6.1 Introduction

Quasi-biennial oscillation (QBO) described by Reed et al. (1961), is a remarkable regular oscillation of the zonal wind between easterlies and westerlies in the equatorial stratosphere with a mean periodicity of about 28 to 29 months. During the one half of this period, easterlies propagate from upper stratosphere to lower stratosphere and during the other half they are replaced by westerly winds. The alternating wind regimes develop at the top of the lower stratosphere and propagate downwards at about 1 km per month until they are dissipated at the tropical tropopause. Westerly shear zones (in which westerly winds increase with height) descend more regularly and rapidly than easterly shear zones. The amplitude of the easterly phase is about twice as strong as that of the westerly phase. At the top of the vertical QBO domain, easterlies dominate, while at the bottom, westerlies are more likely to be found. The amplitude of \( \sim 20 \text{ ms}^{-1} \) is nearly constant from 5 to 40 hPa but decreases rapidly as the wind regimes descend below 50 hPa. (Baldwin et al., 2001).

Lindzen and Holton (1968) showed that QBO could be driven by a broad spectrum of vertically propagating gravity waves (including phase speeds in both westward and eastward directions) and that the oscillation arose through an internal mechanism involving a two-way feedback between the waves and the background flow. The first
part of the feedback is the effect of the background flow on the propagation of the waves (and hence on the momentum fluxes). The second part of the feedback is the effect of the momentum fluxes on the background flow. Holton and Lindzen (1972) assume random forcing by wave energy of eastward moving Kelvin waves and westward moving Rossby-gravity waves from upper troposphere, which interacts with the zonal mean flow in the lower stratosphere will give rise to QBO. But in actual case these forcing has strong seasonal dependence. This mechanism indicates linkage between the tropospheric disturbances and the QBO in the zonal winds of the lower stratosphere.

The QBO exhibits a clear signature in temperature, with pronounced signals in both tropics and extratropics. The tropical temperature QBO is in thermal wind balance with the vertical shear of the zonal winds (Andrews et al., 1987). Besides the equatorial maximum in QBO temperature, there is a coherent maxima over 20°–40° latitude in each hemisphere, which are out of phase with the tropical signal.

Many observational studies reported the presence of a similar quasi-biennial oscillation in many ocean-atmosphere parameters in the tropical region. These include tropospheric winds, temperature, Indian summer monsoon, tropical sea surface temperature (SST), southern oscillation etc (eg; Nicholls, 1978; Rasmusson et al., 1990; Ropeleswki et. al., 1992; Goswami, 1995; Terray 1995; Meehl, 1997). This biennial oscillation observed in the tropical regions, especially over the Indo-Pacific region is known as Tropospheric Biennial Oscillation (TBO). The mechanism of TBO is believed to be quite different from that of stratospheric QBO. Details of TBO and factors controlling TBO are described in the chapter 1 of the thesis. Coupled climate interactions between ocean and atmosphere contribute to a mechanism that produces biennial variability (TBO) in the troposphere and upper ocean in the tropical Indian and Pacific Ocean regions. Detailed explanation of TBO including the role of northwest Pacific warm pool in TBO is given Li et al. (2006).

Because the QBO has its maximum amplitude over the equator, it is natural to inquire whether this oscillation has any effect on the underlying tropical troposphere. But the
zonal wind and temperature anomalies of the QBO do not penetrate significantly below the tropopause. The temperature QBO at the tropopause is small relative to the annual cycle. It is known that the tropical troposphere has a quasi-biennial oscillation of its own, uncorrelated with the stratospheric QBO (Yasunari, 1985; Gutzler and Harrison, 1987; Kawamura, 1988; Lau and Sheu, 1988; Moron et al., 1995; Shen and Lau, 1995). Unlike the stratospheric QBO, the “tropospheric QBO” is irregular in time, asymmetric in longitude and propagates slowly eastward.

Yasunari (1989) suggested a possible link between biennial oscillations in the stratosphere and troposphere over the Asian monsoon region and SST in the equatorial Pacific using station data from Singapore and Pacific SST. Ropelewski et al. (1992) identified the association of stratospheric QBO with interannual variability of coupled air-sea system. Many studies have identified significant relationship between the phases of QBO in the zonal wind in the lower stratosphere (30 hPa) and percentage departure of monsoon rainfall of India. Mukherjee et al., (1985) showed that strong easterly (westerly) phase of the QBO is associated with weak (strong) monsoon. Meehl (1997) emphases the role of Asian summer monsoon on TBO. Sathiyamurthy and Mohanakumar (2000) related the TBO and QBO of zonal wind and temperature over an equatorial Indian station. Thus both these oscillation (QBO and TBO) are linked to the Indian summer monsoon separately. But no physical explanation is available so far for the possible linkage between stratospheric QBO and Tropospheric Biennial Oscillation. Thus both these oscillation (QBO and TBO) are linked to the Indian summer monsoon separately. Mohankumar and Pillai (2008) observed the unique structure of zonal wind over Indian monsoon region in TBO cycle.

6.2 Objectives of the study

The present study is an attempt to identify the characteristics of biennial variation of the stratosphere and troposphere and the possible interaction between these two areas over the Indian monsoon region and north Australian monsoon region, as these two monsoon regions are vigorously involved in TBO. The study further extends to find the
interaction of QBO and TBO associated with the annual cycle of Indian summer monsoon.

6.3 Data and methodology

In the present study zonal wind and temperature data for 23 vertical levels from 1000 hPa to 1 hPa for the period 1960-2002 obtained from European Center for Medium range Weather Forecast (ECMWF) have been used for the investigation of vertical structure of entire troposphere and lower stratosphere. A full description of ECMWF reanalysis (ERA) is available from Gibson et al., (1996, 1997). Indian summer monsoon rainfall (ISMR) data, which is the area averaged June to September rainfall of 306 stations well distributed over India, has been taken from Parthasarathy et al., (1994) and updated for making it to period 1960-2002 is used to define TBO years.

Indian monsoon area considered for the study is confined between the latitudes 10°N and 30°N and longitudes between 65°E and 95°E and Australian monsoon region 5°S and 15°S and longitudes 105° to 135°E. Gridded data of ECMWF is averaged over these areas considered for investigating the vertical structure. These data sets are filtered into biennial scale using a band pass filter developed by Murakami (1979). TBO years are identified using ISMR as done in the chapter 2 of the thesis. Then the vertical structure is studied for the composite of years satisfying the above criteria for ISMR. A Pearson cross correlation analysis is carried out in order to understand the speed of propagation of zonal wind and temperature anomalies and also their relationship with SST anomalies of Pacific and Indian Ocean SST in biennial scale. For sufficiently long data series of \( x \) and \( y \), the general lag \( -k \), Pearson cross-correlation coefficient between them is

\[
(r_{xy})_k \approx \frac{\sum_{i=1}^{n-k} (x_i - \bar{x})(y_{i+k} - \bar{y})}{\sqrt{\sum_{i=1}^{n-k} (x_i - \bar{x})^2 \sum_{i=k+1}^{n} (y_i - \bar{y})^2}}^{1/2}
\]

\[\text{…6.1}\]
6.4 Results

6.4.1 Wavelet analysis of zonal wind over Indian monsoon region

In order to identify the dominating frequencies of zonal wind, wavelet analysis is carried out for monthly mean zonal winds at different levels of troposphere and lower stratosphere. The Morlet wavelet is used and the transform is performed in Fourier space using the method described in Torrence and Compo (1998).

The power spectrum over Indian monsoon region is shown in figure 6.1. In the monsoon region annual oscillation is the major periodicity throughout the entire height ranging from 1000 hPa to 10 hPa. In the entire troposphere only the annual variation is present with maximum power at 100 hPa. Along with the annual oscillation QBO is also predominant in the lower stratosphere. Quasi-biennial periodicity starts at 70 hPa, though it is well below significant level and it becomes significant from 50 hPa onwards. The ratio of QBO to annual oscillation peaks at 30 hPa. The dominance of annual cycle in the troposphere winds over the Indian monsoon region is due to the effect of annually occurring southwest monsoon in this region.

![Wavelet power spectrum of monthly mean zonal wind at different levels over Indian monsoon region. Dotted line indicates 95% significance level.](image)
Thus in order to study the troposphere-stratosphere interaction in the biennial scale, the prominent annual frequency must be removed from the data.

### 6.4.2 Zonal wind structure in QBO scale

The time series of QBO filtered zonal wind over India and Australian monsoon region are presented here.

#### 6.4.2.1 Indian monsoon region

The time series of the vertical profile of zonal wind over the Indian monsoon region is depicted in figure 6.2. In the lower stratosphere, zonal wind propagates downward. The downward propagation of zonal wind anomalies weakens on reaching the tropopause level, but it extends to the lower troposphere. In some years, the stratospheric maximum extends to the troposphere, especially from 1980’s. In both the troposphere and stratosphere the zonal wind anomalies reverses their direction year by year indicating a biennial oscillation.

![Figure 6.2: Time-height plot of zonal wind over Indian monsoon region. Height is in pressure units](image)

#### 6.4.2.2 Australian monsoon region

The time series from 1960-2000 over Australian monsoon region is given in figure 6.3. Over Australian monsoon region the lower stratosphere has easterly/westerly structure extending upto 70 hPa and then weakens and anomalies are larger than over Indian monsoon region. Another maximum anomaly region is found in the surface, which also
Chapter 6

QBO-TBO interaction

alters year by year and it propagates upwards and anomalies are very weak in the middle and upper troposphere.

Figure 6.3: Time-height plot of zonal wind over Australian monsoon region. Height is in pressure units

Thus in both these monsoon regions QBO propagation is almost similar, but is more intense over Australian monsoon region. The interaction with troposphere is different in both the regions. A surface maximum is seen over the Australian monsoon region.

6.4.3 Propagation of zonal wind anomalies in QBO scale

Pearson cross correlation analysis is carried out for both Indian and Australian monsoon region with 20 hPa wind and that at different levels in order to understand the time taken to reach the effect of lower stratospheric winds at different levels. Similar correlation analysis is carried out for tropospheric winds at selected levels with that at levels just below that to get the interaction of zonal winds at troposphere.

6.4.3.1 Indian monsoon region

20 hPa wind has lead correlation of 1 month with that at 30 hPa (correlation coefficient is about 0.97) ie, it takes one month to reach 30 hPa level and about 6 months to reach 50 hPa (cc is 0.83). It will reach at 70 hPa by another two months. It has lead correlation of 12 months with 100 hPa (0.31). Thus zonal wind from 20 hPa reaches tropopause level by about a year time period. Correlation with heights from 200 to 500 hPa is less than 0.2 always. It has a positive lead correlation of 4 months (0.3) with zonal wind at
700 hPa and another 16 month negative correlation (-0.3) also. Similarly with 850 also has +0.3 at 3 months lead and -0.3 with 11 months lead. Correlation is less than 0.2 for 925 and surface with 20 hPa wind leading the lower level by about 5 months.

In order to examine clearly the propagation of anomalies in troposphere we have made similar correlation for tropospheric winds with that at just lower levels and is seen in figure 6.5. Zonal wind spreads from 100 hpa to 200 hpa (max correlation 0.94 is with zero lag) and from 200 to 300 also it spreads with zero lag(0.99) . From 300 to 500 also it spreads (0.94). From 500 hpa to 700 hpa it takes 2 months (0.7) and takes one month to reach 850 hpa (0.61). It again spreads to 925 hpa(0.98) and from there to
surface (0.99). Thus the wind spreads quickly from 100 hPa to 500 hPa (ie, zero lag) and from there to 700 hPa it moves slowly taking 2 months to reach 700 hPa and then spreads to surface.

6.4.3.2 Australian monsoon region

Similar cross correlation analysis is carried out for the zonal wind over Australian monsoon region and the maximum correlation value and lead/lag of 20 hPa with lower levels is given in table 6.1. Here zonal wind at 20 hPa reaches 30 hPa by 2 months (0.99) and with another five months it will reach at 50 hPa (0.97). With in a period of nine months wind from 20 hPa will reach at 70 hPa and it spreads quickly to upper troposphere. Thus zonal wind from 20 hPa will reach the tropopause level three months earlier than that at Indian monsoon region. Correlation pattern reverses in the upper and middle troposphere level here. This reversal of correlation indicates upward propagation of anomalies from troposphere. 20 hPa zonal wind has a lag of 2 months with that of 200 hPa wind (0.3) and the lag is 4 month with 300 hPa. Correlation is negligible upto 850 hPa and surface has a lag correlation of 4 months (-0.26) with surface wind. This

Figure 6.5: cross correlation of tropospheric winds with that at level below that
means that westerly (easterly) phase of surface wind leads the easterly (westerly) phase of 20 hPa by four months. Thus in both these monsoon areas, downward movement of stratospheric winds are different in lower stratosphere and propagation is opposite in troposphere.

Table 6.1: cross correlation of 20 hPa zonal wind over Australia with that at lower levels. Months with + sign indicates 20 hPa wind leads the other and negative indicate lag.

<table>
<thead>
<tr>
<th>U20/levels (hPa)</th>
<th>Lag/lead (months)</th>
<th>Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>50 hPa</td>
<td>+7</td>
<td>0.97</td>
</tr>
<tr>
<td>100 hPa</td>
<td>+9</td>
<td>0.4</td>
</tr>
<tr>
<td>500 hPa</td>
<td>+24</td>
<td>0.22</td>
</tr>
<tr>
<td>850 hPa</td>
<td>-3</td>
<td>-0.22</td>
</tr>
<tr>
<td>1000 hPa</td>
<td>-3</td>
<td>-0.26</td>
</tr>
</tbody>
</table>

Table 6.2: cross correlation of zonal winds of different levels at troposphere with that just below.

<table>
<thead>
<tr>
<th>Levels</th>
<th>Lag/lead (months)</th>
<th>Correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>100 hPa with 200 hPa</td>
<td>0</td>
<td>0.65</td>
</tr>
<tr>
<td>200 hPa with 300 hPa</td>
<td>-1</td>
<td>0.87</td>
</tr>
<tr>
<td>300 hPa with 500 hPa</td>
<td>-1</td>
<td>0.6</td>
</tr>
<tr>
<td>500 hPa with 850 hPa</td>
<td>-1</td>
<td>0.82</td>
</tr>
<tr>
<td>850 hPa with 925 hPa</td>
<td>0</td>
<td>0.98</td>
</tr>
<tr>
<td>925 hPa with 1000 hPa</td>
<td>0</td>
<td>0.99</td>
</tr>
</tbody>
</table>

Over the Australian monsoon region, zonal wind at 100 hPa has zero lag with that at 200 hPa. But 200 hPa wind has one month lag with 300 hPa and this one month lag is present upto 700 hPa wind. Lower troposphere wind has no lag with that at lower levels as seen in table 6.2.
6.4.4 Vertical profile of zonal wind during the TBO years

In order to investigate the stratosphere-troposphere interaction associated with biennial oscillation of ISMR, composite analysis of strong minus weak TBO year, which are defined earlier, is performed from previous year to next year of a strong/weak TBO year.

6.4.4.1 Indian monsoon region

Figure 6.6 is the vertical structure of zonal wind anomalies over Indian monsoon region during TBO cycle. In the Indian summer monsoon region, the stratosphere has downward motion from upper to lower levels and it reverses its sign by the next year. Instead of dissipating at the upper troposphere, the anomalies propagate downwards to troposphere. But the propagation speed varies at different levels. It spreads between lower and middle troposphere and then propagates slowly to 700 hPa level by about nine months, with speed of 0.3 km/month and then extends to lower level quickly. The westerly anomalies seen over the upper stratosphere during the weak monsoon propagates to lower level by the next strong monsoon and easterly anomalies from the stratosphere come to the upper troposphere.

The zonal wind pattern shows three distinct layers in the troposphere. The upper region (above 500 hPa), where zonal wind extends quickly, then a transition region, which lies between 700 and 500 hPa and the bottom region, from 700 hPa to surface. Easterly anomalies prevail in the bottom region during the weak monsoon year. In the upper region westerlies are seen till the spring season before the positive phase of TBO and are replaced by easterlies and continue to the next year spring. The transition layer slowly transits the upper region to the bottom region over the monsoon area, which takes a period of about six months. The existence of the transition zone is quite unique in the monsoon area.

Another interesting fact noted over the monsoon area is that the zonal wind structure remains the same phase in the entire troposphere during the winter season (November-February), which changes its phase in the next year. In figure 6.6, it can be seen that the previous winter season of an active monsoon year, the entire troposphere from surface
to the tropopause level exhibits westerly anomalies. This westerly phase is completely replaced by easterly phase in the next year of a strong monsoon year. Clear downward propagation of easterlies and westerlies from the stratosphere to the troposphere is seen over the monsoon region.

As reported by the earlier observational studies, it is evident in the fig.6.6 that westerlies in the lower troposphere and easterlies in the upper troposphere is an indicator of a strong monsoon year. In a weak monsoon year, TBO has its easterly phase in the lower troposphere and westerly component in the upper troposphere. This biennial tendency has a strong influence in the monsoon circulation. When the TBO exhibits westerly in the lower troposphere it activates the low level jet (LLJ) circulation, thereby intensifying the monsoon circulation. Similarly the easterly anomalies in the upper level enhances the speed of the tropical easterly jet stream (TEJ), which is present about 14 km height over the monsoon region. The intensity and location of TEJ is very much related with the monsoon activity. When TBO is in the opposite phase, the easterlies in the lower troposphere reduce the LLJ and westerlies in the upper troposphere oppose TEJ and both these effects weaken the monsoon circulation.

It can be also seen that the stratospheric QBO has got direct link with the TBO during the monsoon season. Westerly anomalies in the lower stratosphere are transferred to the upper troposphere during the previous year of an active monsoon year. The westerly wind regime is then slowly and steadily transferred to the lower troposphere by the transition layer. In the case of a weak monsoon year, we can see similar transport of the easterly regime from the lower stratosphere to the upper troposphere and then to the lower troposphere. This characteristic supports the role of stratospheric winds on troposphere in biennial scale.

6.4.4.2 Over Australian monsoon region

In the Australian monsoon region also easterly anomalies are seen at the lower stratosphere during the previous year of strong TBO year and it propagates downwards and reaches tropopause level (100 hPa) within a year. At the same time the upper and middle troposphere has westerly anomalies and lower level has easterlies from July
Chapter 6

QBO-TBO interaction

onwards. The easterlies formed propagate upward to upper troposphere by next year April and interacts with downward propagating QBO easterlies at the upper troposphere. At the same time westerlies are formed in the lower stratosphere and propagates downward and westerlies forms in the lower stratosphere by next June, which is strong TBO year and next cycle with reveres anomalies starts.

Figure 6.6: Time –height plot of zonal wind anomalies over the Indian monsoon region (averaged over 10°N-30°N, 65°E-95°E) for strong minus weak TBO years composites, from previous year to next year of reference monsoon. Months with 0 correspond to previous year, with 1 to current TBO year and 2 to the next year.

Figure 6.7: Time –height plot of zonal wind anomalies over the Australian monsoon region (averaged over 5°S-15°S, 105°E-135°E) for strong minus weak TBO year composites.
Chapter 6  

QBO-TBO interaction

Thus during a strong (weak) TBO year Australian monsoon season (December to February) westerlies (easterlies) are seen in the lower troposphere and stratosphere and propagates to upper troposphere and interacts there.

6.4.5 Zonal temperature anomalies in TBO cycle

As the zonal wind over these regions showed difference in structure in order to assess the dynamic features of these regions, we analyzed the temperature structure in these regions by similar composite analysis.

6.4.5.1 Indian monsoon region

Over the Indian monsoon region, the lower troposphere levels have negative temperature anomalies in the previous year upto April month with maximum cooling anomalies over the troposphere region between 500 hPa to 200 hPa (see figure 6.8). But in the upper troposphere the negative anomalies are seen upto June. Positive anomalies sets in the lower troposphere along with the onset of weak monsoon and propagates to midtroposphere and attains maximum in between 500 and 200 hPa by next spring season before the strong monsoon. Cool temperature anomalies are seen in the lower stratosphere at this time. The pattern reverses with the onset of strong monsoon with cooling in the troposphere. Thus in the Indian monsoon region, we have temperature maximum in the mid-troposphere and at lower stratosphere of almost similar intensity as seen in figure 6.8.

6.4.5.2 Australian monsoon region

Vertical temperature in TBO scale over Australian monsoon region is shown in figure 6.9. In the Australian monsoon region cold anomalies in the lower stratosphere in the weak TBO year propagate downward and reach 100 hPa by next year boreal summer monsoon. At this time warm anomalies are seen in the troposphere (figure 6.9). Then the negative anomalies propagate downwards to lower troposphere. Surface was cool from the January itself and it mixes with upward moving negative anomalies below 850 hPa. This anomaly maximum in the lower level is the difference of Australian monsoon pattern from the equatorial pattern.
Chapter 6

QBO-TBO interaction

Figure 6.8: Time–height plot of zonal temperature anomalies over the Indian monsoon region (averaged over 10°N-30°N, 65°E-95°E) for strong minus weak TBO years composites.

Figure 6.9: Time–height plot of zonal temperature anomalies over the Australian monsoon region (averaged over 10°N-30°N, 105°E-135°E) for strong minus weak TBO years composites, from previous year to next year of reference monsoon. Months with 0 correspond to previous year, with 1 to current TBO year and 2 to the next year.

6.4.6 Relationship between zonal wind over India and Australia with sea surface temperature in TBO cycle

Figure 6.10 below shows the time series of wind at different levels over India and Australia and SST anomalies for strong minus weak composites of TBO cycle from
Chapter 6

QBO-TBO interaction

Previous year (months with −1) to the next year (months with +1) of strong TBO year (months with 0). The pattern over Indian monsoon region is given in figure 6.10a and b. In lower levels (1000 hPa and 850 hPa) of the troposphere, wind anomaly over Indian monsoon region becomes westerly by boreal winter and attains maximum with strong monsoon and become easterly in the next winter season. Middle troposphere (500 hPa) pattern is opposite to the lower level pattern with positive maximum during weak monsoon and it becomes negative along with the spring season before the strong monsoon and negative maximum is after the onset of strong monsoon. 100 hPa wind becomes negative by winter and maximum value is during July and again becomes positive by next march. At 20 hPa, zonal wind anomaly becomes westerly after the weak monsoon and attains maximum by April and reverses sign by next winter.

Over the Australian monsoon region, surface winds have easterly anomalies in the previous year and maximum easterly is in the previous year winter and becomes westerly by September and westerly anomaly maximum is at the time of strong Australian monsoon (January of next year). 850 hPa also has similar pattern. Mid-troposphere level also has similar pattern, but delayed by four months (figure 6.10c). 100 hPa has opposite pattern of surface anomalies. 20 hPa has westerlies from the previous year September of strong TBO year and attains maximum during the spring season of strong TBO year and starts reversing at the time of onset of strong monsoon over Australia.

SST anomalies in the north Indian Ocean (0-25°N, 60°-100°E) becomes positive before the weak monsoon and attains its maximum value in the winter season before the strong monsoon and continues to spring and reverses sign by the onset of the strong monsoon and completes the next half in next one year (see figure 6.10d). The south Indian Ocean (0-15°S, 85°-100°E) warms after the weak monsoon attaining maximum warming in the following spring season and cools in the post monsoon season. The Nino3.4 region (10°S-10°N, 170°-120°W ) SST anomaly remains positive from the beginning of previous year and attains maximum in the October month and reverses by the onset of monsoon and remains cool till the end of next weak monsoon.
Chapter 6  QBO-TBO interaction

Figure 6.10: Strong minus weak composite of a) zonal wind over Indian monsoon region, b) over Australian monsoon region and d) SST over north Indian Ocean, south Indian Ocean and equatorial east Pacific. Months with (-1) denotes previous year, 0 current TBO year and +1 next year.

The cross relationship of zonal wind over India and SSTs of both Indian and Pacific Ocean regions are analyzed by Pearson cross correlation analysis of zonal wind at different levels and SSTs of north Indian Ocean (NIO), Southeast Indian Ocean (SEIO) and east equatorial Indian Ocean (Nino3.4) and is shown in figure 6.11. Zonal wind at
20 hPa has insignificant correlation with NIO SST. In the case of 100 hPa wind, the cross correlation is maximum with a lead of 3 months and for 500 hPa it is four month. That is zonal wind at 100 hPa and 500 hPa becomes westerly/easterly a season ahead of north Indian Ocean warming starts. Correlation is insignificant in the surface levels also. Zonal wind at 20 hPa has 1 month lead over SEIO SST (0.28) and the lead increases to two months (0.65) for 100 hPa and is again reduced to one month for midtropsphere level zonal wind. Surface wind has two month lead over SEIO SST. Zonal wind at 20 hPa has lag correlation of 6 months (-0.45) with nino3.4 SST. 100 hPa has lead correlation of 1 month (0.45) and 500 hPa has 2 month lead (0.66). 700 hPa has 1 month lag (-0.64). Lower levels have negligible correlations (<0.15).

Figure 6.11: cross correlation of a) north Indian Ocean SST, b) Southeast Indian Ocean SST and c) Nino3.4 region SST anomalies with zonal wind at different levels over Indian monsoon region.
Chapter 6  

QBO-TBO interaction

It is well known that the autumn season SST anomalies over the southeast Indian Ocean (SEIO) and east Pacific SST (Nino3.4) anomalies are important for Australian monsoon in TBO cycle. So the cross correlation of zonal wind over Australian monsoon region is analyzed with SEIO and Nino3.4 SST anomalies in QBO scale. Lower troposphere winds has a lag of 11 months (0.65) and lead of 14 months with SEIO SST anomalies and lag reduces to 5 months (0.55), for 500 hPa and it has a lead of 15 months also (0.5). 100 hPa has 11 month lag with SEIO SST (0.35). Lower stratosphere has negligible correlation with SEIO SST anomalies. Surface wind has 3 month lag (-0.63) with Nino3.4 SST (both are in opposite phase) and is four month for 850 hPa (-0.71). Midtroposphere wind leads Nino3.4 SST by four months (0.64). 100 hPa has one month and 20 hPa has seven month lag with Nino3.4 SST anomalies.

Figure 6.12: cross correlation of a) Southeast Indian Ocean SST and b) Nino3.4 region SST anomalies with zonal wind at different levels over Australian monsoon region.
6.4.7 Effect of QBO on ISMR in biennial time scale

Peculiarities of zonal wind and temperature structure over Indian monsoon are observed in earlier sections. For zonal wind the QBO scale maximum is in lower stratosphere, while for temperature maximum is between 500 hPa and 200 hPa. The effect if these maxima in Indian summer monsoon are analysed with the help of correlation analysis between ISMR and maximum region U wind and temperature time series.

6.4.7.1 Lower stratospheric winds and ISMR

Lag-lead correlation analysis has been carried out between the zonal wind at 30 hPa (u30) and 50 hPa (u50) with ISMR index on biennial time scale. Figure 6.13 shows the correlation from January of the previous year to the next year December of a monsoon season. A significant negative lag correlation exists between u50 over India and ISMR with maximum value of -0.5 at the previous December. It is interesting to note that the correlation is negligible during the early phase of monsoon and the positive correlation dominates in the later months of the monsoon and attains the maximum in the early month of next year. Similar analysis is also carried out between u30 hPa and ISMR. In this case the correlation becomes positive in the previous winter season before the monsoon and the simultaneous correlation becomes highly significant (0.6). The maximum lag correlation is in the previous year monsoon.

It is also noted that the anomaly of the summer monsoon is very much associated with the phase of QBO at 50 hPa during the previous winter season. During the winter season, before an active monsoon year, the QBO is in its easterly phase at 50 hPa level and in opposite phase for weak monsoon years. The association becomes very strong during the TBO years of Indian summer monsoon. The zonal wind anomaly at 30 hPa level also shows a strong association to the summer monsoon activity over the Indian subcontinent. It appears that the lower stratosphere interact with tropospheric circulation, especially over the monsoon region.
6.4.7.2 Upper troposphere temperature and Indian summer monsoon rainfall

A Lag-lead correlation of the average temperature of maximum area (500 hPa to 200 hPa) with ISMR index is carried out on the biennial scale for 1950-2002 TBO years. Figure 6.16 shows the correlation from the previous year (months with −1) January to next year (months with +1) December of strong TBO year (months with 0). The correlation becomes positive after the weak monsoon and attains positive during the March before the strong monsoon it peaks (0.73) and then reverses after the strong monsoon. From the figure 6.8 we saw that the temperature at this level has maximum positive value in the spring season before the strong monsoon and negative maximum in the spring season of weak monsoon. The correlation maximum is also this time. Thus the temperature maximum at this region has prominent relationship with monsoon rainfall over India.
6.5 Discussion

Both the QBO scale time series and cross correlation analysis indicates the downward propagation of zonal wind from the lower stratosphere to tropopause over Indian and Australian monsoon region. It takes one year to reach tropopause from 20 hPa level over Indian monsoon region and only ten months over Australian monsoon region. Over the Indian monsoon region, the stratosphere winds propagate to the troposphere, though weakened at tropopause level. In TBO cycle this pattern is very clear with three different layers in the troposphere. In the Australian monsoon region pattern is different with downward propagation in the lower stratosphere and upward propagation in the lower troposphere and interaction is at the tropopause level. Temperature pattern over the Indian monsoon region is also peculiar with additional source (sink) in the middle and upper troposphere region, which strengthens in the spring season. The spring season anomaly is important for the TBO mechanism.

In TBO cycle, 20 hPa westerly maximum is at spring season before monsoon and 1000 hPa maximum is during summer, with a three month lead for 20 hPa as observed in Pearson cross correlation. 20 hPa wind has insignificant correlation with North Indian Ocean SST, but 100 and 500 hPa leads north Indian Ocean SST by a season. 20 hPa zonal wind leads southeast Indian Ocean SST by one month and equatorial east Pacific SST leads stratosphere zonal wind by six months. Thus warm SST in the equatorial east Pacific facilitates the zonal wind at 20 hPa six months later. This 20 hPa zonal wind affects southeast Indian Ocean SST a month after. Southeast Indian Ocean SST, which attained maximum warming in the March season before the strong Indian summer monsoon influences the surface Australian summer monsoon wind in the beginning of next calendar year. The relationship of SEIO SST reverses at tropopause level. At the same time equatorial east Pacific SST has 3 month lead over Australian monsoon surface winds with correlation of \(-0.63\). The negative correlation indicates that cool SST in the east Pacific affects the westerlies over Australian monsoon region, a season earlier. The 20 hPa wind over Indian and Australian monsoon region has similar pattern over both Indian and Australian monsoon region indicating the absence of southeast
movement in the stratosphere, which is major property of TBO in the troposphere. By the above mechanism QBO scale zonal winds and TBO scale SST are correlated with a cyclic processes in the Indo-Australian monsoon region and topical Indo-Pacific Oceans.

6.6 Conclusion

The QBO in lower stratosphere wind and temperature is evident in both Indian and Australian monsoon region for TBO cycle. The downward propagating stratospheric zonal winds penetrate deep into the troposphere over the Indian monsoon region, while downward propagating stratospheric zonal winds and upward propagating winds from the lower troposphere meet in the upper troposphere over Australian monsoon region. Propagation of zonal wind anomalies over the Indian monsoon region at different levels is with different speed. An additional source (sink) of temperature is present in between the middle and upper troposphere over Indian monsoon region a season before strong (weak) monsoon. Zonal winds over Indian ant middle and upper troposphere have lead correlation with north Indian Ocean SST, while 20 hPa wind leads southeast Indian Ocean SST. Southeast Indian ocean SST affects the surface wind of Australian summer monsoon season. The east Pacific SST leads lower stratosphere zonal wind of both Indian and Australian monsoon region. Thus QBO scale zonal wind over Indian and Australian monsoon region is closely related to tropical Indian and Pacific Ocean SST anomalies in TBO scale.