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A decadal solar effect in the tropics in July–August

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Abstract

The availability of global gridded precipitation and outgoing long-wave radiation (OLR) data after 1978 makes possible an investigation of the influence of the decadal solar oscillation in the tropics during three solar maxima and two solar minima. The NCEP/NCAR reanalyses starting in the 1950s allows the inclusion of an additional two solar maxima and minima to look for consistency of response across a longer time period. In the northern summer (July–August), the major climatological tropical precipitation maxima are intensified in solar maxima compared to solar minima during the period 1979–2002. The regions of this enhanced climatological precipitation extend from the Indian monsoon to the West Pacific oceanic warm pool and farther eastwards in the Intertropical Convergence Zone of the North Pacific and North American Monsoon, to the tropical Atlantic and greater rainfall over the Sahel and central Africa. The differences between solar maxima and minima in the zonal mean temperature through the depth of the troposphere, OLR, tropospheric vertical motion, and tropopause temperature are consistent with the differences in the rainfall. The upward vertical motion is stronger in regions of enhanced tropical precipitation, tropospheric temperatures are higher, tropopause temperatures are lower, and the OLR is reduced due to higher, colder cloud tops over the areas of deeper convective rainfall in the solar maxima than in the minima. These differences between the extremes of the solar cycle suggest that an increase in solar forcing intensifies the Hadley and Walker circulations, with greater solar forcing resulting in strengthened regional climatological tropical precipitation regimes. These effects are as strong or even more pronounced when warm and cold extremes in the Southern Oscillation are removed from the analyses. Additionally, lower stratospheric temperatures and geopotential heights are higher with greater solar forcing suggesting ozone interactions with solar forcing in the upper stratosphere.

Keywords: Solar cycle; Solar response; Tropical precipitation; Stratosphere

1. Introduction

van Loon and Shea (1999, their Figs. 2b, 3b,c, 4a,b; 2000) pointed out that from 1958 to 1998 the 3-year running summer mean temperatures in the layer between 750 and 200 hPa on the Northern Hemisphere correlated well with the Decadal Solar Oscillation (DSO, this is a better term than 11-year solar cycle, which is neither 11-year nor cycle; the latter implies a fixed period) from 1958 to 1998. Previously, it had been noted that radiosonde stations in the tropics of the Northern Hemisphere showed a marked difference in the vertical distribution of temperature between maxima and

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minima of the DSO (van Loon and Labitzke, 1993, Fig. 3, 1994, Figs. 4–6; Labitzke and van Loon, 1993, Figs. 3–6), with the temperatures being higher in the troposphere and stratosphere and lower or little changed at tropopause levels in the solar maxima compared to the solar minima.

There are other indications from observations that solar forcing could affect the troposphere directly. In a study of reanalysis data in the latter part of the 20th century, Gleisner and Thejll (2003) showed similar positive temperature anomalies in the troposphere of the tropics/subtropics associated with increased solar input, consistent with changes in vertical winds and humidity. Gleisner and Thejll (2003) filtered out other forced signals (volcanoes and ENSO) to isolate a solar signal in the troposphere. Using a comparable analysis technique, Haigh (2003) arrived at similar conclusions. Upper ocean temperature and heat storage were shown to react directly to solar forcing, with higher sea-surface temperatures (SSTs) and greater upper ocean heat storage over most of the tropical oceans for greater solar forcing (White et al., 1997, 1998).

It was also demonstrated that solar forcing affects the stratosphere through the indirect dynamic result of interaction between ultraviolet radiation (UV) and ozone in the upper stratosphere (Balachandran et al., 1999; Shindell et al., 1999) such that higher stratospheric temperatures would occur with increased solar forcing, in agreement with the observational results of van Loon and Labitzke.

Our intention here is to follow up on the earlier van Loon and Labitzke results from radiosonde observations and use, as a hypothesis, their concept that "the vertical motion [in the tropics] might be stronger at the peaks of the solar cycle…the mechanism of the solar effect must include interannual variability of the tropical circulation on the timescale of the sunspot cycle" (van Loon and Labitzke, 1994). That is, the climatological Hadley and Walker Cells could be intensified by greater tropical precipitation because of increased solar forcing. Additionally, from the work of Balachandran et al. (1999) and Shindell et al. (1999), the solar forced signals in temperature and geopotential height should also be seen in the lower stratosphere, with higher temperatures and higher heights in the lower stratosphere owing to greater solar forcing by UV on upper stratospheric ozone.

We must rely on observational data sets of observed precipitation and outgoing long-wave radiation (OLR) that can be studied only for the last several solar cycles. However, the NCEP/NCAR reanalyses extend back to 1957 and allow inclusion of two more solar cycles. However, sampling is obviously an issue and conclusions must rely on the physical concepts in our hypothesis above to make a case for physical significance (Nicholls, 2001). Additionally, we will perform a Monte Carlo significance test. If we are successful in finding such process-oriented physical signals in the system, this will provide further evidence to the original observational results of van Loon and Labitzke, and added to by Haigh, Gleisner and Thejll, White and coauthors, Balachandran, Shindell and others, that solar forcing directly affects the climate system in a measurable way.

2. Temperature, rainfall, vertical motion, and OLR

Zonal mean temperature differences between the extremes in the DSO for the months July–August (JA) are shown in Fig. 1a, which cover the season and longitudes in the Pacific sector of the stations used by van Loon and Labitzke and use all the solar extremes in the period 1957–2002. The period extends van Loon and Labitzke's period by almost two solar oscillations. Temperatures are higher in the upper troposphere (more than 0.6 °C), little changed near the tropopause near 100 hPa, and higher again in the lower stratosphere (more than 1.2 °C) in agreement with the previously obtained results from radiosondes. The data in Fig. 1 is the NCEP/NCAR reanalyses (see Acknowledgments), whereas van Loon and Labitzke, as mentioned above, used observations from single tropical radiosonde stations. Comparing the longer record in Fig. 1a to the period after 1978 shows a similar pattern (not shown). Thus, the implication is that the precipitation pattern in Fig. 2a, solar max minus solar min, derived from the CMAP precipitation data described by Xie and Arkin (1996, 1997) which began in 1979, is similar for the longer period. The Xie and Arkin data are based on a combination of raingauge measurements, satellite estimates, and model calculations.

To assess statistical significance of the signals in this paper, we use a Monte Carlo technique involving random sampling with replacement (i.e. boot strapping; Diaconis and Efron, 1983) of the 23 JA values. Specifically, 100 random samples were used to form a distribution of t-values. The observed t-value was compared to the bootstrapped values. Since we are expecting a specific pattern based on physical arguments from our hypothesis, we test at the 90% confidence level.

Those areas exceeding 90% are shaded blue in Fig. 1 (and in figures to follow), and include the larger differences noted above in the lower stratosphere and upper troposphere.

Since we are computing composite differences from a relatively short sample, it is possible that the Southern Oscillation (SO) events could be aliasing the results. The SO has well-known effects on tropical tropospheric temperatures and precipitation regimes (e.g., van Loon and Shea, 1985; Rasmusson and Carpenter, 1982; Meehl, 1987). However, SST anomalies in SO events

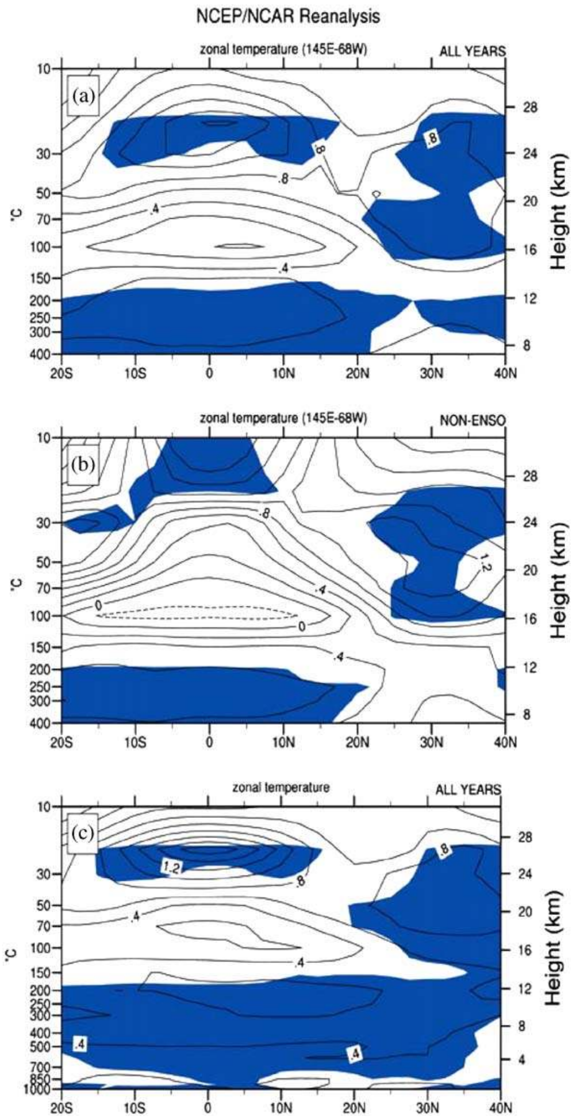

Fig. 1. (a) The zonally averaged temperature difference (°C) between five solar maxima (1957–1958, 1969–1970, 1980–1981, 1990–1991, 2000–2001), and four solar minima (1964–1965, 1975–1976, 1985–1986, 1995–1996) from 145°E eastward to 68°W. Between 400 hPa and 10 hPa in JA; (b) Same as (a) except excluding warm and cold SO extreme years 1957, 1969, and 1991 from the solar maximum composite, and 1965, 1976, and 1986 from the solar minimum composite (denoted here and in other figures as ''non-ENSO''); (c) same as (a) except for full zonal average from surface to 10 hPa, 20°S to 40°N. Blue shading indicates areas exceeding the 90% confidence level from the Monte Carlo test described in text.

act to shift regions of anomalous precipitation. For example, during warm events in the SO, the western equatorial precipitation maximum shifts eastward, and positive precipitation anomalies tend to occur in the central equatorial Pacific with suppressed precipitation over the south Asian monsoon region. For the solar response, we are arguing that precipitation regimes do not shift but are strengthened in place with greater solar forcing. If we remove the warm and cold SO extremes (see discussion of how these years are defined in van Loon et al., 2003) from the composites (Fig. 1b), the features of the differences in Fig. 1a and b are very similar in terms of patterns and magnitudes, indicating that the SO is not biasing the results for the DSO. We will include similar comparisons throughout the rest of the paper to illustrate that the DSO signals seen here are not biased by the SO, and that resampling retains the same basic features of the tropical tropospheric response to the DSO.

The warmer troposphere and cooler or little changed tropopause levels at low latitudes in solar maxima in comparison with the minima are likely owing to warming in the troposphere through release of latent heat in rising air of the tropical convergence zones and diabatic cooling of the tropopause levels in the upward motion (van Loon and Labitzke, 1994). The Hadley and Walker circulation cells might therefore be affected by the DSO such that most of the climatological precipitation maxima in the tropics could be stronger during solar maxima than in the minima. In fact, a plot of zonal mean temperature differences averaged across all longitudes for solar max minus solar min from 20°S to 40°N, extending to the surface (Fig. 1c), shows statistically significant higher temperatures throughout the depth of the troposphere, indicating that these effects are global and involve the entire troposphere and also the lower stratosphere.

The distribution of the difference in rainfall between the solar maxima and minima after 1978 (Fig. 2) shows traits related to the main features of the tropical circulation that are similar for all solar max minus solar min years (Fig. 2a) as well when SO years are not included (Fig. 2c). There is a near-equatorial dipole between the Indian Ocean and the western Pacific warm pool, with higher precipitation over the warm pool, and a belt of positive values stretching across the tropical Pacific and Atlantic Oceans, broken by negative values in South America. The positive values over the warm pool extend northwestward to the South Asian monsoon, and there is thus another, but weaker dipole, this one being north–south, between southern Asia and the equatorial Indian Ocean. The Monte Carlo test identifies areas of enhanced precipitation in the tropics as being statistically significant, with the suppressed precipitation regions in the Indian Ocean significant as well (Fig. 2b). An Indian monsoon rainfall index (Meehl and Arblaster, 2002), defined as area-averaged rainfall over land points in the region 5°N–40°N, 60°E–100°E, shows the solar maxima minus solar minima composite to be $+0.13\,\mathrm{mm\,day^{-1}}$, indicating a slight enhancement of

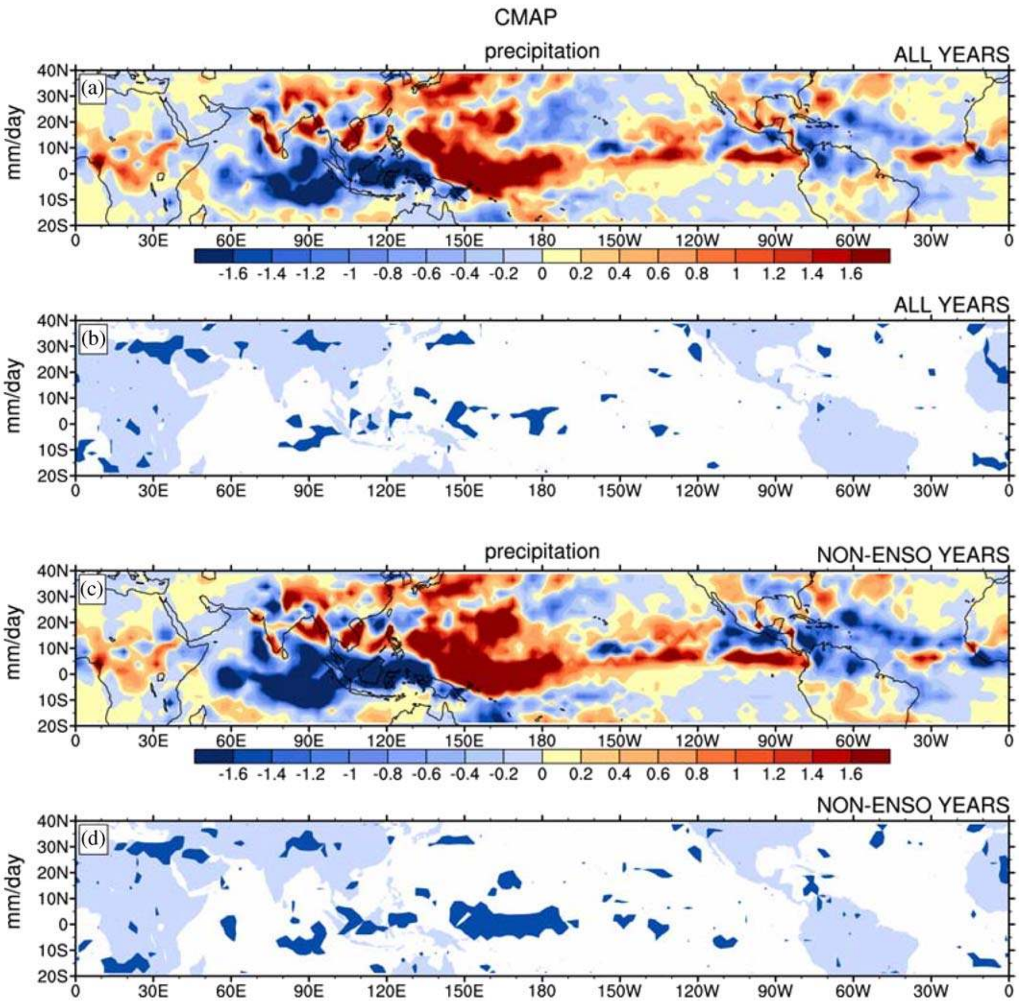

Fig. 2. (a) Difference in rainfall (mm day^{-1}) for the three solar maxima after 1979 minus the two solar minima, JA average; (b) blue shaded areas exceed the 90% confidence level from Monte Carlo test described in the text; (c) same as (a) except excluding SO extremes; (d) same as (b) excluding SO extremes.

area-averaged monsoon precipitation during solar maxima. The interannual standard deviation of this index is 0.38 mm day^{-1}. For the warm minus cold event composite differences calculated for this Indian monsoon index, the SO difference shows suppressed monsoon precipitation with a value of -0.16 mm day^{-1}, which is consistent with previous results for monsoon–SO connections. This can be compared to the value noted above for enhanced monsoon precipitation during solar maxima with a monsoon index difference of $+0.13$ mm day^{-1}. Thus, the SO gives a fundamentally opposite response of the monsoon compared to solar forcing, and results in somewhat larger areas of positive precipitation anomalies over south Asia for the non-SO years in Fig. 2c. Also note the more coherent areas of significance in the western equatorial Pacific in the non-SO years (Fig. 2d). This is due to the fact that during SO warm extremes, as noted above, the areas of enhanced precipitation in the Pacific tend to occur in the central equatorial Pacific (van Loon et al., 2003). Thus, by removing those years from the composites, the positive precipitation maximum in the western equatorial Pacific is stronger and more statistically significant.

The west African monsoon in Fig. 2 strengthens into the Sahel in solar maxima compared to solar minima, shown by positive precipitation differences near 15°N and negative anomalies near the coast at about 5°N, with positive precipitation differences in the climatological precipitation maximum covering central Africa from about 5°S to 5°N. There is also an intensification of the North American Monsoon, with significant positive precipitation anomalies over northern Mexico and the southwestern US The largest magnitude precipitation differences in Fig. 2 are on the order of 1.5 mm day^{-1}, roughly 15% of the climatological values in those areas. These are relatively small signals, but they are physically consistent with the hypothesis posed earlier, since Fig. 2 shows an intensification of the Indian monsoon, the African monsoon, the North American Monsoon, and the western Pacific climatological precipitation maximum. Negative precipitation differences on the coast of west Africa and over the Indian Ocean are typically associated with strong land-based monsoons (e.g. Meehl et al., 2003) and, for the latter, strong convection from the western Pacific warm pool region which could be expected to suppress Indian Ocean precipitation through the western side of the Walker circulation, termed the Western Walker Cell by Meehl and Arblaster (2002).

Though the difference patterns of precipitation in Fig. 2 are spatially coherent and are consistent with the hypothesis that enhanced solar forcing intensifies climatological tropical precipitation regimes in comparison to solar minima, we need to seek other diagnostics to check for consistency among rainfall, vertical motion, and OLR.

Changes in the Western Walker Cell can be illustrated directly by looking at the difference in the vertical motion (omega) from the NCEP/NCAR reanalyses between solar maxima and minima for all years (Fig. 3a), and excluding SO years (Fig. 3b), from the surface to 100 hPa, averaged between 5°S and 8°N, from 70°E eastward to 90°W. Many of the largest differences exceed the 90% confidence level from the Monte Carlo test and are consistent with the differences in the rainfall in Fig. 2. Namely, the west–east precipitation dipole between the equatorial Indian and western Pacific Oceans in Fig. 2 is present in the differences of vertical motion in Fig. 3 as well (positive values denote

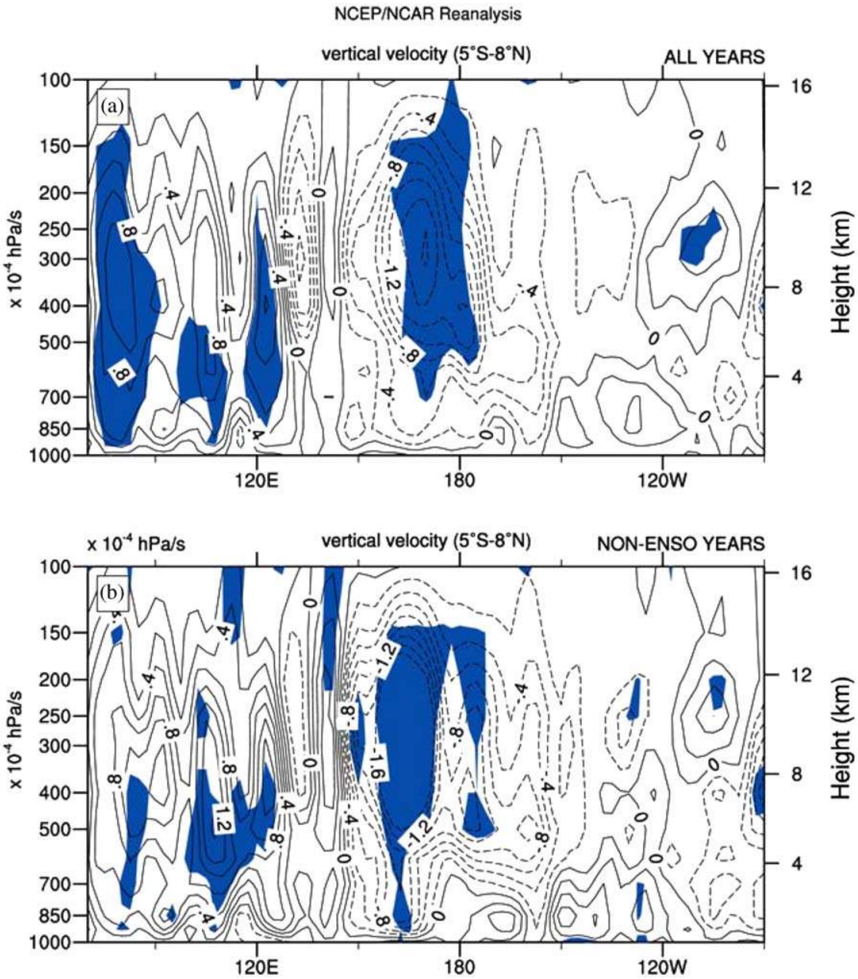

Fig. 3. (a) The difference in vertical motion as represented by omega ($\times 10^{-4}$ hPa s^{-1}) between the solar maxima and minima after 1957 noted in Fig. 1, 70°E eastward to 90°W, averaged between 5°S and 8°N, JA; (b) same as (a) except excluding ENSO years. Blue shaded areas exceed the 90% confidence level.

anomalous downward motion, and negative values stronger anomalous upward motion). In the region of lower rainfall over the equatorial Indian Ocean, the vertical motion differences in the DSO are positive such that there is anomalously downward motion in the solar maxima compared to solar minima. Farther east, in the equatorial western Pacific Ocean, there is stronger upward motion (anomalous negative values) in areas of higher precipitation in the years of high solar activity. Thus, the Western Walker Cell is anomalously strong, with enhanced precipitation, anomalous upward vertical motion, and lower tropopause temperatures over the western equatorial Pacific and ITCZ, with suppressed precipitation, anomalous sinking motion, and higher tropopause temperatures over the equatorial Indian Ocean in solar maxima compared to solar minima.

It is important to note that the vertical motion values in Fig. 3 are independent of the precipitation results in Fig. 2, and show a consistent and significant response to solar forcing. Another observed quantity, the OLR, yields further evidence for the strengthened Western Walker Cell. In the region in Fig. 2 where the rainfall is significantly higher in the solar maxima (western equatorial Pacific), the anomalies of OLR are significantly negative for all solar max minus solar min years (Fig. 4a) as well as when ENSO years are excluded (Fig. 4c). That is, there is more intense deeper convection with higher and significantly colder cloud tops in association with greater precipitation (Fig. 2) and enhanced vertical motion (Fig. 3), and conversely in the region of lower rainfall over the equatorial Indian Ocean. The significance of the OLR differences in the non-ENSO years (Fig. 4d) increases over areas of the west African monsoon and the North American Monsoon.

Thus, the preceding evidence points to a Western Walker Cell that is stronger during solar maxima than in solar minima, with enhanced precipitation, stronger upward motion, and negative OLR differences over the

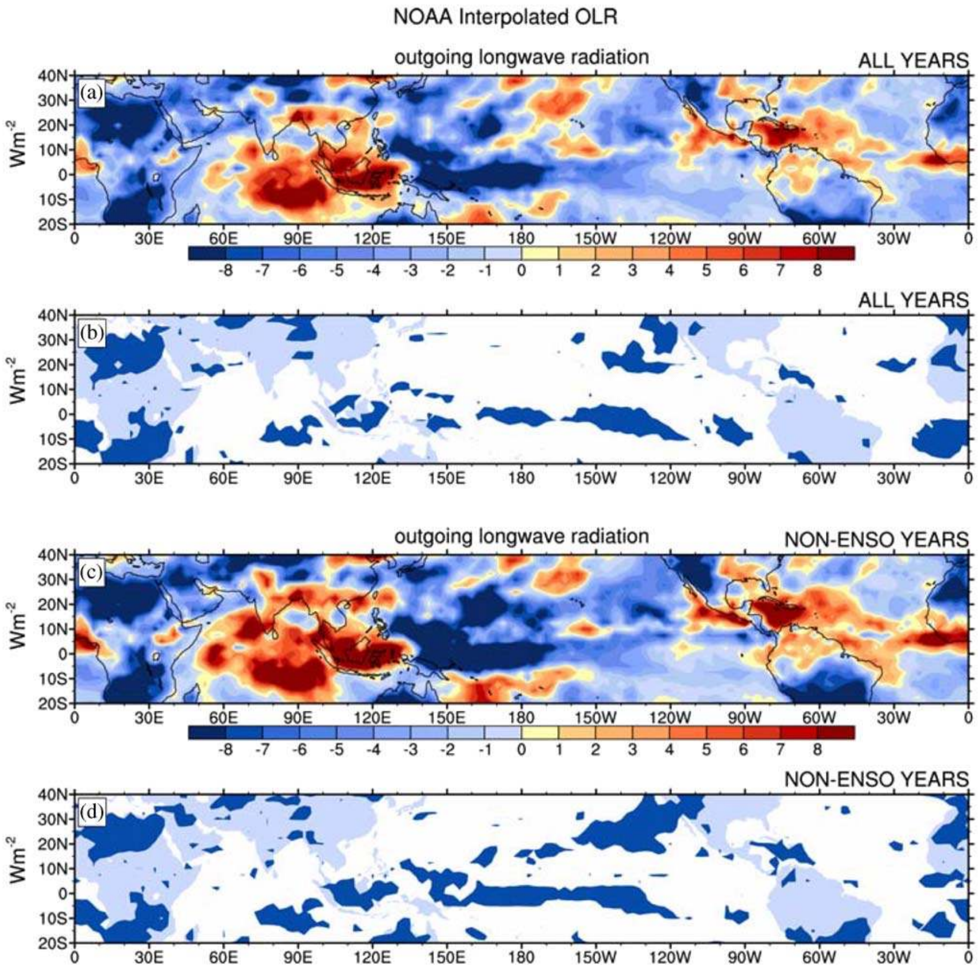

Fig. 4. The difference in OLR (W m^{-2}) between solar maxima and minima after 1979; (b) blue shaded areas exceed the 90% confidence level; (c) same as (a) except excluding SO extremes; (d) same as (b) except excluding SO extremes.

western equatorial Pacific, with suppressed precipitation, stronger sinking motion, and positive OLR differences over the equatorial Indian Ocean, and strengthened Indian, west African, and North American monsoons.

It was hypothesized earlier that, in addition to the differences in the Walker circulations noted above, the Hadley circulation should also be stronger in solar maxima than in solar minima. To document these changes, the zonally averaged differences in vertical motion, solar maxima minus minima since 1957, between 15°S and 35°N for all solar max minus solar min years (Fig. 5a), and excluding SO extremes (Fig. 5b) show statistically significant stronger upward motion (negative anomalies) in the solar maxima between about 4°N and 17°N, bordered by significantly stronger downward motion (positive anomalies) to the south and north. The zone of stronger upward motion in the solar maxima coincides with predominantly higher rainfall in the solar maxima in the south Asian monsoon, west African monsoon, North American monsoon, and tropical western Pacific (Fig. 2). These latitudes contain the upward branch of the Hadley circulation cell. Thus, the vertical motion independently agrees with the precipitation results that the Hadley Cell is strengthened in the peaks of the DSO in comparison with the minima.

To a first order, as noted in Fig. 1 the temperature of the tropopause in conditions of stronger convection tends to be lower than normal. To illustrate further such changes in the Western Walker Cell with greater solar forcing, the differences in the temperature of the tropopause (Fig. 6a), which range from about +1 °C to –1 °C (somewhat less than one standard deviation), are consistent with the rainfall differences in Fig. 2. Located above the higher rainfall in the solar maxima in the regions of the western tropical Pacific and south Asian monsoon, there are lower tropopause tempera-

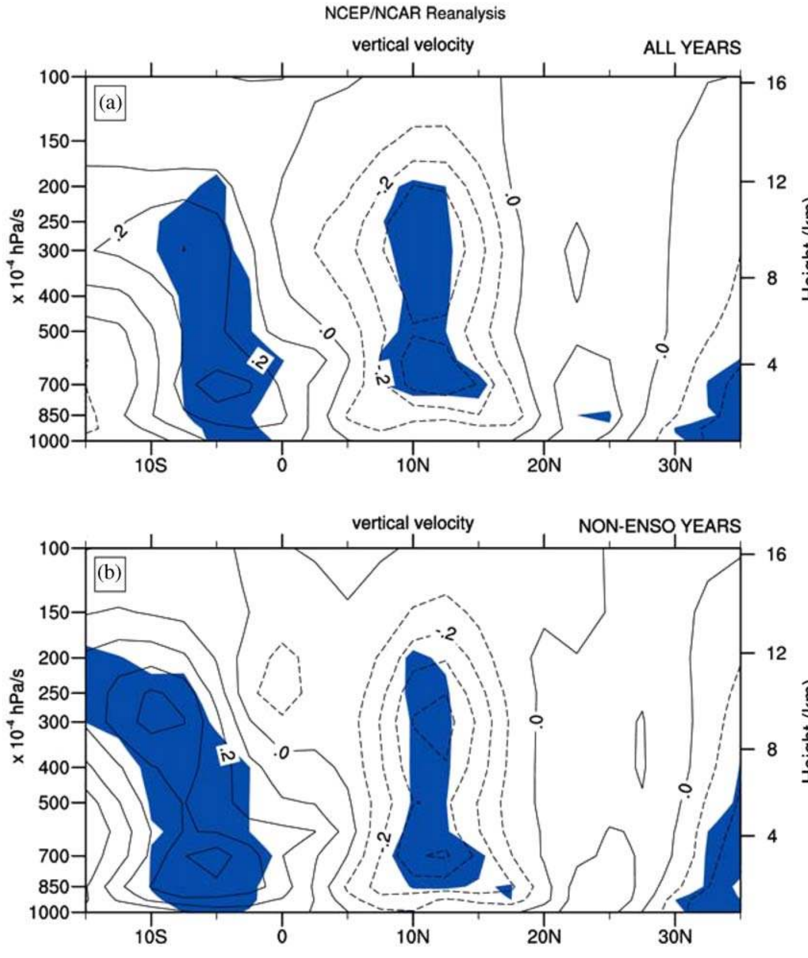

Fig. 5. (a) The zonally averaged vertical motion difference as represented by omega ($\times 10^{-4}$ hPa s^{-1}) between solar maxima and two minima after 1957; 15 S to 35°N, JA; and (b) same as (a) except excluding SO extremes. Blue shaded areas exceed the 90% confidence level.

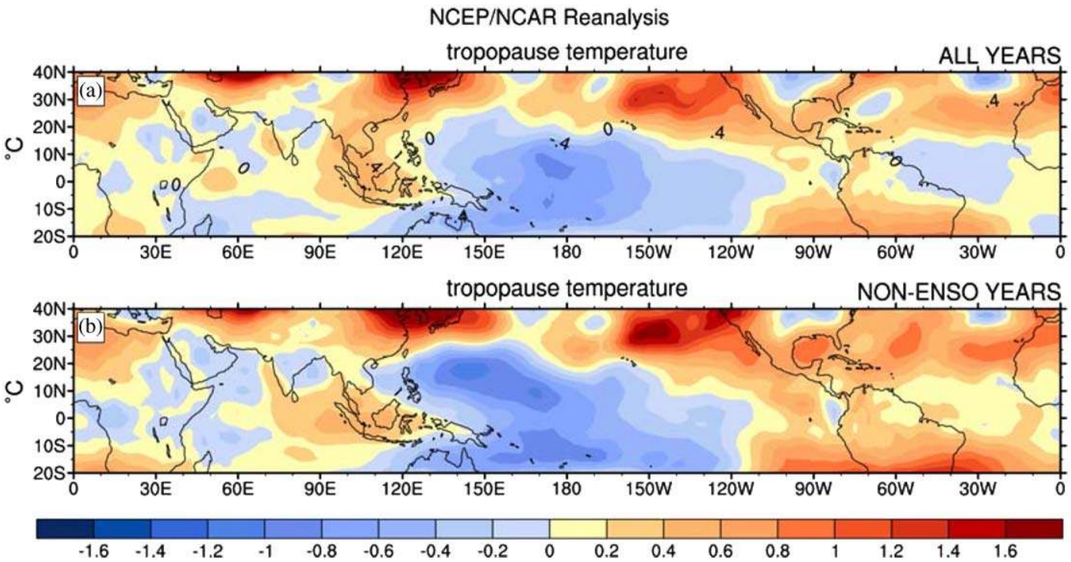

Fig. 6. (a) The difference in the temperature of the tropopause (°C) between the five solar maxima and four minima after 1957 in Fig. 1 and (b) same as (a) except excluding SO extremes.

tures compared to the minima, owing to diabatic cooling over the stronger convection. Above the regions of lower rainfall in the solar maxima in the equatorial Indian Ocean, the tropopause is warmer than in the minima. When warm and cold SO extremes are removed from the solar max minus solar min composites (Fig. 6b), the correspondence with precipitation anomalies in Fig. 2 is even stronger in the western Pacific with negative values extending to the Philippines and a larger area of negative values over the south Asian monsoon.

3. The stratosphere

In Fig. 1 the temperature difference from the reanalyses in the stratosphere between maxima and minima in the DSO had peaks in the lower stratosphere in the northern subtropics , which was known from earlier work by van Loon and Labitzke (e.g., van Loon and Labitzke, 1999, their Fig. 5; van Loon and Labitzke, 2000, their Figs. 8 and 9; Labitzke and van Loon, 1999, their Fig. 6.8). In Fig. 7a the pattern of temperature differences in the tropical stratosphere for solar maxima minus minima is the same after 1979 as in the above-quoted works by van Loon and Labitzke (cf. Figs. 1 and 7a). Near 100 hPa the temperature gradient in the DSO differences in Fig. 7a is directed from higher temperature differences in the outer tropics toward lower temperature differences on the equator. These gradients reverse with height, with statistically significant positive temperature differences (solar maxima minus minima) greater than $+1.6\,^{\circ}$C above 30 hPa being higher on the equator than to the south and north. In other words, the thermal wind differences reverse direction with elevation between 100 and 10 hPa. A similar vertical distribution happens in the temperature differences between the SO extremes (Fig. 7b). But the positive temperature differences between the six warm and five cold extremes in the SO after 1979 in Fig. 7b are less than half that of the solar differences in Fig. 7a and none are significant. In addition, the differences are negative over a larger part of the tropopause region near 100 hPa and again above about 20 hPa for the SO extreme differences in Fig. 7b, whereas the differences for the solar extremes are positive throughout the depth of upper troposphere to middle stratosphere. These substantial differences suggest that, on average, the influence of the DSO on the temperature of the lower to middle stratosphere is stronger than that of the SO during the period considered.

Consequently, the height differences in Figs. 8a and b of the zonally averaged geopotential height, in the DSO and the SO extremes, respectively, are substantially different, with the values for the DSO being appreciably larger, becoming statistically significant and reaching a factor of four larger near 20 hPa. As in the instance of the temperature, the gradients reverse with elevation in both types of extremes, thus the zonal geostrophic wind changes direction with height in both. Therefore, both the DSO and the SO influence the equatorial Quasi-Biennial Oscillation in JA. The DSO and the Quasi-Biennial Oscillation are linked in the northern winter too—as first demonstrated by Labitzke (1987), and later confirmed by Salby and Callaghan (2000).

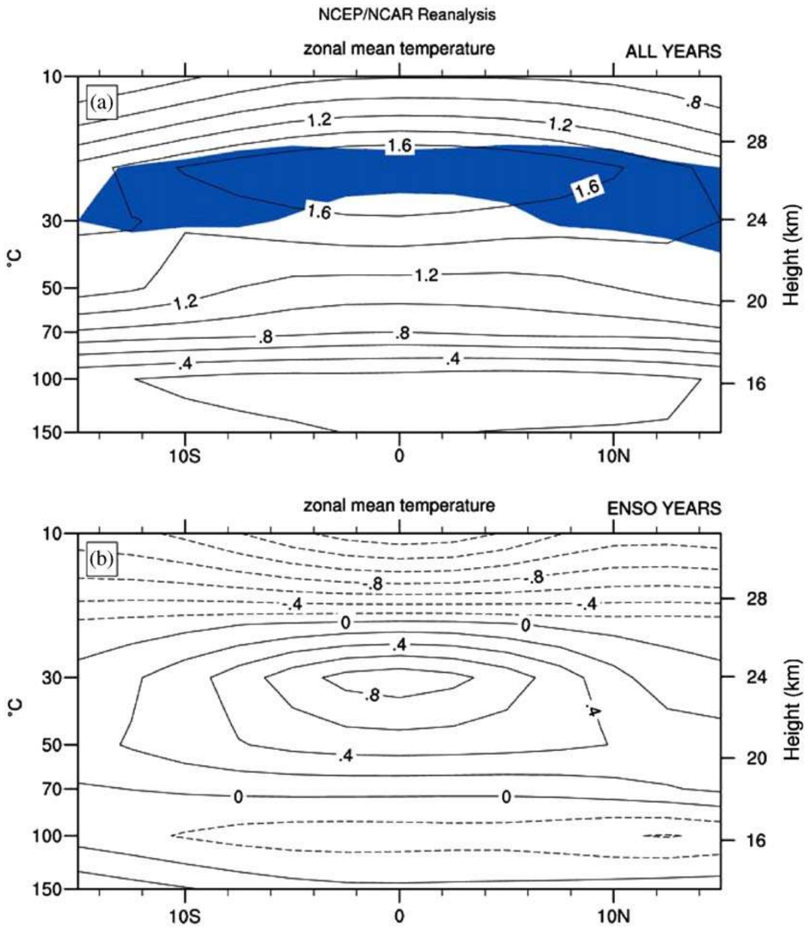

Fig. 7. (a) The difference in zonally averaged temperature (°C) between three solar maxima and two minima after 1979; from 15°S to 15°N, and 150 to 10 hPa, JA. (b) The same as (a), but for six warm events and five cold events in the SO after 1979, JA. Blue shaded areas exceed the 90% confidence level.

4. Discussion

Evidence presented so far from observations tends to confirm the hypothesis that enhanced solar forcing in the DSO intensifies climatological precipitation regimes in the tropics when compared with solar minima. The question remains concerning why this happens. Though we cannot resolve that question in this paper, it is worth noting that in a somewhat different context from a modeling study, Meehl et al. (2003) performed an analysis of ensemble simulations with a global coupled model and noted that an increase in solar forcing over the first part of the 20th century increased tropical SSTs and intensified climatological regional precipitation regimes in the tropics, resulting in stronger regional Hadley and Walker circulations. This was consistent with the earlier van Loon and Labitzke results from radiosonde observations for DSO differences in the late 20th century. The reason for this result in the model was the spatial differentiation of the solar input which affected the surface most directly in the clear sky areas of the subtropics. This had two effects. First, in the monsoon regions there was a greater heating of the subtropical land areas, stronger meridional temperature gradients, and stronger monsoons with greater solar forcing. Second, over ocean, the greater solar input heated the ocean surface, produced more evaporation, and an increased source of low-level moisture which was then carried by the low-level winds into the monsoon and oceanic convergence zones where it fueled more vigorous convection and stronger climatological precipitation regimes.

Though suggestive, several caveats must accompany the idea that a similar mechanism may be operating in the observed system on the time scale of the DSO. First, the time period under consideration for the model is half a century, and the model has sufficient time to respond to the low-frequency variation of solar forcing on that

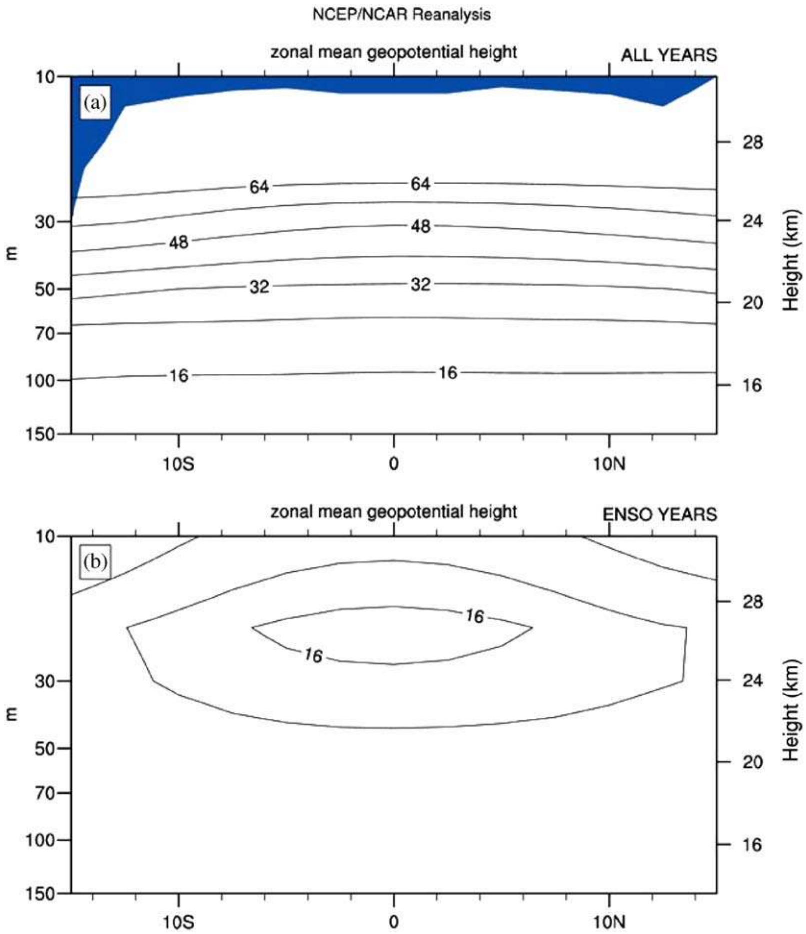

Fig. 8. (a) The same as in Fig. 7a, but for geopotential height (m). (b) The same as in Fig. 7b, but for geopotential height (m).

time scale. This is particularly relevant for the time scale of the ocean response. Second, the model responds most strongly to early 20th century changes of solar forcing when combined with the increases of greenhouse gases during that time period. Third, the modeling study uses only changes in total solar irradiance to force the model, with no wavelength dependence or other effects.

However, even with these caveats, the basic idea that solar forcing is spatially differentiated, with greatest solar input in the regions of the relatively cloud-free subtropical high-pressure regions could still be relevant for the time scale of the DSO. This enhanced energy input into the tropical oceans could be translated, as in the model, into higher SSTs (as noted in observations by White et al., 1997, 1998), greater evaporation and moisture transport into the monsoons, and tropical convergence zones, thus enhancing tropical precipitation. An indication that this could be taking place is given in Table 1. Meehl et al. (2003) argued that in the relatively cloud-free regions in the oceanic subtropics where the trade winds pick up moisture that is carried to the convergence zones, surface energy flux values should show the solar forcing signature with greater energy input at the surface during periods of greater solar forcing. This was the case in the model for the longer time scale in early 20th century period, and this is also the case for the DSO time scale in Table 1.

Using surface flux data from the reanalyses from 1957–2002, after Meehl et al. (2003), we define four relatively cloud-free regions in the tropics in the regions of the trade winds as denoted in Table 1. For solar max minus solar min, differences in net solar radiation at the surface are all positive, ranging from $+0.53$ to $+2.39\,\mathrm{W\,m^{-2}}$. This indicates that there is more energy input to the ocean surface in these regions in solar max compared to solar min. This excess energy is, for the most part, converted into greater evaporation as represented by increases of latent heat flux in three out of the four areas, with those increases ranging from $+0.68$ to $+5.21\,\mathrm{W\,m^{-2}}$. These changes are consistent

Table 1
Area averaged surface energy component differences for solar max minus solar min ($W\,m^{-2}$)

Areas	Net solar	Latent heat flux
Eq.–15°S, 60°E–100°E	+1.32	+3.37
10°N–30°N, 180–150°W	+1.79	+0.68
Eq.–20°S, 150°W–90°W	+0.53	−1.43
Eq.–20°S, 30°W–10°E	+2.39	+5.21

with the mechanism in Meehl et al. (2003) such that greater solar input into the relative cloud-free moisture collection regions of the tropics is translated into increases of evaporation and that moisture is then carried by the trade winds to the regions of low-level convergence over the monsoon and ocean convergence zones to cause increases of tropical precipitation in those regions in solar max compared to solar min.

Thus, though not conclusive, the present results suggest that the mechanism proposed by Meehl et al. (2003) for the climate system response to solar forcing in the troposphere acting over the early 20th century could also work for the DSO time scale in the observations as noted in the hypothesis posed earlier from the work of van Loon and Labitzke. The changes in the stratosphere also shown in this paper point to the likely importance of stratospheric ozone for the response there as suggested by Shindell et al. (1999) and Balachandran et al. (1999). These contribute to changes in tropopause height which could increase or reduce the height to which convection from the troposphere could penetrate.

5. Conclusions

1. In the tropics, the largest rainfall and tropopause temperature differences between maxima and minima in the DSO in northern summer are associated with features inherent in the tropical circulation, such that tropical climatological precipitation regimes are intensified with increased solar forcing as first hypothesized by van Loon and Labitzke (1993) from radiosonde observations.
2. The differences in the vertical motions and outgoing long-wave radiation associated with these rainfall differences point to the DSO enhancing the meridional Hadley and the zonal Walker circulation cells with greater solar forcing.
3. In the tropical stratosphere the pattern of the differences in temperature between the solar extremes are the same in the period when rainfall data were available—from 1979 and on—as when the solar extremes are included from the period before, 1957–1978.
4. The difference between the effect of the DSO and that of the SO on tropical precipitation and in the tropical stratosphere is substantial in the period analyzed, indicating that the effects of the two on the climate system are fundamentally different. Thus, removal of warm and cold SO extremes from the DSO analyses does not change the result.
5. The large temperature differences between solar maxima and minima in the stratosphere, along with the consistent geographical patterns of tropical rainfall, vertical motion, tropopause temperature, and OLR suggest that the response of the climate system to solar forcing likely consists of a combination of dynamical-radiative air–sea coupling (Meehl et al., 2003), such that increased solar forcing produces higher tropical SSTs (White et al., 1997, 1998), intensified climatological precipitation regimes involving the monsoons and the oceanic precipitation convergence zones, and interactions in the upper stratosphere between ozone and UV that indirectly produce warming of the tropical upper troposphere and lower stratosphere (Shindell et al., 1999; Balachandran et al., 1999).

The next step is to combine aspects of the Shindell et al. ozone mechanism in a global coupled model as in Meehl et al. (2003) to help focus on the role of these mechanisms on the DSO time scale. This work awaits a global coupled model with a better-resolved stratosphere and prognostic ozone. Such a model is currently under development and such experiments are a high priority for future studies of this problem.

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