

Solar Signals in Sea Level Pressure and Sea Surface Temperature

Indrani Roy

Supervisor: Prof. Joanna D. Haigh

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Abstract

We investigate solar cycle signals in 150 years of Sea Level Pressure (SLP) and Sea Surface Temperature (SST) data, using multiple regression analysis. We detect a solar signal in both SLP and SST in the North Pacific during DJF, similar to that found by Van Loon et al. (2007) but of smaller magnitude. We do not, however, identify the signal they found in the tropics. Our results do not support mechanisms for a solar influence on climate directly involving tropical SSTs.

We have used different reconstructions of total solar irradiance to investigate the sensitivity of the results. The series of Krivova & Solanki and Foster give similar results to those acquired using sunspot number but the Hoyt & Schatten solar index sometimes produces different results because of mixing of the solar signal with a long-term trend.

Using different approaches Labitzke and van Loon (1992) and Camp and Tung (2007), arrived at different results for a solar influence on winter stratospheric polar temperatures and its relationship to the quasi-biennial oscillation (QBO) in tropical stratospheric zonal winds. We show that these differences appear different largely because of their choices of QBO height. We also show that the effect of the QBO (30, 40 or 50 hPa) combined with solar activity reveals a clear signal in polar annular modes expressed in SLP.

We show that the nature of ENSO was different before 1950s (and after 1997): this may affect any solar influence. Other authors have suggested that tropical circulations were different during the intervening period. Such observation may have implications relating to the sun, tropical circulation and climate change.

During 1958-1997, omission of ENSO from regression gives false warming (cooling) signal of higher (lower) solar on SST in tropics. Such analysis, accompanied by our observation that the years of peak annual sunspot number used by van Loon et al. (2007) generally falls a year or more in advance of the maximum of the smoothed DSO, provides coherence to some apparently conflicting findings.

Finally, an atmosphere-ocean coupling process, (mainly involving the Pacific Ocean) is proposed to account for the solar influences. This coupling appears to be disturbed during the later half of the 20th century, probably due to climate change.

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Chapter 1: Background

1. Background

The main purpose of this thesis is to understand the 11 year solar cycle variability in Sea Level Pressure (SLP) and Sea Surface Temperature (SST). The first part of this chapter provides some background on climatology, the general circulation of the atmosphere and climate variability. The second part discusses previous studies concerning solar climate links.

1.1. Climatology, General Circulation and Climate Variability

In this subsection we describe the climatology of SST and SLP followed by a description of the general circulation and finally a general overview of different forms of climate variability.

1.1.1. Climatology: SLP and SST

Here, we discuss briefly the climatology of SLP and SST during two different seasons: the boreal winter and summer.

The mean SLP during boreal winter (here represented by January) differs to that from summer (here represented by July) and illustrated in Fig. 1.1.1a and b respectively. The seasonal variations in SLP are most apparent in the NH. During winter the high-latitude oceans are characterized by low-pressure centers with the Aleutian Low (AL) and Icelandic Low centered in the northern margins of the Pacific and Atlantic Oceans, respectively; whereas, a high-pressure center lies over Asia. During summer the land-sea pressure contrast is reversed in midlatitudes, with the highest pressures over the oceans and the lowest pressures over the land areas. It is seen from that figure that most high pressure regions persist throughout the year. Although, the Pacific High (PH) and Bermuda High are weaker and displaced farther south in the winter than summer. The dominant low-pressure feature during NH summer is centered over Asia at about 30°N

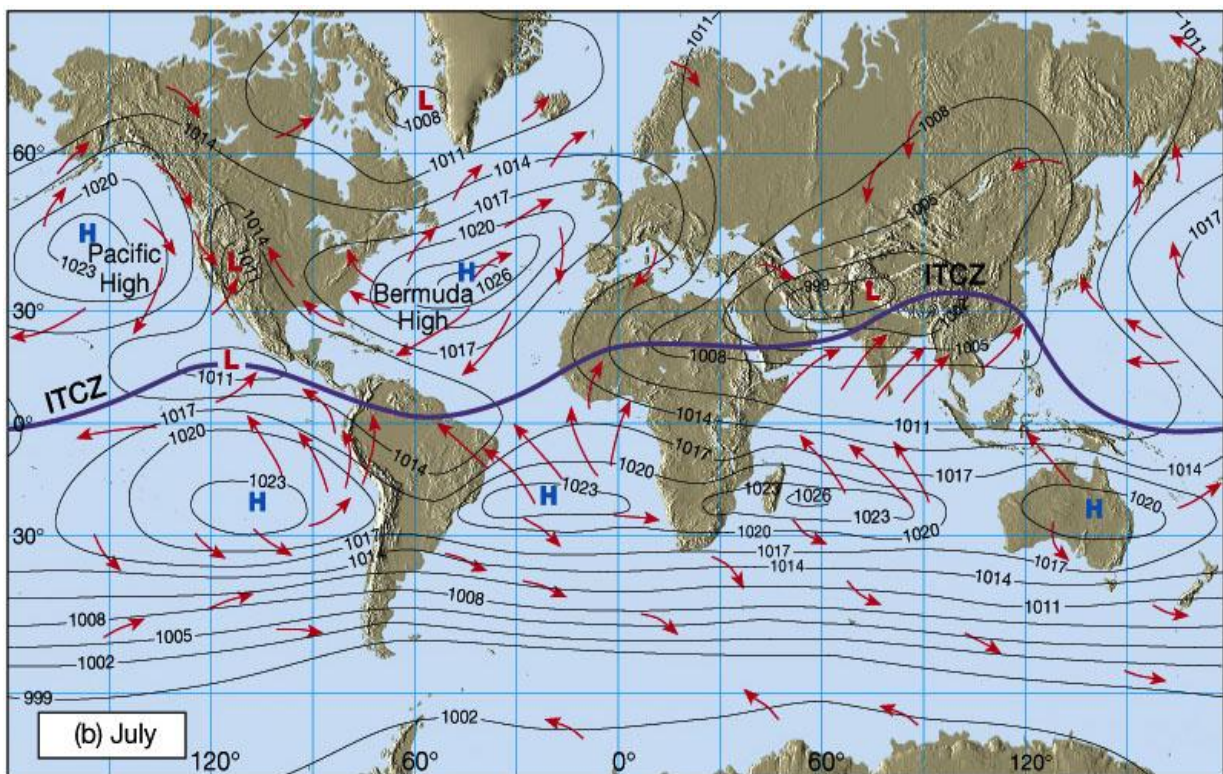
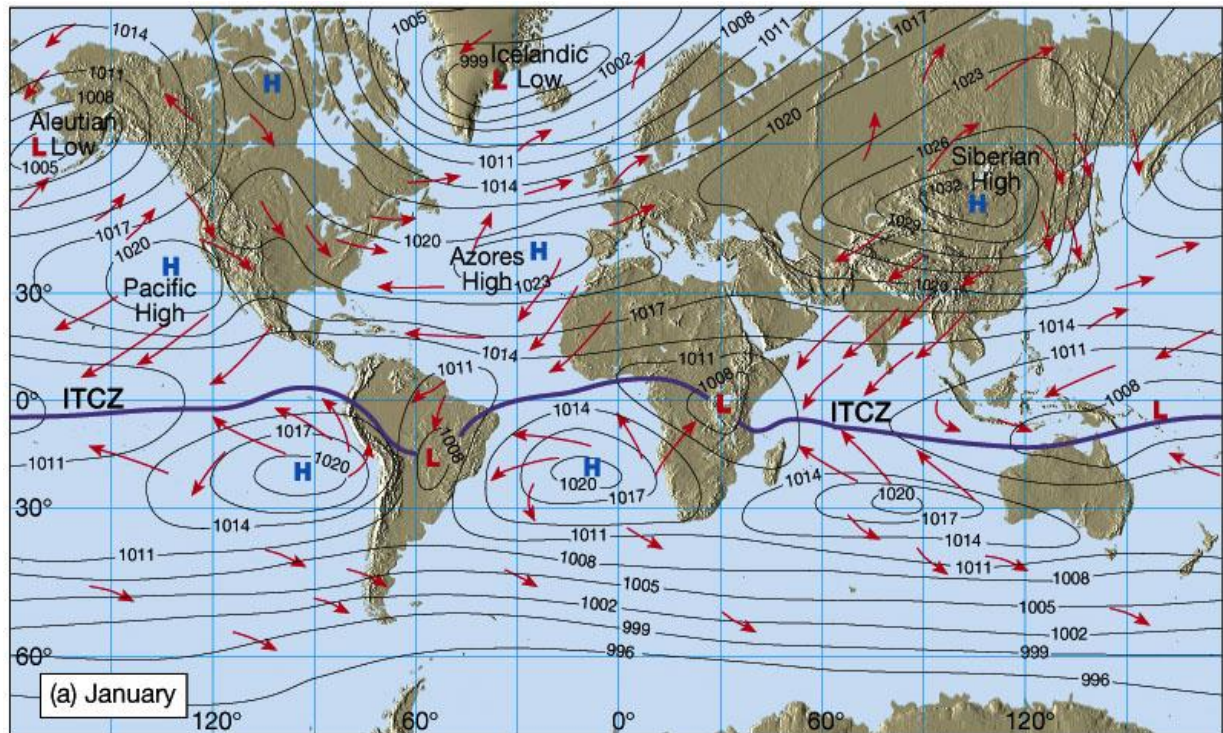


Fig 1.1.1: Mean SLP during a) northern winter and b) northern summer

[Source: <http://www.ux1.eiu.edu/~jpstima/1400/circulation.html>, picture by: Lutgens and Tarbuck]

and is associated with the Asian summer monsoon. Movement of the Inter Tropical Convergence Zone (ITCZ), further north around the Indian Ocean region during summer, (shown in Fig. 1.1.1b compared to 1.1.1a) is clearly noticeable – which is responsible for monsoon around the Indian subcontinent region, causing heavy rainfall.

During winter (shown in Fig 1.1.1a), the AL and Icelandic Low are well developed. The AL extends from the Aleutian Islands into the Gulf of Alaska, and much stormy weather and precipitation in the Western USA are associated with its movement; whereas, the strong circulation around the Icelandic low produces northerly winds and cold weather in the eastern section of N. America. As the AL, ITCZ and PH are frequently in use in our subsequent discussion, we mention their mean position here: the AL covers $\sim 120^{\circ}\text{W}$ - 130°E , 35°N - 70°N ; that of the low of ITCZ in Pacific Ocean is around 140°W - 90°E , 20°S - 20°N ; whereas, the PH centres between 20°N - 50°N , 100°W - 140°E .

Mean SSTs during boreal winter (represented here by December-January-February (DJF)) and summer (represented here by June-July-August (JJA)) are illustrated in Fig. 1.1.2a and b, respectively. During both the seasons, the equator is warmer than the pole. Such temperature gradient is responsible for driving the meridional heat transport through different circulation cells and is described below in the section on general circulation.

1.1.2. General circulation:

Earth's equatorial regions receive more heat, while regions nearer to the pole, experience a net deficit (also shown in Fig. 1.1.2). To prevent the equatorial region getting warmer, and the polar region getting cooler - there must be a transport of heat from the equatorial region towards the pole. Such transport of heat is associated with the following meridional circulation cells (shown in Fig. 1.1.3).

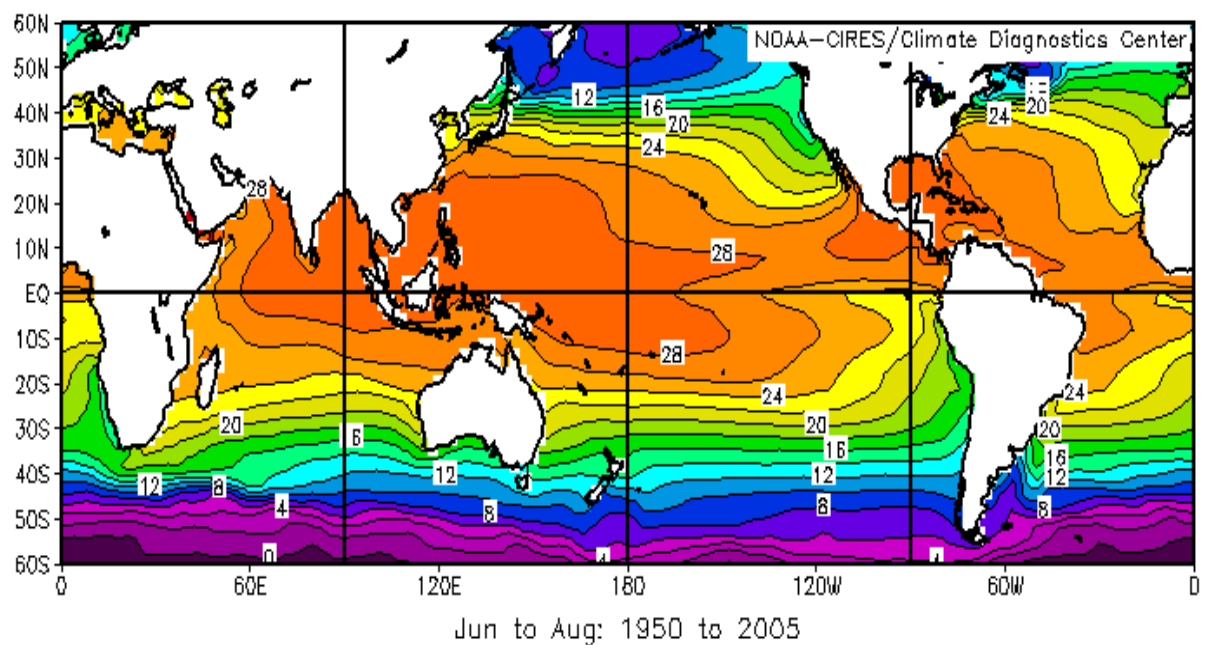
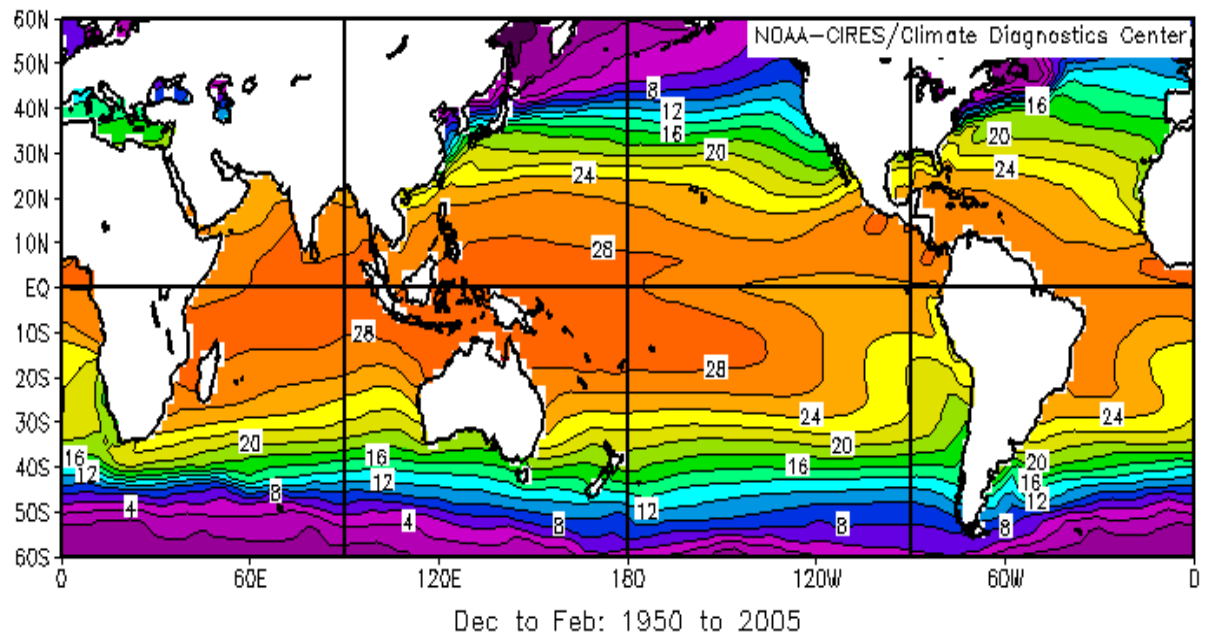


Fig 1.1.2: NOAA extended SST ($^{\circ}$ C) composite mean

[Source: <http://www.atmos.albany.edu/deas/atmclasses/atm305/climomaps.html>]

Hadley Cell: Persistent surface heating around the equator results in a rising and poleward moving air at the equator; as this air moves poleward, it cools radiatively, and sinks near 30° latitude; it finally returns towards the equator, at low levels. This meridional circulation cell is called the Hadley Cell. This is a thermally direct circulation where heat from the sun is converted to motion that transports energy from warm to cold regions. The low pressure near the equator and high pressure bands near 30° latitude are signatures of the rising and sinking portion of this Hadley cell, respectively (shown in Fig. 1.1.3).

Ferrel cell: In midlatitudes the cell that circulates in the opposite direction to the Hadley cell is known as the Ferrel cell and the overall movement of surface air in this cell is from 30° latitude to 60° latitude (shown in Fig. 1.1.3). This cell rises over cold temperature zone and sinks over warm temperature zone. This cell is not driven by thermal forcing but driven by eddy (and associated weather systems) forcing. They are the deviations from the time or zonal average, and are a key component of the general circulation of the atmosphere. In midlatitudes, the biggest contribution to meridional energy transport comes from eddies in the atmosphere and they transport heat poleward. Midlatitude cyclones and anticyclones are the major transient eddies that play an important role in meridional transports of heat, momentum, and moisture. These midlatitude weather systems grow from the baroclinic instability associated with the strong north-south temperature gradients in mid-latitudes. Mid-latitude cyclones are marked by well defined fronts separating the warm air mass from the south and the cold air mass from the north and have typical spatial scales of wavenumbers 5-6 and have typical time scale of 7-10 days (and different from tropical hurricanes, which do not have frontal features).

Polar cell: Air circulates within the troposphere, limited vertically by the tropopause at about 8 km in higher latitudes. In both hemispheres, warm air rises at around 60° latitude and moves poleward through the upper troposphere. When air reaches the polar area, being cooled considerably, it descends as a cold, dry high pressure area, moving away from the pole along the surface and generates the polar cell (shown in Fig.1.1.3).

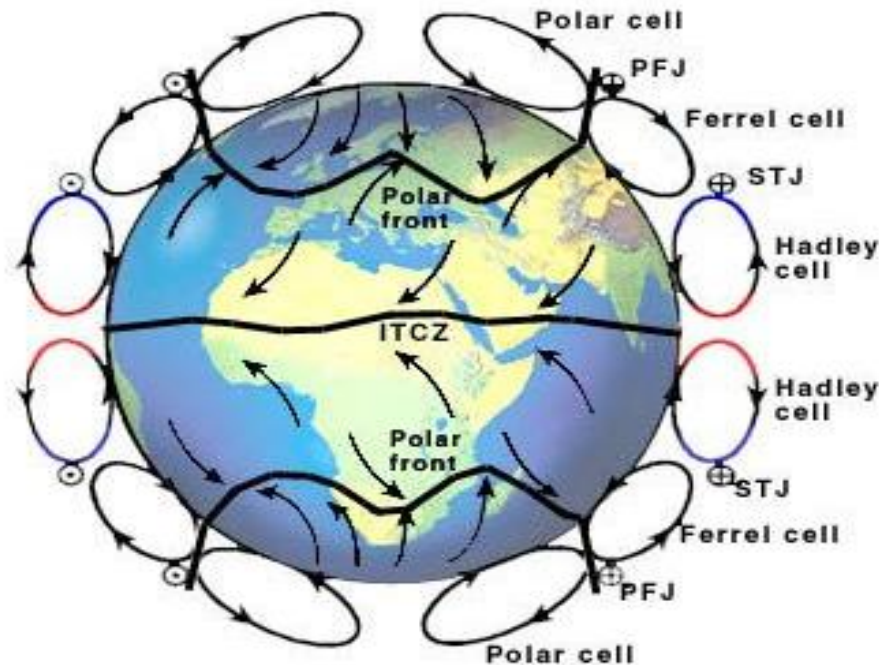


Fig. 1.1.3: Schematic representation of the general circulation showing positions of Polar Front; ITCZ; Subtropical jets (STJ) and Polar Front jets (PFJ).

[\[http://www.metoffice.gov.uk/education/secondary/teachers/atmosphere.html#main\]](http://www.metoffice.gov.uk/education/secondary/teachers/atmosphere.html#main)

Jet streams are seen at approximate boundaries between these cells (shown in Fig. 1.1.3) and are formed around the upper tropospheric region (mainly around the tropopause region –the boundary between the troposphere and stratosphere) due to the thermal wind balance relationship. The subtropical jet (STJ) lies in the interface between the Hadley and the Ferrel cell; whereas the polar front jet (PFJ) also known as the mid-latitude jet or the polar jet are observed between the Ferrel cell and the Polar cell.

Apart from north-south (meridional) circulations, as discussed, there is an east-west (zonal) circulation over the tropical Pacific ocean, known as the Walker circulation, as described below.

Walker Circulation: The Walker circulation is a large scale zonal circulation, with air rising over the Western Pacific and descending around the Eastern Pacific as shown in Fig. 1.1.4. It is generated by the pressure gradient force resulting from a high pressure

system over the eastern Pacific ocean with a low pressure system over Indonesia (the mean position of this high and low pressure system is also shown in Fig. 1.1.1a). The Walker cell varies inter-annually, which is associated with a reversal in direction of winds (from Fig. 1.1.4, shown with arrow) as discussed in the next section.

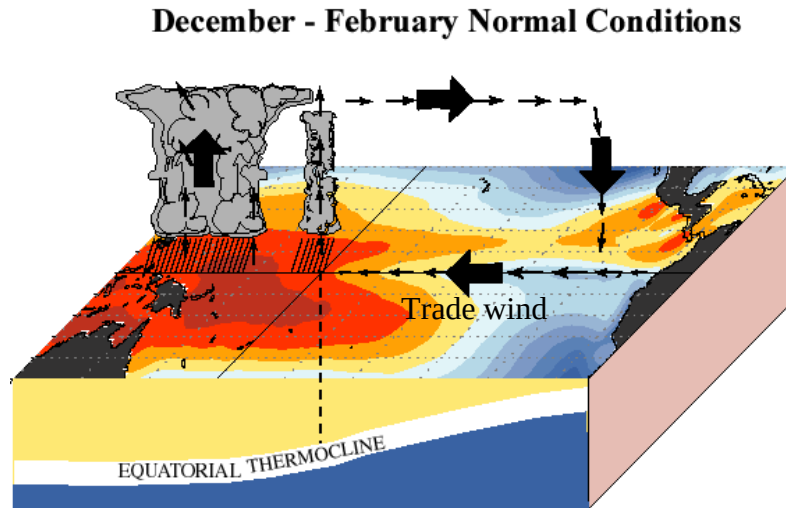


Fig. 1.1.4: Walker Circulation shown with arrow

[http://www.cpc.noaa.gov/products/analysis_monitoring/ensocycle/meanrain.shtml]

1.1.3. Climate variability

There are a number of modes of climate variability, which play important roles in determining the characteristic of different regions. These are classified below as whether they are features of the troposphere or the stratosphere.

1.1.3.1. Variability in the Troposphere:

- El Niño and Southern Oscillation (ENSO):

ENSO is the leading mode in the tropics, although its influence is felt globally. The Southern Oscillation (SO) is a measure of large-scale fluctuation in air pressure between

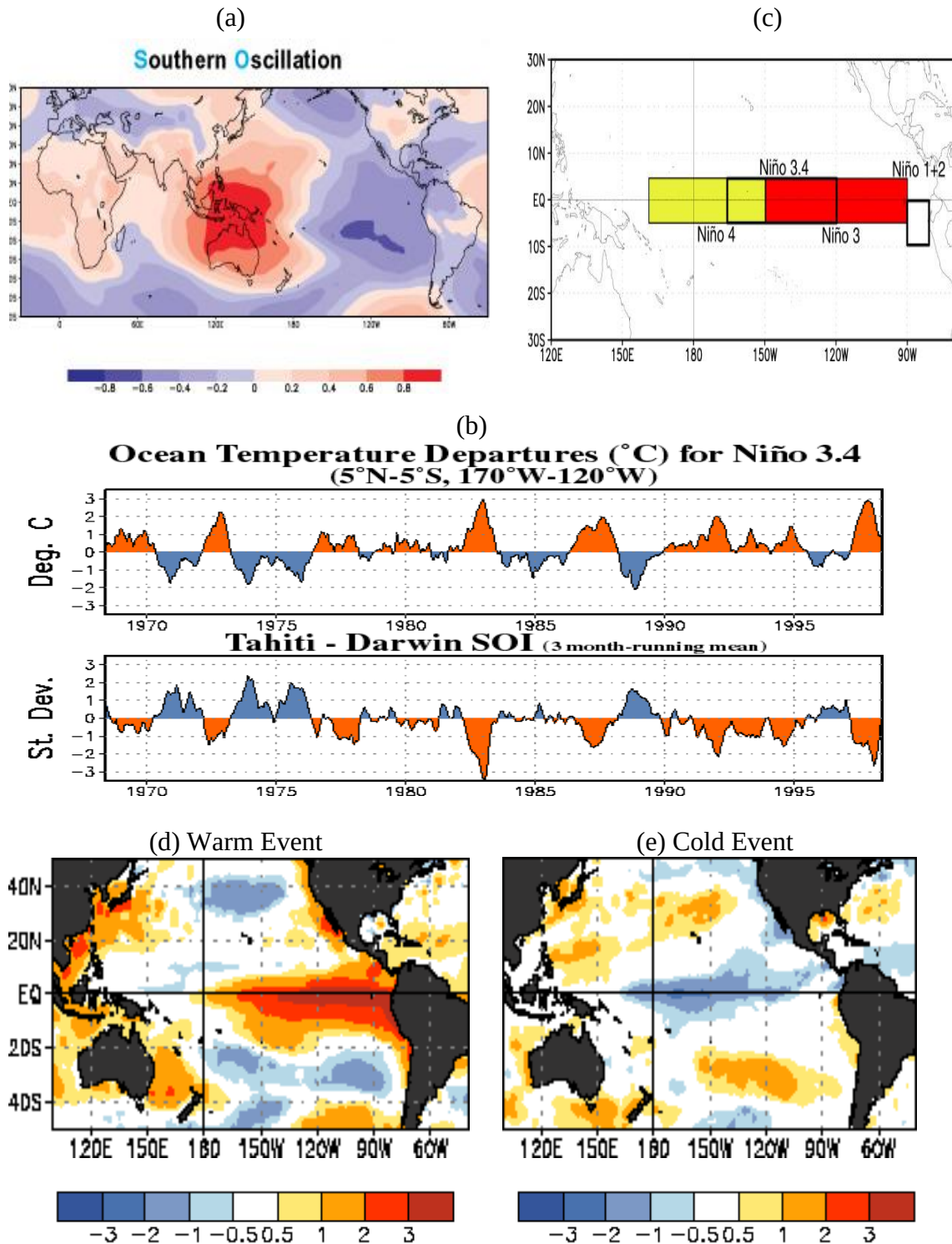


Fig. 1.1.5: Representation of ENSO: a) SLP variation (hPa) showing Southern Oscillation; b) Temperature of Niño 3.4 and Southern Oscillation anti-correlation; c) Different Niño regions; Departure of SST ($^{\circ}\text{C}$) during DJF for Warm events of ENSO (d) and Cold events of ENSO (e).

[Source: http://www.cpc.noaa.gov/products/analysis_monitoring/ensocycle/]

the western and the eastern tropical Pacific about the international dateline. Tahiti (17.5°S, 149.6°W) and Darwin in Australia (12.4°S, 30.9°E) are two opposite ends of this SO's sea-saw and the SO index is calculated based on the difference in air pressure between these two places (shown in Fig. 1.1.5a).

The El Niño is named for the periodic warming which occurs around the eastern tropical Pacific and which adversely affects the fishing industry around the coast of Peru and Ecuador. In general, a smoothed time series of the SO index corresponds very well with changes in ocean temperatures across the eastern tropical Pacific (Fig. 1.1.5b). These two interrelated phenomena, the El Niño and SO, abbreviated as the ENSO, control a large proportion of climate variability in the tropics.

The ENSO index time series is based on oceanic temperatures of the eastern tropical Pacific. Different formulations are in use, based on slight different geographical considerations. The most commonly known indices are Niño 3.4, Niño 3, Niño 4 and Niño 1+2 as shown in Fig. 1.1.5c depicting their geographical coverage.

In the tropical Pacific, trade winds generally drive surface waters westward (also shown in Fig. 1.1.4 to describe the Walker circulation). Surface water that travels from the eastern Pacific all the way long to western Pacific (thus absorbing more solar radiation) become warmer reaching in the western Pacific. El Niño, the warm events of ENSO are observed when the easterly trade winds weaken, which allows warmer waters of the western Pacific to migrate eastward. That eventually reaches at the south American coast causing unusual warming of SST around the east coast of Pacific and thus El Niño (shown in orange colour, Fig. 1.1.5d). In contrast to the El Niño, La Niña, the cold event of ENSO refers to an anomaly of unusually cold SST in the eastern tropical Pacific (shown in blue, Fig. 1.1.5e). During the La Niña, the trade wind and Walker cell both intensify; while during El Niño, they reverse direction.

- Pacific Decadal Oscillation (PDO):

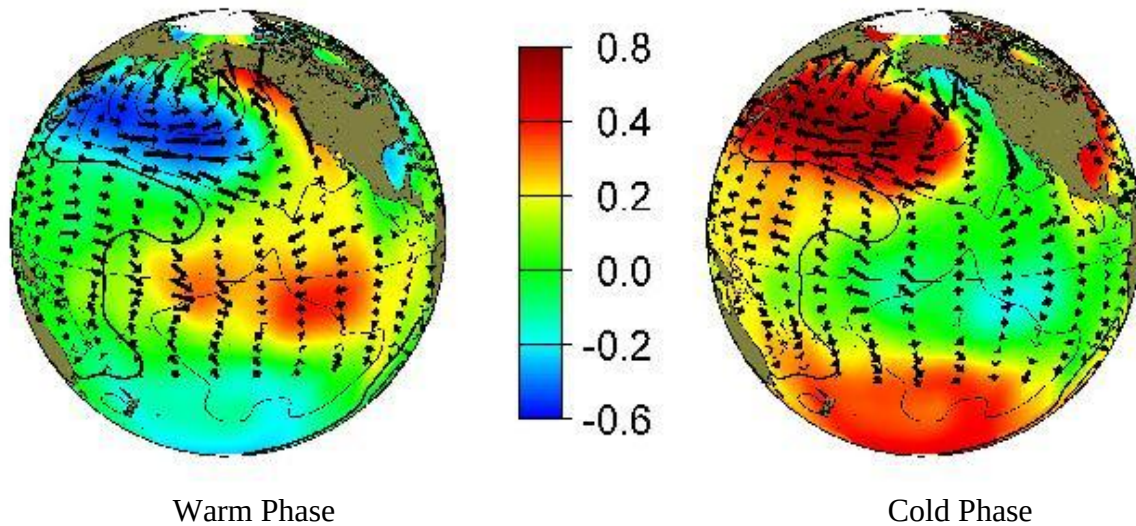


Fig. 1.1.6: Typical wintertime SST in °C (colours), SLP (contours) and surface wind stress (arrows) anomaly patterns during warm and cold phases of the PDO

[Source: <http://jisao.washington.edu/pdo/>]

Apart from the ENSO, there is another mode of strong variability in the Pacific, known as the PDO. The PDO is a long-lived ENSO like pattern of the Pacific climate variability and is defined as the leading principal component of the North Pacific monthly SST variability from pole-ward of 20°N.

Major features of the PDO in relation to the ENSO may be described as follows: a) its main climatic fingerprints are in the north Pacific with secondary signatures (of SST) in the tropics (shown in Fig 1.1.6), which is the other way round for the ENSO (Fig. 1.1.5.d and e), though both their warm and cold phases possess similar sign temperature anomaly like ENSO (and hence the phases have been named); b) it has a long term variability and persists for 20-30 years; whereas for ENSO the variability is for 2-7 years.

- North Atlantic Oscillation (NAO):

Moving from the Pacific region, we are now focusing on the Atlantic Ocean region where another mode of variability, named as the NAO, plays a crucial role. This is a large scale

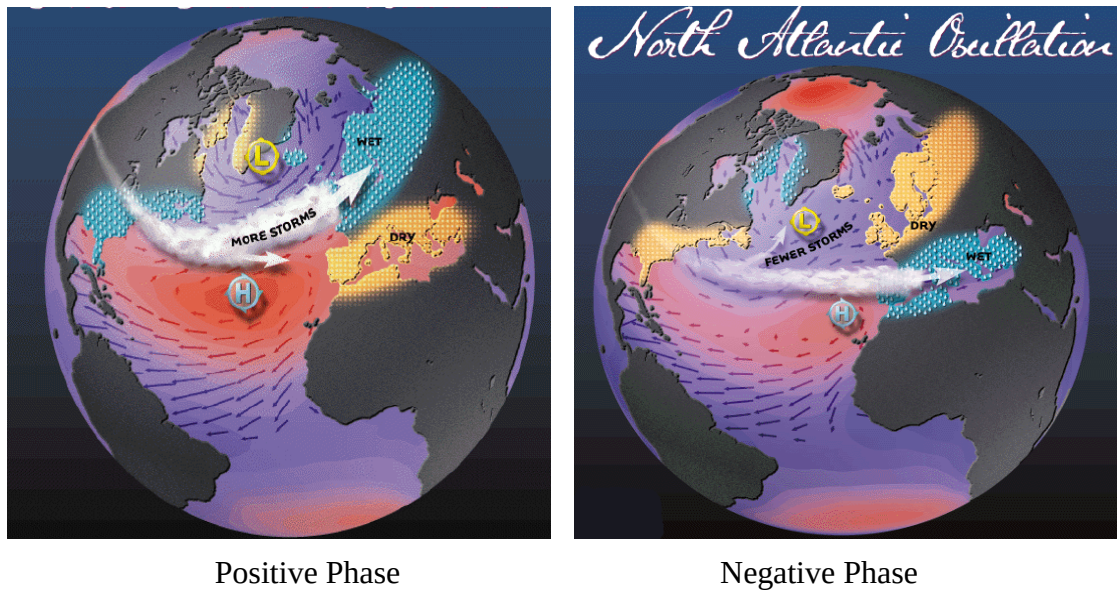


Fig. 1.1.7: Different phases of the NAO

[<http://www.ldeo.columbia.edu/NAO/>, picture by Visbeck, M]

see-saw in atmospheric mass between the polar low and the subtropical high in the north Atlantic region; it is the dominant mode of climate variability during boreal winter around the north Atlantic ranging from Europe to the central north America and much into northern Asia.

The major features of the NAO are as follows, also shown in Fig. 1.1.7: a) the positive phase of NAO shows a stronger than the usual subtropical high pressure centre over the Atlantic, with a deeper than the normal Icelandic low; b) such increased pressure difference generates stronger and frequent winter storms crossing the Atlantic Ocean on a more northerly track; c) it causes dry and cold winters in Greenland and the northern Canada with wet and warm winters in Europe; d) the eastern part of America experiences a wet and mild winter.

Apart from the NAO, which is regionalized and defined only around the Atlantic Ocean region, there is another mode of variability in the NH known as the Arctic Oscillation (AO), that covers more of the Arctic including both the Pacific and the Atlantic Ocean and zonally symmetric, which is described below.

- **Arctic Oscillation (AO)** and **Antarctic Oscillation (AAO)**:

A barometric seesaw between the polar region and mid-latitudes is seen in both hemispheres: for the northern hemisphere it is termed the Arctic Oscillation (AO); whereas, for the southern hemisphere, it is the Antarctic Oscillation (AAO). The AO is usually defined as the first leading mode from the EOF analysis of monthly mean geopotential height anomalies at 1000 hPa, pole-ward of 20° N, in the NH; whereas for the AAO, it is the same at 700 hPa level and measured pole-ward of 20° S in the SH. In SH, the different pressure level is used to partially alleviate the ambiguities introduced by the reduction to sea level over the high terrain of Antarctica. Both of their positive phase shows deeper than normal polar low with stronger than usual mid-latitude high. The surface signature of AO and AAO measured in terms of geopotential height, as observed by Thompson and Wallace (2000) is shown in Fig. 1.1.8.

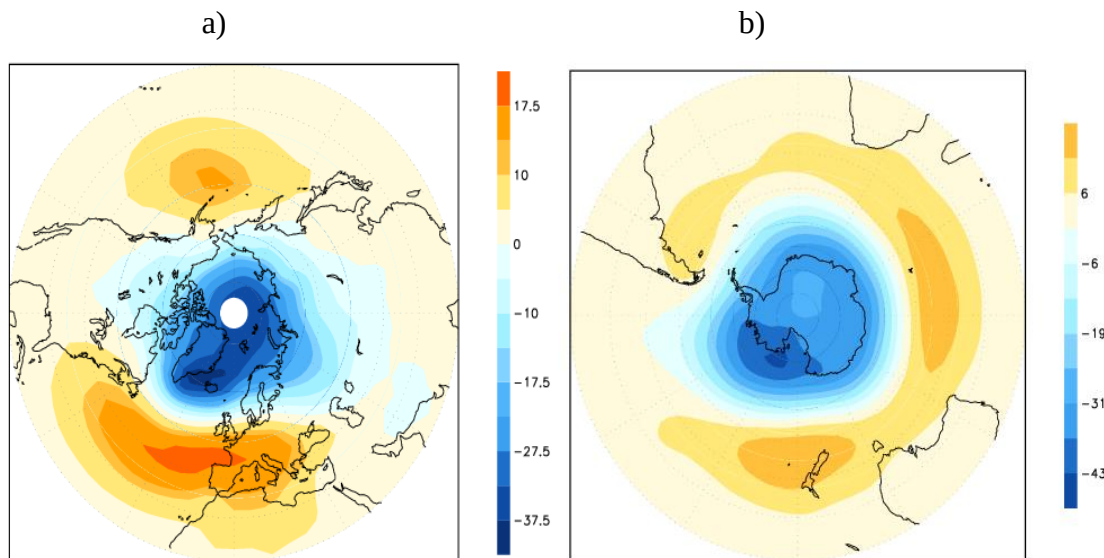


Fig. 1.1.8: The surface signature of positive AO (a) and AAO (b) measured in terms of geopotential height

[after, Thompson and Wallace, 2000]

The AO and AAO patterns are similar from the Earth's surface up to 50 km and in a broader sense are known as the Northern Annular Mode (NAM) and the Southern

Annular Mode (SAM), respectively. In the stratosphere, the NAM (SAM) is a measure of the strength of the polar vortex; whereas at the surface the NAM (SAM) is known as the AO (AAO). The study of Baldwin and Dunkerton (2005) also shown here in Fig. 1.1.9, depicts that the NAM patterns calculated as the leading empirical orthogonal function of November–April (90-day lowpass filtered) geopotential for 1958–2000 at 10 and at 1000 hPa are similar, and to first order, zonally symmetric.

The band of upper-level winds that circulate around the pole in the stratosphere, form the polar vortex (also shown in Fig. 1.1.10 with blue arrows). When the annular mode index becomes positive, the strength of vortex increases and winds constrict around the pole, locking cold air masses in places near the pole. On the other hand, a negative surface annular mode, associated with weak vortex, allows intrusion of cold air masses to plunge southward into the North America, Europe and Asia (for NAM). Different phases of annular modes are thus linked with variations of surface weather patterns in the polar

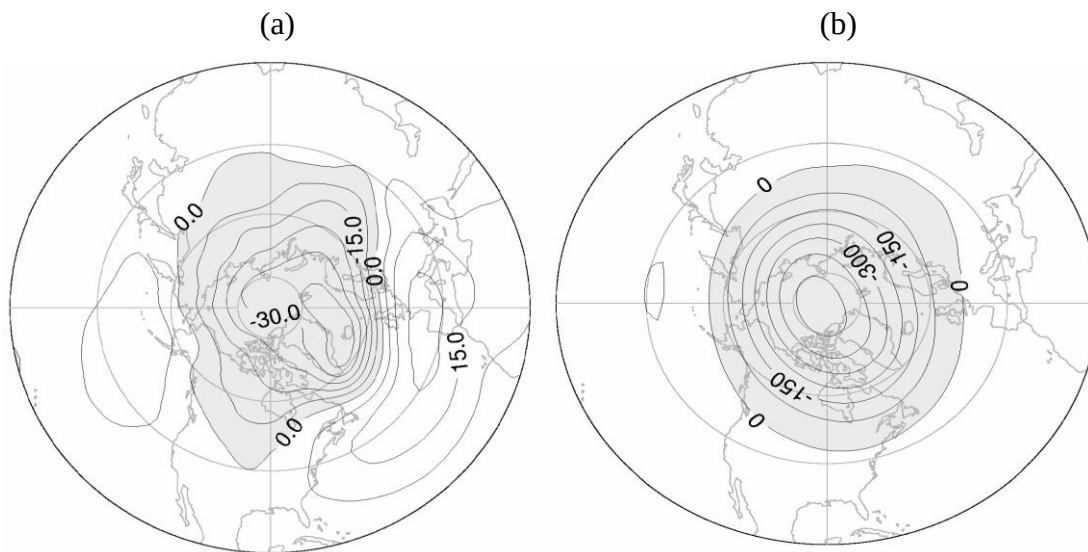


Fig. 1.1.9: Northern Annular Mode (NAM) patterns at (a) 1000 hPa and (b) 10 hPa for November–April geopotential during 1958–2000.

[Baldwin and Dunkerton, 2005]

region. The associated surface climate change with the phase of annular modes are described here with an illustration in Fig.1.1.10. In the positive phase of the surface NAM, the higher pressure at the mid-latitudes drives cyclones farther north toward the Arctic; changes in the circulation pattern then brings wetter weather to the Scandinavia and Iceland, Alaska; alongside it brings drier conditions to the Mediterranean and the western United States. The situation changes during the negative phase (Fig. 1.1.10). Moreover, in the positive phase, cold winter air masses do not extend as far into the north America and Europe as it would during the negative phase of the oscillation. This keeps much of Europe and the east of Rocky Mountains of the United States warmer than normal during periods of high AO. Alongside, it keeps places like Newfoundland, Greenland and Labrador colder than usual.

From the above it is clear that the troposphere and stratosphere are strongly coupled and one possible mode of communication from the stratosphere to troposphere is via the polar mode of variability. Thus, for improving our understanding of tropospheric variability, we may also need to know about the variability of the stratosphere.

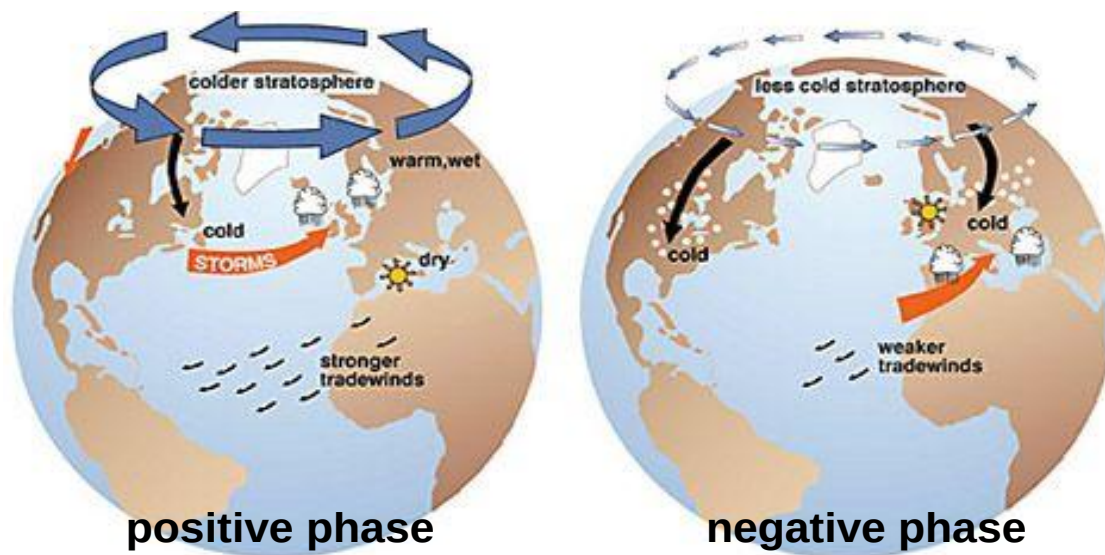


Fig. 1.1.10: Effects of different phases of the Arctic Oscillation

[http://nsidc.org/arcticmet/patterns/arctic_oscillation.html, picture by Wallace, J.]

1.1.3.2. Variability in stratosphere:

There are two major forms of variability in the stratosphere, viz. the **Quasi-Biennial Oscillation (QBO)** and the **Stratospheric Sudden Warming (SSW)**. In the current chapter, we mainly focus on their definition and characteristics; while the associated mechanisms will be addressed in Chapter 4.

- **Quasi-Biennial Oscillation (QBO):**

This is an oscillation in the equatorial stratospheric wind, with zonal winds change between the east and west with a period of just greater than 2 years. The time series of equatorial deseasonalised zonal wind is shown in Fig. 1.1.11, where the colours in the

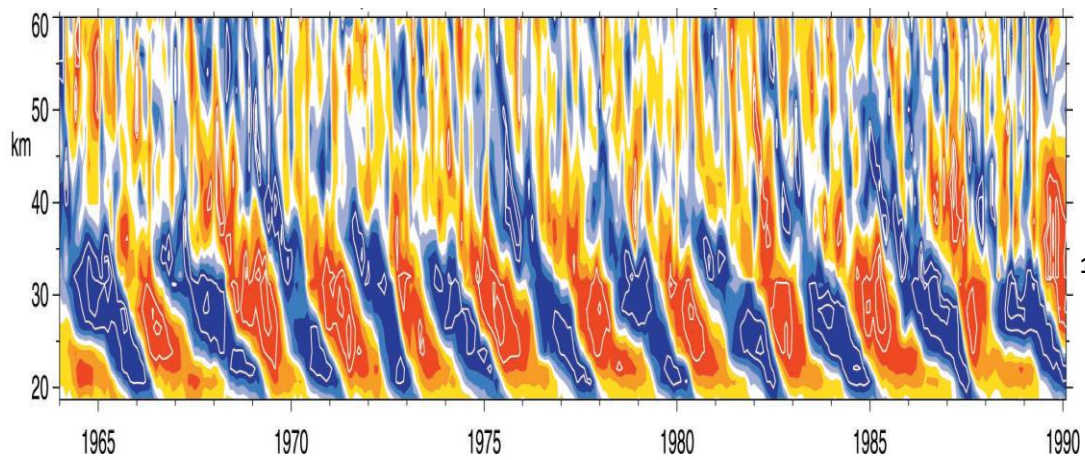


Fig.1.1.11: Time-height section of the monthly-mean zonal wind components (m s^{-1}), with the seasonal cycle removed, for 1964-1990. The contour interval is 6 m s^{-1} , with the band between -3 and $+3$ unshaded.

[Source: Gray et al. 2001a)]

lower and middle stratosphere reveal the two different phases of the QBO: red for westerly and blue for easterly.

The main characteristics of the QBO are as follows:

- a) It is prominent between 10mb and 100mb.
- b) The period of oscillation is 20 to 36 months with a mean of around 28 months.

- c) Wind regimes propagate downward with time with speed roughly 1km/ month.
- d) The maximum peak-to-peak amplitude of 40 to 50 m/s is seen at 20mb.
- e) Easterlies are generally stronger than westerlies.
- f) Easterly winds last longer than westerly winds at lower levels; while the converse is true at higher levels.
- g) There is considerable variability of the QBO, in both amplitude and period.

- **Stratospheric Sudden Warming (SSW):**

It is the most dramatic event in the stratosphere, where the polar vortex of westerly winds in the winter hemisphere, over the course of a few days, abruptly slows down or even reverses direction. This is also accompanied by a rise in the stratospheric temperature by several tens of Kelvins. Fig. 1.1.12 illustrates the latitude time series of zonal average temperatures around 30 km during two different periods, as shown by Gray (2004). The times of sudden warming are marked by 'X' and 'Y' in Fig. 1.1.12a and b respectively, when temperatures were observed to rise abruptly.

SSWs can be classified into three main categories:

- *Major warming*

The major warming occurs when westerly winds at 60° N and 10 hPa reverse, i.e. become easterly from westerly. A complete disruption of the polar vortex is observed and the vortex either splits into two separate vortices, or displaces from its normal location over the pole.

For a major warming to occur the following two defining criteria need to be satisfied as specified by WMO:

- a) An increase in temperature, poleward from 60° latitude at 10 hPa of 40-60 K takes place in less than one week.
- b) Zonal mean zonal wind over the same region reverses.

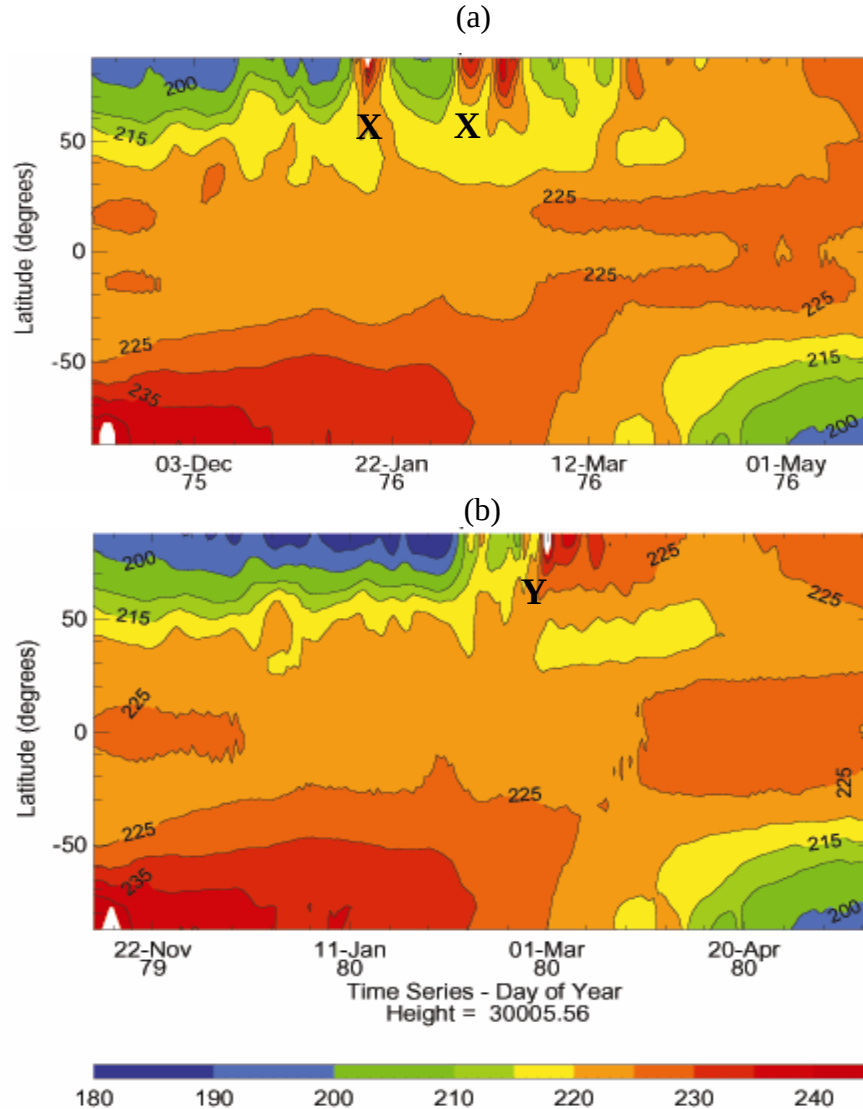


Fig.1.1.12: Latitude time series of zonal average temperatures around 30 km
http://lasp.colorado.edu/sorce/news/2004ScienceMeeting/SORCE%20WORKSHOP%202004/SESSION_2/2_5_Gray.pdf, picture by: Gray, L)

Such warming can be shown in Fig. 1.1.13, where the time series of daily temperature ($^{\circ}\text{C}$) over the north pole is plotted from November 2005 till April 2006. The variations at different levels, viz. 1 hPa, 10 hPa and 30 hPa are shown by different colours. The mean of 30 years North polar temperature at 30 hPa level are marked by a grey line. From this picture, it is evident that there is a sudden rise in temperature for a few weeks after 1st Jan. 2006 at all three pressure levels. Since the temperature rose more than 40° at 10 hPa level (shown with purple colour) and the warming was also accompanied by a reversal of

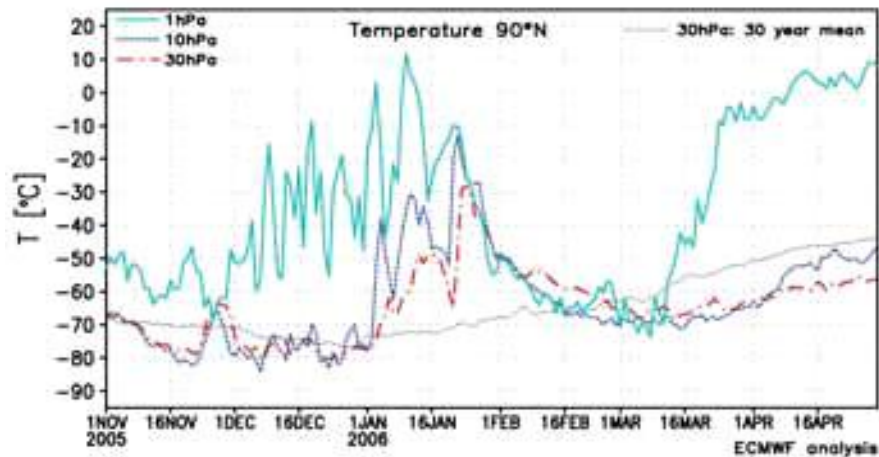


Fig.1.1.13. Time series of the daily temperatures (°C) over the North Pole at different pressure levels, November 2005 till April 2006.

[ECMWF analyses: <http://www.bu.edu/causes/documents/causes-news-v3-n2.pdf>]

the wind (i.e., from westerly to easterly which is not shown here), it may be classified as a major warming.

- *Minor warming*

The minor warming, though similar to the major warming, is less dramatic. According to the definition of WMO, a stratospheric warming is called minor if:

- A significant temperature increase i.e., at least 25 degrees in a period of week or less is observed at any stratospheric level, in any area of the winter time hemisphere and
- The wind reversal from westerly to easterly is less extensive (i.e., the zonally averaged zonal wind does not reverse) and the polar vortex is not broken down.

- *Final warming*

The radiative cycle in the stratosphere means that the mean flow is westerly during the winter, while easterly during the summer. A final warming takes place on this transition, so that winds around the polar vortex reverse direction for the warming and the stratosphere enters the summer easterly phase. It is known as the final warming of the current winter, because winds do not change back until the following winter (shown by 'Y' in Fig. 1.1.12b).

1.2. Solar-climate links

The energy driving the global atmospheric circulation comes from the Sun and the Earth is subject to enhanced solar irradiance at periods of higher solar activity. However, the link between the solar cycle and climate has been a puzzle, because the solar energy output changes by only about .1% during a typical 11 year cycle (Lean and Rind, 2001), a change that is too small to produce significant changes in surface conditions based on energy considerations. The extent to which changes in solar activity affects climate has been the subject of considerable investigations over many years and some of the results are mentioned here.

- **Solar signals in the stratosphere:**

Nowadays, there is a general agreement that the direct influence of the changes in the UV part of the spectrum (6 to 8% between solar maxima and minima) leads to more ozone and warming in the upper stratosphere in solar maxima (Haigh, 1994; Hood et al., 1997; Hood, 2004; Crooks and Gray, 2005). Recent analysis of solar spectral irradiance variability from 200-2400 nm by Harder et al (2009) using Spectral Irradiance Monitor (SIM) on-board the Solar Radiation and Climate Experiment (SORCE) satellite shows an even large change in UV with an opposing trend in the visible range but that remains to be confirmed. An extensive analysis of the latitudinal and seasonal structure of the solar signal at various pressure levels has been carried out by Labitzke, van Loon and co-workers (Labitzke and van Loon, 2000, van Loon and Labitzke, 2000; Labitzke 2001, 2002, 2003, 2004, Soukharev and Labitzke 2000, 2001) using temperature and geopotential height dataset (Labitzke et al., 2002; Labitzke, 2007)). Using Rocketsonde data, Lidar data, model assimilation dataset, satellite data from stratospheric Sounding unit (SSU) and Microwave Sounding Unit (MSU), several analyses were carried out to find the solar influence on the middle atmosphere (Dunkerton et al., 1998; Keckhut et al., 1999, 2001, 2004; Scaife et al. 2000; Hood, 2004). It is seen that apart from a maximum in the upper stratosphere, there also exist a secondary maximum in the lower stratosphere in the signals in both ozone (Fig.1.2.1) and temperature (Fig. 1.2.2). Using a fixed dynamical heating (FDH) model Gray et al. (2009) were able to demonstrate that the lower stratospheric temperature signal is a response consistent with the change in ozone

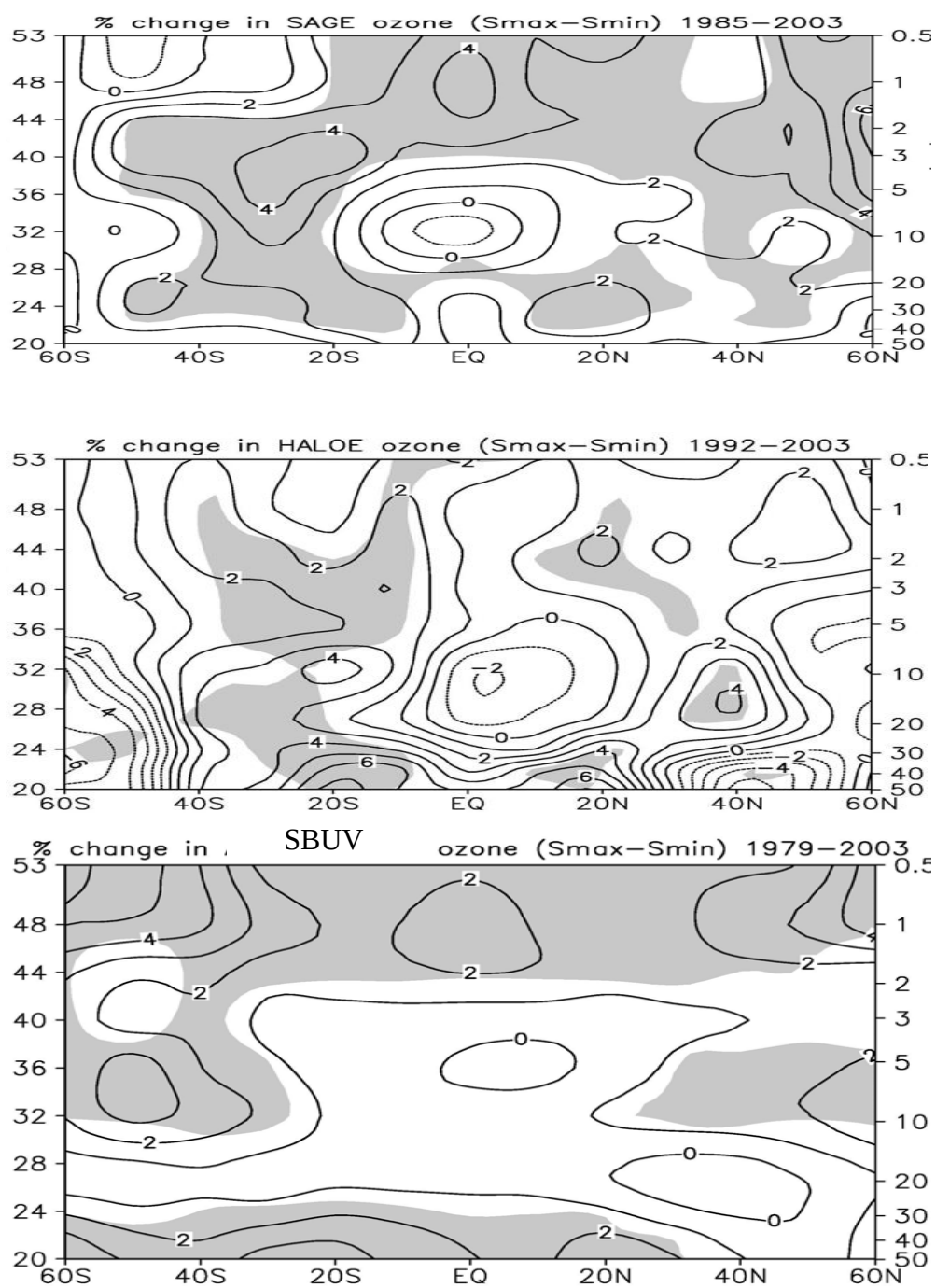


Fig.1.2.1: Ozone solar cycle regression coefficient derived from three independent satellite datasets SAGE (top); HALOE (middle) and SBUV (bottom). Shaded region are significant upto 95 % level.

(Source: <http://www.atm.damtp.cam.ac.uk/shuckburgh/AGU/pdf/hood.pdf>)

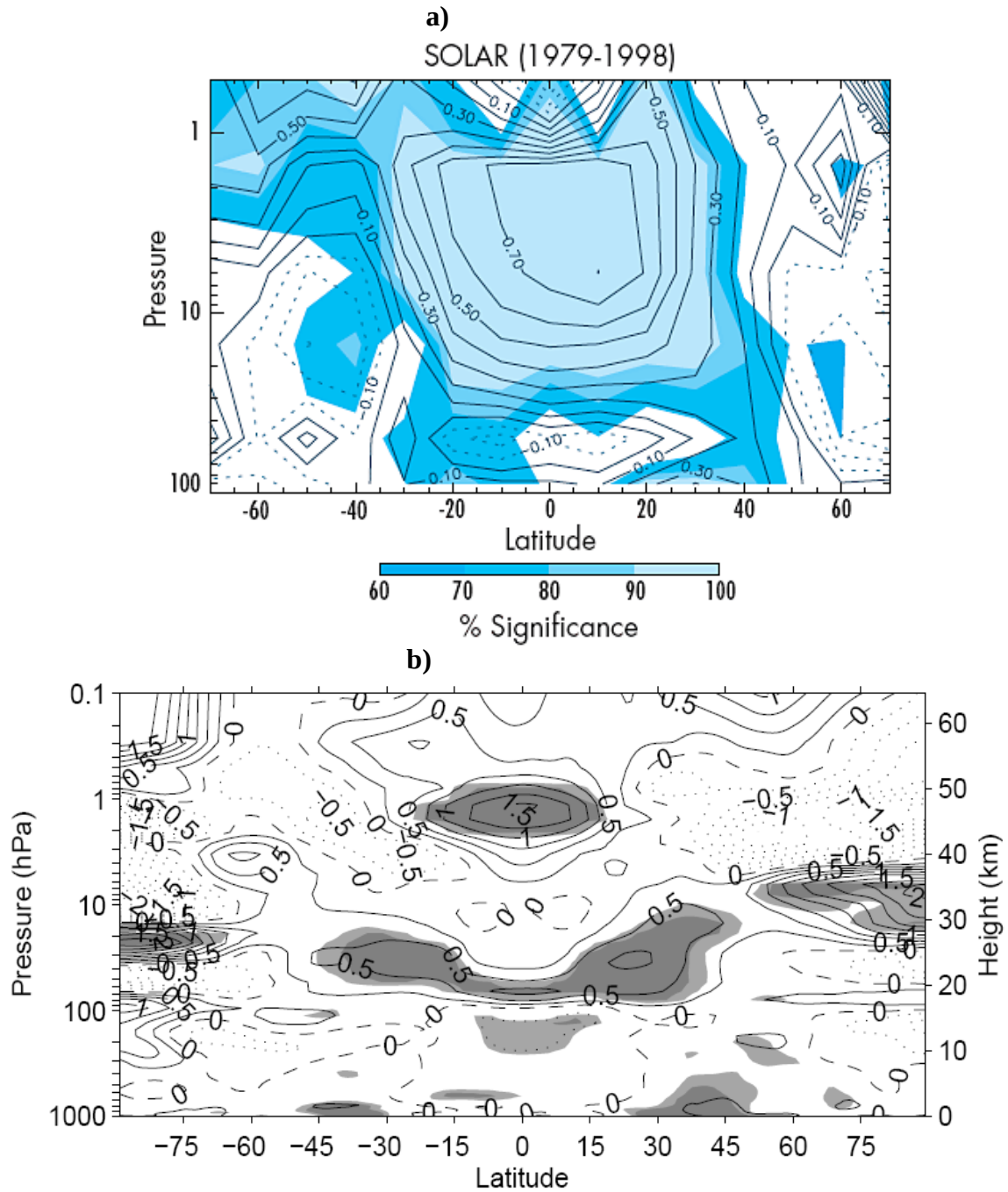


Fig.1.2.2: Annual mean temperature difference (in K) during solar max to solar min from observational data: (a) SSU/MSU satellite data analysis by Keckhut et al. (2004) for period 1979-1998 with various levels of statistical significance; (b) ERA reanalysis data results by Frame and Gray (2009) for period 1978-2008 with dark/light shading indicates 99% / 95% significance level.

at those levels. Now although there is a general consensus regarding the distinct signature of the 11-year solar cycle in stratospheric ozone and temperature; the discrepancies in the detailed analysis of the amplitudes and distribution of the signals, as seen in these figures, still need to be resolved (Gray et al., 2010).

Solar signal and QBO:

Updating their previous work, van Loon and Labitzke (1994, 2000) confirmed that during northern hemisphere winter, it is necessary to group the meteorological data according to the phase of the Quasi-Biennial Oscillation (QBO) in order to find a clear signal of the 11-year sunspot cycle in the stratosphere. The reality and significance of using this approach have been confirmed by Naito and Hirota (1997) and Salby and Callaghan (2004). This phenomenon exists during the whole winter but maximizes during the late winter (February) when the planetary wave activity is largest (van Loon and Labitzke, 2000). Analysis of the 30 hPa temperatures geopotential heights Labitzke (2002, 2003) has shown that this is also true of the summer-time solar signal, which has a very different characteristic in the two QBO phases.

Kodera and Kuroda (2002) and Rind et al. (2002) through modelling studies suggested that solar variability influences the structure of the polar night jet and hence the propagation of planetary-scale waves that travel vertically from the troposphere. This then affects their ability to impact the polar vortex and to produce sudden stratospheric warming. More details of solar cycle/ QBO interaction has been provided in a series of modelling studies by Gray and co-workers (Gray et al. 2001b, 2003, 2004; Gray, 2003). They showed that the propagation of planetary scale waves and the development of NH sudden stratospheric warmings are very sensitive to the zonal flow in the early winter upper stratospheric equatorial/ subtropical region.

Some studies even detected modulating effect of the solar cycle on QBO phases. Salby and Callaghan (2000) proposed that the length of the westerly phase of the equatorial QBO may be affected by the solar cycle, being longer during solar minimum conditions.

Soukharev and Hood (2001) found the lengths of both QBO phases to be longer during the solar minimum. The QBO wind changes sign during the winter of some solar max years has also been observed by Salby and Callaghan (2004). All these studies indicate that there might be some influence of the Sun on the phase of the QBO, which needs to be explored.

- **Solar signals in lower atmosphere/upper ocean:**

The existence of signals of the 11-year solar cycle in meteorological fields of the lower atmosphere/ upper ocean has been confirmed in analyses of observational data (Gray, 2005; Haigh, 2003; Gleisner and Theijll, 2003; Van Loon and Shea, 2000) though the response found is latitudinally non-uniform and larger than would be predicted from

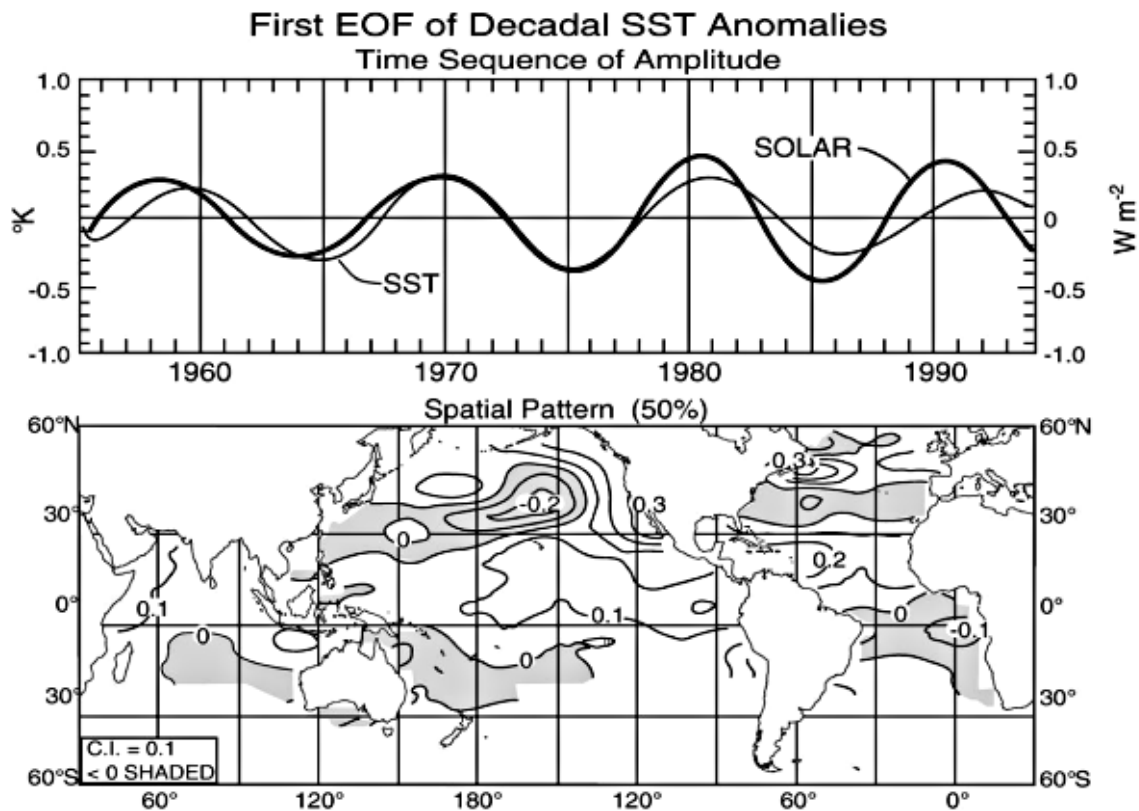


Fig. 1.2.3: Time series (top) of the magnitude and a spatial pattern (bottom) of response in SST associated with solar variability

[White et al. 1997]

radiative forcing considerations alone. E.g. the results of solar signals, using EOF analysis, in sea surface temperatures (SSTs) detected by White et al. 1997, during 1955 – 1994 is shown in Fig. 1.2.3. The upper panel shows the time varying amplitude; whereas, the lower panel, the pattern of response. It is clear from the various studies that SSTs do not warm uniformly in response to enhanced solar activity and the amplitude of change is beyond the predicted range, considering radiative forcing alone.

Christoforou and Hameed (1997), showed that during sunspot maxima, the Center of Action (COA) of the Aleutian Low (AL) pressure system moves north westward (the movement to west by as much as 700 km), while the COA of the North Pacific High (PH) pressure system moves north by as much as 300 km. During a sunspot minima, the COA for these pressure systems moves in the reverse direction (Fig.1.2.4b). Shifts in the COA change storm trajectories and cause large anomalies in regional climate. Apart from varying position, AL even exhibits significant changes in intensity; there is on the average 1.6 mb difference between years of extreme solar activity (Fig.1.2.4a).

Consistent with the observed variation in Pacific COA, Haigh et al. (2005) also detected variation in the Ferrel cell and mid-latitude jets. Using multiple regression analysis of NCEP Reanalysis data for zonal mean zonal winds they showed that when the sun is more active, the mid-latitude jets are positioned further polewards and become weakened (shown in Fig. 1.2.5) and deduce an expansion in Hadley cells and polar shift in the Ferrel cells. Brönnimann et al. (2006) using upper air data also observed poleward displacement of the subtropical jet and Ferrel cell with increasing solar irradiance.

Several studies (e.g. Kodera, 2003; Kuroda and Kodera, 2005; Thejll et al., 2003; Ogi et al., 2003, 2004; Huth et al., 2006; Lee and Hameed, 2007) have even found evidence of a solar influence on polar modes of variability (AAO and AO). Haigh and Roscoe (2006), through multiple regression analysis (using NCEP data during the 2nd half of last century) also showed there is a clear positive solar influence on the NAO.

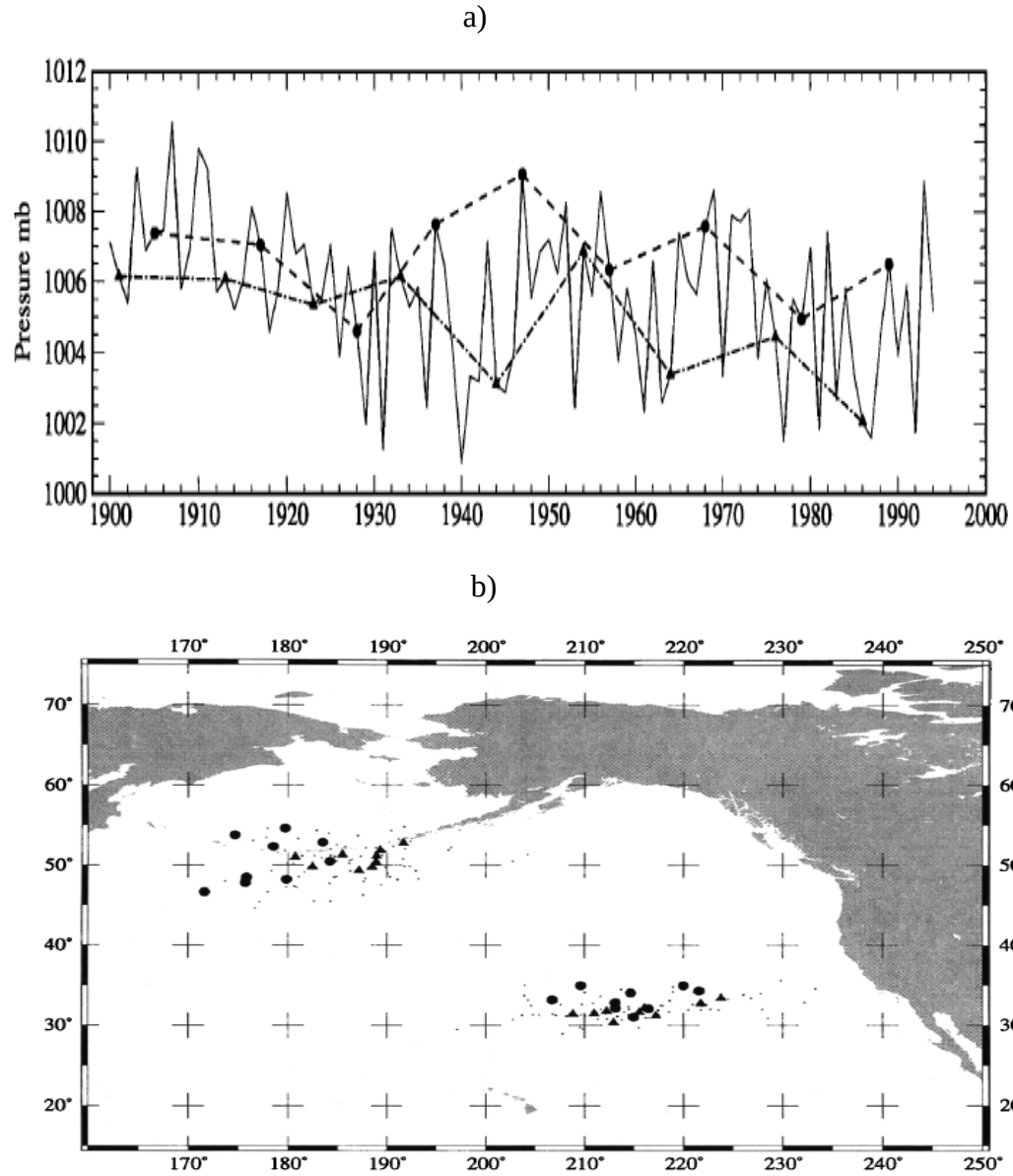


Fig.1.2.4: *Pressure* around AL deviates between years of low and high solar activity (a), along with the *locations* of AL and PH (b); triangles indicate years of low solar activity and circles years of high solar activity.

[Source: Christoforou and Hameed, 1997]

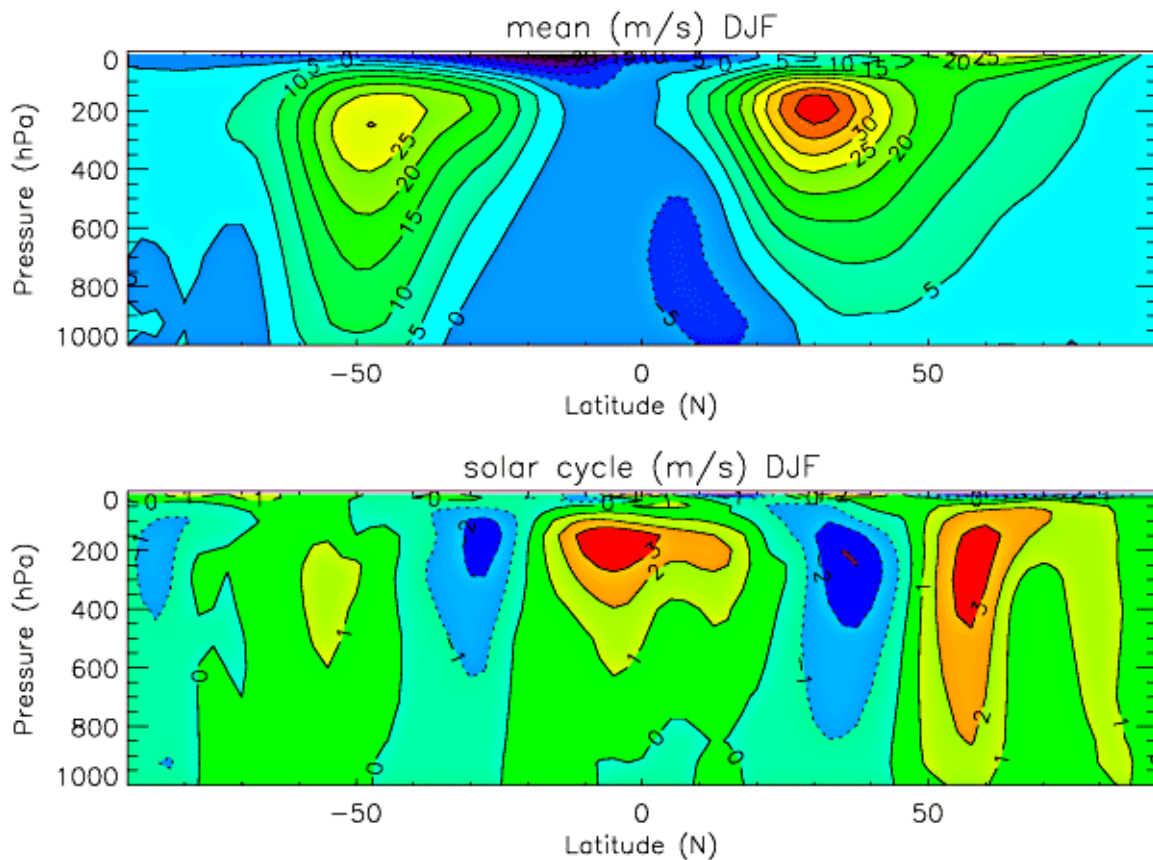


Fig.1.2.5: Results of multiple regression analysis of NCEP Reanalysis data: top) zonal mean zonal wind (u) as a function of pressure and latitude; below) difference in u between solar maximum and minimum.

[after, Haigh et al. (2005)]

- **Solar variability in the troposphere : Potential Mechanisms**

Decadal scale solar signal was detected in the troposphere (e.g., Haigh 1996, 1999, 2003; Lean and Rind 2001, 2008; Rind 2002; Lean et al. 2005; van Loon and Labitzke 1998; van Loon and Shea 2000; Crooks and Gray 2005)); in the ocean (White et al. 1997, 1998, 2008 a & b; Weng 2005); specifically in the Indo-Pacific region (van Loon et al. 2004, 2007, 2008). Meehl and co-workers (2008, 2009) posed the question of whether the effect of solar variability in the troposphere is ‘*bottom-up*’, i.e., forced by solar heating of the surface or ‘*top-down*’, i.e., primarily driven from the stratosphere. Different mechanisms were proposed to support both the views and are briefly described below with some potential routes for amplification.

Bottom-up:

Based on a modelling study Meehl et al. (2003) proposed a mechanism related to air-sea-radiative coupling at the surface in the tropics, whereby the spatial asymmetries of solar forcing, induced by cloud distributions, result in greater evaporation in the subtropics and consequent moisture transport into the tropical convergence zones, thus producing higher precipitation through dynamically coupled ocean-atmosphere interaction. Meehl et al. (2004) also suggest that solar forcing produces coupled dynamical interactions in the tropics that strengthen the Hadley and Walker circulation.

van Loon and co-authors have shown in multiple papers (2007, 2008, 2009) that the response to peaks in solar forcing resembles La Niña conditions. According to them the peak solar conditions are however different from La Niña events in the Southern Oscillation based on reasons: excluding two SunSpot Number(SSN) peak years 1989 and 1905 which are well defined La Niña and El Niño years respectively, the signals essentially remain unaltered; the two signals are different mainly in the equatorial stratosphere (van Loon and Meehl (2008)) which is confined to pressures less than 25 hPa, ie., right at the top of the NCEP/NCAR Reanalysis Dataset domain. Results from two different global coupled models (Parallel Climate Model (PCM) and Community Climate System Model, version3 (CCSM3)) shown by Meehl et al (2008), using multiple ensemble members from 20th century simulations and thus on the order of 40 to 50 realizations of peak solar forcing for each model, show a similar pattern to the shorter observational record, and Meehl et al. (2008, 2009) suggest mechanisms which is illustrated in Fig. 1.2.6. The global average solar forcing during peaks of the DSO compared to solar minimum is $.2\text{Wm}^{-2}$ which is 1 to 2Wm^{-2} at the top of the atmosphere at tropical latitude (Lean et al., 2005). According to Meehl et al. (2008), in the tropics where the sun is most directly overhead, the forcing can be considerably greater than the global average and this extra energy is responsible to trigger coupled air-sea interactions. Increase of net solar at the surface in relatively cloud

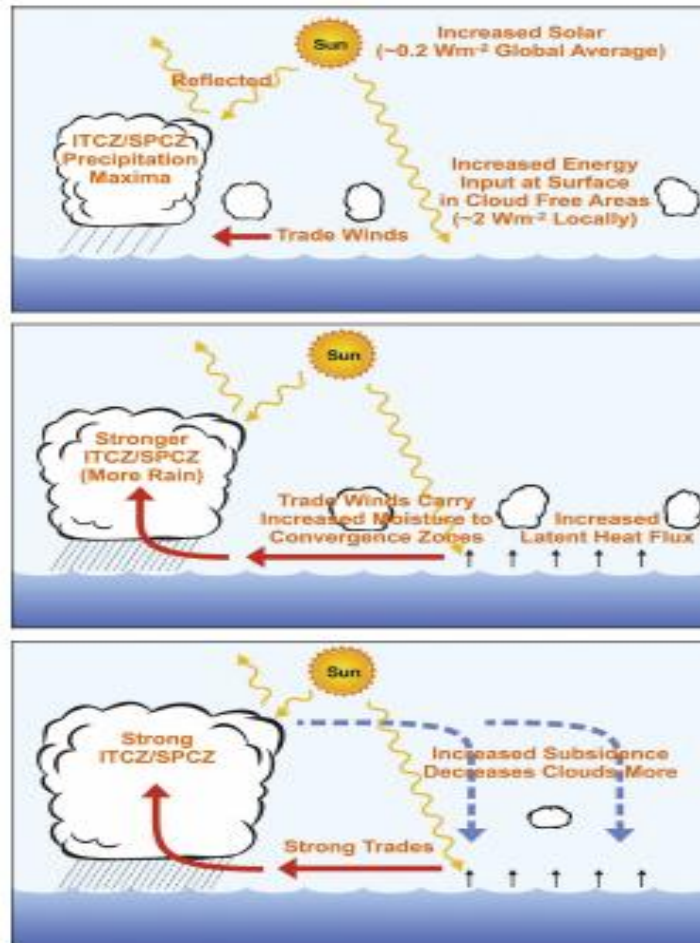


Fig.1.2.6: Schematic showing proposed mechanism involved in the Pacific (N-S direction) during peak solar years.

(Meehl et al., 2008)

free areas of the subtropics increase energy input into the surface that 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 east-west Walker circulations and consequently the north-south Hadley circulation. Thus there are stronger trades 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. It produces consequently even fewer clouds and

more solar radiation making it to the surface (Fig. 1.2.6). Finally, it causes above normal pressure in the north Pacific over the places of Aleutian low.

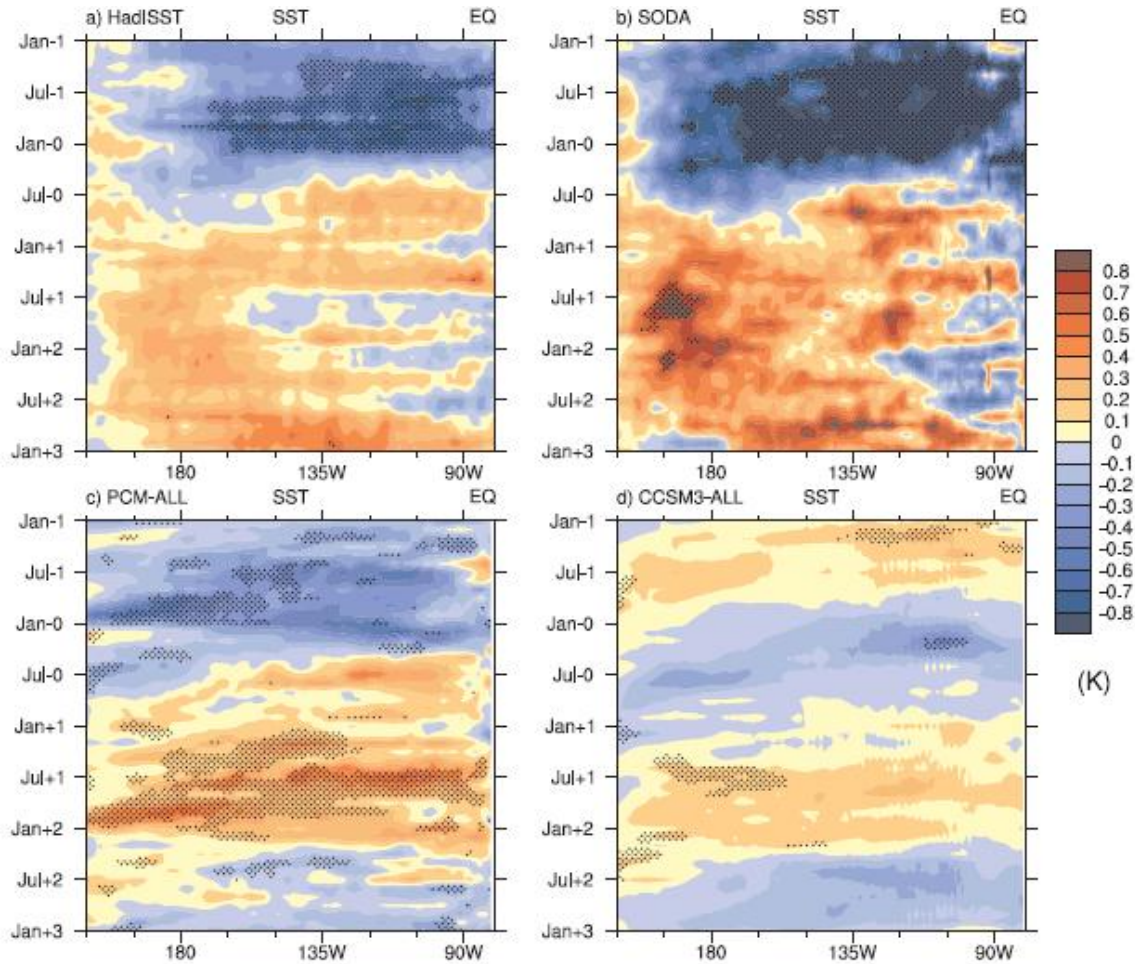


Fig.1.2.7: Composite SST with respect to climatology during peak SSN years (year 0), the year before (year -1) and the three years following the peak SSN years (years +1, +2 and +3): (top) observations from HadISST data (a) and from SODA ocean reanalysis (b); (bottom) composite of four model simulations with PCM (c) and composite of five model simulations with CCSM3 (d).

(Meehl and Arblaster, 2009)

Meehl et al. (2008, 2009) discuss a peak in irradiance at the peak of the decadal solar oscillation producing the La-Niña-like response and this is lagged after a year or two by an El-Niño-like event. Regarding the timing of this forcing and responses, Meehl and

Arblaster (2009) also presented results, but it shows that the results of these two models differ (as shown here in Fig.1.2.7). Among these two models, **Parallel Climate Model (PCM)** and **Community Climate System Model version 3 (CCSM3)** as seen in that figure, the former suggests closer fit to their two different observational results, one from HadlSST and another from Simple Ocean Data Assimilation (SODA) ocean reanalysis product. The discrepancies between the two model results in terms of tropical Pacific SST mainly arise from the different ENSO variability within the models (Meehl and Arblaster, 2009).

However, the observational results of van Loon et al. (2008), unlike Fig. 1.2.7, showed maximum cooling in peak years of DJF, preceded and followed by lesser cooling (and no warming). All these studies indicate that the timings of forcing and response are important to understand the ‘bottom-up’ mechanism.

Top-down:

According to this mechanism, direct variations in irradiance and indirect variations in stratospheric ozone in response to solar ultraviolet (UV) variability, changes the vertical and horizontal temperature structure, resulting in dynamical responses in the stratosphere and troposphere (Haigh, 1996; Balachandran et al., 1999; Shindell et al., 1999). Such changes in the thermal gradients and thus in the wind systems, which in turn lead to changes in the vertical propagation of the planetary waves that drive the global circulation. Moreover, the relatively weak, direct radiative forcing of the solar cycle in the upper stratosphere can possibly lead to a large indirect dynamical response in the lower atmosphere through a modulation of the polar night jet as well as through a change in the Brewer Dobson Circulation (BDC) (Kodera and Kuroda, 2002; Matthes et al., 2004).

Kodera and Kuroda, (2002) proposed a possible mechanism whereby the solar influence in the equatorial troposphere can originate from the equatorial stratosphere through changes in the meridional circulation. According to them, the solar heating anomalies that

changes the strength of polar stratospheric jet can influence the path of upward propagating planetary waves. These waves, depositing their zonal momentum on the poleward side of the jet, weaken the Brewer–Dobson circulation and thus warm the tropical lower stratosphere in solar maximum compared with solar minimum periods.

This type of communication of forcing from the stratosphere downward to the troposphere was also posed by Hameed and Lee (2005). They showed that the apparent penetration of northern hemisphere winter circulation anomalies from the stratosphere to the troposphere travels faster under solar maximum conditions as opposed to solar minimum conditions. The difference is more striking primarily during the QBO westerly phase: no stratospheric signals reach the surface when equatorial 50 hpa winds are from the west under solar minimum conditions. Gray et al. (2006) review recent studies that have shown that perturbations to the equatorial upper stratosphere can perturb the polar lower stratosphere and thus provide a route for QBO-solar modulation of the tropical middle atmosphere to influence the lower stratosphere via a so-called ‘polar route’ during winter.

Haigh et al. (2005) proposed through simplified global circulation model (GCM), (without ocean) that moist feedbacks as suggested by Meehl et al. (2003) are not a crucial component of the observed response in troposphere due to the Sun; it is eddy/ mean flow wind feedbacks that are the primary mechanism. Despite the presence of a uniform stratosphere, the lack of a stratospheric polar vortex, and the use of broad latitudinal-scale perturbations, it is possible to reproduce the tropospheric patterns. Such a study suggests that a detailed representation of the stratosphere is not necessary for understanding the tropospheric aspects of solar influence, although the source of the stratospheric heating remains an important factor. In their simplified atmospheric general circulation model Haigh et al. (2006) shows that imposed changes in the lower stratospheric temperature forcing lead to coherent changes in the latitudinal location and width of the mid-latitude jet stream and its associated storm-track, and that eddy/ mean-flow feedbacks are crucial to these changes. Hence, solar heating of the stratosphere may produce changes in the circulation of the troposphere even without any direct forcing below the tropopause.

They suggested that the impact of the stratospheric changes on wave propagation is key to the mechanisms involved.

Haigh (1999) presents atmospheric circulation model results for the influence of the 11 year solar cycle on the climate of the lower atmosphere. A pattern of response is found in which the tropical Hadley cells weaken and broaden, and the subtropical jets and midlatitude Ferrel cells move poleward for high irradiance. Such a pattern, as also seen in the modelling study by Haigh (1996), is presented here in Fig. 1.2.8, which is very similar to the observational results as shown in Fig. 1.2.5. The changes in dynamics cause subtropical warming and a characteristic vertical band structure of mid-latitude

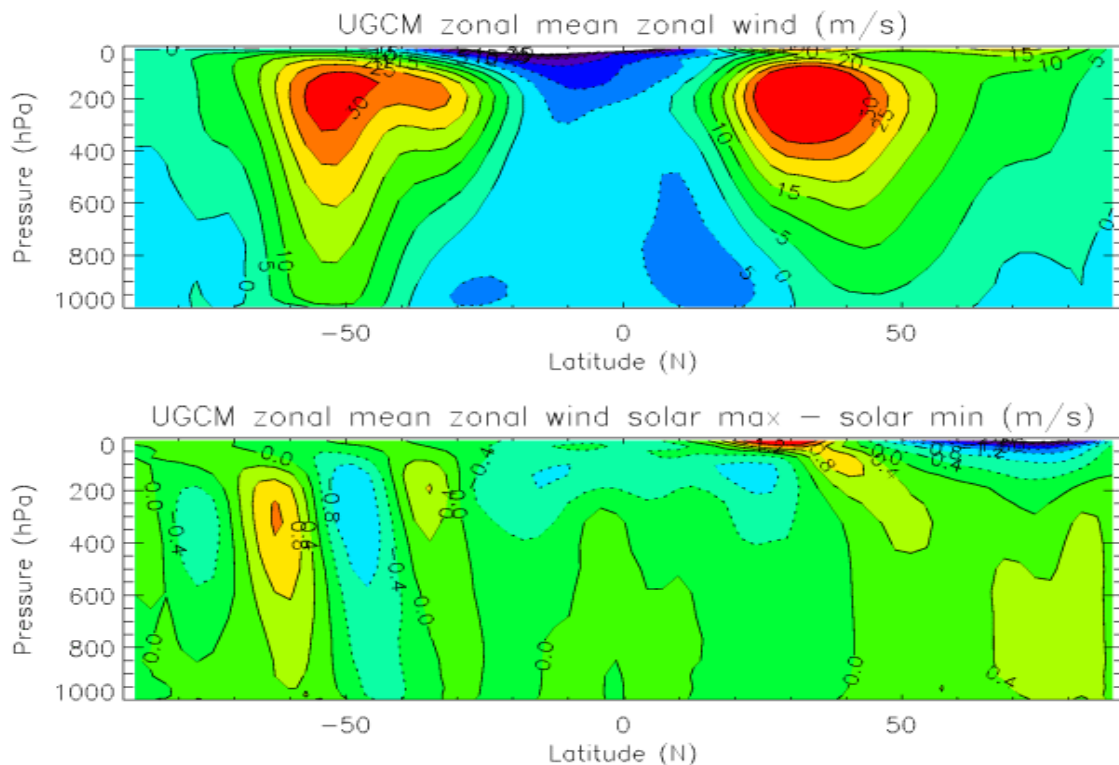


Fig. 1.2.8: Results of GCM simulation for zonal mean zonal wind for control run (top) and for difference between solar maximum and minimum (bottom)

[after Haigh, 1996]

temperature changes. Solar forcing in the lower stratosphere can change the strength and location of Hadley and Ferrel cell along with the subtropical jet have also been identified in the recent modelling study of Simpson et al. (2008).

There are strong similarities between the meridional structures of the annular modes in northern and southern hemisphere (Thompson et al., 2000), despite the sharply contrasting land–sea distributions and stationary wave climatologies of the two hemispheres; which supports some robust influence leading from the top. Baldwin et al. (2001) in their study discussed a dynamical mechanism that might communicate stratospheric circulation anomalies downward to the troposphere and surface via polar modes of variability. Such a pathway can provide a possible route of amplifying solar variability in the surface from the top.

Haigh and Roscoe (2006), through multiple regression analysis (using NCEP data during the 2nd half of last century), showed there is a clear solar influence on the NAO, but no statistically significant signals of solar forcing are found in either the NAM or SAM. However, when a new index, the product of the solar and QBO indices, is used there is a good correlation throughout the atmosphere in the SAM, and at the lower levels in the winter NAM. Thus they pointed out that solar stratospheric influence on the troposphere may well be through two different routes. The first of these is the influence of low latitude lower stratosphere heating on the Hadley circulation and mid-latitude eddies, which stimulate the NAO; the second is modulation, by a combination of solar and QBO forcing, of the polar stratosphere which influences the annular modes in both the northern and southern hemispheres and provides another potential route for the transfer of a solar stratospheric influence down to the troposphere.

- **Atmosphere-Ocean Coupling:**

Dima et al. (2005), using different SST datasets and applying different statistical techniques, identified two distinct modes of climate variability: one mode is associated with the solar sunspots cycle and defined by them as ‘the solar mode’; whereas the other mode is linked to atmosphere-ocean interaction and defined as ‘the internal mode’. They used the term ‘mode’ to refer a set of physical processes that are part of a large-scale coherent spatial structure and that have a quasi-periodic time evolution. According to them, the solar mode dominates SLP and upper atmospheric levels; whereas, in the

oceanic surface temperature, 'the internal mode' explains about three times more variance than that of the solar mode. For the purpose of quantifying the effect of Sun on climate, it is necessary to segregate the contribution resulting from internal climate variability as it may mask the signals of the sun (Rind, 2002). Hence it is really important to understand more about that internal mode, associated with the oceanic surface temperature, where the ENSO, no doubt plays an important role.

ENSO and Polar modes of variability:

Recent works have detected some signature of the ENSO in the polar modes of variability. Haigh and Roscoe (2006), in their multiple regression analysis, with data from the later half of the 20th century, showed that a strong anti-correlation exists between the SAM and ENSO in the lower troposphere. Carvalho et al. (2005), using data analysis for the period 1979 to 2000, observed that during austral summer (DJF), cold events of the ENSO are linked with the dominant positive AAO and vice versa. The alternation of AAO phases were also shown to be allied with the latitudinal migration of the upper level (200 hPa) STJ (around 45°S) and the intensity of mid-latitude polar jet (around 60°S). Positive AAO phases are associated with the pole-ward shift and weakening of the subtropical feature accompanied by an intensification of the high-latitude feature (shown in Fig 1.2.9).

It has been claimed that the northern stratospheric polar vortex is more perturbed and warmer during El Niño winters than during La Niña winters. Camp et al. (2007a) showed that during winter, warm-ENSO years are significantly warmer in the stratosphere at the NH polar and midlatitudes than the cold-ENSO years. Using GCM, Sassi et al. (2004) and Taguchi and Hartmann (2006) showed that the warming difference between El Niño and La Niña years is statistically significant and that SSW are twice as likely to occur in the El Niño winters than in La Niña winters, thus providing a possible connection between the polar stratosphere and ENSO. The ENSO is likely to be associated with changes in the stratospheric Brewer-Dobson circulation with anomalous tropical lower stratospheric heating tending to strengthen the circulation and produce anomalous

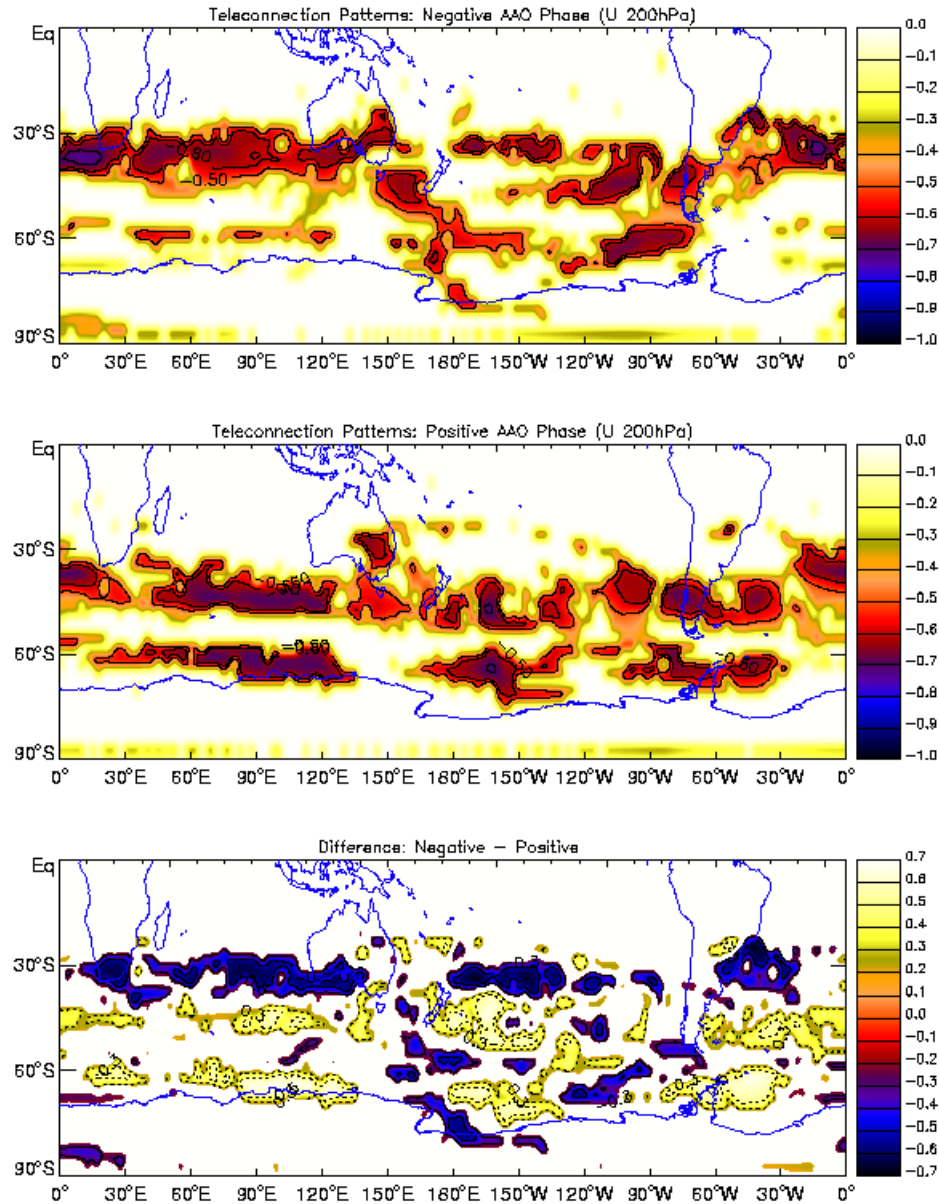


Fig.1.2.9: Patterns of teleconnection obtained for the zonal wind (200 hPa) anomalies during negative (top) and positive (middle) AAO events; difference between these two fields are in the bottom.

[Source: Carvalho et al., 2005]

adiabatic warming at high latitude, thus weakening the polar vortices (Haigh and Roscoe, 2006). The imprint of ENSO in the stratosphere consists of a weak and warm polar vortex in the Arctic stratosphere during El Niño, propagating from the middle to the

lower stratosphere during the course of a winter. Thus the ENSO might affect late winter via a downward propagation of stratospheric anomalies (Randel, 2004). The downward propagation of the ENSO signal from the upper stratosphere in January to the lower stratosphere in February and March is clearly observed and reproduced by models (e.g., Manzini et al., 2006).

ENSO and QBO

Thompson, Baldwin and Wallace (2002) observed that pronounced weakening of the NH wintertime stratospheric polar vortex tend to be followed by episodes of anomalously low surface air temperatures throughout densely populated regions such as the eastern North America, northern Europe, and eastern Asia that persist for ~2 months. Strengthening of the vortex tend to be followed by surface temperature anomalies in the opposite sense. During mid-winter, the quasi-biennial oscillation (QBO) in the equatorial stratosphere has

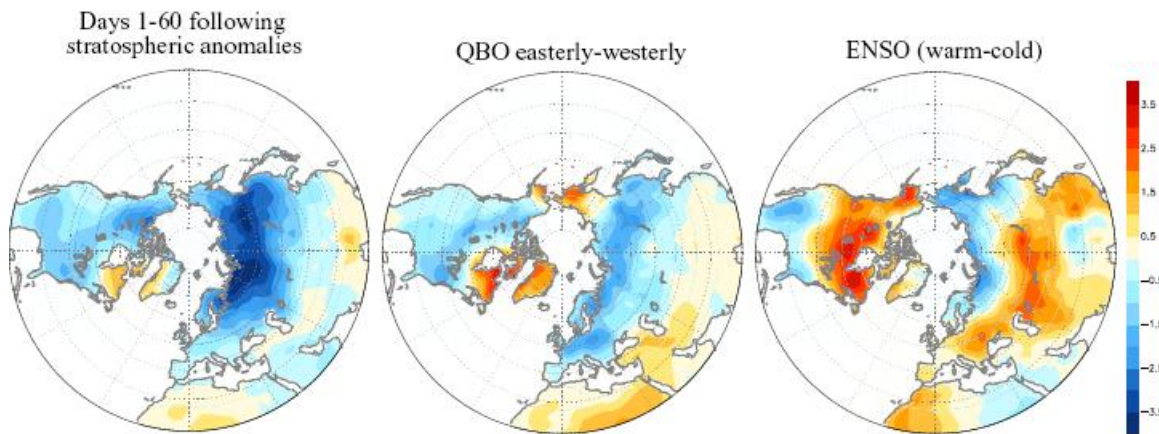


Fig.1.2.10: The difference in daily mean surface temperature anomalies between the 60-day interval following the onset of weak and strong vortex conditions at 10hPa (left panel); between Januarys when the QBO is easterly and westerly (middle panel); and between winters (January-March) corresponding to opposite phases of ENSO (right panel). Contour levels are at 0.5 C.

[Source: Thompson, Baldwin and Wallace, 2002]

a similar but somewhat weaker impact on the NH weather, presumably through its impact on the strength and stability of the stratospheric polar vortex: i.e., the easterly phase of the QBO favours an increased incidence of extreme cold events, and vice versa. The signature of the QBO in NH wintertime temperatures is roughly comparable in amplitude to that observed in relation to the El-Niño/Southern Oscillation phenomenon (shown in Fig. 1.2.10).

ENSO and Solar:

Though apparently correlation/regression analysis could not establish any direct connection between the 11 year solar cycle and the ENSO, recent works have also detected some solar signature on the ENSO. Defining high solar (HS) activity or low solar (LS) activity according to whether the solar index was higher or lower than the long-term mean value (Kodera, 2005, 2004; Kodera et al., 2007), Kodera et al. (2007), found that the ENSO related signal is confined in the Pacific sector during HS years. They also showed that ENSO-related variability extends into the Indian Ocean during period of LS activity which is not the case during higher solar activity. They suggested that such changes in intensity result from the shift in the location of the descending branch of the anomalous Walker circulation. The above result that the ENSO influence extends into the Indian Ocean through a modification of the Walker circulation is quite consistent with a recent model study using an atmosphere-ocean coupled general circulation model (CGCM) by Behera et al. (2006). In continuation of the study of Barnett (1989) who reported that tropospheric biennial oscillation (TBO) is modulated by an 11-year solar cycle, Kodera (2004) suggested that this modulation of the TBO is derived from a difference in the extension of ENSO-related variation into the Indian Ocean. TBO is the tendency for a relatively strong monsoon to be followed by a weaker one and vice versa, for the Asian-Australian monsoon system. Kodera (2002, 2003) found that during solar minimum conditions, the NAO signal is confined to the North Atlantic, while during solar maximum it extends over the Northern Hemisphere. Toniazzo and Scaife, (2006) also detected some footprint of the NAO in the ENSO. According to them, warm events of the ENSO are associated with negative phase of the

NAO and vice versa. More recently, Ineson and Scaife (2009) using a GCM of the atmosphere showed that there is a clear response of the ENSO in European climate via the stratosphere. This mechanism is restricted to years when SSW occurs, leading to a transition to cold conditions in the northern Europe and mild conditions in the southern Europe in late winter during El Niño years. Thus, all these studies are indicative of a global scale teleconnection pattern involving the ENSO, where the role of sun cannot be ignored.

Mendoza et al. (1991) considering the period of 1727-1983 showed that warm events of the ENSO are mainly occurring at decreasing solar activity and in minima of the decadal solar oscillation. Mann et al. (2005), through modelling studies showed considering the timescale of the last 1000 years that there is evidence of a cold event –like pattern during multi-decadal periods of high solar forcing. Their results also suggested that solar-driven changes in the tropical Pacific may have global impacts. As discussed earlier, using the method of solar max compositing of nearly 150 years of data, Meehl and co-authors in multiple papers (2007, 2008, 2009) have showed that for an increase in solar forcing, there is a cold event like pattern in the Pacific during DJF only.

Gleisner and Theijll (2003) following regression analysis and using F10.7 as a measure of solar indices found that there is a significant response of the troposphere to the 11-year solar cycle, and that the apparent solar signals are not merely due to chance covariations with the El Niño or major volcanic eruptions. Their study revealed that solar forcing is strongest in the tropics and at mid-latitudes and the tropical meridional overturning of the atmosphere is somewhat weaker and broader in latitudinal extent during HS years. According to them, solar signals in vertical velocity indicate a spatially heterogeneous modulation of both the Hadley and Walker-type circulation together with a modulation of the Ferrel circulation. Their findings have implications on the issue of how and where the sun exerts its influences in the climate system.

The quasidecadal oscillation (QDO) of 9- to 13-year period in the Earth's climate system has been found in the tropical Pacific Ocean similar to that governing the ENSO of 3- to

5-year period. This global SST and SLP patterns of variability of this QDO, and associated tropical warming, have been found fluctuating in phase with the ~ 11 -year-period signal in the sun's total irradiance during the twentieth century (White et al., 1997, 1998; Allan, 2000; White and Turre, 2003).

White et al. (2003a) and White (2006) found that the tropical global-average temperature of the upper ocean (0.1°C) is not driven by the ~ 11 -year-period signal in surface solar radiative forcing, but rather indirectly (via variable sensible-plus-latent heat flux) by a greater warming of the tropical troposphere temperature ($0.2\text{--}0.5^{\circ}\text{C}$) in response to the ~ 11 -year-period signal in the Sun's UV radiative forcing of the lower stratosphere temperature ($\sim 1.0^{\circ}\text{C}$) via absorption by ozone. Adding an 11-year-period cosine signal of amplitude $\sim 2.0 \text{ W m}^{-2}$ to the solar constant in the fully coupled ocean-atmosphere general circulation model (i.e., Fast Ocean-Atmosphere Model (FOAM)) of Jacob et al. (2001), White et. al. (2008) were able to simulate both the ENSO and the quasi decadal oscillation (QDO); with their model QDO, similar in patterns and evolution with the observed one. On the other hand, in its absence (11 year solar signal), the FOAM can only simulate the ENSO. The recent study of White et al. (2008) using the method of compositing and Singular Value Decomposition (SVD) of 9 solar cycles, covering period 1900-2000, even detected the phase-locking of harmonics of the ENSO time series with

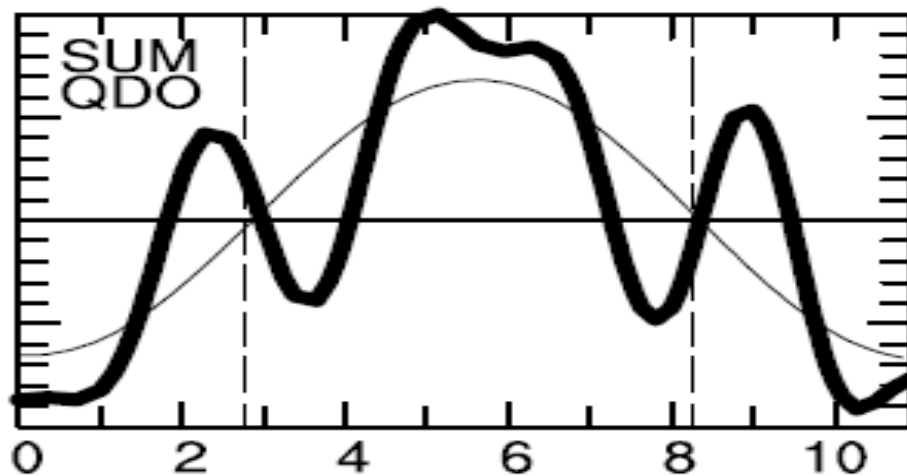


Fig.1.2.11: Phase locking of solar QDO and sum of ENSO signals.

(White and Liu, 2008)

the solar cycle resulting in a warm event like signal for about 3 years around the peak of the DSO with cold events approximately 2 years either side of the peak, and stronger warm events peaking 3-4 years before and after it (shown in Fig.1.2.11).

Apart from White et al., some other research works also detected decadal fingerprint in Pacific SST. Zhang et al. (1997) discovered that Pacific SST possesses ENSO-like inter-decadal variability with periods of about 11 years; analyzing 150 years SST data, Zhao et al. (2003) found that in addition to the inter-annual periodicity of ENSO, it also possesses inter-decadal variability; moreover, Chen et al. (2005) showed that the inter-decadal time scale variability has an important contribution to the thermal and dynamic variability of the tropics, even the entire Pacific.

In summary, these previous studies demonstrate that, whether the effect of solar variability in the troposphere is primarily driven from the stratosphere or forced by solar heating of the surface is a subject of major controversy. As outlined by Brönnimann et al. (2006), stratosphere-troposphere coupling may be a two way interaction and the possible downward propagation is normally preceded by an upward coupling and these two mechanisms may have different impacts on the tropical circulation. A number of works have been devoted in the past few years to understand these coupling mechanisms, but they have tended to raise more questions than they answer.

In this thesis, we will discuss some of these issues, showing how the Sun, ENSO and QBO all play an important part in regulating the climate of the troposphere. Using data analyses, mainly applying a multiple regression technique, we detect a solar link in surface climate and also discuss our results in the context of proposed mechanisms. We show the Sun plays a crucial role in ocean-atmosphere coupling, but this coupling appears to be disturbed during the later half of the 20th century.

Chapter 2: Regression Technique and Data

2. Regression Technique and Data

To understand the effect of solar variability on Sea Level Pressure (SLP) and Sea Surface Temperature (SST), we mainly used the technique of multiple regression. The main advantage of adopting this technique is that it can separate other factors that might influence/contaminate the results of solar signal. In our analysis apart from solar indices, the other independent parameters used include a linear trend, aerosol Optical Depth (OD) and El Niño & Southern Oscillation (ENSO). In the regression, the multiple regression code of Myles Allen (University of Oxford, UK, personal communication) has been incorporated. Alongside a brief description of the regression technique, we discuss in this chapter the different datasets used.

Regression Technique:

Multiple linear regression may be represented as:

$$\mathbf{y} = \beta \mathbf{X} + \mathbf{u}$$

Where, ‘ $\mathbf{y}$ ’ is a vector of rank n containing the time series of the data. ‘ $\mathbf{X}$ ’ is a matrix of order $n \times m$, comprising time series of m indices, which are thought to influence the data. ‘ β ’ is a vector of rank m that contains amplitudes of the indices, that we intend to estimate. ‘ $\mathbf{u}$ ’ is the noise term which is unobserved and may arise due to various sources (e.g., internal noise, all sources of observational error, un-modelled variability etc.). We estimate amplitudes of variability due to various climate factors using autoregressive noise model order one (AR(1)). Using a noise model of higher order does not make much difference to the results. Finally, using the Student’s t -test the level of confidence in the value of β derived for each index is estimated.

In this methodology, noise coefficients are calculated simultaneously with the components of variability so that the residual is consistent with a red noise model of order one. To elaborate: first the autocorrelation and variance of the noise are estimated from

the residual ($y-bX$, where 'b' is an estimate of ' β '); then a red noise function assumed to be of order one is fitted to the residual; afterwards, the values of 'b' and noise parameters are iterated until the noise model fits within a pre-specified threshold. By this process, it is possible to minimise noise being interpreted as a signal. It also produces, using Student's t-test, measures of the confidence intervals of the resultant 'b' values taking into account any covariance between the indices.

DATA:

The observational data provide the dependent parameter (y) for the regression, while the independent parameters (X) are prescribed by factors likely to influence y . The parameters are discussed briefly here in turns.

Dependent parameters:

The dependent parameters we considered in the multiple regression are SLP and SST. The sources of these data that we used are now described.

SLP Data:

For SLP we used data from <http://www.hadobs.org>. This is HadSLP2 an upgraded version of the Hadley Centre's monthly historical mean sea level pressure (MSLP) dataset which is based on a compilation of numerous terrestrial and marine data and has been described in detail by Allan et al. (2006). It covers the whole of the globe and the available time period is from 1850 to 2004.

Unlike HadSLP1, error estimates are available with HadSLP2 to guide the user about the regions of low confidence. Since measurement and sampling errors are large in the high southern latitude due to very low number of observations, caution should be exercised while using HadSLP2 around these regions. Such errors are small over well-observed ocean and over land. Generally, the estimates lie between the observational error

estimates of Kent et al. (1997) of 2.3 ± 0.2 hPa and those of Ingleby (2001) of 1 hPa over most of the ocean basins.

SST Data:

In the regression analysis for SST, we used two different sets of data, one is from Hadley Centre and the other is from NOAA.

Hadley Centre data have been obtained from <http://hadobs.metoffice.com/hadsst2/>. This (HadSST2) is a new sea surface temperature dataset, based on the data contained within the recently created International Comprehensive Ocean Atmosphere Data Set (ICOADS) and has been described in detail by Rayner, et al. (2006). The available time period is from 1850 to 2005. SST anomaly changes ($^{\circ}\text{C}$ relative to 1961-90) have been calculated within a 95% confidence interval. The global SST increase between 1850 and 2004 is $0.52^{\circ} \pm 0.19^{\circ}\text{C}$, with $0.46^{\circ} \pm 0.29^{\circ}\text{C}$ for the Southern Hemisphere, and $0.59^{\circ} \pm 0.20^{\circ}\text{C}$ for the NH.

The other SST data we used is from NOAA and is the extended reconstructed sea surface temperature data set (ERSST.v2). The first version (ERSST) was constructed using the International Comprehensive Ocean-Atmosphere Data Set (ICOADS) SST data and improved statistical methods that allow stable reconstruction using sparse data. ERSST.v2 is an improved extended reconstruction. In the reconstruction the high-frequency SST anomalies are reconstructed by fitting to a set of spatial modes. Compared to the earlier reconstruction, version 1 (v1), the improved reconstruction better resolves variations in weak-variance regions. It also uses sea-ice concentrations to improve the high-latitude SST analysis, a modified historical bias correction for the 1939-1941 period, and it includes an improved error estimate.

In the nineteenth century, the 95% confidence uncertainty for the near-global average is 0.4°C or more. Whereas, it is 0.1°C or less for the last half of the twentieth century and

near 0.2°C before 1950 (Smith et al., 2004). This is available from <http://www.cdc.noaa.gov/cdc/data.noaa.ersst.html> with more details on <http://lwf.ncdc.noaa.gov/oa/climate/research/sst/sst.html>. The available time period is from 1854 to 2007.

Potential problems with SST data:

The two SST datasets differ in their data sources, in analysis procedures and the corrections algorithm applied before 1940s. In terms of sources, the NOAA product uses

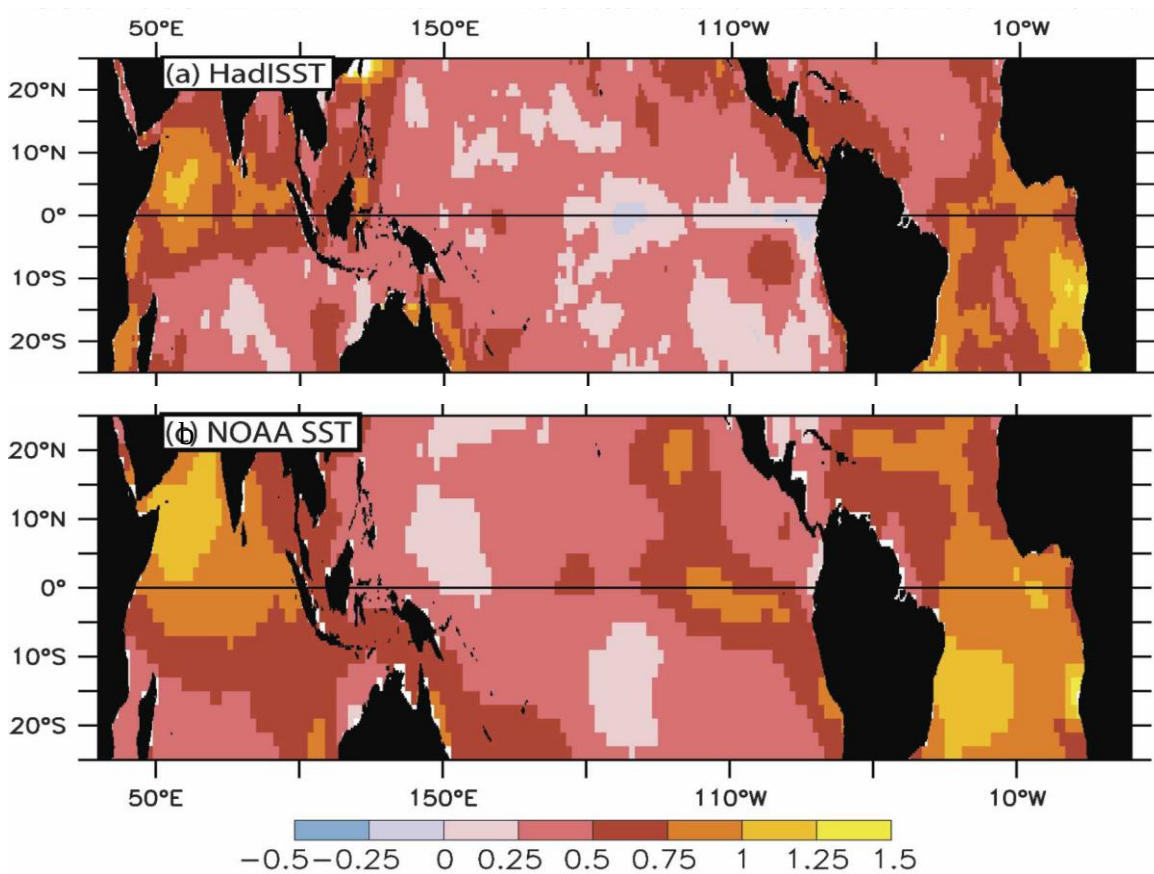


Fig.2.1: Linear trend in historical SST anomaly (in $^{\circ}\text{C}$ between 1880 to 2005), from two different reconstructions: (a) Hadley Centre; (b) NOAA. Monthly mean climatology removed before computing trend.

(Vecchi and Soden, 2007)

only in situ measurements while the Hadley Centre data include satellite-derived SST since early 1980s. The latter also include additional in situ observations from the U.K. Meteorological Office archive which are not included in the former (Vecchi and Soden, 2007). They also showed that, linear trends in tropical Pacific SST over the period 1880-2005 exhibit a “La Niña-like” structure when computed using the HADISST (Rayner et al., 2003) reconstructions. Whereas, trends computed using the NOAA Extended SST reconstruction (Smith and Reynolds, 2004) show an “El Niño-like” structure (see Fig.2.1). Vecchi et al (2008) also describe how inconsistencies of the various SST reconstructions in the equatorial eastern Pacific provide different results between the NOAA and HadISST data. We have carried out our analysis for both datasets but it did not alter our main results and here we mainly present results with the NOAA dataset.

Independent Parameters:

In the regression analysis, the different independent parameters used are the linear trend, Optical Depth (OD), monthly SunSpot Number (SSN) and ENSO. Time series of all these independent parameters used in the regression are shown in Fig. 2.2. On occasions, we also used the PDO, in place of ENSO in the regression and the time series is shown in Fig. 2.3. In parts of our study we have also used the QBO as an independent parameter. To represent solar variability we also used different Total Solar Irradiance (TSI) datasets. The sources of all these data are now briefly described.

Trend:

It is a linear time series that is intended to represent long term climate change. The focus of our work is on 11-year cycle variability so that the choice of long-term trend has essentially no effect on the derived signal.

SSN:

In our study, we mainly used monthly SSN to represent solar cycle variability which is available from

ftp://ftp.ngdc.noaa.gov/STP/SOLAR_DATA/SUNSPOT_NUMBERS/MONTHLY .

The main advantage of using SSN as a solar index is that the measurement is taken directly and also available for longer time period (here, since 1749). It is devoid of any trend and only captures the cyclic variability of the Sun. It is the most commonly used solar indices (for representing decadal variability) for the purpose of analyzing long term climate data. Usually, it is measured noting the number of sunspots and groups of sunspot in the surface of the Sun and can be expressed as follows. $R=k(10g+s)$, where, g is the number of sunspot group, s is the number of individual spots, k is a factor that varies with instrumentation and location, also known as the observatory factor and R is the relative sunspot number.

OD:

Stratospheric aerosol amount is highly variable as a result of sporadic volcanic eruptions. The aerosols affect the Earth's radiation balance, principally by reflecting sunlight to space and, secondarily, by absorbing upwelling terrestrial thermal radiation, and thus plays a potential role in climate forcing. Radiative forcing of the climate system by stratospheric aerosols depends primarily on the aerosol optical depth (OD) for solar radiation (Lacis et al., 1992).

Volcanic aerosols are one of the largest global climate forcing of the past century and hence it is important to account for OD appropriately. Moreover, the volcanic eruption of El Chichon in 1982 and Mt. Pinatubo in 1991 both coincidentally occurred near the peak of solar years and may contaminate the results of solar influence, if not taken into account properly. Hence OD has been used as one of the independent indices in our analysis and the data used are available from http://data.giss.nasa.gov/modelforce/strataer/tau_line.txt. which is up to 1999. It has been extended to 2005 with near zero values.

The method is described in Sato et al. (1993). During period 1850-1882, measurement is based only on crude estimates of aerosol optical thickness, following volume of ejecta from major known volcanoes, supported by qualitative reports of atmospheric optical phenomena. The period 1883-1959 has measurements of solar extinction, but during the time of principal volcanic activity (1883-1915) the data are confined to middle-latitude

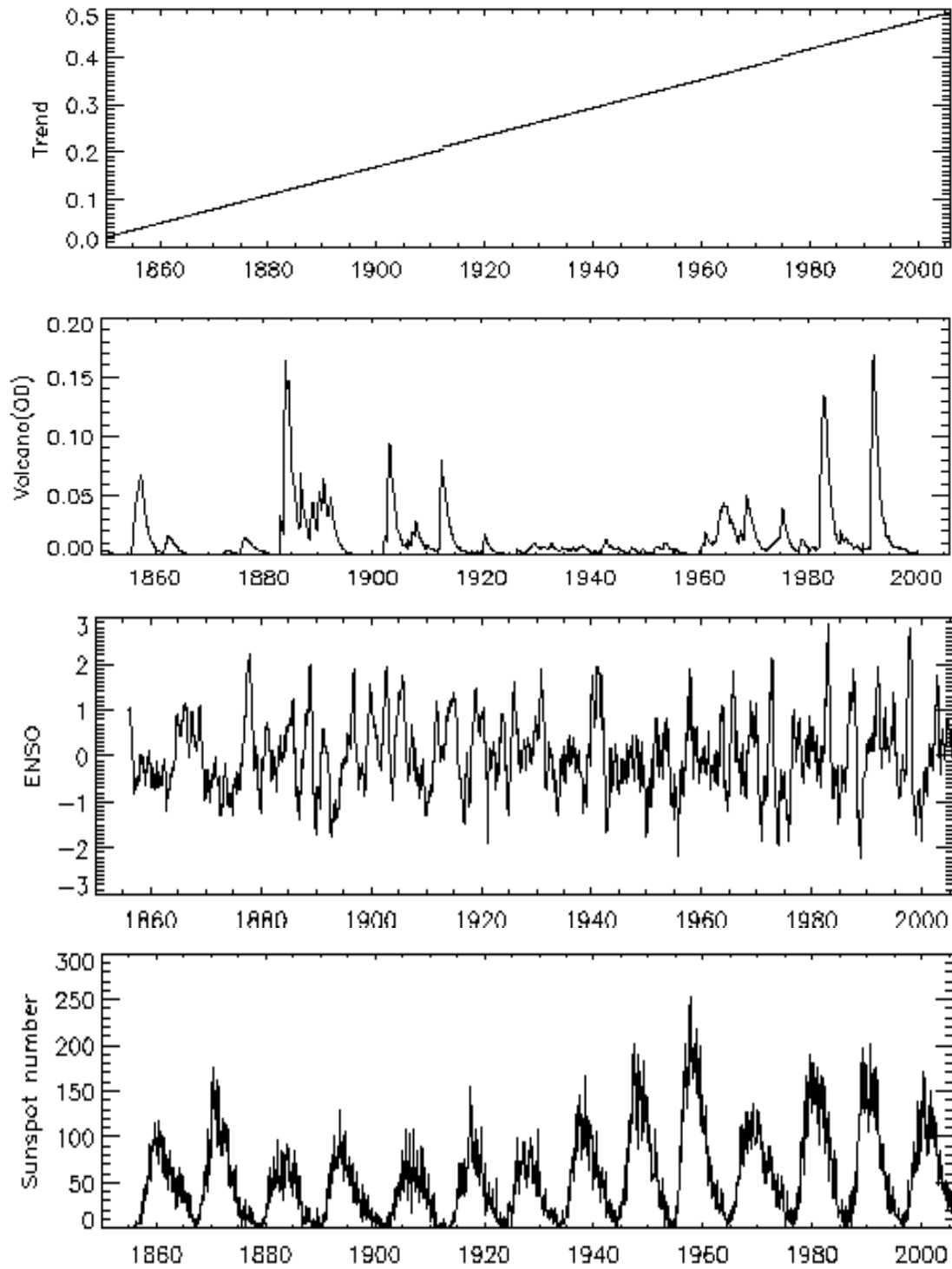


Fig. 2.2: Time series of Trend, Volcano, ENSO and Sunspot number

NH observatories. The period 1960-1978 has more widespread measurements of solar and stellar extinction, lunar eclipses, and some in situ sampling of aerosol properties. Whereas since 1979, precise widespread data from satellite measurements are available.

ENSO:

For ENSO, we used Niño 3.4 indices which is defined as the three month running mean of SST departures in the Niño 3.4 region (5°N-5°S, 120-170°W), calculated with respect to the 1971-2000 base period and is available from www.cpc.noaa.gov/data/indices. Departures are based on a set of improved homogeneous historical SST analyses (Extended Reconstructed SST –ERSST.v3b). The SST reconstruction methodology is described in Smith et al. (2008).

The Climate Prediction Centre (CPC), NOAA considers El Niño or La Niña conditions to occur when the monthly Niño3.4 SST departures meet or exceed $\pm 0.5^{\circ}\text{C}$ along with consistent atmospheric features. By historical standards, to be classified as a full-fledged El Niño or La Niña episode, these thresholds must be exceeded for a period of at least 5 consecutive overlapping 3-month seasons.

PDO:

The PDO dataset is available from <http://jisao.washington.edu/pdo/> and the time period is 1900-2006.

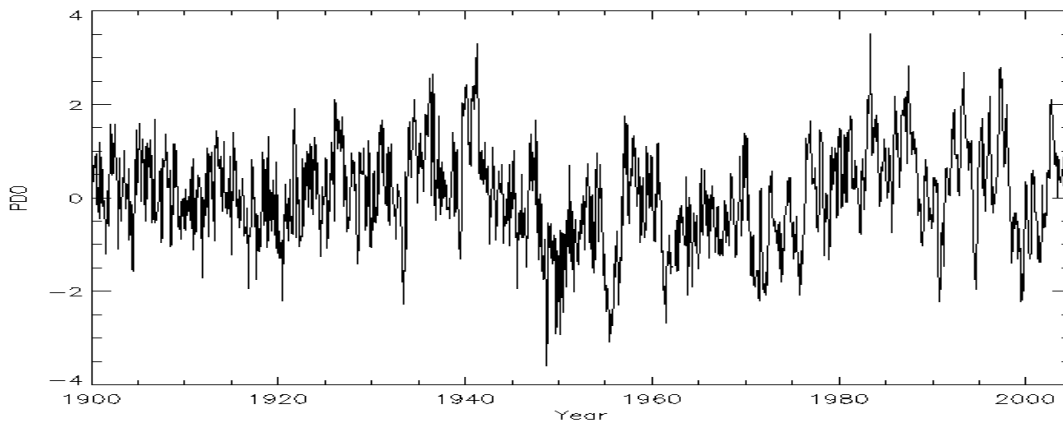


Fig. 2.3: Time series of PDO since 1900.

QBO:

The QBO is directly observed in operational wind measurements by rawinsondes at equatorial meteorological observatories. Ideally these measurements are made within 2 degree latitude from the equator. Barabara Naujokat of the stratospheric research group at the Free University Berlin has collected and processed radio sonde measurements from 1953 onward from Canton Island, Gan (Maledives) and Singapore (Naujokat, 1986; Labitzke et al., 2002). The respective locations and available periods for these three observing stations are: 02°46'S / 171°43'W, Jan.1953-Aug.1967; 00°41'S / 73°09'E, Sept.1967-Dec.1975 ; 01°22' N / 103°55' E, Jan.1976-Dec.2004. The QBO shows a high degree of zonal symmetry which allows the merger of the equatorial zonal wind profiles of these three individual stations into one dataset covering a longer time period. The data of QBO since 1953 at various levels from 10 hPa to 70 hPa is available from http://www.pa.op.dlr.de/CCMVal/Forcings/qbo_data_ccmval/u_profile_195301-200412.html .

Recent reconstructions of the QBO by Brönnimann extending back to 1900 are now available (Brönnimann, 2007). The reconstructions are based on historical pilot balloon data as well as hourly sea-level pressure (SLP) data from Jakarta, Indonesia. The latter were used to extract the signal of the solar semi-diurnal tide in the middle atmosphere, which is modulated by the QBO. The reconstructions are in good agreement with the QBO signal extracted from historical total ozone data extending back to 1924. Further analyses suggest that the maximum phases of the QBO are captured relatively well after about 1910.

Other Solar Indices:

Direct measurements of Total Solar Irradiance (TSI) from outside the earth's atmosphere started with the advent of satellite instruments in 1978. As the recent satellite period covers information only on the short term components of solar variability, to assess the potential influence of the Sun on a long term basis, it is necessary to know TSI further

back into the past. In reconstructing past changes in TSI, proxy indicators of solar variability, for which longer periods of observation are available, are used to produce an estimate of its temporal variation over the past centuries.

However the satellite data for recent decades suggests, significant uncertainties are present among all the existing satellite measurements (as shown in Fig. 2.4). This is relating to calibration of the instruments and degrading over time. It shows an attempt at compositing the measurements (shown in lower panel of Fig. 2.4) that produces the best estimate (Claus Fröhlich: <http://www.pmodwrc.ch/>).

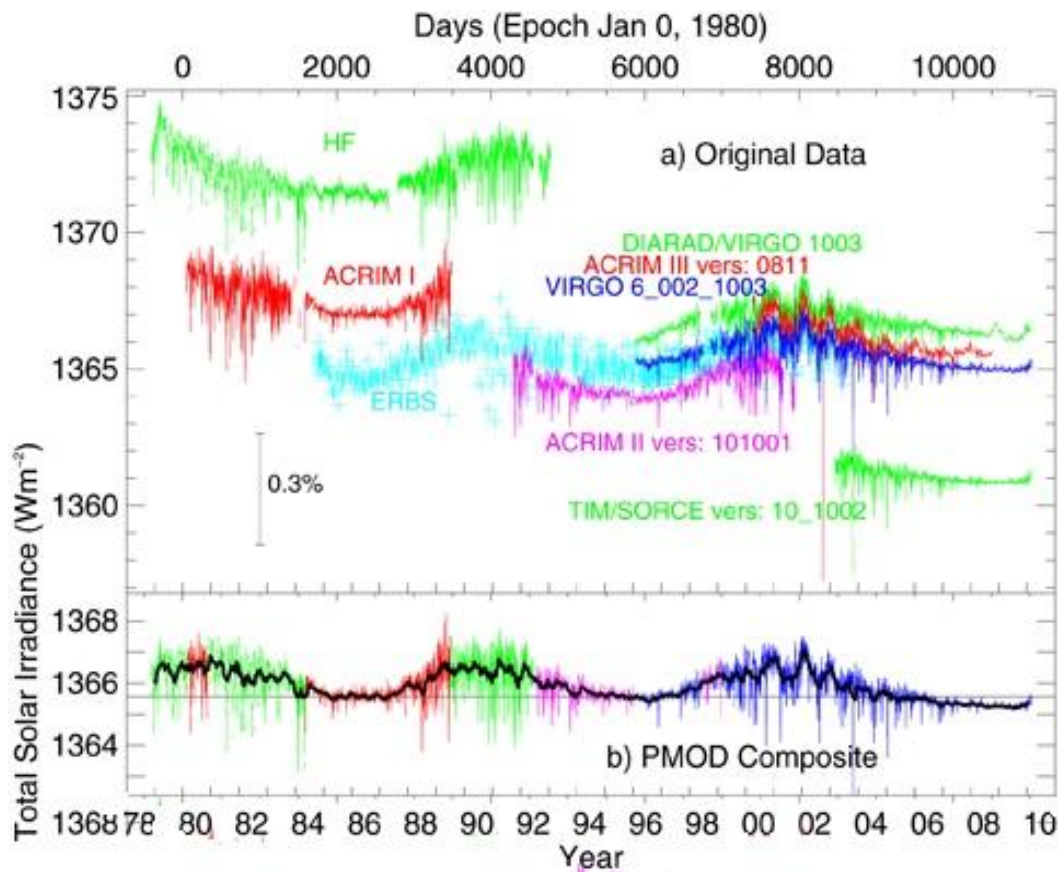


Fig. 2.4: Total solar irradiance daily average: top) measurement from various satellites; bottom) best estimate of TSI using compositing method.

(Source: <http://www.pmodwrc.ch/pmod.php?topic=tsi/composite/SolarConstant>)

Due to the problem of unexplained drift and uncalibrated degradation in the time series of the data obtained from satellite, there is a further uncertainty, in the existence of any underlying trend in TSI over the past 2 cycles. The compositing as shown in Fig 2.4 suggests that between the cycle minima occurring in 1986 and 1996, there is essentially no difference in TSI values. However, the results of Willson & Mordinov (2003) finds an increase in irradiance ($\sim 0.045\%$) between these two consecutive solar cycle minima. Such discrepancies are important because all the available TSI reconstructions are highly dependent on satellite measured data.

There are several different approaches taken for reconstructing the TSI, all employing a substantial degree of empiricism. The first approach uses SSN for TSI reconstruction (Solanki, S.K. and Krivova N.A. (2003), Hoyt D.V. and Schatten K.H. (1993), Foster, S. (2004) and Lean et. al (1995)). The second approach uses aa geomagnetic records (Lockwood and Stamper, 1999), while the third one involves using climate records (Reid, 1997; Beer et al, 2000). The final approach for TSI reconstruction uses numerical models of the Sun to simulate the variations in solar parameters (Sofia and Li, 2001). The review by Gray, L.J, Haigh, J.D. and Harrison, R. G. (2005) discusses the various approaches.

In most SSN-based methods, solar radiative output is determined by a balance between decrease due to the presence of sunspots and increase due to the development of faculae, the bright patches on the Sun's surface. The sunspot darkening depends on the area of the Sun covered by the sunspots, whereas, the facular brightening relates to a number of indices, which might include sunspot number (Lean et al. 1995), solar-cycle length, solar-cycle decay rate, solar rotation rate and various empirical combinations of all these (Solanki and Fligge, 1998; Hoyt & Schatten, 1993). In determining long-term variability, different techniques were employed in different reconstructions. For example, Lean et al. (1995) used sunspot cycle amplitude to determine long-term variability, whereas, Hoyt and Schatten (1993) used mainly sunspot length. Solanki & Fligge (1998) made an additional assumption concerning the long-term contribution of the 'quiet sun' based on observations on sun-like stars (Baliunas & Jastrow, 1990) along with the assumption that during the Maunder minimum the Sun was in a non-cycling state.

Considering longer time series of these reconstruction based on data from Wang et al. (2005), Lean (2000) and Foster (2004) as shown in Fig.2.5, it is clear that the estimates diverge as they go back in time, so that the value of solar radiative forcing over this period is highly uncertain. Moreover, this plot of Fig. 2.5 does not include some estimates which suggest a much larger difference between the present level and that prevailing during the Maunder Minimum in sunspots which occurred during the later part of the 17th century (e.g. Hoyt and Schatten, 1998). Such uncertainties are likely to affect studies relating to solar-climate research.

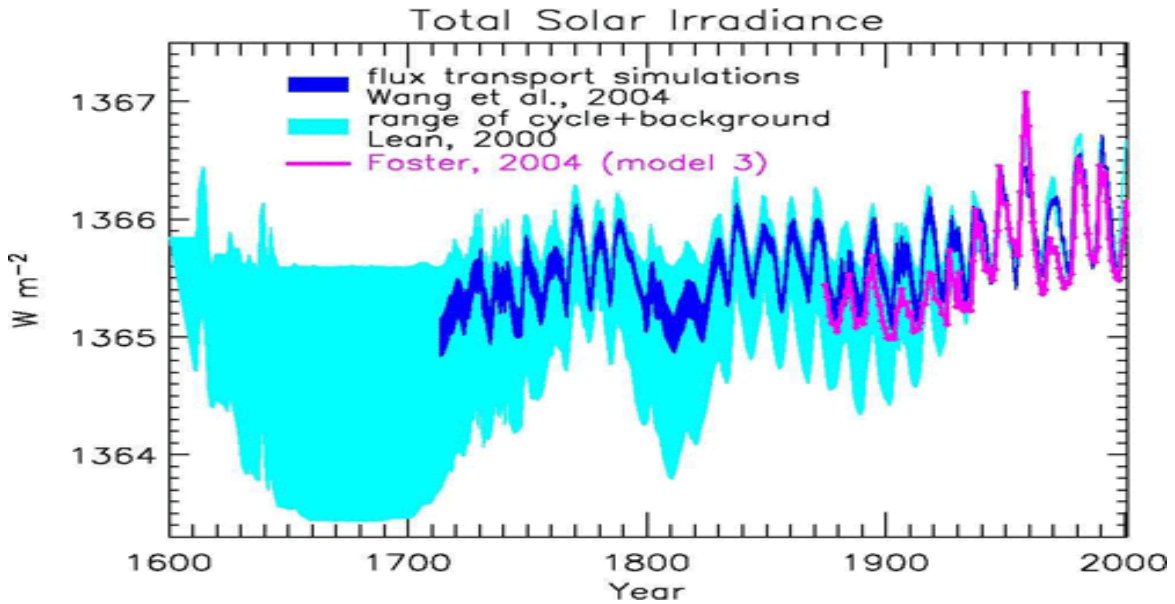


Figure 2.5: Reconstructions of total solar irradiance since 1600 by various authors.

(Lean et al., 2005)

In our analysis apart from SSN, we used three other TSI viz. Solanki & Krivova, Foster and Hoyt & Schatten. The time series of these TSI are shown in Fig. 2.6.

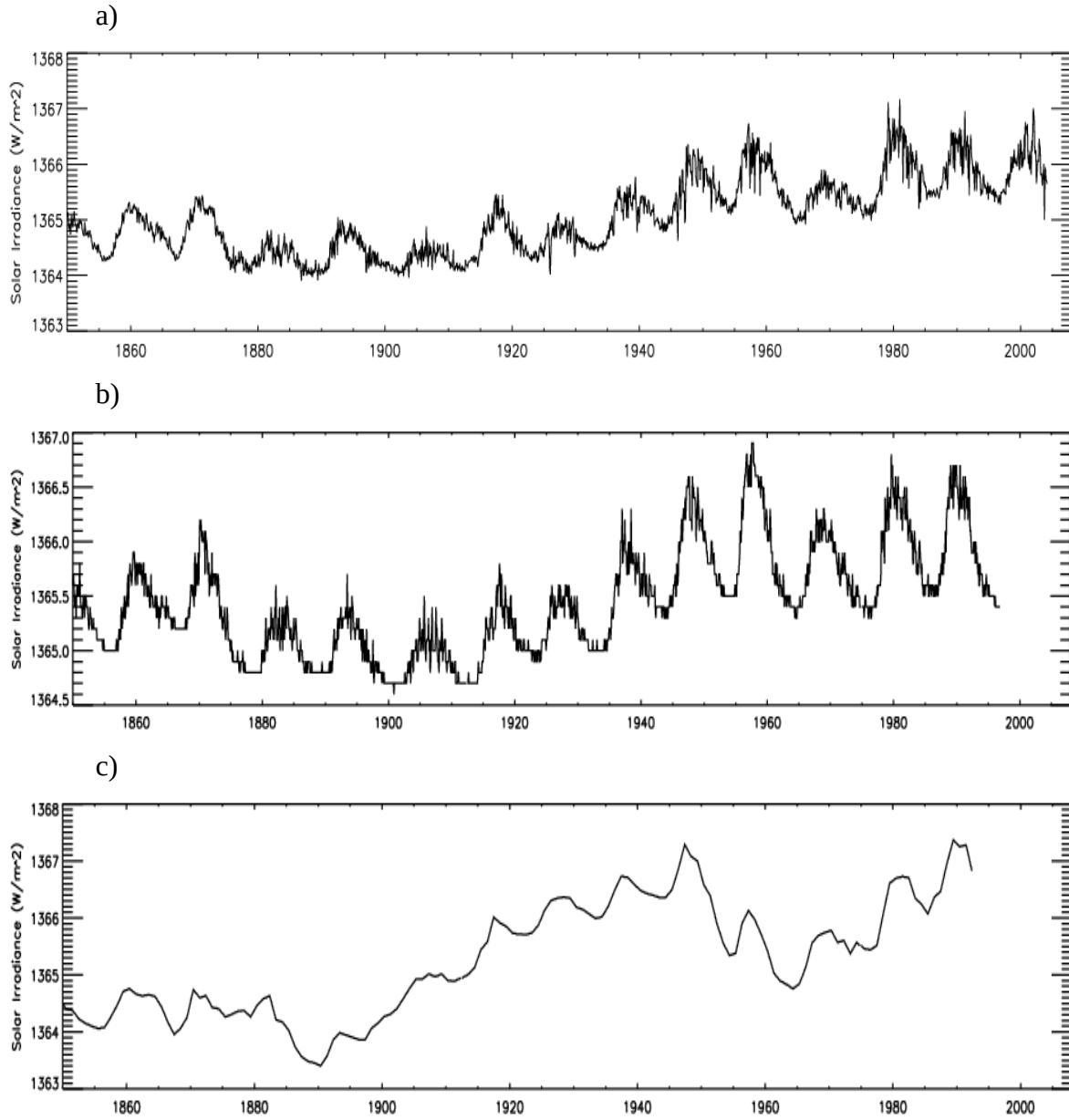


Fig.2.6. Time series of different TSI reconstructions: a) Solanki & Krivova; b) Foster and c) Hoyt & Schatten

The TSI of Solanki & Krivova is mainly based on facular brightening and relative SSN (Solanki et al. 2003). Whereas, Foster's TSI is mainly based on observation on sunspots (Foster, thesis, 2004). The TSI from Hoyt and Schatten mainly considers solar cycle length and thus mainly captures long term solar variations rather than the 11 year solar

cycle variability. It can be found from <http://climexp.knmi.nl/data/isolarconstanths.dat>, also in Hoyt et al, (1993).

Several earlier studies mentioned about uncertainty in using appropriate TSI in climate research (Mann, 2005; Haigh, 2000; Brönnimann, 2005). Stott et al. (2001) discussed the differences for their model experiments using two solar forcing dataset, Lean et al (1995) and Hoyt and Schatten (1993). The Hoyt and Schatten forcing provides a closer fit of the model response to the observed maximum of warming that occurred in the 1940s (Stott et al. 2001; Smith et al. 2003). Apart from TSI the other measures of solar activity which include SSN, solar flux at 10.7 cm and aa index etc also vary widely. For example, F10.7 cm flux and SSN reached their peak values over the last century at the solar max of 1958, whereas, the aa index showed its highest value during 1990. This is because the aa index shows the 11-year cycle imposed on a longer-term modulation, whereas, the SSN return to essentially zero at each solar minimum (Lean & Rind 1998). Such differences between the solar forcing datasets are important for detection/attribution of climate forcing parameters (Meehl, 2002; Haigh, 2006).

For statistical analysis of 155 years of data we are restricted for an indicator of solar activity to sunspot numbers, TSI reconstructions or some geomagnetic indices. The focus of this paper is on 11-year cyclic variability of the Sun, so that the accurate representation of longer term secular variability is less important. In our whole analysis, because we restrict to time period since 1850, the uncertainty in TSI during earlier period as seen in Fig. 2.5 does not have much effect on our results. Moreover, as we focus on solar variability on solar cycle timescales without the effect of any underlying long-term variations in solar irradiance, the uncertainties in TSI regarding trend in recent decades, as we mentioned before, may not have much implications for our results. Overall, our whole work of data analysis, mainly involving SSN or appropriate TSI to detect 11 year solar cyclic variability is free from many of the biases discussed above and hence is powerful in detecting an effective signal.

Chapter 3: Solar signals in SLP and SST- 150 years

3. Solar signals in SLP and SST - 150 years

The main purpose of this thesis is to understand the nature of 11 year solar cyclic variability on SLP and SST. In this chapter, the analysis was mainly carried using data sets covering 150 years. Two methods were employed to detect the solar signal in SLP and SST: one is the method of solar max compositing and the other is multiple regression analysis. First, we discuss the solar signal derived using solar maximum compositing. This is followed by the results from the regression analysis and an investigation into the sensitivity to the choice of solar index.

3.1. Solar signals using solar max compositing

This method first requires that the year of solar max is identified from each whole 11 year solar cycle. In detecting solar max years, usually the years with the highest value of annual average sunspot numbers (SSNs) are chosen. Then the mean value, over all the peak solar years, is calculated for each meteorological parameter of interest (here, SLP or SST). The anomaly of that mean value relative to that for some reference value is identified as the response to enhanced solar activity. The robustness of the signal can be examined through significant testing.

Here, we discuss the solar signal in SLP and SST, during northern winter, following the method of solar max compositing, used by van Loon et al. (2007) (subsequently vL07) and Meehl et al. (2007). In terms of the reference value, these two studies differed: the former used the climatology, which is the mean value of December-January-February (DJF), during 1950-79; whereas, the latter used the composite of solar non-peak years (the mean of their DJF) over the whole data period. Here, we compare their results with some of our results following their method, using different datasets.

3.1.1. Signals on SLP

Studying the variations of the semi-permanent pressure systems, the north Pacific High (PH) and the Aleutian Low (AL), Christoforou and Hameed (1997), showed that the solar cycle can influence their locations to a large degree (Fig. 1.2.4b). During sunspot

maxima, the Center of Action (COA) of the AL pressure system moves west by as much as 700 km (alongside a northward movement), while the COA of the PH pressure system moves north by as much as 300 km. During sunspot minima, the COA for these pressure systems moves in the reverse direction. Movement of the COA of these important pressure systems has a significant impact on storm trajectories which subsequently cause large anomalies in regional climate. Apart from varying position, the AL also exhibits a significant rise in intensity during higher solar years. There is on an average 1.6 mb difference in intensity in the AL, between years of extreme solar activity (Fig. 1.2.4a).

Compositing 11 peak solar years JF (January-February), *relative to the climatology* (1950-1979, DJF), vL07 detected a solar signal in both SLP and SST in regions of the Pacific Ocean. Within the SLP dataset (HadSLP1), dating from 1871 till 1989, they found a very strong positive solar signal around the AL, with a secondary negative footprint in mid-latitudes, crossing the international date line (Fig. 3.1.1a).

First, we ensure that we can reproduce their results following their method, before carrying out a new technique. For SLP, using HadSLP2 (covering the whole of globe) and following their method, over the same time period, we detected a similar strong positive signal around the AL with a negative response (though not significant) north of the equator, covering the international date line (Fig. 3.1.1b). In both the above mentioned figures, (Fig. 3.1.1a & b) a strong negative signal around 70°W-100°W, 70-80°N is clearly noticed. Slight deviations of our results from those of vL07 is due to our use of different datasets. Throughout our whole analysis, (unless mentioned to the contrary) shaded regions are significant at 95% level.

We extended our study outside of the north Pacific region, to the whole of the globe. We identify a number of significant regions covering the latitude belt, 40°N to 40°S (weaker in the Pacific); as well as, identifiable NAM and SAM features in polar regions.

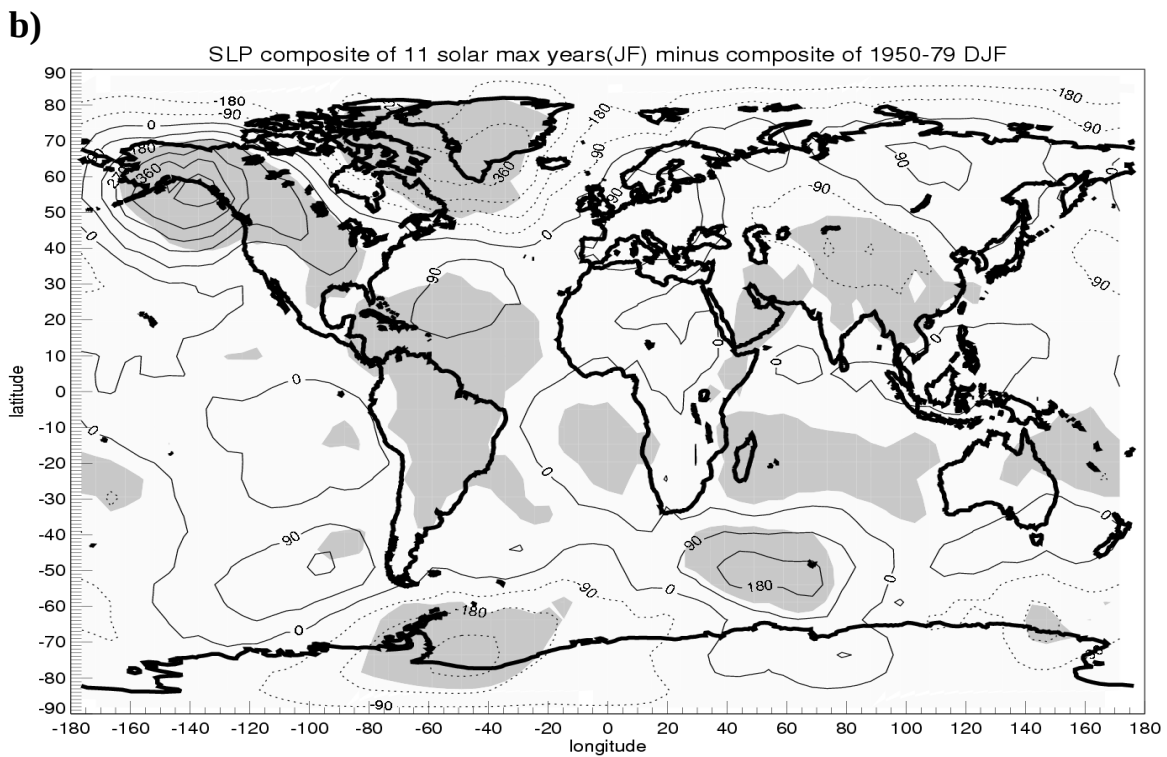
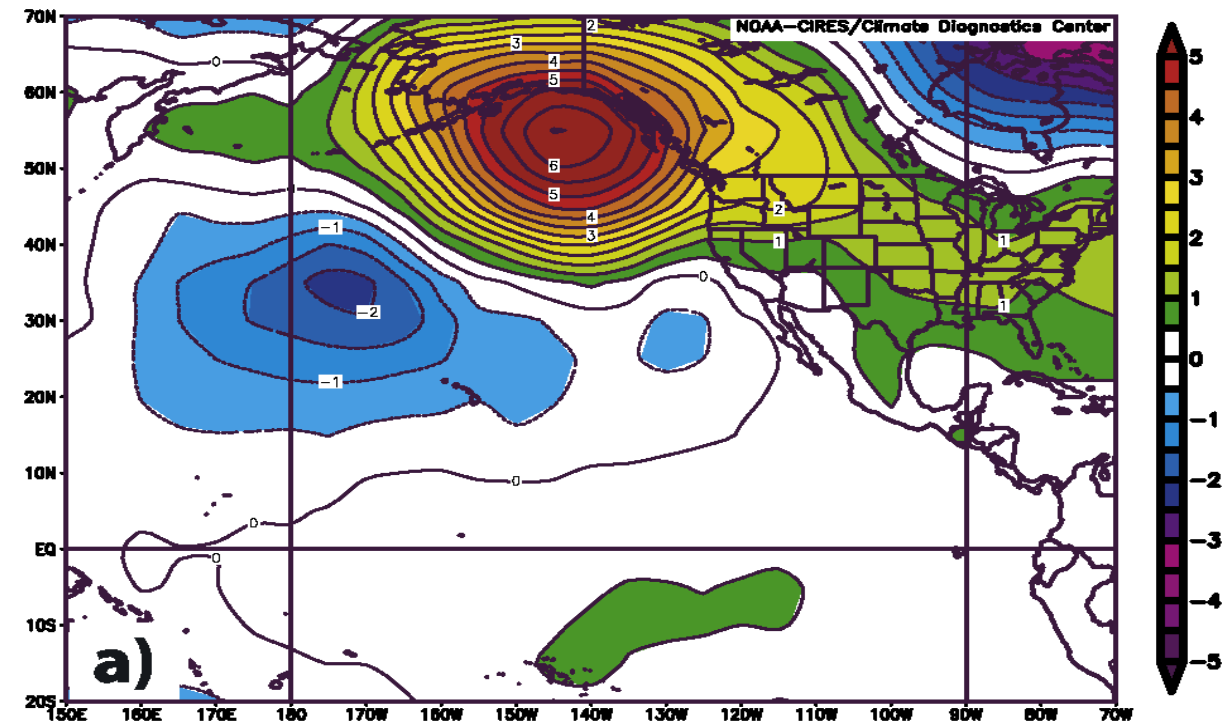
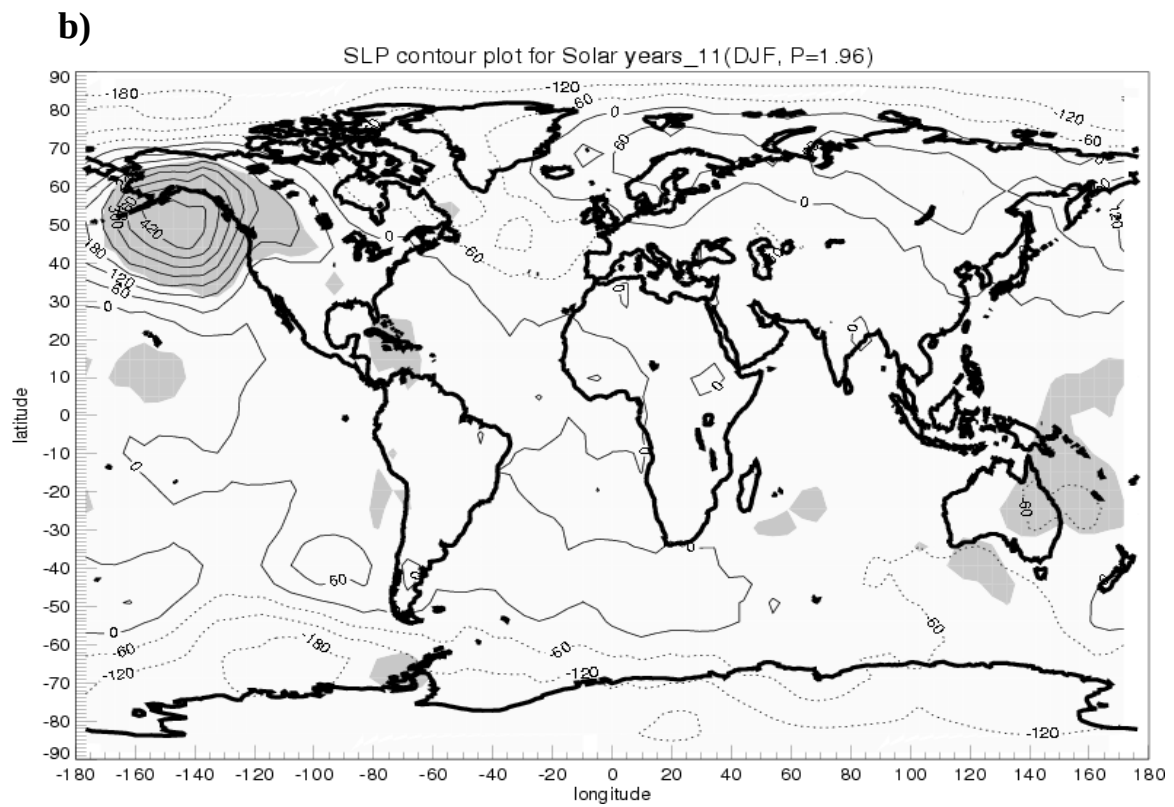
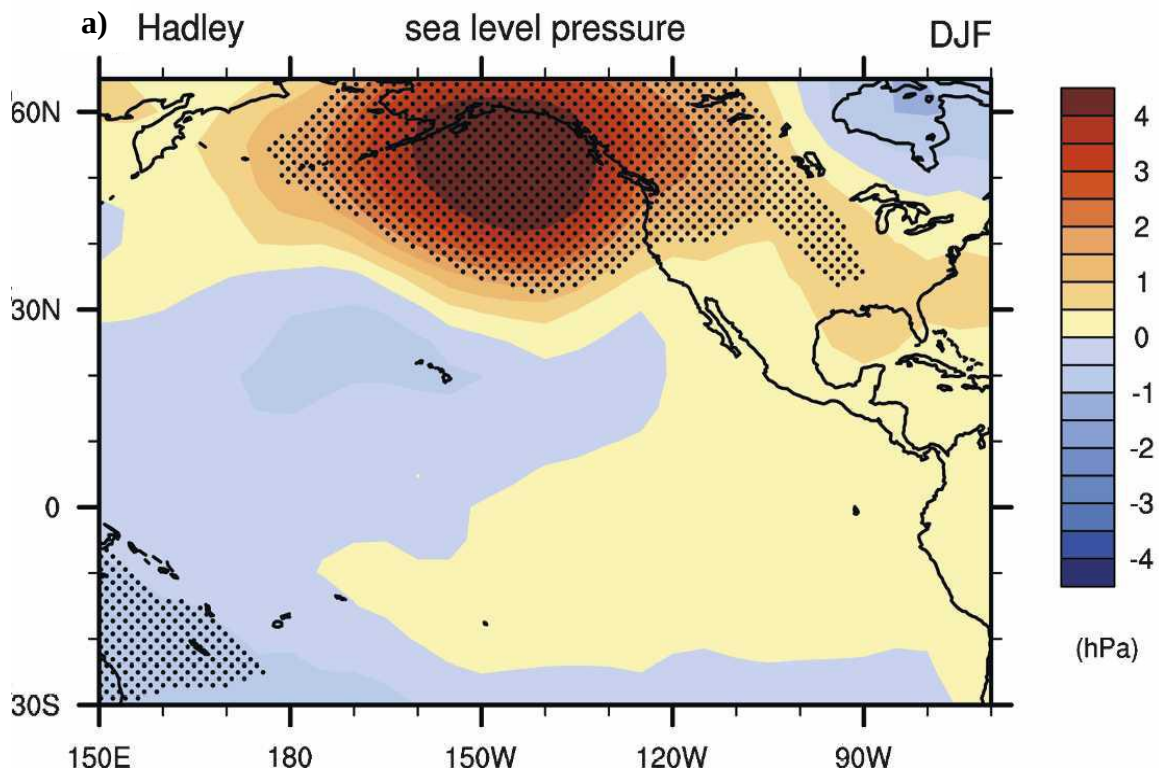


Fig. 3.1.1: Solar signal for SLP using solar max compositing: a) van Loon et. al. (2007) in hPa, using HADSLP1 ; b) our analysis in Pa, using HADSLP2. Shaded regions are significant at 95% level.

Meehl et al. (2007) used a global coupled climate model to identify features of the solar signal in SLP and SST, using the method of solar max compositing. In their compositing, *anomalies in DJF during solar peak years are computed relative to the non-peak years*. Their results for SLP (Fig. 3.1.2a) are similar to those of the earlier study of vL07 (shown in Fig. 3.1.1a), but certain deviations, like shifting the zero line in the Pacific, are evident. First, focusing around the international date line, covering the eastern Pacific, we observe that the zero line has moved from 45°N (Fig. 3.1.1a) to 35°N (Fig. 3.1.2a) and from north of the equator (~5°N) shifted south of the equator (10°S). The signal, that was identified in Fig. 3.1.1a, in the Arctic has also weakened in the Fig. 3.1.2a. Since the same data source (HADSLP1) is used in both analyses, over the same time interval, such departures indicates that results might be sensitive to the way in which the compositing is done. Hence arises the question about the robustness of the signal and robustness of their proposed physical mechanism in which, the zero line played an important role. Moreover, in the tropics, though their detected signal on SLP was statistically insignificant, they proposed their mechanism based on that.

For SLP, using HadSLP2, we also followed the same technique of compositing, with data covering same time period, as Meehl et al. (2007) and could detect a similar signal, which is shown in Fig. 3.1.2b. The signal around 70°W-100°W in the Arctic has weakened and become insignificant (Fig. 3.1.2b compared to Fig. 3.1.1b).

Again we extended our study to the whole globe. Here, apart from a significant negative signal around Darwin, Australia (one end of the Southern Oscillation (SO) see-saw pattern), we could detect no significant regions in other parts of the globe, unlike Fig. 3.1.1b. Positive NAM and SAM features are still noticed here, but are found to be weakened compared to that of Fig. 3.1.1b. All such observations again indicate that signals detected by the method of solar maximum compositing are sensitive to details of the specification of the compositing.



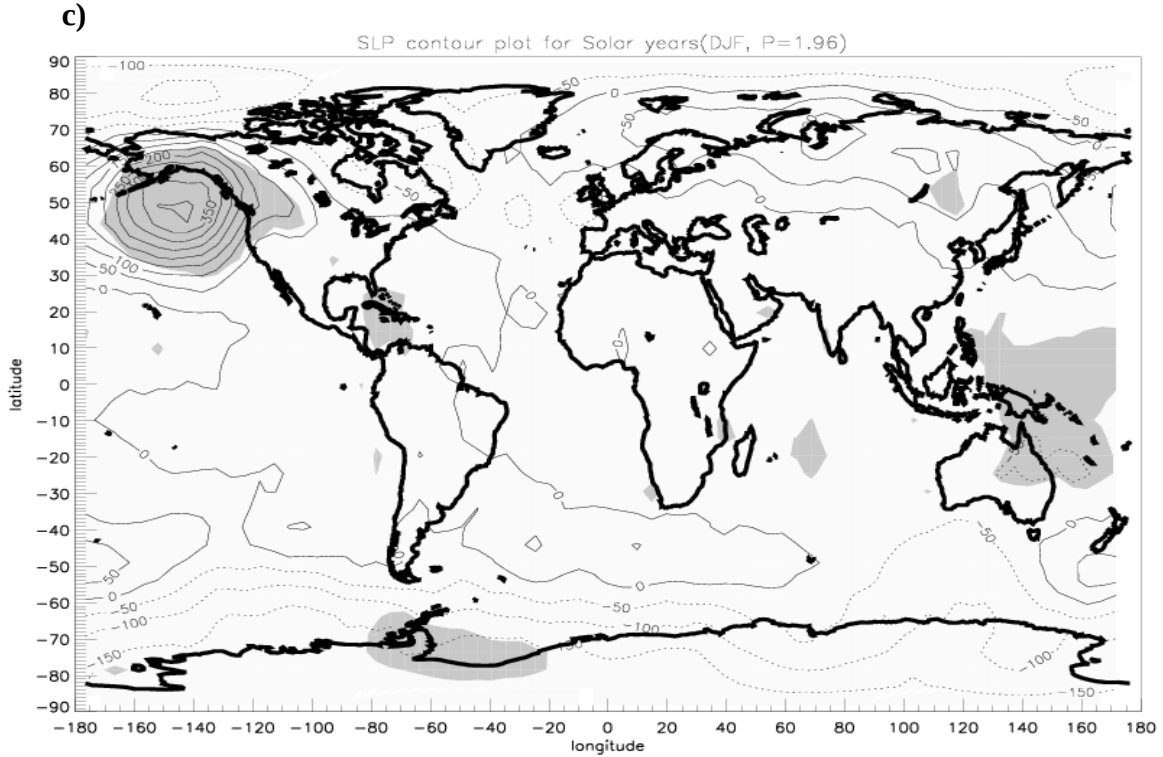


Fig. 3.1.2: Solar signal for SLP using solar max compositing: a) Meehl, 2007 in hPa, using HADSLP1; Our reproduction in Pa, using HADSLP2 for b) 11 solar max years between 1871 to 1989 and c) 14 solar max years between 1856 to 2004.

According to both analyses (Meehl et al. (2007) and vL07), the signals are largely independent of the number of solar cycles included in the analysis. To test that we followed the compositing method of Meehl et al. (2007) with data extends from the mid nineteenth century to 2004 (hence covering extra three solar max years) and showed that (Fig. 3.1.2c) it is in agreement with this conclusion.

3.1.2. Signals in SST

We repeated the compositing analysis of vL07 with the HadSST dataset (in the vL07 paper, both the NOAA and Hadley centre datasets are mentioned, though it emerged later that they actually used data only from NOAA) during Jan-Feb, over the same time period 1871-1989. vL07 detected a very strong tropical solar signal, resembling that of the

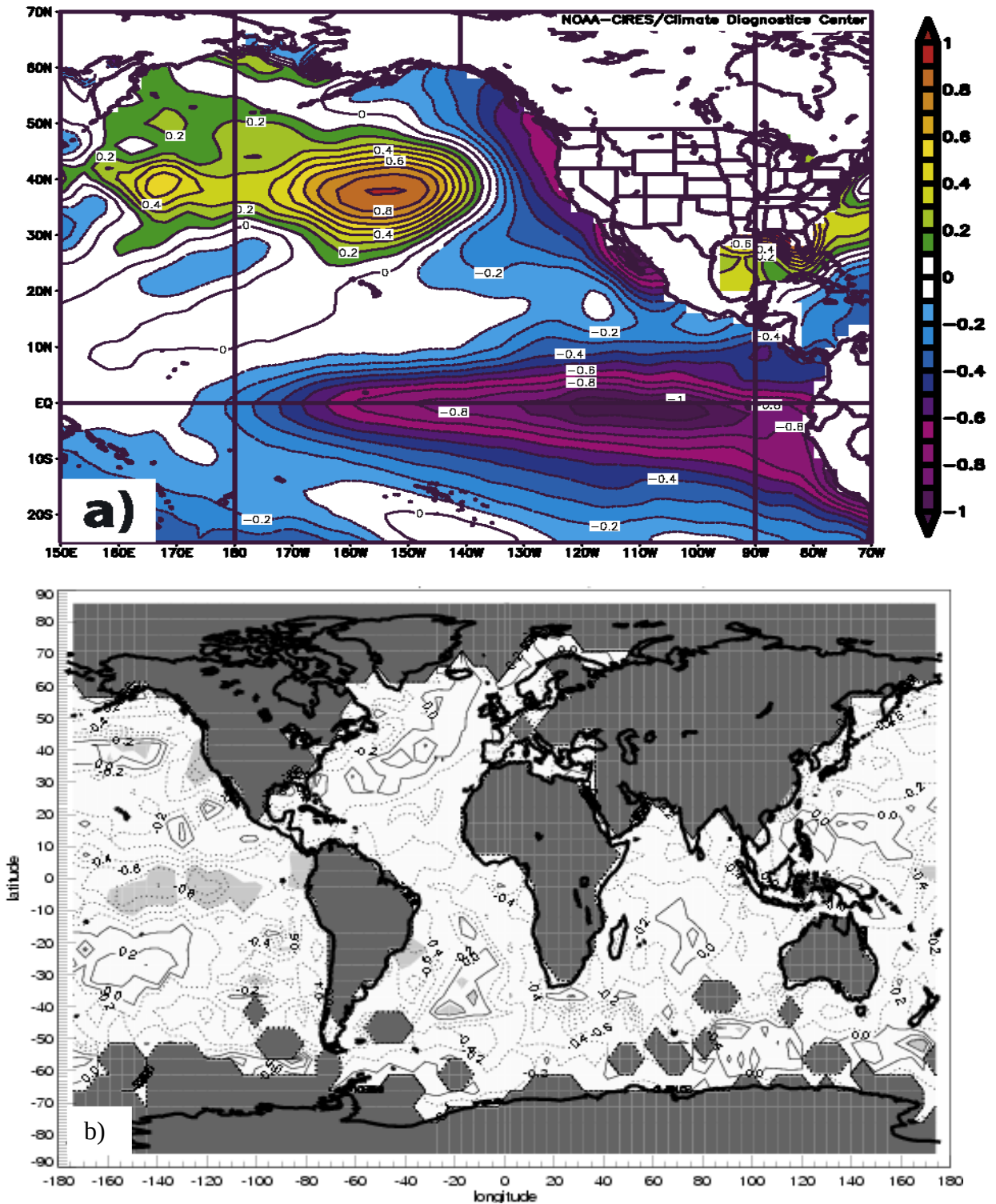


Fig. 3.1.3: Solar signal for SST using solar max compositing in $^{\circ}\text{K}$: a) van Loon et al., (2007) using NOAA; b) our reproduction for SST using HadSST. Shaded regions are significant at 95% level. Regions over land, along with data sparse regions have darker shading.

negative phase of ENSO (Fig. 3.1.3a). Moreover, in the mid-latitude of Pacific, they detected a significant positive solar fingerprint. For SST, using compositing of a total of 14 solar max years, we also captured a very strong negative solar signal around the tropics, covering the eastern part of Pacific (Fig. 3.1.3b). However, our signal around mid-latitude is weaker than that of vL07, which may be due to the use of a different dataset and it is worth mentioning that we are unable to detect any strong signal in other part of the globe. In Fig. 3.1.3b data sparse regions are shaded dark grey, along with the land surface.

Apart from the problem identified of results being sensitive to details of the method, solar maximum compositing is bound to have one very obvious limitation - it cannot identify the solar signal independently, excluding other major climate variability. Hence to extract the actual solar signal, without possible contaminating factors, we adopt the technique of multiple regression on both SLP and SST.

3.2. Regression of SLP

Multiple regression analysis was carried out on SLP, using the HadSLP2 dataset, during DJF, where the mean value during DJF represented the year. Amplitudes of the component of solar variability (using monthly SSN) on SLP, extracted alongside other independent parameters trend, OD and ENSO, are shown in Fig. 3.2. It suggests that a strong solar signal is present around the northern Pacific during DJF; the solar variability component even exceeds 4 hPa in the region of the AL. Including or not including ENSO, the solar signal in that region is robust. This is in agreement with Christoforou and Hameed (1997), who showed that the AL has higher values of pressure during solar max years, compared to those during solar min years. We also showed before that the results of solar max compositing (in the analysis of vL07, Meehl et al (2007) and also our result of 14 solar max years (Fig. 3.1.2c)) identified a similar solar response on SLP around the northern Pacific. Using these two distinct methods (solar max compositing and regression analysis) and comparing the results around the north Pacific lends confidence that the signal is robust and not simply an artefact of the analysis method used.

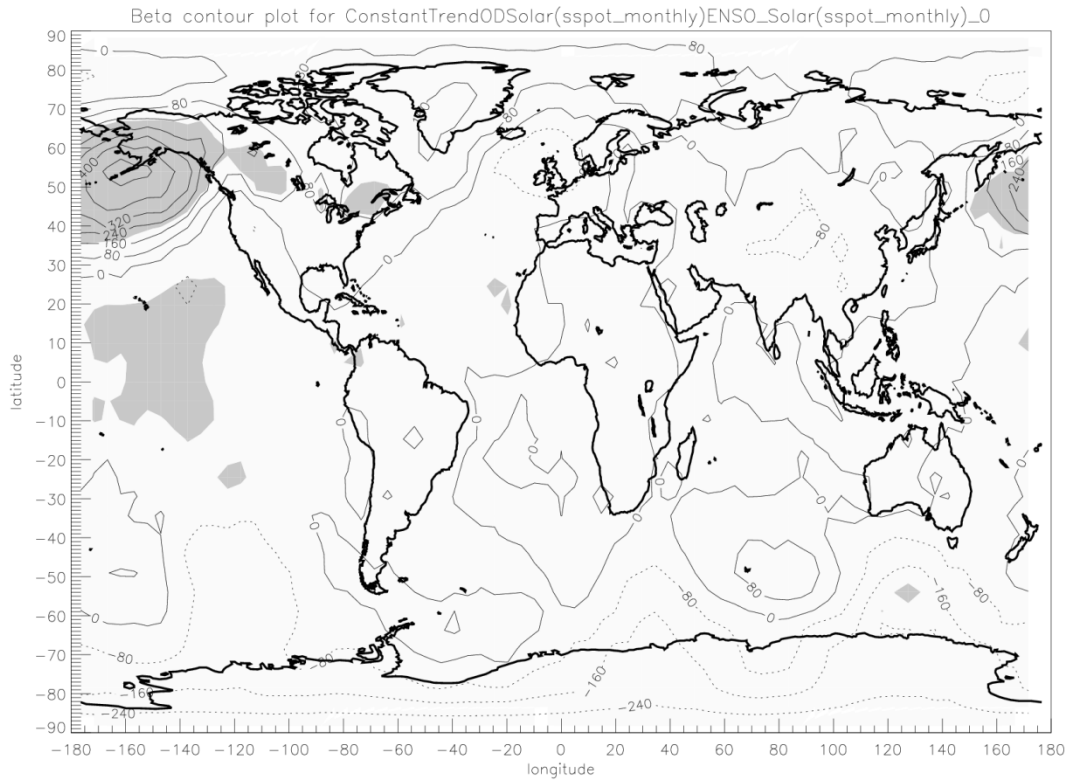


Fig. 3.2: Amplitudes of the components of variability of SLP due to solar (using monthly sunspot number) during DJF (in Pa). Other independent parameters used are: trend, OD and ENSO.

Our results also agree with the shifting of the PH to the south in solar min and to the north in solar max, as observed by Christoforou and Hameed (1997). It not only agrees with vL07 in terms of the SLP signal around the north Pacific with peak SSN but also continues throughout the following years of higher solar activity. It is also consistent with observational studies (Brönnimann et al., 2006b; Haigh et al., 2005; Crooks and Gray, 2005) and modelling studies (Haigh, 1996; 1999; Larkin et al., 2000; Matthes et al., 2006) which have indicated an expansion of the zonal mean Hadley cell, and poleward shift (with an weakening) of the Ferrel cell, at solar maxima.

We cannot, however, reproduce the pattern found by vL07, which placed a negative anomaly around 10-30°N in the mid-Pacific. Instead, we find a negative solar signal

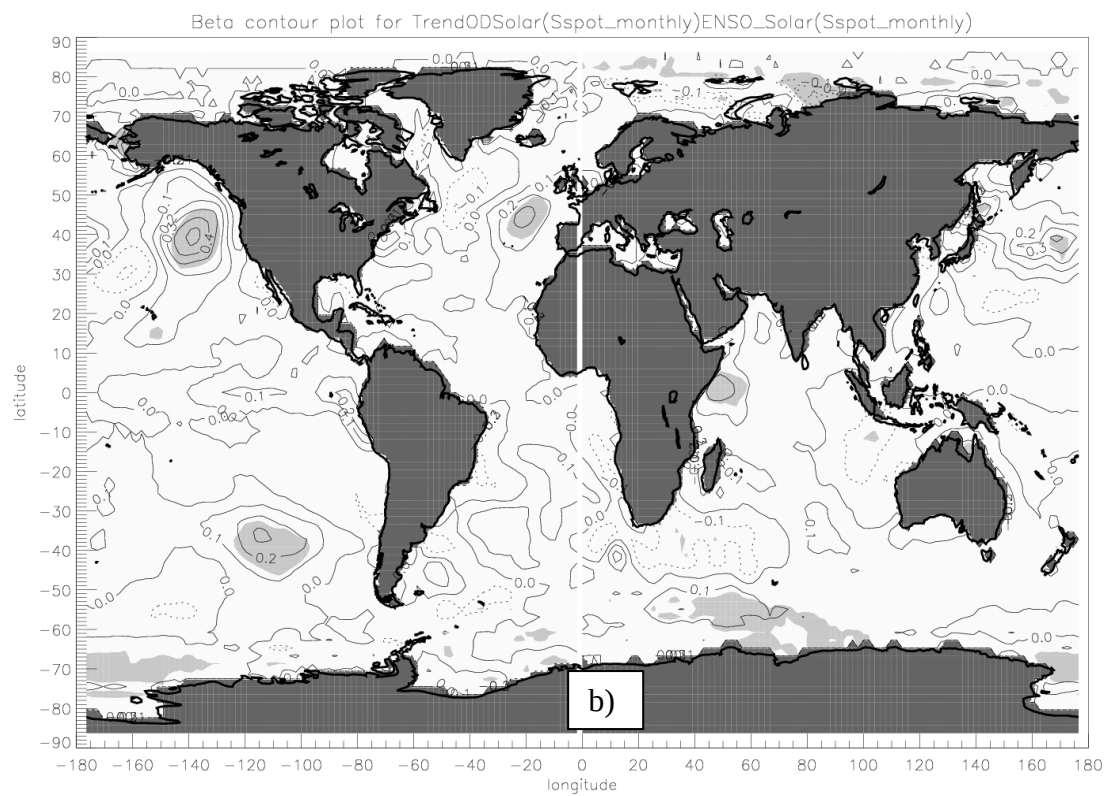
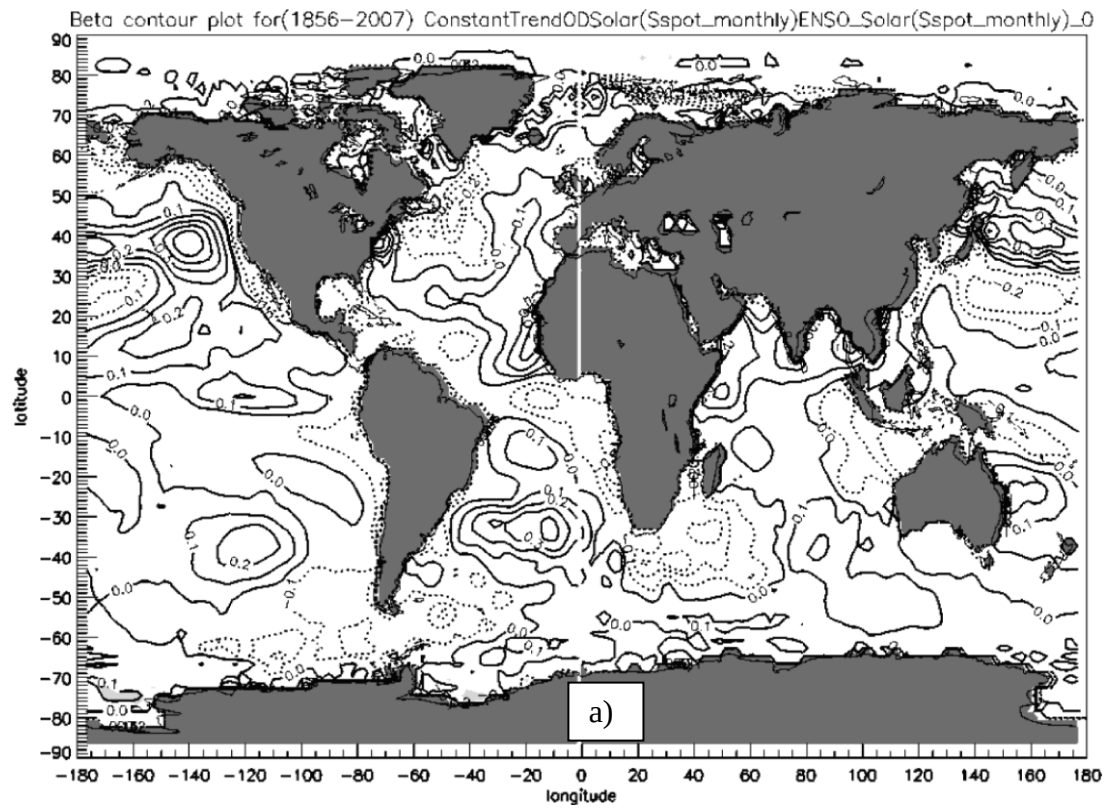
(small in magnitude but significant) in SLP around the eastern part of the international date line.

Moreover, our regression analysis fails to notice any detectable solar signal across the rest of the globe. Here we still observe positive SAM features (though not significant), as found in Fig 3.1.1b and 3.1.2c, but the NAM signal in Fig. 3.2 is of the opposite sign. Thus, our regression analysis also suggests that the signal detected using the compositing method can be spurious.

3.3. Regression of SST

Using the multiple regression technique, the amplitude of the component due to solar variability (using monthly SSN) on SST was carried out with the other independent parameters being the trend, OD and ENSO. Fig.3.3a. is the result using the NOAA dataset, during DJF, where each year corresponds to the average value during DJF only. The results of Fig 3.3b used the same dataset, but the regression was done using all months of the year (represented by monthly average values), removing the annual cycle from each time series. Figure 3.3c uses the same technique as Fig 3.3b on the Hadley centre data.

We detect (Fig. 3.3a) similar (though not that significant) +ve signal of the Sun during DJF, in the region of north Pacific as observed by vL07, using the compositing technique (Fig. 3.1.3) but no negative signal around the equator. A positive signal for SST around the Atlantic Ocean covering the mid-latitude is also noticed, which is stronger in the SH. Our signal for SST is similar in the north Pacific and equator during every season; but a significant response is noticed around the mid-latitude across the Pacific, when regression was carried using all month of the year, removing annual cycle (Fig. 3.3b and c). Including or not including ENSO, as a regression index, the solar signal is found to be the same using any of our two SST datasets: Hadley or NOAA.



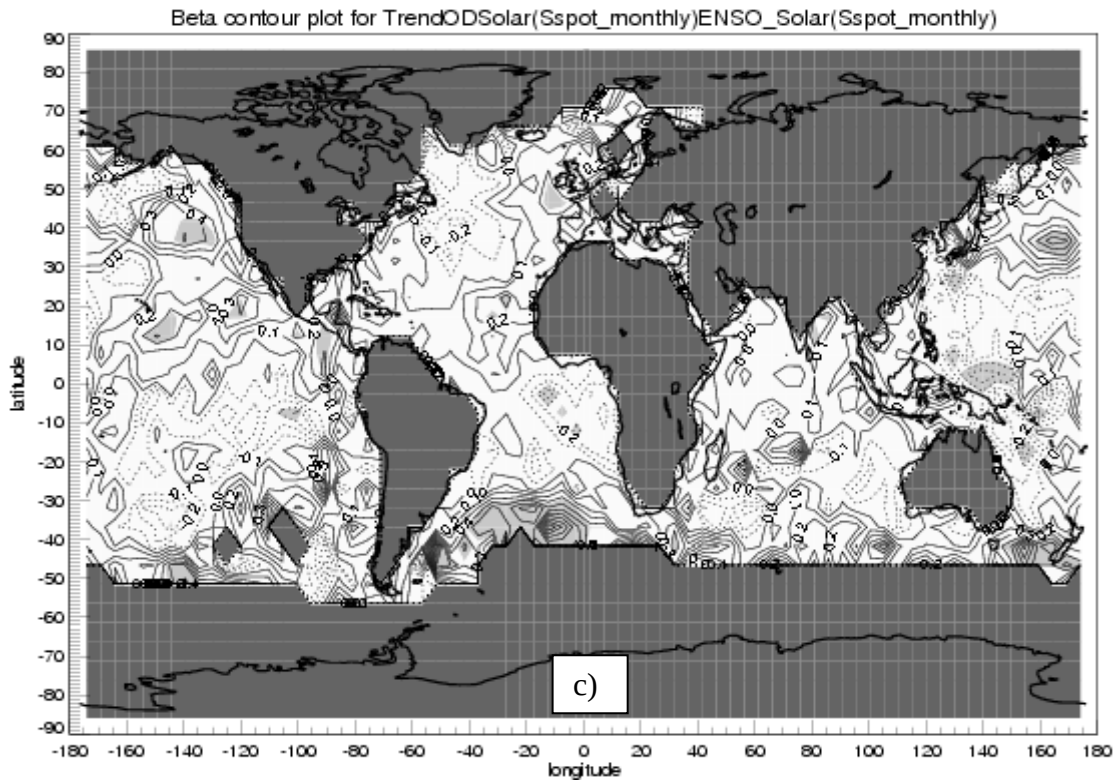


Fig. 3.3: Amplitudes of the components of variability of SST (in °K) due to solar (using monthly sunspot number) with other independent parameters trend, OD and ENSO: a) during DJF (using dataset from NOAA); b) using all months of year, represented by monthly average values, with annual cycle removed (dataset from NOAA); c) same as b) but using dataset from Hadley centre.

Performing a multivariate analysis on near surface air temperature from 1889 to 2006, Lean and Rind (2008) also found that the response of solar forcing is strong and positive at mid-latitudes (near 40°) in both the hemisphere, in the vicinity of Ferrel cells- the interface between the Hadley and Polar cells. The detectable amplitude of surface temperatures around mid-latitude to decadal solar forcing in that study is consistent with prior analyses (Haigh, 2003; van Loon et al., 2004; Crooks and Gray, 2005), which suggests that the response is linked with the large-scale dynamical circulation of the atmosphere. This is also consistent with Haigh et al. (2005) who detect mid-latitude warming during solar max years, even using two different sets of data (SSU/MSU satellite data and ERA reanalysis data) for the period 1979-2001.

3.4. Discussion

To explain the apparent discrepancy between results for the compositing and regression, we investigate the different methodologies employed. vL07 deduced the solar signal by taking a composite of the data corresponding to years identified with the peak of 14 solar activity cycles within the data period, and then associating the anomaly relative to the climatology with effects of the Sun. The pattern was robust to the removal of data from either 1989 or 1905, the only years identified as having a strong ENSO influence.

To explore further any potential link between the apparent solar signal and ENSO, we present a scatter diagram of DJF mean ENSO index against annual average SSN (Fig. 3.4a). Such scatter plot clearly suggests that no obvious relationship exists. A separate multiple regression analysis of the ENSO data (not shown) confirm this to be the case. This would suggest that a signal identified with solar variability in tropical SSTs would be unlikely to express a particular phase of ENSO. van Loon and Meehl (2008), in a subsequent analysis also showed that solar responses are different to that of the ENSO. However, in the same diagram, if we mark solar max years with a red square symbol, see Fig 3.4 b, surprisingly, almost all the solar maximum years coincide with the cold event side of ENSO during DJF. Nine of the fourteen have a value of ENSO index lower than the average, and four of these years (1893, 1917, 1989, 2000) are associated with particularly strong cold events. Only one solar maximum year is associated with a significant positive ENSO signal and this, 1905, is a weak solar cycle. As it is only the solar maximum years that are used by vL07 to characterize the solar signal, it is clear that their result will resemble a cold event (La Niña) pattern. It then remains to be determined to what extent the derived signal (by vL07) can be assigned as due to the Sun rather than mainly due to natural ENSO variability. Any strong signal in SSTs would likely also be seen in near surface air temperature but two independent studies (Stott and Jones (2009), using a optimal detection technique, and Lean and Rind (2008), using multiple regression analysis) both show very little solar signal in the tropics on centennial timescales.

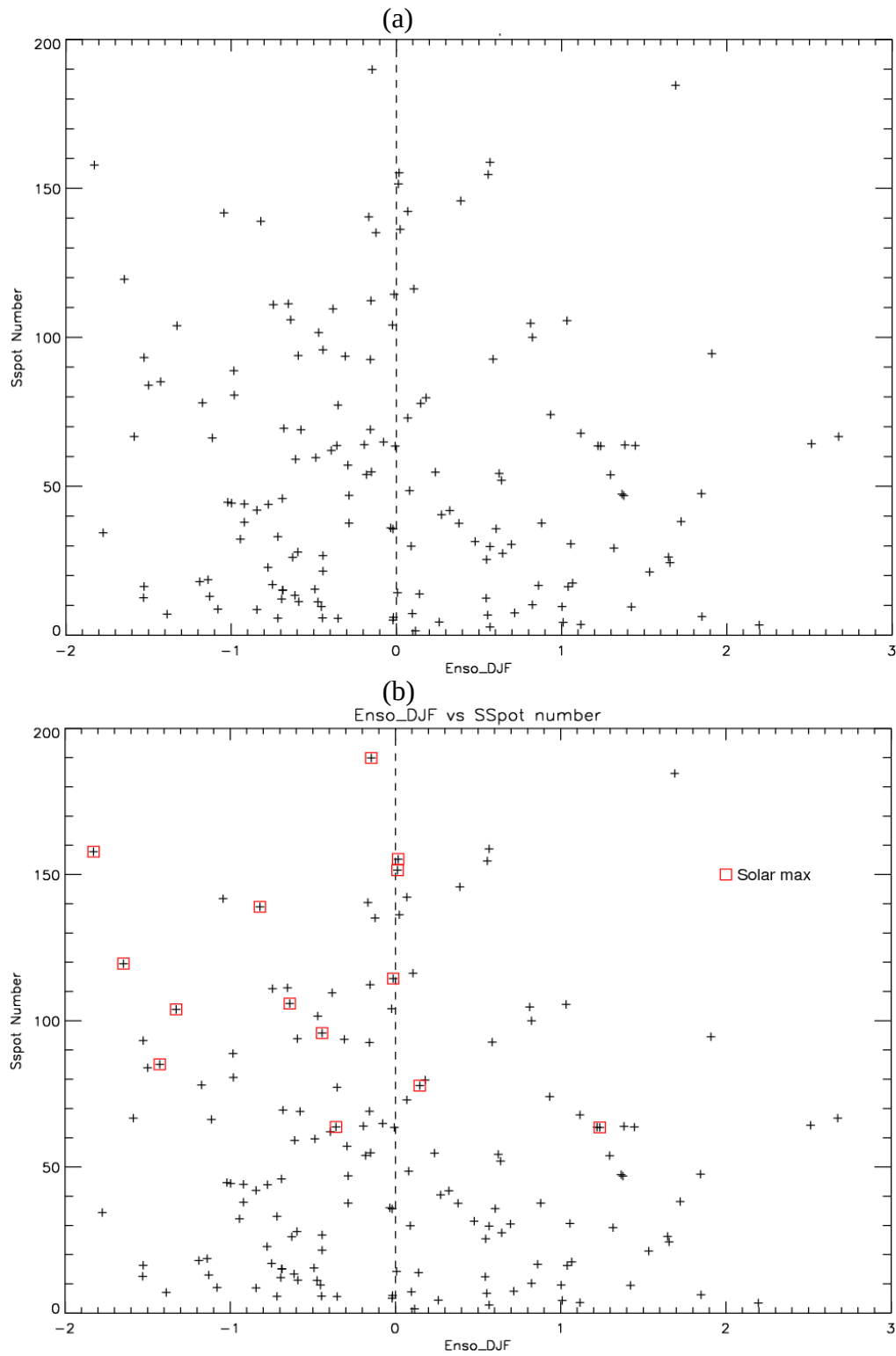


Fig.3.4: (a) Scatter diagram of DJF mean ENSO index vs. annual mean sunspot number; (b) as (a) but with solar maximum years identified.

Thus, our study suggests that using the method of solar max compositing (as used by vL07), the ENSO signal in the tropics may be misinterpreted as a solar signal; nevertheless, as described above, using multiple regression, we were able to detect independent signals of the Sun and ENSO. According to vL07, solar forcing during peak years of the boreal winter is causing a strong negative temperature anomaly around equator, which subsequently causes other related feedbacks including the SLP anomaly around the AL. As we are unable to detect a negative solar signal for SST in the tropics, our results do not support the proposed mechanism of vL07.

Nevertheless, it is not immediately obvious why Figure 3.3 should show a weak warm event (WE)-like response associated with higher SSNs rather than the cold event (CE) pattern of vL07, and as suggested by Figure 3.4b. One possibility is that the multiple regression technique treats the solar and ENSO signals as linearly independent whereas, as discussed by White and Liu (2008) there is non-linear coupling between the two influences and they cannot perhaps be cleanly separated. However, this cannot be the whole story as the solar signal produced by the regression is essentially unaffected by the inclusion of an independent ENSO index and, furthermore, White and Liu (2008) show a WE pattern at the peak of DSO with or without ENSO coupling. Rather, the answer appears to be related to the definition of peak solar activity.

Figure 3.5 shows the time series of monthly SSNs and of their annual means, it also identifies the years of highest annual SSN for each solar cycle, as used by vL07 to label solar peak years (shown with vertical lines). It is apparent that peak years tend to occur very soon after the solar cycle becomes more active - this is true of all the stronger cycles - and at least 1 year before a date that would represent the peak of a broader decadal variation. Thus an analysis based on peak SSN years, such as the vL07 result, represents the solar signal at a particular (rising) phase of the solar cycle, while Figure 3.3 represent more broadly the difference between periods of higher and lower SSN.

Such difference in timing might also provide an explanation for other apparent discrepancies in solar signals. For example, in observational data van Loon et al. (2004)

find a strengthening of the tropical Hadley cell when the Sun is more active while Kodera and Shibata (2006) find it weakened. Peak SSN composites are used in the former paper while correlations between meteorological data and the 10.7cm solar activity index are used in the latter and thus, from the arguments presented above, they represent different aspects of solar cycle variability. Similarly in modelling studies Meehl et al. (2008) with

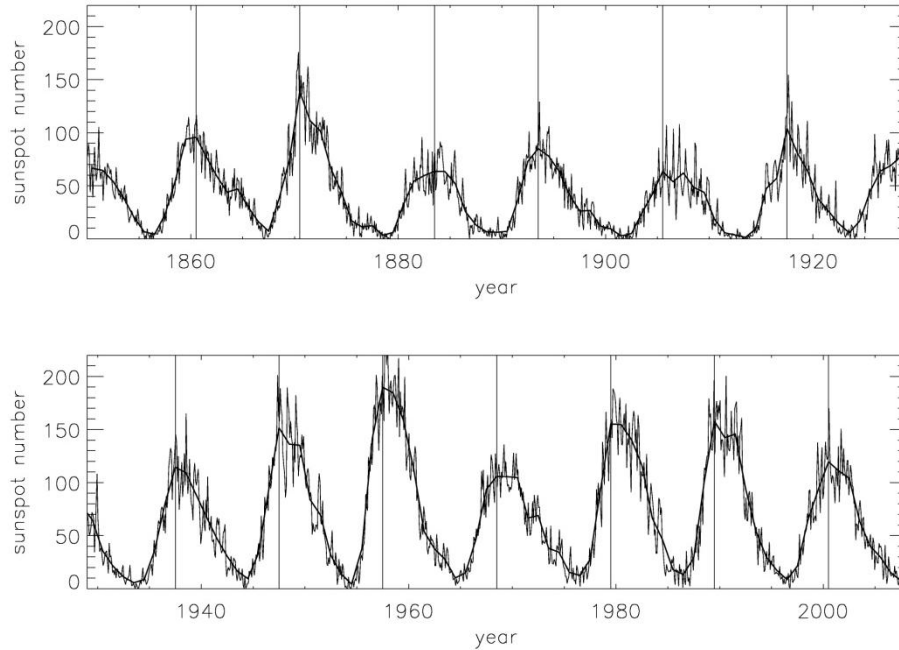


Figure 3.5: Monthly mean sunspot number (thin curve) and the annual mean (thick curve). Vertical lines indicate the year of peak annual sunspot number for each solar cycle.

a transient model run analysed by peak SSN year, find a stronger Hadley circulation; while Haigh (1996;1999), using equilibrated solar maximum and minimum experiments, finds a weaker Hadley cell.

Meehl et al. (2008, 2009) discuss a peak in irradiance at the peak of the decadal solar oscillation producing the La-Niña-like response. They show this is lagged after a year or two by an El-Niño-like event (although van Loon and Meehl (2008) show a weaker cold event at one year lag) and they suggest mechanisms. We also find that there is often a

cold event at peak SSN but this occurs a year or so in advance of the peak in irradiance so that the subsequent warm event coincides with the solar maximum in irradiance, as shown by White (2007) and White and Liu (2008). This difference in timing is crucial in considering mechanisms which involve forcing by solar irradiance. The SLP signal in mid-latitudes varies in phase with solar activity, and does not show the same modulation by ENSO phase as the tropical SST, suggesting that the solar influence here is not driven by coupled-atmosphere-ocean effects but possibly by the impact of changes in the stratosphere resulting in expansion of the Hadley cell in solar max years and poleward shift of the sub-tropical jets (Haigh et al., 2005). It raises another issue with regard to their proposed ‘bottom-up’ mechanisms (also discussed in section 1.1. with Fig. 1.2.6). Associated with the (putative solar-induced) cold event in tropical SSTs, as is standard with an ENSO negative phase, is a positive anomaly in North Pacific SLP in winter and Meehl et al. (2008) discuss how the two are physically related. In the observations, however, this positive anomaly in mid-latitudes is maintained for several years, and continues through the following “warm event” which would normally be accompanied by a negative SLP anomaly. This poses problems for any mechanism proposed for a solar influence on climate which involves an ENSO-type coupled ocean-atmosphere processes.

Results from two different global coupled models (PCM and CCSM3) shown by Meehl et al. (2008), using multiple ensemble members from 20th century simulations and thus on the order of 40 to 50 realizations of peak solar forcing for each model, show a similar pattern to the shorter observational record. Regarding the timing of forcing and responses, however, Meehl and Arblaster (2009) (MA09) showed that the result of these two models during DJF differ: PCM showed cooling that starts one year before and reaches a maximum in the same year with warming around the following year, whereas CCSM3 indicated warming one year before, cooling in the same year and that cooling persists the following year (see Fig. 1.2.7). Such results are not consistent with the observational results of van Loon et al (2008) who showed maximum cooling in peak SSN years, preceded and followed by lesser cooling (and no warming). The discrepancies between the two model results in terms of tropical Pacific SST mainly arise from the different ENSO variability within the models (MA09). Rind et al. (2008), using a model

with various representations of stratospheric ozone, with and without a coupled slab ocean, found that a sea surface temperature response that produces the troposphere warming to be similar to that described by White et al. (1997) and very different from the La Niña response described by van Loon et al. (2007). In determining this SST signal Rind et al. (2008) compared years of higher solar irradiance with those of lower, using a threshold in UV irradiance, and would thus be more likely to find the broad DSO signal than the peak SSN signal, as discussed above. Our analysis thus suggests that the response of SST to variations in solar irradiance and the mechanisms of this response remain to be established in the context of ENSO variability and the uncertainties in timings discussed above.

3.5. Regression with different solar indices

To investigate the robustness of the identified signals on SLP, we used three different reconstructions of TSI (viz., Krivova & Solanki, Foster and Hoyt & Schatten) in our regression analysis. Here we discuss our observations during two different seasons, the northern winter (DJF) and the northern summer (JJA), where each year corresponds to average value during DJF and JJA, respectively.

SLP with different solar indices for DJF:

Fig. 3.5.1 shows, amplitudes of the solar component of variability of SLP during DJF, using different solar indices; the other independent parameters used are trend and OD. In terms of solar indices, apart from monthly SSNs, we used reconstructions from Krivova & Solanki, Foster and Hoyt & Schatten as shown in Fig. 2.6. Using the Krivova & Solanki and Foster reconstructions during DJF, we are able to locate a strong positive signature around the AL, with a secondary negative footprint around the eastern part of the equatorial Pacific (similar to using SSN). Positive SAM features are also recognizable from that figure (Fig. 3.5.1), using those three solar indices. Thus these different measures of solar activity indicate similar results for solar cycle variability. How

these TSI series differ in terms of their reconstruction was discussed in Chapter 2. In that chapter, we also elaborated the fact that the Hoyt & Schatten reconstruction mainly captures long term solar variations rather than the 11 year solar cycle variability. From Fig. 3.5.1, it is clear that the results using the Hoyt & Schatten reconstruction are not in agreement with the rest. We now analyse the β component of the various regression coefficients, to emphasise, how the Hoyt & Schatten TSI can be mixed up with the long term trend to contaminate the results of the regression.

β Component Analysis for SLP during DJF: a) at 140°W, 50°N and b) at 120°W, 60°S

To clarify the disparity between results using the Hoyt & Schatten TSI to that from the rest (in Fig. 3.5.1), we have carried out a β Component Analysis for SLP during DJF, choosing two different points from the globe. One is in the southern Pacific (at 120°W, 60°S and marked with 'X' in Hoyt & Schatten plot in Fig. 3.5.1); while the other is in the northern Pacific (at 140° W, 50° N and marked with 'Y' in sunspot number plot in Fig. 3.5.1). We selected these points based on certain criteria: in the southern Pacific, we considered a place that indicates a strong significant positive signal using the Hoyt & Schatten reconstruction, which is absent/ different using the rest solar indices; whereas, in the northern Pacific, we selected one location, which showed a highly significant response in several indices, but different and insignificant using the Hoyt & Schatten reconstruction.

Here amplitudes of the component of variability (β value) of SLP during DJF are presented for three independent regression indices viz. trend, OD and solar indices. Figure 3.5.2 illustrates their relevant behaviour around 120°W, 60°S; whereas, Fig. 3.5.3 shows the same around 140° W, 50° N. In those figures, results using different solar indices are presented in separate panels.

In Fig. 3.5.2, it is clear that, apart from the analysis using the Hoyt & Schatten reconstruction, the linear trend component is small, while the OD component is identifiably negative. But using the Hoyt & Schatten reconstruction, the trend emerges as

large and negative alongside, practically no influence for OD. The solar signal from the Hoyt & Schatten experiment is large and positive, but its upward trend is clearly compensating for the large negative climate trend signal.

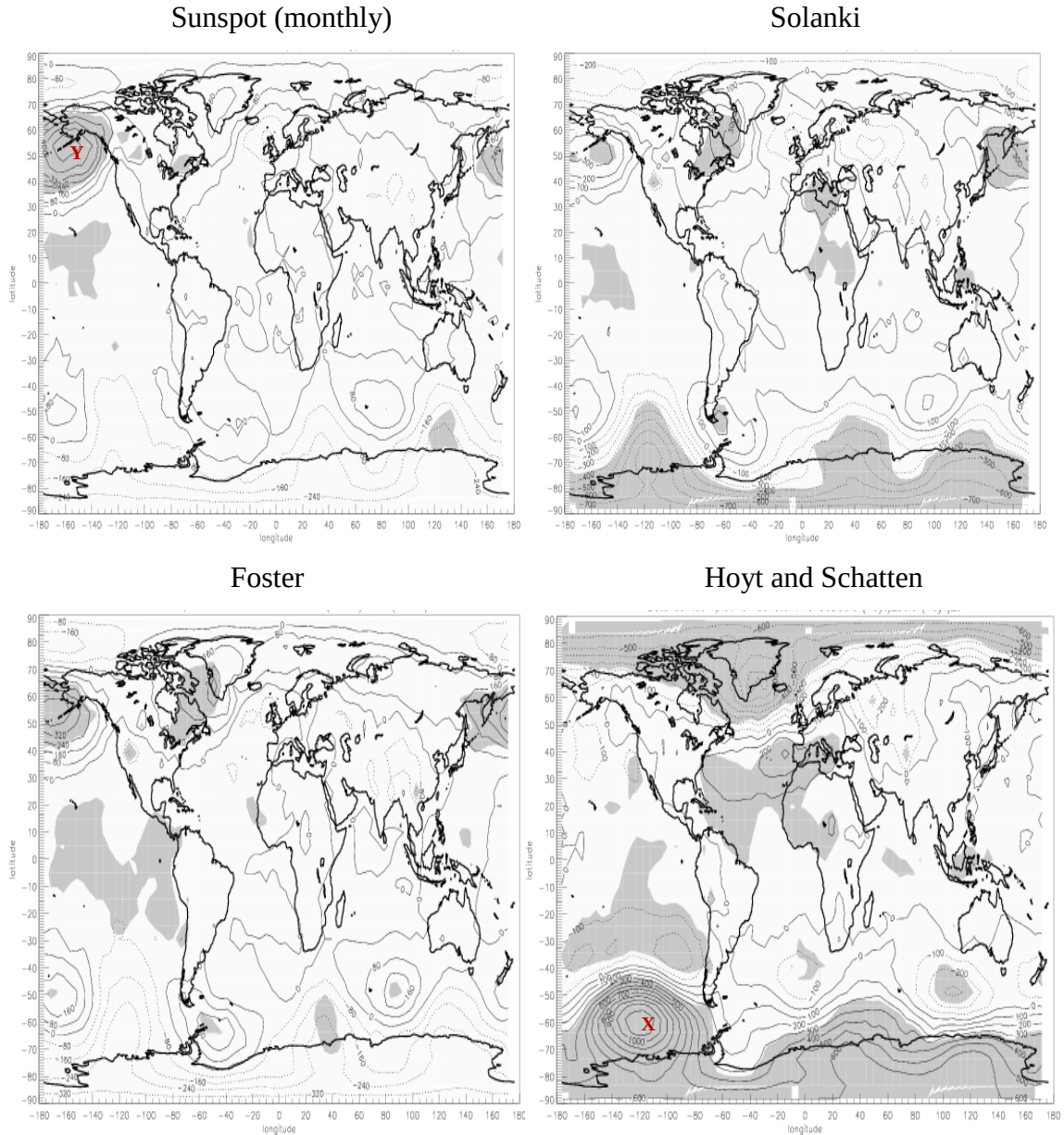


Fig.3.5.1. Amplitudes of the component of variability of SLP during DJF, due to solar (using different solar indices). Other independent parameters used are trend and OD.

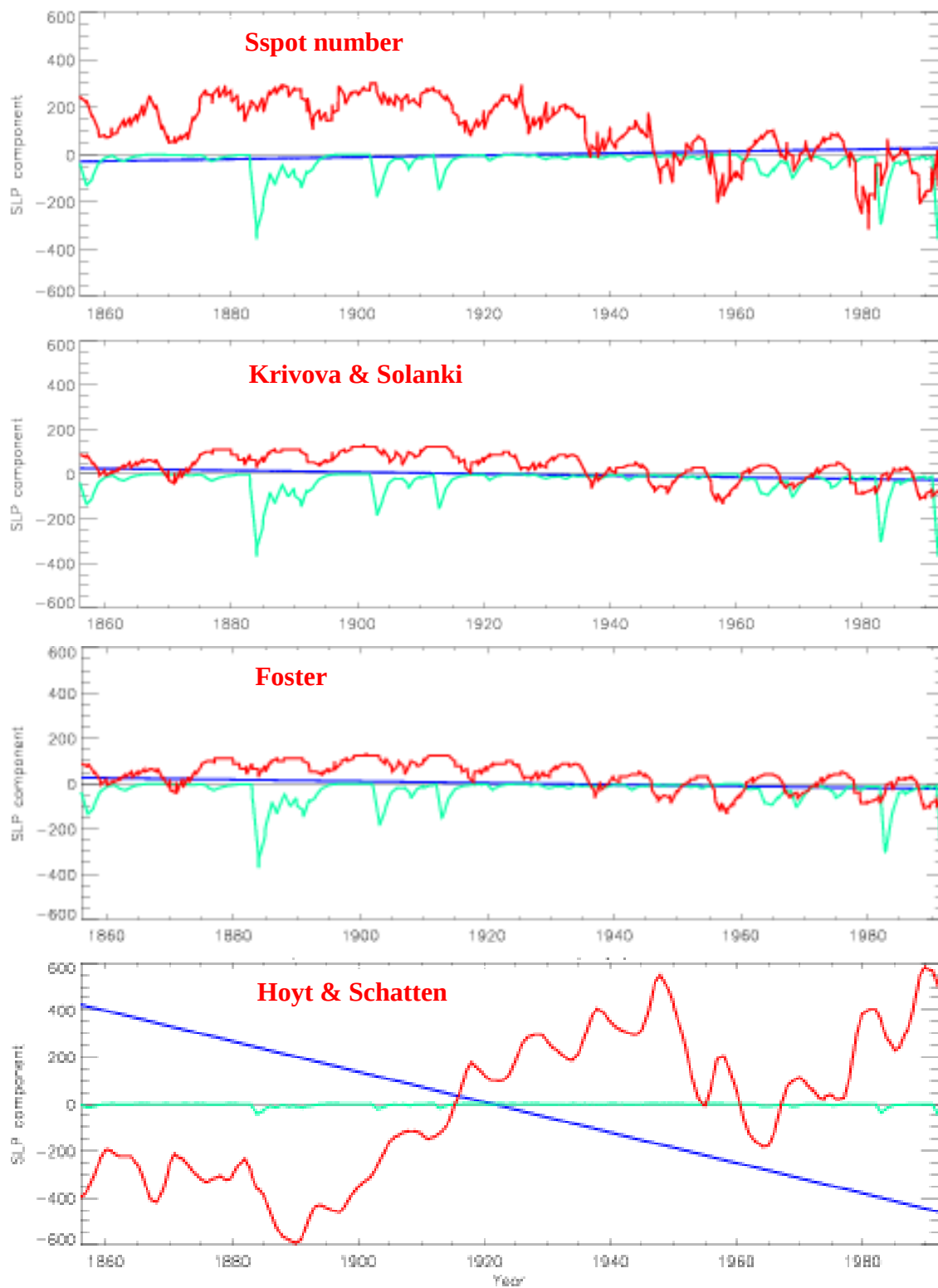


Fig. 3.5.2: Amplitudes of the components of variability of SLP during DJF (*at 120°W, 60°S*, 'X' in Fig. 3.5.1) due to trend (blue), OD (green) and different solar indices (red).

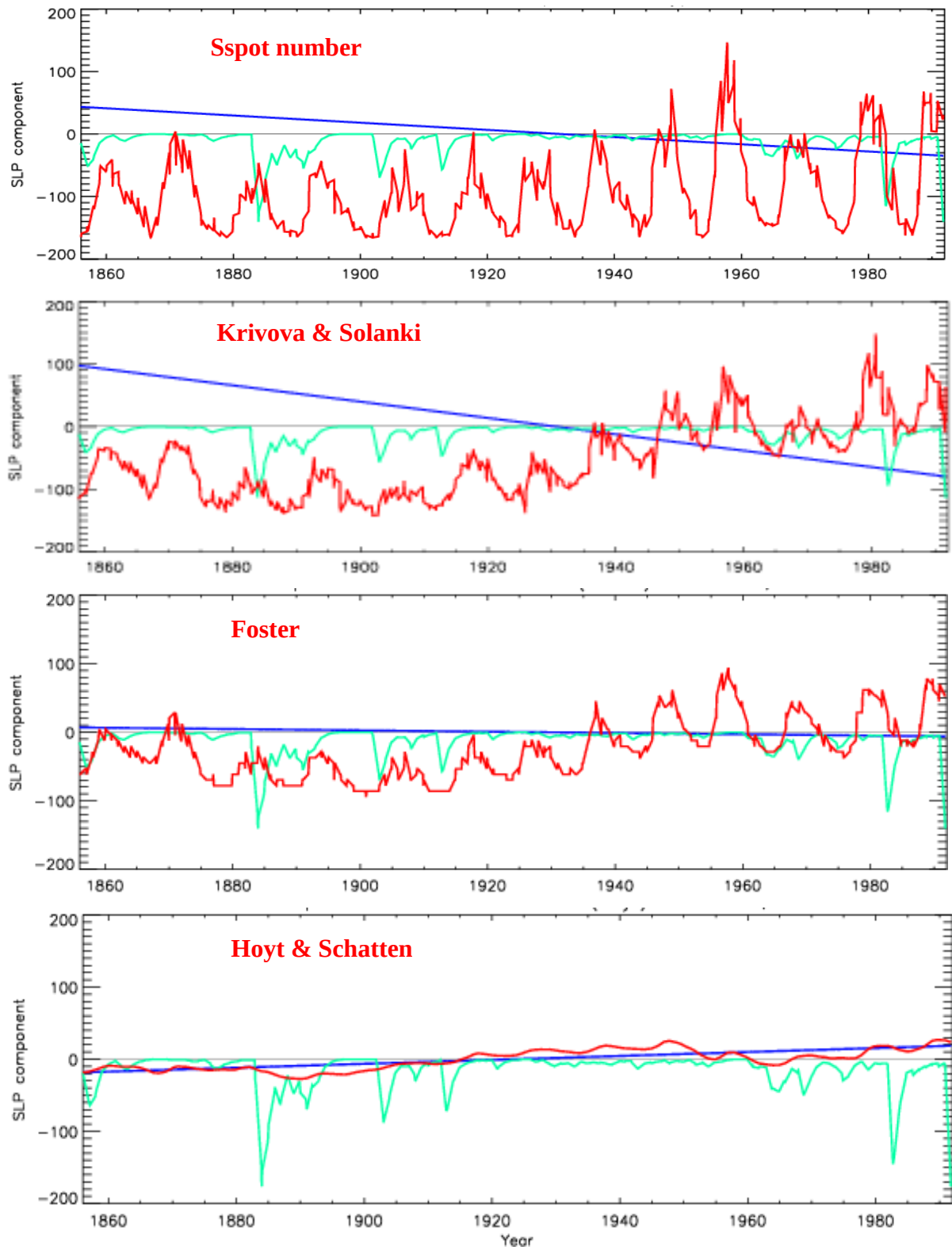


Fig.3.5.3: Amplitudes of the components of variability of SLP during DJF (at 140°W , 50°N , 'Y' in Fig. 3.5.1) due to trend (blue), OD (green) and different solar indices (red).

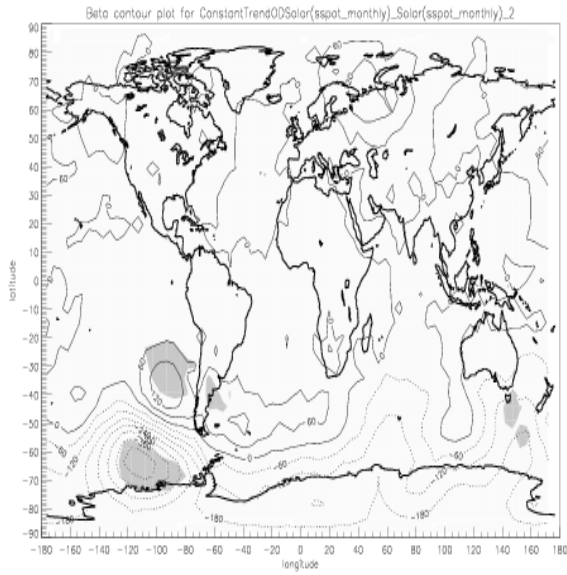
Now we focus on the result of the NH, shown in Fig. 3.5.3. These suggest that using the reconstruction from Foster, there is not much influence for the linear trend (with nominal –ve β value); whereas, solar indices using SSN and TSI from Krivova & Solanki, both reveal a negative β value for the trend. Moreover, all these three indices indicate a negative β value for the OD, with large and positive β value for solar. On the other hand, using the TSI from Hoyt & Schatten, the effect of the actual trend is seen to be the opposite than using the other three indices and the OD though still negative suggests max β value among all four indices. The β value for solar is here found to be a positive but very small suggesting that the inherent trend line in this TSI is influencing both the climate trend signal and OD, indicating different result for the solar signal.

Hence our analysis suggests that the Hoyt & Schatten solar index sometimes produces different results, as the solar signal is mixed with the long-term trend.

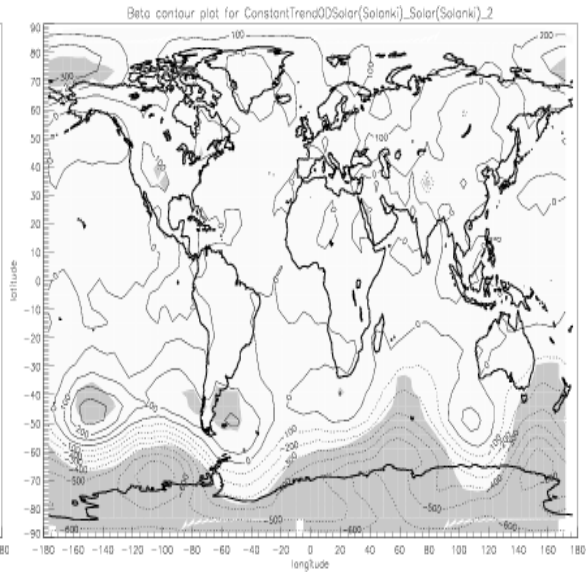
SLP with different solar indices for JJA:

In Fig. 3.5.4, we show amplitudes of the component of solar variability of SLP during JJA, using different solar indices. Other independent parameters we used are the trend and OD. In terms of solar indices, apart from SSN, we also used TSI reconstructions from Krivova & Solanki, Foster and Hoyt & Schatten, (shown with appropriate headings), as were also used during DJF and shown earlier in Fig. 3.5.1. From Fig. 3.5.4, it is clear that the solar signal in SLP during JJA is mainly identified around mid to high latitudes in the SH, resembling that of the SAM. Apart from the Hoyt & Schatten TSI reconstruction, different measures of the solar activity here also show similar results for the 11 year solar cyclic variability. As the Hoyt & Schatten reconstruction is mixed with the long-term trend (as was shown and discussed before), compensating with the climate trend signal, it indicates, quite contrary to the rest - a negative SAM signal.

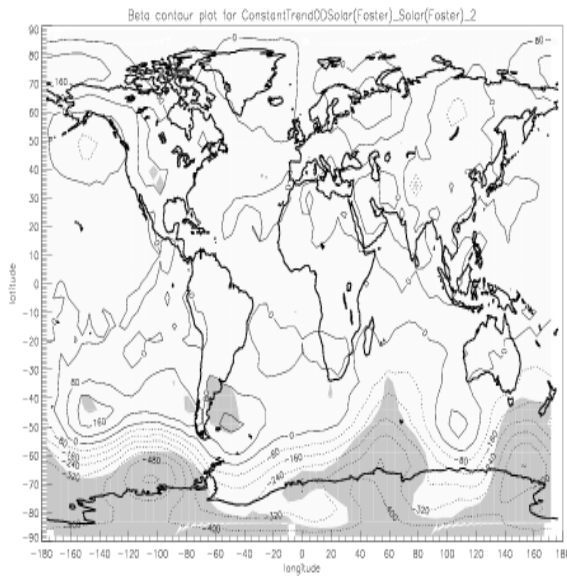
Sunspot (monthly)



Krivova & Solanki



Foster



Hoyt & Schatten

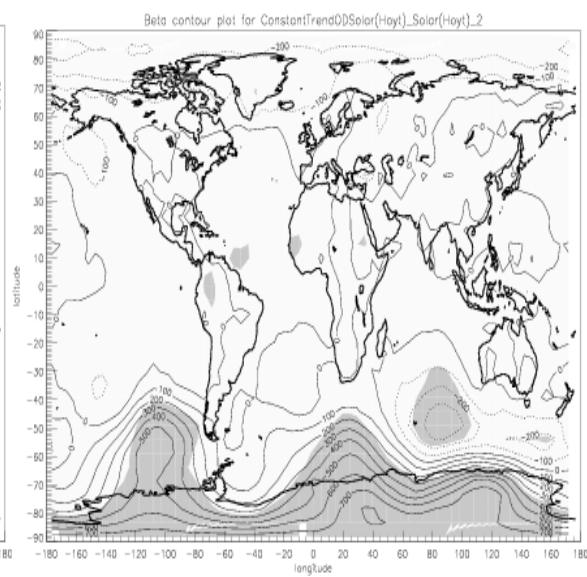


Fig. 3.5.4: Amplitudes of the components of variability of SLP during JJA, due to solar (using different solar indices). Other independent parameters used are trend and OD.

3.6. Summary

Using the method of solar max compositing, we detect a similar solar signal to vL07 and Meehl et al. (2007) in the Pacific region. For SLP, we detected a strong positive signal around the AL, with a negative response (though not significant) north of the equator, covering the international date line. For SST, we also identified a strong La Niña-like pattern in the tropics, with a secondary positive signature around the mid-latitude, though weaker than vL07. We extended both the studies to the whole of globe and compared their results. Our analysis indicated that the results of using this method may not only be biased due to mixing up the signal with other strong variability, like ENSO, but also be sensitive to the choice of details of the compositing method. Thus, our study pointed towards the lack of robustness of the signal detected using compositing method. We also observed that signals, using solar max compositing are almost unaffected by the inclusion or removal of a few solar cycles.

Data analysis, using either solar max compositing or regression analysis, suggests that a positive solar signal is present in the region of Northern Pacific for both SLP and SST. Such a result is consistent with White et al. (1997, 1998); Allan (2000); White and Turre (2003), who also observed a quasi-decadal oscillation (QDO) of 9 to 13 years period in global patterns of SST and SLP, during the late nineteenth and twentieth century, that fluctuates in phase with the ~ 11 year period signal in the Sun's total irradiance.

Our results using multiple regression are different in the tropics to those from the compositing analysis for both SLP and SST. For SLP we find a broad reduction in pressure across the equatorial region but not the negative anomaly in the sub-tropics detected by vL07, using compositing method. For SST, using two sets of data (one from NOAA and another from the Hadley Centre), we are unable to detect a –ve solar signal for SST in the tropics, applying multiple regression technique. Thus our results are not in accordance with vL07, in relation to their proposed mechanisms, directly involving the Sun and tropical SSTs. We find that the peak years of solar sunspot cycle are usually associated with the negative phase of ENSO cycles and thus using the method of solar

max compositing (as used by vL07), the ENSO signal in the tropics might be misinterpreted as a solar signal.

Furthermore, we observe that the peak of the annual average SSN (as used by vL07) generally falls a year or more in advance of the broader maximum of the 11-year solar cycle and thus reflects a different (rising) phase of the solar cycle compared with the peak year of smoothed DSO, causing discrepancies between analysis. Analyses which incorporate data from all years, rather than selecting only those of peak SSN, represent more coherently the difference between periods of high and low solar activity on these timescales.

Meehl et al. (2008, 2009) discuss a peak in irradiance at the peak of the decadal solar oscillation producing the La-Niña-like response which is lagged after a year or two by an El-Niño-like event and they suggest a mechanism to explain this. The difference in timing is crucial in considering any proposed mechanisms which involve forcing by solar irradiance. The SLP signal in mid-latitudes varies in phase with solar activity, and does not show the same modulation by ENSO phase as the tropical SST, suggesting that the solar influence here is not driven by coupled-atmosphere-ocean effects but possibly by the impact of changes in the stratosphere resulting in an expansion of the Hadley cell and poleward shift of the sub-tropical jets (Haigh et al., 2005). Given that climate model results in terms of tropical Pacific SST can be dependent on different ENSO variability within the models, our analyses indicate that the robustness of any proposed mechanism of the response to variations in solar irradiance needs to be analyzed in the context of ENSO variability where timing plays a crucial role.

To investigate the robustness of the identified signal, we applied our technique of regression using three different reconstructions of TSI viz., Krivova & Solanki, Foster and Hoyt & Schatten, in place of SSN. Such analysis, using different measures of solar activity, indicates that results are similar for the 11 year solar cyclic variability. Moreover, we showed that the Hoyt & Schatten solar index sometimes produces different results because the solar signal is mixed with the long-term trend.

Chapter 4: QBO effect – over the last 50 years

4. QBO effect - over the last 50 years

Some studies have found that segregating the meteorological data based on the phase of QBO, is necessary to detect a clear signal of the 11-year solar cycle in polar stratosphere. Moreover, Baldwin and Dunkerton (2001) and Thompson and Wallace (2000) showed that perturbations in the polar stratosphere can affect the troposphere for a few months in the form of polar annular modes. Hence, apart from the Sun, we also incorporate QBO in the following chapter.

First we discuss briefly about the mechanisms and coupling processes in the middle atmosphere. Then there is an attempt to understand the main reason for the apparent inconsistency between two published results relating to the polar temperature, the Sun and the QBO. Finally, using the regression technique, the combined behaviour of the Sun and QBO is captured on SLP.

4.1. Mechanisms and coupling processes in the middle atmosphere

The major climate variability in the middle atmosphere, viz. the QBO and SSW has already been described in Chapter 1; here the related mechanisms are addressed.

4.1.1. *Mechanisms for QBO formation:*

The QBO is mainly governed by equatorially trapped Kelvin and Rossby-gravity waves; the former provide the westerly momentum, the latter the easterly.

In Fig. 4.1.1, the formation of QBO is shown in a series of step (from (a) to (f)) and is described below. Mean flow is shown by black wavy line, with easterly (/westerly) flow shown along the negative (/positive) side of the abscissa. The equatorially trapped easterly wave, propagating upward is marked by blue; whereas, the same for westerly is by red. In Fig (a), both the easterly and the westerly maxima of mean flow are descending, as upward propagating waves deposit momentum just below the maxima.

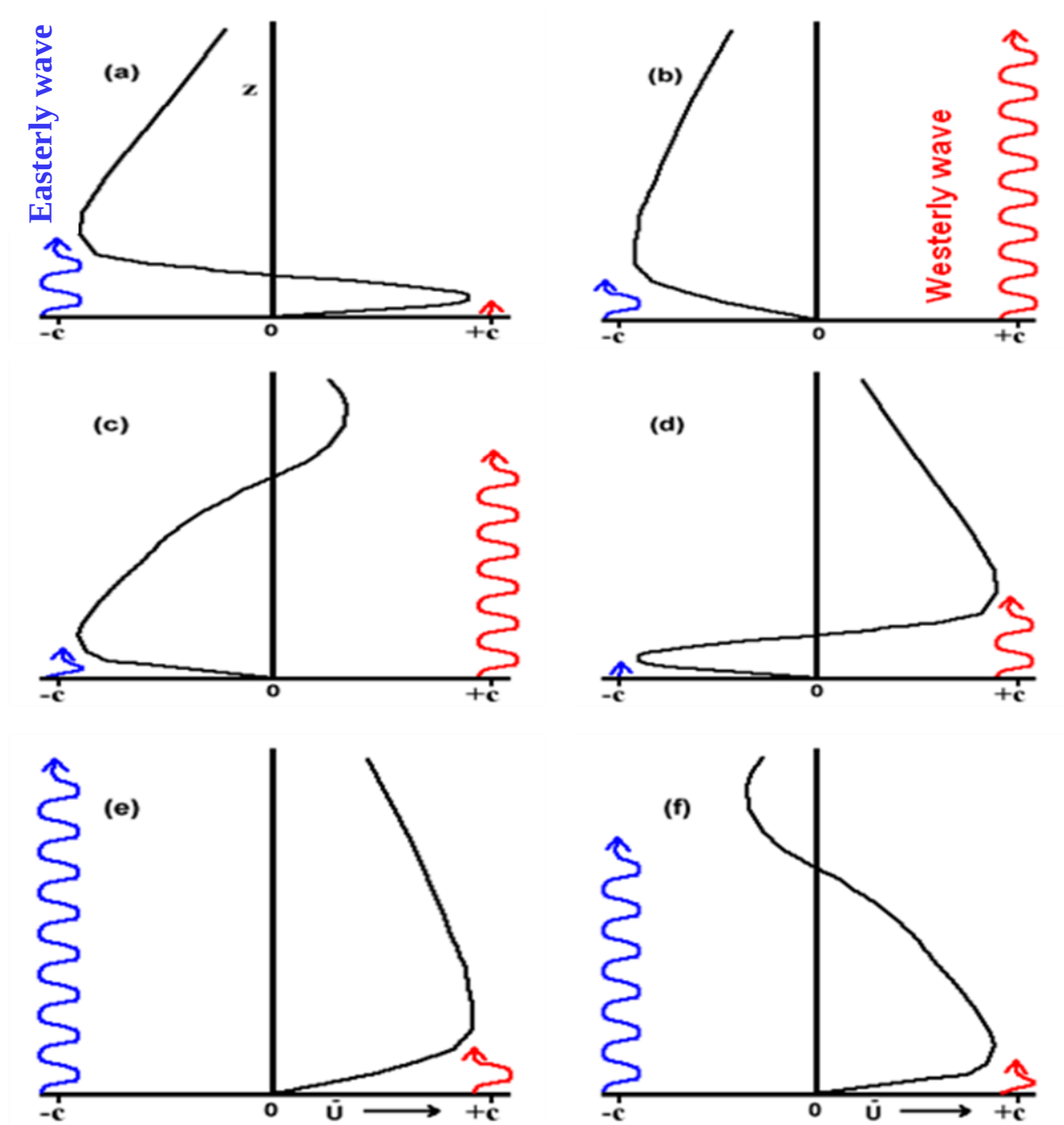


Fig. 4.1.1: Theory of QBO formation

[Source: <http://ugamp.nerc.ac.uk/hot/ajh/qbo.htm>, after Plumb (1984)]

When the westerly shear zone of the mean flow is sufficiently narrow, viscous diffusion destroys the westerlies and upward propagating westerly waves can propagate to high levels through the easterly mean flow, Fig (b). The more freely upward propagating

westerlies are thus dissipated at higher altitudes and produce a westerly acceleration, leading to a new westerly regime of mean flow, Fig (c). Figure (d) shows, both regimes of the mean flow descending downwards until easterly shear zone of the mean flow become sufficient narrow to destroy easterlies and upward propagating easterlies can then propagate to high altitudes through westerly mean flow, Fig (e). Finally, it leads towards the formation of a new easterly regime of mean flow shown in Fig (f); likewise, the same process (as described in Fig. (a) to (f)) continues on.

4.1.2. Mechanism for SSW:

In discussing the mechanism of SSW, we begin with a brief background.

4.1.2.1. Background:

The very basics of the planetary wave (PW) propagation, mainly governed by the Charney Drazin criteria, is shown here by the flow chart of Fig. 4.1.2. It states that the upward propagation of the stationary waves needs to satisfy two main criteria: 1) the background flow must be westerly, though not too strong and 2) only long waves i.e., wave number 1-3 can propagate. Now, winter NH satisfies both the above mentioned criteria: there are westerlies in the high and mid latitude in stratosphere; moreover around the mid-latitude, the orography together with land-sea temperature difference generate long Rossby waves. Thus during the winter NH, there lies the proper breeding ground for the upward propagation of the stationary waves - that allows strong coupling from troposphere to stratosphere be possible.

4.1.2.2. Discussion with Schematic:

Here in schematic 4.1.3, we illustrate the coupling mechanism of middle atmosphere. A picture with latitude vs. altitude (in km) is presented where, different regions, viz. troposphere, stratosphere, mesosphere and thermosphere are marked based on their altitude levels. The zonally averaged (2D) zonal wind (ms^{-1}) for December is shown by solid lines (westerly) or dashed lines (easterly) with respective magnitudes. Zero wind

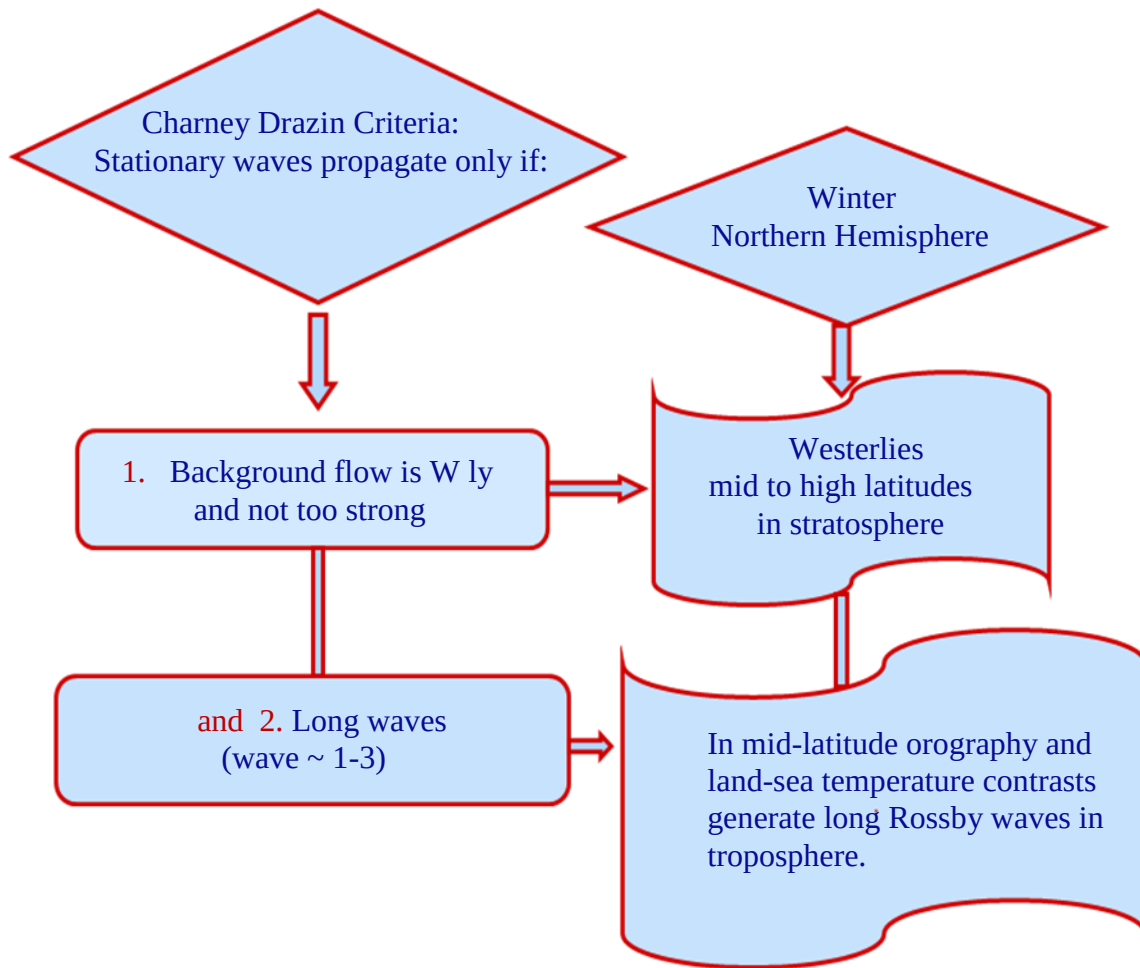
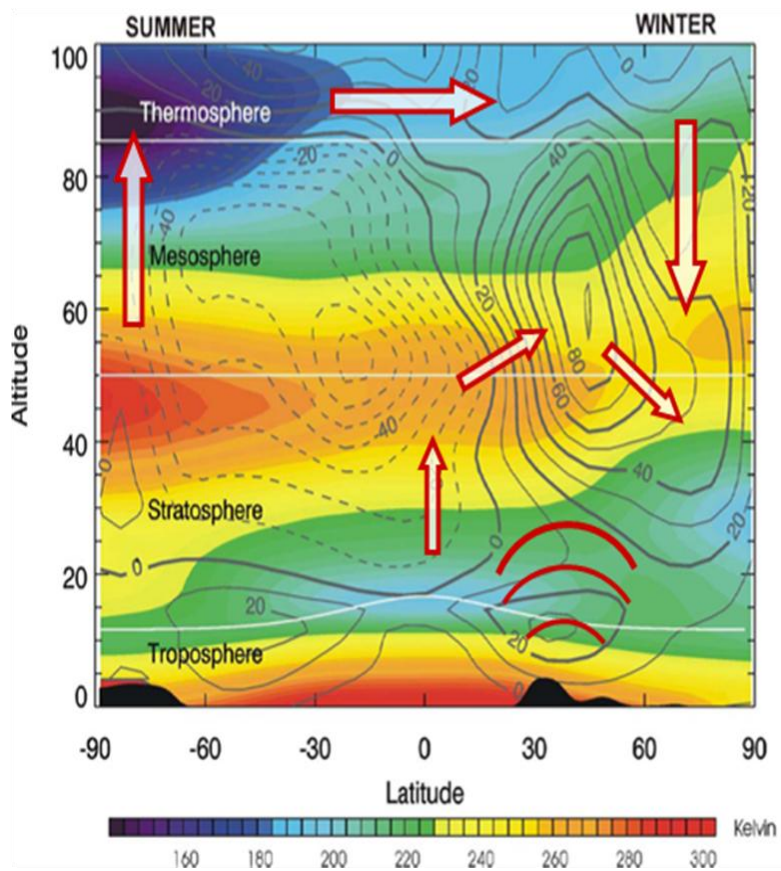


Fig.4.1.2: Flowchart showing the background of planetary wave propagation

line, that separates westerly from the easterly has been marked with label zero. Zonally averaged (2D) zonal temperature (K) is shown with appropriate colours, chosen from the colour bar, placed at the bottom of the picture. From the picture, it is seen that there is an ascending motion in the thermosphere during the summer hemisphere; while descending during the winter hemisphere, shown with thick arrows (however, in our subsequent analysis, we are not attending the thermosphere). But the Brewer-Dobson circulation, (shown with narrower arrows in the schematic) the major circulation in the stratosphere are important in our analysis. As observed in the picture, it is a loop like circulation originating from the equatorial region of the lower stratosphere, travels towards the polar stratosphere (upto stratopause) and then finally comes back in the lower stratosphere.



1. Westerly jet and westerly winds in winter stratosphere due to meridional temperature gradient

2. W ly flow allows planetary-scale waves (PW) to propagate upwards

3. PWs grow and around jet break: decelerating westerlies and warming pole.

[http://lasp.colorado.edu/sorce/news/2004ScienceMeeting/SORCE%20WORKSHOP%202004/SESSION_2/2_4_McCormack.pdf, picture CIRA climatology, Flemming et al., 1990]

PW propagation is sensitive to:

1. Strength of westerlies- **Solar influence**
2. Location of zero wind line- **QBO influence**

Fig.4.1.3: Schematic representation of middle atmosphere coupling

In Fig. 4.1.2, we described how the NH winter can be susceptible to the PW propagation and the upward propagation of these PW's are shown here in Fig. 4.1.3 with red (around

30°N). These PW's move upward and finally break around places of polar stratospheric jet, decelerating westerlies around that region, alongside warming. Now, PW propagation is sensitive to two major influences, one is the solar and the other is the QBO which are discussed below.

*4.1.2.3. Strength of Westerlies: **Solar Influence***

The solar UV at 205 nm increases ~6% from solar min to solar max, which causes more ozone heating in the upper stratosphere (shown with red colour in summer hemisphere around stratopause region in Fig. 4.1.3). That indicates more warming around the equatorial upper stratospheric region during higher solar years compared to the low solar years. Such a strengthening in latitudinal temperature gradient, around the polar stratosphere, is liable to alter the wind structure, strengthening the winter stratospheric polar jet. Such intensified polar stratospheric jet in higher solar years can be responsible for less interaction and breaking of PW; subsequently causing less warming around winter upper stratospheric pole. Though the Sun through interaction with PW can play crucial role in regulating polar temperature, but it is not the sole factor responsible for perturbing polar stratosphere. PW propagation is also susceptible to the phase of QBO and the relevant mechanism involved is described below:

*4.1.2.4. Role of zero wind line: **QBO influence***

The QBO, via modulating the zero wind line, has an influence on PW propagation. It is mainly governed by the well known 'Holton-Tan Effect' (Holton and Tan, 1980, 1982) which states - QBO Ely is associated with the warm polar stratosphere. Schematically, it is shown in Fig. 4.1.4, with a mention about the major steps involved. It is also accompanied by a picture from Baldwin et al. (2001), which is a plot of zonal mean wind differences in a latitude vs. Height/ pressure cross-section. Here easterly wind is marked by blue; PW's are shown by red and various tropical waves originated in the equatorial troposphere (not relevant in describing Holton-Tan effect) are shown by yellow.

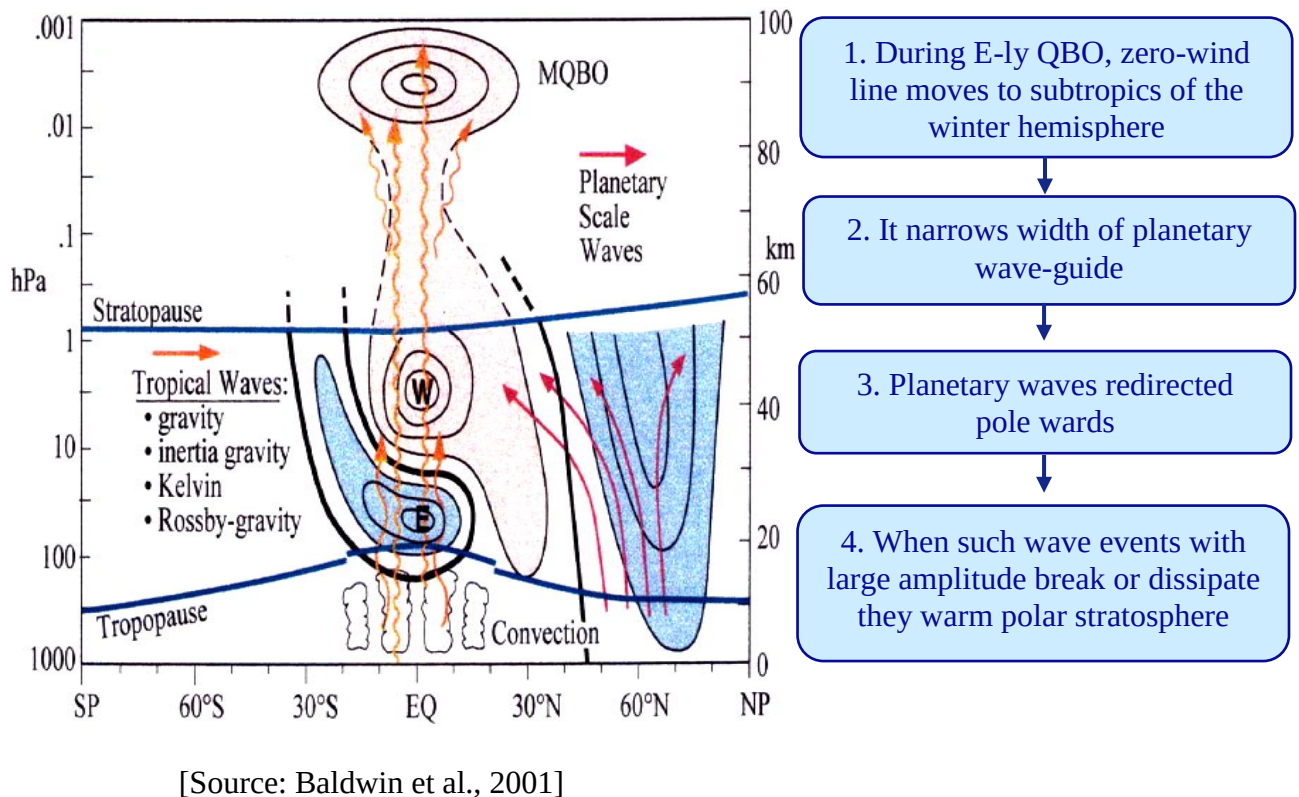


Fig.4.1.4: Zonal mean wind differences and illustration of Holton-Tan Effect

During the easterly QBO, the zero wind line (that separates easterly from the westerly) moves towards subtropics of the winter hemisphere, narrowing the width of PW guide. Thus, PWs are more redirected towards the pole during the E-ly phase of QBO compared to the W-ly phase. When such wave events with large amplitude break or dissipate, they warm the polar stratosphere, depositing their easterly momentum around the polar region (shown with blue colour there). This is the mechanism of the so called Holton-Tan effect – i.e., warm pole during easterly QBO.

We described the main mechanisms relating to the interaction of the QBO and Sun, with upward propagating PW – which was shown to be responsible to modulate the stratospheric polar temperature during the winter hemisphere.

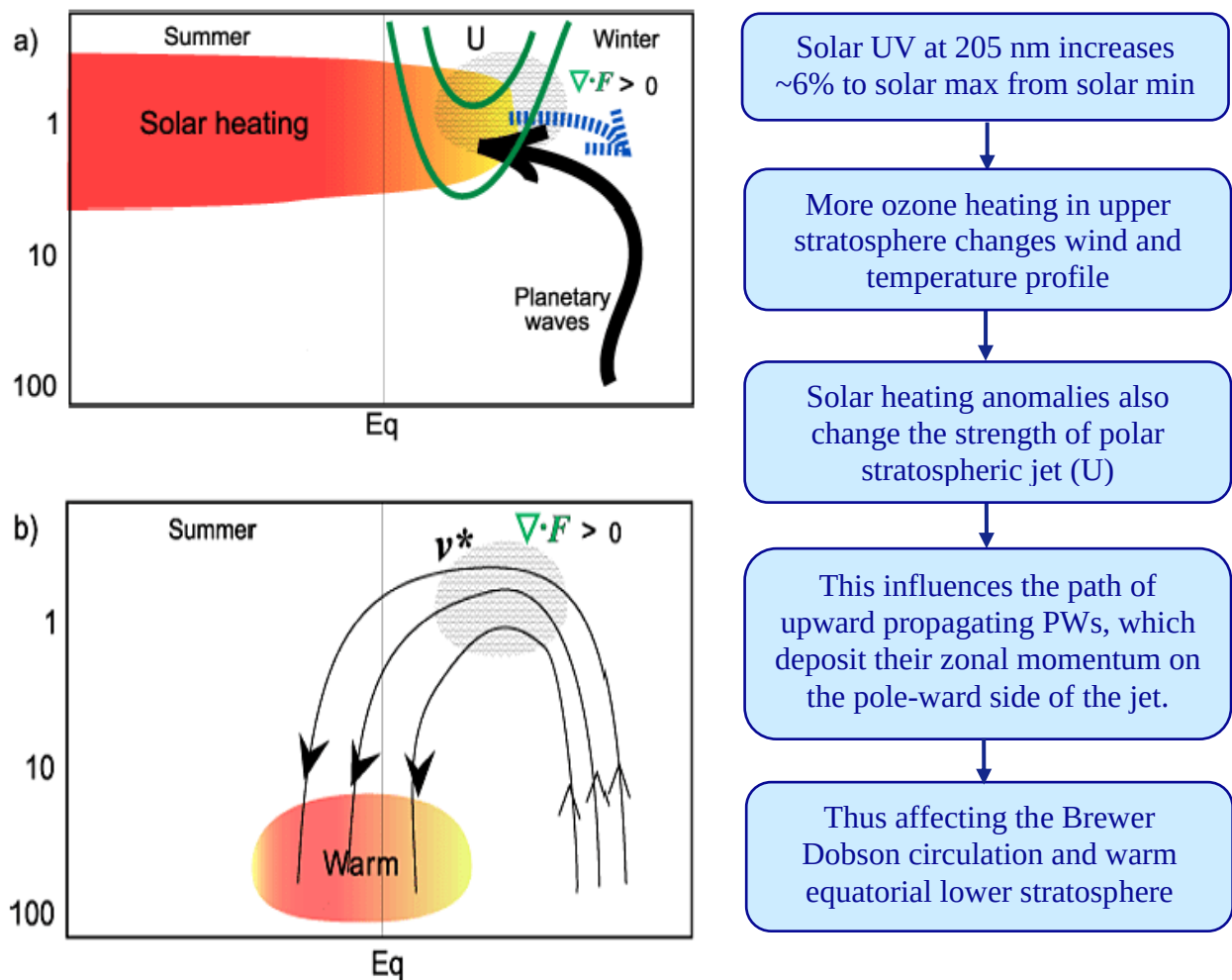
4.2. Polar temperature and importance

Now the study relating to the temperature of the polar stratosphere is very important as it appears to play a crucial role in determining the state of even troposphere. At least, two different mechanisms are proposed in support and mentioned with highlighted box. Here we are going to describe them briefly.

Why is Polar Temperature so important?

- Indicate the strength of the Brewer Dobson circulation with potential influence on equatorial lower stratosphere temperatures: *Kodera and Kuroda (2002)*
 - Potentially affects the lower atmosphere through downwards propagation of the polar annular modes: *Baldwin et al. (2001); Thompson et al. (2000, 2005)*
- *Polar temperature and strength of Brewer Dobson circulation: Kodera and Kuroda (2002)*

The schematic of Fig. 4.2.1 shows a flow chart that depicts a comprehensive overview of proposed mechanism by Kodera and Kuroda (2002) relating to polar stratosphere and finally, warming of equatorial lower stratosphere by solar forcing. The relevant picture from Kodera and Kuroda (2002) is also presented in that schematic. It is a latitude vs. height plot, where the summer hemisphere and the winter hemisphere are also marked. As solar UV increases from solar min to solar max causing more ozone heating in upper stratosphere during higher solar years - it is shown with red colour in the summer hemisphere. Such warming can alter the temperature and wind structure along with changes in the strength of polar stratospheric jet (shown with U). This subsequently, influences the path of the upward propagating planetary waves (shown with thick black lines) which deposit their zonal momentum on the pole-ward side of the jet and thus reducing the strength of the Brewer-Dobson circulation. Such reduction in the strength of the Brewer-Dobson circulation, during HS years, can be responsible for warming of the equatorial region of lower stratosphere. Finally, such warming can be transported to the



[Picture: Koder and Kuroda, 2002]

Fig. 4.2.1: Flow chart showing proposed mechanism by Koder and Kuroda (2002) relating to warming of equatorial lower stratosphere by solar forcing.

troposphere via circulations i.e, the Hadley cell and the Ferrel cell, as elaborately studied by Haigh and co-workers in a succession of papers (1996, 1999, 2005, 2006, 2008).

- *Polar temperature and the polar annular modes : Baldwin and Thompson*

Observations show that changes in the strength of the polar vortex move downward through the stratosphere, and the surface pattern looks like the leading mode of variability, called the AO (Thompson and Wallace, 1998). Fluctuations in the strength of

the stratospheric polar vortices in both hemispheres are observed to couple downward to surface climate (Baldwin and Dunkerton, 1999; Thompson et al., 2005; Baldwin et al., 2005, also shown in Fig 1.1.9).

The analysis of Thompson et al. (2000) based upon monthly mean data from 1958 to 1997 showed that the structure of annular modes in the two hemispheres are remarkably similar, not only in the zonally averaged geopotential height and zonal wind fields, but in the mean meridional circulations as well. Both exist year-round in the troposphere, but they amplify with height upward into the stratosphere during those seasons in which the strength of the zonal flow is conducive to strong planetary wave-mean flow interaction: midwinter in the NH and late spring in the SH. Fig. 4.2.2 shows their results: where zonal mean zonal wind regressed on both the SAM and NAM index are shown on the top panel; whereas, 850 hPa height and 1000 hPa height regressed on the SAM and NAM index respectively are shown in the bottom. The regression was carried out based on all months of the year. Their study depicts that the leading mode of month-to-month variability in the geopotential height field is fundamentally zonally symmetric in both the hemispheres (shown in bottom). Besides, the vertical structures of the zonally symmetric zonal wind anomalies observed in association with the annular modes, shown in the top panels of Fig. 4.2.2 are very similar. In both hemispheres, the annular modes are dominated by meridional dipoles with nodes centred $\sim 45^\circ$. The high-latitude centres of action have maxima centred near 57.5° in the lower troposphere, tilting pole-ward with height to $\sim 65^\circ$ in the upper troposphere-lower stratosphere. The pole-ward wind maximum of the AO amplifies with height upward into the lower stratosphere, whereas the strongest wind anomalies in the SH annular mode are observed in the upper troposphere.

Baldwin and Dunkerton (2001) and Baldwin et al. (2005) suggested that in the NH winter, large-amplitude strengthening and weakening of the stratospheric polar vortex are typically followed by similar signed anomalies in the tropospheric circulation that persists for up to 2 months. Fig 4.2.3 shows results from Baldwin and Dunkerton (2001), where

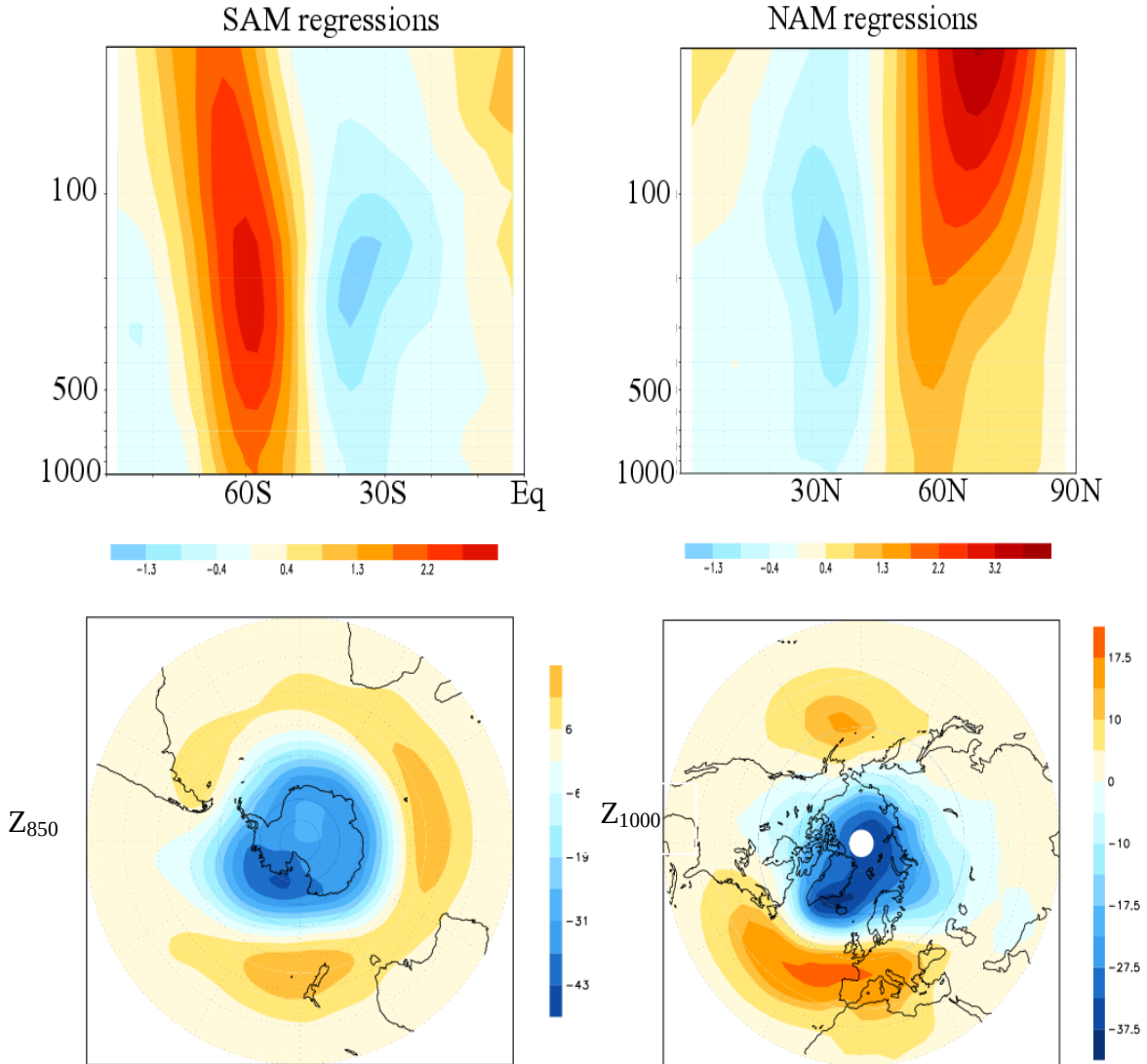


Fig. 4.2.2: Zonal mean zonal wind regressed on the SAM and NAM index (top); 850 hPa height and 1000 hPa height regressed on the SAM and NAM index respectively (bottom). Regression based on all months of the year.

[after, Thompson et al., 2000]

composites of time-height development of the NAM for (A) 18 weak vortex events (corresponding to stratospheric warming) and (B) 30 strong vortex events are presented. Here blue indicates a positive NAM index while red indicates the negative; the thin (also faint) horizontal line above 10 km indicates the approximate tropopause.

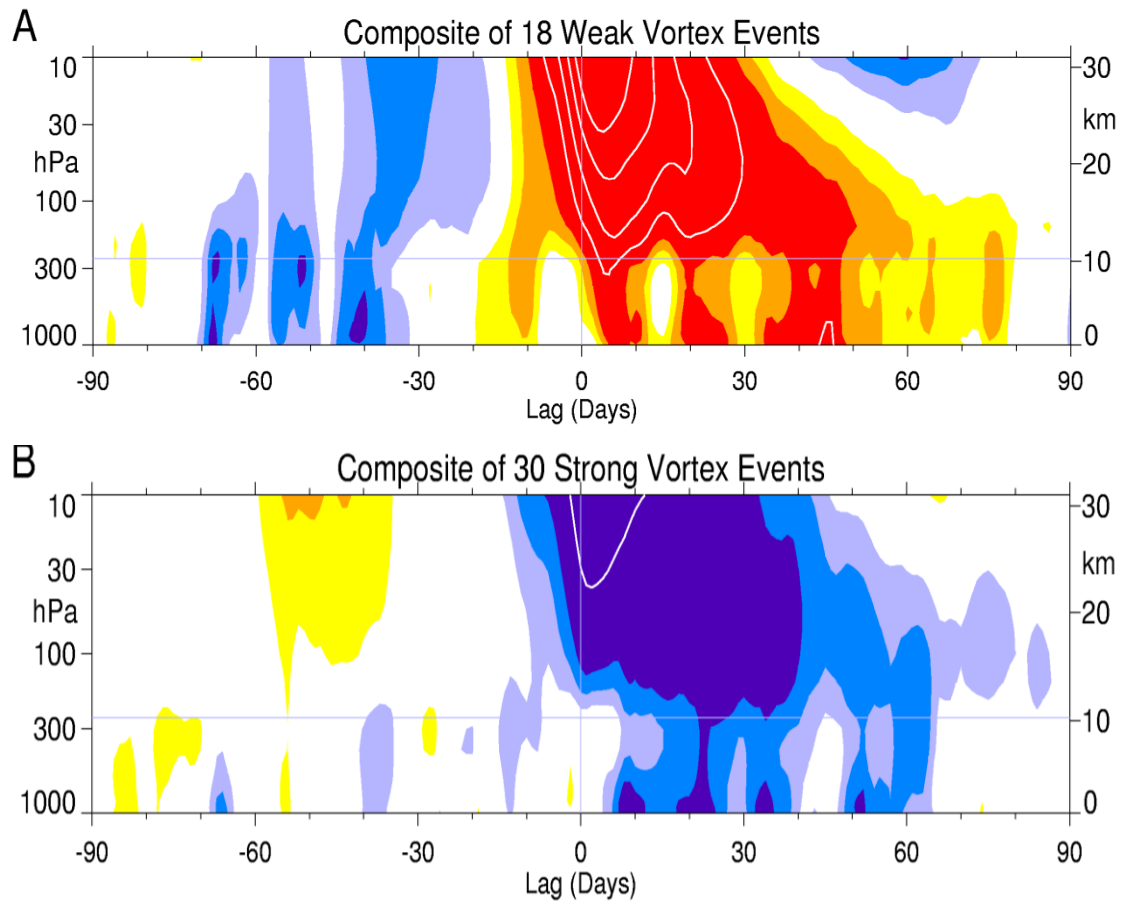


Fig. 4.2.3: Composites of time height development of NAM for A) 18 weak vortex events and B) for 30 strong vortex events.

[Source: Baldwin and Dunkerton, 2001]

Thompson and Wallace (2000) showed that throughout the life cycle of coupled vortex events, the anomalies in the stratospheric and tropospheric circulations in the NH strongly resemble the NAM. In a subsequent analysis, Thompson et al. (2005) noticed that similar situation prevails in the SH as well and anomalies in the strength of the SH stratospheric polar vortex also precede similarly signed anomalies in the tropospheric circulation that persists for more than 2 months. Such observations, also shown in Fig. 4.2.4, reveal that large-amplitude variations in the SH stratospheric polar vortex are followed by persistent anomalies not only in the stratospheric circulation, but in the tropospheric circulation as well. In Fig. 4.2.4, day 0 corresponds to the onset of stratospheric event at 10 hPa and it

is seen that on an average, the ~ 3 month period following the onset of composite stratospheric event is associated with tropospheric anomalies that have the same sign as the overlying stratospheric anomalies. Here red shading denotes positive values in the SAM index. Moreover, Thompson et al. (2000) proposed that regarding development, NAM and SAM follow similar mechanism. All such studies are indicative of the fact that stratospheric variability plays a vital role in driving climate variability at earth's surface on a range of time scales.

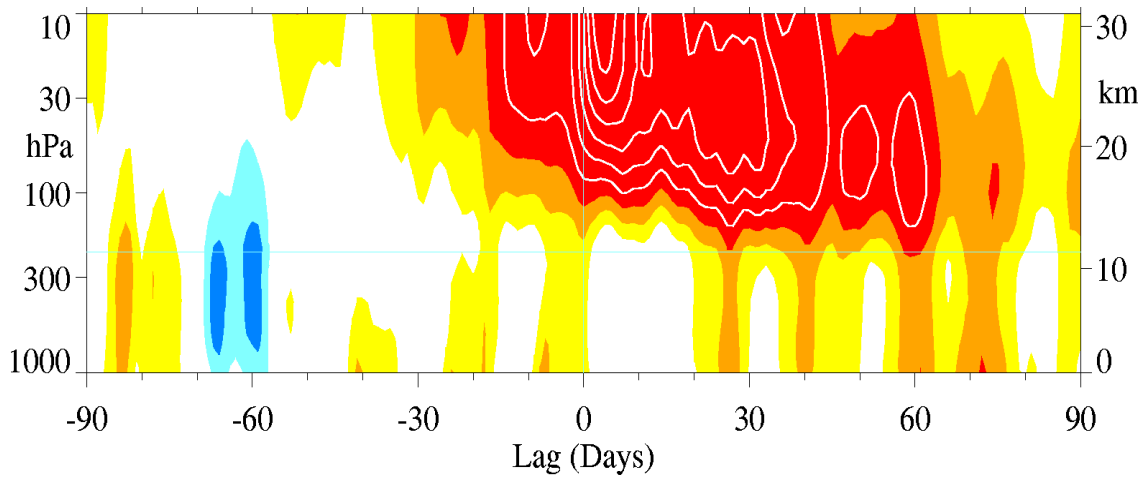


Fig.4.2.4: Composite difference of the SAM index between the weak and strong stratospheric events [Thompson et al. 2005]

The long timescale of stratospheric effects has practical implications for extended-range weather prediction (Baldwin et. al., 2003 a, b). Baldwin and Dunkerton (2005) however, mentioned that the pathway for solar influence involving polar modes of variability, appears to involve interactions with the QBO, but the details are not yet understood. Hence in the current analysis some lights are shed on this issue.

4.3. Data analysis - Solar and QBO

Haigh (2003) using the multiple regression analysis detected independent signals of the Sun and QBO in the zonal mean temperature. Their result detecting amplitudes of the

component of variability in zonal mean temperature due to QBO (at the top) and solar (at the bottom) is shown in Fig. 4.3. Here significant region up to 95% are marked by white. From Fig 4.3, it is evident that no significant signal for the solar or QBO is seen in the north polar temperature around 30 hPa level (~30 km), if considered separately. However, data sorting of the solar and QBO indicates differently and suggests that it is necessary to segregate the meteorological data based on the phase of QBO, to detect clear signal of the 11-year solar cycle in polar stratosphere and discussed below.

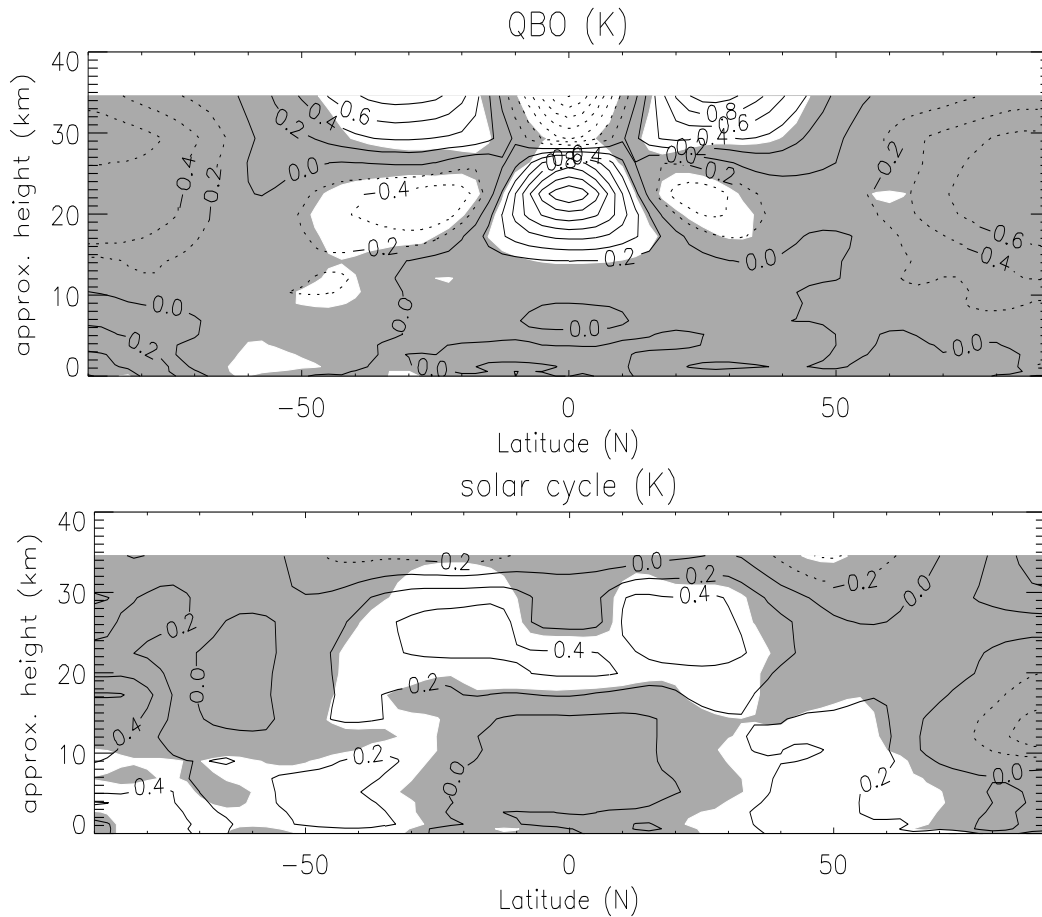


Fig. 4.3: Amplitudes of the component of variability in zonal mean temperature due to: above) QBO; below) solar. Significant region up to 95% are marked by white.

[After Haigh, 2003]

Labitzke and van Loon (1992) plotted a scatter diagram (shown here with Fig. 4.3.1a) that depicts values of 30 hPa temperature at the north pole for each year in Jan/Feb as

ordinate and solar 10.7 cm flux along abscissa. Phase of the QBO was marked with symbols; square for westerlies and triangle for easterlies. The demarcating horizontal and vertical line have been drawn to indicate regions where certain phases of the QBO predominate and shown by the E and W labels. From Fig. 4.3.1a, it is clear that warm polar temperatures tend to occur during the west phase of QBO at solar maximum and east phase at solar minimum. Whereas, the cold polar temperature occur during W-ly QBO at solar minimum and E-ly QBO at solar maximum.

Recent analysis of Camp and Tung (2007) using slight different temporal and spatial coverage of winter north polar temperature, however indicate certain contradiction to that from Labitzke and van Loon (1992) relating to solar/ QBO relationship. They applied Linear Discriminant Analysis of 51 years NCEP-NCAR reanalysis data of north polar temperature during late winter (Feb-Mar). In terms of north polar temperature, they considered the mean temperature in the 10-50 hPa layer, as represented by the difference between 10 hPa and 50 hPa geopotential height surfaces. The summary of their findings are shown in Fig. 4.3.1b in a four-quadrant diagram, with solar 10.7 cm flux as abscissa and QBO (m/s) as ordinate. Following the usual convention, westerly QBO is shown with positive values whereas, easterly with negative. The demarcating horizontal and vertical lines have been drawn to indicate phases of the QBO in relation to solar max or min; thus comprising a total of four groups viz. Solar-max/E-QBO, Solar-min/E-QBO, Solar-max/W-QBO and Solar-min/W-QBO. The arrows indicate the direction of warming. According to the Fig 4.3.1b, it is clear that though the state of the westerly phase of QBO during solar minimum emerges as a distinct coldest state like Labitzke and van Loon (1992), but the easterly phase of QBO during solar maximum emerges very warm indeed. Here lies the contradiction and hence, our initial attempt is to understand the main reason of such inconsistency between the two findings (Labitzke and van Loon (1992); Camp and Tung (2007)), involving the polar temperature, QBO and Sun.

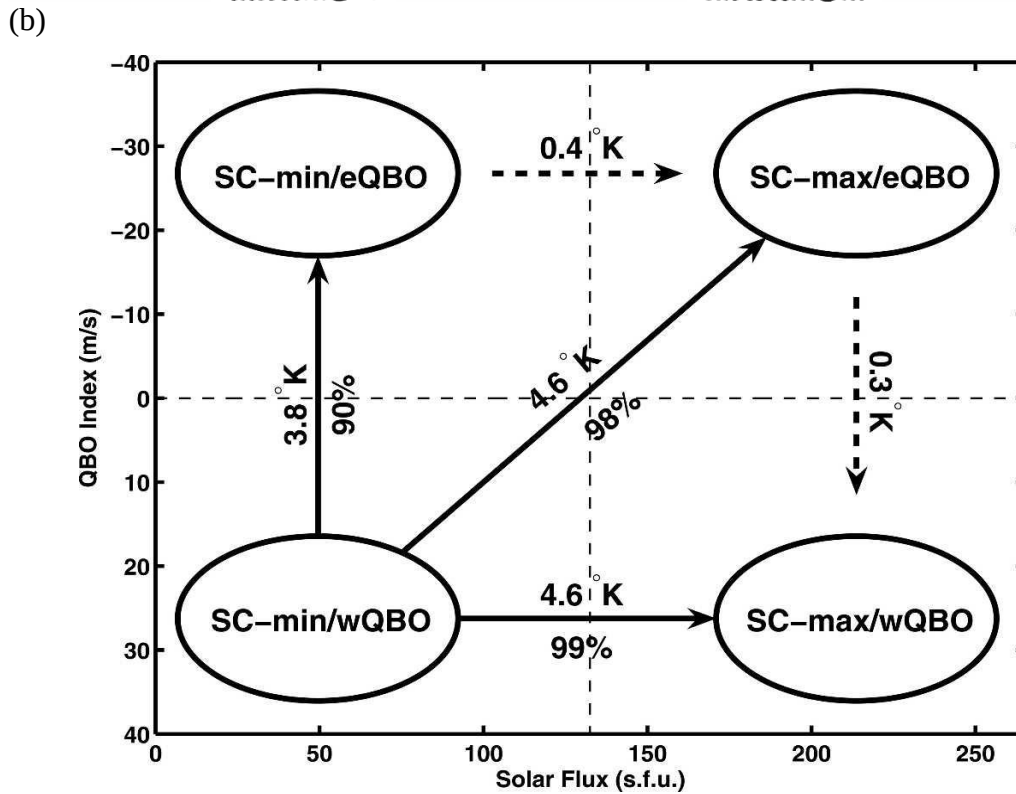
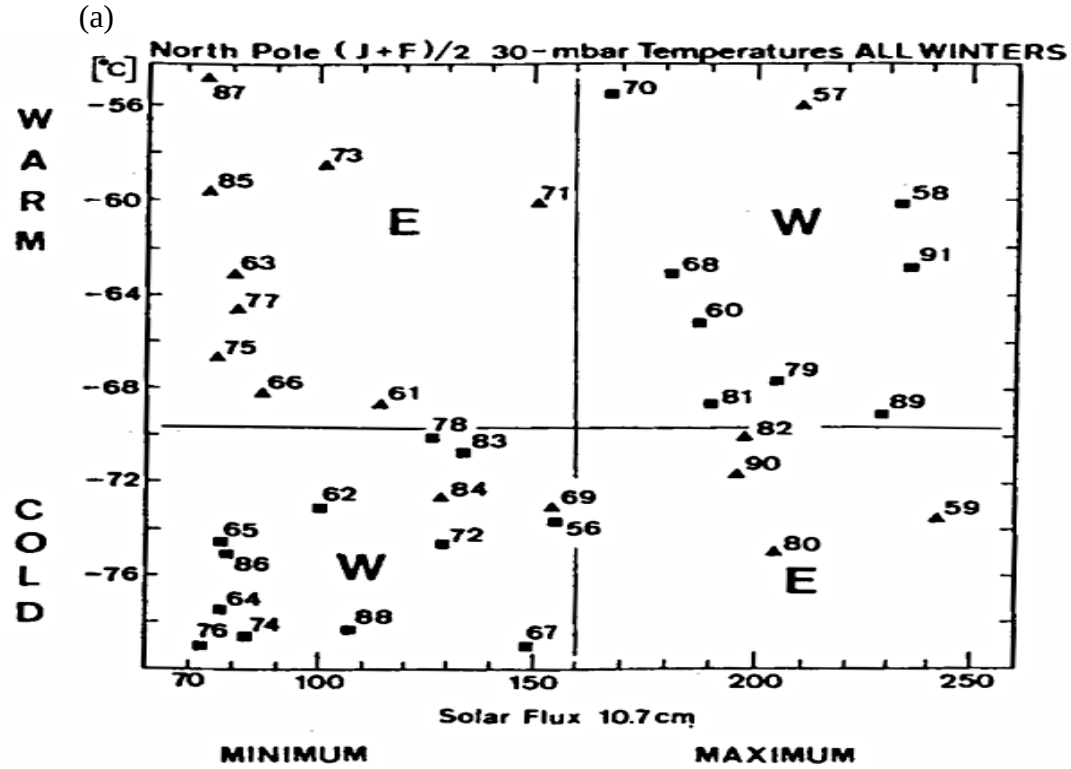
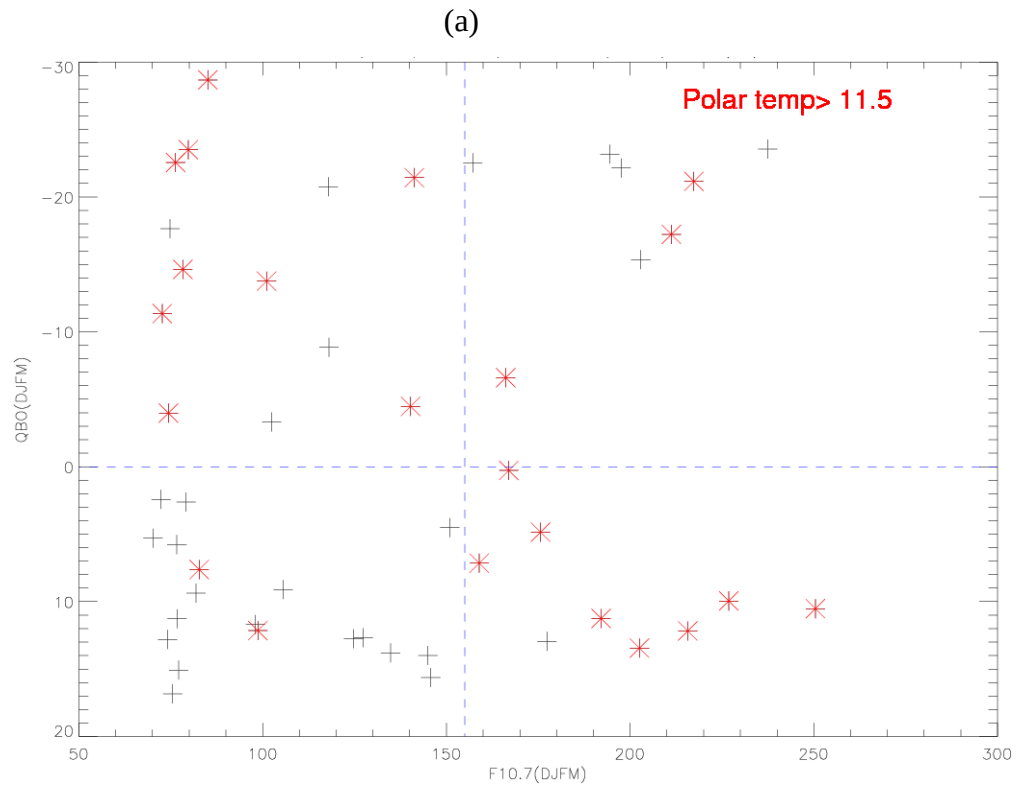


Fig. 4.3.1: Solar- QBO relationship: a) Labitzke and van Loon (1992); b) Camp and Tung (2007).

4.3.1. Polar temperature during JF with respect to QBO (40 hPa) and F10.7:

We started our work keeping as close as possible with the two above mentioned work using north polar temperature from NCEP-NCAR (National Centres for Environmental Prediction- National Centre for Atmospheric Research) reanalysis product, between period 1953-2001. First, we considered similar temporal (JF) and spatial coverage (30 hPa) of north polar temperature dataset as used by Labitzke and van Loon (1992). Following them, we used QBO at 40 hPa level (in m/s) and solar variability with solar 10.7 cm flux. We produced a scatter plot (Fig. 4.3.2a) with QBO (40 hPa) as ordinate and solar 10.7 cm flux along abscissa. Since, Salby and Callaghan (2004) observed that the QBO wind changes sign during the winter of some solar max years, Camp and Tang (2007) used average value of the QBO during DJFM (Dec-Jan-Feb-Mar) to reduce monthly fluctuation. Here, in our analysis, we used the average value of DJFM (Dec-Jan-Feb-Mar) for both the QBO and solar F10.7. A demarcating horizontal line has been drawn to separate the E ly (shown with negative value) to that from the W ly (shown with positive value); whereas, a demarcating vertical line has been drawn above 155 of solar flux F 10.7 (like Labitzke and van Loon (1992)), to segregate higher solar years to that from the lower solar years. These vertical and horizontal line generate a total of four groups representing various combination of the solar and QBO, viz. Solar-min/W-QBO, Solar-min/E-QBO, Solar-max/E-QBO and Solar-max/W-QBO. Polar temperatures anomalies with sufficient high value (say, greater than 11.5K (chosen, to indicate good contrast in schematic)) have been marked by red colour. From Fig. 4.3.2a, it is seen that the quadrant where almost all the points are red lies in the box representing solar max/W-QBO; whereas the quadrant, with least population of red spots represents solar min/W-QBO – thus consistent with Labitzke and van Loon (1992).

To reproduce the same work in a form Fig. 4.3.1b like Camp and Tung (2007), we presented the above mentioned work in a four quadrant box (Fig. 4.3.2b), where each quadrant represents the same state of solar and QBO relationship as demarcated by dotted lines in Fig. 4.3.2a. In each box, inside an oval, we mentioned the state of the solar and QBO and showed the average value (in red) of all the polar temperatures (in K) within



(b)

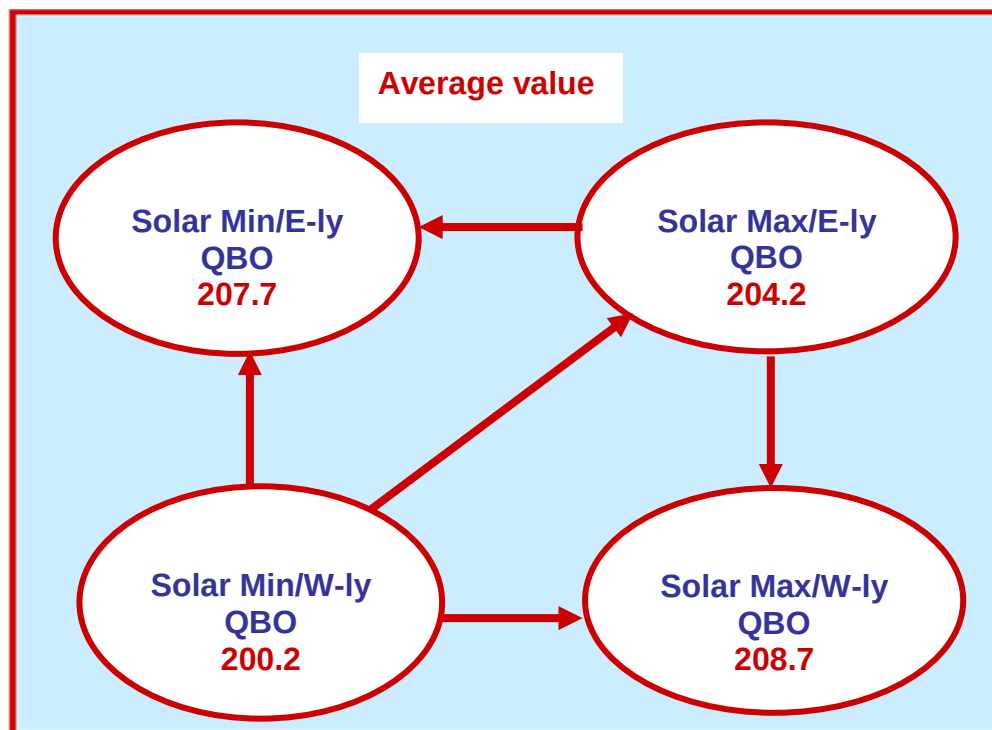


Fig. 4.3.2: Polar temperature during JF in different quarters of solar and QBO (40 hPa): a) absolute value ; b) average value.

the respective quadrant. In calculating the average value, we omitted years that are adversely affected by the volcano (thus excluding '92, '83 and '64, as we restrict to JF); such exclusion follows Labitzke and van Loon (1992). Arrows point from the cold to warm (likewise Camp and Tung, 2007) and thus indicate the direction of warming. From Fig. 4.3.2b it is quite clear that the box representing solar Min/ W ly QBO indicate the coldest polar temperature (with average 200.2K); whereas, the quadrant, representing solar **Max/ W-ly is seen to be the warmest** with an average value of 208.7K, followed by solar Min/ E-ly (with average value 207.7K). Notably, a warming of around 8° in these two quarters in respect to solar Min/ W-ly QBO is clearly noticed in Fig. 4.3.2b. Hence, to summarize the observations of Fig. 4.3.2a & b, it can be stated that using the QBO height at 40 hPa, we notice: cold polar temperature tend to occur during W-ly QBO at solar minimum and E-ly QBO at solar maximum; whereas, warm polar temperatures occur during the west phase of QBO at solar maximum and east phase at solar minimum. Thus our result is in agreement with the work of Labitzke and van Loon (1992) in relation to the north polar temperature, QBO (40 hPa) and solar. In an updated version, Labitzke (2004) using QBO height 45 hPa also verified such observation.

4.3.2. Polar temperature during JF with respect to QBO (30 hPa) and F10.7:

Here, we reproduce the same work, with **only one exception** - in place of QBO at 40 hPa level (as used by Labitzke and van Loon (1992)), we repeated the same analysis using QBO at 30 hPa level (used by Camp and Tung (2007)).

After carrying on with similar analysis, like Fig. 4.3.2a (not shown here), we calculated the average value of north polar temperature during different phases of solar and QBO (**30 hPa**) combination. Such representation is depicted in Fig. 4.3.3 which is done in a similar manner like Fig. 4.3.2b. From Fig 4.3.3, it is evident that the quadrant representing solar Min/ W ly is again the coldest (with average temperature 202.3K). But, surprisingly **here solar Max/ E ly appears to be the warmest** (with average 207.7K), followed by solar Max/ W ly (average 206.4K). Now, let us discuss about the direction of warming, as indicated by red arrows. In Fig 4.3.3, if we focus on the quarter

solar Min/ W ly, we observe that arrows indicating the direction of warming remains the same, as is seen in Fig. 4.3.2b - suggesting that this particular quarter is still the coldest of all. However, if we focus on the quarter solar max/ E ly, we surprisingly observe that the direction of two arrows around this quarter is different to that from Fig. 4.3.2b (hence marked by dotted red lines). Thus, simply using QBO at 30 hPa level, in place of 40 hPa, is seen to have even changed the direction of warming in some quarters, suggesting solar Max/ E ly is the warmest one. Such observation is in accordance with Camp and Tung (2007), who also noticed solar Max/ E ly is remarkably warm. Our analysis thus identifies that the use of different QBO height might be the major reason of indicating some contradicting results, in relation to polar temperature, QBO and solar. Therefore, Labitzke and van Loon (1992) and Camp and Tung (2007) appear different largely because of their choices of QBO height.

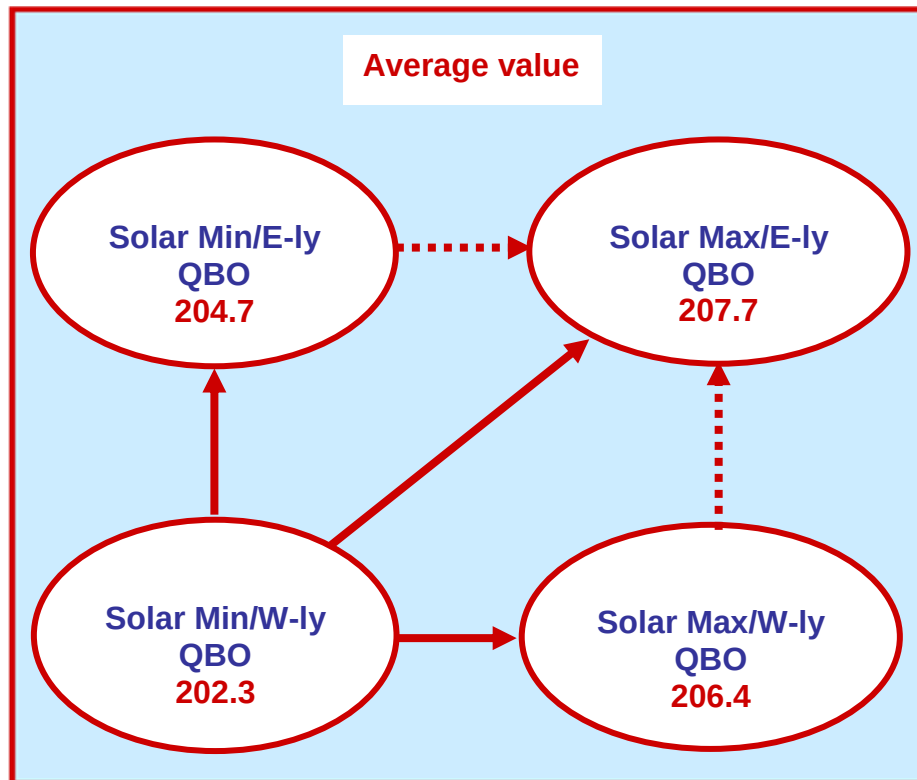


Fig. 4.3.3: Polar temperature (average value) during JF in different quarters of solar and QBO (30 hPa)

To emphasize our observation, we have plotted time series of the QBO, using average of DJFM (Dec-Jan-Feb-Mar), for three different QBO heights (hPa) viz., 30, 40 and 50, as shown in Fig. 4.3.4. In this figure, QBO at 40 hPa is marked with blue, 30 hPa with

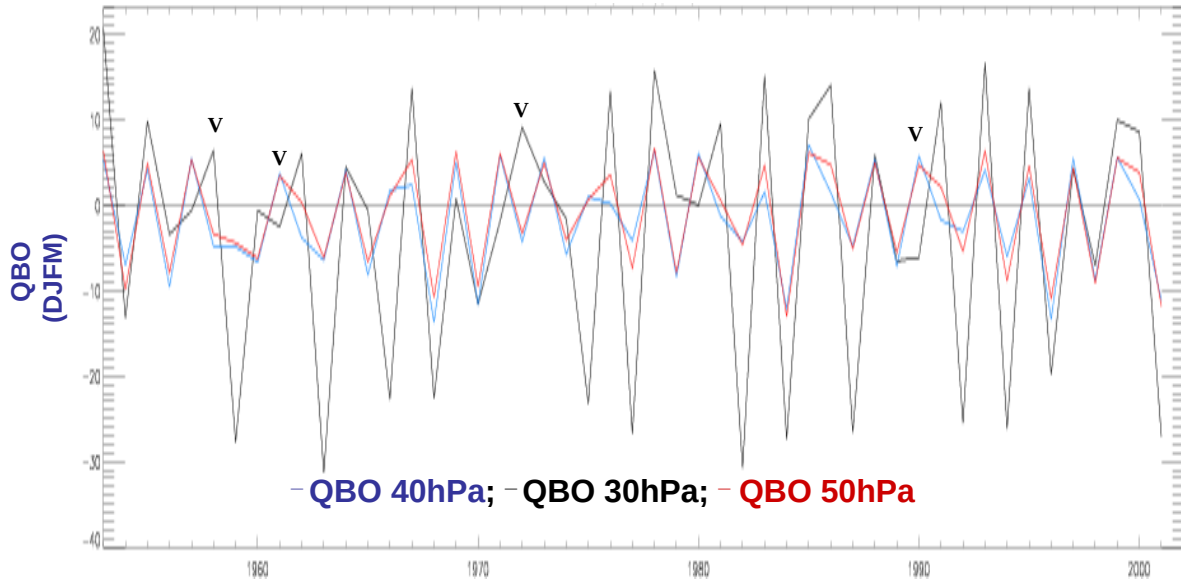


Fig. 4.3.4: Time series of QBO (DJFM) in different QBO height

black and 50 hPa with red. From this figure, it is apparent that there are a number of places (marked by 'V'), where the QBO at 30 hPa is seen to be out of phase with QBO at 40 hPa; that means, on occasions, QBO at 30hPa is seen to be easterly (westerly) though westerly (easterly) using height 40 or 50 hPa level. However, interestingly, QBO at 40 or 50 hPa does not suffer much phase change, indicating that results might not be very sensitive using QBO at 40 or 50 hPa level. Thus Labitzke (2004) using QBO height 45 hPa arrives at same result of Labitzke et al. (1992), who used QBO at 40 hPa.

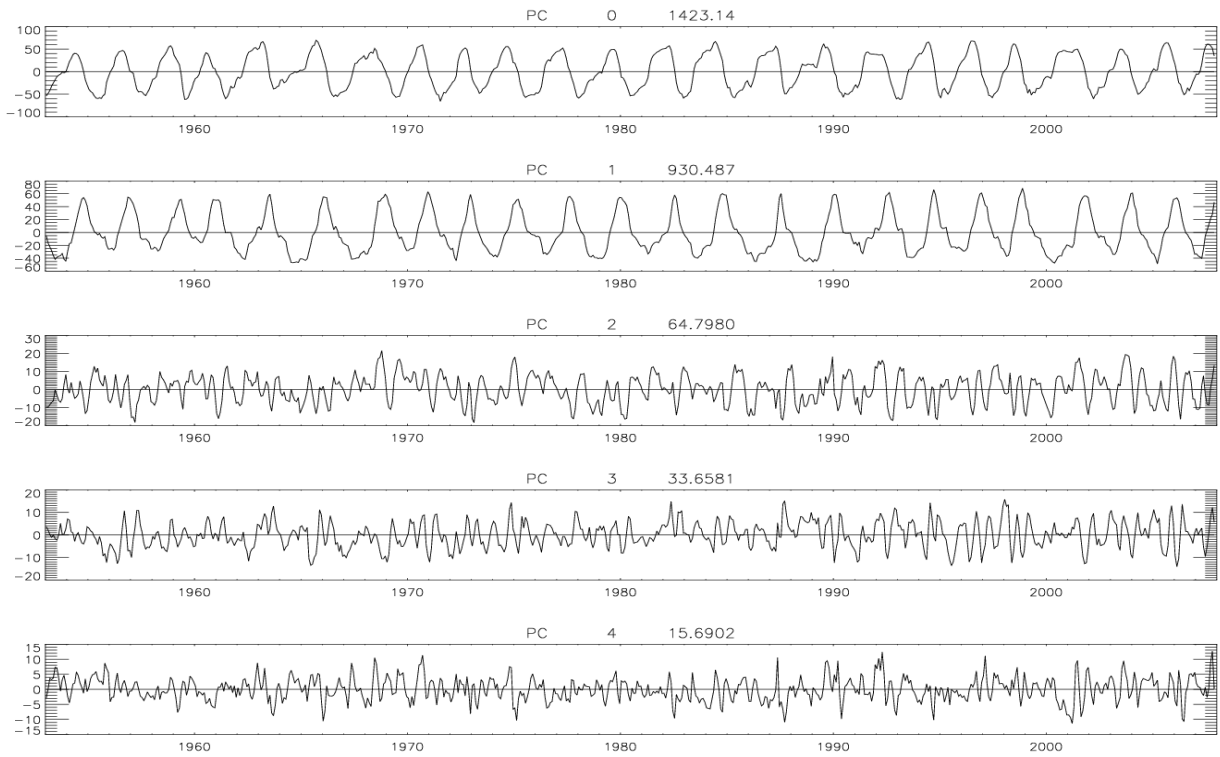
4.3.3. Time series of QBO at different height and EOF analysis:

As mentioned in chapter 1, the QBO propagates downward with a speed of $\sim 1\text{km/month}$. Fig. 4.3.5 shows time series of the QBO, at different heights (in hPa) i.e., 15, 20, 25, 30, 35, 40, 45, 50, 60 and 70. During the course of propagation apart from suffering change



Fig 4.3.5: QBO time series at different levels (hPa); 15, 20, 25, 30, 35, 40, 45, 50, 60, 70.

(a)



(b)

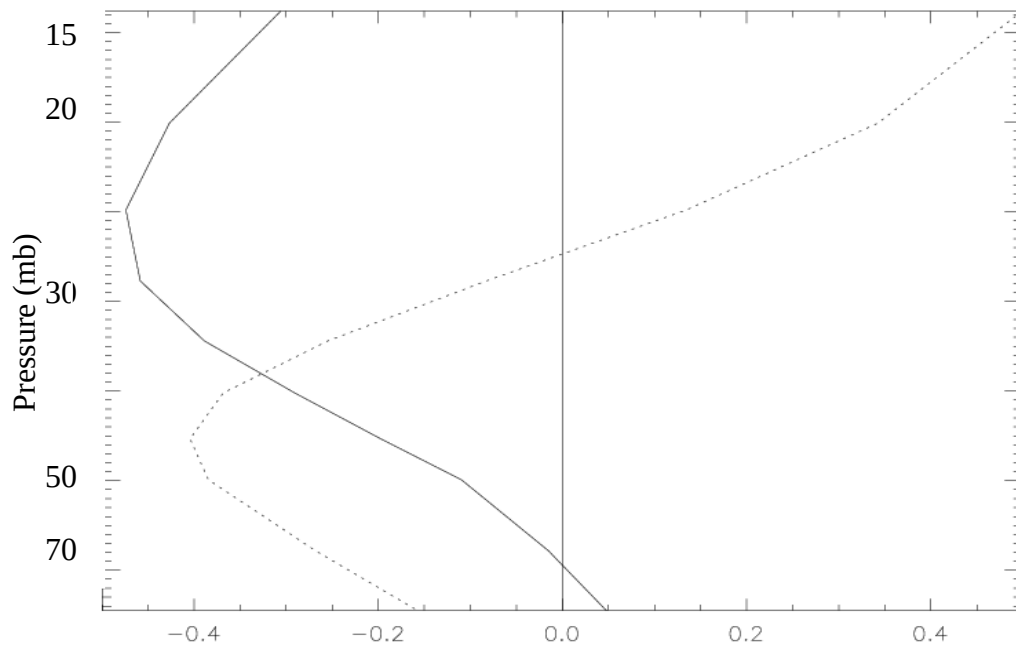


Fig.4.3.6: EOF analysis: a) first few PC time series; b) first two EOF spatial pattern

of phase, they also show considerable variability, in terms of amplitude. It clearly suggests that not only some results, using data sorting, are sensitive to the QBO height (as shown earlier), but also results using other techniques might be sensitive to its height. For instance, the use of different QBO heights in the regression is liable to produce different results and consequently carries different spurious implications. Thus to have an understanding of the spatial and temporal variability of the QBO, EOF technique is applied. Fig. 4.3.6a shows the first few PC time series whereas, Fig. 4.3.6b depicts the first two spatial structures. PC0, which is the first principle component time series, explains 58% variance; whereas PC1, the second principle component time series, explains ~37%. Moreover, from Fig. 4.3.6b, it is observed that their respective spatial patterns, i.e., EOF 0 peaks around 25 hPa level; while EOF 1 around 48 hPa. It is interesting to note: these 2 PCs include 95% of the overall variance so that the remaining PCs carry nominal implications and may be ignored; moreover, being orthogonal in nature and converging approx to the time series near 25 and 48 hPa, indicate that using the time series at ~30 and 40/50 hPa is equivalent to using PC0 and PC1.

The results of the components of variability of QBO on SLP during DJF, using QBO at 30 hPa and 50 hPa level respectively are shown in Fig. 4.3.7. The other independent parameters used are OD, trend, ENSO and solar (using monthly SSN). Using a QBO signal at 30 hPa resembles a positive NAM pattern, whereas at 50 hPa it shows a dipole pattern at mid-latitudes across the Pacific and Atlantic Oceans, along with a weaker signal in the tropical Pacific. The results of the component of variability in the multiple regression on SLP, using PC0 or PC1 separately (with OD, trend, ENSO and solar as independent parameters) also indicate similarly, as using QBO at 30 hPa or 50 hPa respectively, as is expected (not shown here).

Here we present some of our results of the multiple regression analysis on SLP during the last 50 years period, using trend, OD, solar indices (using monthly SSN), ENSO and QBO as independent parameters. Amplitudes of the component of solar variability on SLP, with other independent parameters as trend, OD and ENSO, are shown in Fig. 4.3.8 (top); whereas the same with other independent parameters trend, OD, ENSO, PC0 and PC1 are shown in that figure at the bottom. From Fig. 4.3.8, it emerges that the solar

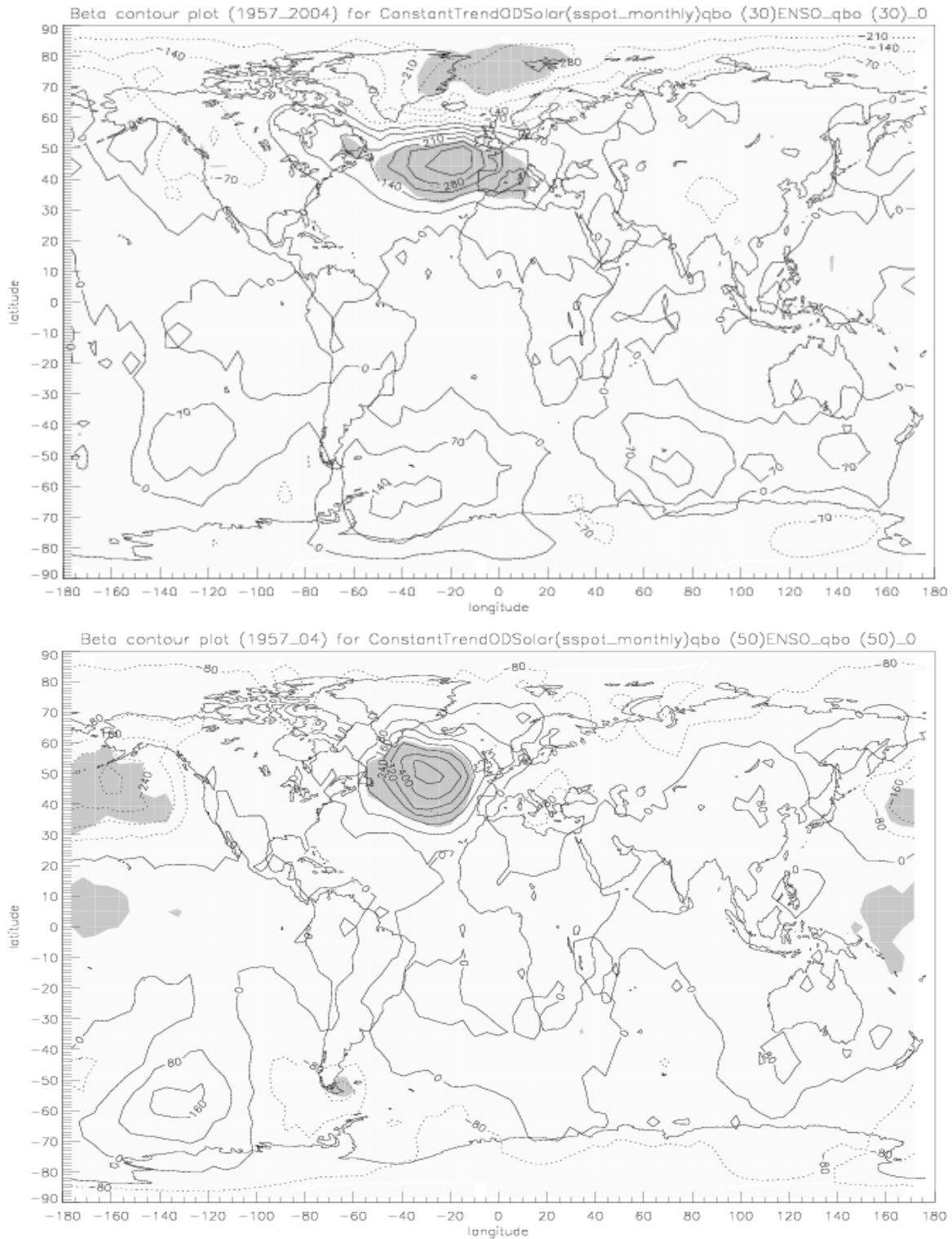


Fig.4.3.7: Amplitudes of the components of variability of SLP due to QBO during DJF for the period 1957-2004 with other independent parameters as trend, OD, solar (using monthly SSN) and ENSO; QBO level used 30 hPa (top) and 50 hPa (bottom).

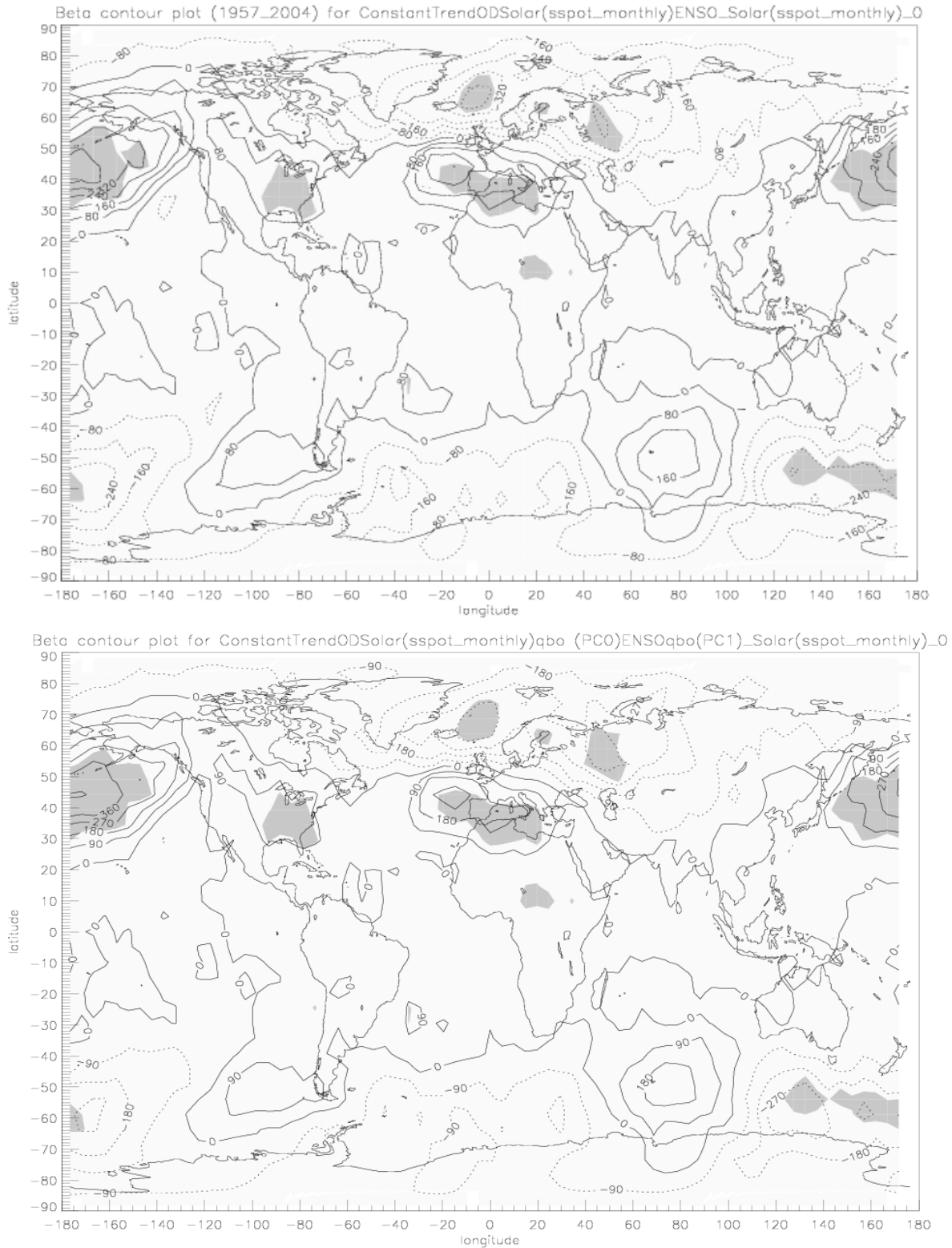


Fig.4.3.8: Amplitudes of the components of variability of SLP due to solar (using monthly SSN) during DJF for the period 1957-2004 with other independent parameters: top) trend, OD and ENSO; bottom) trend, OD, ENSO, PC0 and PC1.

signal with/without both the PC0 and PC1 appears to be the same - thus suggesting the robustness of the solar signal.

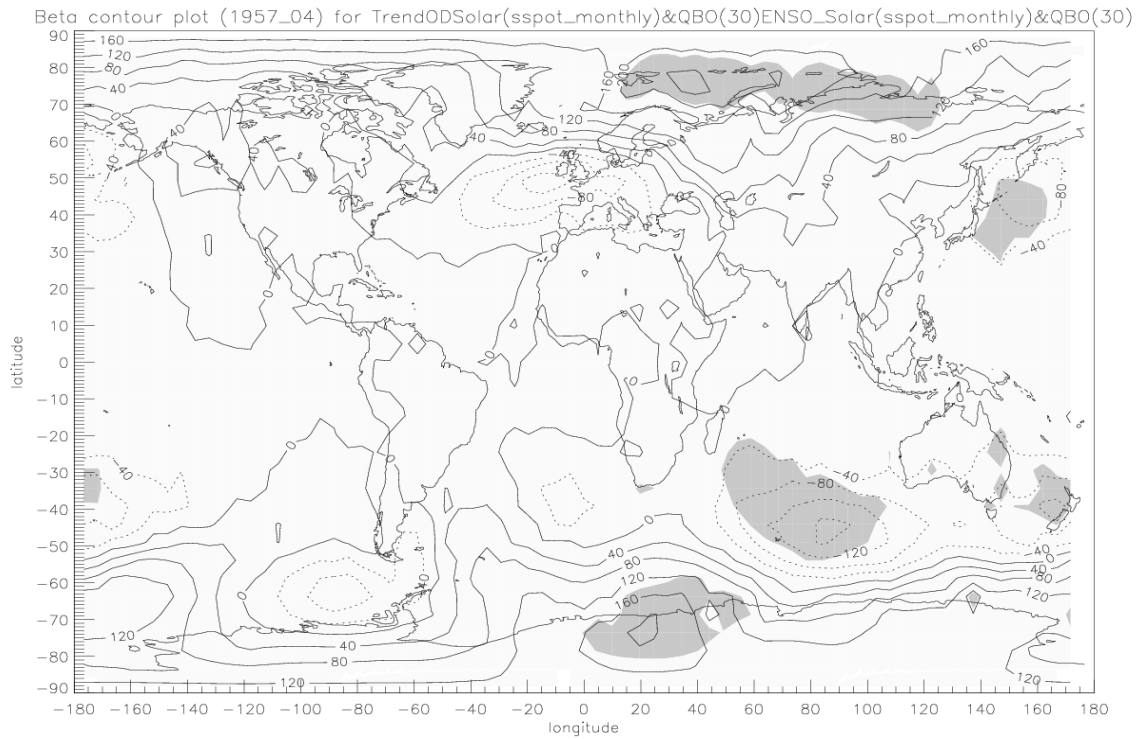
4.4. Combined Effects - Solar with QBO

Haigh and Roscoe (2006), through multiple regression analysis (using NCEP data during 2nd half of the last century), showed that there is no statistically significant solar signal in either the surface NAM or SAM; however, when a new index, the product of the solar and QBO indices, is used there is a good correlation throughout the atmosphere in the SAM, and at the lower levels in the winter NAM. Here following Haigh and Roscoe (2006) we used the index, the product of the solar and QBO in the regression which is described below.

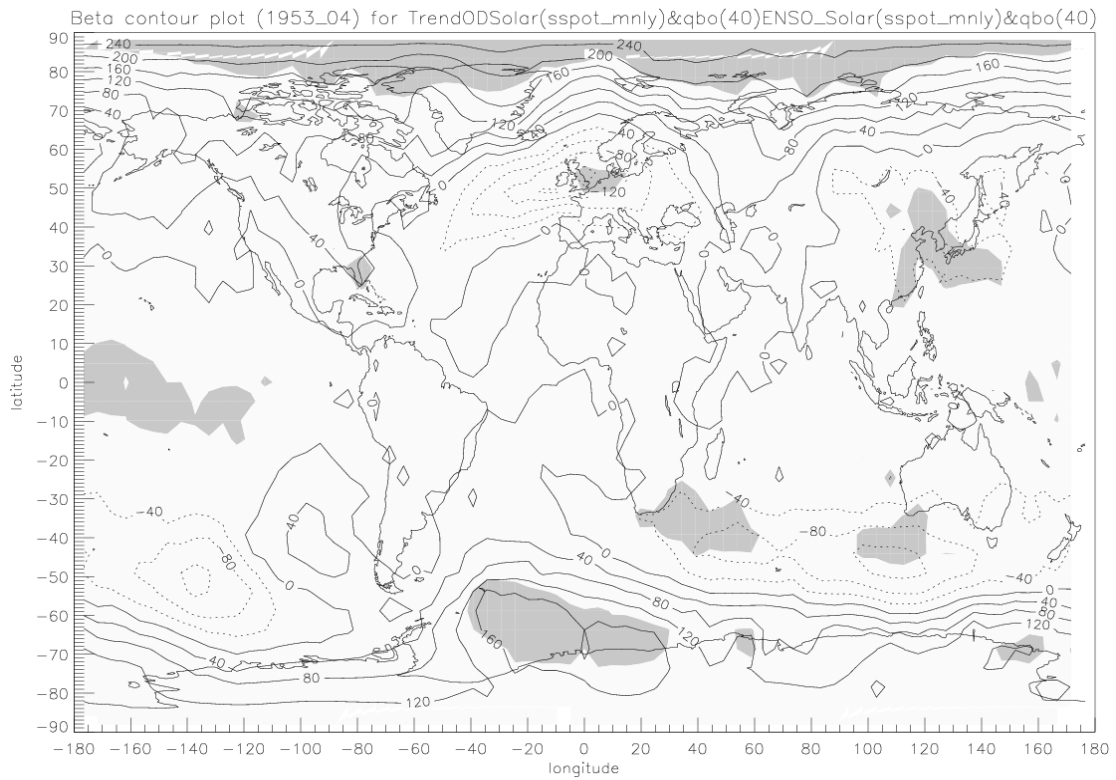
4.4.1. Regression of SLP with ‘Solar*QBO’

Perturbations in the polar vortex appear to be related to the downward propagation via polar annular modes was shown by Baldwin and Dunkerton (2001) through observational analysis. Thompson and Wallace (2000) also identified such downward propagation of both the polar modes. The large-amplitude variations in strength of the stratospheric polar vortex are typically followed with a lag of less than one month by similar signed anomalies in the tropospheric circulation that persists for up to 2 months (in NH) to 3 months (in SH) (shown in Fig. 4.2.3 & 4.2.4 of the thesis). Based on their results we carry the regression analysis, using all months of the year. Here we use a new index, the product of the solar and QBO and termed as ‘solar*QBO’, (also used in the regression analysis of Haigh and Roscoe (2006)) as it well captures the behaviour of the polar temperature in terms of both the solar and QBO (following the analysis of Labitzke et al. (1992) and Labitzke (2004)). This updated version of Labitzke (2004) even showed that the observed behaviour in terms of the polar temperature, QBO and Sun is not only present in the NH but also in the SH.

(a)



(b)



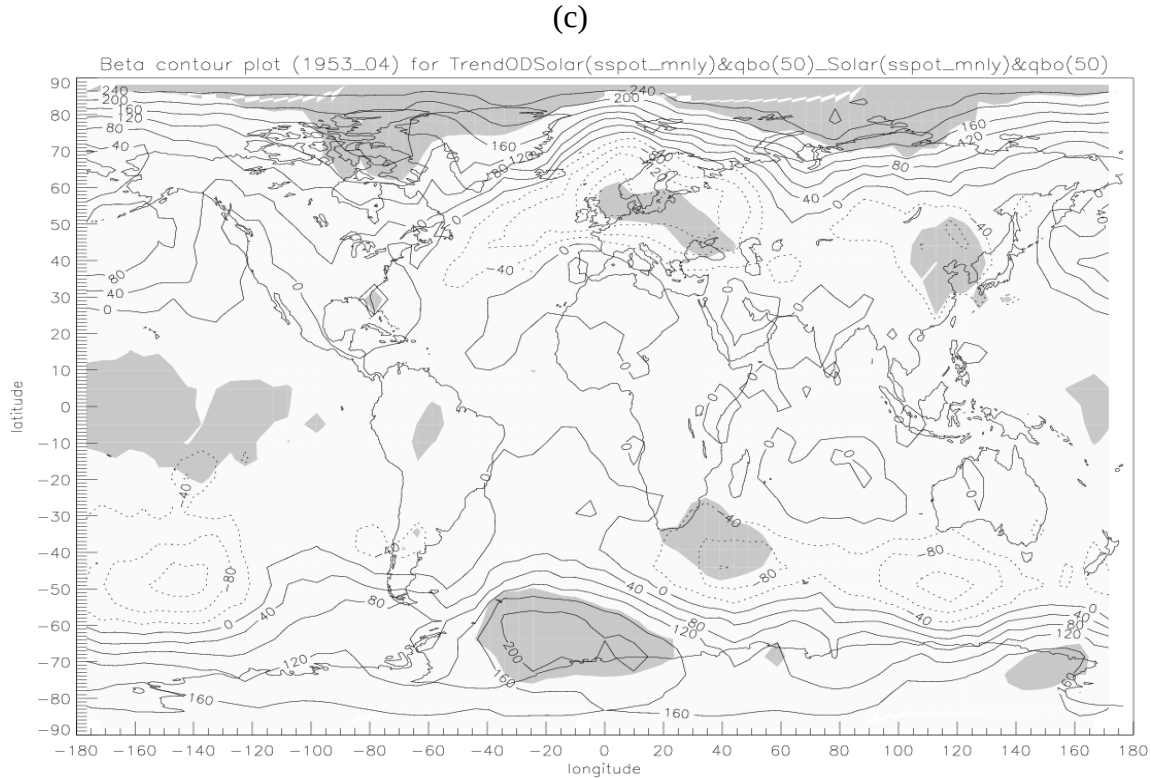
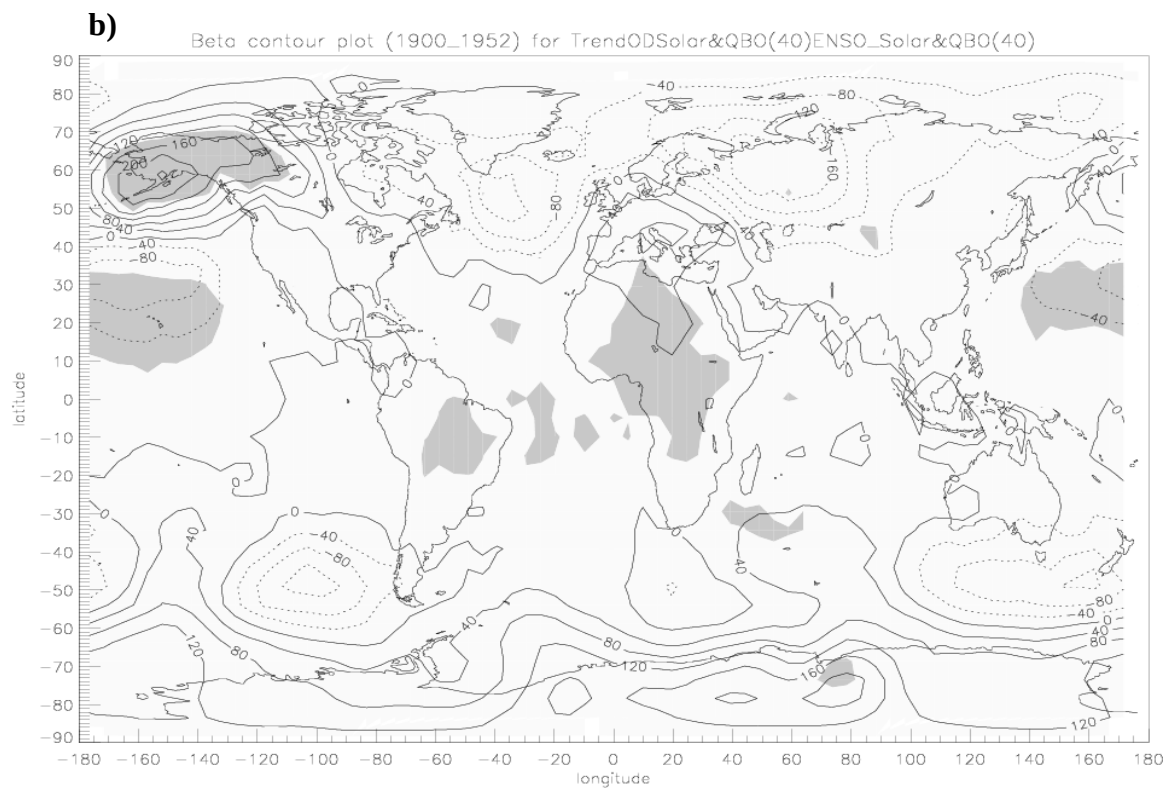
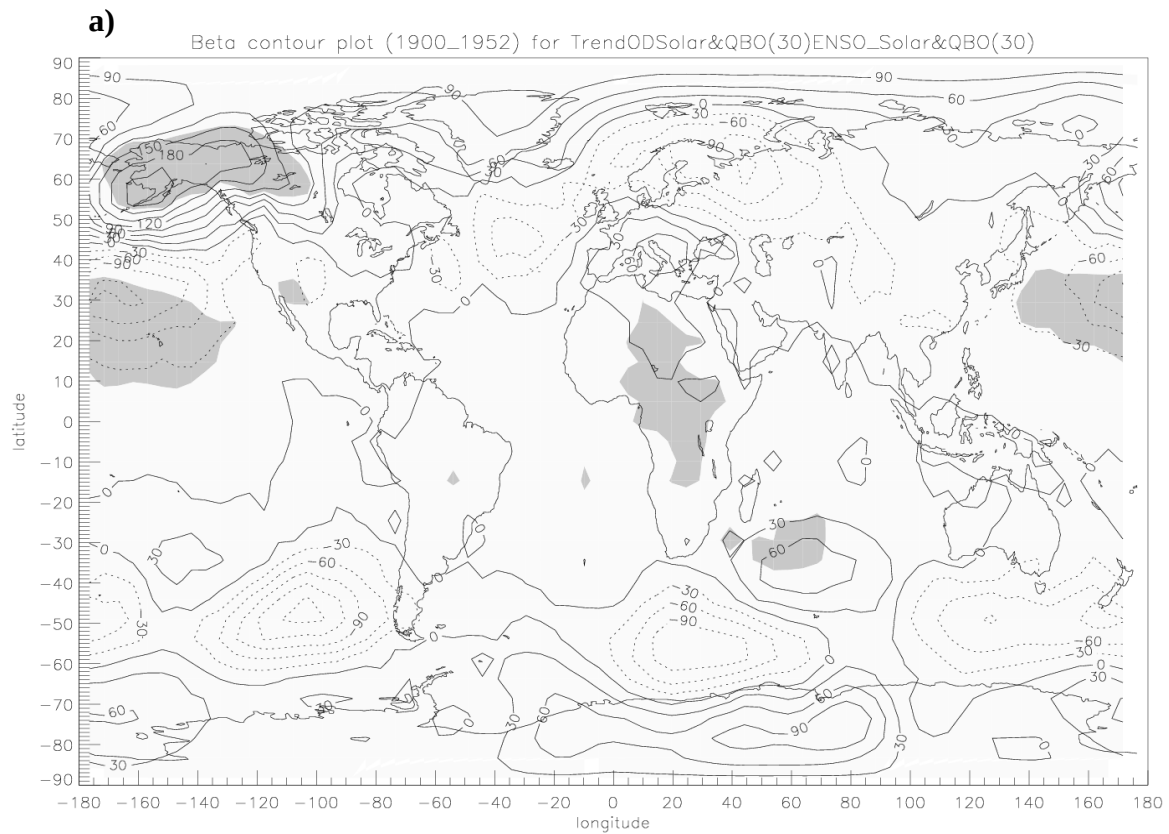


Fig. 4.4.1: Amplitudes of the component of variability of SLP due to Solar*QBO using different QBO height (during 1953-2004): a) 30 hPa; b) 40 hPa and c) 50 hPa. Other independent parameters used are: trend and OD. Regression was done using all months of the year.

Fig. 4.4.1 shows the amplitude of the component of variability on SLP due to this new index the ‘solar* QBO’, with other independent parameters as trend, OD and ENSO. In the analysis, different QBO heights are used: 30 hPa (Fig. 4.4.1a); 40 hPa (Fig. 4.4.1b) and 50 hPa (Fig. 4.4.1c). Interestingly, in all these figures, negative NAM and SAM features are quite evident suggesting that HS years during westerly phase of the QBO and LS years during easterly phase of the QBO, both trigger negative surface NAM and SAM features in zonal sense; whereas the reverse is true during HS years with the easterly QBO phase and LS years during the westerly QBO phase. Small differences between three cases (but generally patterns are same), for three different QBO height, are believed to be due to the shift in phase of QBO with height. Such observations indeed suggest that the effect of QBO, (irrespective of the QBO height at 30, 40 or 50 hPa level) combined with solar activity reveals a negative signal in both the de-seasonalised polar annular



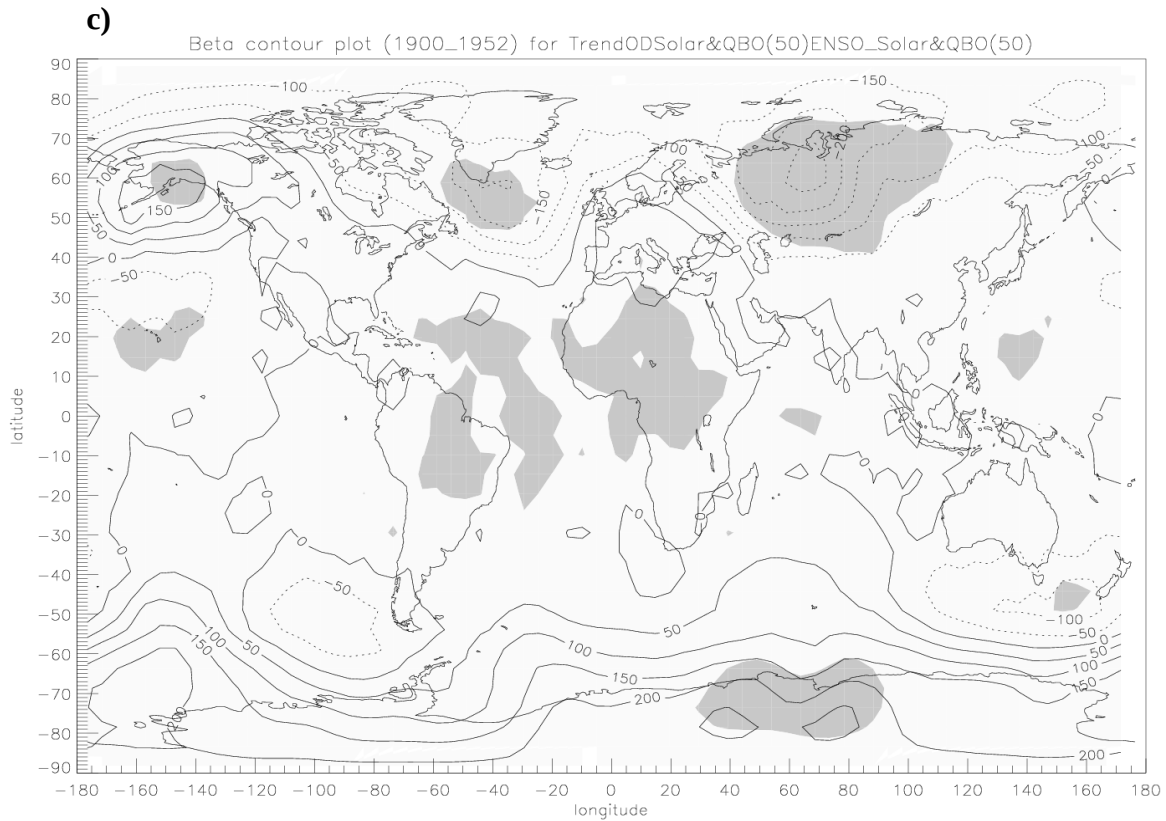
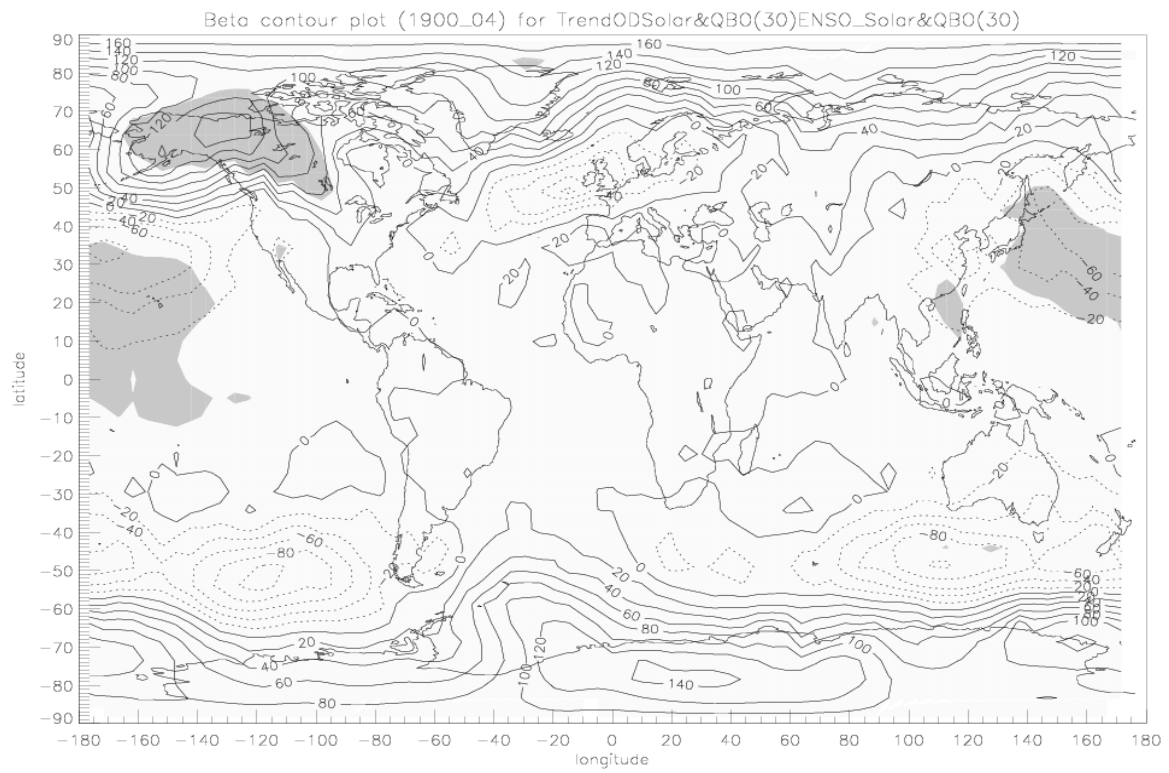


Fig. 4.4.2: Amplitudes of the component of variability of SLP due to Solar*QBO using different QBO height (during 1900-1952): a) 30 hPa; b) 40 hPa and c) 50 hPa. Other independent parameters used are: trend and OD. Regression was done using all months of the year.

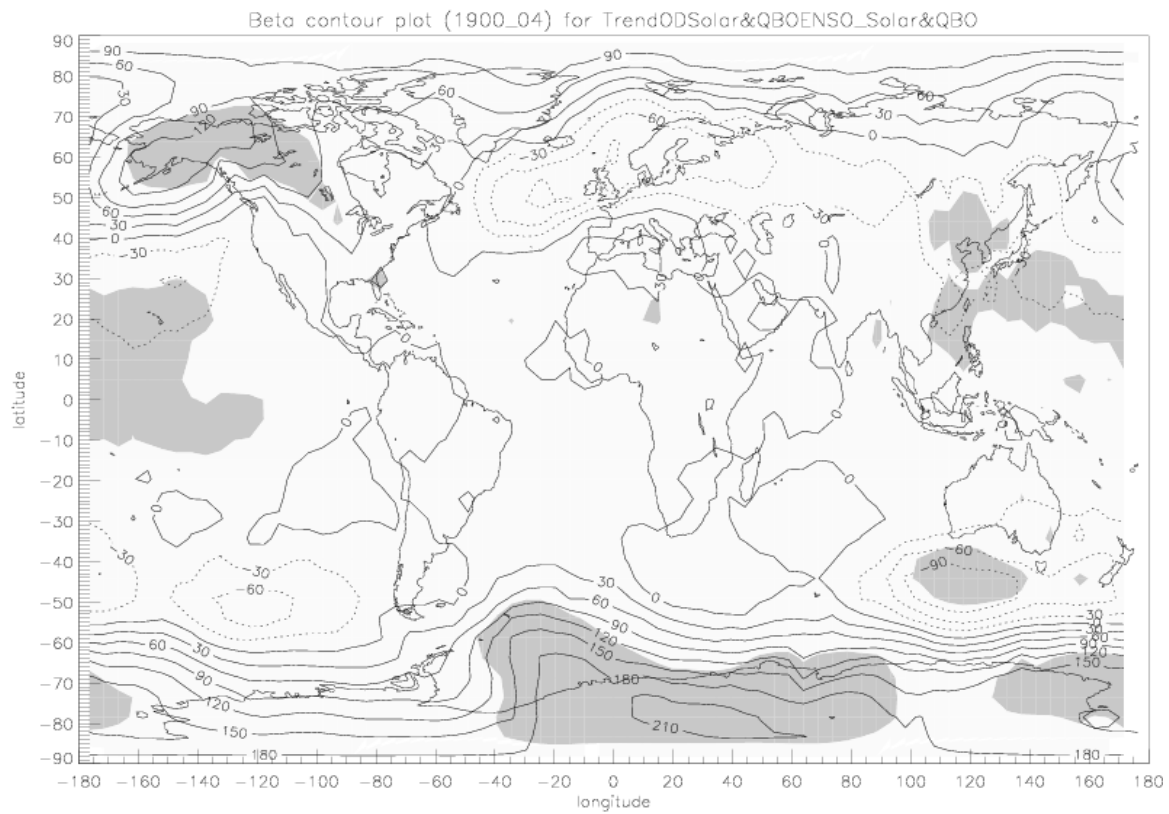
modes. Thus, our result is in agreement with Haigh and Roscoe (2006), who in their index ‘solar*QBO’, using QBO at 40 hPa level, also arrive at a similar sense of zonal features in regard to polar annular modes. In our regression, in place of the ‘solar*QBO’ time series, we have also used solar*QBO(PC0) and solar*QBO(PC1) time series (not shown here). As expected, the former gives similar results using solar*QBO(30); while the latter using solar*QBO(40).

The recent publication of Brönnimann (2007), introducing a reconstruction of the QBO dating back to 1900, enables us to analyse the role of the QBO over a longer term climate record. We have repeated our analysis, combining the QBO and solar indices, over this new time series. This shows that the general behaviour observed during 1953-2004, also prevails during the first 50 year period (1900-1952) in the SH, irrespective of the QBO

a)



b)



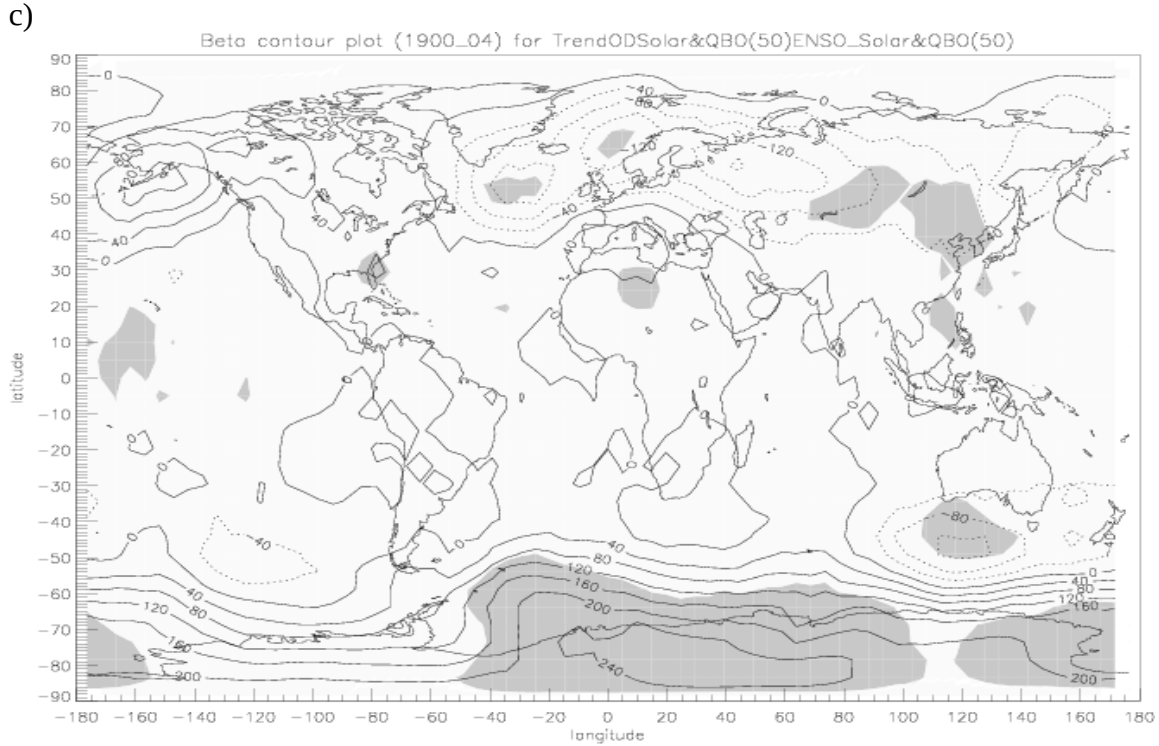


Fig. 4.4.3: Amplitudes of the component of variability of SLP due to Solar*QBO using different QBO height (during 1900-2004): a) 30 hPa; b) 40 hPa and c) 50 hPa. Other independent parameters used are: trend and OD. Regression was done using all months of the year.

height (Fig. 4.4.2 a,b and c). However, in the NH, the behaviour during the last 50 years (Fig.4.4.1) is only identified during the first 50 years if the QBO at 30 hPa level is used (Fig. 4.4.2 a). In the regression, if the whole of 100 years period is considered, the negative features of the polar annular modes are evident using the QBO height 30, 40 or 50 hPa level for SAM and 30, 40 hPa level for NAM (Fig. 4.4.3 a,b and c, respectively).

4.5. Summary

In this chapter, we first discussed about middle atmosphere coupling describing the role of both the QBO and the Sun in relation to stratospheric polar temperature. Following different studies of Baldwin and co workers (1999, 2003) and Thompson et al. (2000)

that indicate perturbations in the polar stratosphere propagate downwards, affecting even the lower troposphere over the next few months, we subsequently focused to detect the combined effect of QBO and solar, in the lower troposphere.

However, our first attempt was to understand the main reason for the apparent inconsistency between two published results relating to the polar temperature, the Sun and the QBO. Using a slightly different temporal and spatial coverage of winter polar temperature data, Labitzke et al. (1992) and Camp and Tung (2007) arrives at some contradictory findings: according to Labitzke et al. (1992), solar max/ QBO E-ly is cold, while Camp and Tung (2007), detects the same as very warm. Our analysis indicates that they appear different largely because of using a different QBO height. The QBO at 30 hPa (used by Camp and Tung (2007)) is seen to be out of phase with QBO 40 hPa (used by Labitzke et al. (1992)) in a number of times; that means, on occasions, QBO at 30hPa is seen to be easterly though westerly using height 40 or 50 hPa level and vice versa. Hence the use of different QBO height appears to be the most likely cause to influence the results of data analysis. However, interestingly, between 40 and 50 hPa, the QBO does not suffer much phase change, indicating that the results might not be very sensitive using QBO at either 40 or 50 hPa level.

Since results appeared susceptible to QBO height, to understand the major variability patterns of the QBO, temporally as well as spatially, we carried out an EOF analysis. Our analysis suggests, PC0, the first principle component time series of QBO explains ~58% of the overall variance; whereas the PC1 (the second component) explains ~ 37%. Moreover, their respective spatial patterns, viz. EOF 0 and EOF1 peaks around 25 hPa and ~48 hPa level respectively. In our multiple regression analysis, we used PC0 as well as PC1 in the same analysis, apart from OD, trend and ENSO, for the purpose of detecting solar variability on SLP. It is seen that in spite of using PC0, PC1, OD, trend and ENSO as other independent parameters in the same multiple regression, the solar cycle signal still preserves its robustness.

Different analysis of Baldwin and co workers (1999, 2001, 2003, 2005) and Thompson et al. (2005) showed that fluctuations in the strength of stratospheric polar vortices in both hemispheres are observed to couple downward to surface climate. The large-amplitude variations in strength of the stratospheric polar vortex are typically followed with a lag of less than one month by similar signed anomalies in the tropospheric circulation that persists for up to 2 months (in NH) to 3 months (in SH). To address this, we carried out the multiple regression analysis on SLP, where the regression was done using all months of the year, with other independent parameters as trend and OD.

Following Baldwin and Dunkerton (2005) who speculated that the pathway involving the polar modes of variability, appears to involve interactions of the solar with QBO, we used a new index in our multiple regression analysis. Here in place of the solar index, we used a new index - the product of solar and QBO (and termed as 'solar*QBO'), to represent the collective behaviour involving both the solar and QBO as observed by Labitzke et al. (1992) and Labitzke (2004). Such index has also been used by Haigh and Roscoe (2006) in their multiple regression analysis. Our study suggests that, the effect of QBO (irrespective of the height 30, 40 or 50 hPa) combined with solar activity reveals a negative signal in the de-seasonalised NAM and SAM for the last 50 years period. That indicates that HS years during the westerly phase of the QBO and LS years during the easterly phase of QBO, both trigger negative NAM and SAM features, in the zonal mean sense; whereas the reverse is true during HS years with the easterly QBO phase and LS years during the westerly QBO phase.

The reconstruction of QBO dating back from 1900 (Brönnimann, 2007) is now available. Similar analysis, combining the QBO and solar was carried with this new time series, which suggested that the combined behaviour, as observed during the later half of the 20th century, still prevails during the earlier period in SH, irrespective of the QBO height (30,40 or 50 hPa). Moreover, in the regression, if we consider the whole of 100 years period, we find that negative SAM features are quite evident (regardless of three different QBO height) and negative NAM features using height 30 or 40 hPa. Such observation, no doubt, can be useful for the purpose of long range surface climate prediction.

Chapter 5: Solar Signal, Climate Change and ENSO

5. Solar Signal, Climate change and ENSO

5.1. Shallow overturning circulation and ENSO

Shallow overturning circulation:

There are two major systems of ocean current: the first one is the surface current which is a wind-driven, shallow and fast moving; the second one is the deep current driven by density differences- which is mostly deep, slow moving and massive.

Throughout most of the ocean, layers form three principal zones: The surface zone (also known as the mixed layer), the pycnocline zone (pycno means density and cline means slope) and the deep zone. The surface zone is usually confined within a layer from the surface ocean to a depth of about 100m. The pycnocline separates the upper water mass and surface layer from the deep water mass and is often termed as the *thermocline* (as temperature also changes rapidly with depth). Since temperature as well as density is decreasing rapidly with depth, this zone is highly stable; mixing is suppressed and the upper and lower current system act independently. Wind acting on the surface of ocean causes a partial transfer of kinetic energy from the wind to water. Wind driven currents decline with depth, and are generally limited by the permanent pycnocline - around 100 to 200m of depth (but in some cases, even as deep as 1000m). The global pattern of wind causes major ocean currents in the surface layer. Deep and shallow waters respond to different forces and have different circulation pattern.

Solar energy governs the circulation of the atmosphere and thus the winds that drive the ocean currents in the surface layer. Atmospheric circulation has time period ranging from days/months to ~year; for ocean surface current, it is weeks/months and for deep ocean circulation, it is 500 to 3500 years (http://instaar.colorado.edu/~lehmans//env-issues/documents/S09_3520_10comp.pdf). Thus, the deep ocean circulation only plays important role in long time scale climate.

The ocean circulation mainly operates through a global scale conveyor belt shown in Fig. 5.1.1. Surface water around the North Pacific, in the vicinity of the Aleutian Low (AL) is

heated by the Sun and moves westward across the Pacific and the Indian Ocean. Crossing the equator, it reaches North of Atlantic and loses much of its heat. Being cold, it then sinks and returns to the Pacific via the Antarctic through a deep ocean current. The ocean conveyor belt is also known as *thermohaline circulation* as temperature (hence thermo) and salinity (hence haline) both drives the conveyor. In Fig. 5.1.1, the warm shallow current is shown by red colour and the cold deep ocean circulation is marked by blue.

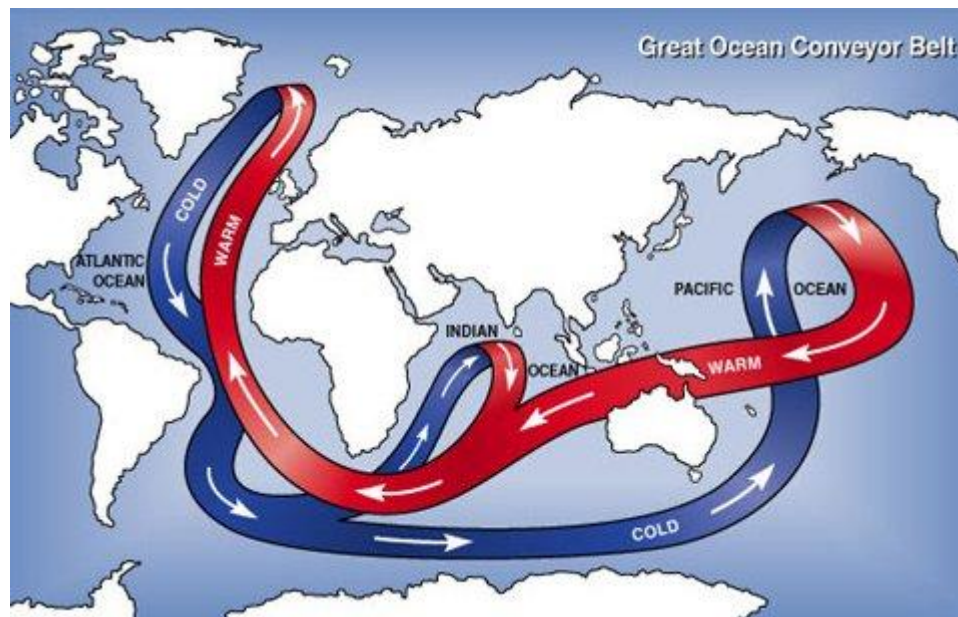


Fig. 5.1.1: Ocean conveyor belt

[http://science.nasa.gov/headlines/y2004/05mar_arctic.htm, picture by Argonne National Laboratory]

On occasion, the thermohaline circulation is referred to as the Meridional Overturning Circulation (often abbreviated as MOC). The term MOC, however, is more accurate, as it is difficult to separate the part of the circulation which is actually driven by temperature and salinity alone, as opposed to other factors such as the wind. The shallow MOC is characterised by equatorward geostrophic volume transport convergence in the interior ocean pycnocline across 9°N and 9°S . It links the tropical pycnocline to the regions of subtropical subduction and is one component of the so-called subtropical cells or STCs (McCreary and Lu, 1994). Subduction occurs over ocean basins predominantly through the gyre-scale circulation. The surface wind-stress drives anticyclonic and cyclonic

recirculations, referred to as subtropical and subpolar gyres, respectively, within a basin (Fig. 5.1.2). The wind forcing induces downwelling of surface fluid over the subtropical gyre.

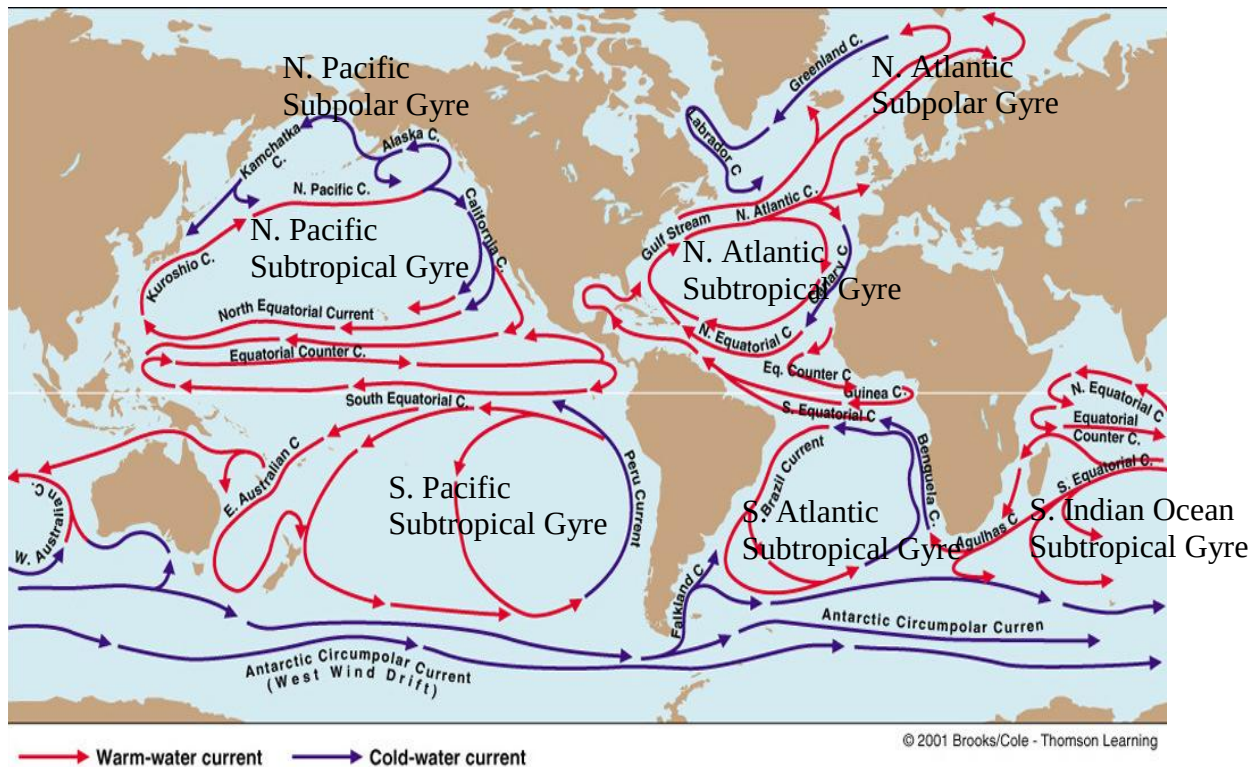


Fig. 5.1.2: Global wind driven ocean circulation

[http://atoc.colorado.edu/~dcn/ATOC1060/Members/Lectures/18_thermohaline.pdf]

In what follows we will mainly focus on the shallow MOC in the Pacific region, which may act as a bridge between tropical and extra-tropical (north) mass exchange. At least two different mechanisms have been proposed for such a connection (Yu et al., 1999): one emphasizes the existence of shallow ocean pathway that allows the north Pacific to influence ENSO through ocean subduction; the other argues that decadal-scale variations in the atmospheric general circulation over the Northern Pacific can extend to the tropics, though the slow changes in the trade wind system that precondition the mean state of the thermocline in the equatorial Pacific. It is observed that ENSO events regularly disrupt the Pacific surface flows and hence it is necessary to know more about the ENSO to understand SST in the Pacific.

ENSO:

Having defined ENSO in Section 1.1, here we address its mechanism, some of its characteristics and its relevance to the tropical-extratropical mass exchange in the Pacific, through the shallow MOC.

For the generation of ENSO, there needs to be a phase-transition mechanism able to provide a negative feedback, to reverse its cycle and account for the period associated with the cycle. The Delayed Oscillator Theory (Schopf and Suarez, 1988; Battisti and Hirst, 1989; Cane et al., 1990) is one of the most well accepted theories, which can successfully explain ENSO behaviour.

ENSO and Delayed Oscillator Theory:

The delayed oscillator suggests that the phase-transition mechanism (i.e. memory) for the ENSO cycle is provided by oceanic Kelvin and Rossby waves, forced by atmospheric wind stress in the central Pacific. The propagation and reflection of waves, along with local air-sea coupling, determine its time period.

Such a phase transition of the ENSO cycle is shown in Fig. 5.1.3 (a-d) for a warm event of the ENSO, though the movement of the thermocline is not captured in that figure. A wind forcing at the central Pacific generates a downwelling Kelvin wave (K) which propagates eastward, alongside a upwelling Rossby wave (R), that propagates westward (Fig. 5.1.3a). Figure 5.1.3b shows the first Kelvin wave reaches the eastern basin (shown with E) and causes shifting down the thermocline there (not shown in that figure). It subsequently, allows warm water of the western basin (shown with W) to move eastward causing the warm event of ENSO the El. Niño. The slow Rossby wave is reflected at the western boundary, at a later time. It then reverts back as a upwelling Kelvin wave and propagates toward the eastern basin causing uplifting the thermocline (not shown) and thus reversing the phase of the ENSO. Likewise, the Kelvin wave being reflected in the eastern basin (E) also relapses as a downwelling slow moving Rossby wave (Fig. 5.1.3c).

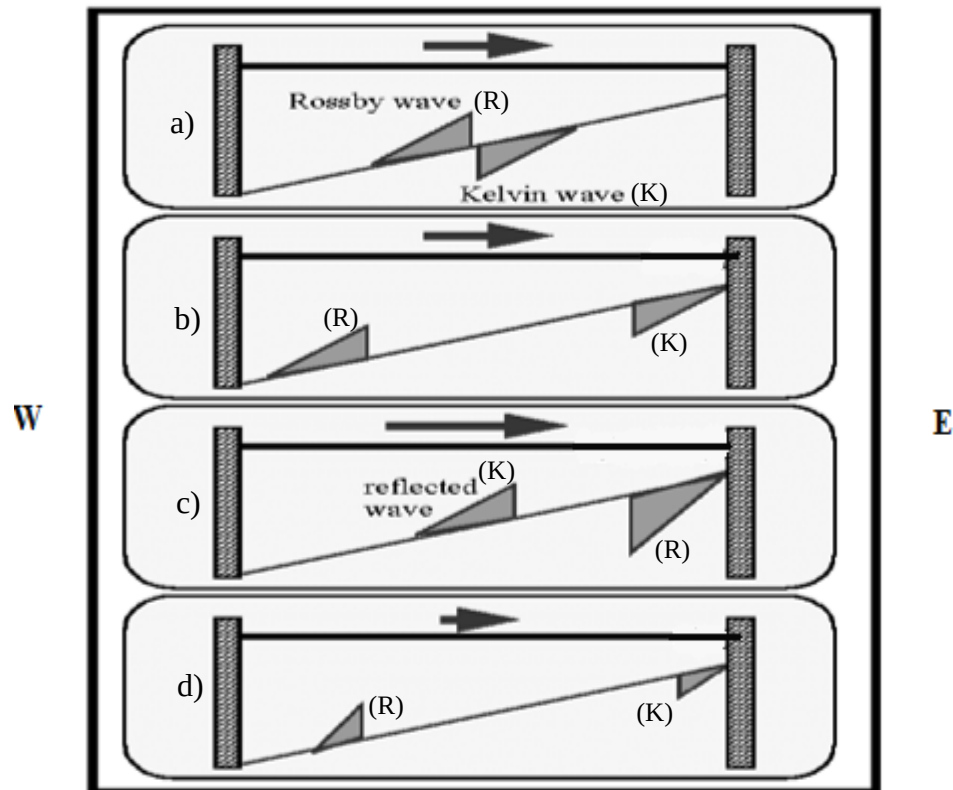


Fig.5.1.3: Phase transition of oceanic Kelvin and Rossby wave

[Source: after <http://www.ess.uci.edu/~yu/class/ess200a/lecture.7.climate.variations.pdf>]

Finally, the westward moving Rossby wave after reflection again hits the east tropical basin as a downwelling Kelvin wave, deepening the thermocline again and causing little warming at the eastern Pacific coast (Fig. 5.1.3d). The period of ENSO is determined by the propagation time of these two waves. Following a similar mechanism, the wind forcing at the central Pacific in a reverse direction can incite the cold event of ENSO.

Based on the delayed oscillator theory of ENSO, the ocean basin needs to be wide enough to produce the delaying by ocean wave propagation and reflection. This is satisfied by the Pacific Ocean. It is generally believed that the Atlantic Ocean could generate ENSO-like oscillatory behaviour, if external forcing is applied. But the Indian Ocean is too small to establish ENSO-like behaviour.

[<http://www.ess.uci.edu/~yu/class/ess200a/lecture.7.climate.variations.pdf>].

ENSO and Phase Locking:

One interesting characteristic of the ENSO is its phase locking with the season. The phenomenon usually peaks around the Christmas holidays and hence the name El Niño

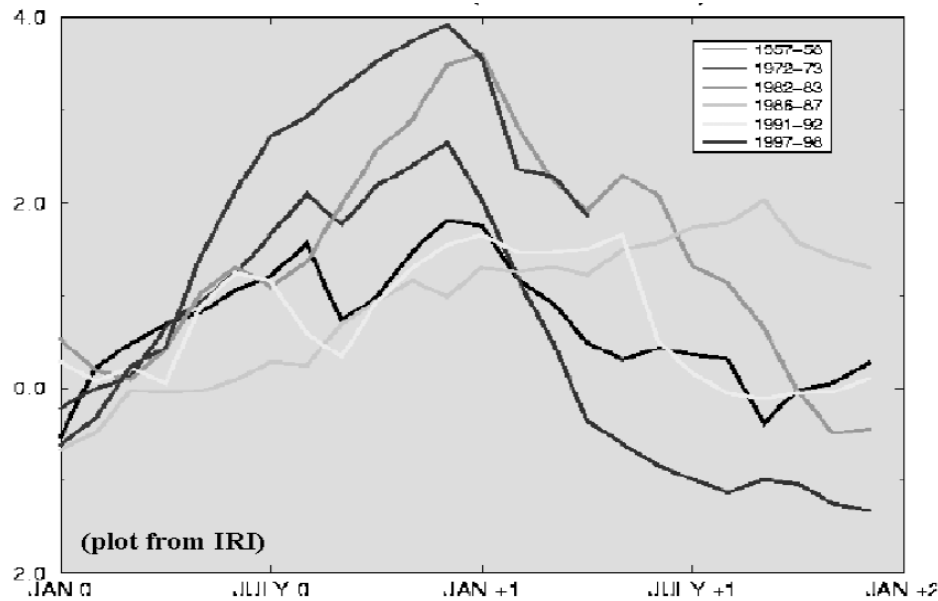


Fig. 5.1.4: Niño3 SST comparison during major El Niño years suggesting phase locking of ENSO with season

[Source: <http://www.ess.uci.edu/~yu/class/ess200a/lecture.7.climate.variations.pdf>]

was chosen which means the 'Christ Child'. The observations of Rasmusson and Carpenter (1982) suggest that ENSO events tend to onset, grow and decay at certain seasons of the year (Fig. 5.1.4).

ENSO, Thermocline and upper ocean heat content:

In Fig. 5.1.5, it is shown that the basin-wide equatorial upper ocean (0-300m) heat content is greatest prior to and during the early stages of a El Niño episode and vice versa during the La Niña. Moreover, the slope of the thermocline is greatest (least) during cold (warm) phases. Here the monthly thermocline slope index has been defined as the

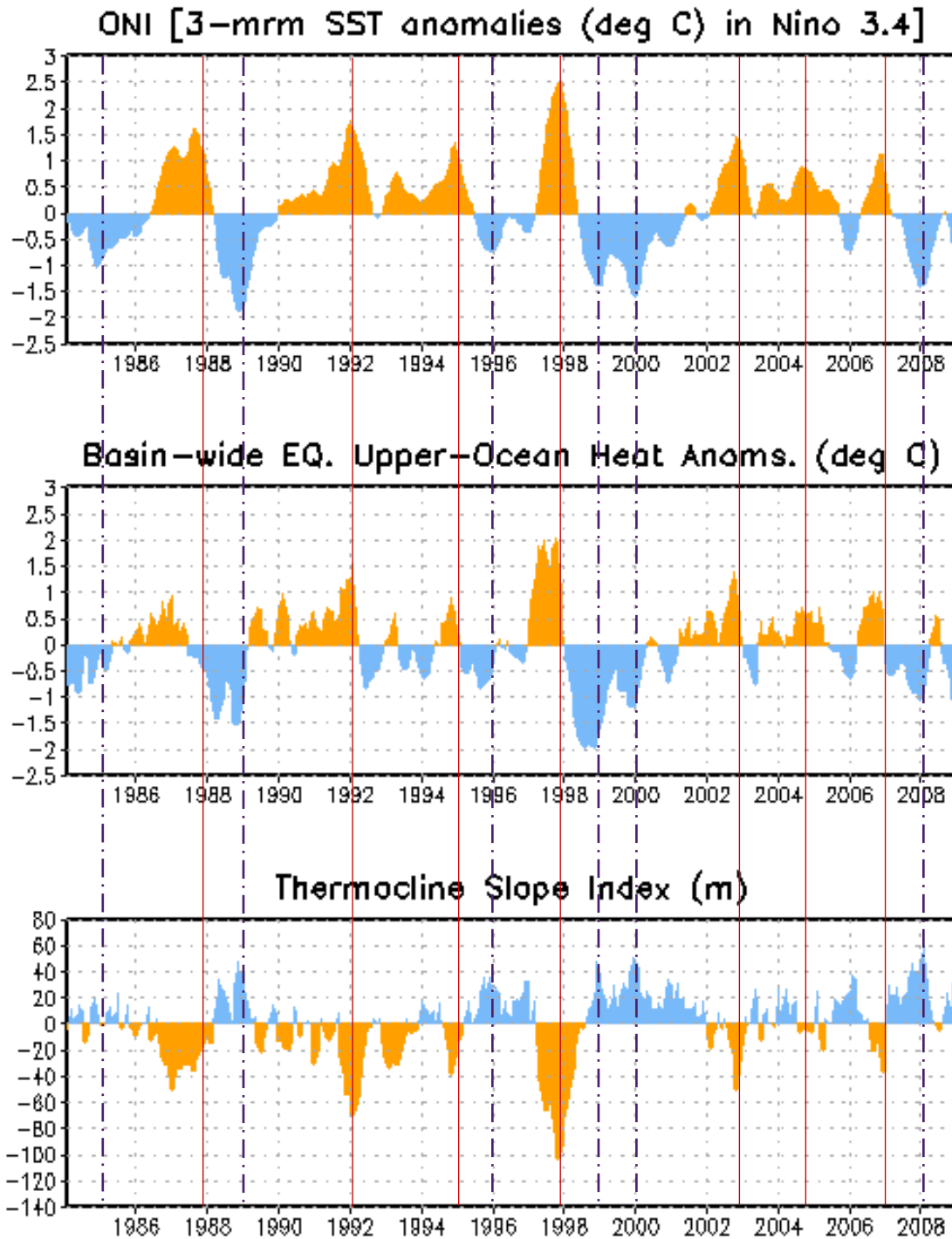


Fig.5.1.5: Upper ocean conditions in the Equatorial Pacific: a) Oceanic nino 3.4 Index; b) upper ocean heat anomaly and c) Thermocline slope index. Warm episodes (red solid lines) and cold episodes (purple dashed line) are marked.

[http://www.cpc.noaa.gov/products/analysis_monitoring/lanina/enso_evolution-status-fcsts-web.pdf]

difference in anomalous depth of the 20° isotherm between the eastern Pacific (90° to 140° W) to that of the western Pacific (160° E to 150° W). Hence the upper ocean heat content is sometimes considered as a precursor of the ENSO formation.

ENSO and shallow MOC in tropical Pacific:

The observational results of Fig. 5.1.6 clearly indicates that the strength of shallow MOC is anti-correlated with the east and central tropical Pacific SST (since transport scale is inverted). Such observations were also explored by Zhang et al. (2006), using 18 model simulations of the 20 th century climate by 14 state-of-the-art coupled climate models. Significant correlation exists between the meridional volume transport convergence and tropical SST in the majority of the models over the last half century. This suggests that, as well as the change in thermocline slope in the eastern Pacific and the upper ocean heat content around the tropical basins of Pacific, cold events of the ENSO are also associated with the strengthening of the shallow overturning circulation.

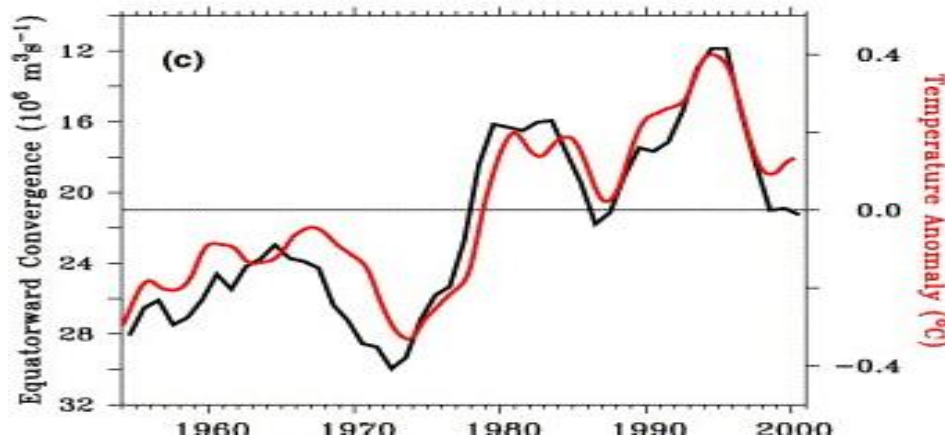


Fig. 5.1.6: Time series of volume transport convergence and SST averaged over the central and eastern tropical Pacific, 90° – 180° W and 9° S– 9° N. The transport scale is inverted to emphasize that a slow down of the circulation corresponds to a warming of SST.

[Zhang et. al. 2006]

ENSO and ambiguities:

Relating to the ENSO, there are certain questions that still need to be resolved and certain factors still need to be accounted for. *First and foremost:* though the theories of ENSO are capable of explaining some of the features of ENSO viz. the onset and evolution - the fundamental cause for its onset is still unclear. *Secondly:* sometimes an ENSO phase seems to be set in, but then terminates halfway (e.g., the weak warm event in 1980 which ended prematurely); on occasions, events which appear to be about to cease then recur (e.g., the weak El Niño in 1993 and 1994/95). Such behaviour can not be explained by existing theories. *Thirdly:* the ENSO cycle is usually analyzed on the basis of ocean-atmosphere interaction around the equatorial Pacific, ignoring the role of tropical-subtropical large-scale mass exchange, which is clearly related. *Finally:* most of the studies consider ENSO as an inter-annual phenomenon, but some recent studies (Zhang et al., 1997; Zhao et al., 2003; Chen et al. 2004, White et al. 2008) suggest that it also possesses an underlying quasi-decadal variability which makes an important contribution to variability around the tropics. In fact, most of the studies relating to ENSO are based on its inter-annual variability and ignore the inter-decadal one.

All these ambiguities imply that we are not yet in a position to completely understand the formation mechanism of the ENSO cycle. Furthermore, most of the current generation of climate models still cannot simulate realistic ENSOs (e.g., Latif et al., 2001; van Oldenborgh et al., 2005). Recently, the overall skill of ENSO prediction in retrospective forecasts made with ten different coupled GCMs was investigated by Jin et al. (2008). They analysed seasonal output from the APCC/CliPAS (**A**sian-**P**acific Economic Cooperation **C**limate **C**enter/**C**limate **P**rediction and its **A**pplication to **S**ociety) and DEMETER (**D**evelopment of a **E**uropean **M**ulti-model **E**nsemble system for seasonal to in**T**ER-annual prediction) projects during the common 22 years from 1980 to 2001 and found the overall prediction skill in need of improvement. Recent modelling studies by Wang et al. (2009) and Jin et al. (2009) have also confirmed this. Given the crucial role play by the Pacific in global climate, advancement of our knowledge relating to the ENSO is obviously important.

In this chapter, we address some of these issues. Our study suggests that the sun plays a role in atmosphere- ocean coupling. But, such coupling is found to be disturbed during the later half of the last century possibly related to climate change.

5.2. Climate change fingerprint

Different studies confirm that both natural and anthropogenic influences contribute to twentieth century climate change (Allen et al., 2006; Vecchi et al., 2006, 2007; Meehl et al., 2003, 2004). Though their regional impacts and relative roles are still controversial, it is well accepted that the influence of anthropogenic gases emerges as the predominant contributor to observed climate change during the later half of the 20th century (Vecchi et al., 2006; Meehl et al. 2003, 2004, 2006; Shindell et al., 2004).

- Climate change and General circulation:

The tropical Hadley circulation and Walker circulation are driven by different forces and may thus respond differently to climate change. For the Hadley circulation, meridional differential heating of the global radiative process is the essential driving force and hence it should exist even under a hypothetical aqua-planet without topography or sea-land contrast. Whereas, to generate/ alter the Walker Circulation, land-ocean distribution plays a paramount role. The Walker Circulation tends to weaken in climate models, forced with increasing greenhouse gases (Meehl et al., 2007; Tanaka et. al., 2004; Vecchi et al., 2006). The Hadley circulation also weakens but less than the Walker circulation (Vecchi and Soden, 2007). Key observations, relating to tropical circulations, are highlighted in the box below as they play a crucial role in our subsequent study.

- Global warming has caused a weakening of tropical circulations: more in the Walker cell than the Hadley cell
- Strong decrease in intensity of Walker circulation after 1950s
- Modest intensification since 1998

(Vecchi and Soden, 2007)

- Climate change and ENSO:

As ENSO is a manifestation of an alteration/ variation in the Walker cell, climate change in the Walker cell will readily be reflected in the ENSO. Moreover, thermocline slope and upper ocean heat anomaly, which are strongly related to the ENSO (as shown in Fig 5.1.5), will also be susceptible to climate change. Changing patterns of the ENSO also bear far reaching consequences in global scale teleconnection patterns. To mention one among several, ENSO in the tropical Pacific influences the interannual frequency of IODs (**I**ndian **O**cean **D**ipole) by modulating the latter's formation mechanism through the Walker circulation (Behara, 2006). The IOD is a coupled ocean-atmosphere phenomenon in the Indian Ocean and is normally characterized by anomalous cooling of SST in the south eastern equatorial Indian Ocean and anomalous warming of SST in the western equatorial Indian Ocean.

Van Oldenborgh et al. (2005) reported that the six most realistic models project no significant change in ENSO amplitudes in the future, and concluded that global warming exerts little influence on ENSO behaviour. Guilyardi (2006) noted a considerable spread of El Niño behaviour among a similar subset of realistic models, with no noticeable change in El Niño frequency and a slight indication of increased El Niño amplitude in a warmer climate. Merryfield (2006) also found a wide scatter of changes in ENSO properties among the IPCC model projections, with no clear consensus on amplitude and a small mean decrease in the ENSO period. However, the results of some numerical coupled ocean-atmosphere model studies suggest that the response of the tropical Pacific to greenhouse-gas forcing resembles a permanent El Niño-like condition (Knutsen et al., 1998; Meehl et al., 1995). More investigations are evidently needed to delineate the effects of global climate change on ENSO variability and its behaviour.

- Climate change and shallow overturning circulation:

Recent studies indicate that, during the 1950s to 1990s, there was a decrease in the shallow overturning circulation in the tropical Pacific (McPhaden and Zhang, 2002; also

shown in Fig. 5.1.6), with a rebound since 1998 (McPhaden and Zhang, 2004). Observational study suggests that convergence of cold interior ocean pycnocline water towards the equator increased to $24.1 \pm 1.8 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ during 1998–2003 from a low of $13.4 \pm 1.6 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ during 1992–98 (McPhaden and Zhang, 2004, see Fig. 5.2.1). There is a trend towards reduced transport convergence of 11 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$), over the period 1953–2001 (Zhang et al., 2006). According to them, the circulation fluctuates significantly on decadal time scales, with decade-to-decade variations of 8–12 Sv about the trend.

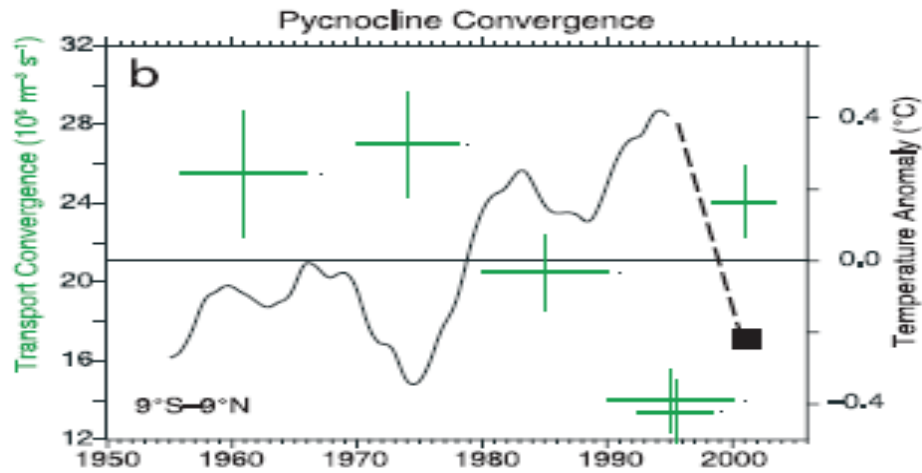


Fig. 5.2.1: Meridional volume transports in the pycnocline across 9°N and 9°S (with green) and equatorial sea surface temperature anomalies (black) for 1950–2003 across same latitude belt and 90°W–180°W, where equatorial upwelling is most prevalent. The temperature time series is smoothed twice with a 5-year running mean to filter out the seasonal cycle and year-to-year oscillations associated with ENSO. The single 5-year average for July 1998 to June 2003 is connected to the smoothed time series with a dashed line. Anomalies are relative to 1950–1999 averages.

[McPhaden and Zhang, 2004]

The timing of the observed decrease in circulation is consistent with the variations in the strength of the Walker circulation. As mentioned above, and supported by Fig. 5.1.6, the ENSO and the shallow overturning circulation are connected via an ocean pathway so that a climate change signal in both ENSO and the Walker circulation is also likely to be detected in the shallow MOC in the Pacific.

- Climate change and NAO:

Processes in the ocean, stratosphere and anthropogenic activity can all affect the phase and amplitude of the NAO. Model results relating to the NAO with imposed climate change factors are still lacking uniform consensus. E.g., utilizing the GISS Global Climate Middle Atmosphere Model, Rind et al. (2005) found that global warming causes a positive phase change in NAO; while Schnadt and Dameris (2003) showed that with increasing GHGs, the model NAO index decreases significantly from 1990 to 2015.

- Climate change and SAM:

The extra tropical climate of the southern hemisphere has changed a lot during the 2nd half of the 20th century. These changes include warming over mid-latitudes and the Antarctic peninsula while cooling over most of the rest of that continent. There has also been an increase in the speed of the westerlies over the Southern Ocean, a shift in mass

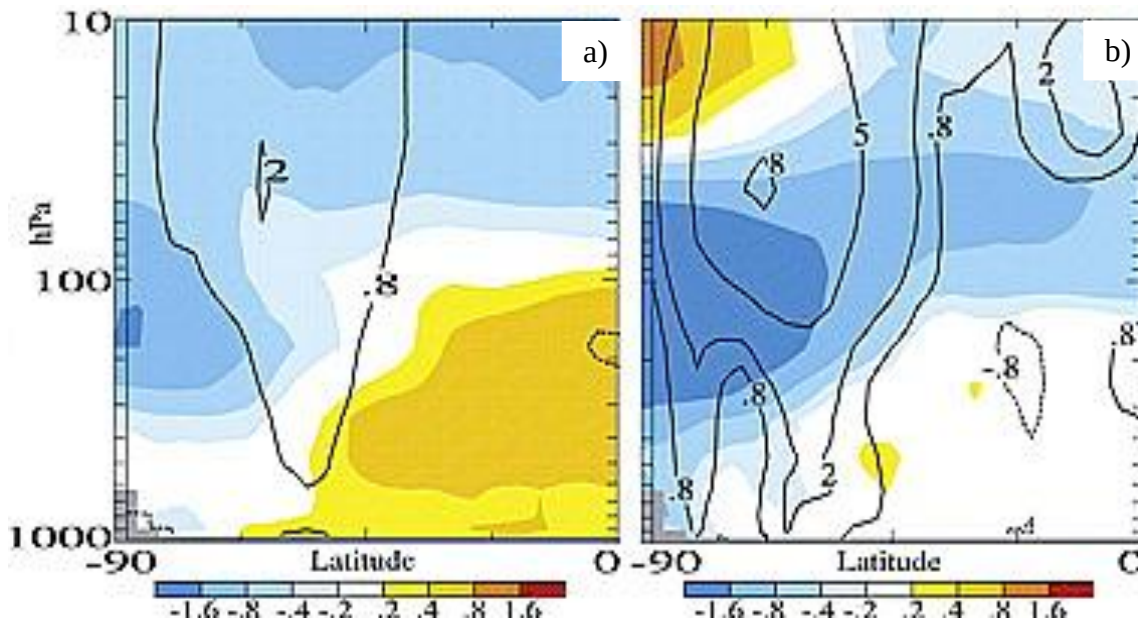


Fig. 5.2.2: Zonal mean Dec–May temperature (in °C, shown with colors) and zonal wind (m/s, shown with contours) linear trends from 1970 to 1999 in the simulations forced by increasing greenhouse gases (a) and by ozone depletion (b).

[Shindell et al., 2004]

from high to mid-latitudes in SLP and consistent changes in geopotential heights and winds extending up into the stratosphere (Thompson et al., 2000; Marshall, 2002; Schneider and Steig, 2002; Marshall, 2003; Jones and Widmann, 2003; Thompson and Solomon, 2002). The leading variability pattern, in all these meteorological fields in the southern extratropics, is the SAM.

Climate change during recent decades is mainly attributed to the combined anthropogenic effects of increasing greenhouse gases and decreasing stratospheric ozone. In the lower stratosphere, as shown by the modeling study of Shindell et al. (2004), the influence of ozone loss is clearly paramount, leading e.g., to a zonal wind anomaly nearly 3 times greater than that caused by greenhouse gases during the last three decades of 20th century (Fig. 5.2.2a and 5.2.2b). Overall, the influence of ozone becomes more important for higher latitudes and altitudes, though the extratropics-wide SLP signals of ozone and greenhouse gases are not statistically different. According to Shindell et al. (2004) the two factors have had comparable surface impacts during that period, though ozone dominates above the middle troposphere.

Projected impacts of the two factors on circulation over the next fifty years oppose one another, resulting in minimal trends. In contrast, their effects on surface climate reinforce one another, causing a departure from the SAM pattern and a turnabout in Antarctic temperatures, which rise more rapidly than elsewhere in the Southern Hemisphere. Observational analyses indicate that the Antarctic vortex has strengthened over the last decades of the 20th century (Waugh et al., 1999; Zhou et al., 2000), and the observed enhancement of the westerlies extends all the way to the surface (Thompson and Solomon, 2002) as in the GCM. Meehl (2005) using a global coupled model showed that ozone changes are the biggest contributor to the observed summertime intensification of the southern polar vortex in the second half of the 20th century, with increases of greenhouse gases being a necessary factor in the reproduction of the observed trends at the surface.

Haigh and Roscoe (2006), in their multiple regression analysis, with data from the later half of the 20th century, showed that a strong anti-correlation exists between SAM and

ENSO in the lower troposphere. Carvalho et al. (2005), using data analysis for the period 1979 to 2000, observed that during austral summer (DJF), cold events of ENSO are linked with dominant positive AAO and vice versa. The alternation of AAO phases were also shown to be allied with the latitudinal migration of upper level (200 hPa) STJ (around 45°S) and the intensity of the mid-latitude polar jet (around 60°S). Positive AAO phases are associated with a pole-ward shift and weakening of the subtropical feature accompanied by an intensification of the high-latitude feature.

Climate change, may act differently in future decades, by virtue of stratospheric ozone recovery, implemented through the Montreal Protocol. However, Shindell et al. (2004) from their future climate change scenario experiments suggest that the recovery of stratosphere ozone in the first half of the twenty-first century will not be sufficient to reverse the positive trend in the SAM index observed during the late twentieth century. Son et al. (2008) however, using the Chemistry-Climate Model Validation (CCMVal) models equipped with fully interactive stratospheric chemistry, predicted that owing to the ozone recovery, the southern hemispheric westerly jet in summer will be decelerated, on the pole-ward side, in contrast to the prediction of most IPCC/AR4 models. According to their study, ozone recovery will overtake greenhouse gas forcing in the 21st century and appropriate representation of ozone recovery is important to understand southern hemispheric climate variation. Due to its strong seasonality, ozone recovery via the southern hemispheric westerly jet may impact ENSO, which is again phase locked with the season (as shown in Fig. 5.1.4).

5.3. Data Sorting - Solar and ENSO

The solar signal, which is possibly modulated in the surface level mainly through the pathway of tropical circulations (Haigh et al., 2005; 2006; 2008), annular modes (Baldwin et al. (2001), Thompson et al., (2000)) and ocean (Meehl, 2004) is also likely to be altered due to climate change during recent decades. Hence, we seek to identify whether there is any influence of climate change (which is mainly attributed to the combined anthropogenic effects of increasing greenhouse gases and decreasing stratospheric ozone)

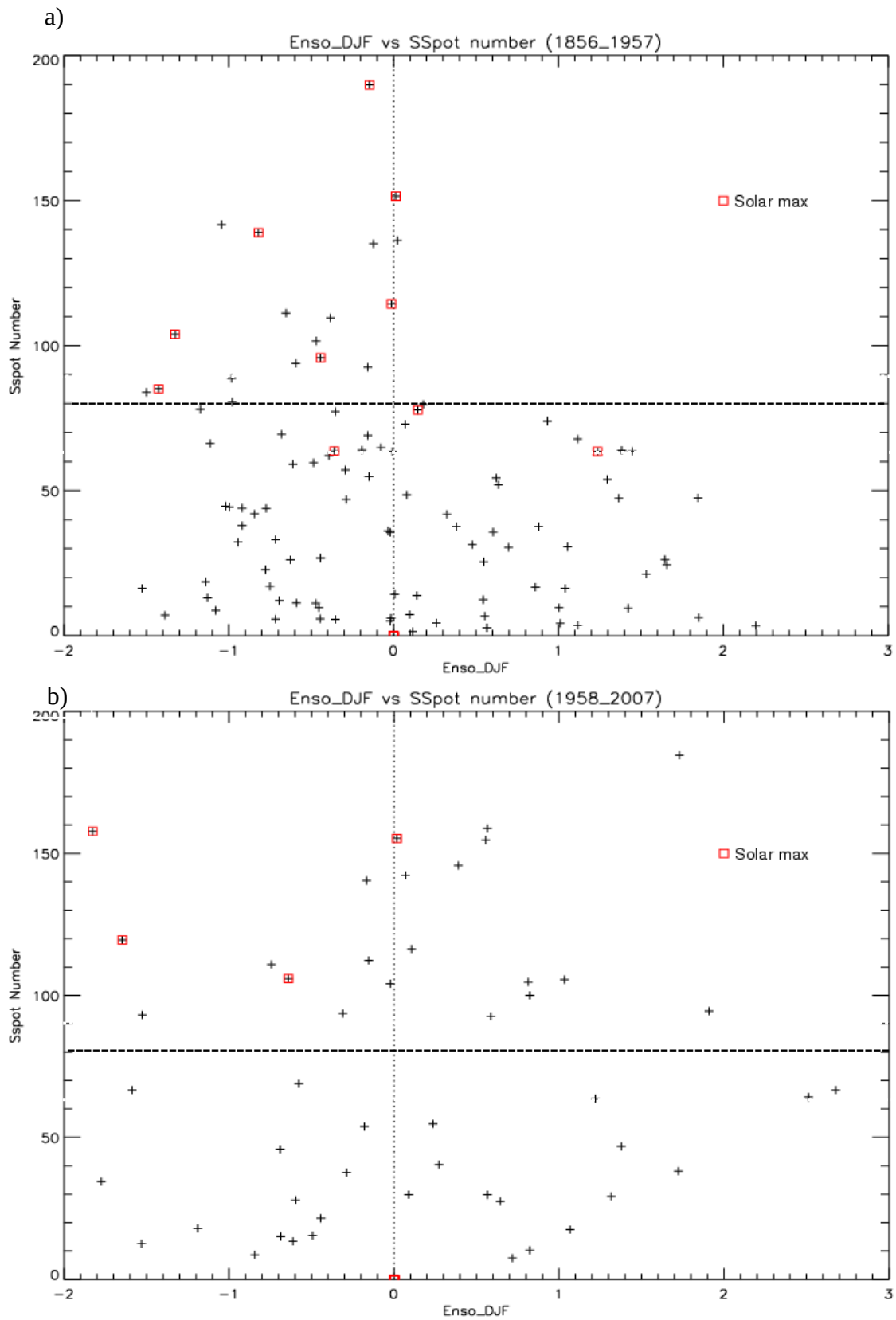


Fig.5.3.1: Scatter diagram of DJF mean ENSO index vs. annual mean SSN during:
a) (1856-1957); b) (1958-2007)

on the solar signal, for our whole 150 years of analysis.

Figure 5.3.1 shows scatter diagrams with the mean ENSO during DJF vs. annual average SSN, as was done before in Fig. 3.4b, except that we have segmented the whole period of analysis in two domains to capture signals, if any, relating to climate change (following Vecchi and Soden (2007), see textbox on page 134) : Fig. 5.3.1a is for the period 1856-1950s; whereas Fig. 5.3.1b is for dates after 1950s. To select the year at which the data would be split we opted for a trial and error method, which indicated that the solar and ENSO relationship during HS years underwent a change around 1958. Before that, all years with higher SSNs (say, above 80) are seen to be with –ve ENSO index (Fig. 5.3.1a); but overwhelmed by innate strong ENSO variability at LS activity. On the other hand, such a bias for HS years is missing since 1958 (Fig.5.3.1b). Such empirical observation, not only point towards some connection between the ENSO and Sun which seems to be present during HS years but also indicates that climate change during recent decades is more likely to disrupt such coherence. Kodera et al. (2007) also showed that the ENSO related signal is confined in the Pacific sector for HS years and is mainly due to the shift in location of the descending branch of anomalous Walker circulation.

Observational analysis suggests that there is a strong decrease in the strength of the shallow meridional overturning circulation (MOC) around tropical Pacific after 1950s with modest intensification since 1998; the same is true for the Walker circulation as well (McPhaden and Zhang, 2004; Vecchi, et al. 2007). Following them, we carried out a similar analysis like Fig. 5.3.1., segregating our whole period of analysis in two different domains: the first domain comprising of 1856-1957 and 1998-2007 (Fig. 5.3.2a); the second the remaining years (Fig. 5.3.2b) and end up with an interesting observation. Our new scatter plots show that the preference, towards the cold event side during HS years, not only persists during 1856-1957, but also during 1998-2007.

However, it is worth noting that the peak SSN years (marked by red squares) with higher SSN (say, above 80) are all biased towards a cold event situation in both the figures

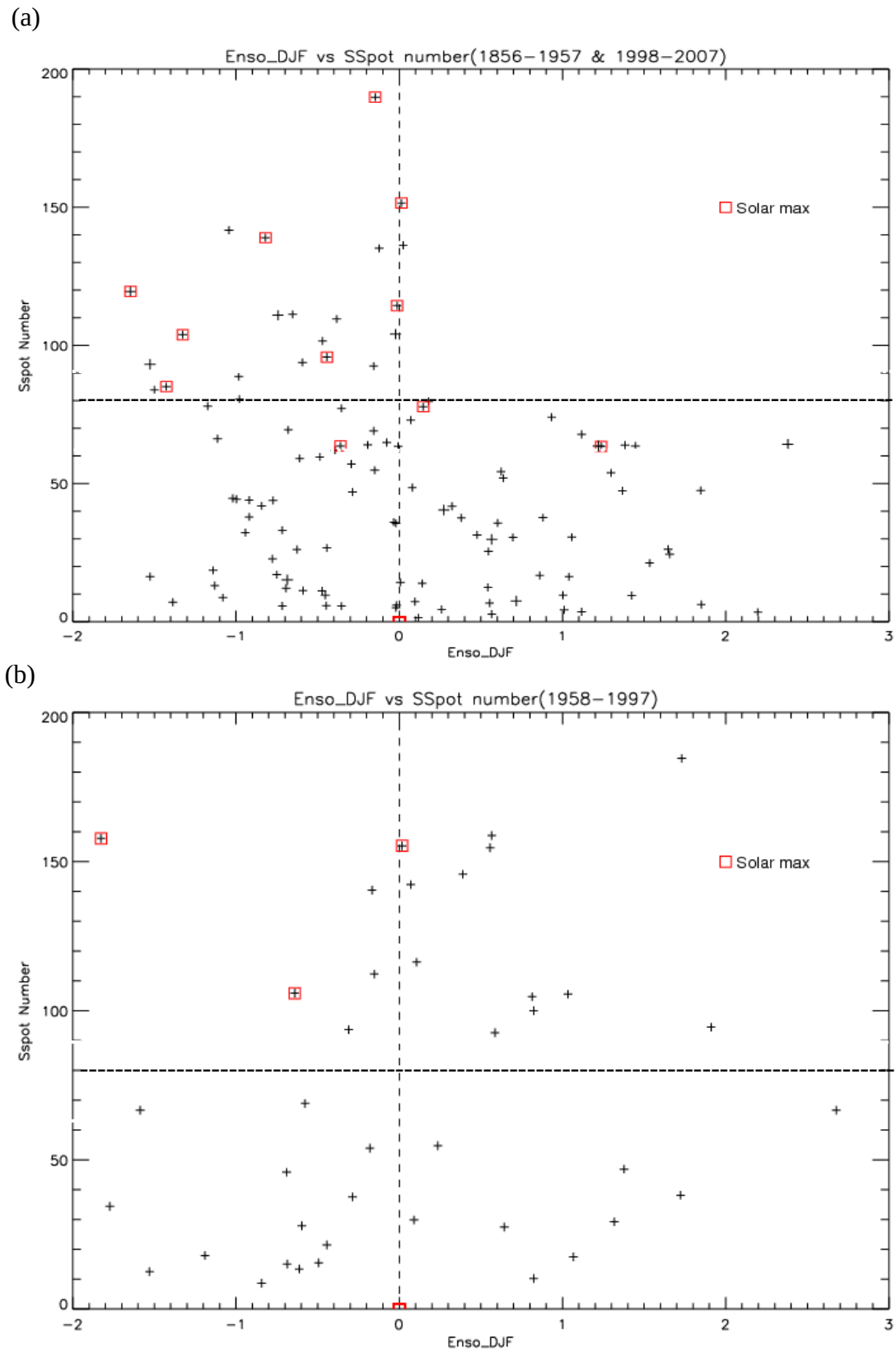


Fig.5.3.2. Scatter diagram of DJF mean ENSO index vs. annual mean SSN during:
a) (1856-1957) & (1998-2007); b) (1958-1997).

(Fig. 5.3.2 a & b). That clearly indicates, why using the method of solar max compositing (as used by vL07 and Meehl, 2007), the solar signal seems to be indifferent and unaffected with the inclusion or removal of a few solar cycles. However, in our subsequent study we will discuss why peak SSN years, during the later part of our analysis are still on the cold event side.

Our study, showing an apparent reversal of the solar and ENSO relationship since 1998, suggests that ozone in the stratosphere may control a certain proportion of the climate variation. This indicates that interactions of the effects of ozone in the stratosphere with other climate change factors in the troposphere may bear far reaching consequences. Such an observation, points to the importance of studying the relative roles played by GHG and ozone depleting substance, along with their individual regional impacts.

5.4. Results using Regression

Our empirical observation suggests that when the Sun is more active it influences SSTs before 1958 and after 1997. This signal is overwhelmed by the innate strong ENSO variability at LS activity (Fig. 5.3.2a). Other authors have suggested that tropical circulations and shallow MOC around the tropical Pacific were different during the intervening period (Vecchi and Soden, 2007; McPhaden and Zhang, 2004) and such observation may have implications relating to the Sun, ENSO, tropical circulation, shallow MOC and climate change which will be addressed here with the application of our regression technique.

Although, the bias of the ENSO (towards cold event side), during HS years, appears to revert back after 1998 to a behaviour similar to that exhibited before 1958, we have not joined these two periods together (1856-1957 and 1998-2007) in our analysis. Here, we have restricted our multiple regression analysis to two time periods: the earlier period comprises of 1856-1957 and the later that of 1958-1997. Such segregation of periods, enables us to capture the robust features (and the difference, if any) of the individual

influences during the later period, owing to the presence of combined anthropogenic effects (shared by GHG and ozone depleting substance), to that from the earlier period.

Regression of SLP (DJF): Solar (using SSN)

In Fig. 5.4, we show the amplitude of the components of variability of SLP, due to the sun (using monthly SSN), during DJF. A linear trend, OD and ENSO were used as the other independent parameters. The results, for the earlier period are shown in Fig. 5.4a and the later in Fig. 5.4b.

From these figures, it can be observed that: no solar signal is present around the ITCZ of the eastern Pacific, during the later period, which is significant during the earlier period; the solar signal around the AL has also weakened (from around 6 hPa to ~3 hPa) during the later period.

Focussing first on the earlier period (DJF) a solar signal is detected in SLP, around the eastern Pacific of tropics. The intensification of the ITCZ around the eastern Pacific, during HS years, may be related to the enhancement of the trade winds, which subsequently cause uplifting of the thermocline (in the eastern Pacific coast) and as a consequence may develop a situation similar to that of the cold event of ENSO. The situation reverses, (via reversal of the trade winds) during LS years, where the solar forcing may act to trigger a warm event like situation of ENSO. Thus a chain of mechanism similar to the ENSO (also described in Fig. 5.1.3) may be initiated via solar forcing on SLP (and consequently on the trade wind), around the ITCZ in eastern Pacific.

As mentioned above, a wind forcing mechanism around the equatorial Pacific is required to trigger ENSO cycles but the source being unclear. Our current study indicates that the Sun might influence the trigger. Such an observation is consistent with some recent studies (White et al., 2008; Zhang et al., 1997; Zhao et al., 2003; Chen et al. 2004), who detected an underlying quasi-decadal variability in the inter-annual ENSO. Moreover, our observations also suggest a decadal signal in the Walker circulation, (simultaneously,

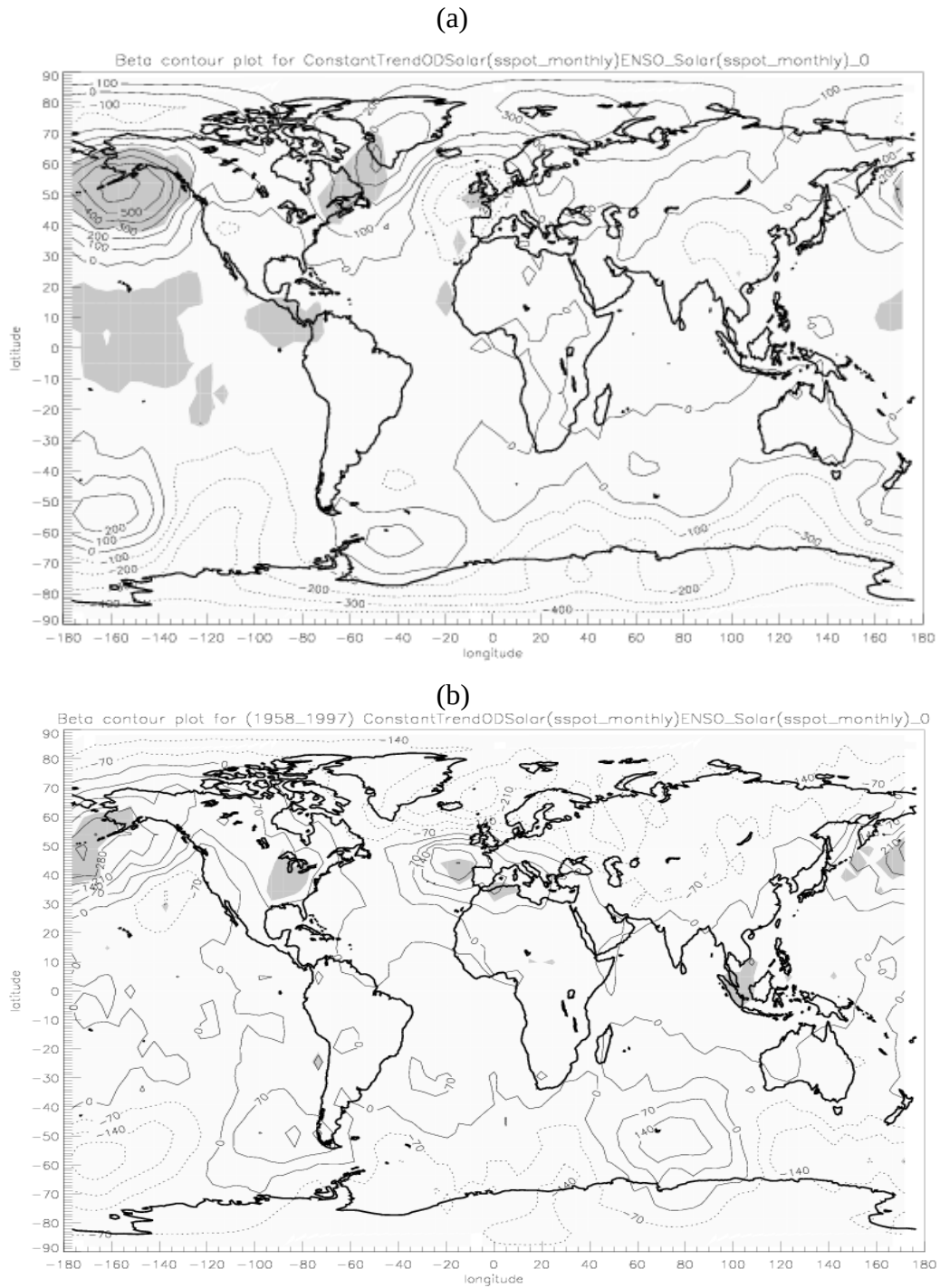


Fig.5.4: Amplitudes of the component of variability of SLP in hPa due to solar (using monthly SSN) during DJF for period: a) 1856-1957; b) 1958-1997. Other independent parameters used are: trend, OD and ENSO.

with the ENSO) related to the Sun, which some other authors (for eg. Vecchi and Soden, 2007) pointed as ‘the source still unclear’. Our scatter plot shown in Fig. 5.3.1a captures such a solar-ENSO behaviour, though only for HS years (say SSN >80). A possible explanation, supported by the work of Dima (2005) may shed some light on why that observation fails to indicate the reverse, during LS years.

Dima et al. (2005) identified one mode of climate, originated by solar variability, which dominates SLP and upper atmospheric levels and at the surface another mode, mainly dominated by atmosphere-ocean coupling, that explains about three times more variance than that due to the solar. Thus ENSO, a measure of the tropical Pacific SST, which is originated via atmosphere-ocean coupling around that region, possesses the potential strength to overpower the influence of solar mode, at the surface.

On the other hand, the solar signal, might make its presence felt differently during HS years and have the potential to overshadow the usual inter-annual ENSO characteristics. Hence our study during the earlier period, indicates that the solar signal during HS years may influence tropical SSTs, but is overwhelmed by the innate strong ENSO variability at LS activity (Fig. 5.3.1a). Nevertheless, the decadal signature in ENSO, generated via solar cannot be ignored. It is possible that this may also shed light on some of the unexplained behaviour of ENSO cycle (viz. premature cessation or prolonged lifetime, as mentioned above).

Now we focus our attention on mid-latitudes, where a strong solar signal in SLP, around the AL (~6 hPa) is detected. Such a signal can be related to a weakening of the Ferrel cell around the north Pacific during HS years, and vice versa during LS years. The mid-latitude (north) and tropics in the Pacific are tied through ocean pathways, and we find that the solar signal in the lower troposphere is also coupled with the ocean around those places, with the Sun being a potential driving factor to regulate the coupling mechanism.

Two fundamentally different routes for a solar influence on the troposphere have been proposed: one is the ‘top-down’ mechanism and the other the ‘bottom up’ one. First we address the ‘top-down’ solar influence which is generated through the stratosphere. In a GCM study (in the absence of a coupled ocean), Haigh et al. (2005) also detected a weakening of the Ferrel cell, associated with enhancement of solar forcing in the lower stratosphere. Such a change should be reflected in the north Pacific sub tropical gyre, which is wind driven. Weakening of the subtropical gyre will impede overturning (thus mixing with cold water) in the north-eastern part of Pacific, augmenting temperature around that place (also seen in Fig. 3.3b). As mentioned above the Pacific is connected between the tropics and mid-latitude (north) via the shallow MOC. Thus the solar signal around the north Pacific can be transported to the tropical Pacific and vice versa. Through such a linkage, the Sun may influence not only the trade winds, but also the tropics via mid-latitude.

In the ‘bottom-up’ pathway, the Sun can directly influence SST without stratospheric feedback. Here we suggest that this pathway is possibly involved with the shallow MOC and originated in the north Pacific (Fig. 5.1.1). The shallow MOC, during HS years, absorbs more heat in the region of north Pacific, and subsequently can cause a weakening of the AL - which in turn, can reduce the strength of Ferrel cell around the north Pacific. The heat absorbed can again be transported to the tropical Pacific via the shallow conveyor belt.

Thus our study suggests that the mid-latitude (north) and tropics in the Pacific are not only tied through ocean pathways, but our solar signal in the lower troposphere is also coupled with ocean around those places. The Sun might be a potential driving factor to regulate such a coupling mechanism and hence may also characterise the decadal fluctuations observed in some of the parameters mainly regulated by the ocean in the surface layer. For example, the study of Zhang et al. (2006), using historical hydrographic data during last 50 years, observed decadal variability in the shallow MOC. The strength of equatorward convergence of the pycnocline volume transport across 9°N and 9°S also characterizes decadal variability and the circulation fluctuates significantly on decadal

time scales, with maximum decade-to-decade variations of 7-11 Sv about the trend (Zhang et al., 2006).

During the later period (1958-1997), the solar signal around the ITCZ is no longer present, suggesting that the initiation of ENSO is governed by an alternate route. Weakening of the Ferrel cell during HS years subsequently causes warming around the north Pacific (alongside, oceanic shallow MOC around that place also absorbs more heat) and via the shallow ocean pathway may cause warming in the tropical Pacific. In Fig 5.1.5 we showed that the oceanic heat content around the tropical Pacific basin is correlated with the ENSO index- with more heat indicates the warm phase of ENSO. Such a route of the solar signal, via transporting heat from the mid-latitude to tropics, may activate warm ENSO in the tropics and will be addressed later. Fig. 5.3.2b also indicates a shift in ENSO, towards warm event side during the later period for HS years, in contrary to the observation of the earlier period. However, in that figure, solar max years are still biased towards the cold event side - this will be discussed below.

During the earlier period, the solar signal (during HS years) is a warming in the north Pacific alongside cooling in the tropics. Such a strengthening of the temperature gradient can enhance the rate of flow of the shallow ocean current from mid-latitudes (north) to the tropics in the Pacific and, consequently, can hasten equatorward convergence. However, the said signal, being missing around the tropics (and weakened in the mid-latitude), during the later period, indicates that the modified solar signal may have some link in decreasing shallow overturning current during the 1950 s to 1998, that has been reported by several authors (Zhang et al. 2006, McPhaden et al., 2002, Vecchi and Soden, 2007). The presence of a weakened solar signal in mid-latitudes during the said period, may be related to the decadal signature still present in the shallow MOC in the tropical Pacific.

Regression of SLP: Solar Indices

To capture the robust features (and the difference, if any) of the 11 year cyclic variability of the Sun in SLP, we applied the regression technique, using different solar indices viz.

SSN (Fig. 5.4.1), Solanki (Fig. 5.4.2) and Foster's TSI (Fig. 5.4.3), during the two different periods. Other independent parameters we used were linear trend and stratospheric optical depth. The regression was done using all months of the year with the annual cycle removed from each time series.

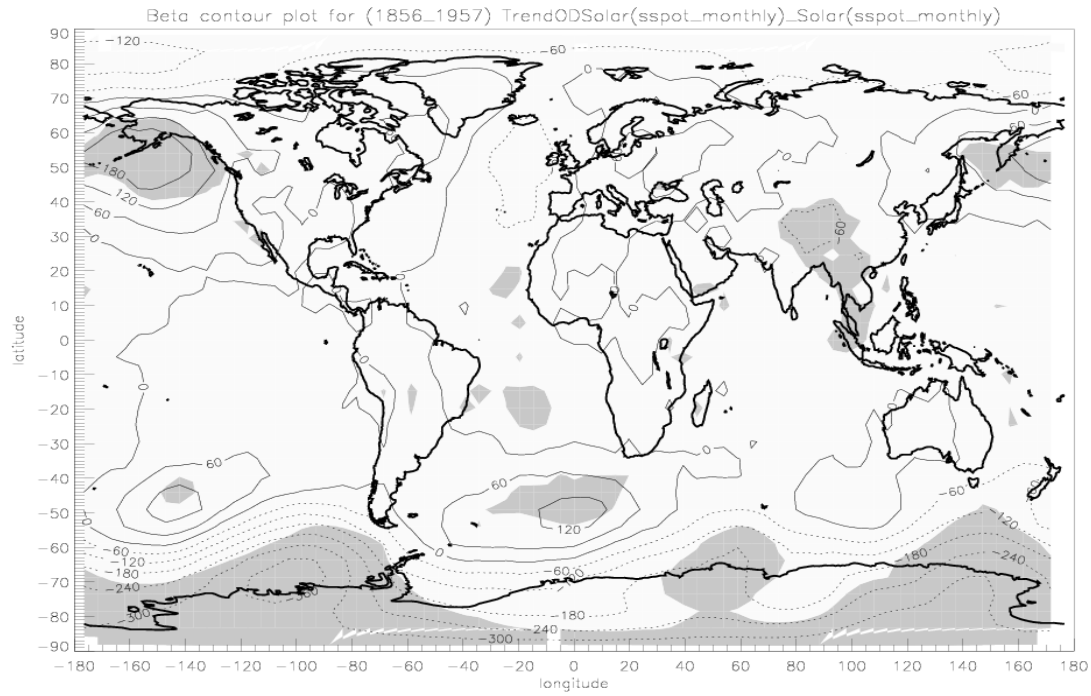
During the earlier period (Fig. 5.4.(1-3)a), positive NAM and SAM features are distinctive in all these pictures which reveals a clear positive solar signal in both the polar annular modes before 1950s regardless of the TSI reconstruction. However, during the later period (Fig. 5.4.(1-3)b), the solar signal does not conform with the signal observed during the earlier one (Fig 5.4.(1-3)a), using any of the solar indices. During this period, the solar signal fails to capture any clear signal in the polar annular modes expressed in SLP. That might be the reason why Haigh and Roscoe (2006), using data of the later period, are unable to perceive significant solar signal in the NAM and SAM.

Recent research suggests that both the annular modes are sensitive to a wide array of forcing mechanisms, including increasing greenhouse gases (Shindell et al., 1999; 2001a; Fyfe et al., 1999; Kushner et al., 2001), feedbacks between greenhouse gases and ozone depletion (Hartmann et al., 2000), increase in tropical sea-surface temperatures (Hoerling et al., 2001), and variations in solar forcing (Shindell et al., 2001b).

Our study, combining most of the potential forcing parameters, suggests one possible pathway of how the polar modes might be affected by climate change. As the ENSO is affected by climate change during the later half of the 20th century, the STJ and mid-latitude jets are also likely to be influenced (Carvalho et al., 2005) and subsequently, through the jets, the SAM and NAM. More recently, Ineson and Scaife (2009), Bell et al. (2009) using a general circulation model of the atmosphere showed that there is a clear response of the ENSO in European climate via the stratosphere. This mechanism is restricted to years when SSW occurs, leading to a transition to cold conditions in northern Europe and mild conditions in southern Europe in late winter during El Niño years. This indicates that the climate change signal in the ENSO may also be captured in the polar stratosphere through modulating the strength of the Brewer-Dobson circulation. Thus the

Regression of SLP: Solar (using SSN)

(a)



(b)

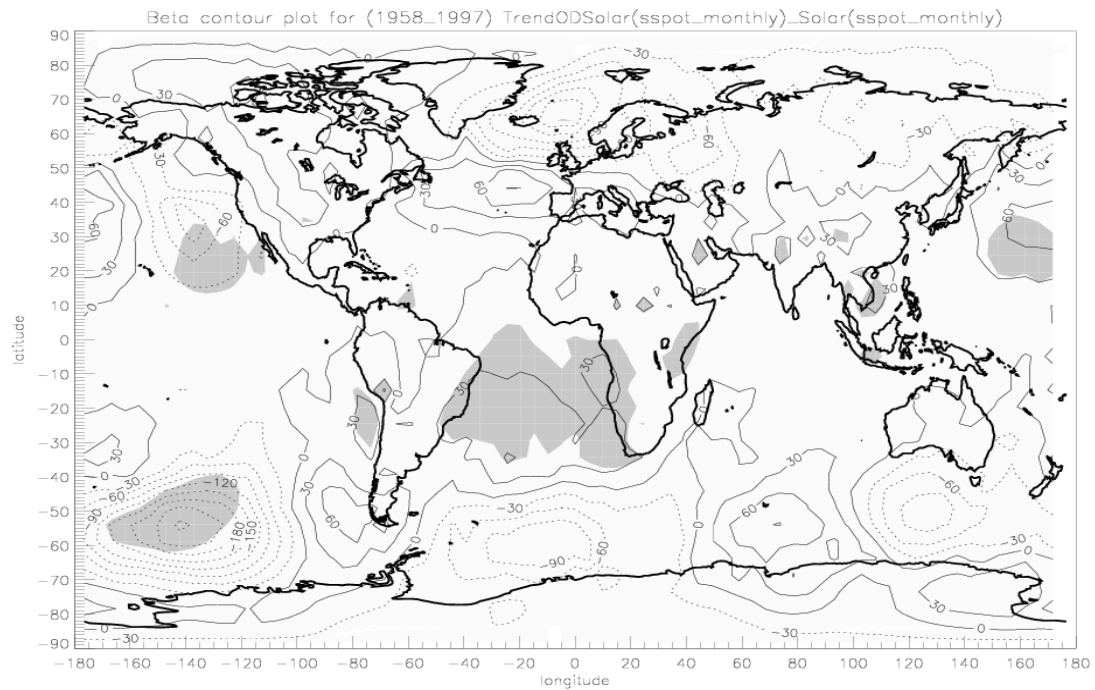


Fig.5.4.1: Amplitudes of the component of variability of SLP due to solar (using monthly SSN) for period: a) 1856-1957; b) 1958-1997. Other independent parameters used are trend and OD.

Regression of SLP: Solar (using TSI Solanki)

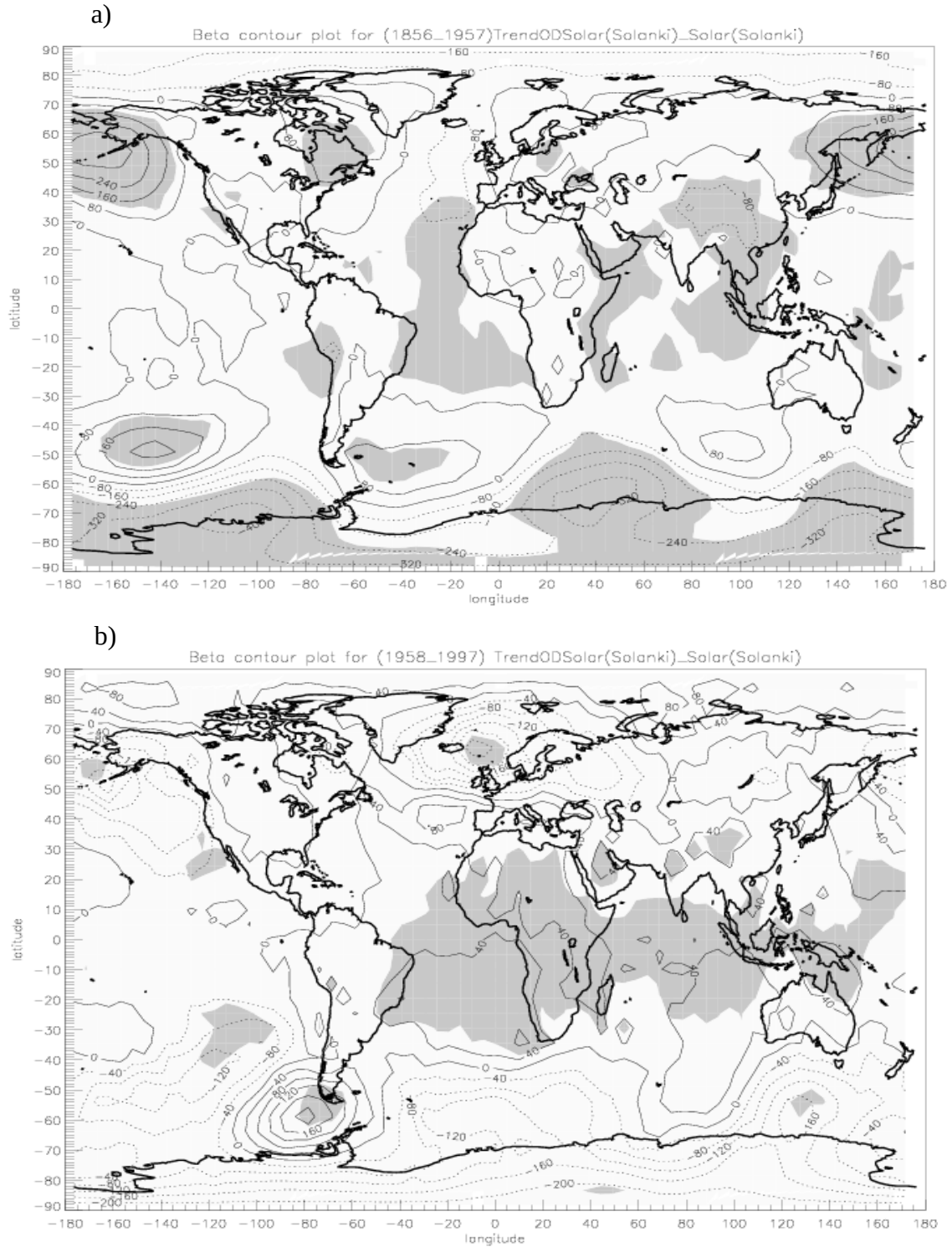


Fig.5.4.2: Amplitudes of the component of variability of SLP due to solar (using TSI of Solanki) for period: a) 1856-1957; b) 1958-1997. Other independent parameters used are trend and OD.

Regression of SLP : Solar (using TSI Foster)

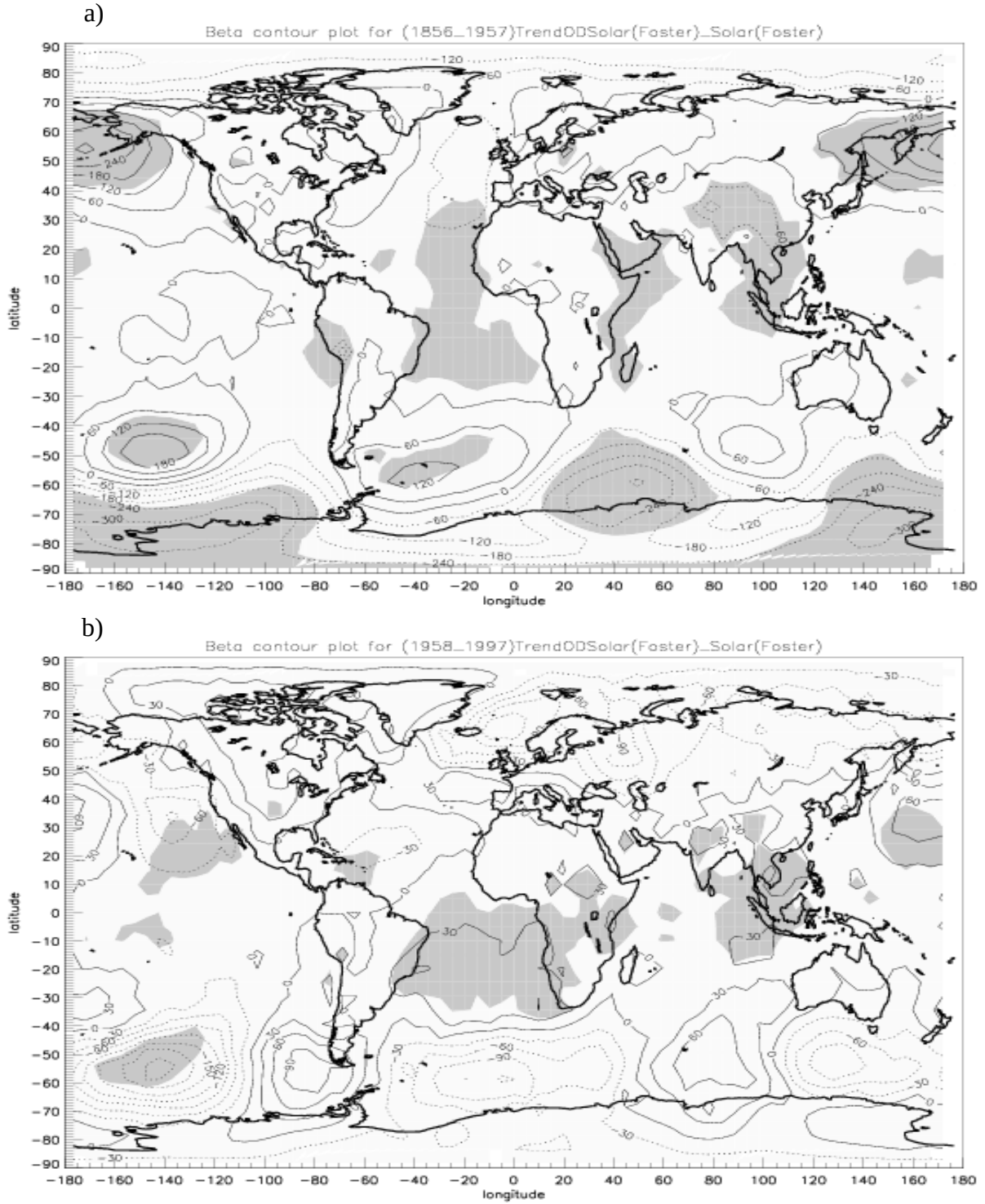


Fig.5.4.3: Amplitudes of the component of variability of SLP due to solar (using TSI of Foster) for period: a) 1856-1957; b) 1958-1997. Other independent parameters used are trend and OD.

clear positive solar signal in the NAM and SAM expressed in SLP, during the earlier period, might be different during the later period.

The effect of solar activity, during the later period, (irrespective of the choice of solar indices), resembles a positive signal of the NAO (Fig. 5.4.(1-3)b). This is in agreement with Haigh and Roscoe (2006), who also detected a positive solar fingerprint in the NAO, using data from the later period.

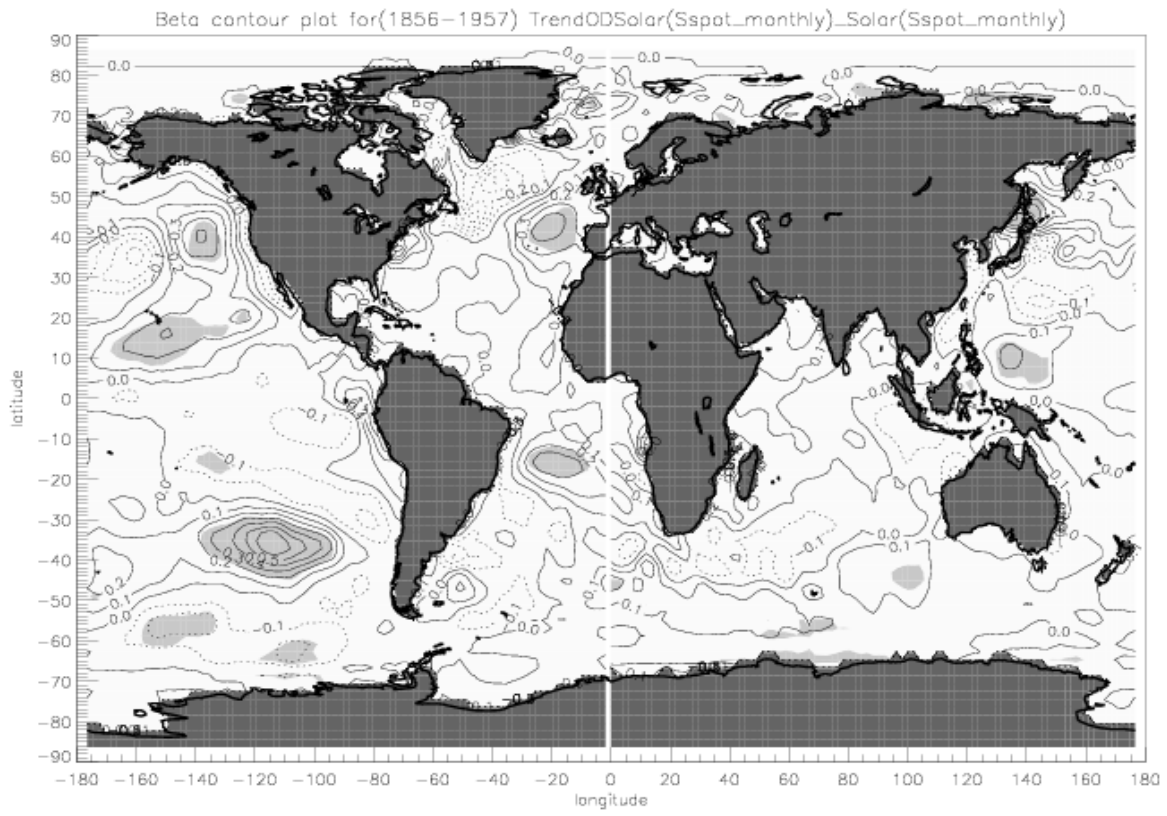
Regression of SST due to solar (using SSN): (1856-1957)

Now we focus our attention on SST, using the data from NOAA. Fig. 5.5 shows amplitudes of the solar component of variability during the earlier period, i.e., from 1856 to 1957. Here we have used monthly SSN for solar cycle variability and the regression was carried annually removing the annual cycle. In Fig. 5.5a., other independent parameters used are trend and OD; those are trend, OD and ENSO in Fig. 5.5b; and are trend, OD and PDO in Fig. 5.5c.

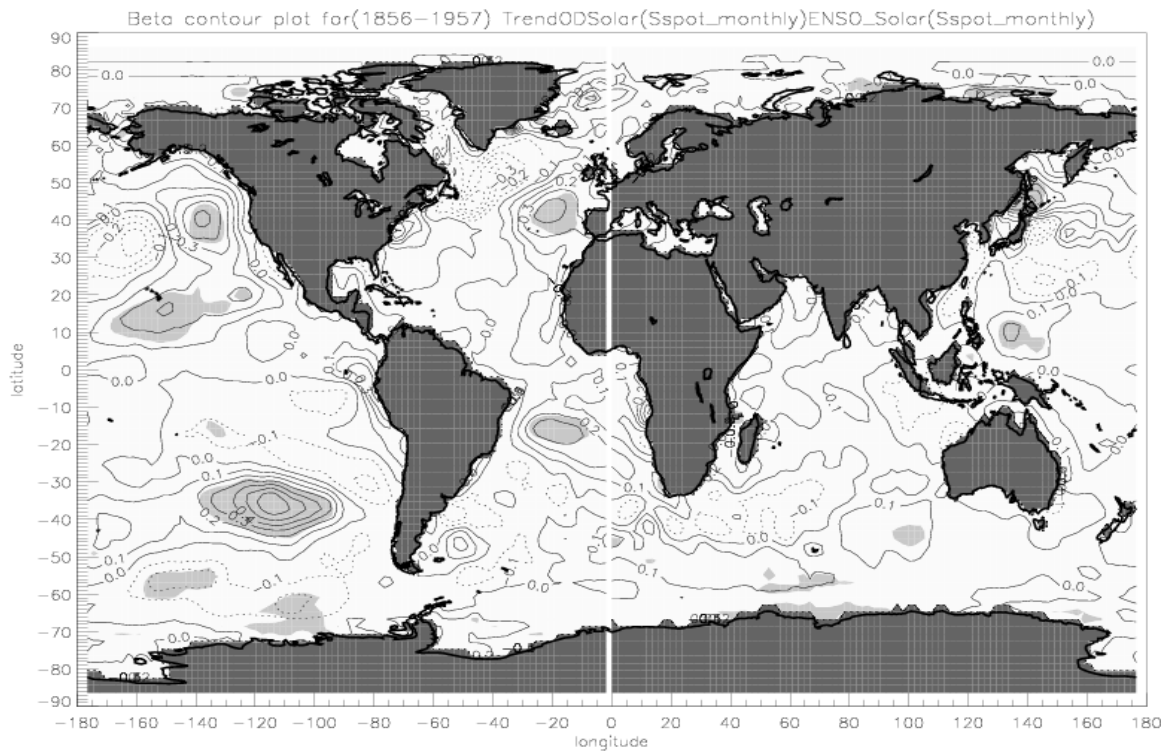
In Fig. 5.5a, we observe that the solar signal (positive) is associated with weak (not statistically significant) cooling in the tropics during the earlier period. Moreover, from Fig.5.5 a & b, it is clear that the ENSO signal in the tropics is slightly mixed with that of the solar. Strong warming around the mid-latitude of Pacific is noticeable. Such observation in Pacific agrees with Haigh (2003), Crooks and Gray (2004) and van Loon et al. (2004) in terms of the possible pathway of solar influence in the lower troposphere via the Ferrel cell.

Our results, during the earlier period, shown in Fig. 5.3.1a capture a cold bias of ENSO during HS years (say SSN >80), though overwhelmed during LS years. The work of Dima et al. (2005), as mentioned above, provided an explanation for such an observation. Here we discuss why the result of regression fails to capture any detectable solar signal in SST around the tropical eastern Pacific, without or with the ENSO as an independent parameter (Fig. 5.5a and b respectively). As very few data points show a bias towards a

(a)



(b)



(c)

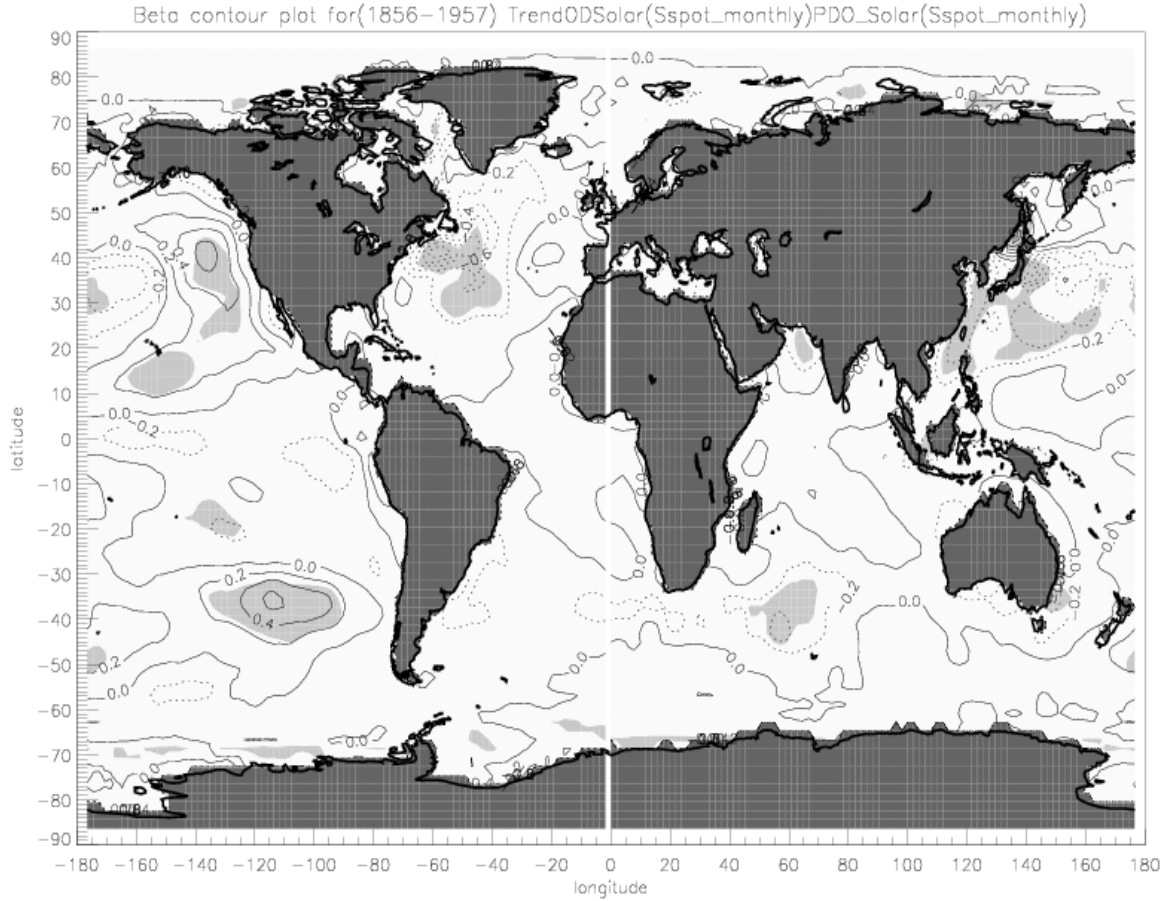
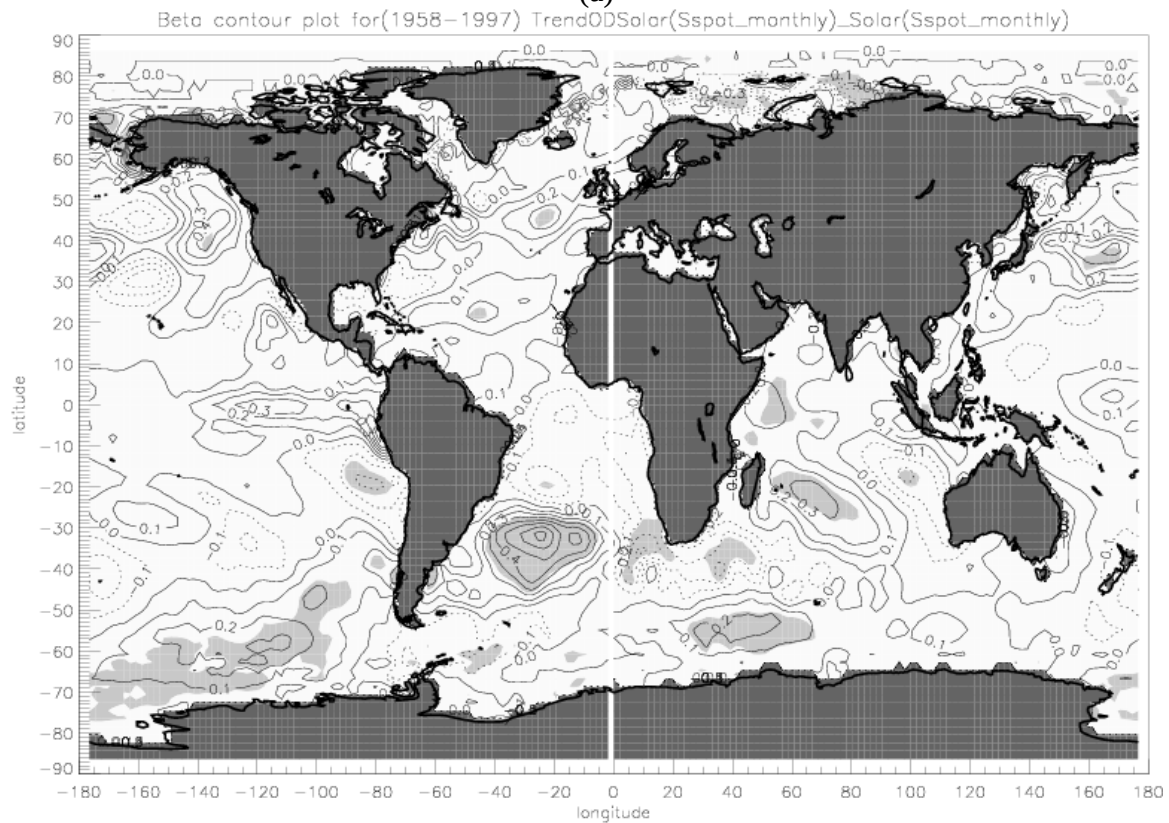


Fig.5.5: Amplitudes of the solar (using monthly SSN) components of variability of SST during (1856-1957). Other independent parameters used are: a) trend and OD; b) trend, OD and ENSO; c) trend, OD and PDO. Regression was done annually removing annual cycle.

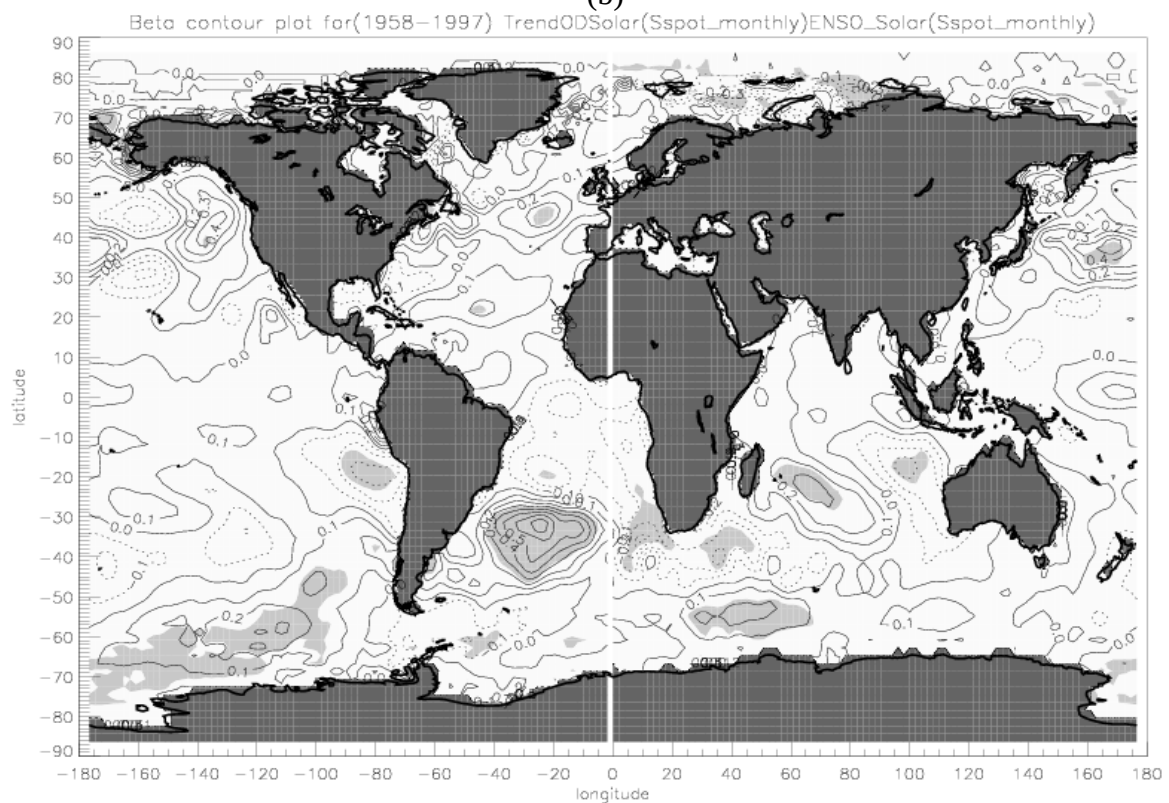
cold event side of the ENSO during HS years, compared to the whole set of observation (shown in Fig. 5.3.1a), the regression fails to indicate any strong result for the Sun and SST in the tropics (Fig. 5.5a). Furthermore, when we include the ENSO as an index in the regression, the results still fail to capture any detectable solar signal (Fig. 5.5b), suggesting again that it is the ENSO which is dominating SST in the eastern tropical Pacific most of the time. Thus, the apparent influence of the Sun on ENSO during the earlier period, may be initiated through triggering the trade wind (as described in details before), but is not detectable by the regression analysis in tropical SSTs.

Regression of SST due to solar (using SSN): (1958-1997)

(a)



(b)



(c)

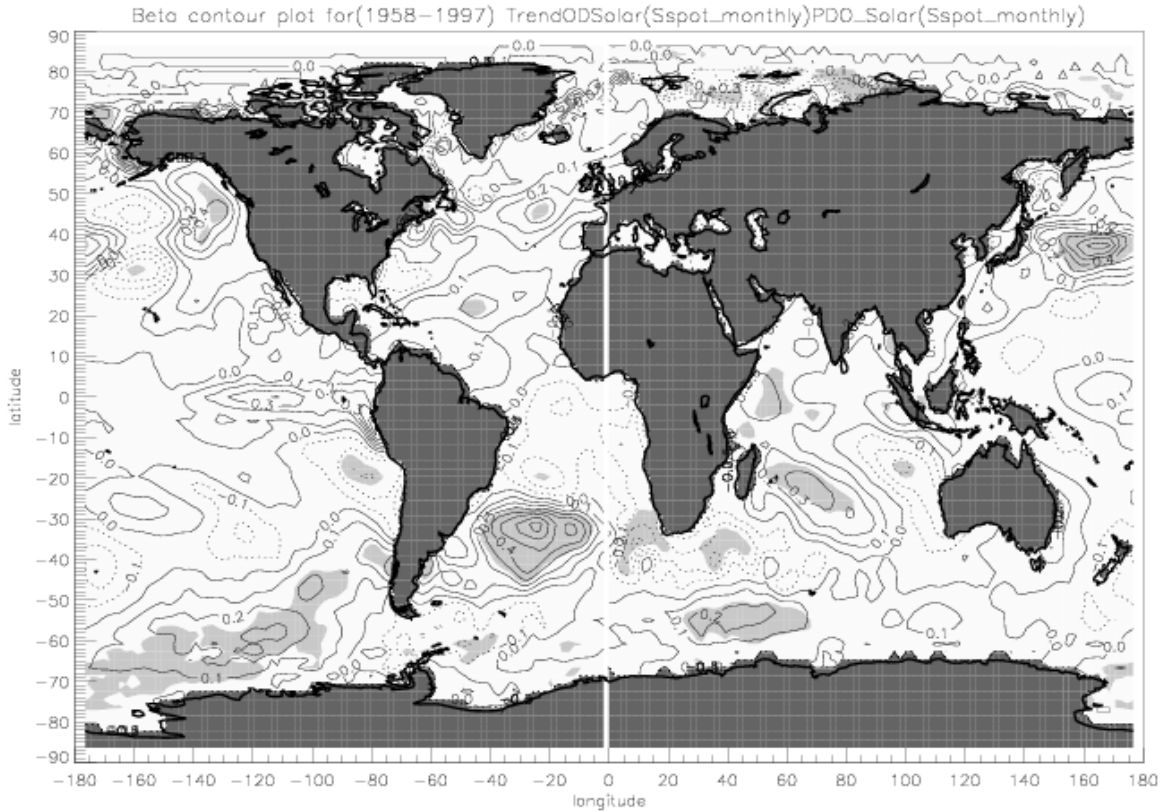


Fig.5.6.(a-c): Same as Fig 5.5.(a-c) respectively; but for the period 1958-1997.

Similar analysis was carried for the later period (during 1958-1997) using the NOAA SST dataset. Fig. 5.6.(a-c) show amplitudes of the solar (using monthly SSN) component of variability of SST. In Fig. 5.6a, the other independent parameters used are trend and OD; those are trend, OD and ENSO in Fig. 5.6b; and are trend, OD and PDO in Fig. 5.6c.

Warming in the tropics is observed around the eastern Pacific during the later period (Fig. 5.6a), unlike the earlier one (Fig. 5.5a). It is also seen that the solar signal during the later period is mixed up with the ENSO in the tropics (Fig. 5.6a and b) and with the PDO in the mid-latitude (Fig. 5.6a and c). If the ENSO signal is not excluded from the solar (Fig. 5.6a), then our regression results for the later half of the 20th century suggest that the solar signal resembles that of the warm event of ENSO; which is not the case during the earlier period. Thus during 1958-1997, omission of ENSO from the regression gives a

false warming (cooling) signal in SST related to higher (lower) solar activity (as observed by White et al. (1997)).

White et al. (1997), using EOF analysis of SST, during the 2nd half of the last century, detected a leading mode of variability, whose spatial pattern consists of a warming around the eastern Pacific in the tropics accompanied by cooling centred around the north Pacific (30°N and 160°W) with a ~11 year period lagging the solar cycle by ~1 year (see Fig. 1.2.3). Such a spatial pattern, with a reminiscence of the warm event of ENSO, contradicts that of vL07 who detect a solar signal analogous to the cold event of ENSO. Moreover, White and Liu (2008) shows a phase-locking of harmonics of the ENSO time series with the solar cycle resulting in a warm event (WE)- like signal for about 3 years around the peak of the DSO (with cold events (CE) approximately 2 years either side of the peak, and stronger WEs peaking 3-4 years before and after it (also shown in Fig. 1.2.11).

Our multiple regression study of the 1958-'97 data is able to explain some of the inconsistencies relating to the solar signal around the tropical Pacific; for instance, the contradictory results of vL07 and White et al. (1997), in addition to White and Liu (2008). Our analysis of regression during the time period, common to all the studies, accompanied by our observation that the years of peak annual SSN, used by vL07, generally falls a year or more in advance of the maximum of the smoothed DSO, provides coherence to these apparently conflicting findings.

First, we identify that the methodology adopted to detect the solar signal by vL07 is not characterising the true behaviour of a solar QDO. In fact, peak years used by vL07 generally falls a year or more in advance of the broader maximum of the 11-year solar cycle and thus reflects a different (rising) phase of the solar cycle to the peak year of smoothed DSO (shown in Fig. 3.5). As the ENSO usually takes 1-2 years to change from its cold to warm phase this explains why the results of vL07 differ to those of White et al. (1997). The regression during 1958-1997, (with omission of the ENSO) indicates why both White et al. (1997) and White and Liu (2008) not only detect warming during HS

solar years, but also *cooling during LS years* (Fig. 5.6a) during the time period, common to all the studies. The trough and peak of tropical ENSO shows a period of about 2 years due to its usual phase transition mechanism, *throughout the solar cycle* as observed by White and Liu (2008) (Fig. 1.2.11); hence consistency between vL07, White et al. (1997) and White and Liu (2008) during the time period, common to all the studies. Such observation also explains why solar max years (as used by vL07), are still on the cold event side during 1958-1997 (shown in Fig. 5.3.2b); which is not any special features of the rising phase of solar cycles – it is simply an artefact of ENSO.

Moreover, our study indicates that the composite of 9 solar cycles (during 1900-2000) as considered by White and Liu (2008) in their analysis, generally captures the behaviour of 1958-1997, the period having stronger solar cycles and strongly affected by the ENSO. It also suggests that the application of their approach to the period prior to 1958 would be unable to detect similar phase locking of the ENSO and solar.

Lean and Rind (2008) using a multivariate analysis, distinguished the impact between natural and anthropogenic forcing on observed near surface air temperature, regionally as well as globally. The zonal response patterns observed by them due to the sun suggests, mid-latitude warming in both the hemispheres are evident and not sensitive to the choice of time period. Covering the period of 1889-2006 (which excludes more than 30 years of data from our earlier period) they also identified the geographical response of surface temperature due to the solar: in the SH, warming is around the mid-latitude in Atlantic; in the NH (over ocean) warming is seen around the north Pacific.

Agreeing with Lean and Rind (2008), our analysis also detects warming around 40° , during both the periods. It is also consistent with Haigh et al. (2004) who detect mid-latitude warming during solar max years, using two different sets of data (SSU/MSU satellite data and ERA reanalysis data) for the period 1979-2001.

The latitudinal distribution of our solar signal is similar for our two time domains, but a shift in longitudinal distribution has been identified. Warming around 30° - 50° S in the eastern Pacific (seen in the earlier period) is no longer present during the later period.

Instead, a pronounced warming (even $\sim 0.6\text{K}$ and significant) is observed around the same latitude belt in the Atlantic. Our results from the later period, relating to the solar influence on SST is consistent to that observed by Lean and Rind (2008) in relation to surface temperature. Whether such observations also have some relevance to the recent turnabout in Antarctic temperatures needs to be addressed.

However, mixing of ENSO and PDO with the solar signal mainly during the later period might modulate the true solar signal (which should be insensitive to the choice of methodology) and needs to be accounted appropriately. This might shed light on some of the discrepancies relating to results of the solar influence on surface temperature.

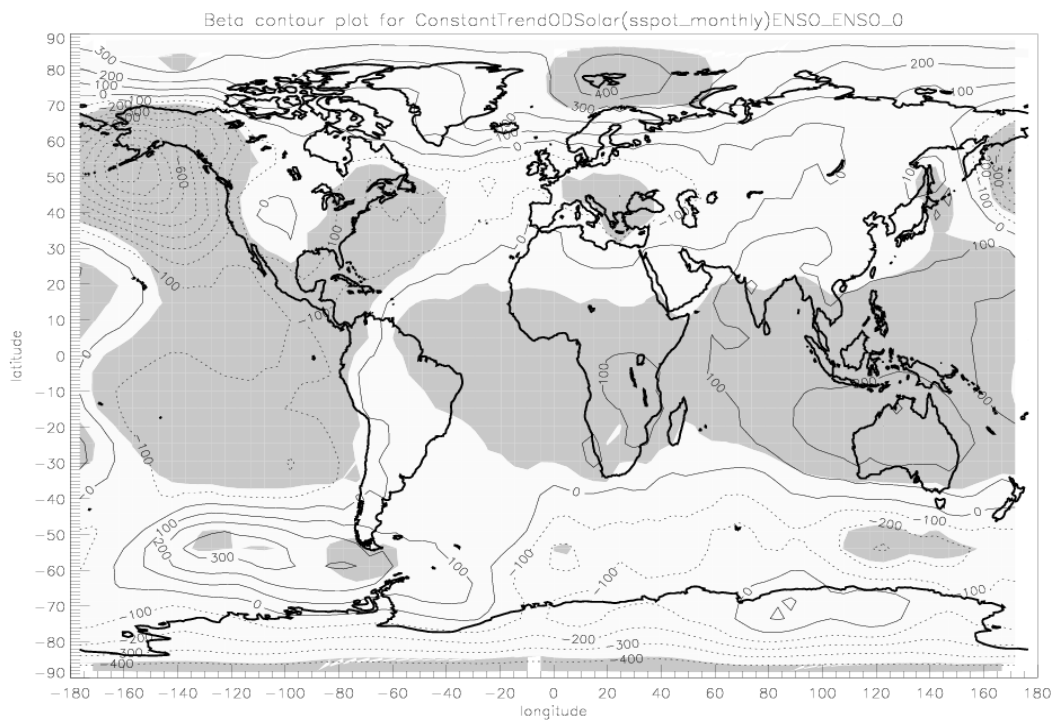
Previous discussions indicated that climate change probably can modulate the ENSO via the Walker circulation. Our results of multiple regression analysis, suggests that the nature of both the ENSO and PDO has changed during the later period as presented below.

Regression of SLP (DJF): ENSO

Amplitudes of the component of variability of SLP due to the ENSO during DJF have been shown for two periods, 1856-1957 (Fig. 5.7a) and 1958-1997 (Fig. 5.7b). Here other independent parameters used are trend, OD and solar (using monthly SSN). The signal of ENSO, during the later period, is associated with a strong NAO pattern (Fig. 5.7b) which is not the case during the earlier period. Such observation is consistent with the study of Toniazzo and Scaife (2006), who also detected a signature of ENSO in the NAO. According to them positive ENSO features are associated with the negative phase of NAO. Moreover, during the later period, the magnitude of the influence of ENSO on SLP, in the two ends of the SO has also increased.

Thompson and Solomon (2002) noticed that the surface temperatures, during last few decades of the 20th century, have decreased over most of the Antarctic continent, with the exception of warming over the Antarctic Peninsula. Our current analysis suggests that there is a strong influence of the ENSO on SLP around the Antarctic Peninsula, (eg., a

(a)



(b)

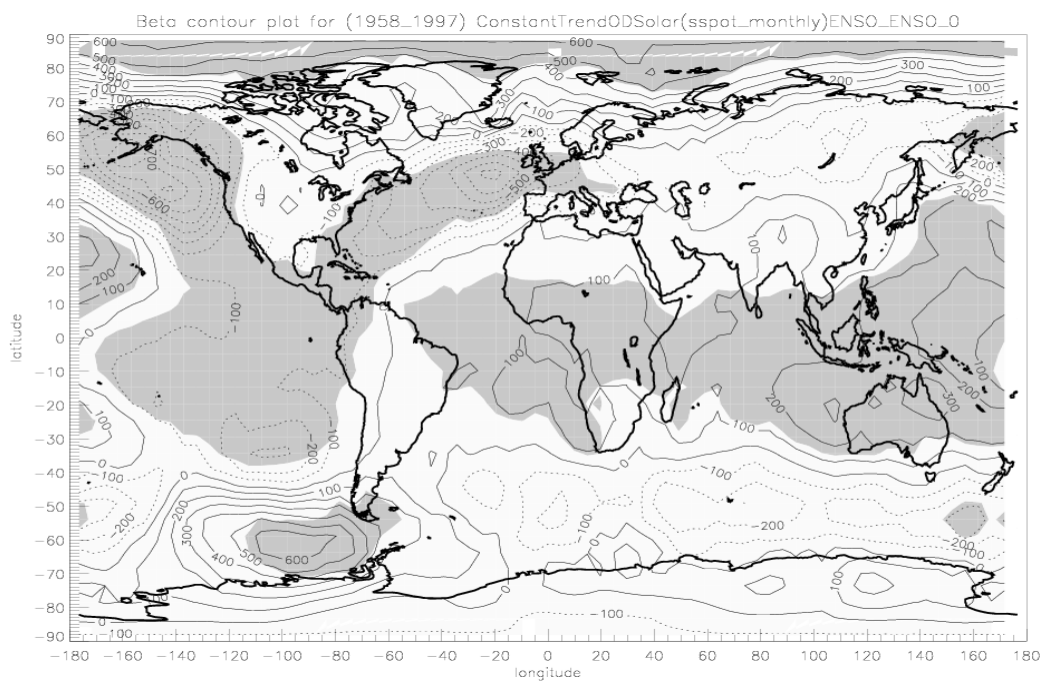
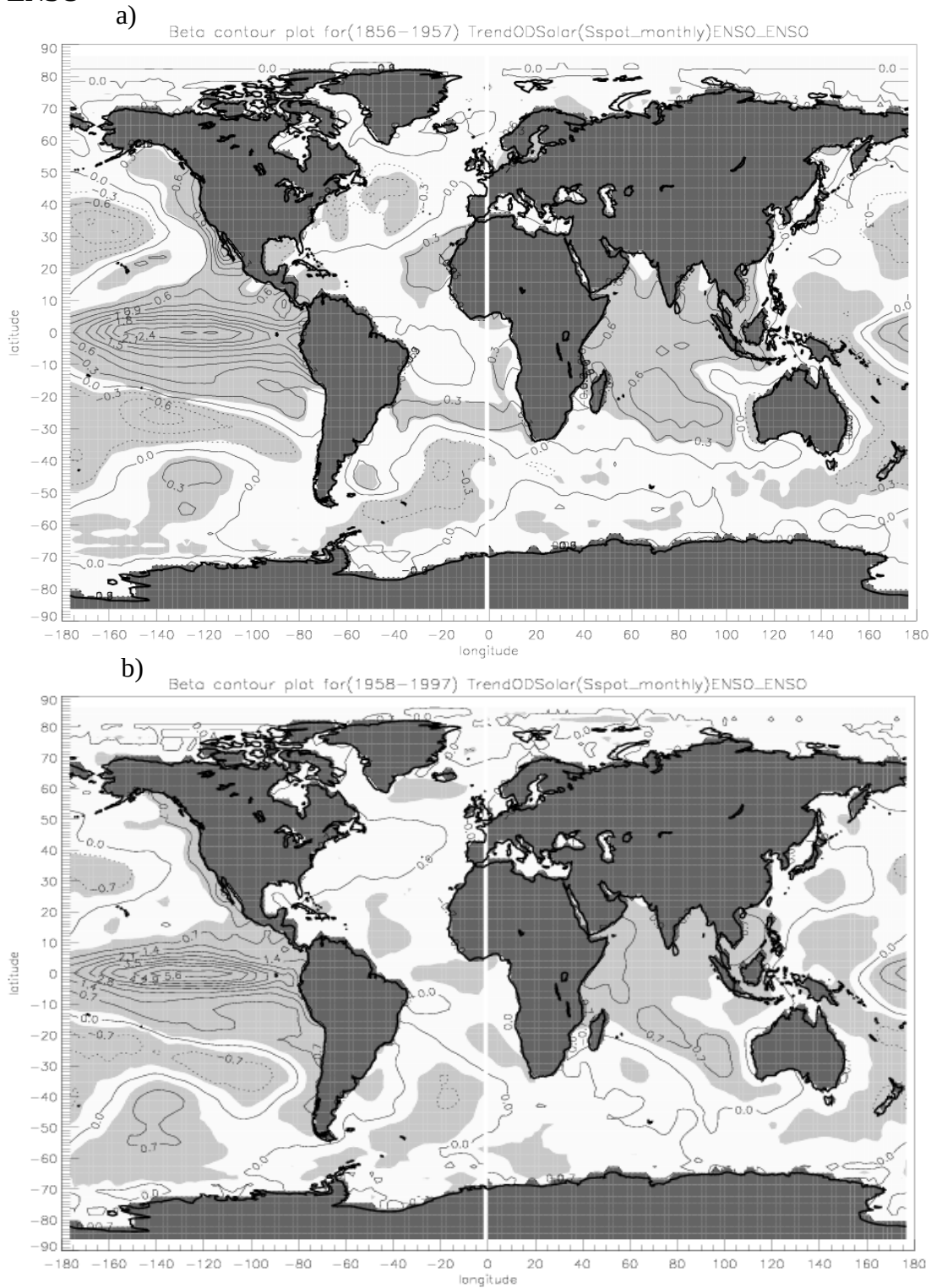


Fig.5.7: Amplitudes of the component of variability of SLP due to ENSO during DJF for period: a) 1856-1957; b) 1958-1997. Other independent parameters used are trend, OD and solar (using monthly SSN).

ENSO



rise from 3 to 6 hPa). That indicates a change in the strength of ENSO, as expressed in SLP, around that region may also have some role for recent warming in the Antarctic Peninsula.

Regression of SST: ENSO and PDO

Amplitudes of the components of variability of SST (using NOAA dataset) due to the ENSO are presented in Fig. 5.8.(a & b). Other independent parameters used are trend, OD and solar (using monthly SSN) and the regression was done using all months of the year (represented by monthly average values), removing the annual cycle from each time series. Fig. 5.8a shows results for the earlier period, while Fig. 5.8b shows that for the later.

From these Fig. 5.8.(a-b), it is clear that amplitudes of SST variability associated with ENSO around the eastern Pacific have also increased significantly (from 2.4° to 6.6°C) from the earlier period to the later. Such enhancement in the variability of ENSO (measured in terms of SST) during the later period is consistent with a rise in variability of the SO, as demonstrated in Fig. 5.7.

Since there is a similarity in the spatial structures of the ENSO and the PDO, we also included the PDO (without the ENSO) in our regression. Fig. 5.9, shows amplitudes of the component of variability in SST (using the NOAA dataset) due to the PDO (annually, removing annual cycle). Fig. 5.9a is during the earlier period, while Fig. 5.9b is that for the later. Other independent parameters used are trend, OD and solar (using monthly SSN). It is seen from Fig. 5.9 that amplitudes of the component of variability of SST due to the PDO around the eastern part of north Pacific have also increased (from 1° to 2.4°C around 40°N and 160°E) during the later period.

Such changes in intensity of the PDO and ENSO indicate that climate change may be responsible for modifying SST, substantially, around the tropical Pacific via the ENSO and around the mid-latitude of North Pacific via the PDO. All of the changes are likely related to anthropogenic influences, as discussed in detail e.g., IPCC (2007).

PDO:

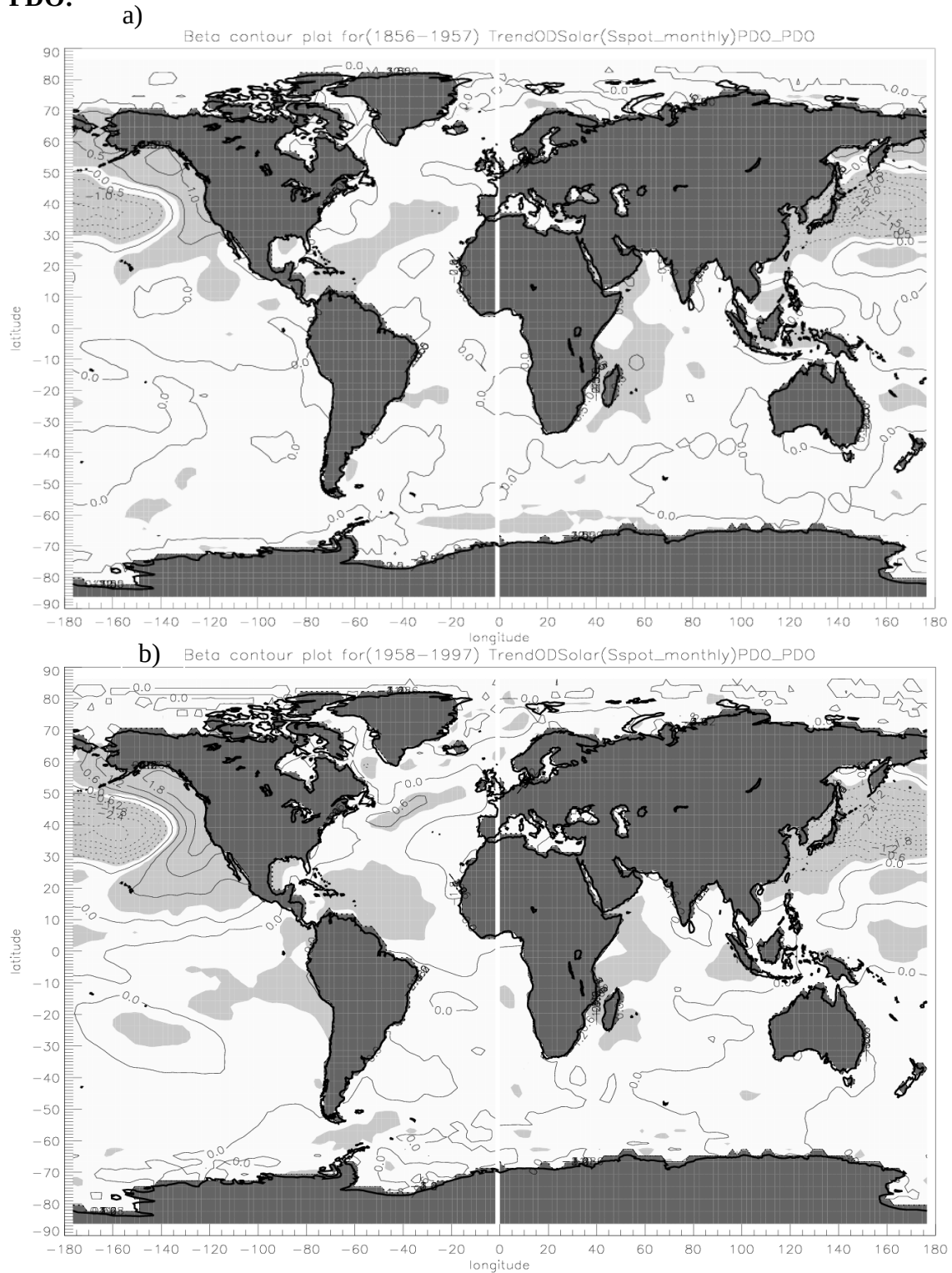


Fig.5.9: Amplitudes of the component of variability of SST due to PDO (annually without annual cycle) for period: a) 1856-1957; b) 1958-1997. Other independent parameters used are trend, OD and solar (using monthly SSN).

5.5. Summary

Our empirical observation suggests that when the Sun is more active it induces changes in SSTs before 1958 and after 1997. This signal is overwhelmed by the innate strong ENSO variability at LS activity (Fig. 5.3.2a). Other authors have suggested that tropical circulations and shallow MOC around the tropical Pacific were different during the intervening period and such observations may have implications relating to the solar influence in the context of changes in ENSO, tropical circulation, shallow MOC and climate change.

Unlike the analysis for the whole 150 years of SLP and SST (Chapter 3), here we restrict our multiple regression analysis to two time periods: the earlier period comprises 1856-1957 and the later 1958-1997. Such segregation of periods, enables us to capture the modified form of the individual influences (mainly solar and ENSO) during the later period, owing to the presence of combined anthropogenic effects (including by GHG and ozone depleting substances), compared to that during the earlier period.

Solar signal and polar modes:

To capture the robust features (and the difference, if any) of the 11 year cyclic variability of the Sun in SLP, during the two periods, we used three different solar indices viz. SSN, Solanki and Foster's TSI (Fig. 5.4.(1-3)a). Here the regression was done using all months of the year (represented by monthly average values), removing the annual cycle from each time series, with other independent parameters being a linear trend and stratospheric aerosol optical depth. Such analysis suggests that the effect of solar activity (regardless of the choice of TSI reconstruction) before the 1950s, is to produce a clear positive signal in both the polar annular modes expressed in SLP. However, during the later period, the solar signal does not conform with the signal observed during the earlier period, using any of the solar indices (Fig 5.4.(1-3)b compared to Fig. 5.4.(1-3)a) and fails to capture any clear signal in the polar annular modes expressed in SLP.

ENSO, PDO and climate change:

Our regression result suggests that ENSO is not the same during the later period as was observed earlier. Here, the signal of ENSO during DJF (where mean value during DJF represent the year), is associated with a strong NAO pattern (Fig. 5.7b) which is not the case during the earlier period. Moreover, during the later, the magnitude of the influence of ENSO on SLP in the equatorial Pacific has also increased.

Thompson and Solomon (2002) noticed that the observed surface temperatures, during last few decades of 20th century, have decreased over most of the Antarctic continent, with the exception of warming over the Antarctic Peninsula. Our current analysis also suggests that there is a strong influence of the ENSO on SLP around the Antarctic Peninsula, (eg., a rise from 3 to 6 hPa during DJF of the later period), indicating that a change in the strength of ENSO, expressed in SLP, around that region, may have some role for recent warming in the Antarctic Penninsula.

It is also seen that amplitudes of SST variability associated with ENSO around the eastern Pacific have increased significantly (from 2.4° to 6.6°C) during the later period, relative to the earlier one. Amplitudes of the component of variability of SST associated with the PDO around the eastern part of the north Pacific have also increased (i.e., from 1° to 2.4°C around 40°N and 160°E) during the later period. Such changes in intensity of both the PDO and ENSO indicate that climate change may be responsible for modifying SST, substantially, around the tropical Pacific via ENSO and around the mid-latitude of North Pacific via PDO. These changes are likely related to anthropogenic influences, as discussed in detail e.g., IPCC (2007).

Solar signal and ENSO:

An underlying quasi-decadal variability in the inter-annual ENSO as detected in some recent studies (White et al., 2008; Zhang et al., 1997; Zhao et al., 2003; Chen et al 2005), raises the question whether the Sun is having any influence on the ENSO. Adding an 11-year-period cosine signal of amplitude $\sim 2.0 \text{ W m}^{-2}$ to the solar constant in the fully coupled ocean-atmosphere general circulation model (i.e., Fast Ocean-Atmosphere Model

(FOAM)) of Jacob et al. (2001), White et al. (2008) were able to simulate both the ENSO and the QDO; with their model QDO, similar in patterns and evolution with the observed one. On the other hand, in its absence (11 year solar signal), the FOAM can only simulate the ENSO. Our observation (Fig. 5.3.1) with regard to the solar-ENSO relationship in HS years (SSN >80), puts more weight on this result and on the necessity to find a plausible mechanism for it.

Proposed Mechanism: Earlier period

Our multiple regression analysis for DJF (mean value during DJF representing the year), detected a solar signal in SLP, around the tropical eastern Pacific and northern Pacific, during the earlier period (Fig.5.4a). A small but significant signal in the tropical eastern Pacific, responsible for intensification of the ITCZ and associated enhancement of the trade wind, during HS years, can account for uplifting the thermocline at the eastern Pacific coast. It consequently can produce a situation similar to that of a cold event of the ENSO, during HS years and vice versa during LS years. Thus inciting the trade wind around the ITCZ of the eastern Pacific, a chain of mechanism similar to that of the ENSO may be initiated via a solar forcing on SLP during DJF.

Earlier, we mentioned that a wind forcing mechanism around the equatorial Pacific is required to trigger the ENSO cycle; the source being unclear. Our current study indicates that the Sun might influence this trigger. Such observation also indicates a decadal signal in the Walker circulation, (simultaneously, with the ENSO) related to the Sun, to which some other authors (e.g. Vecchi and Soden, 2007) have pointed as ‘the source still unclear’.

Our empirical observation shown in Fig. 5.3.1a captures similar solar ENSO behaviour during the earlier period, though only for HS years (say SSN >80). A possible explanation, supported by the work of Dima et al. (2005) may shed some light on why that observation fails to indicate the reverse, during LS years.

Dima et al. (2005) identified one mode of climate, originated by solar variability, which dominates SLP and upper atmospheric levels. However, at the surface, they detected another mode, mainly dominated by atmosphere-ocean coupling, that explains about three times more variance than that due to the solar. Thus ENSO, a measure of the tropical Pacific SST, which is originated via atmosphere-ocean coupling around that region, possesses the potential strength to overpower the influence of solar mode, at the surface.

On the other hand, the solar signal, might make its presence felt differently during HS years and can have the potential strength to overshadow the usual inter-annual ENSO characteristics. Hence our study during the earlier period, indicates that the solar signal during HS years may influence tropical SSTs, but is overwhelmed by the innate strong ENSO variability at LS activity (Fig. 5.3.1a). Nevertheless, the decadal signature in the ENSO, generated via the Sun, as described above, cannot be ignored. Such indication, may also shed light regarding some unexplained behaviour of the ENSO cycle (viz. premature cessation or prolonged lifetime, as mentioned above).

The regression fails to capture any detectable solar signal in SST around the tropical eastern Pacific, without or with the ENSO as an independent parameter (Fig. 5.5 a & b respectively) during earlier period. As very few data points show such bias towards a cold event side of the ENSO during HS years, compared to the whole set of observation (shown in Fig. 5.3.1a), regression fails to indicate any strong result for the Sun and SST in the tropics (Fig. 5.5a). Furthermore, when we include the ENSO as an index in the regression, we still fail to capture any detectable solar signal in SST (Fig. 5.5b), suggesting again that it is the ENSO which is dominating SST in the eastern tropical Pacific most of the time. Thus, the apparent influence of the Sun on ENSO during the earlier period, may be initiated through triggering the trade wind, but is not detectable by the regression analysis in tropical SSTs.

In mid-latitudes, a strong solar signal in SLP, around the AL (~6 hPa) is detected. Such a signal can be related to a weakening of the Ferrel cell around the north Pacific during HS

years and vice versa during LS years. The mid-latitude (north) and tropics in Pacific are tied through ocean pathways, and we find that the solar signal in the lower troposphere is also coupled with the ocean around those places, with the Sun being a potential driving factor to regulate the coupling mechanism.

Two fundamentally different routes for a solar influence on the troposphere have been proposed: one is the ‘top-down’ mechanism and the other the ‘bottom up’ one. First we address the ‘top-down’ solar influence which is generated through the stratosphere. In a GCM study (in absence of ocean), Haigh (1996, 1999) also detected a weakening of the Ferrel cell, associated with enhancement of solar forcing in the lower stratosphere. Such a change should be reflected in the north Pacific sub tropical gyre, which is wind driven. Weakening of the subtropical gyre will impede overturning (thus mixing with cold water) in the north-eastern part of Pacific, augmenting temperature around that place (also seen in Fig. 3.3b). As we mentioned earlier the Pacific is connected between the tropics and mid-latitude (north) via the shallow MOC. Thus the solar signal around the north Pacific can be transported to the tropical Pacific and vice versa. Through such a linkage, the Sun may influence not only the trade winds, but also the tropics via the mid-latitudes.

In the ‘bottom-up’ pathway, the Sun can directly influence SST without stratospheric feedback. Here we suggest that this pathway is possibly involved with the shallow MOC and originated in the north Pacific (Fig. 5.1.1). The shallow MOC, during HS years, absorbs more heat around the place of north Pacific, and subsequently can cause weakening of the AL - which in turn, can reduce the strength of Ferrel cell around the north Pacific. The heat absorbed can again be transported to the tropics of Pacific via the shallow conveyor belt.

Thus our study reveals that the mid-latitude (north) and tropics in the Pacific are not only tied through ocean pathways, but our solar signal in the lower troposphere is also coupled with the ocean around those places. The Sun might be a potential driving factor to regulate such a coupling mechanism and hence may also characterise the decadal fluctuations observed in some of the parameters mainly regulated by the ocean in the

surface layer. For example, the study of Zhang et al (2006), using historical hydrographic data during the last 50 years, observed decadal variability in the shallow MOC. The strength of equatorward convergence of the pycnocline volume transport across 9°N and 9°S also characterizes decadal variability and the circulation fluctuates significantly on decadal time scales, with maximum decade-to-decade variations of 7-11 Sv about the trend (Zhang et. al., 2006).

Proposed Mechanism: Later period

During the later period, the solar signal on SLP in DJF is missing around the tropics, with a weakened yet significant signal around the place of AL. We now consider how the solar cycle can have a different effect on ENSO during the later period. During the earlier period, the solar signal (during HS years) is responsible for warming in the north Pacific alongside cooling in the tropics. Such a strong temperature gradient can be liable to enhance the rate of flow of shallow ocean current from the mid-latitude (north) to tropics in Pacific and consequently, can hasten equatorward convergence. However, said signal being missing around the tropics (and weakened in the mid-latitude), during the later period, indicates that the modified solar signal may have some role in decreasing the shallow overturning current during 1950s to 1998, that has been reported by several authors (Zhang et al., 2006; McPhaden et al., 2002; Vecchi and Soden, 2007). The presence of a weakened solar signal in the mid-latitude, during said period may be responsible for the decadal signature still present in the shallow MOC.

Our multiple regression analysis also suggests that warming in the tropics is observed around the eastern Pacific during the later period (Fig. 5.6a), unlike the earlier one (Fig. 5.5a). It is also seen that the solar signal during the later period is mixed up with ENSO in the tropics (Fig. 5.6a and b). If the ENSO signal is not excluded from the solar (Fig. 5.6a), then our regression results for the later half of the 20th century suggest that the solar signal resembles that of the warm event of ENSO; which is not the case during the earlier period. Thus during 1958-1997, omission of ENSO from the regression gives a false warming (cooling) signal in SST related to higher (lower) solar activity (as observed by White et al. (1997)). We now discuss a plausible route for transporting such a solar signal and also the associated implications.

During the later period, the solar signal around the ITCZ is no longer present. However, weakening of the Ferrel cell during HS years subsequently causes warming around the north Pacific (alongside, oceanic shallow MOC around that place also absorbs more heat) and via the shallow ocean pathway may cause warming in the tropical Pacific. In Fig. 5.1.5 we showed that oceanic heat content around the tropical Pacific basin is correlated with the ENSO index- with more heat indicates warm phase of the ENSO. Such a route of the solar signal, via transporting heat from the mid-latitude to tropics, may activate a warm ENSO cycle in the tropics, during HS years. Fig. 5.3.2b also indicates such a shift in the ENSO, towards a warm event side during the later period for HS years, in contrary to the observation of the earlier. Thus, during the later, the solar signal in the regression is shown to be mixed up directly with the ENSO, producing a comparable effect on tropical SST (Fig. 5.6a); however, if the ENSO is separated from the regression (Fig. 5.6b), the ENSO is very strong in nature and dominates tropical SSTs.

Thus, the Sun which was shown could be a potential factor for triggering a cold event of ENSO at HS years, during the earlier period, may produce a contrary effect during the later. The scatter plot of Fig 5.3.2b, compared to 5.3.2a, also indicates such a bias of HS years on the warm ENSO side during the later period, compared to the earlier one, though solar max years (used by vL07) are still found to be on the cold event side. Relating to solar ENSO behaviour, apart from indicating a plausible mechanism, our analyses are also able to reconcile some of the contradictory previous results.

Contradiction and Reconciliation:

Our multiple regression study of the 1958-'97 data is able to explain some of the inconsistencies relating to the solar signal around the eastern tropical Pacific; for instance, the contradictory results of vL07 and White et al. (1997), in addition to White and Liu (2008). Our analysis of regression (during the time period, common to all the studies), accompanied by our observation that the years of peak annual SSN used by vL07 generally falls a year or more in advance of the maximum of the smoothed DSO, provides coherence to these apparently conflicting findings.

First, in chapter 3, we identify that the methodology adopted to detect the solar signal by vL07 is not characterising the true behaviour of a solar QDO. In fact, peak years used by vL07 generally falls a year or more in advance of the broader maximum of the 11-year solar cycle and thus reflects a different (rising) phase of the solar cycle to peak year of smoothed DSO (shown in Fig. 3.5). We observed that the peak years of solar sunspot cycle are usually associated with the negative phase of ENSO cycles (Fig. 3.4b). As the ENSO usually takes 1-2 years to change from cold to warm phase this explains why the results of vL07 differ to those of White et al. (1997). During 1958-1997, omission of ENSO from the regression gives false warming (cooling) signal of higher (lower) solar on SST in the tropics. It clearly indicates why both White et al. (1997) and White and Liu (2008) not only detect warming in HS years, but also *cooling in LS years* (Fig. 5.6a), during the time period, common to all the studies. The trough and peak of ENSO shows a period of about 2 years due to its usual phase transition mechanism, *throughout the solar cycle*, as observed by White and Liu (2008) (Fig. 1.2.11); hence consistency between vL07, White et al. (1997) and White and Liu (2008) during the time period, common to all the studies. Such observation also explains why solar max years (as used by vL07), are still on the cold event side during 1958-1997 (shown in Fig. 5.3.2b); which is not any special features of the rising phase of solar cycles – it is simply an artefact of the ENSO.

Moreover, our study indicates that the composite of 9 solar cycles (during 1900-2000) as considered by White and Liu (2008) in their analysis, generally captures the behaviour of 1958-1997, the period strongly affected by the ENSO and having stronger solar cycles. It also suggests that the application of their approach to the period prior to 1958 would be unable to detect similar phase locking of the ENSO and solar.

Our analyses suggest that, mixing of ENSO with the solar signal during the later period might modulate the true solar signal (and hence insensitive to some methodology) which needs to be accounted appropriately. This might shed light on some of the discrepancies relating to results of the solar influence on surface temperature.

Chapter 6: Atmosphere-Ocean Coupling

6. Atmosphere-Ocean Coupling

6.1. Coupling processes: Hypotheses and evidences

Recent modelling experiments of Rind et. al. (2008) suggest that both the proposed mechanisms for solar influence on the troposphere, i.e., forcing from above (stratospherically driven) and from below (with an SST influence) are operative; although the tropospheric and stratospheric dynamic responses are likely to be affected by the atmospheric background state. According to them, the use of a coupled atmosphere-ocean model, capable of realistically simulating ENSO phenomena would be necessary to assess the solar impact on the tropical Pacific circulation. Here, we seek to thread a chain of causalities, involved in the coupled atmosphere –ocean system (mainly involving the Pacific ocean) based on evidences and hypotheses, in a holistic context.

We present a flow chart (Fig. 6), depicting a consolidated overview of atmosphere-ocean coupling, supported by observations and mechanisms. Results from our study (as described in Chapters 3-5) have been incorporated in formulating this flowchart. The three major climate variabilities, viz. solar, QBO and ENSO are shown with oval outlines; whereas, the major circulations, responsible for modulating the effect of major variabilities are shown by non-rectangular parallelograms. The pathways of the signals are marked by labels starting from ‘A’, initiated by solar variability and the direction of change in behaviour during the steps are shown by ‘+’ (for increase) or ‘-’ (for decrease). Subscripts in a label are introduced to indicate part of a same process; however, having evidences in each steps. How the solar signal is transmitted to the troposphere via the stratosphere, is covered by ‘A-M’ and shown with the heading ‘**atmosphere**’. Here we have indicated two pathways: one via the **lower stratosphere** (equatorial region) to the troposphere through circulation (A-H). The other from the **stratospheric polar vortex** to the troposphere via polar modes (A, I-M). The role of ‘**atmosphere and ocean**’ have been covered by ‘N-S’. Finally, the ‘**climate change**’ fingerprint in our flowchart is marked by ‘T-Z’ with their pathways shown by dash-dotted lines.

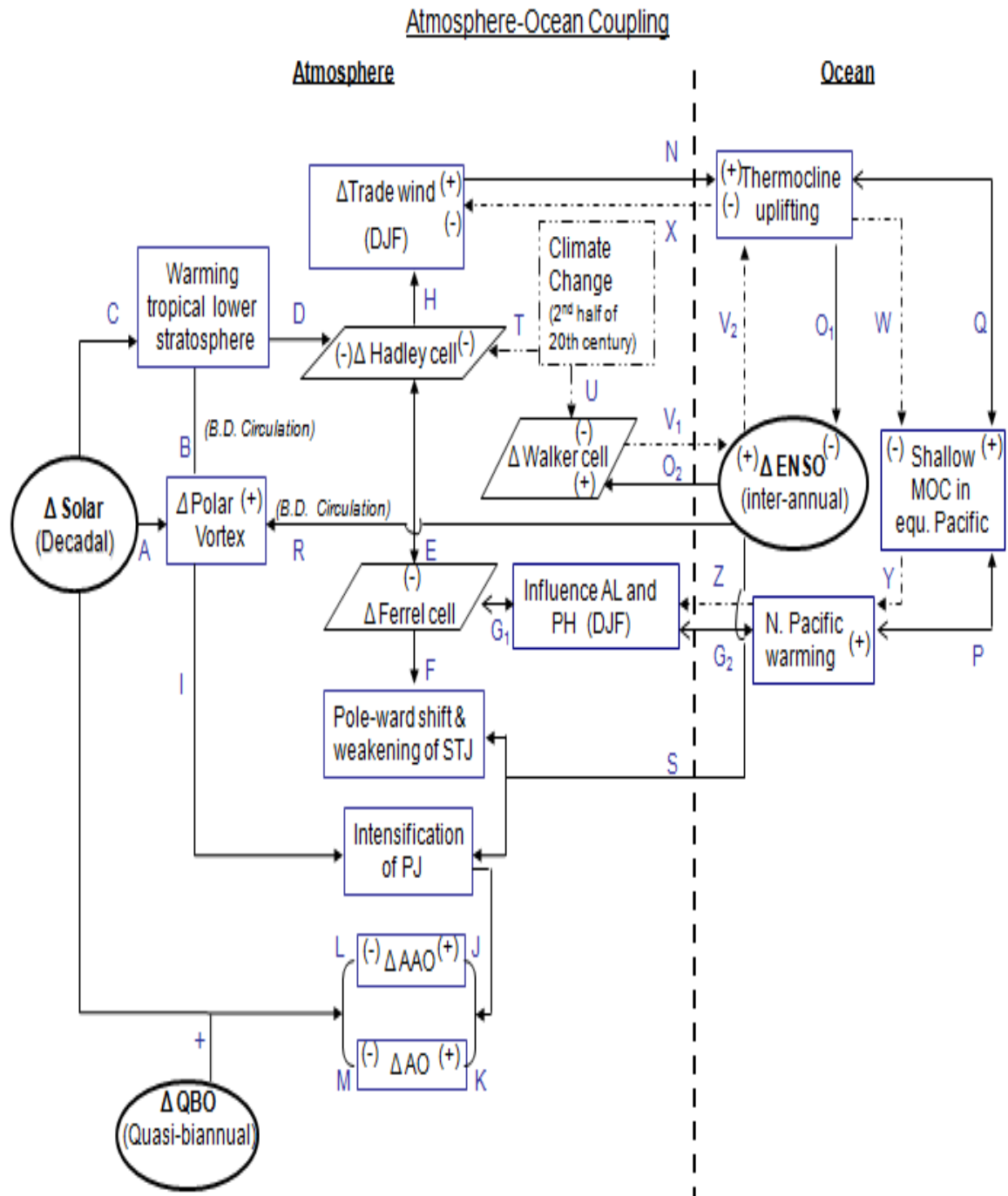


Fig. 6: Flow chart of proposed atmosphere-ocean coupling

Atmosphere (A-M)

Enhanced solar activity is liable to increase the strength of the polar vortex - which is designated by 'A'. The hypothesis that solar cycle/ozone interactions create temperature and wind anomalies in the upper stratosphere near 1 hPa (around 50 km), which is subsequently responsible for developing/ modulating the polar stratospheric jet (also shown in our schematic of Fig. 4.1.3) has been evidenced in observations and also explained by modelling studies. A stronger stratospheric polar jet is associated with a stronger and colder polar vortex (also shown in Fig. 1.1.10 of this thesis) and since HS years generate a stronger stratospheric jet, we can say that the solar influence around the stratospheric polar vortex, which is marked by 'A', is well explained by mechanisms with observational evidence.

Lower stratosphere (A-H)

How the perturbation in the polar vortex is communicated to the tropical lower stratospheric region - is marked by 'B'. Kodera and Kuroda (2002) proposed a mechanism, whereby solar heating around the stratopause may influence the atmosphere below. According to them, the solar heating anomalies that changes the strength of polar stratospheric jet can influence the path of upward propagating planetary waves. These waves, depositing their zonal momentum on the poleward side of the jet, weaken the Brewer–Dobson circulation and warm the tropical lower stratosphere in solar maximum years (also shown in Fig. 4.2.1 of this thesis). 'B' in our flow chart is explained via this mechanism. Moreover, the observational results of Crooks and Gray (2005) and Haigh (2003) also shown in Fig. 4.3b of this thesis also suggests that the tropical lower stratosphere is warmer in the solar maximum than in the minimum, which provides observational evidence for 'B'.

However, warming (a small proportion of observed amplitudes) in the tropical lower stratosphere can also be due to UV absorption by ozone in the lower stratosphere, during solar max – which is shown by 'C'. Haigh (1996) using an AGCM with fixed SSTs

(without ocean) first showed that a detailed representation of the stratosphere may not be necessary for simulating the tropospheric aspects of solar influence, although the source of stratospheric heating remains an important factor. Despite the presence of a uniform stratosphere, *the lack of the stratospheric polar vortex* and use of broad latitudinal-scale perturbations, the model was able to reproduce the tropospheric patterns (at lower amplitude). Thus ‘C’ gives an indication about the mechanism.

In the flow chart (Fig. 6), ‘D’ indicates that warming in the equatorial lower stratospheric region can be responsible for a weakening and expansion of the Hadley cell (shown with ‘-’); a weakening of the Ferrel cell (shown by ‘E’ with ‘-’) and a shift in the STJ poleward, alongside weakening it (shown by ‘F’). Results from both modelling and observational studies, confirm the proposed pathways.

Haigh (1999) presents atmospheric circulation model results for the influence of the 11 year solar cycle on the climate of the lower atmosphere. In her model, solar forcing is represented by changes in both stratospheric ozone concentrations and incident irradiance, where the former has the stronger impact. A pattern of response is found, in which the subtropical jets and midlatitude Ferrel cells move poleward and the tropical Hadley cells broaden and weaken for high irradiance. Despite the presence of a uniform stratosphere, (without the stratospheric polar vortex) and absence of an ocean, the simplified global circulation model (GCM) studies of the solar influence, also showed an impact on tropospheric mean meridional circulation, characterising a weakening and expansion of the tropical Hadley cells, along with a poleward shift of the Ferrel cells (Haigh, 1996; Haigh et al, 2005; Haigh et al, 2006). All these studies are consistent with ‘C-F’. Recently, Simpson et al. (2008), using a simple model, were also able to capture the tropospheric features and indicate more about mechanisms. Thus, all these studies not only agree with ‘C’, but also offer mechanisms to explain ‘D-F’.

‘D-F’ are also evidenced in a number of observational studies. Haigh et al. (2005), using multiple regression analysis of NCEP Reanalysis data for zonal mean zonal wind, showed when the Sun is more active, the subtropical jets are positioned further polewards

and weaker (also shown in Fig. 1.2.5 of this thesis). Brönnimann et al. (2006) using upper air data also observed poleward displacement of the subtropical jet and Ferrel cell with increasing solar irradiance.

The weakening and shifting of the Ferrel cell is associated with subsequent changes in the AL and PH around the Pacific - is marked by 'G₁'. The semi-permanent pressure systems (as shown in Fig. 1.1.1 of this thesis) in the troposphere are an integral part of the tropospheric circulations and play an important role in controlling their behaviour. Studying the variations of semi-permanent pressure systems, the AL and the PH, Christoforou and Hameed (1997) found that solar variability influences the location of these COAs, thus causing changes in storm tracks and large anomalies in regional climatic conditions. Apart from shifting the position, the AL also exhibits significant differences in intensity. Meehl et al. (2007) and vL07, using the method of solar max compositing confirmed this observation. Our multiple regression studies also identify changes in intensity around the AL, due to 11 year solar cyclic variability (Fig.3.2, Fig. 3.5.1), supporting previous results. Such a solar influence around the AL is observed even during the two different time period: period comprising 1850-1957 and period 1958-1997 (Fig. 5.4a & b). Movements of the COA in the Pacific is also in agreement with the observational analysis of Haigh et al. (2005) and Brönnimann et al. (2006), in terms of poleward displacement of the Ferrel cell, Hadley cell and subtropical jet around the NH in Pacific. Though 'G₁' is found in observational results, whether weakening and shifting of the Ferrel cell is associated with subsequent changes in the AL and PH around Pacific, during DJF (and vice versa, shown with open arrow) needs investigation. Hence 'G₁' may be considered as a hypothesised mechanism.

During DJF, HS years are associated with the intensification of ITCZ around the eastern Pacific - which is designated by 'H'. Our study, using the multiple regression analysis, suggests that during DJF, the solar signal (using monthly SSN), is significant around the ITCZ of the eastern Pacific (Fig. 3.2 and Fig. 5.4a). During HS years, the ITCZ intensifies and thus enhances trade winds in the central and eastern Pacific region. Such 11 year solar cyclic variability around the ITCZ of the eastern Pacific has also been

captured using the TSI reconstruction from Krivova & Solanki, and Foster (Fig. 3.5.1). Moreover, Meehl, et al. and van Loon, et al., in a succession of papers (2007, 2008, 2009), based on modelling and observational analyses also showed that solar peak years are associated with the intensification of ITCZ around Pacific, during DJF. Thus ‘H’ is not only evidenced in observations but also explained by a mechanism. Whether it is associated with the weakening and pole-ward positioning of the Hadley cell, as shown in the flow chart (Fig. 6), that needs to be tested and hence may also be considered as a hypothesised mechanism. Thus the evidence for each link may be summarised:

B: Kodera and Kuroda (2002), Haigh (2003), Crooks and Gray (2005)

C: Haigh (1996)

D: Haigh (1996), Haigh (1999), Haigh et al. (2005, 2006), Simpson et al. (2008)

E: Haigh (1996), Haigh (2003), Haigh et al. (2005, 2006), Simpson et al. (2008),
Brönnimann (2006)

F: Haigh et al. (2005, 2006), Simpson et al. (2008), Brönnimann (2006)

G₁: Christoforou and Hameed (1997), vL07, Meehl et al. (2007), Fig. 3.2, Fig.
3.5.1 and Fig 5.4a & b of this thesis

H: Fig. 3.2, Fig. 3.5.1 and Fig. 5.4a of this thesis, Meehl et al. (2008, 2009), vL07

Upper stratosphere (via Polar modes): (A, I-M)

‘A’ has been described above, here we discuss from label ‘I’ onward.

Here ‘I’ indicates that strengthening of the polar vortex is related to an intensification of the upper tropospheric polar jet. Perturbations in the polar vortex that appear to be related to the downward propagation via the NAM and SAM were shown by Baldwin et al. (2001) in observational analysis (also shown Fig. 4.2.3 of this thesis). They discussed a dynamical mechanisms that might communicate stratospheric circulation anomalies downward to the troposphere and surface via polar modes of variability. Thompson and Wallace (2000) also identified such downward propagation of both the polar modes. Thus, intensification of the stratospheric polar vortex is very likely to be associated with

the intensification of upper tropospheric polar jet and hence it can be said that 'I' is evidenced in observations.

Intensification of the upper tropospheric polar jet is also allied with the positive phase of surface SAM (AAO) and surface NAM (AO), which are shown by 'J' and 'K', respectively and discussed here briefly. There are strong similarities between the meridional structures of the annular mode in the NH and SH and they follow similar mechanisms of formation (Thompson and Baldwin (2001); Thompson and Wallace (2000)). In addition, as mentioned above Baldwin et al. (2001) in their study discussed a dynamical mechanisms that might communicate stratospheric circulation anomalies downward to the troposphere and surface, via the polar modes of variability. Consistent with these studies, our results, shown in Fig. 5.4.(1-3)a suggests that the 11 year solar signal is a positive feature, of both the polar modes, expressed in SLP during 1850-1957. The solar signal (during JJA), considering a total of nearly 150 years data also reveals a positive signal in the surface SAM (AAO) (Fig.3.5.4). Thus there is evidence for 'J' and 'K' in observational data.

Baldwin and Dunkerton (2005) mentioned that the pathway for solar influence via the polar modes of variability, appears to involve interactions with the QBO, but the details are not yet understood. Hence we introduced QBO in the flow chart and the combined effect of QBO and Sun is shown by pathways 'L' and 'M', which suggest that the effect of the QBO combined with solar activity reveals a -ve signal in the de-seasonalised AO and AAO. This is based on studies which indicate that during the NH winter, it is necessary to group the meteorological data according to the phase of QBO, in order to find a clear signal of the 11-year solar cycle in the stratosphere. For example, Labitzke and van Loon (1992) using data for the years 1956 – 1991, show that warm polar temperatures tend to occur during the west phase of the QBO at solar maximum and east phase at solar minimum. Moreover, Haigh and Roscoe (2006), through multiple regression analysis (using NCEP data during the 2nd half of last century), showed that there is no statistically significant solar signal in either the NAM or SAM; however, when a new index, the product of the solar and QBO indices, is used there is a good correlation

throughout the atmosphere in the SAM, and at the lower levels in the winter NAM. Our multiple regression analysis (Fig. 4.4.1.(a-c)) is also showing consistency with Labitzke and van Loon (1992), Labitzke (2004), Haigh and Roscoe (2006) and Baldwin (2001). It indicates that the effect of the QBO (irrespective of using height 30, 40 or 50 hPa level) combined with solar activity reveals a –ve signal in the de-seasonalised AO and AAO. All these studies suggest that ‘L’ and ‘M’ are well evidenced through observational analyses. The evidence for how variability from the upper stratosphere is carried to the troposphere, (via the polar modes) may therefore be summarised as:

I: Baldwin and Dunkerton (2001), Thompson and Wallace (2000)

J: Fig. 3.5.4 of this thesis

J-K: Fig. 5.4.(1-3)a of this thesis

L-M: Labitzke and van Loon (1992), Labitzke (2004), Haigh and Roscoe (2006), Fig. 4.4.1.(a-c) of this thesis

Atmosphere-Ocean Coupling (N-S):

The basic formation mechanism of the ENSO has been described in the previous chapter, where we also mentioned that in the process of ENSO, the role of Walker circulation, the trade wind and the thermocline shifting are inseparable. They are very well documented and their coherence is strongly established through various observational analyses and modelling studies. Thus ‘N’ and ‘O’ are not only evidenced through observational results, but also explained by a mechanism. Thus a chain of cause and effect acts: increase in the speed of trade wind (shown by ‘H’ with ‘+’) - uplifting of the thermocline in the eastern Pacific basin (shown by ‘N’ with ‘+’) - increase in the cold event of ENSO (shown by ‘O₁’ with ‘-’) - increase in the strength of Walker circulation (shown by ‘O₂’ with ‘+’) are all linked together and shown by appropriate labels in the flow chart (Fig. 6). Here we have marked cold (/ warm) event of the ENSO with ‘-’ ve (/ +ve) sign. Our regression analysis suggests that the Sun (during DJF), via triggering the trade wind (Fig. 5.4a), can initiate ENSO cold event like situations during HS years (Fig. 5.3.2a), (described in

details in Section 5.4). This observation supports our proposed pathway by which the Sun can stimulate the trade wind and then, via ‘N’ and ‘O’, influence the ENSO.

Another possible pathway, from atmosphere to ocean (around mid-latitudes) is shown by ‘G₂’, which associates a weakening of the AL, with a warming of the north Pacific, during HS years. Due to a weakening of the Ferrel cell and the AL during HS years, wind forcing in the north Pacific gyre circulation reduces in strength, causing less mixing in the north Pacific shallow ocean water which might act to increase warming around the north Pacific.

The results from Haigh (2003), also shown in Fig. 4.3b, suggest that there is a positive solar response in the tropical lower stratosphere which extends in vertical bands throughout the troposphere via mid-latitudes (with a maximum amplitude of 0.5°K around 40-50°N) in both the hemispheres. Frame and Gray (2009), using the multiple linear regression analysis of the ERA-40 dataset for the period 1979-2008 also noted a positive solar response in the annual temperature, at mid-latitudes in the troposphere, in both hemispheres (Fig.1.2.2b). The results of our data analysis, Fig 3.3, Fig. 5.5.(a-c) and Fig. 5.6.(a-c) are in agreement with such mid-latitude warming during HS years. Thus, observational results provide support for ‘G₂’. However, to verify the cause-effect relationship, i.e., whether solar influence during HS years, around the AL and PH are related to warming of north Pacific, that needs to be tested in computer models and hence ‘G₂’ is also shown as a hypothesised mechanism.

Links ‘O’, ‘Q’ and ‘P’ indicate a decadal signature in the Walker cell and ENSO (‘O’), the shallow MOC (‘Q’ via the thermocline and ‘P’ via north Pacific warming). Our observations during the period comprising 1850-1957 suggest that the Sun, during HS years produces an impact in two regions of the Pacific: cooling in the tropics, but warming in the mid-latitude of north Pacific. Enhancement of the trade wind (Fig. 5.4a) causes more uplifting of the thermocline around the eastern tropical Pacific, which is accountable for more cold water in the tropical shallow oceanic basin. The warming in the northern Pacific occurs around the tip of shallow MOC (shown in Fig. 5.1.1) in the

north Pacific during HS years (Fig. 5.5a). This produces an increase in the temperature gradient (between the tropics and North Pacific), which could be responsible for strengthening the shallow MOC (shown by 'Q' with '+') around the tropical Pacific, during HS years. This is in line with the known fact that during a warm ENSO, the shallow MOC is disrupted. Such analysis also agrees with Mantu and Hare (2002) who indicates that Pacific in the mid-latitude and tropics is connected via ocean pathway.

Thus links 'P', which indicates more warming in the north Pacific can strengthen the shallow MOC, but this may be considered as a hypothesised mechanism. The Sun might be a potential driving factor to regulate such a coupling mechanism and may also characterise the decadal fluctuations observed in some of the parameters mainly regulated by the ocean in the surface layer. For example, the study of Zhang et al. (2006) using historical hydrographic data during last 50 years, observed decadal variability in the shallow MOC. The strength of equatorward convergence of the pycnocline volume transport across 9°N and 9°S also characterize decadal variability and the circulation fluctuates significantly on decadal time scales, with maximum decade-to-decade variations of 7-11 Sv about the trend (Zhang et al., 2006). Our analysis indicates that decadal signals in ENSO (White et al., 2008; Zhao et al., 2003 and Chen et al., 2005), in the thermocline around the tropical Pacific, in the strength of the shallow MOC (Zhang et al., 2006; Veechi and Soden, 2007) and the Walker cell (Veechi and Soden, 2007) can be originated via solar cyclic variability as indicated in the flow chart by 'O', 'Q' and 'P'.

We now discuss how signals from the ocean can be transmitted to the atmosphere and thus indicate the 'bottom-up' coupling processes with associated locations. First, we focus on the signals from ocean to the troposphere which are shown with open arrows ('P' and 'G'). The tropical MOC may influence the north Pacific (shown by 'P' with open arrow). As more water circulates via the MOC in the tropical Pacific, during HS years, there is more heat intake around the North of the Pacific through the tip of the shallow ocean conveyor belt (which is around the AL, shown in Fig. 5.1.1) and thus more warming around the north Pacific during HS years, compared to the surroundings.

We also indicate that warming in the north Pacific can cause changes in the AL and PH (open ended 'G₂') and in weakening and poleward positioning of the Ferrel cell (open ended 'G₁'). During DJF, due to the positioning of the AL and PH, warming around the North Pacific, can have an effect (by generating localised high pressure) on these COA (Fig.3.2); consistent with the results of Christoforou and Hameed (1997) in terms of intensity as well as positioning (shown by 'G₂' with open arrow) (also vL07; Fig. 3.2; Fig. 3.5.1 and Fig 5.4a & b of this thesis). Subsequently, this can influence the Ferrel cell as shown by 'G₁' (with open arrow) (Haigh et al., 2005; Brönnimann, 2006) respectively. However, whether warming in the north Pacific causes change in the AL and PH and subsequently weakening and poleward positioning of the Ferrel cell remains to be established so, we consider both the open ended 'G' and 'P' as hypothesised mechanisms.

Bottom-up processes can even be extended to the stratosphere. Link 'R', suggests that the warm events of ENSO are related to a warm polar vortex and vice versa. It has been claimed that the northern stratospheric polar vortex is more perturbed and warmer during El Niño winters than La Niña winters. Camp et al. (2007a) showed that during winter, warm-ENSO years are significantly warmer in the stratosphere at the Northern Hemisphere polar and mid-latitudes than the cold-ENSO years. Using a GCM, Sassi et al. (2004) and Taguchi and Hartmann (2006) showed that the warming difference between El Niño and La Niña years is statistically significant and SSW are twice as likely to occur in the El Niño winters than La Niña, thus providing a possible connection between the polar stratosphere and ENSO. Moreover, Thompson et al. (2002) observed that pronounced strengthening (weakening) of the NH wintertime stratospheric polar vortex is allied with the cold (warm) phase of ENSO (shown in Fig. 1.2.10). Haigh and Roscoe (2006) pointed out that such association of ENSO with the polar vortex is very likely to be related to changes in the Brewer-Dobson circulation. More recently, Ineson and Scaife (2009) using a general circulation model of the atmosphere showed that there is a clear response of the ENSO in European climate via the stratosphere. This mechanism is restricted to years when SSW occurs, leading to a transition to cold conditions in the northern Europe and mild conditions in the southern Europe in late winter during

El Niño years (via pathway ‘R’, ‘I’ and ‘J’). All these studies provide observational evidences for ‘R’ and also offer mechanisms to explain this link.

How the influence of ENSO may be captured on the polar modes is shown by S’, ‘J’ and ‘K’. Carvalho et al. (2005), using data analysis for the period 1979 to 2000, observed that during the austral summer (DJF), cold events of the ENSO are linked with dominant positive AAO and vice versa. The alternation of AAO phases were also shown to be allied with the latitudinal migration of upper level (200 hPa) STJ (around 45 °S) and the intensity of polar jet (around 60 °S). Positive AAO phases are associated with the poleward shift and weakening of the subtropical feature accompanied by an intensification of the high-latitude feature (their results also shown in Fig. 1.2.9 of this thesis). Since we discussed before, the NAM and SAM follow similar mechanism (Thompson et al., 2001); such behavior is expected to be observed in the surface NAM (AO) as well. Their observation is in agreement with the analysis of Haigh and Roscoe (2006), who in their multiple regression analysis, using data from the later half of the 20th century, showed that an anti-correlation exists between the polar modes and ENSO in the lower troposphere. Thus in essence, labels ‘S’, ‘J’ and ‘K’ have proofs through observations. The evidence for atmosphere-ocean coupling may be summarised:

- O₁:** Fig 5.4a and 5.3.2a of this thesis
- O₂:** Veechi and Soden (2007)
- Q and P:** Zhang et al. (2006), Veechi et al. (2007), Mantu and Hare (2002)
- P(<-) and G₂:** Haigh (2003), also in Fig. 4.3b of this thesis; Frame and Gray (2009), also in Fig. 1.2.2b of this thesis; Fig. 3.3, Fig. 5.5 and Fig. 5.6 of this thesis
- G₂ (<-):** Christoforou and Hameed (1997), vL07, Fig. 3.2, Fig. 3.5.1 and Fig. 5.4a & b of this thesis
- G₁ (<-):** Haigh (2005), Brönnimann (2005)
- R:** Camp and Tung (2007), Sassi et al. (2004) and Taguchi and Hartmann (2006), Thompson and Baldwin (2001), Ineson and Scaife (2009)

S: Carvalho et al. (2005)

J: Carvalho et al. (2005), also shown in Fig. 1.2.9; Haigh and Roscoe (2006)

K: Haigh and Roscoe (2006)

Climate Change (T-Z):

Signals relating to climate change have been marked from ‘T’ to ‘Z’ and their pathways indicated by dash-dotted lines. Here, we start with ‘T’ and ‘U’, which suggest climate change during the 2nd half of the 20th century is responsible for weakening of both the tropical circulations, the Hadley cell and Walker cell. We mentioned in the last chapter with emphasis (in a box, page number 134) that observational results suggest that climate change is responsible for weakening of both the Hadley and Walker cell (though more pronounced in the later), during the last half of 20th century (Vecchi and Soden, 2007). Weakening of the Walker circulation in climate models, forced with increased greenhouse gases has also been well documented (Meehl et al., 2007; Tanaka et al., 2004; Vecchi et al., 2006). Thus ‘T’ and ‘U’ have an established mechanism as well as being evidenced in observational analyses. In subsequent discussions, we show that climate change via the Walker circulation probably plays a role in modifying the atmosphere-ocean coupling system that was discussed above.

In heading ‘Atmosphere-Ocean coupling’, we discussed that an increase in the strength of Walker cell (shown by ‘O₂’ with ‘+’) is associated with the cold event side of ENSO (‘O₁’ with ‘-’). Likewise, weakening of the Walker cell (‘U’ with ‘-’) is likely to favour a warm event side of ENSO (‘V₁’ with ‘+’). This, in turn, is allied with less uplifting of the thermocline around the eastern tropical coast of Pacific (‘V₂’ with ‘-’) together with a decelerating effect on the trade wind system (marked as ‘X’ with ‘-’). Thus, the solar signal present around the ITCZ of eastern Pacific during the earlier period (though small in magnitude but significant) is not observed during the later period (Fig. 5.4b compared to Fig. 5.4a). Moreover, the solar forcing which was shown could be a potential factor for triggering a cold event of the ENSO at HS years, during the earlier period (through

pathway ‘N’, ‘O’ from tropics) may provide the opposite during the later. Our observational result shown in Fig 5.3.2b (compared to 5.3.2a), also indicates such a bias of HS years towards the warm event side of ENSO, during the later period. Thus ‘V’ and ‘X’ are explained by the usual ENSO mechanism; moreover, ‘V’ and ‘X’ are also evidenced in observational results.

Now, we focus on ‘W’ and ‘Y’ which indicate that, suppression of the thermocline uplifting weakens the shallow MOC in the tropical Pacific (‘W’ with ‘-’) and that this subsequently reduces warming around the north Pacific (shown with ‘Y’) and can influence the AL (shown with ‘Z’). Deepening the thermocline, favours warm shallow ocean current in the tropical Pacific basin, causing a reduction in the flow of equatorward convergence of ocean current in Pacific, which was associated with the tropics and mid-latitude temperature contrast during the earlier period (through pathway ‘N’, ‘Q’ from tropics, alongside ‘P’ from mid-latitude). Simultaneously, the climate change fingerprint, via deepening of the thermocline in the tropical Pacific is taking place in the shallow MOC (via the pathway ‘V’ and ‘W’). Thus climate change may have a decelerating effect on the shallow MOC, around the Pacific (‘W’ with ‘-’) that has been observed during 1958-1997 (Vecchi and Soden, 2007; Zhang et al., 2006) and hence ‘W’ is evidenced by observations. Moreover, weakening of the shallow MOC can influence the region around north Pacific via less heat intake (compared to the earlier period), which is shown here by ‘Y’ and also detected in our observational analysis (Fig. 5.6 a & b). Being lesser warming around the AL, the solar signal observed there in SLP, during DJF in the earlier period is seen to be weakened during the later (Fig. 5.4b compared with Fig. 5.4a) and shown by the pathway ‘Z’. Thus ‘Z’ is evidenced by observations. We still need to verify whether reduced warming around the north Pacific is related to strengthening of the AL (compared to the earlier period) thus ‘Z’ is considered a hypothesised mechanism. Such a mechanism is the same as the hypothesised mechanism ‘G₂’, mentioned above.

We now focus on the modified solar signal during the later period. The shallow ocean pathway of the conveyor belt not only serves a crucial role in mass exchange between the

mid-latitudes and the tropics of the Pacific, but also can act as a medium for transporting the solar signal. The solar signal (DJF), being missing in the tropics ('X') and weakened around the place of AL ('Y') during the later period compared to the earlier one (shown in Fig. 5.4b compared to Fig. 5.4a), indicates that the Sun can amend the strength of shallow MOC around the tropical Pacific. For example, the Sun, during HS years for the later period, can be responsible for weakening the shallow MOC circulation in the tropical Pacific by reducing the tropics and mid-latitude temperature contrast (being absence of trade wind forcing in the tropics). This also suggests that a decadal footprint (though weakened and different compared to the earlier period) in shallow MOC is still there, during the later period. Such a mention of detecting a decadal signature in the shallow MOC in the tropical Pacific has been reported by several authors (Veechi and Soden, 2007; Zhang et al., 2006). A weakened, yet significant solar signal around the AL (Fig. 5.4b compared to Fig. 5.4a) during that period (supported with the flow chart of Fig. 6) still indicates *more (less) heat* intake from the north Pacific during *HS (LS) years* (and carried towards the tropical basin via the shallow MOC). Bearing in mind the role of the shallow ocean current in the Pacific and incorporating the fact that the upper ocean heat content in the tropical basin is a pre cursor of ENSO (also shown in Fig. 5.1.5 of this thesis), such analysis during the later period, suggests - warm events of the ENSO during HS years and cold during LS years are favoured during the final half of the last century. A similar solar influence on the ENSO has been observed by White et al. (1997) as shown in Fig. 1.2.3 of this thesis. Our multiple regression analysis also suggests that during 1958-1997, omission of the ENSO from regression gives false warming (cooling) signal of higher (lower) solar on SST in the tropics (Fig. 5.6a), unlike the earlier period (Fig. 5.5a). Thus the Sun during the later period is found to be influencing (/mixed up with) the ENSO (Fig. 5.6a and b), which is only via 'P', 'Q' (open arrow) and 'O₁', affected by climate change signal 'Y', 'W' and 'V₂' respectively. If the ENSO is included in the regression, the solar signal in SST in the tropics is nominal, as ENSO is very strong in nature and dominates SST in the east tropical Pacific.

In summary, the evidence for the impact of climate change on the solar signal may be evidenced by:

- T:** Vecchi and Soden (2007)
- U:** Vecchi and Soden (2007), Meehl et al. (2007), Tanaka et al. (2004), Vecchi et al. (2006)
- V₁:** Fig. 5.3.2b of this thesis (compared with Fig. 5.3.2a)
- W:** Vecchi and Soden (2007), Zhang et al. (2006)
- Y:** Fig. 5.6 a and b of this thesis
- X, Z:** Fig. 5.4b of this thesis (compared with Fig. 5.4a)

As ENSO is affected by climate change during the later half of the 20th century, ‘S’ of Fig. 6 is also likely to be influenced; hence are AAO and AO (via ‘J’ and ‘K’). Moreover, the climate change signal in the ENSO will also be captured in the polar stratosphere through modulating the strength of Brewer-Dobson circulation via pathway ‘R’ and subsequently, can influence the polar modes via ‘I’. Thus the solar signal during the earlier period, that captured a positive signal in both the polar modes expressed in SLP, indicates differently during the later period (Fig. 5.4.(1-3) b, compared to 5.4.(1-3) a respectively).

Table 6 indicates how each link is evidenced by observation, explained by a mechanism or only be viewed as a hypothesis at this stage.

Recently, Meehl et al. (2009) investigated two mechanisms, the top-down stratospheric response and the bottom-up coupled ocean-atmosphere surface response, in versions of three global climate models (CCSM3, WACCM-fixed-SST, WACCM-coupled), with mechanisms acting together or alone and compared results with their observations. The Community Climate System Model version 3 (CCSM3) only includes the bottom-up coupled air-sea mechanism with coupled components of atmosphere, ocean, land and sea ice but without a resolved stratosphere or interactive ozone chemistry. Whereas, a version of the Whole Atmosphere Community Climate Model (WACCM), only includes the top-down mechanism. Here, solar variability is changed and the model is run with climatological SSTs keeping other external forcing constant. It has resolved a stratosphere

Table -6: Indicating whether the pathways are evidenced or hypothesised

	Observation	Mechanism	
		Explained	Hypothesised
Lower Stratosphere (A-H)	A,B,D,E,F,G,H	A,B,C,D,E,F,H	G,H
Atmosphere ----- (A-M) Upper Stratosphere (A, I-M)	A,I,J,K,L,M	A	
Atmosphere-ocean Coupling (N-S)	O,P,Q,R,S,G	R,N,O	P,G
Climate Change (T-Z)	T,U,V,W,X,Y,Z	T,U,V,X	Z

with fully interactive ozone chemistry, but no dynamically coupled air-sea interaction. Neither these two models, on their own were able to reproduce the observed pattern of temperature on the decadal solar cycle time scale. However, when a new hybrid model was developed, using dynamical ocean, land and ice modules from CCSM3 coupled with atmospheric component from WACCM, it produced a much closer result to the observations in the Pacific. According to them, each mechanism acting alone can produce a weaker signature, but when the two mechanisms, act together it produces a response in the tropical Pacific that agrees closely to their observations. It indicates that the combination of mechanisms is much more appropriate than the sum of their individual effects. However, there are several issues regarding validation of such findings.

First of all, their ‘bottom-up’ model is recorded to produce frequent ENSO events (IPCC, 2007) and it is not clear that after few runs such a tendency does not dominate.

Moreover, not all experiments are present in all models, making intercomparison of model results difficult, especially, in the absence of major statistical evidences.

Furthermore, their observational results for the solar signal, which are compared with models, are also questionable. They did compositing analysis for peak solar years, considering averages for DJF, with climatology subtracted - which is similar to that of vL07 and discussed in Chapter 3. However, in that chapter, we identified that results of the detected signal, using the method of compositing, may not only be sensitive to the choice of details of the compositing method, but also be strongly biased due to mixing up the signal with other strong variability, like ENSO. We showed that the peak years of solar sunspot cycle are usually associated with the negative phase of ENSO cycles and thus using the method of solar max compositing (as used by vL07 and Meehl et al., 2009), the ENSO signal in the tropics might be misinterpreted as a solar.

The Meehl et al. (2009) results do indicate, however, the importance of atmosphere-ocean coupling which needs to be considered carefully. Our study, based on data analysis and empirical evidence, mainly focused on the technique of multiple regression, which can segregate the solar signal from strong forcing like ENSO, OD and trend. Following the results of our data analysis, supported by evidence from other research, we propose a comprehensive overview including both the atmosphere and ocean, (mainly involving the Pacific Ocean) to account for the solar influences, eliminating the possible biasness of ENSO variability, which often contaminates results, as mentioned before. Such analysis also indicates that the solar influence in the troposphere is governed by both mechanisms, i.e, stratosphere driven as well as via Pacific. However, our proposed pathways, as shown by flow chart (Fig. 6) and discussed here above are different to that from Meehl et al. (2009) and might lead towards better understanding of atmosphere ocean coupling system, accounting for solar cyclic variability. This, no doubt, will be useful for improving our understanding regarding solar climate relationship.

6.2. Outstanding issues

Our analysis still indicates some areas that are unresolved. To mention a few:

- a) Interaction between major modes of climate variability: solar, QBO and ENSO act together with different time cycles. We are still not clear about the true nature of their linear/ non-linear interaction.
- b) Mechanism for solar influence on SSTs: The Sun during HS years (say above SSN 80) has been shown to influence tropical SSTs before 1958 and after 1997, though overwhelmed by the innate strong ENSO variability at lower solar activity (Fig. 5.3.2a). In a coupled atmosphere-ocean system, how the apparent influence is communicated/ reflected is still not fully understood.
- c) Climate change during 1950s-1997: Climate change probably induced the atmosphere-ocean coupling system to behave differently during this period and this also affects the solar signal. Such observations identify the need for quantifying the true solar signal with and without the influence of climate change. Hence, it introduces more complications into the task of characterization of solar influence under global warming.

Overall, we may state that, in spite of the outstanding issues mentioned, we are now in a better position towards understanding the atmosphere-ocean coupling system accounting solar cycle variability. Such analysis, identifying solar influence, not only leads towards better prediction skill but also illuminates scientific communities to address/ mitigate some of the crucial issues that are associated with climate change.

Chapter 7: Conclusions

7. Conclusions

The main purpose of this thesis is to understand the nature of 11 year solar cycle variability on SLP and SST over a 155 years period. We detect a solar climate link in surface climate and discuss our results in the context of others. Our work presents some observational evidences that may help to improve understanding of the decadal scale solar responses and the associated mechanisms in the context of ENSO variability. It suggests that the effect of solar variability in the troposphere is driven both by the stratosphere and by solar heating of the upper ocean; but that coupling is disturbed during the 2nd half of the last century, probably due to climate change.

Chapter 1 discusses the background to our work with mention of previous studies relating to solar-climate links. **Chapter 2** discusses the multiple regression technique and data that we employed. The major findings of our work are presented in Chapters 3-5 and consolidated in Chapter 6 in the context of a broad overview of the mechanisms involved in solar-climate links.

In **Chapter 3**, the analysis was mainly carried out using data sets covering the whole of 155 years. Two methods were employed to detect the solar signal in SLP and SST: the method of solar max compositing and multiple regression analysis. First, we discussed the solar signal derived using solar maximum compositing, and this was followed by the results from the regression analysis and an investigation into the sensitivity to the choice of solar index.

Using the method of solar max compositing, we detected a similar solar signal to vL07 and Meehl et al. (2007) in the Pacific region. For SLP, this was a strong positive signal around the AL, with a negative response (though not significant) north of the equator, covering the international date line. For SST, we also identified a strong La Niña-like pattern in the tropics, with a secondary positive signature around the mid-latitude, though weaker compared to vL07. We extended both the studies to the whole of globe and compared their results. Our analysis indicated that the results of using this method may

not only be biased due to mixing up the signal with other strong variability, like ENSO, but also be sensitive to the choice of details of the compositing method. Thus, our study pointed towards the lack of robustness of the signal detected using compositing method. We also observed that signals, using solar max compositing are almost unaffected by the inclusion or removal of a few solar cycles.

Data analysis, using either solar max compositing or regression analysis, suggests that a positive solar signal is present in the region of Northern Pacific for both SLP and SST. Such a result is consistent with White et al. (1997, 1998), Allan (2000), White and Tourre (2003), who also observed a quasi-decadal oscillation (QDO) of 9 to 13 years period in global patterns of SST and SLP, during the late nineteenth and twentieth century, that fluctuates in phase with the ~ 11 year period signal in the Sun's total irradiance.

Our results using multiple regression are different in the tropics to those from the compositing analysis for both SLP and SST. For SLP we find a broad reduction in pressure across the equatorial region but not the negative anomaly in the sub-tropics detected by vL07, using compositing method. For SST, using two sets of data (one from NOAA and another from the Hadley Centre), we are unable to detect a -ve solar signal for SST in the tropics, applying multiple regression technique. Thus our results are not in accordance with vL07, in relation to their proposed mechanisms, directly involving the Sun and tropical SSTs. We find that the peak years of solar sunspot cycle have usually been associated with the negative phase of ENSO cycles and thus using the method of solar max compositing (as used by vL07), ENSO signal in the tropics might be misinterpreted as a solar.

Furthermore, we observe that the peak of annual average SSN (as used by vL07) generally falls a year or more in advance of the broader maximum of the 11-year solar cycle and thus reflects a different (rising) phase of the solar cycle to the peak year of smoothed DSO, causing discrepancies between analyses. Analyses which incorporate data from all years, rather than selecting only those of peak SSN, represent more

coherently the difference between periods of high and low solar activity on these timescales.

Meehl et al. (2008, 2009) discuss a peak in irradiance at the peak of the decadal solar oscillation producing the La-Niña-like response which is lagged after a year or two by an El-Niño-like event and they suggest a mechanism. This difference in timing is crucial in considering any proposed mechanisms which involve forcing by solar irradiance. The SLP signal in mid-latitudes varies in phase with solar activity, and does not show the same modulation by ENSO phase as the tropical SST, suggesting that the solar influence here is not driven by coupled-atmosphere-ocean effects but possibly by the impact of changes in the stratosphere resulting in expansion of the Hadley cell and poleward shift of the sub-tropical jets (Haigh et al., 2005). Given that climate model results in terms of tropical Pacific SST can be dependent on different ENSO variability within the models (Meehl and Arblaster, 2009), our analyses indicate that the robustness of any proposed mechanism of the response to variations in solar irradiance needs to be analyzed in the context of ENSO variability where timing plays a crucial role.

To investigate the robustness of the identified signal, we applied our technique of regression using three different reconstructions of TSI viz., Krivova & Solanki, Foster and Hoyt & Schatten, in place of SSN. Such analysis, using different measures of solar activity, indicates that results are similar for the 11 year solar cyclic variability. Moreover, we showed that the Hoyt & Schatten solar index sometimes produces different results because the solar signal is mixed with the long-term trend.

Some studies have found that segregating the meteorological data based on the phase of QBO, is necessary to detect a clear signal of the 11-year solar cycle in polar stratosphere. Moreover, Baldwin et al. (2001) and Thompson et al. (2000) showed that perturbations in the polar stratosphere can affect the troposphere for few months in the form of polar annular modes. Hence, apart from the Sun, we also incorporated QBO in our analysis, which is discussed in **Chapter 4**.

Here our first attempt was to understand the main reason for the apparent inconsistency between two published results relating to the polar temperature, the Sun and the QBO. Using a slight different temporal and spatial coverage of winter polar temperature data, Labitzke et al. (1992) and Camp and Tung (2007) arrives at some contradictory findings: according to Labitzke et al. (1992), solar max/ QBO E-ly is cold, while Camp and Tung (2007), detects the same as very warm. Our analysis indicates that they appear different largely because of using a different QBO height. The QBO at 30 hPa (used by Camp and Tung (2007)) is seen to be out of phase with QBO 40 hPa (used by Labitzke et al. (1992)) on a number of occasions; that means, on occasions, QBO at 30hPa is seen to be easterly though westerly using height 40 or 50 hPa level and vice versa. Hence the use of different QBO height appears to be the most likely cause to influence the results of data analysis. However, interestingly, between 40 and 50 hPa, the QBO does not suffer much phase change, indicating that the results might not be very sensitive using QBO at either 40 or 50 hPa level.

Different analyses of Baldwin and co workers (1999, 2001, 2003, 2004) and Thompson et al. (2004) showed that fluctuations in the strength of stratospheric polar vortices in both hemispheres are observed to couple downward to the surface climate. The large-amplitude variations in strength of the stratospheric polar vortex are typically followed with a lag of less than one month by similar signed anomalies in the tropospheric circulation that persists for up to 2 months (in NH) to 3 months (in SH). Based on their results, we extended our analysis. To address this, we applied multiple regression technique on SLP, where the regression was done using all months of the year, with other independent parameters as trend and OD.

Following Baldwin and Dunkerton (2005) who speculated that the pathway involving the polar modes of variability, appears to involve interactions of the solar with QBO, we used a new index in our multiple regression analysis. Here in place of the solar index, we used a new index - the product of solar and QBO (and termed as 'solar*QBO'), to represent the collective behaviour involving both the solar and QBO as observed by Labitzke et al. (1992) and Labitzke (2004). Such index has also been used by Haigh and

Roscoe (2006) in their multiple regression analysis. Our study suggests that, the effect of the QBO (irrespective of the height 30, 40 or 50 hPa) combined with solar activity reveals a negative signal in the de-seasonalised NAM and SAM. That indicates HS years during the westerly phase of the QBO and LS years during the easterly phase of QBO, both trigger negative surface NAM and SAM features, in the zonal mean sense; whereas the reverse is true during HS years with the easterly QBO phase and LS years during the westerly QBO phase.

Similar analysis, combining the QBO and solar signals was carried with the new reconstruction of QBO by Brönnimann (2008) dating back from 1900, which suggested that the combined behaviour, as observed during the later half of the 20th century, still prevails during the earlier period in SH, irrespective of the QBO height (30,40 or 50 hPa). Moreover, in the regression, if we consider the whole of the 100 years period, we find that negative SAM features are quite evident (regardless of three different QBO height) and negative NAM features using height 30 or 40 hPa. Such observation, no doubt, can be useful for the purpose of long range surface climate prediction, considering decadal variability.

Some authors (Veechi and Soden, 2007; McPhaden and Zhang, 2004) have suggested that tropical circulations and shallow MOC around the tropical Pacific were different during period 1950s to 1997. Our empirical study suggests that when the Sun is more active, the solar-ENSO behavior is different during the intervening period. Such observations may have implications relating to the solar influence in the context of ENSO, tropical circulation, shallow MOC and climate change. These issues are discussed in **Chapter 5**.

Unlike the analysis for the whole 155 years of SLP and SST (as in Chapter 3), here in Chapter 5, we restrict our multiple regression analysis to two time periods: the earlier period comprises 1856-1957 and the later 1958-1997. Such segregation of periods, enables us to capture the modified form of the individual influences (mainly solar and

ENSO) during the later period, owing to the presence of anthropogenic effects (including GHG and ozone depleting substances), compared to that during the earlier period.

Solar signal and polar modes: First we studied solar signals on polar modes considering all months of the year. To capture the robust features (and the difference, if any) of the 11 year cyclic variability of the Sun in SLP, during the two periods, we used three different solar indices viz. SSN, Solanki and Foster's TSI. We found that the effect of solar activity (regardless of the choice of TSI reconstruction) before the 1950s, is to produce a clear positive signal in both the polar annular modes expressed in SLP. However, during the later period, the solar signal does not conform with the signal observed during the earlier period, using any of the solar indices and fails to capture any clear signal in the polar annular modes expressed in SLP. Using a new index combining both the solar and the QBO as discussed before (that follows Labitzke (2004)) well captures the behaviour of both the polar annular modes during the later period and also for the SAM during the earlier one.

ENSO, PDO and climate change: We also considered the signals of ENSO and PDO and their relevance to climate change. Our regression result suggests that ENSO is not the same during the later period as was observed earlier. It is observed that the signal of ENSO during DJF (where the mean value during DJF represent the year), is associated with a strong NAO pattern which is not the case during the earlier period.

Our current analysis also suggests that there is a strong influence of the ENSO on SLP around the Antarctic Peninsula, (e.g., a rise from 3 to 6 hPa during DJF of the later period). That indicates, change in the strength of ENSO, expressed in SLP, around that region, may have some role for recent warming in the Antarctic Peninsula, as observed in other studies (e.g., Thompson and Solomon (2002)).

It is also seen that amplitudes of SST variability associated with ENSO around the eastern Pacific have increased significantly (from 2.4° to 6.6°C) during the later period, relative to the earlier one. Amplitudes of the component of variability of SST associated

with the PDO around eastern part of the north Pacific have also increased (i.e., from 1° to 2.4°C around 40°N and 160°E) during the later period. Such changes in intensity of both the PDO and ENSO indicates that climate change may be responsible for modifying SST, substantially, around the tropical Pacific via ENSO and around the mid-latitude of North Pacific via PDO. These changes are likely related to anthropogenic influences, as discussed in detail by e.g., IPCC (2007).

Solar signal and ENSO: An underlying quasi-decadal variability in the inter-annual ENSO as detected in some recent studies (White et al., 2008; Zhang et al., 1997; Zhao et al., 2003; Chen et al., 2005) raises the question whether the Sun is having any influence on the ENSO. Adding an 11-year-period cosine signal of amplitude $\sim 2.0 \text{ W m}^{-2}$ to the solar constant in the fully coupled ocean-atmosphere general circulation model (i.e., Fast Ocean-Atmosphere Model (FOAM)) of Jacob et al. (2001), White et al. (2008) were able to simulate both the ENSO and the QDO; with their model QDO, similar in patterns and evolution with the observed one. On the other hand, in its absence (11 year solar signal), the FOAM can only simulate the ENSO. Our observation with regard to the solar-ENSO relationship in HS years ($\text{SSN} > 80$), puts more weight on this result and on the necessity to find a plausible mechanism for it.

Proposed Mechanisms - Earlier vs. Later Period: Our multiple regression analysis for DJF (mean value during DJF represents the year), during the earlier period, detected a weak yet significant solar signal in SLP around the tropical eastern Pacific, with a strong signal around the region of the AL. Such a signal around the tropics is missing during the later period. Moreover, the signal around the northern Pacific during this period is found to be weakened, though still significant.

Proposed Mechanism - Earlier Period: First we focus on the earlier period. A small but significant signal in the tropical eastern Pacific SLP, responsible for intensification of the ITCZ and associated enhancement of the trade wind, during HS years, can account for uplifting the thermocline at the eastern Pacific coast. It consequently can produce a situation similar to that of a cold event of the ENSO, during HS years and vice versa

during LS years. Thus inciting the trade wind around the ITCZ of the eastern Pacific, a chain of mechanism similar to that of the ENSO may be initiated via a solar forcing on SLP during DJF.

Our empirical observation captures this solar ENSO behaviour (during the earlier period), though only for HS years (say SSN >80). A possible explanation, supported by the work of Dima (2005) may shed some light on why that observation fails to indicate the reverse, during LS years.

Why the regression fails to capture any detectable solar signal in SST around the tropical eastern Pacific, without or with the ENSO as an independent parameter, during earlier period may be described as follows. As very few data points show such bias towards a cold event side of the ENSO during HS years, compared to the whole set of observation, the regression fails to indicate any strong result for the Sun and SST in the tropics. Furthermore, when we include the ENSO as an index in the regression, we still fail to capture any detectable solar signal in SST, suggesting again that it is the ENSO which is dominating SST in the eastern tropical Pacific most of the time. Thus, the apparent influence of the Sun on ENSO during the earlier period, may be initiated through triggering the trade wind, but is not detectable by the regression analysis in tropical SSTs.

In mid-latitudes, a strong solar signal in SLP, around the AL (~6 hPa) is detected. Such a signal can be related to a weakening of the Ferrel cell around the north Pacific during HS years and vice versa during LS years. The mid-latitude (north) and tropics in Pacific are tied through ocean pathways, and we find that the solar signal in the lower troposphere is also coupled with the ocean around those places, with the Sun being a potential driving factor to regulate the coupling mechanism.

Two fundamentally different routes for a solar influence on the troposphere have been proposed: one is the ‘top-down’ mechanism and the other the ‘bottom up’ one. First we address the ‘top-down’ solar influence which is generated through the stratosphere. In a GCM study (in absence of ocean), Haigh (1996, 1999) also detected a weakening of the

Ferrel cell, associated with enhancement of solar forcing in the lower stratosphere. Such a change should be reflected in the north Pacific sub tropical gyre, which is wind driven. Weakening of the subtropical gyre will impede overturning (thus mixing with cold water) in the north-eastern part of Pacific, augmenting temperature around that place. As we mentioned earlier that the Pacific is connected between the tropics and mid-latitude (north) via the shallow MOC. Thus the solar signal around the north Pacific can be transported to the tropical Pacific and vice versa. Through such a linkage, the Sun may influence not only the trade winds, but also the tropics via the mid-latitudes.

In the ‘bottom-up’ pathway, the Sun can directly influence SST without stratospheric feedback. Here we suggest that this pathway involves the shallow MOC and originates in the north Pacific. The shallow MOC, during HS years, absorbs more heat around the region of the north Pacific, and subsequently can cause weakening of the AL - which in turn, can reduce the strength of the Ferrel cell around the north Pacific. The heat absorbed can again be transported to the tropics of Pacific via the shallow conveyor belt.

Proposed Mechanism - Later Period: During the later period, the solar signal on SLP in DJF is missing around the tropics, with a weakened yet significant signal around the place of AL. We now consider how the solar cycle can have a different effect on ENSO during the later period. During the earlier period, the solar signal (during HS years) is responsible for warming in the north Pacific alongside cooling in the tropics. Such strong temperature gradient can be liable to enhance the rate of flow of shallow ocean current from the mid-latitude (north) to tropics in Pacific and consequently, can hasten equatorward convergence. However, said signal being missing around the tropics (and weakened in the mid-latitude), during the later period, indicates that the modified solar signal may have some role in decreasing the shallow overturning current during 1950s to 1998, that has been reported by several authors (Zhang et al., 2006, McPhaden et al., 2002, Vecchi and Soden, 2007). The presence of a weakened solar signal in the mid-latitude, during the said period may be related to the decadal signature still present in the shallow MOC.

Our multiple regression analysis also suggests that warming in the tropics is observed around the eastern Pacific during the later period, unlike the earlier one. It is also seen that the solar signal during the later period is mixed up with ENSO in the tropics. If the ENSO signal is not excluded from the solar, then our regression results for the later half of the 20th century suggest that the solar signal resembles that of the warm event of ENSO; which is not the case during the earlier period. Thus during 1958-1997, omission of ENSO from the regression gives a false warming (cooling) signal in SST related to higher (lower) solar activity (as observed by White et al. (1997)). We now discuss a plausible route for transporting such a solar signal and also the associated implications.

During the later period, the solar signal around the ITCZ is no longer present. However, weakening of the Ferrel cell during HS years subsequently causes warming around the north Pacific (alongside, oceanic shallow MOC around that place also absorbs more heat) and via shallow ocean pathway may cause warming in the tropical Pacific. Such a route of the solar signal, via transporting heat from the mid-latitude to tropics, may activate a warm ENSO cycle in the tropics, during HS years. Fig. 5.3.2b also indicates such a shift in the ENSO, towards a warm event side during the later period for HS years, in contrary to the observation of the earlier. Thus, during the later, the solar signal in the regression is shown to be mixed up directly with the ENSO, producing a comparable effect on tropical SST; however the same is nominal, if the ENSO is separated from the regression, as the ENSO is very strong in nature and dominates tropical SSTs.

Thus, the Sun which was shown could be a potential factor for triggering a cold event of ENSO at HS years, during the earlier period, may produce a contrary effect during the later. Such a bias of HS years on the warm ENSO side during the later period, compared to the earlier one, has also been captured in our scatter plots, though solar max years (used by vL07) are still found to be on the cold event side. Relating to solar ENSO behaviour, apart from indicating a plausible mechanism, our analyses are also able to reconcile some of the contradictory previous results.

Contradiction and Reconciliation: Our multiple regression study of the 1958-'97 data is able to explain some of the inconsistencies relating to the solar signal around the eastern tropical Pacific; for instance, the apparently contradictory results of vL07 and White et al. (1997), in addition to White and Liu (2008). Our regression analysis (during the time period, common to all the studies), accompanied by our observation that the year of peak annual SSN generally falls a year or more in advance of the maximum of the smoothed DSO, provide coherence to these apparently conflicting findings.

In Chapter 3, we indicated that the methodology adopted to detect the solar signal by vL07 is not characterising a true solar decadal variability. In fact, peak sunspot years generally fall a year or more in advance of the broader maximum of the 11-year solar cycle and thus reflects a different (rising) phase of the solar cycle to that of the peak year of smoothed DSO. We observed that the peak years of solar sunspot cycle are usually been associated with the negative phase of ENSO cycles. As the ENSO usually takes 1-2 years to change from cold to warm phase this explains why the results of vL07 differ to those of White et al. (1997). During 1958-1997, omission of ENSO from the regression gives a false warming (cooling) signal in response to higher (lower) solar activity in tropical SSTs. It clearly indicates why both White et al. (1997) and White and Liu (2008) not only detect warming in HS years, but also cooling in LS years, during the time period, common to all the studies. The trough and peak of ENSO shows a period of about 2 years due to its usual phase transition mechanism, throughout the solar cycle, as observed by White and Liu (2008); hence we find consistency between vL07, White et al. (1997) and White and Liu (2008) during the time period common to all the studies. Such observation also explains why solar max years (as used by vL07), are still on the cold event side during 1958-1997; which is not any special features of the rising phase of solar cycles – it is simply an artefact of ENSO.

Moreover, our study indicates that the composite of 9 solar cycles (during 1900-2000) as considered by White and Liu (2008) in their analysis, generally captures the behaviour of 1958-1997, the period having stronger solar cycles and strongly affected by the ENSO. It also suggests that the application of their approach to the period prior to 1958 would be

unable to detect similar phase locking of the ENSO and solar. Our analyses thus suggest that, mixing of ENSO with the solar signal during the later period might modulate the true solar signal which needs to be accounted appropriately. This might shed light on some of the apparent discrepancies between other results of the solar influence on surface temperature.

In **Chapter 6**, we have attempted to construct a chain of causalities involved in the coupled atmosphere–ocean system (mainly involving the Pacific Ocean) based on evidences and hypotheses, in a holistic context. We presented a flow chart (Fig. 6), depicting a consolidated overview of such coupling, supported by observations and mechanisms. Following the results of our data analysis (as described in Chapters 3-5), supported by evidence from other research, we proposed a comprehensive overview including both the atmosphere and ocean, to account for the solar influences, eliminating possible bias due to ENSO variability. The proposed atmosphere-ocean coupling process appears to be disturbed during the 2nd half of the last century, probably due to climate change.

The Meehl et al. (2009) results using the method of solar max compositing, also suggest atmosphere-ocean coupling and indicate that the solar influence in the troposphere is governed by both mechanisms, i.e., stratosphere driven as well as via Pacific. However, our analyses indicate that atmosphere-ocean coupling needs to be considered carefully eliminating the possible bias of ENSO variability, which can contaminate the results (Meehl, 2009). Our study, based on data analysis and empirical evidence can segregate the solar signal from strong forcing like ENSO, OD and trend, but also can detect features relating to the broader DSO. Thus, our proposed pathways, as shown by the flow chart (Fig. 6) might lead to better understanding of atmosphere ocean coupling, accounting the solar cycle variability. This, no doubt, will be useful for improving our understanding regarding solar climate relationship.

Based on our work we propose various **future studies** to further advance understanding of this complex problem.

1. Understanding more about the NAO and PDO, along with their role in atmosphere-ocean coupling processes: In the proposed atmosphere-ocean coupling processes, we mainly focused on the role of QBO and ENSO. How the NAO and PDO fit in the coupling processes, and how they might feature in the flowchart, would be interesting.
2. How ENSO is related to the NAO and PDO: Some connections between ENSO and both the NAO and PDO are mentioned in our analysis, but would be interesting to understand how the possible interactions are taking place, including the role of the Sun.
3. Testing the proposed hypotheses, as mentioned in chapter-6: To understand the proposed mechanisms, modelling studies must play an important role. The hypothesised mechanisms, need to be tested in models to provide a clearer overview of solar influences on atmosphere ocean coupling.
4. Including all the oceans: In considering the feedback of the impact of solar variability from the oceans, we mainly focused on the Walker circulation and the shallow MOC in Pacific. It is necessary to consider all the oceans to achieve a comprehensive overview regarding atmosphere-ocean coupling processes, accounting for the Sun. For example, the MOC in Atlantic Ocean are strongly influenced by climate change. Hence to quantify true decadal solar signature during the last half of the 20th century, the role of the Atlantic must be assessed. Moreover, in response to the changing behaviour of the Hadley circulation with regard to solar variability, the subtropical gyres and sub polar gyres all around the continents are likely to be affected. Additionally, apart from the Walker circulation in Pacific, how the East-West circulation pattern all around the tropics are modulated also needs to be addressed.

Overall our work leads towards a better understanding of the decadal scale solar variability and its impact on surface climate. Such analysis is, no doubt, useful to improve understanding relating to atmosphere-ocean coupling system, identifying the role of the Sun. This might not only serve to improve the prediction skill on decadal time scales, but also to help the scientific community to address/ mitigate some of the crucial issues associated with global responses to climate change.

Acronyms

8. Acronyms

AAO:	A ntarctic O scillation
AGCM:	A tmospheric G eneral C irculation M odel
AL:	A leutian L ow
AO:	A rctic O scillation
APCC:	A sian- P acific E conomic C ooperation C limate C enter .
CCMVal :	C hemistry- C limate M odel V alidation
CCSM3:	C ommunity C limate S ystem M odel version 3
CFC:	C hloro F luoro C arbon
CLiPAS:	C limate P rediction and its A pplication to S ociety
COA:	C entre O f A ction
CPC:	C limate P rediction C entre
DEMETER:	D evelopment of a E uropean M ulti-model E nsemble system for seasonal to in T ER-annual prediction
DJF:	D ecember- J anuary- F ebruary
DSO:	D ecadal S olar O scillation
ENSO:	E l N ino and S outhern O scillation
EOF:	E mpirical O rthogonal F unction
ERSST:	E xtended R econstructed S ea S urface T emperature
FDH:	F ixed D ynamical H eating
FOAM:	F orecasting O cean A ssimilation M odel
GCM:	G eneral C irculation M odel
GHG:	G reen H ouse G as
HADSLP:	H adley centre S ea L evel P ressure
HADSST:	H adley centre S ea S urface T emperature
HS:	H igher S olar
IOD:	I ndian O cean D ipole
ICOADS:	I nternational C omprehensive O cean- A tmosphere D ata S et
JJA:	J une- J uly- A ugust
LS:	L ower S olar
MJO:	M adden J ulian O scillation

MOC:	Meridional Overturning Circulation
MSLP:	Mean Sea Level Pressure
NAM:	Northern Annular Mode
NAO:	North Atlantic Oscillation
NCAR:	National Centre for Atmospheric Research
NCEP:	National Centres for Environmental Prediction
NH:	Northern Hemisphere
NOAA:	National Oceanic and Atmospheric Administration
OD:	Optical Depth
PC:	Principal Component
PCM:	Parallel Climate Model
PDO:	Pacific Decadal Oscillation
PH:	Pacific High
PW:	Planetary Wave
QBO:	Quasi-Biannual Oscillation
QDO:	Quasi Decadal Oscillation
SAM:	Southern Annular Mode
SH:	Sorthern Hemisphere
SIM:	Spectral Irradiance Monitor
SLP:	Sea Level Pressure
SO:	Southern Oscillation
SODA:	Simple Ocean Data Assimilation
SORCE:	Solar Radiation and Climate Experiment
SSN:	SunSpot Number
SST:	Sea Surface Temperature
SSW:	Stratospheric Sudden Warming
STC:	Subtropical Cell
SVD:	Singular Value Decomposition
TBO:	Tropospheric Biennial Oscillation
WACCM:	Whole Atmosphere Community Climate Model
WMO:	World Meteorological Organisation

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