2.0 Introduction

The positioning accuracy achieved with GPS in geodesy, Navigation, etc., is being continuously improved during the last two decades. Historically, the main sources of errors which are progressively being reduced are the orbital error, the antenna phase center modeling, signal multipath, scattering by the environment near by the receiver and the clock error, based on recent technological advances. These advances have limited applications or usage in mitigating the error due to the atmospheric refraction, which can be resolved only through accurate modeling of atmospheric refraction effect. Owing to the large variations of atmospheric properties in space and time, its modeling is rather complicated.

In most of the ranging applications it is assumed that the EM wave travels along straight line with speed equal to that in vacuum. But since the earth's atmosphere is a non-homogeneous medium with a significant variation in its physical properties such as pressure, temperature, humidity etc. with altitude, it introduces a corresponding variation in the index of refraction. An EM wave propagating through the atmosphere undergoes continuous refraction leading to bending of the ray path (from its direct line-of-sight) as well as group retardation because of the decrease in the group velocity (refractive index of the medium being >1). Thus, a refracting medium makes the propagating wave to travel a longer distance taking a longer time to reach the destination than that would have been in the absence of the atmosphere. As the range is inferred from the time-of-flight, this delay manifests as an error in estimated range (a pseudo range).

In general, the region of the earth's atmosphere which affects the propagation of microwaves in the GPS frequencies can be broadly classified into two; (i) the ionosphere and (ii) the neutral atmosphere. The ionosphere is a dispersive medium (i.e. the refractive index of the medium depends on the frequency of the EM wave traveling through it). As its index of refraction varies inversely with square of the frequency, by using higher frequencies (in GHz) and properly modeling the ionosphere or by making measurements in
dual (or multi) frequencies, the range error due to ionosphere can be reduced considerably. At present, in the state-of-art of GPS systems, the ionospheric contribution (in the zenith direction) is reduced to $< 1 - 5$ m.

As the neutral atmosphere is non-dispersive at the GPS frequencies, the delay due to neutral atmospheric refraction cannot be accounted through dual (or multi) frequency measurements. Estimation of this delay needs a strategy entirely different from that used for the ionized region. Since the bulk of the neutral atmosphere resides in the lower region (or Troposphere) and most of the delay caused by this comes from these altitudes, this is usually referred to solely as the “Tropospheric delay”. The geometry of radio refraction in the neutral atmosphere is shown in Figure 2.1. The points $A$ and $B$ in the figure denote the locations of the receiving station and radio beacon source, respectively. The $D$ and $\alpha$ denote the true range and elevation angle of the beacon. If there were no atmosphere a radio signal starting from $B$ would travel along the straight-line path $D$ and reach the point $A$. Due to the presence of the atmosphere, the ray would be refracted and the path is curved along $S$. In this case, the angle of arrival of the radio signal at $A$ is $\alpha'$ which is the apparent elevation angle. The apparent position of the beacon is then at $P$. Here $AP$ is the apparent range that is greater than the true range ($AB$), since ray takes longer time to travel in the refracting medium than that in vacuum. If $n$ is the refractive index, $D$ is the direct path length and $S$ is the curved path length, then range error is given by eq. (1.3).

![Figure 2.1: Schematic diagram showing the range error and angular deviation associated with radio ranging](image-url)
2.1 Effect of Neutral Atmosphere in the Propagation of Microwave

The refraction of radio waves as they pass through the neutral atmosphere is generally referred to as tropospheric delay [Misra and Enge, 2001; Barry and Chorley, 1998]. Debye [1957] has considered the effect of an imposed electric field on the dielectric properties of both non-polar and polar molecules in a liquid. In the presence of an external electric field these molecules develop an induced dipole moment. In addition to that, the polar molecules develop an orientational polarization due to the alignment of the axes of permanent dipoles along the field direction. This adds up to the total induced dipole moment. The expression for the polarization (dipole moment per unit volume) of a polar liquid under the influence of a high frequency radio field as obtained by Debye [1957] can be applied to the case of gases with proper approximation. This polarization \( P(\omega) \) is given as

\[
P(\omega) = \frac{\varepsilon - 1}{\varepsilon + 2} \frac{M}{\rho} = \frac{4\pi A_N}{3} \left[ \gamma_i + \frac{\gamma_p^2}{3KT} + \frac{1}{1 + i\omega \tau} \right]
\]

(2.1)

where \( P \) = Electric Polarization

\[ \frac{\varepsilon - 1}{\varepsilon + 2} \frac{M}{\rho} = \frac{4\pi A_N}{3} \left[ \gamma_i + \frac{\gamma_p^2}{3KT} + \frac{1}{1 + i\omega \tau} \right] \]

\[ (2.1) \]

Debye’s analysis shows that for external fields with frequencies less than 100 GHz, \( \omega \tau \ll 1 \). In this case the above equation can be rewritten as

\[
\frac{\varepsilon - 1}{\varepsilon + 2} \frac{M}{\rho} = \frac{4\pi A_N}{3} \left[ \gamma_i + \frac{\gamma_p^2}{3KT} \right]
\]

(2.2)
For gases, as $\varepsilon$ is close to unity eq. (2.2) reduces to

$$\varepsilon - 1 = \frac{p}{M} 4\pi A_N \left[ \gamma_r + \frac{\gamma_p^2}{3KT} \right] \quad (2.3)$$

For non-polar gases $\gamma_p = 0$; In this case

$$\varepsilon - 1 = \frac{p}{M} 4\pi A_N \gamma_i \quad (2.4)$$

Using the perfect gas law $P = \rho RT$, the above equation can be rewritten as

$$\varepsilon - 1 = k_1 \frac{P}{T} \quad (2.5)$$

where $k_1$ is constant.

For polar gases the eq. (2.3) reduces to

$$\varepsilon - 1 = k_2 \frac{P}{T} \left[ A_0 + B_0/T \right] \quad (2.6)$$

where $k_2$, $A_0$ and $B_0$ are constants. $P$ and $T$ are the pressure and temperature of the polar gas respectively. For a mixture of gases, Dalton's law of partial pressure is assumed to hold with the result that we can sum the effects of polar and non-polar gases. In troposphere, mainly the effects of dry air (non-polar gases) and water vapor (a polar gas) only ($P = P_d + e$) is to be considered in which case eq. (2.6) reduces to

$$\varepsilon - 1 = k_1' \frac{P}{T} + k_2' \frac{e}{T} \left[ A_0 + B_0/T \right] \quad (2.7)$$

where $P_d$ is the pressure of dry air and $e$ the partial pressure of water vapor. In case of air, the dielectric constant ($\varepsilon$) is related to the refractive index ($n$) through $n - 1 = (\mu\varepsilon - 1)/2$; where $\mu$ is the permeability which is assumed to be approximately unity for air. In this case, the above equation can be written in terms of the refractivity ($N$) of the air as [Smith and Weintraub, 1953; Thayer, 1974]

$$N = (n - 1) \times 10^6 = k_1 \frac{P}{T} + k_2 \frac{e}{T} + k_3 \frac{e}{T^2} \quad (2.8)$$

Note here that $P$ is the total atmospheric pressure. In eq. (2.8) $P$ and $e$ are expressed in hPa and $T$ in K. The values of the constants $k_1$, $k_2$ and $k_3$ as reported by Bevis et al. [1992] are, respectively, $77.604 \pm 0.014$ K hPa$^{-1}$, $64.79 \pm 0.08$ K hPa$^{-1}$ and $(3.776 \pm 0.004) \times 10^5$ K$^2$ hPa$^{-1}$. These constants in eq. (2.8) are considered to be good up to 0.5 percent of $N$ in
frequencies up to 30 GHz for normally encountered ranges of pressure, temperature and humidity.

Radio refractivity in troposphere has two principal components, a hydrostatic (Dry) component \( N_D \) due to non-polar molecules (dry air) and a non-hydrostatic (Wet) component \( N_W \) due to polar component of water vapor. The nomenclatures ‘dry’ and ‘wet’ are to some extent misnomers as the dry component also has the contribution from non-polar component of water vapor. In spite of that these terms are very widely used among the GPS community. Thus, from eq. (2.8) the \( N_D \) and \( N_W \) can be written as

\[
N_D = k_1 \frac{P}{T}
\]

\[
N_W = k_2 \frac{e}{T} + k_3 \frac{e}{T^2}
\]

The constant \( k_3 \) of the wet term is much larger than that of the dry term. So, even a small amount of water vapor in dry air makes a large difference in the total refractivity. Using the altitude profiles of \( P, T \) and \( e \) the altitude profiles of \( N \) can be obtained. The range error \( \Delta R \) then is estimated by tracing the ray-path considering the altitude variation of \( N \) given by this profile.

2.2 Ray Tracing

In order to examine the refraction effect of the microwaves in the atmosphere the ray-path is to be traced from the observer at the surface of the earth up to the satellite, taking into account the altitude variation of refractive index and the earth’s curvature. In general, earth’s curvature has a greater influence on path length due to atmospheric refraction for very low elevation angles (grazing incidence), as the wave spends more time in the lower layers where the refractive effects are significant.

2.2.1 Geometry of Ray Propagation

The earth’s atmosphere is assumed to be divided into a series of concentric homogeneous spherical layers, each characterized by a mean refractive index. Applying the Snell’s law for spherically stratified atmospheric layers allows the numerical integration of refractive effect. Consider the trajectory of a ray passing from ‘A’, the point on the surface of the earth (Figure 2.2), to the point ‘B’ at a height \( h \) from the surface of the earth with an initial zenith angle \( \chi_0 \) to the local vertical. On reaching \( B \) the zenith angle becomes \( \chi_m \).
because of the refraction. The total angular deviation is denoted as $\Delta \psi$.

If $Z_i$ and $Z_{i+1}$ are the lower and upper boundaries of a given layer, $\chi_i$ and $\chi_{i+1}$ are local zenith angles at these boundaries. At the height of $Z_{i+1}$ the angle of refraction is $\chi_{i+1}$ and the angle of incidence is $\alpha_i$. Considering a triangular element OPQ

$$\frac{r_e + Z_i}{\sin \alpha_i} = \frac{r_e + Z_{i+1}}{\sin(180 - \chi_i)}$$

or

$$\sin \alpha_i = \frac{(r_e + Z_i) \sin \chi_i}{(r_e + Z_{i+1})}$$

(2.10)

where $r_e$ is the Earth's radius.

Applying Snell's law at boundary $Z_{i+1}$, we have

$$n_i \sin \alpha_i = n_{i+1} \sin \chi_{i+1}$$

(2.11)

where $n_i$ and $n_{i+1}$ are the mean refractive indices of layers above $Z_i$ and $Z_{i+1}$, respectively.

Substituting for $\sin \alpha_i$ in eq. (2.11)

$$n_i (r_e + Z_i) \sin \chi_i = n_{i+1} (r_e + Z_{i+1}) \sin \chi_{i+1}$$

or

$$\sin \chi_{i+1} = \frac{n_i (r_e + Z_i) \sin \chi_i}{n_{i+1} (r_e + Z_{i+1})}$$

(2.12)

Knowing the height profile of $n_i$, the angle of incidence at the upper layer $\chi_{i+1}$ can be estimated from the angle of incidence $\chi_i$ at the lower layer. This procedure can be repeated from the lower boundary at A (where $\chi_i = \chi_0$) to the upper boundary at B (where $\chi_{i+1} = \chi_m$) by dividing the in between region into small layers (like $Z_i, Z_{i+1}$) characterized with its own refractive index and thickness.

The angle $\chi_i$, subtended at the center of the earth by the ray path PQ within the layer between $Z_i$ to $Z_{i+1}$ is given by

$$\chi_i = \chi_i - \alpha_i$$

(2.13)

where $\alpha_i$ is defined by eq. (2.11). The total angles subtended by the ray path at the center of the earth when traversing the atmosphere from A to B is

$$\chi = \sum_{i=1}^{i+1} (\chi_i - \alpha_i)$$

(2.14)
where \( l \) is the number of segments between \( A \) and \( B \). Now the total angular deviation of the trajectory can be estimated knowing the values of \( \chi_m, \chi_0 \) and \( \kappa \). Considering the quadrilateral \( ACBO \), the interior angles

\[
\begin{align*}
\angle OAC &= 180 - \chi_0 \\
\angle ACB &= 180 - \Delta \Psi \\
\angle CBO &= \chi_m \\
\angle OAC + \angle ACB + \angle CBO + \angle AOB &= 360^\circ
\end{align*}
\]

\( \Delta \Psi = \kappa + (\chi_m - \chi_0) \) (2.15)

where, \( \Delta \Psi \) is the total angular deviation of the ray from the straight line path.

![Schematic diagram of refracted ray path used for ray-tracing](image)

Figure 2.2: Schematic diagram of refracted ray path used for ray-tracing

The effective path length, \( D_{si} \), between \( Z_i \) and \( Z_{i+1} \) is given by

\[
\frac{D_{si}}{\sin \kappa_i} = \frac{(r_e + Z_{i+1})}{\sin \chi_i}
\]

(2.16)
or

\[ D_s = (r_x + Z_{i+1}) (\sin \kappa_i / \sin \chi_i) \quad (2.17) \]

for \( 0^\circ < \chi < 180^\circ \) and \( D_s = Z_{i+1} - Z_i \) for \( \chi = 0^\circ \) or \( 180^\circ \). Now, the actual path length traversed by the wave from A to B is

\[ S = \sum_{i} D_s \quad (2.18) \]

The direct path \( AB \) can be estimated by solving the triangle \( ABO \). The geometrical distance \( AB \), or \( D \), is given by

\[ D = \sqrt{r_x^2 + (r_x + h)^2 - 2r_x(r_x + h) \cos \kappa} \quad (2.19) \]

Equations (2.18) and (2.19) are used to estimate \( S \) and \( D \) later in eq. (2.24). As discussed in Section 2.1 (eq. 2.8) the refractive index of the neutral atmosphere is a function of \( P \), \( T \) and \( e \), to estimate a refractive index profile and further \( \Delta R \) the vertical profiles of these parameters are to be known precisely. An easily available source for these data from a fixed location is the regular balloon sounding carried out by the India Meteorological Department (IMD) in accordance with WMO specifications.

### 2.2.2 Estimation of Tropospheric Range Error (\( \Delta R \))

As the refractive index of air is greater than unity, the wave propagating through this medium undergoes a group-retardation. In ranging applications the propagation time is transferred into range by using the velocity of electromagnetic wave in vacuum \((c)\). Thus, delay in propagation introduces an error in the ranging. The other source of error is the refractive bending of ray path due to the spatial variability of the refractive index. Thus, wave takes a curved path through the atmosphere, which is always longer than the direct line-of-sight distance. Considering a wave propagating through a medium with refractive index of \( n \), the time taken by the wave to travel an infinitesimal path distance, \( ds \), is given as

\[ d\tau = \frac{nds}{c} \quad (2.20) \]

where \( c \) is the velocity of light in free space.

The corresponding distance traveled in free space is then

\[ cd\tau = nds \quad (2.21) \]

The total time taken by the wave to cover the path between the source and the destination (e.g., satellite to GPS receiver on earth surface) is
\[ \tau = \int d\tau = \frac{1}{c} \int n ds \]  
(2.22)

The path length traveled by the wave is
\[ G = \int n ds \]  
(2.23)

The wave path length \( G \) is larger than the true geometric distance \( D \), between source and receiver.

The total range error can then be written as
\[ \Delta R = \int n ds - D \]

or
\[ \Delta R = \int (n-1) ds + \int ds - D \]

i.e.,
\[ \Delta R = \int (n-1) ds + [S - D] \]

or
\[ \Delta R = 10^6 \times \int N ds + [S - D] \]  
(2.24)

The first term in the right side of eq. (2.24) represents the error due to retardation of the wave as it traverses through the medium. The second term represents the extra distance traversed by the wave due to the bending of ray path. From eq. (2.9) it is clear that the tropospheric delay (\( \Delta R \)) consists of two components the hydrostatic (or "dry") component, which accounts for approximately 90% of the delay and the "wet" component, which accounts for the remaining 10% of the delay. At nadir (when zenith angle is zero) the second term in eq. (2.24) tends to zero as there will not be any bending for normal incidence. The range error in this case is usually referred to as the Zenith Tropospheric Delay (ZTD). Prior knowledge on hydrostatic and non-hydrostatic refractivity profiles is necessary for tracing the ray path to estimate tropospheric range error, which requires atmospheric parameter profiles obtained from upper atmospheric soundings.

2.3 Mean Neutral Atmospheric Models over the Indian Region

The balloon soundings provide altitude profiles of atmospheric pressure (in hPa), air temperature (in °C) up to ~ 26 km, and dew point temperature (\( T_D \) in °C) up to ~ 12 km. However, depending on the atmospheric conditions this upper limit will vary slightly from
day-to-day. A major problem encountered in the measurement of $T_D$ is the saturation effect of humidity sensor as well as the limitation in the sensitivity at higher altitudes where water vapor is very small. The $T_D$ values are used to derive the water vapor partial pressure ($e$ in hPa) using the relation

$$e(T) = 6.105 \exp \left( 25.22 \left( \frac{T_D - 273.0}{T_D} \right) \right) - 5.311 \ln \left( \frac{T_D}{273.0} \right)$$

(2.25)

The radiosonde raw data is formatted in terms of pressure levels. The temperature and dew point temperature at these levels along with the corresponding altitude are tabulated along with the values of pressure, temperature and dew point temperature at the surface just before the balloon release. This data is reformatted in terms of altitude, to generate the altitude profiles of $P$, $T$, $T_D$ and $e$. These altitude profiles, which are unequally spaced in altitude (as they are reformatted from the data at standard pressure level), are interpolated appropriately to generate suitable spaced profiles at 10 m interval from the surface to 5 km, at 200 m interval up to 20 km and 500 m beyond 20 km. A variable altitude resolution (close resolution at lower altitude where the values of these parameters and their variability is large) for the data is selected mainly to accomplish a fairly uniform accuracy for the derived parameters along the entire altitude region (up to 26 km for $P$ and $T$ and up to 12 km for $e$). Note that typical accuracies achievable in radiosonde measurements as quoted by Bisagni [1989] are, respectively, ~2.0 hPa for the barometric pressure, about +0.4 K for temperature and 4% for relative humidity.

For the present analysis, the radiosonde data from 18 IMD locations, fairly well distributed over Indian subcontinent representing all major climatic and geographical conditions prevailing over the subcontinent, are selected. A list of these stations along with geographical locations is presented in Table 2.1. As mentioned earlier, for all these stations, the altitude profiles of $P$, $T$, $T_D$ and $e$ will not be available up to the same level on all the days. However, the daily profiles of $P$, $T$, $T_D$ and $e$ in each month for five consecutive years from 1995 to 1999 are averaged to generate mean profiles for respective months for every station. The extent of day-to-day variability of these profiles is quantified in terms of the standard deviations at each altitude for different months. Meteorological data from the regular radiosonde observations made by the IMD at these locations (Table 2.1) for a period
Table 2.1: A List of IMD stations selected for the present study.

<table>
<thead>
<tr>
<th>Station</th>
<th>Latitude $^\circ$(N)</th>
<th>Longitude $^\circ$(E)</th>
<th>Height from MSL (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Trivandrum</td>
<td>8.48</td>
<td>76.95</td>
<td>64.0</td>
</tr>
<tr>
<td>Port Blair</td>
<td>11.66</td>
<td>92.71</td>
<td>79.0</td>
</tr>
<tr>
<td>Bangalore</td>
<td>12.96</td>
<td>77.58</td>
<td>921.0</td>
</tr>
<tr>
<td>Mangalore</td>
<td>12.95</td>
<td>74.83</td>
<td>31.0</td>
</tr>
<tr>
<td>Chennai</td>
<td>13.00</td>
<td>80.18</td>
<td>16.0</td>
</tr>
<tr>
<td>Goa</td>
<td>15.48</td>
<td>73.81</td>
<td>60.0</td>
</tr>
<tr>
<td>Hyderabad</td>
<td>17.45</td>
<td>78.46</td>
<td>545.0</td>
</tr>
<tr>
<td>Vishakhapatnam</td>
<td>17.70</td>
<td>83.30</td>
<td>66.0</td>
</tr>
<tr>
<td>Mumbai</td>
<td>19.11</td>
<td>72.85</td>
<td>14.0</td>
</tr>
<tr>
<td>Kolkata</td>
<td>22.65</td>
<td>88.45</td>
<td>6.0</td>
</tr>
<tr>
<td>Ahmedabad</td>
<td>23.06</td>
<td>72.63</td>
<td>55.0</td>
</tr>
<tr>
<td>Bhopal</td>
<td>23.28</td>
<td>77.35</td>
<td>523.0</td>
</tr>
<tr>
<td>Guwahati</td>
<td>26.10</td>
<td>91.58</td>
<td>54.0</td>
</tr>
<tr>
<td>Jodhpur</td>
<td>26.30</td>
<td>73.01</td>
<td>224.0</td>
</tr>
<tr>
<td>Lucknow</td>
<td>26.75</td>
<td>80.88</td>
<td>128.0</td>
</tr>
<tr>
<td>Delhi</td>
<td>28.58</td>
<td>77.20</td>
<td>216.0</td>
</tr>
<tr>
<td>Patiala</td>
<td>30.32</td>
<td>76.45</td>
<td>251.0</td>
</tr>
<tr>
<td>Srinagar</td>
<td>34.08</td>
<td>74.83</td>
<td>1587.0</td>
</tr>
</tbody>
</table>

of five years (1995-1999) are thus used to develop the site-specific monthly mean atmospheric models. As a typical example the monthly mean models for $P$, $T$ and $e$ for the three months of January, May and July (representative of three different seasons) for Trivandrum and Delhi are presented in Figure 2.3. For the near-equatorial coastal station Trivandrum these mean atmospheric parameters show less seasonal variation compared to that at a higher latitude inland station (like Delhi). While the temperature profile remains more-or-less similar throughout the year for Trivandrum it varies significantly for a station like Delhi. In case of water vapor, a clear seasonal variation of $\sim 20$ hPa (at surface) is seen over Delhi while for Trivandrum the profile is almost similar throughout the year. The surface water vapor pressure at Trivandrum is $\sim 30$ hPa which decreases almost exponentially with increase in altitude. Beyond 4 km the amount of water vapor is very low and does not show any pronounced seasonal variation at both these stations. One primary limitation of atmospheric data obtained from radiosonde is that it does not reach beyond 20 hPa or $\sim 26$ km in altitude, even though the prominence of neutral atmosphere extends up to $\sim 100$ km is to be considered for GPS related applications.
Figure 2.3: Altitude profiles of monthly mean $P$, $T$ and $e$ for three seasons at Trivandrum (1) and Delhi (2). The corresponding surface value of these parameters are also marked in each frame.
To achieve this, the upper air data from COSPAR International Reference Atmosphere (CIRA) is incorporated.

2.3.1 The COSPAR International Reference Atmosphere (CIRA) Model

The CIRA provides empirical models of atmospheric temperature and densities as recommended by the Committee on Space Research (COSPAR). Since the early sixties different editions of CIRA have been published and CIRA-86 is the latest revised version of this model [CIRA, 1986]. Above 100 km CIRA-86 compares well with the MSIS model, which is available from NSSDC (MI-91E). In the lower region below 120 km, CIRA-86 consists of tables of the monthly mean values of temperature and zonal wind with almost global coverage (80°N - 80°S). Two sets of files are available, one in pressure coordinates including also the geopotential heights, and the other is in height coordinates generated from several global data compilations including ground-based and satellite (Nimbus 5, 6, 7) measurements [Oort, 1983; Labitzke et al., 1985]. This model reproduces most of the characteristic features of the atmosphere such as the equatorial wind and the general structure of the tropopause, stratopause, and mesopause. Table 2.2 lists all the parameters their vertical coordinates, spatial ranges and resolutions in detail Fleming et al. [1988].

Table 2.2: A List of Different Parameters (vertical coordinates, spatial ranges and resolutions of the zonal mean data) Contained in CIRA Model

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Vertical Coordinate</th>
<th>Vertical Range; Resolution</th>
<th>Latitudinal Range; Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>A Temperature</td>
<td>Altitude</td>
<td>0-120 km; 5 km</td>
<td>80°S-80°N; 10°</td>
</tr>
<tr>
<td>B Pressure</td>
<td>Altitude</td>
<td>20-120 km; 5 km</td>
<td>80°S-80°N; 10°</td>
</tr>
<tr>
<td>C Temperature</td>
<td>Log-pressure</td>
<td>1013-0.000025 hPa; 0.5 sh*</td>
<td>80°S-80°N; 10°</td>
</tr>
<tr>
<td>D Geopotential height</td>
<td>Log-pressure</td>
<td>1013-0.000025 hPa; 0.5 sh</td>
<td>80°S-80°N; 10°</td>
</tr>
<tr>
<td>E Zonal wind</td>
<td>Log-pressure</td>
<td>1013-0.000025 hPa; 0.5 sh</td>
<td>80°S-80°N; 10°</td>
</tr>
</tbody>
</table>

* sh = -ln (p/p₀); where p₀ = 1013 hPa
For the present study $P$ and $T$ profiles from the CIRA model are used from 20 km to 100 km. As radiosonde data is available up to $\sim$ 26 km the CIRA profiles in the overlapping region 20 to 26 km are scaled appropriately to match with the observed radiosonde profiles in this region. Once this is accomplished the modified CIRA profile up to 100 km is combined with the radiosonde data to develop the full profile. The confidence on this profile is further confirmed by comparing with the SS-86 model [Sasi and Sengupta, 1986] developed exclusively for near equatorial stations, viz. Trivandrum.

2.3.2 A Reference Atmosphere for Indian Equatorial Zone from 0-80 km

The SS-86 model is considered as a 'reference atmosphere' valid for Indian Equatorial Zone in the altitude range of 0 (surface) to 80 km and widely used in many atmospheric studies. This model was evolved incorporating radiosonde data from surface to 20 km from four equatorial IMD stations (Chennai, Portblair, Trivandrum, Minicoy) and the M-100 rocket data in the altitude region 20 to 80 km from Trivandrum for a sufficiently long period of 11 years (from 1971 to 1982).

Water vapor in the atmosphere is mostly confined to the troposphere much below the tropopause for which the profiles from radiosonde measurements can be effectively used as no other more reliable measurements are available for this parameter. A comparison of SS-86, CIRA-86 and radiosonde profiles for January at Trivandrum is shown in Figure 2.4(a-c). In Figure 2.4(a) $P$ values from radiosonde are available from surface to 26 km, and for CIRA-86 it is plotted from $\sim$ 20 km to 100 km. The SS-86 model values for $P$ are also plotted in the same figure from 0 – 80 km. The mean absolute difference between the SS-86 and CIRA-86 $P$ values is $\sim$ 0.5 ± 0.4 hPa and the maximum absolute deviation is $\sim$ 1.3 hPa. Figure 2.4 (b) shows a comparison of the $T$ profile obtained from SS-86, radiosonde and CIRA-86 model. The CIRA-86 profile matches reasonably well with the radiosonde (up to 26 km) profile. The mean absolute difference between radiosonde and CIRA-86 is $\sim$ 4.5 ± 3.5 K, where as that between SS-86 and CIRA-86 is $\sim$ 1.95 ± 1.9 K. The maximum deviations are of the order of 13.34 K and 11.32 K, respectively. Figure 2.4 (c) shows a comparison of dry refractivity ($N_0$) from all the three sources. The mean absolute difference in $N_0$ between SS-86 and CIRA-86 model is $\sim$ 1.24 ± 0.8 while the maximum deviation is $\sim$ 3. The Zenith range error values calculated from the above models are, 2.315 m for SASI-86 where as it is 2.294 m for CIRA-86, with a deviation of $\sim$ 2 cm corresponding to an
Figure 2.4: A Comparison of SS-86, Radiosonde and CIRA-86 monthly mean profiles of Pressure (a), Temperature (b) and Dry component of Refractivity (c) at Trivandrum; along with a plot showing the differences between CIRRA-86 and SS-86 model for all the three parameters (d).

uncertainty of ~ 0.01%. Figure 2.4(d) shows the difference between the SS-86 and CIRA-86 model values of $P$, $T$ and $N_D$. The CIRA-86 data used for this analysis are downloaded from http://badc.nerc.ac.uk/data/cira/acquiring.html, the British Atmospheric Data Centre (BADC) website. To estimate the $P$ and $T$ values at the particular stations latitudes ($\phi$) from the zonal mean values for every 10 degree latitude available from the website the following formula are used
\[ P(\varphi, h) = P(\varphi_1, h) - [P(\varphi_1, h) - P(\varphi_2, h)] \left( \frac{\varphi - \varphi_1}{\varphi_2 - \varphi_1} \right) \]

\[ T(\varphi, h) = T(\varphi_1, h) - [T(\varphi_1, h) - T(\varphi_2, h)] \left( \frac{\varphi - \varphi_1}{\varphi_2 - \varphi_1} \right) \]

where \( \varphi_1 \) and \( \varphi_2 \) are, respectively, the values of the latitudes just above and below the latitude (\( \varphi \)) for which the data is to be retrieved. Following this procedure the mean altitude profiles of \( P \) and \( T \) up to 120 km are obtained for all the eighteen stations listed in Table 2.1.

2.4 Altitude Profiles of Dry and Wet Refractivity

The refractive index profiles are generated from the monthly profiles of \( P \), \( T \) and \( e \) using equations (2.9a) and (2.9b). As an example, the dry and wet refractivity for the month of June is presented in Figure 2.5 (a) and (b), respectively, for Trivandrum and Delhi. The horizontal bars indicate the standard deviation representing the amount of day-to-day variability. The value of \( N_w \) is largest near the surface (~130 units) and decreases almost exponentially with increase in altitude. Because of this, the refraction effects will be more at lower altitudes and decreases with increase in altitude. The dry component, \( N_D \), is much larger in magnitude than \( N_w \). The day-to-day variability of \( N_D \) is significantly small compared to that of \( N_w \). Note that though significant contribution of \( N \) comes from \( N_D \) the large variability in \( N_w \) contributes significantly for the day-to-day variation of \( N \).

![Figure 2.5: Typical altitude profiles of \( N_D \) and \( N_w \) for Trivandrum (a) and Delhi (b) in June](image-url)
2.4.1 Relative Contribution of Refraction Delay with Altitude

The refractivity of atmosphere is maximum near the surface and decreases with increasing altitude. It should be worth in this context to examine the relative contribution to total range error and bending angle from different altitudes. The cumulative contribution from different altitudes (in percentage) to $\Delta R$ and $\Delta \Psi$ estimated through ray-tracing for a zenith angle of $80^\circ$ is presented in Figures 2.6 and 2.7, respectively. The cumulative contribution to wet range error ($\Delta R_w$) up to 2 km is about 60% and is about 90% for altitudes below 5 km. The atmospheric region below 9 km contributes ~ 98% of the total $\Delta R_w$. For the dry component ($\Delta R_D$), the atmosphere below 6 km contributes about 50% of total dry range error and atmosphere below 16 km contributes nearly 90%. In case of total range error ($\Delta R$), which includes both dry and wet components, about 98% of contribution comes from the altitude region below 25 km.

Figure 2.7 shows the percentage of cumulative contribution of $\Delta \Psi$ along with its dry and wet components ($\Delta \Psi_D$ and $\Delta \Psi_w$ respectively) at different altitudes. The lower atmosphere below 5 km significantly contributes for $\Delta \Psi_w$ and atmosphere below 15 km for $\Delta \Psi_D$. The atmosphere up to 26 km contributes around 98% of total $\Delta \Psi$. According to the
Figure 2.7: Curves showing the relative contribution (percentage of the total effect due to the entire atmospheric column) from different altitude regions (starting from surface) to the total range error ($\Delta\Psi$). The contribution of dry (-----) and wet (•••••••••) components are also shown separately along with that of the total delay (———).

above analysis, the atmosphere up to 26 km altitude contributes about 98% of the total range error caused by entire neutral atmosphere. However, for GPS-based aircraft navigation, even the remaining 2% error is significant. Taking account of this fact, the neutral atmosphere up to 100 km is considered for estimating $\Delta R$ and $\Delta\Psi$.

2.5 Zenith Tropospheric Delay and its Spatial and Temporal Variability

The Zenith Tropospheric Delay (ZTD) comprises of two components viz., the Zenith Hydrostatic Delay (ZHD) and Zenith Wet Delay (ZWD) [Saastamoinen, 1972], which can be written as

$$ZTD = ZHD + ZWD = 10^{-6} \int_{h_s} N(h) \cdot dh$$

(2.27)

where

$$ZHD = 10^{-6} \int_{h_s} N_d(h) \cdot dh , \quad ZWD = 10^{-6} \int_{h_s} N_w(h) \cdot dh$$

and $h_s$ is the user altitude (it is the altitude of the GPS receiver antenna position). The mean altitude profile of $N$ along with amount of its day-to-day variability (due to the day-to-day variations of $P$, $T$ and $e$)
represented by the respective standard deviations $\sigma_P$, $\sigma_T$ and $\sigma_e$, respectively, are estimated for different months applying the error propagation formula [Ku, 1966]. The day-to-day variability in $N_D$ and $N_W$ is quantified separately based on the respective standard deviations. Thus the standard deviation of $N$ is estimated as,

$$\sigma_N^2 = \left( \frac{\partial N}{\partial P} \right)^2 \sigma_P^2 + \left( \frac{\partial N}{\partial T} \right)^2 \sigma_T^2 + \left( \frac{\partial N}{\partial e} \right)^2 \sigma_e^2$$

(2.28)

Where $n_P$, $n_T$ and $n_e$ are the number of daily values of Pressure, Temperature and water vapor partial pressure at a given altitude. The $\frac{\partial N}{\partial P}$, $\frac{\partial N}{\partial T}$ and $\frac{\partial N}{\partial e}$ are the partial derivative of $N$, about its mean values, which are generated by

$$\frac{\partial N}{\partial P} = \frac{k_1}{T}$$

(2.29)

$$\frac{\partial N}{\partial T} = -\frac{1}{T^2} \left[ k_1 P + k_2 e + 2 k_3 \frac{e}{T} \right]$$

(2.30)

$$\frac{\partial N}{\partial e} = \frac{k_2}{T} + \frac{k_3}{T^2}$$

(2.31)

where $k_1$, $k_2$ and $k_3$ are the coefficients as given in eq. (2.8).

For each month the mean values of ZTD, ZHD and ZWD are estimated separately using the three respective profiles of $N(h)$ along with those of $N(h) + \sigma_N(h)$ and $N(h) - \sigma_N(h)$. The difference between upper and lower limits represents the standard deviation of ZHD and ZWD. Figure 2.8 shows the month-to-month variation of mean ZHD for the eighteen stations along with the standard deviations (vertical bars) representing the amount of day-to-day variability. The annual mean of ZHD is around 2.3 m for all these stations, except for Srinagar, Bangalore, Hyderabad, Bhopal, Jodhpur and Patiala (Table 2.3). At Srinagar, which is situated at an altitude of ~1600 m above MSL, the annual mean value of ZHD is ~1.92 m and for Bangalore, which is situated at an altitude of ~930 m above MSL, it is ~2.08 m. Hyderabad and Bhopal are higher-latitude stations with altitude ~ 500 m above MSL where as Jodhpur and Patiala are mid-latitude stations with altitude ~ 250 m above MSL. The ZHD for all these stations are ~ 2.1 m. The amplitude of the annual variation in ZHD is very small for all the three tropical stations and increases with increase in latitude.
Figure 2.8a: Month to month variation of ZHD for Trivandrum, Portblair, Bangalore, Chennai, Mangalore, Goa, Hyderabad, Vishakhapatnam and Mumbai. Same ordinate scale is used for all plots to illustrate the variation of mean level from one station to the other.
Figure 2.8b: Month to month variation of ZHD for Kolkata, Ahmedabad, Bhopal, Guwahati, Jodhpur, Lucknow, Delhi, Patiala, and Srinagar. Same ordinate scale is used for all plots to illustrate the variation of mean level from one station to the other.
Maximum annual variation is observed at the high latitude stations beyond Guwahati, with a prominent maximum during the period Dec-Jan with an amplitude of \( \leq 15 \text{ cm} \). While the amplitude of the annual variation of mean ZHD for Trivandrum is about 0.34 cm and that for Delhi is \( \sim 1.5 \text{ cm} \), the day-to-day variability of ZHD in a month is around 7.0 cm indicating that in all the cases, the amplitude of the annual variation of mean ZHD is smaller than the day-to-day variability in each month. Note that, a large ZHD variation up to \( \sim 5.0 \text{ cm} \), is reported to be quite common \cite{Bock_and_Doerflinger, 2000} during synoptic-scale and meso-scale disturbances.

Figure 2.9 shows the month-to-month variation of the ZWD for different stations, with vertical bars representing the standard deviation indicating the amount of day-to-day variability caused by the corresponding variations in the atmospheric water vapor. The value of ZWD is small for the dry period (December to February) and large during the monsoon (July and August) period. Amplitude of the annual variation of ZWD is small for the tropical coastal/island stations (Trivandrum and Portblair) compared to that for Delhi, Kolkata, Ahmedabad and Guwahati etc. Note that the absolute values of ZWD at Trivandrum and Portblair are much larger than the corresponding values at other stations during the winter period. At Bangalore and Srinagar, the mean value of ZWD as well as the amplitude of its annual variation is small.

The monthly mean values of ZWD vary in the range \( -27 \text{ cm} \) to \( 37 \text{ cm} \) at Trivandrum and \( -25 \text{ cm} \) to \( 40 \text{ cm} \) at Portblair, while for the higher latitude stations like Delhi, Guwahati, Ahmedabad and Kolkata, the lowest value of ZWD ranges from \( 10 \text{ cm} \) to \( 15 \text{ cm} \), and the highest value exceeds 40 cm. This difference in the annual variation of ZWD at these stations is mainly caused by the large annual variability in atmospheric water vapor content at these stations \cite{Parameswaran_and_Krishna_Murthy, 1990}. It may be noted that at Trivandrum and Portblair as the temperature values are much larger than those at the higher latitude stations during the winter period, the atmosphere can hold relatively more water vapor, leading to much larger values of ZWD than those the other high latitude stations. During July-August period as the air temperature is relatively large at the higher latitude stations the atmosphere can hold relatively more water vapor (compared to near equatorial stations). Over and above the temperature effects, the massive synoptic system, "South-West monsoon", which sets in at the southern tip of the Indian peninsula around first week of June and steadily advances towards north reaching Delhi within a time span of around 20
Figure 2.9a: Month to month variation of ZWD for Trivandrum, Portblair, Bangalore, Chennai, Mangalore, Goa, Hyderabad, Vishakhapatnam and Mumbai. Same ordinate scale is used for all plots to illustrate the variation of mean level from one station to the other.
Figure 2.9b: Month to month variation of ZWD for Kolkata, Ahmedabad, Bhopal, Guwahati, Jodhpur, Lucknow, Delhi, Patiala, and Srinagar. Same ordinate scale is used for all plots to illustrate the variation of mean level from one station to the other.
days, is characterized by cool moist air, advecting from the Arabian Sea and spreading over the entire subcontinent, bringing abundant water vapor over to the continent. Advection of the moist air from the ocean to the heated continental atmosphere causes further increase in the atmospheric water vapor content, especially at the higher latitude stations. This adds to further enhancement in the value of ZWD at these stations during June-September period. The value of ZWD thus depends very much on the capacity of atmosphere to hold water vapor (without precipitating) as well as the supply of water vapor. The picture is quite different for high altitude stations where the atmosphere is relatively dry. At Bangalore the minimum value of ZWD is about 15 cm. Even though the value of ZWD increases during June-August period because of the increase in columnar water vapor, it remains less than 30 cm. Similar is the case for Srinagar, which is located at still higher altitude where the minimum value of ZWD is around 5 cm and the maximum seldom exceeds 25 cm, indicating that the value of ZWD decreases with increase in (station) altitude. At Bangalore the mean ZWD shows a broad peak during June to September while at Srinagar it shows a sharp peak during the July and August. The ZWD varies roughly from 0 to 30 cm between the poles and the equator and from a few cm to about 20 cm during the year at mid-latitudes [Elgered, 1993]. The 2nd and 3rd column of Table 2.3 shows the annual mean values of ZHD and ZWD along with the extent of its month-to-month variation within parenthesis.

As can be seen from this table, the month-to-month variability for both the delays is small for the two tropical stations than those for the other stations. Maximum month-to-month variability in ZHD ($\delta_{ZHD}$) is ≤ 3.8 cm and that of ZWD ($\delta_{ZWD}$) is 13.7 cm (which is almost one order of magnitude larger). The 4th and 5th columns in Table 2.3 show the amount of mean day-to-day variability (mean of standard deviation in each month), which varies from 4.0 cm to 8.5 cm for ZHD and from 7.2 cm to 15.8 cm for ZWD. Considering the day-to-day values of ZTD over a year, it varies by 31.4 cm over the tropics and by 63.6 cm over the mid-latitudes. This variation is quite significant as far as GPS based navigation is concerned. Table 2.3 also shows the maximum and minimum values of ZHD (6th and 7th columns) and ZWD (8th and 9th columns), for each station. Except for the two high altitude stations, the maximum value of ZWD varies from 43 cm (Patiala) to 64 cm (Kolkata) and the minimum value varies from 6.5 cm (Delhi) to 24 cm (Trivandrum). The range of variability of ZHD is relatively small compared to that of ZWD. As seen from the Table 2.3, the day-to-day variability of tropospheric range error due to variations in the atmospheric
parameters could be as large as 15 cm, which is quite significant for navigation and hence calls for the development of fairly accurate empirical models based on easily measurable atmospheric parameters.

Table 2.3: Table Summarizing the Annual Mean Value of ZHD and ZWD for Different Indian Stations along with its Possible Deviation (Standard Deviation) from Month-to-Month. To Illustrate the Maximum Possible Deviation on a Day-to-Day Basis, the Maximum and Minimum Values of ZHD and ZWD Encountered during the Course of an Year is presented in the Last Four Columns

<table>
<thead>
<tr>
<th>Station</th>
<th>ZHD (δZHD) (cm)</th>
<th>ZWD (δZWD) (cm)</th>
<th>Standard deviation of daily ZHD (cm)</th>
<th>Standard deviation of daily ZWD (cm)</th>
<th>Maximum and minimum values of daily ZHD calculated over a period of one year</th>
<th>Maximum and minimum values of daily ZWD calculated over a period of one year</th>
</tr>
</thead>
<tbody>
<tr>
<td>Trivandrum</td>
<td>231.4 (0.3)</td>
<td>34.1 (3.2)</td>
<td>4.5</td>
<td>13.6</td>
<td>Max. (cm) 232.8 230.3 52.8 23.9</td>
<td>Min. (cm) 230.3 234.7 50.0 16.9</td>
</tr>
<tr>
<td>Port Blair</td>
<td>231.3 (0.5)</td>
<td>34.5 (5.0)</td>
<td>4.3</td>
<td>12.7</td>
<td>Max. (cm) 232.1 229.0 50.0 16.9</td>
<td>Min. (cm) 230.0 229.0 50.0 16.9</td>
</tr>
<tr>
<td>Bangalore</td>
<td>208.6 (0.5)</td>
<td>24.2 (4.8)</td>
<td>4.0</td>
<td>11.7</td>
<td>Max. (cm) 209.5 207.1 38.4 20.4</td>
<td>Min. (cm) 206.8 204.7 38.4 20.4</td>
</tr>
<tr>
<td>Chennai</td>
<td>230.2 (0.7)</td>
<td>30.7 (7.9)</td>
<td>4.5</td>
<td>15.1</td>
<td>Max. (cm) 231.7 221.0 51.7 17.3</td>
<td>Min. (cm) 230.0 220.7 51.7 17.3</td>
</tr>
<tr>
<td>Mangalore</td>
<td>230.4 (0.4)</td>
<td>31.7 (5.1)</td>
<td>4.4</td>
<td>13.4</td>
<td>Max. (cm) 231.3 222.0 56.3 22.1</td>
<td>Min. (cm) 229.7 220.3 56.3 22.1</td>
</tr>
<tr>
<td>Goa</td>
<td>229.2 (0.6)</td>
<td>29.9 (7.3)</td>
<td>4.9</td>
<td>14.5</td>
<td>Max. (cm) 231.7 228.6 47.9 18.0</td>
<td>Min. (cm) 230.0 227.3 47.9 18.0</td>
</tr>
<tr>
<td>Hyderabad</td>
<td>211.0 (0.8)</td>
<td>28.5 (7.1)</td>
<td>5.1</td>
<td>14.8</td>
<td>Max. (cm) 219.7 204.0 55.6 14.9</td>
<td>Min. (cm) 218.0 203.2 55.6 14.9</td>
</tr>
<tr>
<td>Vishakhapatnam</td>
<td>227.3 (1.1)</td>
<td>30.5 (6.6)</td>
<td>5.1</td>
<td>15.8</td>
<td>Max. (cm) 229.0 225.6 39.3 20.7</td>
<td>Min. (cm) 226.0 222.3 39.3 20.7</td>
</tr>
<tr>
<td>Mumbai</td>
<td>230.3 (0.8)</td>
<td>30.0 (10.1)</td>
<td>5.5</td>
<td>15.8</td>
<td>Max. (cm) 231.0 228.8 45.3 17.0</td>
<td>Min. (cm) 228.8 226.0 45.3 17.0</td>
</tr>
<tr>
<td>Kolkata</td>
<td>230.7 (1.3)</td>
<td>31.1 (11.4)</td>
<td>5.0</td>
<td>14.6</td>
<td>Max. (cm) 233.2 226.6 63.8 17.9</td>
<td>Min. (cm) 230.0 223.4 63.8 17.9</td>
</tr>
<tr>
<td>Ahmedabad</td>
<td>229.4 (1.2)</td>
<td>24.4 (10.1)</td>
<td>5.5</td>
<td>13.5</td>
<td>Max. (cm) 231.4 224.9 55.0 10.0</td>
<td>Min. (cm) 228.0 221.6 55.0 10.0</td>
</tr>
<tr>
<td>Bhopal</td>
<td>205.4 (0.9)</td>
<td>19.3 (7.9)</td>
<td>6.4</td>
<td>11.9</td>
<td>Max. (cm) 207.2 200.0 51.5 10.6</td>
<td>Min. (cm) 205.0 197.2 51.5 10.6</td>
</tr>
<tr>
<td>Guwahati</td>
<td>229.6 (1.3)</td>
<td>30.7 (11.2)</td>
<td>5.3</td>
<td>11.8</td>
<td>Max. (cm) 231.8 223.4 54.7 15.2</td>
<td>Min. (cm) 229.0 221.2 54.7 15.2</td>
</tr>
<tr>
<td>Jodhpur</td>
<td>217.8 (3.8)</td>
<td>20.6 (9.8)</td>
<td>7.3</td>
<td>13.2</td>
<td>Max. (cm) 228.9 197.0 58.0 11.9</td>
<td>Min. (cm) 226.0 194.7 58.0 11.9</td>
</tr>
<tr>
<td>Lucknow</td>
<td>224.4 (1.5)</td>
<td>25.4 (13.7)</td>
<td>8.5</td>
<td>15.0</td>
<td>Max. (cm) 226.9 222.0 51.5 11.9</td>
<td>Min. (cm) 225.0 220.7 51.5 11.9</td>
</tr>
<tr>
<td>Delhi</td>
<td>224.6 (1.6)</td>
<td>23.5 (12.2)</td>
<td>5.7</td>
<td>13.1</td>
<td>Max. (cm) 228.2 218.0 59.9 6.5</td>
<td>Min. (cm) 226.0 216.7 59.9 6.5</td>
</tr>
<tr>
<td>Patiala</td>
<td>211.8 (1.35)</td>
<td>17.7 (9.3)</td>
<td>7.4</td>
<td>12.2</td>
<td>Max. (cm) 223.0 209.2 42.7 12.8</td>
<td>Min. (cm) 221.0 207.2 42.7 12.8</td>
</tr>
<tr>
<td>Srinagar</td>
<td>191.8 (1.1)</td>
<td>13.5 (6.4)</td>
<td>7.2</td>
<td>7.2</td>
<td>Max. (cm) 192.5 179.8 38.0 11.7</td>
<td>Min. (cm) 190.0 177.6 38.0 11.7</td>
</tr>
</tbody>
</table>

2.6 Summary

When microwave propagates through the neutral atmosphere, it undergoes refraction depending on the real component of the refractive index of the medium and attenuation
depending on the imaginary part. For GPS based aircraft navigation employing microwave ranging, the delay of wave propagation (caused by the wave) due to refraction and subsequent error in positioning is a major challenging task. The delay due to neutral atmosphere consists of two components. While the hydrostatic (dry) component, which depends primarily on atmospheric pressure and temperature, accounts for approximately 90% of the total delay, the wet component, which purely depends on the moisture content of the atmosphere, accounts for the remaining part. The delay caused by the neutral atmosphere depends on the geographical location as well as prevailing climatic conditions. Most accurate estimation of the Tropospheric delay could be achieved by ray-tracing employing the altitude profiles of neutral atmosphere refractivity which is a function of $P, T$ and $e$. The tropospheric range error in the zenith direction as well as its spatial and temporal variabilities over the Indian sub-continent is examined by selecting eighteen appropriately located stations. It is also observed that hydrostatic component (ZHD) and non-hydrostatic component (ZWD) of tropospheric range error at all these stations have shown pronounced day-to-day variation (~ 4.0 cm to ~ 8.5 cm for ZHD and ~ 7.0 cm to ~ 15.8 cm for ZWD). While the month-to-month variations of mean ZHD ranges ~ 0.5 cm to ~ 3.8 cm that of ZWD is about ~ 5.0 cm to ~ 14 cm. Even though the tropospheric range error can be estimated within reasonable accuracies through ray tracing, the information on the altitude profiles of the atmospheric parameters may not be easily available for many places at any given time. To resolve this problem, it is highly essential to develop simple models based on the geographical location and time of the year and/or based on easily available atmospheric parameters at the surface in line with global efforts.