximum anomalous horizontal eddy momentum flux is very similar for all poleward-shifting perturbations (E5,
Figure 5.15: Horizontal eddy momentum flux (left panels) and static stability (right panels) anomalies for the E5, SI, SIII, SV and U5 experiments.
SI-SIV), but their magnitude varies considerably and is almost absent for SI. As seen for the L26Aup and L26Bup experiments the magnitude of the weakening of the horizontal eddy momentum fluxes is not related to the magnitude of the tropospheric response. The magnitude of the poleward shift of the jet is related to the magnitude of the $\overline{uu'}$ strengthening across and poleward of the jet location.

E5, SI and SIV exhibit similar poleward jet shifts as well as similar magnitudes of the positive horizontal eddy momentum flux anomaly. This is also true for the SII and SIII experiments.

The top panel of figure 5.16 gives the magnitude of the horizontal eddy momentum flux anomaly ($m^2/s^2$) at crossing the jet (at 269hPa for L26 and 240hPa for L15) versus the distance between the jet position and the full 5 K heating. Shown are the results for SI-SV for L15 (blue) and L26 (black). Negative values on the x-axis indicate that the full 5 K heating was applied across the jet and only reduced at higher latitudes (e.g. for SV the full heating stretches from the equator to 60 deg), i.e SI to SV are shown from right to left. As previously seen the magnitude of the positive eddy momentum flux anomaly across the jet scales approximately with the magnitude of the jet shift. Fig. 5.16 (top) shows how this positive $\overline{uu'}$ anomaly relates to the latitudinal extent of the heating perturbation. The perturbations go from very restricted to the equatorial regions (SI) to a more uniform heating (SV), similar to the U5 heating used by e.g. Simpson et al. (2009). It can be seen that a more uniform heating, like SV, results in a negative $\overline{uu'}$ anomaly across the jet. This is consistent with the observed equatorward jet shift. For heatings restricted to lower latitudes the dipole pattern of the $\overline{uu'}$ anomaly (see Fig. 5.15, LHS) associated with a poleward jet shift is visible. The negative $\overline{uu'}$ anomaly on the equatorward side is part of the initial response while the positive $\overline{uu'}$ anomaly across and poleward of the jet is part of the feedback response. This positive $\overline{uu'}$ anomaly is captured in Fig. 5.16 (top) for the more equatorially restricted runs (SI-SIV) and shows a systematically stronger anomaly for more restricted heating perturbations up to the point where the heating is almost completely equatorward of the jet (both
Figure 5.16: Top: Magnitude of the horizontal eddy momentum flux anomaly \((m^2/s^2)\) at crossing the jet (at 269 hPa for L26 and 240 hPa) vs. distance between the jet position and the full 5 K heating. Shown are the results for SL-SV for L15 (blue) and L26 (black). The error bars show one standard error. Bottom: Magnitude of the static stability anomaly anomaly \((10^{-5}s^{-2})\) at crossing the jet (at 269 hPa for L26 and 240 hPa) vs. distance between the jet position and the full 5 K heating. Shown are the results for SL-SV for L15 (blue) and L26 (black). The error bars show one standard error.
for L15 and L26 runs). The magnitude of the $\overline{u'v'}$ anomaly does then decrease again. This is the case for the SI and SII runs which give no heating beyond 30 deg and 45 deg, respectively. For both the SI and SII experiments the zonal mean temperature anomalies (see Fig. 5.14 for SI) do not reach the full magnitude of the applied heating (about 2.5 K instead of 5 K for SI, the effect for SII is much smaller with about 4 K instead of 5 K). This is consistent with a weaker eddy feedback response (as apparent in Fig. 5.16, top) and hence a weaker jet shift. Haigh et al. (2005) performed stratospheric temperature perturbation experiments using both 1 K and 5 K perturbations (for U5 and E5 heating shapes). They showed that the response for the 1 K case was much weaker than for 5 K, but the dynamical mechanisms were the same. The differences in L15 and L26 can be explained by the slightly different pressure levels at which the data were obtained and the different control run jet positions. The L26 mean jet is located further equatorward than the L15 jet. It was shown by e.g. Simpson et al. (2010) that mid-latitude jets located more equatorward exhibit stronger jet shifts in response to stratospheric temperature perturbations than more poleward jets. The exact reason for this behaviour is subject to current research (see chapter 1.2.5). The static stability anomalies (RHS of Fig. 5.15) show a strengthening at about tropopause level and a weakening above this for all heating perturbations. The main difference is the latitudinal extent of the anomaly which follows approximately the extent of the temperature anomaly (LHS in Fig. 5.14). The bottom panel of figure 5.16 is similar to Fig. 5.16 top panel, but gives the magnitude of the static stability anomaly ($10^{-5} s^{-2}$) at crossing the jet (at 269 hPa for L26 and 240 hPa) versus the distance between the jet position and the full 5 K heating. Shown are the results for SLSV for L15 (blue) and L26 (black). A clear relation between the latitudinal extent of the applied temperature anomaly and the static stability anomaly at crossing the jet can be seen. The further the heating perturbation stretches towards the pole the stronger the static stability anomaly across the jet. The SI perturbation again shows a slightly different behaviour which can be explained, as above, with the full magnitude of the temperature
anomaly staying well below the 5 K that were applied. For more equatorial perturbations the positive static stability anomaly is mostly restricted towards the equatorward side of the jet (see also Fig. 5.15, RHS) while more uniform heatings (e.g. SV and U5) exhibit a stronger static stability anomaly across the jet that stretched all the way to the pole. For SV and U5 the effect of increasing the stratospheric temperature (almost) uniformly is essentially a lowering of the whole tropopause. Williams (2006) found such a lowering of the tropopause causes the jet stream to move to lower latitudes and the eddies to become weaker as is demonstrated by U5 and SV here.

The analysis of the tropospheric response to heating perturbations with varying latitudinal extent applied down to the reference tropopause shows that both the magnitude of the response and the direction of the mid-latitude jet shift are dependent on the position of the heating with respect to the mean jet latitude. For very narrow heating perturbations such as the SI the mean temperature anomalies do not achieve, in the stratosphere, the applied magnitude and shape, but are rather spread out and hence weakened, resulting in correspondingly weaker changes of the eddy fluxes about the tropopause and therefore a weaker tropospheric response. The perturbations SII to SV show a clear progression with the eddy momentum flux anomalies about the tropopause level going from a clear dipole pattern to a general weakening of the horizontal eddy momentum flux. Consistent with this the tropospheric response changes with increasing latitudinal extent of the heating perturbations from a strong poleward jet shift to an equatorial jet shift. Understanding this behaviour is important not only for the interpretation of model results, but also for understanding real world, temperature perturbations. In the real world temperature perturbations in the stratosphere can occur at different latitudes (e.g. preferentially equatorial solar heating and polar cooling due to water vapor changes) and their combined effects might need to be considered to understand the tropospheric response.
5.4.3 Comparison of the L26 and L15 IGCM2 tropospheric response

In order to compare the tropospheric response to the above heating perturbations in the L26 and L15 versions of the IGCM2.2, each of the experiments has also been run on the L15 version. A list of all L15 experiments is given in Table 4 (section 5.2) and results have been summarised, as for L26, in Table 8 (Appendix D). All experiments plus control were run for 10000 days (plus 1000 days discarded as spin-up) and with the same time step as the L26 experiments (TSPD=96). This allows investigation of the impact of the presence of a resolved stratosphere (L26) on the model response to heating perturbations applied in the lower stratosphere. This section will focus on the tropospheric response and changes at the tropopause level. The L15/L26 comparison of the response in the lower stratosphere will be discussed in section 5.4.5.

First the anomalies of zonal mean temperature, zonal mean zonal wind, horizontal eddy momentum flux and static stability are investigated. The corresponding plots can be found in Appendix B (T and u in Fig. B.1 and $\overline{u'v'}$ and SS in Fig B.2). The analysis is again based on all seven experiments, but for clarity only the results for E5, SI, SIII, SV and U5 are plotted. Comparing these anomalies with the corresponding L26 anomalies it can be seen that both shape and magnitude are very similar. The L15 changes in tropospheric circulation are slightly weaker for the E5, SI - SIV perturbations (poleward shift of mid-latitude jet) while they are slightly stronger for the SV and U5 experiments (equatorward shift of mid-latitude jet). This is consistent with the slight L15/L26 differences in the horizontal eddy momentum flux anomalies where the strengthening of the poleward $\overline{u'v'}$ is weaker than for L26 while the overall weakening of the $\overline{u'v'}$ for SV and U5 is slightly stronger for the L15 experiments. In general the differences in tropospheric response for the L15 and L26 experiments are small.

For a more quantitative comparison the mean jet positions of the respective control runs (L15control and L26complete) and the magnitudes of the mid-latitude jet shifts were determined by the method discussed in Chapter 3.2. On the LHS of Figure 5.17 the magnitudes of the jet shift for each heating perturbation for
both L26 (black) and L15 (blue) are shown. The experiments are offset against each other on the horizontal axis for clarity and easier comparison between L15 and L26. The error bars show one standard error. It is apparent that there is no significant difference in the magnitude of mid-latitude jet shift in response to each heating perturbation between the L26 and L15 runs. Consistent with the anomalies discussed above, the poleward jet shifts for L15 have a slightly smaller magnitude than for L26 while the equatorward jet shifts are slightly stronger.

As shown in Chapter 3, more poleward positions of the mean mid-latitude jet are related with shorter decorrelation times $\tau$. Climatologies with shorter decorrelation times should respond less strongly to the applied heating perturbations (according to the Fluctuation - Dissipation theorem). The mid-latitude jet of the L15 control run is slightly (about 2 deg) more poleward

Figure 5.17: Magnitude of the jet shift (deg) in response to each heating perturbation E5, SI-SV and U5 for both the L15 (blue) and the L26 (black) experiments. Shown are the jet shift vs. experiment ID.

than for the L26 control run. Therefore a weaker tropospheric response might be expected for all experiments. The corresponding decorrelation times for the control runs are however not significantly different from each other (L26: 72
±9 days and L15: 63 ±7 days) which is consistent with observed the jet shifts. Figure 5.18 shows the mean jet positions versus the decorrelation times (LHS) for all experiments conducted for this Chapter, including the control runs and the experiments with heating perturbations retracted in altitude as well as the experiments performed with B1 ($k_s = 1/2$ [1/days]) and B3 ($k_s = 1/8$ [1/days]) control runs. A list of all these experiments is given in Tables 4 and 5. While the relation between $\tau$ and the mean jet position is quite well constrained for jets poleward of $\approx 43$ deg, it is not as clear for more equatorward jets. This indicates that care must be taken when trying to draw conclusions about the tropospheric response based on either the decorrelation time or the mean jet position for climatologies with more equatorward mean jets. The RHS of Figure 5.18 shows the projection of the zonal mean zonal wind response onto the first EOF versus magnitude of the jet shift for all heating perturbation experiments performed for this chapter.

Figure 5.18: LHS: Jet position (deg) vs decorrelation times (days) for all experiments performed for this chapter. RHS: Magnitude of the jet shift (deg) vs. projection onto EOF1 (m/s) for all heating perturbations used for this chapter. The error bars show one standard error. Blue indicates experiments performed with L15 and black experiments with L26.

Both the L15 (blue) and the L26 (black) results exhibit the same linear relation. For experiments where the resulting magnitude of the jet shift is large, the tropospheric zonal wind response projects very well onto the first EOF. Together with
the Fluctuation-Dissipation theorem (i.e. the largest magnitudes of response occur for control run simulation with greater persistence of variability - i.e. long decorrelation times), this is consistent with the eddy feedback mechanism (Simpson et al., 2009 and 2010) whereby the feedback of eddies and the mean flow is thought to be responsible for maintaining the persistence of the variability in the control run as well as for amplifying the response to the heating perturbations. The above comparison of the experiments run on the L15 and the L26 versions of the IGCM2.2 shows that the tropospheric response to the various heating perturbations is not significantly altered by the inclusion of a resolved stratosphere. This indicates that for studies where the focus is exclusively on the tropospheric response to perturbations without vertical structure applied to the lower stratosphere/tropopause region the use of a low-top model, such as the L15 IGCM2.2, is sufficient. However, for the investigation of any stratospheric perturbations which exhibit a vertical structure, an extended model (such as the L26) is needed to capture this vertical structure (e.g. water vapor- or QBO-like perturbations).

5.4.4 Zonally symmetric simulations

In order to better understand the role of eddy processes in altering the forcing region itself and hence the tropospheric response to heating perturbations with varying latitudinal extent, all experiments (E5, S1-SV and U5) have been run in a zonally symmetric mode. In this mode there are no explicit eddies. The eddy fluxes of heat and momentum are prescribed as those required to maintain the control run equilibrium state. Further to avoid artificial vertical advection (the $\frac{\partial u}{\partial z}$ term in the thermodynamic equation) of the anomalies down into the troposphere, the spectral divergence is set to zero after reading in the control zonal mean state. Without this the climatological meridional circulation would act to advect the stratospheric temperature increase down into the subtropical troposphere (as seen in Fig. 8 of Simpson et al., 2009). In the 3D simulations this is counteracted by a stronger anomalous upward vertical motion in the sub-

180
tropics.
The heating perturbation is then applied and each experiment is run for 200 days. This relatively short run length is sufficient to reach an equilibrium response as no eddy feedbacks can occur in the zonally symmetric model, hence the variability is much reduced. Comparing the zonally symmetric results with the full 3D runs allow determination which part of the response to the heating perturbations is attributable to eddy feedbacks. Figure 5.19 shows the zonally symmetric temperature (LHS) and zonal wind (RHS) anomalies for the E5, SI, III, SV and U5 experiments. Shown are the differences between day 0 (equivalent to the control run L26 complete 10000 day mean) and day 200 after applying the respective heating perturbation for the zonally symmetric runs. Contour intervals are 1 K and 1 m/s, respectively and for clarity the zero contour lines are not shown. All plots shown here are for experiments performed with the L26 version of the IGCM2.2. The equivalent zonally symmetric plot for the L15 experiments can be found in Appendix B.
The temperature anomalies (LHS) follow the prescribed perturbations very well for almost all experiments. For the SI experiment however, the heating perturbation is spread out over a far greater latitudinal range than prescribed and the magnitude of the heating at the equator is reduced as also observed for the 3D simulation. All experiments show a temperature maximum that is located at the equator between about 100 - 200 hPa, different from the applied perturbations which consist of a constant temperature anomaly above the tropopause. The zonal mean zonal wind anomalies (RHS of Fig. 5.19) arise from the changes in the temperatures. Westerlies develop in response to the meridional temperature gradients as a result of the thermal wind relation. Progressing from run SI to SV a clear shift of these westerlies towards higher latitudes consistent with the meridional shift in the temperature gradients can be seen. Weak easterlies can be seen at the equatorward side of the westerly anomalies where the temperature anomaly gradients are reversed at the base of the heating.
Figure 5.19: Zonally symmetric temperature (LHS) and zonal wind (RHS) anomalies for the E5, SI, SIII, SV and U5 experiments (see main text).
The extent of the easterlies is closest to the jet latitude (but still equatorward of it) for the SIII experiment. In the full 3D experiment these easterlies provide a strong trigger for eddy feedback to shift the jet (as seen in the 3D version zonal wind response in Fig. 5.14). For the U5 experiment only weak easterlies occur in the lower stratosphere/tropopause region. The tropospheric response observed for the 3D runs is however absent in the zonally symmetric simulations.

The above results also hold true for the experiments performed with the L15 version (Fig. B.3). Using a similar analysis, Simpson (2009) and Simpson et al. (2009) were able to show that changes in the eddies are the driving force of the tropospheric response. The full tropospheric response that occurs in the 3D simulations could not be produced with the zonally symmetric model in which the eddy fluxes are held fixed (as is also obvious from Fig. 5.19 here). The results of the zonally symmetric simulations will be used in the next section to investigate the effects of eddy feedbacks on the forcing region itself and hence to understand the stratospheric response to the heating perturbations.

5.4.5 The stratospheric response

The equilibrated stratospheric response in the 3D model differs from the zonally symmetric runs as shown in the previous section. The L26 IGCM2.2 allows for an investigation as to whether this difference is due to in-situ stratospheric processes or forcing associated with the tropospheric eddy response which might result in a potential feedback on the forcing region. Previously this could not be studied in detail with IGCM2.2 due to the low vertical extent (up to 18.5 hPa) of the L15 version. In this section it is shown that the stratospheric response to the applied heating perturbations can alter the forcing region itself and hence influence the tropospheric response. Investigating the effect on the forcing region will help to develop a more complete picture of stratosphere - troposphere interactions. In addition to this, comparing the response in the lower stratosphere/tropopause region between the L26 and L15 versions allows further assessment of the impact of including a resolved stratosphere on the model results.
The lower stratosphere/tropopause region

Figure 5.20 compares the temperature (LHS) and the zonal wind (RHS) anomalies between the zonally symmetric runs for experiments E5, SI, SIII, SIV and U5 and the full 3D runs. The results of the full 3D runs are represented by the black contours with intervals of 1 K and 1 m/s, respectively. The colors indicate how the zonally symmetric run anomaly compares to the full 3D run anomaly. Red colors indicate that the zonally symmetric run exhibits an equal or stronger anomaly, while blue colors indicate a different sign of anomaly. The numbers on the color bar show the magnitude of the zonal symmetric run anomaly compared to the 3D run anomaly, e.g. 100 indicates the anomalies are of equal strength and 200 that the 2D anomaly has twice the magnitude of the 3D run anomaly. The equivalent zonally symmetric plot for the L15 experiments can be found in Appendix B (Fig. B.4).

First, looking at the temperature anomalies in the polar regions, it is obvious that the zonally symmetric simulations cannot explain the cooling (at poles up to 30 hPa) observed for the full 3D experiments (seen for E5, SI-SIV). This cooling must therefore be created by eddy feedback processes. In order to investigate this, additional eddy diagnostics have been plotted. All fields are plotted with log-pressure scale. Figure 5.21 shows the equivalent plot for the temperature and zonal wind anomalies. The Eliassen-Palm (EP) flux divergence and the Transformed Eulerian Mean (TEM) residual circulation anomalies are shown in Figure 5.22. The TEM meridional and TEM vertical motion anomalies are displayed in Fig. 5.23. These diagnostics are again shown only for experiments E5, SI, SIII, SIV and U5. All equivalent plots for the L15 experiments can be found in Appendix B (Figs. B.5-7). All, L26 and L15, respective control field plots can be found in Appendix C.

An weakening of EP-flux convergence can be seen in the lower stratosphere/ tropopause region at all latitudes for the E5 and SI-SIV experiments. This increase in EP-flux divergence reduces the residual circulation (as seen in the TEM
Figure 5.20: Comparison of the temperature (LHS) and the zonal wind (RHS) anomalies between the zonally symmetric runs for experiments E5, SI, SIII, SIV and U5 and the full 3D runs, see main text.
plots in Fig. 5.22, RHS). As the stratification in the lower stratosphere is strong (large $d\theta/dz$ or $N^2$) a change in the strength of $w*$ changes the vertical advection. The TEM thermodynamic equation (eqn. 2.27) in steady state is a balance between heating and vertical advection by the residual circulation. So if the vertical advection is changed the temperature changes until the heating balances the vertical advection. This results in an anomalous warming or cooling. In the case of the E5, SI-SIV simulations a weaker descent (or stronger ascent) produces a cooler temperature locally as seen in the polar regions. As this modifies the forcing region in the lower stratosphere/ tropopause, it will have an effect on the tropospheric response. The resulting stronger meridional temperature gradient about the tropopause leads to a stronger tropospheric response.

In the lower stratosphere the polar ascent anomaly (Fig. 5.23, RHS) seen in E5 and in the more confined S-experiments is replaced by descent (i.e a downward extension of the deep circulation above) for heating perturbations extending to higher latitudes (SIV and SV). At low latitudes the reverse occurs, especially near the equator at about 100hPa. The variations in equatorial and polar temperature anomalies (Fig. 5.21, LHS) at this pressure are consistent with this. This is also consistent with the comparison of the temperature anomalies between the zonally symmetric and the 3D runs (Fig. 5.20, LHS). In low latitude regions at about 100 hPa the zonally symmetric results cannot account for the full temperature anomaly. This feature is visible for all simulations and gets stronger (i.e. less temperature anomaly explained by zonally symmetric run) and extends higher up, the further the heating extends to the pole (i.e. SI to SV). This further modifies the forcing region itself.

Moreover the above responses to the heating perturbations in the lower stratosphere/ tropopause region look very similar for the L15 simulations. This indicates that in order to capture the processes important for modifying the forcing region a resolved stratosphere is not necessary. This is consistent with the tropospheric response also being very similar for both L15 and L26.
Figure 5.21: Zonal mean temperature (left panels) and zonal mean zonal wind (right panels) anomalies for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are 1 K and 1 m/s, respectively.
Figure 5.22: EP-flux divergence (left panels) and TEM residual circulation (right panels) anomalies for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are half-logarithmic at 1,2 and $5 \times 10^n$ $m^3$ and kg/s, respectively.
The stratosphere above 30hPa

The structure of the anomalies in the higher stratosphere changes significantly. For the E5, SI-SIV experiments the EP-flux anomalies change sign above approximately 30 hPa and are of similar magnitude there throughout all latitudes (Fig. 5.22, LHS). The TEM residual circulations reflect this change in EP-flux divergence (Fig. 5.22, RHS). Throughout the whole stratosphere the EP-flux divergence is in the correct sense to force the deep TEM residual circulation (i.e. the residual circulation strengthens above about 30 hPa).

For all L26 experiments, the heating extends all the way up to 0.1 hPa in the mesosphere, creating a thermal wind on its poleward edge through the depth of the stratosphere. The maximum of this westerly anomaly correspondingly shifts further polewards with the heating extending to higher latitudes (Fig. 5.21, RHS). For the E5, SI-SIV experiments the TEM $w^*$ anomalies (Fig. 5.23, RHS) show ascent in depth in the heating region and descent poleward. The zero line is roughly coincident with the heating gradient. This anomalous $w^*d\theta/dp$ (eqn. 2.27) acts to cool the tropics and warm high latitudes. Thus the warming is effectively spread out and its latitudinal gradient is reduced throughout the middle/upper stratosphere. This is consistent with the zonally symmetric results exhibiting a steeper gradient of the temperature anomaly in the upper stratosphere. This can be seen in the comparison plots for the full 3D temperature anomalies versus the zonally symmetric results (Fig. 5.20, LHS) which show that the zonally symmetric temperature anomaly above 50 hPa is warmer in the equatorial regions (for E5 and SI-SIV) and cooler at the poles.

The comparisons with the zonally symmetric run for both the L26 and L15 (Fig. B.4) experiments show similar results up to 18.5 hPa (the upper limit of L15). The L15 simulations also exhibit the relative reduction of the temperature anomaly gradient in the highest layers (above 30 hPa) and relative strengthening of the temperature anomaly gradient in the lower stratosphere/ tropopause region for the 3D runs compared to the zonally symmetric cases.
Figure 5.23: TEM meridional (left panels) and TEM vertical (right panels) motion anomalies for the E5, SI, SH, SV and U5 experiments with log-pressure scale. The contour intervals are half-logarithmic at 1, 2 and $5 \times 10^6$ m/s and Pa/s, respectively.
The above analysis demonstrates that eddy feedbacks modify both the lower stratosphere/ tropopause region as well as the middle/ upper stratosphere significantly altering the structure of the temperature anomaly in the forcing region. The behaviour of the eddy feedbacks is shown to be influenced in a systematic way by the latitudinal extent of the applied heating perturbation.

5.5 Discussion and Conclusion

The effects of altering the shape and position of heating perturbations applied to the IGCM2.2 have been investigated with the aim of improving the current understanding of stratosphere - troposphere interactions. The main results and conclusions are now summarised.

First, the impact of retracting the perturbations in altitude has been studied. The full heating at the equator was applied above ≈ 85 hPa for the retracted perturbations as opposed to at about 160 hPa for the original E5 perturbation. It has been found that:

- perturbations for which the full heating is moved above the reference tropopause result in weaker anomalies (in zonal mean temperature, zonal mean zonal winds and eddy fluxes) about the tropopause region, but the patterns of these anomalies are the same as for heating perturbations applied directly to the reference tropopause. The tropospheric responses are subsequently weaker than for the non-retracted heating perturbations. This indicates that the mechanisms responsible for creating the tropospheric response are the same for the retracted and non-retracted perturbations.

- The analysis of the retracted perturbation simulations further highlights the role of the mid-latitude parts of these heating perturbations. It has been shown that for heating perturbations retracted in altitude which emphasized the mid-latitude part (i.e. heating retracted mainly in the tropics) of the heating relative to the equatorial part (E5dp, E5dz and Cup) in the
lower stratosphere/tropopause region, no significant tropospheric response was found. Further investigation with retracted heating perturbations that were restricted to be equatorward of the control run mid-latitude jet shows that the mid-latitude part of the heating perturbations effectively counteracts the tropospheric response that would result from the equatorial part of the heating alone. This explains why the L26E5dp/dz and L26Cup experiments do not show a significant tropospheric response. Restricting these heating perturbations in latitude results in a significant tropospheric response of about 35% (E5dz-SIII)/65% (Cup-SII) of the E5-SII (not retracted) perturbation response. It is therefore not necessary to apply a heating perturbation directly onto the reference tropopause in order to achieve a significant tropospheric response.

The impact of the latitudinal extent of the heating perturbation has been investigated in detail by designing five heating perturbations (Sl-SV, in addition to E5 and U5) with varying latitudinal extent (i.e. more confined gradient which is located at different latitudes in 5 cases) and applying these to both the L26 and the L15 control runs. The various simulations were studied with respect to:

- The L26 tropospheric response
  It has been shown that both the magnitude of the response and the direction of the mid-latitude jet shift are dependent on the position of the heating with respect to the mean jet latitude. For very narrow heating perturbations such as the SI the mean temperature anomalies do not achieve the applied magnitude and shape, but are rather spread out and hence weakened, resulting in correspondingly weaker changes of the eddy fluxes about the tropopause and therefore a weaker tropospheric response. It has to be noted, however, that the SI tropospheric response is still of similar magnitude to the original E5 response. Even though the warming is weaker, its confinement to the tropics results in a comparable (to E5) response as not opposing effect from warming the mid-latitudes exists for SI. The perturbations SIII to SV show a clear progression from a preferentially
equatorial heating to a uniform perturbation with the eddy momentum flux anomalies about the tropopause level going from a clear dipole pattern to a general weakening of the horizontal eddy momentum flux. Consistent with this the tropospheric response changes with increasing latitudinal extent of the heating perturbations from a strong poleward jet shift to an equatorial jet shift. The strongest response is found for the SI/III experiments which exhibit maximal tropical heating at the same time as limited heating equatorward of the jet. Such perturbations seem to be an ‘optimal forcing’ for shifting the jet poleward.

• Comparison of the L26 and L15 IGCM2 tropospheric response

It has been shown that the tropospheric response to the various heating perturbations is not significantly different between experiments run on the L15 and the L26 versions of the IGCM2.2. Therefore including of a resolved stratosphere in the model is not necessary to achieve the tropospheric response for the investigated perturbations. This indicates that for studies where the focus is exclusively on the tropospheric response to perturbations (without further vertical structure) applied to the lower stratosphere/tropopause region the use of a low-top model such as the L15 IGCM2.2 is sufficient.

• The stratospheric response

Comparisons with zonally symmetric simulations have been used to show that eddy feedbacks modify both the lower stratosphere/tropopause region as well as the middle/upper stratosphere - thereby significantly altering the structure of the temperature anomaly in the forcing region. The behaviour of the eddy feedbacks was shown to be influenced in a systematic way by the latitudinal extent of the applied heating perturbation.

It has been shown that variations in the shape and position of heating perturbations applied to the stratosphere have a significant influence on the magnitude and direction of the tropospheric response. Further the analysis highlighted the
importance of eddy feedback processes for modifying the forcing region itself and thereby altering the tropospheric response. The results suggest that the mechanisms responsible for creating the tropospheric response are the same for all heating perturbations used for this study.

In addition to this it has been shown that using a model that includes a resolved stratosphere does not significantly alter the response to heating perturbations applied on down to the model reference tropopause. Using a less computationally expensive low-top model, such as the IGCM2.2 L15, is hence sufficient for investigations of the tropospheric response to such perturbations.

These findings will help to inform decisions on which type of model to use in future studies as well as to design further stratospheric perturbation experiments of stratosphere - troposphere interactions.
6 Conclusions and Future Work

The purpose of this research was to improve the understanding of stratosphere-troposphere interactions. For this a simplified general circulation model, the IGCM2.2, was used. As it is crucial to understand the model behaviour in order to correctly interpret any simulation results, the model sensitivities to parameter changes have also been studied as part of this thesis. In order to enable more detailed studies of the impact of varying stratospheric temperature perturbations, a vertically extended version of the IGCM2.2 was developed.

The main conclusions for each chapter are summarised below (section 6.1). This is followed by suggestions for future work (section 6.2).

6.1 Conclusions

The research conducted for this thesis has helped to build up evidence that stratospheric temperature perturbations have a significant impact on the troposphere with the shape and positions of these perturbations being of great importance for the direction and magnitude of the tropospheric circulation anomalies. Further the importance of the position of the mid-latitude jet for the magnitude of response to stratospheric perturbations has been confirmed. In addition to this the work presented here has provided a new tool, the L26 IGCM2.2, with which to study stratosphere-troposphere interactions.

6.1.1 Effects of changing the surface temperature relaxation timescale

The influence of changing the surface temperature relaxation timescale on the model climatology and its effect on the response to a stratospheric heating perturbation was investigated. It was shown that changes to the model's surface temperature relaxation timescale have a significant impact on several aspects of model behaviour.

Both the thermal structure of the lower troposphere and the position of the
mid-latitude jet were found to be influenced by the choice of surface relaxation timescale. Further, it was confirmed that a shorter surface relaxation time improves the temperature structure in the lower equatorial regions as the previously observed unrealistic cold layer did not occur in such experiments. Longer surface relaxation times resulted in a worsening of the cold layer. The observed changes in the jet positions were related to the changes in the low level baroclinicity via the thermal wind relation. A strong temperature relaxation at the surface was shown to lead to a more equatorward jet while weaker relaxations result in more poleward jets.

By changing the surface temperature timescale only the low level temperature is forced, but a deep tropospheric response is obtained. This implied that the annular mode response can therefore be triggered by altering low level baroclinity just as well as by changing winds on the flanks of the jet at the tropopause, i.e. by triggering either the baroclinic or refraction components of the eddy feedback. It was shown that the latitudinal jet variability could not be reduced by relaxing the low-level baroclinicity more strongly. A stronger temperatures relaxation might act to reduce the latitudinal jet variability, any such effect however, is small compared to the effect the mean jet position has on the variability. Investigation of the jet variability showed that model configurations with lower latitude jets exhibit longer timescales of variability than when the jet is located at higher latitudes. This was shown to be in agreement with previous research (e.g. Son and Lee, 2005; Gerber and Vallis, 2007; Gerber et al. 2008; Simpson et al. 2010 and 2011). Two distinct behaviours, propagating and quasi-stationary, were found for the mid-latitude jet. The propagating behaviour strongly dominates the zonal wind anomaly time series for the highest latitude jet, while the lowest latitude jet exhibits mostly quasi-stationary behaviour. This is consistent with results from previous studies (e.g. Son and Lee, 2006; Codron, 2006 Simpson et al., 2009; Sparrow et al., 2009) and is subject to current research (e.g. S. Sparrow, personal communication).

Changes to the surface temperature relaxation timescale were shown to result
in patterns of tropospheric response that were remarkably similar to those observed by Haigh et al. (2005), Simpson (2009) and Simpson et al. (2009). This remarkable similarity of the zonal wind anomaly patterns implies that the same dynamical feedbacks are triggered in the surface temperature relaxation experiments. It was shown that a stronger response is found when the longer timescale quasi-stationary type of variability dominates, i.e. where the jet is located more equatorward. A clear linear relation between the decorrelation times of the unperturbed run and the magnitude of poleward jet shift in response the the heating was found (in agreement with the Fluctuation-Dissipation Theorem). The importance of the jet latitude for the magnitude of response is subject of ongoing research (e.g. Simpson et al., 2011).

Further, the patterns of tropospheric response resulting from the surface parameter changes and their similarity to the response patterns caused by stratospheric perturbations confirm that this particular pattern (i.e. a mid-latitude jet shift) cannot be attributed to any particular kind of forcing (e.g. solar forcing). Rather this is a generic pattern of response that occurs for various forcings and is not restricted to stratospheric temperature perturbations. Hence the occurrence of such a response pattern in observations and climate models could be due to a variety of causes. This has to be noted when trying to draw conclusions from modeling results.

For future model design it should be noted that the surface temperature relaxation time experiments did not succeed in reducing the jet variability by clamping it’s position more strongly. A reduction of the variability is likely to be possible when a stronger relaxation is applied while at the same time the jet is forced to be stationary. This could be achieved by altering $\Delta T_y$ and $T_0$ (in eqn. 2.6) such that the polar $T_{eq}$ and $T(\phi = 0, \sigma_{NL})$ are the same as in the control experiment. However, as $k_s \propto \cos^4 \phi$ and $T_{eq} \propto \cos^2 \phi$, the design of such an experiment is not trivial.

While the above research showed robust results, several limitations apply to this study; First of all, the model used to obtain the data, the IGCM2.2, is a simpli-
fied model and hence is not able to simulate the real world atmosphere. Several of the simplifications have an impact on the models ability to reproduce realistic stratosphere-troposphere interactions, e.g. the absence of surface features like mountains or sea-land temperature contrasts results in large-scale waves being severely underrepresented. In the real world these are thought to play an important role in stratosphere-troposphere interactions. While the model runs for this thesis concerned with stratosphere-troposphere interactions did not intend to represent real world interactions, it has to be noted that the conclusions drawn from these experiments can only cover a piece of the whole picture of stratosphere-troposphere interactions in the real atmosphere.

Further the calculation of the decorrelation times scales that were used to investigate the internal variability and, together with the magnitudes of the jet shifts, interpreted in terms of the fluctuation-dissipation theorem is only a rough measure. While the applied method of calculation is widely used, other methods that might give more accurate results are currently investigated (Sparrow, personal communication). The conclusions of the research for this thesis are however not expected to be influenced by a more accurate calculation of the decorrelation times.

Another limitation lies in the length of the model runs. With limited computational resources, even experiments with simplified models are not feasible to be run for very long periods of time. It was chosen to use 10000 day runs for this research as the model results seemed to be equilibrated after this time and no significant differences to 30000 day runs could be found. It can however not be assumed that longer runs may not be beneficial for future/more detailed analysis. Especially studies concerned with exact decorrelation times and magnitudes of response might need model runs in excess of 100000 days (Simpson, personal communication).
6.1.2 The vertically extended model

A new 26 level version of the IGCM2.2 was developed and investigated in this chapter. Several model updates to both increase the numerical stability and to improve the representation of the stratosphere were shown to be necessary for achieving a stable model. The include:

- Changing the vertical model extend from 18.5 hPa to 0.1 hPa
- Increasing the number of time steps per day from 48 to 96
- Treatment of the model top - additional $\nabla^2$ diffusion for the top three model layers
- A more realistic formulation of the stratospheric temperature relaxation timescale
- A more realistic stratospheric component of the equilibrium temperature $T_{eq}^S$

Overall the new L26 version of the IGCM2.2 was shown to produce a realistic climatology within the constraints of the simplified GCM formulation. In the equatorial stratopause region, however, an anomalous zonal wind signature was observed. While L15 IGCM2.2 control zonal wind in the equatorial regions consists of easterlies which increase in strength with height and the same hold true for the L26t96 experiment, any further addition to the model (aside from reducing the time step) were shown to result in a pattern of zonal winds with minimum easterlies (or even westerlies) roughly at the stratopause level. The cause of the observed anomaly in the equatorial stratopause could not be determined. It was shown that the occurrence and magnitude of this anomaly cannot be linked with the numerical stability of the model nor with the magnitude of the equatorial inertial instability and a cospectral analysis to investigate possible wave driving from below the stratopause did not yield conclusive results. It was however, shown (in Chapter 5) that the occurrence of this stratopause anomaly
does not alter the tropospheric response to heating perturbation applied to the stratosphere.

There are, however, limitations associated with the use of this extended model; The representation of the stratosphere is very crude and not suited for detailed investigations of the stratosphere itself. There is, for example, no realistic stratospheric variability in the model. Further the above mentioned stratospheric equatorial feature limits suggests that further research into its origins is required. The use of the model for studies concerned with this region should take account of this. In addition the limitations for a simplified model mentioned in the last section (6.1.1) still hold true for this extended model.

Despite its limitation this new vertically extended version of the IGCM2.2 was shown to provide an important tool for the research presented in chapter 5 of this thesis and is expected to be very valuable for future research into stratosphere-troposphere interactions.

6.1.3 Effects of altering the shape and position of heating perturbations

The effects of altering the position of stratospheric heating perturbations applied to the IGCM2.2 have been investigated with the aim of improving the current understanding of stratosphere-troposphere interactions. Experiments with stratospheric heating perturbations (a) retracted in altitude and (b) restricted in latitudinal extend were conducted.

a) Heating perturbations retracted in altitude

It was shown that stratospheric perturbations for which the full heating is moved above the reference tropopause result in weaker anomalies (in zonal mean temperature, zonal mean zonal winds and eddy fluxes) about the tropopause region and hence a weaker tropospheric response. The patterns of these anomalies were however shown to be similar as for heating perturbations applied directly to the reference tropopause. This indicates that the mechanisms responsible for creat-
ing the tropospheric response are the same for the retracted and non-retracted perturbations.

The retracted perturbation simulations further highlighted the role of the mid-latitude parts of the perturbations. Heating perturbations retracted in altitude which emphasized the mid-latitude part of the heating relative to the equatorial part in the lower stratosphere/tropopause region were found not to result in a significant tropospheric response. It was shown that the mid-latitude part of the heating perturbations effectively counteracts the tropospheric response that would result from the equatorial part of the heating alone, thus explaining why some of the experiments did not show a significant tropospheric response.

Restricting these heating perturbations in latitude resulted in a significant tropospheric response. It was therefore shown that it is not necessary to apply a heating perturbation directly onto the reference tropopause in order to achieve a significant tropospheric response.

While the sensitivity of the tropospheric response to the amplitude of the forcing was not explicitly investigated, two experiments with higher magnitudes of forcing (10 K instead of 5 K) have been performed. These were the equivalents to E5dp and E5dz, i.e. E10dp and E10dz. Whereas the original 5 K experiments did not yield significant tropospheric responses, the double amplitude experiments resulted in significant jet shifts (1.12 ± 0.45 deg and 1.91 ± 0.34 deg, respectively). Thus the tropospheric response is clearly sensitive to the amplitude of the forcing as would be expected from previous research. For example, Haigh et al. (2005) used both 1 K and 5 K temperature perturbations and found stronger responses for stronger forcings. Further as part of their research on water vapor like forcings Tandon et al. (2011) performed stratospheric temperature perturbation experiments with varying magnitude of forcings and also found a strong sensitivity of the tropospheric response to the forcing amplitude. Their results even indicated an almost linear relation. It might be of interest to perform a systematic investigation of this relationship as this would further help to show to what extent model temperature perturbations can be exaggerated in
comparison to real world forcings in order to produce clearer results (e.g. using 5 K for solar-like forcings instead of 1 K).

**b) Heating perturbations restricted in latitudinal extent**

It was shown that both the magnitude of the response and the direction of the mid-latitude jet shift are dependent on the position of the heating with respect to the mean jet latitude. For very narrow equatorial heating perturbations the mean temperature anomalies do not achieve the applied magnitude and shape, but are rather spread out and hence weakened, resulting in correspondingly weaker changes of the eddy fluxes about the tropopause and therefore a weaker tropospheric response. It has to be noted, however, that the SI tropospheric response is still of similar magnitude to the original E5 response. Even though the warming is weaker, its confinement to the tropics results in a comparable (to E5) response as not opposing effect from warming the mid-latitudes exists for SI. Perturbation with subsequently larger latitudinal extent were shown to lead to the eddy momentum flux anomalies about the tropopause level going from a clear dipole pattern to a general weakening of the horizontal eddy momentum flux. Consistent with this the tropospheric response was shown to change with increasing latitudinal extend of the heating perturbations from a strong poleward jet shift to an equatorial jet shift.

The strongest response is found for the SII/SIII experiments which exhibit maximal tropical heating at the same time as limited heating equatorward of the jet. Such perturbations seem to be an ‘optimal forcing’ for shifting the jet poleward rather than the original heating distribution E5 (Haigh et al., 2005, Simpson, 2009 and Simpson et al., 2009) which extends into mid-latitudes with a broad gradient.

The tropospheric response to the various heating perturbations was further shown to be not significantly different between experiments run on the original L15 and the extended L26 versions of the IGCM2.2. It was therefore concluded that a
resolved stratosphere in the model is not necessary to achieve the tropospheric response and hence that for studies where the focus is exclusively on the tropospheric response to perturbations (without further vertical structure) applied to the lower stratosphere/tropopause region the use of a low-top model such as the L15 IGCM2.2 is sufficient.

Further comparisons with zonally symmetric simulations were used to show that eddy feedbacks modify both the lower stratosphere/tropopause region as well as the middle/upper stratosphere - thereby significantly altering the structure of the temperature anomaly in the forcing region. The behaviour of the eddy feedbacks was shown to be influenced in a systematic way by the latitudinal extent of the applied heating perturbation.

The limitations described in sections 6.1.1 and 6.1.2. regarding the use of a simplified model (including the vertically extended IGCM2.2) and the length of the model runs also apply to this research.

The investigation performed for this chapter were shown to suggest that the mechanisms responsible for creating the tropospheric response are the same for all heating perturbations used for this study. Further the results highlight the importance of eddy feedback processes for modifying the forcing region itself and thereby altering the tropospheric response. These findings thereby help to interpret the results of future stratosphere-troposphere interaction studies and will further help to make informed choices on which type of model to use as well as to design stratospheric perturbation experiments for future studies.

This research further contributed to a better understanding of the influence of the position and extent of stratospheric forcings on the tropospheric response and thus might help to improve understanding of future real world jet shifts in response to stratospheric changes of various shapes that might be expected, e.g. as a result of anthropogenic climate change.
6.2 Future Work

The development of the vertically extended IGCM2.2 offers a range of possibilities for investigation that were not previously possible with the original IGCM2.2. The heating perturbations applied to the stratosphere in Chapter 5 of this thesis could easily be modified to more closely resemble solar variability. The impact of this on both the tropospheric response and on how the forcing region itself is modulated by eddy feedbacks would be very interesting. It is, for example, important to know whether the tropical lower stratospheric heating associated with solar variability leads to a significant tropospheric response in idealised models such as the L26 IGCM2.2, to confirm the relevance of the mechanism identified by Haigh et al. (2005), Simpson (2009) and Simpson et al. (2009) for the solar case, to distinguish it from alternative dynamical mechanism from the tropical upper stratosphere.

Further, it would be of interest to investigate the impact of stratospheric forcings that represent not only solar variability, but also ozone depletion, volcanic aerosols or possible geoengineering proposals. Such forcings have been previously investigated in more complex general circulation models. The focus of such studies usually is to apply a forcing that is as realistic as possible with model that are preferably close to the real world in order to investigate present and future climate. This, however, makes it difficult to identify and understand the basic underlying dynamics by which such stratospheric forcings impact on the troposphere. Hence performing such investigations with an idealised model, such as the L26 IGCM2.2, could yield important insights which would help to improve the interpretation of more complex modeling studies.

As most simplified general circulation model studies of stratosphere - troposphere interactions use temperature perturbations as stratospheric forcing, it would be of great interest to investigate the impact of momentum forcings. This could be done by applying QBO-like momentum perturbations to the equatorial stratosphere and subsequently analysing the possible tropospheric response. With the L26 IGCM2.2 version the full vertical range of the real world QBO could
be applied to the model stratosphere. The research conducted for this thesis highlighted the importance of the mid-latitude part of the heating perturbations for generating a tropospheric response. As the QBO is confined to the equatorial latitudes, it would be very interesting to see if a tropospheric response can be found and if so which mechanisms are responsible for transferring the signal downwards into, and throughout, the troposphere. While the mechanisms for downward propagation of the QBO within the stratosphere are well-known, new information about how the troposphere couples to the stratosphere might be gained. Current research (Garfinkel and Hartmann, 2011a and 2011b) suggests that QBO momentum anomalies induce a meridional circulation (to maintain thermal wind balance) which includes zonal wind anomalies extending from the equatorial stratosphere into the subtropical troposphere where a response throughout the troposphere is induced. Studying the occurrence or absence of such a tropospheric response to stratospheric equatorial momentum forcings in the L26 IGCM2.2. will help to complete the understanding the role of eddy-mean flow interactions in stratosphere - troposphere coupling.

Important future research would be to investigate how the results from this study with a simplified model relate to mechanisms (e.g. planetary wave refraction) derived from more complete studies in order to further improve the understanding of stratosphere - troposphere interactions.
Appendix A

In this section the figures for the experiments with retracted heating perturbations performed with the L26B1 set up ($k_s=1/2$ [1/day]) are presented. Shown are the differences between perturbed and unperturbed experiments in zonal mean temperature and zonal mean zonal wind for the B1E5, B1Aup, B1Bup and B1Cup experiments (Figure A.1). The contour intervals are 0.5 K and 0.5 m/s. Shaded areas are not statistically significant at the 95% level.

Further, the differences between perturbed and unperturbed experiments in horizontal eddy momentum fluxes and static stability for the B1E5, B1Aup, B1Bup and B1Cup experiments are shown in Figure A.2. The contour intervals are $1.0 \, m^2/s^2$ and $10^{-6} \, 1/s^2$. The vertical red line indicates the jet position for the unperturbed control run (B1L26).
Figure A.1: As Fig. 5.4, but for the equivalent run using the B1 ($k_s=1/2$) configuration.
Figure A.2: Differences between perturbed and unperturbed experiments in horizontal eddy momentum fluxes (left panels) and static stability (right panels) for the B1E5, B1Aup, B1Bup and B1Cup experiments. The contour intervals are $1.0 \text{ m}^2/\text{s}^2$ and $1^{-61}/\text{s}^2$, respectively. The vertical red line indicates the jet position for the unperturbed control run (B1L26).
Appendix B

In this section the figures for the experiments with heating perturbations with varying latitudinal extent performed with the L15 version of the IGCM2.2 are presented.

Shown are:

- Fig. B.1, zonal mean temperature and zonal mean zonal wind anomalies.
- Fig. B.2, horizontal eddy momentum flux and static stability anomalies.
- Fig. B.3, zonally symmetric temperatures and zonal winds anomalies.
- Fig. B.4, comparison of the temperature (LHS) and the zonal wind (RHS) anomalies between the zonally symmetric runs for experiments E5, S1, SIII, SIV and U5 and the full 3D runs.
- Fig. B.5, zonal mean temperature and zonal mean zonal wind anomalies with log-pressure scale.
- Fig. B.6, EP-flux divergence and TEM residual circulation anomalies.
- Fig. B.7, TEM meridional and TEM vertical motion anomalies.
Figure B.1: As Figure 5.12, but for experiments run with L15.
Figure B.2: As Figure 5.13, but for experiments run with L15.
Figure B.3: As Figure 5.19, but for experiments run with L15.
Figure B.4: As Figure 5.20, but for experiments run with L15.
Figure B.5: As Figure 5.21, but for experiments run with L15.
Figure B.6: As Figure 5.22, but for experiments run with L15.
Figure B.7: As Figure 5.23, but for experiments run with L15.
Appendix C

In this section the figures for the experiments with heating perturbations with varying latitudinal extent performed with the L15 version of the IGCM2.2 are presented.

Shown are:

• Fig. C.1, Zonal mean temperature and zonal mean zonal wind control fields for L26.

• Fig. C.2, Zonal mean temperature and zonal mean zonal wind control fields for L15.

• Fig. C.3, EP-flux divergence and TEM residual circulation control fields for L26.

• Fig. C.4, EP-flux divergence and TEM residual circulation control fields for L15.

• Fig. C.5, TEM meridional and TEM vertical motion control fields for L26.

• Fig. C.6, TEM meridional and TEM vertical motion control fields for L15.
Figure C.1: L26, Zonal mean temperature (left panels) and zonal mean zonal wind (right panels) for the E5, SI, SHII, SV and U5 experiments with log-pressure scale. The contour intervals are 5 K and 4 m/s, respectively.
Figure C.2: L26, EP-flux divergence (left panels) and TEM residual circulation (right panels) for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are half-logarithmic at 1,2 and $5 \times 10^n$ m$^3$ and kg/s, respectively.
Figure C.3: L26, TEM meridional (left panels) and TEM vertical (right panels) motion for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are half-logarithmic at $1.2 \times 10^n$ m/s and Pa/s, respectively.
Figure C.4: L15, Zonal mean temperature (left panels) and zonal mean zonal wind (right panels) for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are 5 K and 4 m/s, respectively.
Figure C.5: L15, EP-flux divergence (left panels) and TEM residual circulation (right panels) for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are half-logarithmic at 1,2 and $5 \times 10^n$ m$^3$ and kg/s, respectively.
Figure C.6: L15, TEM meridional (left panels) and TEM vertical (right panels) motion for the E5, SI, SIII, SV and U5 experiments with log-pressure scale. The contour intervals are half-logarithmic at 1.2 and $5 \times 10^n$ m/s and Pa/s, respectively.
Appendix D

In this section an overview of jet position, decorrelation times and magnitude of jet shift in response to the various heating perturbations applied in Chapter 5 is presented. Table 7 shows the results for all experiments and control run performed with the L26 version of the IGCM2.2. The results for the L15 simulations are given in Table 8.

The errors indicated are one standard error for the jet positions and the jet shifts. The errors for the decorrelation times $\tau$ were calculated following the method described in Chapter 3. The jet shifts are always given relative the the respective control run (i.e. L26complete, B1L26 or B3L26).
<table>
<thead>
<tr>
<th>experiment</th>
<th>jet position [deg]</th>
<th>$\tau$ [days]</th>
<th>jet shift [deg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>L26complete</td>
<td>43.64 ± 0.23</td>
<td>72 ± 9</td>
<td>-</td>
</tr>
<tr>
<td>L26E5</td>
<td>45.46 ± 0.23</td>
<td>49 ± 5</td>
<td>1.82 ± 0.33</td>
</tr>
<tr>
<td>L26Aup</td>
<td>44.45 ± 0.26</td>
<td>53 ± 5</td>
<td>0.81 ± 0.35</td>
</tr>
<tr>
<td>L26Bup</td>
<td>44.76 ± 0.20</td>
<td>60 ± 7</td>
<td>1.12 ± 0.30</td>
</tr>
<tr>
<td>L26Cup</td>
<td>44.04 ± 0.42</td>
<td>93 ± 13</td>
<td>0.40 ± 0.48</td>
</tr>
<tr>
<td>L26E5dp</td>
<td>43.80 ± 0.19</td>
<td>98 ± 14</td>
<td>0.16 ± 0.30</td>
</tr>
<tr>
<td>L26E5dz</td>
<td>43.62 ± 0.47</td>
<td>84 ± 11</td>
<td>-0.02 ± 0.52</td>
</tr>
<tr>
<td>B1L26</td>
<td>38.18 ± 0.39</td>
<td>93 ± 13</td>
<td>-</td>
</tr>
<tr>
<td>B1L26E5</td>
<td>43.17 ± 0.47</td>
<td>130 ± 21</td>
<td>-4.99 ± 0.61</td>
</tr>
<tr>
<td>B1L26U5</td>
<td>36.37 ± 0.58</td>
<td>101 ± 14</td>
<td>-1.81 ± 0.70</td>
</tr>
<tr>
<td>B1L26Aup</td>
<td>41.17 ± 0.76</td>
<td>151 ± 26</td>
<td>2.99 ± 0.85</td>
</tr>
<tr>
<td>B1L26Bup</td>
<td>40.45 ± 0.43</td>
<td>115 ± 17</td>
<td>2.27 ± 0.58</td>
</tr>
<tr>
<td>B1L26Cup</td>
<td>38.32 ± 0.50</td>
<td>151 ± 26</td>
<td>0.14 ± 0.63</td>
</tr>
<tr>
<td>B3L26</td>
<td>46.82 ± 0.45</td>
<td>39 ± 3</td>
<td>-</td>
</tr>
<tr>
<td>B3L26E5</td>
<td>48.85 ± 1.14</td>
<td>31 ± 2</td>
<td>2.03 ± 1.23</td>
</tr>
<tr>
<td>B3L26U5</td>
<td>45.13 ± 0.09</td>
<td>48 ± 5</td>
<td>-1.69 ± 0.46</td>
</tr>
<tr>
<td>L26Cup SII</td>
<td>45.79 ± 0.10</td>
<td>66 ± 8</td>
<td>2.15 ± 0.25</td>
</tr>
<tr>
<td>L26E5dz SII</td>
<td>45.03 ± 0.21</td>
<td>63 ± 7</td>
<td>1.39 ± 0.31</td>
</tr>
<tr>
<td>L26SI</td>
<td>45.11 ± 0.18</td>
<td>62 ± 7</td>
<td>1.47 ± 0.29</td>
</tr>
<tr>
<td>L26SII</td>
<td>47.14 ± 0.66</td>
<td>39 ± 4</td>
<td>3.50 ± 0.70</td>
</tr>
<tr>
<td>L26SIII</td>
<td>47.49 ± 0.08</td>
<td>33 ± 3</td>
<td>3.85 ± 0.24</td>
</tr>
<tr>
<td>L26SIV</td>
<td>45.51 ± 0.23</td>
<td>50 ± 5</td>
<td>1.87 ± 0.33</td>
</tr>
<tr>
<td>L26SV</td>
<td>42.58 ± 0.21</td>
<td>97 ± 14</td>
<td>-1.06 ± 0.31</td>
</tr>
<tr>
<td>L26U5</td>
<td>42.18 ± 0.28</td>
<td>74 ± 9</td>
<td>-1.46 ± 0.36</td>
</tr>
</tbody>
</table>

Table 7: Results for the experiments run with the L26 IGCM2.2 for Chapter 5. The errors indicated are one standard error for the jet positions and the jet shifts. The errors for the decorrelation times $\tau$ were calculated following the method described in Chapter 3. The jet shifts are always given relative the the respective control run (i.e. L26complete, B1L26 or B3L26).
<table>
<thead>
<tr>
<th>experiment</th>
<th>jet position [deg]</th>
<th>$\tau$ [days]</th>
<th>jet shift [deg]</th>
</tr>
</thead>
<tbody>
<tr>
<td>L15control</td>
<td>45.41 ± 0.40</td>
<td>63 ± 7</td>
<td>-</td>
</tr>
<tr>
<td>L15E5</td>
<td>47.04 ± 0.16</td>
<td>44 ± 4</td>
<td>1.63 ± 0.43</td>
</tr>
<tr>
<td>L15SI</td>
<td>46.59 ± 0.26</td>
<td>55 ± 6</td>
<td>1.18 ± 0.48</td>
</tr>
<tr>
<td>L15SII</td>
<td>48.22 ± 0.07</td>
<td>30 ± 2</td>
<td>2.81 ± 0.41</td>
</tr>
<tr>
<td>L15SIII</td>
<td>48.32 ± 0.18</td>
<td>37 ± 3</td>
<td>2.91 ± 0.44</td>
</tr>
<tr>
<td>L15SIV</td>
<td>46.86 ± 0.26</td>
<td>48 ± 5</td>
<td>1.45 ± 0.48</td>
</tr>
<tr>
<td>L15SV</td>
<td>43.75 ± 0.46</td>
<td>90 ± 12</td>
<td>-1.66 ± 0.61</td>
</tr>
<tr>
<td>L15U5</td>
<td>43.34 ± 0.41</td>
<td>74 ± 9</td>
<td>-2.07 ± 0.57</td>
</tr>
<tr>
<td>B1L15</td>
<td>40.63 ± 0.50</td>
<td>131 ± 21</td>
<td>-</td>
</tr>
<tr>
<td>B1L15E5</td>
<td>45.28 ± 0.42</td>
<td>96 ± 13</td>
<td>4.65 ± 0.65</td>
</tr>
<tr>
<td>B1L15U5</td>
<td>37.6 ± 0.49</td>
<td>139 ± 23</td>
<td>-3.03 ± 0.70</td>
</tr>
</tbody>
</table>

Table 8: Results for the experiments run with the L15 IGCM2.2 for Chapter 5. The errors indicated are one standard error for the jet positions and the jet shifts. The errors for the decorrelation times $\tau$ were calculated following the method described in Chapter 3. The jet shifts are always given relative the the respective control run (i.e. L15control or B1L15).
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