Chapter 2

Major Modes of Variability

Abstract This chapter focused on major modes of variability which serve the key role in controlling the regional climate. In terms of tropospheric variability, it defined and discussed ENSO (El Niño Southern Oscillation), NAO (North Atlantic Oscillation), AO and AAO (Arctic Oscillation, Antarctic Oscillation), Indian Monsoon, Indian Ocean Dipole (IOD), PDO (Pacific Decadal Oscillation) and AMO (Atlantic Multidecadal Oscillation). Later it attended stratosphere variability; this constitutes QBO (quasi-biennial oscillation) and SSW (stratospheric sudden warming). Main characteristic features of each of these modes were elaborately discussed.

Keywords Modes of variability · ENSO · NAO · AO · AAO · PDO · AMO · Indian Summer Monsoon (ISM) · Indian Ocean Dipole (IOD) · QBO · SSW

2. Major Modes of Variability:

There are various modes of climate variability, which play important roles in determining the characteristic of different regions of the climate of the earth as shown in Fig.2.1

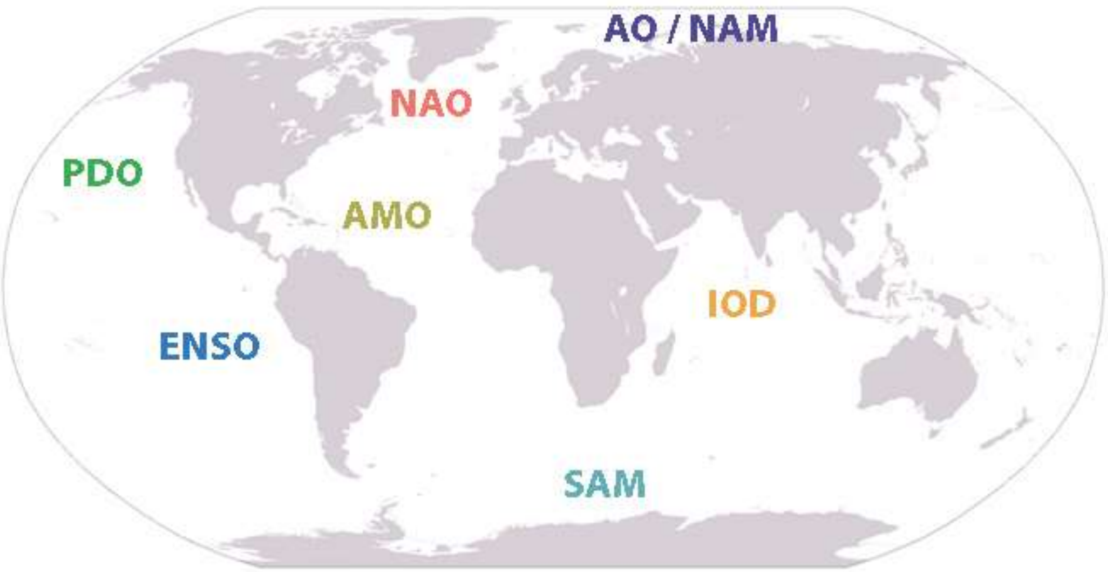


Fig. 2.1. Various modes of climate variability and the region of their influence.

Those are classified below as whether they are features of the troposphere or the stratosphere.

- Troposphere: ENSO (El Niño Southern Oscillation)
  NAO (North Atlantic Oscillation)
  AO and AAO (Arctic Oscillation, Antarctic Oscillation)
  PDO (Pacific Decadal Oscillation)
  AMO (Atlantic Multidecadal Oscillation)
- Stratosphere: QBO (Quasi-Biennial Oscillation)
  SSW (Stratospheric Sudden Warming)

2.1. Variability in the Troposphere:

- **El Niño and Southern Oscillation (ENSO)**

ENSO is the leading mode of variability in the tropics, although its influence is felt globally. The Southern Oscillation (SO) is a large-scale fluctuation in air pressure between the western and the eastern tropical Pacific about the international dateline. Tahiti (17.6^{o}S, 149.4^{o}W) and Darwin in Australia (12.3^{o}S, 130.5^{o}E) are two opposite ends of this SO's seesaw, and the SO index is calculated based on the difference in air pressure between these two places (shown in Fig. 2.2a).

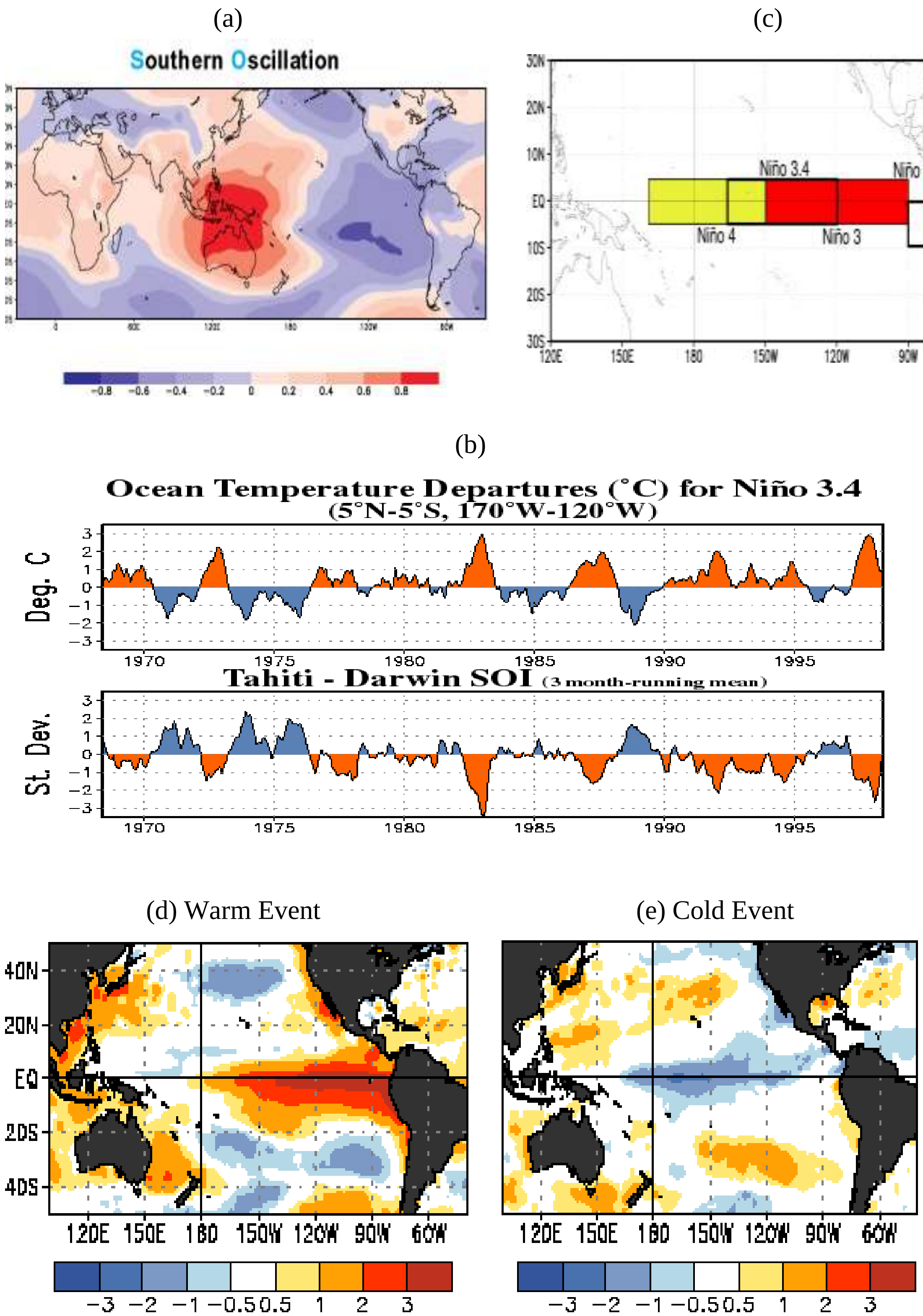


Fig. 2.2: Representation of ENSO: a) SLP variation (hPa) showing Southern Oscillation; b) Temperature of Niño 3.4 and Southern Oscillation anti-correlation; c) Different Niño regions; Departure of SST (° C) during DJF for Warm events of ENSO (d) and Cold events of ENSO (e).

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

The 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. 2.2b). 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 1+2, Niño 3, Niño 3.4 and Niño 4 as shown in Fig. 2.2c depicting their geographic coverage. Geographic coverage of various commonly used Niño Regions are as follows:

- Niño 2 and 1 : (0-10S, 90W-80W)
- Niño 3: (5N-5S, 150W-90W)
- Niño 3.4: (5N-5S, 170W-120W)
- Niño 4: (5N-5S, 160E-150W)

In the tropical Pacific, trade winds drive surface waters westward (also shown in Fig. 1.4 to describe the Walker circulation). Surface water that travels from the eastern Pacific all the way extended 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 the South American coast causing unusual warming of SST around the eastern seaboard of Pacific and thus El Niño (shown in orange colour, Fig. 2.2d). 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. 2.2e). During the La Niña, the trade wind and Walker cell both intensify; while during El Niño, they reverse direction.

Fig. 2.3 Sir Gilbert Walker (1868–1958).
(Source: https://pt.wikipedia.org/wiki/ Gilbert_Walker, link accessed on 10/12/2017)

Sir Gilbert Walker (photo: Fig. 2.3) made a pioneering discovery about Walker Circulation while he was the Director General of India Meteorological Department (IMD), in India (1904-1924). The Walker Circulation, which he first described, is named after him. He retired from Kolkata, IMD in 1924 and joined as a professor in Imperial College London. He also did some ground breaking work on climate variability in North Atlantic Region, which is known as the NAO.

- **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. It is a large-scale seesaw 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. It is usually measured as the pressure difference between Azore High and Icelandic Low (regions shown in Fig. 1.1, top).

The major features of the NAO are as follows, also shown in Fig. 2.4: 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 gradient generates frequent and stronger winter storms that advances 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 humid and mild winter.

Positive Phase Negative Phase

Fig. 2.4: Different phases of the NAO

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

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 mid-latitudes and polar region is seen in both hemispheres: for the northern hemisphere it is termed the Arctic Oscillation (AO); whereas, in 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 20º N, in the NH; whereas for the AAO, it is the same at 700 hPa level and measured pole-ward of 20º S in the SH. In SH, the different pressure level is used to alleviate partially the ambiguities introduced by the reduction of sea level over the high terrain of Antarctica. Both of their positive phases 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. 2.5.

a) b)

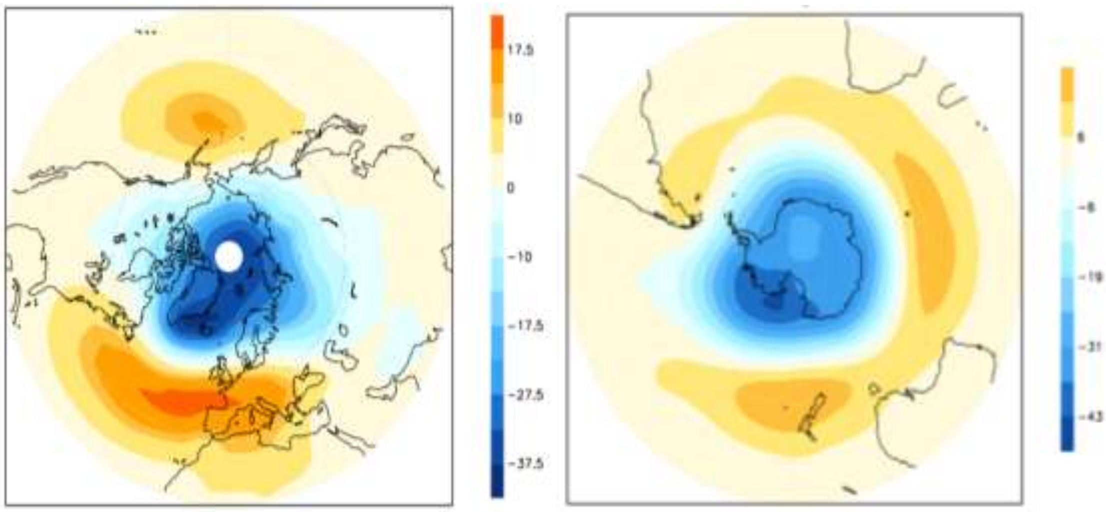


Fig. 2.5. The surface signature of positive AO (a) and AAO (b) measured in terms of geopotential height

[after, Thompson and Wallace, 2000]

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

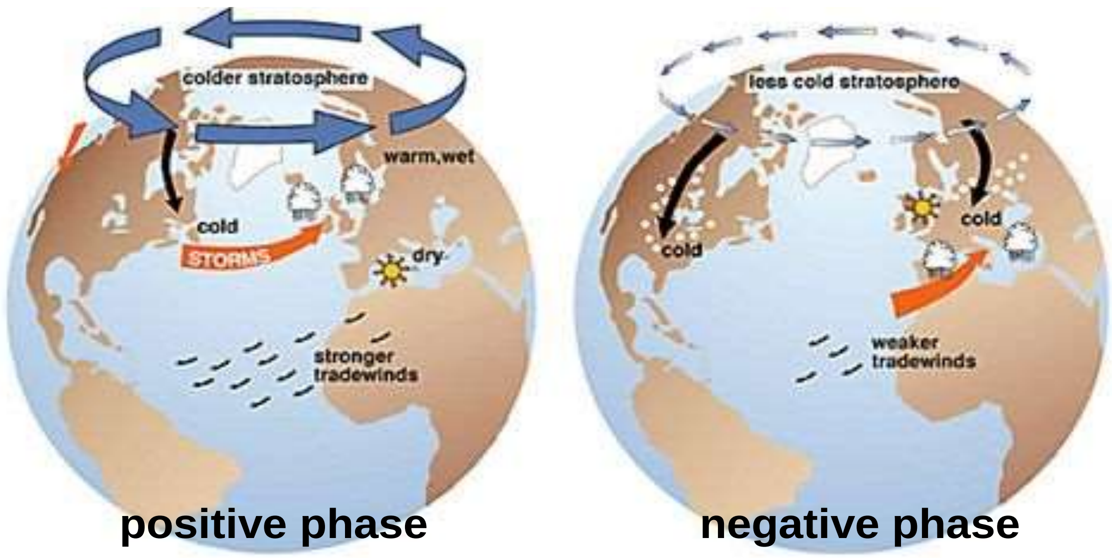


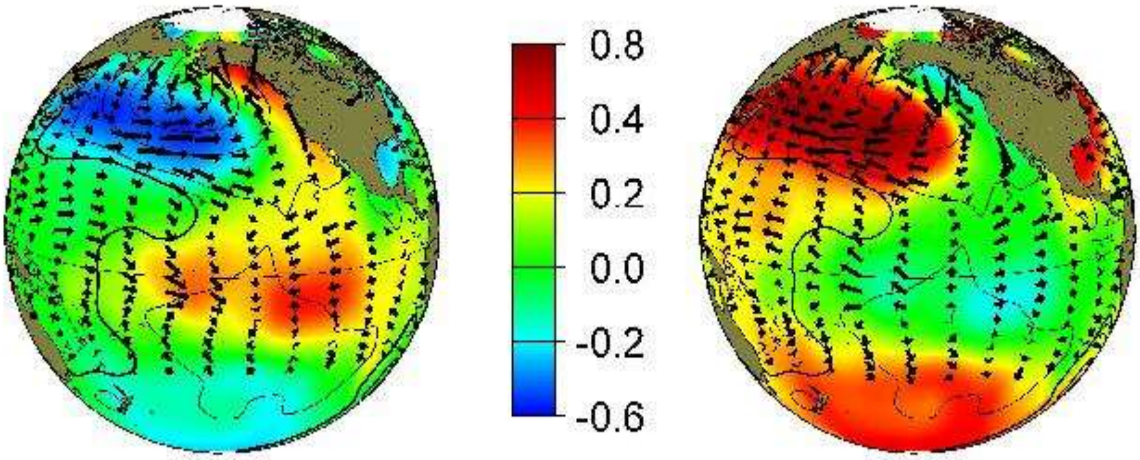
Fig. 2.6. Effects of different phases of the Arctic Oscillation [http://nsidc.org/arcticmet/patterns/arctic_oscillation.html, picture by Wallace, J.]

The band of upper-level winds that circulate the pole in the stratosphere generate the polar vortex (also shown in Fig. 2.6 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 region. The associated surface climate change with the phase of annular modes are described here with an illustration in Fig.2.6. In the positive phase of the surface NAM, the higher pressure at the mid-latitudes drives cyclones farther north toward the Arctic; that alters the pattern of circulation. It then brings wetter conditions to the Scandinavia and Iceland, Alaska; alongside it brings drier weather to the Mediterranean and the western United States. While the situation changes during the negative phase (Fig. 2.6). Moreover, in the positive phase of oscillation, cold air masses during winter do not enter as far into the Europe and North America as it would during the negative phase. It preserves much of Europe and the eastern 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.

- **Pacific Decadal Oscillation (PDO)**

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

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 2.7), which is the other way round for the ENSO (Fig. 2.2.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.



Warm Phase Cold Phase

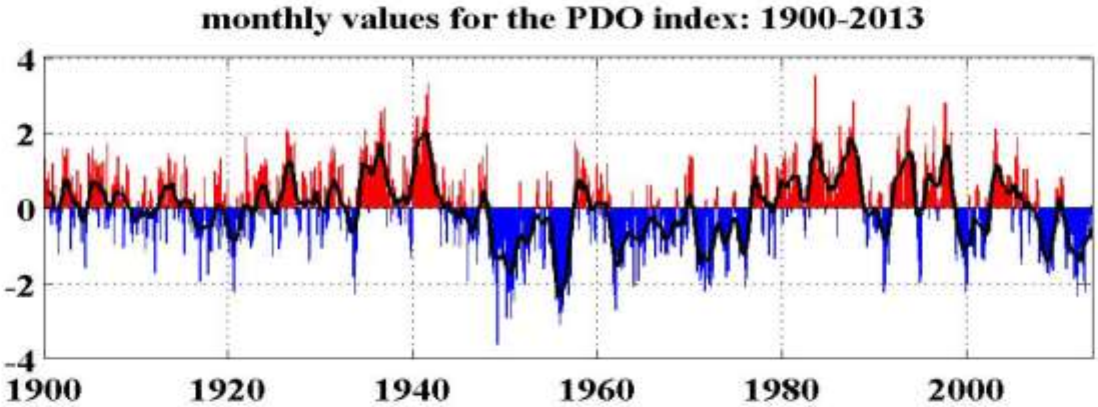


Fig. 2.7. Typical wintertime SST in ^{o}C (colours), surface wind stress (arrows) anomaly and SLP (contours) during warm and cold phases of the PDO (Top). Time series of the PDO are shown in bottom.

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

- **Atlantic Multidecadal Oscillation (AMO)**

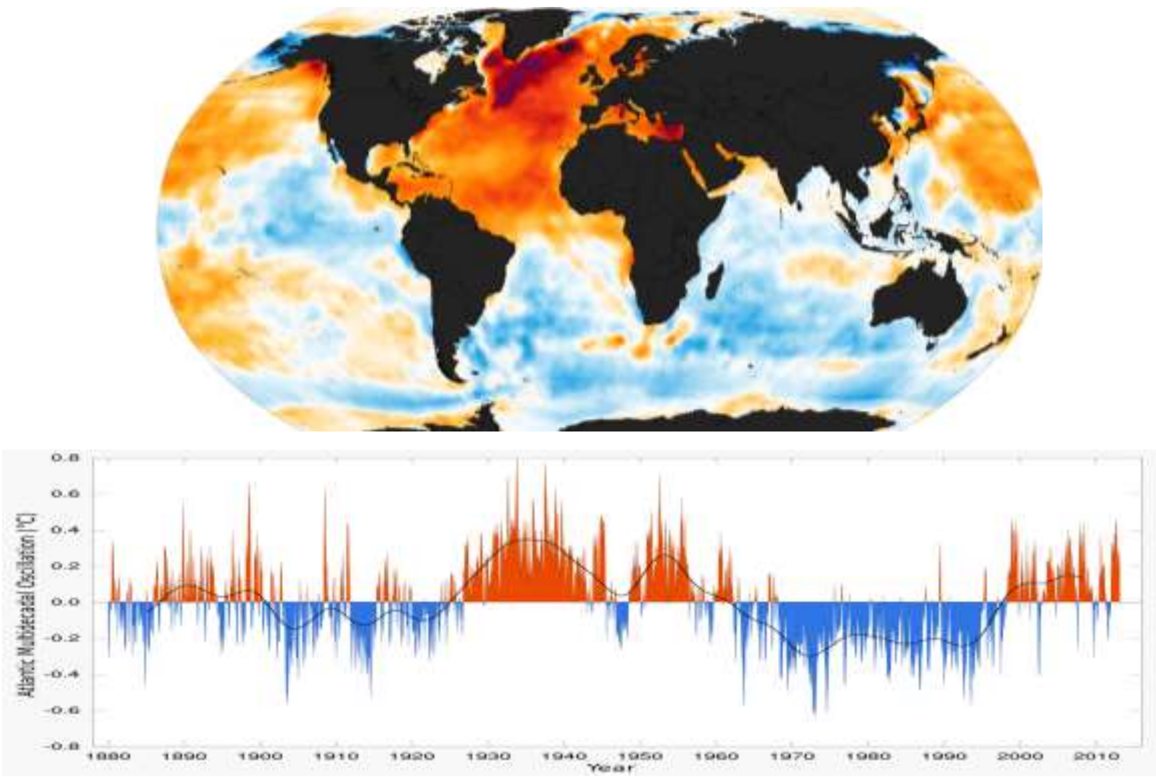


Fig. 2.8: The spatial pattern of the AMO shown at the top and temporal behaviour at the bottom. [https://en.wikipedia.org/wiki/Atlantic_multidecadal_oscillation]

The AMO is defined as the area average over the entire North Atlantic of the low pass filtered annual mean SST anomalies after removing linear trend (Fig 2.8). The time scale of the AMO is ~ 20-40 years.

Apart from those major variability, as discussed above, here are descriptions of two more important climate variability known as the Indian Summer Monsoon and the Indian Ocean Dipole.

- Indian Summer Monsoon (ISM):

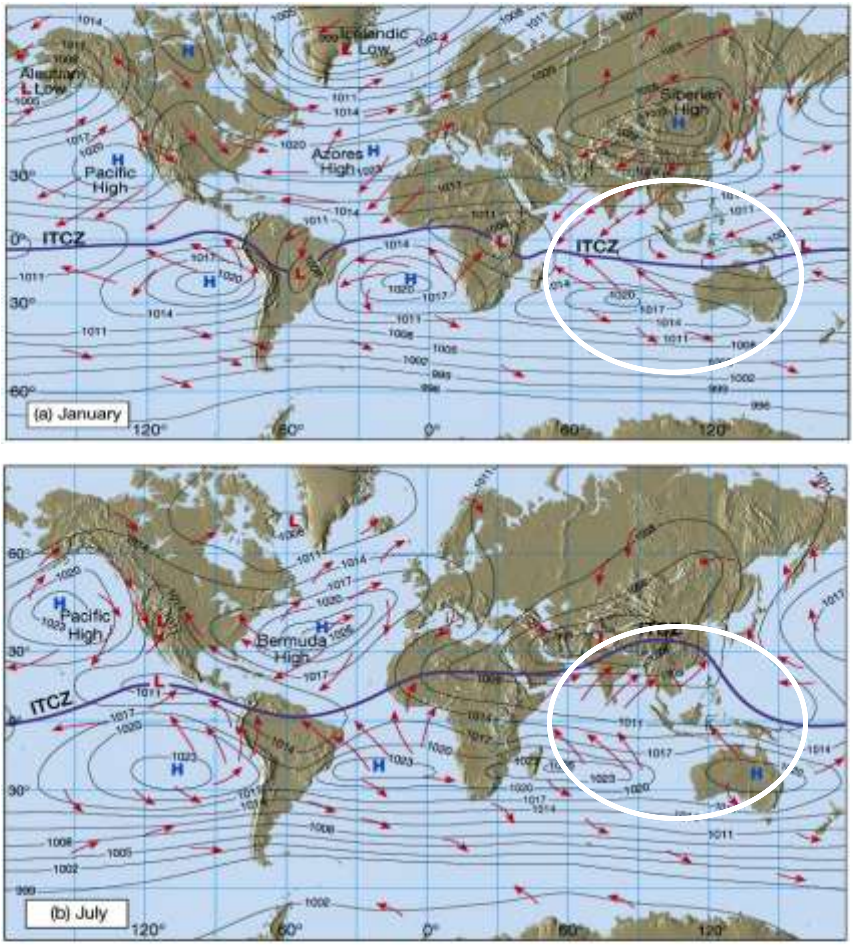
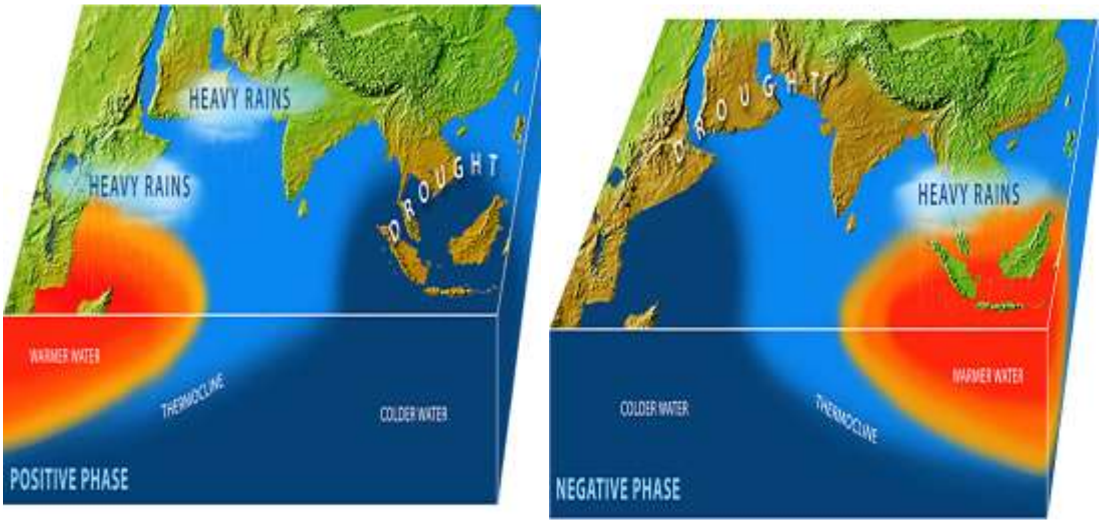


Fig. 2.9. The climatology of SLP during northern winter (top) and northern summer (bottom) is shown again (as in Fig 1.1) with particular emphasis on ISM region. A shift in the ITCZ and associated seasonal wind reversal is marked by white oval, covering Indian subcontinent.

Monsoon means seasonal wind reversals and for ISM, it is from North Easterly (NE-ly) during northern winter to South Westerly (SW-ly) during northern summer as shown in Fig 1.12. It is associated with the movement of Inter-Tropical Convergence Zone (ITCZ) which separates northern hemisphere to that from the southern hemisphere. Indian subcontinent experiences heavy rainfall during summer (June-July-Aug-Sept) as moisture rich air from ocean enters the land region. The Walker circulation and Hadley circulation both play a role in controlling ISM rainfall variability.

- **Indian Ocean Dipole (IOD)**



Positive Phase **Negative Phase**

Fig. 2.10. Different phases of the Indian Ocean Dipole. [http://www.whoi.edu/]

The IOD is a dipole behaviour in the Indian Ocean that regulates rainfall around Australia, East Africa and Indian subcontinent as shown in Fig. 2.10.

During the positive (negative) phase of the IOD, Darwin in Australia experiences drought (heavy rain) which are also associated with cold (warm) ocean water. On the other hand, in India or East Africa, there is heavy rainfall (drought).

2.2. Variability in Stratosphere:

While discussing the polar modes of variability, it is described that the troposphere and stratosphere are strongly coupled. Thus, for improving understanding of tropospheric variability, it is important to know about the variability in the stratosphere. There are two primary forms of variability in the stratosphere, viz. the **Q**uasi-**B**iennial **O**scillation (**QBO**) and the **S**tratospheric **S**udden **W**arming (**SSW**) and discussed here briefly.

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

It is an oscillation in the equatorial stratospheric wind, with zonal winds change between the east and west with a period of just greater than two years. The time series of equatorial deseasonalized zonal wind is shown in Fig. 2.11, where the colours in the

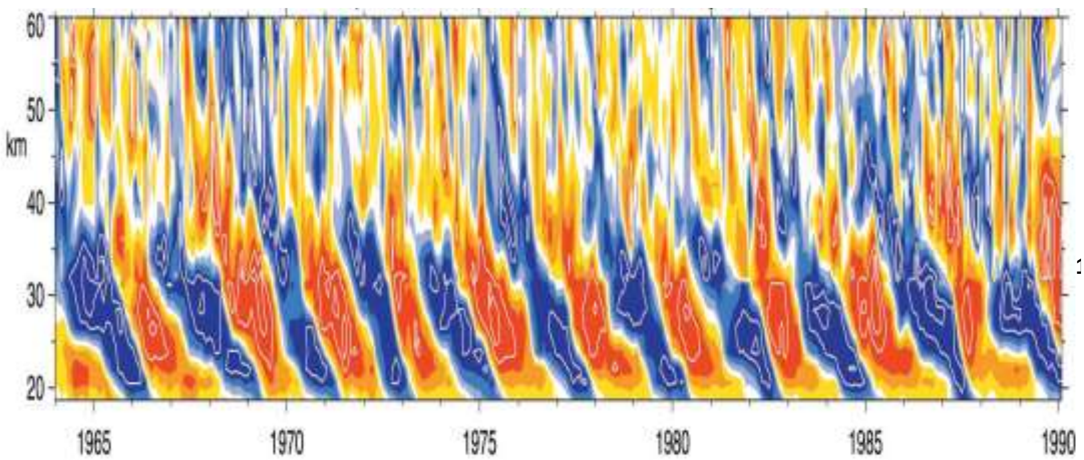


Fig. 2.11: Time-height section of the monthly-mean zonal wind components (m s^{-1}), with the seasonal cycle, removed, for 1964-1990. The contour interval is 6 m s^{-1}, with the band between -3 and +3 unshaded.

[Source: Baldwin et al. 2001]

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 around 20-36 months with a mean of around 28.

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 usually stronger than westerlies.

f) Easterly winds last longer at lower levels than westerlies; while the reverse is true at higher levels.

g) QBO shows considerable variability, in both amplitude and period.

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. 2.12, the formation of QBO is shown in a series of step (from (a) to (f)) and is described below. Black wavy line shows mean flow, 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 of red. Both the westerly and easterly maxima of mean flow are descending [Fig (a)]. Upward travelling waves are seen depositing their momentum just below the maxima. Viscous diffusion destroys the westerlies, when the shear zone of the mean westerly flow is sufficiently narrow. Upward propagating westerly waves can travel to high levels [Fig (b)] through the mean easterly flow. The more freely upward travelling westerlies lead to a new westerly regime of mean flow, as those dissipate at higher altitudes and produce a westerly acceleration [Fig (c)]. Figure (d) shows, both systems of the mean flow descending downwards until the easterly shear zone of the mean flow become sufficient narrow to destroy easterlies and upward propagating easterlies can then spread to high altitudes through westerly mean flow, Fig (e). Finally, it leads towards the formation of a new easterly regime of the mean flow shown in Fig (f); likewise, the same process (as described in Fig. (a) to (f)) continues.

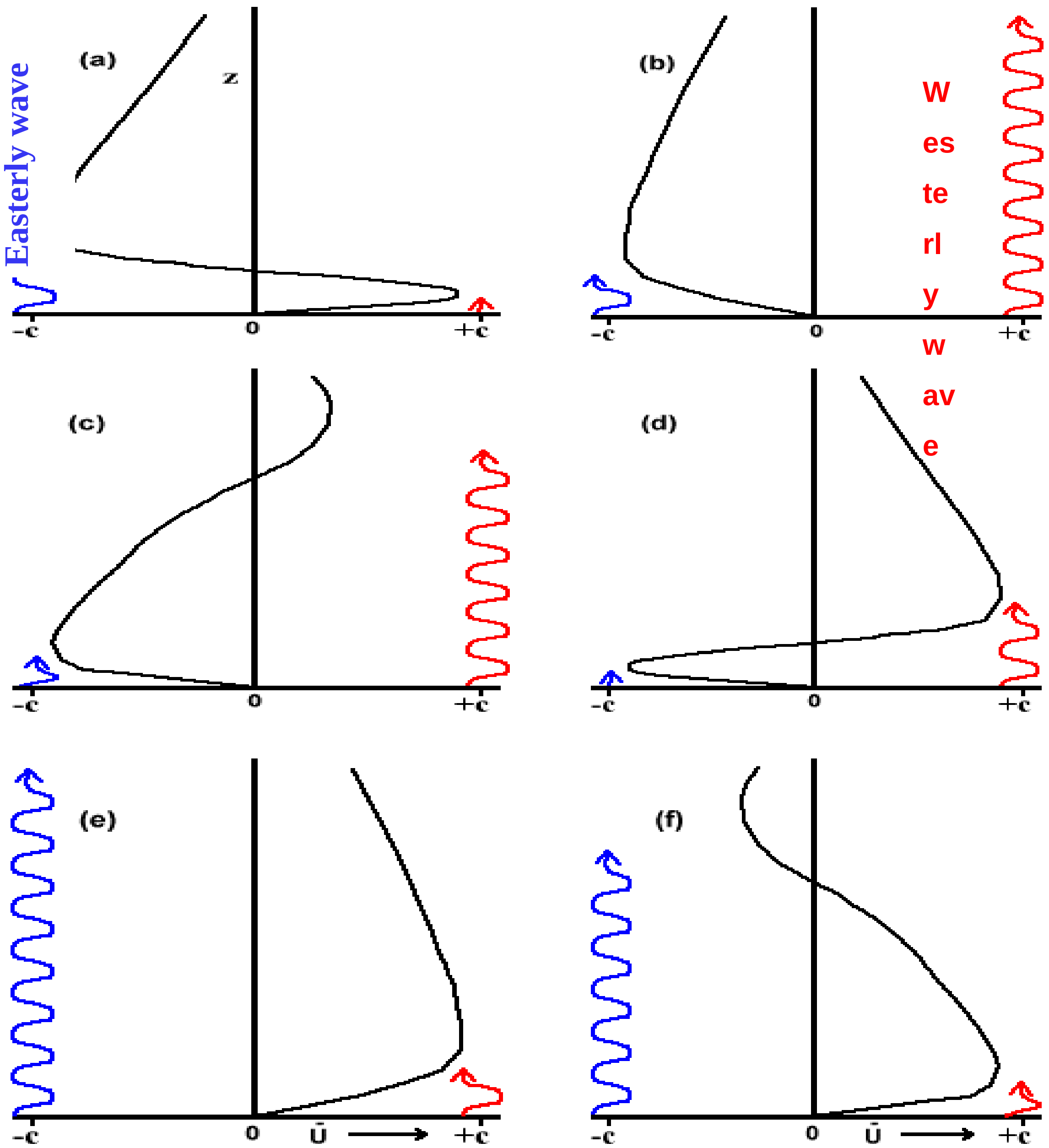


Fig. 2.12: Theory of QBO formation

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

The QBO is directly measured in operational wind measurements by radiosondes at equatorial meteorological observatories. Observations are taken within 2-degree latitude from the equator. A stratospheric research group at the Free University Berlin, Barabara Naujokat, has processed and collected radiosonde measurements from 1953 onward. It is from Canton Island, Gan (Maldives) and Singapore (Naujokat, 1986; Labitzke et al., 2002). The respective locations and available periods for these three observing stations are: 171°43'W/ 02°46'S, Jan.1953-Aug.1967; 73°09'E/ 00°41'S, Sept.1967-Dec.1975; 103°55' E/ 01°22' N, Jan.1976-Dec.2004. The QBO suggests 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 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.

Reconstructions of the QBO dating back to 1900 are now available (Brönnimann, 2007). It is based on the data from historical pilot balloon and the data of hourly SLP from Jakarta, Indonesia. The latter was considered to extract the solar semi-diurnal tide related signal. This is because in the middle atmosphere, it is modulated by the QBO. There is a good match with this reconstructions with the extracted QBO signal from historical total ozone data dating back to 1924. Moreover, the maximum phases of the QBO are also well captured after about 1910.

Wind and Temperature (Vertical structure)

Before discussing SSW, it is better to have some knowledge about the region of the polar vortex and the vertical distribution of wind and temperature profile as presented in Fig. 2.13. In the Troposphere, the temperature decreases with height up to Tropopause Region. This region is a boundary between the Troposphere and Stratosphere. It is 18 km around the tropics, through 8 km in Polar region. Due to an abundance of Ozone in the Stratosphere, the temperature increases there with height. It reaches maximum in the Stratopause area (50 km and 1 hPa level). It then decreases with height in Mesosphere, which is up to 80 km. Above 80 km, which is Thermosphere, temperature again increases with height. The polar region around the Stratopause, where polar vortex are formed in both the hemispheres is responsible for SSW as discussed below.

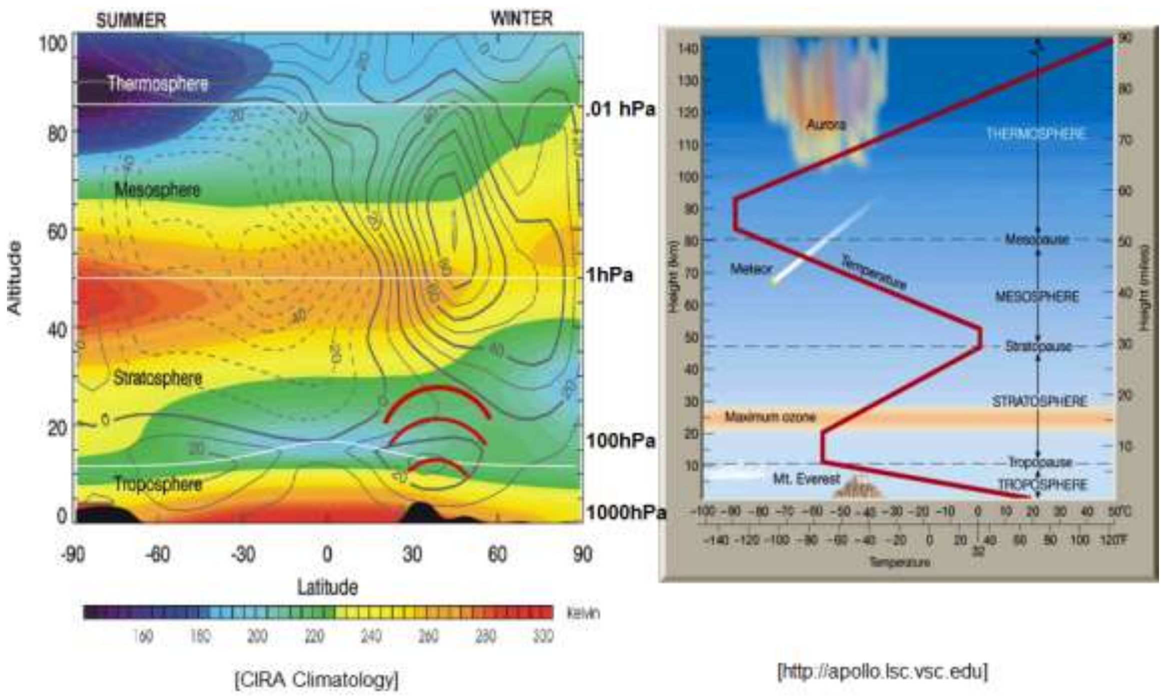


Fig. 2.13. The Wind and Temperature distribution in the top of atmosphere

- **Stratospheric Sudden Warming (SSW)**

It is the most dramatic event in the stratosphere during winter hemisphere, where the polar vortex of westerly winds, over the course of a few days, abruptly slows down or even reverses direction. it is also associated with a rise in the stratospheric temperature by several tens of Kelvins. Fig. 2.14 illustrates the latitude time series of average zonal temperatures around 30 km during two different periods, as shown by Gray (2004). The periods of sudden warming are marked by ‘X’ and ‘Y’ in Fig. 2.14a and b respectively, when temperatures were observed to rise abruptly.

SSWs can be classified into three broad categories:

- Major warming

The major warming occurs when westerly winds at 60ºN and 10 hPa reverse, i.e. become easterly from westerly. A thorough disruption of the polar vortex is noticed and

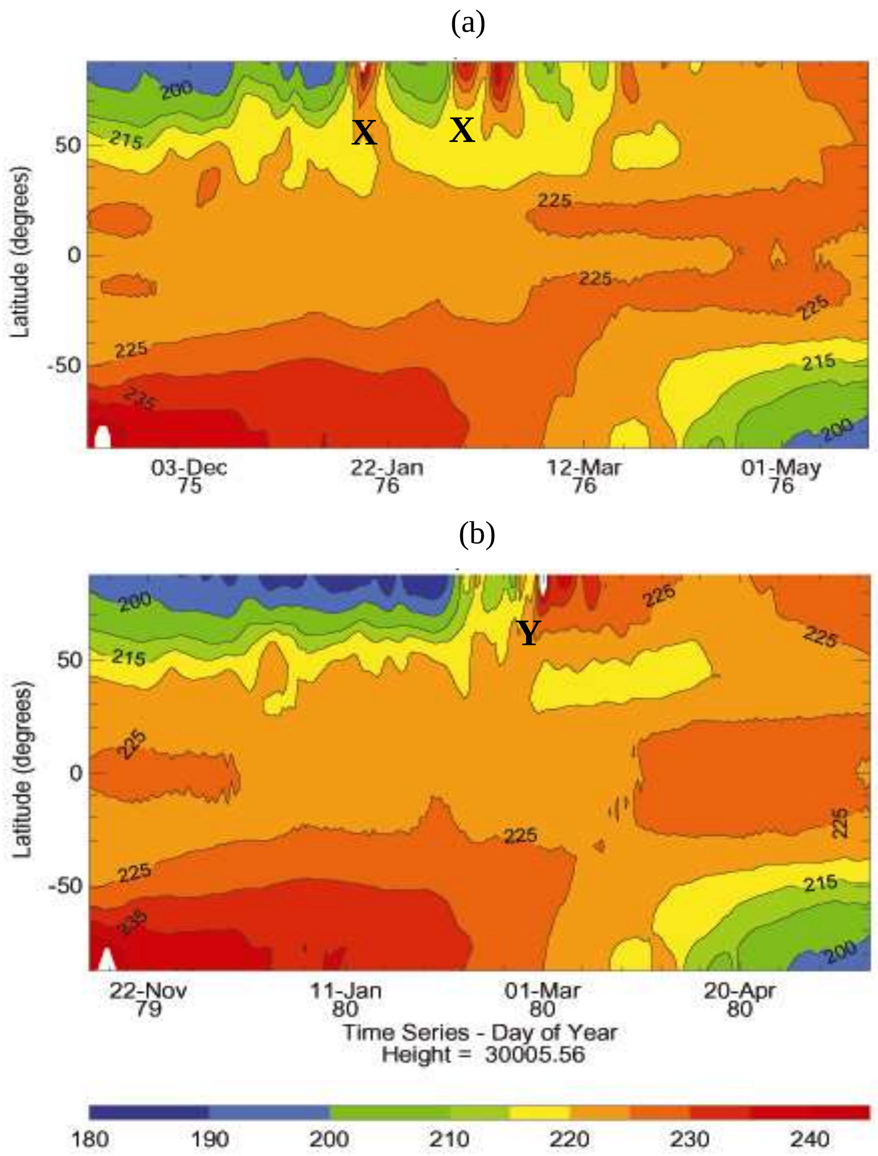


Fig. 2.14: Latitude-time series of average zonal temperatures around 30 km (http://lasp.colorado.edu/sorce/news/2004ScienceMeeting/SORCE%20WORKSHOP%202004/SESSION_2/2_5_Gray.pdf)

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, pole ward from 60^o 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.

Such warming can be shown in Fig. 2.15, where the time series of daily temperature (^oC) over the north poles 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. A grey line marks the mean of 30 years North polar temperature at 30 hPa level. From this picture, it is clear 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^o at 10 hPa level (shown with purple colour) and the warming was also accompanied by a reversal of the wind (i.e., from westerly to easterly which is not shown here), it may be classified as a major warming.

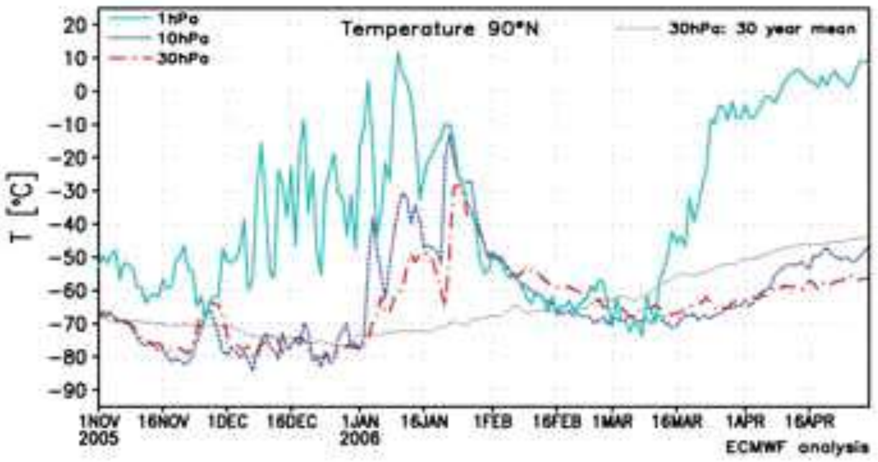


Fig.2.15. Time series plot of the daily temperatures (°C) over the North Pole during November 2005 till April 2006. Different pressure levels are marked with colours.

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

- *Minor warming*

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

a) A significant temperature rise which is at least 25 degrees in a period of week or less at any stratospheric level and
b) 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. 2.15b).

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