The Effects of Greenhouse-gas and Surface Thermal Forcing on the Stratosphere

Barbara Winter

Doctor of Philosophy

Department of Atmospheric and Oceanic Sciences

McGill University
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DEDICATION

Gewidmet den drei Winters noch ohne eigenen Doktortitel:

Layth, Felix, und Anouk.

Die Jahre der Diss sind die Jahre mit euch.
Acknowledgements

My heartfelt thanks go most of all to Michel Bourqui, who supervised this thesis. When I first asked Michel whether I could work with him, he said yes at once, which I always felt was a little reckless. It set the tone for all interactions within our group, where his optimism has become legendary. While it might be argued that it is a supervisor’s job to provide guidance and crack the whip as needed and see the forest when one bumps into trees, Michel’s infectious enthusiasm for all projects, his openness to suggestions and his unlimited availability for discussions and questions on any topic at any time go well beyond the call of duty. It has been a welcome and novel experience for me to have such access to a supervisor, and, as a bonus, to work for someone who knows GCM source code inside out. I have been spoilt.

The road to a PhD began with the warm welcome I received in the atmospheric science group at Dalhousie University, which led me to complete an MSc. It was then my great privilege to work as research staff with George Philander at Princeton University, an unparalleled work environment as stimulating as it was friendly. Part of the job involved teaching undergraduates; this, and the desire to better understand the science behind my climate model-running work, led me to think about going back to school myself. It was very difficult to walk away from the job I had; I did regret it at times, but can truthfully say that the work I got to do with Michel pretty exactly matched what I came back to school for.

I want to thank Jacques Derome and Charles Lin, whose speedy and encouraging replies to my initial emails convinced me to apply to McGill, and who allowed me to come here. I am grateful to the administrative staff of the department for shepherding me through all the paperwork, with special thanks to Karin Braidwood, Vaughn Thomassin, Paula Domingues and Ornella Cavaliere. Furthermore, there is no thesis...
without computers and there are no computers without the incomparable Michael Havas, sysadmin extraordinaire.

On a personal level, the first year here was sad and difficult, and my friends of the ages, Chione and Rachel, were there when it mattered, in spite of intervening continents or oceans. In happier times and closer at hand, thank you to Alex and Lily, Lisa and Irena, for generous friendship and support, encouragement, advice, gossip and beer (or tea), and for laughing at the sheer absurdity of it all. For absorbing me into the grad student mass and cheerfully ignoring the generation gap, thank you to my classmates from first year, to the CAOS executive back in our militant days, to my brothers-in-arms in Michel’s group; pour l’amitié et la solidarité, câlins à Bruno, Louis-Philippe et Jean-François. Big thank-you to my fellow Cathedral Singers: your friendship and our music was the single biggest factor keeping my head above water in these years. Giving it up to have more time in the final months has been the most difficult aspect of thesis-writing.

Finally, the years of my PhD are inextricably both the years with my nephews and niece and the years without my mother. It puts a different spin on the role of family in the life of the grad student. And so in memory and in person, in long phone calls and emails, in home-cooked meals and in surprise visits to my office, in birthday cakes covered with blue smarties and in artwork for my walls, in reminiscences about their own PhD experiences and in advice for mine, I have been surrounded and supported by my family and I am deeply grateful.
Contributions of Authors

The research for the manuscripts of which this thesis is composed was carried out by myself, under the supervision of Prof. Michel Bourqui, who is the sole co-author on these papers. All texts are the result of numerous discussions about both the structure and the content of the papers, and were entirely written by me and edited by Prof. Bourqui. The source code for the Coupled Chemistry Climate Model used in this research was written at the University of Reading, and analysis codes were written mostly by myself, though with additional routines written by Prof. Bourqui. Apart from such measurables as comments on the drafts or written code, Prof. Bourqui contributed to the research presented here through many discussions and much valuable feedback.

The topic for the first manuscript (“Wave-Forcing in the Stratosphere under Doubled-CO$_2$ Conditions in a 100-year Coupled-Chemistry GCM Study”, Chapter 2) was suggested by Prof. Bourqui. The version included here is the final, published version and reflects the suggestions and comments made on the original manuscript by three anonymous reviewers.

The idea for the second manuscript (“Sensitivity of the Stratospheric Circulation to the Latitude of Thermal Surface Forcing”, Chapter 3), was mine, presented and approved as my thesis proposal (ATOC 700) in November 2006. The manuscript included here has been submitted to the Journal of Climate and is presently being revised following peer review.

The idea for the third manuscript (“The Impact of Surface Temperature Variability on the Climate Change Response in the Northern Hemisphere Polar Vortex”, Chapter 4) arose from discussions between Prof. Bourqui and myself, and follows from the results found in Chapter 2.
The temperature in the high-latitude stratosphere is governed by both radiative cooling and dynamical heating, and is highly variable. An increase in well-mixed greenhouse gases will lead to an increase in radiative cooling in the stratosphere, but the response of stratospheric dynamics to increased greenhouse gas loading is still not fully known. Feedbacks between ozone chemistry, radiation and dynamics also affect the temperature in the stratosphere, and the competing effects between greenhouse gas increases and ozone recovery further complicate the prediction of future stratospheric temperatures.

The interest in the future of stratospheric temperature is dominated by two issues. First, the seasonal destruction of ozone occurs inside the stratospheric polar vortex at the end of the winter. In the Northern Hemisphere, the conditions for an ozone hole are met only intermittently, partly because temperatures inside the vortex are not continuously low enough for the formation of Polar Stratospheric Clouds (on which the heterogeneous chemistry responsible for ozone destruction depends), and partly because the vortex is weaker than its Southern Hemisphere counterpart and allows the mixing-in of more ozone-rich air from lower latitudes. Greenhouse gas-induced changes to the polar stratospheric temperature and dynamics therefore impact future ozone loading, independently of changes in the concentrations of ozone-depleting substances. Second, it is now known that the coupling between stratosphere and troposphere is such that, under appropriate conditions, events in the stratosphere (such as Sudden Stratospheric Warmings) can be used as predictors for weather regimes in the troposphere on seasonal timescales. Changes in stratospheric dynamics therefore have the potential to feed back on changes in tropospheric dynamics.

Focus on the stratosphere has consequently been intense in the last decades, in a wide
range of studies from satellite observation programmes to general circulation model simulations. In this thesis, a number of experiments carried out with a Chemistry Climate Model (the IGCM-FASTOC, a general circulation model of the atmosphere coupled to a chemistry model) are used to study the stratosphere subject to different forcings, either by varying surface conditions or by increasing atmospheric CO$_2$ concentrations. Results are presented in three manuscripts, whose new contributions can be summarized as follows:

**Wave-Forcing in the Stratosphere under Doubled-CO$_2$ Conditions in a 100-year Coupled-Chemistry GCM Study** (Chapter 2):

- In order to reliably assess the statistical significance of the climate response, particularly in an area of highly variable conditions, large ensembles are needed. While the different modelling studies in the last decade agree that the Brewer-Dobson Circulation will increase in a changed climate, there is less agreement on the magnitude or even the sign of the accompanying temperature response. Few of the available studies find a statistically significant response in the high-latitude stratosphere, and only one other study (thus far) has used a sufficiently long integration for the application of statistical tools to be meaningful. The first contribution by this thesis to the study of the climate change response in the stratosphere is the extension to 100 years (timeslice) of the length of the simulations, reducing the average variability in high latitudes and making the statistical evaluation of results reliable.

- In the majority of climate-change experiments, the stratosphere-resolving general circulation model is forced with prescribed interannually varying monthly-mean sea surface temperatures (see Section 1.4). A comparison was made by
Rind et al. (1998) of responses in the stratosphere when sea surface temperatures are calculated interactively (by a mixed-layer slab ocean coupled to the atmospheric model) and when they are prescribed; it is the only such comparison, and the authors warn of some inconsistencies in the way the prescribed surface temperatures were obtained. The biggest new contribution of the first chapter in this thesis is to present results from a CCM coupled to a mixed-layer ocean, such that the surface temperatures are in thermal equilibrium with the atmosphere at all gridpoints and timesteps, and to compare these to results when the surface temperatures are prescribed.

**Sensitivity of the Stratospheric Circulation to the Latitude of Thermal Surface Forcing** (Chapter 3):

- It is known that a large part of the climate change response in the high-latitude Northern Hemisphere winter stratosphere is due to the warming of the surface caused by increased greenhouse gas heating of the troposphere. It has also been shown that tropospheric dynamics (for instance, the location of the tropospheric jet) are sensitive to the latitude of imposed thermal surface forcing. The first contribution of the work presented in Chapter 3 is to combine these questions and show that surface thermal forcing imposed in any isolated band affects the generation of planetary waves from the surface to the tropopause. In addition, it is shown that forcing imposed in some bands affects the baroclinicity of the troposphere through changes in the vertical shear of zonal winds particularly in the jet, and that this response can contribute to the anomalous generation of planetary waves, which propagate to the polar vortex in the stratosphere.
• Studies of systematic band-wise forcing are generally undertaken with highly idealized general circulation models (for example, dry aquaplanets). Although this allows precise targeting of specific processes for study, it can lead to responses that are not valid in a more realistic climate. The surface temperature response to greenhouse gas increases in realistic models is a global warming of different magnitudes in different latitudes. By using a realistic model for studies of idealized, localized surface forcing, it is established that surface heating in the equatorial and tropical latitudes determines the sign of the response in the high-latitude stratosphere, even in the presence of surface warming at higher latitudes. This is the second major contribution of the work presented in Chapter 3.

The Impact of Surface Temperature Variability on the Climate Change Response in the Northern Hemisphere Polar Vortex (Chapter 4):

• As stated above, there has been one previous comparison of the climate change response in the stratosphere obtained in a model with interactively calculated, rather than prescribed, sea surface temperatures, but the prescribed temperatures in that study were output from a slightly different model (Rind et al. (1998)). There have also been comparisons between the climate change response when the sea surface temperatures are prescribed and climatological (i.e. fixed), versus prescribed and interannually varying (Braesicke and Pyle (2004); Olsen et al. (2007)). Chapter 4 of this thesis provides the first comparison of the climate change response in a complete set of experiments in which the sea and/or land surface temperatures are either interactively calculated, prescribed and interannually varying, or prescribed and climatological. The
sensitivity of the climate change response to these surface representations has not been discussed before.
Abstract

This thesis is concerned with the possible future changes in the stratosphere as a result of climate change. In this context, the climate change forcing can be separated into two components: (i) radiative forcing due to a doubling of the atmospheric CO$_2$ mixing ratio, and (ii) surface thermal forcing. My focus is on the response in the circulation of the Northern Hemisphere stratosphere in winter. All experiments are carried out with a chemistry-climate model (CCM, i.e. an atmospheric general circulation model coupled to a chemistry model), the IGCM-FASTOC, and all results shown are averages over 100-year or 50-year simulations in timeslice mode (i.e. every year can be considered as one member of an ensemble having 50 or 100 members). This allows statistically robust results in a region of high variability in the temperature and wind fields.

When the IGCM-FASTOC is coupled to a mixed-layer slab ocean, the Arctic lower stratosphere in winter warms by up to 4K under 2$\times$CO$_2$ conditions, with associated weakening of the polar vortex and enhancement of the Brewer-Dobson circulation (BDC). This change is related to a significant increase in the Rossby wave forcing near the vortex core starting in January, followed by an increased wave forcing at the lower edge of the polar vortex in February. Maximum wave forcing is found both to begin earlier in the winter and to be distributed over a longer period of time in the 2$\times$CO$_2$ climate. Results from four additional pairs of simulations (control and 2$\times$CO$_2$), in which sea and/or land surface temperatures were either calculated interactively, prescribed as a climatological cycle, or prescribed as interannually varying monthly-mean fields, demonstrate that the interannual variability in sea and land temperatures, and the adjustment of oceans and lands to the atmosphere
and to one another, are essential in order to maintain realistic stratospheric forcing by Rossby waves and to adequately capture the stratospheric response to global warming. Specifically, when the land is interactive but the slab ocean is replaced by prescribed interannually varying monthly-mean temperatures, the stratospheric response is qualitatively similar to the fully interactive case, but has lower amplitude and is statistically significant over a smaller area. In the experiment without interannual variability, there is no response in stratospheric dynamics.

To assess the impact of surface temperatures on the stratospheric circulation, a separate suite of experiments was carried out in which a 2K temperature anomaly was added to the control surface temperature at all gridpoints within latitudinal windows of 10 or 30 degrees. Thermal surface forcing applied anywhere equatorwards of 20°N, or continuously from the equator to 30°N, increases the generation of planetary waves in the troposphere, resulting in increased upward propagation. Consequently, a greater flux of wave activity enters the mid- to high latitude stratosphere and breaks in the polar vortex, increasing the BDC and leading to a warm anomaly in the polar stratosphere. Ozone concentration increases at high latitudes and decreases at low latitudes. Thermal surface forcing imposed between 30°N and 60°N has the reverse effect and leads to a stronger and colder vortex. Thermal forcing applied polewards of 60°N has little effect on tropospheric baroclinicity, but results in a sufficient decrease of the vertical flux of planetary wave activity that the vortex becomes anomalously strong and cold. In all cases when surface forcing is imposed only polewards of 30°N, the ozone concentration decreases at high latitudes but is not affected at low latitudes. Combining the forcing in an equatorial and a mid-latitude band leads to a response similar to that of the equatorial forcing, demonstrating that the subtropical surface temperature changes dominate the sign of the surface-driven response in the vortex.
Cette thèse étudie les possibles futurs changements dans la stratosphère dus au forçage par les changements climatiques. Dans ce contexte, les changements climatiques peuvent être séparés en deux composantes: (i) le forçage radiatif dû à un doublage de la concentration de CO₂ dans l’atmosphère et (ii) le forçage thermique à la surface. L’emphase est mise sur la réponse dans la circulation de la stratosphère de l’Hémisphère nord en hiver. Toutes les expériences sont réalisées au moyen d’un modèle climat-chimie (un modèle de la circulation générale de l’atmosphère couplé à un modèle de chimie), le IGCM-FASTOC, et les résultats présentés sont les moyennes sur 100 ou 50 ans de simulations en mode répété (ou stationnaire, i.e. chaque année peut être considérée comme un membre d’un ensemble ayant 100 ou 50 membres). Ceci permet d’aboutir à des résultats qui sont statistiquement significatifs dans une région où la température et les vents sont hautement variables.

Lorsque le IGCM-FASTOC est couplé à un océan homogène de 25m de profondeur et que les températures de surface sont calculées de façon interactive, la réponse dans la basse stratosphère de l’Arctique à un doublage de CO₂ est un réchauffement de 4K, accompagné d’un affaiblissement du vortex polaire et d’une augmentation de la circulation Brewer-Dobson (CBD). Ces changements sont reliés à une importante augmentation du flux vertical d’ondes Rossby, qui déclèrèrent le centre du vortex dès le mois de janvier, puis son bord inférieur en février. Le forçage maximal du vortex par les ondes débute plus tôt dans la saison hivernale et dure pour une plus grande période de temps dans le climat 2×CO₂. Quatre paires de simulations supplémentaires (contrôle et 2×CO₂) ont été réalisées, dans lesquelles les températures à la surface de l’océan et/ou de la terre étaient soit calculées de manière
interactive, soit prescrites comme cycle climatologique fixe ou ayant une variabilité interannuelle. Ces expériences démontrent que la variabilité interannuelle, tout comme les ajustements des températures de surface à l’atmosphère ou des températures de la terre à celle des océans, sont essentielles pour maintenir un forçage réaliste du vortex par les ondes Rossby et donc pour capturer la réponse aux changements climatiques de façon adéquate. Quand la surface terrestre est interactive mais l’océan couplé est remplacé par des températures imposées qui varient à l’échelle interannuelle, la réponse dans la stratosphère est qualitativement semblable à celle de l’expérience ayant l’océan interactif; par contre, cette réponse a une amplitude inférieure en plus d’être statistiquement significative sur une région moins importante. Les expériences sans variabilité interannuelle ne montrent aucun changement de la circulation dans la stratosphère.

Pour évaluer l’importance du rôle de la température de surface dans la circulation stratosphérique, j’ai entrepris une nouvelle série d’expériences dans lesquelles une anomalie thermique de 2K est imposée à tous les points de grille à l’intérieur de bandes zonales larges de 10° ou de 30° de latitude. Lorsque le forçage thermique est imposé à la surface entre l’équateur et 20°N, ou de façon continue entre l’équateur et 30°N, un plus grand flux d’activité ondulatoire accède à la stratosphère dans les moyennes et hautes latitudes. Les ondes se cassent dans le vortex, menant à une augmentation de la CBD et à une anomalie positive dans température de la stratosphère polaire. La concentration d’ozone augmente dans les hautes latitudes mais diminue dans les basses. Un forçage thermique imposé à la surface entre 30°N et 60°N donne le résultat inverse, soit un vortex plus fort et plus froid. Le forçage thermique imposé à la surface au nord de 60°N a peu d’effet sur la troposphère, mais affaiblit suffisamment le flux vertical d’activité ondulatoire pour que le vortex en résulte plus froid et plus fort également. La concentration d’ozone diminue dans
les hautes latitudes sans être affectée dans les basses. Lorsque le forçage thermique est appliqué à la fois dans une bande équatoriale et dans une bande des moyennes latitudes, la réponse dans la stratosphère ressemble à celle au forçage dans la bande équatoriale uniquement. Ceci démontre que les changements de température à la surface dans les subtropiques dominent le signe de la réponse dans le vortex polaire.
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</table>
1.1 Discovery of the Stratosphere

The discovery of the stratosphere has much to do with rubber. By the end of the 19th century, sending aloft hot-air balloons with scientific instrumentation on board had become regular practice, routinely reaching close to 10 km in height, and providing temperature profiles of the atmosphere above their launching sites in western Europe or the United States. The balloons were made of silk or paper, and thus had limited elasticity; once they reached maximum size, further ascent was impeded and they drifted with the wind. This compromised the reliability of the instruments, which required ventilation in order to function in the daytime, and so, to calibrate and operate the instruments, humans were sent up with them (not all survived). Larger balloons made higher ascents possible, and night launches eliminated the problem of overheating by radiation, but the balloons nonetheless eventually drifted, and without proper ventilation the measured temperatures risked being polluted by temperature from the balloon itself. Léon Philippe Teisserenc de Bort, working at an observatory near Versailles, launched unmanned paper balloons at night and noticed a temperature inversion at 11 km, but was wary of his findings because of the ventilation problem (Labitzke and van Loon (1999)).

The new century saw new technology applied to an old substance: in 1901, Richard Aßmann in Berlin launched the first instrumented rubber balloon, alongside a manned regular silk version (Labitzke and van Loon (1999)). Both carried instruments and rose to 10.5 km, by which point the crew in the silk balloon had lost
consciousness. The unmanned instruments in the rubber balloon performed well, as did the balloon itself, and new altitudes became open to exploration: the rubber balloons launched from Berlin rose steadily to reach heights up to 25 km before bursting. The expansion of the rubber balloon and the height at which it burst could be calculated in advance, and the balloon’s ascent controlled, guaranteeing continuous ventilation for the instruments and reliable temperature data. Aßmann’s rubber balloons recorded not only a temperature inversion between 10 and 15 km, but indeed a slight temperature increase in this layer (Labitzke and van Loon (1999)).

The discovery of the stratosphere is attributed jointly to Teisserenc de Bort and to Aßmann, who both published their findings on the 11 km inversion in the spring of 1902.

These early pioneers did not know the full vertical extent of their inversion layer, nor the reason for its existence. After joining Oxford University in 1920, Gordon Dobson, in studying meteor trails with a colleague, found a warm layer in the upper atmosphere at about 50 km, and attributed it to heating by ozone (Walshaw (1989); see also Fig. 1–1). Ozone became Dobson’s main focus of research, but during the Second World War he was tasked with elucidating the conditions for airplane condensation trails. Alan Brewer was hired to make the necessary aircraft-based measurements of humidity in the stratosphere; after the war, Dobson and Brewer continued their collaboration and returned the focus to ozone (among other species; see Brewer (1949)). The distributions of the chemical species they measured, ozone generally but also helium in high latitudes, combined with the striking dryness Brewer had found in the stratosphere, led them to posit the existence of what is now known as the Brewer-Dobson Circulation: a slow ascent of air into the stratosphere in a narrow range of low latitudes, balanced by descent spread more broadly over mid- to high latitudes (Brewer (1949); Dobson (1956)). If either Dobson or Brewer
pondered the dynamical origins of the meridional circulation they had discovered in the stratosphere, it is not recorded.

Meanwhile, balloons continued to rise at Berlin, outfitted now with radiosondes and launched on a daily basis beginning in 1951. By then, synthetic rubber had replaced latex in their fabrication, and allowed ascents to 30, even 40 km (Labitzke and van Loon (1999)). The climatological view of the stratosphere obtained from the early occasional balloon launches or infrequent aircraft field campaigns – that of a stably stratified layer with a very slow meridional overturning circulation – was soon augmented by surprising data in the new daily record: in the winter of 1952, the temperature above Berlin, at the height of nearly 30 km, rose by about 30°C in an interval of two days, then cooled again almost as rapidly (Labitzke and van Loon (1999)). Proudly baptized the “Berlin Phenomenon”, this first recorded instance of a midwinter Sudden Stratospheric Warming event was followed by a second less than a month later, and a meticulous record of these warmings has been kept at Berlin ever since. The year 1952 marked the golden anniversary of the stratosphere’s discovery, and observations had mapped its vertical extent, its climatological dynamical structure (at least in the mid-latitude solstice seasons) and distribution of key radiating chemical species, and its dramatic high-frequency temperature changes in high-latitude winter.

1.2 Mathematical Models of Stratospheric Dynamics

“[I]t can be shown that the transmissivity of the upper atmosphere to planetary waves is exceedingly sensitive to the mean zonal wind structure. Indeed, [...] it is primarily the variation of zonal mean wind with height that gives rise to the energy trapping [in the lower atmosphere].”
Figure 1–1:

This quote from the seminal 1961 paper by Jule Charney and Philip Drazin (*Charney and Drazin* (1961)) underscores the fact that the prime motivation for theoretical studies of vertical propagation of planetary waves (undertaken by Charney
as of 1949) was not the search for a mechanism to explain the observed Brewer-Dobson circulation (BDC). Instead, it was to answer the question of why the kinetic energy of the troposphere largely remained trapped in the troposphere.

Charney and Drazin considered quasi-geostrophic motion on a $\beta$-plane, expressed in terms of mean and perturbation components. The mean motion was assumed to vary only in the vertical, allowing the perturbations $\psi'$ to be written as products of a function of height and linear wave components in the horizontal:

$$\psi'(x, y, z, t) = \Psi(z) e^{i(k x + l y - k c t)}$$  \hspace{1cm} (1.1)

where $k$ and $l$ are the zonal and meridional wavenumbers. The equations of motion can then be rewritten as a wave equation in $\Psi(z)$:

$$\frac{d^2 \Psi}{dz^2} + m^2 \Psi = 0$$  \hspace{1cm} (1.2)

whose vertical wavenumber $m^2$ is defined, for instance, as

$$m^2 = \frac{N^2}{f_0^2} \left[ \frac{\beta}{\bar{u} - c_x} - (k^2 + l^2) \right] - \frac{1}{4H^2}$$  \hspace{1cm} (1.3)

Here, $k$ and $l$ as above are the zonal and meridional wavenumbers, $c_x$ is the zonal phase speed, $\bar{u} = \bar{u}(z)$ is the zonal mean wind, $N$ is the Brunt-Väisälä frequency and $H$ the scale height (about 7 km at a characteristic average temperature of the stratosphere). This form of $m^2$, from Holton (2004), is nearly identical to that of Charney and Drazin (1961), with an additional exponential factor in equation 1.1 for simplification, but no additional assumptions or constraints. Vertical propagation of the perturbation can occur only when $m^2 > 0$, and this is the first key result from the Charney and Drazin paper: vertical propagation is constrained by the strength of the Doppler-shifted mean flow, $\bar{u} - c_x$, and by the size of the perturbation ($k^2 + l^2$). Small wavenumbers are favoured, and the mean flow must be westerly, and not too
strong, for perturbations (waves) to propagate vertically. Charney and Drazin had set out to demonstrate why kinetic energy did not accumulate at the top of the atmosphere (Charney and Drazin (1961)), and equation 1.3 proves that atmospheric wind profiles, for most of the year and most wavenumbers, trap the perturbations in the troposphere. Yet for students of the stratosphere it is precisely the “leaks” in the trap that are of interest: the winter stratosphere allows vertical propagation of Rossby waves to the level at which the zonal mean wind is too strong and \( m^2 \) becomes negative, and the waves are said to break or to be absorbed.

Wave-breaking, i.e. the consideration of non-linear effects, led to the second key result in Charney and Drazin (1961). With non-linear terms retained, changes in the zonal-mean streamfunction become a function of the divergence of the eddy heat and eddy momentum fluxes that constitute the perturbation. Charney and Drazin (1961) demonstrate that, for steady, conservative eddies, there is no change in the zonal-mean streamfunction: under such conditions, a meridional circulation is induced that cancels the effects of the eddies. This result is now known as the Charney-Drazin non-acceleration theorem.

Arnt Eliassen, a long-term collaborator of Charney’s whose informal contribution to the non-acceleration theorem is acknowledged in Charney and Drazin (1961), studied vertical wave propagation in terms of wave energy with Enok Palm (Eliassen and Palm (1960)). Considering stationary mountain waves in the same zonal-mean framework as Charney and Drazin (1961), but allowing the mean motion to vary in latitude as well as in height, Eliassen and Palm (1960) derive expressions “relating the horizontal and vertical wave energy flux to the meridional eddy flux of sensible heat and to the meridional and vertical flux of momentum” (Eliassen and Palm (1960), p.20). Today this wave energy flux, although commonly derived somewhat differently, is referred to as the Eliassen-Palm (EP) flux.
In the 1960s and 1970s, wave breaking in the polar vortex was studied mainly in the context of Sudden Stratospheric Warmings (e.g. Matsuno (1971)). (Berlin may have the longest daily record of observations, but Sudden Warming events were well-known and documented elsewhere. The cold-war climate of the late 1950s, when the arrival of satellites made the study of the upper atmosphere strategically important, prompted the US Air Force and Canada’s Department of Defence to invest jointly in polar stratospheric research. With this funding, an often-cited 1961 study of polar vortex dynamics during a warming event was undertaken by Prof. Byron W. Boville and his Arctic Meteorology Research Group here at McGill [Boville, 1961].) Absent from the papers cited thus far in this section is any attempt to link vertical propagation and breaking of waves to the slow meridional circulation discovered by Brewer and Dobson.

*Murgatroyd and Singleton* (1961) are credited with the first effort to reproduce the observed meridional tracer motions (the Brewer-Dobson circulation) in a zonal-mean model (*Dunkerton* (1978)), with horizontal and vertical velocities calculated from the heat and continuity equations (*Murgatroyd and Singleton* (1961)). For simplicity, they neglected all eddy fluxes, but qualitatively their modelled equator-to-pole transport was consistent with the relatively high tracer concentrations observed at the poles (*Murgatroyd and Singleton* (1961)). In contrast, *Muench* (1965) (among others; see *Dunkerton* (1978)), retaining all eddy fluxes in a zonal-mean framework, arrived at a meridional structure in the stratosphere analogous to the troposphere’s Ferrel cell, with ascent at the pole and descent in midlatitudes, and this, too, was supported by observational data.

The explicit statement that Lagrangian-mean transport, which can account for the observed tracer accumulation at the pole, is not inconsistent with a zonal mean (a.k.a. Eulerian mean) eddy-driven Ferrel-like structure was made by *Dunkerton*
(1978). By then, Andrews and McIntyre (1976) had introduced the formulation of a “residual mean” circulation, defined as that part of the zonal-mean vertical velocity whose effect on the temperature is not balanced by the meridional gradient in the eddy heat flux (see Appendix A for the equations). In Andrews and McIntyre (1976) this is presented simply as a mathematically convenient transformation of variables, and the resulting set of governing equations is known as the Transformed Eulerian Mean (TEM) framework. A physical interpretation is given by Dunkerton (1978): a meridional circulation from mid- to high latitudes largely cancels the effects of the eddy heat and momentum fluxes on the zonal-mean flow, such that thermal wind balance is maintained; however, the distribution of diabatic sources of heat in winter forces a weaker hemispheric meridional circulation that accounts for mass transport from low to high latitudes. In the TEM framework, diabatic heating and eddy flux divergences both act on the residual mean circulation and on the zonal-mean flow. The eddy fluxes in this framework differ by some coefficients from Eliassen and Palm’s original derivation of the wave energy flux (Eliassen and Palm (1960)), but have become the most commonly used form and are collectively known as the EP flux components. The divergence of the EP flux constitutes a body force on the mean flow.

The main elements of the mathematical language of stratospheric dynamics were established by the papers cited above, and the refining and generalizing of the concepts in them has been on-going. Matsuno (1970) extended the Charney-Drazin criterion for vertical propagation to basic zonal flows varying in both latitude and height. The vertical momentum number $m^2$ (equation 1.3) was referred to as a “refractive index” by Charney and Drazin (1961), but Matsuno’s form of this wavenumber (Matsuno (1970)) is the refractive index commonly used today (Chen and Robinson (1992); Harnik and Lindzen (2001); Liberato et al. (2007); Rivière
Karoly and Hoskins (1982) study the refraction of planetary waves in detail with ray-tracing calculations.

Eliassen and Palm (1960) showed that the regions in which the wave energy flux converged were sinks of wave energy. Twenty years later, a number of papers by McIntyre and Palmer elucidated the concept of wave breaking (McIntyre and Palmer (1985)) and coined the term “surf zone” (McIntyre and Palmer (1983)) to describe the region of strong lateral mixing outside the polar vortex produced when upward-propagating waves reach their critical levels (i.e. where $m^2 \leq 0$ in equation 1.3). The relationship between wave forcing and the residual mean circulation was made explicit in the TEM framework of Andrews and McIntyre (1976), and an important aspect of this interaction was detailed by Haynes et al. (1991): the wave forcing affects only the residual circulation beneath it (this is now known as the “downward control” principle). The implication is that distribution of diabatic heat sources alone cannot drive the residual circulation; Plumb and Eluszkiewicz (1999) demonstrate that wave-breaking must occur within 20° of the equator to account for the observed upwelling in the tropics, while the downwelling in high latitudes is caused by wave forcing at the equatorward edge of the polar vortex.

Stratospheric Sudden Warming events (SSWs) continued to be a focus of research. Holton (1980) offers a review of SSW studies, re-expressing earlier work in the TEM terminology. Davies (1981); Schoeberl (1982) and McIntyre and Palmer (1983) suggest that pre-conditioning of the polar vortex by the “regular” wintertime wave breaking is necessary in order for the vortex to be subsequently destroyed in a sudden warming event.

In all papers cited in this section, some attempt was made to confirm the theoretical results with observational data, as available. (Conversely, lack of any observational eddy flux data was used by Murgatroyd and Singleton (1961) to justify
the exclusion of eddy fluxes in their calculations.) By the 1990s, general circulation models of the atmosphere (GCMs) were sufficiently complex to give an adequate representation of the atmosphere well into the stratosphere, and became an essential partner to the detailed study of selected phenomena. Global coverage of GCM output can compensate, to some degree, for incomplete or infrequent ground-based observational data, and modelled atmospheres, unlike the real one, allow for multiple experiments of different climates with controlled components.

1.3 The Stratosphere as a Component of the Climate System

The studies of the 1970s and 1980s outlined in Section 1.2 consider the stratosphere in isolation, and are concerned mainly with the mathematics of wave-mean flow interaction in highly idealized models. In order to minimize computational cost, many GCMs, which became standard research tools in climate sciences in these years, did not extend beyond the tropopause, in spite of recognition that even a coarse stratosphere led to better simulation of tropospheric dynamics than no stratosphere at all (Boville (1984)). Boville (1984) notes that this has implications for the accuracy of climate predictions made by models whose lid is at the tropopause, a point made again more than 20 years later by Baldwin et al. (2007), following the publication of the 2007 IPCC report. Of the 19 models which contributed to this report and to its predictions for climate scenarios of the future, only three have a well-resolved stratosphere (Cordero and Forster (2006); IPCC (2007)).

The presence and distribution of ozone in the stratosphere have been known since Dobson’s time and as of 1985 received considerable new interest due to the discovery of the Antarctic ozone hole (Farnam et al. (1985); Solomon (1987) and references therein). The importance of potential feedbacks between ozone chemistry, radiation and local stratospheric dynamics were addressed in Kiehl and Solomon
(1986) and *Shine* (1987), however, it was too computationally demanding to include a representation of chemistry, or even of chemical species as passive tracers, in GCMs at the time. Ozone was generally included either as a constant or as a climatological distribution, and the inaccuracies in the circulation that can ensue as a result of such simplification are discussed in *Shine* (1987); *Sassi et al.* (2005); *Cordero and Forster* (2006); *Curry and Scinocca* (2006).

In short, it has been agreed since the mid-1980s that accurate representation of ozone chemistry is important for correct modelling of the stratospheric dynamics; that inclusion of the stratosphere is essential for a reliable simulation of the troposphere; and that a realistic present-day troposphere is a pre-requisite for reliable predictions of future climates. Climate change experiments, in other words, at the least should be undertaken with a stratosphere-resolving GCM, and at best include interactive ozone chemistry. The importance of the stratosphere in climate research was formalized when the World Climate Research Program (WCRP) designated the stratosphere as the focus of one of its core research projects in 1992, and created SPARC – Stratospheric Processes And their Role in Climate – as an umbrella for international working groups with corresponding research goals. A necessary starting point for SPARC was an assessment of what already existed in terms of stratosphere-resolving models; where and why they differed in their climatological fields, how well they reproduced observations of the current climate, and which aspects required the greatest or most immediate attention. This was provided by the model intercomparison study by *Pawson et al.* (2000), whose Fig. 1 is reproduced here (as Fig. 1–2). It is clear that the models in the study agree well with one another and with the ERA-15 reanalysis data in the annual, global mean temperature profile throughout most of the stratosphere, although they are all too cold (Fig. 1–2, top panel). However, zonal-mean temperatures at 100 hPa (about tropopause level) show enormous spread
among models, particularly in the tropics and in high latitudes (bottom panel). The “cold pole” problem was identified as one of the key deficiencies common to all the models, none of which included interactive ozone chemistry (Pawson et al. (2000)).

Further intercomparison studies specifically targeting coupled Chemistry-Climate Models (CCMs: AGCMs coupled to a module calculating ozone chemistry interactively) were published by Austin et al. (2003) and Shine et al. (2003). In Austin et al. (2003), the objective was to assess the fate of ozone and its eventual recovery to pre-1980 levels within the context of secular climate change. All the models adhered to the IS92a scenario of the IPCC for the future evolution of greenhouse
gases (GHGs) and aerosols, but start and end times varied widely between groups, as did prescribed surface forcing, model resolution, complexity of the modelled ozone chemistry and parameterizations of gravity wave drag on the stratospheric flow. In the authors’ words, model results are “crucially depend[ent]” on all of these factors, and consensus on the details of ozone recovery was not reached (Austin et al. (2003)). Turning the focus to accurate modelling of the recent past, Shine et al. (2003) compared output from a number of models to observations of ozone loading for the period 1979–1999. Agreement in modelled temperature with observations was good at those levels that are dominated by ozone and CO$_2$ but less good where water vapour plays a role. Not all models in this intercomparison were CCMs; some were AGCMs with imposed ozone trends, and these were found to perform just as well as the CCMs when compared to observations. Comparing models to one another revealed the same spread among models as noted in Austin et al. (2003).

Two points emerge from these intercomparison studies. First, although modelling ozone chemistry with the precision only an interactive ozone chemistry scheme can give is optimal for climate change predictions in principle, in practice the CCMs of 2003 were not an improvement over prescribing ozone profiles in AGCMs. Second, the spread in model results in climate change studies, particularly in high latitudes, is mainly attributable to dynamical processes rather than chemistry: there is no agreement in Austin et al. (2003) whether wave propagation into the stratosphere will increase or decrease in future. This is an essential aspect of modelling ozone recovery: the rate of transport of ozone from low to high latitudes by the Brewer-Dobson circulation – and consequently the local density of ozone – and the temperature of the polar stratosphere in winter govern the ozone-destroying reactions. The strength of the BDC and the polar temperature are both determined by the strength of the wave forcing of the zonal-mean flow in the stratosphere.
To address the first point, a protocol for validating CCMs was outlined in *Eyring et al.* (2005), which identified key processes to target and provided minimum criteria for model complexity as well as performance benchmarks. The same authors then presented an intercomparison paper of projected stratospheric ozone (*Eyring et al.* (2007)), and identified the aspects of the chemistry schemes that lead to the greatest inter-model differences. Forcings and experimental set-up were by design very similar among models, unlike in the earlier CCM intercomparison by *Austin et al.* (2003), and model-to-model agreement was higher.

The second point, i.e. the spread in the wave forcing response among models simulating climate change, is addressed in intercomparisons carried out by *Cordero and Forster* (2006) and *Butchart et al.* (2006). *Cordero and Forster* (2006) consider the models whose output would be used in the 2007 IPCC report and note the high variability in the responses to climate change, and the large discrepancies between modelled and observed past trends at the top of the troposphere and in the lower stratosphere. As mentioned above, only a few of the IPCC AR4 models had a well-resolved stratosphere. The intercomparison by *Butchart et al.* (2006), in contrast, focused specifically on the strength of the BDC in a changing climate by examining a variety of stratosphere-resolving AGCMs. The model outputs used were available *a priori* and therefore varied somewhat because forcing scenarios and inputs were not uniform, but there was agreement that, under conditions of increased greenhouse gases, the BDC increases in strength, regardless of whether or not there is interactive ozone chemistry in the model. That being said, there is no consensus on the magnitude of the increase, as can be seen in Fig. 1–3, reproduced from *Butchart et al.* (2006) (their Fig. 9).
Figure 1–3: (Fig. 9 from Butchart et al., 2006, ©Springer). Trends in the downwelling mass flux (kT/s/year) in the extra-tropical lower stratosphere in 14 AGCM and CCM climate-change experiments.

1.4 Motivation for this Thesis

1.4.1 The Northern Hemisphere High-latitude Stratosphere in a Changing Climate

To extend the intercomparison by Butchart et al. (2006) somewhat, Table 1–1 presents 15 other model studies of climate change in the stratosphere, eight of them undertaken since the publication of Butchart et al. (2006). They are listed in chronological order, and the final column gives a brief description of the temperature response in the polar lower stratosphere in the Northern Hemisphere winter. This is a more commonly available diagnostic than the streamfunction of the TEM meridional circulation and is related to the influence of changes in the strength of the descending branch of the wave-driven BDC. In spite of the consensus noted by Butchart et al. (2006) among the models in that paper, i.e. that there is an increase in the strength of the BDC, there is little consensus in Table 1–1. However, considering only the
Table 1: List of CO$_2$-doubling or climate change experiments not included in the intercomparison by Butchart et al. (2006). Column 1: References in chronological order. Column 2: Model name, horizontal resolution, number of vertical levels and lid height. Column 3: Surface fields used; coupled slab oceans are in boldface. Column 4: CO$_2$ loading or IPCC GHG scenario used. Column 5: Number of years used for analyses or averaging. Column 6: Zonal-mean temperature response in the NH high-latitude lower stratosphere; when statistical significance was met at the 95% level, it is indicated; no comment means it was not available.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Model details</th>
<th>SST</th>
<th>GHG, CO$_2$</th>
<th>Years in Analysis</th>
<th>∆T in NH winter (polar lower stratosphere)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rind et al., 1990</td>
<td>GISS GCMAM 10° x 8° L23, 0.005 hPa</td>
<td>prescribed: prior A-O run</td>
<td>315 ppmv, 630 ppmv</td>
<td>timeslice, 3 years each</td>
<td>cooling 0-3K</td>
</tr>
<tr>
<td>Rind et al., 1998</td>
<td>GISS GCMAM 10° x 8° L23, 0.005 hPa</td>
<td>coupled slab</td>
<td>315 ppmv, 630 ppmv</td>
<td>timeslice, 10 years each</td>
<td>warming 0-1K</td>
</tr>
<tr>
<td>Butchart et al., 2000</td>
<td>UKMO UM 3.75° x 2.5° L49, 0.1 hPa</td>
<td>prescribed: prior A-O run</td>
<td>1992-2051: IS92a</td>
<td>2 runs</td>
<td>cooling 0.1K/decade (annual mean, not significant)</td>
</tr>
<tr>
<td>Gillett et al., 2003</td>
<td>HadSM3L64 3.75° x 2.5° L64, 0.01 hPa</td>
<td>coupled slab</td>
<td>289 ppmv, 578 ppmv</td>
<td>timeslice, 25 years each</td>
<td>warming 2K (not significant)</td>
</tr>
<tr>
<td>Jonsson et al., 2004</td>
<td>CMAM (CCM) T32 L65, 0.00067 hPa</td>
<td>prescribed: prior A-O run (same for all experiments)</td>
<td>“current”, 2 x current</td>
<td>timeslice, 10 years each</td>
<td>cooling 0-2K (not significant)</td>
</tr>
<tr>
<td>Sigmund et al., 2004</td>
<td>MAECHAM4 T42 L39, 0.01 hPa</td>
<td>prescribed: prior A-O run</td>
<td>353 ppmv, 706 ppmv</td>
<td>timeslice, 30 years each</td>
<td>warming 0-2K</td>
</tr>
<tr>
<td>Bourqui et al., 2005</td>
<td>IGCM-FASTOC (CCM) T21 L26, 0.1 hPa</td>
<td>prescribed: climatological (same for all simulations)</td>
<td>1979, 2060: A2</td>
<td>timeslice, 15 years each</td>
<td>cooling 0-0.25K/decade (significant)</td>
</tr>
<tr>
<td>Fomichev et al., 2007</td>
<td>CMAM (CCM) T32 L65, 0.00067 hPa</td>
<td>prescribed: prior A-O run</td>
<td>348 ppmv, 696 ppmv</td>
<td>timeslice, 30 years each</td>
<td>0-1K warming (significant at 90%)</td>
</tr>
<tr>
<td>Huebener et al., 2007</td>
<td>EGMAM T30 (2.8° ocean) L39, 0.01 hPa</td>
<td>coupled OGCM</td>
<td>1861-2000: obs 2001-2010: A2</td>
<td>3 runs of each climate</td>
<td>warming 0-1K in high lats (significant)</td>
</tr>
<tr>
<td>Kodama et al., 2007</td>
<td>MRI M598 T42 L45, 0.01 hPa</td>
<td>prescribed: HadISST (past) prior A-O run (future)</td>
<td>348 ppmv, 696 ppmv</td>
<td>timeslice, 30 years each</td>
<td>cooling</td>
</tr>
<tr>
<td>Garcia and Randel, 2008</td>
<td>WACC3M (CCM) 5° x 4° L66, 4.5x10^{-6} hPa</td>
<td>prescribed: observations (past) prior A-O run (future)</td>
<td>1950-2003: obs 1980-2050: A1B</td>
<td>3 runs of each climate</td>
<td>warming 0-0.1K/decade (not significant)</td>
</tr>
<tr>
<td>Li et al., 2008</td>
<td>AMTRAC (CCM) 2.5° x 2° L48, 0.002 hPa</td>
<td>prescribed: observations (past) prior A-O run (future)</td>
<td>1960-2004: obs 2005-2100: A1B</td>
<td>3 runs of each climate</td>
<td>cooling 0-3K in high lats; warming at 60°N</td>
</tr>
<tr>
<td>McLandress and Shepherd, 2009</td>
<td>CMAM (CCM) T31 L71, 0.00067 hPa</td>
<td>prescribed: prior A-O run</td>
<td>1950-2099: A1B</td>
<td>last 20 years minus first 20, 3 members each</td>
<td>warming 0-1K annual mean, warming 2K at 70 hPa (significant)</td>
</tr>
<tr>
<td>Winter and Bourqui, 2010</td>
<td>IGCM-FASTOC (CCM) T31 L26, 0.1 hPa</td>
<td>1. coupled slab 2. prescribed: output from coupled (climatological or interannually varying)</td>
<td>335.8 ppmv, 576 ppmv</td>
<td>timeslice, 100 years each</td>
<td>warming 4K (significant)</td>
</tr>
</tbody>
</table>

1. coupled slab
2. prescribed: output from coupled (climatological or interannually varying)
studies published later than Burchart et al. (2006) with experiments of comparable length and model complexity, a warming response is found in all but one case. Only one of the warming responses (aside from Winter and Bourqui (2010)) is statistically significant at 95% or better. In these studies, the changed climate is in the future, with considerably higher GHG loadings than present, and cannot be compared to observational data. Yet the effects of increased GHG loading since the onset of industrialization may have been significant in the stratosphere already by the early 20th century (Schwartzkopf and Ramaswamy (2008)), and other aspects of climate change, such as the widening of the tropics or the poleward shift of the Southern Hemisphere tropospheric jet, have been observed (Thompson and Solomon (2002); Seidel and Randel (2007); Lu et al. (2009)). However, changes in the transport of air from low to high latitudes in the BDC, measured as the age of air, thus far do not indicate an increase in the strength of the wave-driven meridional circulation (Engel et al. (2009)).

The temperature and wind fields of the polar stratosphere in winter are highly variable and it is difficult to obtain a statistically significant response to climate forcing there. Furthermore, the standard tool for assessing statistical significance in these studies is the bilateral Student’s T-test, which supposes a normal distribution of the data. By virtue of the Central Limit Theorem, sample sizes must be at least 30 before a normal distribution of the data may be assumed; more if the data are skewed (Wilks (2006)), as is the case for polar stratospheric temperatures (towards warmer values). Therefore in order to establish statistical significance, and to ensure that it is real, 30 years of monthly data are a barely sufficient minimum. In the post-2006 studies on Table 1–1, only McLandress and Shepherd (2009) use more (60 years in total for each climate). The first contribution made by this thesis to the study of the climate change response in the stratosphere is the extension to 100 years
(timeslice) of the length of the simulations, reducing the overall variability in high latitudes and making the statistical evaluation of results reliable.

The large majority of climate change experiments cited in the intercomparison studies of Section 1.3 and in Table 1–1 prescribe interannually varying sea surface temperatures and sea ice, using output from prior runs of a coupled atmosphere-ocean GCM for the future and observations or prior-run output for the past. The coupled AOGCMs generally have the same basic atmospheric component as the middle-atmosphere models used in the experiments, but with a lower lid. Rind et al. (1990) suggest that the discrepancy in lid height affects the climate change response, however, Sigmond et al. (2008) point out that other parametrizations which are tuned to optimize a representation of basic states with different vertical extent have a greater impact on the response than lid height per se. The argument avoids the more important issue of having surface temperatures that are in thermal balance with the atmosphere at every timestep. Rind et al. (1998) repeat the experiment of Rind et al. (1990) with SSTs calculated interactively by a mixed-layer slab ocean. The response in the high-latitude lower stratosphere in this case is a warming, whereas with the prescribed SST it was a cooling. There has been no other comparison between interactive and prescribed SSTs in middle-atmosphere models, and few other middle-atmosphere climate change experiments with a coupled slab ocean (e.g. Gillett et al. (2003)) or coupled ocean GCM (Huebener et al. (2007)). The biggest new contribution of the first paper in this thesis (Chapter 2) is providing results from CO₂-doubling experiments carried out with the same CCM, having either interactively calculated (by a coupled mixed-layer slab ocean) or prescribed (with different timescales of variability) surface temperatures. (The CCM used in the experiments presented here is the IGCM-FASTOC, described in Section 1.5 as well as in Chapters 2 and 3.) The pair of interactive simulations give a strong and statistically
significant warming response in the polar lower stratosphere, and the simulations with prescribed interannually varying surface temperatures give a weaker and not as significant warming (consistent with the findings of Rind et al. (1998)).

1.4.2 Response to Thermal Forcing in Different Latitudes

Climate response to increases in GHGs is a combination of the response to the radiative forcing of the higher GHG loading throughout the troposphere and stratosphere, and of the response to the increased surface temperature that results from GHG-induced tropospheric warming. The relative importance of these two effects in determining the overall response in the stratosphere was examined by Sigmond et al. (2004) and Fomichev et al. (2007). Both groups found the two effects to have different degrees of influence in different regions of the stratosphere. The response in the BDC is for the most part attributed to the increase in tropospheric CO$_2$ (Sigmond et al. (2004)), and so to the higher surface temperatures. It was also demonstrated by Olsen et al. (2007) that, in the absence of any changes in greenhouse gas loading, increased SST alone results in greater wave forcing of the vortex through changes in the strength of the tropospheric jet and related changes in the propagation characteristics of planetary waves.

Fig. 1–4 shows the global annual average surface temperature response to a doubling of CO$_2$ in our experiments (top panel), and to scenario A1B from the IPCC’s AR4 (bottom panel, IPCC (2007)). The response is globally positive but not uniform, having a zonal structure dictated by the distribution of land masses. This “bandedness” in the surface temperature response provides the motivation for the second paper of this thesis (Chapter 3): how sensitive is the response in the stratospheric circulation to the surface warming at different latitudes? And what processes determine the sign and magnitude of the response?
Unlike the stratospheric response to GHG increase, the response to thermal forcing applied in specific ranges of latitude has received relatively little attention. A number of studies considers the effects of realistic, often localized, surface forcings in realistic GCMs (e.g. El Niño, as in Brönniman et al. (2004); Fischer et al. (2008); Lu et al. (2008); or snow anomalies, as in Cohen et al. (2007); Fletcher et al. (2009)) while others focus on imposing idealized forcings on an idealized basic state (e.g. dry aquaplanets, as in Chen and Zurita-Gotor (2008); Brayshaw et al. (2008); Ring and Plumb (2007); Butler et al. (2010)). The second paper of this thesis bridges these two approaches by using the same fully three-dimensional CCM (IGCM-FASTOC)

Figure 1–4: The annual-mean surface temperature response (°C) to (a) CO₂-doubling in the IGCM-FASTOC and to (b) SRES A1B GHG loading in the period 2080-2099 in a multi-model mean from the IPCC (2007) (Fig. 10.8).
to investigate the response in the Northern Hemisphere winter stratosphere to latitudinal variations in thermal surface forcing. In the idealized studies cited above, the main focus is on the shift of the tropospheric jet as a response to the forcing, and the response in the stratosphere is taken to be a consequence of the changed profile of the zonal winds. However, it is pointed out in Gerber and Polvani (2009) that the absence of topography in aquaplanet models is a large factor in producing a shift of the jet. When topography is present, shifts in the jet are very small.

By applying the idealized forcing to a model atmosphere in which the effects of topography and humidity are included, the study presented here in Chapter 3 gives a more realistic response than what is found in the literature to date. It is pertinent to ask why this matters when the forcings themselves are highly idealized. The goal is to attribute, if possible, the response found in global-change experiments undertaken with full CCMs or AGCMs to contributions from processes arising from isolated zonal surface forcing, and in order to do this the model used in the experiments must necessarily be as complex as other climate change models. The first new contribution of the work in Chapter 3 is to show that surface thermal forcing imposed in isolated bands affects the direct generation, at the surface, of planetary waves having zonal wavenumbers 1 and 2, and that these anomalies are seen continuously from the surface into the stratosphere. Furthermore, the thermal surface forcing affects the baroclinicity of the troposphere through changes in the vertical shear of zonal winds particularly in the jet, and this can also lead to anomalous generation of planetary-scale waves through nonlinear interaction of synoptic waves. Second, by using a realistic model for studies of idealized, localized surface forcing, it is established that surface heating in the equatorial and tropical latitudes determines the sign of the response in the high-latitude stratosphere, even in the presence of surface warming at high latitudes.
1.4.3 The Sensitivity of the Climate Change Response to Surface Temperature Variability

The response of a model atmosphere to climate change is highly dependent on its basic state, and equally realistic basic states can respond very differently (in magnitude and even in sign) to the same applied forcing (?). Studies by Braesicke and Pyle (2004) and Olsen et al. (2007) have shown that the climate change response in the high-latitude stratosphere when the sea surface temperatures are prescribed and climatologically fixed is much weaker than when they are prescribed but have interannual variability. The results of Chapter 2 show that interactively calculated surface temperatures lead to a much stronger and more statistically robust climate change response in the polar stratosphere than when the surface temperatures are prescribed globally and have interannually variability, which is consistent with the findings of Rind et al. (1998).

This sensitivity in the climate change response motivates the third paper presented in this thesis (Chapter 4), in which results from a completed set of five pairs of simulations (control and 2×CO₂), are presented. Land or sea surface temperatures in these experiments are either calculated interactively, prescribed interannually varying, or prescribed as a climatological seasonal cycle. The differences between the experiments lie in the timescales of surface temperature variability: when monthly-means are prescribed, daily values are obtained by linear interpolation, thereby reducing daily variability. When climatological values are prescribed, interannual variability is lost. We find that a statistically significant warming response in the lowest part of the high-latitude winter stratosphere is found only when the sea surface temperatures vary at least interannually. When the sea surface temperatures are calculated interactively, this warming response is stronger and is significant up to an altitude of 30 km.
Chapter 4 demonstrates the importance of both (i) the interannual variability in ocean (and land) temperatures and (ii) the adjustment of oceans (and lands) to the atmosphere, in order to maintain the realistic stratospheric forcing necessary to adequately capture the stratospheric response to global warming. The implications of the results are that the representation of the oceans could be an important contributor to the large range of stratospheric responses to CO$_2$-doubling currently found in the literature among the various climate-chemistry models (e.g. Austin et al. (2003); Shine et al. (2003); Eyring et al. (2007)). They also suggest that multi-model averages of the stratospheric response to global warming may underestimate the increase in stratospheric wave driving, since these are dominated by models with prescribed interannually varying ocean temperatures (see Table 1–1).

1.5 The IGCM-FASTOC Model

All numerical experiments undertaken for this thesis were carried out using a Chemistry-Climate Model (CCM), i.e. a general circulation model (GCM) of the atmosphere (here, the Reading University IGCM) coupled to a stratospheric chemistry module (FASTOC). The letter “I” in Reading’s “IGCM” stands for “Intermediate”; this epithet was bestowed for historical reasons, since the model was originally conceived as an intermediate stage between a basic dry, orography-free GCM that already existed, and a planned state-of-the-art GCM with very complex physics (though this project was then abandoned). Note that the designation “intermediate” is also found in the term “Earth System Model of Intermediate Complexity” (EMIC); these are much simpler GCMs without physics parametrizations (having linear radiative relaxation and prescribed physics forcings instead). The IGCM does not belong to this category as it has a comprehensive representation of the physics. The IGCM is built on the spectral dynamical core of Hoskins and Simmons (1975).
and retains its centred-difference ("leapfrog") timestepping for the dynamics, but calculates all diabatic processes in a forward timestep after the dynamics timestep is complete. The updated diabatic terms are then incorporated with the dynamics at each new timestep. For the experiments presented in this thesis, the IGCM’s spectral representation is truncated at wavenumber 31, which corresponds to an effective resolution of $5.8^\circ \times 5.8^\circ$, that is, to about 645 km at the equator and 460 km at mid-latitudes. The corresponding transform grid has 96 longitudes and 24 gaussian latitudes per hemisphere. The model has 26 vertical levels, with a lid at 0.1 hPa; 13 levels are in the stratosphere. Level spacing is between 1 and 1.5 km in the troposphere and between 2 and 4.5 km in most of the stratosphere, increasing to about 10 km between the three uppermost levels. The model timestep used is 15 minutes, and CFL criteria are monitored through all simulations. Details on the model construction and physics parametrizations are given in the appendix of Forster et al. (2000).

In its role as a bridge between simple models and the most expensive ones, the IGCM’s physics parametrizations are intentionally kept at “intermediate” complexity. All the physical processes proper to a full three-dimensional GCM are represented, but their parametrizations are generally simpler (and faster) than in the most advanced models used for policy-related climate predictions or contributing to the IPCC’s assessment reports. For instance, the land-surface scheme is a three-layer bucket model and loses accuracy in regions that are either very dry or very wet; moist and dry convection are calculated by the Betts-Miller convection scheme, which performs best outside the tropics; damping of the dynamics near the model lid is done via Rayleigh friction applied at the top three model levels, and no additional provisions for the effects of gravity wave drag on the flow are made. Long- and shortwave radiation is handled by the Morcrette radiation scheme (Morcrette (1991)), which
includes seven radiatively active gases (O$_3$, H$_2$O, N$_2$O, CH$_4$, CO$_2$, CFC-11 and CFC-12, Taylor and Bourqui (2005)). The concentrations of these species are prescribed (note that the water vapour concentration is only specified in the stratosphere; in the troposphere, it is calculated by the model). CH$_4$, N$_2$O and H$_2$O are prescribed from annual-cycle climatologies representative of the year 1979. The well-mixed greenhouse gases CFC-11, CFC-12 and CO$_2$ are prescribed as global, annual averages. Running a CO$_2$-doubling experiment therefore simply requires changing the input value of the CO$_2$ concentration.

In all experiments presented in this thesis, the IGCM was coupled to the chemistry module FASTOC (FAsSTratospheric Ozone Chemistry). The FASTOC scheme is introduced and explained in Bourqui et al. (2005) and Taylor and Bourqui (2005), and currently includes four chemically active species (O$_3$, NO$_x$, N$_2$O$_5$ and HNO$_3$), subject to gaseous chemistry modelled for three chemical families (O$_x$, HO$_x$ and NO$_y$) (Bourqui et al. (2005); Taylor and Bourqui (2005)). This means there is presently no chlorine chemistry in the model and thus no ozone hole (the CFCs listed above are radiatively, not chemically, active). Effects of ozone depletion or recovery are therefore absent from the idealized climate change experiments presented here; the choice of 1979 as reference year for the input climatologies is consistent with this. The FASTOC scheme operates between the climatological tropopause and 4 hPa, and the chemical species are restored to their climatological values outside of this region. FASTOC represents the daily average chemistry and interacts with the IGCM once per day.

The IGCM can be coupled to a mixed-layer slab ocean and thus calculate sea surface temperatures (SSTs) interactively, rather than running with prescribed SSTs. This ensures that the model atmosphere is at all times in thermal equilibrium with the surface; the importance of this equilibrium in the model’s response to climate change
is addressed in particular in Chapter 4. In the coupled ocean-atmosphere experiments presented in this thesis, the slab ocean is 25 m deep and has (uniform) seawater density and heat capacity of 1000 kg/m³ and 4190 J/kg/K respectively. These are not true seawater values and reflect the “tuning” required to obtain accurate heatfluxes into the atmosphere. A flux correction is also applied as part of the surface radiative flux calculation that goes into determining the SST at each point. The role of this correction is to compensate for the absence of any dynamical heat transport in a slab ocean, and the correction terms are calculated separately, using SSTs and radiative flux output from a prior run of the IGCM. They are therefore model-consistent. The same flux correction is applied to the control and to the perturbation simulations, to satisfy the standard assumption that ocean heat transport does not change under doubled-CO₂ conditions. A slab ocean necessarily adds to the length of any spinup time required by the model to come into equilibrium. All results presented in this thesis follow a 10-year spinup, although 15 and 20 years were also tested. After 10 years, the coupled control climate surface temperature drifted by 0.07 degrees over the next 100 years, and the doubled-CO₂ climate drifted by 0.1 degrees over the next 100 years. Drifts after the 15 and 20-year spin-ups were very similar to one another and were smaller than after 10 years, as expected. Given the insensitivity of the stratospheric circulation to such minute fluctuations in the surface temperature and the importance of retaining as many years of the simulations as possible for analysis, the 10-year spinup was considered sufficient (the response fields did not change, in magnitude or level of significance, when the spinup length was doubled to 20 years).

The IGCM was designed to be computationally fast, as was the FASTOC module. It is always necessary in climate modelling studies to strike a balance between the most accurate representation of the climate and sufficiently long simulations to give statistically robust results. The IGCM-FASTOC was validated in Taylor
and Bourqui (2005) for its T21 version and in Fischer et al. (2008) and Winter and Bourqui (2010) in its T31 version. Its main dynamical features are consistent with observations and therefore, within the limitations of the model, we can use the IGCM-FASTOC to investigate the response of the stratosphere to the various forcings considered in this thesis. Furthermore, thanks to the computational efficiency of the IGCM-FASTOC, 50-year simulations can be performed in two to three weeks on a single CPU, even with interactive stratospheric chemistry and a coupled mixed-layer slab ocean. The ability to run 50- or 100-year simulations is crucial, as it allows attaining statistical significance where other studies have not succeeded. Note that for the purposes of the studies presented here, a parallelization of the code was not considered useful. Instead, a large number of simulations were performed in parallel on our group’s high-performance computer (‘hydroxyl’), an SGI Altix 350 server with 32 CPUs.
In this chapter, the response in the stratosphere to a doubling of atmospheric CO₂ is described. There is enhanced forcing of the stratospheric polar vortex by Rossby waves propagating vertically from the troposphere, leading to a robust warm anomaly in the Arctic lower stratosphere accompanied by considerable weakening of the polar vortex. The feedbacks between ozone chemistry and radiative cooling, together with the stronger Brewer-Dobson circulation, lead to increased ozone concentrations in the Northern Hemisphere polar stratosphere. The manuscript which constitutes this chapter was published in the Journal of Geophysical Research: B. Winter and M.S. Bourqui (2010), Wave-Forcing in the Stratosphere under Doubled-CO₂ Conditions in a 100-year Coupled-Chemistry GCM Study, J. Geophys. Res., 115, doi: 10.1029/2009JD012777. ©American Geophysical Union.
Wave-Forcing in the Stratosphere under Doubled-CO$_2$ Conditions in a 100-year Coupled-Chemistry GCM Study

B. Winter $^1$ and M.S. Bourqui$^1$

$^1$Department of Atmospheric and Oceanic Sciences, McGill University, Montréal, QC, Canada

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Abstract

The impact of doubling atmospheric CO$_2$ on the resolved-wave forcing of the stratospheric flow, and thus on the Brewer-Dobson Circulation (BDC), is investigated with 100-year timeslice simulations using a chemistry-climate model, the IGCM-FASTOC, coupled to a mixed-layer slab ocean. The Arctic lower stratosphere in winter warms by up to 4K, with associated weakening of the polar vortex and enhancement of the BDC. This change is related to a significant increase in the wave forcing near the vortex core starting in January, followed by an increased wave forcing at the lower edge of the polar vortex in February. Maximum wave forcing is found both to occur earlier in the winter and to be distributed over a longer period of time in the 2×CO$_2$ climate. The sensitivity to surface conditions is studied by repeating the CO$_2$-doubling experiments with prescribed interannually varying and fixed annual cycle monthly-mean surface temperatures. In the absence of interannual variability, the BDC response is strongly attenuated. With interannual variability, the monthly-mean prescribed surface temperatures lead to a similar dynamical response in the stratosphere as found with the interactive surface, but with reduced magnitude.
2.1 Introduction

In spite of an emerging consensus among stratosphere-resolving models that the strength of the Brewer-Dobson Circulation (BDC) increases under conditions of increased greenhouse gas (GHG) concentrations, there is considerable spread in the estimated magnitude of the increase (Butchart et al. (2006); Austin et al. (2003)). Observations to date can neither support nor reject an increase in the BDC (Engel et al. (2009)), yet knowledge of the dynamical response to climate change in the stratosphere is essential for such practical considerations as informing projections of the evolution of the ozone layer and its effect on human health. GHG increases may affect the dynamics of the stratosphere via increased infra-red radiative cooling and changes in concentrations of ozone-destroying catalysts. In addition, tropospheric sources of upward-propagating waves may change in a changed climate, thereby modifying the momentum and energy transport into the stratosphere. GHG-induced changes in the surface temperatures can alter the meridional temperature gradient, thereby impacting the strength of zonal-mean winds and the propagation direction of planetary waves. The goal of the present paper is to assess the impact of the GHG increase on the resolved-wave forcing of the stratospheric circulation and on the strength of the BDC. For the purposes of this study, climate change is restrained to its GHG radiative forcing component and the subsequent dynamical changes, and will be represented by a doubling of atmospheric CO$_2$ with respect to pre-industrial concentrations. Stratospheric photochemical changes induced by increasing concentrations of chemically active GHGs will not be considered (see for instance Bourqui et al. (2005)).

Early studies of the stratospheric dynamical response to climate change, using a full general circulation model (GCM), were undertaken by Rind et al. (1990). They doubled atmospheric CO$_2$ and forced their simulations with surface temperatures
taken from a prior doubled-CO$_2$ run of a lower-lid, coupled ocean-atmosphere version of the same GCM. They noted an increase in the Northern Hemisphere BDC, more early-winter sudden stratospheric warmings, and a warming near the north polar stratopause in winter. A number of other groups have performed similar experiments since then, using GCMs and Coupled Chemistry-Climate Models (CCMs). Their results are summarized in the intercomparison studies of Austin et al. (2003) and Butchart et al. (2006). An increase in the Northern Hemisphere BDC is noted in nearly all participating models, ranging from 0.1 to 17.8 kt/s/yr among models (Butchart et al. (2006)). Inspection of Fig. 6 in Butchart et al. (2006) reveals the participating models to span this range evenly, without clustering near a preferred value. Thus while all model responses have the same sign, there is substantial disagreement concerning the magnitude of the BDC trend.

Computing cost remains a limiting factor to ensemble sizes and simulation lengths. Rind et al. (1990) ran their model for three years, and more recent experiments typically consist of 15-30 year transient or timeslice runs (e.g. Butchart et al. (2006); Jonsson et al. (2004); Fomichev et al. (2007); Deckert and Dameris (2008)). Simulations of this length typically have limited robustness of results in areas of high variability (Austin et al. (2003); Fomichev et al. (2007)), especially in the Northern Hemisphere polar lower stratosphere. This area is of particular interest when considering future ozone trends. Recent model studies report winter-time temperature changes in the range extending from a cooling of 1–2 K (Fomichev et al. (2007); Li et al. (2008)) to a warming of similar magnitude (Sigmond et al. (2004)), with statistical significance not reaching the 10% level. Longer transient runs are now being carried out, though necessarily in small ensemble sizes (McLandress and
The experiments presented in this paper consist of 100-year times-lapse simulations for each of our climate simulations and permit a statistically robust analysis.

In all past and recent experiments cited above, the sea surface temperatures are prescribed to the model. Simulations with comprehensive IPCC climate change scenarios (e.g. Deckert and Dameris (2008); McLandress and Shepherd (2009) and references in Butchart et al. (2006)), or with idealized CO$_2$ experiments (e.g. Rind et al. (1990); Fomichev et al. (2007) and references in Butchart et al. (2006)), have surface temperatures taken from a prior run of a lower-lid version of the same GCM coupled to an ocean model. Rind et al. (1998), in a follow-up to their earlier study, coupled a slab ocean to their middle-atmosphere GCM rather than prescribing the surface temperatures. In both cases the BDC was enhanced in the changed climate, but much more so when the surface temperatures came from the interactive slab ocean. The authors note that prescribed surface temperatures coming from a lower-lid model may not be internally consistent with a middle-atmosphere model. Gillett et al. (2003) also couple a mixed-layer slab ocean to their middle-atmosphere GCM, but do not make a comparison with a forced simulation.

The absence of further experiments in which an atmospheric GCM is coupled to a slab ocean for a study of stratospheric dynamics is the first motivation for the work presented in this paper. In our simulations, the chemistry-climate model is coupled to a mixed-layer slab ocean. Surface temperatures are therefore at all times in thermal balance with the atmosphere. To evaluate the impact of allowing interactive surface temperatures, we then repeat our experiments without the mixed-layer ocean, using both prescribed fixed and prescribed interannually varying monthly-mean surface temperatures to force the model. These surface temperatures are taken from the coupled-slab ocean run. Details on the experiments are given in Section 2.2. To better
isolate key processes, climate change is represented by a doubling of atmospheric 
CO$_2$, and ozone does not undergo catalytic destruction by CFCs. However, it is 
radiatively and chemically interactive in our simulations, and has the potential to 
affect both local temperatures and larger-scale dynamics.

The second motivation for this study comes, as mentioned above, from the high 
uncertainty regarding climate responses in the polar stratosphere. Each of our sim-
ulations is run for 100 years, and our key findings, such as polar lower-stratospheric 
warming in the Northern Hemisphere winter under doubled-CO$_2$ conditions, are sig-
nificant at or beyond the 5% level.

The structure of this paper is as follows: The model and experiments are pre-
sented in Section 2.2, and results from the coupled-ocean simulations are described 
in Section 2.3. In Section 2.4, we discuss the sensitivity of our results to prescribed 
surface temperatures. Concluding remarks are in Section 2.5.

### 2.2 Model and Numerical Experiments

Our experiments are carried out using the IGCM-FASTOC, which consists of an 
atmospheric general circulation model (IGCM, Hoskins and Simmons (1975); Forster 
and Shine (1999); Rosier and Shine (2000); Forster et al. (2000)) coupled to a chem-
istry model (FASTOC, Bourqui et al. (2005); Taylor and Bourqui (2005)). Validation 
of the IGCM may be found in Forster and Shine (1999) and Rosier and Shine (2000), 
and of the IGCM-FASTOC in Taylor and Bourqui (2005). The version we use here 
has a higher horizontal resolution of T31, with 26 vertical levels, of which 13 are in 
the middle atmosphere. The model lid is at 0.1 hPa. This new version is briefly 
discussed in Fischer et al. (2008). Figure 2–1 shows solstice surface temperatures, 
zonal-mean temperatures and zonal-mean winds from the control simulation (see 
below for experimental details) as contours, with NCEP sea surface temperatures
Chapter 2: Wave-Forcing in the Stratosphere Under $2 \times CO_2$ Conditions

Figure 2–1: Comparison of control simulations (contours) with reanalyses (shaded), January-February (left) and July-August (right). Top row: surface temperatures ($^\circ C$) compared to NCEP. Middle row: zonal-mean temperatures ($^\circ C$) with ERA40. Bottom row: zonal-mean zonal winds (m/s) with ERA40.

and ERA40 zonal-mean fields\(^1\) shaded for comparison. In order to make the comparison meaningful, a climatology was constructed of ERA40 or NCEP data coming

\(^1\)ECMWF ERA-40 data used in this study were obtained from the ECMWF data server. NCEP sea surface temperatures are taken from http://www.cdc.noaa.gov/data/gridded/data.noaa.oisst.v2.html
from 15 years in which there was neither significant depletion of ozone, nor a large ENSO event (1965 through 1971 and 1974 through 1981). Surface temperatures and all features of the general circulation are well reproduced by the IGCM-FASTOC, though the polar vortex in both winter hemispheres is too strong, and thus breaks down too late in the spring, particularly in the Southern Hemisphere. Consequently, polar winter temperatures in the stratosphere are too cold by 10–12 K. The cold-pole problem is common to many stratosphere-resolving GCMs, and is related to the difficulty of representing the drag on the mean flow caused by breaking of sub-gridscale gravity waves in the upper stratosphere and above.

Rayleigh friction in the top levels of GCMs has been used since the early days of stratospheric models to damp upward propagation of waves, as well as a “brute-force” representation of the gravity-wave drag (see for instance Butchart et al. (1982)). The limitations of this approach were first formally discussed in Shepherd et al. (1996), who showed in particular that there is a high risk of inducing a spurious counter-circulation between the level of the deposition of resolved waves and the model lid, whose effects are felt on the “real” circulation below. In order to improve the reproduction of the present-day climate and thus \textit{a priori} the accuracy of the prediction of future climates, state-of-the-art GCMs increasingly use gravity wave drag parameterizations (Eyring et al. (2005)). These parameterizations have limitations of their own, stemming from both the tuning parameters used (Sigmond et al. (2008)), and the handling of momentum fluxes at the model’s top (Shepherd and Shaw (2004); Shaw and Shepherd (2007)). In the model intercomparison carried out recently by Butchart et al. (2006), most of the models shown had implemented a parameterized orographic gravity wave drag (OGWD). Sigmond et al. (2008) showed that OGWD settings influence the tropospheric circulation response to climate change. A new study by Sigmond and Scinocca (2010) demonstrates that this impact of OGWD on
the tropospheric circulation response is indirect through the effect of OGWD on the lower stratospheric basic state and resolved wave response. However, the authors caution that this conclusion cannot at present be extended to the impact on the Brewer-Dobson circulation response in the stratosphere. In light of these uncertainties, therefore, it remains useful to perform process studies attempting to isolate particular aspects of the stratospheric climate without the additional effects of a GWD parameterization (for instance as in Gillett et al. (2003); Braesicke and Pyle (2004); Kodama et al. (2007)).

The IGCM is a three-dimensional GCM with a multi-layer spectral dynamical core and a complete set of physics parameterizations. In order to keep this model computationally fast, and thereby appropriate for long climate simulations, some of its physics packages (e.g. the Betts-Miller moist and dry convection schemes) are deliberately kept simpler than in most GCMs. A consequence of overly simple convection parameterization in the tropical troposphere is an excess of vertical motion there, with boreal summer eddy heat fluxes near the tropical tropopause higher than in the ERA40 reanalysis dataset. This impacts the mass flux into the stratosphere from April through October, and the ascending branch of the Brewer-Dobson circulation in the lower tropical stratosphere in those months. The problem does not persist through the boreal winter months, and for this reason we limit analyses of wave forcing to the mid- and high northern latitudes.

Rayleigh friction acts on the top three model levels (from the lid at 0.1 hPa down to 1 hPa), decreasing downward in strength. Because of the exponential fall-off of density with height, the effects of a spurious circulation induced by the Rayleigh drag are strongest at the model top, and, as noted by Shepherd et al. (1996), drop to about 10% of their maximum value by about two scale heights (of 7 km) beneath the applied drag. The results on which we focus in Section 2.3 occur in the middle and
lower stratosphere, between two and four scale heights beneath the applied Rayleigh drag. In addition, we present our results mainly in the form of differences between a control and a perturbed climate, both of which are affected by the Rayleigh drag. The Rayleigh drag is therefore not expected to contaminate the fields analysed in this study. Nevertheless, it is important to keep in mind that this analysis includes the effects of resolved waves only, and that the absence of sub-gridscale wave drag may impact our results.

The chemically active region for the FASTOC scheme is between the tropopause and 4 hPa, and the advected chemically active species are O\textsubscript{x} (hereafter referred to as O\textsubscript{3}), N\textsubscript{2}O, NO\textsubscript{x} and HNO\textsubscript{3}. This is sufficient for an adequate representation of the pre- and post-ozone hole stratosphere (Taylor and Bourqui (2005)). Chlorine and bromine chemistry is not active, so that anthropogenic modification of ozone concentration is limited to the consequences of GHG-induced changes in O\textsubscript{x}, NO\textsubscript{x} or HO\textsubscript{x} chemistry, and in temperature or dynamics. Concentrations of other GHGs such as N\textsubscript{2}O, CH\textsubscript{4}, H\textsubscript{2}O or CFCs remain at climatological levels in the chemically active region in all our experiments. The potential photochemical changes in the stratosphere due to these other GHGs are studied in Bourqui et al. (2005).

Four pairs of simulations are performed in total, summarized in Table 2–1. In our main experiment, subject of Section 2.3, the IGCM-FASTOC is coupled to a 25-m mixed-layer ocean and interacts with a land surface scheme (a two-layer “bucket” model; see Forster et al. (1999) for details). Sea-ice is formed when sea surface temperatures are below the freezing point. This simulation will be referred to as INTocINTls, for “INTeractive ocean INTeractive land surface”. Three additional sets of simulations are then carried out with the mixed-layer ocean switched off, and the effects of different surface temperature forcings and the corresponding stratospheric responses are discussed in Section 2.4. These three additional experiment pairs are
Table 2–1: Summary of numerical experiments. In interactive (INT) simulations, the atmospheric GCM is coupled to a mixed-layer ocean (INToc) and/or land surface scheme (INTls) which calculates the surface temperatures. In the forced simulations, surface temperatures taken from the fully interactive INTocINTls simulations are prescribed as monthly means: these either vary interannually (prefix VAR) or have a fixed annual cycle (prefix FIX). Each experiment consists of a control and a doubled-CO$_2$ simulation, whose atmospheric CO$_2$ loadings are 335.8 ppmv and 576.0 ppmv, respectively. See Section 2.2 for details on the experiments and choice of CO$_2$ loadings.

<table>
<thead>
<tr>
<th>Experiment name</th>
<th>Ocean</th>
<th>Land</th>
<th>Length (timeslice)</th>
</tr>
</thead>
<tbody>
<tr>
<td>INTocINTls</td>
<td>INT</td>
<td>INT</td>
<td>100 years</td>
</tr>
<tr>
<td>VARocVARls</td>
<td>VAR</td>
<td>VAR</td>
<td>100 years</td>
</tr>
<tr>
<td>FIXocFIXls</td>
<td>FIX</td>
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<td>FIXocINTls</td>
<td>FIX</td>
<td>INT</td>
<td>100 years</td>
</tr>
</tbody>
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<table>
<thead>
<tr>
<th>Experiment type</th>
<th>Atmospheric CO$_2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>control</td>
<td>335.8 ppmv</td>
</tr>
<tr>
<td>2xCO$_2$</td>
<td>576.0 ppmv</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Surface temperature prefix</th>
<th>Meaning</th>
</tr>
</thead>
<tbody>
<tr>
<td>INT</td>
<td>calculated INTeractively each timestep</td>
</tr>
<tr>
<td>VAR</td>
<td>prescribed VARying annual cycle</td>
</tr>
<tr>
<td>FIX</td>
<td>prescribed FIXed annual cycle</td>
</tr>
</tbody>
</table>

all forced with prescribed monthly mean surface temperatures, which were calculated in the fully interactive INTocINTls run, and are then linearly interpolated to provide daily temperatures.

First, the prescribed surface temperatures have a varying annual cycle (VARoc-VARls, or VARying ocean VARying land surface). This is similar to the technique adopted by many of the climate studies cited above, in which surface temperatures from a coupled ocean-atmosphere transient experiment are used to force a transient middle-atmosphere model (see Butchart et al. (2006)). By using the same model for generating the surface temperatures as we do when prescribing those temperatures, we avoid the problems associated by Rind et al. (1998) with varying lid heights. The difference between the INTocINTls and VARocVARls simulations – that is, the
consequence of a coupled ocean and coupled land – lies in the daily variability of the surface temperatures.

Second, the prescribed monthly-mean surface temperatures have a fixed (climatic) annual cycle, thereby eliminating the interannual variability of the surface. These runs are called FIXocFIXls (FIXed ocean, FIXed land surface). The effect of interannual variability in the surface forcing is examined by Braesicke and Pyle (2004), who find a strong attenuation of the stratospheric response in their experiments when the interannual variability is absent. The difference between the VARocVARls and the FIXocFIXls simulations is, as in Braesicke and Pyle (2004), the interannual variability. Since the surface temperatures in the FIXocFIXls simulations are averages over 100 years, they are, by design, considerably more smooth than their VARocVARls counterparts.

Third, we present a hybrid simulation in which the sea surface temperatures again have a fixed annual cycle (i.e. no interannual variability), but the land surface scheme is coupled to the atmospheric model, so that land surface temperatures are fully interactive. This simulation is called FIXocINTls (FIXed ocean, INTeractive land surface), and serves to highlight the role played by the variability of land surface temperatures.

All four pairs of simulations consist of a control run and a doubled-CO$_2$ run. In the control run, atmospheric CO$_2$ concentration is 335.8 ppmv, representing the year 1979. In the CO$_2$-doubling experiment, atmospheric CO$_2$ concentration is 576 ppmv. The “doubling” is with respect to the level of 1850, considered a standard pre-industrial value. The two runs are otherwise identical. To situate 576 ppmv in the spectrum of CO$_2$ loadings considered by other experiments, it is 20% lower than a doubled present-day value (such as used in Rind et al. (1990), for instance), and corresponds to an estimated loading for the year 2060 under both the A1b and
A2 emissions scenarios \((IPCC\ (2001),\ Table\ II.2.1)\). Note that in our experiments, the only GHG whose concentration is changed is \(\text{CO}_2\). Gillett et al. (2003) finds a doubling of pre-industrial \(\text{CO}_2\) to represent the effects of combined GHG increase by the year 2040, following scenario IS92a.

All simulations are 100-year integrations of the model in timeslice mode: each of the 100 years can be considered as one member of an ensemble, which has taken its initial conditions from the final day of another member. Results shown are always climatological means over the 100 years. The integration was preceded by a spinup of five years for the dynamics alone and five additional years for the chemistry. Tests were also undertaken with longer spinup times and showed that these 10 years are sufficient for the mixed-layer ocean even in the doubled-\(\text{CO}_2\) experiment. Having a long timeslice run of 100 years makes it possible to draw conclusions about \(\text{CO}_2\)-induced climate change in the high-latitude stratosphere – both one of the most interesting regions, and the one fraught with the highest variability – with some confidence. A bilateral Student’s \(t\)-test is used for our statistical analyses. For clarity, grey shading on all plots overlays regions that are not significant at the 5% level, leaving all significant results easily visible.

2.3 Results of the Fully Interactive Simulations

2.3.1 Solstice Season Response

The surface temperature response to doubled-\(\text{CO}_2\) forcing in the IGCM-FASTOC, shown on Figure 2–2, is consistent with the model-mean results presented in the IPCC’s 4th Assessment Report \((IPCC\ (2007),\ their\ Fig.\ 10-8;\ the\ A1B\ and\ A2\ panels\ of\ the\ middle\ column\ most\ closely\ correspond\ to\ our\ \text{CO}_2\)-doubling, as mentioned in Section 2.2\). The temperature difference is 3–4K in the high latitudes, around 2K in the extra-tropics and mid-latitudes, and 1K in the equatorial Pacific. Our
high-latitude Southern Hemisphere warming is stronger by 2–3K than in the multimodel mean of the figure presented in IPCC (2007). In the Northern Hemisphere, the equatorial Pacific and some areas of the mid-latitude land masses are around 1K cooler than in the figure in IPCC (2007), while the colder spot south of Greenland is not fully captured by the IGCM-FASTOC.

The vertical profile of zonal-mean temperature change, plotted on Figure 2–3, panels A and C, also shows the preferential warming at high latitudes. The tropical upper troposphere warms as expected (Fomichev et al. (2007)), and the tropical tropopause is higher. This confinement of the warming below the tropopause, and the associated change in the meridional temperature gradients around 10 km altitude, imply an upward and equatorward intensification of the jet stream by direct application of the thermal wind relationship. In addition, a purely radiative response to a uniform increase in CO₂ concentration induces a uniform cooling of the stratosphere, increasing with height. This simple response is reasonably followed in the summer hemispheres in Fig. 2–3, but clearly breaks down in the winter hemispheres. The largest deviations are the weakening of the polar vortex in the Northern Hemisphere (NH) and the equatorward shift of the Southern Hemisphere (SH) polar vortex. The
NH reduction of the polar vortex is in thermal wind balance with a significant warm anomaly extending upwards into the stratosphere north of 60°N: the high-latitude warming decreases the meridional temperature gradient, and we find, as expected from the thermal wind relation, a corresponding decrease in the vertical wind shear, i.e. a weakening of the polar vortex (Fig. 2–3, panel B). Note also a slight strengthening of the westerly winds in the latitudes 30°N to 40°N, extending from the jet stream throughout most of the stratosphere, suggesting some equatorward expansion of the polar vortex. The high-latitude surface warming seen in Fig. 2–3 extends throughout the year (Fig. 2–4A). It decreases the meridional temperature gradient
in winter, and is reflected in the weakened mid-latitude tropospheric winds in the winter months (Fig. 2–3 and Fig. 2–4C).

In the SH winter (Fig. 2–3C) a slight warming is also present in the lower stratosphere up to 15 km, poleward of 70°S. This is similar in character to the warm anomaly found in the Northern Hemisphere winter, but with considerably weaker magnitude. At the same height (15 km) but at 60°S, a cold anomaly starts at the tropopause and extends throughout most of the stratosphere. Beginning at 15 km there is thus first an enhanced meridional temperature gradient from 40°S to 60°S, extending up to 55 km, with an associated increase of the vortex (Fig. 2–3D). Then polewards of 60°S, the meridional temperature gradient reverses and is weaker, consistent with a slight weakening of the winds of the inner side of the vortex. Finally, an intensification of the summer easterlies is also present on Fig. 2–3, and reflects the slight poleward slope with height of the isopleths of temperature differences between the equator and 30° latitude and from 20 km to 35 km.

The dynamical causes for these departures from a simple radiative response are discussed in Section 2.3.3. In Section 2.3.2, a description of the annual cycle in the wind and temperature differences is presented for the two hemispheres.

### 2.3.2 Annual Cycle Response

The annual cycle of the Northern Hemisphere temperature differences is shown as an area-weighted average for the high latitudes in Fig. 2–4A. The annual cycle of NH zonal-wind differences is shown on Fig. 2–4B, C and D for three regions: high, middle and tropical latitudes. In the upper troposphere / lower stratosphere, the warm anomaly in high latitudes mentioned above occurs in February and March (Fig. 2–4A), concurrently with a strong cold anomaly in the upper stratosphere. Earlier, in January, the warm anomaly’s presence is already seen throughout the
stratosphere, even though it is not statistically significant in the lower stratosphere. Consistently, a sequence of strong decelerations of the polar vortex relative to the control climate starts in the first week of January and persists throughout February and into March, extending to lower levels with time.

A shorter sequence of such decelerations is also seen in mid-latitudes in the upper stratosphere in January (Fig. 2–4C). This region corresponds to the southern edge of the weakening of the vortex (cf. Fig. 2–3B). In March at mid-latitudes the
upper stratospheric zonal winds are less decelerated than in the control case, but the springtime zonal-wind reversal of stratospheric winds in this region occurs at the same time in both climates, from early April at the stratopause to late April at the 30 km level (not shown).

The tropical band from the equator to 30° latitude covers the average extent of the tropospheric Hadley cell. The control wind profile in the troposphere in this region is characterized by the equatorial easterlies, or trade-winds, in the lower troposphere and the equatorward flank of the mid-latitude jet in the upper troposphere. The differences in the zonal winds between the doubled-CO$_2$ and control climates in the tropical region are shown in Fig. 2–4D. A strengthening of the trade winds near the surface is seen throughout the year, and is strongest in the solstice seasons. This is related to the higher tropical tropopause seen year-round under doubled-CO$_2$ conditions. Increased convection in the equatorial latitudes results in a stronger and wider Hadley circulation in both summer hemispheres, though it is more pronounced in the NH (not shown). The stronger Hadley cell implies enhanced trade winds.

In the stratosphere, easterly winds prevail throughout the summer while in winter the zero-wind line at 20°N separates the westerly winds to the north from the southern hemisphere easterlies. The equatorward expansion of the NH polar vortex, seen on Fig. 2–3B and described above, appears also on Fig. 2–4D as a positive wind anomaly from November through February, though it does not meet statistical significance in early January. Lower down, a positive wind anomaly near the tropopause level from mid-February until April reflects the equatorward and upward expansion of the jet stream.

In the Southern Hemisphere annual cycle (Fig. 2–5), at high latitudes the vortex is not significantly changed (Fig. 2–5B), except for a slight weakening in its lower part. Nor does the temperature signal deviate much from a simple radiative response
until the SH spring in October-November. However, the equatorward edge of the polar vortex, seen in Fig. 2–5C, reaches a peak amplitude earlier in the winter and remains stronger for a longer duration. This is seen as a strong positive anomaly in July and August on Fig. 2–5C.

The same late-fall equatorward shift of the zero-wind line as in the NH is seen on Fig. 2–5D in the SH tropical latitudes, beginning in April in the stratosphere and
descending into the troposphere from June until October. The summertime strengthening of equatorial easterlies in the stratosphere follows from October through February.

### 2.3.3 Changes in NH Wave Forcing

To investigate the departure from a simple radiative response seen in the NH winter temperature and winds on Fig. 2–3A and B, we turn now to the Transformed Eulerian Mean (TEM) diagnostics (Andrews et al. (1987)) to analyze the forcing of the stratospheric mean state winds by waves propagating upwards from the troposphere. Left-hand panels on Fig. 2–6 show the difference in total Eliassen-Palm (EP) flux divergence ($\nabla \cdot \vec{F}$), representing the forcing by resolved waves. Vectors indicate the direction of the EP flux response to CO$_2$ doubling. The right-hand panels show the difference in residual mean mass streamfunction, representing changes to the BDC.

It is clear from Fig. 2–6A that both vertical wave propagation and wave convergence increase significantly in winter under doubled-CO$_2$ conditions. Convergence of the EP flux acts to decelerate the flow, and the attendant increase in the meridional residual circulation, required to maintain geostrophic balance, is seen in Fig. 2–6B. This increased meridional circulation extends downwards in accordance with the downward control principle (Haynes et al. (1991)). The induced increased adiabatic descent is responsible for the warm anomaly in the lower stratosphere. Note that the increase in the meridional circulation reaches the high latitudes, close to the pole, and its magnitude is over 50% of the control value there. This is a very substantial modification of the high-latitude stratosphere. An increased meridional circulation in the NH due to climate change has also been seen in other studies (e.g. Rind et al.
Figure 2–6: Differences (2×CO₂ - control) for the INTocINTls experiments. A) January-February \( \bar{\nabla} \cdot \bar{F} \) (10⁻⁵ m s⁻²) with \( \bar{F} \) vectors; B) January-February residual streamfunction (Tg/s); C) July-August \( \bar{\nabla} \cdot \bar{F} \) (10⁻⁵ m s⁻²) with \( \bar{F} \) vectors; D) July-August residual streamfunction (Tg/s). Vectors are confined to the region between 5 km and 35 km vertically in the winter hemisphere. Contours on the streamfunction plots are control values. Shading covers all areas not significant at the 5% level.

(1998); Butchart et al. (2006)). Attempts to attribute the increased BDC to particular types of waves are not unanimous: in Butchart et al. (2006) (their Figs. 6 and 9), the fraction of the trend in tropical upwelling that is balanced by extra-tropical downwelling due to changes in resolved-wave forcing varies greatly from one model to another. (Note that the IGCM used in the Butchart et al. (2006) study was not the same version as the one used here, described in Section 2.2.) We used the downward control principle to calculate the residual mean mass streamfunction arising
from resolved-wave forcing alone ($\Psi_{DC}$) and compared it to the TEM streamfunction ($\Psi^*$), obtained from vertical integration of the residual meridional velocity. In the lower stratosphere in the control simulation, $\Psi_{DC}$ ranged from 70% of the $\Psi^*$ value in mid-latitudes to 100% at high latitudes. $\Psi^*$ represents all contributions to the residual circulation, including Rayleigh drag effects at the upper model levels, but the comparison with $\Psi_{DC}$ shows that its impact in the lower stratosphere is minor compared to resolved wave forcing. The total increase in the residual streamfunction shown in Fig. 2–6B is entirely due to the contribution from $\Psi_{DC}$ from 30° N to 75° N (not shown). The enhancement of EP flux convergence in the high-latitude stratosphere seen on Fig. 2–6A is therefore attributed to increased forcing by large-scale resolved waves.

We now consider the annual evolution of total wave forcing between the latitudes 50°N to 70°N at two different heights: 4 hPa (Fig. 2–7A), corresponding to the strongest part of the polar vortex in our domain, and 45 hPa (Fig. 2–7B), corresponding to the lower extent of the vortex. At 4 hPa, wave forcing is significantly stronger throughout December, January and February under doubled-CO$_2$ conditions. Signs of reduced wave forcing are apparent in March, but not statistically significant. Lower in the stratosphere, at 45 hPa, increased wave forcing is significant only in February, and a significant reduction is seen in April. This is consistent with the zonal wind anomalies seen on Fig. 2–4B, which show the vortex to be eroded from the top down throughout January-March under CO$_2$-doubling. Contours on panels A and B of Fig. 2–7 give control values of $\nabla \cdot \vec{F}$. These show that the strongest increases in wave forcing under doubled-CO$_2$ conditions occur one or two months ahead of the maximum control forcing.

To illustrate this change in the timing of wave forcing in greater detail, the histograms on panels C and D of Fig. 2–7 show how many, out of the 100 timeslice
Figure 2–7: **Upper panels:** \((2\times\text{CO}_2 - \text{control})\) differences in \(\nabla \cdot \vec{F}\) \((10^{-5} \text{ m s}^{-2})\) for the INTocINTls experiments, at \(A\) 4 hPa and \(B\) 45 hPa. Contours are control values. Shading covers all areas not significant at the 5% level. **Lower panels:** Histograms of annual maximum wave forcing events \((\nabla \cdot \vec{F})\) between the latitudes 50°N and 70°N for the INTocINTls experiments at \(C\) 4 hPa and \(D\) 45 hPa. Blue is control and red is doubled-CO\(_2\).

years, have their maximum annual wave forcing in a particular month. For these plots, total EP-flux divergence was averaged over the latitudes 50°N to 70°N, then the month having the largest monthly-mean value of negative EP-flux divergence
was identified for each winter, defined as the period November-April. Not only the
timing, but the character of annual wave forcing is affected by the changed climate.
At 4 hPa, in the control case, 42% of years have the strongest annual wave forcing in
March, and 21% in February. January, at 11%, is not much more likely than April
to experience the strongest wave forcing. In the 2xCO$_2$ case however, strongest wave
forcing is nearly equally likely in January, February, and March, with respectively
26%, 24%, and 29% of all years. Thus the strongest forcing of the upper vortex
over the winter period is not only likely to occur earlier in the winter, but is also
distributed over a longer period of time.

Lower down, at 45 hPa, March and April experience the most wave forcing
among the control years, followed by the January-February period. Among the
2xCO$_2$ years, February and March are most often the months with the strongest
wave forcing, followed by a sharp drop-off into April. In both climates, there is a
two-month period in which about half of all years experience the strongest wave
forcing, but it occurs one month earlier in the 2xCO$_2$ case.

A conventional metric for assessing the wave flux entering the stratosphere,
in particular the wave flux that affects the polar vortex, is the zonal-mean eddy
heat flux $\overline{v'T'}$ at the 100 hPa level (*Newman and Nash* (2000)). The shift towards
the earlier part of the NH winter can be seen here as well, as shown in Fig. 2–8.
While the latitudinal extent over which $\overline{v'T'}$ is positive does not change significantly
under doubled-CO$_2$ conditions, the months during which it is strongest shift from
February-March to December-January, with an enhanced peak in January. Waves
entering the stratosphere propagate vertically along the polar vortex and break first
at the highest levels, thereby altering the zonal-mean wind profile. Subsequent waves
break at lower levels. The shift in the 100 hPa wave flux towards December-January
Figure 2–8: Differences (2×CO₂ - control) in the NH winter zonal-mean eddy heat flux ($v'T'$, K m/s) at 100 hPa for the INToclINTls experiments. Contours are control values, spaced every 4 K m/s until 16 K m/s, then every 2 K m/s. Shading covers all areas not significant at the 5% level.

seen on Fig. 2–8, and particularly the high January value, is thus consistent with the strong January wave forcing at 4 hPa illustrated on Fig. 2–7A and C.

The eddy heat flux also attains larger values in the changed climate, in addition to reaching its maximum amplitude earlier in the year. The highest values in the control simulation, seen as contours on Fig. 2–8, are below 19 Km/s, occurring in February-March. The highest values in the doubled-CO₂ climate are above 20 Km/s (contours not shown), and occur in January. The combined effect of higher amplitude and shift in the seasonal cycle is reflected in Fig. 2–6A, where the vectors give the differences in EP flux. These differences are essentially in the vertical component of the flux, which is proportional to the zonal-mean eddy heat flux. Note that the EP flux on Fig. 2–6 has been normalized by the density, which enhances the plotted vector lengths at high altitudes. Overall, then, the increased wave-induced BDC seen as a January-February average on Fig. 2–6B in the NH winter is due both to the stronger forcing occurring in the doubled-CO₂ climate, and to the earlier onset of strongest annual forcing.
2.3.4 Ozone

Ozone is produced at low latitudes and transported polewards by the Brewer-Dobson Circulation, thus any climate-induced change in the latter will necessarily affect the amount of ozone found over the poles. As outlined in Section 2.2, the IGCM-FASTOC includes simplified ozone chemistry, in the sense that destruction by bromine and chlorine is absent. Changes seen in the ozone field in the doubled-CO$_2$ experiment are therefore due mainly to two factors:

1) The observed enhancement, in strength and in duration, of the BDC in the boreal winter months leads to more ozone in the polar lower stratosphere.

2) The CO$_2$-induced radiative cooling of the mid- to upper stratosphere decreases the efficiency of ozone destruction by NO$_x$, resulting in more ozone there at nearly all latitudes.

Figure 2–9 shows the change in ozone number density (panel A) and mixing ratio (panel B) under doubled-CO$_2$ conditions, for the January-February period. Above 30 km, ozone abundances are mainly photochemically controlled and Factor 2 above leads to increases in ozone by about 10%. Below 30 km, transport is the dominant factor (Dessler (2000)), such that the two factors act together to increase ozone found at the winter pole: greater transport of air with higher concentrations of ozone.

In the lower stratosphere, ozone behaves as a GHG and heats the air radiatively between 20 km and 30 km in the tropics and in the summer hemispheres. In the subarctic winter, radiative heating by ozone is mainly confined between 10 km and 20 km (Liou (2002)). Between 20 km and 25 km, in the subarctic winter, ozone is radiatively neutral (Liou (2002)). Changes in high-latitude winter ozone abundances could therefore modify the meridional temperature gradient in the extra-tropical lower stratosphere, and in turn the polar vortex. Note that such effects were detected
Figure 2–9: Upper panels: Differences (2xCO₂ - control) in the January-February ozone abundances expressed as A) number density (10¹⁷ molecules/m³) and B) mixing ratio (ppmv). Contours on A and B give control values. Lower panel: Annual cycle of differences (2xCO₂ - control) in the ozone number density (10¹⁷ molecules/m³; colours) and in the residual vertical velocity \( \bar{w}^* \) (hPa/day; contours). Shading covers all areas not significant at the 5% level. On the lower panel, white shading pertains to the ozone values and grey shading to the \( \bar{w}^* \) values.

in particular in the SH summer by Son et al. (2008) and Li et al. (2008) as a result of the persistance of low polar ozone levels from the spring into the summer.

The comparison of NH high-latitude annual cycles of the differences in ozone number density, vertical residual velocity \( \bar{w}^* \) (both on Fig. 2–9C), and temperature (Fig. 2–4A) provides some insight on the importance of the radiative heating by changes in ozone. First note that the strongest anomalous downwelling under
doubled-CO$_2$ conditions around 20 km altitude begins in January, peaks in February and tapers off sharply in March, consistent with the peak wave forcing by month described in Section 2.3.3. The temperature anomaly follows this cycle without apparent lag, but seems to last longer into March (Fig. 2–4A). The pattern of the ozone anomaly on Fig. 2–9C closely follows the pattern of $\overline{w^r}$ from October into February, but with a lag of about one month and a vertical shift. From late fall through February, the coincident peaks in $\overline{w^r}$ and temperature anomalies, and the shift in the ozone anomaly, suggest that radiative heating by ozone is unimportant. In March, the downwelling anomaly weakens very sharply while the ozone anomaly persists (Fig. 2–9C), as does the warm anomaly (Fig. 2–4A): this indicates that radiative heating by ozone is likely more important at this time than wave-induced downwelling. This tentatively suggests that the ozone anomaly makes only a minor contribution to the NH winter warm anomaly, but may help extend its duration slightly. A more formal separation of the contribution to the warm anomaly made by the changes in ozone concentration from the contribution made by the enhanced adiabatic descent would require a separate model simulation without chemistry.

It must be kept in mind however that the northern high-latitude warm anomaly in the lower stratosphere would conceivably be smaller if polar ozone destruction by chlorine and bromine were taken into account (that is, within ozone hole conditions). Since the warm anomaly is in thermal wind balance with the polar vortex, and since vortex strength affects both the spectrum of waves that propagate into the stratosphere and the height at which they break, a lower concentration of ozone in winter would have consequences for the vertical profile of zonal winds and even for final vortex break-up dates.
2.4 Results with Prescribed Surface Temperatures

Experiments such as those presented in Section 2.3, with a mixed-layer slab ocean coupled to a middle-atmosphere-resolving GCM, were also carried out by Rind et al. (1998) and Gillett et al. (2003), but are otherwise in the minority. Both of these studies exhibit northern high-latitude, lower-stratospheric warming in the winter resembling what we present here; in Rind et al. (1998), it is of similar magnitude and latitudinal extent, but reaches higher into the stratosphere. In Gillett et al. (2003), this winter warm anomaly is at the same location as in our study, but weaker, and below statistical significance (perhaps due to their relatively short timeslice integration).

More common are studies in which monthly-mean sea surface temperatures are prescribed, and are either interannually varying (Eyring et al. (2007)), or have a fixed annual cycle (e.g. some of the models presented in Butchart et al. (2006)). This raises the question of the effect of the nature of the surface forcing on the stratospheric flow. As mentioned in Section 2.1, Rind et al. (1998) compare results from an atmospheric GCM forced with prescribed interannually varying surface temperatures to results from a GCM coupled to a mixed-layer slab ocean. The prescribed surface temperatures in that experiment came from a coupled ocean-atmosphere GCM having a lower lid than the middle-atmosphere GCM the authors used, and they attribute the different stratospheric responses to this difference in lid height.

Braesicke and Pyle (2004) compare results from model simulations forced with interannually varying surface temperatures to results when the prescribed surface temperatures have a fixed annual cycle, and find the stratospheric response to be qualitatively the same, but weaker, without interannual variability.

In this section, we present results from three pairs of simulations identical to those described in Section 2.3, except that the atmospheric model no longer interacts
freely with the surface. Instead, monthly-mean surface temperatures coming from
the fully interactive run (INTocINTls, for INTeractive ocean, INTeractive land sur-
face) are used to force the atmosphere. These additional simulations are described in
Section 2.2 and listed in Table 2–1. Figure 2–10 shows the zonal-mean temperature,
zonal wind, resolved wave forcing ($\vec{\nabla} \cdot \vec{F}$) and residual mass streamfunction ($\Psi^*$) re-
response to doubling of atmospheric CO$_2$ for the three new simulations: VARocVARls
(with prescribed interannually VARying ocean, VARying land surface temperature)
in the left-hand column, FIXocFIXls (with prescribed FIXed annual cycle for both
ocean and land surface temperature) in the centre column, and FIXocINTls (with
prescribed FIXed annual cycle for the ocean but INTeractive land surface tempera-
ture) in the right column. Fig. 2–10 is a January-February average and shows only
the northern hemisphere.

The top row of Fig. 2–10 shows the zonal-mean temperature response to dou-
bling of CO$_2$ and should be compared to the right half of Fig. 2–3A. It is striking
that none of the experiments with prescribed surface temperatures fully reproduces
the strong northern high-latitude, lower stratosphere warm anomaly seen on Fig. 2–
3A. However, a warm anomaly of up to 2$^\circ$C is seen to a height of 15 km in the high
latitudes in the VARocVARls simulations (Fig. 2–10, top left panel). Consistent
with the temperature response, the zonal-mean zonal winds, displayed in the second
row of Fig. 2–10, exhibit a slight weakening of the lower vortex in the VARocVARls
simulations but none at all in the simulations having a fixed annual cycle. The third
row of Fig. 2–10 shows the change in wave forcing (shading) and in the EP flux
vectors (vectors are scaled to represent the same magnitude on all graphs). Com-
paring this row with the right-hand side of Fig. 2–6A suggests a similarity between
VARocVARls and INTocINTls responses: in both cases, there is enhanced EP flux
from the troposphere between 30$^\circ$N and 50$^\circ$N, which curves north and upwards into
Figure 2–10: January-February response in the northern hemisphere to CO₂-doubling in simulations with prescribed surface temperature: **Top row:** zonal-mean temperature (°C); **second row:** zonal-mean zonal wind (m/s); **third row:** change in $\nabla \cdot \vec{F}$ ($10^{-5}$ m s⁻²) with overlaid vectors of the EP flux response; **bottom row:** residual mass streamfunction (Tg/s). In all panels, contours are control values and shading covers all areas not significant at the 5% level. See Sections 2.2 and 2.4 and Table 2–1 for details on these simulations.

the stratosphere, associated with strong convergence above 35 km between 50°N and 70°N where the flux turns towards the equator. The difference between the VARoc-VARIs and the INTocINTls responses to CO₂ doubling lies in the magnitude of the
EP flux, which is everywhere stronger in the fully interactive INTocINTls case, particularly in the middle stratosphere between 40°N and 70°N. The associated stronger convergence of EP flux, i.e. the stronger wave forcing of the vortex, extends further down into the middle stratosphere in the INTocINTls experiments. The FIXocFIXls and FIXocINTls responses, on the other hand, do not exhibit a significant increase in the EP flux from the mid-latitude tropopause into the polar vortex, and are therefore associated with weaker upper stratospheric EP flux convergence.

The changes in residual mean mass streamfunction (bottom row of Fig. 2–10) are consistent with the changes in wave forcing in the upper stratosphere. The streamfunction increases in all three cases on Fig. 2–10, but more so in the VARocVARls simulations than in the other two. The increase in the descending branch reaches farthest north in the VARocVARls, but does not reach statistical significance in any of the high latitudes, unlike the INTocINTls case.

It is important to keep in mind that the basic state of the control climate plays a key role in determining the response to a climate perturbation; see for instance Sigmond and Scinocca (2010). Comparisons of the January-February zonal-mean zonal winds of the four control simulations (not shown) reveal that the VARocVARls and INTocINTls are nearly identical, though in VARocVARls the lower-stratospheric polar vortex, the equatorward flank of the NH tropospheric jet, and the mid-stratospheric easterlies at 20°S are about 10% weaker than their counterparts in the INTocINTls basic state. Aside from this weakening, all features of the circulation are in the same location in the two control states. When the annual cycle is fixed in the sea surface temperature forcing field, as in the FIXocFIXls and FIXocINTls simulations, there is a large weakening of the equatorial easterlies centred at 40 km with respect to the INTocINTls control state, and the zero wind line in the NH shifts
equatorward by nearly 20° latitude. Holding the annual cycle fixed in the land surface temperature forcing as well as in the SSTs, as in the FIXocFIXls simulation, has the additional effect of strengthening the polar vortex. It is beyond the purpose of this paper to analyse these changes in the control states, but the differences between the two FIXoc simulations and the INTocINTls run preclude further comparison of their respective climate change responses.

The differences between simulations with prescribed or interactive surface temperature discussed on the basis of Fig. 2–10 focus on the January-February period, and the question naturally arises of a possible difference in the timing of these responses. The annual cycle of the high-latitude response in the VARocVARls simulations show the development of a warm anomaly in the lower polar stratosphere in February, which persists into March (not shown). This response is delayed by a month, and is weaker (by about 2K), compared to the INTocINTls experiments shown on Fig. 2–4A, where the warm anomaly appears in January, reaches its peak in February, and persists into March. A similar comparison of annual cycles of the wave forcing and EP flux responses shows the biggest signal in January and February in the INTocINTls runs (as on Fig. 2–7), and in February and March in the VARocVARls simulations (not shown). A similar and consistent delay is found in the eddy heat flux response at the 100 hPa level, averaged over 40°N to 80°N.

The INTocINTls and VARocVARls pairs of experiments thus present qualitatively the same response to a doubling of CO₂ but the response in the VARocVARls simulations is weaker and lags the INTocINTls response by about a month. The cause of this quantitative difference lies in the daily variability of surface temperatures. In the VARocVARls runs, the daily surface temperatures are obtained by interpolating between the prescribed monthly-mean values taken from the INTocINTls runs. This smooths out the daily INTocINTls surface temperatures, so that, for instance,
a given January in the INTocINTls run may be characterized by two or three major storms – resulting in strong stratospheric wave forcing events – but its monthly-mean surface temperature may be only a little above climatology. However, a mechanistic description of the causal relationship between this difference in the stratospheric response and the daily variability of surface temperatures is beyond the scope of this study.

2.5 Conclusions

We have used the IGCM-FASTOC, a chemistry-climate model coupled to a mixed-layer slab ocean, to show that a doubling of pre-industrial CO$_2$ levels (to 576 ppmv) in the atmosphere leads to strongly increased resolved-wave forcing of the Northern Hemisphere polar vortex. The enhanced descending branch of the Brewer-Dobson Circulation (BDC) at high northern latitudes leads to significant adiabatic warming in the lower stratosphere in January and February, as seen on Figs. 2–3 3A and 2–4A. Specifically, the lower-stratosphere NH winter pole warms by up to 4 K. Other studies that examine the response to CO$_2$-doubling in an atmospheric GCM coupled to a slab ocean include Rind et al. (1998) and Gillett et al. (2003), who find a similar high-latitude warm anomaly (of 2 K in each case, for a CO$_2$ doubling value of 760 ppmv in Rind et al. (1998) and 578 ppmv in Gillett et al. (2003)). The enhanced BDC together with the radiative cooling of the upper stratosphere by CO$_2$ lead to an increased ozone concentration in the winter polar lower stratosphere, which extends into the spring. This increase in concentration is of about 10%, and its radiative impact is minor. Note, however, that GHGs other than CO$_2$ were not included in this study. Bourqui et al. (2005) showed that anthropogenic increases in N$_2$O lead to an ozone change of roughly the same magnitude, but of opposite sign, as the cooling effect of CO$_2$ in the upper stratosphere does. With respect to this, the present study
provides an upper bound on ozone changes in the winter pole. Our analysis suggests a mechanism through which the severity of an Arctic ozone hole might be reduced by the effects of increasing CO$_2$ concentrations on the stratospheric dynamics and vortex temperatures.

The magnitude of the BDC increase as represented by the mean mass stream-function is about 50% of its control value from 70°N polewards, with an estimated winter-time hemispheric trend between 3.3 kt/s/yr and 4.4 kt/s/yr, calculated considering the IPCC’s SRES A1B over 80 years and SRES IS92a over 60 years, respectively (see Section 2.2). This falls within the range given in Butchart et al. (2006).

A key result of this study is that the timing of the wave forcing events changes under doubled-CO$_2$ conditions. In the control climate, the amount of forcing of the mean flow at the top of the stratosphere builds gradually throughout the winter months, towards a maximum in March. In the doubled-CO$_2$ climate, maximum wave forcing begins in January and persists through March. The vortex is therefore subject to erosion by waves through a longer period of time. The implications for the frequency and timing of stratospheric sudden warmings will be the focus of a future publication.

We have also demonstrated that surface temperatures generated interactively by the model’s coupled mixed-layer ocean and soil scheme have a different impact on the stratosphere than prescribed surface temperatures, even when these have a varying annual cycle. Here, model differences, such as suggested by Rind et al. (1998), cannot account for the change in response between the fully interactive (INTocINTls) and the forced simulations having interannual variability (VARocVARls) since the same model is used. We find that the magnitude of the response, as well as its timing, is affected by prescribing monthly-mean surface temperatures.
In the absence thus far of conclusive observational evidence for an increase in the Brewer-Dobson Circulation due to enhanced GHG forcing (Engel et al. (2009)), it is important to treat the predictions made by climate-chemistry models with caution. This study shows that the stratospheric response to climate change is highly sensitive to the representation of surface temperatures, on sub-monthly as well as interannual timescales. The smoothing out of sub-monthly temperature variations, particularly in the winter months, seems to attenuate the wave forcing of the stratospheric flow, which is sensitive to forcings of short duration (such as those leading to Sudden Warming events). The fact that the response in simulations forced with monthly-mean (but interannually varying) surface temperatures lags the response in the fully interactive simulations has implications for comparisons with other studies, when results are presented as averages over two or three months.

Braesicke and Pyle (2004) compare the climate-change response from simulations forced with monthly-mean, interannually varying surface temperatures, and simulations forced with a fixed annual cycle of climatological temperatures. These authors find the responses to be similar, but with lower amplitude when the fixed annual cycle is used. The same comparison can be made between the VARocVARls and FIXocFIXls simulations in our study by considering the left-hand and middle columns of Fig. 2–10, though keeping in mind that there are significant differences in the control basic states of the tropical stratosphere between these two cases. The responses in the temperature and zonal wind fields are similar, but weaker for the FIXocFIXls simulations, in agreement with Braesicke and Pyle (2004). However, this is no longer true for the response in the EP flux and its divergence, i.e. for the wave forcing response. When there is no interannual variability in the prescribed surface temperatures (centre column on Fig. 2–10), the wave forcing response is very weak and the response in EP flux is limited to a slight strengthening of its vertical
component in the mid-latitudes. When interannual variability is present only in the land surface temperatures, not in sea surface temperatures (experiment FIXocINTls, right-hand column of Fig. 2–10), then there is additional strengthening of the vertical EP flux in high latitudes. This is very different from the VARocVARls response, in which both the strength and the direction of propagation of the EP flux are affected by doubling of CO$_2$. An implication is that climate change experiments running with fixed annual cycle surface temperatures risk an underestimation of the wave forcing response.

This study also re-emphasises the importance of considering the seasonal cycle of the response to climate change. Without this information, the comparison of results from different models may be misleading. More investigations are required to better understand the mechanisms linking the various timescales of surface variability and the different responses of the stratosphere. Finally, the absence of gravity wave drag parameterization in the model implies that such a parameterisation has not been used to correct the mean flow to better match observations. Subsequent biases in the model (shown in Fig. 2–1), particularly in the polar vortex, may have some influence on the response of the stratosphere found in this study. Further investigations using the same model but with a gravity wave drag parameterisation are needed to address this issue.

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CHAPTER 3

Sensitivity of the Stratospheric Circulation to the Latitude of Thermal Surface Forcing

The winter response in the Northern Hemisphere high-latitude stratosphere is known to be in large part caused by the greenhouse gas-induced changes in the troposphere, where a key effect of greenhouse gas warming is to raise surface temperatures. The surface temperature increase is global but not uniform: high latitudes show the greatest surface temperature anomalies in both hemispheres, and Northern Hemisphere mid-latitudes undergo greater warming than tropical latitudes. There is therefore a zonal aspect to the surface warming, particularly in higher latitudes where the presence of landmasses dominates the surface temperature signal. In this chapter, a surface thermal anomaly is imposed in a sequence of latitude bands, in order to establish the effect of each band in isolation on the stratospheric circulation, and to identify which latitudes govern the stratospheric response when several bands are forced simultaneously.

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Chapter 3: Sensitivity to Latitude of Surface Forcing

Sensitivity of the Stratospheric Circulation to the Latitude of Thermal Surface Forcing

B. Winter\textsuperscript{1} and M.S. Bourqui\textsuperscript{1}

\textsuperscript{1}\textit{Department of Atmospheric and Oceanic Sciences, McGill University, Montréal, QC, Canada}

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Abstract

Using the Chemistry Climate Model IGCM-FASTOC, we analyze the response in the Northern Hemisphere winter stratosphere to idealized thermal forcing imposed at the surface. The forcing is a 2K temperature anomaly added to the control surface temperature at all gridpoints within a latitudinal window of 10 or 30 degrees. The band-wise forcing is applied systematically throughout all latitudes of the Northern Hemisphere. Thermal forcing applied anywhere equatorwards of 20°N, or continuously from the equator to 30°N, increases planetary-wave generation in the troposphere and enhances the flux of wave activity propagating vertically into the stratosphere. Consequently, a greater flux of wave activity breaks in the polar vortex, increasing the Brewer-Dobson circulation and leading to a warm anomaly in the polar stratosphere. Ozone concentration increases at high latitudes and decreases at low latitudes. Thermal surface forcing imposed between 30°N and 60°N has the reverse effect – decreased planetary-wave generation in the lower troposphere and reduced vertically propagating wave flux entering the stratosphere – and leads to a stronger and colder vortex. Thermal forcing applied polewards of 60°N has little effect on the tropospheric mean state, but nonetheless decreases the planetary-scale eddy heat flux from the surface to the tropopause, resulting in a sufficient decrease of the vertical flux of wave activity for the vortex to be anomalously strong and cold. When surface forcing is imposed only polewards of 30°N, ozone concentration decreases at high latitudes but is not affected at low latitudes. Combining the forcing in an equatorial and an extra-tropical band leads to a response similar to that of the equatorial forcing, demonstrating that the subtropical surface temperature changes determine the sign of the surface-driven response in the vortex.
3.1 Introduction

Annually averaged surface temperatures, under climate change conditions, are expected to increase by 1-2°C near the equator to 4-6°C at mid- to high latitudes by the end of the 21st century, depending on the greenhouse gas emissions scenario (IPCC (2007), Fig. 10-8 therein). The response in the winter lower stratosphere in high northern latitudes to similar emissions scenarios applied in stratosphere-resolving models is projected to include greater wave forcing in the polar vortex and thus an increased Brewer-Dobson circulation, with associated reduced vortex strength and higher temperatures in the polar lower stratosphere (Rind et al. (1998); Sigmond et al. (2004); Braesicke and Pyle (2004); Butchart et al. (2006); Olsen et al. (2007); Winter and Bourqui (2010)). The question has then been asked whether the stratosphere in such experiments is responding to the radiative effect of the increased greenhouse gas loading throughout the atmosphere or to the changed surface temperatures, or both. Olsen et al. (2007) showed that, in the absence of any changes in greenhouse gas loading, increased SST alone results in greater wave forcing of the vortex through changes in the strength of the tropospheric jet and associated changes in the propagation characteristics of planetary waves. The study by Sigmond et al. (2004) compares the response in the stratosphere to a doubling of CO₂ applied throughout the atmosphere to that of a CO₂-doubling applied either in the troposphere alone or in the stratosphere alone, and attributes two-thirds of the warming in the Arctic winter lower stratosphere to the tropospheric CO₂ doubling. In these experiments, SSTs were adjusted to be consistent with the corresponding radiative effect of CO₂, and were therefore higher in the tropospheric than in the stratospheric doubling case. In the Northern Hemisphere, the stratospheric response to tropospheric CO₂ doubling was found to be driven by the higher SSTs (Sigmond et al. (2004)).
The potential importance of the surface temperature in modulating the strength of the northern polar vortex in winter has also been investigated with a focus on localized phenomena with a strong signal in the surface temperature, such as the El Niño/La Niña oscillation or Siberian snow cover changes. The signature of El Niño in the SST field is a warm anomaly in the central to eastern equatorial Pacific of about 2°C (though episodically up to 5°C) (Philander (1990)). Its effect on the northern stratospheric winter vortex is found to be an increased forcing by planetary waves (Brönniman et al. (2004); Taguchi and Hartmann (2006); Bell et al. (2009)). Localized surface temperature changes due to seasonal snow cover anomalies in Siberia were studied by Cohen et al. (2007) and Fletcher et al. (2009). In these studies, early-winter increases in the snow cover extent over Siberia, by inducing a surface temperature anomaly of about -3°C via albedo changes, affect the local diabatic heating in such a way as to excite a Rossby wave response, which propagates to the stratosphere. Siberia was chosen for these experiments because it is a key area for generating vertically propagating Rossby waves (Cohen et al. (2007)).

To isolate the processes involved in transmitting the surface signal to the stratosphere, other studies have focused on idealized surface forcings imposed on an idealized climate state. Brayshaw et al. (2008) imposed SST anomalies of finite length and width at different latitudes in an aquaplanet GCM and investigated the effects on the stormtrack (at 850 hPa) and on the tropospheric jet of varying both the gradient and the location of the anomaly. The changes in SST gradient patterns were found to be more relevant to the response in the wind field than the magnitude of the anomalies themselves, and the position of the anomaly gradient relative to the location of the subtropical jet was a key factor in determining whether the jet underwent a poleward or an equatorward shift. In Chen et al. (2007) and Chen and Zurita-Gotor (2008), the applied forcing was a surface friction and a westerly torque
(not confined to the surface), respectively. The torque was moved incrementally from low to high latitudes, and the authors noted that the response of the jet was to shift equatorwards for low-latitude forcing and polewards for high-latitude forcing, with a transition from one to the other when the forcing was at 35°N. The authors explained these jet displacements on the basis of shifts in the critical latitudes for tropospheric eddies (Chen and Zurita-Gotor (2008)). More recently, Chen et al. (2010) investigated the different effects of imposed SST warming in high or low latitudes on an aquaplanet model having a moist atmosphere. The emphasis in these idealized studies is on the displacements of the tropospheric jets in general, and particularly on the eddy processes leading to such shifts. Responses in the stratosphere to surface and tropospheric forcing are not explicitly considered (rather, changes in the lower-stratospheric winds are the starting point leading to the tropospheric jet shifts (Chen and Zurita-Gotor (2008); also Wittman et al. (2007)). Note that it is pointed out in Gerber and Polvani (2009) that the absence of topography in aquaplanet models is a large factor in producing a shift of the jet. When topography is present, shifts in the jet are very small.

Between studies targeting specific processes in idealized climates and studies examining realistic surface perturbations in comprehensive GCMs there is a gap. Experiments with idealized surface forcings imposed on a realistic basic state atmosphere would help untangle the link between surface temperature changes and the stratospheric response to increased greenhouse gas forcing. Similarly to fully idealized studies focussing on the troposphere, such experiments can establish the surface region of largest influence on the stratosphere and the sign of the change. However, such experiments are not well represented in the literature. It is the aim of the present paper to help fill this gap, with particular attention to the Northern
Chapter 3: Sensitivity to Latitude of Surface Forcing

Hemisphere stratosphere in winter. Given that the surface temperature increase associated with climate change is global but has latitudinal structure, our underlying motivation is to relate the stratospheric response to a surface temperature change in a particular range of latitudes. We therefore investigate the effect on the stratospheric circulation of surface temperature changes in given latitude bands. This is the simplest way of controlling the location of the forcing and the subsequent change in the meridional thermal gradient, and allows the forcing to be applied systematically to a series of different latitudinal windows. Although the surface forcing is zonally uniform, the control surface field on which it is imposed is not; it has a realistic distribution of topography and specific humidity and thus of temperatures, whose zonal gradients are retained when the thermal anomaly is added.

The experimental set-up and model are described in detail in Section 3.2. Results for the Northern Hemisphere winter stratosphere are presented in Section 3.3, and in Section 3.4 we discuss the processes by which the surface forcings are communicated to the stratosphere. Section 3.5 investigates the results from the combination of forcing in two or more surface bands, and conclusions are given in Section 3.6.

3.2 Model and Numerical Experiments

Our experiments are performed using a Chemistry Climate Model (CCM), the IGCM-FASTOC (Bourqui et al. (2005)). Its atmospheric general circulation model component is the University of Reading IGCM, a general circulation model built on the spectral dynamical core of Hoskins and Simmons (1975). The model physics are described in Forster and Shine (1999) and Rosier and Shine (2000), and include orography, a land-surface scheme, moist and dry convection, and a radiation scheme (Morcrette (1991)), and are designed to be computationally fast. Here the IGCM is run at horizontal resolution T31 and has 26 vertical levels, with a lid at 0.1 hPa; 13
levels are within the stratosphere. Rayleigh friction is applied in the top three model levels (1 hPa or above).

The IGCM is coupled to the computationally fast chemistry scheme FASTOC introduced in Bourqui et al. (2005) and Taylor and Bourqui (2005). The chemically active region for the FASTOC scheme is between the tropopause and 4 hPa, and the advected chemically active species are O$_3$, NO$_x$, N$_2$O$_5$ and HNO$_3$. The focus of our study is on idealized surface forcing in an otherwise undisturbed control climate, representing pre-ozone hole conditions. Advected chemical species are relaxed to climatological values outside of the chemically active region. Chemistry and dynamics in the model interact through the feedbacks between ozone concentration and radiation. A description of the T31 version of the IGCM-FASTOC, including validation of zonal-mean fields, is given in Winter and Bourqui (2010). Note that the computational efficiency of this model is a critical factor in the present study, as it allows us to perform simulations of sufficient length for our results to meet statistical significance.

In all the simulations presented here, the IGCM-FASTOC is run for 50 years in timeslice mode, i.e. the 50 years are not chronological in real time, but each year can be regarded as a single member of an ensemble, taking its initial conditions from the last timestep of another member. In the control case, global surface temperatures and specific humidities are prescribed using climatological monthly-mean output from a separate 100-year IGCM-FASTOC simulation in which the atmosphere interacted with a 25m mixed-layer ocean and the land surface scheme. The prescribed surface temperatures and specific humidities are linearly interpolated between monthly means to give daily values. Note that the daily surface temperatures thereby become less variable than when the land surface scheme operates. The consequence on the model climate of limiting daily variability and removing interannual variability is to
reduce the amount of wave driving in the polar vortex (Winter and Bourqui (2010); Olsen et al. (2007); Braesicke and Pyle (2004)). However, this is unavoidable here because of the nature of our experiments. In the perturbed simulations, we add a temperature anomaly of 2K to all gridpoints (land and sea) within latitudinal bands of 10° or 30° width, keeping surface temperatures outside the bands identical to the control values. The transform grid of the IGCM-FASTOC at T31 resolution has 24 gaussian latitudes per hemisphere. In order to best represent the full width of the 10-degree bands, it was necessary to use some latitudes as endpoints in more than one band (e.g. the latitude 50°N is included in both the N5 and the N6 band; see Table 1). Six of the 10-degree bands correspond to four gaussian latitudes in width and the others to three gaussian latitudes. The wide 30-degree bands do not have endpoints in common. To prevent step-like gradients in the surface temperature fields, a running mean over five points was applied to the perturbed surface temperature field (the running mean was reduced to three points for 10-degree bands spanning only three gaussian latitudes). This means that contiguous bands are not perfectly independent of one another, and the imposed smoothed temperature anomaly is not absolutely identical for all bands. It does not affect our results because the response above the boundary layer is insensitive to slight variations in width (this was tested in an additional run in which the N3 band was made wider by one gridpoint; the response was unchanged). However, the band pairs N1, N2 and N7, N8 have greater latitudinal overlap than the others, and this probably contributes to the similarities in their responses. The forcing value of 2K was chosen to be consistent with real-world examples of surface heating (see Section 3.1). Note that we also tested a 3K anomaly and found the response to scale linearly in amplitude. The surface specific humidity field was adjusted in the bands to be consistent with the higher
Table 3–1: Summary of numerical experiments. A surface temperature anomaly of 2K is imposed at all gridpoints of each latitude band. The meridional extent of the bands is given below. All experiments were run for 50 years in timeslice mode, and climatological means are presented.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Extent of 2K forcing</th>
<th>Gaussian latitudes</th>
</tr>
</thead>
<tbody>
<tr>
<td>control</td>
<td>no forcing</td>
<td>no forcing</td>
</tr>
<tr>
<td>NP</td>
<td>60°N – 90°N</td>
<td>61°N – 87°N</td>
</tr>
<tr>
<td>NM</td>
<td>30°N – 60°N</td>
<td>31°N – 57°N</td>
</tr>
<tr>
<td>EQ</td>
<td>equator – 30°N</td>
<td>2°N – 28°N</td>
</tr>
<tr>
<td>wide bands</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N1</td>
<td>equator – 10°N</td>
<td>2°N – 13°N</td>
</tr>
<tr>
<td>N2</td>
<td>10°N – 20°N</td>
<td>9°N – 20°N</td>
</tr>
<tr>
<td>N3</td>
<td>20°N – 30°N</td>
<td>20°N – 28°N</td>
</tr>
<tr>
<td>N4</td>
<td>30°N – 40°N</td>
<td>31°N – 39°N</td>
</tr>
<tr>
<td>N5</td>
<td>40°N – 50°N</td>
<td>39°N – 50°N</td>
</tr>
<tr>
<td>N6</td>
<td>50°N – 60°N</td>
<td>50°N – 61°N</td>
</tr>
<tr>
<td>N7</td>
<td>60°N – 70°N</td>
<td>61°N – 72°N</td>
</tr>
<tr>
<td>N8</td>
<td>70°N – 80°N</td>
<td>68°N – 79°N</td>
</tr>
<tr>
<td>N9</td>
<td>80°N – 90°N</td>
<td>79°N – 87°N</td>
</tr>
<tr>
<td>narrow bands</td>
<td></td>
<td></td>
</tr>
<tr>
<td>N1N4</td>
<td>eq. – 10°N, 30°N – 60°N</td>
<td>2°N – 13°N, 31°N – 39°N</td>
</tr>
<tr>
<td>EQNM</td>
<td>equator – 60°N</td>
<td>2°N – 57°N</td>
</tr>
<tr>
<td>EQNMP</td>
<td>equator – 90°N</td>
<td>2°N – 87°N</td>
</tr>
<tr>
<td>combination bands</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

temperatures, with the assumption that surface relative humidity does not change under climate warming conditions (Held and Soden (2006)).

The perturbed experiments are as follows: in a first series, the thermal surface forcing is applied in three wide bands: an equatorial band (EQ) from the equator to 30°N; a mid-latitude band (NM) from 30°N to 60°N; and a polar band (NP) from 60°N to 90°N. In a second series, the width of the bands is reduced to 10 degrees, so that there are nine bands in the Northern Hemisphere, from N1 (equator to 10°N) to N9 (80°N to 90°N). Finally, in a third series, we apply the surface forcing in a combination of bands: N1N4, forced from the equator to 10°N and from 30°N to 40°N; EQNM, forced continuously from the equator to 60°N, and EQNMP,
forced continuously from the equator to 90°N. The main focus of this paper is the results from latitudinal forcing in the Northern Hemisphere, although some results for the Southern Hemisphere are also presented in Section 3.3.3. Table 1 summarizes these experiments. The 50-year duration of individual experiments provided good statistical significance on most results, and therefore longer simulations were not made. Results are presented as differences between the 50-year climatological fields of the perturbation experiments and the control. Bilateral Student’s T-test statistics are used to determine the robustness of results, and for clarity on the graphics, we apply grey shading to those areas of the plot not significant at the 5% level.

3.3 Response in the Winter Stratosphere
3.3.1 Seasonal-mean Response in the Northern Hemisphere

We begin by considering the January-February climatology of the zonal-mean temperature and zonal wind in Northern Hemisphere, as shown on Figures 3–1, 3–2 and 3–3 by difference fields (perturbed band experiment - control). The temperature and wind anomalies of January-February are expected to be related to prior wave forcing anomalies: Newman et al. (2001) showed that temperatures in the polar vortex are correlated most strongly with wave flux entering the stratosphere between 40°N and 80°N a month earlier. We therefore show, in the right column of Figs. 3–1, 3–2 and 3–3, the December-January average difference in the Eliassen-Palm (EP) flux vectors and in its total divergence. To facilitate the graphical representation of these vectors, the EP flux components above 20 km are divided by density. Vector scaling is therefore different in the lower and upper portions of each panel, but is consistent across all graphs. A convergence (negative divergence) of the EP flux acts to decelerate the zonal mean flow, and geostrophic balance is maintained by an induced poleward circulation with a descending branch near the pole. The adiabatic
warming related to this descent, partly balanced by radiative relaxation, is in thermal wind balance with the deceleration in the polar vortex.

The top row in Figs. 3–1, 3–2 and 3–3 gives the results for the wide-band experiments (EQuatorial, Northern Midlatitude and North Polar on Figs. 3–1, 3–2 and 3–3 respectively). The lower rows give the results for the narrow 10-degree bands into which each wide band can be subdivided. A green or black line at the bottom of each panel indicates where the 2K thermal forcing was applied. Refer to Table 1 for a summary of all the experiments and the extent of the bands. In Fig. 3–1, the thermal forcing is applied within 30° latitude of the equator. The thermal anomaly extends upwards at the equator (EQ and N1 cases, left column, upper two rows) and follows isentropes upwards and polewards in the tropics (EQ, N2 and N3 experiments). To maintain thermal wind balance, the changes in the meridional temperature gradient are accompanied by changes in the tropospheric jet. When the thermal forcing is confined within 10° latitude of the equator (N1, second row), the tropospheric jet is enhanced. When the surface forcing extends away from the equator but remains in the tropics (EQ, N2 and N3 experiments), the tropospheric jet undergoes a poleward shift (Fig. 3–1, centre panels, rows 1, 3 and 4).

In the stratosphere, the effect of the thermal surface forcing in the equatorial and tropical latitudes in Fig. 3–1 is seen as a large warm anomaly at high latitudes, and a significant weakening of the vortex. This response is found for experiments EQ, N1 and N2, but is absent when the thermal forcing is applied between 20°N and 30°N (N3 case). While a weakening and slight poleward shift of the tropospheric jet is noted in the N3 case, there is no significant response in the stratosphere to the surface forcing.

The right column of Fig. 3–1 provides a key to the response in the stratospheric vortex winds. For experiments EQ, N1 and N2, there is increased EP flux into
Figure 3–1: January-February response in zonal-mean temperature ($T$, °C, left) and wind ($u$, m/s, centre), and December-January response in EP flux divergence and EP flux vectors ($\nabla \cdot \vec{F}$, m/s/day and $\vec{F}$, m$^2$/s/day, right) to forcing in the northern equatorial and tropical bands EQ, N1, N2 and N3. Contours of control values are overlaid. In the right column, the zonal wind for each experiment is overlaid (green contours), and the EP flux and its divergence above 20 km are divided by density. Vectors have the same scaling on all panels in Figures 1, 2, and 3. Grey shading covers all areas not significant at the 5% level for a bilateral T-test. A green or black line on the bottom of each panel shows the latitudes where surface forcing was applied. See Table 1 for experiment details, as well as explanations in the text.
the stratosphere in December-January, especially polewards of 60°N, and stronger convergence of the flux into the upper stratosphere. Since EP flux convergence acts as a force on the flow, increased convergence is accompanied by the deceleration of the vortex seen in the centre column of Fig. 3–1. The deceleration in turn allows greater propagation of waves into the core of the vortex, in accordance with the Charney-Drazin criterion. The zonal wind field for each experiment is overlaid (green contours) on the right panels of the figure. In the N3 experiment (bottom row of Fig. 3–1), the upward EP flux entering the stratosphere between 40°N and 80°N is slightly decreased, with weaker convergence into the upper stratosphere. The accompanying acceleration of the vortex, however, is not above the statistical significance criterion in this experiment. The stratospheric response to thermal surface forcing thus changes in character between the N2 and N3 cases. The link between tropospheric and stratospheric responses will be analyzed in Section 3.4.

Fig. 3–2 presents the same diagnostic fields as Fig. 3–1, but for the cases when thermal surface forcing was applied between 30°N and 60°N, either over this entire span (experiment NM, top row), or over 10-degree sections at a time (experiments N4, N5 and N6, in rows two, three and bottom, respectively). The effect throughout the mid- to high-latitude troposphere is to warm the extratropical lower troposphere and thereby decrease the meridional temperature gradient, with the attendant slowing of the tropospheric jet expected by thermal wind balance. When the surface thermal forcing is applied only north of 50°N, in the N6 case (Fig. 3–2, bottom row), the meridional temperature gradient is affected only in the lowest troposphere and the decrease in the jet is small and limited to its northern flank. The stratospheric response to the band-wise surface forcing is consistent in all panels of Fig. 3–2: there is a cold anomaly at high latitudes throughout the stratosphere, accompanied by an increase in vortex winds. The stratospheric response is of the same sign in all
Figure 3–2: As Fig. 3–1, but for the northern mid-latitude band experiments NM, N4, N5 and N6.

experiments, though weaker and not statistically significant in the N6 case. Applying the surface forcing to a narrow band of latitudes, such as the N4 and N5, yields a stronger response than applying the surface forcing to all latitudes from 30°N to 60°N (ie, the NM case). The changes in wave forcing are seen in the right-hand column of Fig. 3–2. In the NM, N4 and N6 cases, the December-January EP flux entering the stratosphere is decreased equatorwards of 60°N but increased polewards of 60°N,
while in the N5 case, the EP flux entering the stratosphere is decreased from 40°N to 80°N. In all cases, the December-January upward EP flux in the high-latitude middle stratosphere is reduced (shown by the downward arrows), leading to its decreased convergence in the upper stratosphere, and consequently to stronger vortex winds.

Finally, Fig. 3–3 presents the same diagnostics for the northern polar bands, covering the range 60°N to 90°N. In the troposphere, the thermal response follows
the pattern seen in the N6 experiment (Fig. 3–2, bottom row), and becomes limited to the surface in the N8 and N9 experiments. Therefore little or no weakening of the tropospheric jet is noted. The response in the stratosphere is not statistically significant in either the wide NP band experiment (thermal forcing is applied from 60°N to 90°N) or in the N7 experiment (forcing applied from 60°N to 70°N), but its sign in the latter is nevertheless consistent with that of the N8 and N9 responses (bottom two rows of Fig. 3–3). The EP flux entering the stratosphere shows a slight decrease equatorwards, and increase polewards, of 60°N in the NP and N7 cases, and a decrease everywhere polewards of 40°N in the N8 and N9 cases (Fig. 3–3, right column). In the NP case, the mid-stratospheric vertical EP flux is stronger on the poleward flank of the vortex, but the associated increase in flux convergence is very small and the weakening of the January-February vortex (Fig. 3–3, top centre) is not strong or statistically significant. In the N7, N8 and N9 cases, there is a strong reduction of EP flux into the vortex, particularly on the poleward side. The large cold anomaly evident in the high-latitude stratosphere in the N8 and N9 experiments, accompanied by a stronger vortex, is consistent with the reduced vertical component of the EP flux in these experiments and the associated reduced convergence of EP flux in the upper stratosphere.

On Fig. 3–2, the mid-latitude 10-degree experiments gave results whose average would resemble the results of the 30-degree NM experiment. This is not the case on either of Figs. 3–1 or 3–3. On Fig. 3–1, the proximity of the surface forcing to the equator determines the character of the response. On Fig. 3–3, the three 10-degree experiments each present a response to the applied forcing that is consistent with the others, but the NP experiment (Fig. 3–3, top row) essentially shows no response to the surface forcing, aside from a slight weakening of the winds at the northern flank of the tropospheric jet. From these results, it is clear that the latitudinal width of
thermal forcing at the surface, in addition to the location of the forcing itself, plays a role in determining the stratospheric response in high latitudes.

Wave forcing in the vortex drives the Brewer-Dobson Circulation (BDC), in which ozone is transported from low to high latitudes in winter. We find the BDC to be significantly enhanced in the lower stratosphere in mid- to high latitudes in the EQ, N1 and N2 experiments, and weakened throughout the stratosphere from low to mid-latitudes in the NM, N4 and N5 experiments (not shown). A slight but statistically significant weakening is also noted in the high-latitude upper stratosphere in the N8 and N9 cases. By affecting the strength of the BDC, changes in the wave forcing impact the concentration of ozone at the poles directly through changes in the transport of ozone, as well as indirectly through dynamically induced temperature changes that alter the reaction rates of ozone chemistry. Changes in ozone concentration feed back on local temperature and dynamics through changes in radiative heating or cooling. It is therefore of interest to consider the response in the ozone concentration to the thermal surface forcing in different bands. This is presented in Figure 3–4. (Note that the flat bottom of the anomalies is due to the chemistry scheme being active only in the stratosphere.) We expect high-latitude ozone concentration to be higher when the wave forcing response and BDC are stronger, and this is the case for experiments EQ, N1, N2 and N3 (Fig. 3–4, left column). The low-latitude decrease in these experiments suggests a more vigorous upwelling branch of the BDC there, which is also consistent with the lower temperatures seen in the tropical upper troposphere on Fig. 3–1, left column. In cases EQ, N1 and N2, the strongest increase in ozone concentration occurs polewards of the climatological maximum. In experiments NM, NP and N4 through N9, decreased wave forcing of the vortex is seen as a decrease in ozone concentration near the climatological
Figure 3–4: January-February average response in the Northern Hemisphere concentration of O₃ (number density; $10^{17}\text{ m}^{-3}$). Contours from the control simulation are overlaid. Grey shading covers all areas not significant at the 5% level and green lines indicate the latitudes of the imposed surface forcing.

maximum. In all cases, the response in the high-latitude ozone concentration is statistically robust.
3.3.2 Seasonal Cycle Response in the Northern Hemisphere

In the preceding analysis (Figs. 3–1, 3–2 and 3–3), the changes in the mean flow were discussed for the period January-February because it corresponds to the annual maximum in the wave forcing near the top of the stratospheric polar vortex, and therefore to the period in which the mean flow is most influenced by wave drag. However, as mentioned earlier, the state of the winds in January and February is the result of prior wave forcing beginning in the fall. Early-winter wave breaking, by affecting the vertical profile of zonal winds, influences the amount and level of breaking in subsequent months, a “pre-conditioning” of the vortex that is also invoked to explain why the polar vortex in some cases is more susceptible to destruction by Sudden Stratospheric Warming events (for instance, Butchart et al. (1982); Charlton and Polvani (2007)).

To provide more insight into the seasonal evolution of the changes in the mean flow and wave forcing, seasonal cycles of the same fields as in Figs. 3–1, 3–2 and 3–3 are shown in Figures 3–5, 3–6 and 3–7. On these height-time plots, we show the area-averaged response in the high latitudes, in order to capture the extent of the polar vortex. The two contrasting responses to the imposed forcing discussed for Figs. 3–1, 3–2 and 3–3 in fact persist through the seasonal cycle: when the forcing is between the equator and 20°N (cases N1 and N2), or continuously from the equator to 30°N (case EQ), the polar stratosphere warms. When the forcing is exclusively north of 30°N (cases NM, NP and N4 through N9), the polar stratosphere cools.

However, the seasonal cycles show interesting variations from case to case. The warming in cases EQ, N1 and N2 is seen throughout the winter from November through April, with a peak in November-December. Consistently, anomalously large wave forcing is also seen early in the winter (November and December). Subsequent wave forcing can be reduced or no longer statistically significant, yet the vortex
Figure 3–5: Area-weighted zonal-mean winter (October through April) response in temperature ($T$, °C, left), wind ($u$, m/s, centre), and EP flux divergence ($\nabla \cdot F$, m/s/day, right) to thermal surface forcing in the NH equatorial and tropical bands EQ, N1, N2 and N3. Area-averaging is done over 60°N to 90°N for the temperature and over 50°N to 80°N for wind and wave forcing. Contours are control values and grey shading covers all areas not significant at the 5% level.

is weakened for the remainder of the winter. When the forcing is applied in the mid-latitude NM, N4 or N5 bands, the polar stratosphere is colder from November onwards, with a minimum in February. The largest decrease in wave driving of the vortex occurs in January and February. These are normally the months of maximum wave drag, and if the drag is anomalously weak at this time, the vortex remains stronger throughout the winter and spring. The N3 case, with surface forcing imposed from 20°N to 30°N, serves as a transition between these two opposite responses, with a warm anomaly only from October through December (associated with an early and weak pulse of anomalous wave activity), and cold anomalies in the winter months thereafter. Finally, when the surface forcing is applied only polewards of 60°N, the
same reduction of January-February wave forcing as for the mid-latitude bands is seen, but is of shorter duration. In a sense, there is another transition marked by the N6 and N7 cases, as the surface forcing is applied to either side of 60°N: the response goes from a cooling throughout the stratosphere continuously from the fall into the spring to a cooling localized in the mid-stratosphere mainly in January and February. Furthermore, once the forcing is applied polewards of 60°N, there is no longer any statistically significant response in the troposphere. The NP case does not exhibit any response resembling the N8 and N9 cases. Note that the response in the mean flow in January-February gives a good representation of the overall response in all experiments. Similarly, December-January is a good representation of the overall change in the wave forcing. This justifies the choices made in Figs. 3–1, 3–2 and 3–3.

Figure 3–6: As Fig. 3–5, but for the northern mid-latitude band experiments NM, N4, N5 and N6.
3.3.3 Seasonal-mean Response in the Southern Hemisphere

It is beyond the scope of this paper to investigate in detail the Southern Hemisphere’s winter stratospheric response to band-wise forcing imposed in the Southern Hemisphere. In the absence of strong land-sea contrasts and the stationary Rossby waves they engender, we do not expect an imposed zonally uniform warm surface anomaly to have as large an impact on the polar vortex as in the Northern Hemisphere. Nevertheless, we ran a complementary series of 10-degree latitude band experiments in the Southern Hemisphere, analogous to the cases N1 to N9 presented in this study. The response in the austral winter zonal-mean wind field (Figure 3–8) differs from the Northern Hemisphere experiments in two significant ways: 1) All cases exhibit the same response in the stratosphere, i.e. there is no transition zone; and 2) the zonal winds essentially weaken along the entire axis of maximum
Figure 3–8: July-August average response in the zonal-mean wind $u$ (m/s) for the Southern Hemisphere band experiments. Note that the scale on the EQ panel (upper left) is twice as big as on the others. Contours of control values are overlaid and grey shading covers all areas not significant at the 5% level. A green line at the bottom of each panel indicates where the surface forcing was imposed.

windspeed, from the troposphere to the stratosphere, with strong positive anomalies occurring on either side (with some variation by band). This suggests an annular mode-like response in the winds rather than a change in the strength of wave forcing of the vortex.
3.4 From the Troposphere to the Stratosphere

Having established, in Section 3.3, that the polar vortex is anomalously disrupted by wave activity when there is thermal surface forcing equatorwards of 30°N, and anomalously strong when the surface forcing is polewards of 30°N, we now seek the origins of these responses in the troposphere. In this section, we limit our analysis to three of the experiments: EQ, whose response fields are similar to the low-latitude cases N1 and N2; NM, which is characteristic of the mid-latitude cases N4, N5, and N6; and N8, whose response fields are similar to the high-latitude cases N7 and N9, and are sufficiently different, in the troposphere, from the NM responses to merit a separate category (the wide high-latitude band experiment NP did not give a statistically significant response.)

In Cartesian coordinates under quasi-geostrophic scaling, the horizontal and vertical components of the EP flux are defined, respectively, as (Holton (2004)):

\[
F_y \equiv -\rho_0 \overline{u'v'}, \quad F_z \equiv \frac{\rho_0 f_0}{N^2 H} \overline{v'T'}
\]  

(3.1)

\(R\) is the dry air gas constant, \(\rho_0(z)\) is the reference density profile, \(N^2\) is the squared Brunt-Väisälä frequency and \(H\) is the scale height. The left column on Fig. 3–9 presents the zonal-mean eddy heat flux \(\overline{v'T'}\) at 100 hPa for the Northern Hemisphere. Note that this quantity is, through Eq. 1, proportional to the vertical component of the EP flux entering the stratosphere. Green horizontal lines on Fig. 3–9 demarcate the region of imposed thermal forcing at the surface. The control eddy heat flux is greatest between 40°N and 80°N (see the black contours for the control simulation on Fig. 3–9) and its average within this sector was found to be strongly correlated with temperatures in the polar cap (60°N to 90°N) at 50 hPa a month later by Newman et al. (2001). As discussed in Section 3.3.1 (Figs. 3–1, 3–2 and 3–3), waves entering the stratosphere north of 40°N propagate upwards and polewards,
Figure 3–9: Seasonal cycle of the response in zonal-mean eddy heat flux $\overline{\upsilon^\prime T^\prime}$ (K m/s) at 100 hPa (left), tropospheric November-December average response in the eddy heat flux (second column) and in its wavenumber 1 and 2 components (two right columns). Experiments EQ (top row), NM (middle row), and N8 (bottom row) are shown. In all panels, contours of the control values are overlaid; contour spacing is 4 Kms$^{-1}$, 4 Kms$^{-1}$, 2 Kms$^{-1}$, and 1 Kms$^{-1}$ from the left column to the right. Grey shading in the left column covers all areas not significant at the 5% level. Horizontal green lines on all panels indicate the latitudinal extent of the imposed surface forcing.

and thereby affect the stratospheric polar temperature by breaking in the vortex.

Comparing the temporal evolution of $\overline{\upsilon^\prime T^\prime}$ anomalies at 100 hPa in Fig. 3–9 (left column) with the seasonal cycle in EP flux divergence anomalies on Figs. 3–5, 3–6 and 3–7 confirms that $\overline{\upsilon^\prime T^\prime}$ at 100 hPa is a good proxy for stratospheric forcing. For the EQ experiment (Fig. 3–9, top left), the eddy heat flux entering the stratosphere is increased from October to February north of 50°N, with a maximum positive anomaly in December-January. The January-February temperature response field (Fig. 3–1, top left) accordingly exhibits a strong warm anomaly. On the middle left
panel of Fig. 3–9, the NM results show the opposite, i.e. a reduction north of 40°N in November and January, followed by a spring negative anomaly in March-April. The reduced flux in January leads to the negative temperature anomaly seen in the January-February average on Fig. 3–2 (top left). Similarly, the reduction in $\overline{v'T'}$ polewards of 40°N in the N8 case (Fig. 3–9, bottom left) from November into January leads to the cold anomaly in January-February seen in the area of the polar vortex in Fig. 3–3 (left column, second from bottom). In all three cases, the January-February average response in the vortex discussed in Section 3.3 is set up at the beginning of winter. A change of opposite sign is generally seen in the eddy heat flux south of 40°N, but these waves propagate preferentially towards the equator, and do not affect the polar vortex in the stratosphere.

The variations in wave flux entering the stratosphere seen in our experiments can be due to changes in tropospheric wave sources, redirection of vertically propagating waves below 100 hPa, or a combination thereof. Low wavenumbers are favoured for propagation into the winter stratosphere in general (e.g. Charney and Drazin (1961)), and stationary Rossby waves in particular play an important role in wave forcing at high latitudes (McLandress and Shepherd (2009)). The sources for such waves are primarily the land-sea thermal contrasts and mountain ranges of the Northern Hemisphere. To investigate changes in the stationary waves, we show, in columns two to four of Fig. 3–9, the eddy heat flux and its zonal-wavenumber 1 and 2 (denoted s1 and s2) components from the surface to 16 km (about 100 hPa). These are averaged over November and December to capture the early-winter response in the troposphere. (As stated in Section 3.2, all graphs in this paper show the climatological mean response over the 50 timeslice years of each experiment, therefore eddies can only be stationary.)
The control values of the November-December eddy heat flux (black contours on the second column of Fig. 3–9, spaced every 4 Kms\(^{-1}\)) indicate maxima in the troposphere (centred at around 45°N and 4 km) and the stratosphere (centred at around 60°N), and a mid-latitude minimum at around 12 km. In the EQ case, both maxima are strengthened and extended upwards and northwards in the troposphere and stratosphere, respectively. The minimum is weakened and extended upwards. In the NM experiment (second column, middle row), the eddy heat flux maxima are weakened, particularly on the equatorward side of the tropospheric maximum, while the minimum at 12 km becomes slightly stronger. It indicates that the centres of greatest poleward transport of heat in stationary eddies are all attenuated as a result of surface forcing in the NM band, contrary to the response to EQ band forcing. In the N8 case (Fig. 3–9, second on bottom), both eddy heat flux maxima are again decreased, though only weakly in the troposphere. However, in contrast to case NM, the minimum at 12 km becomes more negative. The stationary thermal eddies shown on Fig. 3–9 are one mechanism by which heat is transported from low to high latitudes in a balanced climate. A thermal surface anomaly imposed in low latitudes (as in the EQ case) requires a greater poleward transport of heat in order for the climate to remain in equilibrium and, conversely, when a thermal forcing is imposed in higher latitudes, as in the NM and N8 cases, the poleward transport of heat must decrease. It is seen on Fig. 3–9 that the response to the surface forcing in the stationary eddy field is consistent with this energy balance requirement.

Turning to the s1 component of the eddy heat flux shown in the third column of Fig. 3–9, the control values (black contours, spaced every 2 Kms\(^{-1}\)) show a maximum around 60°N, which gradually increases from the surface to the lower stratosphere. This is the latitude of the greatest zonal variability of mean winter temperature,
stemming from the stationary eddies (*Peixoto and Oort* (1992)). In the EQ experiment (third column, top row), an increase in the \( s_1 \) component of the eddy heat flux is seen continuously from the surface to the lower stratosphere following two upward branches: one around 45\(^\circ\)N, corresponding to an equatorward expansion throughout the troposphere of the region of the maximum, and the other branch around 65\(^\circ\)N, corresponding to an upward expansion of the surface maximum there and a poleward strengthening of the upper-level maximum. Note that the 10-degree bands N1, N2 and N3, which occupy the same latitudes as the EQ band, also exhibit a bottom-to-top increase in the \( s_1 \) component of the eddy heat flux (not shown). In the N1 case, however, the equatorward expansion of the 45\(^\circ\)N branch seen in the EQ case is also present from the surface upwards, while no accompanying increase in the surface maximum at 65\(^\circ\)N occurs. The reverse is true for the N2 and N3 cases: each of these exhibits a stronger and higher surface maximum at 65\(^\circ\)N as well as a strengthening at the poleward edge of the maximum near the tropopause, but no equatorward expansion at any level. In the NM case, the early winter decrease in total eddy heat flux at 100 hPa (middle row, left-most panel) can be connected to a decrease in the \( s_1 \) component (second panel) throughout most of the troposphere. Here we note a weakening along the equatorward edge of the control high-\( s_1 \) zone from the near-surface upwards and northwards. A similar response is found for the ten-degree bands N4 through N9 (only N8 is shown, Fig. 3–9, bottom row), but not for the NP case (not shown), in which there is no response.

The stationary \( s_2 \) control values (black contours on the third column of Fig. 3–9, spaced every 1 Kms\(^{-1}\)) show two maxima around 50\(^\circ\)N to 60\(^\circ\)N, in the lower stratosphere and near the surface. At the surface the maximum corresponds roughly to the zonal-mean signature of the storm tracks. In all experiments the response in the \( s_2 \) component of the eddy heat flux in November-December is considerably weaker
than the s1 response, and is often not of the same sign at the top and the bottom of the troposphere. For the bands having surface forcing near the equator (EQ, N1 and N2), the s2 response is positive at the top and bottom of the troposphere, and thus reinforces the s1 response. In most other cases the s2 response near the tropopause is of opposite sign to the s1 response.

Comparing the amplitude of the control values of $v'$ on these last three columns of Fig. 3–9 (keeping in mind that the contour spacing changes) confirms that the vertically propagating wave flux entering the stratosphere at 60°N (about 12 Kms$^{-1}$) is almost entirely composed of the s1 and s2 components (8 Kms$^{-1}$ and 3 Kms$^{-1}$, respectively), although these components account for only a fraction of the tropospheric maximum eddy heat flux. Yet an overall picture that emerges from considering the stationary s1 and s2 components in the troposphere is that the response found in the early part of the seasonal cycle in the total flux at 100 hPa (left-hand column) at least partly originates from the near-surface in the s1 component averaged over November and December. The change in the s2 component is secondary. This suggests that the application of thermal surface forcing in the sensitive latitude bands south (north) of 30°N leads to an increase (a decrease) in the tropospheric stationary eddies at higher latitudes, which then propagate from the near-surface upwards to the stratosphere. Poleward transport of heat via the stationary synoptic-scale eddies can be represented by changes in the baroclinicity, as quantified by the growth rate of the maximum Eady mode. Plots of the climatological Eady growth rate calculated at about 5 km (Fig. 3–10, top two rows) are consistent with the second column of Fig. 3–9: in the EQ case, baroclinicity increases at around 40°N, where a strong positive anomaly occurs in the eddy heat flux, and the reverse takes place in the NM case (decreased baroclinicity where there is a negative heat flux anomaly). This is expected, since the tropospheric thermal eddy maximum is dominated by
Figure 3–10: Response in the maximum Eady growth rate at 500 hPa (day$^{-1}$) for experiments (top two rows: EQ, NM and N8 and (bottom two rows: N1N4, EQNM and EQNMP. The stereographic views show the averaged November-December response from 30°N to 90°N, marked every 15°. Longitudes are marked every 30°.

contributions from higher wavenumbers, as indicated above. In the N8 case, there is no evident signal in the Eady growth rate, which may be due to limitations of this
diagnostic at the high latitudes where the N8 surface forcing is imposed. However, it is clear from comparing the three left panels on the bottom of Fig. 3–9 that the s1 and s2 components of the eddy heat flux response make a larger contribution in the troposphere in this experiment than in the EQ and NM cases. The response in baroclinicity reflects the response in both stationary and transient synoptic-scale eddies. Both have the potential to affect the climatological storm track response, and thereby to affect the stationary eddy heat flux on a planetary scale, for instance by modifying the strength and location of the storm track exit regions either in the Pacific alone (leading to an s1 component in the response) or in both the Pacific and the Atlantic (s2). In this manner synoptic eddies can lead to planetary eddies, which are able to propagate to the stratosphere. Elucidating the processes that cause these changes in the lower troposphere is, however, beyond the scope of this paper. Another potential link between anomalies in synoptic and planetary eddies is the non-linear interaction between shorter waves that can result in transfer of energy to lower wavenumbers. It was shown by Scinocca and Haynes (1998) that synoptic-scale waves can become re-organized into packets having envelopes of zonal-wavenumber 2 (the absence of s1 waves in that paper was attributed to the absence of topography in the model used). These packets can propagate to the stratosphere (Scinocca and Haynes (1998)).

It is clear from the results presented thus far that there is a transition between responses of opposite sign when the surface forcing is moved from south to north of 30°N. An argument can also be made for a transition zone near 60°N: the NP, N6 and N7 experiments show little or no response in the averaged January-February stratospheric circulation (Fig. 3–3). The low sensitivity to surface forcing in these experiments may stem from a more variable seasonal cycle in the 100 hPa eddy heat flux response polewards of 40°N: in experiments NM and N4 through N9, this
seasonal cycle begins with a negative anomaly in the late fall (only cases NM and N8 are shown, on Fig. 3–9). In most experiments, the negative anomaly persists for the entire winter. In cases N6 and N7 alone (not shown), there is a brief period of positive 100 hPa eddy heat flux anomalies in high latitudes in January. This brief interruption is not statistically significant, but may sufficiently undermine the effect of the negative early-winter wave flux to give the weak and statistically insignificant responses in the vortex seen in these two experiments.

3.5 Combined Bands

Given the opposite responses to the surface forcing in the N1 and N4, and in the EQ and NM, experiments, we conducted three further experiments with a surface thermal forcing applied simultaneously in two or more latitude bands: case N1N4, with surface forcing from the equator to 10°N and from 30°N to 40°N; case EQNM, with surface forcing continuously from the equator to 60°N; and EQNMP, with the forcing applied continuously from the equator to the north pole, covering the entire Northern Hemisphere. Responses to the forcing in these additional experiments are presented in Figure 3–11 (zonal-mean temperature, zonal wind, EP flux and its divergence, analogously to Figs. 3–1, 3–2 and 3–3) and Figure 3–12 (eddy heat flux), analogously to Fig. 3–9).

3.5.1 Experiment N1N4

The stratospheric response in the N1N4 experiment is similar to, but slightly weaker than, the N1 response: a warm anomaly is found in the vortex at the same location as – but about 1°C weaker than – in case N1, and correspondingly the vortex wind response is negative (compare Fig. 3–11, bottom row and Fig. 3–1, second row, but note that the colour scales are different). The tropospheric response again combines elements from both the N1 and the N4 cases. The troposphere is
warmer throughout all latitudes and the jet is stronger, but not as much as in the N1 case. The December-January (average) increased EP flux convergence in the upper stratosphere in the N1N4 experiment occupies a smaller area than in the N1 case, and is due almost entirely to the vertical component of the EP flux. The anomalous northward flux in the lower stratosphere seen in the N1 case is almost absent in the N1N4 response (compare the second row right panels of Figs. 3–1 and 3–11).

The seasonal response in the zonal-mean eddy heat flux at 100 hPa is positive polewards of 50°N, but significant only from late December to early February. There is no significant early winter response when surface forcing is in both the N1 and N4 bands. Note that, forced individually, the N1 and N4 early winter (i.e. November-December) eddy heat flux response (not shown) is, respectively, positive as for EQ and negative as for NM. It is possible that cancellation of these early winter effects leads to the lack of a response in the combined-band N1N4 experiment (Fig. 3–12, top row left). The response in November-December average of the eddy heat flux (Fig. 3–12, second column) resembles that of the EQ case (Fig. 3–9) in the troposphere, with a stronger and expanded maximum and a weakened minimum. However, the response in the lower stratospheric maximum at 60°N is weak (and statistically insignificant, cf. top left) in these early winter months. A continuous response from the top of the troposphere to the near-surface in the s1 component of the eddy heat flux can nonetheless be observed in the November-December average (top row, third panel), though it is partly offset by the s2 response (top row right).

3.5.2 Experiment EQNM

The EQNM response in the stratosphere appears to be an enhanced version of the EQ response, with a stronger high-latitude, lower stratospheric warm anomaly accompanied by a decrease in vortex strength (note that the scale for the zonal-mean
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Figure 3–11: January-February average response in zonal-mean temperature ($T$, °C, left) and wind ($u$, m/s, centre), and December-January average response in the EP flux divergence and EP flux vectors ($\vec{\nabla} \cdot \vec{F}$, m/s/day and $\vec{F}$, m$^2$/s/day, right) to forcing in the combined-band experiments N1N4 (top), EQNM (centre) and EQNMP (bottom). Contours of control values are overlaid. In the right column, the zonal wind for each experiment is overlaid (green), and the EP flux and its divergence above 20 km are divided by density. Note that the colourscale on the temperature and wind plots is larger than on Figs. 1, 2, or 3 while EP vectors are scaled the same. Grey shading covers all areas not significant at the 5% level. Green or black lines at the bottom of the plots indicate the latitudes where surface forcing was applied. See Table 1 for experiment details, as well as explanations in the text.

The tropospheric response patterns retain characteristics from both the EQ and the NM experiments. The greatest temperature anomaly is located around 30°N as in the EQ
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Figure 3–12: As Fig. 3–9, but for the combined-band experiments N1N4 (top), EQNM (centre) and EQNMP (bottom).

case (Fig. 3–1), but with a stronger warming throughout the lower troposphere which extends to the high latitudes, as in the NM case (Fig.3–2). Consequently, the tropospheric jet undergoes a weaker poleward shift than in the EQ experiment. The eddy heat flux response in the EQNM experiment resembles that of the EQ experiment but begins about a month later, with a maximum response in February. Consequently, the vertical profile of the eddy heat flux averaged over November-December in the EQNM experiment (middle row, second) is weaker than its EQ counterpart while largely exhibiting the same features. A notable difference is the weak positive response throughout the lower troposphere polewards of around 50°N; in the EQ case (Fig.3–9), the response in this region was weakly negative. Decomposed by wavelength, the s1 component in the November-December average is somewhat weaker
while the $s_2$ response is stronger. Both responses are continuous from the surface to the stratosphere.

### 3.5.3 Experiment EQNMP

When surface forcing is applied continuously from the equator to the north pole (experiment EQNMP, Fig. 3–12, bottom row), the response in the 100 hPa eddy heat flux retains the character of the EQ and EQNM cases, with a strong December increase around $70^\circ$N preceded by a small November increase at $60^\circ$N. The November-December average in the eddy heat flux response (total as well as the $s_1$ and $s_2$ components) strongly resembles that of the EQ and EQNM cases. It was noted in Section 3.4 that the surface heating imposed only in low latitudes, as in the EQ case, required poleward heat transport in order to maintain a balanced climate, and that the stationary eddies provide an important contribution to this transport. In the EQNMP experiment, the surface thermal forcing is applied at all gridpoints of the Northern Hemisphere, yet the eddy heat flux response is nearly identical to the EQ case. Because the IGCM-FASTOC model used in these experiments includes comprehensive physics and topography, a uniformly imposed surface thermal anomaly does not necessarily lead to a uniform near-surface temperature increase. As described in Section 3.3, surface heating applied only in the high latitude bands remains confined to the surface, while heating applied in low latitudes leads to convective heating throughout the tropical troposphere. The resulting changes in the meridional temperature gradient in the middle and upper troposphere therefore require additional poleward heat transport even when surface heating is also imposed at higher latitudes. The EQNMP results demonstrate that the response to heating imposed in the tropics dominates the response when surface heating is also imposed at higher latitudes.
3.6 Conclusions

Using a fully three-dimensional Climate Chemistry Model, we investigated the stratospheric response in the Northern Hemisphere winter to latitudinal variations in thermal surface forcing by imposing 2K temperature anomalies in a series of zonal bands of 10° and 30° latitude width. Our work provides a link between studies imposing idealized forcings on an idealized basic state (e.g. dry or moist aquaplanets, as in Chen et al. (2010); Chen and Zurita-Gotor (2008); Brayshaw et al. (2008) and Ring and Plumb (2007)), and those considering realistic, often localized, surface forcings in realistic GCMs (e.g. El Niño, as in Brönniman et al. (2004); Fischer et al. (2008); Lu et al. (2008); or snow anomalies, as in Cohen et al. (2007); Fletcher et al. (2009)).

We note two fundamental responses to our surface forcing. First, when the forcing is applied continuously between 30°N and 60°N, or within 10° latitude windows centred at 35°N, 45°N, 75°N or 85°N, the wave activity entering the winter stratosphere at 100 hPa polewards of 40°N is reduced, and so is the wave breaking in the vortex. The anomalously weak upward flux at 100 hPa begins in the early winter, allowing an anomalously strong vortex to form. The reduction in wave forcing higher in the stratosphere occurs after the vortex has reached its annual peak, so that it persists at full strength for a longer time. The associated response in zonal-mean temperature is a strong cooling in the high-latitude stratosphere. Responses of the same sign, but strongly reduced and not statistically significant, are found when the surface forcing is centred at 55°N and 65°N or imposed continuously from 60°N to 90°N. Second, when surface forcing is applied equatorwards of 30°N, the response is of opposite sign. There is a greater upward flux of waves into the stratosphere polewards of 40°N from the fall into the winter, leading to anomalously strong vortex disruption in January and February. An anomalously warm high-latitude lower to
mid-stratosphere accompanies the weakened vortex. The increased wave-forcing response occurs before the vortex is fully established, and has the potential to weaken the vortex for the entire winter, even when the wave forcing anomaly subsequently decreases. The upward flux of waves from the troposphere into the stratosphere is represented by the eddy heat flux at 100 hPa, whose response at that level can be traced to the near-surface response of its zonal wavenumber-1 component. This is an indication that the thermal surface forcings we apply lead to anomalous generation of stationary Rossby waves in the lower troposphere, which propagate into the stratosphere and lead to changes in vortex strength. Sources of these waves include changes in the land-sea thermal contrasts and changes in the strength and location of the climatological storm tracks and their exit regions, the latter mechanism linking responses in the baroclinicity (and thus in synoptic-scale waves) and responses in planetary waves. In addition to this direct response of planetary waves to the applied surface forcing, a contribution involving planetary wave formation from the non-linear interaction among synoptic-scale waves cannot be ruled out.

In the “fully idealized” experiments of Chen et al. (2010); Chen and Zurita-Gotor (2008); Brayshaw et al. (2008) and Ring and Plumb (2007) discussed in Section 3.1, the position of the imposed surface or tropospheric forcing, or the location of the strongest gradient in the forcing relative to the position of the tropospheric jet, are of key importance in determining the strength and sign of the response. Chen and Zurita-Gotor (2008) identify 35°N as the latitude where the response changes sign. In our 10-degree band experiments, a transition between responses of opposite sign takes place at the N3 band (20°N to 30°N): there is an increased wave forcing response in the fall and early winter, as in cases EQ, N1 and N2, followed by a decreased wave forcing response through the winter, as in cases NM and N4 through N9. In the 30-degree band experiments, a transition occurs between the EQ and NM
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bands, in which the surface forcing is applied south and north of 30°N, respectively. However, when the forcing extends continuously from the equator to a latitude north of 30°N, as in the EQNM and EQNMP experiments, the stratospheric response is similar to cases in which only tropical latitudes are forced. It was pointed out in Olsen et al. (2007) that the wave driving of the polar vortex is highly correlated with the zonal-mean difference between observed SSTs in latitudes 30°N to 40°N and those in latitudes 0° to 10°N. Our N1 experiment increases the difference between the surface temperatures in these two regions, and, consistently with the results of Olsen et al. (2007), greater wave forcing in the vortex ensues. However, when surface heating is applied to both regions as in both our N1N4 and EQNM experiments, the wave driving of the vortex nonetheless remains similar to the N1 experiment. This highlights the non-linear relationship between the thermal surface forcing in both of these bands and the stratospheric response.

The surface forcing we apply in this study – a 2K warm anomaly over all longitudes in a range of latitudes – is different from natural surface forcings such as El Niño, in which the surface forcing is zonally varying. However, climatological distributions of wave sources in our model are also zonally varying and govern the longitudinal distribution of the response, so that imposed forcing in a complete latitude band can have a similar effect on the stratospheric circulation as a natural, localized surface forcing such as El Niño. It has been shown that El Niño leads to more wave forcing of the polar vortex (Sassi et al. (2004); Taguchi and Hartmann (2006), who focus on vortex variability; or Fischer et al. (2008), who compare ENSO-related increases in the Brewer Dobson Circulation in different models, including the IGCM-FASTOC used in the present study). At the surface, our EQ, N1, and N2 experiments, like El Niño, produce a classic positive PNA pattern in the mean sea level pressure (not shown). The relationship between changes in the PNA pattern
induced by both phases of the ENSO and circulation in the stratosphere is discussed in Sassi et al. (2004).

Unlike this similarity between imposed low latitude heating and El Niño, surface forcing in mid- or high latitudes alone does not have a counterpart in a realistic climate. In climate change experiments, these latitudes are shown to warm the most, but the surface heating is nonetheless global, and expected to range from 1-2°C at the equator to to 4-6°C in high latitudes (see for instance IPCC (2007), their Fig. 10-8). The 2K surface thermal forcing applied at low latitudes in this paper is therefore too strong to invite direct comparison to climate-change experiments, while our applied high-latitude forcing is too weak. Imposing a surface thermal forcing that respects the temperature gradient expected from greenhouse-gas induced climate change might attenuate the dominance of the equatorial response that we find in the combined-band experiments N1N4, EQNM and EQNMP, although qualitatively the stratospheric response in these experiments resembles what is found in CO$_2$-doubling experiments (Winter and Bourqui (2010); McLandress and Shepherd (2009)). More important, the cause of the surface temperature increase in climate-change simulations is the higher atmospheric concentration of greenhouse gases, which leads to air temperature increases throughout the troposphere and to a warming of the surface. The stratospheric circulation responds to the combination of in-situ radiative cooling, tropospheric radiative heating, and the increased surface temperatures. These effects are not fully separable, although one may dominate the others in specific regions or seasons (Sigmond et al. (2004)). The experiments presented here are concerned only with temperature forcing at the surface.

Finally, the results we present are of course subject to the model’s limitations. In particular, ocean points north of 70°N are sufficiently cold as to be ice-covered during
the winter months, with or without our imposed 2K thermal anomaly. The IGCM-
FASTOC does not allow for thinning of the ice or creation of open “leads”, therefore
ice albedo is unaffected by the imposed surface forcing in the N8 and N9 experiments,
and fluxes of surface heat and humidity respond only to the imposed temperature
anomaly without causing ice albedo feedbacks. At lower latitudes, the simple soil
scheme of the IGCM-FASTOC may not respond adequately to the imposed surface
heating in regions that are very dry or very wet. In CO$_2$-doubling experiments, for
instance, we have found that surface warming in the IGCM-FASTOC matches the
multi-model results from *IPCC* (2007) very well over the ocean, but is weaker (by
1-2K) over land in the Northern Hemisphere mid- and high latitudes (see Winter and
Bourqui (2010)). This has the potential to reduce land-sea temperature contrasts,
and thus to decrease an important source of planetary waves. The implication for
the experiments presented here is that our results likely represent a lower limit on
the response in wave activity. Wave activity is also inhibited because surface temper-
atures at all gridpoints are prescribed monthly means, interpolated linearly to give
daily values, and thus have neither daily nor interannual variability. Consequences
of suppressing this variability are discussed in Winter and Bourqui (2011b). In the
stratosphere, an important limitation of the model is its use of Rayleigh friction at
the top three model levels (at or above 1 hPa). It is known that Rayleigh friction can
induce an artificial counter-circulation between the level of the deposition of resolved
waves and the model lid, which may affect the modelled circulation below (*Shepherd
et al. (1996)*). However, this contamination of the circulation is strongest at the
model top, and falls to about 10% of its maximum value by about two scale heights
beneath the applied drag (*Shepherd et al. (1996)*). The middle stratosphere where
we see the strongest response in zonal-mean temperature is located about three scale
heights beneath the lowest level in which Rayleigh friction is applied. The generation
and vertical propagation of planetary waves in the troposphere, on which we focus in Sections 3.4 and 3.5, are not affected by Rayleigh friction at the model top. Finally, the model’s spectral truncation at wavenumber 31 limits its capacity to capture all synoptic-scale baroclinic eddies.

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CHAPTER 4

The Impact of Surface Temperature Variability on the Climate Change Response in the Northern Hemisphere Polar Vortex

The climate change response in the stratosphere depends strongly on the basic state of the control climate. Depending on the choice of surface temperature representation as the model’s lower boundary condition, different timescales of variability are present: when surface temperatures are calculated interactively at all gridpoints, all timescales of variability are present. When surface temperatures are prescribed monthly means having a climatological seasonal cycle, daily and interannual variability are lost. In this chapter, a comparison is made of the climate change response in five experiments, each having a different combination of interactively calculated, prescribed and interannually varying, or prescribed and climatological land and/or sea surface temperatures. The climate change response in the high-latitude stratosphere exhibits a robust warming signal and associated weak winds only when at least interannually variability is present in both land and sea surface temperatures, and is strongest when surface temperatures at all gridpoints are calculated interactively.

This chapter consists of a manuscript to be submitted to Geophysical Research Letters: B. Winter and M.S. Bourqui (2010), The Impact of Surface Temperature Variability on the Climate Change Response in the Northern Hemisphere Polar Vortex
Chapter 4: The Impact of Surface Temperature Variability

The Impact of Surface Temperature Variability on the Climate Change Response in the Northern Hemisphere Polar Vortex

B. Winter ¹ and M.S. Bourqui¹

¹Department of Atmospheric and Oceanic Sciences, McGill University, Montréal, QC, Canada

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Abstract

This study investigates the importance of the timescales of variability of land and ocean surface temperatures in the stratospheric response to increased atmospheric greenhouse gas concentrations. We present results from five pairs of 100-year (timeslice) simulations – control and 2×CO$_2$ – carried out with the coupled chemistry-climate model IGCM-FASTOC, in which land and/or sea surface temperatures are either calculated interactively, prescribed and interannually varying, or prescribed with a climatological seasonal cycle. The strongest response to CO$_2$-doubling in the Northern Hemisphere high-latitude winter stratosphere is found when surface temperatures are calculated interactively by a coupled slab ocean and a land surface scheme. In this case, the response is a statistically significant 3-4°C warm anomaly in the high-latitude stratosphere below 25 km and an associated weakening of the polar vortex. When the land is interactive but the slab ocean is replaced by prescribed interannually varying monthly-mean temperatures, the response is similar but weaker, with a warm anomaly statistically significant only below 20 km. When there is no interannual variability in the surface temperature field, there is no dynamic stratospheric response to CO$_2$-doubling. Both the interannual variability in ocean and land temperatures and the adjustment of oceans and lands to the atmosphere and to one another are important in order to maintain realistic stratospheric forcing by planetary waves and to adequately capture the stratospheric response to global warming.
4.1 Introduction

The impact of increasing atmospheric CO$_2$ concentrations on the Brewer-Dobson Circulation, and therefore on the temperature and winds of the high-latitude winter stratosphere, depends strongly on the basic state of the control climate (Sigmond and Scinocca (2010)). Rossby waves that propagate to the stratosphere and shape the control climate have a variety of sources in the troposphere, in particular land-sea temperature contrasts and baroclinicity. The generation of these Rossby waves can be significantly modulated by perturbations to these thermal gradients, whether they are localized temperature anomalies (Brayshaw et al. (2008); Fletcher et al. (2009)) or zonally symmetric anomalies (Brönniman et al. (2004); Lu et al. (2008); Winter and Bourqui (2011a)). It is therefore expected that the timescales of variability of the land and ocean surface temperatures are an important factor governing the magnitude of a climate’s response to increased atmospheric greenhouse gas concentrations. It is shown by Braesicke and Pyle (2004); Olsen et al. (2007) and Winter and Bourqui (2010) that removing interannual variability from monthly sea surface temperatures (SSTs) by prescribing climatological fields strongly dampens (or entirely suppresses) the warming response to CO$_2$-doubling in the Northern Hemisphere high-latitude lower stratosphere that is found when SSTs are either prescribed and interannually varying, or interactively calculated.

The present study was motivated by Winter and Bourqui (2010), in which the CO$_2$-doubling response in a coupled atmosphere-slab ocean model was analyzed and was shown to be highly sensitive to the way the surface is represented in the model. In the present paper we present results from a completed set of five pairs of simulations (control and 2×CO$_2$), in which land or sea surface temperatures are either calculated interactively, prescribed interannually varying, or prescribed as a climatological seasonal cycle. The differences between the experiments lie in the timescales
of surface temperature variability: when monthly-means are prescribed, daily values are obtained by linear interpolation, thereby reducing daily variability. When climatological values are prescribed, interannual variability is lost. We find that a statistically significant warming response in the lowest part of the high-latitude winter stratosphere is found only when the sea surface temperatures vary at least interannually. When the sea surface temperatures are calculated interactively, this warming response is stronger and is significant up to an altitude of 25 km.

The model and experiments are described in Section 4.2; results are presented and discussed in Section 4.3, and conclusions in Section 4.4.

4.2 Model and Experiments

Our five pairs of simulations are performed using the three-dimensional chemistry-climate model IGCM-FASTOC \cite{Bourqui2005, Taylor2005} in its T31 version with 26 vertical levels \cite{Fischer2008, Winter2010}. Each experiment is run for 100 years in timeslice mode. Each pair uses different land and sea surface temperatures:

- **INTocINTls**: INTeractive ocean and INTeractive land surface; the IGCM-FASTOC is coupled to a 25m mixed-layer slab ocean and its land surface scheme is switched on.
- **VARocVARls**: Interannually VARying monthly-mean surface temperatures are prescribed at all ocean and land gridpoints; these surface fields come from the 100-year output of the INTocINTls simulation.
- **FIXocFIXls**: Interannually FIXed monthly-mean surface temperatures (i.e. with seasonal, but without interannual, variability) are prescribed at all ocean and land gridpoints; these are the climatological means from the INTocINTls output.
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- **VARocINTls**: The land surface temperatures are calculated interactively by the land surface scheme but SSTs are prescribed, using the interannually varying fields from INTocINTls.

- **FIXocINTls**: The land surface temperatures are calculated interactively but SSTs are prescribed, using the fixed climatological fields from INTocINTls.

The atmospheric CO$_2$ mixing ratio is 335.8 ppmv for the control and 576 ppmv for the doubling experiments (“doubling” is with respect to a standard pre-industrial value). The prescribed temperatures in the control (doubling) experiments are taken from the control (doubled-CO$_2$) INTocINTls run.

### 4.3 Results and Discussion

Forcing of the Northern Hemisphere vortex by upward-propagating Rossby waves is strongest in January and February and we begin by considering the control and response fields averaged over these months. Figure 4–1 shows the basic state (i.e. the control) surface temperatures (left column) and Northern Hemisphere zonal-mean temperature and winds (middle and right columns) of the five experiments. For reference, NCEP SSTs and ERA40 temperatures and winds are overlaid$^1$ on the graphs. Since the prescribed SSTs in rows one to four are derived from the INTocINTls control field (row five), the shown SSTs are nearly identical (there are differences to within 1°C with respect to the INTocINTls case, with warmer winter SSTs in the four prescribed-SST cases). All match the NCEP SSTs (contours) very well. Land surface temperatures in the FIXocFIXls and VARocVARls experiments are also derived from, and very similar to, the INTocINTls case. In the FIXocINTls

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$^1$ECMWF ERA-40 data used in this study were obtained from the ECMWF data server. NCEP sea surface temperatures are taken from [http://www.cdc.noaa.gov/data/gridded/data.noaa.oisst.v2.html](http://www.cdc.noaa.gov/data/gridded/data.noaa.oisst.v2.html)
Figure 4–1: January-February average surface temperature (°C, left), zonal-mean temperature (°C, centre) and zonal-mean wind (m/s, right) for the five control simulations. NCEP SSTs and ERA40 zonal temperatures and winds are overlaid.
and VARocINTls experiments, the land surface scheme adjusts to the way the ocean is prescribed and in the FIXocINTls case this results in colder temperatures (by up to 5°C) than in the INTocINTls case in mid- to high latitudes. The main dynamical features of the atmosphere are reasonably reproduced, although all five simulations exhibit the classic “cold pole” problem in the stratosphere, generally attributed to insufficient forcing by resolved and unresolved waves. The polar vortex is particularly excessive in the two simulations without an interactive land surface (FIXocFIXls and VARocVARls, second and third rows). This highlights the importance of short-term variability in the land surface temperatures in producing adequate forcing of the vortex.

The January-February average response to CO$_2$-doubling in the zonal-mean temperature and wind fields of the five experiments is presented in the left and middle columns of Figure 4–2, together with the seasonal-cycle response in the eddy heat flux at 100 hPa (right column). Each experiment’s control values are overlaid. Note that this figure repeats some of the panels from Figs. 2–3 and 2–10. The response in the troposphere is very similar in the five experiments – it warms everywhere, particularly at high latitudes – while the stratospheric responses show remarkable differences. In the top two rows, the stratosphere cools as a result of the increased CO$_2$. Like the tropospheric warming, this is the expected radiative response to increases in greenhouse gas loading, and suggests that there are no strong or significant changes in the wave forcing of the vortex when SSTs are climatological, whether or not the land surface is interactive. In the lower three rows, a high-latitude lower-stratospheric warming response becomes increasingly evident, accompanied by a weakening of the vortex. The strongest and most robust response is found when the surface temperatures are calculated interactively at all land and sea gridpoints (i.e. when the model is coupled to a mixed-layer slab ocean and has its land surface scheme switched on).
Figure 4–2: January-February average response in zonal-mean temperature $T$ (K, left) and wind $u$ (m/s, centre), and seasonal cycle response in the 100 hPa eddy heat flux $\nu' T'$, Km/s, right) for the five experiments. Grey shading indicates covers areas not statistically significant at the 5% level.

The state of the vortex in January and February is the result of forcing by waves that entered the stratosphere in the months immediately preceding (e.g. Newman et al. (2001)). The upward component of the Eliassen-Palm (EP) flux is proportional to the zonal-mean eddy heat flux, whose magnitude at 100 hPa in the
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latitudes 40°N to 80°N is strongly correlated with subsequent wave forcing in the vortex (Newman et al. (2001)). The right column in Fig. 4–2 gives the seasonal cycle of the response in eddy heat flux at 100 hPa, with control values overlaid, for each of the five experiments. The strongest positive responses in eddy heat flux between 40°N and 80°N are found in February for the VARocINTls case and from December into early February in the INTocINTls case (Fig. 4–2, right, two at bottom). In the INTocINTls case, the positive eddy heat flux anomaly is located on the control eddy heat flux maximum and extends into the fall. Strong early-winter vertical wave flux in this experiment leads to the strong deceleration of the vortex seen in the January-February average (Fig. 4–2, centre bottom). The anomalous positive eddy heat flux in December in the VARocINTls case does not extend into January, and the erosion of the vortex is weaker than in the INTocINTls case (Fig. 4–2, centre, second from bottom). The very strong positive heat flux anomaly at 100 hPa in February in the VARocINTls experiment does not affect the upper-stratospheric winds until later, and is likely linked to the steep spring breakup of the vortex winds seen on Fig. 4–3 (right column, second from bottom).

The December-January response to CO₂-doubling in the EP flux (vectors) and its divergence (shading) is shown for the five experiments in the left column of Figure 4–3. Above 20 km, the EP flux is divided by density for clarity on the graph. Contours of the zonal-mean zonal wind in the 2×CO₂ simulations are overlaid in green. In all cases except FIXocFIXls the response is an increased EP flux and greater convergence of the flux in the vortex (the increased flux does not lead to a statistically significant change in the vortex in all cases, as seen on Fig. 4–2). In most simulations the increase is mainly in the vertical component of the flux coming out of the troposphere between the latitudes 50°N and 80°N. In the VARocINTls experiment, the vertical flux anomaly at the top of the troposphere is negative north
Figure 4–3: **Left:** December-January average response in EP flux $F$ and its divergence ($\nabla \cdot F$, m/s/day) with zonal-mean winds from the doubled-CO$_2$ simulation overlaid (m/s); **right:** seasonal cycle of the zonal wind variances at 65°N, 10 hPa ($m^2/s^2$) for the control (blue) and doubled-CO$_2$ (red) simulations. EP flux values above 20 km are divided by density.

of 60°N, but there is anomalous northward propagation between 45°N and 60°N at the bottom of the stratosphere, which becomes a positive vertical flux anomaly in
the vortex. The doubled-CO$_2$ vortex in the VARocINTls experiment is stronger than in INTocINTls (see green contours), consistent with the weaker EP flux convergence and propagation compared to INTocINTls. The stronger wind is both a consequence (through reduced wave breaking) and a cause (through greater windspeed-related filtering of waves) of the weaker EP flux.

In order to illustrate the effect of wave forcing separately in the control and 2×CO$_2$ climates, variances of the zonal-mean wind averaged from 65°N to 70°N and from 7.5 to 12.5 hPa are given for each simulation in the right column of Fig. 4–3. This location was selected to be consistent with the WMO definition of Sudden Stratospheric Warming events, and corresponds to the lowest extent of the climatological vortex core. The variance on each day of the year is calculated from the 100 years of the simulation, assuming that each year is independent from the others. These timeseries provide a simple measure of the degree to which the vortex is disturbed by wave forcing in each climate simulation. In all simulations there is lower variability throughout the fall and early winter, followed by a rapid transition to greater variability in January or February, as the vortex begins to be affected by the upward-propagating Rossby waves. Later in the spring the variances taper off as the vortex is broken down. Little else is common to all plots, underlining the fundamental difference between the basic states when the character of the surface representation changes. When surface temperatures are prescribed everywhere as climatological fields (panel A), there is little variability in the flow (control and 2×CO$_2$), consistent with the strong vortex winds noted in Fig. 4–1 (top). The control curves (blue) of the VARocVARls and INTocINTls experiments, in which land and sea surface temperatures have the same timescale of variability and vary at least interannually (panels C and E on the right of Fig. 4–3), show an early onset of variability in the beginning of December, followed by a transition to maximum variance during the month of
February and the start of the decay in early April. The VARocINTls case (panel D), in which the ocean temperatures have interannual variability while the land surface interacts on a daily basis with the atmosphere, follows the same pattern of onset, but the variance decreases in February before reaching its maximum in March. In the FIXocINTls case (panel B), the variance remains low throughout the fall, and the transition to maximum variance occurs earlier, in January. The control vortex (blue curve) thus keeps its maximum variability for a longer period of time. In this experiment, where the sea surface temperatures have no interannual variability while the land surface is interactive, the land-sea temperature differences are stronger than in any of the other cases (not shown). This enhanced land-sea temperature contrast, through its effect on the generation of planetary waves, is likely to induce greater variability in the middle of the winter.

When comparing the control and doubled-CO$_2$ variances of each pair of simulations, several interesting observations can be made. First, the variances in the FIXocFIXls experiments do not undergo significant changes. Therefore, making the assumption that large variances in the vortex wind speed are the result of planetary wave breaking, the surface is not able to induce a change in the generation of planetary waves in response to the 2×CO$_2$ increases despite the monthly mean temperature changes between the control and the 2×CO$_2$ experiments. In the FIXocINTls case the transition to maximum variability is delayed by about a month in the 2×CO$_2$ simulation, a possible consequence of the reduction (albeit statistically insignificant) of the 100 hPa eddy heat flux polewards of 50°N during January (Fig. 4–2, second row left). By March the two climates again exhibit the same comparatively low level of vortex variance, as in the FIXocFIXls experiments. Although land surface temperatures in the FIXocINTls case are interactive, it is likely that the climatologically fixed SSTs nonetheless dampen the generation of planetary waves in both climates.
Second, in the two experiment pairs having interannual, but not daily, variability in the ocean temperatures (VARocVARls and VARocINTls, panels C and D on Fig. 4–3), the early-winter vortex variability is the same for both the control and 2×CO₂ climates until the February transition towards the annual maximum described above. Beginning in February, the 2×CO₂ polar vortex winds show greater variance, consistent with the greater early-winter 100 hPa eddy flux north of 40°N seen on Fig. 4–2, and with the stronger vertical EP flux into the vortex in the December-January average shown in the left column of Fig. 4–3. The biggest difference between these two cases is the February “dip” in the VARocINTls control variances. The low wave flux in the February control climate, rather than a particularly exceptional 2×CO₂ climate, is seen in the strong positive response seen in February in the 100 hPa eddy heat flux between 60°N and 80°N (Fig. 4–2, second from bottom). Third, the INTocINTls case stands out as undergoing the earliest change in the vortex variance. The 2×CO₂ vortex becomes increasingly more perturbed after the early onset of variance until the beginning of February, when it reaches its maximum variance a month earlier than in the control simulation. This is consistent with the large positive anomaly discussed above in the eddy heat flux (Fig. 4–2, bottom). As a consequence, the duration of maximum disturbance of the vortex is extended by a month in the 2×CO₂ simulation in this experiment, and in this experiment only.

The combination of these factors – the time of onset, the duration, and the magnitude of anomalous wave forcing – determines the late winter response in the concentration of ozone in the polar lower stratosphere, where it accumulates as a result of transport by the wave-driven Brewer-Dobson circulation. In Figure 4–4 we present the ozone response (as a number density), averaged over February and March. The almost uniform meridional anomaly above 30 km is due to changes in the efficiency of the ozone-depleting NOₓ chemistry caused by the temperature
Figure 4–4: February-March average response in ozone number density ($10^{17}$ m$^{-3}$) for all five experiments. Control values are overlaid. Grey shading covers all areas not statistically significant at the 5% level. Note that ozone chemistry is only active between the tropopause and 4 hPa.

changes at these altitudes. Transport controls the concentration of ozone mainly below this level and in high latitudes. The statistically significant positive ozone anomaly present in the FIXocFIXls case, where no changes in wave forcing were noticed, represents the effect of the transport of anomalously high mid-stratospheric ozone abundances by an unchanged Brewer-Dobson circulation. With the exception again of the FIXocFIXls and FIXocINTls cases, the response to a doubling of CO$_2$ is increased ozone concentrations in the polar lower stratosphere, consistent with the increased early-winter wave forcing noted above. We also note the strongest increases for the VARocINTls and INTocINTls cases (lower panels), as expected from the strong eddy heat flux and wave-forcing responses in these experiments. Despite a smaller response in these features in the VARocVARls case and the weaker and less significant signal found in its vortex strength (Fig. 4–2), there is nonetheless significantly more polar ozone as a consequence of CO$_2$-doubling (Fig. 4–4, top right).
4.4 Conclusions

This study compares five pairs of 100-year timeslice simulations to analyse the sensitivity of the stratospheric CO$_2$-doubling response to the nature of the surface temperature representation in a climate-chemistry model. By nature, the most realistic among our experiments is the one in which both the interactive land surface scheme and the slab ocean were switched on (INTocINTls). Only this experiment has the surface temperatures in thermal equilibrium with the atmosphere in all locations (lands and oceans) and at all times, and has land and ocean temperatures consistent with one another. The response to doubling atmospheric CO$_2$ in the INTocINTls experiment is the largest response among all our experiments. It shows a statistically significant 3-4°C warm anomaly in the high-latitude stratosphere below 25 km and an associated weakening of the polar vortex. This response is caused by an earlier onset of the maximum wave forcing period by a month, which makes the vortex more susceptible to subsequent forcing. When the land is interactive but the slab ocean is replaced by prescribed interannually varying monthly-mean temperatures (VARocINTls), the daily two-way adjustment of ocean surface temperatures and the atmosphere is lost. The stratospheric response exhibits similar patterns as in the INTocINTls experiment, but with lower magnitudes, and the lower stratospheric polar temperature increase (2-3°C) is only statistically significant below 20 km altitude. In the experiment with prescribed land and ocean temperatures using interannually varying monthly-mean fields (VARocVARls), the stratospheric response in high latitudes is a warming of 1-2°C, statistically significant below 15 km. When the prescribed ocean is fixed to climatology (i.e. there is no interannual variability), and the land is kept interactive (case FIXocINTls), the dynamic stratospheric response essentially vanishes as the change in wave forcing becomes too weak, and the experiment without interannual variability (FIXocFIXls) similarly showed no stratospheric
response (other than the increased radiative cooling from the doubled CO$_2$ common to all experiments).

The ozone density response includes the expected general increase due to the slow-down of the NO$_x$ catalytic chemistry in all experiments. In addition, it includes a positive anomaly at the lower-stratosphere winter pole in the VARocVARls, VARocINTls and INTocINTls experiments. This anomaly is largest in the INTocINTls case.

This study clearly demonstrates the importance of both (i) the interannual variability in ocean (and land) temperatures and (ii) the adjustment of oceans (and lands) to the atmosphere, in order to maintain a realistic stratospheric forcing and to adequately capture the stratospheric response to global warming. It is beyond the scope of this paper to establish the processes by which the daily variability in surface temperature feeds into the stationary thermal eddies that most affect the stratospheric vortex. However, the results shown here suggest that the representation of the oceans could be an important contributor to the currently large range of stratospheric responses to CO$_2$-doubling found in the literature among the various climate-chemistry models (e.g. Austin et al. (2003); Shine et al. (2003); Eyring et al. (2007)). It also suggests that multi-model averages of the stratospheric response to global warming may underestimate the increase in stratospheric wave driving, since they are dominated by models with prescribed interannually varying ocean temperatures (Butchart et al. (2006); Cordero and Forster (2006)), analogous to the VARocINTls design presented here.

The long simulations needed in this study in order to achieve statistical significance were made with the climate-chemistry model IGCM-FASTOC in its T31, 26-level version. Although the results are statistically robust, the limitations of the IGCM-FASTOC should be kept in mind. The model does not include chlorine and
bromine chemistry, as the experiments are meant to represent pre- and post-ozone hole conditions where these families play a secondary role in the ozone chemistry. The model has a classical cold bias which is likely related to the lack of gravity wave drag representation. Adding a gravity wave drag parametrization would improve the wind profiles and might thereby change the vertical propagation of planetary waves. However, the main results found here are robust throughout the troposphere and the stratosphere and are therefore unlikely to be extremely sensitive to the details of the vertical propagation of planetary waves in the stratosphere. The relatively low resolution may hinder the representation of small baroclinic waves in the troposphere, which through non-linear interaction may form planetary waves. However, this source of planetary waves is expected to be minor in regard to the larger-scale sources.

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5.1 Summary

The overarching goal of this thesis is to assess the response of the stratosphere to anomalous thermal forcing at the surface and radiative forcing by greenhouse gas increases related to global warming. My analysis focuses on the response in the Northern Hemisphere polar vortex forcing by planetary waves propagating upward from the troposphere, and the impact this has on the strength of the Brewer-Dobson circulation, temperature and wind fields, and concentration of polar ozone. The thermal forcing applied at the surface is a warm anomaly added to the climatological surface temperature which constitutes the model’s lower boundary, and the radiative forcing applied throughout the atmosphere results from a doubling of atmospheric CO$_2$. The purpose of using the CO$_2$-doubling scenario is to facilitate the comparison with other greenhouse gas increase studies, as reviewed in Chapter 1.3. While most past and current efforts in this field are directed \textit{a priori} towards more precise predictions of the future climate, the sensitivity of these predictions with respect to the choice of greenhouse gas scenario or surface forcing is not known. The goal of this thesis is to isolate specific aspects of the forcing – CO$_2$ changes without ozone depletion or recovery; latitude of the surface warming; timescale of the variability of the basic state of the control climate – and thereby to uncover the sensitivity of the results to these factors and to address the implications of this sensitivity for climate prediction studies.
All experiments were carried out with a full three-dimensional chemistry-climate model, the IGCM-FASTOC. A general circulation model with complete representation of atmospheric physics is necessary in order to obtain as realistic a control climate as possible. In the stratosphere, the feedbacks between ozone photochemistry, radiation and dynamics have non-negligible effects on the local temperature, making an interactive chemistry model an important tool for assessing the response to climate change.

In Chapter 2, the IGCM-FASTOC is coupled to a mixed-layer slab ocean, and it is shown that a doubling of pre-industrial CO$_2$ levels (to 576 ppmv) in the atmosphere leads to strongly increased resolved-wave forcing of the Northern Hemisphere polar vortex. The enhanced descending branch of the Brewer-Dobson Circulation (BDC) at high northern latitudes leads to significant adiabatic warming in the lower stratosphere in January and February; in particular, the lower stratosphere NH winter pole warms by up to 4K. The enhanced BDC together with the radiative cooling of the upper stratosphere by CO$_2$ lead to an increased ozone concentration in the winter polar lower stratosphere, which extends into the spring.

A key result of this chapter is that the timing of the wave forcing events changes under doubled-CO$_2$ conditions. In the control climate, the amount of forcing of the mean flow at the top of the stratosphere builds gradually throughout the winter months, towards a maximum in March. In the doubled-CO$_2$ climate, maximum wave forcing begins in January and persists through March. The vortex is therefore subject to erosion by waves through a longer period of time. This emphasises the importance of considering the seasonal cycle of the response to climate change. Without this information, the comparison of results from different models may be misleading if some are based, for instance, on perpetual-January simulations while others present seasonal means.
Chapter 2 suggests that the severity of a potential Arctic ozone hole would be mitigated by the effects of increasing CO$_2$ concentrations on the stratospheric dynamics and vortex temperatures: the low temperatures required for the formation of polar stratospheric clouds are less likely to occur, and the longer period during which the vortex undergoes strong forcing by planetary waves implies more frequent mixing-in of ozone-rich air from lower latitudes.

Doubling the atmospheric CO$_2$ concentration leads to a global warming of the surface, and the response found in the stratosphere (summarized above) is due both to the *in situ* radiative forcing by the CO$_2$ and to the increased surface temperature. Chapter 3 investigates the surface temperature forcing alone (i.e. without CO$_2$-doubling) by imposing 2K temperature anomalies in a series of zonal bands of 10° and 30° latitude width.

Two fundamental responses to the surface forcing are noted: (1) When the forcing is applied continuously between 30°N and 60°N, or within 10° latitude windows centred at 35°N, 45°N, 75°N or 85°N, the wave activity entering the winter stratosphere at 100 hPa polewards of 40°N is reduced, and so is the wave breaking in the vortex. The anomalously weak upward flux at 100 hPa begins in the early winter, allowing an anomalously strong vortex to form. The reduction in wave forcing higher in the stratosphere occurs after the vortex has reached its annual peak, so that it persists at full strength for a longer time. The associated response in zonal-mean temperature is a strong cooling in the high-latitude stratosphere. Responses of the same sign, but strongly reduced and not statistically significant, are found when the surface forcing is centred at 55°N and 65°N or imposed continuously from 60°N to 90°N. (2) The reverse is found when surface forcing is applied equatorwards of 30°N: there is a greater upward flux of waves into the stratosphere polewards of 40°N from the fall into the winter, leading to anomalously strong vortex disruption in January.
and February. An anomalously warm high-latitude lower to mid-stratosphere ac-
companies the weakened vortex. The increased wave-forcing response occurs before the
vortex is fully established, and has the potential to weaken the vortex for the entire
winter, even when the wave forcing anomaly subsequently decreases.

A transition from one to the other of these opposite responses occurs when the
imposed surface forcing is moved from south to north of around 30°N. The location
of the applied thermal surface forcing affects the meridional temperature gradient in
the troposphere, and in order for the climate to remain in equilibrium, the poleward
transport of heat must adjust to the imposed forcing. Stationary thermal eddies
are one mechanism by which heat is transported to high latitudes in any climate,
and those stationary eddies having zonal wavenumber 1 or 2 constitute the main
forcing of the polar vortex. A response in these stationary Rossby waves is seen
from the near-surface to the top of the troposphere when surface forcing is imposed.
The transient synoptic eddies are another mechanism by which heat is transported
polewards. In the climatological results shown in this thesis, the effect of transient
synoptic eddies is included in climatological measures of baroclinicity, which can be
quantified for instance by the Eady growth rate. The response in the baroclinicity to
a given surface forcing is of the same sign as the response in the stationary s1 and s2
waves, when this surface forcing is imposed south of around 60°N. When it is imposed
polewards of this latitude, no significant response is seen in the baroclinicity, although
a response in the planetary waves occurs, whose effect is seen in the polar vortex.
A transition zone can therefore also be said to exist near 60°N, between latitudes
whose surface temperature changes impact the baroclinicity and those having no such
impact. Combining the surface forcing in an equatorial and an extra-tropical band
demonstrates that the equatorial response dominates and causes the surface forcing
in these cases to weaken the polar vortex, though not as much as when an equatorial band is forced alone.

The stratospheric response to CO$_2$-doubling qualitatively resembles its response to the forcing in equatorial zonal bands, which is consistent with the dominant role found to be played by the low-latitude surface forcing when there is no increase in atmospheric CO$_2$. CO$_2$-doubling clearly leads to changes in the tropospheric jet and thus to changes in the baroclinicity and in the planetary wave generation in the troposphere. However, this link between the results from Chapters 3 and 2 should be taken with caution. By experimental design, the surface temperatures in the “band forcing” simulations are prescribed at all land and sea surface gridpoints and have no interannual variability. The experiments discussed in Chapter 2, on the other hand, are carried out with interactively calculated surface temperatures at all land and sea gridpoints. The representation of surface temperatures has an enormous impact on the climate change response in the stratosphere. This was noted in Chapter 2, then addressed in detail in Chapter 4.

In Chapter 4, the 2$\times$CO$_2$ experiment is carried out five times, each time with a different representation of the surface temperatures in both the control and the doubled-CO$_2$ simulations. The five experiments differ by the timescale of the variability in the surface temperatures. When land and ocean temperatures are both calculated interactively (the INTocINTls pair of simulations), they adjust to one another and to the atmospheric temperature on a daily basis. The doubled-CO$_2$ response in this experiment is the subject of Chapter 2, and it is the largest response among all the experiments. When the land is interactive but the slab ocean is replaced by prescribed interannually varying monthly-mean temperatures (VARocINTls), the daily two-way adjustment of sea surface temperatures and the atmosphere is lost. The stratospheric response exhibits similar patterns as in the INTocINTls case, but with
lower magnitudes, and the lower stratospheric polar temperature increase is statistically significant only below 20 km altitude. In the experiment with prescribed land and ocean temperatures using interannually varying monthly-mean fields (VARoc-VARls), the stratospheric response in high latitudes is only half as strong as in the INTocINTls case, and statistically significant below 15 km. When the prescribed ocean is fixed to climatology (i.e. there is no interannual variability), and the land is kept interactive (case FIXocINTls), the dynamic stratospheric response essentially vanishes as the change in wave forcing becomes too weak, and the experiment without any interannual variability (FIXocFIXls) led to no response in the high-latitude stratosphere other than increased radiative cooling.

The implications of the results found in Chapter 4 are that the representation of the oceans in particular could be an important contributor to the large range of stratospheric responses to CO₂-doubling currently found in the literature among the various climate-chemistry models (e.g. Austin et al. (2003); Shine et al. (2003); Eyring et al. (2007)). It also suggests that multi-model averages of the stratospheric response to global warming may underestimate the increase in stratospheric wave driving, since they are dominated by models analogous to VARocINTls, with prescribed interannually varying ocean temperatures (Butchart et al. (2006); Cordero and Forster (2006)).

5.2 Future Work

5.2.1 Short-term variability

The degree by which the Northern Hemisphere polar vortex is disturbed by upward-propagating planetary waves under different forcing conditions is a key element in all the results presented in this thesis. However, all the results shown are climatological means over the 100 or 50 years available for each simulation, and
these smooth out the effects of the particularly severe wave forcing events that lead to Sudden Stratospheric Warmings (SSWs). In the case of a SSW, the vortex typically breaks down completely and the winds may become easterly. The associated anomalously large downwelling in the polar descending branch of the Brewer-Dobson circulation is accompanied by the large adiabatic temperature increase that gives the Sudden Warming its name. SSWs occur at a rate of about one every other winter and are linked to the Quasi-Biennial Oscillation (QBO) in the tropical upper troposphere (Labitzke and Naujokat (2000)), which plays a role in determining whether the polar vortex in a given year is more or less stable (and consequently more or less susceptible to severe disruption). Because planetary wave driving of the vortex is strongly related to the land-sea temperature contrasts found only in the Northern Hemisphere, SSWs are rarely seen in the Southern Hemisphere (in fact, there is only one recorded instance, which took place in 2002).

Whether the frequency of SSWs would increase in a future climate having greater greenhouse-gas concentrations is an open question. It is of interest because, as described in Baldwin and Dunkerton (2001), large variations in the strength of the stratospheric circulation can propagate downwards not only to the tropopause but indeed to the surface, and have the potential to contribute to medium-range weather prediction. More generally, a climate change response in the frequency or severity of SSWs would reflect both a change in wave forcing (which has been shown to occur; e.g. Butchart et al. (2006); Winter and Bourqui (2010)) and a change in the state of the vortex (this was quantified as the vortex variance in Chapter 4). Part of the difficulty in answering the question of a change in SSW occurrence lies in the inability of most stratosphere-resolving GCMs, including the IGCM-FASTOC, to reproduce the QBO, as this would require very high vertical resolution. However, this mainly acts
to reduce the frequency of modelled SSWs; it does not prevent them from happen-
ing. The biggest obstacle to detecting a climate change response is the low number of events available for analysis in most climate simulations. Another obstacle of sorts is the rigid definition of what constitutes an SSW event that became the standard after SSWs were first observed in the stratosphere and a classification scheme was needed. The definition, formalized by the WMO, is based on the zonal winds at 65°N and 10 hPa; this corresponds to the latitude of the vortex axis at a height reached by the majority of radiosondes (Harvey and Hitchman (1996)). However, vortex axis locations differ among models and may themselves be affected by climate change.

It would therefore be interesting to return to the 100 years of both the control and 2×CO$_2$ simulations in the experiments in which the IGCM-FASTOC is coupled to a slab ocean (as these show the strongest climate change response), and determine the nature and magnitude of the SSW response. As already shown in Chapter 4, the variability of the polar vortex reaches its maximum earlier in the season under doubled-CO$_2$ conditions in this experiment. Examination by eye of the zonal-mean winds averaged over the central lower vortex in each of the 100 years (of the control and of the doubled-CO$_2$ simulations) indicates an increase in SSW occurrence of nearly 50% when CO$_2$ is doubled, but this needs to be confirmed in a quantitative analysis. In order to avoid definitions tied to present-day observations, SSWs should be identified within the context of their own climate, for instance by the sign and magnitude of the time derivative of the zonal wind averaged over the vortex core, and the magnitude of the change compared to the standard deviation of the winds.

5.2.2 Links between surface and radiative forcing

It is known that the surface temperature response to global warming is partic-
ularly strong at high latitudes (IPCC (2007)), but the combined-band experiments
in Chapter 3 demonstrate that the stratospheric response to a surface heat anomaly applied in the sub-tropical latitudes dominates the response to an anomaly applied at higher latitudes. The band combinations, however, included one low-latitude and one mid-latitude band, because individually these led to the strongest responses in the stratosphere. It would be interesting to extend these experiments by combining low-latitude with high-latitude surface forcing.

Furthermore, comparing the response in the hemispheric combined-band forcing experiment EQNMP presented in Chapter 3 to the response in the FIXocFIXls CO$_2$-doubling experiment in Chapter 4 (both experiments have climatologically fixed surface temperatures at every land and ocean point) raises two issues: (1) A temperature anomaly applied only at the surface leads to dynamic changes in the tropospheric circulation that are not present when the entire troposphere is radiatively heated by increased greenhouse gases. (2) A uniform 2K surface temperature anomaly is stronger than the expected surface temperature response to greenhouse gas forcing in low latitudes and weaker in high latitudes. An experimental design by which a thermal anomaly is added to the air temperature at every model level would allow identifying which isolated bands of latitude dominate the response when the entire atmospheric column within a range of latitudes is forced (testing the radiative heating effect), and would allow imposing meridional gradients in the applied thermal forcing at all vertical levels (testing the sensitivity to such gradients). Nudging the model-level air temperature towards an imposed profile allows the model to be run in fully interactive mode, avoiding the under-representation of planetary wave forcing found in simulations having climatologically fixed prescribed boundary conditions (see Chapter 4).
5.2.3 The Southern Hemisphere

Although the response in the Southern Hemisphere to CO$_2$-doubling was addressed in Chapter 2 and to band-wise surface thermal forcing in Chapter 3, the focus in this thesis has been primarily on the Northern Hemisphere. Rossby wave forcing of the polar vortex is considerably stronger in the Northern Hemisphere because of the land-sea temperature constraints provided by the continents. Upward propagation of wavenumber 1 or 2 waves is not absent in the Southern Hemisphere, but is less organized; vortex forcing by higher wavenumbers takes on greater importance, particularly as these can become organized into packets having envelopes of size wavenumber 1 or 2 (Scinocca and Haynes (1998)). Consequently, the Southern Hemisphere polar vortex is much stronger than its NH counterpart, both less affected by, and more resistant to, wave forcing. The latitude-band results in Chapter 3 suggest that the response to surface thermal forcing is annular mode-like in character, in that the tropospheric jet and stratospheric vortex appear to weaken along the axis of their cores, or to shift with respect to this axis. This is true also for the response found in the 2×CO$_2$ experiment of Chapter 2, which was accompanied by slight warming in the lower polar stratosphere. An analysis of the Brewer-Dobson circulation response in the high southern latitudes would be of interest mainly because of its role in transporting ozone to the pole. In Antarctica, necessary conditions for the formation of polar stratospheric clouds are always met, and the slight warming noted in the lower stratosphere under doubled-CO$_2$ conditions would not change this. However, if the vortex is more disturbed in a changed climate, the potential for mixing in ozone from lower latitudes increases. Without chlorine and bromine chemistry, the IGCM-FASTOC results reflect a pre- or post-ozone hole scenario, and would isolate the effects of CO$_2$ changes on the ozone concentration in the southern high latitudes.
5.2.4 Other greenhouse gases

Greenhouse-gas forcing in this thesis has been represented by a doubling of atmospheric CO$_2$ with the other radiatively active greenhouse gases held constant at 1979 levels. It was shown in Bourqui et al. (2005) that the anthropogenic increases in other greenhouse gas concentrations can act to compensate some of the effects of increased CO$_2$. That study was carried out with the IGCM-FASTOC run at lower resolution and without the coupled slab ocean, and it is worth investigating the impact of other greenhouse gas increases – in isolation or in combination with CO$_2$ increases – in the more advanced version of the model as it is available now.
Appendix A: Transformed Eulerian Mean Equations

The Transformed Eulerian Mean (TEM) form of the momentum and energy equations has become the standard for describing the dynamics of the winter meridional circulation in the stratosphere driven by wave forcing of the polar vortex. As they are less frequently applied to diagnostics of the tropospheric circulation, they are reproduced here for reference. The version in Cartesian coordinates is from Holton (2004); see Andrews et al. (1987) for a transformation to spherical coordinates on log-pressure levels. The zonally uniform character of the stratospheric large-scale flow lends itself to representation by Eulerian-mean system of governing equations (from Holton (2004), section 10.2.1, for quasi-geostrophic scaling):

\[
\frac{\partial \mathbf{u}}{\partial t} - f_0 \mathbf{v} = - \frac{\partial (u'u')}{\partial y} + X
\]

(1)

\[
f_0 \mathbf{u} = - \frac{\partial \Phi}{\partial y}
\]

(2)

\[
\frac{\partial \Phi}{\partial z} = \frac{RT}{H}
\]

(3)

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

(4)

\[
\frac{\partial T}{\partial t} + \frac{N^2 H}{R} \frac{\mathbf{w}}{\bar{w}} = - \frac{\partial (v'T')}{\partial y} + \frac{\mathbf{J}}{c_p}
\]

(5)

The vertical coordinate used in the above is the log-pressure height, defined as \( z \equiv -H \ln(p/p_s) \), where \( H \) is a scale height (7 km for a representative stratospheric temperature of 240 K) and \( p_s \) is a reference pressure. \( R \) is the dry air gas constant, \( c_p \) is the specific heat for dry air and \( N \) is the Brunt-Väisälä frequency. \( X \) designates all small-scale zonal forcing and \( \mathbf{J} \) is diabatic heating. Eulerian mean motion, however, cannot account for the poleward transport of air parcels from low to high latitudes.
throughout the winter. The TEM equations were developed to approximate the actual Lagrangian mean motions of air parcels. In Andrews and McIntyre (1976), this is done by defining a residual meridional circulation \((v^*, w^*)\) as follows:

\[
\begin{align*}
  v^* &\equiv \bar{v} - \frac{R}{\rho_0 H} \frac{\partial}{\partial z} \left[ \frac{\rho_0 (v'T')}{N^2} \right], \\
  w^* &\equiv \bar{w} + \frac{R}{H} \frac{\partial}{\partial y} \left[ \frac{(v'T')}{N^2} \right]
\end{align*}
\] (6)

These are definitions, not derivations, and are not the only way in which the diabatically driven part of the zonal circulation may be expressed; see for instance Kodama et al. (2007) and references therein. However, these definitions by Andrews and McIntyre (1976) are the most widely used in the western scientific literature and have become standard. The asterisk indicates the residual circulation associated with diabatic heating. Substituting these residual terms \((v^*, w^*)\) for the “regular” Eulerian zonal-mean terms \((\bar{v}, \bar{w})\) in the set of equations 1 to 5 gives the Transformed Eulerian Mean equations:

\[
\begin{align*}
  \frac{\partial \mathbf{v}^*}{\partial t} - f_0 \frac{\bar{v}}{\rho_0} \vec{\nabla} \cdot \vec{F} + \mathbf{X} = \frac{1}{\rho_0} \nabla \cdot \vec{F} + \mathbf{X} \\
  f_0 \bar{u} = \frac{\partial \Phi}{\partial y} \\
  \frac{\partial \Phi}{\partial z} = \frac{RT}{H} \\
  \frac{\partial \mathbf{v}^*}{\partial y} + \frac{1}{\rho_0} \frac{\partial (\rho_0 \bar{w}^*)}{\partial z} = 0 \\
  \frac{\partial T^*}{\partial t} + \frac{N^2 H}{R} \frac{\bar{w}^*}{\rho_0} = \frac{\dot{J}}{c_p}
\end{align*}
\] (7-11)

The vector \(\vec{F}\) appearing in the first term on the right of equation 7 is the Eliassen-Palm (EP) flux, whose components are proportional to the zonal-mean eddy momentum and heat fluxes:

\[
(F_y, F_z) = (-\rho_0 \bar{u}'v', \frac{f_0 R \rho_0}{N^2 H} \frac{v'T'}{v'T'})
\] (12)
The divergence of this flux, not the individual components of the flux itself, acts as a body force on the flow, and $\mathbf{X}$ as before represents all forcing on the flow other than by eddies. In general circulation model output, $\mathbf{X}$ also includes all forcing by eddies the model cannot resolve, such as gravity waves. The usefulness of the EP flux components as well as of the divergence of the EP flux as diagnostics of the stratospheric circulation is outlined in Edmon et al. (1981). All plots in this thesis showing TEM diagnostics use their complete form, not the quasi-geostrophic approximations. In spherical coordinates and on isobaric levels $(\lambda, \phi, p)$, the complete TEM equations are:

\[
\frac{\partial \pi}{\partial t} - \bar{v}^* \left[ f - \frac{1}{a \cos \phi} \frac{\partial (\pi \cos \phi)}{\partial \phi} \right] + \bar{\omega}^* \frac{\partial \pi}{\partial p} = \frac{1}{a \cos \phi} \nabla \cdot \mathbf{F} + \mathbf{X} \tag{13}
\]

\[
\bar{u} \left[ f - \frac{\pi \tan \phi}{a} \right] + \frac{1}{R} \frac{\partial \Phi}{\partial \phi} = G \tag{14}
\]

\[
\frac{\partial \Phi}{\partial p} = -\frac{RT}{p} \tag{15}
\]

\[
\frac{1}{a \cos \phi} \frac{\partial (\bar{v}^* \cos \phi)}{\partial \phi} + \frac{\partial \bar{\omega}^*}{\partial p} = 0 \tag{16}
\]

\[
\frac{\partial \Theta}{\partial t} + \bar{v}^* \frac{\partial \Theta}{\partial \phi} + \bar{\omega}^* \frac{\partial \Theta}{\partial p} = -\frac{\partial}{\partial p} \left[ \frac{1}{a} \frac{\partial \Theta}{\partial \phi} \frac{\partial \Theta'}{\partial \phi} \right] + \bar{Q} \tag{17}
\]

Here, $a$ is the radius of the Earth, $G$ represents all departures from gradient wind balance (see Andrews et al. (1987)) and $\bar{Q}$ is the diabatic heating in the notation for using potential, rather than absolute, temperature. The components of the residual meridional circulation in these coordinates are defined as follows:

\[
\bar{v}^* \equiv \bar{v} - \frac{\partial}{\partial p} \left[ \frac{\bar{v}' \Theta'}{(\partial \Theta' / \partial p)} \right] , \quad \bar{\omega}^* \equiv \bar{\omega} + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left[ \frac{\bar{v}' \Theta' \cos \phi}{(\partial \Theta' / \partial p)} \right] \tag{18}
\]
The components of the EP flux are:

\[ F_\phi = a \cos \phi \left[ \frac{v' \Theta'}{(\partial \Theta/\partial p)} \frac{\partial \pi}{\partial p} - u' v' \right], \quad F_p = a \cos \phi \left[ \frac{v' \Theta'}{(\partial \Theta/\partial p)} \zeta - \omega' u' \right], \quad (19) \]

where

\[ \zeta = f - \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (\pi \cos \phi) \quad (20) \]
## Appendix B: Acronyms

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGCM</td>
<td>Atmospheric General Circulation Model</td>
</tr>
<tr>
<td>AR4</td>
<td>Assessment Report 4 (of the IPCC)</td>
</tr>
<tr>
<td>BDC</td>
<td>Brewer-Dobson Circulation</td>
</tr>
<tr>
<td>CCM</td>
<td>Chemistry-Climate Model</td>
</tr>
<tr>
<td>CPU</td>
<td>Central Processing Unit</td>
</tr>
<tr>
<td>ENSO</td>
<td>El Niño - Southern Oscillation</td>
</tr>
<tr>
<td>EP</td>
<td>Eliassen-Palm</td>
</tr>
<tr>
<td>FASTOC</td>
<td>FAst STRatospheric Ozone Chemistry (chemistry module for the IGCM)</td>
</tr>
<tr>
<td>GCM</td>
<td>General Circulation Model</td>
</tr>
<tr>
<td>GHG</td>
<td>Greenhouse Gases</td>
</tr>
<tr>
<td>IGCM</td>
<td>Intermediate General Circulation Model (the Reading GCM)</td>
</tr>
<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
</tr>
<tr>
<td>NH</td>
<td>Northern Hemisphere</td>
</tr>
<tr>
<td>PNA</td>
<td>Pacific-North America (pattern)</td>
</tr>
<tr>
<td>QBO</td>
<td>Quasi-Biennial Oscillation</td>
</tr>
<tr>
<td>SH</td>
<td>Southern Hemisphere</td>
</tr>
<tr>
<td>SPARC</td>
<td>Stratospheric Processes And their Role in Climate</td>
</tr>
<tr>
<td>SRES</td>
<td>Special Report on Emission Scenarios</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperature</td>
</tr>
<tr>
<td>SSW</td>
<td>Sudden Stratospheric Warming</td>
</tr>
<tr>
<td>TAR</td>
<td>Third Assessment Report (of the IPCC)</td>
</tr>
<tr>
<td>TEM</td>
<td>Transformed Eulerian Mean</td>
</tr>
<tr>
<td>WCRP</td>
<td>World Climate Research Programme</td>
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</table>
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