Signatures of naturally induced variability in the atmosphere using multiple reanalysis datasets

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A multiple linear regression analysis of nine different reanalysis datasets has been performed to test the robustness of variability associated with volcanic eruptions, the El Niño Southern Oscillation, the Quasi-Biennial Oscillation and with a specific focus on the 11-year solar cycle. The analysis covers both the stratosphere and troposphere and extends over the period 1979–2009. The characteristic signals of all four sources of variability are remarkably consistent between the datasets and confirm the responses seen in previous analyses. In general, the solar signatures reported are primarily due to the assimilation of observations, rather than the underlying forecast model used in the reanalysis system. Analysis of the 11-year solar response in the lower stratosphere confirms the existence of the equatorial temperature maximum, although there is less consistency in the upper stratosphere, probably reflecting the reduced level of assimilated data there. The solar modulation of the polar jet oscillation is also evident, but only significant during February. In the troposphere, vertically banded anomalies in zonal mean zonal winds are seen in all the reanalyses, with easterly anomalies at 30°N and 30°S suggesting a weaker and possibly broader Hadley circulation under solar maximum conditions. This structure is present in the annual signal and is particularly evident in NH wintertime. As well as the ‘top-down’ solar contribution to Northern Annular Mode variability, we show the potential contribution from the surface conditions allowing for a ‘bottom-up’ pathway. Finally, the reanalyses are compared with both observed global-mean temperatures from the Stratospheric Sounding Unit (SSU) and from the latest general circulation models from CMIP-5. The SSU samples the stratosphere over three different altitudes, and the 11-year solar cycle fingerprint is identified in these observations using detection and attribution techniques.

**Key Words:** natural variability; solar; stratosphere; S-RIP; reanalyses; regression

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1. Introduction

There are a large number of natural sources that give rise to variability in the atmosphere. They include so-called externally driven variability associated with solar and volcanic variability, internally driven variability associated with phenomena such as the ENSO* and the QBO and also internally driven variability associated with fluid instabilities and wave–mean flow
interactions. These modes of variability do not act in isolation and there are many examples where interactions are known to occur. A full understanding of the influence of these sources of variability and their interactions is not possible by studying observations alone, because it is difficult to assign cause and effect given the complexity of the system. For this reason comprehensive general circulation models of the atmosphere–ocean system are required to explore mechanisms and their interactions. However, these models require validation to increase our confidence that the individual sources of variability have been adequately modelled.

An opportunity to characterise the various sources of variability in reanalysis data and to help validate models is presented by the SPARC-RIP. This is a coordinated activity to compare the available reanalysis datasets (Fujiiwara and Jackson, 2013); it aims to compare, contrast and understand the causes of differences among reanalyses. It focuses on reanalysis fields in the upper troposphere, stratosphere and lower mesosphere for the period since 1979, when greater availability of satellite observations for data assimilation had a marked influence on the reanalysis fields. Nine reanalysis datasets are available from five different reanalysis centres (ECMWF, NOAA-NCEP, JMA, NASA and NOAA-CIRES) as shown in Table 1, which provides a brief description of each dataset, including the resolution of the forecast model used in the reanalysis system and the most relevant references for further information. Several of the datasets are different evolutionary versions of the same product. For example, the ERA-Interim dataset is a more recent and improved product than the ERA-40 dataset. Similarly, JRA-55 is an updated version of JRA-25.

The primary purpose of this article is two-fold. Firstly, to broadly characterise the annual zonal-mean temperature and zonal wind signatures of major sources of atmospheric variability, namely the 11-year SC, volcanic, ENSO and QBO in order to provide a test bed for atmospheric model validation. Secondly, to examine in more detail the seasonal evolution of the 11-year SC signal by analysing individual months to investigate aspects of solar influence that are believed to operate through the stratosphere and may penetrate downwards to influence the troposphere and surface. In particular, we examine the robustness of the temperature maxima in the equatorial upper and lower stratosphere identified by Crooks and Gray (2005) (also Shibata and Deushi, 2008; Frame and Gray, 2010) and the poleward/downward propagation of the small upper-stratospheric subtropical wind anomaly in winter identified by Kodera and Kuroda (2002).

The study employs a linear multiple regression technique and the results from each reanalysis dataset are compared and contrasted in order to identify which responses are common to all datasets and may thus be considered more robust. The multilinear regression analysis includes indices to represent the 11-year SC, volcanic eruptions, ENSO, the QBO and a linear trend. A residual term is also included to account for internal climate variability. There are many drawbacks to employing a linear technique, especially when it is clear that there are nonlinear interactions between the different sources of variability (e.g. Lee and Smith, 2003; Chiodo et al., 2014). These drawbacks must be borne in mind when interpreting the results. Nevertheless this approach remains a useful way to characterise the main features of the responses and should provide a useful benchmark by which to test the capability of models, especially if the model output is analysed in a similar manner. The technique has previously been employed (e.g. Haigh, 2003; Crooks and Gray, 2005; Frame and Gray, 2010) but only to individual reanalysis datasets. A thorough comparison of different regression techniques, including this one, is required.

The layout of the article is as follows. Details of the regression technique are provided in section 2 along with a description of the reanalysis datasets. The main signatures of volcanic, ENSO and QBO influences are described in section 3. The solar signal is described in section 4 and results of the analysis are summarised in section 5.

2. Data and methods

2.1. Reanalysis datasets

Reanalyses blend observations with a model, taking into account uncertainties in both. Each reanalysis consists of a data assimilation scheme and a forecast-model component. Table 1 provides brief details of each of the reanalysis datasets employed, including the type of model, horizontal and vertical resolutions and the height of the upper lid of the model. All except the MERRA model employ spectral coefficients to represent the analysed fields. There is some variation in the height of the lid of the model. While most of the models extend to the lower mesosphere and thus include the whole of the stratosphere, the NCEP-1 and NCEP-2 models extend only to 3 hPa (and data are available at heights below 10 hPa). Nevertheless they are included in this analysis because they provide additional evidence for robustness of features in the lower stratosphere and troposphere. Similarly, the 20CR dataset extends only to the mid-stratosphere (with data available up to 10 hPa); it is unusual because it assimilates only surface pressure observations (and employs monthly averaged sea-surface temperature and sea-ice distributions as boundary conditions in a similar manner to the recent reanalysis datasets). It therefore does not include signals from directly assimilated measurements of the atmosphere. Nevertheless, it may show signals in the atmosphere that arise because of the variability in the assimilated surface fields and hence can potentially indicate the relative importance of surface influences on the atmospheric signal, in agreement with (e.g.) Misios and Schmidt (2013).

All the reanalysis systems under consideration (apart from 20CR) assimilate atmospheric observations from both in situ and satellite instruments, where the in situ measurements come from radiosondes, commercial aircraft and station data. The satellite observations are included either directly by assimilating observed radiances (MERRA, ERA-40, ERA-Interim, JRA-25, JRA-55, NCEP-CFSR) or indirectly by first carrying out an offline temperature retrieval and then assimilating the retrieved temperature profiles (NCEP-1, NCEP-2). In this way it is possible for the reanalysis fields to capture some of the influences of the 11-year SC, volcanic, QBO and ENSO on atmospheric temperatures (and hence on zonal winds). In addition, some of the forecast model components include time-varying irradiances and aerosols (Table 2) in the radiation scheme so that the 11-year SC in irradiances and variations in stratospheric aerosol due to volcanic eruptions are directly represented in the forecast model itself and thus in resulting background fields presented to the assimilation scheme. There is also an 11-year SC and volcanic signal observed in ozone amounts (Gray et al., 2010); however, although ozone observations are assimilated in the majority of the reanalyses, those ozone values are not generally passed through to the radiation scheme (Table 2), so this additional mechanism for generating a volcanic or solar signal in the temperatures was not generally included.

The regression analysis has been carried out for the period 1979–2009 since this is the data period available from the majority of the reanalysis datasets, with the exception of the ERA-40 dataset which has been analysed only for the period 1979–2001 (i.e. the ERA-40 data have not been extended using operational data as in the case of Frame and Gray, 2010). The regression analysis was carried out using monthly mean datasets provided by the reanalysis centres which are available on pressure surfaces. Where possible (i.e. for ERA-Interim and JRA-55), additional model-level data provided at 6 h intervals were downloaded so the atmosphere could be analysed at higher altitudes. These were interpolated onto pressure levels 0.7, 0.5, 0.2, 0.1 hPa and averaged to monthly data. The number of pressure levels provided by each...
Table 1. Technical details of the reanalysis datasets.

| Reanalysis product | Centre | Period of available data | Analysis system version | Horizontal resolution | No. of model vert. levels | Lid height of forecast model (hPa) | Reference |
|--------------------|--------|--------------------------|-------------------------|-----------------------|--------------------------|-----------------------------------|-----------|
| MERRA              | NASA   | 1979–present             | GEOS 5.0.2 (2008)       | 2/3°lat × 1/2°lon     | 72                       | 0.01                              | Rienecker et al. (2011)          |
| ERA-40             | ECMWF  | Sep 1957–Aug 2002        | IFS Cycle 23r4 (12 Jun 2001) | T159                  | 60                       | 0.1                               | Uppala et al. (2005)            |
| ERA-Interim        | ECMWF  | 1979–present             | IFS Cycle 31r2 (5 Jun 2007) | T255                  | 60                       | 0.1                               | Dee et al. (2011)               |
| JRA-25/ JCDAS      | JMA    | 1979–Jan 2014            | JMA GSM (Mar 2004)      | T106                  | 40                       | 0.4                               | Onogi et al. (2007)             |
| JRA-55             | CRIEPI | 1948–1979                | JMA GSM (Dec 2009)      | TL319                 | 60                       | 0.1                               | Ebita et al. (2011)             |
| NCEP-1 (R-1)       | NOAA/NCEP | 1948–present            |                          |                       |                          |                                   |                                      |
| NCEP-2 (R-2)       | NOAA/NCEP | 1948–present            |                          |                       |                          |                                   |                                      |
| NCEP-CFSR          | NOAA/NCEP | 1979–present            | 2007 (CFS)              | T382 (T574 from 2010) | 64                       | 0.266                             | Saha et al. (2010)              |
| NOAA-CIRES (20CR_v2) | NOAA/ESRL and CIRES Univ. Colorado | 1871–2010 | NCEP GFS Apr 2008 | T62 | 28 | 2.511 | Compo et al. (2011) |

Table 2. Further relevant details of the reanalysis datasets.

| Reanalysis product | SCI? | SCO? | Time-varying aerosol |
|--------------------|------|------|----------------------|
| MERRA              | No   | Yes  | No                   |
| ERA-40             | No   | No   | No                   |
| ERA-Interim        | No   | No   | No                   |
| JRA-25/ JCDAS      | No   | Yes  | Only as monthly mean annual cycle |
| JRA-55             | No   | No   | No                   |
| NCEP-1 (R-1)       | No   | No   | No                   |
| NCEP-2 (R-2)       | No   | No   | No                   |
| NCEP-CFSR          | Yes  | Yes  | Yes                  |
| NOAA-CIRES (20CR_v2) | Yes | No   | Yes                  |

SCI? = Solar cycle in irradiances passed to radiative part of forecast model?  
SCO? = Solar cycle in ozone passed to forecast model?  
*SBUV = Solar Backscattered Ultraviolet.

of the reanalysis centres for the standard data is not uniform. The main annual response analysis (Figures 1–8) was carried out on the pressure levels provided by each centre, to maintain as much vertical information as possible. We note that MERRA, unlike all others, does not extrapolate data underground (i.e. to 1000 hPa over mountainous regions) so the results at the very lowest levels should be disregarded for the MERRA dataset. Similarly, the horizontal resolution of the available data is not uniform and the latitude grid provided by the operational centres has been retained in the analysis to maintain as much horizontal resolution as possible. When features are examined in detail on a month-by-month basis, only averages of the three datasets MERRA, JRA-55 and ERA-Interim are shown for conciseness, since these are considered to be the most up-to-date analyses that provide data up to the lower mesosphere. Where averaged signals are shown (as in Figures 9, 10 and 14) the individual dataset regression coefficients were interpolated onto a standard set of pressure levels (1000, 975, 950, 925, 900, 875, 850, 825, 800, 775, 750, 725, 700, 650, 600, 550, 500, 450, 400, 350, 300, 250, 200, 150, 100, 70, 50, 40, 30, 20, 10, 7, 5, 4, 3, 2, 1 hPa) and a standard latitude grid (every 10° from Pole to Pole) before the averaging was performed.

There are a number of issues surrounding the changes in observational data coverage and instrument changes that can result in biases, jumps and drifts in the reanalysis datasets. Various bias corrections are employed by the different reanalysis...
centres but they are not uniformly applied across the different reanalysis data centres. (The individual references provided in Table 1 give details of these.) In the stratosphere the major source of discrepancy occurs when observations from new satellite instruments are introduced. The discrepancies are linked to relative biases between the new observations and previous observations and/or the forecast component of the assimilation system. An example of this was in October 1998 when data from the ATOVS on the AMSU-A replaced the TOVS SSU data stream (e.g. Rienecker et al., 2011). Although it is possible to include additional forcing indices with a step change at the appropriate date to account for these (as done for example by Crooks and Gray, 2005, where different operational streams were joined), it is less practicable to do this when so many diverse reanalysis datasets are being analysed. Using different indices to try to account for these different adjustment treatments would also mean that a different set of indices would be used for each dataset analysis, making a valid comparison difficult. For these reasons, the results we present have employed identical forcing indices (although the QBO index is calculated separately for each reanalysis zonal wind) and no additional indices have been included to account for these transitions. Step changes in the data can cause issues, and so are discussed in more detail during the analysis (specifically in section 4.4). For this study it is desirable to report on the raw data, therefore we have chosen not to carry out any adjustments to the datasets ourselves.

2.2. Regression technique

The study employs the linear multiple regression analysis technique described by Crooks and Gray (2005) (also Gray et al., 2013). The data are first de-seasonalised by removing the relevant climatology (derived from the reanalysis dataset in question) for the period 1979–2009. The data are then fitted to 18 independent indices using a total least-squares regression model. For simplicity, no time lags were included in any of the forcing indices. The monthly averaged time series of TSI was employed to represent 11-year SC variations using the reconstructions from Wang et al. (2005), an update to Lean et al. (2005) (also Lean et al., 1997; Lean, 2000; Lean and Rind, 2009). The dataset is available from http://solarisheppa.geomar.de/solarisheppa/cmip5 (accessed 13 November 2014) and is referred to as the NRLSSI dataset. The updated monthly mean Sato et al. (1993) aerosol index was employed to represent volcanic eruptions. The updated average monthly mean Niño 3.4 index (5°S−5°N; 120°−170°W) derived from the ERNST version 3b dataset (Smith and Reynolds, 2003) was used to represent ENSO. (http://www.cpc.ncep.noaa.gov/data/indices/; accessed 13 November 2014).

Two indices were used to represent the QBO. These were derived by first carrying out a preliminary multilinear regression analysis of the monthly averaged zonal wind field from each dataset, employing all indices apart from the QBO (so the QBO variability in this case is assigned to the residual term). The first two Empirical Orthogonal Coefficients (EOCs) of the residual of this preliminary analysis were then calculated for the equatorial region (5°S−5°N) between 10 and 100 hPa. The full regression analysis was then performed by repeating the analysis but including the two EOC time series as the QBO indices. In this way, any solar, ENSO or volcanic influence on the QBO index is minimised (Crooks and Gray, 2005, give more details). Employing two (orthogonal) QBO indices allows us to take into account the descent of the QBO with time, e.g. to account for the fact that the phase of the 50 hPa QBO is approximately out of phase with the QBO signal at 10 hPa. Twelve indices were also used to remove any recurrent annual variations in the forcing indices; they consisted of a series of spiked square pulses superimposed on a zero line with each pulse representing a single month (Crooks and Gray, 2005, give more details; although these indices were included for the sake of rigour, the results were found not to be sensitive to their exclusion). The final index was a simple linear trend term.

An autoregressive AR(1) noise model was included in the analysis. A two-tailed Student’s t-test was used to determine the probability that the regression coefficients are significantly different from the noise. Before the regression analysis is performed, all forcing index time series are normalised by dividing the time series by its standard deviation, so the regression coefficients from the analysis provide estimates of the variability associated with one standard deviation of the index. However, for presentation purposes, the regression coefficients have been first normalised by the standard deviation and then rescaled by multiplicative factors to give a more meaningful estimate. For the 11-year SC, the coefficients have been multiplied by the maximum peak-to-trough values of the TSI index to obtain an estimate of the maximum likely response to variations in the Sun’s output; they therefore represent a ‘solar max minus solar min’ difference, amounting to a change in TSI of 1.4 W m\(^{-2}\). For the volcanic results, the regression coefficients have been multiplied by the difference between the largest and smallest ENSO and QBO index values and thus represent an estimate of the maximum likely ‘El Niño minus La Niña’ and ‘QBO-west minus QBO-east’ signals, respectively.

The analysed data are monthly mean, zonal-mean temperature and zonal winds from the nine reanalysis datasets in Table 1. Results for individual months (Figures 9 and 10) were derived by performing the regression analysis using indices and data only for that month, e.g. by selecting only the January from each year and fitting it to the January of each index (in this case the 12 additional monthly indices were not required since there is no chance of a contaminating seasonal cycle being present in the index). The results referred to as the annual response (Figures 1–8) were derived by fitting all the available months of the data to all months of the indices.

3. The ENSO, volcanic and QBO signals

In the following sections the signatures associated with the ENSO, volcanic, solar and QBO indices employed in the regression are described. Shaded regions in the Figures denote statistical significance at the 95% and 99% levels and these are the prime regions to be compared and contrasted. Since the focus of this article is on the solar signal, the ENSO, volcanic and QBO signals are discussed quite briefly.

3.1. ENSO signals

Figures 1 and 2 show the annual ENSO response in zonal-mean temperatures and zonal winds respectively. The tropospheric signals are highly statistically significant and remarkably consistent between the datasets. The primary signals consist of anomalously positive temperatures in the equatorial troposphere of up to 1 K, associated with increased latent heat release (Figure 1), and strengthened westerly winds in the Subtropics of approximately 5 m s\(^{-1}\), consistent with an increased Hadley circulation (Figure 2). There is also a signal in the Southern Hemisphere (SH) zonal winds indicative of an ENSO influence on the SAM, with a dipolar positive/negative response centred around 30 and 55°S that is statistically significant right down to the surface. A similar structure is present in the Northern Hemisphere (NH) centred at around 30 and 45°N but the significance does not extend down to the surface.

There is also evidence of an ENSO influence in the stratosphere. There is a statistically significant negative temperature anomaly over the Equator centred around 50 hPa which may indicate an
ENSO influence on the QBO (e.g. Calvo et al., 2010), possibly through anomalous forcing of the equatorial waves responsible for the QBO. However, interaction of ENSO and the QBO is likely to be highly nonlinear and the meridional structure of the anomaly looks similar to that of the QBO, suggesting that the linear technique may not distinguish accurately between the two signals (Crooks and Gray, 2005; Garfinkel and Hartmann, 2007). There is also a statistically significant positive anomaly in the NH polar mid/upper stratosphere of up to 5 K and an associated small easterly wind anomaly in the same region, although the latter is barely significant because of high variability in the region. Analysis of monthly data (not shown) confirms that this NH polar response is primarily present in winter and is thus indicative of increased wave propagation into the winter stratosphere during strong El Niño events (Garfinkel and Hartmann, 2008). All of these signals are in good agreement with previous analyses and modelling studies (e.g. Sassi et al., 2004; Garcia-Herrera et al., 2006; Camp and Tung, 2007; Calvo et al., 2009; Bell et al., 2010; Garfinkel et al., 2010; Mitchell et al., 2011).

3.2. Volcanic signals

The temperature response to volcanic eruptions is shown in Figure 3 for all nine reanalysis datasets. There is good agreement between the datasets in terms of the position, amplitude and structure of the response with a positive response in the tropical lower stratosphere and a smaller generalised cooling in the troposphere. The positive temperature anomaly of up to 3 K in equatorial/subtropical regions is centred at around 50–70 hPa, in response to the anomalous additional heating associated with increased absorption of both near-infrared solar radiation and terrestrial long-wave radiation by the volcanic aerosol (Robock, 2000). The structure of the anomaly is triangular in shape, with the anomaly over the Equator extending higher into the stratosphere than at the Subtropics. This is similar to that seen in observations of aerosol extinction ratio, e.g. after the eruption of Mount Pinatubo (Trepte and Hitchman, 1992), and is consistent with the expected spread of the aerosol, with greater lofting at equatorial latitudes associated with the upwelling branch of the Brewer–Dobson circulation and a distribution at midlatitudes following isentropic surfaces, along which mixing occurs in the lower stratosphere. Despite assimilating only surface observations, the 20CR dataset captures the same broad structure of the anomaly, albeit at much reduced amplitude and with a much broader region of equatorial warming, because it includes a representation of volcanic aerosol in its radiative scheme. Its inability to capture the structure of the lower stratospheric temperature response is likely associated with the low lid of
the model which means that it cannot adequately represent the Brewer–Dobson circulation and this inadequacy is not compensated by the assimilation of observations, as in the case of the NCEP reanalyses.

At high latitudes, there are positive temperature anomalies of around 2–3 K present in the different reanalyses, with up to 8 K present for the JRA-25 reanalysis. These high temperature responses occur at both Poles in the upper stratosphere, generally extending between ∼1 and 50 hPa, although the significant regions for these signals are not very consistent among the reanalyses, especially in the SH. These signals were also noted in the ERA-40 analysis of Crooks and Gray (2005) using regression analysis, and Mitchell et al. (2011) using composite analysis. They were found to come primarily from the winter months and assumed to be associated with changes in planetary wave propagation and thus circulation strength. Although the positive temperature response in the polar upper stratosphere is suggestive of increased planetary wave influence at these altitudes, the signal does not extend deep into the lower stratosphere and indeed the volcanic response at NH polar latitudes changes sign in the lower stratosphere and generally shows weak polar stratospheric cooling below about 50 hPa, although only the JRA-25 analysis shows any statistical significance of this feature. There is also a negative volcanic temperature response anomaly at ∼3 hPa over the Equator, above the main primary equatorial response, although the structure of the anomaly is not consistent between the analyses and there is only minimal statistical significance. It is likely that there is some contamination of the signal, for example by the QBO, since several of these distributions have a meridional structure similar to that of the QBO (e.g. Lee and Smith, 2003). A likely contamination with the QBO is supported by the zonal wind anomalies in Figure 4, which show a highly significant westerly wind response over the Equator centred at around 10 hPa in all of the datasets. This overlies a small easterly wind anomaly that is centred approximately at the height of the maximum heating. The easterly anomaly can be accounted for by increased ascent associated with the (diabatic) aerosol heating but the large westerly anomaly above it can only be a result of anomalous wave propagation (also Graf et al., 2007), since it implies super-rotation which requires additional momentum transfer into the region. The signal is therefore most likely associated with a volcanic influence on the propagation of equatorial waves (primarily Kelvin, inertia-gravity and gravity waves) which are important in the generation of the SAO and the QBO. Increased upwelling in the equatorial lower stratosphere after a volcanic eruption associated with the local anomalous heating could also have an influence on the descent rates of the QBO phases, leading to a stalling of the descent (Baldwin et al., 2001). Examination of the equatorial zonal wind time series (not shown) indicates a prolonged stalling of the easterly QBO phase descent after the

Figure 2. As Figure 1, but for zonal wind speed (m s$^{-1}$).

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The annual temperature response to volcanic eruptions for each of the nine reanalyses, obtained using multiple linear regression. The units are in K per standard volcanic eruption (where we choose Mt Pinatubo as the ‘standard’ eruption). Note that MERRA does not extrapolate data down to 1000 hPa, so some data are missing due to topography.

Figure 3. The annual temperature response to volcanic eruptions for each of the nine reanalyses, obtained using multiple linear regression. The units are in K per standard volcanic eruption (where we choose Mt Pinatubo as the ‘standard’ eruption). Note that MERRA does not extrapolate data down to 1000 hPa, so some data are missing due to topography.

eruption of Pinatubo, lasting from summer 1991 through the NH spring of 1992. However, a similar stalling is not seen following the eruption of El Chichon.

The pattern of positive temperature anomaly at equatorial latitudes and negative anomaly at polar latitudes in the lower stratosphere gives rise to a westerly wind anomaly at midlatitudes through thermal wind balance, and this is also evident in the zonal wind distributions shown in Figure 4. There is a small, statistically significant westerly wind response of around 2 m s$^{-1}$ in the NH polar vortex and this signal extends deep into the troposphere. This response has been noted by previous studies (e.g. Robock, 2000; Stenchikov et al., 2006; Driscoll et al., 2012) and gives rise to a NAM response together with a distinctive surface temperature and wind response associated with a positive phase of the NAO in the one or two winters following an eruption. Analysis of the separate months (not shown) confirms that, although a small wind anomaly is present year-round, consistent with a thermal wind response, the signal is stronger in the winter months, suggesting a role for volcanic influence on planetary wave propagation. There is also some evidence of a similar small westerly wind anomaly between 10 and 100 hPa in the SH polar region but there is no statistical significance.

In addition to the clear NAM response in the NH mid- to-high latitudes that extends throughout the stratosphere and troposphere, there is evidence of an inverted horseshoe distribution in the tropical troposphere zonal wind anomalies, similar to that noted by Haigh et al. (2005), with a negative anomaly extending from the equatorial lower stratosphere into the subtropical troposphere at 30–40°N and 50–60°S. This pattern of response was investigated by Simpson et al. (2010) who tested the tropospheric wind response to idealised heating anomalies in the lower stratosphere. However, from the responses in Figures 3 and 4, it is unclear whether the volcanic response signal in midlatitude tropospheric zonal winds comes primarily from the high-latitude polar response (i.e. positive NAM in winter) or from the anomalous heating in the equatorial lower stratosphere, or indeed a mixture of both.

3.3. QBO signals

As explained in section 2, two terms were employed to represent QBO variability in the regression analysis, referred to as QBO-A and QBO-B. For conciseness, only the QBO-B distributions are shown in Figure 5 for each individual reanalysis dataset, to facilitate an assessment of any differences between the individual reanalysis results. However, the average structure of the QBO-A distributions is also shown later (Figure 14).
As expected, the QBO temperature responses (Figure 3(a)) show an alternating positive/negative structure in height over the Equator, with all datasets displaying the triple anomaly structure in height highlighted by Pascoe et al. (2005), where a third equatorial anomaly is also present in the upper stratosphere, centred around 3 hPa. This third upper-stratospheric anomaly is indicative of a QBO influence on the SAO, which dominates this height region (Gray, 2010, and references therein). The meridional structure in Figure 5 is also very characteristic of the QBO. The equatorial temperature anomaly (Figure 5) on average extends to \(\sim 15^\circ\) with a reversal in sign in the Subtropics which extends to \(\sim 45–50^\circ\). This is consistent with the local induced (adiabatic) meridional circulation (Plumb and Bell, 1982; Baldwin et al., 2001). In the zonal wind analysis (Figure 6), the signature consists of a westerly signal of around 30 m s\(^{-1}\) centred at 50 hPa over the Equator that is broader than the corresponding temperature anomaly and extends to \(\sim 25–30^\circ\). Above this, the next incoming easterly anomaly is evident, with amplitudes reaching 40 m s\(^{-1}\) around 10 hPa. In the Subtropics between 30 and 100 hPa there is a region of opposite-signed anomalies consistent with the induced secondary meridional circulation (Plumb and Bell, 1982; Baldwin et al., 2001; Gray, 2010). In the NH at upper levels this subtropical westerly anomaly extends to polar latitudes with a statistically significant amplitude of approximately 2 m s\(^{-1}\). This high-latitude wind signal is present primarily in the winter months (not shown) thus confirming the Holton–Tan relationship (e.g. Anstey and Shepherd, 2014), in which an equatorial easterly QBO anomaly is associated with a winter polar vortex that is warmer and more disturbed (and vice versa for equatorial westerly anomalies).

The regression coefficients for the QBO are distinctly different for the 20CR than the other reanalyses, and this is because the underlying driving model of the assimilation is not adequate to generate an internal QBO. This is most likely due to the relatively coarse vertical resolution of 20CR, and high vertical resolution is a prerequisite for QBO generation (e.g. Baldwin et al., 2001). While the resolution of most of the other reanalyses is also coarse, they are assimilating observational data in this region, which will contain information about the QBO.

In the troposphere, there is a suggestion in some of the temperature analyses (Figure 5) that the QBO signal extends down into the troposphere but none of these signals are statistically significant. However, in the corresponding zonal wind analysis (Figure 6) this is more evident, with an inverted horseshoe-shaped response so that the equatorial lower-stratospheric westerly anomaly extends deep into the tropospheric Subtropics and the subtropical mid-stratospheric easterly anomaly extends deep into the tropospheric midlatitudes. This is in agreement with previous analysis of the QBO signature (e.g. Anstey and Shepherd, 2014). The resulting vertical banding structure in the troposphere shows high statistical significance in the majority of the reanalysis datasets.
4. 11-year solar cycle signals

4.1. Summary of proposed mechanisms

A number of mechanisms for 11-year SC influence on the atmosphere have been proposed. The direct influence of variations in the visible and infrared part of the solar spectrum penetrates down to the Earth’s surface. This is usually referred to as the bottom-up mechanism. It primarily influences tropical SSTs and thus the pattern of evaporation, convection and latent heat release (Meehl et al., 2009). This in turn can influence the large-scale global atmospheric (Hadley and Walker) circulations, the generation of waves in the tropical atmosphere and their propagation both horizontally into the mid and high latitudes and vertically into the stratosphere.

On the other hand, variability in the UV part of the Sun’s spectrum primarily influences the temperature and composition of the stratosphere (Haigh, 1994). The UV variability directly influences the tropical mid to upper stratosphere, increasing the temperature at solar maximum not only because of the increased irradiance but also because of the resulting increase in production of ozone, which is the main absorber of radiation in this height region. For this reason, the UV solar influence is generally referred to as the top-down mechanism. Gray et al. (2009) proposed that the increased temperatures at solar maximum over the equatorial stratopause could be attributed approximately 60% to the direct irradiance changes and 40% to the ozone changes. However, their estimates employed solar irradiance variations in the UV region of the solar spectrum from the NRLSSI dataset. Recent observations from the SORCE satellite instrument have suggested much greater variability in the UV spectral range (Harder et al., 2009), and the NRLSSI estimates are now considered to be at the lower end of the probable range (Ermolli et al., 2013), so there is still much uncertainty and the estimates of Gray et al. (2009) may need to be revised. It should also be noted that the SORCE measurements are still not certain, and therefore may still need further, substantial bias correction.

Additionally there are a number of indirect mechanisms by which small temperature variations in the tropical stratosphere may influence the underlying weather patterns at the surface, several of which interact or describe different aspects of the same process, so that separation of the mechanisms is difficult. One of the major mechanisms involves changes in the background winds associated with the tropical temperature variations in the stratosphere. Kuroda and Kodera (2001) (also Kodera, 1995; Kodera and Kuroda, 2002) analysed NCEP and MO data for the period 1979–1999 to form composite differences between NH zonal wind distributions under solar maximum and solar minimum conditions (Figure 3 of Kuroda and Kodera, 2001). They identified a small westerly anomaly in the subtropical upper stratosphere in November that grew in amplitude in December.
and moved poleward and extended downwards by January. They proposed that the small anomaly in subtropical winds early in winter could be explained by thermal wind balance in the presence of horizontal temperature gradients associated with the equatorial stratospheric anomaly. These small subtropical wind anomalies could then be amplified by wave–mean flow interaction during wintertime, thus providing a positive feedback mechanism through which the small solar influence in the upper stratosphere could be amplified and extend its influence to lower levels and ultimately down into the troposphere. This is therefore primarily a wintertime response.

An 11-year SC influence on planetary wave propagation will influence not only the strength of the polar vortex but also the strength of the mean meridional Brewer–Dobson (BD) circulation since this is forced primarily by planetary wave momentum transfer. This will therefore give rise to 11-year variations in upwelling over the equatorial region and also an 11-year solar signal in the transport of ozone and other long-lived trace gases (Hood and Soukharev, 2003; Hood, 2004; Gray et al., 2009). Both the (adiabatic) variability in BD equatorial upwelling and the resulting variability in ozone amounts will give rise to a temperature anomaly in the lower equatorial stratosphere, which could therefore explain the secondary temperature anomaly observed in the equatorial lower stratosphere by Crooks and Gray (2005). It could also just be a regression artifact, from

the predictors being non-orthogonal (Chiodo et al., 2014). Although the BD circulation is forced by planetary waves in the winter hemisphere, this mechanism could result in a year-round temperature anomaly at equatorial latitudes due to forcing from each winter in turn. Alternatively, the observed temperature (and ozone) anomaly in the equatorial lower stratosphere could be the result of a direct SC modulation within the tropical region, through its influence on the descent of the phases of the QBO. This manifests itself in a SC modulation of the timing of the QBO phase reversals (Pascoe et al., 2005) and possibly in the length of the individual QBO phases (Salby and Callaghan, 2006). This could also result in a year-round equatorial lower-stratospheric temperature anomaly.

The presence of an 11-year solar variation in temperature in the lowermost part of the equatorial stratosphere provides a further possible mechanism for the transfer of the 11-year solar signal to the troposphere. The associated changes in latitudinal temperature gradient will influence the strength of the midlatitude zonal winds in the lower stratosphere and hence the synoptic-scale wave propagation that influences the tropospheric midlatitude storm tracks (e.g. Haigh et al., 2005; Simpson et al., 2010). This influence will also be a year-round influence, because the tropical lower-stratospheric temperature anomaly is present year-round and synoptic-scale waves are also active year-round.

Figure 6. As Figure 5, but for zonal wind speed (m s$^{-1}$).

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In addition to the irradiance mechanisms described above, there are also proposed mechanisms that involve energetic particles from the Sun that can influence stratospheric temperatures in polar regions during winter and can also affect ozone chemistry (Randall et al., 2006; Jackman et al., 2007). There are also 11-year SC variations in the number of GCRs that impact the Earth’s atmosphere, as a result of variations in the level of shielding afforded by the Sun’s magnetic field (e.g. Gray et al., 2010). These GCR variations have been proposed as a mechanism for directly influencing cloud cover and aspects of the global electrical circuit, but the processes have yet to be included in climate models and an assessment of their impact is therefore not yet possible.

### 4.2. Annual response

The 11-year solar signals in annual zonal–mean temperature from the regression analysis are shown in Figure 7. As described in section 2, the regression coefficients have been re-scaled by the standard deviation and multiplied by the maximum peak-to-trough values of the TSI index to obtain an estimate of the maximum likely response to 11-year variations in the Sun’s output over the period. The distributions confirm the presence of the localised, double peak, temperature response in the lower equatorial/subtropical stratosphere first identified by Crooks and Gray (2005) (also Shibata and Deushi, 2008; Frame and Gray, 2010), with higher temperatures at solar maximum. The amplitude of the lower stratospheric signal reaches approximately 0.5 K over the Subtropics between 20 and 25 km and there is a localised minimum over the Equator. The only dataset not to display this lower stratospheric feature is 20CR, which is understandable given that this dataset only assimilates observations at the surface. The meridional structure of this signal is similar to that of the QBO and the signal is in the region where the QBO dominates the variability, so contamination of the signal by the QBO cannot be ruled out, especially in such a short time series (Lee and Smith, 2003; Smith and Matthes, 2008).

In the upper stratosphere, there is less agreement between the temperature responses in Figure 7. The JRA-25, ERA-40 and ERA-Interim datasets show a local equatorial maximum of 1–1.25 K centred in the region of the stratopause at around 1 hPa. On the other hand, the MERRA dataset shows a rather complicated structure with a maximum of 1 K lower down at around 3 hPa. The CFSR and JRA-55 datasets have a very weak positive signal in the equatorial upper stratosphere with no statistical significance. The better agreement in the lower stratosphere suggests that the relative abundance of observations available for assimilation is a significant factor for imparting robustness in the signal.

![Figure 7](image_url)
In a comparison of 11-year SC temperature responses predicted by a range of climate models to imposed NRLSSI and SORCE variations, Ermolli et al. (2013) show a 0.25–0.5 K response to NRLSSI forcing at \( \sim 1 \) hPa averaged between 25°S and 25°N over the period 2004–2007 compared with a 1–2 K range of responses when forced by the SORCE variations over the same period (their Figure 10). While it is difficult to convert these differences over three years to variations over a complete SC, the analysis shown in Figure 7 suggests that climate models employing the NRLSSI spectral variations may underestimate the upper-stratospheric temperature changes seen in the reanalysis datasets while those employing the SORCE spectral variation overestimate them. Additionally, the majority of the climate models employing the SORCE spectral variability showed a positive temperature response throughout the depth of the equatorial stratosphere, while those employing the NRLSSI variations showed a near-zero response in the mid-stratosphere, with localised maxima in the lower and upper stratosphere similar to that seen in Figure 7 (although we note again that this localised minimum in the height distribution is not statistically significant\(^1\)). A significant temperature anomaly is also observed in the NH troposphere, and is consistent among all reanalyses.

The 11-year SC variability in annual zonal-mean zonal winds (Figure 8) displays three main regions of statistically significant response. Firstly, in the upper stratosphere (1–3 hPa) there is a subtropical response in both hemispheres of a few m s\(^{-1}\), which is most likely a thermal wind response to the equatorial stratopause temperature response. This is evident even in those datasets whose temperature response is not statistically significant. Secondly, there is a statistically significant response over the Antarctic region between 10 and 300 hPa in all datasets apart from 20CR (which is not expected to capture the stratospheric response, since it only assimilates surface observations). In all of the datasets, this negative South Polar response is mirrored by an opposite signed response at around 50°S, although this secondary response only achieves statistical significance in one of the datasets (ERA-40). The main climatological region of westerly winds is situated at around 60°S at 10 hPa, so this solar signal anomaly suggests a broadening of the vortex winds under solar maximum conditions and an equatorward shift of the zonally averaged stratospheric jet maximum. On the other hand, in the NH there is a broad positive wind anomaly extending from 60°N to the Pole in all datasets, but none of the signals is statistically significant at the 95% level, presumably because of the much greater variability of the NH region. The time evolution of these polar anomalies is discussed

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\(^1\)The significance test we use only tests significance at an exact point, rather than the spatial structure of the response, and therefore the temperature minimum may be significant in this respect.
in more detail in the next section, to explore whether they are present only in the wintertime or all year round.

Finally, there is a clear solar signal in tropospheric zonal winds (and also in the lower stratosphere), most notably throughout the depth of the subtropical troposphere where the signal reaches 95% significance levels in all of the reanalysis datasets. The JRA-25, MERRA, ERA-40, JRA-55, NCEP-1, NCEP-2 all show 99% significance in the SH Subtropics and JRA-25, MERRA, ERA-40, JRA-55 show 99% significance in both SH and NH Subtropics, suggesting that this feature is robust. The signal again shows the similar inverted horseshoe shape displayed by the volcanic and QBO signals. At the Poles, although all datasets show an extension of the stratospheric wind anomaly reaching deep into the troposphere, i.e. a positive anomaly at solar maximum in the NH and a negative anomaly in the SH polar region, none of these tropospheric extensions is statistically significant. Because of the various possible solar mechanisms and interactions outlined in the previous section, it is impossible to determine whether this highly significant subtropical solar signal in the troposphere is primarily a response to

(i) solar variability in the high-latitude stratospheric polar vortex (the polar mechanism; Kuroda and Kodera, 2001; Kodera and Kuroda, 2002; Matthes et al., 2006; Ineson et al., 2011),
(ii) solar variability in equatorial lower stratospheric temperature and winds (the equatorial mechanism; Simpson et al., 2010) or
(iii) solar variability in surface temperatures (the surface mechanism; Meehl et al., 2009).

Note that there is also crossover between these.

Nevertheless, some discussion of the results is instructive. For example, the 20CR dataset assimilates only surface data so it may indicate how much of the atmospheric solar signal can be attributed to surface feedback mechanisms (Misios and Schmidt, 2013, give a more in-depth discussion). The structure of the 20CR solar signal in the troposphere compares reasonably well with the other reanalysis signals, especially in the NH Subtropics where there is a 95% statistically significant negative (easterly) anomaly under solar maximum conditions and also at NH midlatitudes where there is a weak significant positive (westerly) anomaly, giving rise to a positive NAM-like response in solar maximum. However, it does not reproduce the SAM-like signal suggesting that additional atmospheric mechanisms are important in the SH.

The 20CR results indicate that the assimilated surface pressure observations and/or the imposed monthly mean SSTs used as lower boundary conditions include an 11-year SC variability that generates the NAM-like response in the NH but not the SH. This suggests that the surface plays a relatively greater role in generating the tropospheric signal in the NH than in the SH. However, we caution that this result does not mean that the NH tropospheric response is therefore driven by the bottom-up mechanism involving the variability in the incoming visible/infrared irradiances absorbed directly by the surface Meehl et al. (2009). This is because the 11-year SC variations in surface pressure and temperature fields could be a result of changes in the stratosphere that impact tropospheric dynamics and hence the NAO, which can then influence the underlying midlatitude (Atlantic) SSTs (Gray et al., 2013; Scaife et al., 2013).

### 4.3. Seasonal evolution

An examination of the seasonal evolution of these SC signals may help to distinguish the relative importance of the various proposed mechanisms.

Figures 9 and 10 show the monthly evolution of the solar signal in zonal-mean zonal winds and temperatures respectively in NH winter (October–March) and SH winter (April–September). The figures show the averaged response signal from only three of the reanalysis datasets (MERRA, ERA-Interim, JRA-55). These were selected on the basis that they are the most up-to-date reanalysis data products available that include coverage of the stratosphere up to 0.1 hPa. Although not shown, all of the other reanalysis datasets show a similar behaviour, apart from 20CR which shows a much weaker response, as discussed in the previous section. The data have been interpolated onto a regular grid (as described in section 2). Significances were derived by taking the square root of the averaged squared standard error from each dataset analysis.

The solar signal most widely discussed in the NH winter stratospheric winds in previous literature is that identified by Kuroda and Kodera (2001). It can be seen in Figure 9 in the form of a subtropical NH positive wind anomaly in the upper stratosphere in November that moves polewards and then extends deep into the lower stratosphere by January (also Matthes et al., 2004, 2006; Ineson et al., 2011). The initial anomaly in early winter (November) can be interpreted as a straightforward
1.00 anomaly to the north (south) of 60° polar stratosphere in September and October. The dipolar pattern in September, together with a significant warming anomaly in the stratosphere in June that amplifies in July and extends deep into the troposphere and remains throughout the spring months to the final warming, with a delayed onset of summer easterlies under solar maximum conditions. April is a transition month when the winds gradually change from weak westerlies to easterlies. The spring/summer signal in the NH stratosphere is particularly interesting and does not appear to have attracted as much attention as the winter signal. There is a persistent, highly significant westerly anomaly of around 5 m s$^{-1}$ from April through to July that extends from the stratosphere deep into the troposphere and remains throughout the spring months to midsummer (July). It is suggestive of a solar modulation of the final warming, with a delayed onset of summer easterlies under solar maximum conditions. April is a transition month when the winds gradually change from weak westerlies to easterlies. The monthly climatological winds are only of order 10–15 m s$^{-1}$, so a solar modulation of 5 m s$^{-1}$ is relatively large.

In the troposphere, the vertical banded structure of the subtropical/midlatitude signal in the annual response (Figure 7) suggests a solar influence on the midlatitude jet streams. In the monthly wind response (Figure 9) there is evidence of a similar banded midlatitude signal primarily in late winter of both hemispheres, for example the positive anomaly of $\sim$2 m s$^{-1}$ at around 45°N in January–March and the positive anomaly at 30–45°S in September. The SH signal in September clearly extends throughout the depth of the atmosphere, connecting with the winter stratospheric dipolar signal discussed above and there is a significant negative response near 60°S. The response resembles a positive surface SAM anomaly under solar maximum conditions and vice versa under solar minimum conditions (Haigh and Roscoe, 2006; Roscoe and Haigh, 2007), although it does not appear to persist through to October/November when the stratospheric SAM connection to the troposphere is thought to be strongest. In the Subtropics the annual response (Figure 8) shows an inverted horseshoe pattern with a negative thermal wind response to higher temperatures in the equatorial stratopause region (Figure 10) and is statistically significant in all three individual regression analyses (not shown). However, the subsequent poleward and downward progression of the signal in December and January is not statistically significant, either in the wind or the temperature response, probably because the variability in these midwinter months is so high and the data record is relatively short. On the other hand, the stratospheric solar signal in late winter is much more evident, with a statistically significant negative (easterly) anomaly at polar latitudes in February (Figure 9) and a strong, significant positive anomaly over the Pole (Figure 10). This evolution of the NH stratospheric anomalies is consistent with earlier studies that show the early winter (November–January) stratosphere is less disturbed (more westerly) under solar minimum conditions than under solar minimum conditions (Labitzke, 1987; Kodera and Kuroda, 2002; Gray et al., 2004; Lu et al., 2008). However, the significant positive temperature/negative wind anomaly in February suggests that the reverse is true in the late winter. This was also noted in the study of Gray et al. (2004) who interpreted the behaviour in terms of the timing of major stratospheric disturbances, noting that they were more likely to occur earlier in the winter at solar minimum and later in winter at solar maximum. This response can be understood if the occurrence of stratospheric warming events is considered to be a cumulative result of many wave disturbances, each contributing to the destabilisation of the polar vortex, with increased wave forcing under solar minimum conditions resulting in earlier major warmings.

A corresponding SH anomaly evolution can also be seen, with a statistically significant wind anomaly in the subtropical upper stratosphere in June that amplifies in July and extends deep into the lower stratosphere in August and into the troposphere in September, together with a significant warming anomaly in the polar stratosphere in September and October. The dipolar pattern in tropospheric winds in September with a positive (negative) anomaly to the north (south) of 60°S suggests a weakening and broadening of the polar night jet under solar maximum conditions in late winter, similar to the NH response.

In terms of mechanisms, the role of the subtropical wind anomaly in the upper stratosphere may be important if planetary wave propagation is particularly sensitive to relatively small wind anomalies in this region (Kodera, 1995; Kodera and Kuroda, 2002; Gray, 2003). Thus at solar minimum, a small easterly wind anomaly in the subtropical upper stratosphere can help to guide planetary wave disturbances towards polar regions and so helps to speed up the process leading to polar warming events, while at solar maximum the westerly anomaly allows wave propagation further equatorward and hence slows down the process. However, we also note that the presence of a statistically significant signal in the subtropical upper stratosphere does not necessarily imply causality. It may simply be that variability in this region is generally relatively small so that the signal-to-noise ratio is improved in that region when compared with the polar region in midwinter or the equatorial lower stratosphere, both of which are regions of high interannual variability. An alternative possibility is that the winds in the equatorial lower stratosphere may be influenced by the SC (as discussed in the previous section; also Salby and Callaghan, 2006), influencing the QBO and thereby modulating its influence on planetary wave propagation and the polar vortex (Holton and Tan, 1980, 1982). However, the variability is also large in this equatorial region and the data record is too short to test this mechanism.

The spring/summer signal in the NH stratosphere is particularly interesting and does not appear to have attracted as much attention as the winter signal. There is a persistent, highly significant westerly anomaly of around 5 m s$^{-1}$ from April through to July that extends from the stratosphere deep into the troposphere and remains throughout the spring months to midsummer (July). It is suggestive of a solar modulation of the final warming, with a delayed onset of summer easterlies under solar maximum conditions. April is a transition month when the winds gradually change from weak westerlies to easterlies. The monthly climatological winds are only of order 10–15 m s$^{-1}$, so a solar modulation of 5 m s$^{-1}$ is relatively large.

Figure 10. As Figure 9, but for temperature (K).
response centred on 30°S and 30°N. In the monthly response (Figure 9) this feature is rather intermittent. It is strongest in January/February and June, when it reaches −2 m s⁻¹, suggesting a wintertime mechanism but it is also weakly present at other times of the year, possibly indicating a year-round response with winter amplification. At present the data record does not appear to be sufficiently long to adequately characterise the nature of this tropospheric signal.

4.4. Comparing reanalysis with satellite data

While the insight gained from reanalyses datasets is critical for some areas of the atmosphere, it is always desirable to compare the reanalysis response with the response from ‘pure’ observations. As such, we compare the global-mean temperature from the SSU which spans 1979–2005. Note that the reanalysis do assimilate the SSU data (excluding 20CR), so some correlation is expected. Figure 11 compares global average time series of the SSU (coloured) and reanalyses (black) sampled in exactly the same way (satellite weighting functions are given in, e.g. Randel et al., 2009). Blue and red lines show the MO and NOAA SSU reconstructions, respectively, the differences of which are not the focus of this article (Thompson et al., 2012, gives a complete discussion of differences between these datasets). The largest difference between the reanalysis is observed around the stratopause region in Figure 11(a) where both the MERRA and ERA-Interim datasets have a spurious jump which maximises at around the year 2000. This almost step-like behaviour has been well documented in some reanalyses and is due to the assimilation of new data from the AMSU instrument which was launched in 1998. The most recent reanalysis in this figure, JRA-55 (dot-dashed line), shows only a minor step change around this time but otherwise a far more realistic evolution. Over the other two channels, the reanalyses and SSU data are in better agreement, especially considering the uncertainty between the two different SSU reconstructions.

Figure 12 shows the SC component of annual temperature time series (i.e. from Figure 11) using the regression techniques used throughout this article. The corresponding p-values are also

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**Figure 11.** Global mean temperature anomaly averaged over the three stratospheric regions sampled by the SSU satellite over the observational period (1979–2005): (a) channel 3, (b) channel 2, (c) channel 1. Blue and red lines show two reconstructions of the SSU dataset as detailed in Thompson et al. (2012). Black lines show the three reanalyses sampled using the SSU weighting functions.

**Figure 12.** Solar regression coefficients for the two reconstructions of the SSU data and the three reanalyses shown in Figure 11. p-values are given below each symbol, rounded to the nearest percent.
given in parentheses (rounded to the nearest integer percent). The two SSU reconstructions are in good agreement, showing a temperature change of around 0.40 K for an 11-year SC in the lower channel (∼25–35 km), to around 0.55 K in the upper channel (∼40–50 km) (Randel et al., 2009), give the solar regression of the highest channel from the MO SSU reconstruction). The reanalyses are generally in agreement, although they show a smaller temperature change over the 11-year SC, on average, for all three channels. However, in channel 1, there is a clear outlier in that the ERA-Interim data (red diamond) has an amplitude of over twice that of any other dataset. This overestimate comes from the artificial step change in 1998, which projects positively onto the solar regression pattern, but also increases the uncertainty (reflected in the p-value of 85%). If the step change is accounted for in the the ERA-Interim dataset, then the regression coefficient is substantially reduced and corresponds to a SC temperature change of ∼1 K. Caution must be taken in this approach, because it introduces a human bias into how the step change is accounted for, and this is particularly problematic for datasets such as MERRA, where the change is less abrupt (i.e. Figure 11).

4.5. Detection and attribution of the solar signal in stratospheric temperatures

One clear drawback from regression studies that use reanalysis alone (or observations alone), as in the previous sections, is that they make substantial assumptions to separate the components of observed changes due to forced variability and due to natural internal variability (Hegerl and Zwiars, 2011). Detection and attribution studies get around this by using climate models to explicitly simulate both components of variability, giving an expected forced response (the so-called fingerprint), along with the uncertainty in that response (e.g. Allen and Nakamura, 2003). Using this framework, the solar forcing component of the SSU temperature evolution is examined using general circulation models taken from CMIP-5 models (Taylor et al., 2012). This allows for an identification, or not, of the solar signal in global stratospheric temperatures. While the technique is still regression-based, the method is slightly different from that detailed in section 2.2 in that the predictors are not considered noise-free, and the predictors themselves are model responses to various forcings, rather than simply a time series of the forcings (i.e. the detection and attribution method gets around the drawback of the observations-alone method, used previously). Instead of regressing the modelled temperature response to all known forcings onto different indices, the modelled temperature responses to individual forcings are regressed onto the observed temperature (Allen and Tett, 1999; Stott et al., 2003). Two issues arise with this method:

(i) the analysis is constrained to the observed period, which in the stratosphere is from 1979 onwards, and

(ii) the analysis requires many more model simulations because it not only requires all forcing scenarios (i.e. the historical scenarios from CMIP-5), but also requires simulations forced only with solar irradiance.

In this way, the observations are regressed against both all-forcing simulations and solar-only simulations in a two-way regression. We use a TLS method, following exactly the methodology of Allen and Stott (2003):

\[ y_t = \sum_{i=1}^{n} \beta_i (x_{it} - \bar{x}_t) + \epsilon_{0,t} . \]  

(1)

The \( \epsilon_{0,t} \) term represents the internal variability of the model simulations, which is estimated from pre-industrial control simulations of the appropriate model (in the previous analysis this term was not used, and the \( x \) vector was assumed to be noise-free), and the \( \epsilon_{0,t} \) term represents the observed noise. \( x_{it} \) is now a matrix containing the modelled predicted responses to various external forcings, and \( y \) is the observed response. Equation (1) cannot be explicitly solved because estimation of \( \beta_i \) requires inversion of the covariance matrix associated with \( \epsilon_{0,t} \) (Allen and Tett, 1999). To estimate the noise covariance matrix we therefore project all data onto the leading EOFs, and then estimate \( \beta \) at a specific truncation such that the variance of the original data is maximal. Note that results are not sensitive to the exact choice of truncation above 18 (and in all cases we choose truncations higher than this). As a consistency check, we also compare the variance of \( \epsilon_i \) with the variance of internal model variability using an F test (Allen and Tett, 1999). To check the robustness of the method, \( \beta \) was also estimated using a different detection and attribution technique, ROF (Ribes et al., 2013). This method has the advantage that data are not projected onto the leading EOFs, and therefore no bias is introduced when a truncation is selected. In all further analyses the TLS and ROF methods are in agreement with each other, and therefore add confidence to our results. In subsequent figures, only the TLS method is shown, for conciseness.

Figure 13 shows the time series of the monthly mean global-mean temperature for the four models with the required solar-only simulations; CanESM2, HadGEM2-CC, GISS-E2-H and GISS-E2-R. The temperatures are averaged over the height regions sampled by the SSU instrument. The all-forcing and solar-only forcing simulations are plotted in blue and green, respectively (note that CanESM2 is excluded from Channel 3 as the model lid height only extends to 1 hPa). Black and grey lines show the MO and NOAA SSU reconstructions. While differences in the long-term trend are substantial between these two reconstructions, the difference in the 11-year SC component may not be. We regress our model simulations onto both reconstructions separately in order to investigate this, and to address whether or not the 11-year SC response is (i) detected in observations and (ii) of the correct amplitude.

The global-mean modelled temperature responses to solar forcings (Figure 13, green lines) show a clear 11-year cycle with an amplitude of ∼0.5 K in the upper stratosphere (channel 3) decreasing to ∼0.25 K in the lower-middle stratosphere (channel 1). The global-mean temperature in the upper stratosphere has a negligible contribution from volcanic eruptions and is characterised principally by the long-term anthropogenic trend, with the superimposed 11-year solar variations, whereas in the middle and lower stratosphere the volcanic influence becomes more important (analysis not shown).

In order to understand how these modelled temperature responses compare with the observed temperature responses, we consider the solar-only scaling factors, \( \beta \), for each of the SSU channels. The scaling factors give an estimate of how the amplitude of the temperature patterns needs to be adjusted for consistency with observations. Figure 14 shows the solar-forced model-scaling factor, averaged over all models, using the NOAA and NOAA reconstructions. The bars show 5–95% confidence intervals. It is clear that, despite the difference in the long-term component of the MO and NOAA reconstructions, the solar irradiance component is consistent. The scaling factors are also not consistent with zero, i.e. within the 5–95% range they do not encompass zero. This means a positive detection of the solar signal in all three SSU channels. Finally, for all channels, the \( \beta \) factors are consistent with 1, which means that within the 5–95% confidence interval, the modelled solar variability is in agreement with the observed solar variability in the middle and lower stratosphere. This means that, at least in the global-mean sense, the models can be trusted to represent solar variability in these height regions, and therefore may have some skill in surface climate predictions associated with solar variability from the stratosphere. The best-guess scaling factors for channel 3 are collectively less than one.
The signals were derived by taking the average of the annual regression analysis response using the three most up-to-date reanalysis datasets that provide data up to 0.1 hPa: MERRA, JRA-55 and ERA-Interim (section 2). The estimates of variability associated with volcanic eruptions, ENSO, and the QBO are all remarkably consistent between the datasets and in general confirm the responses seen in previous regression analyses.

The analysis also suggests the presence of an 11-year SC response in the equatorial lower stratospheric temperatures (20–30 km; Crooks and Gray, 2005) although this is weaker and less statistically significant than the response to other forcings in this region of the atmosphere. These temperatures were found to be up to 0.5 K higher in solar maximum than in solar minimum and the anomaly was present in eight of the nine reanalysis datasets (the exception was the 20CR which only assimilates surface pressure data). As the result was not present in the 20CR, yet was present in NCEP-R1/-R2 reanalyses (which would not represent top-down mechanisms), it is likely that the signature is due to assimilation of data, rather than the forecast model.

The 11-year solar signal in the upper stratospheric temperatures of approximately 1–1.5 K was found to be less robust, although this may simply reflect the sparsity of assimilated data at those heights, and differences in the way in which the data are assimilated, e.g. how tidal variations are accounted for at these upper levels.

A comparison of global mean temperatures with SSU satellite data suggests that, near the stratopause, there are unphysical jumps in the reanalyses which project onto the 11-year SC and therefore may increase the magnitude of regression coefficients in this region. A detection and attribution analysis between the SSU data and CMIP-5 models supports this, and suggests a temperature change of ~0.4–0.45 K over an 11-year SC, which is in broad agreement with the reanalysis-only based regression.

The NH subtropical wind anomaly in the upper stratosphere in early winter identified by Kodera and Kuroda (2002) that moves polewards and downwards from November through to January is evident but is not statistically significant for the whole three-month evolution. In contrast, the main statistically significant polar stratospheric solar signal was found to be in late winter of each hemisphere (February/March in NH; August/Sept in SH). In solar maximum years, the NH polar stratospheric vortex was found to be more than 25 m s$^{-1}$ weaker in February than in solar minimum years and this solar signal was statistically significant at the 99% level. This pattern of behaviour is consistent with solar influence through anomalous planetary wave propagation and the timing of major stratospheric vortex disturbances, which appear to occur later under solar maximum conditions than under solar minimum conditions.

5. Summary and discussion

Results for the period 1979–2009 from a multilinear regression analysis of nine reanalysis products from four different reanalysis centres have been compared and contrasted. Figure 15 shows a summary figure of zonal-mean annual temperature and zonal wind responses to ENSO, volcanic, QBO and solar forcing.
There is an additional relatively strong statistically significant signal in zonal winds in NH spring/summer time (April through to July) which extends throughout the stratosphere and troposphere. This signal has not previously been noted. April is a transition month when the winds gradually change from weak westerlies to easterlies, so the 11-year solar signal of ~5 m s$^{-1}$ in this month is comparatively large compared with the background climatology (~10–15 m s$^{-1}$). The 11-year solar signal in spring and summertime has received much less attention than the midwinter signal and merits further investigation.

In the troposphere, the horseshoe-shaped anomaly identified in the zonal mean zonal winds by Haigh et al. (2005) is confirmed with relatively high statistical significance (95%). However, the monthly evolution of this signal is weak and intermittent and it is not possible to distinguish between the various proposed mechanisms for this signal based on the analysis of such a short dataset.

The 20CR response captures the tropospheric zonal wind response in the NH but not in the SH, suggesting a role for surface anomalies in the NH but not in the SH. However, this does not necessarily imply that the proposed bottom-up mechanism for solar influence (via TSI impact on SSTs; Meehl et al., 2009) is responsible, since the 11-year solar anomalies in the surface pressure and temperature fields assimilated in the 20CR reanalysis could also have been generated by processes involving the stratosphere.

As noted in the introduction, care should be taken in the interpretation of these results since the available data period is so short and the regression technique assumes linearity when we know that there are substantial nonlinear interactions. For instance, the predictors are often non-orthogonal, which could cause issues among, e.g. solar and volcanic signals in the lower stratosphere (Chiodo et al., 2014). The reanalysis datasets also suffer from abrupt jumps in the time dimension, especially when observational data from new satellite instruments are introduced. This is especially problematic in the mid and upper stratosphere where radiosonde observations are not available. Nevertheless, characterisation of the response to various forcings and comparison with the available reanalysis is a necessary validation requirement for climate models. For this reason the results presented here should be valuable, especially if model simulations are analysed using the same tools.

The results also highlight the difficulty of interpreting and attributing the solar response from reanalyses alone. Interpretation of the tropospheric response is especially difficult since there is an added complication of time-lags, which have not been considered here. For example, it has been shown that anomalies in the stratosphere can precede tropospheric anomalies which persist for several months after the occurrence of the stratospheric anomaly (Baldwin and Dunkerton, 2001; Mitchell et al., 2013). Because of this, a tropospheric signal seen in any particular month could be a response to stratospheric anomalies in previous months. Similarly the ocean introduces the possibility of a lagged response (Scaife et al., 2013, in the context of the solar response). Indeed, observational analysis of a long record of surface temperature and mean-sea-level pressure fields (140 years) suggest that the 11-year SC response at the surface in NH winter midlatitudes maximaes at lags of up to 3–4 years (Gray et al., 2013). Both these sources of lag in the solar response mean that the interpretation of the tropospheric response is more difficult.

Finally, we also note that care is required in the interpretation of statistical significance tests to highlight the main response regions. Although statistical tests highlight certain regional and seasonal responses, this does not necessarily mean that these are the primary regions for driving the solar response; the available data record is short and the true solar signal in other regions and months can be obscured by high levels of variability.

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Appendix

Table A1. Abbreviations and acronyms.

| Abbreviation | Description |
|--------------|-------------|
| AMIP         | Atmospheric Model Intercomparison Project |
| AMSU         | Advanced Microwave Sounding Unit |
| ATOVS        | Advanced TIROS Operational Vertical Sounder |
| CanESM       | Canadian Earth System Model |
| CFS          | Climate Forecast System |
| CFSR         | Climate Forecast System Reanalysis |
| CIERES       | Cooperative Institute for Research in Environmental Sciences |
| CMIP         | Coupled Model Intercomparison Project |
| CRIEPI       | Central Research Institute of Electric Power Industry (Japan) |
| DOE          | Department of Energy (USA) |
| ECMWF        | European Centre for Medium-range Weather Forecasts |
| ENSO         | El Niño Southern Oscillation |
| EOF          | Empirical orthogonal function |
| ERSST        | Extended reconstructed Sea-Surface Temperature |
| ESRL         | Earth System Research Laboratory |
| GCR          | Galactic cosmic ray |
| GFS          | Global Forecast System |
| GISS         | Goddard Institute for Space Studies |
| GMAO         | Global Modeling and Assimilation Office |
| GSM          | Global Spectral Model |
| HadGEM       | Hadley Centre Global Environmental Model |
| IFS          | Integrated Forecasting System |
| JCDAS        | JMA Climate Data Assimilation System |
| JMA          | Japan Meteorological Agency |
| JRA          | Japanese Reanalysis |
| MERRA        | Modern Era Retrospective Analysis for Research and Applications |
| MO           | Met Office (UK) |
| MRF          | Medium-Range Forecast |
| NAM          | Northern Annular Mode |
| NAO          | North Atlantic Oscillation |
| NASA         | National Aeronautics and Space Administration |
| NCAR         | National Center for Atmospheric Research |
| NCDC         | National Climatic Data Center |
| NCEP         | National Centers for Environmental Prediction |
| NOAA         | National Oceanic and Atmospheric Administration |
| NH           | Northern Hemisphere |
| NRLSSI       | Naval Research Laboratory |
| OAR          | Oceanic and Atmospheric Research |
| QBO          | Quasi-Biennial Oscillation |
| ROF          | Regularised optimal fingerprinting |
| S-RIP        | Stratospheric-Reanalysis Intercomparison Project |
| SAM          | Southern Annular Mode |
| SAO          | Semi-annual oscillation |
| SC           | Solar cycle |
| SH           | Southern Hemisphere |
| SORCE        | Solar Radiation and Climate Experiment |
| SPARC        | Stratospheric Processes and their Role in Climate |
| SST          | Sea-surface temperature |
| SSU          | Stratospheric Sounder Unit |
| TSI          | Total solar irradiance |
| UV           | Ultraviolet |
| 20CR         | Twentieth Century Reanalysis |

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