CHAPTER – 1
1. **Earth’s Atmosphere**

Earth’s atmospheric origin, composition and organization are a subject of debate from a long time. One can certainly be fair of is that, when the Earth was formed, it was assumed to be hot enough to keep hold of its hydro carbon gases in embryonic atmosphere. Since from its formation, expulsion of gases had taken place from volcanoes, emitting large quantities of water vapour and major gases from Earth’s interior part that led to the formation of clouds which produces rain which are in turn gathered by water flow channels thereby acting as absorbers for collected carbon dioxide that leads to the formation of limestone and sedimentary rocks while chemically inactive nitrogen remains back in the atmosphere with absence of oxygen in early Earth’s atmosphere.

Earth’s modern environment contains gas coverlet encircled by nitrogen (78%) and oxygen (21%). Existence of life cannot be imagined without the existence of this coverlet. The gas coverlet begins at Earth’s surface level and reaches up to six hundred miles into the sky, forming different layers with a thickness of six miles. The uppermost terminal of Mount Everest is supposed to peek sharp of its layer while left over five hundred and ninety four miles doesn’t allow sustenance of life. Nonetheless, the atmospheric layers shield us from the hazards of sun’s radiation, in addition to downward drifting of space debris and others.

1.1 **EARTH’S ATMOSPHERIC LAYERS**

To understand the nature and different processes of weather and climate within the Earth’s atmosphere, both horizontal and vertical structures will be taken into consideration. The vertical structure of Earth’s atmosphere can be studied by considering that atmosphere is organized in different layers. From analytical point of view, most often these layers are studied by considering their vertical structure which exhibits variability in height with respect to physical parameters like temperature, pressure and density, while horizontal scales are used to study horizontal motion of clouds and other associated phenomena. From thermal structure point of view i.e. variation in temperature with height, Earth’s atmosphere can be separated into four different layers and that starts with troposphere followed by stratosphere, mesosphere and thermosphere as shown in following figure 1.1.1, where the layers are listed beginning with the lowest followed by corresponding description of each layer.
1.1.1 Troposphere

The first and foremost layer from the Earth’s surface is called the troposphere. The word “tropo” is derived from Greek word "τροπή", which signifies turn or overturn, as it promotes rising and falling of air by vertical mixing of gases stirred by storms, convection and wind systems. Troposphere contains eighty percent of atmospheric mass. The troposphere starts at ground level and extends up to four to twelve miles (6-20km). In this layer, as the density of gases decreases with height, air becomes thinner with the increase of height. Energy from the surface mostly heats up the troposphere, and hence, the lowest part of the troposphere is warmest (re-radiation) and then it decreases with altitude (temperature inversion) at an average rate of 6 C/km. The height of troposphere differs between seasons as well as latitudes. To elaborate, Earth’s equator having the highest troposphere (~17 km/56,000 ft./ 5½ miles) at 50°N and 50°S and lowest troposphere at the poles with four miles (~7-8 km/23,000 ft.). Tropopause is the boundary that segregates the troposphere and the stratosphere.
1.1.2 Stratosphere

The second layer above the troposphere separated by tropopause is known as stratosphere and it commences at the six-mile point and ends up nearly thirty one mile. It seizes nineteen percent of the atmospheric gases, but little water vapour. Importantly, stratosphere contains an ozone sub layer that blocks large amounts of dangerous solar UV radiation by absorption, where energy from solar radiation is converted into kinetic energy, thereby becoming warm and leading to temperature inversion. Positive gradient in temperature is a general phenomenon (temperature increases with height) in the stratosphere as radiation is increasingly absorbed by oxygen molecules in the process of ozone layer formation and it rises from an average of \(-76^\circ F\) (-60°C) at tropopause to a maximum of about \(44^\circ F\)(7°C) at the stratopause. This increase in temperature with height restricts turbulence and mixing (means no convection occurs at all) since the vertical movement of gases is restricted by this temperature inversion. The transition boundary separating the stratosphere from the mesosphere is called the stratopause, where the pressure will be around \(1/1000^{th}\) of sea level and stratospheric temperature stops increasing.

1.1.3 Mesosphere

The third layer of Earth’s atmosphere is known as the mesosphere. Above the stratospheric layer there comes the layer of mesosphere. It begins from thirty two miles above the Earth’s surface and limits at around 80km, wherein warmest temperatures present at lowest level and colder ones at highest level (-100°C). This decrease in temperature makes the mesosphere unstable and leads to large vertical convective currents. A small fraction of the neutral particles gets ionized by the solar radiation in ionospheric part of mesosphere termed as D region (~60-90 km). Both stratosphere and mesosphere along with tropopause and stratopause constitute middle atmosphere. The transition layer where the temperature approaches minimum value is called mesopause which will be around 90 km from Earth’s surface.

1.1.4 Thermosphere

The fourth and outer most layer located above mesosphere of the Earth’s atmosphere is known as thermosphere and it starts at fifty mile. Thermospheric temperatures become too hot (600-2000K) and this increase in temperature is associated with absorption of intense solar radiation by residual molecular oxygen. Although thermosphere has a very high temperature due to faster movement of molecules, it exhibits a low heat since the molecules presented in this layer are spaced so far apart according to their atomic masses. In this layer, the chemical composition of air in the atmosphere becomes mostly dependent on altitude and filled with lighter gases like atomic oxygen, hydrogen.
and helium. Here, a small but significant fraction of neutral particles gets ionized by solar activity resulting in E and F regions of the ionosphere extending from 90 km to greater than 1000 km that leads to wide range of electrodynamics phenomena due to ionosphere-thermosphere interactions.

Although the atmosphere is constantly evolving in its details, some aspects of its observed composition and gross vertical structure are observed to be relatively constant. Atmosphere is bound to the Earth by gravity. It is only because of the uneven addition and subtraction of heat energy to and from the atmosphere, equilibrium level cannot be reached, so that atmosphere remains in a perpetual state of both horizontal and vertical motion, otherwise real weather doesn’t exist. The ability of air to compress or expand substantially in response to changes of pressure is also directly or indirectly responsible for a variety of atmospheric phenomena.

1.2 Atmospheric Dynamics - Different wave activities

Waves can be seen all around us. Perhaps, the most common are gravity waves on water and compressibility waves in air. To understand different wave activities which are very much essential in the observational analysis of the Earth’s atmosphere, it is required to learn various wave motions. The identification and analysis of these waves will allow us to focus on certain wave types and to better understand the capability and effectiveness of commonly made approximations of atmospheric wave dynamics. Some waves those are common to rotating bodies with sensible atmospheres like our planet Earth are Rossby-Haurwitz, gravity, tidal, Kelvin, acoustic, inertial and
specifically observed oscillations such as quasi biennial oscillation (QBO), Madden Julian oscillation (MJO), El-nino and Semi Annual Oscillation (SAO) and etc. Figure 1.2.1 represents different types of waves that are prevalent in our Earth’s atmosphere. Waves in Earth’s atmosphere may be internal or external. Internal waves exhibit wave-like behaviour with transfer of energy in the vertical direction. Conservative waves maintain nearly constant energy density with altitude which implies that wave amplitude grows exponentially with altitude whereas small amplitude waves originating in the lower atmosphere may achieve large amplitudes in the upper atmosphere. External waves have constant phase with height but their wave amplitudes temporarily decays with height.

In lower thermosphere and mesosphere, major payment to the momentum balance is mainly due to either breaking or dissipation of non-mountain gravity waves, resulting in hindrance which decelerates the wave motion and generates a Coriolis torque thereby modulating the meridional circulation by inducing additional summer pole to winter pole drift and an ascent over summer pole while a descent over winter pole. This adiabatic expansion and compression influences the temperature structure down to the lower stratosphere in Polar Regions. Dunkerton (1997) have shown that non-mountain gravity wave forcing is also important in driving the interannual variability of the tropical stratospheric circulation, most importantly QBO and SAO. Further, equatorial waves as well as tropical convection and their associations are the primary mechanisms for the climatic systems at tropics.

In order to study the wave nature where energy exchanges are oscillatory such as ENSO, MJO, QBO and SAO, it is essential to observe planetary scale equatorial waves (Kelvin waves) and westward propagating (Rossby-gravity waves) along with small and intermediate scale gravity waves. Here, an attempt is made to identify the basic atmospheric wave motions in order to understand some important properties and characteristics of waves. It is very true that to retain all wave types in this chapter is unmanageable and we, therefore, restrict ourselves to those waves that have significant impact on climate change.

1.2.1 Gravity waves

In a stably stratified fluid, high frequency and short horizontal scale oscillations arise when atmospheric parcels are displaced vertically (Baldwin et al., 2001). Gravity waves are those disturbances whose intrinsic frequencies are smaller than the Brunt Väisälä frequency. The wave perturbations, in general, will be described by downward phase progression and increasing amplitude with height. Atmospheric gravity waves are ubiquitous in nature with small scale propagation having wavelengths from tens to thousands of kilometres and periods of minutes to
hours. Physically they appear as billows and bands. As they transport momentum vertically, these waves are capable in dictating large-scale momentum balance of the stratosphere and mesosphere. In general, gravity wave amplitudes are limited due to the dissipation and saturation processes. Saturation implies that the wave field has reached an amplitude such that either secondary instabilities or nonlinear interactions, such as the parametric sub-harmonic instabilities can occur, which limit the growth of the waves. The saturation mechanism is considered to be associated with the generation of convective or Kelvin-Helmholtz (KH) instabilities.

Researchers showed that, in the absence of gravity wave hindrance, sinking at the southern winter polar stratopause was reduced by a factor of two which results in middle atmosphere temperatures too close to radiative equilibrium, particularly with unrealistic cold winter polar stratosphere temperatures. Such biases are more intense in the southern winter due to weak planetary wave forcing. It has been disclosed by the early works made by research community that the size of the cold bias was extremely sensitive to horizontal resolution in the absence of gravity wave hindrance while the sensitivity of the cold bias to vertical resolution was shown to be limited. As gravity waves propagate vertically their amplitude increases and eventually becomes unstable due to decrease of ambient air density with height which results in nonlinear breakdown and turbulent dissipation. Although the shape of wave spectrum is consistent with saturation theory but still remained as an unresolved question. In contrast to that, gravity wave variations in both space and time are observed in the troposphere with respect to their source and are frequently modelled as a spectrum of horizontal winds proportional to transitional wavenumber estimated with the constant $s=1$.

1.2.2 Kelvin waves

Kelvin waves were first discovered by Lord Kelvin in 1879, which are commonly identified on rotating bodies with sensible atmospheres such as planets like Earth. Kelvin wave is nothing but a one type of gravity wave that exhibits large scale motion due to Earth’s rotation. They were trapped symmetrically about the Earth’s equator or along lateral vertical boundaries such as mountain ranges or coastlines with eastward propagation and a negligible meridional velocity component. Among various atmospheric waves, planetary scale eastward propagating waves with periods of ~10 - 20 days (slow), zonal wavenumbers of 1 - 2 and phase speeds of ~20 - 30 ms$^{-1}$ are known as Kelvin waves (Wallace and Kousky, 1968). Holton (1972), Salby and Garcia (1987) described that KWs are produced by convective heating in the tropical troposphere. A dominant presence of gravity waves with periods of ~10 h and ~4–5 h near tropopause were noticed by Dhaka et al. (2011).
Figure 1.2.2.1 Longitude-time diagram of outgoing long wave radiation (OLR) over 10°N–10°S during October 2001 to March 2002. Blue and green colours correspond to deep convection. The arrow (on 27 January) denotes the time period shown in Figure 1.2.2.2 (adopted from Randel and Wu, 2005).

Figure 1.2.2.2 Relationship between convective forcing and Kelvin wave temperature response (adopted from Randel and Wu, 2005)

The above figures 1.2.2.1 and 1.2.2.2 (fig.12 and 15 of Randel and Wu, 2005) represents the relationship between convective forcing and Kelvin wave temperature response near the tropopause for an event on late January 2002 identified by Randel and Wu, (2005). The location and altitude of convection is derived from OLR data shown in figure 1.2.2.1 (fig.12 of Randel and Wu, 2005).
contours of temperature anomalies are ± 0.5, 1.5, 2.5, while the thick line denotes cold point tropopause with vertical arrow representing group velocity of Kelvin wave i.e. perpendicular to the phase lines. Dhaka et al. (2011) identified emergence of gravity waves with short vertical wavelength (~2km) and short period (~2-3h) which were confined in the close vicinity of tropopause during and after convection, while absence of them during non-convective days, suggesting convection to be the source of gravity waves.

Takayabu and Murakami (1991) suggested that the Kelvin waves might be characterized by different phase speeds in different regions of the world. Wheeler et al. (2000) showed that Kelvin waves produce the dominant coherent signal in eastward propagation in the 3-20 day band. An exclusive feature of the Kelvin wave is its unidirectional propagation. It moves equator-ward along a western boundary whereas pole-ward along an eastern boundary and cyclonically around a closed boundary i.e. clockwise in the southern hemisphere and counter clockwise in the northern hemisphere. The relationship between horizontal wind and convection in the northern hemisphere is consistent with that found in Kiladis (1998) who indicated that intensified convection appears in the region of the pole-ward flow ahead of upper-level troughs (cyclonic anomalies) in the wave train over the central Pacific. Generally, the phase speed associated with Kelvin waves is nearly 17 ms⁻¹ (Wheeler and Kiladis, 1999; Straub and Kiladis, 2002), meanwhile the phase speeds reported by Dunkerton and Crum (1995) was 10-13 ms⁻¹ over the Indian ocean. Wang and Xie (1997), Hoskins and Yang (2000) showed that Kelvin waves are stronger in an easterly wind region while they are stronger in westerly region during westward motion. Yang and Hoskins (2007a, b) showed that the Kelvin wave is more sensitive to the boundary thermal condition and closely coupled with convection than other equatorial wave modes.

To eliminate the ambiguity between observed waves and pure Kelvin waves, Chao (2007) labelled convectively coupled Kelvin waves as “chimeric Kelvin waves” in order to include “parts of different origin” than pure Kelvin waves. Wheeler and Kiladis (1999), Straub and Kiladis (2002, 2003a) reported that phase speeds associated with convectively coupled Kelvin waves as 17ms⁻¹ which are consistent with Dunkerton and Crum (1995) while analysing super clusters over the Indian Ocean. Nakazawa (1988), Takayabu and Murakami (1991), Dunkerton and Crum (1995), Wheeler and Kiladis (1999), Straub and Kiladis (2002) showed that synoptic-scale eastward moving super clusters were linked to the convectively coupled waves. In contrast to studies made by Wheeler and Kiladis (1999), Roundy and Frank (2004a), nonlinear interactions between atmospheric waves would invalidate the strict superposition principle for convectively coupled waves. It appears that convection filtered in a larger spectral domain of eastward-moving activity, as used in Yang and Hoskins (2007a, b) is better
in capturing the coupling with Kelvin waves. This lower level coupling feature in the eastern hemisphere (EH) was also shown in Yang and Hoskins (2007a, c). For example, it has been well noticed that the spectral peak associated with other oscillations (e.g. MJO) intersects with the theoretical Kelvin wave dispersion line at zonal wavenumber 1 while spectral power associated with them is spread broadly across wavenumbers 0 to 10.

![Figure 1.2.2.3](image1.png)

**Figure 1.2.2.3** Symmetric component of OLR with odd meridional mode numbered equatorial waves at frequency spectral bandwidth of 1.96 cpd (adopted from Wheeler and Kiladis, 1999)

![Figure 1.2.2.4](image2.png)

**Figure 1.2.2.4** Normalized OLR spectra for Northern Hemisphere (adopted from Roundy and Frank, 2004a)
Figure 1.2.2.5 Filtered cross equatorial symmetric data (adopted from Roundy and Frank, 2004a)

Figure 1.2.2.6 Spectral bands of the five filters used in the project are outlined with solid curves and plotted over a smoothed tropical OLR spectrum. Where these filters differ from those of Wheeler and Kiladis, 1999, their corresponding filters are enclosed by dotted lines (adopted from Wheeler and Kiladis, 1999; Roundy and Frank, 2004a)

Figure 1.2.2.7 Spectral Bands for Filtering (adopted from Wheeler and Kiladis, 1999 and Roundy and Frank, 2004a)
The wavenumber frequency spectra shown in figure 1.2.2.3 (Fig. 3b of Wheeler and Kiladis, 1999) and figures 1.2.2.4, 1.2.2.5 and 1.2.2.6 (Figures 2b, 2c and 4 of Roundy and Frank, 2004a) show that the Kelvin wave peak following the h=50m dispersion line near wavenumber 2 and the 25m dispersion line near wavenumber 7, exhibiting the tendency for small scale Kelvin waves to propagate more slowly than large scale waves as shown in following figure 1.2.2.7.

The theory of equatorial waves was first developed during the late 1960s comprising development of solutions based on modes which are formulated by considering lateral propagation, refraction, and reflection of waves within an equatorial waveguide, the horizontal extent of which depends on wave properties, for example, turning points where wave intrinsic frequency equals the local inertial frequency. The solutions included a Rossby mode, and a mode which became known as mixed Rossby-gravity mode. A third solution, an eastward propagating gravity mode, was called the equatorial Kelvin mode. In general, the equatorial region receives relatively high solar radiation compared to Polar Regions. Consequently, the equatorial region acts as a source for many atmospheric wave modes. Amongst them, Kelvin waves are most important equatorial wave modes that are forced in the tropical troposphere by convective processes described by Pires et al. (1997), Lindzen, (2003), Randel and Wu (2005), Brahmanandam et al. (2010).

![Figure 1.2.2.8 Kelvin wave signatures during different seasons from Fall 2006 to Winter 2007 [(a) - (f)]. (adopted from Brahmanandam et al.2010)
Figure 1.2.2.8 (Fig.8 of Brahmanandam et al., 2010) shows the overall space-time characteristics of the planetary-scale equatorial waves in GPS data that are consistent with Kelvin waves satisfying the linear dispersion relation for eastward propagating hydrostatic gravity waves while downward phase progression is clearly evident in all the cases, indicating that the possible source of these waves and their interaction with mean flow of background wind conditions could be from lower altitudes due to tropospheric convective heating (Ratnam et al., 2006). The planetary-scale (wavenumber, n=1 or 2), eastward propagating and equatorially trapped Kelvin waves often show periods ranging from few days to few tens of days with vertical wavelengths of 5-8 km (Ratnam et al., 2006). Yang and Hoskins (1996) specified that midlatitude non-stationary Rossby waves with eastward phase speeds can propagate into equatorial region with reasonably strong westerly flow indicating stronger association of Kelvin wave amplitude with an upper-tropospheric forcing of extra tropical Rossby waves. Yang and Hoskins (2007a, 2011, and 2012) showed that equatorial wave activity exhibits zonal and vertical variations which are partly due to variations of the ambient zonal flow.

Deep convection is a predominant source of wave variability in the tropical atmosphere while tropical Kelvin waves are primarily forced by transient deep convection (Randel and Wu, 2005). During the process of tropical cyclogenesis from an equatorial wave, a region of active convection forms on the northwest side of a near equatorial wave trough within the region of maximum low-level convergence for a mixed Rossby-gravity wave.

Figure 1.2.2.9 Satellite imagery of an equatorial MRG wave at 0530 UTC 31 Oct 2007 illustrates the impact of the convection on the wave structure. (adopted from http://www.goes-r.gov)
This convective region moves away from the equator and eventually develops into a tropical storm as shown in figure 1.2.2.9, which exhibits westernmost convection centre of the wave moving northwards east of Somalia as weakening Tropical Cyclone 05A. Further, the seasonality of this tropical cyclogensis has been related to the convective potential of the region under consideration (Ratnam et al. 2013), while dynamical factors contribute to the daily potential for genesis.

Convective Available Potential Energy (CAPE) is good a measure of conditional stability (Ratnam et al., 2013). In general, high CAPE values are identified to be associated with high rainfall in a study conducted by Dhaka et al. (2010) during the study period of 1980-2006 over four different specific locations in India which showed that temperature variations in the upper troposphere are jointly controlled by CAPE and the solar cycle.

Figure 1.2.2.10 Time series of large-scale temperature variance at 16.5 km (top curve) and OLR averaged over 10°N–10°S, 60°–180°E (near Indonesia, bottom curve). Note the relationship between maxima in wave variance and transient convection, as indicated by the arrowed events (adopted from Randel and Wu, 2005).

Figure 1.2.2.11 Correlation maps between OLR variations over 10°N–10°S, 60°–180°E (the location is indicated by the thick arrows; the time series is shown in Figure 1.2.2.10) and mapped GPS temperature data at each longitude and altitude. The temperature time series have been lagged by 2 days with respect to OLR to maximize the overall correlations. The sign of the correlations is such that negative (dashed) lines refer to cold temperature anomalies during enhanced convection (low OLR). Contours are ±0.2, 0.3, 0.4. The thick line is the cold point tropopause. Results are shown for the periods (a) November 2001 to March 2002 and (b) April–September 2002 (adopted from Randel and Wu, 2005).
Large scale variations of Kelvin wave amplitude in the upper troposphere are mainly linked with the magnitude of transient convection near Indonesia as shown in figure 1.2.2.10 (Figure 13 of Randel and Wu, 2005) whose spatial patterns of temperature correlations show a characteristic Kelvin wave structure shown by figure 1.2.2.11 (Figure 14 of Randel and Wu, 2005) indicating that the Kelvin waves extend into the lower stratosphere when QBO easterlies are present and confined to upper troposphere during westerly phase of QBO as shown in figures 1.2.2.11 and 1.2.2.2 (figures 14 and 15 of Randel and Wu, 2005) and confirmed by Ratnam et al (2006). In both phases, transient convection near Indonesia exhibits planetary scale temperature response near the tropical tropopause with a dipole pattern of warm and cold anomalies with cold temperatures over convection, in agreement with earlier studies (Garcia and Salby, 1987). Thus global scale wave response and convective cooling may contribute to the observed cooling of the tropopause region above deep convection and interdependent on each other, an important consideration to recognize the impacts of convection on the tropopause region.

Equatorial waves as well as tropical convection and their coupled behaviour are fundamental components of the tropical climate system. Redelsperger et al. (1996) showed a substantial fraction of large scale variability in convection at time scales less than 30 days associated with equatorial waves. Earlier, researchers in the field of meteorology and atmospheric sciences have used equatorial waves to explain some fundamental phenomena of the tropical climate such as Walker circulation, atmospheric teleconnection patterns, MJO, QBO and semi-annual oscillation (SAO) in the stratosphere respectively.

A broad spectrum of waves in the tropics will contribute to QBO. On the basis of observations of wave amplitudes, one can believe that Kelvin waves do play a major role in posing required momentum flux to drive QBO (Dunkerton, 1997). Convection in the tropical troposphere produces Kelvin waves with a variety of vertical and horizontal wavelengths and phase speeds which propagate into the stratosphere that transport easterly and westerly zonal momentum and deposited at stratospheric levels thereby driving the zonal wind anomalies of the QBO (Ratnam et al., 2006). Holton and Lindzen (1972) refined their earlier work in 1968 by simulating a one dimensional model that represents a QBO driven by vertically propagating Kelvin waves which contribute a westerly force. The vertical profile of the zonal wind determines the critical level at or below which the momentum is deposited which further depends on shear zones of the QBO. Kelvin waves which are equatorially trapped and having periods of ≥ 3 days with wave numbers 1–4 of zonal wavelengths ≥ 10,000 km can usually contribute to QBO.
Figure 1.2.12 Kelvin wave signatures in temperature fluctuations during eastward phase [(a), (b)] and westward phase [(c), (d)] of QBO (adopted from Brahmanandam et al. 2010).

For example, figure 1.2.12 shows a study conducted by Brahmanandam et al. (2010) that reveals characteristic Kelvin wave signatures in temperature fluctuations during different phases of QBO. The vertical wavelengths in Figures 1.2.12a and 1.2.12b are found to be ~10 and 12 km while the phase lines of temperature curves in figures 1.2.12c and 1.2.12d are extending towards upper troposphere with more upright behaviour through longer vertical wavelengths around 13 and 14 km. This increase and decrease in Kelvin wave clearly signifies that there is an enhancement in its characteristic activity during the westward phase of the QBO while diminishes during the eastward phase (Ratnam et al., 2006).

Above 30km, a larger fraction of monthly mean zonal wind variance, known as “semi-annual oscillation”, had been identified, especially near the equator where the annual cycle is small, which has been first, inferred by Reed (1966). According to Hirota (1979), it can be viewed as a two linked oscillations that are out of phase with each other, one peaking near the stratopause and the other near the mesopause. Holton (1975) has suggested that the appearance of westerlies above the equator may be due to Kelvin wave absorption because of downward propagation of westerlies. Based on the evidence of existence of long vertical wavelength Kelvin waves provided by Hirota (1979), it has been showed that, for reasonable values of the wave parameters and thermal damping rate, Kelvin waves could indeed give rise to westerly accelerations of SAO.
Figure 1.2.13 ENSO SST anomaly and Bjerknes feedback

**Figure 1.2.13a:** El Niño anomalies in SST (color shading and scale in °C), surface atmospheric pressure (contours), and surface wind stress (vectors) in the Pacific basin. Pressure contour interval is 0.5 mb, with solid contours positive and dashed contours negative. Wind stress vectors indicate direction and intensity, with the longest vector equivalent to ~1 N m. The patterns in this graphic are derived from a linear regression against SST anomalies averaged over 6°N–6°S, 90°W–180° in the eastern and central equatorial Pacific. All quantities scale up or down with the intensity of anomalies in this index region, that is, higher for strong El Niños and lower for weak El Niños. Anomalies of opposite sign apply to La Niña events, although there are some differences in the spatial patterns of El Niño and La Niña that this linear analysis does not capture (Larkin and Harrison, 2002; An and Jin, 2004)( adopted from www.pmel.noaa.gov).

**Figure 1.2.13b:** Flow graph illustrating the Bjerknes feedback. A positive sign on an arrow show that an increase in one variable produces an increase in the other. The positive sign inside a circle indicates that the overall feedback is positive (adopted from www.elic.ucl.ac.be).

During El Niño event, the trade winds weaken along the equator due to rise in atmospheric pressure along western Pacific and fall in eastern Pacific resulting in anomalous warming in the central and eastern Pacific, as warm water in the western Pacific migrates eastward followed by reduced upwelling (figure 1.2.13a). Now, the Bjerknes feedback (figure. 1.2.13b) runs in reverse with reduced trade winds and decreased sea surface temperature (SST) warming tendencies along the equator reinforcing one another as El Niño develops. Further, equatorial wave induced cooling which represents a delayed negative feedback will bring an end to El Niño and, if strong enough, initiates La Niña. Thus the combination of both Bjerknes and equatorial wave feedbacks controls the magnitude and duration of individual ENSO events and the interval between them.

Figure 1.2.14 Latitude-pressure sections of horizontal eddy momentum flux of wind perturbations at 200 hPa regressed onto the extrema of equatorial mean Kelvin waves averaged over longitudinal displacements, (a) during El Niño (b) during La Niña (adopted from Yang and Hoskins, 2013)
Figure 1.2.14 clearly shows the fact that an El Niño event enhances the Kelvin wave-related convection and winds itself and a La Niña event suppresses them, whose peak is identified to be around 200hpa in contrast to earlier cases where it is above 125-150hpa, consistent with the local maximum of the wave amplitude and associated extra tropical forcing (figure 1.2.14b). Although Kelvin wave amplitude around 200 hpa is observed to be stronger in La Niña winters, due to stronger extra tropical forcing, there is less convectively coupled wave signal propagating upwards into the lower stratosphere, consistent with less vertical coupling wave observations of their study.

1.2.3 Rossby-gravity waves

The flow over the middle latitudes consists of a series of waves in the form of troughs and ridges. The distance between trough and trough or ridge and ridge is normally thousands of kilometres, known to be long wave or planetary wave. At any given instant of time, there are at least three to six long waves looping around our planet Earth. These planetary waves are also known as Rossby waves named after C.G. Rossby who studied their motion. On a dispersion relation diagram, they can be found at n=0. For large positive zonal wavenumbers, these waves behave like a gravity wave but for large negative zonal wave numbers they behave like Rossby waves and hence known to be Rossby-gravity waves. In general, they are just similar to Kelvin waves, as they are equatorially trapped i.e. they rapidly decay as their distance increases away from the equator and their Brunt-Vaisala frequency does not remain constant while trapping scale is analogous to Kelvin waves. Usually they carry energy eastward but their crests and troughs may propagate westward during long wave periods while their group velocity is always directed towards the east with a maximum for gravity waves. Based on the wavelet analysis, these waves exhibit three distinct solutions that correspond to equatorially trapped gravity wave, equatorially trapped Rossby wave and mixed Rossby-gravity waves that hold a set of characteristics of the former two. Here, it should be emphasized that, on a dispersion relation diagram, both equatorial gravity waves and equatorial trapped Rossby waves can move either westward or eastward and correspond to n=1. Mixed Rossby-gravity waves are quite common in deep tropics which can be identified on longitudinal scale of 1000-4000 km range having periods of 4-5 days that can move at speeds of ~8-10 ms⁻¹. These waves are strongly modulated by moist convection.

Mixed Rossby gravity waves also exhibit vertical propagation when their Brunt Vaisala frequency does not change. These waves were first identified by Maruyama and Yana (1967) as westward propagating mixed waves. When mass-zonal wavenumber dispersion diagram is taken into consideration, the group velocity would be directed at right angles to the n=0 for mixed Rossby-gravity waves and at n=1 for Rossby-gravity waves and continues to increase in the direction of
increasing angular frequency with speeds of \(1\text{ cm/s}\) for gravity waves and \(2\text{ mm/s}\) for Rossby waves. Their characteristics include periods of 4-5 days with horizontal wavenumbers of 4 of 10,000 km wavelengths and upward group velocity with vertical wavelengths of 4-8 km.

1.2.4 Tidal waves

The study of atmospheric tidal oscillations provides an interesting insight into the way in which the Earth’s atmosphere responds to an extra-terrestrial influence such as the input of solar energy and the local heating of the planetary atmosphere by absorption in regions with strongly absorbing atmospheric trace constituents. These oscillations are the best examples of internal waves in the atmosphere. In general, atmospheric tide indicates planetary scale oscillations of the atmosphere either excited gravitationally or thermally. Tides generated due to gravitational forces of the moon are known to be lunar tides while the tides generated due to thermal action of the sun are known to be solar tides.

General atmospheric motion that leads to generation of tides is mainly due to variations in net global heating between northern and southern hemispheres that result in meridional energy exchange. Due to Earth’s rotation, tidal strength in the atmosphere at any point repeatedly changes that leads to seasonal variations. Thus the tide is a dynamical phenomenon. For the sake of observational analysis, the wave periods corresponding to atmospheric tides are considered as integral fractions of a solar or lunar day, where diurnal refers to harmonic variation with a period of one day (24hrs), semi diurnal (12hrs) refers to a period of half a day and terdiurnal (8hrs) refers to a period of one third of a day, quarter diurnal (6hrs) refers to a period of one fourth of a day, to study tidal forcing.

Diurnal wind oscillations are found to be associated with local topographic features at the Earth’s surface at temperate and tropical latitudes. In general, diurnal tides are mainly due to heating caused by the absorption of infrared radiation (IR) by water vapour and clouds present in the troposphere and also due to ultraviolet radiation by ozone in the lower stratosphere. The feedbacks to these two driving sources are in general twelve hours out of phase so that they destructively interfere with one another resulting in reduction of diurnal tide amplitude when compared with individual IR forcing. Generally, the amplitude of diurnal tides increases with altitude up to a peak range of 105 to 115km at \(\pm 30^0\) latitude (zonal), \(\pm 20^0\) (meridional), at the equator for temperature and vertical velocity fields. Atmospheric tidal oscillations with a period of half a day i.e. 12 hours are known as semidiurnal tidal waves. These waves exhibit two maxima and two minima i.e. two cycles per day with respect to longitude in contrast to single set of maxima and minima in case of diurnal
wave where as its vertical wavelength is much longer (48-58km) than that of diurnal tide (20-25km). Atmospheric tidal oscillations with a period of one third of a day i.e. 8 hours were known to be ter-diurnal waves. Ter-diurnal waves are third harmonic oscillation of either nonlinear processes due to global diurnal tide or locally confined gravity wave component or normal mode Lamb waves often found in temperature and wind fields of atmospheric observations. This kind of tidal forcing is mainly due solar UV absorption heating by water vapour and ozone. Observational reports indicate that their appearance in winter is more evident and stable with longer vertical wavelengths than in summer.

1.3 ATMOSPHERIC OSCILLATIONS

1.3.1 QUASI BIENNIAL OSCILLATION

Quasi-periodic oscillation in the zonal wind at the equator between easterlies and westerlies which descend through the middle and lower stratosphere at a rate of ~1 km per month in the tropical stratosphere until they dissipate at the tropical tropopause with a mean period of ~26 to 29 months is known as the quasi biennial oscillation (QBO). Usually the downward motion of easterlies are irregular than that of westerlies. At the top of vertical QBO domain, easterlies dominate, while at the bottom westerlies are more pronounced. Their peak to peak range is 42 ms⁻¹ at the equator and decreases symmetrically away from the equator with a half width of ~12° latitude. The amplitude of easterly phase is twice as strong as that of westerly phase and their oscillation doesn’t appear to be directly linked to the annual cycle.

The QBO was discovered six decades back in 1950’s at the UK Meteorological Office but unable to recognize its cause. Later, Reed et al. (1961) as well as Veryard and Edbon (1961) have noticed this observation which has been explained by Lindzen and Holton (1968) that the periodic wind reversal was driven by atmospheric waves originating from tropical troposphere that travels upward and dissipated in the stratosphere by radiative cooling. Though recent studies indicate that gravity waves might be a major contributor for this kind of oscillation, the precise nature of these waves are yet to be unveiled and a growing number of climate model simulations have been adopted to understand their phenomena.

1.3.2 Madden Julian Oscillation and ElNino Southern Oscillation

Way back to 1971, Roland Madden and Paul Julian had noticed around 40-50 day oscillation of surface and upper level winds when they were analysing zonal wind anomalies in the tropical Pacific. Later, this low frequency oscillation in tropics is known as Madden and Julian Oscillation (MJO). Intra-annual fluctuations of these oscillations can reveal weather variations in the tropics since MJO
is swayed by variations in wind, SST, cloudiness and rainfall. Although it affects entire tropical troposphere, its evidence can be clearly seen over Indian and Pacific Oceans. MJO affects the intensity and break periods of Asian as well as Australian monsoons and interacts with El Niño. In general, there are periods of enhanced and suppressed activities within a particular season. At the centre of suppressed convection, Stronger than normal trade wind inversion associated with clear skies allow more shortwave radiation to reach ocean surface which causes slight increase in SST as the wave travels eastward. As trade winds are stronger than normal, enhanced evaporation from the sea surface takes place. These easterly winds and the evaporation rate weaken near the western edge of suppressed convection region leading to low level moisture convergence that triggers deep convection which in turn sways the oscillation of other half of outgoing long wave radiation i.e. the region of enhanced convection. This region comprises of one or more super cloud clusters (SCCs) that moves eastward along with Madden Julian wave. Within these SCCs, westward moving small cloud clusters form at the eastern edge of SCC and cease at the western edge with a lifetime of one or two days whereas individual mesoscale convective systems within these smaller clusters move eastward by discrete propagation with a lifetime of 6 to 12 hours. The SCCs travel eastward at 5-10ms$^{-1}$ as a travelling oscillating wave (MJO). MJO exhibits a wavenumber of 1-2 which signifies that at any given time there are one or two areas around the equator with enhanced convection and suppressed convection activities.

A pattern of change in Earth’s climate due to interaction between atmosphere and ocean is known as El Niño-Southern Oscillation (ENSO) (Trenberth, 1997). El Niño refers to the warm phase of ENSO while La Niña refers to the cold phase of ENSO in the Equatorial Central and Eastern Pacific region. The period when neither El Niño nor La Niña is present is known to be ENSO-neutral. This Earth’s natural coupled ocean-atmospheric interaction helps to stabilize the climate system by rolling between phase states for every 4-5 years. El Niño develops due to coupled ocean-atmospheric phenomenon and the amount of warm water in the tropics gets redistributed thereby depleted and again gets restored during La Niña. Often, a mini global warming follows El Niño event as a consequence of heat from ocean which affects the atmospheric circulation that in turn leads to temperature changes throughout the world. In effect, inter-annual variations takes place in the energy balance of the combined atmosphere-ocean system and are transformed as important changes in weather regimes and climate throughout the world. Recent observational reports show that the pace, intensity and duration of ENSO events were quickened and extremes became more extreme due to accumulation of heat in oceans that intensifies weather anomalies thereby leading to the modification of natural ENSO mode. Overall, the MJO tends to be more active during ENSO neutral years while passive during moderate to strong El Niño and La Niña events. Although it cannot
be the root cause for El Niño or La Niña events, observational studies indicate the MJO effects the ENSO cycle by contributing to their development speed and intensity.

1.3.3 Semi Annual Oscillation

The existence of semi-annual oscillation (SAO) was first reported by Reed (1962). SAO can be defined as an oscillation with a period of approximately six months. In fact, it can be viewed as two linked oscillations approximately out of phase with each other with one peaking near the stratopause and the other near the mesopause (Hirota, 1978). The semi-annual cycle decreases in amplitude in the lower mesosphere, with a minimum near 65km. In the layer between 55 and 70 km, the average time flow is westerly and westerlies in this layer are often interrupted by brief intervals of easterly flow near solstices. Above 65km, the amplitude of SAO tends to increase thereby attaining its second maximum near the mesopause where easterly phase is stronger than that of westerly phase while the average time flow remains easterly. Above 90 km, weak SAO with easterly phase throughout the year while steady westerly flows throughout the year above 105 km have been reported (Garcia, 1997). SAO experiences large inter-annual variations due to the influence of gravity waves, inertia gravity waves, Kelvin waves, diurnal tides and QBO.

1.4 Inter-tropical convergence zone (ITCZ)

Acronym for inter tropical convergence zone (ITCZ) can be defined as a discontinuous band of clouds with thunderstorms in the low pressure belt or equatorial trough that encircles parallel to equator whose location varies over time. A large thermal trough replaces the subtropical ridge over a continent, allowing air to enter into equatorial trough that leads to cloudiness in the ITCZ. It can be viewed as the boundary zone that separates northeast trade winds of the northern hemisphere from the southeast trade winds of the southern hemisphere and its location lies about north of the geographic equator. The north to south variation of the ITCZ as shown in figure 1.4.1 over the oceans exhibits a zonal wind speed of only 4ms\(^{-1}\) through the year which is less than that of continents since the oceans exhibit greater thermal stability.
1.5 SIGNIFICANCE OF KELVIN WAVES

1.5.1 Importance of Kelvin Waves and their implications on upper layers

Obviously convection leads to the generation of tropical waves. Out of several wave activities, Kelvin wave activity is a large scale wave motion generated in the tropical region of Earth’s atmosphere in response to tropical atmospheric circulation from a heat source. It plays an important role in the adjustment of tropical atmosphere to convective latent heat release so that the longitudinal variation of equatorial wave activity depends on zonal flow and convective forcing. When an imposed heating centred on the equator is switched on at some initial time, they carry information rapidly eastward thereby creating easterly trade winds in that region leading to rising motion over the heat source region and sinking motion to its east (Walker circulation). Since the easterly winds are in geostrophic equilibrium, formation of trough along the equator takes place so that winds along the equator tend to move directly down the pressure gradient. This kind of wave motion exhibited by Kelvin wave influences upper atmospheric wave activities such as gravity waves, Rossby-gravity waves, inertia-gravity waves, QBO, ENSO and SAO either by wave-wave interaction or by convective forcing that leads to variation in weather regimes and climatic conditions. It has been reported that Kelvin waves significantly affect the thermal structure around tropical tropopause (Tsuda et al., 1994, Ratnam et al., 2006), such as dehydration as well as cirrus formation (Fujiwara et al., 1998). Production of enhanced turbulence related to Kelvin waves breaking near the tropopause could play an important role in Stratospheric – Tropospheric exchange (Ratnam et al.,2006) which causes a downward mixing of the stratospheric ozone into the troposphere (Fujiwara et al., 2003). Though it is not the right time to generalize their interactive properties but its nature of interaction and their influence on other atmospheric wave activities can be revealed by improved monitoring observations of prevailing climatic conditions either from ground based or satellite based data retrieval systems over a stipulated period of time as have been done earlier and presented by many researchers. Further research on long term basis is still needed.

1.5.2 Importance of them in climate studies

Discrimination of Kelvin waves among different wave activities is a great scientific challenge because of interactions that are associated with other wave activities for improved climate assessment and prediction. Since Kelvin waves are associated with all other major climate forcing wave activities which we have discussed in Kelvin waves section either by driving other oscillations or get forced by other oscillations, it is important to study their nature of interaction with other wave activities in order to improve knowledge of climate forcing and feedback mechanisms to overcome uncertainties.
that are still existing and to constitute a convincing evidence of a cause and effect relationship. Major improvements are needed to understand the role of Kelvin waves in our climate system and our ability to predict climate change. Further, as theoretical derivations of Kelvin wave sensitivity depend so sensitively on other wave activities, such as gravity waves, inertia gravity waves, Rossby-gravity waves, QBO, ENSO, SAO, the best opportunity for major improvement in our understanding of Kelvin wave sensitivity is probably monitoring of its vertical and horizontal propagation with respect to temperature and pressure at different altitudes over global scale. Such measurements would be needed along different latitudinal and longitudinal sections covering both land and oceans. In principle, the measurements would only be needed at decadal intervals, but continuous measurements are highly desirable to average out the global effect on local climate fluctuations.

1.5.3 Instigative envisage

Of great significance is the investigation of the Kelvin wave flow structure at different stages of its passing over layers of atmosphere, taking in to account the specific features of interaction mechanism with other wave activities of that layer. It is necessary to determine the variation of the vertical gradient of temperature near the underlying surface influenced by pavements of the crests and troughs of that layer in order to observe the formation of temperature inversion above that layers and to compare the direction and speed of these waves flowing along different layers and associated interaction mechanism with other wave activities so that we can evaluate the degree of Kelvin wave breaking at different stages of their flow from troposphere to thermosphere and to determine the friction associated with them during their propagation. One of the goals of the investigation is the determination of the limits of the heat island in horizontal and vertical directions, the analysis of their dependence on these limits upon the width of the region and upon the advective factors.

The investigation of temperature field anomalies and of different wave activities under different layers requires combined application of ground and satellite based measurements along with radiosonde, rawinsondes, dropsondes, radar, lidar and airborne measurements. The most effective method of studying atmospheric diffusion of wave activities is satellite based remote sensing due to their vertical propagation. Data on the above measurements will make it possible to obtain information about vertical turbulent transfer of heat and momentum thereby energy potentials associated with waves.
1.5.4 Need of study and scope of the research

Recent studies have speculated that higher wavenumber (s~4-7) eastward and westward propagating equatorially trapped waves along with gravity waves contribute significantly to the total momentum flux transfer, which in turn play a key role in the dynamics of upper atmosphere. However, quantification of the contributions of each of the different waves to the dynamics of middle atmosphere is still very uncertain, primarily due to lack of global database.

Several techniques have been used to study these large-scale wave-structures including ground, rocket-borne and satellite based techniques. It is very true that a significant progress has been achieved in understanding the morphological features of ionospheric irregularities. In order to address effectively the most dynamic features of these waves that need to be studied with an experimental facility that can provide global perceptive along with great resolutions, such as the radio occultation (RO) technique. It is proposed to use COSMIC data base including temperature profiles and to study the possible coupling between lower and upper atmospheric regions and vice-versa. Also proposed to use radiosonde database that is freely available from the Wyoming University (USA)(http://weather.uwyo.edu/upperair/sounding) in order to study the local dynamics at a few strategic locations.

1.5.5 Future studies

Since an RO technique made with multi-satellite GPS receivers placed at low-Earth orbits (LEO) such as COSMIC satellite constellation (which consists six satellites at ~ 800 km altitude) is a relatively powerful technique, it can dramatically change the way the space environment is being sensed now. So, it is envisaged that several unique and important global features of lower and upper atmospheric waves will be revealed. Further, since the COSMIC GPS RO technique can provide increasingly dense temperature profiles from near the surface of the Earth to 40 km altitude, which are distributed fully in longitude and local time (higher spatial and temporal occurrence density and such an higher and globally symmetric occurrence density has not been achieved with any existing experimental techniques), it is also possible to study the global scale wave activities including Kelvin waves and planetary wave activity and gravity wave activity over short-time scales. Therefore, it is proposed to make use of abundant COSMIC GPS RO lower atmospheric data products to study comprehensively typical global characteristics of atmospheric waves, dynamical coupling between lower and upper atmospheric regions, and their interactive efficiencies with the background atmosphere at higher altitudes. Henceforth, the scope of the present research is expected to be so wide.
1.6 IMPORTANT ATMOSPHERIC INDICES:

CONVECTIVE POTENTIAL ENERGY (CAPE):

CAPE is a measure of the degree of convective instability that can potentially lead to deep convection. It can be expressed in Jkg\(^{-1}\) for a reference air parcel. EL designated as equilibrium level, located above the LFC (level of free convection) along the moist adiabat and defined as the level at which the reference parcel undergoing free convection above the LFC is no longer warmer than its environment.

According to the American Meteorological Society (AMS), CAPE is given by,

\[
\text{CAPE} = \int_{P_{LFC}}^{P_B} (\alpha_p - \alpha_e) dp
\]

\[(1)\]

Where, \(\alpha_p\) = specific volume (volume per unit mass for given p and T) of the rising air parcel
\n\(\alpha\) = specific volume of the environmental air at the same level
\nP_{LFC} = pressure where the level of free convection occurs, and
\nP_{EL} = pressure at which the parcel becomes neutrally buoyant.

For real-time measurements of the atmosphere’s profile, eq. 1 must be expressed as a finite number of pressure levels as shown below.

\[
\text{CAPE} = \left( \sum_{P_{LFC}}^{P_B} \frac{\alpha_p - \alpha_e}{\Delta \Delta p} \right) \Delta p
\]

\[(2)\]
Figure 1.6.1 shows the schematic illustration of CAPE with shaded area. Here, the pressure level from which the buoyancy becomes positive is represented as $P_{LFC}$, a level of intersection between the state curve of sounding and the saturated adiabatic attained by a parcel from the ground level ascending through a layer of air that is warmer than itself before it reaches its pressure at lifting condensation level and then continues to go up following pseudo adiabatic curve thereby becoming buoyant with respect to its surroundings and starts accelerating freely upwards, which occurs at the level of free convection. $P_{TOP}$ represents pressure at thermal equilibrium that corresponds to top of CAPE zone. As shown in figure, the pseudo-adiabatic curve meets the curve of state of the particle at $P_{TOP}$.

**CONVective INHIBITION (CIN):**

CIN describes kinetic energy required to lift an air parcel to overcome through a stable surface layer of air that is warmer than itself before it reaches its instable layer which can be calculated from,

$$CIN = GRAVTY * SUMN (DELZ * (TP - TE)/TE)$$

where \(SUMN\) = sum over sounding layers from top of the mixed layer to LFCT i.e level of free convection at virtual temperature for which (TP - TE ) is less than zero; \(DELZ\) =incremental depth; \(TP=\)temperature of a parcel from the lowest 500 m of the atmosphere raised dry adiabatically to the