Influence of Solar Activity Cycles on Earth's Climate

Task 5 – Hypothetical Physical Mechanisms

WP 501b – Role of UV and energetic particles

Joanna D. Haigh
Imperial College London

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1 Introduction

In this Work Package the physical mechanisms whereby changes in spectrally resolved solar irradiance may affect climate are discussed. Both direct radiative effects and indirect dynamical effects, produced in response to the solar heating, are considered. Radiative heating by ultraviolet radiation depends on stratospheric ozone concentrations so the potential implications of the effects of solar proton events on ozone are also briefly discussed.

2 Radiative heating

The processes whereby solar radiation is absorbed by, and heats, the atmosphere were outlined in Section 4.2.1 of the report on WP104. Here we consider the effect on heating rates of increases in radiative input when the Sun is more active. Figure 1 (upper panels) shows vertical profiles of the solar fluxes and heating rates for solar minimum conditions resolved into three spectral regions (ultraviolet, visible and near infrared). The magnitude of the incoming UV flux is much smaller than in the other two regions but its absorption by ozone causes the largest heating rates throughout the atmosphere. The weaker absorption of visible radiation in the lower stratosphere and of near infrared radiation in the troposphere give much smaller heating rates.

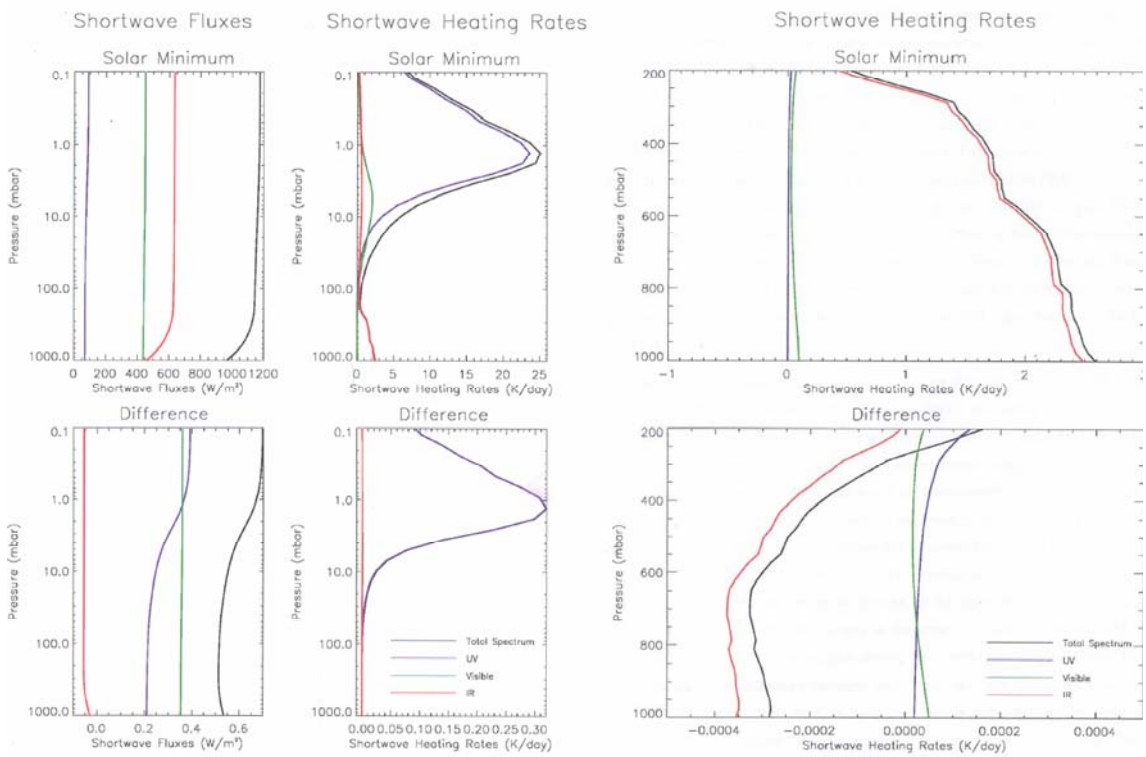

Figure 1 Solar fluxes and heating rates for an overhead Sun as a function of altitude (log pressure). Solar minimum conditions (above) and difference between solar maximum and solar minimum (below) at ultraviolet (blue curves), visible (green curves) and near infrared (red curves) wavelengths. Fluxes (W m^{-2}, left), heating rates (K d^{-1}, centre and, on an expanded scale as a function of pressure, right). Spectral data from Judith Lean (personal communication), calculation from *Larkin* (2000).

The lower panels of Figure 1 show the differences in fluxes and heating rates between 11-year solar cycle (SC) minimum and maximum conditions. At the top of the atmosphere the increases in incoming flux in the visible and UV regions are of similar magnitude (see also Figure 7.4 of the WP104 report) but the stronger absorption of UV produces much larger heating rates, with the extra visible radiation largely passing through to the Earth's surface. The same data are shown with a linear pressure scale for the ordinate (to emphasise the troposphere) in the two panels on the right hand side of Figure 1. [N.B. The spectral changes prescribed for these calculations were such that the near-infrared radiation was reduced at solar maximum giving decreases in heating rate; this is contentious but the changes are anyway very small being about one part in ten thousand.]

Clearly much larger variations of within-atmosphere heating rate, related to the 11-year cycle in solar irradiance, occur in the stratosphere than in the troposphere. Heating rates in the troposphere are much smaller but the variability here is less well known due to uncertainties in the near-infrared solar spectrum. The UV heating rates depend not only on the incident solar flux but also on the concentration of ozone which absorbs it. Thus changes in ozone will modify the distribution of absorbed energy and feedback on the heating rates. All these factors must be accurately represented when assessing the potential for climate to be influenced by variations in solar irradiance.

3 Stratospheric ozone

3.1 Response to changes in UV radiation

As discussed in section 7.3 of WP104, ozone is produced by short wavelength solar ultraviolet radiation and destroyed by radiation at somewhat longer wavelengths. The amplitude of solar variability is greater in the far ultraviolet (Fig.7.4 of WP104) so that ozone production is more strongly modulated by solar activity than the destruction mechanisms and thus there are higher concentrations of stratospheric ozone during periods of higher solar activity. Both observational records and model calculations show approximately 2 % higher values in ozone columns at 11-year cycle maximum relative to minimum. However, as discussed in section 7 of WP403, there are some discrepancies between satellite observations and model predictions in the vertical and latitudinal distributions of the response.

3.2 Response to Solar Proton Events

The chemical composition of the atmosphere also responds to the incidence of high energy solar particles. Precipitating electrons and solar protons affect the nitrogen oxide budget of the middle atmosphere through ionization and dissociation of nitrogen and oxygen molecules. NO catalytically destroys ozone so reductions in ozone concentration are seen in response to particle events. As the solar particles follow the Earth's magnetic field lines these effects occur primarily at high latitudes but ozone depletions may propagate to mid-latitudes and down to the middle stratosphere for particularly energetic events (see example in Figure 2). It is interesting to note that the effect of energetic particle events on ozone is in the opposite sense to that of enhanced ultraviolet irradiance. As particle events are more likely to occur when the Sun is in an active state the combined effect on ozone may be complex in its geographical, altitudinal and temporal distribution.

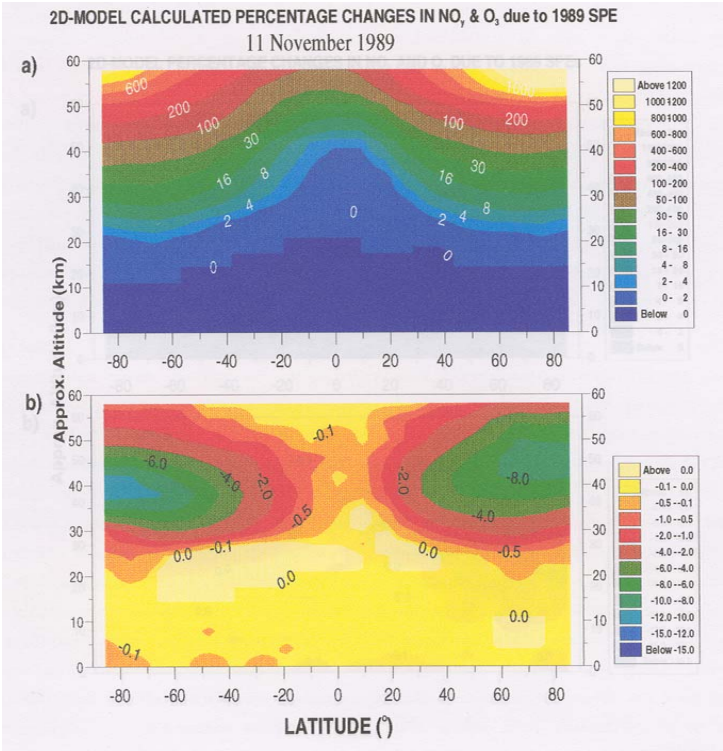

Figure 2 Latitude-height section of percentage change in NO_y (top) and O_3 (bottom) calculated using a 2D model for November 1989, following the major solar proton event of October 1989. (Vlachogiannis and Haigh, 1998)

4 Temperature response

Increases in the solar radiative heating, including those due to the ozone variations, clearly produce a tendency to warm the atmosphere. This is expressed as the diabatic heating term, Q, in the thermodynamic equation (WP104 eq. (2.5)) which may be written approximately as:

$$\frac{\partial T}{\partial t} = Q - \boldsymbol{u}.\nabla T - \frac{g}{C_p} w \qquad (1)$$

where $\boldsymbol{u}$ is the vector velocity and w its vertical component. Enhanced solar heating (an increase in Q) will tend to warm the atmosphere locally but the actual temperature change depends also on dynamical factors, through advection and adiabatic expansion (the second and third terms on the right hand side of eq.(1) respectively). The changes in wind are themselves coupled to temperature changes (see WP104 section 2.1), and the advection term incorporates heat fluxes due to the mean circulation of the atmosphere and also eddies (waves), so the temperature response at any particular place will depend on non-local heating rates.

For this reason it is likely that at least part of the climate response to enhanced solar ultraviolet radiation will be manifest through dynamical feedbacks. And, indeed, the patterns of temperature response to solar activity derived in section3 of WP403 cannot be explained by purely radiative effects. The necessity of correctly representing the dynamical effects also offers an explanation as to why model simulations of changes to the middle atmospheric temperature (and ozone) structure have varied widely and had only limited success in reproducing signals derived from observational data (see WP104 sections 7.2 and 7.3).

We now consider mechanisms whereby dynamical feedbacks may act to amplify, at least in certain regions, the radiative effects of solar variability. Specifically we suggest that the relatively large heating of the middle atmosphere by solar UV radiation may produce a climate signal at lower altitudes through dynamical coupling, as discussed by Gray et al (2005). We consider first how solar radiative heating of the tropical stratopause region may influence the polar lower stratosphere. Secondly we consider dynamical mechanisms whereby solar heating (either direct radiative or indirect dynamical) of the lower stratosphere may influence circulations in the troposphere.

5 Dynamical coupling within the middle atmosphere

5.1 Planetary wave propagation and solar/QBO interaction

A series of papers (Kodera and Kuroda 2000a,b, 2002, Kuroda and Kodera 2002, Kodera et al. 1990, 2000, 2003) highlighted a slowly-propagating zonal wind anomaly, in response to solar activity, that moves poleward and downward throughout the winter months in each hemisphere (see Figure 7.7 of WP104). Kodera (1995) had already shown that a similar zonal wind anomaly is also seen when winters are divided between QBO east and west phase, instead of solar minimum and solar maximum years.

Early modelling work by Rind and Balachandran (1995) and Balachandran and Rind (1995), and more recently discussed by Rind et al. (2002), was able to simulate these zonal wind anomalies. They suggested that solar variability influences the structure of the polar night jet and hence the propagation of planetary-scale waves that travel vertically from the troposphere. This then affects their ability to impact the polar vortex and to produce sudden stratospheric warmings. Specifically, Rind and co-workers noted that the 11-year SC temperature anomaly in the equatorial upper stratosphere gives rise to an anomalous horizontal temperature gradient and hence to a corresponding anomaly in the vertical wind shear in the region of the polar night jet at upper levels. As a result of the consequent anomalous planetary wave propagation, this zonal wind anomaly gradually descends with time into the lower stratosphere (see also Dunkerton 2000). In addition, they noted that the QBO influences the latitudinal wind shear in the lower stratosphere (Holton and Tan 1982). Both these factors affect the structure of the polar night jet and thus there is an interaction of the solar and QBO influences through their combined influence on wave propagation. However, the details of how the solar and QBO interaction occurred were not clear, especially the precise mechanism by which the 11-year SC influence in the upper stratosphere impacts the QBO influence in the lower stratosphere.

More details of this SC/QBO interaction has been provided in a series of mechanistic modelling studies (Gray et al. 2001b, 2003, 2004, Gray 2003). They confirmed that the propagation of planetary scale waves and the development of NH sudden stratospheric warmings is very sensitive to the zonal flow in the early winter upper stratospheric equatorial/subtropical region. They interpreted the zonal wind anomaly identified by Kodera and co-workers as the manifestation of the Aleutian High which is a common pre-cursor to a sudden warming. Their model suggested that the easterly anomaly at upper levels associated with solar minimum years will tend to speed up the development of the Aleutian High and the subsequent sudden warming, resulting in an early-to-mid winter warming, while the westerly anomaly associated with solar maximum years will slow it down, resulting in a mid-to-late winter warming. This is also consistent with a recent interpretation of the zonal wind anomaly by Kodera and Kuroda (2002a). They noted that in a climatological-mean state the

circulation of the upper stratosphere evolves during the winter from a radiatively controlled state to a dynamically controlled state (i.e. via planetary wave propagation from the troposphere). The transition is characterised by the poleward shift of the subtropical westerly jet. They suggest that there is a solar influence on this transition time, which is earlier in solar minimum, thus creating the easterly anomaly in the subtropics as the jet axis moves poleward.

Gray (2001a) also noted that in addition to the well-known QBO in the low and mid equatorial stratosphere (Baldwin et al. 2001) the QBO signal actually extends as high as the upper stratosphere (and possibly beyond). There are two phase changes with height which means that the QBO near the stratopause (~50 km) has the same phase as the lower stratospheric QBO at 20-30 km, with a much smaller amplitude of around 5-10 ms^{-1} (Pascoe et al. 2005). This amplitude is similar to the solar zonal wind anomaly in the same region and Gray et al. (2004) have proposed that there is a QBO influence on sudden stratospheric warmings from the upper stratosphere in addition to one from the lower stratosphere. They proposed that this can explain the observed interaction of the solar and QBO influences on the NH winter (see Labitzke and van Loon 1987) which gives rise to a reversal of the Holton-Tan effect in solar maximum years. Both solar and QBO anomalies in this region influence planetary wave propagation and their effects can either reinforce or cancel each other, thus influencing the timing of the SSWs. In solar minimum / easterly QBO phase years the easterly anomalies reinforce and speed up the onset of a sudden stratospheric warming. In solar maximum/westerly QBO phase years the westerly anomalies reinforce to slow down the development of the warming.

Salby and Callaghan (2004) have noted an 11-yr SC modulation of the frequency of the QBO in the lower stratosphere and this has been successfully simulated in a 2D model by McCormack (2003). It is therefore possible that the observed interaction of the 11-yr SC and QBO influence on NH sudden warmings occurs through the equatorial lower stratospheric winds, without any additional influence from the equatorial upper stratosphere. On the other hand, Matthes et al. (2004) found that in order to achieve a simulation of the solar/QBO signal interaction in their model they had to impose a QBO that extended throughout the whole depth of the stratosphere and not just the lower stratosphere, suggesting that an interaction near the stratopause region has some role to play.

5.2 Solar signal penetration to the lower stratosphere

The sensitivity of planetary wave propagation in winter to the background temperature and wind structure, as described above, provides a mechanism for the solar signal to penetrate to the lower stratosphere, although the precise details of how that influence is achieved is still not well understood. Sudden stratospheric warmings initially commence in the upper stratosphere and extend downwards until they encompass the whole vertical extent of the stratosphere, thus directly changing the temperature and circulation in the polar winter lower stratosphere.

During the winter, the same planetary wave activity induces a large-scale meridional circulation (Haynes et al. 1991) that is strengthened during the more disturbed periods of stratospheric warming events. The meridional circulation is therefore a prime route for winter polar events to influence the lower stratosphere not only in polar latitudes but throughout the winter hemisphere and even the equatorial and summer subtropical latitudes, through a modulation of the strength of equatorial upwelling.

A dynamical feedback via the meridional circulation would serve to amplify the solar signal since the less disturbed early winter conditions in solar maximum will lead to a weaker meridional circulation, weaker equatorial upwelling and hence a warmer equatorial lower stratosphere in solar maximum than solar minimum. This mechanism might explain the secondary temperature maximum in the lower stratosphere shown in observational analyses (see Figure 7.5 WP104 and Figure 2 WP403). It might also produce a second ozone maximum in the lower stratosphere, as observed, since at these heights ozone has a long lifetime and is dynamically controlled. Solar modulation of the equatorial upwelling in this way is also likely to interact with the QBO which is the dominant signal of variability in the equatorial lower stratosphere, and thus may be the cause of the solar modulation of the frequency of the QBO observed by Salby and Callaghan (2004). The meridional circulation also extends into the summer subtropics and may therefore be a route for amplification of the summer hemisphere solar signal.

6 Stratosphere-troposphere dynamical coupling

It is feasible that a solar signal in lower stratospheric temperature and circulation will impact the troposphere below. A number of mechanisms have been discussed, not only in the context of solar influence but in the wider context of examining to what extent stratospheric processes might influence tropospheric weather and climate. A general overview of stratosphere troposphere coupling has been provided by Shepherd (2002). Here we consider separately approaches focusing on polar and tropical processes.

6.1 Polar influence

It has been suggested that variations in the strength of the stratospheric polar vortex influence surface climate, most often expressed as a modulation of the AO/NAM/NAO in the Northern Hemisphere (Baldwin and Dunkerton 1999, 2001) or SAM in the south (Thompson and Wallace 1998, Thompson et al, 2005), see the discussion in section 5 of WP403. Figure 3 presents an analysis of NCEP geopotential height data which suggests that anomalies in the strength of the northern hemisphere winter polar vortex propagate down through the stratosphere to at least the tropopause and possible to the surface. It has been pointed out by Plumb and Semeniuk (2003) that the apparent downward migration does not necessarily indicate any controlling influence of the stratosphere on the troposphere but modelling studies (see below) have shown that perturbations forced in the stratosphere may be manifest in changes at lower altitudes.

Nevertheless, there is as yet no consensus on the mechanism of this stratospheric influence on the troposphere and surface (Baldwin and Dunkerton 2005). A simple transfer of pressure anomalies in the stratosphere directly to the surface can perhaps explain some of the influence (Black, 2002); wave reflection in the upper stratosphere is another possibility (Perlwitz and Harnik 2003). Shindell et al. (2001) and Kodera (2002b) and Castanheira and Graf (2003) propose a similar modulation of planetary wave propagation to that described in section 5.1 in the upper stratosphere, but operating in the very lowermost stratosphere and thus also influencing the troposphere.

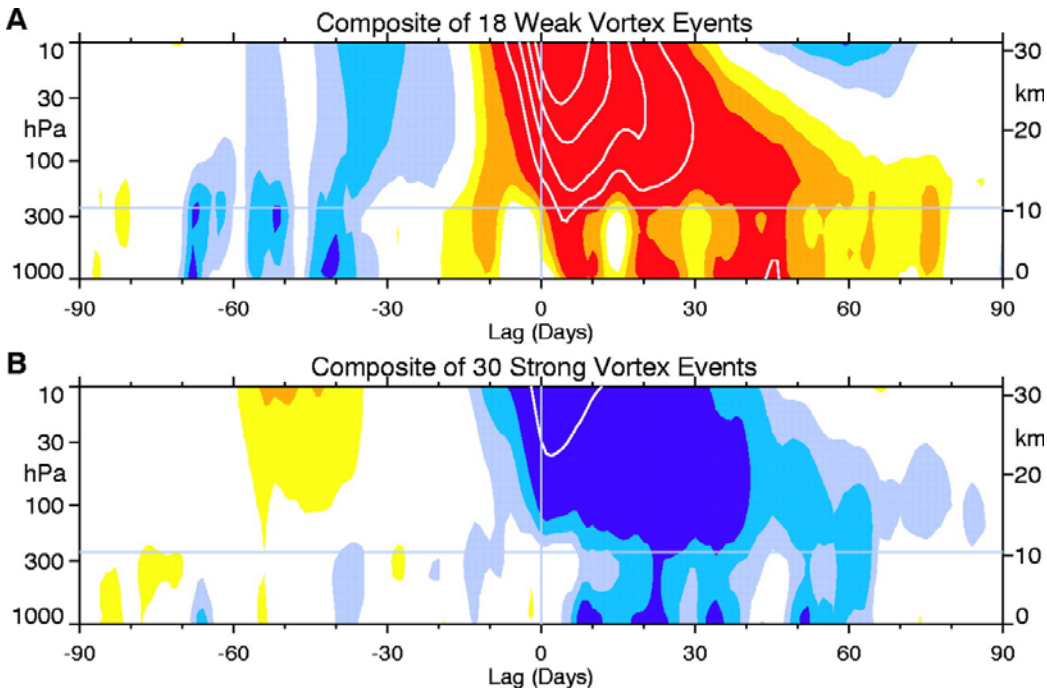

Figure 3 Composite of time-height development of the NAM for (A) 18 weak vortex events and (B) 30 strong vortex events. Blue corresponds to high NAM index (strong vortex, cold pole) and red to low NAM index (weak vortex, warm pole). (Baldwin and Dunkerton, 2001)

6.2 Tropical influence

As discussed in WP501a??, some model studies have shown changes in the tropical tropospheric Hadley circulation in response to varying TSI via thermodynamic processes. Those studies, however, used a spectrally flat increase in irradiance so that preferential warming of the stratosphere could not occur. Other model studies, in which the UV component is enhanced, and stratospheric ozone increased, produce rather different changes in the Hadley cells. As first shown by Haigh (1996), using an AGCM with fixed SSTs, solar heating of the tropical lower stratosphere causes the Hadley cells to weaken and broaden, the sub-tropical jets to move poleward and vertical banding of temperature anomalies in mid-latitudes when the sun is more active. This signal, which is very similar to that seen in NCEP data analysis by Haigh (2003) and Gleisner and Thejll (2003), has been reproduced in different models by Larkin et al (2000), Palmer (2002) and Matthes (2003).

Salby and Callaghan (2005), in a study of the NCEP reanalysis dataset, show that anomalous warming in the tropical lower stratosphere is correlated with cooling in the polar lower stratosphere, consistent with a weakening of the Brewer-Dobson circulation. They go on to show that the signatures of anomalous vertical motion represent changes which operate coherently across the tropopause, coupling the stratosphere and troposphere through a transfer of mass. This modifies vertical motion inside the tropical troposphere and influences organized convection. They find that much of the variability is associated with the QBO, modulated by the solar cycle.

These modelling and data studies are intriguing because they suggest that a solar influence primarily in the stratosphere may be indirectly responsible for the solar signal detected in the troposphere. (Although this does not, of course, preclude the possibility of responses to other solar forcing factors such as TSI (see WP501a) or GCRs (WP503).)

However, while the apparent response to UV is indicative of a stratosphere-troposphere coupling mechanism, it does not explain the detail of how such effects take place. Experiments with a simplified GCM (sGCM) have started to provide some indication as to how these responses may arise (Haigh et al 2005). In these experiments idealized heating perturbations were applied only in the stratosphere but effects were shown throughout the troposphere with vertical banded temperature anomalies and changes in circulation. Heating the lower stratosphere increases the static stability in this region, lowers the tropopause and reduces wave fluxes here. This leads to coherent changes through the depth of the troposphere, involving the location and width of the jetstream, storm-track and eddy-induced circulation. The success of the sGCM in reproducing qualitatively the observed tropospheric response to imposed stratospheric perturbations also suggests that moist feedbacks are not a crucial component of the response: it is eddy/mean-flow feedbacks that are the primary mechanism. Furthermore, despite the presence of a uniform stratosphere, the lack of a stratospheric polar vortex and use of broad latitudinal-scale perturbations the sGCM is able to reproduce the tropospheric patterns. This suggests that a detailed representation of the stratosphere may not necessary for simulating the tropospheric aspects of solar influence, although the nature and sources of the stratospheric heating remain important factors.

7 Summary

The effect on the atmosphere of variations in solar irradiance is, prima facie, a response in radiative heating. The physical basis of this is well understood and uncertainties in heating rate estimates lie in the specification of the input data. The least well known component of variability in the solar spectrum lies in the near-infrared which has greatest impact on tropospheric heating rates. The ultraviolet spectrum has been better measured and thus variations in the direct solar heating of the middle atmosphere are fairly well understood. However, the response of the ozone distribution to UV variability is not well established and this implies an uncertainty in heating rates. The response of ozone to intermittent solar proton events will further complicate the picture although it is dubious whether their effects would be of sufficient magnitude to significantly affect heating rates and temperatures in the lower stratosphere.

That solar UV variability impacts the middle atmosphere is not questioned. However, the direct impact of radiative heating within the atmosphere is unlikely to be responsible for the (statistically) observed effects of solar variability on the lower atmosphere. Much less well understood than the direct effects of solar irradiance are their indirect effects on the lower atmosphere through modification of atmospheric dynamics. While there is no consensus on the details of the mechanisms involved, current research is suggesting various potential avenues whereby anomalous solar heating of the middle atmosphere may result in a response in tropospheric climate. These mainly involve changes in wave momentum transport and deposition and the subsequent modification of atmospheric circulations. Further modelling studies are required to elucidate the exact mechanism of influence, to determine the relative influence and coupling mechanisms between the various atmospheric levels and to make further improvements to model simulations of these effects. Potential modelling studies are the subject of Work Package 6.

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