1 INTRODUCTION

1.1 Motivation

In the realm of Earth’s atmosphere, an inquest of winds and waves is prerequisite to understand the climate variability and for weather prediction. One of the most important dynamical properties of the atmosphere is its ability to support wave motions. Earth’s atmosphere acts as a source and sink for the waves of various scales ranging from planetary to micro scales. These waves transport energy and momentum from one region to another without the transport of material medium thereby impinging the signatures of the source region on to the sink. This is very clear from McIntyre’s [1993] narration that what makes the dynamics of a rotating stratified atmosphere special is the ubiquitous occurrence of wave motions and the fact that wave propagation is generally accompanied by a transport of energy and momentum. Waves effectively couple the Earth’s atmosphere not only horizontally but also vertically through different processes, which have not been explored in depth so far. While the horizontal coupling is important for the conventional meteorology, vertical one unites the Earth’s atmosphere as one single entity which extends from ground to upper atmosphere and farther into space. Each part of the atmosphere is coupled mainly in three ways, viz., chemical, dynamical, and electro dynamical coupling. Among these, chemical coupling involves the exchange of constituents such as water vapor and O$_3$ etc., between the atmospheric regions while electro dynamical coupling deals with the coupling between neutral atmosphere and ionized plasma. In the dynamical coupling, atmospheric waves play an insurmountable role by transporting energy and momentum mainly from below and depositing on the way in their propagation. A better understanding of the vertical coupling by these wave activities will provide a deeper insight into the processes that controls the dynamics and energetics of the whole atmosphere. These facts motivated to take up a detailed study on role of atmospheric waves in coupling the various regions of the Earth’s atmosphere. In
particular, the present study aims to shed light on the dynamical coupling in the Earth’s middle atmosphere (10-100 km), the atmospheric region that separates the domain of meteorology (the troposphere) from the domain of space (the upper atmosphere and above) through gravity waves, which are cynosures of the atmospheric science.

The present chapter highlights the basic structure and dynamics of the atmosphere along with various wave activities and their sources, especially gravity waves and eventually provides a detailed prologue to the dynamical coupling in the middle atmosphere through gravity waves.

1.2 Atmosphere and its Dynamics

The atmosphere is a shallow envelope of compressible gas surrounding an approximately spherical planet. It comprises a stable mechanical mixture of gases, the most important of which are nitrogen (78.08%), oxygen (20.95%), argon (0.93%), water vapor (0-4%) and carbon dioxide (0.036%) These gases occupy 99.98% of the atmosphere by volume below 90 km. Other gases such as ozone, hydrogen, methane and the rare gases like neon, helium, krypton, xenon etc., account for the remaining 0.02% of the air [Wallace and Hobbs, 1977]. Further, the state of the atmosphere is characterized by certain physical quantities like pressure, density, temperature and wind velocity, called field variables, which are assumed to be continuous functions of space and time.

1.2.1 Vertical Structure of the Atmosphere

Atmospheric field variables such as pressure, density and temperature vary with height in the atmosphere. The variation can be over many orders of magnitude and is very much larger than horizontal or temporal variations. The vertical variation of pressure (p) with height (z) may be derived approximately using the hydrostatic equation and ideal gas law as [Wallace and Hobbs, 1977].

\[ p(z) = p(0) e^{-\frac{z}{H}} \]  \hspace{1cm} (1.1)

where \( p(z) \) is the pressure at height \( z \) above sea level, \( p(0) \) is the sea level pressure, and \( H \) is a constant, called the scale height. The variation in density can also be expressed in the similar manner.
According to the vertical structure of the temperature as shown in figure 1.1, the atmosphere can be divided into four distinct layers namely as 'troposphere' (~0-15 km), 'stratosphere' (~15-50 km), 'mesosphere' (50-90 km) and 'thermosphere' (above 90 km). The transition regions between these layers are named as 'tropopause' (~15 km), 'stratopause' (~50 km), and the 'mesopause' (~85 km) respectively. However, these boundaries are not very well defined. The height and temperature of the transition regions exhibit significant temporal, seasonal and latitudinal variations.

![Figure 1.1 Vertical Structure of the atmosphere based on temperature.](image)

In the lower most part of the atmosphere, called troposphere, the temperature decreases with altitude at a nearly constant lapse rate of about 6.5 K/km. Troposphere accounts for more than 80 % of the mass and virtually all of the water vapor, clouds, and precipitation in the Earth’s atmosphere. The tropopause is around 16 km high at the equator while it is at around 8 km at the poles. In the stratosphere temperature increases with altitude and maximizes at stratopause. The observed increase in temperature in the stratosphere is due to the absorption of ultraviolet radiation by ozone (O₃). In addition to the noticeable change in temperature, an abrupt change in the
concentrations of the variable trace constituents is also observed in stratosphere. The temperature again decreases in the height region above the stratopause, which is called the mesosphere. The minimum temperature as low as 180 K over the tropics and about 140 K over the poles is detected at the mesopause height of around 85 km. The mesopause region is the coldest region of the Earth’s atmosphere. This low temperature due to the decrease in ozone heating rate balanced largely by the radiative cooling by CO$_2$ [Andrews et al., 1987; Rodgers et al., 1992]. Nevertheless the radiative cooling by H$_2$O and OH cannot be precluded. The thermosphere lies above the mesopause, where the temperature of the neutral atmosphere rapidly increases because of the absorption of solar extreme ultraviolet radiation and X-rays. The thermosphere, though primarily neutral in terms of dominant constituents, exhibits a significant presence of ionized species, which forms the ionosphere. Ionosphere is commonly applied to the layers above 80 km although sometimes it is used only for the region of high electron density between 100 and 300 km.

In addition to these an important boundary in the atmosphere namely the turbopause occurs at an altitude of ~110 km. Below the turbopause, the atmosphere is in a well mixed condition due to eddy diffusion. The term homosphere or turbosphere is also often used denote this region. The atmosphere above the turbopause, where turbulence minimizes, is called the heterosphere. In this region the gases are in diffusive equilibrium. In other words they are distributed in altitude according to their own weight and thus lighter gases float to the top. Geographically, the atmosphere within ± 30° latitude is termed as ‘equatorial/low’ latitude, from ± 30° to ± 60° as ‘mid’ latitude and ± 60° to ± 90° as ‘high’ latitude atmosphere.

1.2.2 Atmosphere in Motion

Sun is the prime source of energy for all the activities that takes place in the earth. The Earth intercepts a small fraction of the energy that is continually radiated away from the Sun. Because of the spherical shape of the Earth, the amount of energy decreases with increasing latitude. Thus the uneven distribution of the incoming radiation leads to the differential heating of the atmosphere, thus to the creation of a global temperature gradient. The differential heating thus forces the atmosphere into motion,
and its flow patterns. Dynamics of the atmosphere deals with the motions in the atmosphere. For such motions, the atmosphere can be regarded as a continuous fluid medium or continuum ignoring the discrete molecular nature.

There are different forces, which influence the atmospheric motions. The forces, which are of primary concern, are the pressure gradient force, the gravitational force, the frictional force and apparent forces (pseudo forces) like the centrifugal force and the Coriolis force. Further, the basic state of the atmosphere can be expressed by the equations comprising conservation of momentum and energy, the continuity equation and the hydrostatic equation which are governed by the fundamental laws of fluid mechanics and thermodynamics [Holton, 1975] which are given as follows

\[ \frac{DU}{Dt} = -2\Omega \times U - \frac{1}{\rho} \nabla p + g + F_r \]  
(1.2)

\[ \frac{1}{\rho} \frac{D\rho}{Dt} + \nabla \cdot U = 0 \]  
(1.3)

\[ c_v \frac{DT}{Dt} + p \frac{D\alpha}{Dt} = J \]  
(1.4)

\[ dp = -\rho gdz \]  
(1.5)

Equation (1.2) is the statement of Newton’s second law for motion relative to a rotating coordinate frame. The first term in the right hand side of the equation (1.2) represents the Coriolis force, second one the pressure gradient force and \( F_r \) designates the frictional force, and the centrifugal force has been combined with gravitation in the gravity term \( g \). \( \Omega \) is the angular speed of rotation of the earth. It states that the acceleration following the relative motion in the rotating frame equals the sum of the Coriolis force, the pressure gradient force, effective gravity, and friction. This form of the momentum equation is basic to most work in dynamic meteorology. Equation (1.3) is the velocity divergence form of the continuity equation. It states that the fractional rate of increase of the density following the motion of an air parcel is equal to minus the velocity divergence. Equation (1.4) is the usual form of the thermodynamic energy equation. It states that first law of thermodynamics indeed is applicable to a fluid in motion. The second term on the left, representing the rate of working by the fluid system (per unit
mass), represents a conversion between thermal and mechanical energy. This conversion process enables the solar heat energy to drive the motions of the atmosphere. Equation (1.5) known as the hydrostatic approximation, states that in the absence of atmospheric motions the gravity force must be exactly balanced by the vertical component of the pressure gradient force. This condition of hydrostatic balance provides an excellent approximation for the vertical dependence of the pressure field in the real atmosphere.

The governing equations of the atmosphere can be written in terms of perturbations in the field variables. These equations are then linearised by neglecting the second order terms of products of perturbations. General solutions to these equations are complex. The complexity of the equations describing the different scales of atmospheric oscillations can be simplified by scale analysis, which involves the investigation of relevant orders of magnitude of various terms for a particular type of motion. The different modes of atmospheric oscillations are assumed to be wavelike and are obtained as eigen solutions of the linearised perturbation equations.

1.3 Earth’s Middle Atmosphere

Earth’s middle atmosphere (approximately the region between tropopause and mesopause), as mentioned earlier, is the coupling region between the troposphere where the weather processes and most of the dynamical processes are generated and the upper atmosphere where the atmospheric structure, composition and dynamics are directly controlled by the solar EUV radiation. This region holds the key to the understanding of the lower atmosphere, having tremendous influence on the human activities, health and way of life. This intermediate region while getting influenced by different physical processes occurring above and below, processes taking place here, in turn, influence the lower and upper atmosphere. The energy transport and dispersal in the middle atmospheric region are through wave motions. Atmospheric waves generated by dynamical processes propagate vertically and horizontally, dissipate, interact non-linearly, and profoundly influence the flows of momentum, energy, and constituents on a global basis.
1.4 Equatorial Middle Atmosphere

Many studies suggested that the equatorial oscillations influence the global atmospheric circulation and structure in prominent ways. The behavior of the equatorial middle atmosphere is also important on the global change of the Earth’s environment. Dynamics of the equatorial middle atmosphere differs significantly from mid and high latitudes. The Coriolis parameter is very small in the equatorial (±20°) latitudes. The small Coriolis parameter at low latitudes leads to a breakdown of the validity of the geostrophic approximation for the wind and invalidates the quasi-geostrophic theory that is so useful in explaining the large-scale circulation of the extra tropical atmosphere. The solar insolation and hence the heating of the low latitude atmosphere by the absorption of water vapor and ozone is maximum. The distribution of sea and land masses in different longitudes in the equatorial latitudes produces uneven heating and this produces east-west circulation in the vertical plane along the equator.

Particularly the Indonesian region is spread out in longitude along the equator. Similarly the African subcontinent and American subcontinent in the equatorial region are heated up more than the ocean areas in between. These localized heat sources generate the localized circulation cells in the vertical plane along the equator. The release of latent heat in cumulus convection is considered as a primary energy source for the maintenance of the equatorial disturbances. There is an interaction between the cumulus scale and large-scale disturbances, wherein, the large-scale convergence provides moisture for the convection and cumulus cells provide the large-scale heat source [Horel and Wallace, 1981]. Because of the special nature of the driving force, as well as, the smallness of the Coriolis parameter, the large-scale equatorial atmospheric dynamics have certain distinct characteristic structural features, which are quite different from those of the mid-latitude system.

1.4.1 Atmospheric Waves

As briefed in the previous section, various types of wave motion found in the atmosphere represent solutions of a simplified form of the equations of motion in which the nonlinearities are removed. These waves can be categorized in various ways,
according to their physical, geometrical properties and restoring mechanisms. The restoring mechanisms for the waves are associated with physical properties such as the density stratification of the atmosphere, which give rise to buoyancy forces, the rotation of the earth (leading to Coriolis force) and more subtly, the vorticity coupled with the near-spherical geometry of the planet. Depending upon the wave parameters it is possible to understand the physics of each class of waves by suitably neglecting the forces which are relatively unimportant. The most familiar fluid dynamical wave is the sound wave, in which the restoring mechanism is due to the compressibility of the gas. The planetary waves, tides and gravity waves are the three most important wave types that result from the various forces in the atmosphere. Of these, planetary waves result from the beta-effect (change of Coriolis force with latitude), the buoyancy or internal gravity waves owe their existence to stratification while inertia-gravity waves result from a combination of stratification and Coriolis force. These wave motions, with quite different physics, strongly influence atmospheric behavior in various ways. Propagating waves can be characterized by the amplitude and phase as well, but with phase depending on the time and also on space coordinates. The phase speed is given by \( \omega/K \) where \( \omega \) the frequency of the wave and \( K \) is the wave number. For propagating waves \( \omega \) generally depends on \( K \) as well as on the physical properties of the medium. Waves for which \( \omega \) depends on \( K \) are referred to as dispersive. In the non dispersive waves, a transient consisting of a number of sine waves group will preserve its shape as it propagates. For dispersive waves, the group speed is generally different from the phase speed given by \( \delta\omega/\delta K \).

These atmospheric waves can also be classified on the basis of the following properties. (1) extra tropical modes against equatorially trapped modes (2) free modes and forced modes (3) external modes and internal modes and (4) modes that interact with the mean flow through wave dissipation. Waves of primary importance for middle atmospheric dynamics are forced internal modes which are excited by various processes in the troposphere and propagate vertically into the middle atmosphere. These waves carry information over thousands of kilometres horizontally and tens of kilometres vertically, so that forces imposed, say, at ground level may ultimately lead to deviations of the flow in the stratosphere and even in the mesosphere. Studies have been carried out
on atmospheric waves from the point of view of energy coupling from the middle atmosphere to the upper atmosphere [e.g., Andrews et al., 1987]. The atmospheric wave motions of various scales which are dominant in the equatorial middle atmosphere are elaborated in subsequent section.

1.4.2 Rossby Waves

The zonally asymmetric component of the circulation in the winter hemisphere middle atmosphere is dominated by quasi-stationary Rossby waves of zonal wave number 1 and 2. Rossby waves are of large enough scale to be influenced by the curvature of the Earth as well as its rotation. In a barotropic atmosphere, the Rossby wave is ‘absolute vorticity’ conserving motion, which owes its existence to the variation of Coriolis force with latitude, the beta effect. They have horizontal wavelengths ranging from hundreds of kilometres to the circumference of the earth and vertical scales of tens of kilometres. Rossby waves propagates westwards relative to mean flow. These waves are merely the upward extensions of tropospheric planetary waves generated by topographic forcing and by land-sea diabatic heating contrasts. Although both type of forcing are strongest during the winter season, the confinement of these waves to the winter hemisphere in the middle atmosphere is not due to seasonal changes in the forcing, but rather to the strong dependence of the vertical wave transmission on the mean zonal wind. Likewise, oscillations in wave amplitude and phase may occur not only as a result of oscillations in the forcing but also in response to changes in the transmission characteristics of the middle atmosphere due to mean zonal wind changes.

1.4.3 Equatorial Waves

In the equatorial latitude, there are two significant forced oscillations, one propagating eastward (westerly mode) and another westward (easterly mode). The westerly mode, known as the Kelvin wave is symmetric about the equator. The easterly mode is antisymmetric about the equator and is called Mixed Rossby Gravity waves (MRG). Both the wave modes are latitudinally trapped, i.e., their propagation which is essentially zonal in the horizontal plane is confined to about ±20° latitude. The observed characteristics of equatorial waves [Andrews et al., 1987] are summarized in Table 1.1.
<table>
<thead>
<tr>
<th>Sl. No.</th>
<th>Characteristics</th>
<th>Kelvin waves</th>
<th>MRG waves</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Zonal wave number</td>
<td>1 – 2</td>
<td>4</td>
</tr>
<tr>
<td>2</td>
<td>Period (days)</td>
<td>10 – 20</td>
<td>4 – 5</td>
</tr>
<tr>
<td>3</td>
<td>Phase speed (m s(^{-1}))</td>
<td>25</td>
<td>-23</td>
</tr>
<tr>
<td>4</td>
<td>Direction of propagation</td>
<td>Westerly</td>
<td>Easterly</td>
</tr>
<tr>
<td></td>
<td>(Eastward)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Horizontal wavelength (km)</td>
<td>30,000</td>
<td>10,000</td>
</tr>
<tr>
<td>6</td>
<td>Latitudinal wavelength (km)</td>
<td>1,000</td>
<td>1,000</td>
</tr>
<tr>
<td>7</td>
<td>Vertical wavelength (km)</td>
<td>6 – 10</td>
<td>4 – 8</td>
</tr>
<tr>
<td>8</td>
<td>Amplitude</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(a)</td>
<td>Zonal wind ‘(u)’ (m s(^{-1}))</td>
<td>8</td>
<td>2 – 3</td>
</tr>
<tr>
<td>(b)</td>
<td>Meridional wind ‘(v)’ (m s(^{-1}))</td>
<td>0</td>
<td>2 – 3</td>
</tr>
<tr>
<td>(c)</td>
<td>Vertical wind ‘(w)’ (cm s(^{-1}))</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>(d)</td>
<td>Geopotential height ‘(z^*)’ (m)</td>
<td>4</td>
<td>30</td>
</tr>
<tr>
<td>(e)</td>
<td>Temperature ‘(T)’ (°K)</td>
<td>2 – 3</td>
<td>1</td>
</tr>
<tr>
<td>9</td>
<td>Symmetry about the equator</td>
<td>Even</td>
<td>Odd</td>
</tr>
<tr>
<td>(a)</td>
<td>Zonal wind</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(b)</td>
<td>Meridional wind</td>
<td>-</td>
<td>Even</td>
</tr>
<tr>
<td>(c)</td>
<td>Geopotential height</td>
<td>Even</td>
<td>Odd</td>
</tr>
<tr>
<td>10</td>
<td>Phase relation between the</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>perturbation parameters</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(a)</td>
<td>Pressure ‘(P)’ and ‘(u)’</td>
<td>Out of phase</td>
<td>In phase</td>
</tr>
<tr>
<td>(b)</td>
<td>Pressure ‘(P)’ and ‘(T)’</td>
<td>‘(T)’ leads ‘(P)’ by 90°</td>
<td>‘(T)’ leads ‘(P)’ by 90°</td>
</tr>
<tr>
<td>(c)</td>
<td>Pressure ‘(P)’ and ‘(v)’</td>
<td>-</td>
<td>‘(P)’ leads ‘(v)’ by 90°</td>
</tr>
<tr>
<td>(d)</td>
<td>‘(u)’ and ‘(w)’</td>
<td>In phase</td>
<td>In phase</td>
</tr>
</tbody>
</table>

Table 1.1 Characteristics of the dominant observed planetary-scale waves in the Equatorial lower stratosphere [After Andrews et al., 1987].

1.4.4 Atmospheric Tides

Another important wave motion in the equatorial middle atmosphere is the atmospheric tides. Atmospheric pressure, temperature, density and winds are all subject to variations with 24-hour (diurnal) and 12-hour (semidiurnal) periods. The atmospheric tides are excited by the heating of the atmosphere due to the absorption of solar radiation by water vapor primarily in the troposphere [Lindzen, 1967], ozone in the stratosphere
and mesosphere [Lindzen, 1967] and ionized oxygen in the ionosphere [Harris and Mayr, 1975]. The heating of the atmosphere has a strong altitude and latitude variation as well. The small but measurable, variations of atmospheric parameters with lunar semidiurnal period are caused by the gravitational attraction between the moon and the earth.

The amount of atmospheric minor constituents that absorb solar radiation can fundamentally be assumed not to show a significant variation in the longitudinal distribution. Therefore, the corresponding tidal generation also becomes zonally homogeneous, so the tides propagate westward, synchronizing with the apparent motion of the sun. This sun-synchronous component is commonly called the migrating tide. In addition to the sun-synchronous tides, non-migrating tides can be generated by a localized excitation source, such as longitudinal non-uniformity of water vapor content and the cloud convective activity as well as a land-sea contrast in the heat transfer process within the planetary boundary layer [Kato et al., 1982; Forbes and Groves, 1987; Tsuda and Kato, 1989; Hsu and Hoskins, 1989; Williams and Avery, 1996 a, b; Ekanayake et al., 1997, Hagan et al., 1997]. These non-migrating tidal modes do not follow the motion of the sun. The non-migrating tides consist of various zonal wave numbers, propagating both westward and eastward, or standing [Tsuda and Kato, 1989]. The propagating wave components of atmospheric tides transport momentum and wave energy upward from the regions of excitation, they then dissipate by various instabilities and wave-wave or wave-mean flow interactions. Theoretical [Chapman and Lindzen, 1970] modeling [Forbes, 1982a, b; Miyahara et al., 1993; Hagan et al., 1995] and observational [e.g., Portnyagin et al., 1993; Pancheva et al., 2002] studies have provided significant insight into atmospheric tides.

1.4.5 Gravity Waves

Gravity waves in the atmosphere arise from the coupling of gravity with the continuous vertical density stratification. They are also referred to as buoyancy waves since they have the buoyancy force as their restoring force. Atmospheric gravity waves can exist only when the atmosphere is stably stratified, so that a fluid parcel displaced vertically will undergo buoyancy oscillations. In a fluid such as the ocean which is bounded both above and below, gravity waves propagate primarily in the horizontal plane.
since vertically propagating waves are reflected from the boundaries to form standing waves. However, in atmosphere, gravity waves propagate vertically as well as horizontally.

Gravity waves may be trapped or vertically propagating depending on the background wind, buoyancy characteristics of the atmosphere and their horizontal wavelength. Horizontal propagation occur when the waves are trapped between the ground surface and some upper level. These types of waves have been called ‘cellular’ to denote waves that propagate horizontally and present a standing pattern in the vertical direction. In vertically propagating waves, the phase is a function of height. Such waves are referred to as internal gravity waves. Internal gravity waves are observed in the atmosphere, at scales ranging from few metres to few kilometres.

As mentioned earlier, the existence of gravity waves requires stable atmospheric conditions. Atmospheric stability is defined as the ability of the atmosphere to enhance or to resist atmospheric motions. The stability of the atmosphere can be determined in a simpler way, using the dry adiabatic lapse rate ($\Gamma_d$), the rate at which the temperature of the parcel of air changes with height when it rises up adiabatically. The environmental lapse rate is designated as $\Gamma$. Table 1.2 and Figure 1.2 illustrate the stable, unstable and neutral atmospheric conditions.

<table>
<thead>
<tr>
<th>LAPSE RATE</th>
<th>ATMOSPHERIC STABILITY</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\Gamma &gt; \Gamma_d$</td>
<td>Unstable</td>
</tr>
<tr>
<td>$\Gamma = \Gamma_d$</td>
<td>Neutral</td>
</tr>
<tr>
<td>$\Gamma &lt; \Gamma_d$</td>
<td>Stable</td>
</tr>
</tbody>
</table>

Table 1.2 Stability of the atmosphere based on lapse rate

From figure 1.2, if the environmental lapse rate exceeds the dry adiabatic lapse rate of 10°C/km (curve a), then a parcel of air will remain warmer than its environment and continue to be buoyant. Such an environment is said to be unstable. If,
on the other hand, the environmental lapse rate is less than 10°C/km (curve b), a rising parcel will become cooler than its environment and settle back to its original height. We therefore say that an environment which has a lapse rate which is less than the adiabatic lapse rate, is stable. Finally, if the environmental lapse rate equals the adiabatic lapse rate, the environment is said to be neutral since any vertical displacement of the parcel will not change its temperature relative to the environment. In such stable condition the parcel moves adiabatically, which means it heats and cools according to the adiabatic lapse rate, and undergoes buoyant oscillations about its equilibrium position. If the atmosphere is unstable, gravity waves cannot exist, since a parcel of air which is displaced vertically will continue to rise, rather than oscillate.

![Figure 1.2](image_url)  
**Figure 1.2 Temperature change with height in a stable, unstable and neutral atmosphere and for a parcel cooling adiabatically while ascending.**

The restoring buoyancy force acting on a vertically displaced fluid particle is characterized by the Brunt-Väisälä (or buoyancy) frequency N defined (in an incompressible fluid) by

\[
N^2(z) = \frac{g(z)}{T_0(z)} \left[ \frac{\partial T_0(z)}{\partial z} + \frac{g(z)}{C_p} \right]
\]

(1.5)
where \( g(z) \) is acceleration due to gravity, \( T_0(z) \) is mean temperature at an altitude \( z \) and \( C_p \) is the specific heat capacity of the air at constant pressure. \( N \) represents the natural frequency of oscillation of a parcel displaced vertically from its equilibrium altitude. From equation (1.2) it is seen that \( N \) is larger (smaller) when temperature increases (decreases) with altitude. Thus the stratosphere and lower thermosphere are more stratified than the troposphere and mesosphere. For average tropospheric conditions, \( N \approx 1.2 \times 10^{-2} \text{ s}^{-1} \), so the period of oscillation is 8 min. Thus internal gravity waves whose frequencies lie in the range \( N \geq \omega \geq f \) can only exist in the atmosphere, where \( \omega \), the frequency of the gravity wave, \( f \) is the coriolis frequency and is given by \( f = 2\Omega \sin \phi \) where \( \Omega \) is the angular frequency of the Earth and \( \phi \), the latitude.

Gravity waves with horizontal scales greater than a few hundred kilometres and periods greater than a few hours are hydrostatic, but they are influenced by the Coriolis effect. When the flow is both inertially and gravitationally stable, parcel displacements are resisted by both rotation and buoyancy. The resulting oscillations are called inertia–gravity waves. Inertia–gravity wave frequencies must lie in the range \( f \leq |\omega| \leq N \). For typical mid latitude tropospheric conditions, inertia–gravity wave periods are in the approximate range of 12 min to 15h.

Gravity waves play a key role in the middle atmosphere for several reasons. They are ubiquitous throughout the atmosphere, arising in response to a variety of sources in the troposphere and middle atmosphere. More importantly in the middle atmosphere, gravity waves propagate readily over large distances and depths, account for substantial fluxes of energy and momentum and experience filtering and dissipation due to interaction with the environment through which they propagate. Thus, the present thesis work mainly focuses on the gravity waves, their vertical propagation characteristics and the processes through which they couple middle atmosphere with rest of the atmosphere. Figure 1.3 shows the signature of gravity waves in low-level tropospheric clouds (a), noctilucent clouds (b) and airglow images (c) in the mesospheric region.
Figure 1.3 Observed gravity wave and instability structures in (a) Mountain waves (b) NLC [After Fritts et al., 1993a], and (c) Airglow images.

Gravity waves play a very intriguing role in transporting momentum and energy right from the troposphere to the upper atmosphere, which contribute to the shaping and structure of atmospheric circulation [Hines, 1960; Eliassen and Palm, 1961]. The gravity wave drag effect on the large-scale flow is important for the dynamic coupling of the lower and middle atmosphere. Vertically propagating gravity waves carry vertical flux of horizontal momentum with horizontal momentum
transferred to the mean flow whenever the wave transient or experience attenuation (by dissipation, saturation, wave breaking, or absorption at a critical level). Momentum deposition by gravity waves in the middle and upper atmosphere provides significant forcing to the mean flow. Also, break down of gravity waves and resulting turbulence has shown that this mechanism is crucial in understanding the general circulation of the middle atmosphere [Lindzen, 1981; Holton, 1982; Mastuno, 1982]. Three-dimensional general circulation models are, however, sensitive to the details of the geographical and temporal distribution of wave activity and so require a more detailed description of the gravity waves and their sources. Various sources of gravity waves and their effects in the middle and upper atmosphere are illustrated in figure 1.4.

![Figure 1.4 Schematic diagram showing the sources of gravity wave and their effects in the middle and upper atmospheric region.](image)

### 1.5 Major Sources of Gravity Waves

The majority of gravity waves is tropospheric or stratospheric in origin, and can be attributed to one of three major sources: orographic forcing, convective activity and shear instability of the general flow. In addition to this, other sources like unbalanced flows in the vicinity of jet streams and frontal systems and wave-wave
interaction also generate these waves. Gravity waves can also be generated from the transient sources like nuclear explosion, earthquake [Row, 1967] and solar eclipse [Chimonas, 1970]. Since the mean winds and patterns of precipitation exhibit seasonal changes, it is reasonable to expect seasonal changes in the generation of gravity waves. It is seen that waves generated by topography and convection are characterized by a wide range of frequencies, vertical and horizontal scales and phase speeds.

1.5.1 Topography

Topography is a geographically limited source of gravity waves. When the airflow is forced to rise and pass over the mountain, the air is displaced from its original equilibrium level. Once the air parcel passes the crest of the mountain, restoring forces will push it back toward its equilibrium level. Passing over the mountain initiates a buoyancy oscillation. Since the parcel continues to move horizontally, the oscillation occurs over a distance as a wave. This wave is termed as an orographically forced gravity wave. These waves, in the presence of a steady wind, will have no traveling component and are likely to generate a discrete wave spectrum. By definition they have non-zero intrinsic frequencies, if Doppler shifted by the mean zonal winds. Thus, even though their source is fixed to the Earth, wave activity can propagate upstream and downstream relative to the mean wind. If the mean flow varies with time, a broad spectrum of waves can be generated. Gravity waves generated by orography have been studied extensively over the past three decades using theoretical, numerical, and observational methods. In a very recent work, Alexander et al., [2008] gave a clear evidence for the propagation of mountain waves up to the mesosphere using High Resolution Dynamics Limb Sounder (HIRDLS) observations.

1.5.2 Convection

Convection plays a vital role in exciting gravity waves in the tropical latitudes. Deep convection involves strong transient vertical motions in a conditionally stable environment, which is an important source mechanism for the generation of gravity waves. Satellite based estimates of deep convective activity in the tropics have been correlated with gravity wave energy in the stratosphere. The major difficulty in
characterizing these convectively generated gravity waves arises due to the inherent intermittency of the source and triggered waves have wide spectrum of phase speeds, wave frequencies, vertical and horizontal wavelengths unlike the waves generated by topography which have a single prominent phase speed or frequency. Some seminal works on generation of gravity waves have shown that the deep fast convection excites gravity waves having large vertical scales, high intrinsic frequencies and phase speeds [Fovell et al., 1992; Dewan et al., 1998; Piani et al., 2000; Lane et al., 2001; Horinouchi et al., 2003; Dhaka et al., 2002; Roettger et al., 2003; Sentman et al., 2003; Kumar, 2007a]. Three mechanisms are proposed to describe the convectively generated gravity waves (CGWs) as follows

1. Thermal forcing
2. Obstacle effect
3. Mechanical oscillator effect

In thermal forcing, the latent heat released within the convective storm acts as forcing mechanism of gravity waves. Through a number of numerical experiments, Salby and Garcia [1987] showed that the dominant tropospheric vertical wavelength of waves generated by a thermal forcing is approximately equal to twice the vertical extent of the heating. This dependence of vertical wavelength of gravity waves on the depth of the heating region has been confirmed by the observations as well as by other numerical studies [Alexander et al., 1995; Pandya and Alexander, 1999; Vadas and Fritts, 2001; McLandress et al., 2000; Piani et al., 2000]. Another likely mechanism for the generation of CGWs is the obstacle effect which is akin to the topographic effect. In the obstacle effect, the pressure field produced by an updraft or a downdraft within a convective system, acts as an obstacle to the mean horizontal flow and thus generates gravity waves. This particular mechanism requires persistent updrafts or downdrafts and a mean flow relative to the convective system [Clark et al., 1986]. Subsequent modeling and observational studies gave further evidence for the waves generated by obstacle effect [Pfister et al., 1993; Vincent and Alexander, 2000; Alexander and Vincent, 2000]. On similar lines, Clark et al., [1986] suggested that oscillating updrafts and downdrafts impinging on the interface between the unstable and
stable regions cause displacements of isentropes, which are oscillatory in nature at the base of the stable layer and will excite vertically propagating gravity waves. This mechanism can be called as mechanical oscillator effect since it is very similar to mechanical oscillatory forcing [Fovell et al., 1992]. The gravity waves generated by the mechanical oscillator effect have frequencies equal to that of the oscillation frequency of the periodic and localized momentum source that generate these waves. In a very recent study using VHF radar observations during deep convective events, Kumar [2007a] experimentally corroborated the role of mechanical oscillator effect in exciting convectively generated gravity waves.

Among these mechanisms proposed for the generation of gravity waves, Clark et al., [1986] pointed out that the obstacle effect is the dominant mechanism while Fovell et al., [1992] using numerical simulation studies concluded that the mechanical oscillator effect is more prominent one. However all the three mechanisms strongly depend on the local shear, vertical profile and time dependence of the latent heating. Each of this mechanism may be helpful in getting deeper insight into the characteristics of wide spectrum of gravity waves generated. Though for decades, it is known that convection can excite gravity waves, there is still controversy and ongoing research aimed at understanding this generation mechanism.

1.5.3 Wind Shears

The excitation of gravity waves from wind shears is another interesting mechanism, which has been investigated for so many years, but least understood when compared to convective and topographically generated gravity waves. Unlike other tropospheric sources of gravity waves, namely, convection and topography, wind shears act as a potential source of gravity waves right from the troposphere to the thermosphere. Shear instability can generate gravity waves when the vertical shear of horizontal velocity exceeds some critical value with respect to the vertical restoring force (Richardson number \( \leq 0.25 \)). Studies have shown enhancement in the gravity wave variances in the vicinity of the jet stream shears and suggested significant contribution to the momentum budget of the middle atmosphere [Fritts and Nastrom,
Other sources of gravity waves which are important for the middle atmospheric dynamics are the frontogenesis and baroclinic instability \cite{Sullivan1995, Griffiths1996, Reeder1996}. Gravity waves arising from the local body forces due to dissipating gravity waves were considered theoretically by \textit{Zhu and Holton} \cite{Zhu1987} and \textit{Vadas and Fritts} \cite{Vadas2001}. \textit{Fritts and Nastrom} \cite{Fritts1992b} provided a comparison of gravity wave velocity and temperature variances with various sources computed using Global Atmospheric Sampling Program (GASP) data near the tropopause and concluded that all these sources contribute substantially to gravity wave excitation where these sources are prevalent.

\textbf{1.6 Gravity Wave Saturation and Breaking}

As mentioned earlier, gravity waves grow in amplitude as they propagate upwards into higher regions of the atmosphere. In the absence of any dissipation, the wave amplitude grows with altitude as the wave propagates into the higher regions of the atmosphere. Wave breaking occurs when the wave approaches the critical level (where its phase velocity is equal to the background wind velocity) or when the wave amplitude grows sufficiently and become saturated \cite{Fritts1984}. In general, saturation refers to any mechanisms that act to limit or reduce wave amplitudes due to their superposition or growth with height. The breaking or dissipation of atmospheric gravity waves due to their exponential growth with height was first suggested by \textit{Lindzen} \cite{Lindzen1967}. There exist basically three theoretical treatments of the wave saturation, which can be categorized as linear and non linear.

The simplest theory conceptually on the gravity wave saturation and breakdown is the linear saturation theory developed by \textit{Lindzen} \cite{Lindzen1981}. The basic premise of the linear theory is that gravity waves grow exponentially with height until the perturbation lapse rate becomes sufficiently large to cancel the mean (stable) lapse rate. Once this condition is attained, convective instabilities occur which act to dissipate the wave motion and cancel further wave growth. The induced wave dissipation results in a
vertical divergence of the gravity wave momentum flux, causing the acceleration/deceleration of the medium towards the phase velocity of the wave. The other two involve non linear interactions among the components of the gravity wave spectrum. One of the theories invokes the wave-wave interactions at the small scale end of the gravity wave spectrum to produce non linear diffusion of the energy from the lower scale waves [Weinstock, 1990]. The other theory by Hines [1991a, b] employs non linear Doppler spreading of medium scale waves into small scale end of the spectrum by the winds of the larger scales coupled with dissipation at the smallest scale of the spectrum. These theories distinguish between the unsaturated and saturated parts of the vertical wave number (m) spectrum. At small m (large scale) the spectrum is unsaturated and the gravity waves in this regime are free to grow exponentially with altitude due to decrease in density. On the other hand at larger m (short scale) regions of the spectrum the saturation produces an m⁻³ spectral form. Wave amplitudes can grow large enough in the mesosphere that waves with large intrinsic phase speeds can break, while in the stratosphere, where the wave amplitudes are smaller, breaking or dissipation will tend to occur only as waves approach their critical levels unless the amplitudes are large. Theoretical and observational studies have emphasized that upward propagating gravity waves which carry energy and momentum flux from lower atmosphere to middle atmosphere play an important role in maintaining the general circulation by providing dynamical stress due to breaking [e.g., Lindzen, 1981; Holton, 1982; Matuno, 1982; Tsuda et al., 1990].

1.7 Wave-Mean Flow Interaction

Once the gravity wave breaks or saturates, it cannot grow further and deposits the energy and momentum flux to the background mean flows (i.e. winds) of varying magnitudes and directions which in turn accelerate or decelerate the mean flow. The interaction between the waves and the mean flow in general involve four main processes: (i) reflection of waves, (ii) dissipation and deposition of wave momentum, (iii) critical level absorption of wave energy/momentum and (iv) destabilization of atmosphere by waves. Dissipation results from processes such as radiative damping [Fels, 1984; Zhu, 1994], wave-wave and wave-mean flow interaction [Fritts and Alexander,
wave breaking and instability processes. The relative importance of each process varies greatly in space and time, depending upon the details of the wave structure and mean flow structure at a given location at a given time. The quasi-stationary Rossby waves interact with the mean flow primarily through the mechanism of wave transience (local changes in amplitude) while the equatorial waves interact with the mean flow primarily through the mechanism of wave dissipation (thermal or mechanical damping). But the above interactions play a fundamental role in the global scale, mesoscale and micro scale dynamics of the atmosphere.

The interaction of gravity waves with the background flow at higher altitudes is believed to be partly responsible for the generation of typical global-scale oscillations, such as Quasi-Biennial Oscillation (QBO) in the stratospheric and mesospheric region (i.e. SQBO and MQBO) [e.g. Baldwin et al., 2001] and Semi Annual Oscillation (SAO) [e.g., Garcia et al., 1997] in the stratospheric and mesospheric regions (i.e. SSAO and MSAO).

**Figure 1.5** Vertical distribution of the amplitude of the MQBO, MSAO, SSAO, SQBO, and annual component at the equator [After Baldwin et al., 2001].
The amplitude structure of various wave induced long period oscillations in the equatorial middle atmosphere is depicted in figure 1.5 along with the annual oscillation resulting from radiative forcing. In the figure, MQBO is based on Upper Atmosphere Research Satellite (UARS)/ High-Resolution Doppler Imager (HRDI) observations [Burrage et al., 1996], SAO is based on rocket observations at Ascension Island [Hirota, 1978], and the annual component is after COSPAR International Reference Atmosphere (1986). The amplitude of the wave-induced oscillations (QBO and SAO) is found to be more than annual oscillations. These interactions are quite important because it is the small-scale processes of atmospheric waves generated in the lower atmosphere that control the large-scale process of global circulation in the middle atmosphere [Garcia and Solomon, 1985]. The basic mechanism in the wave mean flow interaction is theoretically explained by Plumb [1977, 1984], which is illustrated in figures 1.6a and 1.6b.

Figure 1.6 Schematic representation of the instability of zonal flow in a stratified fluid with standing wave forcing applied at a lower boundary. Zonal mean velocity ($\bar{U}$) vs. height ($z$) [After Plumb, 1984]. (a) Onset of instability from a small zonal flow perturbation. (b) Early stages of subsequent evolution. Double arrows: Approximate location and direction of maximum acceleration. Wavy lines: Schematic representation of penetration of wave components.

In his model, Plumb considered a vertically unbounded stratified fluid subject to standing wave forcing at its lowest boundary by two traveling waves of equal
amplitude and oppositely directed phase speeds. This forcing generates vertically propagating internal gravity waves, which propagate upward into the fluid wherein they dissipate by a weak Newtonian cooling with constant rate $\alpha$. In figure 1.6a, a westerly meanflow perturbation has been introduced. In the regions of westerly flow, the Doppler shifted phase speed $(c-\overline{U})$ of the westerly wave is reduced ($c$ is the phase velocity of the wave and $\overline{U}$ is the mean zonal flow velocity). The vertical group velocity is proportional to $(c-\overline{U})^2$. When the Doppler shifted phase speed decreases, the vertical group velocity decreases and there will be a larger time available for the energy to be damped for a given vertical propagation distance. Conversely the easterly wave is attenuated less rapidly, since $c+\overline{U}$ is increased. Therefore westerly acceleration dominates in this region, as shown in figure 1.6a. As a result of this selective attenuation of the westerly wave, however, at high levels, the easterly wave dominates and therefore the net acceleration becomes easterly there. Plumb [1977] showed that the level of maximum acceleration is below the level of maximum $\overline{U}$, consequently, the level of maximum $\overline{U}$ must descent with time. This instability can lead to a finite amplitude mean wind oscillation resembling QBO as shown in figure 1.6b.

The subsequent evolution of the flow is depicted schematically in figure 1.7. The shear zone separating the low-level westerly regime from the easterly flow above becomes increasingly narrow with time, (figure 1.7a) so that viscous diffusion across the shear zone destroys the low-level westerly. This has the effect of destroying the westerly regime from above and therefore ‘switching’ the low-level flow into an easterly regime (figure 1.7b). When this occurs, the westerly wave is no longer attenuated at low levels and penetrates to greater heights. Consequently a high level westerly acceleration of the mean flow takes place. Eventually (figure 1.7c) a westerly regime develops there, which moves downward until the interior shear layer again becomes narrow enough for diffusion to act (figure 1.7d). The easterly waves can then penetrate to higher levels (figure 1.7e), where it initiates a new easterly plane of oscillation (figure 1.7f). Thus the interaction between upward propagating dissipating waves and the mean flow generates a new oscillation with finite amplitude, which depends primarily on the
momentum flux and vertical decay scale of the waves. This mechanism was demonstrated in a laboratory experiment by Plumb and McEwan [1978].

Figure 1.7. Schematic representation of evolution and structure of fully developed flow (following from figure 1.6). Six stages in one complete cycle of the mean flow oscillation. Double arrows show wave-driven acceleration, and single arrows show viscously driven accelerations. Wavy lines indicate relative penetration of eastward and westward waves [After Plumb, 1984].
1.7.1 Quasi-Biennial Oscillation (QBO)

QBO is the most exciting outcome of the wave-mean flow interaction in the tropical lower stratosphere. Plate 1.1 provides an overview of the QBO, its sources, and its global dynamical effects. The diagram spans the troposphere, stratosphere, and mesosphere from pole to pole and shows schematically the differences in zonal wind at 40 hPa easterly and westerly phases of the QBO.

Plate 1.1 Dynamical overview of the QBO during northern winter [After Baldwin et al., 2001].

Convection in the tropical troposphere, ranging from the scale of mesoscale convective complexes (spanning more than 100 km) to planetary-scale phenomena, produces a broad spectrum of waves (orange wavy arrows), including gravity, inertia-gravity, Kelvin, and Rossby-gravity waves. These waves, with a variety of vertical and horizontal wavelengths and phase speeds, propagate into the stratosphere, transporting easterly and westerly zonal momentum. Most of this zonal momentum is deposited at stratospheric levels, driving the zonal wind anomalies of the QBO. For each wave, the vertical profile of the zonal wind determines the critical level at or below which the momentum is deposited. The critical levels for these waves depend, in part, on the shear zones of the QBO. Some gravity waves propagate through
the entire stratosphere and produce a QBO near the mesopause known as the mesospheric QBO, or MQBO.

To the north of the equator in the lower stratosphere, tropical winds alter the effective wave-guide for upward and equatorward propagating planetary-scale waves (curved purple arrows). Equatorward propagating waves originating outside the tropics, such as planetary Rossby waves from the winter hemisphere, may have some influence in upper levels of the QBO [Ortland, 1997]. The lower region of the QBO (20–23 km) near the equator is relatively well shielded from the intrusion of extratropical planetary waves [O’Sullivan, 1997]. Though QBO is believed to be driven by dissipation of planetary scale waves [Baldwin et al., 2001], studies have shown that wave momentum provided by these planetary scale waves is not sufficient to drive QBO [Dunkerton, 1997; Piani et al., 2000] and suggested an additional wave momentum.

1.7.2 Semi Annual Oscillation (SAO)

The semiannual oscillation (SAO) of mean zonal winds in the tropical stratosphere and mesosphere was first documented by Reed [1961, 1962, 1966]. The SAO is actually two linked oscillations approximately out of phase with each other, one peaking near the stratopause and the other near the mesopause, as shown by Hirota [1978] (figure 1.5). Satellite measurements made by the HRDI onboard the UARS have extended the range of observations into the lower thermosphere and documented an apparent coupling between the stratospheric QBO and the mesopause SAO [Burrage et al., 1996]. Garcia et al., [1997] have summarized existing observations and discussed the behavior of the mesopause SAO, including its possible link to the stratospheric QBO. Even though the mean structure of the SAO is well documented and there is a growing body of data on its variability, the physical mechanism that gives rise to the oscillation is not completely understood. It is generally agreed that the easterly phase in the stratosphere is due to advection of zonal mean easterly momentum by the meridional circulation, with a contribution from planetary Rossby waves propagating into the tropics from the winter hemisphere. (The fact that both advection and Rossby wave forcing are tied to the seasonal cycle, producing easterly accelerations at the solstices, explains the
semianual periodicity of the oscillation). The stratospheric westerly phase is known to be driven, at least in part, by planetary-scale Kelvin waves [Hirota, 1980; Hitchman and Leovy, 1988]. However, Hitchman and Leovy has also shown that large-scale Kelvin waves alone cannot account for the observed westerly acceleration (they estimate their contribution to be between 30 and 70%). This conclusion is consistent with the results of General Circulation Models (GCMs) [e.g., Sassi et al., 1993], which do not produce a completely realistic stratospheric SAO, although the computed planetary-scale Kelvin wave amplitudes are comparable or even larger than those observed.

Less is known about the mechanism of the mesospheric SAO, except that both the westerly and easterly phases appear to be wave driven since mean advection at mesopause altitudes cannot account for the observed easterly accelerations. A number of hypotheses have been put forward to explain the mesospheric oscillation; these have generally relied on momentum transport by a spectrum of waves propagating from the lower atmosphere and filtered by the wind system of the stratospheric oscillation. That is, zonal mean westerlies in the stratosphere suppress waves that transport westerly momentum but allow propagation of those that transport easterly momentum, and vice-versa. The result is a mesospheric oscillation out of phase with the stratospheric one, as observed. Although this aspect of the mechanism is almost certainly correct, the nature of the waves involved continues to be the subject of speculation. In Dunkerton's [1982] pioneering study of the mesopause SAO it was assumed that momentum was transported by planetary-scale Kelvin waves and small-scale gravity waves (zonal wavelength < 1000 km). A similar scheme was used in a model of the zonally averaged circulation of middle atmosphere by Garcia et al., [1997], who dispensed with the planetary-scale Kelvin waves. Mengel et al., [1995] have used a parameterization of small-scale gravity waves to model both the QBO and SAO. Although these mechanistic models have succeeded in producing more or less realistic oscillations in both the stratosphere and mesosphere, the mesospheric semi annual oscillation has never been simulated with a GCM [e.g., Sassi et al., 1993; Hamilton et al., 1995]. The state of theoretical understanding of the SAO can thus be described as mixed. There is scant evidence about the nature of the waves that drive the SAO.
1.7.3 Gravity Wave Parameterization

Internal gravity waves can be considered as short-time and short-scale perturbations in natural media. For this reason, they are filtered out from weather forecast models, and their effect upon the larger scales must be modeled through parameterizations. The need for a suitable modeling is assessed by the global role of internal gravity waves in transporting momentum from the lower to the upper atmosphere [e.g., Holton et al., 1995]. The horizontal scales of internal gravity waves are too small to permit explicit representation in global-scale numerical models. To include the effect of these waves sub-grid scale parameterizations are required. A number of parameterizations have been used to successfully model the role of gravity waves in driving the mean flow in the MLT region [Holton, 1982; Garcia and Solomon, 1985; Akmaev, 1994].

Gardner [1996] provides a good review of the theoretical basis behind many of these parameterizations. Several parameterizations were considered for incorporation into the Global Scale Wave Model (GSWM) [Hagan et al., 1995; Hagan and Forbes, 2003]. These include the linear saturation approach of Lindzen [1981], the atmospheric transmission concepts of Matsuno [1982], the Doppler spread basis of Hines [1997a, b] and the observation based technique of Fritts and Lu [1993b]. In a seminal work, McLandress [1998] discussed the importance of gravity waves on the general circulation of the middle atmosphere and also described different ways of parameterizing their effects on the global atmospheric models. These parameterizations fall into two general categories, those that employ a discrete spectrum of gravity waves and those that use a continuous spectrum. Each parameterization requires some form of the wave spectra as input: phase speeds, initial magnitudes, vertical and horizontal wavelengths. Thus studies on characteristics of gravity waves such as vertical wavelengths, phase speeds, etc., over various seasons could be helpful in better parameterization of these waves in the GCMs.
1.8 Dynamical Coupling in the Equatorial Middle Atmosphere Through Gravity Waves: A Prelude

Studies on the dynamical coupling in the middle atmosphere has got momentum with the classic work of Hines [1960] who attributed the irregularities and irregular motions in the upper atmosphere to a single physical mechanism, namely, internal atmospheric gravity waves. He concluded that these waves are mainly generated in the lower atmosphere and propagate upwards, thereby carrying the signatures of the background wind on the way of their propagation. Thus, Hines’ work set the stage for the studies on the dynamical coupling through gravity waves in the middle and upper atmosphere, which are otherwise thought to be of only meteorologically important. Since then considerable amount of work has been done to study the coupling processes in the Earth’s atmosphere. Studies on the coupling of low, middle and upper atmosphere through gravity waves are extensively reviewed by so many authors [Sato, 1996; Hamilton, 1999; Hocking, 1996]. In order to have a complete knowledge of the coupling processes, information on the gravity wave source characteristics, their propagation as well as interaction with the background wind and other low frequency waves is needed.

Though the understanding of gravity waves in the middle atmosphere has improved a lot during last 3-4 decades, interest in this field gets increased still now and their effects get better and better understood. In a recent seminal work, Fritts and Alexander [2003] reviewed gravity wave dynamics and their effects in the middle atmosphere. The authors focused on various theoretical and observational studies on gravity waves conducted till then and also discussed the unresolved issues in which further studies are required. However, recent ground based and space based observational studies as well as numerical and theoretical studies have contributed significantly towards the gravity wave source characteristics, spectra, fluxes, energy transfer, wave-mean flow interaction, wave-wave interaction and their implications for atmospheric dynamical structures etc., [Wilson et al., 1991 a, b; Beatty et al., 1992; Tsuda et al., 2000; Kawatani et al., 2004]. Theoretical and observational studies using radiosonde and rocket flights investigated the seasonal and latitudinal variations of
gravity wave activity in detail [Eckerman et al., 1995; Alexander and Vincent, 2000; Vincent and Alexander, 2000].

All the above-mentioned studies emphasized one important aspect of middle atmosphere coupling, which is the interaction between the gravity waves and the mean flow. During such interaction, transfer of momentum takes place from gravity waves to the mean flow. Hence studies of gravity wave momentum fluxes as a function of height are becoming very important, since they impact the mean wind in a dramatic manner. Thus, it is an interesting and important fact that internal gravity waves originating in the troposphere have such significant and wide-ranging influences on the dynamics of the middle atmosphere. It demonstrates the strong connections between the troposphere, a source region, the stratosphere and lower mesosphere, whose refractive properties control wave transport, and the upper atmosphere, where the wave effects can be the dominant physical mechanism at work. It is also interesting that different models of gravity wave physics predict similar effects in the upper atmosphere with regard to the mean flow, but vary in their predictions on interactions with other dynamical features such as tides and planetary waves. Increasing our understanding of the spectrum of gravity waves, including magnitudes, phase speeds and wavelengths will increase our capability to understand and describe their effects on the atmosphere as a whole. A better understanding of the short-term variability of both gravity wave sources and the background atmosphere through which they propagate will also facilitate a better description of their influence on the dynamics of the middle and upper atmosphere.

Realizing the importance of gravity waves in atmospheric coupling, a series of experiments were conducted at the present observational site, Gadanki, using radar and lidar observations [Dutta et al., 1999; Dhaka et al., 2002; Kumar, 2006; Sivakumar et al., 2003, 2006]. Using lidar observations, detailed studies were carried out on middle atmospheric thermal structure and gravity wave characteristics such as their growth, saturation and associated potential energy over Gadanki [Sivakumar et al., 2003, 2006]. However, none of these studies aimed to divulge the gravity wave-mean flow interactions and to quantify the role of gravity waves in driving the well-known
middle atmospheric oscillations such as QBO and SAO. In this regard, a comprehensive program known as Middle Atmospheric Dynamics Program (MIDAS) was carried out to study the various aspects of low latitude middle atmospheric dynamics using various ground based experimental facilities. A brief description of MIDAS is given in the following section.

1.8.1 Middle Atmospheric Dynamics Program (MIDAS): Coupling Studies in the Indian Perspective

During Indian Middle Atmosphere Programme (IMAP), several new results on the equatorial middle atmosphere have been reported based on a series of rocket campaigns [e.g., Sasi and Sen Gupta, 1986; Chakravarthy et al., 1992; Sasi, 1994a; Mohankumar, 1994]. During the past two decades, five equatorial wave campaigns were conducted, involving several Indian stations, to study the characteristics of equatorial waves and gravity waves and also to estimate the momentum fluxes associated with these waves [Sasi et al., 2005]. Later, two unique campaigns of coordinated measurements have been conducted during 1999 and 2000 for better understanding of equatorial waves with emphasis on estimation of their momentum fluxes. The first campaign involved 45 days of simultaneous measurements of co-located Rayleigh lidar and Indian MST radar [Krishna Murthy et al., 2000] and the second one employing lidar, MST radar and rockets [Krishna Murthy et al., 2002]. The experience and expertise gained from these campaigns set tone for another five year long national program known as ‘ISRO’s Middle Atmosphere Dynamics Program’ “MIDAS (2002-2007)” which is proposed and executed by Space Physics Laboratory (SPL) of Vikram Sarabhai Space Centre (VSSC), Trivandrum. This program started on 21 November 2002, to commemorate the first rocket (Nicke-Apache) launch from Thumba Equatorial Rocket Launching Station (TERLS) at VSSC. The MIDAS program mainly consists of three themes to address the following scientific objectives:

**Theme-I**

To obtain gravity wave climatology in the middle atmosphere over low-latitude Indian region and study the impact of gravity waves in the evolution of easterly
and westerly phases of stratopause and mesopause zonal wind Semi-Annual Oscillations (SAO).

**Theme-II**

To study the phase difference between the high-latitude 'stratospheric warming' and low-latitude 'stratospheric cooling' and evolution of middle atmospheric circulation changes during stratwarm events.

**Theme-III**

To study the mutual coupling between the middle atmospheric circulation changes and 'monsoon' phenomenon.

Under Theme-I of this program, regular co-ordinated measurements of winds and temperatures were made fortnightly using in-situ measurements like High Altitude Balloon (HAB) flights and Rohini Sounding Rocket - RH 200 flights, ground based measurements using Partial Reflection Radar (PRR) and SKiYMET Meteor Wind radar from TERLS, Trivandrum (8.5°N, 77°E), PRR measurements from Equatorial Geophysical Research Laboratory (EGRL), Tirunelveli (8.7°N, 77.8°E) and MST radar and lidar measurements at National MST Radar Facility (NMRF)/National Atmospheric Research Laboratory (NARL), Gadanki(13.5°N, 79.2°E). Table 1.3 shows the details of experiments conducted regularly under this program.

<table>
<thead>
<tr>
<th>Instrument</th>
<th>RH-200 Rockets (TERLS)</th>
<th>High-altitude Balloons (TERLS)</th>
<th>Lidar (Gadanki)</th>
<th>PR Radar (TERLS and Tirunelveli)</th>
<th>MST Radar (Gadanki)</th>
<th>SKiYMET Meteor Radar (TERLS)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Measured Parameters</td>
<td>Wind</td>
<td>Wind</td>
<td>Temperature</td>
<td>Wind</td>
<td>Wind</td>
<td>Wind</td>
</tr>
<tr>
<td>Height (km)</td>
<td>20-65</td>
<td>0-35</td>
<td>30-70</td>
<td>65-95</td>
<td>3-20 70-90</td>
<td>80-110</td>
</tr>
<tr>
<td>Time (hours) IST</td>
<td>1930</td>
<td>1500</td>
<td>1900-0100</td>
<td>0900-2000 0000-2400</td>
<td>1200-1400</td>
<td>00-24</td>
</tr>
<tr>
<td>Frequency of Measurement</td>
<td>Fortnightly</td>
<td>Fortnightly</td>
<td>Fortnightly</td>
<td>Fortnightly</td>
<td>Fortnightly</td>
<td>Daily</td>
</tr>
</tbody>
</table>

**Table 1.3 Experiments under MIDAS program**

Objectives under Theme (II) and (III) were realized through 15 day long campaign based observations, using all the above experiments shown in Table 1.3, in
addition to lidar observations from Mount Abu (24.5°N, 72.7°E) and HAB flights from five additional Meteorological stations over Indian sub continent. Two such campaigns were conducted during July-August 2005 (monsoon campaign) and January – February 2006 (stratwarm campaign).

1.9 Scope of the Present Thesis

The central objective of the present thesis work is to study the dynamical coupling in the equatorial middle and upper atmosphere through gravity waves using ground based and in-situ observations viz., radars, lidars and rocketsondes. As discussed earlier, through exchange of energy and momentum, vertically propagating gravity waves effectively couple the equatorial middle and upper atmosphere in profound ways. The characteristics of the gravity waves in the middle atmospheric region and their sources are less explored over low latitude. In this regard, the present thesis aims to explore the role of gravity waves in coupling the low latitude middle atmosphere with following specific objectives,

(1) To delineate the seasonal variation in the gravity wave activity viz., potential energy and momentum fluxes in the stratospheric region and to find out the relation between the wave activity and their lower atmospheric sources namely the convection and wind shears.

(2) To quantify the role of gravity waves in driving easterly and westerly phases of the stratospheric QBO

(3) To study the characteristics of SSAO and to quantify the role of gravity waves in driving its easterly and westerly phases

(4) To estimate the gravity wave momentum fluxes and their seasonal variation in the mesospheric and lower thermospheric region, hence to quantify the role of gravity waves in forcing the easterly and westerly phases of MSAO.

Chapter 2 describes various instruments used for the present study, which includes Rayleigh Lidar, Meteor wind radar, MST radar, rocketsondes and high altitude balloons.
The system details along with signal processing and derivation of atmospheric field variables like wind and temperature are also described.

**Chapter 3** elaborates the characteristic features of the prominent periods of gravity waves and seasonal variation in the wave activity in the middle atmospheric region using temperature data from Rayleigh lidar observations over Gadanki (13.5°N, 79.2°E). The variations in wave activity are compared with variations in the strength of sources viz., convection and wind shears, in the lower atmospheric region. Thus the present study not only brought out the seasonal variations of gravity wave activity over Gadanki but also related these variations with their source strength, which is a first step towards the parameterization of these waves.

**Chapter 4** discusses gravity wave forcing towards the generation of easterly and westerly phases of Quasi-Biennial Oscillation (QBO). This chapter provides a comprehensive review of the origin, dynamics, structure as well as various forcing mechanisms involved in the easterly and westerly phases of QBO. The seasonal variation of the gravity wave momentum fluxes is divulged for the first time over this tropical latitude, Gadanki. The monthly mean zonal winds from NCEP/NCAR reanalysis data over the observational site are used to study the QBO characteristics in the lower stratospheric region and hence the mean flow acceleration. Forcing of gravity wave towards the generation of four cycles of QBO is estimated and compared with observed mean flow acceleration. This comparison could give a better insight into the role of gravity waves in driving the easterly and westerly phases of QBO.

**Chapter 5** elucidates the role of gravity waves in the generation of easterly and westerly phases of Stratospheric Semi Annual Oscillation (SSAO). The characteristics as well as the forcing mechanisms of SSAO are explained in detail. The simultaneous wind measurements using rocketsondes over Trivandrum (8.5°N, 77°E) under MIDAS program are also extensively used to study the background wind and the mean flow acceleration in the 30-60 km altitude region. The divergence of gravity wave momentum fluxes is estimated during three different cycles of SSAO and compared with observed mean flow acceleration. The present study could be useful in quantifying the role of shorter period gravity waves in the evolution of SSAO.
Chapter 6 portrays the measurement of gravity wave momentum fluxes in the Mesospheric Lower Thermospheric (MLT) region over a low latitude site Trivandrum, employing a novel technique using meteor wind radar observations. Using three years of round the clock wind measurements, the seasonal variation in the gravity wave momentum fluxes and the characteristics of the Mesospheric Semi Annual Oscillation (MSAO) are examined. From the divergence of gravity wave momentum fluxes, the mean flow acceleration is estimated and compared with observed mean flow acceleration computed using the monthly mean zonal winds in the 82-94 km height region. Forcing of gravity waves towards the evolution of westerly and easterly phase of six MSAO cycles are estimated, which is first of its kind over this tropical latitude, Trivandrum.

Chapter 7 presents the summary and conclusions of the major findings from this research work and also includes the future scope of the study.

Thus the present thesis work quantifies the role of gravity waves in the generation of equatorial oscillations thus corroborating the results from the modeling studies and providing a stepping stone towards an improved parameterization of gravity waves in general circulation models.