# Chapter 2 Literature Review

## 2.1 The Quasi-Biennial Oscillation

### 2.1.1 Observed Features

Evidence of a downward propagating annual wind reversal in the lower equatorial stratosphere was observed in the early 60's (Reed et al., 1961; Veryard and Ebdon, 1961).

The lower stratospheric zonally symmetric easterly and westerly wind regimes alternate regularly with periods ranging from 24 to 30 months (Holton, 1979). Figure 2.1 shows this spectacular alternating easterly and westerly wind pattern generated by data obtained in the lower stratosphere above the Malaysian region. There were a total of 17.5 oscillations in the period of 41 years from January 1958 to December 1998 giving rise to a mean period of 28.1 months. The fastest observed oscillation had a period of about 20 months while the slowest had a period of about 35 months.

Successive regimes first appear above 30 km (~10 hPa), but propagate downward at a rate of 1 km/month. The downward propagation occurs without loss of amplitude between 30 and 23 km (~10 and 35 hPa), but there is rapid attenuation below the 23-km level (Holton, 1979). Below the 23-km level, the westerly wind regimes propagate downward without much variation, however there is a stalling of the descent of the easterly wind regimes (Kinnersley and Pawson, 1996). Due to this stalling effect, the time between maximum easterlies and maximum westerlies is much shorter than the opposite sequence (Naujokat, 1986). In both wind regimes the velocity decreases as height decreases. The amplitude of the oscillation is between 40 and 50 m/s. The maximum amplitude occurs at the equator at around 27 km (~20 hPa). However, the amplitudes as well as the length of the periods vary considerably. The easterlies are generally stronger than the westerlies, but the westerlies propagate downward faster than the easterlies. At the higher levels the easterly phase lasts longer than the westerly phase, and conversely at the lower levels.

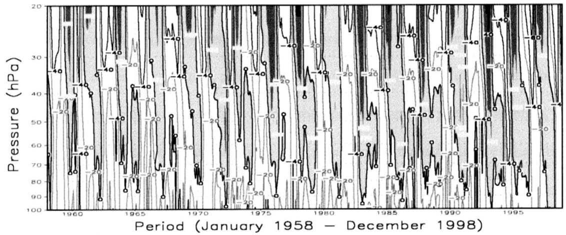


**Figure 2.1** Vertical time section of monthly mean zonal wind above the Malaysian region showing the QBO in the lower stratosphere. The westerlies are shaded. Contours are drawn at 20 m/s.

### 2.1.2 Theory of the QBO

The earliest theoretical explanation of the QBO were generally based on thermal forcing by variations of solar ultraviolet radiation caused by a small hitherto neglected sunspot maximum with a period of 25 months (Shapiro and Ward, 1962). Later, it was suggested that the QBO was a forced geostrophic wave resulting from downward propagation of heat supplied by a solar source of ultraviolet radiation with a period of about 26 months (Staley, 1963). Another theory suggested that the QBO was a subharmonic response to the annual heating cycle (Reed, 1964). Both Staley (1963) and Reed (1964) presented theoretical models to support their suggestions. However, neither of the models was able to account for the downward propagation of the oscillation without loss of amplitude.

In the theories based on thermal forcing, the zonal wind field is coupled to the heating field through a system of zonally symmetric circulation cells driven by differential heating producing a local change in the zonal wind by advecting absolute angular momentum in the meridional plane. In a rotating frame of reference the local changes in angular momentum arise mainly due to Coriolis torque produced by the mean meridional circulation. However, this mechanism is not very effective near the equator, where the Coriolis parameter is small. In addition, the mean meridional circulation is incapable of producing any sizeable westerly momentum near the equator. Therefore, the theories of thermal forcing are unable to fully describe the main features of the QBO. Fluctuations in the horizontal or meridional momentum fluxes were then suggested as a possible momentum source for the oscillation (Tucker, 1964; Wallace and Newell, 1966). Tucker (1964) found evidence of long-term fluctuations in the meridional flux divergence of zonal momentum at the 25-km level in the tropical stratosphere that seemed to be related to the QBO. Wallace and Newell (1966) also indicated a biennial periodicity in horizontal eddy momentum fluxes in the middle latitude above the 30-hPa level. However, the evidence for these fluxes in the tropical latitudes was not very convincing, and the amplitudes of such variations were too small to account for the QBO. Moreover, this type of forcing does not explain how an oscillation in momentum fluxes could generate the downward propagation of the QBO.

Later, a completely different theoretical explanation emerged. New theory shows that vertically upward-propagating, internal, short-period planetary-scale waves in the equatorial upper troposphere play an important role in generating the QBO in the lower stratosphere (Lindzen and Holton, 1968). The vertical flux of momentum due to a spectrum of gravity waves is deposited in the mean flow at critical levels where the waves are absorbed by the mean flow. The region above the critical level at which the absorption takes place is shielded from the action of the waves by a zone of strong wind shear that is produced by the absorption. As a result, the absorption will then take place at a lower level and the shear zone will propagate downward. When the shear zone reaches the tropopause, the shielding effect disappears, so the waves can propagate upward again for another round of interaction.

With a better understanding of gravity waves and their interaction with the mean flow, Holton and Lindzen (1972) had formulated an updated theory. Evidence now indicates that these waves consist primarily in a single westerly Kelvin wave during the descending westerly phase of the QBO. And in a single easterly mixed Rossby-gravity wave during the descending easterly phase, although the mixed Rossby-gravity wave forcing alone is not sufficient in driving the QBO. The assumption of absorption taking place at critical levels is replaced by a more realistic mechanism; the damping of Kelvin and Rossby-gravity waves by infrared cooling produces the necessary momentum flux for accelerating the mean flow. As a result, there is no longer the need for a full spectrum of gravity waves.

The Kelvin wave is an eastward travelling wave, which has distributions of pressure and zonal velocity symmetrical about the equator but has essentially no meridional velocity component. On the other hand, the westward travelling mixed Rossby-gravity wave has distributions of pressure and zonal velocity that are antisymmetric about the equator but has a symmetrical distribution of meriodional velocity. The horizontal distributions of pressure and velocity characteristic of these waves are shown in Figure 2.2. Both these waves transport zonal momentum upward; the Kelvin waves transport westerly momentum while the mixed Rossby-gravity waves transport easterly momentum. The dynamics of these waves can easily be deduced theoretically as given by Holton (1979) (see Appendix A) as well as by Gill (1982). Both the Kelvin and the mixed Rossby-gravity waves are primarily absorbed by existing dissipation mostly due to infrared cooling as they propagate upward. This absorption leads quite obviously to a divergence of the total momentum flux and a consequent acceleration of the mean flow. The greatest absorption tends to occur near the transition between easterlies and westerlies. When this occurs, the deposition of momentum in the mean zonal flow produces an acceleration that is westerly for the Kelvin wave and easterly for the mixed Rossby-gravity wave. This acceleration then alters the original mean zonal wind distribution in such a manner as to cause a downward propagation of the shear zone. This in turn causes the waves to be absorbed at lower levels so that the shear zone is lowered further. This process continues until that shear zone reaches the tropopause.

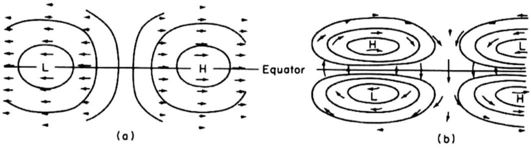


**Figure 2.2** Velocity and pressure distributions in the horizontal plane for (a) Kelvin waves, and (b) mixed Rossby-gravity waves. (After Matsuno, 1966)

There is strong observational and theoretical support for all the crucial requirements of Holton and Linden's (1972) mechanism; however, there are still more uncertainties to be cleared up. Despite these remaining uncertainties, the theory has been able to explain the major features of the QBO and produce some understanding of its origin.

More recent studies reviewed that both the Kelvin and the mixed Rossby-gravity waves cannot explain fully the QBO mechanism due to two main reasons (Sato, 1998). First, there is an upward mean motion in the equatorial stratosphere which is about 0.3 mm/s according to the result obtained from recent Upper Atmosphere Research Satellite observations (Mote et al., 1996). This upward speed is comparable to the magnitude of the downward phase speed of the QBO, which is about -0.5 mm/s at the maximum. Thus even long-period Kelvin waves provide only about a half of the westerly acceleration (Dunkerton, 1991). The easterly acceleration contributed by the Rossby-gravity waves is no more than 20%. Second, the QBO extends latitudinally with its amplitude reduced by half at about 10 degrees. On the other hand, amplitude of the Kelvin waves becomes half, meaning that now they contribute only one forth of the westerly acceleration. The meridional expanse of Kelvin waves shrinks further when the phase velocity approaches the velocity of the mean wind produced by the wave acceleration (Alexander and Holton, 1997). Thus more waves that accelerate the off-equatorial mean zonal wind are necessary. The most possible candidate that contributes to the QBO acceleration is gravity waves. Piani et al. (2000) found that the momentum-flux divergence associated with convectively generated gravity waves could account for up to 30% of the forcing required for downward propagation of the westerly phase of the QBO, and up to 15% of the required forcing in the easterly phase of the QBO.

## 2.2 Equatorial Gravity Waves - Some Observational Facts and Characteristics

Recently the momentum fluxes associated with eastward- and westward-propagating gravity waves were estimated separately (Sato and Dunkerton, 1997). These momentum fluxes of short-period gravity waves with periods of 1-3 days are found to be at least 3 times larger than that of Kelvin waves. This means that gravity waves have a higher potential to drive the QBO.

Recent statistical studies on the meridional distribution of gravity waves around the equator had shown that gravity waves have large amplitudes in the equatorial region as well as mid-latitude regions (Allen and Vincent, 1995). Even if the forcing induced by gravity waves is widely spread, the QBO can keep its own meridional expanse as Coriolis force balances the wave-induced forcing in the middle and higher latitudes. Thus the wave-induced forcing can produce the QBO only in the equatorial region (Haynes, 1998). The latitudinal width of the QBO as given by Haynes (1998) is

$$L \approx (\sigma/\alpha)^{1/4} (ND/\beta)^{1/2}$$

where $\sigma$ is the frequency of the QBO at the equator, $\alpha$ is the radiative damping rate, $N$ is the buoyancy frequency, $D$ is the depth scale of the QBO forcing and $\beta$ is the horizontal gradient of the Coriolis parameter at the equator. This width is controlled, not by the width of the momentum fluxes arising from any particular wave, but by the internal dynamics of the symmetric circulation itself with radiative damping, in particular, playing an important role.

There are a variety of gravity waves existing in the equatorial region (Sato, 1998). First, the inertial frequency, the low frequency cut-off of gravity waves, becomes zero in the equatorial region. Thus the intrinsic periods of gravity waves are distributed from 5 minutes to a few days. Second, equatorially trapped inertia-gravity waves exist as well as plane gravity waves. There are four possibilities of equatorially trapped waves. The zonal wavelengths of these waves are 3,000 km for Kelvin or plane gravity wave, 4,500 km for eastward propagating inertia-gravity wave, 10,000 km for eastward propagating inertia-gravity wave, and 5,000 km westward propagating inertia-gravity wave. The corresponding zonal phase velocities are 20 m/s, 27 m/s, 60 m/s, and –30 m/s respectively.

## 2.3 The Tropospheric Biennial Oscillation

### 2.3.1 Observed Features

2.3.1.1 General

One important feature associated with the TBO over the Asian-Australian monsoon sector is that the atmospheric convection exhibits a characteristic spatial and temporal structure. Meehl (1987, 1993) noted a specific TBO pattern with a distinct seasonal sequence. Spatially, anomalies in convection start over the Indian monsoon region during northern summer and propagate southeastward in the course of the seasonal march from northern summer to winter. Temporally, while the anomalies associated with the TBO move and develop continuously from northern summer to the following winter season, they become stationary and decay between northern winter and the following summer (Yasunari, 1990; Tomita and Yasunari, 1996).

In a related study, Ropelewski et al. (1992) observed the temporal and spatial structure of the biennial and lower-frequency oscillations in both SST and surface wind fields over the equatorial Indian and Pacific oceans.

Ose (2000) related the variability of the precipitation in the tropical western Pacific to the biennial oscillation in the South China Sea SST.

2.3.1.2 Sea Surface Temperature Anomaly

Shen and Lau (1995) found that the East Asian summer monsoon (EASM) rainfall possesses a strong biennial signal, which is particularly pronounced over the southeast China. Their results indicate that the biennial variations in the SST and EASM rainfall are closely linked. The most pronounced SST signals are found in the equatorial eastern Pacific and Indian Ocean about two seasons preceding and following the EASM rainfall.

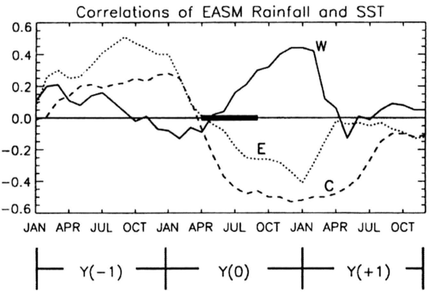


**Figure 2.3** Lag cross-correlations between the EASM rainfall and the SST at the western Pacific (120°E-140°E, 0°-8°N) (solid), the central Pacific (170°W-150°W, 0°-8°N) (long dashed) and the eastern Pacific (90°W-82°W, 0°-10°S)(short dashed). Thick bar denotes the rainy season (April to September). The Y(0) denotes the year that the rainy season within, and Y(-1), Y(+1) denote the year before and after Y(0), respectively. (Adopted from Shen and Lau, 1995)

Their lag correlations (Figure 2.3) show that a wet EASM is preceded by warm eastern and central equatorial Pacific and Indian Ocean SST anomalies (SSTA) in the previous winter. These SSTA decay and switch sign in northern spring, leading to cool anomalies after the monsoon rain starts. The cooling continues through northern fall and peaks in the following winter. Thus the EASM will be strong after an El Nino-like condition in the preceding winter and in the following year a weak summer monsoon will follow a La Nina-like condition in the winter. They also suggested that the relationship between the Asian monsoon and tropical SST is more robust in the biennial than the ENSO time scale, hence raising the possibility that the biennial oscillation may be more fundamentally related to monsoon-ocean-atmosphere interaction than ENSO itself.

A different study (Lau and Yang, 1996) shows that prior to a strong Indian monsoon, SSTA tend to be positive in the equatorial Indian Ocean as well as the equatorial eastern and central Pacific. But they tend to be negative in the equatorial western Pacific north of Australia. Specifically, the equatorial eastern-central Pacific is occupied by warm anomalies for the entire preceding calendar year until about March of the same year, and the equatorial Indian Ocean experiences warm anomalies from the preceding October through the boreal spring. By contrast, cold anomalies prevail in the equatorial western Pacific during most part of the preceding 18 months.

In another study by Chang et al. (2000), the EASM rainfall anomalies had been related to different phases of the El Nino and La Nina cycles, the western Pacific SST and the TBO. They show some important differences in the relationships between the EASM rainfall and the tropical Pacific SST. The relationship resembles a TBO pattern during the first interdecadal period of 1951-77 (ID1) when the SST changes sign in northern spring. But in the second interdecadal period of 1978-96 (ID2) the sign change occurs in northern fall and longer time scales replace the TBO pattern in the equatorial eastern Pacific SST. The cause of the interdecadal change of the SST-monsoon relationship is attributed to the change of interdecadal basic state SST and wind changes in the Pacific. The relationships between the EASM and the eastern Pacific SSTA, in both the interannual and interdecadal scales, are mostly due to the variations of the western North Pacific subtropical ridge. In a wet monsoon year, the anomalous easterly winds south of the monsoon-enhanced anomalous anticyclone starts to move slowly eastward towards the eastern Pacific, carrying with them a cooling effect on the ocean surface. In 1951-77 this effect is insignificant as the equatorial eastern Pacific SSTA already changed from warm to cold in northern spring. In 1978-96 the equatorial eastern Pacific has a warmer mean SST. Thus a stronger positive feedback with the Walker circulation during a warm phase tends to keep the SSTA warm until northern fall, when the eastward propagating anomalous easterly winds reach the eastern Pacific and reverses the SSTA. The interdecadal SST differences for March are shown in Figure 2.4.

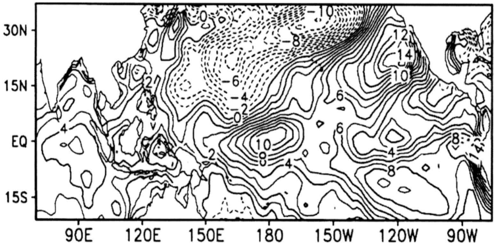


**Figure 2.4** Interdecadal SST differences (0.1°C) for March (ID2 wet composite minus ID1 wet composite). (Adopted from Chang et al., 2000)

2.3.1.3 Wind Anomaly

Corresponding to the pre-summer SSTA, the wind anomaly is mostly westerly in the equatorial central Pacific and easterly in the equatorial Indian Ocean. Near the start of the strong Indian monsoon, easterly anomalies develop over the Pacific between approximately 130°E-130°W simultaneously with the development of significant negative SSTA in the equatorial eastern-central Pacific and positive SSTA in the equatorial western Pacific. Over the equatorial Indian Ocean, westerly anomalies are established following the strong monsoon and are accompanied by the decrease of the local SST.

### 2.3.2 Theory of the TBO

Early theory proposed that the TBO originates from the modulation or amplification of the seasonal cycle due to distinctive feedbacks between the surface wind and pressure during wet and dry seasons (Nicholls, 1978).

Later, Meehl (1987, 1993) proposed a conceptual mechanism that includes a large-scale east-west circulation linking the strength of the monsoon with the equatorial eastern Pacific SST and a memory effect of the tropical ocean west of the South Pacific Convergence Zone. To identify the processes that can help a reversal of the monsoon anomalies from one summer to the next, Meehl (1994, 1997) further considered the impact of land surface and mid-latitude processes. The TBO is actively linked to the South Asian monsoon. The linking mechanism depends on coupled land-atmosphere-ocean interactions in the Indian sector, large-scale atmospheric east-west circulations in the tropics, convective heating anomalies over Africa and the Pacific, and tropical-midlatitude interactions in the Northern Hemisphere. Convective heating anomalies elsewhere in the tropics associated with the coupled ocean-atmosphere biennial mechanism can contribute to altering seasonal midlatitude circulation. These changes in the midlatitude longwave pattern are forced by a combination of tropical convective heating anomalies over East Africa, Southeast Asia, and the western Pacific. They are then able to maintain temperature anomalies over south Asia via advection through winter and spring to set up the land-sea meridional tropospheric temperature contrast for the subsequent monsoon. Anomalous heat sources and sinks associated with SST and convective anomalies produced by air-sea interaction and east-west atmospheric circulation over the tropical Indian and Pacific sectors involved with the atmosphere-ocean biennial mechanism may be forcing circulation anomalies in the midlatitudes via remote Rossby wave response. Such forcing could alter the midlatitude circulation in such a way as to maintain the surface temperature anomalies over South Asia that are essential to the atmosphere-land-ocean biennial mechanism.

Tomita and Yasunari (1996) also proposed that anomalous SST in the South China Sea served as the agent to connect the Asian summer monsoon and midlatitude winter circulation pattern. More recently, Clarke et al. (1998) suggested that the interaction between the mean seasonal wind cycle and evaporation may excite an unstable biennial mode in the western equatorial Pacific that may then drive a biennial oscillation in the Indian Ocean.

Chang and Li (2000) had formulated a theory to explain why a strong South Asian summer monsoon will lead to a strong Australian monsoon. And how the reversed sea surface temperature anomalies following a South Asian summer monsoon can be sustained for several seasons to reverse the phase of the monsoon TBO in the next year. In their studies, they apply a simple tropical atmospheric-ocean-land model over two land regions representing the South Asian and Australian monsoon regions, and three oceans representing the equatorial Indian ocean as well as the western and eastern Pacific. In this model, the five regions interact with each other through the SST-monsoon, evaporation-wind, monsoon-Walker circulation and wind stress-ocean thermocline feedback.

Warm SSTA in July in the equatorial Indian Ocean cause an increase of surface moisture convergence into South Asia, leading to a stronger monsoon. The monsoon heating induces a westerly wind anomaly in the Indian Ocean, and at the same time, intensifies a planetary-scale east-west circulation leading to anomalous easterlies over the central and western Pacific. The westerly anomaly over the Indian Ocean decreases the local SST due to evaporation-wind feedback while the easterly anomaly in the central Pacific increases the SST of the western Pacific due to wind stress-ocean thermocline feedback. The persistence of this warm SSTA in the western Pacific through northern fall leads to a stronger Australian monsoon. Meanwhile, the warming in the western Pacific also induces a stronger local Walker cell and thus maintains a surface westerly anomaly over the Indian Ocean. This westerly anomaly helps the cold SSTA to persist through the succeeding seasons, leading to a weaker Asian monsoon in the following summer. During northern winter, the westerly anomaly associated with the stronger Australian monsoon reinvigorates the warm SSTA in the western Pacific through anomalous ocean downwelling and reduction of evaporation. This further intensifies the eastern Walker cell and helps to keep the eastern Pacific cold. A schematic diagram indicating the interactive processes leading to the TBO is shown in Figure 2.5.

This theory indicates that the TBO is an inherent result of the interactions between northern summer and winter monsoon and the tropical Indian and Pacific oceans. This TBO mechanism is very different from ENSO. The ENSO mechanism involves ocean-atmosphere interaction processes mainly in the tropical Pacific. It does not require monsoon interactions while the Asian-Australian monsoon is an essential part of the TBO. Thus, the TBO is another important component of the tropical ocean-atmosphere interaction system, other than the ENSO. While the eastern Pacific plays only a passive role in the TBO mechanism, the western Pacific is crucially important. It serves as a bridge in space and time, both in connecting the convection anomaly from the northern summer to the northern winter monsoon and in channelling the feedback of the northern winter monsoon to the Indian Ocean.

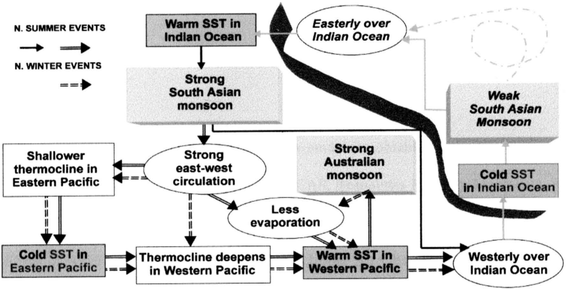


**Figure 2.5** Schematic diagram indicating the interactive processes leading to the TBO. Land and ocean regions are shaded. Nonshaded boxes indicate atmospheric (oval) and oceanic (rectangular) processes involved in the interactions. Northern summer-driven events follow red solid thin and double arrows; northern winter-driven events follow black-dashed double arrows. The strong monsoon phase starts with warm SST in the Indian Ocean leading to a strong South Asian monsoon in northern summer. In addition to a surface westerly anomaly over the Indian Ocean during northern summer (thin arrow) that cools the SST, complex interactive processes (double arrows, both summer and winter driven) involving both western and eastern equatorial Pacific and the Australian monsoon are required to reinvigorate this westerly anomaly until the next summer. These processes also lead to a strong northern winter monsoon that follows the strong summer monsoon. The reversed phase is sketched in the upper right side of the ribbon. (Adopted from Chang and Li, 2000)