JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 111, D06108, doi:10.1029/2005JD006283, 2006

Transfer of the solar signal from the stratosphere to the troposphere: Northern winter

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Received 28 May 2005; revised 26 September 2005; accepted 9 December 2005; published 29 March 2006.

[1] The atmospheric response to the solar cycle has been previously investigated with the Freie Universität Berlin Climate Middle Atmosphere Model (FUB-CMAM) using prescribed spectral solar UV and ozone changes as well as prescribed equatorial, QBO-like winds. The solar signal is transferred from the upper to the lower stratosphere through a modulation of the polar night jet and the Brewer-Dobson circulation. These model experiments are further investigated here to show the transfer of the solar signal from the lower stratosphere to the troposphere and down to the surface during Northern Hemisphere winter. Analysis focuses on the transition from significant stratospheric effects in October and November to significant tropospheric effects in December and January. The results highlight the importance of stratospheric circulation changes for the troposphere. Together with the poleward-downward movement of zonal wind anomalies and enhanced equatorward planetary wave propagation, an AO-like pattern develops in the troposphere in December and January during solar maximum. In the middle of November, one third of eddy-forced tropospheric mean meridional circulation and surface pressure tendency changes can be attributed to the stratosphere, whereas most of the polar surface pressure tendency changes from the end of November through the middle of December are related to tropospheric mechanical forcing changes. The weakening of the Brewer-Dobson circulation during solar maximum leads to dynamical heating in the tropical lower stratosphere, inducing circulation changes in the tropical troposphere and down to the surface that are strongest in January. The simulated tropospheric effects are identified as indirect effects from the stratosphere because the sea surface temperatures are identical in the solar maximum and minimum experiment. These results confirm those from other simplified model studies as well as results from observations.

Citation: Matthes, K., Y. Kuroda, K. Kodera, and U. Langematz (2006), Transfer of the solar signal from the stratosphere to the troposphere: Northern winter, *J. Geophys. Res.*, *111*, D06108, doi:10.1029/2005JD006283.

1. Introduction

[2] The influence of solar variability on climate is an important research topic intending to estimate the natural versus the anthropogenic climate change. *Hines* [1974] and *Bates* [1981] proposed that the 11-year solar UV changes would have an impact on tropospheric climate through changes in planetary wave propagation. However, the limitation of observational data makes it difficult to follow the large initial solar signal from the upper stratosphere down to the troposphere. A variety of observations indicate that the troposphere seems to be influenced by the 11-year solar cycle [e.g., *Labitzke and van Loon*, 1988; *Kodera*, 2002; *Gleisner and Thejll*, 2003; *Haigh*, 2003; *van Loon et al.*, 2004; *Crooks and Gray*, 2005]. The transfer of the solar signal from the upper to the lower stratosphere can be explained by the modulation of the polar night jet (PNJ) and the Brewer-Dobson (BD) circulation through wave-mean flow interactions [e.g., *Kodera and Kuroda*, 2002] and has been recently confirmed in a general circulation model (GCM) study [*Matthes et al.*, 2004]. The explanation of the transfer from the lower stratosphere to the troposphere remains however a difficult task and is subject of this paper.

[3] Recent studies with GCMs and chemistry climate models (CCMs) [e.g., *Haigh*, 1996, 1999; *Shindell et al.*, 1999; *Rind et al.*, 2002; *Tourpali et al.*, 2003; *Egorova et al.*, 2004; *Rozanov et al.*, 2004] try to confirm the observed influence of the 11-year solar cycle on the troposphere down to the surface. *Shindell et al.* [1999] and *Rind et al.* [2002] explain this with changes in planetary wave propagation in the stratosphere and troposphere, similar to estimates from observations [e.g., *Kodera et al.*, 1990]. However, these model studies did not show the detailed time evolution of the transfer of the solar signal from the upper stratosphere to the lower stratosphere and troposphere which is the purpose of this paper.

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[4] Many current observational and modeling studies are aimed at understanding of stratosphere-troposphere coupling in order to improve the predictability of surface weather. The influence of the troposphere on the stratosphere is basically understood. Planetary Rossby waves generated in the troposphere propagate vertically upward to the stratosphere, where they break or dissipate under certain conditions, and deposit their energy and momentum [*Matsuno*, 1971]. Changes in tropospheric waves can therefore directly influence the stratospheric circulation. The other direction, i.e. an influence of the stratosphere on the tropospheric circulation, is more difficult to determine. Some studies suggest a downward dynamical coupling through downward propagating zonal mean anomalies involving wave-mean flow interactions [e.g., *Kodera et al.*, 1990; *Christiansen*, 2001; *Norton*, 2003; *Song and Robinson*, 2004] and/or provide statistical evidence for stratosphere-troposphere coupling through typical atmospheric modes [*Perlwitz and Graf*, 1995; *Thompson and Wallace*, 2000; *Baldwin and Dunkerton*, 2001; *Thompson et al.*, 2002; *Black and McDaniel*, 2004]. *Perlwitz and Harnik* [2003, 2004] additionally discuss the reflection of waves from the stratosphere back to the troposphere.

[5] Studies with middle atmosphere (MA) general circulation models (GCMs) provide a useful tool to test the observed findings and explain the mechanism by which stratospheric circulation changes can affect surface climate. The aim of this paper is to show the importance of stratospheric circulation changes for the troposphere with the example of 11-year solar cycle model experiments.

[6] The paper is structured as follows. Section 2 describes the model experiments and section 3 discusses the downward transfer of the solar signals in NH winter, with special emphasis on the transition period from significant stratospheric effects in October and November to significant tropospheric effects in December and January. Section 4 investigates the tropospheric high-latitude signal and section 5 the tropospheric low-latitude signal. Section 6 discusses and summarizes the results.

2. Model Experiments

[7] We use two 15-year experiments with the Freie Universität Berlin Climate Middle Atmosphere Model (FUB-CMAM) which were performed under constant solar maximum (max) and constant solar minimum (min) conditions with spectrally discriminated solar UV changes, prescribed solar-induced O_3 changes and prescribed idealized equatorial winds throughout the stratosphere from rocketsonde data (for details, see *Matthes et al.* [2004]). These model experiments showed an improved stratospheric response to the 11-year solar UV signal during Northern Hemisphere (NH) winter. The polar night jet and the mean meridional circulation are modulated by the solar cycle comparable to observations [*Kodera and Kuroda*, 2002] and stratospheric warmings occur in the west phase of the QBO during solar max conditions. These improvements were ascribed to a better wind climatology due to the imposed relaxation toward more realistic equatorial winds throughout the stratosphere which allowed a more realistic feedback of the weak solar signal. Thus these model experiments are used to examine in detail the transfer of the solar signal from the lower stratosphere into the troposphere. We take the max and min experiments with a prescribed QBO easterly phase in the lower stratosphere, which show a stronger influence on the tropospheric circulation than the experiments with a prescribed QBO westerly phase. Here, we do not discuss the dependence of the results on the phase of the Quasi-Biennial Oscillation (QBO). It should be noted that the experiments include neither a realistic time-varying 11-year solar cycle nor a realistic time-varying QBO. Such experiments are now becoming possible with increased available computer resources. We investigate the transfer of the solar signal from the stratosphere to the troposphere during NH winter with long-term mean (15-year mean) differences between max and min fields.

[8] The sea surface temperatures (SSTs) in the experiments are monthly mean varying climatological values which are identical for max and min conditions. Surface and tropospheric variability are therefore reduced and direct solar influences on the surface covered with water are suppressed.

[9] A comparison with previous 20-year equilibrium simulations from the FUB-CMAM [e.g., *Labitzke and Matthes*, 2003] revealed that the statistical significances are robust in the tropics, subtropics and midlatitudes. The large interannual variability of the FUB-CMAM at high latitudes during winter prevents statistically significant signals for the 15-year integrations presented here, as well as for longer integrations. To overcome the limitation of computer resources for longer integrations, an intercomparison with other models (a quasi-ensemble approach, e.g., *Matthes et al.* [2003]) as well as with observations is useful to understand the robustness of the signals. Model simulations can be used to test the observed signals and vice versa.

3. Downward Transfer of the Solar Signal in Northern Winter

3.1. Zonal Mean Wind

[10] According to the previous analysis of these experiments [*Matthes et al.*, 2004], the enhanced short wave heating during solar max in the tropical upper stratosphere leads to a warming and, through the thermal wind relationship, to a significant acceleration of the subtropical zonal mean wind in October. This westerly wind anomaly further strengthens and moves poleward and downward with time through the interaction between planetary waves and the zonal mean flow, in agreement with observations. In December and January, a significant influence has been found in the troposphere and down to the surface at midlatitudes.

[11] To show the transfer of the solar signal from the stratosphere to the troposphere, we will discuss 10-day mean differences that combine all 15 years for the transition period from significant stratospheric effects in November to significant tropospheric effects in December.

[12] In the first and second periods of November (Nov1 and Nov2), the strong and significant westerly wind anomaly can be found in the midlatitude upper stratosphere/lower mesosphere moving poleward with time (Figure 1a). A rapid change occurs from Nov3 to the first period of December (Dec1). Whereas in Nov3 a strong and no longer significant westerly wind anomaly exists at the NH high-

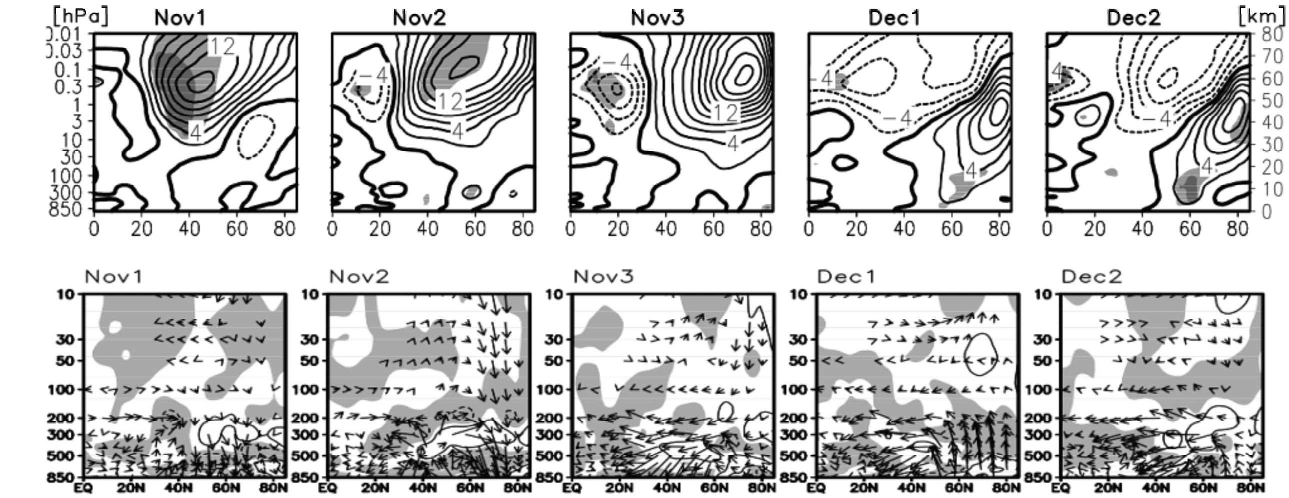

Figure 1. (a) Long-term 10-day mean differences of the mean zonal mean wind between the solar max and min experiments for the NH from Nov1 to Dec2, contour interval: 2 m/s. Light (heavy) shading indicates the 95% (99%) significance level calculated with a Student's t-test. (b) Long-term 10-day mean differences of the Eliassen-Palm Flux vector between the solar max and min experiments (arrows, scaled by the inverse of pressure) and its divergence (only 1 m/s/d contour is shown, negative values are shaded) for the NH from 850 to 10 hPa from Nov1 to Dec2.

latitude mesosphere, westerly wind anomalies exist at NH high latitudes throughout the stratosphere, troposphere, and down to the surface in December. Easterly wind anomalies dominate the tropical, subtropical, and midlatitude upper stratosphere and mesosphere. Statistical significances occur around 60°N in the troposphere, which become more significant.

3.2. Wave-Mean Flow Interaction

[13] The westerly wind anomaly in the upper stratosphere and mesosphere changes the propagation conditions for planetary waves. In the work of *Matthes et al.* [2004], we showed that the positive feedback between zonal mean wind and waves leads to a poleward downward propagation of the west wind anomalies from the upper subtropical stratosphere and mesosphere in October and November to the midlatitude and high-latitude lower stratosphere in winter. We will now discuss how wave-mean flow interactions also alter the tropospheric circulation. Therefore the following discussion will focus only on signals in the middle (10 hPa) and lower stratosphere and in the troposphere.

[14] Figure 1b shows the long-term 10-day mean differences of the Eliassen-Palm Flux vector (EPF), which is a measure for the direction of planetary wave propagation, and its divergence, which is a measure for wave-mean flow interactions. In Nov1, there is a downward difference in the EPF at high latitudes (i.e., less upward propagation of planetary waves from the troposphere during solar max) from the upper into the middle stratosphere which intensifies and extends further down to 100 hPa in Nov2. Planetary waves propagating from the midlatitude troposphere upward to the stratosphere during NH winter are refracted in the vicinity of the west wind anomaly in November (see Figure 1a). Through a positive feedback between waves and the zonal mean flow, the west wind anomalies move poleward and downward with time. In Nov1, waves dissipate at lower altitudes (relative convergence at high-latitude middle stratosphere, Figure 1b) and decelerate the zonal mean wind in the midlatitude upper stratosphere (relative easterly winds, Figure 1a). Most of the upward propagating waves are refracted toward the equator (not explicitly shown here, see *Matthes et al.* [2004, Figure 6a]) where they dissipate and decelerate the zonal mean wind in the subtropical upper stratosphere (significant easterly wind anomalies in Figure 1a from Nov2 on). Therefore is less wave dissipation at high latitudes and the west wind anomaly can extend downward. The downward movement of EPF anomalies continues in Nov3 but weakens and turns into enhanced upward propagation of waves in Dec1. The enhanced upward propagation of waves in Dec1 is related to the stronger and more compact PNJ in Nov3 which guides the waves and leads to a weaker jet in the upper stratosphere/lower mesosphere and to a stronger PNJ in the lower stratosphere and troposphere for Dec1. From Nov3 onward, there is more wave propagation equatorward in the subtropical and midlatitude troposphere at the significant west wind anomaly and additional wave dissipation in the troposphere (relative convergence at 300hPa/70°N in Dec1). The simultaneous occurrence of stratospheric and tropospheric changes in Nov2 and Nov3 indicates the transition from dominant changes in stratospheric waves through Nov2 to dominant changes in tropospheric waves from Nov3 on.

3.3. Stratospheric and Tropospheric Circulation Patterns

[15] The time evolution of the anomalies in the stratosphere and troposphere implies a relation between anomalous stratospheric and tropospheric circulation patterns from November onward. This is supported by the evolution of the geopotential height differences from the middle stratosphere through the troposphere to the surface (Figure 2).

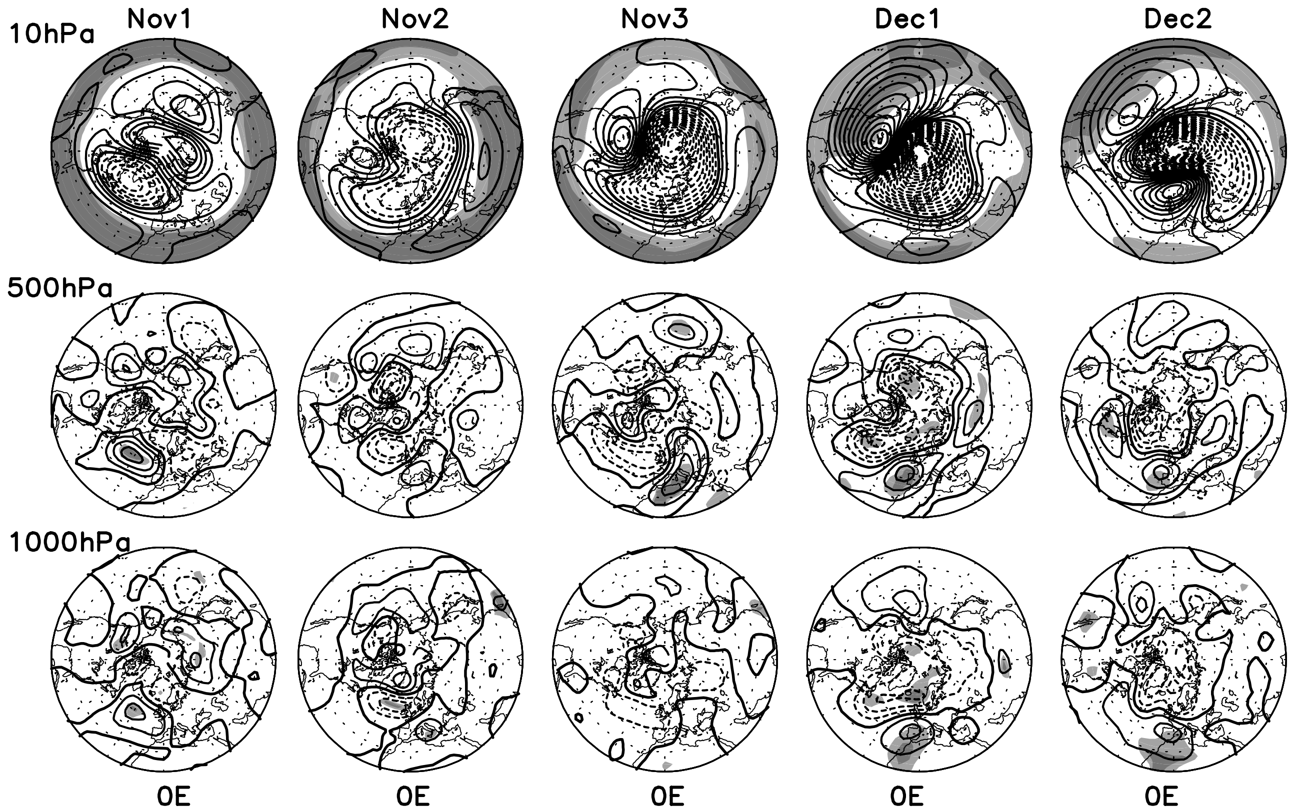

Figure 2. Polar stereographic projection from 20°–90°N of the long-term 10-day mean differences of the geopotential heights at 10, 500, and 1000 hPa (from top to bottom) between the solar max and min experiments from Nov1 (left column) to Dec2 (right column), contour interval: 20 gpm. Shading as in Figure 1a.

[16] The geopotential height differences at 10 hPa show a negative annular-like pattern (positive anomalies at the pole surrounded by negative anomalies) in Nov1 (Figure 2, top). In Nov3, the polar vortex during solar max is stronger, corresponding to the stronger PNJ. In Dec1 a strong wave-1-pattern develops due to the enhanced upward propagation of planetary waves in the middle stratosphere (Figure 1b). The wave-1-pattern changes into a positive annular-like pattern (negative anomalies at the pole surrounded by positive anomalies) that persist through Dec3.

[17] In Nov1, there is no similarity between the circulation differences in the stratosphere and those in the troposphere. This completely changes in December, when the differences in the circulation pattern in the stratosphere and troposphere resemble each other (Figure 2, middle and bottom). A deep equivalent barotropic spatial anomaly extends from the stratosphere down to the Earth's surface. The strongest anomalies are in the stratosphere and the magnitude decreases with decreasing altitude. Note that the statistical significances in the stratosphere are confined to lower latitudes. Some of the tropospheric anomalies are statistically significant at high latitudes, too.

[18] To show the evolution of the near-surface signal during winter, monthly mean differences of the geopotential height and temperature at 1000 hPa are displayed in Figure 3. In November there is no clear and significant surface signal in the geopotential height differences (Figure 3a). This changes in December into a well ordered and significant AO positive signal which still persists in January but weakens and starts to change sign in February. The preferred occurrence of a positive AO signal in December and January during solar max agrees with observations [*Kuroda and Kodera*, 2002, Figure 7]. The significant tropospheric and near-surface signal exists during the time of the modulation of the stratospheric PNJ, i.e., until March (not shown). The near-surface temperature differences (Figure 3b) correspond to the geopotential height changes. Significant positive surface temperature differences of the order of 2–3 K start to occur in December over NH land masses (Siberia and North America). From December until February the Siberian maximum disappears, while the maximum over North America intensifies.

4. Tropospheric High-Latitude Signal: Surface Pressure

[19] The temporal evolution of the zonal wind anomalies and the wave-mean flow interaction as well as the spatial evolution of the geopotential height and temperature differences discussed in the previous section imply that the tropospheric anomalies originate in the stratosphere. The fixed SSTs in the model also support the interpretation of a solar origin for the tropospheric signals but may also damp it. Here, we analyze in more detail the source of the forcings for changes in the meridional circulation (MC) and the surface pressure (SP) between solar max and min. Therefore

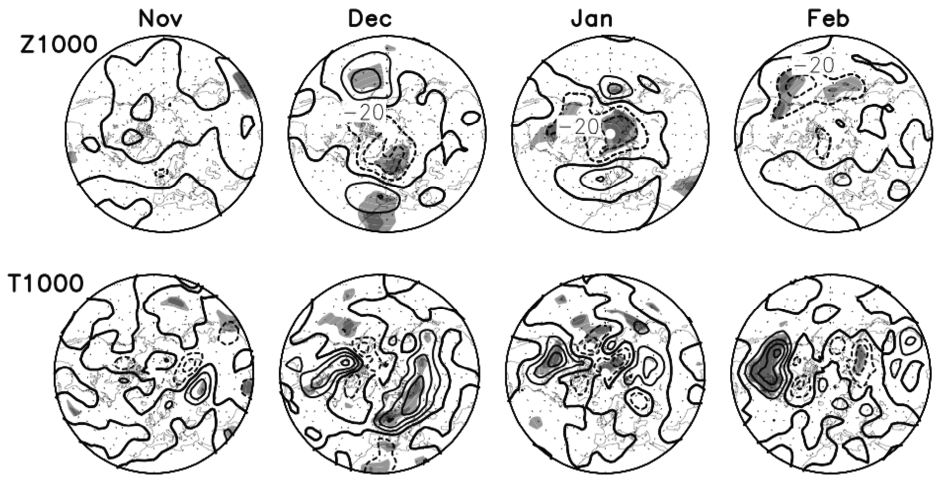

Figure 3. Polar stereographic projection from 20°–90°N of the long-term monthly mean differences at 1000 hPa between the solar max and min experiments for November, December, January, and February of (a) the geopotential heights, contour interval: 20 gpm, and (b) temperature, contour interval: 1 K. Shading as in Figure 1a.

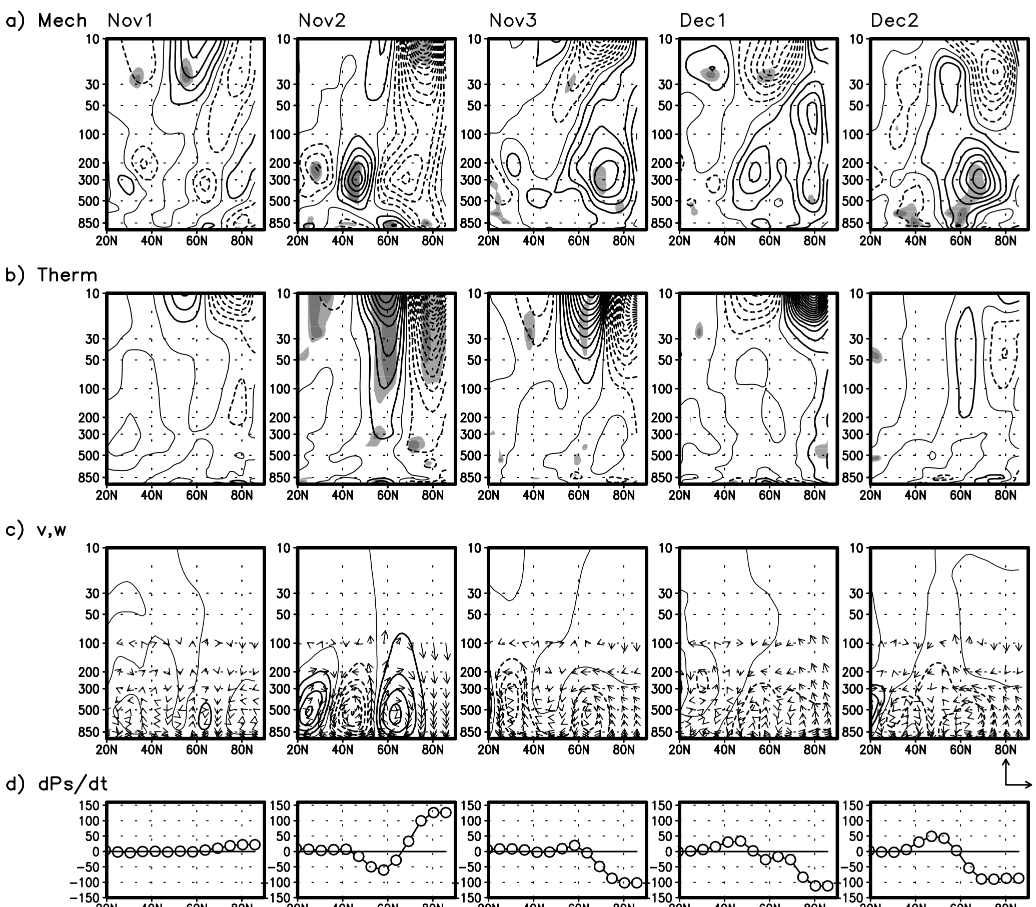

Figure 4. Long-term 10-day mean differences between the solar max and min experiments of (a) the mechanical and (b) the thermal forcing between solar max and min (contour interval: 0.2 m/s/day and 0.2 K/day), shading as in Figure 1a; (c) the difference in the mass stream function (contour interval: 10^9 kg/s) and the meridional circulation (arrows; horizontal reference arrow corresponds to 0.25 m/s, vertical component scaled with 200), and (d) the surface pressure tendency in Pa/day. The calculation is based on the total 10-day mean heat and momentum flux and therefore includes transient as well as stationary waves.

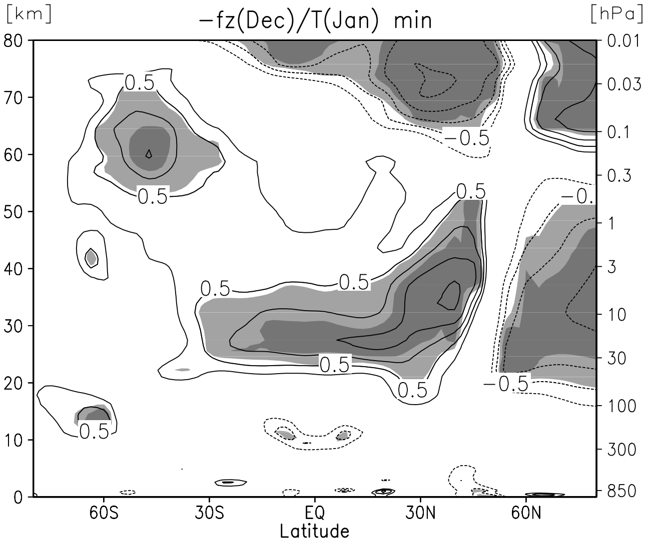

Figure 5. Correlation between the negative vertical component of the EPF at 60°N, 10 hPa in December and the zonal mean temperature field in January for the solar min experiment. Only correlations larger than 0.4 are shown, contour interval: 0.1. Shading as in Figure 1a.

the Eulerian Mean (EM) model on the sphere with a new application of the *Haynes and Shepherd* [1989] method is used as a diagnostic tool (see *Kuroda and Kodera* [2004] for details). The EM formulation is more appropriate to study lower troposphere and surface changes than the TEM formulation as it employs a less complex boundary condition. In the EM framework, the eddy forcing is supplied by the mechanical or momentum forcing, which is proportional to the meridional gradient of the northward eddy momentum flux, and the thermal forcing, which is proportional to the meridional gradient of the northward eddy heat flux. The eddy fluxes from the solar max and min experiments were taken to calculate with the described method changes in the thermal and mechanical forcing as well as changes in the MC and the SP (Figure 4). Figures 4a and 4b show the differences of the mechanical and thermal forcing between the solar max and min experiment together with the statistical significances. Figure 4c shows the anomalous meridional circulation (MC) and Figure 4d the anomalous SP tendency that is produced by the eddy forcings in Figures 4a and 4b. Note that in Figures 4c and 4d, contributions from other forcings (diabatic heating, friction, etc.) are not included and therefore the actual MC and SP anomalies may be slightly different. However, the calculated MC and SP tendencies (Figure 4) are good indicators for the actual anomalies (not shown).

[20] Whereas the thermal forcing anomalies (Figure 4b) show the largest values in the stratosphere, the mechanical forcing anomalies (Figure 4a) have two centers, one in the stratosphere (around 10 hPa, 60N), and the other one in the troposphere around 300 hPa. In Nov2, the statistically significant thermal forcing anomalies extend into the lower stratosphere and upper troposphere. Statistical significant anomalies of the mechanical forcing are mostly confined to the troposphere from Nov2 onward. Figure 4c shows that these eddy forcing anomalies induce an anomalous MC in the troposphere. The vertical velocity anomaly is downward if the thermal forcing anomaly is negative, and vice versa for an upward anomaly of the vertical velocity. The meridional velocity anomaly is poleward if the mechanical forcing is negative, and vice versa for an equatorward anomaly of the meridional velocity. In Nov2 a strong MC anomaly is seen with clockwise changes north of 60°N and anomalous downward motion at high latitudes as well as two smaller circulation anomaly cells around 40°N and 20°N. The strong relative downwelling at high latitudes during that time is related to a strong increase in SP tendency (Figure 4d), whereas the strong relative upwelling around 60°N is related to a decrease. From Nov3 onward, the changes in the SP tendency are gradual and develop into a negative signal at high and a positive signal at middle latitudes. The time evolution of the SP tendency corresponds well with the formation of the anomalous AO-positive signal in Figure 2. It should be noted that the response in Nov2 is exaggerated due to the use of 10-day mean data but does not affect the main conclusions.

[21] Additionally, the method is used to apply the wave forcing for the stratosphere and troposphere separately to distinguish their dominance in the surface pressure signal [*Kuroda and Kodera*, 2004]. Therefore eddy fluxes from the solar experiments below and equal to 300 hPa for all latitudes were taken to represent the troposphere and fluxes above and equal to 200 hPa were taken to represent the stratosphere. In Nov2, the stratosphere contributes one third to the changes shown in Figure 4d (not explicitly shown). The stratosphere therefore has an important effect on total SP changes when a deep MC anomaly is present. As expected, the troposphere contributes more, i.e. two third, to the total surface pressure changes. During the AO stage (Nov3–Dec2) most of the SP changes are related to changes in the tropospheric mechanical forcing. The stratospheric effect is small as expected from the shallow structure of the anomalous MC. The enhanced changes in the tropospheric mechanical forcing are related to zonal wind changes (Figure 1a) which affect the wave propagation and result in anomalous horizontal wave propagation (Figure 1b). The largest contribution to the described changes arises from stationary waves. These findings are in agreement with *Kuroda and Kodera* [2004] and provide further evidence that the stratosphere seems to trigger tropospheric changes. Our results also support the findings of *Black and McDaniel* [2004] who showed that a signal extending into the lower stratosphere and a preconditioning of the troposphere are needed for a stratospheric influence on the troposphere.

5. Tropospheric Low-Latitude Signal: Vertical Motion and Precipitation

[22] After investigating the significant high-latitude changes in the troposphere in December and January, we will now investigate possible changes in the tropics. Figure 5 shows the correlation between the negative monthly mean vertical component of the EPF at 10 hPa/60N for solar min in December, which is a measure for the wave activity in the stratosphere, and the January zonal mean temperature field. The statistically significant correlations in the stratosphere indicate that weaker wave forcing

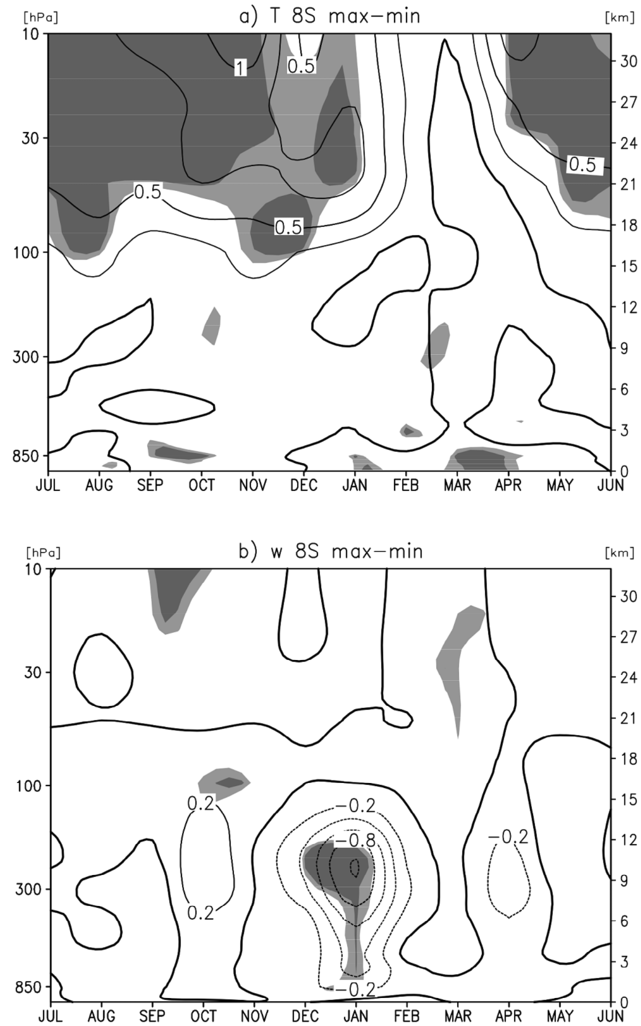

Figure 6. Monthly mean differences of the height-time section from June to July at 8°S between the solar max and min experiments of (a) the zonal mean temperature, contour interval: 0.25 K, and (b) the vertical velocity, contour interval: 0.2 mm/s. Shading as in Figure 1a.

at high latitudes corresponds to lower temperatures at high latitudes (relative upwelling) and higher temperatures at low latitudes (relative downwelling). During NH winter this corresponds to a weakening of the Brewer-Dobson (BD) circulation in the stratosphere. The effects are reversed in the mesosphere. The weakening of the BD circulation in the stratosphere in early winter during solar max [*Matthes et al.*, 2004] thus leads to a dynamically induced positive temperature anomaly in the tropical lower stratosphere seen in Figure 6a. The significant positive temperature anomaly in the tropical lower stratosphere is strongest from October until January when a stronger PNJ is present at high latitudes. With a weaker PNJ in February and March (not shown) this anomaly disappears.

[23] The short wave heating rate differences do not have a relative maximum in the tropical lower stratosphere (not shown) indicating that the relative temperature maximum is not due to the absorption of enhanced UV radiation. This further indicates that the relative warming of the tropical lower stratosphere is dynamically induced through a relative downwelling due to a weakening of the BD circulation. Concurrent negative zonal mean temperature anomalies occur in the tropical troposphere in December and January, statistically significant in some areas (not shown).

[24] The warming of the tropical lower stratosphere suggests an increase in the static stability and therefore a lowering of the tropopause [e.g., *Shepherd*, 2000]. Owing to the vertical resolution of the FUB-CMAM around the tropopause ($\sim$1 km), such small changes of the tropopause height cannot be resolved and calculated in the model. However, simplified GCM experiments from *Thuburn and Craig* [2000] support the above findings. They found a lowering of the tropopause as well as an influence on tropical convection when they artificially imposed a dynamical heating source in the tropical lower stratosphere.

[25] Figure 6b suggests that temperature changes in the tropical lower stratosphere are connected with vertical motion changes in the tropical troposphere. Changes in vertical motion are strongest and most significant in January, maximizing at 9 km. The negative changes at 8°S (maximum of the negative dipole anomaly, see Figure 7b) can be interpreted as a weakening of the absolute upwelling south of the equator. At the same time positive changes occur over the equator which indicate a weakening and broadening of the absolute downwelling. The timing of the largest and most statistically significant changes in solar max compared with solar min in lower stratospheric temperatures and tropospheric vertical velocities indicates that they are closely related.

[26] The large and statistically significant changes of vertical motions in the tropics in January (Figure 7b) occur simultaneously with changes in precipitation (Figure 7d) and cloud cover (not shown). During NH winter the Intertropical Convergence Zone (ITCZ) lies south of the equator with a maximum in upwelling and precipitation in January (Figures 7a and 7c). The ITCZ seems to be weakened and shifted northward in January. In general it moves less northward and southward and therefore has a more zonally uniform pattern during solar max years. The latitude-longitude differences in vertical velocity and precipitation patterns reveal that these changes are not uniformly distributed around the globe (Figure 8). The largest changes occur over the Indian Ocean and the western Pacific, areas which are covered by water. This confirms recent observational findings from *van Loon et al.* [2004] and *Kodera* [2004] which showed not only a solar influence on zonally symmetric features like the Hadley circulation in the tropics but also an influence on longitudinal motions like the Walker circulation.

6. Discussion and Summary

[27] To show the transfer of signals from the lower stratosphere to the troposphere, we investigated atmospheric circulation anomalies for solar max and min model experiments for easterly QBO phase only. In these experiments the solar signal is transferred from the upper stratosphere to the lower stratosphere through wave-mean flow interactions, leading to a poleward and downward movement of mean zonal mean wind and EPF anomalies [*Matthes et al.*, 2004]. Here we further investigated the transfer of signals

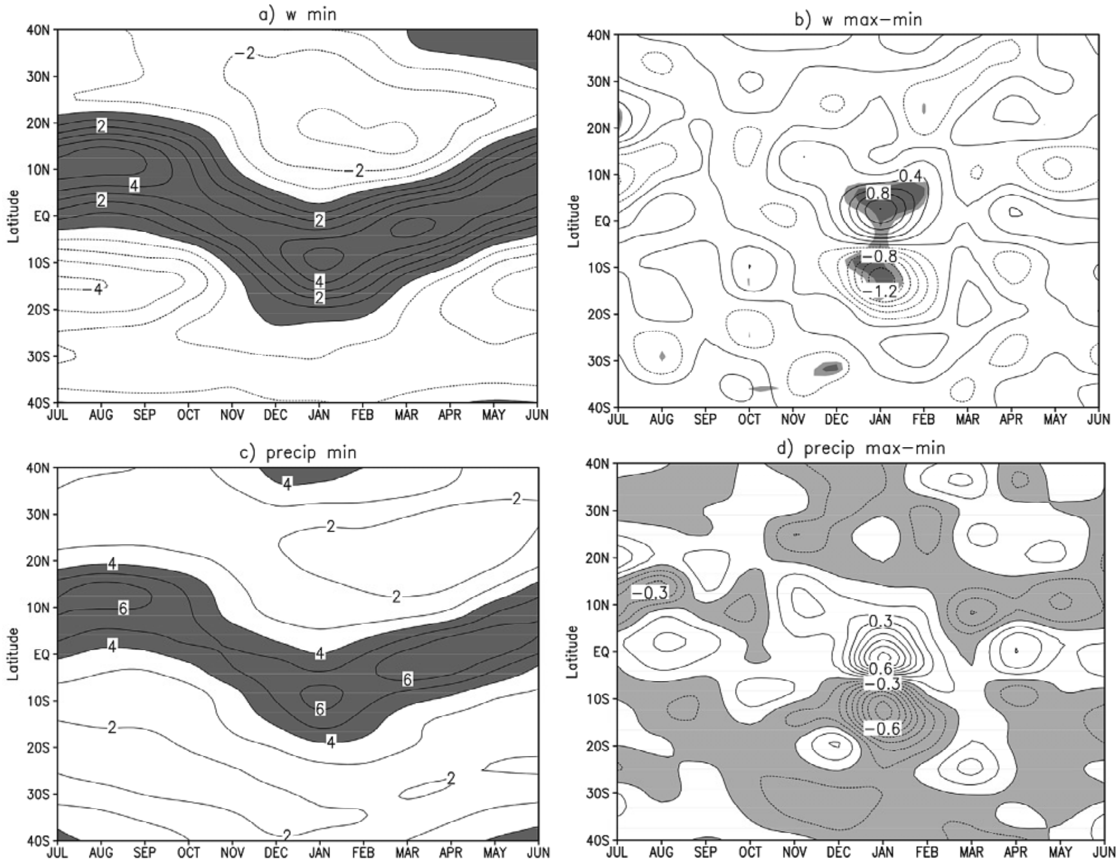

Figure 7. Monthly mean latitude-time sections from 40°S–40°N from July to June at 300 hPa (~8 km) of (a) the absolute vertical velocity field for the solar min experiment, contour interval: 1 mm/s, positive values are shaded; (b) the difference between solar max and min of the vertical velocity, contour interval: 0.2 mm/s, shading as in Figure 1a; (c) the absolute precipitation field for the solar min experiment, contour interval: 1 mm/d, values larger than 4 mm/d shaded; (d) the difference between solar max and min of the precipitation, contour interval: 0.1 mm/d, negative differences are shaded.

from significant stratospheric effects in October and November toward significant tropospheric effects in December and January.

[28] We suggest the following transfer mechanism: After the poleward-downward movement of zonal mean wind and EPF anomalies which begins in November (Figures 1a and 1b), a significant AO positive pattern develops in the troposphere in December during solar max (Figures 2, 3, and 4). The AO signal persists through January and then weakens and changes sign in February (Figure 3), when a weaker PNJ dominates the stratosphere. The appearance of the AO-like anomalies corresponds with the downward propagation of zonal mean wind anomalies (Figure 1a) and enhanced equatorward propagation of the EPF in the troposphere from Nov3 (Figure 1b), similar to what was seen by *Kuroda and Kodera [2004*, Figure 4]. The stronger polar night jet (PNJ) during solar max alters the propagation conditions for planetary waves: in October and November, there is less wave propagation upward from the troposphere. In December and January, there is more wave propagation equatorward at the westerly wind anomalies in the midlatitude troposphere (Figures 1a and 1b). At this time a deep equivalent barotropic structure exists throughout the stratosphere and troposphere (Figure 2). The effect of stratospheric and tropospheric eddy forcing changes on surface pressure was estimated using the EM equations with the new application of the *Haynes and Shepherd* [1989] model [*Kuroda and Kodera*, 2004]. The analysis showed that eddy forcing changes in the stratosphere and troposphere both contribute to tropospheric MC changes, which in turn induce SP tendencies at middle and high latitudes (Figure 4). The time evolution of the eddy forced SP tendencies corresponds well with the formation of the AO positive pattern in December and January. In Nov2, a deep MC anomaly is present and therefore one third of the SP changes can be attributed to stratospheric wave forcing changes. Most of the polar SP changes during the AO stage from Nov3 to Dec2 are related to tropospheric mechanical forcing changes. The largest part of the eddy forcing changes is due to stationary waves.

[29] These findings are in agreement with *Kuroda and Kodera* [2004]. However, the time evolution of solar induced anomalies in the model is, with the exception of the time after the AO formation, about two times slower than that of the PNJ oscillation in observations [*Kuroda and Kodera*, 2004]. Our results also support the results of *Black and McDaniel* [2004] who showed that a signal extending into the lower stratosphere and a preconditioning of the

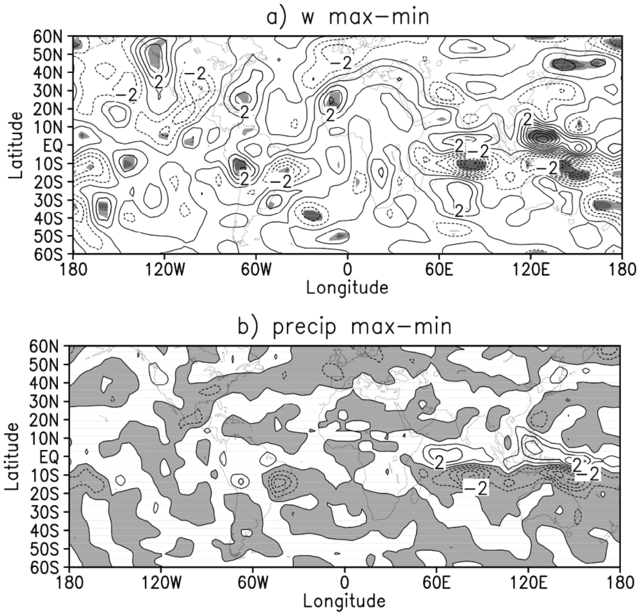

Figure 8. January monthly mean latitude-longitude sections from 60°S–60°N and 180°W–180°E at 300 hPa (∼8 km) of (a) the difference between solar max and min of the vertical velocity, contour interval: 1 mm/s, shading as in Figure 1a; (b) the difference between solar max and min of the precipitation, contour interval: 1 mm/d, negative differences are shaded.

troposphere are needed for a stratospheric influence on the troposphere.

[30] The transfer of the solar signal from the tropical lower stratosphere to the tropical troposphere is suggested to work as follows: Through changes in planetary wave propagation the BD circulation is weakened [*Matthes et al.*, 2004] and induces dynamical heating in the tropical lower stratosphere in December and January (Figures 5 and 6). The relative warming above the tropopause (Figures 6) suggests a lowering of the tropopause and increased stability in the troposphere [*Shepherd*, 2000], which in turn influences vertical motions, precipitation, and convection patterns in the tropical troposphere (Figures 7 and 8) in agreement with findings by *Thuburn and Craig* [2000], *Gleisner and Thejll* [2003], *van Loon et al.* [2004], and *Kodera* [2004]. The largest signal in the tropical troposphere appears in January, approximately 2 months after the largest signals in the upper stratosphere. Changes in the tropical troposphere are not uniformly distributed with longitude but show largest changes over the Indian ocean and western Pacific (Figure 8), indicating an influence on the Walker circulation as shown in observational studies [*Gleisner and Thejll*, 2003; *van Loon et al.*, 2004; *Kodera*, 2004].

[31] However, in observations the strongest effect in the tropical troposphere on the Hadley and the Walker circulations is observed during SH winter (July/August) [e.g., *Labitzke and van Loon*, 1988; *van Loon et al.*, 2004; *Kodera*, 2004] and not during NH winter (January) as in the model. This can be explained in two ways. First, the transfer of the solar signal in the model works only during NH winter [*Matthes et al.*, 2004]. During SH winter, stratospheric winds are unrealistically strong and suppress the small initial solar signal. Therefore the BD circulation is not modulated during SH winter and changes in the tropics do not occur. On the other hand, changes in observed tropical circulation patterns during NH winter could be masked by other processes such as the El Niño Southern Oscillation phenomenon or volcanic eruptions. We estimated a pattern of vertical velocity changes between solar max and min years from observational data where the ENSO effect was eliminated. These changes are very similar to that in the model (not shown). Owing to the limited amount of observational data and the difficulty in extracting the vertical velocity in the tropics this needs to be further investigated.

[32] Furthermore, the model results support several recent model and observational studies of troposphere-stratosphere coupling. They provide further evidence of an influence of stratospheric circulation pattern on the troposphere and surface climate as discussed, e.g., by *Baldwin and Dunkerton* [2001], *Thompson et al.* [2002], *Norton* [2003], *Baldwin et al.* [2003], and *Kuroda and Kodera* [2004].

[33] In summary, our results highlight the importance of stratospheric circulation changes for middle to high latitudes and tropical tropospheric changes. The SSTs in the two model experiments are identical, which means that changes seen in the troposphere are caused by indirect effects induced through direct changes in the upper stratosphere. To test our findings and to investigate the tropospheric signals in more detail, future model studies should include a fully coupled ocean-atmosphere module to allow a feedback between atmosphere and ocean. This would enhance the tropospheric variability and its response to the solar signal.

[34] **Acknowledgments.** We thank M. Kunze for the assistance in data handling and the development and maintenance of the diagnostic programs. We thank A. Smith and two anonymous reviewers for valuable comments. This study has been funded by the European Commission under the contract EVK2-CT-1999-00001 (SOLICE) and the Grand-in-Aid (17340143) for Scientific Research of the Ministry of Education, Culture, Sports, Science, and Technology of Japan and is also part of the Ph.D. thesis of K. Matthes. The FUB-CMAM was run at the Konrad-Zuse-Zentrum für Informationstechnik, Berlin.

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