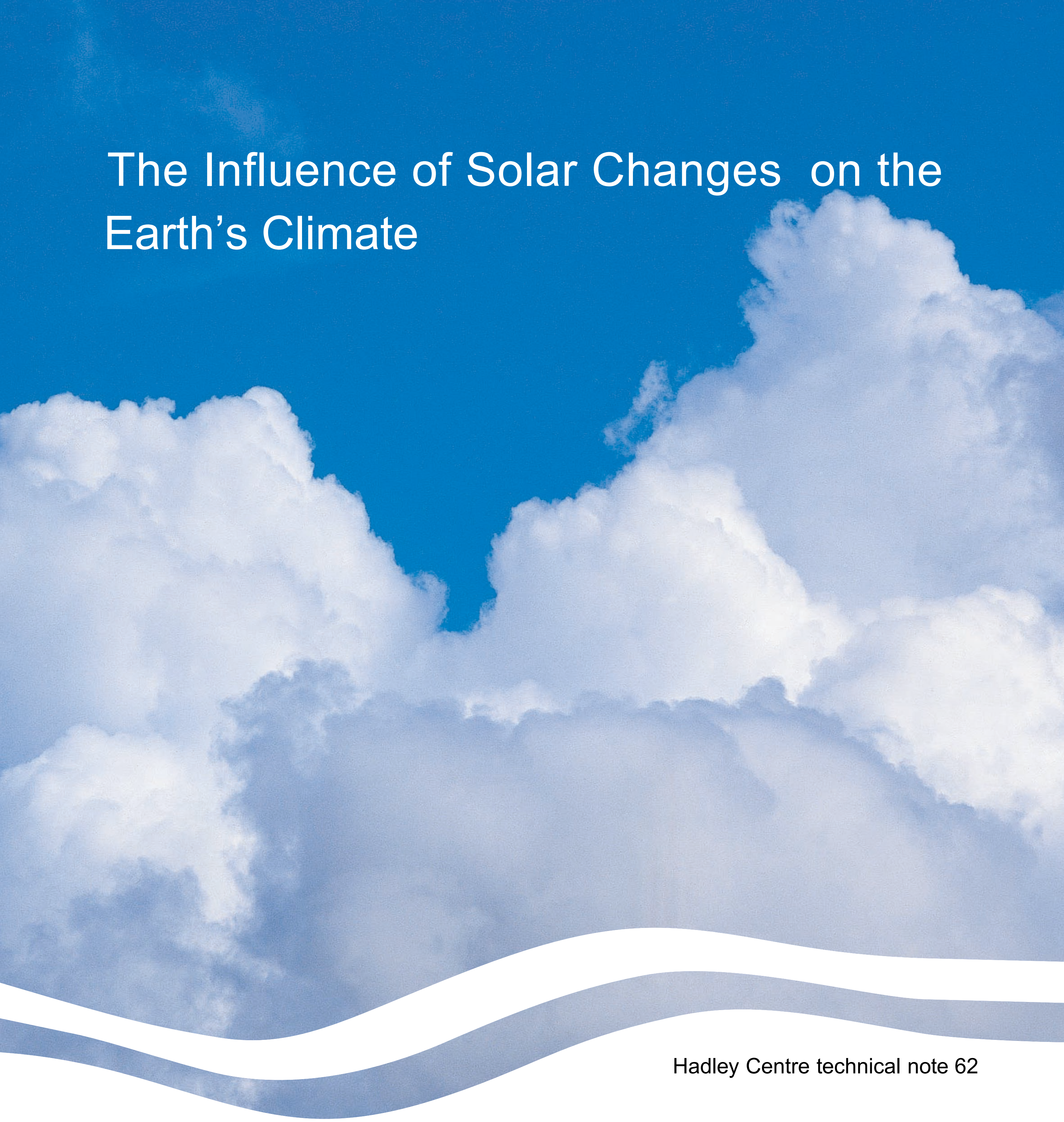
The Influence of Solar Changes on the Earth's Climate



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A Review of

The Influence of Solar Changes on the Earth's Climate

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EXECUTIVE SUMMARY

This report provides a review of the major advances in this field since the previous report by Harrison and Shine (1999).

- **The existence of an underlying trend in total solar irradiance (TSI) over the past two solar cycles is still uncertain. New work on the relationship between solar magnetic activity and irradiance has reduced our confidence in the reliability of reconstructions of past TSI.**
- **Additional observations through the last solar maximum, during which there was no major volcanic activity, have confirmed the solar response found previously in the middle atmosphere. Models are able to reproduce the gross features of the solar response but do not reproduce details, including the response in the lower stratosphere, that may be important for the transfer of the signal to the surface.**
- **The statistical robustness of some observed solar signals in the troposphere has been strengthened. Mechanisms proposed to explain these signals involve stratosphere – troposphere coupling, though details of this coupling are not well understood.**
- **Galactic cosmic rays have been shown to be closely correlated with continuous satellite (ISCCP) retrievals of low cloud from 1983-1994, and possibly to 2001. Modelling and observations now support atmospheric production of ultra-fine aerosol from cosmic ray produced ions. Establishing effects of cosmogenic ultrafine aerosol in the atmosphere is necessary before suggested links between the satellite analysis and the aerosol microphysics can be confirmed.**
- **Theory shows that charged aerosols are preferentially removed by cloud droplets, presenting the possibility of a long-range influence through the global atmospheric electrical circuit.**

Detailed summary:

Reconstructions of Past Total Solar Irradiance (TSI)

- Comparison of measurements from different satellite observing systems shows that the present day value of total solar irradiance is only known to approximately 0.15% (~2 W m^{-2}). Its variation over the past two solar cycles is known, with greater accuracy, to be ~0.08% (~1.1 Wm^{-2}) but difficulties encountered in the inter-calibration of the instruments mean that the existence of any underlying trend in TSI over the past two solar cycles is uncertain.
- Most reconstructions of past values of TSI rely on (i) empirically-derived relationships between proxy indices of solar activity and measurements of TSI and (ii) assumptions concerning long-term secular changes in the "quiet Sun". The former are fairly well established and, to a certain extent, can be reproduced by physical models. The latter are not well known and continue to be highly controversial.

Observations of solar response: 11-year solar cycle

- Updated observational analyses have confirmed earlier 11-year solar cycle signals in zonally averaged stratospheric temperature, ozone and circulation with increased statistical confidence. The addition of the past few years of data have been especially important since they have provided observations through a solar maximum period without any major volcanic activity. This has enabled the solar cycle signal to be more effectively distinguished from the volcanic signature.

- There are several differences between the various observational analyses that need to be resolved. These include (i) differences in analysed temperature signal from the SSU dataset, most probably caused by difference corrections for instrument drift, which need to be resolved (ii) the presence of a negative response in the equatorial low / mid stratosphere in some of the temperature and ozone analyses but not in others. This may be a real signal, but if so the differences between the datasets needs to be resolved, or it may be an artefact of the linear regression technique employed.
- Details of the 11-year solar signal in tropospheric circulations have been confirmed. The analyses show that the subtropical tropospheric jets are weaker and further poleward at solar maximum. Vertical bands of warming, of amplitude about 0.5K, extend throughout the troposphere to the surface. One analysis has also shown a warmer, moister atmosphere in solar maximum, with the response being strongest in latitudinal bands of deep tropical convection and mid-latitude cyclone activity.
- Correlation analyses of the North Atlantic Oscillation (NAO) index and sea level pressure fields suggest a solar modulation of the pattern of correlation. In solar minimum the familiar dipole pattern is present over the Atlantic / European sector while in solar maximum the pattern extends across into the Eurasian sector.

Mechanisms: Direct TSI Changes

- Uncertainties in the reconstruction of historical TSI mean that the value of solar radiative forcing from 1750 to the present is even less well established than was perceived for the IPCC (2001) TAR but, based on the current consensus that the secular trend in TSI is smaller than previously assumed, it may be appropriate for the value to be reduced from the TAR value of 0.30 Wm^{-2} to about 0.28 Wm^{-2}. Stratospheric adjustment of this value is not well known because of uncertainties in the vertical profile of solar-induced ozone changes.
- Models are generally able to reproduce, within the bounds of observational uncertainty, natural internal variability and model noise, the global mean surface temperature variation from 1860 to the present day if both natural (solar and volcanic) and anthropogenic (greenhouse gas and sulphate aerosol) forcings are included. Largest discrepancies occur during the first part of the twentieth century when increasing solar irradiance may be influential.
- Detection/attribution assessments, using General Circulation Models (GCMs) or Energy Balance Models (EBMs) with geographical distributions of surface temperature trends, suggest that the solar influence on climate is greater than would be anticipated from radiative forcing estimates. This implies that either the radiative forcing is underestimated or there are some processes inadequately represented in those models.
- One GCM study, using spectrally uniform variations in TSI, shows that horizontally heterogeneous (because it acts more strongly in areas where sunlight reaches the surface) solar forcing induces positive feedbacks involving temperature-gradient driven circulation regimes which reduce cloud cover. This effect is enhanced in the presence of greenhouse gas warming because the latter induces a basic state warming which enhances the feedbacks.

Mechanisms: Indirect UV effects

- Models that are forced with spectrally-resolved changes to TSI, and whose upper lids have been extended to around 80 km in order to fully resolve stratospheric processes, are able to reproduce the gross features of the 11-year solar cycle temperature response in the stratosphere, namely increased temperatures in solar maximum centred over the equator in the upper stratosphere. However, they cannot reproduce several important features, the most pertinent one being the secondary maximum in temperature in the lower stratosphere. This secondary maximum is likely to be an important part of the mechanism that transfers the solar signal from the lower stratosphere to the troposphere.

- Recent advances in the detailed understanding of the sensitivity of planetary propagation to the background circulation have enabled improvements in the modelling of the interaction of the 11-year solar signal and the QBO. However, this has not significantly improved the models' ability to reproduce the lower stratospheric temperature signal.
- Models appear to underestimate the ozone signal in the upper stratosphere / lower mesosphere and do not reproduce the equatorial negative signal in the low / mid stratosphere seen in some observational datasets. More data are required to confirm whether this is a model deficiency or a data artefact. A possible model deficiency is the neglect of the effects of energetic electron precipitation in most models.
- Some GCMs are able to reproduce the structure of the observed 11-year solar cycle response in the troposphere, with a poleward shift of the subtropical jets and a weakening of the Hadley cells, but they underestimate some aspects of the response. Experiments with a simplified GCM show that idealised heating of the tropical lower stratosphere can also produce these patterns of response, suggesting that the secondary temperature maximum in the equatorial lower stratosphere is an important feature of the transfer of the solar signal to lower levels.

Mechanisms: High Energy Particles

- High energy particles, such as galactic cosmic rays and solar protons are modulated by the 11-year solar cycle and by short-term (hours to days) solar changes. In some cases, high energy particles penetrate to the surface, which presents the possibility of atmospheric radiative changes linked to the particles' passage through the stratosphere and troposphere. Changes in atmospheric ionisation at the surface can be caused by substantial solar changes.
- Aerosol microphysics**,** such as particle nucleation, coagulation, and scavenging can be expected to be modified through changes in cosmic ray ion production.
- There is detailed theoretical support for ultrafine aerosol production in the atmosphere from ions, in the regions where there is condensable vapours with no substantial competing vapour sinks. Large cluster ions, which are expected in the initial phase of particle formation, have also been detected in the free troposphere. The combination of theoretical prediction and supportive (but preliminary) experimental atmospheric results now gives reasonable confidence that particle formation occurs from ions in the troposphere, and therefore from cosmic rays. The geographical distribution of particle formation and its frequency is not known with any confidence.
- Charging of aerosol particles and droplets occurs on the particle and cloud boundaries, related to global circuit properties. Theory supports the enhanced removal of charged particles to cloud droplets, even at low particle charge levels**.** There is some basis for an effect of charged aerosol on freezing of super-cooled clouds through electro-scavenging, but this has not been experimentally demonstrated in the atmosphere.

- Currents flow in the global atmospheric electrical system as a result of the partial electrical conductivity of air, caused mainly by cosmic ray ionisation. The global atmospheric electrical circuit drives vertical electric currents in fair weather regions. It provides a global teleconnection, communicating atmospheric electrical changes throughout the stratosphere and troposphere, down to the surface.
- Cloud-retrievals from the ISCCP satellite program show a strong correlation between low liquid water clouds with galactic cosmic rays from July 1983 to September 1994. Without detailed model calculations combining the aerosol and cloud microphysics and estimating the effects, it is not possible to be sure whether the correlation results from a direct cosmic ray effect on low clouds, or has another origin. If a calibration correction to the data is accepted, on the basis of the absence of a polar orbiting satellite, the correlation continues until September 2001; there is some recent evidence that the correlation is stronger at high latitudes than low latitudes.
- Cosmic rays and Total Solar Irradiance variations are often closely correlated. This relationship has been used to infer past TSI changes. However it also means that cosmic ray and TSI signals may be ambiguous on some timescales: which has implications for the use of cosmogenic isotopes as proxies. The proxies actually measure the cosmic ray variations. Consequently if cosmic-ray induced aerosol microphysics changes couples strongly through to clouds, climate signals attributed through proxies to solar TSI changes could also be explained by a direct atmospheric effect of cosmic rays.

1. Background

Changes in the direct energy output of the sun (once known as the "solar constant") are believed to be important in historical simulations of climate. Beyond changes in the sun's direct energy output, the possibility that a significant part of the warming of the last century might be due to other changes in solar characteristics still cannot be completely discounted, although the scientific consensus remains that most of the observed warming over the last 50 years is likely to have been due to the increase in greenhouse gas concentrations. Possible changes in the sun's influence on climate remains one of the most commonly raised objections to global warming attribution and projections. Under a previous contract, Harrison and Shine (1999) provided a review of the influence of solar changes on the Earth's climate. This formed a useful basis for discussion in the IPCC Third Assessment Report. Since this report, further research has been carried out, including some important new results which cast doubt on the likelihood of appreciable century-scale variations in solar irradiance. Consequently there remains considerable uncertainty about the sun's role.

This review is an update of the Harrison and Shine (1999) report, and provides information on recent advances in this area of science to be considered in preparation of the IPCC Fourth Assessment Report. It also provides advice on the available reconstructions of past direct solar output (now known as the "Total Solar Irradiance" or TSI), to inform the decision by the Met Office on which solar reconstruction to base the transient climate simulations for the Fourth Assessment Report.

2. Reconstructions of past solar activity

2.1 Comparison of TSI reconstructions

Direct measurements of Total Solar Irradiance (TSI) from outside the Earth's atmosphere began with the availability of satellite instruments in 1978. Previous surface-based measurements did not provide sufficient accuracy, as they were subject to uncertainties and fluctuations in atmospheric absorption that may have swamped the small solar variability signal. In reconstructing past changes in TSI, proxy indicators of solar variability, for which longer periods of observation are available, are used to produce an estimate of its temporal variation over the past centuries. A variety of indicators and methods have been employed, and these are discussed further below. The most recent reconstructions are summarised in table 2.1. It is, however, important to note that even in the satellite era, significant uncertainties in TSI values remain. These are related to the calibration of the instruments and their degradation over time. As a result, the *absolute* value of TSI is only known to approximately 2 Wm^{-2}, arising from the spread in values obtained by instruments measuring simultaneously over the past 8 years. The *variation* over the past two "11-year" (Schwabe) cycles is known to greater accuracy showing approximately an 0.08% (~1.1 Wm^{-2}) variation, although this does depend on inter-instrument calibration (see below). Nevertheless, it is on these measurements that many of the irradiance models discussed below are based.

There is a further uncertainty, however, in the existence of any underlying trend in TSI over the past 2 cycles with two TSI composites showing different results. The results of Fröhlich & Lean (1998, updated to present on http://www.pmodwrc.ch) show essentially no difference in TSI values between the cycle minima occurring in 1986 and 1996. The results of Willson & Mordinov (2003), however, show an increase in irradiance of 0.045% between these dates. The discrepancy hinges on assumptions made concerning the degradations of the Nimbus7/ERB and ERBS/ERBE instruments, data from which fills the interval, July 1989 to October 1991, between observations made by the ACRIM I and II instruments. If such a trend were maintained physically, it would imply an increase in radiative forcing of about 0.1 Wm^{-2} per decade. Compared in terms of climate forcing, this is appreciable, being about one-third that due to the increase in greenhouse gases averaged over the past 50 years. These discrepancies are important because all the available TSI reconstructions discussed below either use directly, or are scaled to fit, the recent satellite measurements, using one or other of the TSI composites.

The recent satellite period contains information only on the short term components of solar variability but, in order to assess the potential influence of the Sun on centennial-scale climate change, it is necessary to know TSI further back into the past. There are four different approaches taken to "reconstructing" the TSI, all employing a substantial degree of empiricism.

2.1.1 First approach to TSI reconstruction: use of sunspot numbers

The first approach attempts to include some understanding of the physical processes determining the irradiance variations: the components of TSI variability due to facular brightening, sunspot darkening and long-term changes in the "quiet sun" are estimated separately and then summed to give an overall value. This is the method employed by Lean and co-workers, by Solanki and co-workers (SF) and by Hoyt and Schatten (HS). All these authors use sunspot number (or sunspot group number) interpreted as sunspot area to indicate the fractional darkening of the solar surface. Relationships between facular brightening and sunspot darkening are then found empirically using recent observational data. The different authors use different methods to accomplish the fitting but obtain similar results for short-term variability. A further sophistication employed by SF98 is that the relationship is allowed to vary with time according to a facular index derived from very detailed analysis of high spatial resolution solar images for five different proxies, including Ca II plage areas.

The next step in this first approach is to invoke some method for determining the longer-term (multi-decadal) variation in the "quiet sun" *i.e.* the solar surface with faculae and sunspot contributions to variations removed. Lean *et al* use sunspot cycle amplitude (defined as a running average of sunspot number, not a conventional wave amplitude) as a proxy for this variability, Hoyt and Schatten use sunspot cycle length and Solanki and Fligge compare series using both of these assumptions. Progressing backwards in time, the two proxies diverge markedly.

The final aspect of this approach is that the secular variation given by cycle amplitude (or, less directly, the cycle length) is scaled so that the difference between the present day value of TSI and that assumed valid for the late 17^{th} century Maunder Minimum (MM) in solar activity is prescribed. This assumed MM value is the least well known of all those required for this approach to TSI reconstruction and is the subject of ongoing debate. Based on a comparison of solar activity with "sun-like" stars, Baliunas and Jastrow (1990, BJ90) suggested that the sun had brightened by 2-8 Wm^{-2} since the MM and the TSI papers discussed above all used values in this range. HS93 and LBB95 used a 3.3 Wm^{-2} increase, SF98, SF99 used 4 Wm^{-2}, and Lean (2000) used 2.7 Wm^{-2}. Since then, various reassessments of the BJ90 range of values have been made, mostly based on cosmogenic isotope and geomagnetic indices (see below), which have decreased the MM TSI difference. Furthermore, the BJ90 work has been called into question with regard to the statistical robustness of their sample of stars, their treatment of the data, as well as the central concept of a "sun-like" star solar analogue (D. Soderblom, personal communication).

2.1.2 Second approach to TSI reconstruction: use of geomagnetic records

The second approach to TSI reconstruction uses records of geomagnetic and cosmogenic indices. These have the appeal that the reconstruction can be based on a single observational record, without recourse to any assumptions of secular change since the MM. Thus, for example, Lockwood and Stamper (1999) use the *aa*-index, a direct measure of the geomagnetic field found from two antipodal stations, scale it to recent satellite measurements and produce a record of TSI back to 1868. Similar techniques using the ^{14}C and ^{10}Be records have been employed by Bard *et al* (2000) and Jirikowic and Damon (1994). The main problem with this approach is that the assumption of a direct relationship between solar open magnetic flux (which modulates cosmic rays) and TSI has not been physically validated. Indeed, the idealised modelling work of Lean *et al* (2002) suggests that TSI depends on the solar total magnetic field, of which only a small proportion is the open flux. These authors argue that, although there has been an upward trend in open flux, the total flux returns essentially to zero at each solar minimum giving no long-term trend in solar irradiance. This work is controversial; other authors (*e.g.* Solanki *et al*, 2002) are of the opinion that open flux is related to total flux so that either may be used as a proxy for irradiance and that the overlap (in time) of consecutive solar cycles causes the total flux not to return to zero at solar minimum.

2.1.3 Third approach to TSI reconstruction: deduction from climate records

The third approach to TSI reconstruction (Reid, 1977; Beer *et al*, 2000) involves using climate records to determine the solar influence which gives the best fit. Although this work is interesting, we reject these reconstructions here because of the need to consider the Sun as an independent climate forcing agent, and therefore to avoid the use of measures which could potentially be contaminated by atmospheric changes. The fourth approach (*e.g.* Sofia and Li, 2001) involves using numerical models of the sun to simulate the variations in solar parameters. In the future this may well provide the most physically satisfactory approach but at the moment the science is in its infancy. Consequently the results are not able to provide secular changes in irradiance.

Entry number	Source	Description	Physical theory	Method	Reliance on other work	Comments
1.	**Hoyt & Schatten, 1993** **(HS93)** A discussion of plausible solar irradiance variations, 1700-1992, JGR1993	TSI for 1700-1992	Convective energy transport drives luminosity variations. Solar rotation, rate of sunspot decay, sunspot structure and the fraction of sunspots with only penumbrae are proxies of this energy transport.	Uses sunspot Schwabe (~11year) cycle length as proxy for sunspot decay rate. Long term trend from 1880's to 1980's =0.14% of solar irradiance (taken from Lean 1992.) 11 year cycle variations from sunspot amplitude (scaled to satellite era data) added to long term to obtain Schwabe irradiance variations.	Uses Maunder Minimum (MM) - to - present solar irradiance mean change of 0.24% from Lean 1992.	• Used in IPCC 3AR • Dataset recently extended to 2000 (Meehl *et al*, 2003) • Sunspot cycle length is a more highly derived quantity than sunspot number • See table 2 for further comments
2.	**Lean, Beer & Bradley, 1995** **(LBB95)** Reconstruction of solar irradiance since 1610: Implications for climate change, GRL 1995	TSI for 1610 to present. With UVSI variations.	Bright facular areas and dark sunspot areas cause irradiance variations over a solar cycle. Over a solar cycle the bright facular regions dominate, so with more sunspots, more faculae increase brightness. During solar minima, sunspots not present, but bright facular regions still present. Long-term variations of these bright facular regions determine long term irradiance changes. Evidence from sun like stars that surface magnetic activity varies and is very low for non-cycling stars (Baliunas & Jastrow).	Use sunspot record to provide proxies of concentrations of bright faculae and dark sunspots 11 year cycle contribution to irradiance. Assume that on longer timescales facular quiet network (bright facular during solar minima) tracks average solar cycle amplitude. Difference in TSI from present mean cycle irradiance to MM to be 0.24%, from Lean 1992.	Uses MM - to - present solar irradiance mean change of 0.24% from Lean 1992.	• Used in IPCC 3AR • Dataset superseded by Lean 2000

3.	**Lean 2000.** Evolution of the Sun's Spectral irradiance since the Maunder Minimum, GRL 2000	TSI for 1600 to present with irradiance for UV, visible/near IR and IR.	As LBB95	Use sunspot record to estimate darkening effect on irradiance. Use Measurements of MgII, CaII and daily Ca indices (derived from the strength of spectral lines in the solar UV spectrum) and mean sunspot group numbers to estimate bright facular contribution to irradiance for 11 year cycles. Use 15 year moving mean of annual sunspot group numbers, scaled to give a MM - to - present mean of 0.2%. (cf 0.24% from Lean 1992).	Uses MM - present solar irradiance mean change of 0.2%, due to revision of work. No obvious ref given.	• Scaling to MM not ideal • Sunspot number slightly less derived quantity than cycle length • See table 2 for further comments
4.	**Lean, Wang & Sheeley, 2002.** **(LWS 2002)** The effect of increasing solar activity on the Sun's total and open magnetic flux during multiple cycles: Implications for solar forcing of climate, GRL 2002	TSI as above, but suggest that secular changes from quiet network variations may not happen.	Models of magnetic surface and open flux on Sun suggest that open flux may have secular changes but surface flux, and thus quiet network, may not have. N.B. open magnetic flux is what directly affects things outside the sun - including cosmic rays & the aa index - but only makes very small contribution to total surface flux	11 year cycle defined as above.	Uses model of magnetic field. Also refers to Foukal & Milano 2001, which suggests no change in quiet network over 20th C.	• Scaling to MM not ideal • Very new results, which remain to be confirmed or otherwise
5.	**Bard, Raisbeck, Yiou & Jouzel, 2000.** Solar irradiance during the last 1200 years based on cosmogenic nuclides, Tellus 2000	Range of possible TSI over last 1200 years.	Assume geomagnetic proxies, ^{14}C and ^{10}Be, can be used to simulate solar irradiance changes.	Smooth the record of cosmogenic nuclides, and scale by different estimates of irradiance change from MM - to - present.	Uses MM - to - present TSI estimates or 0.25% (Lean 1995), 0.4% (Zhang 1994, Solanki & Fligge 1998), 0.55% (Cliver 1998), 0.65% (Reid 1997)	• Scaling to MM not ideal • Cosmogenic isotopes are produced in the atmosphere and measured in ice cores.: There is a possible contamination by atmosphere signals but because the processes of formation and removal of ^{14}C and ^{10}Be are different where their records show the same variability that is assumed to indicate the cosmic ray signal.

6.	**Solanki & Fligge 1998.** **(SF98)** Solar irradiance since 1874 revisited, GRL 1998.	TSI 1874-1995	11 year cycle variations in TSI due to changes in magnetic sunspot/facular regions. Network (facular during minima) and possible convection dominant over longer timescales.	Active region contribution is fitted to sunspot numbers using quadratic equation. This relationship is modulated over time using facular index (includes 5 different solar proxies). Quiet sun modulation due to (a) sunspot cycle amplitude (similar LBB95) and (b) cycle length (similar HS93) plus long term change since MM.	Assumes 4 Wm^{-2} (0.29%) increase in 11-year averaged irradiance since MM	• Scaling to MM not ideal • Reconstruction produced by solar modellers gives it good physical foundation • See table 2 for further comments
7.	**Solanki & Fligge 1999.** **(SF99)** A reconstruction of total solar irradiance since 1700, GRL 1999	TSI 1700-1995 A spectral variation version is available in Fligge and Solanki, (GRL 2000).	As SF98	As SF98 but extends further back in time; mean amplitude scaled to fit satellite measurements over past 2 solar cycles	As SF98	As SF98
8.	**Solanki S.K. and Krivova N.A., 2003** Can solar variability explain global warming since 1970? JGR, 2003**.**	TSI 1856-1999	As SF99 but discusses evidence for range of possible values for increase in quiet sun between 1700 and 1978.	Uses 2 and 4 Wm^{-2} with each of cycle length and cycle amplitude factors and each of Willson and Fröhlich & Lean recent composites (*i.e.* total of 8 reconstructions). The TSI reconstructions are derived independently of the temperature record except that the cycle length version is lagged to give a better match.	Temperature-matching makes cycle-length versions unacceptable.	• Given that facular and sunspot contributions are now quite well established this paper goes furthest to assess evidence for long-term variability. • See table 2 for further comments
9.	**Lockwood & Stamper, 1999.** Long-term drift of the coronal source magnetic flux and the total solar irradiance, GRL 1999	TSI since 1860s for a range of possible scaling factors.	Correlation of coronal source magnetic flux with irradiance over the last two cycle, assumed valid for all times. The coronal source surface is where the solar magnetic field becomes approximately radial and lies at a heliocentric distance of about 2.5 solar radii	Use geomagnetic *aa* index as proxy for solar coronal source flux which correlates with irradiance over last couple of cycles. Regression of flux against the different satellite TSI measurements provides range of scaling factors	Only assumes solar magnetic flux correlates with irradiance on all time scales.	• Independent method with substantial data content (monthly) but assumes a connection between *aa*-index and TSI that is not fully supported by solar physics community • See table 2 for further comments

10.	**Reid, 1997.** Solar forcing of global climate change since the mid-17th Century Climatic Change 1997	TSI since 1600 to present	Assume sunspot record is proxy for irradiance. Scale to match simple climate model to observations.	Use sunspot amplitudes to get 11 year cycle and long term secular trend (Similar to Lean). Use climate model, scale irradiance to fit to observed temps. The MM - to - present is about 0.58%!	Circular reasoning prevents this being useful.	• Circular arguments
11.	**Sofia & Li, 2001**. Solar variability and climate, JGR 2001	TSI from 11 year cycle but only suggesting possible secular changes	Using model of magnetic field within the sun can create the solar cycle variations in irradiance. 11year cycle due to internal magnetic field changes at about 0.96 solar radius. Solar radius measurements suggest that longer term irradiance due to deeper magnetic field variations.	Use of sunspots, magnetic model and satellite irradiance measurements to get TSI over 11 year cycle. However not yet enough solar radii measurements to get temporal info for longer term irradiance variations: they may exist (Paris Observatory)	Using Magnetic model for near surface and for deeper, near base of convection zone, effects on irradiance. No assumptions about sun-like stars etc.	• Promising physically based approach, but no actual numbers available yet • Reconstruction produced by solar modellers gives it good physical foundation
12.	**Jirikowic & Damon, 1994.** The Medieval solar activity maximum, Climate Change 1994	TSI for long periods into past and future!	From work done by Damon & Jirikowic on ^{10}Be and ^{14}C find cycles in solar activity and thus irradiance. Scale to observed temperatures.	Irradiance described as $I = a$*f(11 years)+b*f(88 years) + c*f(208 years).		• Circular arguments not helpful
13.	**Beer, Mende & Stellmacher, 2000.** The role of the sun in climate forcing, QSR 2000	TSI since 1750	Choose method because "we find a better fit with the temperature data if we assume a linear relationship between cycle frequency and irradiance".	Get TSI "sun melody" from frequency of the Schwabe (11 year) cycle. Demodulate and filter.		• Circular arguments not helpful

Table 2.1. Summary of recent TSI reconstructions (based on an original Table by Gareth Jones)

2.2 TSI reconstruction recommendations for IPCC Fourth Assessment Report.

We note that the current state-of-the art in terms of available TSI data reconstructions has not advanced substantially since the IPCC Third Assessment Report. There are, however, new ideas and developments in understanding since then.

Based on the comparison of the various approaches described in section 2.1, the 5 main candidate TSI reconstructions for use in the Fourth Assessment Report have been identified and listed in table 2.2. Additional comments on each reconstruction have been included in the table and the entries have been ranked in terms of their suitability.

For the reasons discussed above we recommend employing a reconstruction that is based on the first approach discussed in section 2.1. Specifically, we recommend the reconstruction of the Solanki and Krivova (2003, SK) which uses (i) solar cycle amplitude for secular variation with (ii) a 2 Wm^{-2} increase since the Maunder Minimum and (iii) the Fröhlich and Lean reconstruction for the satellite era.

Our reasons for choosing cycle amplitude are as follows. Very little justification is given for using cycle length except that (a) it fits well with the global average temperature record before 1970 and (b) shorter cycles are more intense because there is not time for a previous cycle to decay before the subsequent one starts. We reject reason (a) on the basis that it is important to seek reconstruction methods independent of the climate system. (b) suggests that cycle length is only an indirect measure of cycle amplitude. Furthermore, the value of cycle length is more difficult to determine, as is evidenced by the differences between the time series of cycle length produced by Friis-Christensen and Lassen (1991), Fligge et al (1999) and Lockwood (2001). For this pragmatic reason, and until a physical basis is established for the use of cycle length, we reject it in favour of the direct use of amplitude.

SK discuss the possible values for the long-term increase in TSI since the Maunder Minimum (MM). They present an overview of current results and opinions concerning possible values, including estimates based on UV spectral lines, open magnetic flux and magnetograms and conclude that it probably lies in the range 2-4 W m^{-2}. We recommend the value at the low end of the range because (i) recent reanalysis of the Baliunas and Jastrow work (Foukal, 2003; Foukal et al 2004) suggests that they may have overestimated it; (ii) in an independent study Foster (2004) deduced from an analysis of SoHO magnetograms a base level quiet sun irradiance which is higher than the MM values in all reconstructions produced up to 2000 and (iii) it is moving in the direction suggested by Lean et al (2003, LWS). We do not adopt the zero value presented by LWS because, while this result was obtained by careful consideration of the relationship between open flux, total flux and TSI, the assumption that the total flux returns to zero between every solar cycle remains to be confirmed.

Using our recommended parameters the SK reconstruction is actually very similar to that of Lean (2000). Figure 2.1 shows the two curves overlaid and reveals a close correspondence, although the magnitude of the 11-year cycle is generally somewhat larger in SK, especially earlier in the record. We consider that the differences would probably have negligible impact on GCM simulations of the type needed for 4AR and that use of either reconstruction would be acceptable. Nevertheless our slight preference for SK is based on our view that the physical components of the facular brightening have been considered in more detail in this latest work, developing from its predecessor SF98.

Finally, the choice of the Fröhlich and Lean (FL) reconstruction over that of Willson is a difficult one. It is unclear how the disagreement will be resolved as it turns on analyses of measurements made during 1989-1992. Longer time series of well-calibrated TSI measurements are clearly needed for some time into the future to indicate if the upward trend suggested by Willson will be maintained. In the meantime we have opted for FL mainly because Fröhlich (2004, http://www.pmodwrc.ch/) has identified some glitches in the ERB data not accounted for by Willson. Furthermore, if the TSI reconstruction methodologies are extended to include the satellite period they do not reproduce the Willson trend. Of course this does not prove that the latter is incorrect but inclusion of it in our TSI series would make for an inconsistency in our approach.

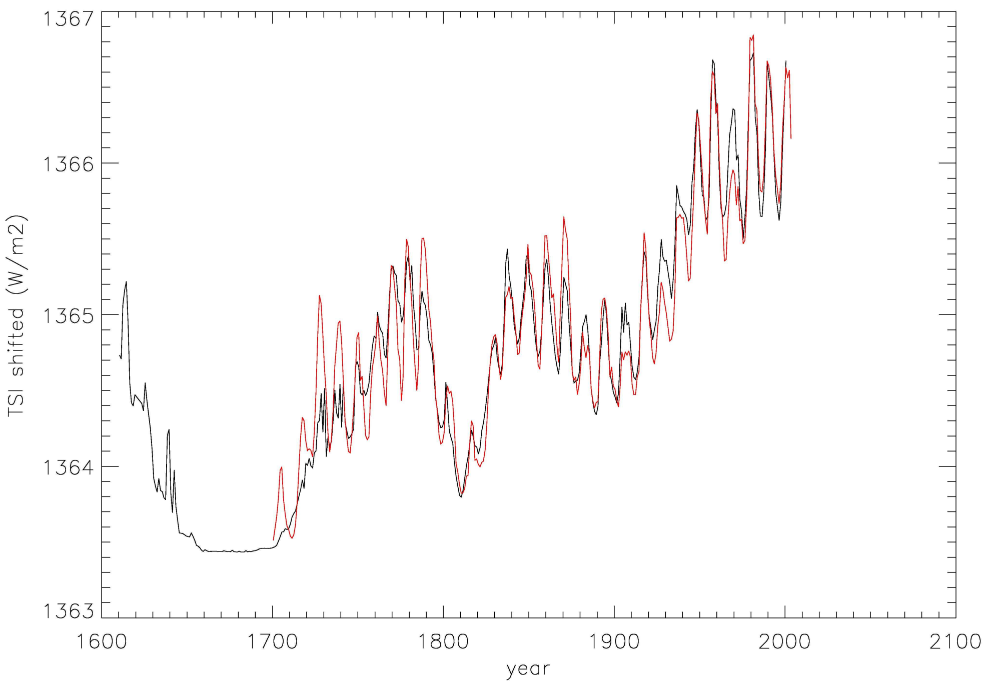


Figure 2.1. TSI reconstructions. Black: Lean (2000). Red: Data from Natalie Krivova (personal communication) based on the methods described by Solanki and Krivova (2003); this version has secular variations determined by the cycle amplitude and a secular increase since the Maunder minimum of 2.15 Wm^{-2}, with the whole curve shifted upwards by 0.327 Wm^{-2} to match the long-term average of the Lean data.

Table 2.2. Prioritised list of available TSI reconstructions

Priority	Paper	Approach	Comments
1	**Solanki and Krivova (2003)** (cycle amplitude versions) Subsumes SF99 & SF98.	Non-linear relationship between facular brightening contribution to TSI and relative sunspot number. Imposed long-term quiet sun increase.	Facular and sunspot contributions are found by empirical fitting to daily data. Longer term variability is based on sunspot cycle amplitude or length. The magnitude of the long-term trend is assessed by reviewing the current state of understanding and a range of possible values deduced. The max (4 Wm^{-2}) and min (2 Wm^{-2}) are used separately and each matched to the Willson and Fröhlich& Lean satellite data composites. Our recommendation is to use this index with 2 Wm^{-2} MM value and the Fröhlich & Lean composite TSI observations for recent values.
2.	**Lean (2000)** Subsumes: Foukal and Lean (1990), LBB95	Sunspot number, faculae brightness (from CaII line) and group sunspot numbers.	Explains variation in solar output since 1980; possible discrepancy around time of the International Geophysical Year (IGY, 1957-58). Faculae are physically important in determining solar uv variations, and the astronomical technique permits comparison with other sun-like stars or to compare Maunder Minimum sun with non-cycling stars.
3.	**Lockwood and Stamper (1999)**	*aa*-index (a geomagnetic index) calibrated to satellite measurements of TSI	Uses the instrumental record of *aa*-index extending back continuously to 1868, available as monthly data (Mayaud, 1972). An important feature of the data set is that it is geomagnetic, and therefore uncontaminated by atmospheric changes. However the correlation between TSI and *aa*-index has been criticised: the open flux may provide a good basis on which to determine cosmic rays fluxes, but possibly not TSI. It also looks to perform less well in second half of 20th century: recent work indicates that the major open flux increase occurred during the first half of the twentieth century (Cliver and Ling, 2002; Richardson *et al*, 2002). In its favour, this approach is entirely independent of the approaches in (1) and (2) above.
4	**Hoyt & Schatten, 1993**	Solar cycle length	Hoyt and Schatten use solar cycle length as a proxy for amplitude of the solar cycle, based on Clough (1943). They combine this with the Lean *et al*. (1992) estimate of the Maunder Minimum TSI (see also 2 above and 5) for calibration. The original series was only to 1992, although Meehl *et al* (2003) indicate it has been extended.
5.	**Lean, Wang and Sheeley (2002)**	Cosmic ray generation of isotopes, measured in ice cores and calibrated to TSI	This paper raises fundamental doubts over the use of cosmogenic isotope variations as a proxy for changes in solar irradiance: it is clear, for example that ^{10}Be continues to cycle through the MM, so magnetic cycling continued. Consequently it is possible that MM solar irradiance not significantly lower than contemporary solar cycle minima. Questions remain about the assumptions made in the model run used as a basis for this work but if it is confirmed, there are large implications for paleoclimate modelling in which the long cosmogenic isotopes series available are used to indicate TSI.

3. Observations of solar response: 11-year solar cycle

In the following sections, major results are reviewed and assessed from analyses that have been carried out since the last review (Harrison and Shine 1999). The availability of just a few additional years of data since then have made a substantial impact in this research area, because the last few years have included a solar maximum period in which there were no major volcanic eruptions. This has enabled the 11-year solar cycle to be separated from the volcanic signal with increased confidence. There is now a general consensus that a distinct signature of the 11-year solar cycle is present in stratospheric temperature, ozone and circulation fields, although there are still discrepancies in the detailed analysis of the distribution and amplitudes of the signals that need to be resolved. Further details of the 11-year solar signal in tropospheric circulations have also been confirmed.

3.1 Middle Atmosphere

There are only a limited number of available datasets of the middle atmosphere that are long enough for a meaningful analysis of 11-year solar cycle (SC) variations (Chanin et al. 1998). This is especially true for the height region above ~30 km (10 hPa), the upper limit of radiosonde soundings. The primary datasets employed are:

1. Rocketsonde data: available for the period 1960's – mid 1990's, depending on the station site. The most suitable sites for 11-year SC analysis are mostly concentrated in the tropics and subtropics, with just a few in the northern middle latitudes (Dunkerton et al. 1998, Keckhut et al., 1999, 2004).
2. Lidar data: from Haute Provence Observatory in southern France, available for the period 1979-present (Keckhut et al. 2004). Although this dataset provides information at only a single location, it is an absolute measurement of temperature and does not require external calibration.
3. Satellite data: primarily the stratospheric Sounding Unit (SSU) and Microwave Sounding Unit (MSU) instruments on board a series of NOAA operational satellites. Data are available for the period 1979 – present (Scaife et al. 2000, Keckhut et al. 2001, 2004, Hood 2004). Also AMSU since 1998. The radiance data can be used to derive temperatures and geopotential heights (see e.g. Scaife et al. 2000). Although providing global coverage, care is required in the use of this dataset to account for discontinuities between successive instruments and also drift in each satellite due to instrumental and orbital effects.
4. Model assimilated datasets: global temperature and other data are available from the assimilation of various observations (e.g. radiosonde, satellite) into general circulation models. In the 'assimilation' process the model fields are nudged towards the available observations. In data-sparse regions the model data alone are used. In this way, a 'best estimate' is obtained with global coverage and other dynamically consistent fields such as winds can also be derived using the model. The two commonly employed assimilation datasets are the NCEP reanalysis dataset (Haigh 2003) and the ECMWF Reanalysis (ERA-40) dataset (Crooks and Gray 2005). Both datasets assimilate the SSU/MSU observations. The NCEP analysis assimilates derived temperatures while the ERA-40 analysis assimilates the radiance data directly. We note that neither the rocketsonde nor the lidar observations have been assimilated into these products, so they are valuable independent datasets.

3.1.1 Temperature

Rocketsonde / lidar Analyses

Seven rocketsonde sites were analysed by Keckhut et al. (2004) for the period from the 1960s to mid-1990s using a multiple regression analysis. In addition to the F10.7 cm flux, the analysis included seasonal changes, a linear trend and a QBO signal. The volcanic signal was eliminated by excluding data during the 1-2 years following a major eruption. Figure 3.1 shows the results of their analysis grouped into three latitude bands. At tropical latitudes there is a positive response reaching +2K just below the stratopause region at ~ 40 km. In the subtropics there is a smaller response that is still positive. At mid-latitudes the signal has reversed and shows a negative response over the height region 25-50 km with a maximum of 3K. A similar negative correlation at mid-latitudes was found in a corresponding analysis of the lidar dataset (44^{o}N), which also showed that the negative response originates primarily from the winter season.

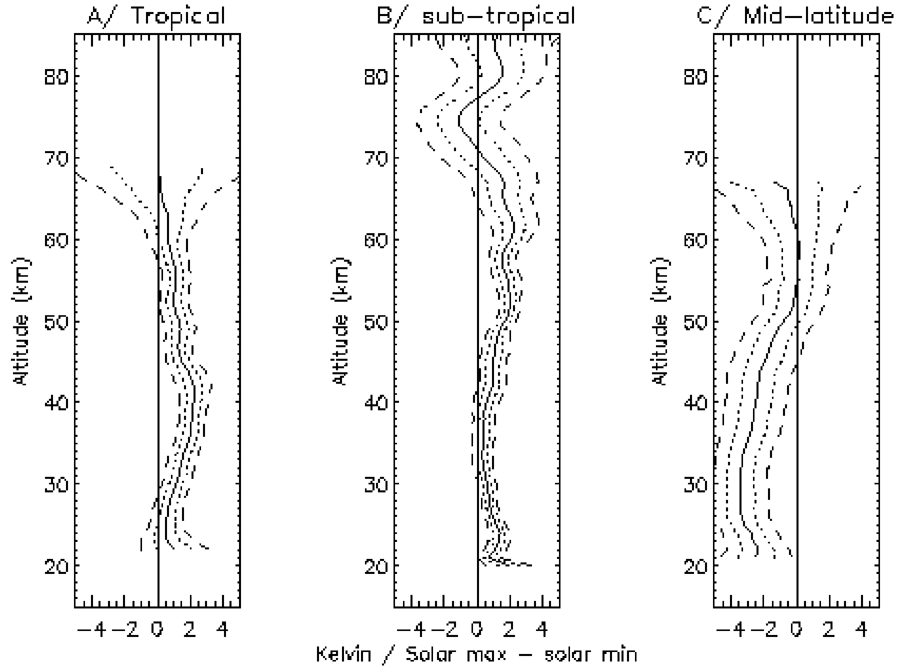


Figure 3.1: Annual-mean solar maximum minus solar minimum signal (K) from the rocketsonde temperature analysis of Keckhut et al. (2004).

Satellite Data Analyses

Figure 3.2 (a) shows the annual-mean height-latitude solar signal from the analysis of SSU/MSU data by Keckhut et al. (2004) for the period 1979-1998 (see also Scaife et al. (2000) for an earlier analysis of this dataset). The overall structure of the SSU/MSU signal is similar to the rocketsonde analysis, with a positive maximum over the tropics peaking just below the stratopause and an indication of a negative signal at mid-latitudes. However, the maximum amplitude of the signal estimated by the Keckhut SSU/MSU analysis is significantly smaller than the rocketsonde estimates. This difference likely reflects the differences in spatial sampling of the rocketsonde and satellite instruments, the zonal averaging of the satellite data and, possibly, the different time-periods of the two analyses. The reversed signal at mid-latitudes in fig 3.2 (a) is not as statistically significant as the tropical signal but this is consistent with it coming primarily from the winter months, since there is more internal variability during those months.

Figure 3.2 (b) shows the annual-mean height-latitude solar signal from a similar analysis of the SSU dataset by Hood (2004). Note the different height axis of this figure compared with Fig 3.2 (a). The range is slightly different and the linear height scale tends to accentuate the signals in the lower atmosphere compared with the logarithmic pressure scale. This analysis employed a version of the SSU/MSU dataset corrected and compiled by the NCEP climate

Prediction Center (CPC). This dataset extends to 1 hPa and should not be confused with the NCEP/NCAR assimilated dataset (e.g. as used by Haigh 2003) that extends only to 10 hPa. The analysis of Hood (2004) shows some interesting differences from the Scaife et al. (2000) and Keckhut et al. (2004) analyses. The maximum positive signal over the tropics has greater amplitude and is situated higher in the atmosphere, indicating a maximum value of > 2K near the stratopause level at 50 km (1 hPa). At mid-latitudes the signal is reduced but is still positive, in contrast to the other two analyses. Finally, there is a region of negative values centred over the tropics at around 10 hPa (30 km). Although a similar negative feature is present in the other two analyses it is lower down, nearer 50 hPa in figure 3.2(a) than the 10 hPa in figure 3.2(b). Also, the negative feature in the Hood analysis is statistically significant and larger in amplitude.

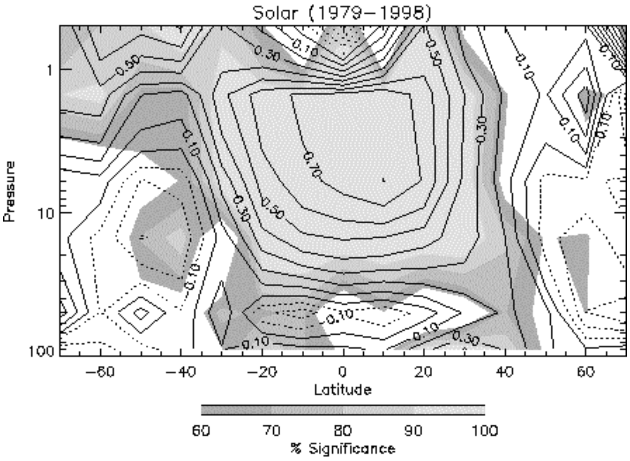


Figure 3.2(a): Annual-mean solar maximum minus minimum signal (K) from the SSU/MSU temperature analysis of Keckhut et al. (2004) for the period 1979-1998. Shading indicates statistical significance levels as indicated in the figure legend.

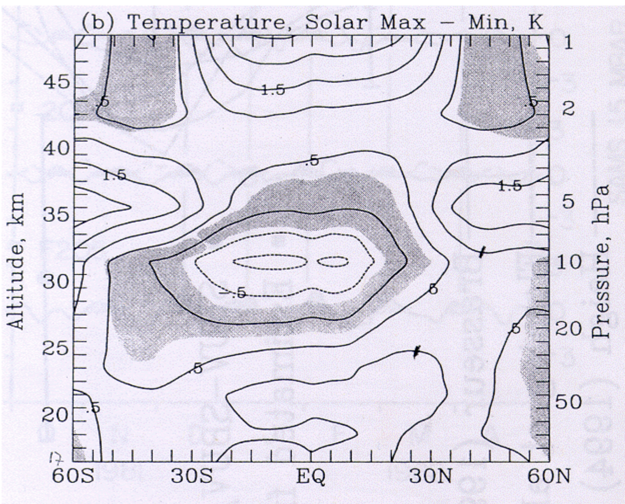


Figure 3.2(b): Annual-mean solar maximum minus solar minimum signal (K) from the SSU analysis of Hood (2004) for the period 1979-1997. Shading indicates regions that are <u>not</u> statistically significant at the 2 sigma level.

It is not clear why the results of Hood differ from the other two SSU studies in these respects, although the most likely difference lies in the treatments applied to the data to correct for instrument drift. As noted by Hood (2004) the SSU dataset compiled by NCEP/CPC may still exhibit significant offsets near the times of satellite transitions. Chenin et al. (1998) concluded that the time series analysed by Scaife et al. (and thus also, by implication, the one employed by Keckhut et al.) was more nearly free of instrument-related errors. On the other hand, the magnitude (but not the height) of the tropical maximum in the Hood (2004) analysis agrees more closely with the rocketsonde analyses. More work is evidently required to understand the different treatments applied to the SSU/MSU dataset so that the differences in these solar signal analyses can be resolved.

Assimilated Data Analyses

Figure 3.3 shows the annual-mean height-latitude solar signal from a regression analysis of the ERA-40 reanalysis (Crooks and Gray, 2005) and the NCEP/NCAR reanalysis (Haigh, pers. comm.). Note that, as previously mentioned, the NCEP reanalysis dataset used for figure 3.3(b) is not the same as the NCEP/CPC SSU dataset employed for figure 3.2(b). The multiple regression analysis of these two datasets included seasonal variations, a long-term trend term, the solar variation estimated using the 10.7cm radio flux, and estimates of the variations associated with volcanic aerosols, the QBO, ENSO and the NAO. Apart from minor details in the representation of the QBO signal, the regression analyses were identical. Note the different vertical extents of figures 3.3 (a) and (b). The NCEP/NCAR reanalysis extends to 10 hPa while the ERA-40 reanalysis extends to 0.1 hPa (although analysis above 1 hPa should be treated with caution).

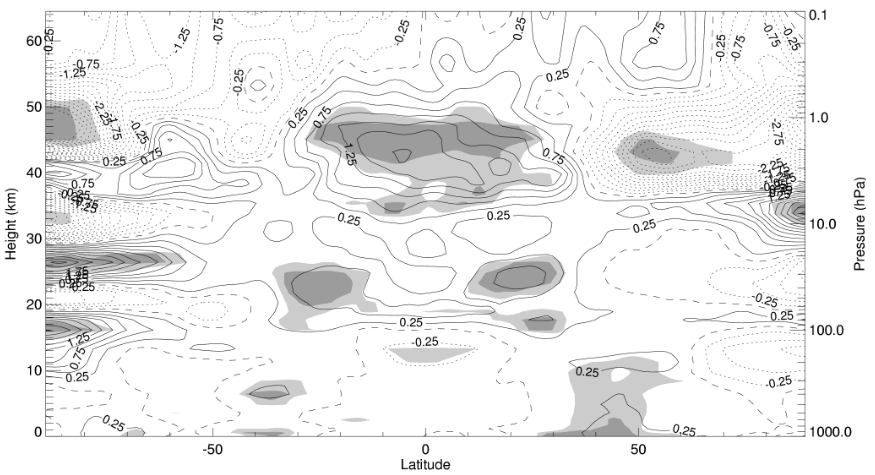


Figure 3.3(a): Annual-mean solar maximum minus minimum signal (K) from the ERA-40 temperature analysis of Crooks and Gray (2005) for the period 1979-2001. Light and dark shading indicates statistical significance levels at the 95% and 99% levels.

The broad structure of the ERA-40 temperature signal in the middle atmosphere is similar to the SSU analyses (figure 3.2). A positive region is located over the tropics reaching a maximum value of 1.75 K at around 45 km. The signal decreases in the subtropics and reverses in sign at mid-latitudes. The amplitude of the maximum positive signal is closer to the value of Hood (2004) although the height of the maximum just below the stratopause is closer to the results of Scaife and Keckhut. The negative signals seen in the SSU analyses over the tropics near 50 hPa in the Keckhut analysis and near 10 hPa in the Hood analysis is not present, but there is nevertheless a local minimum centred around 10-30 hPa and the statistical significance of the signal is very small here, probably reflecting the large inter-annual variability associated with the QBO which dominates this region.

There is an interesting structure in the subtropics of the ERA-40 analysis between ~20-30 km which, although hinted at in the SSU analyses, is much more prominent in the ERA-40 and NCEP/NCAR analysis. It consists of two separate lobes of significant positive signal with a local minimum between them over the equator. This pattern is typical of the meridional structure of the QBO and suggests either that the solar response is similar to the QBO response, with differential upwelling between the tropics and subtropics or that there is some non-linear interaction between the solar and QBO signals in this region that the linear regression analysis could not account for (Crooks and Gray, 2005). Above this region, the Haigh analysis indicates a negative signal above about 10 hPa, similar to the feature in the Hood SSU analysis, but this should be treated with caution since it is located at the upper limit of the data assimilation model and the signal is not statistically significant.

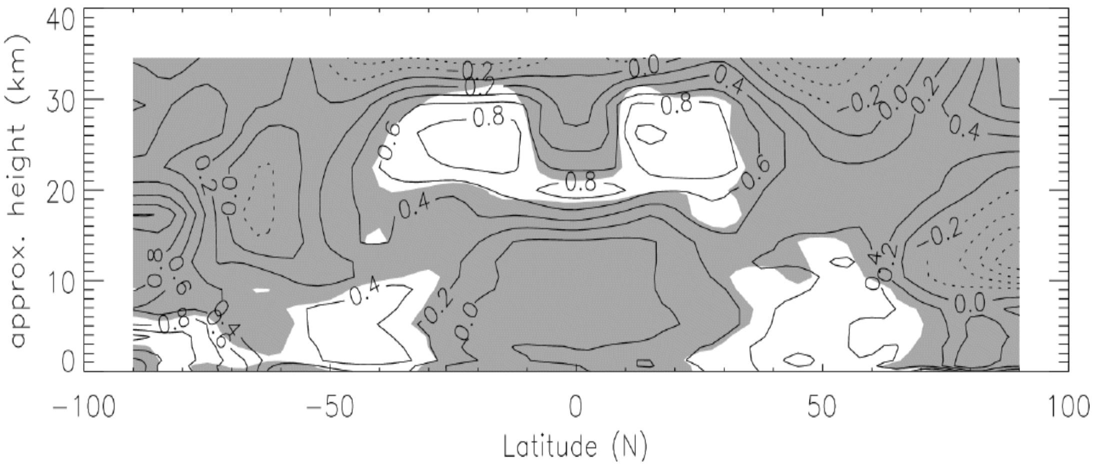


Figure 3.3(b): Annual-mean solar maximum minus minimum signal (K) from the NCEP/NCAR temperature analysis of Haigh (pers. comm.) for the period 1979-2001. Shading indicates regions that are not statistically significant at the 95% level.

We note here the potential problematic impact of non-linear interactions on these linear regression studies between the 11-year solar signal and other signals such as the QBO (Salby and Callaghan 2000, McCormack 2003, Lee and Smith 2003, Gray et al. 2004), the volcanic signal (Lee and Smith 2003) and the seasonal cycle (Salby and Callaghan 2004). They constitute a substantial complication when attempting to extract the 11-year solar signal from observations. Coughlin and Tung (2004a) have employed a novel method to deal with the non-linearity of these signal interactions, using NCEP data. They find relatively high correlations with the 11-year solar signal, although details of the overall spatial pattern are more uniform than the pattern in figure 3.3(b).

Interaction with the Quasi Biennial Oscillation (QBO)

An update of the extensive analysis of the longitudinal and seasonal structure of the solar signal at various pressure levels has been carried out by Labitzke, van Loon and co-workers (Labitzke and van Loon 2000, van Loon and Labitzke 2000, Labitzke 2001, 2002, 2003, 2004, Soukharev and Labitzke 2000, 2001) using the FUB temperature and geopotential height dataset (Labitzke et al. 2002)) and, more recently, the NCEP/NCAR reanalysis. Previous results have been updated (e.g. Labitzke 1987, 2002, Labitzke and van Loon 1988, 2000, van Loon and Labitzke 1994, 2000) to confirm that during northern hemisphere (NH) winter, the solar signal is only readily apparent if the data are first divided according to the phase of the QBO winds in the equatorial lower stratosphere. This result was also confirmed by Salby and Callaghan (2004). More recent analysis of the 30 hPa temperatures and geopotential heights (e.g. Labitzke 2002, 2003) has shown that this is also true of the summer-time solar signal, which has a very different characteristic in the two QBO phases.

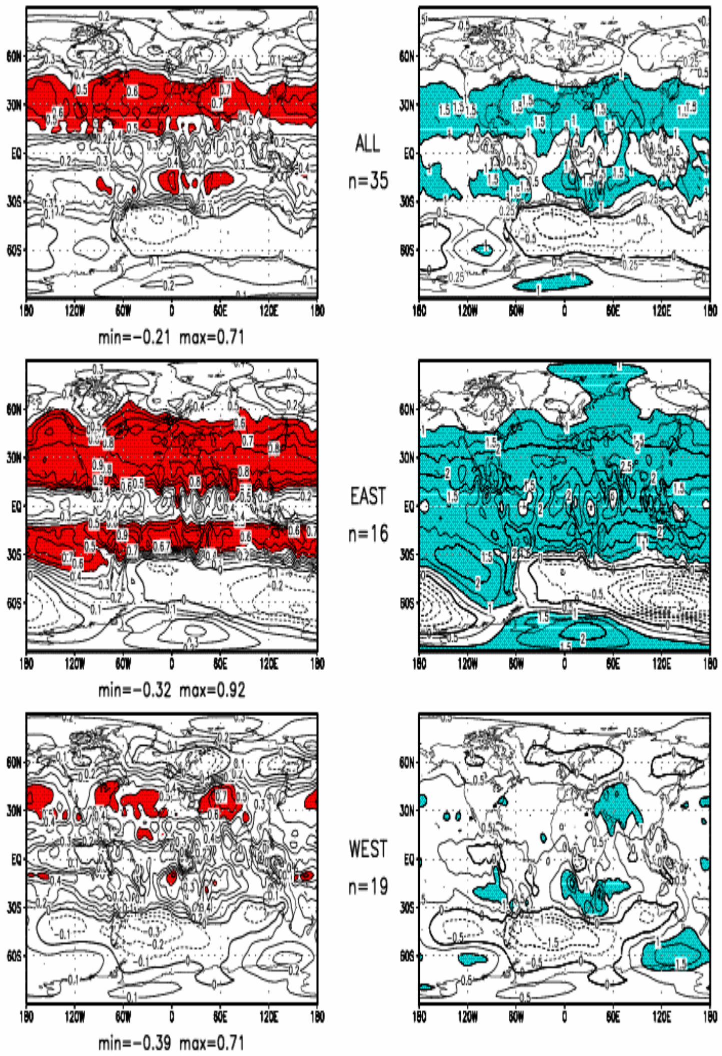


Figure 3.4: Left panel: correlations between de-trended 30 hPa temperatures and 10.7 cm radio flux from the study of NCEP/NCAR 1968-2002 reanalysis dataset by Labitzke (2003). Correlations above 0.5 are shaded for emphasis. Right panel: solar maximum minus solar minimum temperature signal (K) from the same analysis. Upper row: results using al years of the dataset. Middle panel: results using only QBO east phase years. Lower row: results using only QBO west phase years.

Figure 3.4 shows some results from their analysis of the solar signal in 30 hPa NCEP/NCAR temperature dataset. The right hand panel shows the solar maximum minus solar minimum signal (derived using a very simple regression fit) and a correlation analysis is shown in the left panel. The top row shows results using all years, while the middle and lower rows show results for the east phase and west phase QBO years respectively. As indicated in the zonally-averaged analyses in figure 3.3 discussed earlier, there is a 3-fold structure across the equator and subtropics, with 2 separate lobes of positive signal in the subtropics and a local minimum over the equator. The separation into QBO phase indicates that most of this structure comes from the East QBO phase years (Labitzke 2003).

In an analysis of the geopotential height differences between solar minimum and solar maximum, Labitzke (2004) has used the NCEP/NCAR reanalyses to confirm and extend earlier results (e.g. Labitzke 1987) that showed a puzzling relationship between NH stratospheric sudden warming events in solar max/min and QBO east/west phases. It is well known that stratospheric sudden warmings most often occur during the QBO east phase (Holton and Tan 1982) and this can be explained in terms of planetary wave propagation that is guided by the presence of zero-winds in the winter subtropics (see e.g. Baldwin et al. 2001). Planetary waves are unable to propagate vertically in easterly flow, which is present in the summer hemisphere. In the winter hemisphere, waves propagating in the background westerlies are refracted away from the subtropical region when the QBO is in its easterly phase. This confines the waves to higher latitudes and hence closer to the polar vortex in east QBO phase winters. This means that they more effectively cause the breakdown of the polar vortex and a mid-winter sudden stratospheric warming is more likely to occur. However, Labitzke (1987) showed that while this was usually the case at solar minimum, the situation was reversed at solar maximum and sudden warmings were just as likely in QBO west phase years. Labitzke (2004) and studies referenced therein have confirmed this result using longer datasets and have shown that this behaviour is true in the Southern Hemisphere as well as in Northern Hemisphere winters. While this observation has puzzled theoreticians for many years, recent modelling studies have succeeded in modelling this solar / QBO interaction (see section 5) and this may provide an important clue to understanding how the solar cycle influence extends to lower altitudes.

3.1.2 Ozone

The availability of additional years of observations has enabled further analyses of the solar signal in middle atmosphere ozone, with increased statistical significance emerging (Lee and Smith 2003, Hood 2004, Haigh et al. 2004). As with the temperature analyses, the additional data around the period of the last solar maximum (since 1995) has proved especially valuable since there has been no significant volcanic activity during this period and it provides the opportunity to distinguish between solar and volcanic signals more accurately.

Figure 3.5 shows the results of analyses of the SBUV/SBUV satellite dataset for the period 1979-1994 from Hood (2004) and of the SAGE II satellite dataset for the period 1984-2002 from Haigh et al. (2004). Both show good agreement in terms of the amplitude (note that the SAGE values need to be multiplied by 1.3) and broad structure, with a positive signal throughout most of the middle atmosphere, implying more ozone during solar maximum than solar minimum. This result is also reflected in the analysis of long-term observations of the total column ozone (WMO 2002), which during at least the last 3 solar cycles has been in phase with the 10.7 cm radio flux. The amplitude of the column ozone solar signal varies with latitude between 2.5-7 DU/100 units of F10.7 cm flux (see figure 4-17, WMO 2002), which equates to a change of 2-3% on the column ozone at tropical latitudes, where the statistical significance of the analysis is greatest. Model studies (e.g. Fleming et al. 1995, Lee and Smith 2003) obtain estimates of the solar signal in column ozone that has a broad latitudinal structure with an amplitude of 1.5-2% (corresponding to 3-5 DU/100 units of 10.7 cm flux), which is in reasonable agreement with the observations. Lee and Smith (2003) included volcanic influences in their model and their inclusion slightly increased the solar signal in their model, but primarily outside of the tropics.

Shindell et al. (1999) and Hood (2004) note that the amplitude of the solar signal in the SBUV ozone observations is largest above 40 km (see also his figure 4), in contrast to model studies (both 2D and 3D models, including those with sophisticated coupled chemistry) that predict a maximum signal at ~ 5 hPa and much lower values above this (Brasseur 1993, Haigh 1994, Fleming et al. 1995, Shindell et al. 1999, Takigawa et al. 1999, Austin et al. 2002, Labitzke et al. 2001).

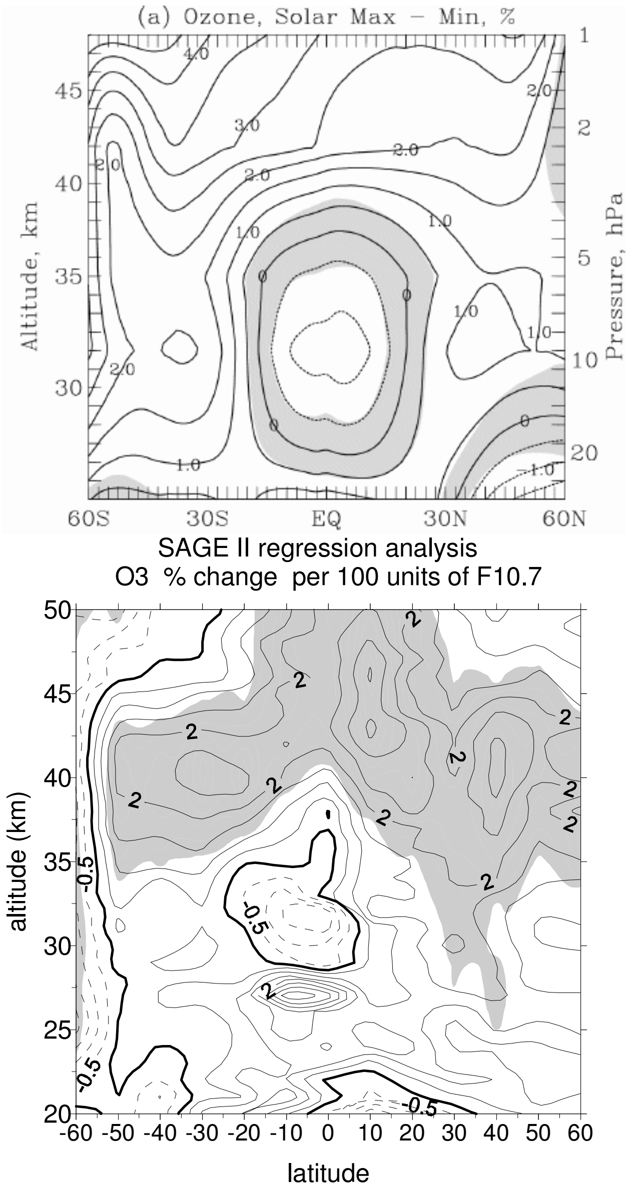


Figure 3.5: Analysis of the 11-year solar signal in ozone from the BUV/SBUV dataset (top) and the SAGE II dataset (bottom). Units are % change from solar max to solar min in the top panel and % change per 100 units of F10.7cm radio flux in the bottom panel. Shading indicates regions of low statistical significance in the top panel and high statistical significance in the bottom panel.

Hood suggests that the models may be missing an important mechanism for the ozone solar signal at these higher levels. On the other hand, the SAGE II analysis in figure 3.5 indicates peak values at around 40-45 km, which suggests there may not be a discrepancy between models and observations. Wang et al. (1996) note that the discrepancy between SAGE II and SBUV above 10 hPa is likely to be caused by incorrect adjustments to the SAGE II data. Clearly, this difference between the SBUV and SAGE analyses needs to be understood, in order to ascertain the observed height profile of the solar signal above 40 km with more confidence. If the SBUV analysis is found to be more accurate, an additional mechanism for solar influence on ozone above 5 hPa will be required for inclusion in models (but see also section 4.2.1).

A notable feature in figure 3.5 is the region of negative values centred over the tropics near 30-35 km that is present in both analyses. A negative signal in this region is not generally predicted by models. Lee and Smith (2003) suggests that this structure does not arise directly from the solar forcing, but from an interference from the volcanic and QBO influences. They included solar cycle, volcanic and QBO influences in their model. When only the solar forcing is included, a single monopole structure is generated, similar to most other modelling studies. The negative region was only reproduced when volcanic and QBO terms were included. This suggests that the feature is an artifact and that the linear regression analysis employed on the satellite data cannot completely account for non-linear interactions of these various terms in this region. On the other hand, recent experiments with the fully coupled chemistry models by Langmatz et al. (see section 4.2.1 and Haigh et al. 2004) have reproduced a similar negative signal in this region and suggest that it is related to stronger chemical and dynamical ozone destruction. An analysis of 4 Umkehr stations for the period 1957-2001 (see Haigh et al. 2004) show a negative anomaly between 30-40 km at the Mauna Loa station, which appears to substantiate the negative signal in the satellite results (figure 3.5). However, these stations were also analysed using a linear multiple regression analysis so may also suffer from an inability to account for non-linear interactions between the solar, volcanic and QBO terms, as suggested by Lee and Smith (2003).

The vertical profile of the solar signal in ozone in the very lowest part of the stratosphere cannot be assessed from the SAGE or SBUV data. However, by subtracting the available height profiles from the column amounts measured by TOMS, an estimate can be acquired of the amplitude of the solar signal coming from below about 25 km (Hood 1997). In contrast to most model predictions, the majority of the column ozone response appears to come from this lower-most part of the stratosphere, where ozone is controlled primarily by dynamics rather than by photochemistry. This suggests that despite advances in model simulations of the 11-year solar cycle response (see section 5), the modelled dynamical response in the lower stratosphere may still be inadequately simulated. This is most likely due to the inadequate modelling of the mean circulation response to solar influence on planetary wave propagation and sudden warmings, since the mean upwelling at equator latitudes is governed by the wave forcing at high winter latitudes, but the inability of models to capture this is not well understood (Hood 1997).

3.1.3 Circulation

Changes in the circulation of the middle atmosphere associated with the 11-year solar cycle have been investigated primarily using the FUB and NMC/NCEP geopotential height datasets (e.g. van Loon and Labitzke 2000, Labitzke and van Loon 2000, Soukharev and Labitzke 2000, Labitzke 2003, 2003) and derived quantities such as the balanced winds and E-P flux diagnostics (e.g. Kodera and Kuroda 2002a) and also using zonal winds from the ERA-40 assimilated dataset (Crooks and Gray 2005, Gray et al. 2004).

Recent analyses have demonstrated a stronger statistical significance of the well-characterised solar signal in 30 hPa geopotential height (e.g. van Loon and Labitzke 2000, Soukharev and Labitzke 2000, Labitzke 2001,2002, Labitzke et al. 2002) and improved characterisation of the interaction between the 11-year solar cycle and the QBO (Labitzke and van Loon 2000, Labitzke 2003, 2004). The ability of models to simulate this complicated relationship is likely to be an important test of their ability to simulate the mechanism for solar influence in the lower atmosphere, so this further characterisation is valuable.

In the upper stratosphere, the work of Kodera and co-workers using the NCEP dataset has identified zonal wind anomalies in the upper stratosphere and stratopause region associated with the 11-year solar cycle (Kodera et al. 1990, Kodera 1995, Kuroda and Kodera 2001, Kodera and Kuroda 2002a). Specifically, they observe an easterly / westerly anomaly in the subtropical winter jet in the upper stratosphere and lower mesosphere in solar minimum / solar maximum that moves polewards and downwards as the winter progresses. This solar signal in the subtropical lower mesosphere jet is statistically significant at the 99% level (Crooks and Gray 2005). However, the statistical significance of the polar signal is weaker, most likely because the inter-annual variability in this region is much greater, and more years of data will be required to improve the statistics in this region.

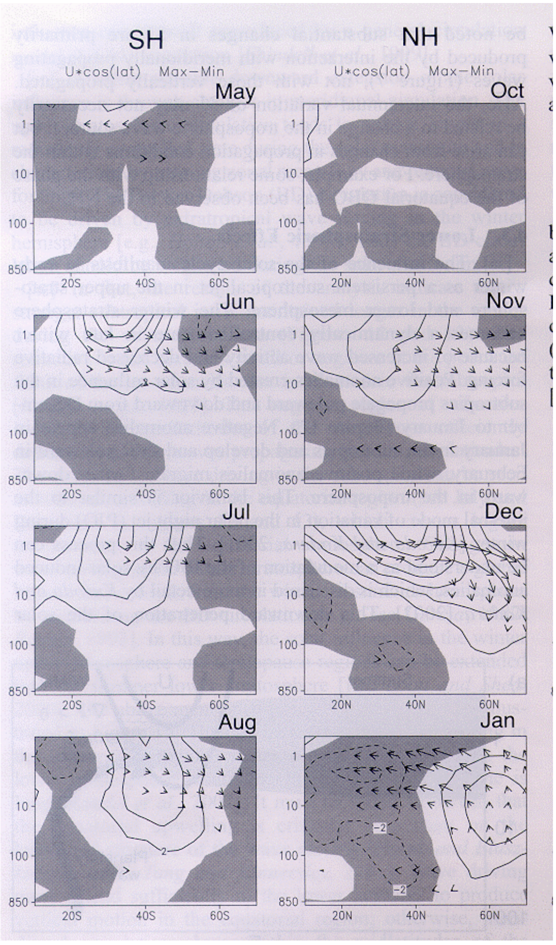


Figure 3.6: Pressure (hPa) versus latitude plots of Solar maximum minus solar minimum composite difference of zonal mean zonal winds (contour interval 2 ms^{-1}) and E-P flux vectors (arrows) from the NCEP analysis of Kodera and Kuroda (2002a). Left: Southern Hemisphere winter. Right: Northern Hemisphere winter. All variables are weighted by the cosine of the latitude. E-P fluxes are scaled by the inverse of the pressure. Zero lines are suppressed and negative values shaded.

A similar poleward and downward propagation of a subtropical easterly anomaly is associated with the evolution of a sudden stratospheric warming event, when the polar night jet is weakened from the top downwards by the action of planetary waves that propagate vertically from the troposphere. Figure 3.6 shows composite differences (solar maximum minus solar minimum) in zonal wind and E-P flux diagnostics from Kodera and Kuroda (2002a). The poleward and downward propagation of the zonal wind anomaly is clearly visible, especially in the NH. Figure 3.7 shows the associated temperature and residual circulation anomalies (also from Kodera and Kuroda 2002a). As expected, the weaker wave forcing during solar maximum results in a weaker meridional circulation compared with solar minimum years.

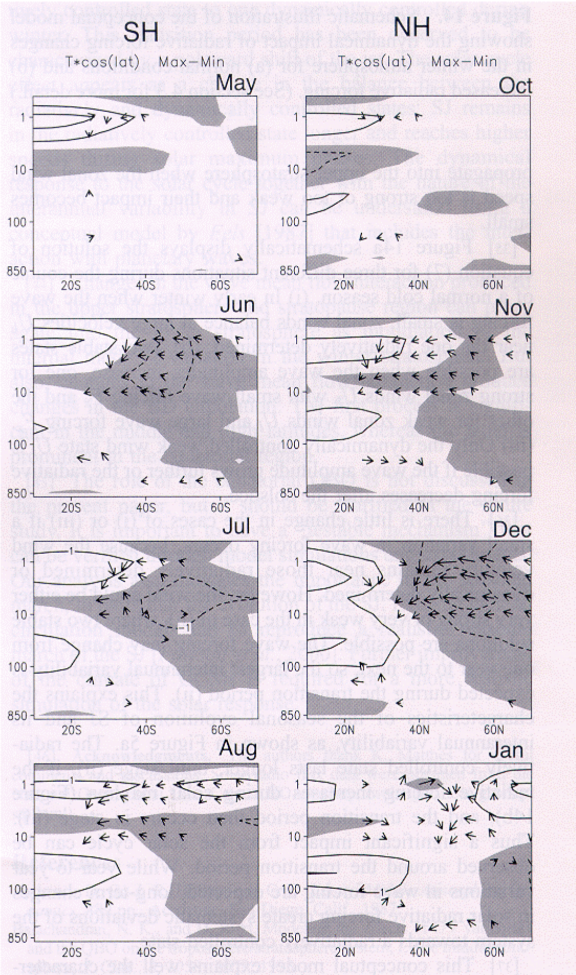


Figure 3.7: As figure 3.6 except solar maximum minus solar minimum composites of temperature (contour interval 1K) and meridional circulation (arrows).

An easterly anomaly in the subtropical upper stratosphere in solar minimum compared with solar maximum years is a manifestation of the early development of the Aleutian High which is a recognised pre-cursor to the development of a sudden warming at lower altitudes (e.g. Soukharev and Labitzke 2001). Gray et al. (2004) have used the ERA-40 dataset to clarify the nature of the zonal wind anomalies in the NH. They show that the solar cycle appears to modify the timing of sudden warmings, with earlier warmings (and hence an early-winter easterly zonal wind anomaly) in solar minimum compared with solar maximum. Figure 3.8 shows 11-year solar cycle composites of zonal wind from their study. Strong polar easterly anomalies associated with major sudden warmings clearly occur in early winter (December/January) in solar minimum years and in late winter (February / March) in solar maximum years. The solar maximum minus solar minimum differences (3rd column) compare well with the NCEP analysis of Kodera and Kuroda (2002a) in figure 3.6.

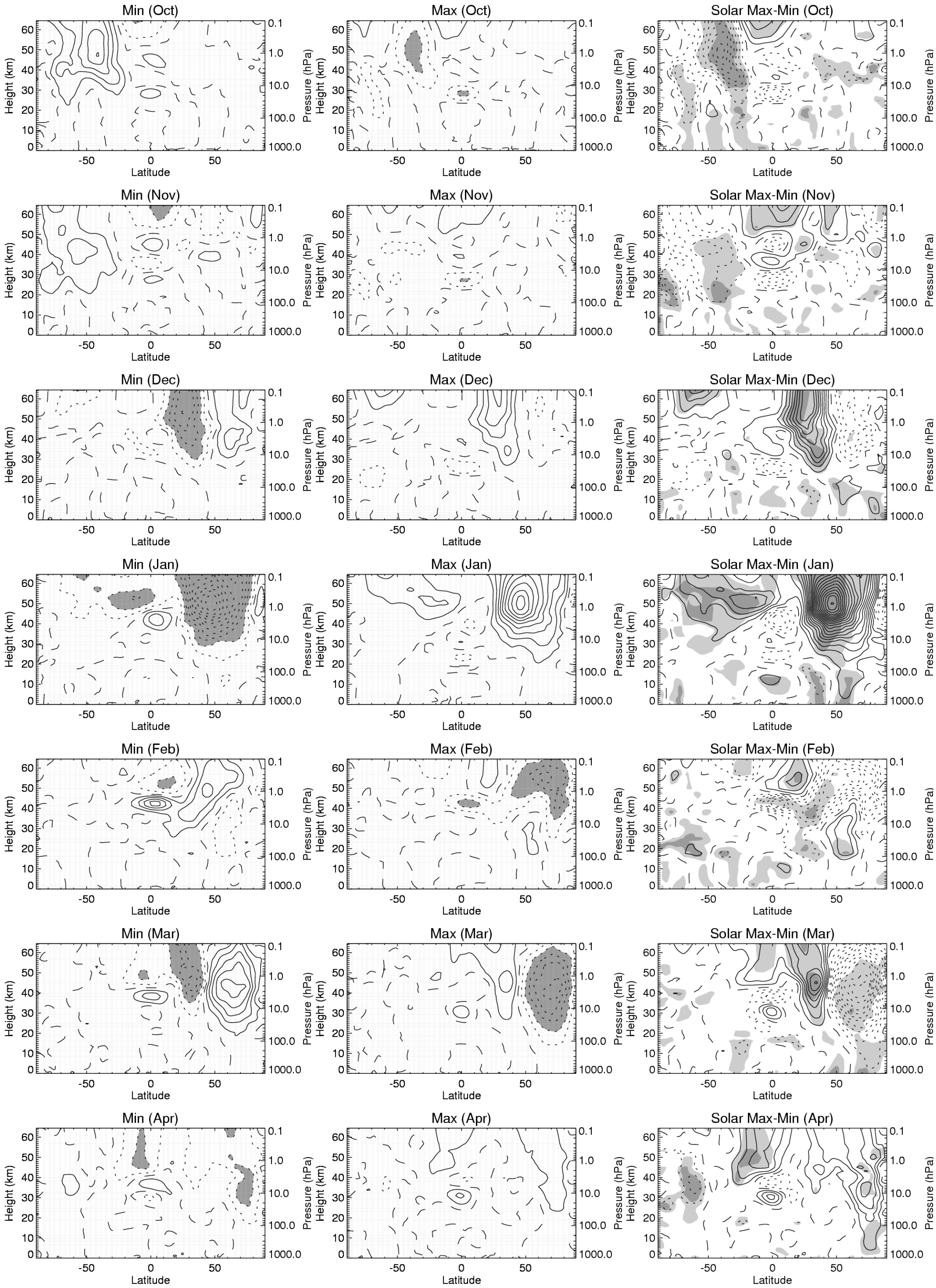


Figure 3.8: Solar cycle composites from the ERA-40 dataset. Monthly averaged latitude height sections of zonally averaged zonal wind anomalies (ms^{-1}) for the period 1979-2001. Contour interval 2 ms^{-1}. Dotted contours show negative values. In 1st and 2nd columns shaded regions show values less than 4 ms^{-1}. In 3rd column light and dark shading indicate 75% and 95% statistical significance.

The easterly / westerly polar anomalies in December and January in solar minimum / maximum years (figure 3.8) can be traced backwards in time to an anomaly of the same sign in equatorial / subtropical latitudes near the stratopause in November and even possibly in October (see figure 3.8). Gray et al. (2004) also note a similar pattern of zonal wind anomaly when the same data are composited into QBO east and west phases and they suggest this may be a clue to the observed interaction of the solar and QBO influences. This is discussed in more detail in section 5.

A further possible clue to the interaction between the 11-year solar cycle and QBO has come from the observational studies of Salby and Callaghan (2000, 2004). They examined the observed QBO in equatorial zonal winds at 50 hPa for the period 1956-1996. They showed that while the length of the easterly phase remained essentially constant throughout the period, the length of the westerly phase varied by more than a factor of two and displayed a quasi-decadal variation. Salby and Callaghan have proposed that this quasi-decadal variation is a manifestation of a modulation by the 11-year solar cycle, although Hamilton (2002) examined a slightly longer period of data (1950-2001) and found that the correlation was not so good. Nevertheless, Hamilton noted that the variation did not appear to be random and that understanding the source of this modulation should be a high priority. If this variation is indeed associated with the 11-year solar cycle, it is a good example of the complicated nature of interaction, since it is a frequency modulation of the QBO by the solar cycle and not a straightforward amplitude modulation as in the case of the temperature and ozone signals discussed previously.

3.2 Surface and lower atmosphere

3.2.1 Temperature

The solar signal in tropospheric zonal averaged temperatures and circulation has been studied in the NCEP reanalysis (Labitzke 2002, Haigh 2003, Gleisner and Thejll 2003) and in the ERA-40 reanalysis (Crooks and Gray 2005). The results are generally in good agreement with each other. Figure 3.9 shows results from the Haigh (2003) study. A positive solar response is seen in the tropical lower stratosphere i.e. warmer tropospheric temperatures in solar maximum than in solar minimum, and this signal extends in vertical bands throughout the troposphere in both hemispheres at latitudes 20^o-60^o with a maximum amplitude of 0.5 K around 40-50^oN where there are strong gradients so substantial local changes can be produced by modest North-South shifts.. Although the results are broadly similar, there are minor differences between the results of the studies that highlight the need to account for signals associated with other modes of variability. For example, Labitzke (2002) includes only the solar and linear trend, Gleisner and Thejll include the solar term, the linear trend, ENSO and volcanic signals while Haigh (2003) and Crooks and Gray (2005) include the solar term, the linear trend, seasonal cycle, ENSO, QBO, NAO and volcanic signals. As figure 3.9 shows, the QBO signal is largely confined to the tropical lower stratosphere while the ENSO signal is seen clearly throughout the tropics. Volcanic eruptions cause the stratosphere to warm and the troposphere to cool while the NAO is mainly confined to NH mid-latitudes as expected.

3.2.3 Circulation

The solar signal in tropospheric zonally-averaged zonal winds have also been analysed (e.g. Gleisner and Thejll 2003, Haigh et al. 2004, 2005, Crooks and Gray 2005). Figure 3.10 shows selected results from the Haigh et al. (2005) study. They show that the subtropical jets are weaker and further poleward at solar maximum than at solar minimum, although the magnitudes are relatively small compared with the variations associated with the NAO and volcanic signals. Gleisner and Thejll (2003), who also examined the solar signal in zonally-averaged vertical velocities and the longitudinal distribution of the solar signal in vertical velocities at the equator, note a stronger subsidence at mid-latitudes and a poleward movement of the peak subsidence. Although they do not find a significant solar signal in the strength of the ascending branch of the Hadley circulation, they do find evidence for a latitudinal broadening of the equatorial upwelling region, in good accord with the results of Haigh (2003, 2004) and Crooks and Gray (2005). In addition, they also find evidence of stronger ascending motion in the rising branch of the Ferrel circulation at mid-latitudes.

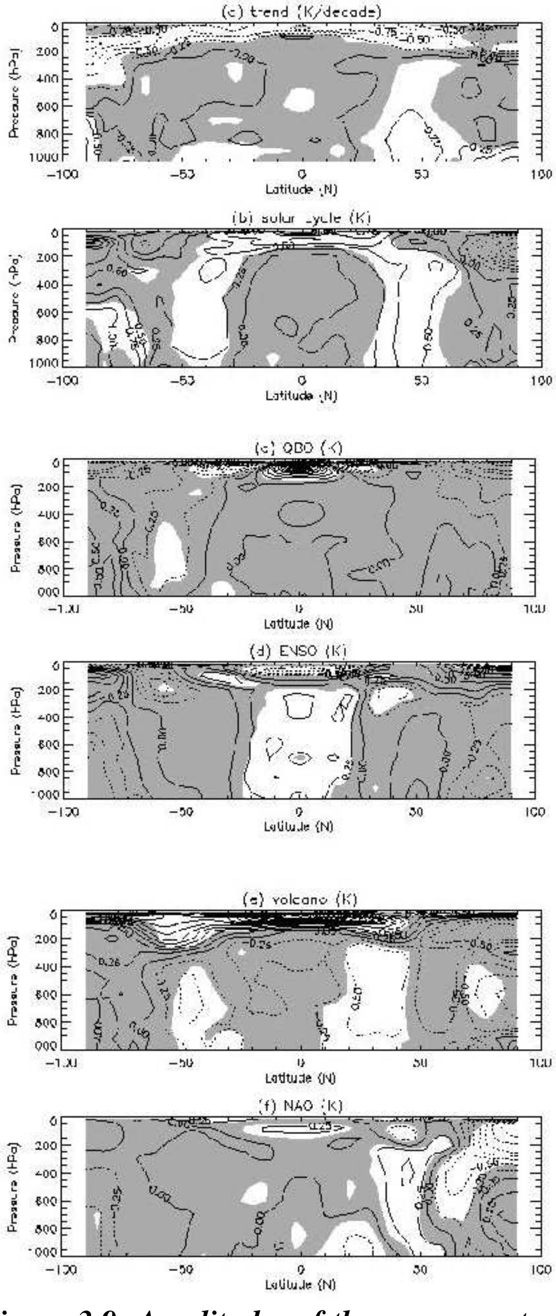


Figure 3.9: Amplitudes of the components of variability in zonal mean temperatures from the NCEP 1979-2000 study by Haigh (2003). (a) trend, (b) solar, (c) QBO, (d) ENSO, (e) volcanoes, (f) NAO. The units are K/decade for the trend, otherwise maximum variation over the data period i.e. equivalent to solar maximum minus solar minimum. Shaded areas are NOT statistically significant at the 95% level.

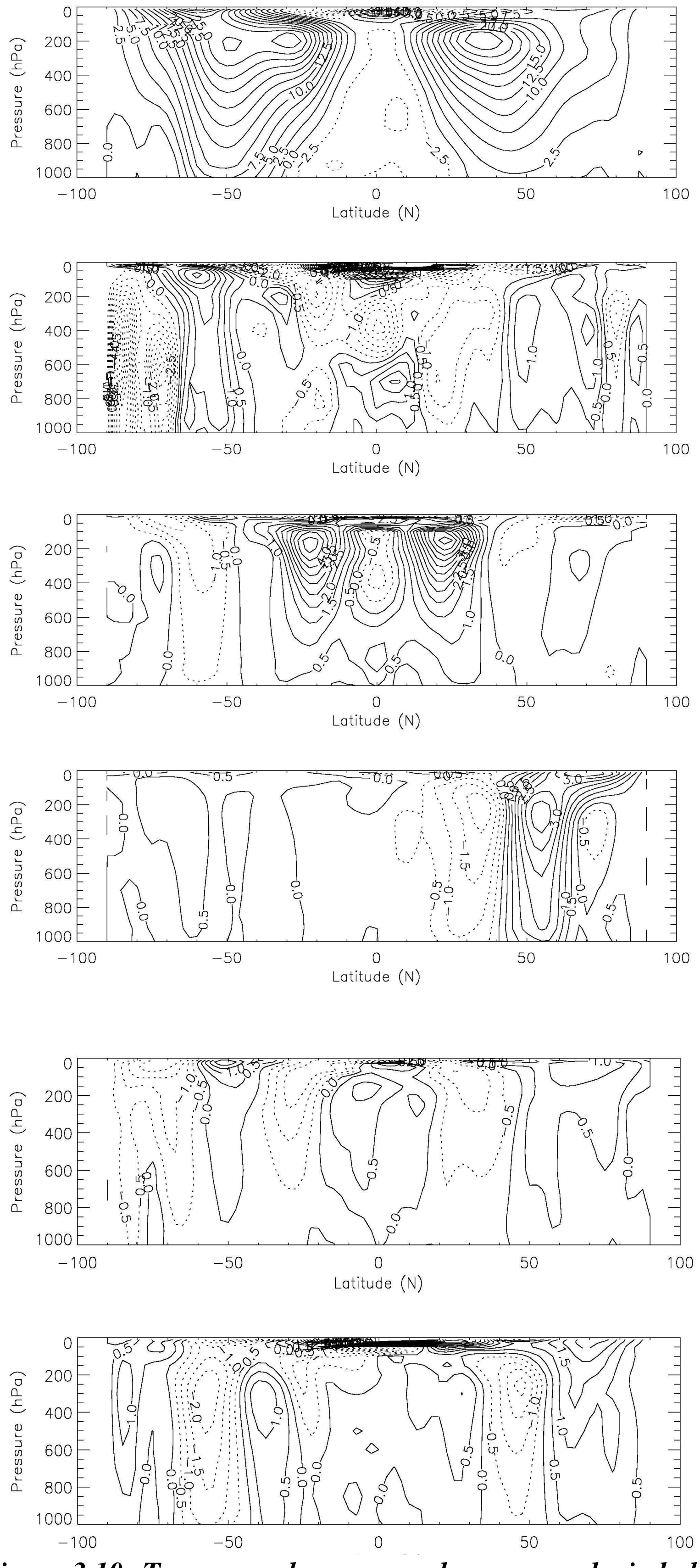


Figure 3.10: Top: annual mean zonal mean zonal wind climatology, 1979-2000 from the NCEP dataset (Haigh et al. 2005). Remaining panels: amplitudes of the components of variability in zonal mean zonal wind associated with 2nd panel:linear trend, 3rd panel: ENSO, 4th panel: NAO, 5th panel: solar, 6th panel: volcanic.

Gleisner and Thejll (2003) also examined the solar signal in specific humidity of the lower troposphere. Their results show a warmer, moister atmosphere during solar maximum. The response was spatially heterogeneous, being strongest in latitudinal bands of deep tropical convection and of mid-latitude cyclonic activity.

3.2.3 AO / NAO / surface parameters

An analysis of the solar signal in the North Atlantic Oscillation (NAO) and related surface pressure patterns has been carried out by Kodera (2002b, 2003, 2004). He found that the spatial pattern of correlations between the DJFM NAO index and global sea level pressure (SLP) differed between solar minimum and solar maximum years. Figure 3.11 shows the correlation pattern when all years of NCEP/NCAR assimilated dataset for the period 1959-1997 are included. The familiar dipole pattern centred in the Atlantic region is evident. However, when the years are divided into solar minimum and solar maximum, as shown in figure 3.12, the patterns change. In solar minimum conditions the high correlations are similarly concentrated in the Atlantic sector but in solar maximum conditions the pattern is extended across to the Eurasian sector. Kodera notes, however, that although the pattern is more hemispheric in nature in solar maximum, it does not resemble the Arctic Oscillation (AO) or the surface Northern Annular Mode (NAM) (Thompson and Wallace 2000, Baldwin and Dunkerton 2001) since the pattern does not extend across the Pacific. The authors interpret these differing patterns in terms of the different structure of the polar vortex (stronger and vertically more coherent in early winter at upper levels in solar maximum than in solar minimum), and hence its influence on the refraction of planetary waves propagating upwards from the troposphere.

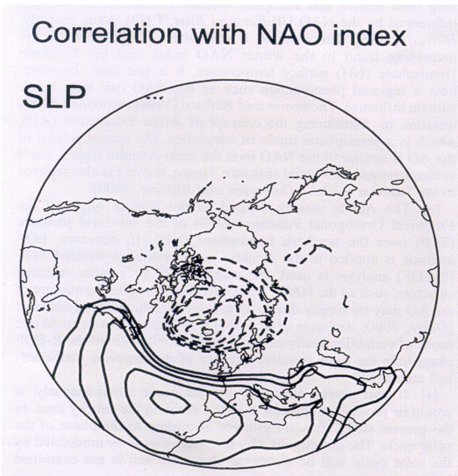


Figure 3.11: correlation coefficients between DJFM-mean NAO index and sea level pressure in the NCEP/NCAR 1959-1997 dataset, taken from Kodera (2002b). Contour interval is 0.1 and absolute values below 0.5 are omitted. Negative values indicated by dashed contours.

In addition to this modulation of the winter-time NAO by the solar cycle, Ogi et al. (2003) have also shown that the winter NAO is well correlated with the following spring / summer circulation during solar maximum years whereas in solar minimum years they found no significant correlation.

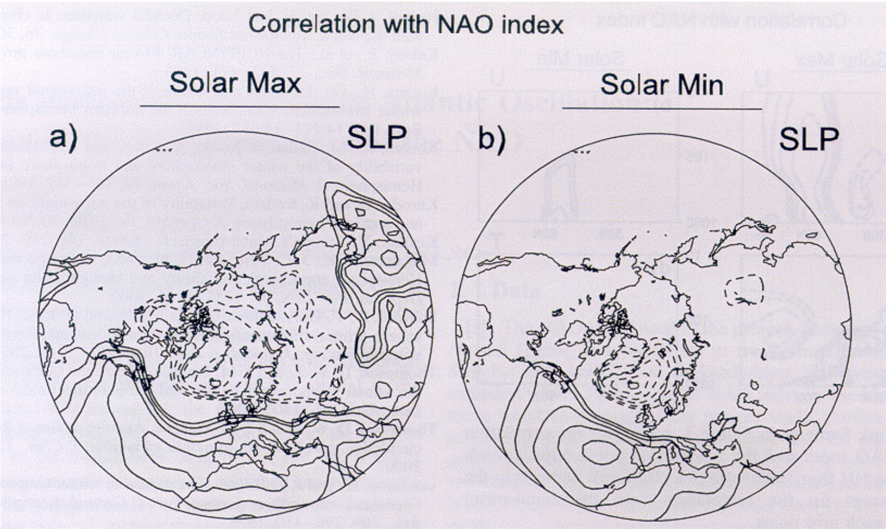


Figure 3.12: as figure 3.11 except that the data have been divided into (a) solar maximum and (b) solar minimum years.

In a similar study using the same dataset, Castanheira and Graf (2003) carried out a linear regression / correlation analysis of SLP. However, they did not divide the years explicitly into solar maximum and solar minimum years. Instead, they divided the dataset into periods with a strong or weak stratospheric vortex and showed that under a weak vortex regime (WVR) the NAO correlations with SLP are confined to the Atlantic sector whereas under a strong vortex regime (SVR) there is an additional correlation over the North Pacific, as shown in figure 3.13 and 3.14. They suggest that there is a teleconnection between SLP over the North Pacific and the North Atlantic when the stratospheric vortex is strong but not when it is weak.

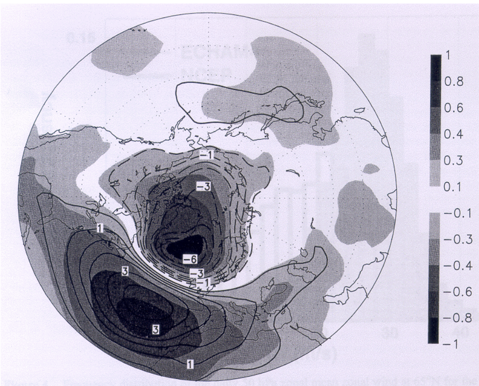


Figure 3.13: Correlation / regression pattern (denoted by shading / isolines) between the NAO index and SLP for Nov-April for the NCEP.NCAR 1950-1998 dataset, taken from Castanheira and Graf (2003). Regression values are in hPa per one standard deviation of the NAO index. Dashed contours indicate negative values. Contour interval os 0.5 for absolute values smaller than 2.0 and 1.0 for absolute values greater than 2.0.

Like Kodera, they interpret these changes in terms of the ability of tropospheric planetary-scale waves to penetrate into the stratosphere, but their interpretation is based more directly on the state of the lower stratospheric vortex strength throughout the extended winter, whereas Kodera links the solar cycle influence to the strength of the upper level vortex in the early winter (December). Although these studies are not immediately comparable, since the winters are separated on a different basis, the studies both indicate an influence on tropospheric weather patterns from the stratosphere. Further research is required using model studies to understand the mechanism and relative importance of this influence.

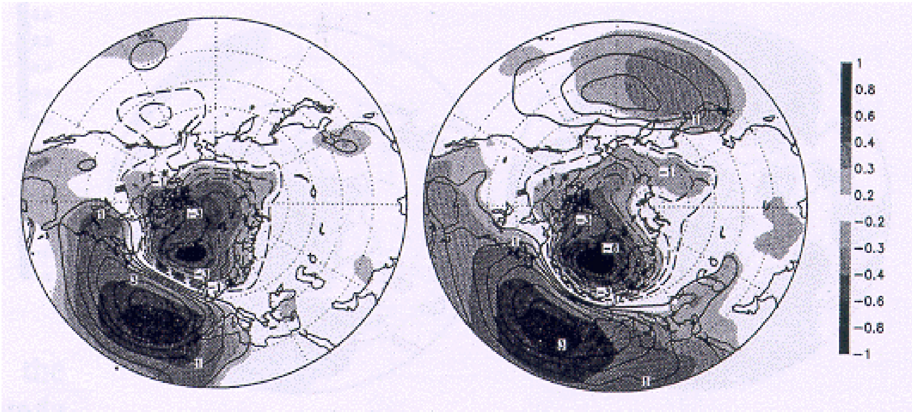


Figure 3.14: As figure 3.14 except that the dataset has been divided into years with a weak stratospheric vortex (left) and a strong stratospheric vortex (right). Weak vortex years were selected when the zonal average zonal winds at 50 hPa, 65^{o}N were greater than zero but less than 10 m^{-1}. Strong vortex years selected similarly, but when wind values were greater than 20 ms^{-1}.

4. Irradiance Mechanisms

4.1 Direct effect of solar irradiance changes

4.1.1 Radiative forcing

Top-of-atmosphere (TOA) solar radiative forcing (RF) may be deduced from anomalies in total solar irradiance using a scaling factor of 0.18 to take account of reduction by global albedo and by global averaging. Thus a 1.7Wm^{-2} increase in TSI since 1750 (as suggested by the LBB and HS reconstructions) translates into a TOA RF of 0.3 Wm^{-2}. In this context the apparently increasing uncertainty in TSI reconstructions, discussed in section 2, means that solar radiative forcing may be less well known now than was perceived during the writing of the IPCC TAR. We note in passing that the solar value included in the well-known IPCC (2001) RF bar chart might be considerably larger or smaller if a different year were assumed for the pre-industrial start-date.

The IPCC definition of radiative forcing differs from the TOA estimate in that it uses the RF value at the tropopause with the state of the stratosphere (but not the troposphere) adjusted to the forcing. This is justified by the faster equilibration time of the stratosphere and also because it has been shown that the adjusted forcing is a better indicator of global average surface temperature response.. For solar radiative forcing the first impact of this is to reduce RF because the existence of molecular oxygen and ozone in the stratosphere reduces the solar radiation reaching the tropopause by about 5%. Secondly, however, the RF value has to be adjusted to take account of the effects of any solar-induced changes in the stratosphere. Heating of the stratosphere by enhanced solar UV produces additional downward LW radiation at the tropopause, thus enhancing RF. Changes in ozone also impact the radiation fields – additional O_3 reduces the downward SW fluxes but increases the LW fluxes. Thus a precise determination of solar RF depends on the response of stratospheric temperatures and ozone to the changes in solar irradiance and, as discussed in section 3.1, there are still large uncertainties in these. Table 4.1 presents an overview of recent published estimates of solar radiative forcing and the extent to which it is amplified by changes in ozone. The amplification varies between -18% and +67%, thus even the sign is unknown, with the major uncertainty factor being the vertical profile of the ozone anomaly: upper stratospheric increases result in predominantly shortwave (-ve) impacts while lower stratospheric changes are dominated by LW (+ve) effects.

4.1.2 Modelling studies of impacts of long-term (centennial) variations in TSI

Since the TAR several modelling centres have carried out investigations into the contributions of different factors to climate variability and climate change on centennial timescales. The models include the coupled atmosphere-ocean GCMs at GISS (Rind et al, 1999; Shindell et al, 2003), UKMO (Tett et al, 2002; Stott et al, 2000, 2001, 2003), NCAR (Meehl et al, 2003) and GFDL (Broccoli, 2003) and the energy balance models (EBMs) of Crowley (2000) and North and Wu (2001). It should be noted that for these experiments the GCMs are generally run with relatively poor resolution in the stratosphere so that the effects of stratosphere-troposphere coupling may be underestimated – see section 4.2.1 for a discussion of this.

Simulations with the GCMs comprise experiments representing the climate from ~1860 to 2000 with time-evolving natural (solar and volcanic) and anthropogenic (greenhouse gases, sulphate aerosol) forcings. In order to provide estimates of natural internal variability each experiment comprises an ensemble of perhaps 3-5 runs starting from slightly different initial conditions. The GCMs are generally able to reproduce, within the bounds of observational uncertainty and natural variability, the temporal variation of global average surface temperature over the twentieth century with the best match to observations obtained when all the above forcings are included. Separation of the effects of natural and anthropogenic forcing suggests that the solar contribution is particularly significant to the observed warming over the period 1900-1940. However, uncertainties remain with the solar effect, particularly regarding the impact of the choice of solar reconstruction. The early twentieth century warming should be easier to reproduce using the HS93 reconstruction than that of LBB95 or Lean (2000) but the Broccoli et al (2003) results do show this warming using the latter and the Meehl et al (2003) results underestimate the warming despite using the former, although note that Meehl et al. do not include a representation of volcanic activity. Rind (2000) suggests, given some anthropogenic warming and large natural variability, that solar forcing is not necessarily involved but analysis by Stott et al (2000) and Meehl et al (2003) shows that it is detected.

Going back over the past millennium the EBM study of Crowley (2000) shows that natural forcings can explain the amplitude of natural variability in northern hemisphere surface temperature over the pre-industrial period. In particular he points out that the higher levels of volcanism prevalent during the 17th century, as well as lower solar irradiance, may contribute to the cooler northern hemisphere temperatures experienced at that time. This period is sometimes referred to as "The Little Ice Age" although how global it was is contentious. Hegerl et al (2003) present multiple regression analyses of two different N.H. temperature reconstructions, and repeat the EBM calculations of Crowley (2000) using a modified volcanic time series and an unidentified solar time series. They find a small but detectable solar response of about 0.15K over the past millennium.

Historically many authors have suggested, based on analyses of observational data, that the solar influence on climate is larger than would be anticipated based on radiative forcing arguments alone. The problems with these studies (apart from any question concerning the statistical robustness of their conclusions) is that (i) they frequently apply only in certain locations and (ii) they do not offer any advances in understanding of how the supposed amplification takes place. Recently some interesting developments have been made to address these problems based on two different approaches: the first uses detection/attribution techniques to compare model simulations with observations. These studies (North and Wu, 2001; Stott et al 2003, Crooks et al. 2005) show that the amplitude of the solar response, derived from multiple regression analysis of the data using model-derived signal patterns and noise estimates, is larger than predicted by the model simulations by up to a factor 4. This technique does not offer any physical insight but suggests the existence of deficiencies in the models. It also shows how the solar signal may be spatially distributed although the modelled large-scale spatial near-surface temperature response to TSI forcing is very similar to that shown for GHG forcing. Use of better vertical resolution in the stratosphere and better spectral resolution in the solar irradiance changes may affect the patterns.

The second approach proposes feedbacks in the climate system (that may already exist in GCMs) and uses these to explain features found both in model results and in observational data. Dynamical coupling between the middle and lower atmosphere is discussed in the next section; other mechanisms are generally concerned with thermodynamic processes, water vapour feedback and clouds. It has been suggested that the solar influence might be larger where there is less cloud (so that more radiation is absorbed at the surface) and that this would lead to changes in circulation associated with anomalies in horizontal temperature gradient. Rind et al (1999) found no evidence that it was possible to differentiate solar from greenhouse warming on the basis of latitudinal temperature gradients and little consistent relationship between changes in global temperature and low latitude temperature gradient.

The Meehl et al (2003) model results, however, showed an amplification of the solar influence when this acted in combination with greenhouse gases which was explained by the spatial heterogeneity of the solar surface forcing. In cloud free regions of the sub-tropical oceans enhanced solar forcing produced greater evaporation, greater moisture convergence and strengthening of regional monsoon, Hadley and Walker circulations. This result is consistent with observed trends in tropical rainfall over the 20th century but not with some analyses of the zonal mean Hadley circulation on 11-year timescales which shows a weakening when the Sun is more active (see discussion in section 5.2.3).

4.1.3 Modelling studies of impacts of 11-year cycle variations in TSI

Little work has been published in this area since the IPCC TAR probably because (i) such cyclical variations are not perceived to be significant in the context of climate change and (ii) studies of 11-year cycle influence have focussed on the middle atmosphere and its potential to influence the troposphere below. In the middle atmosphere the spectral composition of the variation in solar irradiance is key. This work is now discussed in some detail in the next section, as any mechanisms operating on 11-year timescales will presumably also be involved in the centennial-scale solar influence.

Table 4.1 ***Direct solar radiative forcing calculated by different authors and the adjustments due to the inclusion of solar-induced changes in stratospheric ozone, all in Wm^{-2}. The final column shows the percentage amplification by ozone of the direct forcing.***

author	solar change	direct RF at tropopause	ΔO_3	O_3 SW effect	O_3 LW effect	net O_3 effect	RF amplification by O_3 (%)
Haigh 1994	**11-year amp**	**0.11**	**+ve peak near 40km**	**-0.03**	**+0.02**	**-0.01**	**-9**
Hansen et al 1997	**11-year amp**	**0.15**	**+ve 10-150hPa**			**+0.05**	**+33**
Myhre et al 1998	**11-year amp**	**0.11**	**+ve**	**-0.08**	**+0.06**	**-0.02**	**-18**
Wuebbles et al 1998	**c1680-c1990**	**1.1**	**+ve peak near 40km**			**-0.13**	**-12**
Larkin et al 2000	**11-year amp**	**0.18** **0.18**	**+ve (as Haigh 94)** **+ve (SBUV/TOMS)**	**-0.06** **-0.03**	**+0.06 - +0.14** **+0.04 - +0.12**	**0.00 - +0.12** **+0.01 - +0.09**	**0 - +67** **+6 - +50**
Shindell 2001	**1680-1780 (low H_2O)**	**0.26 - 0.33**	**-ve (upper strat)**			**+0.02**	**+6 - +8**

4.2 Indirect effect of UV

4.2.1 Model simulations

As noted in the previous report, the observed variations in solar irradiance are not evenly distributed across the solar spectrum, but are concentrated in the ultra-violet. This can lead to changes in stratospheric ozone production and hence, through the associated changes in stratospheric heating, to changes in circulation patterns. This is generally referred to as the indirect solar modulation of climate via UV.

Modelling studies of the solar modulation of climate via UV have traditionally been carried out by first estimating the predicted ozone changes between solar minimum and solar maximum (or 20th century versus Maunder Minimum values) using 2-d models that include relatively sophisticated chemical schemes. These monthly-averaged, zonally-averaged ozone changes are then used as input to the radiation schemes of full GCM simulations. In this way, the temperature and circulation response to the ozone changes can be assessed, although there is no possibility for those temperature and circulation changes to interactively feed back onto the ozone distributions. The prime methodology employed to assess the impact of including these additional ozone changes is to carry out model simulations with only TSI changes and compare them with simulations with both TSI and ozone changes. Most 'process' modelling studies, that seek to simulate the solar signal and thence explore the process mechanisms, employ models that extend from the ground to around 80 km in order to fully resolve the ozone distribution in the stratosphere.

Because of computing resource constraints, these GCM simulations usually consist of two 20-30 years in which the model is run under perpetual solar minimum conditions and perpetual solar maximum conditions and the difference between the two runs is used as an estimate of the peak-to-peak solar signal. By running for many years in 'perpetual' mode like this, the statistical significance of the results is substantially improved. In order to gain the equivalent statistical significance from a single simulation in which the time-varying 11-year solar signal is imposed, the simulation would need to be many hundreds of years long. Employing a coupled ocean-atmosphere model in which the ocean temperatures can adjust to the imposed solar changes is inappropriate for integrations of this type employed for 11-year solar cycle studies, because the atmosphere is never actually in solar minimum or maximum for long enough for the ocean to adjust to any great extent. In this case, the models are used with observed or climatological-mean sea surface temperatures imposed at their lower boundaries. On the other hand, estimates of longer term solar changes, such as those between 20th century and Maunder Minimum values, are more appropriately carried out using a coupled model in which the ocean temperatures are able to respond and feedback onto the atmosphere component of the model.

Recently, improved computing capabilities have allowed the development of GCMs that include fully-coupled chemistry schemes so that improved feedback is possible not only from the ozone changes on to the temperature and circulation patterns but vice versa. However, the use of these fully-interactive chemistry GCMs for studies of the solar cycle influence is still at a relatively immature stage (e.g. Labitzke et al. 2002, Tourpali et al. 2003, Egorova et al. 2004, Haigh et al. 2004, Langematz et al. 2004, Austin et al.??).

Ozone

Figure 4.2.1 shows the percentage ozone changes predicted by two different models as a result of specifying the appropriate TSI and UV solar irradiance spectrum changes. They agree reasonably well with each other and with the observed ozone changes (see figure 3.5) although there is significantly less latitudinal structure and the values in the upper stratosphere are underestimated (see Shindell et al. 1999). Neither model predicts a region of negative signal in the equatorial lower stratosphere. However, as discussed in section 3.5 this may be an anomalous signal in the observational results due to the regression analysis employed (Lee and Smith 2003). On the other hand, recent results from the fully-coupled FUB GCM (Langematz et al. 2004, see also the summary in Haigh et al. 2004) reproduce this feature and attribute it to the inclusion in their model of an additional source of NOx due to Energetic Electron Precipitation (EEP), which is episodic but occurs more frequently during solar minimum (Callis et al. 2001). More measurements and improved methods of data analysis are needed to ascertain whether this feature in the observations is real. Similarly, the FUB results have suggested sensitivity to the inclusion of EEP that needs further investigation.

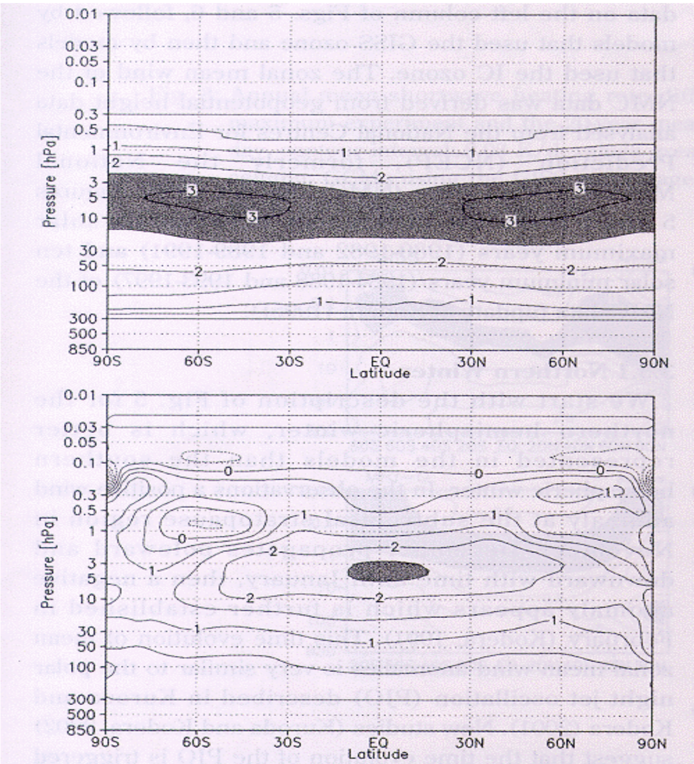


Figure 4.2.1: Solar maximum minus solar minimum difference in the annual-mean ozone in percent calculated with (a) the Imperial College 2-D chemical transport model (Haigh 1994), and (b) the chemistry GISS GCM (Shindell et al. 1999). Contour interval: 0.5%. Values greater than 2.5% have been shaded for emphasis.

Temperature

An extensive GCM inter-comparison was carried out (Matthes et al. 2003) as part of the GRIPS project (GCM Reality Intercomparison Project for SPARC – Stratospheric Processes and their Relation to Climate). See also Haigh (1999), Larkin et al. (2000), Shindell et al. (1999, 2001) for descriptions of individual models that contributed to it. The models were run under solar minimum and solar maximum conditions with identical TSI changes and with specified, monthly-averaged percentage ozone change distributions imposed via the models' radiation schemes (i.e. the models were not interactive coupled chemistry GCMs).

Although the resulting heating changes were similar in each of the models, the modelled annual-mean temperature change between solar maximum and solar minimum, shown in figure 4.2.2, were different and did not compare particularly well with observations, although this disagreement is probably no worse in these models than in the coupled-chemistry model simulations (see below). Although all models reproduced the sign of the change, with warmer temperatures in solar maximum than in solar minimum, the models generally underestimate the amplitude of the temperature change in the upper stratosphere and do not reproduce the secondary temperature maximum in the lower stratosphere. Neither is the reversal in the signal at high latitudes reproduced. The main reason for these discrepancies was identified as the models' basic climatologies and associated variability. The paper stressed the importance of achieving a good background temperature and zonal wind climatology and variability, since this has a crucial impact on wave propagation, wave-mean interaction and hence on the mean meridional circulation. By comparing results from the same model with the different imposed ozone distributions it was also shown that an individual model's temperature response was very sensitive to the imposed ozone change, so improved observations are essential in order to validate these model simulations.

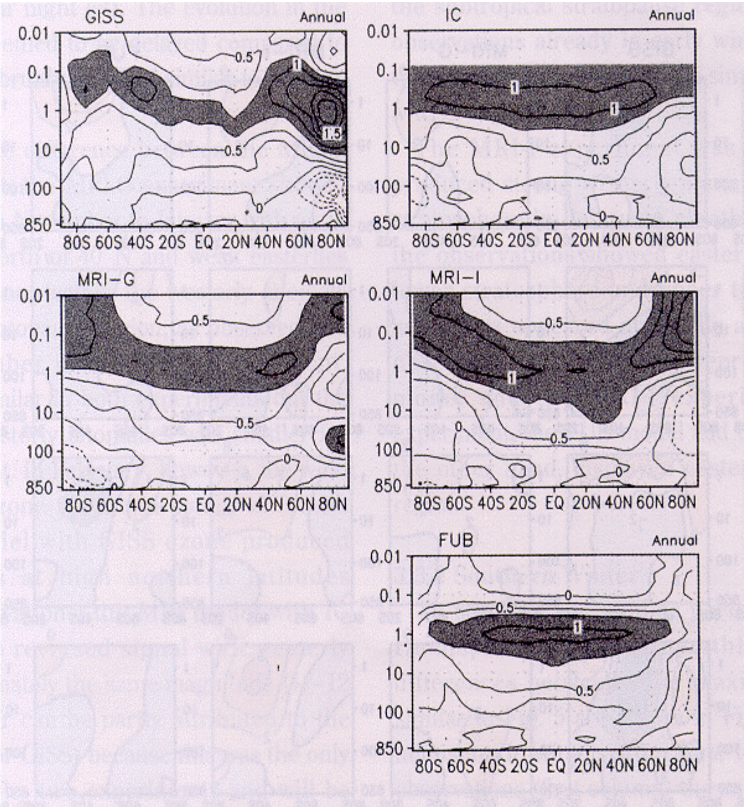
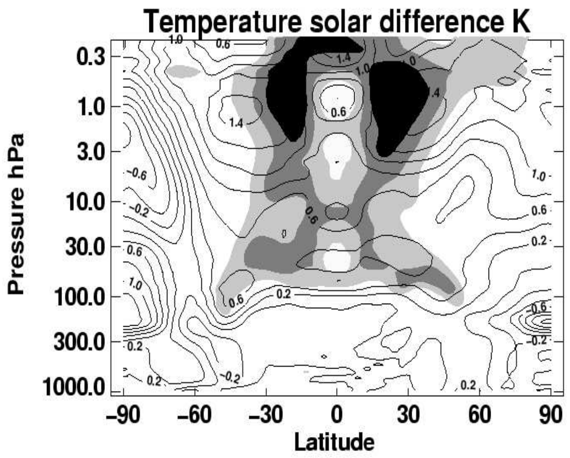


Figure4.2.2: Annual-mean solar maximum minus solar minimum temperature differences (K). The GISS and MRI-G models (left panel) used the GISS percentage ozone changes shown in the lower panel of figure 4.2.1. The IC, MRI-I and FUB models employed the IC percentage ozone changes shown in the top panel of figure 4.2.1. Contour interval is 0.25. Values greaer than 0.75 have been shaded for emphasis. Figure taken from Matthes et al. (2003).

Fully coupled chemistry climate models have also recently been employed for studies of the 11-year solar cycle influence. A comparison of the 11-year solar cycle annual-mean temperature signal from 2 fully coupled chemistry climate models (see Austin and Butchart 2003, Langematz et al. 2004) was carried out as part of the EU SOLICE project (Solar Influence on Climate and the Environment) and is summarised in Haigh et al. (2004). Figure 4.2.3 shows the solar signal in annual-mean temperature from the two models. Again, neither model was able to adequately reproduce the secondary temperature maximum in the lower stratosphere nor the reversed temperature signal at high latitudes (see figure 3.2 and 3.3).



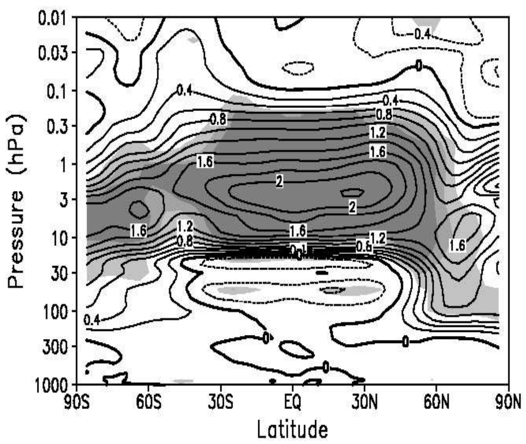


Figure 4.2.3: Annual-mean, zonal-mean solar maximum minus solar minimum temperature difference (K) from top: the UK Met Office UMETRAC model and bottom: the FUB-CMAM-CHEM model. The shading denotes 80%, 95% and 99% statistical significance.

4.2.2 Middle atmosphere dynamical influence

Planetary wave propagation and solar / QBO interaction.

In section 3.1.3 a series of papers was discussed in which Kodera and co-workers confirmed their earlier work (e.g. Kodera et al. 1990, Kodera 1995) that highlighted a slowly-propagating zonal wind anomaly that moves poleward and downward throughout the winter months in each hemisphere (Kodera and Kuroda 2000a,b, 2002, Kuroda and Kodera 2002, Kodera et al. 2000, 2003). Figure 3.6 shows this anomaly in the Northern Hemisphere from their NCEP data analysis and also the associated EP-flux changes, illustrating the change in planetary wave propagation. 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 in Rind et al. (2002) was able to simulate 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 solar cycle 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 will gradually descend with time into the lower stratosphere (see also Dunkerton 2000). In addition, they note that the QBO influences the *latitudinal wind shear* in the *lower* stratosphere (Holton and Tan 1982). This also influences the structure of the polar night jet and therefore allows an interaction of the solar and QBO influences to occur, 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 solar cycle influence in the upper stratosphere impacts the QBO influence in the lower stratosphere.

More details of this solar cycle / QBO interaction has been provided in a series of mechanistic modelling studies, Gray and co-workers (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 (see figures 3.6, 3.8). Hence, 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 and section 3.1.1) 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.

As discussed in section 3.1.3 Salby and Callaghan (2004) have noted an 11-yr SC modulation of the frequency of the QBO in the lower stratosphere (see section 3) and this has been successfully simulated in a 2-d model by McCormack (2004). 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.

Although the majority of research in this area has concentrated on a mechanism for influencing planetary waves in the stratosphere, Arnold and Robinson (2000, 2001) have suggested a role for changes in the thermosphere to influence planetary waves propagation. They found they could achieve a qualitatively similar response to solar forcing when only changes to the thermospheric temperatures in their models were introduced, although further investigation of this result is required to demonstrate statistical significance.

In summary, there is general consensus that a solar modulation of the background temperature and zonal wind structure will influence planetary wave propagation. Although recent advances have been made (Matthes et al. 2004) there are still several aspects that are not adequately reproduced. The most important of these is the inability to reproduce the lower stratospheric temperature anomaly (at all latitudes and seasons), since this is likely to be the most important influence on tropospheric climate, as discussed in the next section. 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.

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 direct TSI and indirect UV 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 could therefore explain the secondary temperature maximum in the lower stratosphere seen in the observations (section 3). It would 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.

The fact that most of the modelling studies are unable to reproduce the secondary temperature (and ozone) maximum in the lower stratosphere through a modification of the meridional circulation as outlined above suggests an underestimation of the modelled dynamical feedback which requires further investigation. Very few of the models employed include an adequate gravity wave parametrization scheme and this may be a factor that requires improvement (Arnold and Robinson, 2003). Despite recent progress in modelling the NH winter interaction between the solar and QBO influences, there are still many deficiencies in the model simulations (Labitzke et al. 2002) that need addressing further.

Solar signal penetration to the troposphere

A solar signal in the lower stratospheric temperature and circulation is likely to 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 assess separately the approaches focussing on polar and tropical processes.

(a) Polar influence

Variations in the strength of the stratospheric polar vortex have been shown to influence surface climate in both hemispheres, most often expressed in terms of a modulation of the AO/NAO (Thompson and Wallace 1998, Baldwin and Dunkerton 1999, Thompson et al. 2004). In a series of papers, Kodera (2002b, 2003, 2004 - see section 3.2.3) has examined the relationship between the NAO and surface pressure in winters separated into solar maximum and solar minimum years (see also Ogi et al. (2003) and found a significant signal. Similarly, Castanheira and Graf (2003) separated winters into those with a strong stratospheric vortex versus those with a weak stratospheric vortex and found a significant difference, although understandably not identical to that of Kodera.

There is, however, as yet no consensus on the mechanism of this stratospheric influence on the troposphere and surface (Baldwin and Dunkerton 2004). A simple transfer of pressure anomalies in the stratosphere directly to the surface can perhaps explain some of the influence (Oortland and Dunkerton 2004, Baldwin and Dunkerton 2004). Wave reflection in the upper stratosphere is another possibility (Perlwitz and Harnik 2003). Several authors (Shindell et al. 2001, Kodera 2002b, Castanheira and Graf 2003, Coughlin and Tung 2004b) have proposed a similar modulation of planetary wave propagation to that described earlier, but operating in the very lower-most stratosphere, thus influencing the troposphere also. Shindell et al. (2001) note that since planetary waves are generated by topography and land-sea contrasts they are longitudinal asymmetric. They find that the impact of solar forcing in their model is not as symmetric as that of greenhouse gas forcing because of this.

(b) Tropical influence

As discussed in section 5.1, some model studies show 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) in 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), Matthes (2003) and Palmer (2004).

Experiments with a simplified GCM (sGCM) provide some indication as to how these responses 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 (see section 7), although the source of the stratospheric heating remains an important factor.

5. High energy particles and clouds

In the previous review of solar changes on the earth's climate (HS99), the possibility of a cosmic ray influence on climate through clouds was extensively considered, motivated by the recent work showing a correlation between satellite data cloud measurements and cosmic rays. The conclusion was that the possibility should not be excluded, but that there was low confidence that a significant link existed between cosmic rays and clouds. There were several caveats made, including drawing attention to the lack of definitive experiments on ion-aerosol microphysics in the atmosphere. However several laboratory experiments did indicate that nucleation of particles from ions could occur in suitable circumstances. At that point it was considered that there was partial evidence for a physical influence of electrically-charged aerosol on super-cooled clouds, possibly leading to changes in freezing.

Since then there has been further work on the related topics, and new ideas and interpretations have developed. In what follows, Section 5.1 provides an overview of the more general material, and the ideas more recently developed. Section 5.2 gives a discussion of studies on cosmic rays, which maybe relevant to the atmospheric problem. Much of the recent work is concerned with further atmospheric data analysis of satellite cloud data, and is summarised in sections 5.3 and 5.4. Only a few further atmospheric measurements on the ion-aerosol microphysics have been made, which are included in a discussion of mechanisms in section 5.5.

5.1 General comments on the cosmic ray climate problem

The fact that ion formation by cosmic rays occurs throughout the lower atmosphere to the surface is now more widely appreciated by atmospheric scientists than when HS99 was written. In part, this may due to review papers written to summarise physical processes of ions and aerosols relatively unknown by modern scientists. The lack of knowledge is understandable as ion-aerosol physics has historically been a sub-topic of atmospheric electricity, which itself falls between meteorology, atmospheric chemistry and solar-terrestrial physics. Because the working boundaries of these scientific communities seem to be based on fairly arbitrary historical distinctions, the modern review papers are of particular value in linking the ideas and topics between the disparate scientific communities.

At a 1999 workshop on *Solar Variability and Climate* organised by the International Space Science Institute in Bern, important links were made between scientists working in solar and atmospheric physics. The workshop resulted in a special issue of *Space Science Reviews* in which the cosmic rays-cloud-climate results from data were discussed alongside possible physical mechanisms. A review of cosmic ray variability (Bazilevskaya, 2000) emphasised the difference between galactic cosmic rays and solar cosmic rays, and how this could be used to investigate possible atmospheric responses. The microphysics of aerosol on cloud boundaries as a link between high-energy particles and the global circuit was summarised by Tinsley (2000), with calculations on effects of image charge on aerosol collection by droplets given in Harrison (2000). Bazilevskaya (2000) summarised how ions could affect atmospheric chemical reactions, droplet nucleation, ice nucleation and the global atmospheric electrical circuit. Perhaps resulting from some of this work, the widely-held objection to the cosmic ray hypothesis of no known physical mechanism linking cosmic rays and clouds seems to have become less prevalent.

A summary of two physical mechanisms potentially relating cosmic rays to cloud modulation is given in Carslaw *et al* (2002). These two mechanisms are very different: one (the "ion-aerosol clean-air" mechanism) is suggested to occur in cloud-free clean air in which there is an abundance of condensable vapour, and the other (the "ion-aerosol near cloud" mechanism) argues that a cloud layer can couple changes in the global atmospheric electrical circuit to charged aerosol. This work emphasised that, as well as the 11-year cycle in solar variability, longer-term solar changes would also lead to cosmic ray changes on long timescales. These may also influence clouds. Such changes can be inferred from cosmic ray proxies such as cosmogenic isotopes and geomagnetic measurements (see figure 5.1).

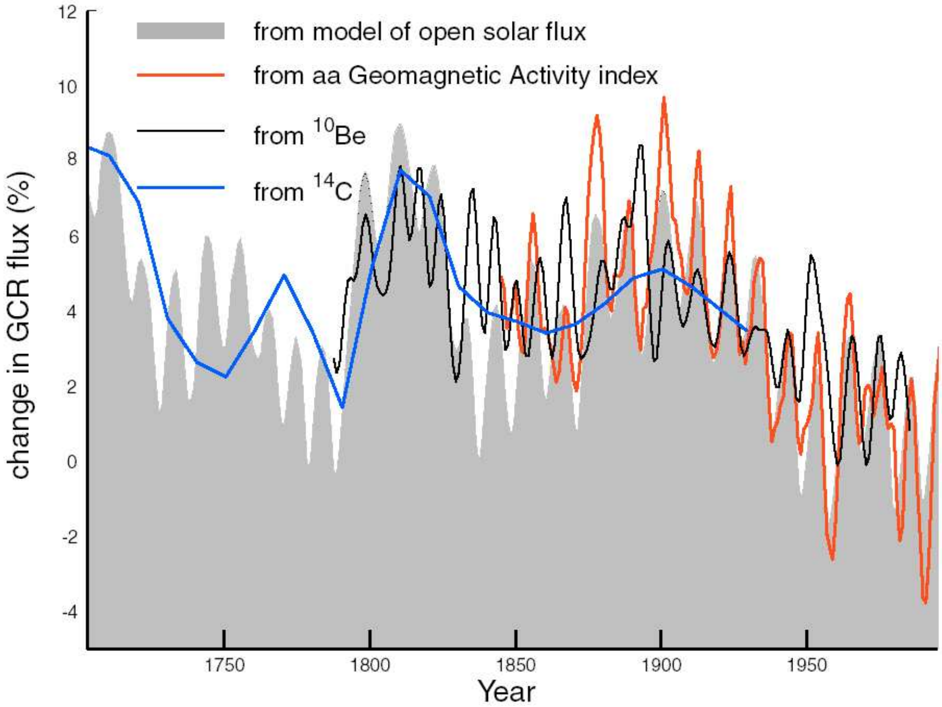


Figure 5.1. Change in cosmic ray intensity between 1700 and the present day from four independent proxies. Intensities have been scaled to the13 GeV cosmic ray data from Huancayo, Hawaii and then normalized to the 1990-2001 mean. The plot shows % deviation from this mean (from Carslaw et al, 2002).

The microphysics of the clear-air mechanism depends on the laboratory experiments and theoretical models showing the formation of ultrafine particles from ions. Modelling work (Yu and Turco, 2001) illustrate that the particles formed can grow to sizes sufficient to act as cloud condensation nuclei Harrison and Carslaw (2003) discuss further aspects of the aerosol microphysics in detail, and address the effects of aerosol electrification on aerosol coagulation and charged aerosol scavenging by water droplets. Charge can effect aerosol coagulation, but the charge distribution and coagulation timescales in the atmosphere are often similar. Drawing general conclusions is therefore difficult: for an aerosol carrying a certain charge distribution, the net effect may be a decrease in Brownian coagulation. A further review by Tinsley and Yu (2004) indicates that both the clean-air and near-cloud mechanisms could be used to explain the cosmic ray-cloud linkage, and there is no reason why the global circuit and direct ionisation effects could not act in combination, or indeed in combination with TSI changes. More modelling work is needed to link the emerging observations of aerosol nucleation rates with CCN populations, if the effects on clouds are to become usefully quantified.

5.2 Cosmic rays

(a) Changes in galactic cosmic rays

Cosmic ray data is generally obtained from surface neutron counter measurements, made at a selection of sites globally, or from balloon-carried sensors (Bazilevskaya, 2000). These respond to galactic cosmic rays and solar cosmic rays, although generally the galactic cosmic rays (GCR) are considered to be important in the studies because of their modulation over the solar cycle. For almost all of the atmospheric studies, the neutron counter measurements, which began at Climax in Colorado in the early 1950s (Simpson, 2000), have been used: these are widely available. In many cases, there are similar changes in galactic cosmic rays recorded at all neutron stations at different geomagnetic latitudes, but for some events, *e.g.* producing solar protons, only some of the stations are affected. Before the neutron measurements, surface ion chambers provide cosmic ray data back to 1937 (Ahluwalia 1997), with disparate measurements before then at different sites, including the global cruises of the geomagnetic survey *Carnegie* beginning form 1915. The geomagnetic *aa*-index extends back to 1868 on a monthly basis, which has been used to infer solar changes: before this series begins, cosmogenic isotopes are the principal source of evidence of cosmic ray variations, although for isotopes found in ice cores (Beer, 2000) there may be some atmospheric effects on the variations found.

From these various indicators, there seems to be a general consensus on increases in solar activity during the twentieth century (Lockwood, 2001), but it is contentious (Lockwood, 1999; Cliver and Ling, 2002; Richardson et al, 2002) whether the changes occurred principally in the first half, or have continued through both halves. The reconstruction of Usoskin et al (2003) suggests that the period of high solar activity during the last 60 years could be unique in the past 1150 years. If confirmed, this may not, however, be a good indication of a high TSI, following the discussion in section 2.1. Solar changes cause changes in cosmic ray fluxes in the terrestrial atmosphere, by modulation through the heliospheric field (Lockwood, 2001). There is an inverse relation: at solar maximum, the atmospheric cosmic ray fluxes are reduced.

Because of this, there are close negative correlations between GCR fluxes and TSI, with $r = -0.93$ for one-year running means. The TSI lags 2.8 (0.8-8.0 months at the 90% level) behind the open solar flux (Lockwood, 2002). This means that distinguishing between climate effects caused by TSI and GCR is difficult, and depends on identifying responses unambiguously caused by one and not the other. This can be achieved using different spectral properties in the time domain, or by studying events which have geophysical consequences rather than radiative, such as Forbush decreases (FD) in cosmic rays. Another example which may permit separating radiative and cosmic ray effects is the study of terrestrial geomagnetic changes, which will only affect the cosmic rays. Feynman and Ruzmaikin (1999) have argued that the tropospheric region open to cosmic rays of all energies was a relatively small high-latitude region in the early part of the century, and that, with time, the size of this region increased by over 25%. Their 6.5 degrees equatorward shift in the yearly-averaged latitudinal position of the sub-auroral region may have changed the region in which, if there is a GCR modulation, cloud cover may be affected.

There is considerable difficulty in separating cosmic ray and solar flux variations using purely statistical approaches, as the two quantities are closely inversely correlated. This is illustrated in figure 5.2, in which, following van Loon and Labitzke (2000), correlations with the 30mbar geopotential height are compared for 1958-1998. Figure 5.2(a) reproduces the results in figure 3 of van Loon and Labitzke (2000), for the spatial correlation variation between the 10.7 solar flux and the 30mbar height. In figure 5.2(b), the monthly surface neutron counts from the University of Chicago's monitor at Climax are used instead. Figure 5.2(a) and figure 5.2(b) are, as expected, inversely correlated, although there are slight differences in position of the regions of high correlation, and magnitude of the correlation. As in each case there are physical reasons for expecting a correlation (*e.g.* the 30mbar height is close to the meson producing region, the temperature of which is related to the surface neutron yield), detailed theoretical predictions are needed to distinguish between the TSI and GCR effects.

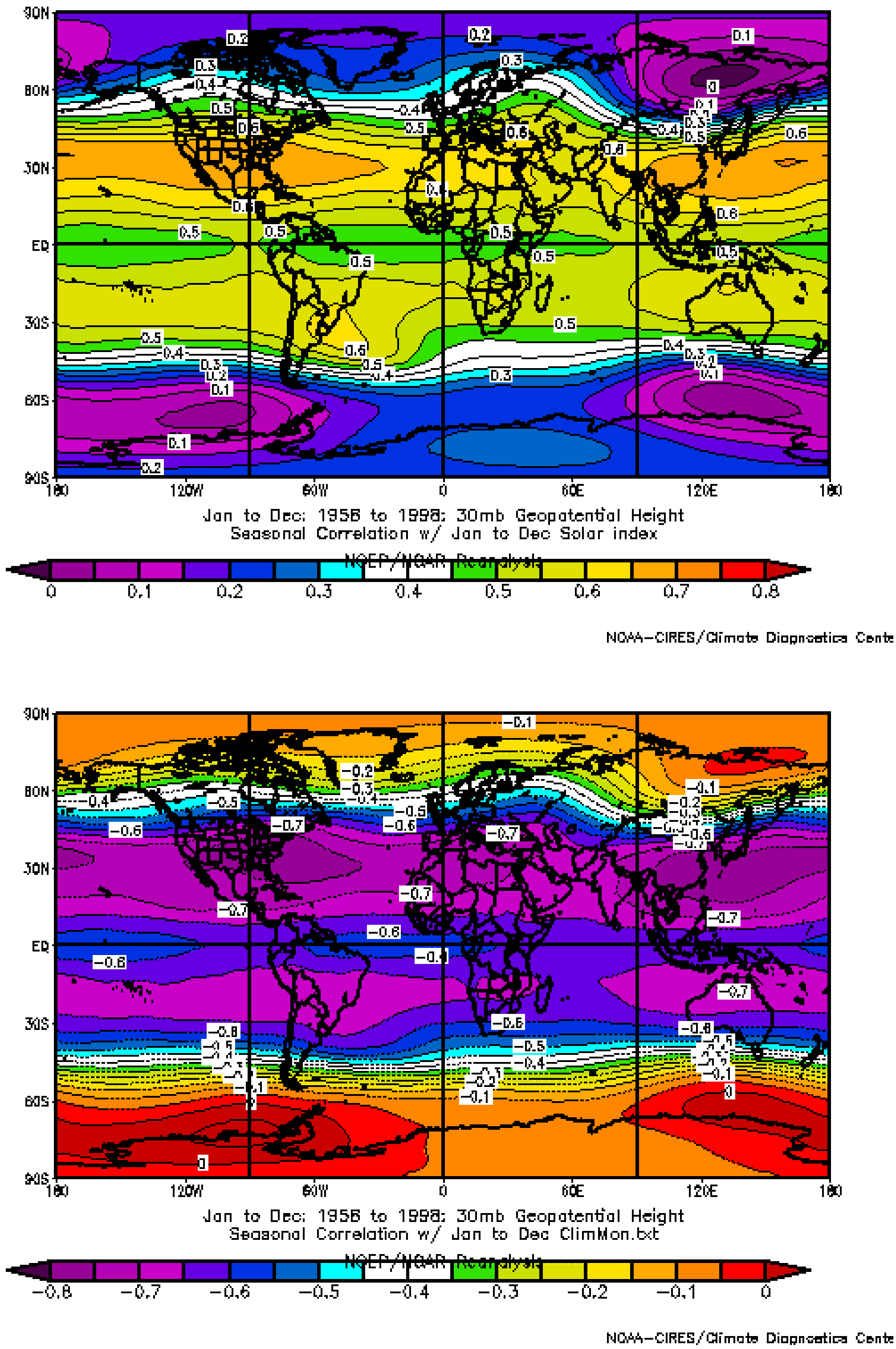


Figure 5.2 Spatial variation of correlation between 30mb geopotential height (using the NCEP/NCAR reanalysis at http://www.cdc.noaa.gov) and (a) (upper) 10.7cm solar flux and (b) (lower) surface neutron counter at Climax, from monthly averages in each case. (Note that in (a) the magnitude of the correlation runs from 0 to +0.8 and in (b) 0 to –0.8.) (From Aplin and Harrison, 2005).

(b) Solar cosmic ray effects

Solar cosmic rays associated with solar energetic particle (SEP) events, lasting a few hours show a sharp onset and cut-off, and are quite well-suited to searching for atmospheric responses. SEP events detected by the surface neutron network are described as ground level enhancements (GLEs) and are important as the solar particles are clearly penetrating the troposphere. The number of SEP events varies is modulated by the solar cycle, increasing at solar maximum, and they may also coincide with FDs. Over the solar cycles 19 to 23, the number of GLEs during each solar cycle varied between 10 and 15.

In a study of SEPs, Raspopov *et al* (1998) showed an increase in aerosol concentration of 1.4-1.8 times at 12-18km altitudes, with associated changes in atmospheric transparency of 5-6%. Murata and Muraki (2002) found that SEPs do not play a major role in producing the year-to-year variability in nitrate abundance in the polar region over the 11year solar cycle. Veretenenko and Thejll (2004) studied solar proton effects on characteristics of the lower atmosphere over the north atlantic. They found pressure and temperature decreases at high-latitude stations in the cold (October-March) half of year with increases in tropospheric relative vorticity, with the greatest effects near south-east Greenland and Iceland. Analysis of synoptic charts associated the intensification with well-developed cold cyclones in the same region. They suggested that radiative forcing caused by cosmic ray induced changes in cloudiness was influencing the cyclone evolution over the northern part of the atlantic.

(c) Other atmospheric changes associated with cosmic rays

Beyond the hypothesised tropospheric aerosol and cloud effects, other proposals have been made for cosmic ray effects on the atmosphere. There is, for example, the possibility of direct radiative effects of hydrated ion clusters in the troposphere, following their production by cosmic rays (Aplin, 2003; Aplin and McPheat, 2005). Laboratory work indicates absorption bands in the 9-12μm wavelengths, which could provide a route to atmospheric radiative forcing, or may influence satellite retrievals using these wavelengths. Vanhellemont *et al* (2002), suggested, on statistically arguments, that cosmic rays influence stratospheric aerosol concentrations with a time response of a few months. Merzlyakov (2000) found long-term similarities in global air temperature and cosmic rays, with a temperature response lagging by 1.8 years, on average. Beginning with the 1980s, an additional temperature rise by 0.015 K/yr, was claimed for the period from 1983 to 1995. Egorova *et al* (2000) reported a reduction of pressure and a warming following a FD, using data from Vostok. This was attributed to be a consequence of electrofreezing. Lam and Rodger (2002) were unable to reproduce the Egorova *et al* findings.

A further solar-terrestrial atmospheric connection has been suggested through thunderstorms and lightning, for which there is, in principle, much data available because of the routine meteorological reporting of thunderdays. These data have been investigated for possible solar influences in the past, and close statistical correlations between sunspots and thunderdays shown to be apparent in some regions (Brooks, 1934), including the UK (Stringfellow, 1974). More modern statistical data and the earlier data (figure 5.3) are presented in Schlegel *et al* (2001), using lightning data detection system for middle Europe. In Germany, a significant inverse correlation was found between lightning incidence and cosmic rays, for 1992-2000.

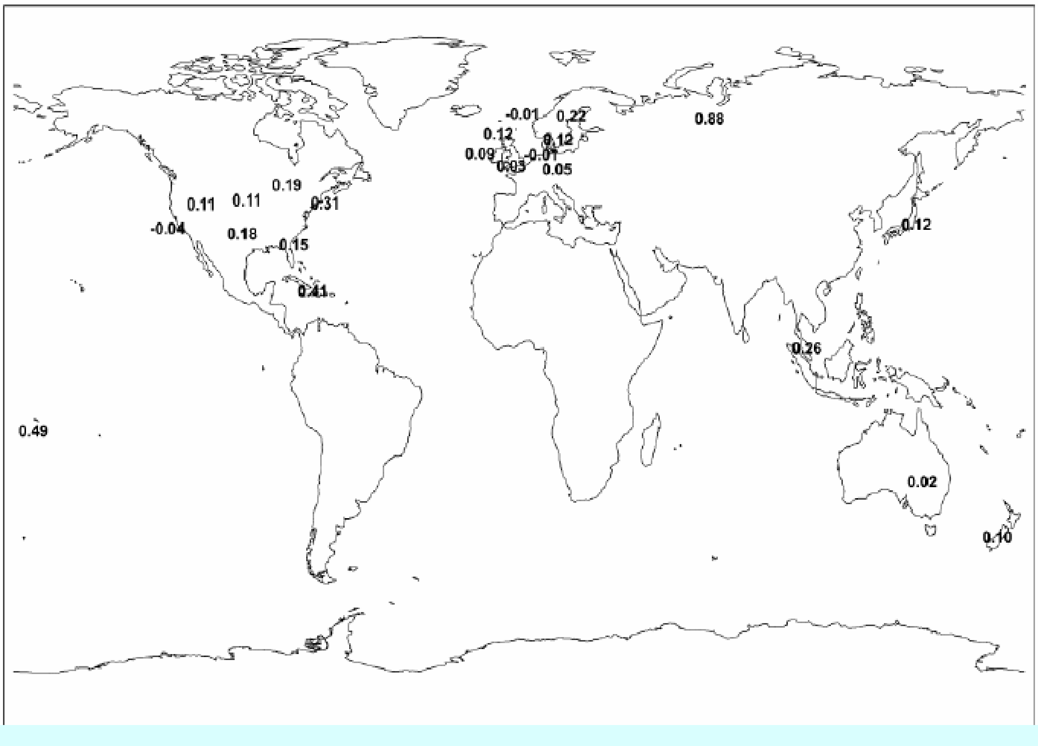


Figure 5.3 Regional distribution of correlation between annual totals of thunderdays and sunspots found by Brooks (1934) and presented by Schlegel et al (2001).

Cosmic ray proxies permit investigations of effects in the distant past. Estimates of cosmic ray variation during the Phanerozoic (the past 500 million years) have been found from meteorite data (Shaviv and Veizer, 2003), which show a correlation with proxies for temperature. However the methods used to obtain agreement between the two time series have been questioned (*e.g.* EOS, 27th Jan 2004[1]), both because (i) the CO_2 and temperature paeleo-data is poor, and (ii) the GCR reconstruction from 50 meteorites could be interpreted differently.

5.3 Danish findings on cosmic rays and clouds

Since HS99, which was primarily concerned with the publication of Svensmark and Friis-Christensen (1997 – SFC97), there have been substantial developments from the early Danish results. Marsh and Svensmark (2000) found that the correlation was present in the low cloud data (below 3.2 km, or at pressures above 680 hPa), as identified by the ISCCP analyses, with a high statistical significance. The criticism made in HS99 Section 4, about use of a composite of satellites, is no longer valid, as the ISCCP-D2 data used is a single dataset, running from July 1983 to September 1994. Marsh and Svensmark (2003) suggest that the correlation is due to a change in transparency associated with low liquid clouds (< 3.2 km). Because GCR are responsible for the majority of atmospheric ionisation below 35 km, ion-induced particle formation could produce aerosol (with diameters 0.001-1.0 μm) able to act as cloud condensation nuclei (CCN). Marsh and Sevenmark consider that cosmic ray related CCN changes would influence the cloud's radiative properties, and that this has been detected in the ISCCP analyses. The physics of the formation of aerosol from ions is central to any interpretation of Marsh and Svensmark results in terms of a cosmic ray cloud modulation, which is discussed further in section 5.4.

[1] Role of cosmic rays in climate change refuted *Bull Am Met Soc*., 85 (4): 490-491

More recently, the available ISCCP-D2 data has been increased, extending the time series over which the correlation can be calculated. Marsh and Svensmark (2003) analysed the ISCCP-D2 data extended to September 2001 and found that the correlation was weakened. They argued, however, that there was an inter-calibration problem with the ISCCP cloud data between September 1994 and January 1995, in the absence of a polar satellite, and suggested a correction (figure 5.4).

Whether this correction is appropriate is important. HS99 pointed out that there were calibration issues with the ISCCP data and therefore that it should only be used for short periods for correlation studies. This comment was influenced by the work of Kernthaler *et al* (1999) and (Brest *et al*, 1997), following calibration difficulties with the ISCCP data through the use of three different polar orbiters. Transitions were identified in earlier ISCCP-C2 data at the change from one satellite sensor to another (Klein and Hartmann, 1993). This was the reason for the short period of comparison from 1985 to 1988 used by Kernthaler *et al* (1999), during which there was consistent calibration from the use of a single polar-orbiting satellite. Since then, Robinson (2004) has argued that detailed analysis of ISCCP thermal infra-red data reveals sensor noise, as well as navigation and calibration errors. Systematic brightness temperature differences between early geostationary and polar orbiting satellites of 2 to 5K have been found, which have implications for the accuracy of ISCCP D2 data products. Robinson (2004) points out that the loss of the NOAA-11 reference polar orbital satellite, may explain the ISCCP-D2 anomaly (September 1994- January 1995), highlighted by Marsh and Svensmark (2003).

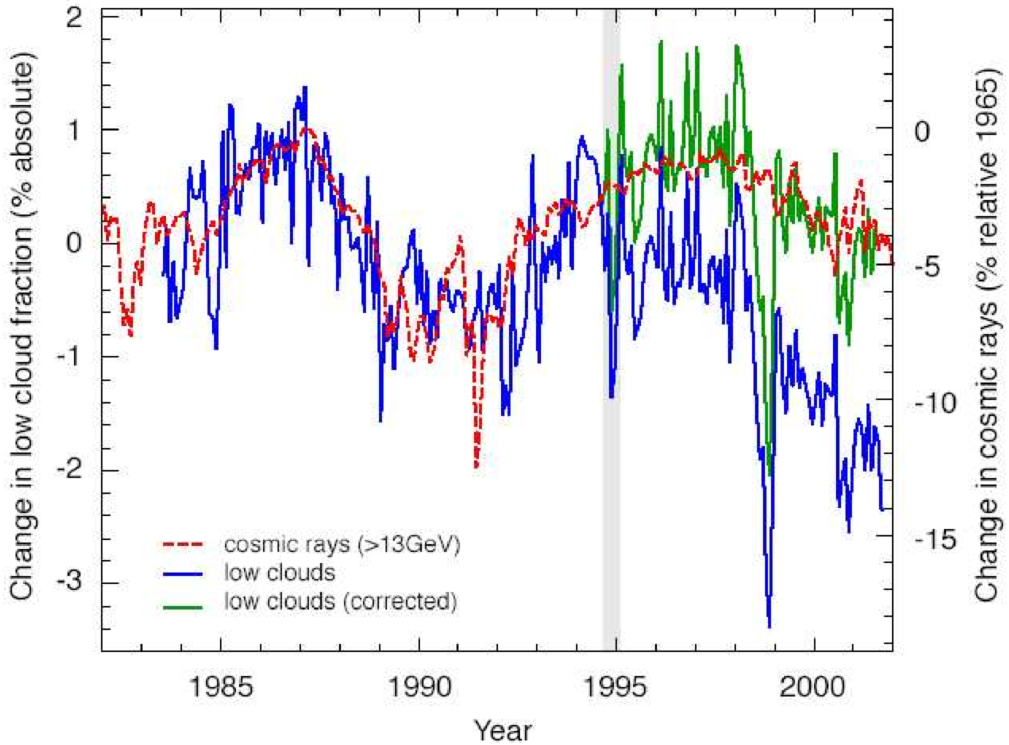


Figure 5.4. Variation of low cloud cover (ISCCP-D2 data) and cosmic rays between 1984 and 2002. The green curve shows data obtained by applying satellite calibrations. (Redrawn from Marsh and Svensmark, 2003).

In recent papers currently in press, and presented at conferences during 2004, Usoskin *et al* (2004b) have compared the ISCCP-D2 cloud retrievals with a global model of cosmic ray ion production (Usoskin *et al*, 2004a). The global model of ion production uses neutron monitor data with the geomagnetic field variations to find the height-latitude-time dependence of ion production. It uses the ion-balance equation, but rather than considering the ion-removal to aerosol explicitly, uses an effective recombination coefficient to represent the ion losses. Whether this is an effective approach to include the effects of aerosol, which is a major factor is determining atmospheric ion concentrations, is controversial, and depends on the form of the ion balance equation used (Bazilevskaya *et al*, 2000). Few atmospheric measurements of ion profiles exist with which to validate the model. Using a different approach determining the ion-production rate, which unlike ion concentrations is not affected by pre-existing aerosol, Palle *et al* (2004) also find a latitude effect in the correlation.

Figure 5.5 shows the global distribution of the correlation of cosmic ray-induced ionisation with low cloud found by Usoskin *et al* (2004b). Many of the regions of significant correlation are in ocean regions. Usoskin *et al* (2004b) show that the geographic latitude variation of the correlation in the data shows a minimum around the equator (figure 5.6). This is consistent with geomagnetic effects on cosmic rays, giving more ready penetration of lower energy cosmic rays at the poles, and less at the equator. Usoskin *et al* (2004b) argue that this finding strongly supports a GCR effect on clouds.

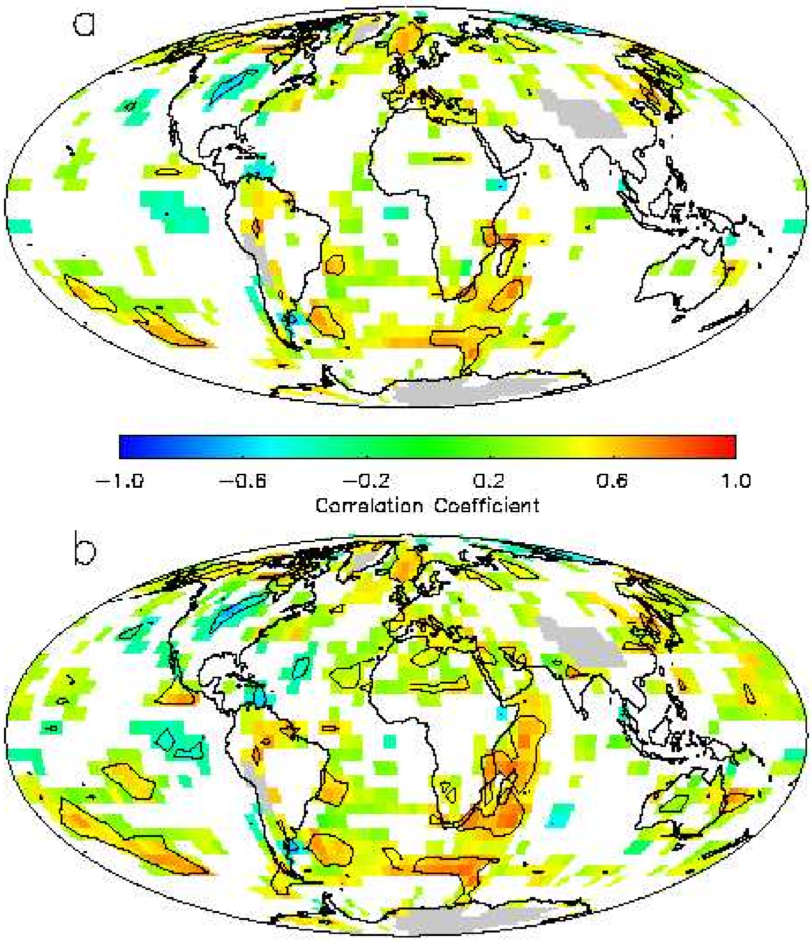


Figure 5.5. Global distribution of the correlation between cosmic ray ionisation and low cloud amount for 1984-2000, using (a) raw and (b) de-trended cloud data. Areas with significant correlations (>90%) are shown using contour lines (from Usoskin et al, 2004b).

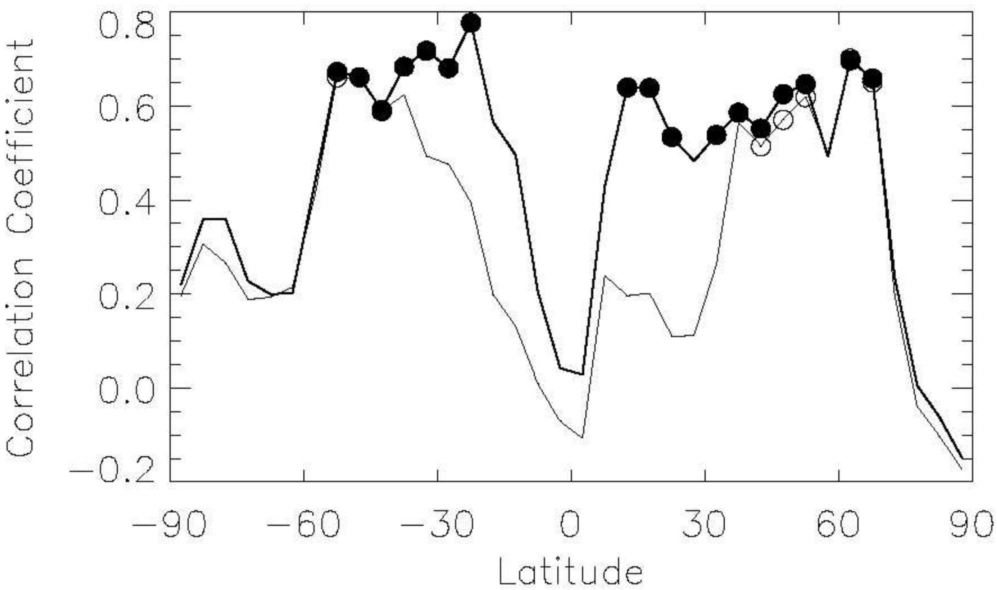


Figure 5.6 Latitudinal dependence of the cross-correlation coefficient between ISCCP low cloud amount and cosmic ray induced ionisation, for 1984-2000. The thin and thick lines show raw and detrended data respectively, and the symbols statistically significant (90%) correlations. (from Usoskin et al, 2004b).

5.4 Discussion of the Svensmark results

Work criticising and expanding on the Danish results has usually considered limitations of their analysis, or suggested possible physical deficiencies of their interpretation. Other than briefly mentioning the paper of Laut (2003) first, a chronological survey of the papers published is given here, with any closely-related subsequent work if it exists. Laut (2003) points out very specific problems with the data processing in many of the SFC97 graphs at the centre of the discussions about cosmic rays and cloud cover, but emphasised that they did not rule out the existence of important links between solar activity and terrestrial climate. (Some of the presentational aspects had also been mentioned in HS99.)

Farrar (2000) showed that ISCCP-C2 data for July 1986-June 1991 had global and regional cloud variations which could be associated with the El Nino of 1986-1987, and were particularly apparent in the Pacific. (These results agreed well with contemporaneous satellite observations of broadband shortwave infrared cloud forcing measured by the ERBE.) It was argued that the patterns of cloud variation were expected for the atmospheric circulation changes typically associated with ENSO, rather than GCR. Subsequently Marsh and Svensmark (2003) analysed the low-cloud properties (ISCCP-D2 data) for July 1983 to August 1994. They also found statistical relationships with both GCR and ENSO, but showed that the GCR relationship explained a greater percentage of the total variance. Because low cloud data can be contaminated by high cloud, they argued that the good correlation with ENSO was due to the lower likelihood of high cloud contamination in the ENSO regions.

The work of Kristjannsson and Kristiansen (2000) used ISCCP and ERBE data but did not support the coupling between cosmic rays, total cloudiness, and radiative forcing of climate. An exception to this was in low marine clouds at midlatitudes, but it was argued that there was no known physical mechanism for a cosmic ray effect in such clouds. They pointed out the difficulties in such studies from instrument calibrations and satellite changes. In an attempt to circumvent these issues, a 43-year series of marine synoptic observations was compared with cosmic rays: a weak negative correlation found. (They also pointed out that the net radiative effect 1985-1989 of clouds was cooling, despite a reduction in both total and low cloud cover, which contradicts the simple cosmic-ray-cloud-climate radiative relationship of SFC97.) In subsequent work, Kristjansson *et al* (2002) compared the ISCCP data with both TSI and GCR data and found that TSI actually correlated better and more consistently with low cloud cover than did GCR. The correlations were lower with daytime multi-channel retrievals than IR-channel retrievals, but it has to be said that the statistical significance of the results is marginal because of large autocorrelations. The correlations found were explained as amplification of the TSI by SSTs, affecting cloud cover. Negative correlations between SSTs and low cloud cover were shown to exist in subtropical regions.

Udelhofen and Cess (2001) found a statistically significant (0.7, 95%) cloud cover signal with 11 year periodicity, using sunspot number (SSN) as the measure of solar variability for correlation with surface cloud observations over the US. There was a *positive* correlation with SSN for the majority of the area, although some small regions of negative correlation existed (figure 5.7). They attributed the observed cloud variability to a solar-uv-ozone-induced circulation modulation.

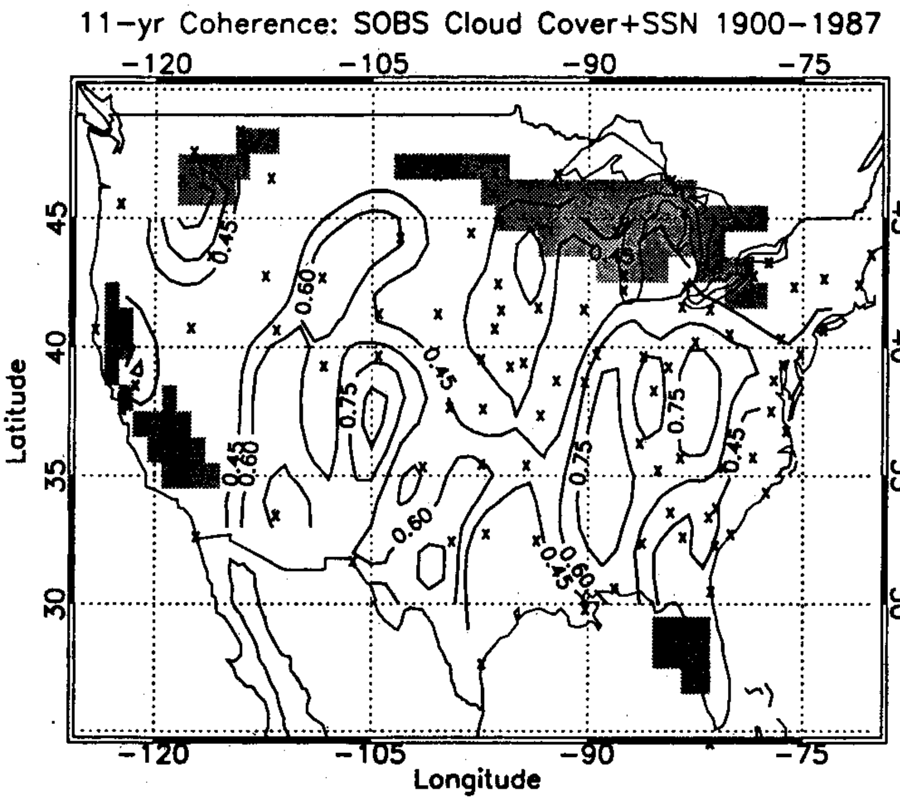


Figure 5.7. Spatial variation of the 11 year coherence between Sunspot Number (SSN – inversely correlated with cosmic rays) and surface cloud observations. In shaded areas cloud cover variations are in phase with cosmic ray variations (from Udelhofen and Cess, 2001).

Wagner *et al* (2001) emphasised that there was unlikely to have been a cosmic ray modulation of cloud cover in the recent geological past (20-60 kyr B.P.) because of the absence of a relationship between the cosmic ray proxies Be^{10} and Cl^{36} at Summit, Greenland and the regional climate proxies. At around this time (the Laschamp event), geomagnetic changes led to at least a doubling of Be^{10} in Greenland ice core. In the same core, changes in climate proxy parameters, (δO^{18}, methane, dissolved ions) were not at all exceptional. Although the absence of dramatic combined effects in the same core is an interesting result, changes in climate proxies at other locations have been found at about this time.

Palle and Butler (2001) compared a long series of historical sunshine records with data on solar variability. They found a decrease in sunshine hours for four stations in Ireland operating since the late 19^{th} century. They argued that this could have been related to an increase in cloud through enhanced evaporation rates over the Atlantic from rising sea surface temperatures. In later work, Palle and Butler (2002) sought 20^{th} century low cloud decreases expected from the cosmic ray decrease associated from rising solar activity, again using sunshine recorder and synoptic cloud data. Both data sources supported an increase in global total cloud cover during the past century, which they used to argue that GCR cloud effects had not dominated on centennial time scales. It was, however, pointed out that the data used could not discriminate between the alternative possibilities of an actual absence of a low cloud decrease, or a low cloud decrease swamped in surface effects by an increase in medium level or high cloud.

Sun and Bradley (2002) were unable to support a solid relationship between cosmic ray flux and low cloud amount. Their analyses used both long-term surface-based cloud data from national weather services and ISCCP-D2 data for 1983-1993. No meaningful relationship was found between cosmic ray intensity and cloud cover over tropical and extra-tropical land areas back to the 1950s using the synoptic data, but there was, however, a high cosmic ray-cloud cover correlation in the period 1983-1991 over the Atlantic Ocean. Over this region the correlation was statistically significant, but it was reduced when data set was extended to 1993. In additions, the surface observations of cloud cover data from North Atlantic ship observations did not show any relationship with solar activity 1953-1995. There was, however, a large discrepancy between ISCCP-D2 data and the surface marine observations.

Balling and Cerveny (2003) pointed out that data analysis work was hampered by the relatively short time period with accurate cloud and cosmic ray flux records. They assembled surface and radiosonde data for the United States over the period 1957-1996 along with sunspot records as a (inverse) proxy for cosmic ray flux. They found that periods with low sunspot number (times with high cosmic ray flux) were associated with significantly higher dew point depressions, a higher diurnal temperature range, and less cloud cover. This does not support suggestions of increased cloud cover during periods of high cosmic ray flux.

Marsden and Lingenfelter (2003) used a model to predict CCN formation rate from GCR and then searched ISCCP data (1983-99) for common variations in the yearly mean visible cloud opacity and visible cloud amount due to cosmic rays. After separating out temporal variations in the data due to the Mount Pinatubo eruption and El Nino-Southern Oscillation, systematic variations in opacity and cloud amount due to cosmic rays were identified, and they concluded that “ionisation plays a crucial role in tropospheric cloud formation”. The effects found were on low cloud. Fractional changes in visible cloud amount were only positively correlated for low clouds and become negative or zero for the higher clouds. The opacity trends found suggested ion-aerosol physics opposite to the current predictions of ion-mediated nucleation (IMN) models.

5.5 Mechanisms for cosmic ray effects

In this section the mechanisms linking ions to clouds proposed in Carslaw *et al* (2002) are discussed.

(a) Clear air mechanism

As identified in HS99, a variety of laboratory measurements show that, in the presence of suitable vapour concentrations, ultrafine particles can be formed from ions. A summary of the range of conversion rates in different experiments is given in Figure 5.8. Detailed modelling of ion-aerosol microphysics was undertaken to investigate work on the properties of aircraft exhaust (Yu *et al*, 1998). It was found that the principal factor controlling the population of ultrafine plume particles was the number of charged ionic species (or *chemi-ions*), emitted by the aircraft exhaust. This work led on to studies of particle formation in the background atmosphere (Yu and Turco, 2000): an important result was finding that charged molecular clusters can grow significantly faster than neutral clusters.

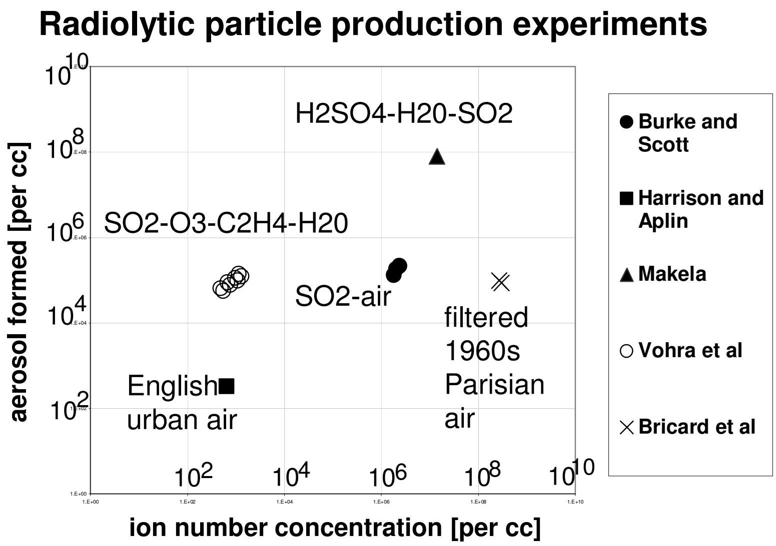


Figure 5.8. Radiolytic ultrafine aerosol particle production (in number of particles per cm^{-3}) as a function of ion concentration, for different gas compositions (from Harrison (2002).

Using a box model, Yu and Turco (2001) explained the diurnal variation in the surface ion mobility spectrum reported by Horrak *et al* (1998), in which intermediate-size ions (charged aerosol) were found after the presence of small ions had been observed hours earlier. The model simulates the evolution of ion clusters through to sub-micron particles, including the processes of condensational growth and coagulation among neutral and charged particles. Particular care was taken in describing the rate of coagulation, using coagulation kernels appropriate for charged and neutral particles from molecular cluster scale to micron sizes. The enhanced growth of ion clusters by uptake of condensable vapours was parameterised in terms of an increased accommodation coefficient, recognising that the thermodynamic data needed for calculating vapour condensation are not available. The simulations were restricted to uptake of H_2SO_4 and H_2O vapours, with the H_2SO_4 concentration prescribed to vary over a diurnal cycle, with a peak at midday. The model results emphasise that charged clusters have a *growth advantage over neutral clusters* and can therefore grow more rapidly into nanometer sized particles with sizes large enough to become stable particles. Charge may also act to stabilise small clusters, reducing evaporative loss. The ultrafine aerosol concentration is controlled by a competition between new particle formation and scavenging by the larger pre-existing background aerosol. The ion-induced particle formation is greatest when the nucleation rate is limited principally by the availability of ions, such as in and above the marine boundary layer.

Yu (2002) argued that increases in CR would lead to differential particle production, with increases in the lower troposphere but decreases in the upper troposphere. This is through the altitude-dependence of the sensitivity of particle production to the ionization rate. It depends on a combination of the ionisation rate, the precursor gas concentration and ambient conditions. The observed different correlations between cosmic ray variations and low, middle and high cloud anomalies are consistent with the variations found. The work was extended to suggest that a systematic change in global cloudiness could change the atmosphere heating profile, possibly providing sufficient external forcing to reconcile the observed surface and troposphere temperature trends.

Some atmospheric evidence has emerged which may be related to particle formation from ions. Nuclear weapons tests in the 1950s and early 1960s resulted in large-scale changes in atmospheric ionisation. The largest weapons were detonated in autumn 1962. At Lerwick, where a surface radioactive contamination occurred, there was an increase in overcast days 1962/3 at the site, found from the routine radiative measurements (Harrison, 2002). A similar anomaly in short wave diffuse and direct measurements was reported in data from the Hohenpeißenberg Observatory in Germany by Winkler *et al* (1998). (In both cases, the anomalies occurred before the eruption of Mt Agung in March 1963.) If either of these changes were caused by the extra radioactivity in the atmosphere, it would arguably have been more likely to be caused by radioactive ionisation, rather than the supply of particles from the explosions. This is because the later tests were detonated well above the surface, to reduce the radioactive fallout generated. For the large (megatonne) nuclear explosions occurring at the time, the radioactivity reached the stratosphere, which provided a steady supply of ionisation to the troposphere.

More direct atmospheric evidence can be obtained in experiments in which the ion production rate, ion concentrations and aerosol concentrations are measured simultaneously. Existing aerosol instrumentation has a low diameter cut-off at about 3nm, below which ion measurements can be used to determine the charged fraction present; the ion production rate can be found using instrumentation to detect the contributions of radon, gamma and cosmic rays. In one of the first experiments combining instrumentation to investigate the different terms in the ion balance equation, modern ion mobility instrumentation (Aplin and Harrison 2001) was deployed with a condensation nuclei counter and Geiger counters in urban air (Harrison and Aplin, 2001). Increases in CN were found which coincided with increases in the ion production rate. In recently-reported work, Laakso *et al.* (2004) also describe an integrated suite of measurements, but in a boreal forest. They found preferential condensation of sulphuric acid onto negative cluster ions. At the surface, the ion production is not dominated by cosmic rays, although these results support the formation of ultrafine particles from ions produced elsewhere in the atmosphere from cosmic rays. Above the continental boundary layer, cosmic rays dominate ion production, hence an important finding was the discovery in the free troposphere of large positive and negative gaseous ions (Eichkorn *et al* 2002; Wilhelm *et al*, 2004). These large ions were interpreted as precursors of new aerosol particles formed by ion-induced nucleation. From the ion data obtained, the total concentrations of condensable gases inferred were consistent with those expected for gaseous H_2SO_4 in the troposphere.

In summary, model results show that ultrafine particle formation can be expected in the atmosphere where there is sufficient condensable vapour: experimental results at the surface and in the free troposphere strongly support the expectation of ion growth by condensation of sulphuric acid.

(b) Near-cloud mechanism

Adjacent to a horizontal layer of cloud or aerosol, there are electric field changes observed which arise from the vertical conduction current flowing in the global atmospheric electrical circuit. These changes cause the upper part of a thin stratiform cloud to become more positively charged than the clear air above it, with a gradual return to quiescent fair weather away from the cloud (figure 5.9a). This occurs because of the decrease in conductivity associated with the layer of particles or droplets, which remove ions. To maintain a constant conduction current, the vertical electric field will increase in the low conductivity region. Perturbations in aerosol charge, although of smaller magnitude, are found around layers of aerosol at the top of the polluted boundary layer. Steady-state droplet charges at cloud boundaries can be large, as the combination of unipolar charges and low conductivity environment around the cloud prevents the neutralisation of such droplets. Aerosol particles in this region are also relatively highly charged. Although electrical effects are very much weaker in stratiform clouds than in thunderstorms, they are certainly present and the local electric fields and charge densities are modulated by cosmic rays and global circuit changes.

Following the suggestion of Tinsley and Heelis (1993) that electrification could enhance the effectiveness of aerosol as ice-forming nuclei ("electrofreezing"), the microphysical consequences of particle charging in the near-cloud region has received attention. There are some difficulties with establishing what level of aerosol charging can be regarded as typical. Observed charge densities (Clark, 1958) are typically ~100*e*.cm^{-3}, rather less than the 10^3-10^4*e*.cm^3 estimated by Tinsley (2000) on evaporative residues. Tripathi and Harrison (2002), calculated that, for 0.5μm radius aerosol charging in typical atmospheric ion concentration asymmetries, there would be some particles (> 1 per thousand) carrying 10*e* charges, but a negligible number carrying 20*e* charges. (At temperatures warmer than –20degC, the fraction of all micron-size aerosol particles able to act as Ice Nuclei is 10^{-4} or less.)

New measurements of charge densities using radiosondes (Harrison, 2001), following the recommendations of HS99, have been made recently (Harrison, 2004a; Harrison 2005a). In measurements over Reading, fluctuations in space charge at aerosol boundaries in fair weather conditions occur over relative small distances, typically ~100m, and in the range 10 to 100 *e*.cm^{-3} (figure 5.9b).

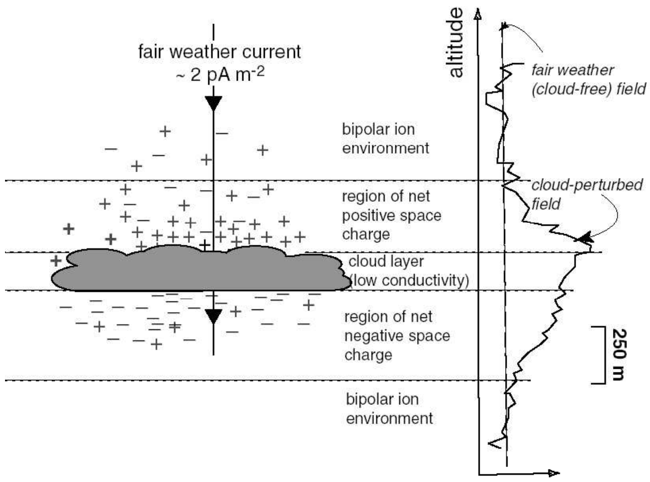


Figure 5.9 Schematic of the charge structure around a stratiform layer of cloud, deduced from a measured electric field profile (from Harrison and Carslaw, 2003).

Since HS99, the suggestion that charging enhances the effectiveness of aerosol for ice nucleation has been explored using theory to describe the enhancement of aerosol collection. Electrically-assisted aerosol scavenging (or electro-scavenging) has been shown to have substantial effects on the collection of charged aerosol (Tripathi and Harrison, 2001). This effect increases as the aerosol radius decrease below 1μm (Tinsley *et al*, 2001; Tripathi and Harrison 2002). Because of the electrical image force dominating close to the droplet (Tinsley *et al*, 2000), this effect is independent of sign of the aerosol charge: there is also little effect of droplet charge. Quite apart from any direct effect of charging on freezing, the electrical enhancement of particle collection rate is likely to increase the number of contact ice nuclei encountered by a supercooled drop. Any direct effect of ionisation on freezing of supercooled water droplets now seems unlikely following the work of Seeley *et al* (2001). They found no direct effect of ionisation on freezing of a drop, by cooling it in the presence and absence of an alpha source. In both cases the mean nucleation temperatures was -34.5 deg C, consistent with previous studies of homogeneous nucleation.

There are many uncertain steps between the electroscavenging and cyclone intensification in the near-cloud mechanism, as emphasised in HS99, so progress has been made either by looking at specific microphysics (e.g. scavenging theory), or from statistical studies linking cosmic rays with cloud-related parameters, such as precipitation. Kniveton and Todd (2001) show that variability in cosmic ray flux is a possible explanation of the observed inter-annual variability in precipitation and precipitation efficiency in the southern hemisphere mid and high latitudes. The absence of a meridional dipole in the correlation structure and the peak in precipitation at cosmic ray maximum is used to support a cosmic ray–related mechanism such as electrofreezing, rather than through irradiance changes. Todd and Kniveton (2001) investigated short-term changes in cloud cover during Forbush decreases (FD) in GCR. They did a superposed epoch study using the ISCCP D1 data, and found a small but significant (at 0.001) reduction of up to 1.4% in the global cloud cover immediately before and after FD events. Analysis using geomagnetic latitude showed significant cloud anomalies at high latitudes and over Antarctica, the changes were as much as -30% at 10-180 mbar (figure 5.10). They found no effect for FDs which occurred simultaneously with SPEs. In later work, Todd and Kniveton (2004) again found the largest cloud reductions over Antarctica, for FDs without SPEs, but found that these were inconsistent with NCEP re-analysis of surface and tropospheric temperatures. Todd and Kniveton (2004) point out ICCSP data quality is poor over the Antarctic plateau.

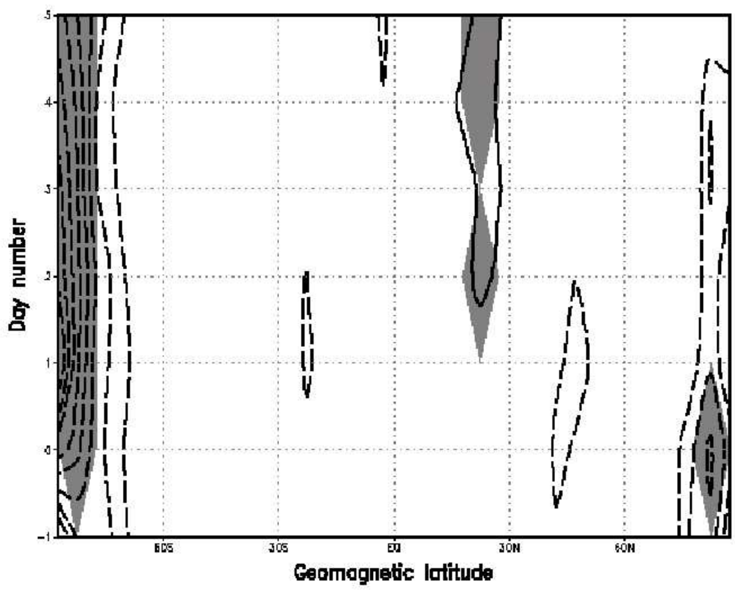


Figure 5.10. Zonal mean (averaged over 5deg geomagnetic latitude bands) total cloud proportion anomalies (relative to base period) for days –1 to 5 following a Forbush decrease, positive anomalies dotted and significant (0.05 level) anomalies shaded (from Todd and Kniveton, 2001).

Another aspect of the near-cloud mechanism is that, through aerosol charging, it couples global atmospheric electrical circuit changes into aerosol-cloud microphysics at long distances away from the source disturbance or through the troposphere down to the surface. Effects on clouds have also been postulated by Stozhkov (2003). Short-term solar changes may be communicated through the global atmospheric electrical system: Farell and Desch (2002) have shown theoretically that solar proton events can modify the fair weather field at the surface. Consequently information on the global circuit is needed, and, more specifically the vertical air-earth current which passes through cloud layers and down to the surface. The global circuit provides a teleconnection network from tropical regions, through the ionosphere, and down through the stratosphere and troposphere.

Surface atmospheric electrical measurements (*e.g.* of the vertical Potential Gradient, PG) provide one method of investigating the global circuit, although the method of choice would be ionospheric potential measurements obtained from soundings, which is a global quantity. Surface measurements were regularly made in the UK at Meteorological stations until the early 1980s, but none are made now routinely (Harrison, 2003). A difficulty with surface PG measurements is the complicating effects of weather and pollution, but they have been found to vary with the ionospheric potential on hourly and annual timescales and, in addition, long-range variations have been found (Harrison, 2004c), which represent an atmospheric electric teleconnection (see figure 5.11). Compiling these measurements from the UK sites of Eskdalemuir for 1911-1981 (Harrison 2004b) and Kew 1909-1979 (Harrison and Ingram, 2004) is leading to an archive of past changes in the global circuit. A common feature in several sets of measurements is a twentieth century decline (Marcz and Harrison, 2003), which may be related to the cosmic ray decrease.

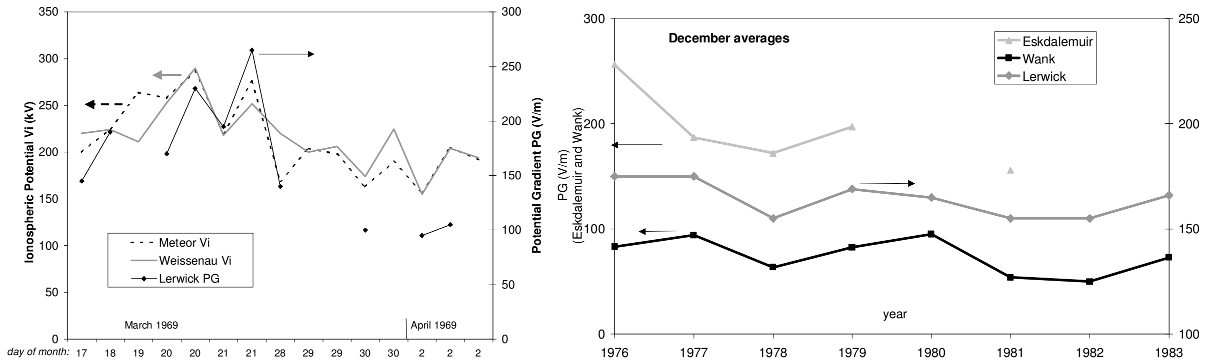


Figure 5.11. (a) (left) Ionospheric potential measured by Mühleisen and Fischer using synchronous balloon soundings on the Meteor (Atlantic Ocean) and at Weissenau (S.Germany) (data from Budyko, 1970). The PG at Lerwick, when available for synchronous hours with no hydrometeors, is also shown (from Harrison 2005b).) (b) Comparison of monthly average PG for December at Eskdalemuir, Lerwick and Mount Wank (Bavaria), for days selected for a correlation with the Carnegie curve greater than 0.8. (from Harrison 2004b).

5. 6 Discussion

Since SFC97 there have been several key developments:

- The cosmic ray cloud correlations has been identified in the low cloud ISCCP data, at least up until September 1994.
- Theory and experimental evidence supports ion-induced ultrafine particle formation, with the possibility that CCN may be produced
- Theory supports the enhanced scavenging of electrically-charged aerosol over neutral aerosol

Taken together however, it is not possible to say that variations in clouds are linked to cosmic rays or even that the cosmic ray-cloud correlation has ion-induced particle formation as its physical cause. The microphysics indicates that cosmic ray produced ions can influence atmospheric aerosol processes, but our ability to estimate magnitudes of effects is poor. In the absence of a model linking ion, aerosols and clouds, it is not possible to be sure if the observed satellite effect is quantitatively what would be expected. Only when the model predictions are matched by the experimental findings will this stage be reached. Nothwithstanding that caveat, it is reasonable to expect that global averages from stable and accurate satellite instruments could provide data where small cloud effects would be detected.

A very specific problem exists in the later ISCCP-D2 data: a calibration step. Whilst it was pointed out in HS99 that this could be a difficulty in extracting small signals close to the limits of measurements from the satellite system, it is now central to the continuation of the ISCCP low cloud-cosmic ray correlation. This requires further independent analysis.

There remain many uncertain aspects of the ion-aerosol physics, such as the effect on particle coagulation, and the consequences for clouds have not been directly investigated. The indications are that ion-induced effects on clouds will not be spatially uniform, as the availability of vapour and the presence of aerosol as vapour sinks are important in the determining nucleation rates. The nucleation effect requires a combination of physical constraints on ion production, aerosol concentration and vapour availability to be met, before any effect on cloud: it is therefore not considered likely that it will be ubiquitous. The suggestion of the theory is that marine clouds will be affected, and to some extent, the satellite data supports this. Independent sources of data, such as from synoptic records, would provide evidence that the satellite correlation is not an artefact of the observing system: such evidence as has been presented indicates discrepancies between marine data and satellite cloud retrieval.

The global circuit links the upper atmosphere to the surface, as there is no doubt that changes in the electrical potential of the ionosphere are communicated directly by the air-earth current flowing through cloud layers to the surface. The physical consequences of this coupling for climate are less clear. Electrically-influenced aerosol microphysics on the boundaries of clouds may be important, such as coagulation and scavenging, but this area is more complicated, and currently less developed, than the particle nucleation topic. Figure 5.12 summarise the range of processes that can result from solar modulation of cosmic rays, with an indication of the level of certainty. Section 6 includes a summary of each of the cosmic ray physics and ion-aerosol physics, with an opinion on the possible consequences for climate.

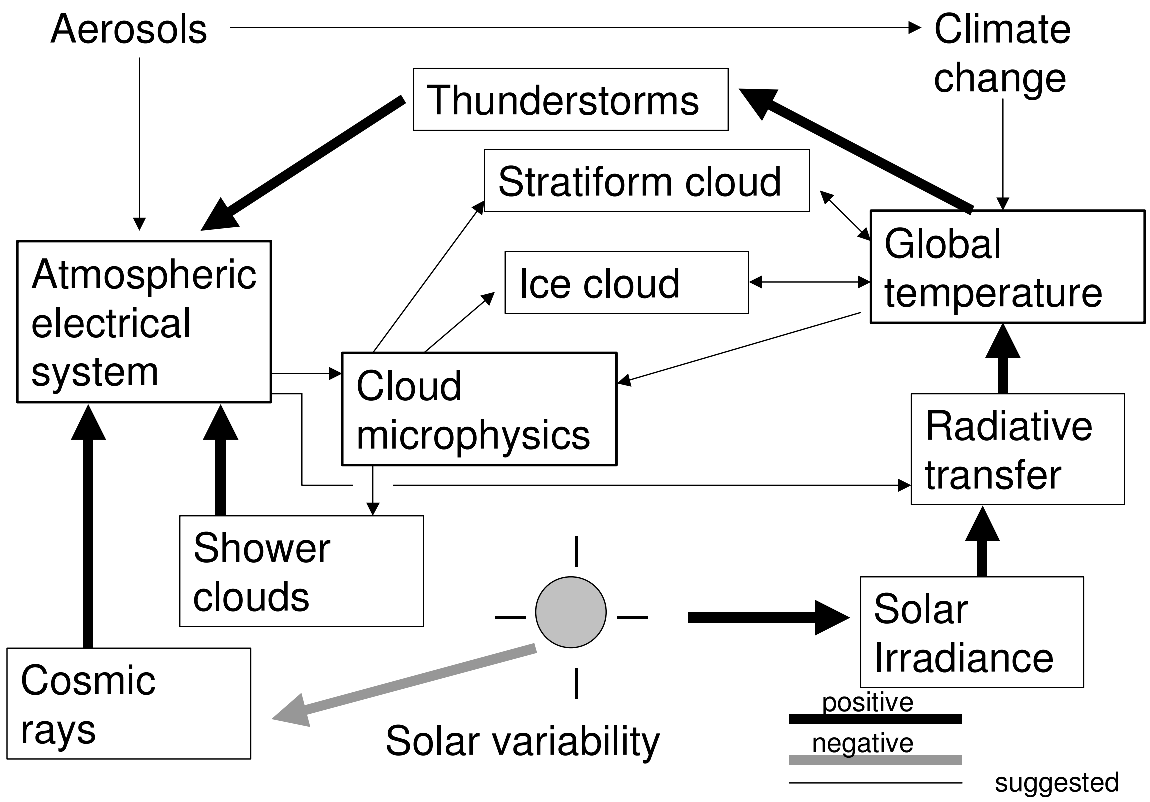


Figure 5.12. Summary of the processes linking the global atmospheric electrical circuit with global climate. Thick lines indicate established links, thin lines indicate suggested links. (from Harrison 2005b).

6. Summary and assessment of mechanisms

Aspect	**Comments**
DIRECT TSI CHANGES	
Spatially inhomogeneous heating of the surface, due to variations in cloud cover, produces horizontal temperature gradients which feedback on cloudiness through changes in circulation.	Contentious. One GCM experiment purports to show this.
DIRECT UV EFFECTS	
Irradiance changes are not uniformly distributed across the EM spectrum. Changes in UV (of a few percent) will lead to a response in stratospheric ozone and hence temperature	Well established. Temp and ozone signals confirmed in in-situ and satellite data analyses. However, disagreement in details of the observed signals needs resolving.
INDIRECT UV EFFFECT THROUGH MIDDLE ATMOSPHERE DYNAMICS	
Both direct TSI changes and indirect UV effects result in temperatures changes in equatorial upper stratosphere and hence corresponding stratospheric zonal wind anomalies. This influences winter hemisphere planetary wave propagation.	Not contentious. Recent analyses have confirmed anomalies in zonal wind and EP-fluxes in both winter hemispheres.
Changes in stratospheric planetary wave propagation in winter influence frequency of stratospheric sudden warmings and thus transfer solar signal from upper equatorial stratosphere to winter polar lower stratosphere.	Not contentious, although there is strong interaction with QBO influence which complicates extraction of solar signal in polar winter latitudes. Recent progress in understanding details of planetary wave sensitivity has enabled models to reproduce the observed solar cycle / QBO interaction.
Changes in planetary wave activity associated with changes in sudden warming frequency also cause changes in global meridional circulation strength. This is a mechanism for solar cycle influence to reach equatorial lower stratosphere and summer hemisphere.	Probable. Changes in lower stratospheric temperature, ozone and zonal winds that are consistent with this hypothesis are observed in the data but models are unable to reproduce the amplitude of these observed changes.
Changes in stratospheric temperature and circulation penetrate downwards into the troposphere.	Possible. Some observational evidence, and models reproduce some features, but details of the mechanism unclear. Possibilities include: affecting propagation of stationary planetary waves; affecting growth of internally generated long waves; modifying baroclinic lifecycles.
THERMOSPHERIC TEMP CHANGES	
Irradiance changes are not uniformly distributed across the EM spectrum. Large changes in soft X-ray part of spectrum lead to a large temperature response in the thermosphere.	Well established
Thermospheric and upper mesospheric	Possible. Limited modelling studies suggest

temperature changes lead to changes in stratospheric planetary wave propagation.	enhancement of stratospheric temperature response to solar cycle, but precise mechanism needs clarifying and testing in GCMs.
Wave effects propagate downwards from upper to lower atmosphere.	Contentious. Modelling studies not statistically conclusive.
ATMOSPHERIC ION PRODUCTION	
Cosmic ray ionisation is ubiquitous throughout the atmosphere and provides a source of ions, modulated by cosmic rays.	Well established. Cosmic rays lead to ionisation and, above the continental boundary layer are the principal source of tropospheric small ions. The cosmic ionisation rate depends on latitude (increasing towards the poles) and increases with altitude.
Variations in the solar wind modulate the cosmic ray flux in antiphase with the sunspot number	Not contentious. Neutron fluxes, ion production rates and air conductivity have been observed to vary inversely with the solar cycle.
Cosmic ray and solar proton ionisation penetrate the troposphere at mid and high latitudes.	Not contentious. Cosmic rays are the major source of ions at the surface in marine air. Solar proton events reach the surface.
ION-AEROSOL-CLOUD PHYSICS	
Ions provide a source of atmospheric condensation nuclei (CN).	Not contentious. Laboratory and atmospheric measurements show ultrafine particles (CN) can be formed from ions. Variations in the trace condensable species and aerosol vapour removal control the efficiency of the conversion, which will lead to spatial changes in the process.
Ions provide a source of Cloud Condensation Nuclei (CCN).	Probable. Detailed microphysical models show that particles large enough to act as cloud condensation nuclei can grow from ion-induced ultrafine aerosol production on the timescale of hours. There will usually, however, be competing effects, so the fraction of new aerosol that ultimately grow to CCN sizes is uncertain, but is likely to be small.
Electrification of aerosol increases its effectiveness as Ice Nuclei (IN)	Possible. Limited laboratory evidence suggests that charging alters the ice nucleation efficiency of aerosol, but there is no atmospheric experimental evidence. Theory, however, shows electrically enhanced aerosol scavenging increases the acquisition of contact ice nuclei by supercooled droplets. Only a small fraction of atmospheric aerosol is able to act as IN, depending on temperature.
Cosmic ray ionisation variations lead to variations in aerosol and water droplet	Probable. Atmospheric aerosol undergoes charge exchange through ion-aerosol

electrification	collisions and charged atmospheric aerosol particles are observed. Variations in the magnitude of aerosol charge on layer boundaries through which the conduction current J_z passes are expected from theory and have been observed.
Aerosol electrification modifies coagulation and scavenging rates.	Not contentious. Theory indicates coagulation rates are modified by charging, but this depends on aerosol charge distributions. Charge distributions depend on ion asymmetry; charging timescales depend on ion concentrations. Theory also shows that aerosol scavenging rates are strongly sensitive to aerosol charge.
CCN changes modify cloud properties, cloud lifetimes and precipitation.	Not contentious. Reducing uncertainties in understanding the effects of aerosol changes on clouds (the *aerosol indirect effect*) is a major area of active climate research.
Cosmic rays significantly modulate liquid water clouds.	Possible. Quantitative evidence is lacking.
SOLAR VARIABILITY EFFECTS ON CLOUDS	
Solar modulation of global clouds, through production of CCN from cosmic ray ionisation, has been observed.	Contentious. A significant positive correlation between cosmic rays and cloud cover has been obtained from ISCCP satellite cloud data over 1983-1994. If a correction is accepted to the satellite data, the correlation continues to 2001. Different suggestions have been made to explain spatial pattern of the correlation, some of which do not require CCN production from cosmic rays.
Solar-induced changes in ionisation will influence freezing in *all* clouds.	Unlikely. The variability in atmospheric clouds and aerosol is such that it cannot be predicted which clouds will be affected by electrofreezing. It seems likely that only a subset will be affected
A solar influence on electrofreezing intensifies cyclone development and modifies storm tracks, through the changes in cloud temperature and latent heat release.	Contentious. Modification of cyclones is highly selective, depending on where and at what stage in the cyclone's development the latent heat is released. The magnitude, and even sign, of any effect must be regarded as very uncertain

7. Recommendations for future work

7.1 Identified areas requiring further research

Solar variability

Uncertainties remain in knowledge and understanding of the magnitude and variability of solar irradiance. Aspects requiring further work include:

- Continuous measurement from inter-calibrated satellite instruments with high absolute accuracy of solar total and spectral irradiance.
- Understanding of instrumental limitations and application to resolve the current disparity in TSI composites.
- Advances in understanding of the variation in the quiet sun, including the relationship between magnetic activity and radiance, and application to reconstructions of past TSI.

Atmospheric response to solar variability

Advances have been made in the detection of solar signals in the middle and lower atmospheres but further work is needed to establish firmly:

- The existence of a secondary maximum in the lower stratosphere (minimum in the middle stratosphere) in the signals in ozone and temperature.
- The existence of a response in the NAO.
- The existence of a response in tropical convection, the local Walker and Hadley cells and the monsoons.
- The response of cloud.

In addition, some details of the current detection analyses require further work:

- Differences between temperature response analyses from the SSU/MSU dataset need to be resolved. Additional years' data since 2001, including that from the AMSU instrument, need to be merged in a consistent and accurate way so that the solar response analysis can be updated.
- Differences between the solar response in ozone from the SAGE and SBUV instruments need to be understood. More accurate assessment is needed of the height profile of the ozone changes, since models suggest that the temperature and circulation responses are very sensitive to this.

Modelling the impacts of the 11-year solar cycle

- Further model studies are required to investigate the apparent under-estimation of the ozone response in the upper stratosphere. The possibility that energetic electron precipitation may account for this merits investigation.
- More work is needed to understand the coupling between the stratosphere and troposphere: does the tropospheric circulation respond through changes in the mean circulation brought about by changes in tropopause height, through modulation of stationary or transient planetary waves or through modifications to baroclinic lifecycles?
- As computing resources become available, improvements in experiment design are needed to more accurately model the time-evolving solar cycle and to include the ocean response.

Modelling the impacts of long-term (centennial) solar variations

This is discussed in more detail in section 7.2

High Energy Particle Mechanisms

Advances are being made by the studies of ion-aerosol-cloud processes, which should be expanded. Further work is needed to assess the importance of ion-induced particle formation. The following list is in order of priority:

Investigation of ionisation effects on aerosol formation in the background atmosphere. Experiments should combine measurements of ion production rates and aerosol nucleation rates with evolving aerosol size and charge distributions. Precursor trace vapour concentrations will also be required. Aircraft (or balloon-carried) experiments to investigate the aerosol formation rates from cosmic ray ionisation are necessary. These are required to constrain the fraction of nucleation mode aerosol formed by cosmic rays in the troposphere.

- Measurements of aerosol and charge properties are needed to characterise the complex and highly-variable aerosol electrical state around clouds. This is suited to an instrumented aircraft or balloon-borne instrumentation, with an emphasis on high spatial and temporal resolution measurements.
- Theoretical studies need to be performed to investigate the repopulation of the aerosol size distribution from nucleation mode aerosol are required, including the sensitivity of the CCN population to particle nucleation events.
- An aerosol-cloud model is needed, which includes aerosol charge, and therefore permits an investigation of the sensitivity of the cloud properties to the charge state of the aerosol. It is likely that the aerosol coagulation, droplet scavenging, and ice nucleation rates will need to become a function of aerosol charge.
- Improvements are needed in understanding the effect of ion composition on aerosol nucleation, with laboratory measurements of basic thermodynamic quantities for the nucleation models.

Other areas which are closely-linked are prioritised as:

- Reconstruction of past global circuit changes, to assess the possible large-scale response to electrical cloud microphysics.
- Consideration of the sensitivity of satellite sensors to direct cosmic ray effects and the sensitivity of satellite cloud retrievals to long wave absorption by ion clusters
- Studies on the effects of the atmosphere on surface neutron measurements
- Analysis of atmospheric data from the nuclear weapons test period for tropospheric evidence of radiolytic particle formation
- Modelling of the spatial and temporal changes in ion production rates (especially from cosmic rays) in the cloud-forming regions of the atmosphere, for comparison of cloud and aerosol properties in the same regions.

7.2 Advice on future model implementation

For implementation in a coupled GCM to be used for simulating climate change, such as the Hadley Centre runs for IPCC, we recommend incorporation into the model:

- TSI variations as discussed in section 2.
- Several spectral bands, extending shortwave to at least 200nm, in the SW radiation scheme.

Some representation of solar-induced ozone changes. It is difficult to be specific about these but probably the least controversial choice would be to enhance lower stratospheric ozone in line with the signal found in TOMS data over the past 2.5 solar cycles, to derive a scaling between ozone and TSI from this period and then to use this scaling to derive past ozone from the chosen TSI reconstruction.

For scientific investigations with GCMs we suggest experiments to investigate:

- The importance of representing the middle atmosphere, including vertical resolution near the tropopause and the need to include fully coupled chemistry schemes.
- Non-linear interactions between increasing GHG effects and solar effects.
- The role of feedbacks between the ocean surface temperature, evaporation and cloud. This might involve mesoscale studies to investigate the importance of horizontal inhomogeneity.
- The sensitivity of liquid clouds to changes in background ultrafine aerosol concentrations
- The parametrisation of aerosol microphysics to link aerosol production to ion production rates.

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