A Coupled Air–Sea Response Mechanism to Solar Forcing in the Pacific Region

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ABSTRACT

The 11-yr solar cycle [decadal solar oscillation (DSO)] at its peaks strengthens the climatological precipitation maxima in the tropical Pacific during northern winter. Results from two global coupled climate model ensemble simulations of twentieth-century climate that include anthropogenic (greenhouse gases, ozone, and sulfate aerosols, as well as black carbon aerosols in one of the models) and natural (volcano and solar) forcings agree with observations in the Pacific region, though the amplitude of the response in the models is about half the magnitude of the observations. These models have poorly resolved stratospheres and no 11-yr ozone variations, so the mechanism depends almost entirely on the increased solar forcing at peaks in the DSO acting on the ocean surface in clear sky areas of the equatorial and subtropical Pacific. Mainly due to geometrical considerations and cloud feedbacks, this solar forcing can be nearly an order of magnitude greater in those regions than the globally averaged solar forcing. The mechanism involves the increased solar forcing at the surface being manifested by increased latent heat flux and evaporation. The resulting moisture is carried to the convergence zones by the trade winds, thereby strengthening the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ). Once these precipitation regimes begin to intensify, an amplifying set of coupled feedbacks similar to that in cold events (or La Niña events) occurs. There is a strengthening of the trades and greater upwelling of colder water that extends the equatorial cold tongue farther west and reduces precipitation across the equatorial Pacific, while increasing precipitation even more in the ITCZ and SPCZ. Experiments with the atmosphere component from one of the coupled models are performed in which heating anomalies similar to those observed during DSO peaks are specified in the tropical Pacific. The result is an anomalous Rossby wave response in the atmosphere and consequent positive sea level pressure (SLP) anomalies in the North Pacific extending to western North America. These patterns match features that occur during DSO peak years in observations and the coupled models.

1. Introduction

Analyses of observations have suggested that the effect on the earth’s climate system of the small-amplitude variations in solar radiation that occur during the approximately 11-yr solar cycle [the decadal solar oscillation (DSO)] may produce significant responses in the troposphere (e.g., Haigh 1996, 2001, 2003; Lean and Rind 2001; Rind 2002; Lean et al. 2005; van Loon and Labitzke 1998; van Loon and Shea 1999, 2000; Gleisner and Thejll 2003; van Loon et al. 2004; Crooks and Gray 2005; Wang et al. 2005; Bhattacharyya and Narasimha 2005; Lim et al. 2006) and ocean (White et al. 1997, 1998; Bond et al. 2001; Weng 2005). Specifically for the Indo-Pacific region, van Loon et al. (2004, 2007) have demonstrated that the state of the climate system during peaks in the DSO resembles cold events (also known as La Niña events) in the Southern Oscillation. Furthermore, the monsoon regimes over South Asia...
and the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ) in the Pacific are intensified, with concomitant strengthening of the trade winds and equatorial Pacific cold tongue. The resulting convective heating anomalies are associated with an anomalous high pressure center over the North Pacific, presumably through a Rossby wave response that affects precipitation in northwestern North America. Consistent with this result, Christofoorou and Hameed (1997) found that the Aleutian low moved westward and the Pacific subtropical high moved northward during solar maxima for the period 1900–94. Moreover, on the time scale of the last 1000 yr, there is evidence that there is a cold event–like pattern during decadal periods of high solar forcing (Mann et al. 2005). But without mechanisms to account for these presumed responses to solar forcing, it could be argued that these associations in the observations are purely coincidental.

The purpose of this article is to address the mechanism that has been postulated from analysis of model simulations for the middle versus early twenty century (Meehl et al. 2003), and further proposes to explain the observed associations mentioned above from van Loon et al. (2004, 2007). The key difference from the earlier Meehl et al. (2003) results is that, for the first time, we document a response directly comparable to the observations for the DSO, which has roughly half the globally averaged forcing of the multidecadal forcing spanning the first half of the twentieth century (about 0.2 W m$^{-2}$ for DSO amplitude versus about 0.4 W m$^{-2}$) discussed in Meehl et al. (2003). Output from two global coupled climate models is analyzed here to identify the processes that are involved with this mechanism. Further atmospheric model simulations with specified heating anomalies are performed to substantiate one aspect of the hypothesized mechanism, namely the excitation of a Rossby wave response by precipitation anomalies in the ITCZ and SPCZ in the Pacific that produces an anomalous high pressure center in the North Pacific with accompanying precipitation anomalies over the northwestern United States.

The mechanism we address here depends solely on coupled interactions at the air–sea interface and is a fast response of the Pacific climate system (time scales on the order of months). In the simulations analyzed in this article, the coupled models respond only to variations of total solar irradiance (TSI). These models do not have well-resolved stratospheres, and there are no changes in stratospheric ozone in response to solar variability. A powerful amplifier that is not represented in our study’s models but that could enhance the response we address here involves variations in stratospheric ozone in response to solar UV variability. Such changes affect the vertical and horizontal temperature structure and lead to dynamical responses in the stratosphere and troposphere, with the latter extending to the tropics to enhance the precipitation maxima there (Balachandran et al. 1999; Shindell et al. 1999, 2006; Matthes et al. 2006). Indications from these studies suggest that the coupled air–sea response we document here could be augmented and enhanced by such processes in the stratosphere. Such a connection is beyond the scope of this article and needs to be addressed realistically with coupled climate models that include a resolved stratosphere and detailed stratospheric ozone chemistry. The indication that coupled processes at the surface and dynamical responses in the stratosphere could work in the same sense suggest that the signals in this article could be amplified further through these other processes.

2. Models and observed data

We analyze results from two global coupled climate models. The first, the Parallel Climate Model (PCM), is described by Washington et al. (2000) and used in the studies of Ammann et al. (2003), Meehl et al. (2003, 2004), and Santer et al. (2003a,b), among others. The resolution of the atmosphere is T42, or roughly 2.8° × 2.8°, with 18 levels in the vertical and a majority in the troposphere [model layer midpoints (hPa): 4.8, 13.1, 32.6, 63.9, 99.0, 138.7, 189.2, 251.2, 324.8, 409.0, 501.3, 598.2, 695.2, 786.5, 866.4, 929.3, 970.4, and 992.5]. Resolution in the ocean is roughly 1/4°, tapering down to 1/5° in the equatorial tropics, with 32 levels. No flux adjustments are used in the model, and, at least in terms of global-mean temperature, a relatively stationary climate is simulated. For example, a 1000-yr-long control integration shows only a small cooling trend of globally averaged surface temperatures of roughly 0.03 K per century. The interannual climate variability related to ENSO is in good agreement with observations (Meehl et al. 2001; Dai et al. 2001). The amplitude of El Niño events in PCM is close to observed, but the frequency is in the 3- to 4-yr range as opposed to observations that show more of a 3- to 7-yr spectral peak (Meehl et al. 2006a). Thus coupled processes relevant to this article are simulated, though ENSO frequency is not a factor for the mechanisms herein.

Here we analyze four-member ensembles of twentieth-century climate that branched from a preindustrial control run. This is an ensemble of opportunity, being the largest number that could be run with available computer resources. Using this rather small ensemble size increases the noise of the model response to solar forcing. In results below, this aspect is noted by high-
lighting the differences that are consistent across the model ensemble members. The twentieth-century simulations were started at intervals of roughly 10-20 yr from a long control run with constant preindustrial conditions for the year 1870 [the same technique was used for the Community Climate System Model version 3 (CCSM3) below]. The simulations use combinations of observed anthropogenic and natural forcings, including greenhouse gases (GHGs), the direct effect of sulfate aerosols, tropospheric and stratospheric ozone, solar variability, and volcanoes (Meehl et al. 2004). Of interest here is that the reconstruction of total solar irradiance used in the PCM is from Hoyt and Schatten (1993).

The second global coupled model is the CCSM3 described by Collins et al. (2006). We analyze twentieth-century simulations from the T85 version of CCSM3, with the Community Atmospheric Model version 3 (CAM3) as the atmospheric model component (and also used later for the heating anomaly experiments but with a resolution of T42). The T85 version of CAM3 has grid points in the atmosphere roughly every 1.4° latitude and longitude, and 26 levels in the vertical with a majority of the levels in the tropopause [model layer midpoints (hPa): 3.5, 7.4, 14.0, 23.9, 37.2, 53.1, 70.1, 85.4, 100.5, 118.3, 139.1, 163.7, 192.5, 226.5, 266.5, 313.5, 368.8, 433.9, 510.5, 600.5, 696.8, 787.7, 867.2, 929.6, 970.6, and 992.6]. The ocean is a version of the Parallel Ocean Program (POP) with a nominal latitude–longitude resolution of 1° (1/6° equatorial tropics) and 40 levels in the vertical. No flux adjustments are used in the CCSM3.

As in the PCM, the CCSM3 was run for a preindustrial (1870) control run, which provided initial states for the twentieth-century simulations. A five-member ensemble of CCSM3 twentieth-century simulations is analyzed here. The twentieth-century simulations were started from different times in the 1870 control run separated by 20 yr with the first ensemble member branching from the control run at yr 360. The natural and anthropogenic forcings are similar to those used in the PCM but also include black carbon aerosols as described in Meehl et al. (2006b). Of interest here is that the solar forcing in CCSM3 is the Lean et al. (1995) TSI reconstruction. The response to the two different solar forcing datasets was shown by Meehl et al. (2006b). The differences between the two lie in the time evolution of the multidecadal variability of the solar forcing. That is, Hoyt and Schatten (1993) show an earlier multidecadal peak of the solar forcing in the mid-1940s, while Lean et al. (1995) show a somewhat later peak in the 1950s. The DSO variability, which is of interest here, is comparable between the two.

The observed data used in our study are the same as those employed by van Loon et al. (2007), namely, the reanalyses by the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR), and sea surface temperature (SST) and sea level pressure (SLP) datasets from the Hadley Centre in the United Kingdom. The sea level pressure data from the Hadley Centre are available from 1871 to 1998, and cover land as well as sea. The global precipitation dataset from the Global Precipitation Climatology Project (GPCP; http://cics.umd.edu/GPCP/) is also used in our investigation. The National Oceanic and Atmospheric Administration (NOAA) Extended Reconstructed SST (ERSST) data are provided by the NOAA–Cooperative Institute for Research in Environmental Sciences (CIRES) Climate Diagnostics Center in Boulder, Colorado.

3. The coupled air–sea response mechanism to solar forcing

As noted above, Meehl et al. (2003) first proposed a coupled ocean–atmosphere response mechanism to solar forcing based on a global coupled climate model’s simulation of a significant signal in the tropics associated with low-frequency solar forcing in the early part of the twentieth century. Here we test whether that mechanism is also applicable to the tropical climate system response to the 11-yr solar cycle (DSO) in two global coupled climate models. As demonstrated shortly, the response of these models is similar to the climate anomalies that occur during DSO peaks in observations, thus lending support to this mechanism working in a similar way in nature.

As noted above, the models considered here have poorly resolved stratospheres and no 11-yr ozone variations, so the mechanism must depend almost entirely on the increased solar forcing at peaks in the DSO acting on the ocean surface in clear sky areas of the equatorial and subtropical Pacific. The solar forcing at the surface in those tropical regions can be considerably larger than the globally averaged solar forcing at the top of the atmosphere (Meehl et al. 2003). In the experiment reported by Meehl et al. (2003), solar forcing at the surface from the multidecadal increase of total solar irradiance over the first half of the twentieth century in the solar forcing reconstruction of Hoyt and Schatten (1993) was associated with increased evaporation from the tropical oceans, and that moisture was carried to the tropical low-level convergence zones (the summer monsoon regimes and ocean convergence zones) to strengthen the associated climatological precipitation regimes.
Meehl et al. (2003) noted that all the major tropical low-level convergence zones were affected by solar forcing in the model. Van Loon et al. (2004) showed for northern summer that the low-level convergence associated with the South Asian monsoon and the ITCZ in the tropical Pacific, in particular, were intensified during peaks in the DSO. Thus there was observational evidence to support coupled surface feedbacks that could amplify a relatively small solar signal, certainly much smaller than the multidecadal increase in the early twentieth century studied by Meehl et al. (2003).

Subsequently, van Loon et al. (2007) and van Loon and Meehl (2008) showed a similar response in observations for northern winter. The focus in those studies showed a large response in the intensification of the Pacific Ocean convergence zones, the ITCZ and SPCZ, to increased solar forcing at peaks in the DSO. These were shown to be most notable in the tropical Pacific at that time of year in part because of the large expanse of the air–sea interaction across the Pacific Ocean, as well as the nature of the air–sea coupling at that time of year occurring mostly over ocean regions in the Pacific. These same effects are occurring in other regions, but are most strongly and consistently seen in the Pacific in northern winter. Thus, in this article, we focus on the northern winter season and Pacific region as in van Loon et al. (2007). As noted earlier, what is new here is that we document a response directly comparable to the observations for the DSO, which has roughly half the globally averaged forcing of the multidecadal forcing spanning the first half of the twentieth century discussed in Meehl et al. (2003). Since there has been some question regarding the multidecadal amplitude of the total solar irradiance change in the first half of the twentieth century (e.g., Foukal et al. 2004), the DSO is much more firmly documented, albeit with smaller amplitude. Thus this is a more severe test of the climate system response to the solar forcing coincident with peaks of the DSO. Possible lagged responses shown in other studies will be discussed later.

4. Observed coupled response to solar forcing in the Pacific region

Before seeing if this mechanism is working in the models used in our study, we first examine the climate system response to solar forcing in the Pacific region in observations. Anomalies (computed relative to the nonpeak years) for SST and SLP for the peak solar years when these data are available (11 peak yr in the DSO: 1883, 1893, 1905, 1917, 1928, 1937, 1947, 1957, 1968, 1979, and 1989—from the NOAA ERSST dataset (available online at http://www.cdc.noaa.gov/cdc/data.noaa.ersst.html). (b) The average tropical rainfall anomalies (mm day$^{-1}$) for January–February (GPCP gridded precipitation dataset) in the solar peaks in 1979, 1989, and 2000, in comparison to all other years. Dashed line is the 6 mm day$^{-1}$ contour from the long-term mean climatology. (c) As in (a) but for the average anomalies of SLP (hPa) (Hadley Centre SLP dataset); shading indicates significance at or above the 95% level, indicating the relative magnitude of the anomalies compared to the noise. For further details regarding observed data sources, see van Loon et al. (2007).
February (DJF; the year being that of January 1883, 1893, etc.) show anomalously cold water in the central and eastern equatorial Pacific, as documented by van Loon et al. (2007), with the largest SST anomalies at about 1°C, about half the amplitude of typical cold events in the Southern Oscillation. The most notable features of the SLP anomalies for these 11 peak solar years are small-amplitude positive anomalies just south of the equator; negative SLP anomalies stretching across the subtropics west of Baja California; and positive, statistically significant anomalies over the North Pacific extending to western North America (van Loon et al. 2007). Though not statistically significant, as noted by van Loon et al. (2007), the SLP anomalies in the tropics are physically significant since they contribute to stronger trade winds and greater upwelling that help produce the negative SST anomalies in Fig. 1. Taking subsets of the 11 yr in the composite shown in Fig. 1 produces similar patterns, but trying to interpret differences in amplitude of the patterns based on relative size of the DSO peaks is beyond what can be justified with the limited sample. A subsequent paper shows that years on either side of the peaks in the DSO produce similar but diminishing amplitude patterns, and that La Niña or cold event years are fundamentally different from the DSO years (van Loon and Meehl 2008). Additionally, as discussed later, lagging a couple of years after the peak years in the DSO, there appears to be more of a warm event–like pattern and a general warming of the tropics (White et al. 1997, 1998, 2003; White and Tourre 2003; White 2006; White and Liu 2008).

The corresponding precipitation anomalies in Fig. 1b show a poleward-shifted and enhanced ITCZ and SPCZ (i.e., positive values somewhat poleward of the climatological positions of these features as indicated by the dashed lines in the figure) and reduced precipitation in the central equatorial Pacific. This was noted in van Loon et al. (2007) and also occurs during cold events in the Southern Oscillation. Note that because of dataset limitations these precipitation results are based on only the three solar peak years since 1979, and only the months of January and February are used because December 1978 rainfall data are not available.

The equatorially precipitation deficit in Fig. 1b and negative surface pressure anomaly in Fig. 1c are also reminiscent of cold (La Niña) events. This fact raises the possibility that the DSO composites could simply be a reflection of a few La Niña events that happened to fall during DSO peaks. But La Niña events are extremes in the Southern Oscillation and have about twice the amplitude compared to the signals associated with the DSO (van Loon et al. 2007). In the 11 solar peak yr considered, there was only one cold (La Niña) event (1989) and one warm (El Niño) event (1905). Note that in the convention of Kiladis and Diaz (1989), year 0 for these events would be 1988 and 1904, the years when the events begin in northern spring. We use the succeeding years, when the tropical events typically peak, to compare with the DSO index. If either is removed, the pattern remains the same (not shown). Therefore, this pattern is not caused by the Southern Oscillation. Rather van Loon et al. (2007) have invoked the mechanism proposed by Meehl et al. (2003), which is outlined in the previous section, to argue that the pattern is actually forced by the enhanced solar irradiance during peak DSO years. A subsequent paper (van Loon and Meehl 2008) contrasts La Niña events in the Southern Oscillation with the anomalies connected to the DSO discussed here and in van Loon et al. (2007).

5. Coupled response in the Pacific in the models

Turning to the model experiments described in section 2, Fig. 2 shows various atmospheric features that occur in peak solar years in the ensemble mean differences, solar peak years minus long-term climatology. For PCM, climatology is averaged from 1890 to 1999; for CCSM3, it is 1870–1999; 1883 is included as a solar peak year for CCSM3 but not for PCM since the PCM runs started in 1890. Thus in the CCSM3 there are 11 peak solar yr in each of the 5 ensemble members (or 55 solar peak yr contributing to the ensemble mean), and 10 for PCM (or 40 solar peak yr for the ensemble mean). Stippling in Fig. 2 designates areas where the multimember ensemble mean divided by the interensemble standard deviation exceeds 1.0, and is an indication of consistency across the model ensembles (e.g., Meehl et al. 2003). Computing anomalies by taking solar peak years minus solar minimum years yields similar patterns with somewhat larger and more consistent amplitudes (not shown). Here we show the anomalies compared to climatology to be consistent with van Loon et al. (2007).

Figures 2a,b show surface temperature anomalies (very comparable to SSTs over the ocean) for comparison to the observations in Fig. 1a. Both models simulate a negative anomaly in the equatorial Pacific, though the anomalies (up to 0.5°C) are about half the amplitude of the observed (up to about 1°C). The CCSM3 has lower-amplitude composite negative anomalies in this region compared to PCM, with a correspondingly less consistent response across the ensemble members (i.e., less stippling in the equatorial Pacific, as seen in Fig. 2b compared to Fig. 2a). The precipitation anomalies in the models (Figs. 2c,d) associated with those SST anomalies are also similar to the observed (Fig. 1b),
with somewhat of a poleward shift and an enhancement of precipitation in the ITCZ and the models’ version of the SPCZ that actually lies in a more zonal direction compared to the observations (e.g., Meehl et al. 2006b). The DSO peak pattern of precipitation (and associated convective heating anomalies) is associated with an area of anomalous high pressure in the North Pacific in both models (Figs. 2e,f) that is consistent across the model ensembles (as denoted by the stippling) and qualitatively similar to the observations (Fig. 1c), but again with about half the amplitude (about 3 hPa in the PCM and about 4 hPa in CCSM3, compared to nearly 6 hPa in the observations). Therefore, the models are simulating the patterns of variability that are observed to accompany variations in solar forcing in nature, but with about half the amplitude seen in the observations. Note that the magnitude of the CCSM3 ensemble mean anomalies of SST and precipitation are smaller than that in PCM, but the North Pacific SLP response is somewhat larger. As will be seen later, the teleconnec-
tion pattern depends on both the amplitude and the geographical placement of the convective heating anomalies in the tropics. Thus there is a certain model-dependence to the details of this response.

Given our analysis design, the natural conclusion to draw is that the anomalies of Fig. 2 represent the response of the models to the peak years of solar forcing. And given their similarity to nature, one would also like to conclude that the patterns of Fig. 1 correspond to the response of the natural system to the peak years of solar forcing. But one might question how a response of this amplitude could occur with a globally averaged solar forcing of only 0.2 W m\(^{-2}\) (Lean et al. 2005) during peaks of the DSO compared to solar minima. However, this typically cited value of 0.2 W m\(^{-2}\) is a globally averaged number for the forcing, and the actual amplitude of the DSO is 1–2 W m\(^{-2}\) at the top of the atmosphere at tropical latitudes (Lean et al. 2005). Therefore, in relatively cloud-free regions in the tropics, the actual amplitude of the solar forcing (total solar irradiance) at the surface can be considerably greater than the globally averaged number. In fact, in the model simulations the mean net solar radiation anomalies at the surface for peaks in the DSO are greater than 1 W m\(^{-2}\) in some of the relatively cloud-free areas in the subtropics northeast of Hawaii in the Northern Hemisphere and southeast of Tahiti in the Southern Hemisphere (Figs. 3a,b). Increases of net solar radiation at the surface greater than 2 W m\(^{-2}\) also occur in parts of the equatorial Pacific where there are anomalously cooler SSTs and correspondingly fewer clouds in peak years of the DSO.

The increases of net solar radiation produce greater energy input into the ocean surface, and this is translated, in the subtropical regions mentioned above, into increases of latent heat flux on the order of several W m\(^{-2}\) (Figs. 3c,d) and thus greater evaporation and low-level moisture. This contributes to decreases of SST in those regions (Figs. 3e,f), with boxes highlighting the subtropical regions where these coupled processes are important for the coupled response to increased solar forcing. The numbers in the boxes for Figs. 3e,f are the area-averaged net solar differences derived from Figs. 3a,b to highlight the magnitude of the changes in net solar at the surface for the peak years of the DSO.

There are a couple of factors that could produce such large regional solar forcing at the surface. First, there is relatively clean air in the solar peak years because the major volcanic eruptions of the twentieth century did not coincide with solar peak years (Lean et al. 2005). To investigate the possibility that the signal here could be related to just the volcanic forcing, an ensemble of four twentieth-century simulations with the PCM run only with time-varying volcanic aerosols (Meehl et al. 2004) was sampled in the same way as the simulations in Figs. 2 and 3, compositing only the peak solar years. If the volcanic forcing was causing the patterns in Figs. 2 and 3, then these simulations would show a similar signal. But the composite patterns from those simulations show a quite different response with little signal in the tropical Pacific (not shown). Therefore, it is likely that the patterns in Figs. 2 and 3 are actually the response coincident with peaks in the solar forcing.

Second, these subtropical areas are typically characterized by relatively fewer clouds on average than other tropical regions. Furthermore, cloud feedbacks—because of the large-scale descent in these regions associated with increases in the intensity of the Hadley and Walker circulations driven by the greater precipitation and stronger vertical ascent in the precipitation convergence zones (van Loon et al. 2007)—produce even fewer clouds and more incoming solar radiation at the surface during peak solar years (Figs. 3g,h). That is, increases of net solar at the surface in relatively cloud-free areas of the subtropics increase energy input into the surface, which is translated into increased latent heat flux and surface moisture that is carried into the ITCZ and SPCZ to strengthen those features. That correspondingly strengthens the north–south Hadley and east–west Walker circulations, as noted in van Loon et al. (2007). Thus there are stronger trades (Figs. 3c,d) that contribute to the increased latent heat flux and reduction of SSTs in the subtropics. The stronger trades in the equatorial Pacific are additionally associated with a dynamical response with increased upwelling that contributes to decreased SSTs. There is also increased atmospheric upper-level outflow from the ITCZ and SPCZ, and stronger subsidence in areas of the subtropics. Consequently, as noted above, this produces even fewer clouds (Figs. 3g,h) and more solar radiation makes it to the surface (shown schematically in Fig. 4). For example, the areas highlighted in Figs. 3e,f show consistent decreases in total cloud (Figs. 3g,h), with decreases typically 1%–3%. These decreases in total cloud, combined with the relatively small amount of volcanic aerosol in these peak solar years, produce changes of planetary albedo in these regions that amount to decreases of a little over 1% for peak solar years.

Therefore, regionally in the tropics where the sun is most directly overhead, the solar forcing at the surface can be considerably greater than what could be expected from a globally averaged perspective. This extra energy input to the surface triggers coupled air–sea interactions (see schematic in Fig. 4) that produce the patterns shown in Figs. 2 and 3 in the models. These model results are along the same lines as the earlier
analyses of Meehl et al. (2003) that addressed a model response to a much higher-amplitude and lower-frequency sustained variation in solar forcing than the DSO solar variations considered here. The response to peaks in the DSO in the models shown here also compares favorably to observations during the latter part of the twentieth century (Fig. 1).

6. Tropical forcing and the response over North America and the North Pacific

Because of the complex, coupled nature of the processes invoked by Meehl et al. (2003) and van Loon et al. (2007) to explain the apparent response of the climate system to solar variations, it is difficult to consider the effect of each process individually. One that can be considered in isolation is the influence of tropical precipitation anomalies on midlatitude atmospheric circulation. To quantify the apparent teleconnections between the tropical Pacific precipitation and SLP anomalies in the North Pacific in observations (Fig. 1) and the two models (Fig. 2), specified heating anomaly experiments with the Community Atmospheric Model version 3 are performed. Recall that the CAM3 is the atmospheric component to the CCSM3, and the CAM3 is run here at T42 resolution with climatologically varying SSTs.

To examine the effect of each major feature of the tropical precipitation associated with the DSO in nature and in our models, we find how local heat sources and sinks in the same positions as these precipitation features affect CAM3’s climate. Not all of these tropical precipitation features occur in both GCM experiments and in nature, so in some cases we consider precipitation anomalies that are prominent in just one or two of these situations.

In the first experiment a negative heating anomaly is placed in the CAM3 in the shape of an elongated ellipse centered on the equator at the date line. This heating has a central value of $-2^\circ$C day$^{-1}$ and decreases linearly with distance until it vanishes on an elliptical boundary with a 3000-km east–west semimajor axis and a 500-km north–south semiminor axis. In the vertical, the heating has the profile $\sin(\pi p/p_s)$, where $p$ is pressure and $p_s$ is surface pressure. This profile is an approximation to the vertical distribution of long-term heating anomalies that tend to occur in this model. This idealized heating is intended to correspond to the negative equatorial precipitation anomaly in about the same position in nature (Fig. 1b) and PCM (Fig. 2c). A second region with strong negative precipitation anomalies in nature and in both GCM experiments is a zonally elongated strip just south of the equator and centered east of the date line. Our second experiment approximates its influence with a negative elliptical heat source centered at 5$^\circ$S and 150$^\circ$W, with a 5000-km semimajor axis and a 500-km semiminor axis. This source has the same vertical structure and central value as that in the first experiment. In nature another area of large (positive) precipitation anomalies is in the SPCZ, so our third experiment places a positive heating anomaly in the form of a circle centered at 20$^\circ$S and 170$^\circ$E, with a radius of 1500 km, a central value of $+2^\circ$C day$^{-1}$, and the same vertical structure as the imposed heating in the other experiments. At their centers the magnitude of these idealized sources produces the same depth-averaged heating as about a 5 mm day$^{-1}$ precipitation anomaly. This is larger than the observed and modeled anomalies of Figs. 1b and 2c,d and represents a compromise between a desire to force CAM3 with a heating similar in magnitude to the prominent precipitation features associated with DSO peaks and the requirement that the CAM3 response be statistically significant. Each experiment was run for 20 yr, and results are shown for DJF as anomalies from a 100-yr control that uses climatologically varying sea surface temperatures.

As described by Webster (1972), one expects a midtropospheric tropical heat source to induce upward motion, low-level convergence, and upper-level divergence. Given the ready availability of moisture in the tropics, this circulation will encourage precipitation near the heat source. Results in Fig. 5, which displays the precipitation anomalies produced in the three CAM3 heating experiments, show just such a behavior. For the first experiment with just the negative heating anomaly in the equatorial Pacific, negative precipitation anom-
FIG. 4. Schematic showing processes involved with the Pacific air–sea response coincident with the peak years of solar forcing.
lies (with maximum values of about $-3 \text{ mm day}^{-1}$) result as expected in the area of the prescribed negative heating. The dynamical response of the atmosphere to the combined imposed heating and resulting collocated precipitation anomaly produces small positive precipitation anomalies in the ITCZ to the north and the SPCZ to the south (Fig. 5a), thus reproducing part of the pattern of enhanced precipitation in these regions in Figs. 1 and 2. Likewise, the negative heating in the second experiment generates negative precipitation anomalies south of the equator with maximum values of about $-2 \text{ mm day}^{-1}$ and positive anomalies just north of the equator and east of the date line (Fig. 5b). (It is because of the induced positive precipitation in these first two experiments that we have not included results for an experiment with positive heating anomalies at about $10^\circ$ and $10^\circ \text{N}$ as seen in Figs. 2c,d.) In the third experiment, for the positive heating anomaly in the SPCZ (Fig. 5c), as anticipated, there are large positive precipitation anomalies produced in that region (maximum values of about $8 \text{ mm day}^{-1}$), with indirectly forced negative anomalies in the ITCZ region to the north. We note that with the exception of the third experiment the positive feedbacks resulting from moist processes are weaker than the imposed heating. On the other hand, from the standpoint of statistical significance, the precipitation anomalies in each experiment are impressive, as can be ascertained from the fact that in the control integration the standard deviation of 20-yr DJF averages is nowhere greater than $0.5 \text{ mm day}^{-1}$ in the plotted domain.

Examining the idealized heating experiments further, we can estimate the influence on the concurrent midlatitude circulation anomalies in various regions when tropical anomalies are prominent during DSO peak years. For the negative equatorial anomalies in experiment 1, the midlatitude response shows a positive SLP anomaly in the North Pacific (Fig. 6a) with a maximum value of roughly 3 hPa. Similarly, the negative heating to the south of the equator in the second experiment induces a positive SLP feature of even stronger magnitude and similar position (Fig. 6b). For the positive anomaly in the SPCZ, Fig. 6c shows the response in the North Pacific to again be a positive SLP anomaly with maximum values of roughly 3 hPa. Similarly, the negative heating to the south of the equator in the second experiment induces a positive SLP feature of even stronger magnitude and similar position (Fig. 6b). For the positive anomaly in the SPCZ, Fig. 6c shows the response in the North Pacific to again be a positive SLP anomaly with maximum values of 4 hPa. Also note that when global SLP plots (not shown) are examined, the response to each of the two negative sources gives a positive enhancement of the Northern Annular Mode (NAM), with generally positive SLP anomalies stretching around the Northern Hemisphere midlatitudes and negative SLP anomalies over the arctic. This more zonal response of the NAM is also seen for years with

![Fig. 5. Precipitation anomalies (mm day$^{-1}$) for the specified heating anomaly experiments described in the text: (a) negative heating anomaly centered at 180° and equator; (b) negative heating anomaly centered at 5°S and 150°W; and (c) positive heating anomaly centered at 20°S and 170°E. Contour interval is 0.25 mm day$^{-1}$ in (a), (b) and 1.0 mm day$^{-1}$ in (c). The zero line is omitted. Blue areas are decreased precipitation; red areas are increased precipitation.](image-url)
enhanced solar forcing in observations (Huth et al. 2006; Bochníček and Hejda 2002; van Loon et al. 2007).

As a gauge of significance, shading in Fig. 6 indicates regions where anomalies have amplitudes greater than 1 standard deviation of 20-yr DJF averages in the control integration. Furthermore, by examining a 20-yr extension of the first experiment we have added to our confidence in the general characteristics of the forced features we have described. But this extension also suggests we are not warranted in comparing details like the exact position of the North Pacific pressure anomaly or its central value.

Because we have used idealized heating, these experiments cannot have exact correspondence to our earlier coupled solutions or to nature. For example, the amplitudes of the circulation anomalies are only about twice as strong as those in the coupled experiments even though the maximum combined imposed heating and heating from moist feedbacks is between 3 and 5 times as strong as the maximum heating implied by tropical precipitation anomalies in the coupled experiments. On the other hand, because the precipitation feedbacks in the idealized experiments were positive and largely confined to the regions of imposed heating, we did succeed in producing calculations in which we controlled the heating that is driving circulation anomalies. Therefore these experiments can be interpreted as indicating that it is likely that each of the major tropical precipitation features we have identified can contribute to the North Pacific SLP anomalous high observed in the coupled models and in nature during peak DSO years. Which of these three precipitation features is primarily responsible varies among the models and observations, but in all cases tropical Pacific rainfall anomalies make large contributions.

As noted earlier, previous studies have attempted to link a warm event–like response to solar forcing with a lag of a couple of years to the peaks in the DSO (White et al. 1997, 1998, 2003; White and Tourre 2003; White 2006). A subsequent study with a reduced complexity global coupled climate model with idealized solar forcing showed some of these same features with a similar lag (White and Liu 2008). However, in those studies the data were filtered and smoothed, and it was unclear if a similar signal would emerge using simple composites as in the present paper and van Loon et al. (2007). Preliminary analysis using lagged composites indicates a warm event–like response and tropical warming in observations as well as the PCM and CCSM3 (not shown) consistent with the earlier studies cited above. These results suggest that the cold event–like response shown here is the fast response of the Pacific climate system to the peaks of solar forcing, while a slower adjustment

**FIG. 6.** SLP anomaly (hPa) in response to the convective heating anomalies described in the text for (a). The negative heating anomaly centered at 180° and equator; (b) the negative heating anomaly at 5°S and 150°W; and (c) the positive heating anomaly centered at 20°S and 170°E. Contour interval is 0.5 hPa. Red areas denote increased SLP; blue areas are decreases. Shading indicates regions where the amplitude of an anomaly is larger than one std dev of 20-yr DJF averages in an AGCM control integration with climatological SSTs.
occurs with a lag of a couple of years with a warm event–like response. These analyses are being studied further and will be the subject of a subsequent paper.

7. Summary and concluding remarks

The observed composite signal in DJF of peak years in the 11-yr solar cycle documented by van Loon et al. (2007) is characterized by negative SST anomalies and reduced rainfall in the equatorial Pacific, enhanced precipitation in the ITCZ and SPCZ, and anomalous high pressure in the North Pacific extending to North America. A mechanism for this response involving solar radiation falling on the cloud-free regions of the subtropical Pacific and resulting coupled air-sea interactions in both the subtropics and tropics was proposed by Meehl et al. (2003). They based their hypothesized mechanism on their analysis of an AOGCM experiment that concerned the climate reaction to low-frequency solar variability that occurred during the first half of the twentieth century. That variability was of stronger magnitude and lower frequency than the variability in the 11-yr solar cycle. In the present study it is shown that this same mechanism is operative in peak years of the 11-yr solar cycle in simulations carried out with PCM and CCSM3. These experiments consisted of four- and five-member ensemble simulations, respectively, of the twentieth century with combinations of anthropogenic and natural forcings. Additional AGCM experiments with specified heating anomalies demonstrated that the solar maximum midlatitude pattern of anomalous high pressure in the North Pacific and North America is remotely driven by precipitation and associated convective heating anomalies in the tropical Pacific associated with the coupled air–sea response mechanism. Reduced precipitation on or south of the equator in the central Pacific can contribute to the North Pacific pressure feature as an enhanced precipitation in the SPCZ.

In work not described in the results sections of this article, we have also examined twentieth-century simulations in the global coupled climate models archived at the Program for Climate Model Diagnosis and Intercomparison (PCMDI) as part of the World Climate Research Programme (WCRP) Coupled Model Intercomparison Project Phase 3 (CMIP3) multimodel dataset (Meehl et al. 2007). We did this with the intent of determining whether the coupled air–sea response to solar forcing found in our PCM and CCSM experiments could be generalized to other models. But the different models in this archive used widely disparate solar forcing, so it was not feasible to extract a meaningful signal from the multimodel average response. It may be possible to examine some of the other models individually, depending on the nature of the solar forcing they used in the twentieth-century simulations, but this must be done on a model-by-model basis until modeling groups use a more uniform solar forcing for the twentieth century and is thus beyond the scope of the present study.

As noted previously, other investigations (e.g., Matthes et al. 2006; Shindell et al. 2006) have found that the response to solar forcing in models with interactive ozone chemistry in the stratosphere indicates that a mechanism linking ozone changes with solar forcing could be manifested in a similar way in the tropical Pacific to the present coupled air–sea mechanism (i.e., an enhancement of the climatological precipitation maxima in the SPCZ and ITCZ). Matthes et al. (2006) and Kuroda and Kodera (2002, 2004) have found that the stratospheric mechanism could also produce a positive phase of the northern annular mode similar to that evidenced as a response to convective heating anomalies in the Pacific (e.g., Fig. 6). Therefore, different mechanisms, one acting at the surface and one in the stratosphere, could react in the same sense to solar forcing, and thus combine to produce an even larger response in the troposphere to solar forcing than the one documented here. This possibility is consistent with our finding that the coupled air–sea mechanism acting alone only produces a response that is about half that observed. It remains for a model with ozone chemistry and high vertical resolution in the stratosphere to be coupled to a dynamic ocean model and to be run in long enough experiments to determine if this is a possibility.

The lagged response of the tropics (on the order of a couple of years from the peaks of the DSO) with more of a warm event–like pattern is noted to occur in preliminary composite analyses of the observations and both models, consistent with previous studies using filtered and smoothed data. We tentatively interpret the cold event–like pattern shown here to be the fast response of the Pacific climate system, with the warm event–like pattern a possible slower adjustment to the solar forcing. Further analyses of these responses in models and observations will be the subject of a subsequent paper.

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