MODELLING PERTURBATIONS
PROPAGATING THROUGH THE MESOPAUSE
INTO THE EARTH'S UPPER ATMOSPHERE

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"Scientific discoveries cannot be justified - they can only be criticised."

Karl Popper
Global oscillations formed in the terrestrial troposphere, stratosphere and mesosphere propagate into the thermosphere and ionosphere where they change the dynamics, energy and composition. This thesis presents a series of studies which examine in detail the nature and influence of solar tides and the planetary 2-day wave above 80 km altitude. The Coupled Thermosphere-Ionosphere Model (CTIM) calculates self-consistently the dynamics, energy and composition of the terrestrial thermosphere and ionosphere in three dimensions and is used as the main tool in these studies. In order to simulate the upwardly propagating perturbations which are formed outside the height range of the model, the lower boundary of the CTIM at 80 km height was modified to allow the global profiles of pressure-, wind- and temperature oscillations to be specified. In principle, following the modification, any such profile can be used for the external forcing as long the parameters at the lower boundary are self-consistent. One effective method of achieving this is to specify global perturbations of geopotential height, using Hough functions for the latitudinal structure, and calculating the simultaneous wind- and temperature oscillations at the lower boundary analytically with expressions from Classical Tidal Theory. The necessary formalism for this has been fully implemented. For validation of the new code a series of comparisons with other numerical models and Incoherent Scatter Radar measurements at equinox and solstice are presented and show that CTIM is capable of reproducing many tidal features found in the “real” thermosphere. A further study is presented which investigates processes causing planetary wave signatures in the ionosphere. It is found not only that CTIM reproduces some key properties of upwards propagating planetary waves found in other theoretical and modelling studies, but also that upwards propagating tides may, through modulation of their amplitudes, carry planetary wave signatures into the 200 km height regime where they are transferred into the ionosphere by chemical processes. The new CTIM thus offers the possibility of carrying out many unprecedented studies exploring the nature of the Earth’s upper atmosphere.
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FIGURE VI.1.: Neutral composition of the upper atmosphere, under average solar- and geomagnetic conditions. The plot shows globally averaged number densities in units of $\log_{10} [m^2]$ for atomic oxygen (solid), molecular oxygen (dashed) and molecular nitrogen (dotted).

FIGURE VI.2.: Illustration of the role of horizontal winds in causing composition changes. The blue and read filled circles represent air parcels which have been displaced vertically through tidally induced up- and downwelling. At time $t_A$ the parcels have a distance of $A$, at time $t_B$ their horizontal distance is $B$. If time scales for horizontal wind transport are smaller than the perturbation period and furthermore the horizontal winds are different for the red and blue parcels, the distances $A$ and $B$ are different, thus causing departure from diffusive equilibrium.

FIGURE VI.3.: Height of pressure level 7 (dashed) in a regular simulation (top) and when setting horizontal gradients in the model artificially to zero (bottom). Also shown are heights of levels of constant $O/O_2$ ratios (solid). The ratios have values of 2.1 (top) and 3.0 (bottom).

FIGURE VI.4.: Heights of pressure level 6 (top) and 8 (bottom) from the regular CTIM run used in Figure VI.3. (top). Solid curves are the heights of constant $O/O_2 = 0.85$ (top) and $O/O_2 = 4.2$ (bottom) ratios.

FIGURE VI.5.: Semidiurnal amplitudes of the $O/O_2$ ratio, as modelled by CTIM with (2,4) mode forcing. The values shown are 12 h amplitudes in percent of the mean background aspect ratio.

FIGURE VI.6.: The change (in %) of background $O/O_2$ ratio caused by lower boundary (2,4) forcing (top). The bottom plot shows the change of
O/O₂ ratio at latitude 0° (solid) as well as the changes of O density (dashed) and O₂ density (dotted). Positive values denote increase of the ratio or densities due to the tides.

FIGURE VI.7.: Temperature of pressure level 5 (solid) and at fixed heights (dashed) of 106.1 km (top) and 107.2 km (bottom). Values in the bottom plot are taken from the CTIM simulation in which horizontal wind gradients were set to 0.

CHAPTER VII

FIGURE VII.1.: Quasi-2-day wave amplitudes at 80 km used at CTIM’s lower boundary. Values are based on simulations by [Hagan et al., 1993].

FIGURE VII.2.: Amplitudes of quasi-2-day meridional- and zonal winds, as modelled by CTIM in Run A, using planetary wave forcing at the lower boundary and no tides. The simulation was carried out for January conditions.

FIGURE VII.3.: Same as Figure VII.2., but for temperatures and electron densities. Electron density values are given as ratios of amplitude to background value (in %).

FIGURE VII.4.: Same as Figure VII.2., but for Run C, using modulated tidal amplitudes at the lower boundary.

FIGURE VII.5.: Same as Figure VII.3., but for Run C, using modulated tidal amplitudes at the lower boundary.
CHAPTER I. BACKGROUND

ABSTRACT

This chapter introduces the basic properties of the Earth’s atmosphere, its vertical structure and the main processes governing each of the vertical layers. The terms of lower and upper atmosphere are introduced, with the latter denoting the regions above 80 km altitude. Upper atmosphere processes such as ion-neutral interactions at high latitudes and the low-latitude dynamo effect are described in more detail as well as their relevance to the propagation of tides. A review is then presented which outlines the history of tidal studies in the oceans and atmosphere, describing some key problems which have occupied researchers for the past century, such as Lord Kelvin’s Resonance Theory. The most common techniques used nowadays for atmospheric measurements are also described, with particular emphasis on remote sensing methods using radar instruments. Finally, the historic development of numerical models which simulate the Earth’s atmosphere is presented. Differences between various families of models are outlined and their various advantages and disadvantages discussed. By describing the important atmospheric models used nowadays the Coupled Thermosphere Ionosphere Model (CTIM) is put into context with other available models. The chapter concludes with a discussion of practical difficulties typically encountered when operating large numerical models.

I. 1. INTRODUCTION

In the terrestrial environment a large number of periodic events occur, many of which have a profound influence on life. Perhaps the most apparent of these are the variations of day and night and of season. The periods of oscillations span a wide spectrum ranging from seconds, minutes and days to 11 years (the duration of one solar cycle) and more, up to galactic time scales. The main subject of this thesis are tides, a group of oscillations with periods of one day and fractions thereof which occur in the Earth’s oceans as well as its atmosphere. In Chapter VII, planetary waves are studied, which are global oscillations with periods in the order of several days. Although this thesis will investigate atmospheric perturbations only, the mathematical framework presented may in principle be applied to the ocean as well. This chapter will outline the main properties of the Earth’s atmosphere, review the historic development of tidal research and describe techniques used
in the past and today for studying the atmosphere. The use of numerical models will also be addressed by discussing their advantages and limits.

I. 2. THE TERRESTRIAL ATMOSPHERE

Among the planetary atmospheres found in our solar system the Earth's is extraordinary in that it is highly non-uniform both horizontally and vertically. The chemistry, composition, dynamics and temperature change significantly with altitude and the study of these properties thus faces many challenges. Various criteria can be used to describe the atmosphere's vertical structure, each defining the regions with respect to a different property. The most common definition uses the reversal of the vertical temperature gradient to define the layers, as shown in Figure 1.1. A distinction is made between five main regions, the troposphere, stratosphere, mesosphere, thermosphere and exosphere. In the following, the terms lower and upper atmosphere refer to the regions below and above 80 km, respectively. Since studies presented in this thesis will focus on the upper atmosphere, the main properties of the lower regions are only addressed briefly.

![Figure 1.1](image_url)

**Figure 1.1:** Regions of the atmosphere as defined by the temperature gradient (left side) and degree of mixing of constituents (middle). The structure of the ionosphere is shown on the right side, as defined by the electron density. Electron density values are for average daytime conditions at mid-latitudes for high solar activity. [from Rishbeth and Garriott, 1969].
I. 2. 1. THE LOWER ATMOSPHERE

In the atmosphere below around 80 km, the average vertical temperature gradient changes sign twice and thus defines three main regions. The **troposphere** defines the lowest layer in the atmosphere, ranging from the ground to around 12 km altitude, a value which can vary by up to 50%, depending on the latitude. Temperature in the troposphere falls with altitude from around 280°K at ground level to around 200°K at its upper boundary, the **tropopause**. The principal source of energy in the troposphere is the absorption of solar radiation, either directly by water vapour in the air or indirectly by the ground which absorbs at higher frequencies and then re-emits in the infrared spectrum. Temperature typically decreases vertically at a rate of 7°K/km, which is smaller than the critical value of 10°K/km, the **adiabatic lapse rate** for dry and transparent air. The adiabatic lapse rate can be used to evaluate the stability of the troposphere. If the temperature decreases more strongly with height than the adiabatic lapse rate, a parcel transported upwards adiabatically will have a higher temperature than the surrounding gas and continue to rise. This describes an unstable situation. If the rate of temperature decrease is smaller than the adiabatic lapse rate, the parcel transported upwards adiabatically will have a lower temperature than the surrounding gas and fall again. The atmosphere is then stable. This stability is normally maintained in the troposphere and only occasionally upset during nighttime when the heat released by the ground can lead to a sharp temperature gradient near ground level. Convective motion such as turbulence and vertical winds still occur though, since the warmer air is at lower altitudes and thus easily rises. The dynamics of the troposphere are driven exclusively by pressure gradients and Coriolis force. A complex system of large circulation cells is generated which extend over the entire height regime of the troposphere. The exact shape of these cells depends largely on ground texture and the varying surface albedo. When near-surface horizontal winds encounter changes in the texture height (such as mountains), they are driven up- or downwards and cause turbulence. This generates gravity waves which then propagate upwards in the atmosphere and play an important role in higher regions, as outlined later in this chapter.

Above the tropopause lies the **stratosphere**, a layer which is characterized by its positive vertical temperature gradient. This rise in temperature with altitude is caused by the absorption of solar radiation energy through ozone (O₃). In spite of the ozone’s relatively small concentration, its presence is important in that it not only causes this temperature behaviour in the stratosphere but
also absorbs most solar ultraviolet radiation of less than 290 nm wavelength, which is lethal to all life. The stratospheric heating is furthermore essential for the generation of tides. Since the temperature gradient is positive in the stratosphere, warmer air occurs at higher altitudes. The stratosphere is thus very stable and vertical mixing occurs considerably less than in the troposphere and mesosphere. The most effective vertical stratospheric mixing happens in the winter hemisphere and is caused by planetary waves propagating upwards from the troposphere. The radiative heating also implies that temperatures are higher in the summer- than in the winter hemisphere, thus causing a north-south temperature gradient. This gradient with the Coriolis forces causes strong zonal winds of up to 60 m/s which are eastward in the summer- and westward in the winter hemisphere.

Above around 50 km the vertical temperature gradient reverses again to a negative value and defines the mesosphere. This reversal in the mesosphere is caused by the sharply decreasing ozone concentration and hence associated heating above around 50 km. The thermal structure of the mesosphere is controlled mainly by the balance between 15 μm band cooling by carbon dioxide (CO₂) and relatively weaker radiative heating by O₃. In addition, though, considerable vertical mixing occurs, with upwards transport in the summer- and downwards transport in the winter hemisphere [Murgatroyd and Singleton, 1961]. The mesopause lies at around 80-85 km and forms the upper boundary of the mesosphere. One characteristic property of this altitude is the mesopause temperature anomaly which consists of the summer hemisphere being considerably colder than the winter hemisphere. With temperature values of as little as 140°K the summer mesopause is the coldest region in the atmosphere [Rishbeth and Garriott, 1969]. This temperature minimum, although forming an energy sink, is of little importance to the overall energy of the mesosphere since it is outweighed in magnitude by ozone heating further down. In contrast, the thermosphere with its considerably lower density is affected by the temperature minimum and heat is conducted downwards into the region.

I. 2. 2. THE UPPER ATMOSPHERE

Above around 80 km the atmosphere changes considerably, as compared to the lower regions. The main energy source is solar radiation in the ultraviolet (UV) and extreme ultraviolet (XUV) while cooling occurs mainly through downwards conduction of heat. Below around 300 km heating dominates and the average temperature increases considerably with height, whereas at higher
altitudes the heat conduction becomes more effective and the temperature reaches a saturation value. This exospheric temperature is dependent on the time of day and the level of solar activity. At sunspot minimum, values can reach around 1200°K during daytime and 800°K during the night. At high solar activity level, daytime values can lie around 2200°K and nighttime values around 1600°K. The thermosphere defines the region above the mesopause in which neutral particles collide frequently enough to influence each other considerably. As density decreases with altitude, particle collisions become less frequent. Above an altitude of around 400-600 km, depending on solar activity, neutral particles hardly interact and move undisturbed, mainly under the influence of gravity. This upper region of the neutral atmosphere is referred to as the exosphere.

![Figure 1.2: Average daytime ionospheric and atmospheric composition for sunspot minimum conditions](image)

In Figure 1.2, average daytime composition is shown for the atmosphere above 90 km. One important feature of neutral composition in the upper atmosphere is the separation of constituents. Up to the turbopause at an altitude of around 100 km, constituents are uniformly mixed by turbulence. Although the concentration of chemically active constituents such as ozone changes with altitude the relative abundance of major constituents is constant at about 78% by volume molecular nitrogen (N₂) and 21% molecular oxygen (O₂). Above the turbopause, however, turbulent mixing diminishes rapidly and constituents are separated by diffusion. Each constituent is distributed vertically according to its individual scale height to satisfy the hydrostatic equation (II.2). Figure 1.2 shows that the density of O₂ decreases more with height than the N₂ density and that atomic oxygen (O) appears to replace O₂ above around 120 km. The reason for this behaviour is that O₂ is dissociated into O, while N₂, due to its much smaller photo-dissociation cross-section,
is to a less extent affected by the radiation. In Chapter VI, neutral composition of the thermosphere is discussed in more detail. In the lower thermosphere, dynamics are largely controlled by pressure gradients and Coriolis force, but furthermore ion- and viscous drag play an important role. Ion-neutral interactions are discussed in sections 1.2.3 and 1.2.4.

It was mentioned earlier that the two main processes controlling the heat balance in the thermosphere are absorption of solar radiation in the UV and XUV bands and downwards conduction of heat. Further important sources of energy and momentum in the thermosphere are dissipation of upwards propagating oscillations and Joule heating. The latter is caused by ionospheric electric currents and occurs mainly at high latitudes where the strongest currents are found. Tides and planetary waves propagate upwards from the lower- and middle atmosphere into the thermosphere and through processes such as turbulence, ion- and viscous drag are damped effectively below around 200 km height. The associated release of energy and momentum has a major influence on thermospheric energy, momentum and composition, as shown in Chapters IV-VII. The same is valid for gravity waves but their influence is somewhat more complicated since they not only dissipate but under certain circumstances also break in the lower thermosphere. The mechanism of breaking involves heavy turbulence and is triggered by local temperature gradients being forced by the wave to exceed the adiabatic lapse rate [Schoeberl et al., 1983]. As a gravity wave is damped by the turbulence it becomes an important source of momentum [Fritts, 1984]. Hines [1965] estimated the heat release due to gravity waves to be of a similar order to the heat input from solar radiation below 150 km. Garcia and Solomon [1985] showed that breaking gravity waves also influence the composition in the lower thermosphere by changing the rate of turbulent diffusion. In order to understand processes in the thermosphere, therefore, it is important to consider the propagating oscillations. The aim of this thesis is to study the effects of larger scale oscillations, while gravity waves are not addressed.

With the absorption of high energy solar radiation, molecules are not only split into atoms but neutral particles are also ionized and form a conducting layer in the atmosphere. The existence of a conducting layer was first suggested by C.F.Gauss in 1839 when he observed variations of the geomagnetic field and associated them with atmospheric electric currents. Although Lord Kelvin also suggested the existence of a conducting layer, no progress on the issue was made until in 1900 J.J.Thomson discovered the electron. In 1901, G.Marconi successfully transmitted radio signals across the Atlantic and proved that diffraction of the waves alone could not have explained this.
The term *ionosphere* was first suggested by *R.A. Watson-Watt* in 1926 and is somewhat loosely defined as the region in the upper atmosphere where electron- and ion densities are large enough to influence the propagation of radio waves. The ionosphere ranges down into mesospheric heights, around 60 km, but overlaps mostly in terms of height with the thermosphere. Its upper boundary is not well defined, but above around 1000 km the ionosphere merges into the *magnetosphere*, an outer region where ion motion is controlled entirely by the Earth’s magnetic field. The ionospheric structure is defined by the electron density, as shown on the right side of Figure I.1. The layers are given names suggested by *E.V. Appleton*, using the letters *D, E* and *F*, and the daytime *F*-layer is further subdivided into the *F1* and *F2* layers. The choice of these letters for naming the layers originates from the fact that the first layer to be discovered was named *E*-layer as a short form for “electric field layer”, and other letters were then assigned to the surrounding layers in an alphabetically logical order. Each layer represents a local maximum of the vertical electron density profile and the term ‘region’ is used for the heights surrounding the layers. Of these, the *F* (or *F2*) layer usually has the highest electron density and also represents the highest level at which radio waves are reflected back to the Earth’s surface. This is because radio waves are reflected when the plasma frequency is equal to the wave frequency. Since the plasma frequency is proportional to the square root of the electron density [*Rishbeth and Garriott, 1969*], the highest frequencies will be reflected by the region of highest electron density, which is the *F*-region. As a consequence, any structure above the *F*-layer will be invisible to the propagation of radio waves from below. Since the existence of charged particles is directly linked to solar radiation, the ionospheric structure is different during day and night. The electron- and ion densities decrease in general and a number of features such as the *D*-region and the subdivision of the *F*-layer into the *F1* - and *F2* layers disappear. However, transport of ions and electrons from the day- to the night side (especially at high latitudes), long recombination times and a number of other processes ensure that the *E*- and *F*- layers remain also during nighttime.

I. 2. 3. NEUTRAL-ION INTERACTIONS BY COLLISIONS

It was mentioned earlier that neutral particles and ions in the upper atmosphere interact and influence each other. The relative influence of neutrals and ions is expressed by the ion-neutral and the neutral-ion collision frequencies. Of these, the former expresses the rate at which ions encounter neutral particles and the latter gives the rate at which neutrals encounter ions. A detailed investigation of collisions between ions and neutrals was carried out by *Chapman and Cowling*
[1952], and Rishbeth and Garriott [1969] present a simplified approach which, although not accurate in the detail, illustrates the basic principles. One may define an effective collision frequency in terms of effective transfer of momentum during a collision, as opposed to a definition in terms of the number of collisions between neutral and charged particles only. Following [Chapman, 1956], one may also define a rate \( K_n \), the collision rate coefficient corresponding to transfer of momentum between ions and neutrals (per unit volume), and obtains the effective collision frequencies by multiplying this factor with the relevant number density. The ion-neutral collision frequency is then given by \( \nu_{in} = n K_n \) and the neutral-ion collision frequency by \( \nu_{ni} = N K_n \), where \( n \) and \( N \) are the number densities of neutrals and ions, respectively. Using the expression for \( K_n \) given by Chapman [1956], one can calculate effective collision frequencies for day- and nighttime conditions in the thermosphere, as shown in Table I.1 for three heights. In order to obtain information about ion motion, it is necessary to take into account another quantity, the gyrofrequency. The ion gyrofrequency gives the rotation rate of ions around the magnetic field lines in the atmosphere and can be calculated from \( \omega_i = B i/m \), where \( B \), \( i \) and \( m \) are the magnetic field strength, ion charge and ion mass, respectively. If no neutral particles were present in the atmosphere the ions would move in spirals along the magnetic field lines. In the presence of neutral particles the movement of ions depends on the collision frequency with neutrals. If the collision frequency is much smaller than the gyrofrequency (ie. \( \nu_{in}/\omega_i \ll 1 \) ), ions will hardly be affected on their trajectories around the magnetic field lines by the neutral winds. If the collision frequency is larger than the gyrofrequency (ie. \( \nu_{in}/\omega_i > 1 \) ) the ions will collide with neutrals before completing a cycle around the field lines and effectively follow the motion of the neutrals. Table I.1 shows typical values of \( \nu_{in}/\omega_i \) in the upper atmosphere. It is evident that at E-region heights (around 120 km) the gyro- and collision frequencies roughly balance each other, while at higher altitudes ions move mainly around the magnetic field lines.
Background

Chapter I

Table 1.1: Neutral-ion ($v_{ni}$) and ion-neutral ($v_{n}$) collision frequencies in the upper atmosphere. Also shown are average ion gyrofrequencies ($\omega_i$), neutral- ($n$) and ion ($N$) densities as well as the ratio of ion-neutral collision frequency to gyrofrequency ($v_{ni}/\omega_i$). Adapted from [Rishbeth and Garriott, 1969].

<table>
<thead>
<tr>
<th></th>
<th>100 km</th>
<th></th>
<th>200 km</th>
<th></th>
<th>300 km</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>day</td>
<td>night</td>
<td>day</td>
<td>night</td>
<td>day</td>
</tr>
<tr>
<td>$n$ [m$^{-3}$]</td>
<td>1.2·10$^{19}$</td>
<td></td>
<td>7.6·10$^{15}$</td>
<td></td>
<td>9.8·10$^{14}$</td>
</tr>
<tr>
<td>$N$ [m$^{-3}$]</td>
<td>1.7·10$^{11}$</td>
<td>0.1·10$^{11}$</td>
<td>3.5·10$^{11}$</td>
<td>0.3·10$^{11}$</td>
<td>11·10$^{11}$</td>
</tr>
<tr>
<td>$\omega_i$ [rad s$^{-1}$]</td>
<td>160</td>
<td>300</td>
<td>300</td>
<td>300</td>
<td>300</td>
</tr>
<tr>
<td>$v_{ni}$ [s$^{-1}$]</td>
<td>8.3·10$^{-5}$</td>
<td>0.5·10$^{-5}$</td>
<td>20·10$^{-5}$</td>
<td>1.6·10$^{-5}$</td>
<td>65·10$^{-3}$</td>
</tr>
<tr>
<td>$v_{n}$ [s$^{-1}$]</td>
<td>5800</td>
<td>4.1</td>
<td>0.57</td>
<td>0.4</td>
<td></td>
</tr>
<tr>
<td>$v_{ni}/\omega_i$</td>
<td>36</td>
<td>1.4·10$^{-2}$</td>
<td>1.9·10$^{-3}$</td>
<td>1.3·10$^{-3}$</td>
<td></td>
</tr>
</tbody>
</table>

The table also illustrates that neutral-ion collision frequencies are much smaller than the ion-neutral frequencies. This is because ion densities are lower than the neutral densities throughout the atmosphere (see Figure 1.2) and an ion will thus be much more likely to encounter a neutral particle than a neutral encountering an ion. The influence of ions on neutral particles depends not only on the ion-neutral collision frequency but also on the difference between ion- and neutral velocities (see also III.3.3). Because of this, the ion-neutral coupling depends on altitude, as shown above, but also on latitude. While at low latitudes the ion velocities are fairly similar to the neutral velocities (except in the region of the equatorial electrojet), they are much larger at high latitudes where the convection electric field accelerates the ions to average velocities of the order of a few hundred m/s. In the real atmosphere the convection occurs in “bursts” according to the direction of the Interplanetary Magnetic Field (IMF) $B_z$ component and the phase of the substorm cycle, giving peak ion velocities of up to a few km/s. Therefore, the neutral particles in the atmosphere above around 150 km are at high latitudes influenced considerably by the ion motion.

1.2.4. THE DYNAMO EFFECT

The dynamo effect, or dynamo theory, refer to neutral-ion coupling in the lower thermosphere at mid- to low latitudes. The theory was first suggested by Balfour Stewart in 1882 and developed
quantitatively by A. Schuster in 1908. More recent comprehensive accounts include those by Maeda and Kato [1966], Volland [1984] and Richmond [1991]. Winds at mid- to low latitudes move the charged particles through the Earth’s magnetic field and generate electromotive forces. These electromotive forces are different for electrons and ions, and therefore an effective current flows. The highly anisotropic conductivities do not allow the resulting currents to flow freely, and therefore polarization charges build up (see also III.3.7) which lead to an electric field. The dynamo field is generated mainly in the E-region (around 120 km) since Hall- and Pedersen conductivities are only there large enough to enable the electric current flow which leads to the build-up of polarization charges. However, the highly conducting magnetic field lines transmit the field into the F-region where it leads to electromagnetic drifts, similar to the high latitude convection field being mapped down into the F region from above, as described in III.3.8. A weaker dynamo electric field is also generated in the F-region [Rishbeth, 1971 a,b]. Modelling this complex dynamo coupling presents a major challenge and so far only two self-consistent models have been developed, by Richmond et al. [1992] and Namgaladze et al. [1990]. An empirical model of the low-latitude electric dynamo field was presented by Richmond et al. [1980]. Considering that neutral winds at E-region altitudes are strongly tidal it is evident that these tidal oscillations are reflected in the dynamo electric field. This may then cause plasma drifts in the F-region and essentially carry the tidal oscillations into those altitudes. Semidiurnal patterns have been found in measurements of ionospheric electric fields [Richmond et al., 1976; Takeda, 1990], but to-date the measurements of semidiurnal effects in the ionosphere have not yet been successfully reproduced by any modelling attempts which generally over-estimate the effect [Richmond, 1995].
In Figure 1.3, a typical potential profile of the dynamo electric field is shown at 300 km altitude, as generated by the empirical model of Richmond et al. [1980]. The semidiurnal pattern at mid-latitudes can be seen and was found to be generated by semidiurnal winds in the lower thermosphere with amplitudes described by Fesen et al. [1986]. The dynamo effect is also thought to cause observed day-to-day irregularities in ionospheric electrodynamics. A large number of authors have investigated the reason for these ionospheric irregularities and generally agreed that these effects are caused by irregularities in the lower thermosphere winds, either caused by solar and lunar tides and gravity waves [Stening, 1975; Marriott et al., 1979] or by planetary waves [Stening, 1977]. The dynamo effect therefore needs to be considered when examining ionospheric properties.

It is of interest to examine also whether the dynamo-induced plasma winds will notably influence the neutral thermospheric winds. It was shown previously (section 1.2.3) that in the lower thermosphere at mid-to low latitudes the ions followed the neutrals, rather than the opposite. Although the F-region dynamo electric field does influence the F-region ions, this is at low- to mid-latitudes unlikely to be fed back to the neutrals. It is therefore unlikely that the dynamo effect will
cause any feedback on the neutral winds in the lower thermosphere regions primarily studied here. Since this thesis investigates mainly the neutral atmosphere, it is not essential to consider the dynamo forcing. At high latitudes, ionospheric forcing of neutral winds is important and therefore included in the model simulations.

I. 3. HISTORY OF TIDAL RESEARCH

Atmospheric tides are global oscillations of parameters such as pressure, density, temperature and winds which are forced by two different mechanisms, the thermal heating by solar radiation and the gravitational forces of the Moon and the Sun. The discovery of atmospheric tides is relatively young and detailed studies have only been carried out in the second half of this century. Most of the mathematical theory describing tidal oscillations in the atmosphere is based on a study by Laplace which investigated ocean tides. In order to outline the development of atmospheric tidal theory, one therefore needs to begin by looking back to times when the well known periodic changes of the sea level were first investigated analytically.

I. 3.1. THE EARLIEST STUDIES

In the 6th century B.C. the Pythagorean school in Greece triumphed with revolutionary findings in science and mathematics. Ironically, however, no mathematical study on sea tides has been delivered by that school although it might have seemed an obvious field of study. The reason for this might be that the Mediterranean Sea, around which the old Greek and Roman Cultures evolved, behaves more like a lake and shows only very weak tidal effects. The earliest known study on tides was carried out around 320 B.C. by a famous explorer-mariner from Marseilles named Pytheas. He first made tidal observations on a long voyage outside the Mediterranean, leading northward along the British coasts. With detailed observations he found the link between tides and the Moon's position in the sky.

In 1678 J.Newton in his *Principia Mathematica* outlined a detailed mathematical description of sea tides. With his theory of gravitational forces Newton found that the Moon and Sun both generated tides by attracting the water on the Earth's side facing them. He furthermore described the interaction of the solar and lunar tide. His finding that the solar- modulates the lunar tide successfully explained the changing amplitudes of the sea tides which peak at full and new Moon.
He was also the first to predict atmospheric tides but thought they would be too weak to measure.

1.3.2. THE FOUNDATIONS FOR MODERN TIDAL THEORY

In 1799 and 1825 Laplace with his *Mécanique Céleste* published a mathematical study of ocean tides. In this he mostly assumed the ideal case of an ocean of uniform depth on a rotating sphere. Such an ocean would have an infinite number of free oscillation modes, some of which would be in resonance with the periodic forcing. Ironically, the assumption of uniform ocean depth makes his theory weak when applied to real oceans, but the absence of lateral boundaries in the atmosphere makes the theory more applicable to atmospheric tides. Laplace carried out a study in which he applied his theory of ocean tides to the atmosphere. He assumed an atmosphere of uniform scale height, $H$, and showed that the tidal oscillations were the same as those of an ocean of uniform depth $H$, giving $H$ the name *equivalent depth*, a term which was in 1936 defined differently by Taylor. Although Laplace assumed isothermal changes of pressure and density, he also introduced correction terms which accounted for adiabatic temperature changes occurring under rapid pressure- and density variations. Attempts to measure lunar atmospheric tides which his theory predicted were partly successful in that they indicated the existence of a very weak tide at ground level but Laplace, using error theory calculations which he pioneered, concluded that 40000 observations were necessary in order to determine that the weak measured oscillations was a lunar tide with sufficient certainty. Laplace predicted also the thermal semidiurnal tide but made no attempt to develop a theory of thermally excited oscillations. Hough [1897, 1898] published studies which improved Laplace’s theory in that he introduced the spherical harmonic functions commonly referred to as *Hough functions*. Laplace's Tidal Equation (11.28) with the alterations by Hough is the core of modern tidal theory and describes the global response of the atmosphere and oceans to perturbations.

1.3.3. THE RESONANCE THEORY AND FIRST MEASUREMENTS

Almost a century after Laplace, Thomson (who was later known as Lord Kelvin) presented his Resonance Theory in which he built on Laplace's results. In 1882 he addressed the Royal Society in Edinburgh with the following words: "The cause of the semi-diurnal variation of barometric pressure cannot be the gravitational tide-generating influence of the Sun, because if it were there would be a much larger lunar influence of the same kind, while in reality the lunar
barometric tide is insensible, or nearly so. It seems, therefore, certain that the semidiurnal variation of the barometer is due to the temperature." The scientific community agreed that the semidiurnal oscillations they found were of mainly thermal origin, but this brought up a new question which was to occupy researchers for almost another century: why at ground level the dominant pressure oscillations were semidiurnal while the strongest temperature changes were diurnal. Kelvin’s Resonance Theory argues that the atmosphere’s free oscillation modes have periods closer to 12 hours than 24 hours and that the strongest resulting oscillation would thus be semidiurnal.

Critics of the Resonance Theory, like F. Whipple, argued that variations of pressure and temperature associated with weather, seasonal- and annual changes would influence the 12 hour resonant frequency, and thus the period of the measured semidiurnal tide. This was in 1927 proven wrong by J. Bartels. In order to examine the Resonance Theory, much effort was put into determining the free oscillation periods of the atmosphere. Although Laplace had derived equations to calculate these resonance modes and their periods, his formulae were based on a simplistic model of the atmosphere which assumes constant scale height and temperature with altitude. Therefore, finding resonance frequencies was closely linked to the knowledge of the atmosphere’s vertical structure. Since measurements at greater heights could only be carried out since the middle of this century, the Resonance Theory could not be tested properly until then. Rocket experiments which were carried out in the 1950s finally proved the Resonance Theory wrong since no measured vertical temperature profiles could explain the large amplification of semidiurnal pressure oscillations, $S_2(p)$, necessary to match observations. Despite being proven wrong, the merit of Kelvin’s Resonance Theory lies in the fact that confirming or disproving it has lead to great advances in knowledge about the atmosphere and oscillations.

With the failure of the Resonance Theory the problem of explaining why at ground level $S_2(p)$ are larger than $S_1(p)$ (the diurnal pressure variations) still persisted. Some progress on the issue was only made when Siebert [1954] suggested that the absorption of solar radiation by water vapour might cause the large $S_2(p)$ amplitudes. Later, he suggested that ozone in the stratosphere might have the same effect, but by assuming an isothermal and very cold stratosphere his calculations found a very weak influence of ozone absorption on $S_2(p)$ at ground level and he later dismissed this idea [Siebert, 1961]. The significance of Siebert’s approach lies in that he considers thermal absorption in the atmosphere, and not only at ground level, as was assumed previously. Butler and
Small [1963], by assuming a more realistic stratosphere profile, showed that ozone absorption did contribute to the ground level \( S_2(p) \). At this stage, having proven why \( S_2(p) \) are large, it was still necessary to show why the \( S_j(p) \) are small at ground level. Lindzen [1966] re-examined Laplace’s tidal equation and found that solutions existed not only for positive values of equivalent depth, but also for negative values (see also Chapter II, section II.4.4). While the former describe oscillations with propagating energy, the case of negative equivalent depth describes oscillations confined to the region of excitation, the energy being trapped. Lindzen [1967] used this result when evaluating the detailed response of the atmosphere to diurnal thermal excitation by water vapour- and ozone absorption. In this study he found that most of the absorbed energy generates a diurnal trapped mode (with negative equivalent depth) and some goes into a propagating semidiurnal mode, followed by a diurnal propagating mode. With the ground level being too far from the regions of peak absorption in the troposphere and stratosphere, no trapped mode can be found there and thus the strongest component is the semidiurnal. This explanation of the original problem which challenged Kelvin’s Resonance Theory is now universally accepted. A further, though second-order, effect is that of destructive interference. With the regions of ozone absorption ranging over 30-40 km in altitude, which is comparable to the vertical wavelength of some generated propagating modes, destructive interference can take place within the region of excitation. This however does not happen with the tides generated by the water vapour absorption, since that region in the troposphere is much thinner.

I. 3. 4. CLASSICAL TIDAL THEORY AND ITS LIMITATIONS

The analytical theory describing the atmosphere’s response to perturbations has been described in detail by Chapman and Lindzen [1970] as well as Holton [1975] and is outlined in Chapter II. In this theory, commonly referred to as Classical Tidal Theory, perturbations are applied to a simplified version of the momentum- and energy equations of atmospheric gas particles and two differential equations are derived which define the latitudinal structure of tides (Laplace’s tidal equation, II.28) and their vertical structure in the atmosphere (Vertical structure equation, II.27). The simplifying assumptions made, as outlined in II.2, are necessary to ensure that the latitude- and height dependencies of oscillations are separable and analytical solutions to the differential equations are possible, but they also form the principal limitations of the theory. Many authors have discussed the implications of ignoring the vertical temperature changes and dissipative processes such as turbulence, viscosity, conductivity and ion drag, as Classical Tidal Theory does.
Similarly, it was investigated how far these limitations could be reduced while maintaining the analytical treatment, or the separability of latitude- and height dependencies. The separability would hold in a non-isothermal atmosphere and also when considering a special form of height-dependent dissipation [Lindzen and McKenzie, 1967] or a constant Coriolis parameter [Lindzen, 1970]. Authors investigating how realistic the approach of Classical Tidal Theory was were, amongst others Wilkes [1949] and Hines [1960] who with a theoretical approach found that dissipative processes were important when their time scales are smaller than the tides' time scale for horizontal and vertical propagation. More detailed studies were later carried out by Yanowitz [1967], Lindzen [1968] and Lindzen and Blake [1972] who investigated effects of vertical temperature structure and molecular and turbulent dissipation processes on the propagation of tides. Salby [1979, 1980] carried out similar studies for planetary waves. Lindzen and Blake found that dissipation had no significant effects on tides below 100 km altitude. For migrating thermal tides, the zonal phase speed is of the order of 400 m/s and with average zonal winds in the stratosphere and mesosphere being much smaller no considerable influence of background winds on tides should be expected [Chapman and Lindzen, 1970]. Therefore, Classical Tidal Theory may be applied to tides below 100 km. However, the region above 100 km is, amongst other things, characterized by diffusion, larger background winds and ion drag, all of which prohibit the separability of latitude and height in the tidal equations and require a numerical solution. The advances in computer technology brought a tool which has become of considerable importance to atmospheric tidal research.

I. 4. MEASUREMENTS OF ATMOSPHERIC TIDES

One difficulty with studying global oscillations such as tides and planetary waves is that understanding their nature requires information from globally distributed locations at many heights for continuous time intervals extending over the period of the tides, 8, 12 or 24 hours, or planetary waves (several days). Tidal activity was found to be important below 200 km height, and although measurements satisfying some of these criteria are now possible the problem of satisfying all still persists. The principal techniques used for measuring tides will be reviewed below.

I. 4.1. THE EARLIEST DATA

The first known measurements of atmospheric oscillations were carried out by Laplace in the late
19th century when he tried to find tidal oscillations of the surface air pressure. Atmosphere measurements at altitudes other than ground level only became possible in the middle of this century. The earliest developed techniques enabled measurements up to an altitude of around 100 km, with some additional data between 100-120 km. Therefore, not much was known about the ionosphere. Johnson [1955] and Harris [1959] described data for heights between ground level and 30 km, using radiosonde measurements. They successfully measured temperatures and winds being dominated by tidal oscillations at those heights. The same techniques were used by Wexler [1959] to measure for the first time the springtime warming of the Antarctic lower stratosphere during the International Geophysical Year 1957-58. The first rocket experiments were capable of carrying out measurements between 40 and 60 km altitude, as described by Miers [1965] and Reed et al. [1969], who also found strong tidal oscillations. Elford [1959] and Greenhow and Neufeld [1961] described the earliest meteor trail measurements in the upper mesosphere, around 80 to 100 km height. The technique of measuring the Doppler shift of gas emissions caused by meteors which enter the atmosphere is still used today and requires relatively little effort in terms of facilities. Some wind oscillations were measured also above 100 km, but these were based on many empirical assumptions. The basic idea behind these measurements was that wind-induced electric currents at 110-120 km altitude were assumed to be responsible for the daily variations of the geomagnetic field at ground level. By assuming basic structures of parameters such as winds and electrical conductivity the neutral winds could be derived from ground-measurements of the geomagnetic field. Such measurements were described by Maeda [1955] and Kato [1956]. The uncertainty about the measurements were clearly the assumptions about various parameter structures.

### 1.4.2. BASIC TECHNIQUES

In general, one may distinguish between measurements which are ground-based and those which are performed in-flight. Ground based instruments use remote sensing techniques to measure at heights other than ground level, while in-flight instruments either measure their immediate surrounding in-situ or use remote sensing techniques to measure at other locations. The carriers for in-flight instruments are either aeroplanes (for tropospheric heights, in some cases up to mesosphere), rockets (up to thermospheric heights) and balloons (up to stratospheric heights) or satellites (thermospheric and exospheric heights).

In-situ measurements have the advantage of giving more information than remote sensing would
provide, while at the same time their disadvantage is that data are obtained only along the path of the instrument carrier. Balloon- and rocket measurements are suitable for measuring vertical profiles and have been used since the early 1950s. Still today they provide a relatively cheap means of obtaining information quickly. Since flight times are limited, no continuous measurements can be carried out for the same set of locations. Air crafts as carriers provide more flexibility than balloons and rockets in that they allow a larger payload and a controllable flight path. Their disadvantages are the higher costs for operation and limited flying altitudes which allow tropospheric, or in exceptional cases, stratospheric and mesospheric measurements only. They are therefore routinely used only for meteorological work. Satellites also carry out in-situ measurements but cannot fly below around 100 km altitude since their high speed would cause them to burn up below that where the density is larger. Typical instruments for in-situ measurements are mass spectrometers, Langmuir probes, magnetometers and interferometers. These are capable of measuring neutral gas properties such as temperature, winds and composition as well as ionospheric parameters.

Remote sensing offers the considerable benefit of measuring over large distances. This makes it possible not only to penetrate into regions which are difficult to access with aeroplanes, rockets or balloons but also to give a considerable spatial coverage by measuring almost simultaneously at many locations. This is achieved by changing the direction of the instrument’s line of sight while it measures. Therefore, remote sensing techniques are particularly useful for measuring tidal structures in the atmosphere. All measurements used in this thesis were obtained from remote sensing instruments. The carrier of the information about the gas is electromagnetic radiation of different wavelengths which is emitted or scattered by the particles. All instruments used for remote sensing therefore in principle need to detect and analyse electromagnetic radiation. One may distinguish between active and passive instruments, of which the latter measure radiation which is naturally emitted or scattered by constituents while the former first excite the particles with radiation of specific wavelengths and then measure the emissions. A series of powerful techniques have been developed over the past 20 years for remote measurements. The most commonly used instruments include Fabry-Perot interferometers, various types of radars and sounders. Since data used in this thesis derive from some of these instruments their basic properties and measurement techniques are outlined below.
I. 4. 3. MODERN INSTRUMENTS

With the technological advances, the methods for exploring atmospheric properties have advanced considerably during the past 20 years. In what follows, the most common modern instruments are described.

i) FABRY-PEROT INTERFEROMETERS

In the upper atmosphere, photochemical reactions are accompanied by the emission of radiation in the infra-red, visible and ultra-violet parts of the electromagnetic spectrum. The emissions are referred to as airglow and are always present at all latitudes. Outside the auroral (high-latitude-) regions airglow is virtually unstructured, whereas magnetospheric disturbances can cause highly structured airglow profiles in auroral regions. The regions of airglow emission move with the neutral air and thus carry information about the neutral winds. Observing the apparent wavelength of a prominent airglow line makes it possible to determine its Doppler-shift and thereby the line-of-sight neutral air velocity. Fabry-Perot Interferometers (FPI’s) measure these airglow emissions for specific (known) emission frequencies and thus give as primary parameters the neutral gas temperature (through the emissivity or spectral width) and the line-of-sight neutral wind (through the Doppler-shift). From these, the vector velocities can be derived as secondary parameters. Since airglow of one constituent peaks at one specific height, the measured winds and temperature are values for that altitude only. However, different constituents have their airglow emission peaks at different heights. By using a selection of filters one FPI can therefore measure temperatures and winds for different altitudes.

ii) RADARS

Atmospheric gas is a medium of differently sized particles and irregularities. As radiation propagates through the gas it is frequently scattered in all directions. If scattering particles are spaced at half a wavelength along the ray path of the radiation, the scattered and weak signals will add up to a strong and measurable signal. This volume scattering is used to measure atmospheric properties with radars. Essentially, a radar instrument emits well-defined radiation signals into the atmosphere and measures the backscattered signal. This signal is then analyzed in terms of
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Chapter I

intensity, frequency and echo shape. The intensity is controlled by the backscatter mechanism, the
frequency will through its Doppler shift give information about the line-of-sight velocity of the
scattering medium and the signal shape can provide information such as the temperature. The
basic technique of radars can be applied in many different ways which differ mainly in the
properties of the signal sent out. A distinction is therefore made between four different types of
radars, the coherent- and incoherent radars, MST/MLT radars and lidars. Of these, the coherent-
and incoherent scatter radars measure in the ionosphere, whereas the other two measure neutral gas
parameters below 100 km altitude. While lidars measure in the visible spectrum, the other radars
all use electromagnetic radiation in the radio frequency spectrum.

Coherent scatter radars measure the physical structures of the ionized gas which change slowly
enough for successively arriving scattered pulses to be almost identical. This means that the
initially weak scattered pulses add up to a strong signal, thus improving the sensitivity of
measurements. Measurements are carried out with all radars by sending out a signal pulse and
receiving the backscatter signal with the antenna. When most of the backscatter signals have
arrived (with travel times depending on the distance or altitude at which they are scattered) a new
signal is sent out. The scattering particles are electrons and the returned signal provides
information about the altitude of measurement (through the travel time of the signal), line-of-sight
plasma velocity (through its Doppler-shift) and plasma density (through its backscatter intensity).

Incoherent scatter radars measure thermal processes in the ionosphere with short coherence times.
The received signal strength is much weaker since the scattering electrons move thermally and the amplification of the waves cannot take place as for coherent scatter signals. Incoherent scatter radars therefore require high power transmitters, a large antenna equipped with the most sensitive receivers and sophisticated data processing equipment. Their main disadvantages are therefore the large costs associated with such a facility. The first incoherent scatter radar was constructed in the early 1960s, and currently less than a dozen radar facilities operate worldwide. They are mostly national or international facilities in order to cover the high costs. Still, though, incoherent scatter radars are the most powerful ground based technique for measuring ionospheric parameters. Incoherent scatter occurs in all directions and not only perpendicular to the magnetic field lines. The basic differences between an incoherent- and a coherent scatter measurement are therefore the strength of the signal pulse sent out and the direction of viewing (which is flexible for incoherent scatter radars). In principle, therefore a facility can be used for both coherent and incoherent measurements. However, the signal processing techniques used are different as well as the parameters measured and the altitudes investigated. Since incoherent scatter does not depend on specific events in the atmosphere (while the coherent scatter relies on turbulence) measurements can be carried out for the D-, E- and F regions at all latitudes. The parameters primarily derived from the signal analysis are the electron- and ion temperatures, their densities and line-of-sight velocities. By making a series of assumptions, neutral air parameters such as winds and temperatures can be derived as well. Examples for networks of incoherent scatter radars are the globally distributed instruments for the Coupling, Energetics and Dynamics of Atmospheric Regions (CEDAR) program and the high-latitude facilities for the EISCAT consortium.

*MST (Mesosphere Stratosphere Troposphere) and MLT (Mesosphere Lower Thermosphere) radars* work in a similar manner to the incoherent scatter radars but measure in the neutral atmosphere instead of the ionosphere. Radio waves are scattered by neutral particles in the homosphere (below 100 km) where turbulence occurs. The backscatter returns are strongest from altitudes around 10-12 km and 60-75 km, but measurements are carried out also between these regions. A preferred operation frequency lies at 50 MHZ, implying that the backscatter occurs from irregularities with a spatial period of around 3 m. These turbulence features are often associated with gravity wave phenomena and MST/MLT radars are thus often used for studies of gravity waves and also tides. The primary parameter measured is the *neutral line-of-sight velocity* which again is derived from the Doppler shift of the returned echo. Since the upper mesosphere region is difficult to observe by other techniques, MST/MLT radars offer a unique means for
measuring that region. A large number of instruments are operated by the CEDAR community.

Lidar instruments rely on similar principles to incoherent scatter radars, but the radiation used is in the visible spectrum. The signals undergo Raleigh scattering on particles much smaller than the wavelength. The brightness of the backscattered signal gives the line-of-sight neutral velocity and neutral density from which, by using the hydrostatic equation, the temperature and pressure can be derived, provided the composition is known. Alternatively, the temperature can also be derived from the Doppler broadening of the returned signal. While early instruments relied on searchlights, the recent ones use lasers and thus achieve a height range of up to 100 km. Lidars have since the early 1970s been used for tidal and gravity wave studies. The lack of an absolute calibration of atmospheric transmission implies that the density profiles obtained need to be fitted either to theoretical models or experimental data. Lidar instruments are relatively cheap and often operate continuously.

I. 5. NUMERICAL MODELS

I. 5.1. GENERAL PROPERTIES AND APPLICATIONS

Numerical models are computer programs which use known mathematical relations to simulate complex time-dependent situations. The complexity of these situations in most cases does not enable an analytical solution of the underlying mathematical equations, and numerical models offer the only known alternative method for finding a solution. Nowadays, numerical models are also used in many disciplines other than science which require understanding or predictions of complex dynamical situations such as demographic, economic and financial developments as well as systems such as structures of buildings, cars and aeroplanes.

A numerical model only contains the scientific information known to its developers. In practise, the results produced by a model furthermore depend on a number of boundary factors such as accuracy of calculations (time and space resolution), the numerical technique used and the initial-and background values of parameters implemented. Two models, therefore, which in principle contain the same scientific framework, can produce different results, depending on the various, often empirical, factors. In recent years, the considerable increase in computing capacities has allowed significant improvements in the capabilities of numerical models.
Being a complex system with many physical processes interacting linearly and non-linearly, the atmosphere is particularly suitable for numerical modelling. Along with measurements, modelling nowadays plays a central role in atmospheric research since it helps to interpret the often localized measurements in a global context. More importantly, it enables us to experiment with different conditions, thus allowing an insight into the balance of physical processes which measurements alone could not provide. Modelling the Earth’s atmosphere is complex insofar as atmospheric layers (see 1.2) differ considerably in their properties and governing physical processes. Limitations in computer capacities have always made it necessary to simplify certain processes or ignore them entirely if they are of no importance to the region studied. At different altitudes, different processes can be ignored and therefore 2- or 3-dimensional global atmospheric models have been developed for specific height regions only, as opposed to one global model for all altitudes. Suitable lower- and upper boundary regions for an atmospheric model are those heights at which either no processes take place which are of importance to the region studied or at which the relevant vertical flow of momentum, energy and mass is known and can be parameterized.

In general, one may distinguish between two families of atmospheric models of which one category are the general circulation models and the other are perturbation models. While the former calculate the complete set of parameters, the latter calculate perturbations only which are relative to a fixed (given) background atmosphere. Within the family of general circulation models, three types of models may be distinguished with respect to their height coverage. Meteorological models very accurately simulate the troposphere only and are used regularly for weather predictions, while middle atmosphere models have their lower boundary either at the tropopause or at ground level and usually simulate processes up to the mesopause, in some cases up to lower thermospheric heights. A third type are the upper atmosphere models which either simulate the thermosphere only and use parameterized charged particle properties, or the ionosphere only with parameterized neutral gas properties. A small fraction of the upper atmosphere models simulate self-consistently the coupled thermosphere and ionosphere. In contrast, the perturbation models often cover a larger height range since their calculations are less complex.

I. 5. 2. THE HISTORY OF ATMOSPHERIC MODELLING

Reviews of early atmosphere modelling work can be found by Creekmore et al. [1975], Roble et al. [1982] and others. The following review will describe some of the milestones in the history of
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With regard to tidal and upper atmosphere modelling. Most of the earliest atmospheric models calculated steady-state situations and were unable to simulate time-dependent phenomena such as tides or high-latitude energy bursts. The first thermosphere model was that by Kohl and King [1967]. It used pressure gradients as the only driving force for winds in the thermosphere and calculated ion drag from a globally uniform ionosphere parameterization. The pressure gradients were calculated from semi-empirical density models which were largely based on satellite drag observations. Considerable simplifications were necessary for solving the Navier-Stokes equations. The model did not include calculations of solar heating, thermal transport and infrared (IR) cooling but nevertheless successfully reproduced upper thermospheric winds at mid- and low latitudes above around 300 km, showing that they were to large extent thermally driven. Significant advances were made with the introduction of harmonic (or spectral-) models and those which carry out solar EUV calculations, by Creekmore et al. [1975] and Blum and Harris [1975]. Still, though, they depended on empirical descriptions of pressure- and temperature fields for the calculations. Furthermore, all of these early models had to ignore some of the major terms in the energy and momentum equations, such as viscosity, Coriolis forces and nonlinear terms. A further problem was that the poor knowledge of the exact magnitude of driving forces such as solar EUV and heating efficiency, particle precipitation and plasma convection lead to the use of unrealistic values in the models. Still though, some early models, like the two-dimensional models by Richmond and Matsushita [1975], Dickinson et al. [1975] and Richmond [1979a] with their results contributed to the development of the more advanced models a few years later.

A major new advance was made with the development of global 3-dimensional, time-dependent general circulation models, such as those by Fuller-Rowell and Rees [1980] and Dickinson et al. [1981]. These models solve self-consistently the time-dependent coupled equations of energy and momentum in three dimensions. The solar EUV and UV heating is considered, using height-dependent heating functions. Both models also consider ion drag and high-latitude Joule heating. In their initial versions many aspects had to be simplified, such as using a parameterized ionosphere and assuming a single component gas only or parameterized composition. These and other simplifications have in the following decades been replaced with more realistic and self-consistent treatment. In their latest versions [Fuller-Rowell et al., 1996; Roble and Ridley, 1994] both models simulate the coupled thermosphere and ionosphere, and the model by Roble and Ridley (which is generally referred to as TIME-GCM, see also Chapter V, section V.3.1) extends down to an altitude of around 30 km. They today form the two principal time-dependent three-
dimensional upper atmosphere models. The model used in this thesis is an extended version of the model by Fuller-Rowell et al.

As outlined in I.3.4, investigating the propagation of tides into the thermosphere can only be done numerically when dissipation is considered, and the earliest perturbation models for these studies were developed by Volland and Mayr [1973], Forbes and Garrett [1976, 1978], Harris and Mayr [1975] and Mayr and Harris [1977]. These models ignored the effect of mean winds as well as latitude-dependent temperature and composition. Some of these simplifications were overcome in the modelling work by Forbes [1982a, b] and Forbes and Hagan [1982] who extended the model by Forbes and Garrett by using more realistic heating rates of water vapour and ozone and considering eddy diffusion, thermal conductivity and viscosity as well as a more realistic background atmosphere. This model solves equations of motion and energy for perturbations from the ground level to around 400 km altitude. It uses parameterized background values for temperature, zonal wind (ignoring meridional wind), composition, tidal forcing, Newtonian cooling and eddy- and molecular diffusion. The treatment is therefore not fully self-consistent in that the effects of tides on mean winds are ignored. Studies by Teitelbaum and Vial [1991], Holton [1982], Garcia and Solomon [1985] and others showed that the dissipating tides considerably influence mean winds in the thermosphere. The model by Forbes [1982a] is therefore suitable for investigating the perturbations only and not their influence on the background atmosphere. The model by Hagan et al. [1993] is an extension of the model by Forbes [1982a] and uses, amongst other, more realistic background atmosphere values as well as extending the original model to simulate planetary waves. These models have been used for many analyses of tides and planetary waves and as input for other models, such as the NCAR TIGCM [Roble et al., 1988] during various campaign studies, such as the Lower Thermosphere Coupling Study (LTCS) [Fesen et al., 1993a]. Some of these models are described in more detail in Chapter V, section V.3.1.

A different approach is used by the harmonic (or spectral-) tidal models, like those by Volland and Mayr [1973], Straus et al. [1975], Harris and Mayr [1975] and Mayr and Harris [1977]. Essentially, physical quantities in these models are expanded into spherical harmonics. The demand on computing time is less, and therefore even the earliest of these models like that of Volland and Mayr simulate the lower atmosphere and thermosphere in three dimensions, including upwards propagating tides and some of the dissipative mechanisms. The model by Straus et al. [1975] is an extension of the model by Creekmore et al. [1975] and included non-linear processes such as
viscosity and ion drag as well as solar EUV heating. The Harris and Mayr [1975] model and its extension by Mayr and Harris [1977] considered a binary gas in three dimensions with self-consistent diffusive interaction between the components.

The first upper atmosphere model to simulate tidal perturbations in the thermosphere and their self-consistent interaction with the background atmosphere was that by Fesen et al. [1986], an extension of the model by Dickinson et al., [1981]. It does not simulate the formation of tides in the stratosphere and mesosphere but uses global tidal perturbations at its lower boundary (97 km) as the external tidal source. The model has since been extended [Roble and Ridley, 1994] and now includes self-consistent modelling of the tides' generation as well. The modelling carried out in this thesis uses an extended version of the Coupled Thermosphere-Ionosphere model (CTIM) by Fuller-Rowell et al. [1996] which itself is based on the thermosphere model by Fuller-Rowell and Rees [1980] and ionosphere model by Quegan et al. [1982].

1.5.3. THE DIFFICULTIES WITH NUMERICAL MODELS

Despite the substantial benefits of numerical models it is also of interest here to describe some of the difficulties which a modeller typically encounters. The value of any numerical model is determined by its ability to reproduce as accurately as possible measured parameters under as many different conditions as possible. During the development stage of a model, comparisons with existing measurements are carried out. Once a model has been tested and verified it can be used to predict conditions which have not yet been measured. As long as the model successfully reproduces measurements within tolerable error boundaries it is assumed that the physical processes have been correctly described mathematically. Strictly, however, a deep understanding of the physics is not always provided by models. In reality, model output can only be compared to a limited number of observations and will only match a selection of these, while other measurements can differ substantially from the simulated predictions. The modeller does not expect perfect agreement since he is often aware of some of the limitations of his model. The most common sources of error include ambiguous and over-simplified parameterizations of certain variables, incomplete descriptions of processes such as chemistry, controversial parameters such as chemical reaction rates and problematic numerical schemes. If a model delivers inaccurate results one key problem is to identify which of the above listed categories is most likely to be responsible. This can often take long to identify, and essentially the question is whether the
problem is of "scientific" or "technical" nature. The first three of the above listed possible error sources would fall into the category of scientific errors, while the last would be technical.

Technical difficulties in models often cause results which contradict expected behaviour. One particular category are problems with numerical stability. Finite difference techniques which are used when differentiating and integrating numerically (see also III.3.9) may become unreliable under various circumstances. When parameters change very little compared with their magnitude, gradients become very small and the digit-precision insufficient. Small gradients can easily fluctuate by factors of 10 or more which in some calculations may produce similar or larger fluctuations of other important parameters. These random noise errors either become evident immediately in the data output or they build up slowly and suddenly appear in the form of often localized "bursts" of a parameter value. Occasionally, these cause a simulation to stop running. Numerical smoothing is necessary to remove these errors. When discovered, the modeller will have to trace back the source of an error since the most likely case is that the variable which became unstable in the output was influenced by another variable which might have influenced many other model parameters even though they may not appear unrealistic at the time. Tracing back the sources of numerical errors and correcting them is often a very time consuming process.

Identifying and reducing scientific errors can similarly be difficult if the exact source of the faults is not known. Comprehensive atmosphere models, such as that used in this thesis, contain a large number of coupled equations which implies that the results of one calculation depend on the output of another. If the variable passed between two sets of calculations is a measurable physical quantity its values can be verified and the error source more easily identified. This, however, is not always the case and the modeller may in some cases prefer to parameterize certain quantities in order to decouple as far as possible the calculations, thus making the code more easily traceable. Still, a certain amount of ambiguity is often inevitable and models sometimes contain empirical factors which are introduced to match the observations better. The scientific reason behind such a factor is not always known.

From these general remarks it is evident that there is not only ambiguity in the sources of errors produced by numerical models but that a correct model output is no guarantee for the accuracy of calculations. If a systematic error is present it will usually be identified when compared to several measurements, preferably from different geographic locations, but it cannot fully be ruled out that
several errors compensate each other under many simulation conditions and might not be noticed by the modeller for a long time. It is therefore important to thoroughly test new models or improvements to existing code under various conditions. This has become easier nowadays with faster and cheaper computers, while previously modellers often had to work with a minimum number of runs not only because of long run times but also because high costs for computers constrained them to systems shared with other users. Chapter V is dedicated to validation of the changes made to the CTIM code, as described in Chapter III, section III.5
CHAPTER II. CLASSICAL TIDAL THEORY

ABSTRACT

This chapter outlines the mathematical framework of Classical Tidal Theory which analytically describes the response of the atmosphere to perturbations, ignoring amongst other factors any momentum- and energy damping of the waves as well as non-linearities. The core of this theory are two coupled differential equations referred to as Laplace’s Tidal Equation and the Vertical Structure Equation, both of which have as solutions an infinite series of eigenfunctions. The former of these equations describes the latitudinal structure of oscillations, while the latter gives their vertical propagation profiles. Tidal theory states that any global oscillation profile found in an (idealized) atmosphere can be decomposed into an infinite series of eigenfunctions both in vertical- and horizontal directions. The horizontal eigenfunctions are referred to as Hough modes and are used to specify externally the tidal forcing profile in the Coupled Thermosphere-Ionosphere Model (CTIM), as outlined in detail in Chapter III. Since studies of this thesis are carried out for the upper atmosphere, where most of the initial assumptions of tidal theory are no longer valid, the vertical structure is calculated numerically by CTIM, and analytical solutions found in this chapter are used only at CTIM’s lower boundary for the horizontal plane.

II. 1. INTRODUCTION TO TIDAL THEORY

Classical Tidal Theory analytically describes the response of the atmosphere to global perturbations of pressure, density, winds and temperature. Linear perturbations are applied to an idealized version of the equations of momentum and energy, leading to a pair of differential equations referred to as Laplace’s Tidal Equation and the Vertical Structure Equation. They are solved analytically by an infinite set of eigenfunctions, and any global oscillation profile found in the idealized atmosphere can be decomposed horizontally and vertically into these eigenfunctions. Classical Tidal Theory does not attempt to describe the exact formation process of tidal oscillations with the relevant chemistry. The role of numerical models lies in solving numerically the perturbation equations under circumstances where the idealizing assumptions are no longer valid. Although in the present context the analytical results from the theory are later applied only in the horizontal plane when specifying the Coupled Thermosphere-Ionosphere Model’s (CTIM’s) lower
Calculations described below show that one may distinguish between various categories of atmospheric tides. On the broadest level, there are two types of waves, the *forced* and *free* oscillations. While *forced* tides result from energy- and momentum input through various mechanisms, *free* tides are the resonant normal mode oscillations of the atmosphere. Forced oscillations may be regarded as free when they are at large distance (in terms of their vertical wavelength) from the region of forcing.

![Figure II.1: Vertical structure of tidal heating, from [Forbes, 1995]](image)

Figure II.1, taken from [Forbes, 1995], shows the three main height regimes in which atmospheric tides are thermally generated through absorption of solar radiation by various constituents. Water molecules absorb in the near-infrared band throughout the troposphere, while ozone absorbs in the ultraviolet (UV) band in the stratosphere-mesosphere regime between around 40 and 60 km altitude. In the upper atmosphere tides are generated through absorption in the extreme-UV by molecular oxygen and nitrogen. The main height regime for this is on average between around 120 and 180 km, but in the auroral regions ion motion under the influence of the high-latitude electric convection field leads to generation of perturbations at larger heights as well. At 80 km altitude, the lower boundary of the CTIM, oscillations propagating upwards can to a first approximation be regarded as free.

Forced oscillations are further subdivided into two categories, vertically propagating (internal) and
vertically trapped (external or evanescent) waves. While the former propagate upwards over larger
distances from their source (depending on their vertical wavelength and damping forces), the latter
are confined vertically to near the region of excitation.

II.2. LINEARIZED EQUATIONS OF MOMENTUM AND ENERGY

In order to be able to describe analytically the atmosphere's response to perturbations, a series of
simplifying assumptions need to be made which are listed in the following.

a) The atmosphere is a perfect gas with constant composition, so that

\[ p = \rho R_{\text{air}} T \]  \hspace{1cm} (II.1)

where \( p \) is the pressure, \( \rho \) the mass density and \( R_{\text{air}} \) the gas constant for dry air
\( (R_{\text{air}}=287.05 \text{Jkg}^{-1}\text{K}^{-1}) \). Note that \( R_{\text{air}} \) is not the universal gas constant, \( R \), (with a value of
\( R = 8314.51 \text{Jkg}^{-1}\text{K}^{-1}\text{mol}^{-1} \)) but \( R_{\text{air}}=R/M \), with \( M=28.9 \) being the average molecular weight
(in atomic units) of air below around 100 km. The use of \( R_{\text{air}} \) instead of \( R \) is merely a
technical matter which is often found in the literature and therefore adopted here for
consistency.

b) The atmosphere is in hydrostatic equilibrium. Vertical accelerations are small compared
to gravity, as expressed by the hydrostatic equation (with \( g \) denoting the gravity
acceleration at the height in question):

\[ \frac{1}{\rho} \frac{\partial p}{\partial z} = -g \]  \hspace{1cm} (II.2)

The hydrostatic equation follows from the equation of momentum if the wind terms are
set to zero.

c) Molecular and turbulent viscosity, conductivity, ion drag, radiative transfer, the Earth's
ellipticity and surface topography are ignored.

d) The background atmosphere is assumed horizontally stratified; zonal background winds
A key parameter used in tidal theory is the Geopotential, \( \Phi \), which is defined as
\[
d \Phi = g \, dz
\]  
and is a measure for the energy needed to move a particle or air cell vertically by \( dz \) through the Earth's gravitational field. Equations (II.2) and (II.3) give a relationship between the partial differentials of pressure and geopotential:
\[
\frac{\partial \Phi}{\partial p} = -\frac{1}{\rho} \frac{\partial p}{\partial z}
\]  
One may furthermore introduce \( f \) to represent the Coriolis term:
\[
f = 2 \Omega_e \sin \theta
\]  
where \( \theta \) denotes the latitude and \( \Omega_e \) is the Earth's angular velocity. The scale height, \( H \), is defined as
\[
H = \frac{R \, T}{g}
\]  
When moving upwards by the amount of one scale height the atmosphere pressure is reduced by a factor of \( e^{-2.71828} \). For average conditions in the atmosphere between 80-100 km height, with \( T=210^\circ K \), the scale height has a value of around \( H=6.2 \) km. In the upper atmosphere values can reach more than 20 km.

For all tidal analyses it is most convenient to introduce the log-pressure coordinate system which defines the vertical coordinate, \( z^* \), as:
\[
z^* := -H \ln \left( \frac{p}{p_0} \right)
\]  
where the scale height, \( H \), is assumed constant, \( p \) denotes the pressure at \( z^* \) and \( p_0 \) the pressure at
a ground level $z_0^*$. The vertical velocity in the log pressure coordinate system is defined as:

$$ w^* = \frac{dz^*}{dt} \quad (\text{II.8}) $$

One can easily derive from the definition of $z^*$ (II.7), the perfect gas law (II.1), the hydrostatic equation (II.2) and the definition of scale height (II.6) that $\partial z^* = \partial z$ when neglecting the non-linear terms (second-order derivatives) in the total differentials $dz^*$ and $dz$. All horizontal components in the log pressure coordinate system are given as in the spherical coordinate system. The only difference between the two coordinate systems is therefore their vertical coordinate.

The momentum equations for atmospheric gas particles are under the above assumptions given by the expressions

$$ \frac{\partial u}{\partial t} - f v = - \frac{1}{r_E} \frac{\partial \Phi}{\partial \theta} \quad (\text{II.9}) $$

$$ \frac{\partial v}{\partial t} + f u = \frac{1}{r_E \cos \theta} \frac{\partial \Phi}{\partial \varphi} \quad (\text{II.10}) $$

for southward ($u$) and eastward ($v$) velocities in spherical coordinates [Rishbeth & Garriott, 1969]. Here, $\varphi$ denotes the longitude and $r_E$ the Earth's radius. The second term on the left side of equation (II.9) represents acceleration due to Coriolis force and the term on the right side is acceleration due the horizontal zonal pressure gradient. Terms in equation (II.10) are the corresponding for eastward wind. It is necessary to consider Coriolis force since the oscillations investigated here are of global scale. One sees from (II.9) and (II.10) that velocities are expressed in terms of geopotential, $\Phi$, instead of pressure.

The energy for the same gas is described by the relation

$$ c_p \frac{dT}{dt} = \frac{1}{\rho} \frac{dp}{dt} + J \quad (\text{II.11}) $$

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Equation (II.11) is essentially the first thermodynamic law, and thus expresses the conservation of energy for a gas cell. The term on the left gives the total energy change caused by adiabatic temperature changes, the first term on the right side is the heat produced through changes of pressure in the cell and the second, $J$, represents external heating per unit mass. If the gas cell moves with velocity $V$, the total differential on the left side term of (II.11) becomes:

$$c_p \frac{dT}{dt} = c_p \left( \frac{\partial T}{\partial t} + V \cdot \mathbf{q} T \right) \quad (II.12)$$

Ignoring horizontal motion (see approximation (d) from above) however allows to simplify the left side of (II.12) by considering only the vertical motion:

$$c_p \frac{dT}{dt} = c_p \left( \frac{\partial T}{\partial t} + w \cdot \frac{\partial T}{\partial z^*} \right) \quad (II.13)$$

Similarly, the first term on the right side of (II.11) becomes, after neglecting horizontal velocity terms:

$$\frac{1}{\rho} \frac{dp}{dt} = \frac{1}{\rho} \left( \frac{\partial p}{\partial t} + w \cdot \frac{\partial p}{\partial z^*} \right) \quad (II.14)$$

For the gas cell in question one assumes pressure changes only to occur as a result of its vertical motion, giving $\partial p/\partial t = 0$ in (II.14), and thus one may after simple rearranging write for the energy equation (II.11):

$$c_p \frac{\partial T}{\partial t} + w \left( c_p \frac{\partial T}{\partial z^*} + g \right) = J \quad (II.15)$$

where the hydrostatic equation (II.2) was used to replace $\partial p/\partial z^*$.

Finally, this form of the energy equation is converted into one using the geopotential, $\Phi$, instead of temperature, $T$, since this is the most common in the literature. From the definitions of geopotential (II.3) and scale height (II.6) it is known that
As outlined above, the gas cell is heated adiabatically. This implies that temperature changes may take place, but the heating is assumed to occur over a height extent which is large compared with the scale height, $H$. Therefore, one may assume the scale height to be roughly constant and obtains:

$$\frac{\partial \Phi}{\partial t} = \frac{R_{\text{air}}}{H} \frac{\partial T}{\partial t}$$

(II.17)

and similarly,

$$\frac{\partial \Phi}{\partial z} = \frac{R_{\text{air}}}{H} \frac{\partial T}{\partial z}$$

(II.18)

Inserting (II.17) and (II.18) into (II.15) gives

$$\frac{\partial}{\partial t} \left( \frac{\partial \Phi}{\partial z} \right) + w^* \left( \frac{\partial^2 \Phi}{\partial z^2} + \frac{\kappa g}{H} \right) = \frac{\kappa}{H} J$$

(II.19)

where $\kappa = R_{\text{ad}}/c_p \approx 2/7$. Here, in accordance with assumption (e) from above, the second derivative term on the left side is ignored, thus giving

$$\frac{\partial}{\partial t} \left( \frac{\partial \Phi}{\partial z} \right) + w^* \frac{\kappa g}{H} = \frac{\kappa}{H} J$$

(II.20)

Finally, the *buoyancy frequency*, $N$, for an isothermal atmosphere is introduced which is written as $N^2 = \kappa g/H$, giving an alternative form of the energy equation:

$$\frac{\partial}{\partial t} \left( \frac{\partial \Phi}{\partial z} \right) + N^2 w^* = \frac{\kappa}{H} J$$

(II.21)

Equation (II.21) is the energy equation for an isothermal atmosphere, whereas (II.15) does not make the isothermal assumption.
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Some factors in equation (II.21) may vary in the literature, where occasionally the equation of state is used in a different form with \( R \), the universal gas constant, instead of \( R_{\text{air}} \) for dry air. It is therefore important to pay attention to the exact definition of all parameters involved.

II.3. LAPLACE’S TIDAL EQUATION AND THE VERTICAL STRUCTURE EQUATION

Linear perturbations are now applied to the momentum- and energy equations (II.9), (II.10) and (II.21). A perturbation of pressure, velocity and temperature may under the linear assumption be expressed as

\[
a' = a_0 + a
\]

where the variable \( a' \) represents any of the perturbation parameters wind, pressure and temperature and is given by the sum of an underlying constant value \( a_0 \) and the perturbation component \( a \). The quadratic and all higher order terms are neglected. This assumption is valid if \( a \) is small compared with \( a_0 \). When assuming the perturbation \( a \) to be a function of time, latitude, longitude and altitude it may be expressed as

\[
a = a_0 \cdot e^{\omega t} \cdot e^{i(\theta_0 + \varphi)}
\]

where \( \omega \) denotes the perturbation frequency, ie. \( 2\pi/\omega \) is in the case of tides one solar (24 hours) or lunar (24.8 hours) day, or a fraction thereof, and \( s \) is the longitudinal wavenumber

\[
s = 0, 1, 2, ...
\]

A value of \( s = 1 \) then describes diurnal- and a value of \( s = 2 \) semidiurnal tides. For \( \omega > 0 \) the wave travels westward and for \( \omega < 0 \) eastward. Only westward travelling solar tides are considered in the modelling here. Lunar tides are not considered further.

In (II.22), the exponential term defines the time-dependence and phase. The first term on the right side is a time- and longitude-independent amplitude. This amplitude expresses the latitudinal- and height structure of the tide and thus plays a key role. Perturbations of pressure, winds and temperature are all given in the general form of (II.22) and tidal theory, as outlined below, describes that amplitude terms of these parameters not only have a distinct global structure but
furthermore are related to each other. Once the global pattern of one parameter has been specified, other parameters are derived from this analytically. Herein lies one of the motivations for applying relations from tidal theory in the CTIM model. At its lower boundary, only the geopotential height perturbations are specified by the user and the simultaneous oscillations of winds and temperature are calculated analytically. The altitude-dependence of $a(\theta, z)$ is of no importance in the modelling carried out in this thesis since the model calculates vertical progression numerically, considering also dissipation and other more realistic conditions for the thermosphere and ionosphere which are ignored in Classical Tidal Theory.

From (II.22) the partial time- and longitude derivatives for the perturbation $a$ can be expressed as:

\[
\frac{\partial a}{\partial t} = i \omega a
\]
\[
\frac{\partial a}{\partial \theta} = i s a
\]

which, applied to the geopotential and velocities in (II.9) and (II.10), replace the time- and longitude derivatives, giving:

\[
u = \frac{1}{r_E \omega \left(1 - \frac{f^2}{\omega^2}\right)} \left[ \frac{s f}{\omega \cos \theta} - \frac{\partial}{\partial \theta} \right] \Phi
\]  
\[
u = \frac{1}{r_E \omega \left(1 - \frac{f^2}{\omega^2}\right)} \left[ \frac{s f}{\omega \cos \theta} - \frac{\partial}{\partial \theta} \right] \Phi
\]

where variables in (II.9) and (II.10) were furthermore separated. Conventionally, the geopotential, $\Phi$, is chosen as the primary forcing parameter for tides, implying that other perturbations are given as a function of geopotential. However, any other parameter could in principle also be chosen. In accordance with (II.22), $\Phi$ is given in the form:

\[
\Phi = \Phi(\theta, z) \ e^{i(\omega t + s \phi)}
\]

The following calculations will focus on the amplitude term in (II.25). One may perform a
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separation of the latitude- and height variables:

\[ \Phi_s(\theta, z) = \sum_n \Phi_{s,n}^0 \Theta_{s,n}(\theta) G_{s,n}(z) \]  

(II.26)

where \( \Phi_s(\theta, z) \) is \( \Phi(\theta, z) \) for fixed \( s \) (with \( \Phi = \sum \Phi_s \)), \( \Theta_{s,n} \) specify the latitudinal- and \( G_{s,n} \), the vertical structure; \( n \) is the latitudinal wavenumber. Various versions of this equation can be found in the literature, giving \( \Theta_{s,n} \) and \( G_{s,n} \) different physical units. For simplicity it was decided here to introduce the factor \( \Phi_{s,n}^0 \), being the geopotential amplitude of each individual mode \( (s,n) \). Therefore, \( \Phi_{s,n}^0 \) is in units of geopotential \( [m^2/s^2] \) and \( \Theta_{s,n} \) as well as \( G_{s,n} \) are unitless quantities.

Inserting these definitions into the equations of momentum (II.9) and (II.10) as well as energy (II.21) produces after more complex calculations which are not reproduced here the following differential equations for \( G_{s,n} \) and \( \Theta_{s,n} \) [Forbes, 1995]:

\[ i \omega H \left[ \frac{1}{\rho} \frac{d}{dz} \rho \frac{d}{dz} G_n(z) \right] + \frac{i \omega \kappa}{h_n} G_n(z) = - \frac{1}{\rho} \frac{d}{dz} \left( \rho \kappa J_n \right) \]  

(II.27)

\[ \frac{d}{d\mu} \left[ \frac{(1 - \mu^2)}{(\eta^2 - \mu^2)} \frac{d\Theta_n(\theta)}{d\mu} \right] - \frac{1}{\eta^2 - \mu^2} \left[ \frac{s}{\eta} \left( \frac{\eta^2 + \mu^2}{(\eta^2 - \mu^2)} \right) + \frac{s^2}{1 - \mu^2} \right] \Theta_n(\theta) = - \epsilon_n \Theta_n(\theta) \]  

(II.28)

with \( \mu = \sin \theta, \eta = \omega/2 \Omega_E \) and \( \epsilon_n = (2 \Omega_E r_E)^2/gh_n \). Equation (II.27) is referred to as the vertical structure equation, while (II.28) was first derived by P. Laplace and is therefore called Laplace's Tidal Equation. Both equations define an eigenfunction-eigenvalue problem and \( h_n \) is the set of eigenvalues usually referred to as "equivalent depth". The term equivalent depth was in this context introduced by Taylor [1936]. In Laplace's original work he used these equations to describe ocean tides and \( h \) was the ocean depth (see also I.3.2). For simplicity, equations (II.27) and (II.28) have been written above for fixed \( s \) and indexed accordingly, replacing the "\( s,n \)" indices from (II.26) with \( n \).
II.4 VERTICAL STRUCTURE

Although an analytical description of the vertical structure will not be used in the modelling with CTIM, it forms an essential part of Classical Tidal Theory and is used to classify different types of tides. Furthermore, expressions using the equivalent depth, $h_m$, are needed in III.5.3 when deriving an analytical expression for temperature perturbations.

II.4.1 CANONICAL FORM OF THE VERTICAL STRUCTURE EQUATION

Equation (II.27) is often written not with the vertical variable $z^*$ but with a dimensionless height variable, $x$, which is defined as

$$x := - \ln \left( \frac{p}{p_0} \right) = \frac{z^*}{H} \tag{II.29}$$

where $p$ denotes the pressure at $x$ and $p_0$ the pressure at a ground level $x_0$. When defining furthermore a variable

$$G_n' (\theta) = G_n (\theta) \sqrt{\frac{\rho H}{\kappa g}} \tag{II.30}$$

equation (II.27) can be simplified significantly, forming the canonical form of the vertical structure equation for an isothermal atmosphere [Forbes, 1995]:

$$\frac{d^2 G_n' (\theta)}{dx^2} + \alpha^2 G_n' (\theta) = - \frac{i}{\omega} \sqrt{\frac{H}{\rho \kappa g}} \frac{d}{dx} (\rho J_{\omega}) \tag{II.31}$$

with

$$\alpha^2 = \frac{\kappa H}{h_n} - \frac{1}{4} \tag{II.32}$$

The general solution for $G_n' (\theta)$ is of the form
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\[ G_n(x) = A e^{iux} + B e^{-iux} \]  \hspace{1cm} (II.33)

Here, the term \( A e^{iux} \) describes the upward- and \( B e^{-iux} \) the downward propagating component of the wave energy, and thus the above expression refers to the wave’s vertical group propagation. The values for the eigenvalues \( h_n \), and thereby \( u \), depend on the boundary conditions which are discussed in the following.

II.4.2 BOUNDARY CONDITIONS

A necessary upper boundary condition is that no source of energy exists at \( x=\infty \), implying that there is no downward propagating tidal component. This is often referred to as the radiation condition. It also implies that internal reflections of waves are ignored at the upper boundary. Therefore \( B=0 \), giving:

\[ G_n(x) = A e^{iux} \]  \hspace{1cm} (II.34)

It is convenient to define the lower boundary below the region of tidal formation. Note that the lower boundary referred to here is not that of the CTIM model. At the lower boundary a constant pressure is assumed:

\[ \frac{dT}{dz} = 0 \]  \hspace{1cm} (II.35)

Expanding the total differential in (II.35) and again neglecting the horizontal wind component gives:

\[ \frac{\partial \Phi}{\partial t} + w \cdot \frac{RT}{H} = 0 \]  \hspace{1cm} (II.36)

Finally, (II.36) is inserted into the energy equation (II.21) in order to eliminate \( w^* \):

\[ \frac{\partial}{\partial t} \left( \frac{\partial \Phi}{\partial z} - \frac{N^2 H}{RT} \Phi \right) = \frac{\kappa}{H} J \]  \hspace{1cm} (II.37)

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This expression is thus a direct consequence of the lower boundary condition.

II.4.3 FREE SOLUTION

If in (II.31) the external energy input is assumed to vanish, \( J_n = 0 \) and the equation is an unforced differential equation. The solution is in this case referred to as free. Although atmospheric tides are forced in the terrestrial troposphere, stratosphere and at higher altitudes, they may be regarded as free at a sufficient height above the region of forcing, throughout some of the mesosphere and above (see also Figure II.1). The free solution is therefore an important case. It describes the resonant normal mode oscillations of the atmosphere.

By assuming \( J_n = 0 \) and using the definitions (II.26) and (II.30), (II.37) can be rewritten in terms of \( G_n' \):

\[
\frac{\partial G_n'}{\partial z^*} - \left( \frac{N^2}{g} + \frac{N}{\sqrt{\rho}} \frac{d}{dz^*} \left[ \sqrt{\rho} \frac{N}{N} \right] \right) G_n' = 0
\]  

(II.38)

Assuming again a constant scale height, \( H \), the expression may be further simplified. Applying also the hydrostatic equation (II.2) along with the definition of the buoyancy frequency, \( N^2 = \kappa g / H \), gives:

\[
\frac{\partial G_n'}{\partial z^*} - \frac{1}{H} \left( \kappa - \frac{1}{2} \right) G_n' = 0
\]  

(II.39)

which, in terms of \( x \) rather than \( z^* \) becomes with (II.29):

\[
\frac{\partial G_n'}{\partial x} - \left( \kappa - \frac{1}{2} \right) G_n' = 0
\]  

(II.40)

The lower boundary condition, therefore, along with the assumption of no external heating, \( J_m \), lead to a first degree linear differential equation for \( G_n' \). The solution for this differential equation is proportional to an exponential function,

\[
G_n' \propto e^{\left( \kappa - \frac{1}{2} \right) x}
\]  

(II.41)
Comparing this with (II.34), the general solution for the vertical structure equation, one finds that the lower boundary conditions require

\[ i \alpha = \left( \kappa - \frac{1}{2} \right) \quad (\text{II.42}) \]

or:

\[ -\alpha^2 = \left( \kappa - \frac{1}{2} \right)^2 \quad (\text{II.43}) \]

The importance of this result is that only one eigenvalue, \( h_n \), is allowed in the solution of (II.31) when assuming \( J_n=0 \) and \( d\Phi/dn=0 \) at the lower boundary. Inserting the definition for \( \alpha \) (II.32) into (II.43) gives for the equivalent depth, \( h_n \):

\[ h_n = \frac{H}{1 - \kappa} \quad (\text{II.44}) \]

For average conditions between 80 and 100 km height, \( H=6.2 \) km and \( h_n=8.7 \) km for all \( n \).

Finally, the energy density is calculated. The waves described here are assumed to be non-dissipative and unforced, implying that energy remains constant. However, the energy density, \( \rho V^2 \), changes with height, as derived in the following. The mass density, \( \rho \), falls vertically with

\[ \rho = \rho_0 e^{-\frac{z}{H}} = \rho_0 e^{-\kappa x} \quad (\text{II.45}) \]

where \( \rho_0 \) is the lower boundary density. From equations (II.23), (II.24), (II.26), (II.30) and (II.41) it follows that in the free solution the horizontal velocity increases exponentially with height:

\[ V \propto e^{\kappa x} \quad (\text{II.46}) \]

The energy density, \( \rho V^2 \), thus behaves vertically as

\[ \rho V^2 \propto e^{(2\kappa - 1)x} \quad (\text{II.47}) \]

Since \((2\kappa-1)<0\) one sees that energy density decreases with height even though horizontal velocities...
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(II.46) increase. Free waves would in theory propagate upwards infinitely. In reality, molecular viscosity and thermal conductivity are important damping forces in the upper mesosphere and lower thermosphere with ion drag becoming important above 100 km height. Therefore, tides in the “real” atmosphere will lose energy to the surrounding atmosphere as they propagate upwards, which is shown in more detail in Chapters IV and V.

II.4.4 FORCED SOLUTION

When in (II.31) \(J_n > 0\) the solution is referred to as forced. Here, \(J_n\) describes for tides the thermal forcing occurring in the atmosphere through absorption of solar radiation. For exact solution of the forced differential equation (II.31), \(J_n\) need to be specified analytically. Since realistic vertical profiles of the heating function, \(J_n\), depend on the chemistry, absorption and transport processes in the atmosphere they are given in parameterized form only and thus (II.31) can only be solved numerically. Although the general form of the solution for vertical structure is that of (II.33), the values for \(\alpha\) now depend on \(J_n\). Therefore, the eigenvalues \(h_n\) are in this case also determined by \(J_n\) and not limited to only one value, as in the free solution (II.44).

However, even without specifying the exact form of \(J_n\) one may distinguish two possible cases for which solutions are different:

i) \(\alpha^2 < 0\)

Waves with \(\alpha^2 < 0\) are called evanescent or external. The condition is satisfied for \(h_n > 4\kappa H\) or \(h_n < 0\). One may write

\[\alpha = i \cdot |\alpha|\]

Inserting this into (II.33) gives two exponential terms, of which \(e^{-|\alpha|x}\) is bound for \(x \to -\infty\), while \(e^{i|\alpha|x}\) is unbound. Since conservation of energy requires the solution to be bound, the only possible solution is

\[G_n(x) \propto e^{-|\alpha|x}\]

The exponential decay of amplitude with height shows that waves with \(\alpha^2 < 0\) are confined vertically to near the region of forcing.
In the above discussions as well as those of sections II.2 and II.3 it was assumed that the scale height, $H$, is constant over the height range of interest. This approximation becomes critical near the mesopause, where strong changes of the vertical temperature gradient, and thus the scale height, are found. When considering a vertical scale height gradient, $dH/dx$, the derivations of Laplace's Tidal Equation (II.28), the Vertical Structure Equation (II.27) and its canonical form (II.31) proceed in identical manner, but expression (II.32) for $\alpha^2$ changes to

$$\alpha^2 = \frac{\kappa H + \frac{dH}{dx}}{h_n} - \frac{1}{4}$$

which implies that $\alpha^2 < 0$ for $h_n < 0$ or $h_n > 4\kappa H + 4dH/dx$. One sees that the condition of evanescence is satisfied more easily when $dH/dx$ is negative. The equivalent depth, $h_n$, of the semiurnal $(2,2)$ mode has a value of around 7 km, which is larger than the $h_n$ values of the diurnal $(1,1)$- as well as higher order semiurnal modes (see Table II.1). Therefore, the $(2,2)$ mode more easily becomes evanescent in the atmosphere than these other modes. Under circumstances often encountered near the mesopause, with a scale height of around 6 km and $dH/dx<0$, the above expression gives $\alpha^2<0$ and the $(2,2)$ mode is evanescent. It has been proposed that the $(2,2)$ mode "tunnels" upwards through this region of non-propagation near the mesopause and is strongly dampened until it reaches altitudes where the mode is propagating again [P. Williams, private comm., 1997].

ii) $\alpha^2>0$

For real $\alpha$ values the waves are called internal. The condition is satisfied for $0<h_n<4\kappa H$. A general solution is again given by (II.33) and now both terms are bound. If however, as previously, only upwards propagating tides are allowed, the only possible solutions is, in accordance with (II.34):

$$G_n'(x) = e^{i\alpha x}$$

Internal waves thus propagate vertically away from the region of forcing, unlike the evanescent waves which were found to be vertically bound. The sign of $\alpha$ depends on the horizontal direction of the tides. In II.6. it will be shown that $\alpha<0$ for westward- and $\alpha>0$ for eastward propagating tides.
The case $\alpha^2 > 0$ is thus the only forced solution in which the waves propagate vertically over larger distances. One may therefore here define a vertical wavelength as

$$\lambda_z = H \cdot \lambda_x = H \cdot \frac{2\pi}{\alpha}$$

(II.48)

which is a function of the equivalent depth, $h_n$, through $\alpha$. In (II.48), two vertical wavelengths, $\lambda_z$ and $\lambda_x$, appear. While the latter is the vertical wavelength of $G_n(x)$, expressed in terms of the dimensionless variable $x$, the former defines the same vertical wavelength in terms of the height coordinate, $z^*$ (defined in (II.29)). The concept of a vertical wavelength is plausible only for internal waves since the free solution (II.41) describes a steady decrease of amplitude with height and not a vertical oscillation.

II.4.5 VERTICAL WAVELENGTHS OF TIDAL MODES

Table II.1 shows equivalent depths and vertical wavelengths for various tidal modes (described in II.5). Vertical wavelengths are given for a constant scale height of $H = 6.2$ km, which is roughly the value found between 80 and 100 km altitude, based on a temperature of $T = 210^\circ$K. Equivalent depth values are taken from Chapman & Lindzen [1970].

<table>
<thead>
<tr>
<th>Hough mode</th>
<th>(1,1)</th>
<th>(2,2)</th>
<th>(2,3)</th>
<th>(2,4)</th>
<th>(2,5)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$h_n$ [km]</td>
<td>0.69</td>
<td>7.07</td>
<td>3.67</td>
<td>1.85</td>
<td>1.37</td>
</tr>
<tr>
<td>$\lambda_z$ [km]</td>
<td>25.4</td>
<td>1652.3</td>
<td>99.2</td>
<td>46.5</td>
<td>38.4</td>
</tr>
</tbody>
</table>

Table II.1: Vertical wavelengths $\lambda_z$ and equivalent depths $h_n$ of the main propagating diurnal- and semidiurnal Hough modes. Wavelengths are based on a scale height of 6.2 km. Equivalent depths are taken from Chapman and Lindzen, 1970.

Table II.1 illustrates the dependence of vertical wavelength on longitudinal- as well as latitudinal wavenumbers. One sees that the semidiurnal (2,2), (2,3), (2,4) and (2,5) modes have larger vertical wavelengths than the diurnal (1,1) mode. Among the semidiurnal modes those of higher latitudinal wave order have shorter vertical wavelengths. Observations show that the semidiurnal (2,2) mode propagates to higher altitudes (around 180 km) than the (2,3), (2,4) and (2,5) modes (below 150
Classical Tidal Theory

Chapter II

km). This is explored in depth in Chapter IV. It is important to note that the above stated values for vertical wavelengths are based on equation (II.48) which assumes idealized conditions. In the real atmosphere the wavelengths become smaller with altitude as a result of damping processes. In particular, the (2,2) mode wavelength is in the thermosphere much smaller than that given in Table II.1, with values of around 80 km. The large value of the (2,2) mode vertical wavelength suggests that the mode is almost evanescent at the height in question, in agreement with earlier discussions (see section ). Values of Table II.1 do not consider the negative vertical scale height gradient found in the real atmosphere near the mesopause.

II.5 LATITUDINAL STRUCTURE

The solution of Laplace's tidal equation (II.28) gives the latitudinal structure of tidal oscillations. The pioneering work of Hough [1897, 1898] gave the eigenfunctions \( \Theta_n \) their name Hough functions or Hough modes. Analytically they are given by the expression

\[
\Theta_n^{\omega,s} (\theta) = \sum_{m=1}^{\infty} C_{n,m}^{\omega,s} P_{m,s} (\sin\theta)
\]

Here, \( P_{m,s} (\sin\theta) \) are associated Legendre Polynomials of degree \( m \) and order \( s \), and \( C_{n,m} \) are the expansion coefficients for the polynomial series of a given Hough mode. The modes are often referred to by the pair of numbers \( (s,n) \), \( s \) being the longitudinal wave number and \( n \) the latitudinal wave number. The latter may have values of \( n=\pm1, \pm2, \pm3, \ldots \), where even numbers describe latitudinally symmetric modes and odd numbers antisymmetric modes about the geographic equator. Positive values of \( n \) denote propagating waves and negative values symbolize trapped waves. Always, \(|n| \geq s\) since the function \( \Theta_n \) has \(|n|-s \) nodes between the poles, except for \( s=1 \). The Hough modes of interest here are diurnal- and semidiurnal-, symmetric- and antisymmetric-propagating waves.

Figure II.2 shows the normalized latitudinal structures of the diurnal (1,1) and semidiurnal (2,2), (2,3), (2,4) and (2,5) modes. The graphs illustrate some of the mode properties described earlier verbally, that latitudinal complexity increases with \( n \) and even values of \( n \) denote the symmetric modes. Equations (II.23) and (II.24) express horizontal winds as a function of these profiles, and from those expressions one sees that the corresponding wind profiles are strongest where the geopotential height gradients in Figure II.2 are largest. Therefore, the profiles of wind oscillations
resulting from the above basic Hough modes extend to higher latitudes than seen in Figure II.2, especially for the (2,4) and (2,5) modes which produce large wind amplitudes at up to around ±70°.

It is important to mention one further aspect which becomes evident from the plots in Figure II.2, that the Hough modes shown there are all bound at the poles. This condition holds as long as the modes are propagating, i.e. for \( n > 0 \). Non-propagating modes however have different latitudinal structures, with largest values at the poles and smaller amplitudes towards the equator. One typical example is the diurnal non-propagating (1,-2) mode which is also described in Chapter IV, section IV.5.3 as well as in Figure IV.34. If strong tidal patterns are found at the poles they can analytically be described through non-propagating Hough modes. As outlined in IV.5.3, the (1,-2) mode plays an important role in the thermosphere.
In principle, Laplace's tidal equation (II.28) states that for each set of $s$ and $\omega$ there are a set of $\Theta_n$ and $\varepsilon_n$ which solve the equation. The structure of the equation shows that the $\varepsilon_n$ are eigenvalues of the wave modes. For a fixed longitudinal wave number, $s$, one can display the possible solutions for $\varepsilon_n$ as a function of $\omega$ in a diagram to help classifying the families of oscillations referred to in
II.1 and II.4. Such a diagram is shown in Figure II.3.

**Figure II.3:** Eigenvalues $\varepsilon_n$ of wave modes for zonal wave number $s=1$ versus normalized frequency $\omega/\Omega$ for $T=210$ K. Adapted from [Volland, 1988].

The plot (adapted from [Volland, 1988]) shows a diagram for $s=1$ where the $\varepsilon_n$ are shown against the normalized frequency, $\omega/\Omega$. Positive values of the normalized frequency denote westward- and negative values denote eastward propagating oscillations. Since Figure II.3 is for fixed longitudinal wave number ($s=1$), it shows a family of curves with different latitudinal wavenumbers, each of which are indicated on the curves. Each line represents one tidal $(1,n)$ mode. A diagram for $s=2$ looks similar in principle and is not shown here.

Figure II.3 also distinguishes between Class I and Class II waves. Class I waves are also called atmospheric gravity waves, while Class II are Rossby- or planetary waves. For gravity waves, the main force attempting to restore a perturbation is gravity and reacts to vertical displacements of air cells. Rossby waves, in contrast, are a consequence of the change of Coriolis force with latitude. Here, the restoring force attempts to conserve vorticity and acts on the latitudinal displacement of the cells.
The dashed horizontal line in Figure II.3 separates the regions for which \( \epsilon_n > 12.4 \) and \( \epsilon_n < 12.4 \). As the definitions of \( a^2 \) (II.32) and \( \epsilon_n \) (see II.3) show, the upper region is that for which \( a^2 > 0 \), thus \( \alpha \) is real, while the lower region is for \( a^2 < 0 \), i.e. imaginary \( \alpha \). The upper region incorporates all internal waves, while the lower contains the evanescent (or external) waves. There are numerous ways of interpreting the information in the diagram, as shown in the following.

a) FORCED OSCILLATIONS

If the oscillation is forced, its frequency is known and thus \( \omega / \Omega_E \). If in the diagram a vertical line is drawn at this \( \omega / \Omega_E \) value, the points of intersection with various curves will determine which latitudinal modes this forcing generates. As an example, the points of intersection for the diurnal tide (\( \omega / \Omega_E = 1.0 \)) are indicated (right panel in Figure II.3). A number of these are for trapped modes (\( n < 0 \)) and the rest for propagating modes (\( n > 0 \)). For the latter, the scale on the top left side of the diagram gives vertical wavelengths (based on \( T = 210^\circ K \)). Thus the diagram also provides information about the vertical structure of the forced wave.

b) FREE OSCILLATIONS

If an oscillation is unforced its equivalent depth, \( h_n \), has a fixed value of \( h_n = 8.7 \) km. This was derived in (II.44), assuming \( T = 210^\circ K \), which also gives a fixed value for \( \epsilon_n = 10.1 \). In Figure II.3 the horizontal set of dots are the points of intersection between the horizontal line at \( \epsilon_n = 10.1 \) and the curves, thus indicating the complete set of diurnal modes occurring in an unforced environment, or at sufficient distance from the region of forcing. Apart from numerous trapped modes there are two propagating modes, namely (1,1) and (1,2).

II.6 GROUP AND PHASE VELOCITY

The previous sections of this chapter have outlined an analytical description of upwards propagating tides. Curiously, however, contour plots of various parameters versus local time and height show downwards propagating features. In order to understand this, it is necessary to distinguish between group- and phase velocity of a wave. This is relevant only to propagating waves, with \( \alpha^2 > 0 \).
The vertical group velocity for the propagating wave, as defined in (II.4.4), is given by

$$c_{\text{group},x} = \frac{\partial \omega}{\partial x}$$  \hspace{1cm} (II.50)

This expression may be re-written in the form

$$c_{\text{group},x} = \frac{\partial \omega}{\partial x} = 2 \alpha \left[ \frac{\partial \omega}{\partial x} \right]$$  \hspace{1cm} (II.51)

where the relation \(\partial(\alpha x) = 2 \alpha \partial x\) was used. Inserting the definitions of \(\epsilon_{n}\) and \(\alpha x\) from (II.32) gives

$$c_{\text{group},x} = \alpha \frac{8 \Omega_{E}^2 r_{L}^2}{\kappa g H} \frac{\partial \omega}{\partial x}$$  \hspace{1cm} (II.52)

Since the waves considered are propagating, it is known that \(\alpha x > 0\), so there is uncertainty about the sign of \(\alpha\). The radiation condition at \(x = \infty\) (see II.4.2) requires upwards propagation of energy, and thus \(c_{\text{group},x} > 0\). In Figure 11.3 one sees that for westward travelling waves (right panel in diagram) the slope is negative \((\partial \omega/\partial x < 0)\) while for eastward travelling waves (left panel in diagram) \((\partial \omega/\partial x > 0)\). In order to satisfy (II.52), this implies that for westward travelling waves \(\alpha x < 0\) and for eastward travelling waves \(\alpha x > 0\).

An expression for the vertical phase velocity may be derived when considering the total phase term for a tidal oscillation is given by \((\omega x + x \Phi + \alpha x)\). At a fixed longitude the phase is thus given by \((\omega x + \alpha x)\). The phase velocity is then

$$c_{\text{phase},x} = \frac{\omega}{\alpha}$$  \hspace{1cm} (II.53)

For westward travelling waves it is known that \(\omega < 0\) and \(\alpha < 0\), while for eastward travelling waves \(\omega > 0\) and \(\alpha > 0\). In both cases, therefore,

$$c_{\text{phase},x} < 0$$
At fixed longitude the phase therefore propagates *downwards* with time while the energy propagates upwards:

\[ c_{\text{group},x} > 0 \]

For the CTIM simulations of Chapter IV, typical values of \( c_{\text{phase},x} \) in the lower thermosphere were found to lie around -0.7 hours/km for diurnal tides and -0.06 hours/km for semidiurnal tides, but values depend very much on the vertical wavelength of the wave and on background atmosphere conditions.
CHAPTER III. THE MODEL

ABSTRACT

This chapter presents the main tool used for studies of Chapters IV-VII, the Coupled Thermosphere-Ionosphere model (CTIM). Since an important proportion of tidal oscillations found in the lower thermosphere originate from the middle atmosphere, but these regions lie outside the CTIM's height regime, they need to be specified externally. For this purpose, the previously fixed lower boundary of CTIM was developed into a flexible layer which allows specification of global tidal- and planetary wave profiles. While in Chapter II the theoretical background for analytical lower boundary perturbation profiles was presented, this chapter presents the set of coupled equations for wind-, temperature and pressure oscillations which have been implemented into the model. This is preceded by a detailed description of the code in its present form, distinguishing between the thermosphere model and the fully coupled ionosphere code.

III. 1. INTRODUCTION

The modelling studies presented in this thesis are carried out with a global three-dimensional general circulation model of the terrestrial atmosphere above 80 km. One of the main characteristics of this region is the simultaneous occurrence of charged (ionospheric) and neutral (thermospheric) gases. The large differences found in the behaviour of ions and neutral particles have historically implied that modellers tended to concentrate on simulating either the thermosphere or the ionosphere alone. The model used here fully integrates two originally separate models of the thermosphere and high-latitude ionosphere which were developed by different groups in the UK, and represents the only example of a fully integrated upper atmosphere model of its kind on this continent. This chapter will, in sections III.2, III.3 and III.4, outline the main characteristics of the model and in section III.5 describe in detail the changes made to the code in order to simulate upwards propagating tides.

III. 2. THE INTEGRATED MODEL

The model used here is available to users in three basic versions. In its most basic version, the
thermosphere code is run on its own, using the empirical ionosphere model by Chiu [1975] for the “background” charged particle properties. This version is suitable for carrying out thermospheric studies only and has the advantage of running at more than twice the speed of the fully integrated model. For high-latitude ionosphere studies another model version is used which consists of the coupled thermosphere- and high-latitude ionosphere code. It is in the literature referred to as the Coupled Thermosphere-Ionosphere Model (CTIM) and has for many years been the main version of the model. Between latitudes 34°N and 34°S, the empirical model by Chiu [1975] is still used, implying that the calculations of charged particle momentum-energy and composition there are not treated self-consistently. The advantage of this is a saving of around 30% running time, compared with the third version. The latter incorporates the thermosphere-, ionosphere- and a low-latitude plasmasphere model and is in the literature referred to as the Coupled Thermosphere-Ionosphere-Plasmasphere Model (CTIP). This version is particularly useful for studies of the low-latitude charged particles and currently requires around 6 hours of run-time for a full 24-hour simulation on a DEC Alpha AXP 3000/400 system with 64 Mb of memory.

All runs carried out here use CTIM in order to simulate ionospheric parameters accurately at high latitudes. Although the upwards propagating tides are particularly important at low latitudes, the plasmasphere code has not been used here for two reasons. One aspect is that the low-latitude electric dynamo field is not calculated self-consistently in any of the model versions currently available, implying that the ion-neutral coupling at low latitudes is not treated sufficiently self-consistent to justify extra run-time for the large amount of simulations presented here. The other aspect is that at the time of writing the plasmasphere code was being modified considerably and the new code had not been fully tested and verified. Therefore, the plasmasphere code is not described separately in this chapter. The basic equations it uses are the same as in the ionosphere code (section III.4), and calculations are carried out along closed magnetic flux tubes between ±55° magnetic latitude. The main difference to the ionosphere code is that the plasmasphere is co-rotating with the Earth, while the ionosphere is not. A detailed description of the earlier version of the plasmasphere code can be found in [Millward et al., 1996].

III. 3. THE THERMOSPHERE MODEL

The thermosphere model, originally developed by Fuller-Rowell and Rees [1980], solves the non-linear equations of momentum, energy and continuity for the three neutral components O, O₂ and
Calculations are carried out on a global spherical pressure coordinate system with grid-points spaced 2° in geographic latitude, 18° in geographic longitude and 1 scale height vertically. The grid points are fixed relative to the Earth’s surface and thus form a co-rotating, non-inertial frame of reference. Calculations are performed at time steps of 60 seconds by default, though the code automatically uses a 30 sec time step when necessary. The model’s lower boundary is at a pressure level of 1.0376 Pa (around 80 km) and the upper boundary lies 14 scale heights above, at 8.63 \times 10^{-7} Pa. Table III.1 shows globally averaged heights of CTIM’s 15 pressure levels [in km] under weak and stronger solar activity (F10.7=75 and F10.7=200, respectively). One sees considerable variability of the pressure level heights with solar activity, and at high latitudes the magnetic activity plays a similarly important role. This altitude range covers the mesopause vicinity and the E- and F-regions.

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Table III.1: Globally averaged heights of the thermosphere model’s pressure levels [in km] under two different conditions of solar activity. The values denoted “low” are for F10.7=75 and those referred to as “high” are for F10.7=200. The magnetic activity is moderate in both simulations (Kp=2-).

III. 3.1. THE PRESSURE COORDINATE SYSTEM

The use of a spherical pressure coordinate system instead of a height coordinate system considerably simplifies the fundamental equations in CTIM. The difference to a height coordinate system is that the height coordinate variable, \( z \), is replaced by a vertical variable, \( p \), denoting the pressure at that height. The important assumption made in this transformation is that hydrostatic equilibrium holds throughout the region of interest (the thermosphere), and that gravitational acceleration, \( g \), is constant vertically as well as horizontally. In hydrostatic equilibrium, vertical gravitational acceleration balances the vertical pressure gradient force, as expressed by the hydrostatic equation

\[ \frac{\partial p}{\partial z} = -\rho g \]
\[
\frac{\partial p}{\partial z} = -\rho g \tag{III.1}
\]

where \(p\) denotes the pressure, \(z\) the vertical height coordinate and \(\rho\) the mass density. In the thermosphere, thermal heating may cause departures from hydrostatic equilibrium, but the time scales of vertical motion introduced by this are long enough to assume quasi-hydrostatic equilibrium \cite{Rishbeth1969}. In the following the height coordinate system is referred to as "\(z\)" and the pressure coordinate system as "\(p\)".

A scalar, \(S\), is transformed from the \(z\)-system into the \(p\)-system by

\[
\left( \frac{\partial S}{\partial a_x} \right)_p = \left( \frac{\partial S}{\partial a_x} \right)_z + \left( \frac{\partial z}{\partial a_x} \right)_p \left( \frac{\partial S}{\partial z} \right) \tag{III.2}
\]

where \(a_x\) is a horizontal coordinate. As in Chapter II, the geopotential, \(\Phi\), is introduced and given by

\[
d\Phi = g\, dz \tag{III.3}
\]

From equation (III.3) one sees that

\[
\left( \frac{\partial z}{\partial a_x} \right)_p = \frac{1}{g} \left( \frac{\partial \Phi}{\partial a_x} \right) \tag{III.4}
\]

Inserting (III.4) into (III.2) gives

\[
\left( \frac{\partial S}{\partial a_x} \right)_z = \left( \frac{\partial S}{\partial a_x} \right)_p + \rho \cdot \left( \frac{\partial \Phi}{\partial a_x} \right)_p \left( \frac{\partial S}{\partial p} \right) \tag{III.5}
\]

where (III.1) was used to substitute
Similarly, the time derivative is transformed through

\[
\left( \frac{\partial S}{\partial t} \right)_z = \left( \frac{\partial S}{\partial t} \right)_p + \rho \cdot \left( \frac{\partial \Phi}{\partial t} \right)_p \cdot \left( \frac{\partial S}{\partial p} \right)
\]

(III.7)

Let \( \nabla_z \) be the two-dimensional del-operator in the \( z \)-frame on a level of constant height and \( \nabla_p \) the del-operator in the \( p \)-frame on a constant pressure surface. Applying (III.5) also to the \( y \)-direction gives for the transformation of horizontal gradients between the coordinate systems

\[
\tilde{\nabla}_z S = \tilde{\nabla}_p S + \rho \cdot \tilde{\nabla}_p \Phi \cdot \frac{\partial S}{\partial p}
\]

(III.8)

The vertical gradient transformation is given by

\[
\frac{\partial S}{\partial z} = \frac{\partial S}{\partial p} \cdot \frac{\partial p}{\partial z} = -g \rho \frac{\partial S}{\partial p}
\]

(III.9)

III. 3. 2. VERTICAL WIND

The vertical wind in the \( p \)-system, \( w \), is defined as

\[
w = \frac{dp}{dt}
\]

(III.10)

and represents the velocity relative to a level of constant pressure. \( w \) is positive for a descending- and negative for an ascending air parcel. One of the thermosphere model's output parameters is the vertical velocity in the height frame, \( v_z \), which is given by

\[
v_z = \frac{dz}{dt}
\]

(III.11)

In order to derive the relationship between \( w \) and \( v_z \), one needs to express the total time derivative \( d/dt \) in the \( p \)-system:
where $V_p$ is the 2-dimensional velocity vector on a level of constant pressure. Applying (III.12) to operate on the geopotential, $\Phi$, gives with (III.9)

$$\frac{d\Phi}{dt} = g \frac{dz}{dt} = \left( \frac{\partial \Phi}{\partial t} \right)_p + \vec{V}_p \cdot \vec{\nabla}_p \Phi - \frac{w}{\rho}$$

Therefore, the vertical velocity in the height frame is given by

$$v_z = \frac{1}{g} \left[ \frac{\partial \Phi}{\partial t} \right]_p + \frac{1}{g} \vec{V}_p \cdot \vec{\nabla}_p \Phi - \frac{w}{\rho g}$$

Rishbeth et al. [1969] distinguish between two vertical wind components, the barometric velocity, $v_b$, and the divergence velocity, $v_d$. The latter represents motion relative to a fixed pressure level, the former is the motion of the pressure level itself, and their sum is the vertical velocity $v_z$ in the height frame, $v_z = v_b + v_d$. Barometric wind is given by the first two terms on the right side of equation (III.14),

$$v_b = \frac{1}{g} \left[ \frac{\partial \Phi}{\partial t} \right]_p + \frac{1}{g} \vec{V}_p \cdot \vec{\nabla}_p \Phi$$

The divergence wind is expressed by the third term in equation (III.14) and thus directly related to the vertical wind in the pressure coordinate frame, $w$.

$$v_d = -\frac{w}{\rho g}$$

The divergence wind is the vertical wind velocity in the $p$-frame, $w$, transformed to the $z$-system. The relationship (III.16) can thus also be understood from equation (III.6).

From the continuity equation in the $p$-frame (III.26) one sees that divergence wind is coupled to horizontal wind gradients. In contrast, barometric vertical wind is the result of thermal expansion. The concept of distinguishing between these two components is useful in various studies and applied in Chapter VI where tidally induced composition changes are investigated.
III. 3.3. MOMENTUM EQUATION

The basic assumption made for the momentum equation is that the thermosphere, although being a tenuous gas, is considered to be dominated by collisions and thus isotropic. The thermospheric gas is therefore treated as a fluid and the Navier-Stokes expressions are applied. The change of neutral wind velocity is given by the balance of forces acting on a parcel of air:

\[
\frac{d\vec{V}}{dt} = \vec{g} - \frac{1}{\rho} \nabla \rho - 2 \Omega \times \vec{V} - v_n (\vec{V} - \vec{U}) + \frac{1}{\rho} \nabla (\mu \nabla V) \tag{III.17}
\]

with \( \vec{V} \) denoting the 3-dimensional velocity vector in the \( z \)-frame, \( \vec{g} \) denoting the gravity vector, \( \rho \) the neutral density, \( \Omega \) the Earth’s angular rotation rate, \( v_n \) the neutral-ion collision frequency, \( \vec{U} \) the (3-dimensional) ion drift velocity and \( \mu \) the sum of the molecular and turbulent viscosity coefficient. The terms on the right side of equation (III.17) describe the gravity force, pressure gradient, Coriolis force, ion drag and viscosity, respectively. The pressure coordinate system used in the model is an Eulerian frame of reference, where latitude and longitude are independent variables. Transforming (III.17) to the pressure coordinate system gives for the change of the southward (\( V_{\phi} \)) and eastward (\( V_{\theta} \)) winds:

\[
\frac{\partial V_{\theta}}{\partial t} = - \frac{V_{\theta}}{R} \frac{\partial V_{\theta}}{\partial \theta} - \frac{V_{\phi}}{R \sin \theta} \frac{\partial V_{\phi}}{\partial \phi} - \omega \frac{\partial V_{\phi}}{\partial \rho} - \frac{g}{R} \frac{\partial h}{\partial \theta} + \left( 2 \Omega + \frac{V_{\phi}}{R \sin \theta} \right) V_{\phi} \cos \theta + g \frac{\partial}{\partial \rho} \left( \mu_m + \mu_f \right) \frac{P}{H} \frac{\partial V_{\phi}}{\partial \rho} - v_n (V_{\phi} - U_{\phi}) \tag{III.18}
\]

\[
\frac{\partial V_{\phi}}{\partial t} = - \frac{V_{\theta}}{R} \frac{\partial V_{\phi}}{\partial \theta} - \frac{V_{\phi}}{R \sin \theta} \frac{\partial V_{\phi}}{\partial \phi} - \omega \frac{\partial V_{\phi}}{\partial \rho} - \frac{g}{R \sin \theta} \frac{\partial h}{\partial \phi} - \left( 2 \Omega + \frac{V_{\phi}}{R \sin \theta} \right) V_{\theta} \cos \theta + g \frac{\partial}{\partial \rho} \left( \mu_m + \mu_f \right) \frac{P}{H} \frac{\partial V_{\phi}}{\partial \rho} - v_n (V_{\phi} - U_{\phi}) \tag{III.19}
\]

The dependency of the ion drag terms on ion velocity, \( U \), and neutral-ion collision frequency, \( v_n \), can be eliminated by introducing the electric current density, \( J \), as described in detail in section III.3.7. Equations (III.18) and (III.19) with (III.37) are used in the thermosphere model to evaluate winds on levels of constant pressure at every time step.
The molecular viscosity term, $\mu_m$, used in the model is taken from [Dalgarno and Smith, 1962] and given by

$$\mu_m = 4.5 \cdot 10^{-5} \left( \frac{T}{1000} \right)^{0.71} \text{[kg m}^{-1}\text{s}^{-1}]$$  \hspace{1cm} (III.20)

Turbulent viscosity, $\mu_t$, is calculated from the turbulent heat conduction, $K_t$, using the relationship [Fuller-Rowell, 1984]

$$\mu_t = K_t \cdot \frac{2}{c_p}$$  \hspace{1cm} (III.21)

The values used in the model for $K_t$ are given in section III.3.4.

### III.3.4. ENERGY EQUATION

The energy balance in the thermosphere is determined by external energy sources and sinks as well as internal energy distribution processes. The external heating sources and sinks considered in the thermosphere model are solar heating, Joule heating through charged particles which are driven by the convection electric field at high latitudes, tidal forcing from the mesopause and radiative cooling. Internal energy flux occurs via ion drag, heat conduction, viscous forces and advection. The model version used in these studies neglects radiative cooling through CO$_2$ and NO which occur at altitudes of around 110 km and 170 km, respectively.

The energy equation used in the thermosphere model is in the p-system given by

$$\frac{\partial e}{\partial t} + \vec{v}_p \cdot \vec{\nabla}_p (e + g Z_p) + w \frac{\partial(e + g Z_p)}{\partial p} = Q_{\text{EUV}} + Q_{\text{JR}} + g \frac{\partial}{\partial p} \left( \frac{K_m + K_t}{H} \right) p \frac{\partial T}{\partial p}$$

$$+ \frac{1}{\rho} (K_m + K_t) \nabla^2_p T - g \frac{\partial}{\partial p} \left( \frac{K_t g}{C_p} \right) + \frac{J \cdot E}{p}$$  \hspace{1cm} (III.22)

where $e = \frac{1}{2} \vec{V}^2 + c_p T$ is defined as the specific enthalpy plus kinetic energy density per unit of mass of gas. Furthermore, $Z_p$ is the height of the pressure level, $Q_{\text{EUV}}$ and $Q_{\text{JR}}$ denote solar heating and infrared cooling, respectively, $K_m$ and $K_t$ are the molecular and turbulent coefficients of heat conduction and $c_p$ is the specific heat at constant pressure. $E$ is an external electrical field, and the model allows the option of choosing amongst various statistical convection fields (see III.3.8).
Values for $K_m$ are consistent with those by Bauer [1973], with initial conditions taken from the Jacchia [1971] model atmosphere for 1000 °K exosphere temperature. The $K_T$ values are consistent with those by Iwasaka [1973], Johnson and Gottlieb [1970], Lloyd et al. [1972] and Justus [1967]. Values have however been modified in the past to obtain better vertical temperature profiles below 120 km height. The $K_m$ and $K_T$ coefficients are used as pre-calculated values and remain by default constant during a simulation. Studies in Chapter V will show the effect of changing the molecular diffusion coefficient. The cooling term, $Q_{IR}$, assumes the 63μm radiative cooling by atomic oxygen. Values used in the model are based on the formula by Bates [1951]:

$$Q_{IR} = \frac{1.67 \cdot 10^{-25} \exp(-228.0/T_n)}{1 + 0.6 \exp(-228.0/T_n) + 0.2 \exp(-325.0/T_n)} \cdot [O] \quad [W m^{-2}] \quad (III.23)$$

where $T_n$ is the neutral temperature (°K) and [O] the atomic oxygen number density (m⁻³). The expression is here given in c.g.s. units since that corresponds to the form used in the model code. However, Creekmore et al. [1972] found that this expression overestimated the cooling rate by up to 50% at 130 km altitude. Therefore, expression (III.23) is used with the damping factor

$$Q_{IR, new} = Q_{IR} \cdot \left(1 - \exp\left[\frac{80.0 - Z}{20.0}\right]\right)^4 \quad (III.24)$$

where the altitude Z is used in units of [km].

The solar heating term in equation (III.22), $Q_{EUV}$, is of considerable importance to the energy balance since it forms the major source of energy in the thermosphere. The wavelength bands considered in the model are the extreme ultraviolet (3-131 nm) and in the ultraviolet the Schumann-Runge continuum (131-240 nm). In order to evaluate the heating term, three main parameters need to be specified, the total number of absorbing O₂ molecules in a unit column along a ray path from the Sun at a given point, the solar flux and the heating efficiency factor. The number of O₂ molecules in the column along the ray-path is calculated using a routine by Creekmore et al. [1972], and the O₂ density is evaluated self-consistently in the model. Values for the solar flux at wavelengths > 180 nm are taken from Straus et al. [1975]. Wavelengths shorter than 180 nm are based on values by Hinteregger [1970] which were revised following Schmidke et al. [1977] and Torr et al. [1979]. The heating efficiency factor was originally chosen as 0.3 but, following results from Torr et al. [1980 a,b] and Roble et al. [1987], is now dependent on the pressure level. More
recent solar flux values for the wavelengths > 180 nm were presented by Feng et al. [1989] and Ogawa [1990], but have not been implemented into the current version of CTIM. The level of solar activity is specified through the F10.7 solar index which gives the flux of solar radiation at 10.7 cm wavelength. This wavelength is of no importance in the upper atmosphere, but is simple to measure and was found to correlate very well with the UV and EUV fluxes.

III. 3.5. CONTINUITY EQUATION AND COMPOSITION

The continuity equation describes the conservation of mass and in the z-frame is given by

$$\frac{\partial \rho}{\partial t} = - \nabla \cdot \rho \vec{V}$$

(III.25)

where, as previously, $\rho$ is the neutral density, $\vec{V}$ the 3-dimensional velocity vector and $\nabla$ the 3-dimensional delta-operator. Using (III.3), (III.7), (III.8), (III.9) and (III.14) one may transfer (III.25) into the p-system [Fuller-Rowell, 1981] and obtains

$$\frac{\partial w}{\partial p} + \vec{V}_p \cdot \vec{V}_p = 0$$

(III.26)

This equation forms the basis for equation (III.30) which is used in the model to calculate composition. While in the thermosphere model’s original form the neutral gas was assumed of uniform composition, a later extension introduced the two-component gas [Fuller-Rowell and Rees, 1983] which recently was replaced with the three-component composition [Fuller-Rowell et al. 1994]. The three major species in the thermosphere model are molecular nitrogen, atomic oxygen and molecular oxygen. Their concentrations are determined by chemical production and loss, mutual molecular diffusion, horizontal and vertical advection and horizontal as well as vertical turbulent mixing. The model solves the three-dimensional time-dependent continuity equation for each of these major species. In order to determine the diffusion velocities for each constituent it simultaneously solves a generalized form of the diffusion equation for a multiple-species non-uniform gas which was derived by Chapman and Cowling [1952] to be

$$\frac{1}{n} \sum_{ij} \left( \frac{\chi_i}{m_i} D_{ij} \frac{n_j m_j C_j}{n_i m_i C_i} - \frac{\chi_i}{m_j} D_{ij} \frac{n_j m_j C_j}{n_i m_i C_i} \right) = \vec{V}_i \chi_i + \frac{\chi_i}{m} \vec{V} m + \left( 1 - \frac{m_i}{m} \right) \frac{\chi_i \vec{V}_p}{p}$$

(III.27)

where $\chi_i = n_i m_i / p$ are mass mixing ratios for species $i$ ($i = O, O_2, N_2$), $m_i$ is the molecular mass
of species \( i \), \( n_i \) is its number density, \( D_{ij} \) is the mutual molecular diffusion coefficient between species \( i \) and \( j \) and \( C_i \) is the diffusion velocity. The mean molecular mass, \( m \), is given by

\[
m = \frac{n_1 m_1 + n_2 m_2 + n_3 m_3}{n} \quad \text{(III.28)}
\]

with

\[
n = n_1 + n_2 + n_3 \quad \text{(III.29)}
\]

denoting the total number density. The continuity equation (III.26) can be extended to allow an air parcel to be influenced by the surrounding parcels. Written in terms of the mass mixing ratio, \( \chi_i \), for species \( i \), it was derived by Fuller-Rowell [1984] to be

\[
\frac{\partial \chi_i}{\partial t} = \frac{1}{\rho} m_i S_i - \vec{V}_p \cdot \vec{\nabla}_p \chi_i - w \frac{\partial}{\partial p} \chi_i - \frac{1}{\rho} \vec{\nabla}_p \cdot (n_i m_i C_i) + \frac{1}{\rho} \vec{V}_p (D_{\text{EDDY}} n \vec{\nabla}_p m \chi_i) \quad \text{(III.30)}
\]

where, \( S_i \) denotes sources and sinks of the species by the chemistry outlined below and \( D_{\text{EDDY}} \) is the eddy diffusion coefficient. The terms on the right side of (III.30) are chemical changes, horizontal advection, vertical advection, molecular diffusion and eddy diffusion, respectively. The model code solves (III.30) for each constituent \( i \) and uses the three coupled diffusion equations (III.27) to obtain the diffusion velocities of the species. The molecular diffusion coefficients \( D_{ij} \) used in the model are assumed symmetric, \( D_{ij} = D_{ji} \), and are taken from the expressions by Colegrove et al. [1966]:

<table>
<thead>
<tr>
<th>( i - j )</th>
<th>( D_{ij} ) [m²s⁻¹]</th>
</tr>
</thead>
<tbody>
<tr>
<td>( O-O_2 )</td>
<td>( \frac{[T/273.0]^{1.75}}{[1.01325 \cdot 10^5/p]} \cdot 2.60 \cdot 10^{-5} )</td>
</tr>
<tr>
<td>( O-N_2 )</td>
<td>( \frac{[T/273.0]^{1.75}}{[1.01325 \cdot 10^5/p]} \cdot 2.60 \cdot 10^{-5} )</td>
</tr>
<tr>
<td>( O_2-N_2 )</td>
<td>( \frac{[T/273.0]^{1.75}}{[1.01325 \cdot 10^5/p]} \cdot 1.81 \cdot 10^{-5} )</td>
</tr>
</tbody>
</table>

Table (III.2) Molecular diffusion coefficients used in the thermosphere model, after [Colegrove et al., 1966]
where the temperature, $T$, is assumed in units of Kelvin and the pressure, $p$, in Pascal. The eddy diffusion coefficient, $D_{EDDY}$, is taken from [Reber, 1973] and given by

\[
D_{EDDY} [\text{m}^2 \text{s}^{-1}] = \begin{cases} 
100.0 \cdot \exp\left[3.0 \cdot (Z-105.0) \cdot 10^{-5}\right] + 50.0 \cdot \exp\left[-3.0 \cdot (Z-105.0)^2 \cdot 10^{-8}\right] & \text{for } 80 \text{ km} \leq Z \leq 105 \text{ km} \\
150.0 \cdot \exp\left[-5.0 \cdot (Z-105.0)^2 \cdot 10^{-8}\right] & \text{for } 105 \text{ km} < Z \leq 150 \text{ km} \\
0 & \text{for } Z > 150 \text{ km}
\end{cases}
\]

<table>
<thead>
<tr>
<th>Height</th>
<th>$D_{EDDY}$ [m$^2$ s$^{-1}$]</th>
</tr>
</thead>
<tbody>
<tr>
<td>$80 \text{ km} \leq Z \leq 105 \text{ km}$</td>
<td>$100.0 \cdot \exp\left[3.0 \cdot (Z-105.0) \cdot 10^{-5}\right] + 50.0 \cdot \exp\left[-3.0 \cdot (Z-105.0)^2 \cdot 10^{-8}\right]$</td>
</tr>
<tr>
<td>$105 \text{ km} &lt; Z \leq 150 \text{ km}$</td>
<td>$150.0 \cdot \exp\left[-5.0 \cdot (Z-105.0)^2 \cdot 10^{-8}\right]$</td>
</tr>
<tr>
<td>$Z &gt; 150 \text{ km}$</td>
<td>0</td>
</tr>
</tbody>
</table>

**Table (III.3)** Eddy diffusion coefficient used in the model, after [Reber, 1973].

While the eddy diffusion is dominant below 100 km, it quickly vanishes above 105 km and molecular diffusion becomes more important. The height at which both coefficients have the same value is defined as the turbopause. With values for the coefficients from **Table (III.2)** and **Table (III.3)** the turbopause lies at around 100 km altitude. More detailed studies discussing molecular diffusion are presented in Chapter VI.

The neutral chemistry considered in the model is given by the following 5 reactions

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Rate Constant</th>
</tr>
</thead>
<tbody>
<tr>
<td>(A) $O_2 + h\nu \rightarrow O + O$</td>
<td>see Table III.5</td>
</tr>
<tr>
<td>(B) $O + O + M \rightarrow O_2 + M$</td>
<td>$k_B = 2.7 \cdot 10^{-46} \exp\left(710/T\right) \text{ m}^6 \text{ s}^{-1}$</td>
</tr>
<tr>
<td>(C) $O + OH \rightarrow O_2 + H$</td>
<td>$k_C = 4.2 \cdot 10^{-17} \text{ m}^3 \text{ s}^{-1}$</td>
</tr>
<tr>
<td>(D) $O + HO_2 \rightarrow O_2 + OH$</td>
<td>$k_D = 3.5 \cdot 10^{-17} \text{ m}^3 \text{ s}^{-1}$</td>
</tr>
<tr>
<td>(E) $O + O + O_2 + h\nu$</td>
<td>$k_E = 1.0 \cdot 10^{-26} \text{ m}^3 \text{ s}^{-1}$</td>
</tr>
</tbody>
</table>

**Table (III.4)** Neutral chemistry in the thermosphere model.

The rate coefficient for photo dissociation of $O_2$ is given in parameterized form, based on studies by R. Roble [private comm., 1983] which were carried out with the model by Roble and Emery [1983]. The vertical profile is depicted in [Fuller-Rowell, 1984], and values for a selection of
heights are given in Table III.5.

<table>
<thead>
<tr>
<th>Height [km]</th>
<th>80</th>
<th>120</th>
<th>160</th>
<th>220</th>
<th>320</th>
</tr>
</thead>
<tbody>
<tr>
<td>(O_2) dissociation rate [sec(^{-1})]</td>
<td>(1.0 \times 10^{-8})</td>
<td>(8.0 \times 10^{-7})</td>
<td>(1.8 \times 10^{-6})</td>
<td>(2.0 \times 10^{-6})</td>
<td>(2.3 \times 10^{-6})</td>
</tr>
</tbody>
</table>

**Table III.5:** Photo dissociation rates of \(O_2\) as used in the CTIM. The rates are by R Roble [private comm., 1983].

When investigating the relative importance of reactions in Table III.4 with the given rate coefficients and typical constituent densities one finds that the photodissociation of \(O_2\) (A) is the dominant neutral reaction in the thermosphere above 110 km height, while reactions (B), (C), (D) and (E) are important only between 80 and around 100 km height. Reactions (C) and (D) are at present not calculated fully self-consistently in the model; parameterized vertical profiles of OH and HO\(_2\) are used. The chemistry is further discussed in Chapter VI.

### III. 3. 6. BOUNDARY CONDITIONS

The model's lower boundary is defined at pressure level 1 (1.0376 Pa) and lies at 80 km altitude. In the thermosphere model's original version this lower boundary assumed a globally uniform temperature of 180°K with horizontal and vertical winds being set to zero. Considerable modifications of lower boundary conditions have been implemented and are subject of section III.5. The upper boundary is the 15\(^{th}\) pressure level (8.6 \times 10^{-7} Pa) and lies for average solar and geomagnetic conditions at around 400 km altitude. One may assume no sources of momentum and energy at that height [Fuller-Rowell, 1981] and therefore may specify the upper boundary as follows: vertical gradients of the total velocity, \(v_z\), and the vertical velocity in the p-system, \(w\), are set to be zero in order to prevent vertical mass flow and keep the model self-contained. Furthermore, vertical flux of heat across the upper boundary is prevented by setting the vertical temperature gradient to zero. In practise, this is done by setting the temperatures and horizontal winds on pressure level 15 globally equal to those at pressure level 14. At both geographic poles values of any parameter are obtained by zonally averaging the values from the next equatorward latitude.
III. 3. 7. ION-NEUTRAL COUPLING TERM

In the upper atmosphere, an important fraction of the gas is ionized by the solar extreme ultraviolet (XUV) and ultraviolet (UV) radiation. In the absence of a magnetic and electric background field the presence of charged particles would have no considerable influence on the dynamics of the region. However, the Earth’s magnetic- and the high-latitude electric fields both influence the dynamics of the charged particles. Since the sizes of the charged and neutral atoms and molecules are similar, collisions between the charged and neutral gas particles occur, implying that their dynamics are coupled. This coupling is considered in the thermospheric momentum equation (III.17) through the ion-neutral coupling term $v_{ni}(V-U)$, where $v_{ni}$ is the neutral-ion collision frequency, $U$ is the ion- and $V$ the neutral gas velocity. In order to understand the influence of ion-neutral coupling on the dynamics of the neutral gas, one may distinguish between two cases, one assuming the presence of an electric field and the other assuming no electric field (or a very weak one). In the absence of an external electrical field, the charged particles are constrained to move in spirals around magnetic field lines. The force acting on a charged particle is given by $q(U \times B)$ and is perpendicular to the particle velocity, $U$, implying that no work is done on the particle and it receives no kinetic energy from the magnetic field. In the presence of ion-neutral coupling the circular ion movement around magnetic field lines causes an overall damping of the neutral gas velocity. This is relevant at mid latitudes where the electric field is weak. At high latitudes, where the convection field is strong, charged particles gain kinetic energy and are accelerated by the electric field. The ions now drag the neutrals and thus cause overall acceleration of neutral gas. The neutral-ion coupling term in equation (III.17) is in CTIM replaced by another expression which uses the current density instead, assuming the concept of layer conductivities. The conversion between the terms is derived in the following.

The single-species equation of motion for charged particles (ions and electrons) is given by Rishbeth [1972] as

$$\frac{d\vec{U}}{dt} = g + \vec{\nabla} \cdot \left( \frac{P_i + P_e}{Nm} \right) - \frac{e}{m_i} (\vec{E} + \vec{U} \times \vec{B}) - v_{ni}(\vec{U} - \vec{V})$$

(III.31)

where $N$ is the ion (or electron-) density (assumed equal), $m_i$ is the ion mass, $P_i$ and $P_e$ are ion- and electron partial pressures, $e$ is the electron charge and $v_{ni}$ is the ion-neutral (or electron-neutral)
The collision frequency. Note that the latter is not the same as the neutral-ion collision frequency, \( v_{ni} \), which describes the rate at which neutral particles encounter (and collide with) an ion. The ion-neutral collision frequency gives the rate at which ions (or electrons) encounter neutral particles. The two rates are different, with \( v_{ni} \) being larger than \( v_n \), throughout the Earth's atmosphere since the concentration of neutral particles is always larger than that of ions (see Figure I.2) and the chances of an ion encountering a neutral are thus much larger than of a neutral encountering an ion.

In (III.31), the Coriolis term \(-2\Omega \times U\) does not appear since it is negligible compared with the ion-neutral collision term, implying that the mean free path of ions is too small on global scale for the Coriolis effect to matter. When assuming a steady-state condition, the horizontal momentum equation derived from (III.31) may be written for ions and electrons as:

\[
e N_i (\vec{E} + \vec{U} \times \vec{B}) - \rho_i v_{in} (\vec{U} - \vec{V}) = 0 \tag{III.32}
\]

\[
e N_e (\vec{E} + \vec{U} \times \vec{B}) - \rho_e v_{en} (\vec{U}_e - \vec{V}) = 0 \tag{III.33}
\]

where \( N_e, U_e \) and \( v_{en} \) are electron density, -velocity and electron-neutral collision frequency, respectively, and \( \rho_e, \rho_i \) the mass densities of electrons and ions. When adding equations (III.32) and (III.33) while assuming charge-neutrality \( (N_i + N_e = 0) \), one obtains

\[
0 = e N_i (\vec{U} \times \vec{B}) - e N_e (\vec{U}_e \times \vec{B}) - \rho_i v_{in} (\vec{U} - \vec{V}) - \rho_e v_{en} (\vec{U}_e - \vec{V}) \tag{III.34}
\]

The current density, \( J \), defines the effective flow of charges and may be written as

\[
J = e N_i \vec{U} - e N_e \vec{U}_e \tag{III.35}
\]

Thus, (III.34) becomes

\[
v_{in} (\vec{U} - \vec{V}) = \frac{1}{\rho_i} J \times \vec{B} - \frac{\rho_e}{\rho_i} v_{en} (\vec{U}_e - \vec{V}) \tag{III.36}
\]

Since the number densities of ions and electrons is assumed equal, but the electron mass is considerably smaller than the ion mass, \( \rho_e \ll \rho_i \), and one may neglect the second term in (III.36),
Expression (III.37) can be used to substitute the ion-drag term in (III.17) if the current density, \( J \), is evaluated. The generalized form of Ohm's law defines \( J \) as

\[
\bar{J} = \sigma \cdot (\bar{E} + \bar{V} \times \bar{B})
\] (III.38)

where \( \sigma \) is the 3x3 conductivity tensor, \( E \) an external convection- or internal polarization field and \( \bar{V} \times \bar{B} \) describes the dynamo field. With \( E, V \) and \( B \) being known quantities, the latter through equation (III.49), the only unknown in equation (III.37) is \( \sigma \). The conductivity tensor is in its general form given by

\[
\sigma = \begin{pmatrix}
\sigma_0 \sin^2 \delta + \sigma_2 \cos^2 \delta & \sigma_2 \sin \delta \cos \delta & (\sigma_0 - \sigma_1) \sin \delta \cos \delta \\
-\sigma_2 \sin \delta & \sigma_1 & \sigma_2 \cos \delta \\
(\sigma_0 - \sigma_1) \sin \delta \cos \delta & -\sigma_2 \cos \delta & \sigma_1 \cos^2 \delta + \sigma_0 \sin^2 \delta
\end{pmatrix}
\] (III.39)

where \( \delta \) is the magnetic dip angle and \( \sigma_0, \sigma_1 \) and \( \sigma_2 \) are the parallel-, Pedersen- and Hall conductivities, respectively, which are derived in CTIM from the electron densities, using relations by Rishbeth and Garriott [1969] (equations III.44, III.45). The parallel conductivity applies where electric- and magnetic field components are parallel. It describes the mobility of charged particles along the magnetic field lines. For electric field components perpendicular to the magnetic field, one may distinguish between motion along the electric field lines, described by the Pedersen conductivity, and perpendicular to the electric and magnetic field lines which is determined by the Hall conductivity. The latter two "transverse" conductivities have peaks in the \( E \)-region, the Hall-at around 100 km and the Pedersen conductivity at around 130 km altitude. At greater heights the parallel conductivity dominates both transverse conductivities and charged particles move mostly along the magnetic field lines, particularly above around 400 km. The large value of the parallel conductivity leads to an important assumption made in this and most other models, that magnetic field lines are treated as equipotentials and no electric potential gradients occur along them. The ion-movement along the magnetic field lines, therefore, is not caused by any electrical field but is
The shape of the height profiles of transverse conductivities implies that separate horizontal layers of enhanced conductivity exist in the $E$-region. This allows to replace the $3\times3$ conductivity tensor (III.39) with a $2\times2$ tensor, as suggested by Rishbeth and Garriott [1969], by using the concept of layer conductivities. Within a region of enhanced conductivity any vertical electrical current will lead to accumulation of charges at the lower- and upper boundaries of the layer since the region immediately outside the layer has a lower conductivity. An internal polarization field builds up until all charge flow is horizontal within the layer. Therefore, one may assume that currents in the $E$-region flow mostly horizontally. The layer conductivity tensor then becomes

$$\bar{\sigma}_{\text{layer}} = \begin{pmatrix} \sigma_{xx} & \sigma_{xy} \\ \sigma_{yx} & \sigma_{yy} \end{pmatrix}$$

(III.40)

where

$$\sigma_{xx} = \frac{\sigma_0 \sigma_1}{\sigma_1 \cos^2\delta + \sigma_0 \sin^2\delta} = \frac{\sigma_1}{\sin^2\delta}$$

(III.41)

$$\sigma_{yx} = \sigma_{xy} = \frac{\sigma_0 \sigma_2 \sin\delta}{\sigma_1 \cos^2\delta + \sigma_0 \sin^2\delta} = \frac{\sigma_2}{\sin\delta}$$

(III.42)

$$\sigma_{yy} = \frac{\sigma_1 \sigma_0 \sin^2\delta + (\sigma_1^2 + \sigma_2^2) \cos^2\delta}{\sigma_1 \cos^2\delta + \sigma_0 \sin^2\delta} = \sigma_1$$

(III.43)

for larger magnetic dip angles [Rishbeth and Garriott, 1969], and the Pedersen and Hall conductivities are expressed by
\[ \sigma_1 = \frac{N_i e}{B} \frac{r}{1 + r} \]  

\[ \sigma_2 = \frac{N_i e}{B} \frac{r^2}{1 + r^2} = \sigma_1 \cdot r \]

respectively, where

\[ r = \frac{m_i v_m}{e B} \]

and, as previously, \( N_i \) is the electron or ion density (assumed equal), \( m_i \) is the mean ion mass and \( e \) the electron charge. The magnitude of the Earth’s magnetic field, \( B \), is calculated using equation (III.49). The parameter \( r \) in equation (III.46) is the ratio of the collision- to gyro-frequencies of the ions. Instead of the neutral-ion collision frequency one therefore now needs the ion-neutral collision frequency, \( v_m \), to calculate the ion drag term in the momentum equation. In CTIM, the ion-neutral collision frequency is for molecular ions \( \text{NO}^+ \) and \( \text{O}_2^+ \) taken from [Schunk and Walker, 1973]:

\[ v_{m,\text{mol}} = 4.34 \times 10^{-16} [N_2] + 4.28 \times 10^{-16} [O_2] + 2.44 \times 10^{-16} [O] \]  

The \( \text{O}^+ - \text{O}, \text{O}^- - \text{O}_2 \) and \( \text{O}^+ - \text{N}_2 \) collision frequency is taken from [Salah, 1993] and given by

\[ v_{m,\text{O}^+} = 6.82 \times 10^{-16} [N_2] + 6.66 \times 10^{-16} [O_2] \\
+ 3.42 \times 10^{-17} [O] \sqrt{T} \left( 1.08 - 0.139 \cdot \log_{10} T + 4.51 \cdot 10^{-3} (\log_{10} T)^2 \right) \]  

The collision frequencies in expressions (III.47) and (III.48) are given in S.I. units and number densities are thus inserted in units of \([\text{m}^3]\). The collision frequency for molecular ions (III.47) is correct only for temperatures of around 800°K, and the variations of the term under lower temperatures is at present not considered in CTIM.

In summary, therefore, the ion drag term in the momentum equation (III.17) can be calculated using relations (III.37) and (III.37). The conductivity tensor (III.40) is calculated at every time
step by using the electron density.

It is important to note that the thermosphere code considers the horizontal ion movement only in its ion-neutral coupling calculations. Although in the ionosphere charged particles do move along the field lines the vertical acceleration of neutral particles by ions is not calculated. As shown in section III.3.3, the momentum equation used in the thermosphere model is for horizontal motion only, and a vertical momentum equation is not implemented. The reason for this is that quasi-hydrostatic equilibrium is assumed in the atmosphere. This fundamental assumption has been found to adequately describe the thermosphere [Rishbeth and Garriott, 1969] and would have to be given up when modelling vertical acceleration of neutral gas by ions moving along the magnetic field lines.

III. 3.8. MAGNETOSPHERIC FORCING

Along with solar heating and tidal forcing, a third important source of energy and momentum in the thermosphere and ionosphere is the magnetosphere. The term magnetosphere is used to describe the region in which the particle movement is controlled mainly by the Earth's magnetic field. The terrestrial magnetosphere starts at around 150 km altitude and is often considered to be the upper boundary of the Earth's atmosphere since it ranges furthest into Space. The motion of charged particles along magnetic field lines has been referred to in more detailed in section III.3.7. The bipolar terrestrial magnetic field originates mainly from within the Earth and is here assumed constant with time. The field strength is in the model given by

\[
B = 2.93 \cdot 10^{-5} (1 + 3 \sin^2 \theta)^{\frac{1}{2}} \quad [\text{Tesla}]
\]

at geomagnetic latitude \( \theta \). As the solar wind plasma blows across the magnetic field lines at high altitudes a horizontal electric field is set up which is mapped along the conducting magnetic field lines onto the high latitude ionosphere, causing a polar cross-cap potential of around 50 kV under average conditions. In addition to this high latitude electric convection field, charged particles of high energy from the solar wind precipitate into the atmosphere at the poles. The magnetosphere thus provides the thermosphere and ionosphere with energy in two ways. The convection electric field drives electric currents which cause Joule-heating in the thermosphere and the precipitating particles collide with atmosphere particles, causing transfer of momentum and energy. Particle
precipitation and the convection field are both considered in the thermosphere code by using statistical models based on measurements. In principle, any such statistical model can be used, and at present two options are available. The electric field models used are either by Foster et al. [1986] or Heppner and Maynard [1987]. These are used in conjunction with particle precipitation models by Fuller-Rowell and Evans [1987] or Hardy et al. [1985]. The former is based on the TIROS/NOAA auroral particle measurements and the latter on DMSP observations. Both allow specification of the precipitation index which is related to the magnetospheric $K_p$ index, ranging from magnetically quiet ($K_p = 1$-) to disturbed conditions ($K_p = 6$-). The best match of electric field and particle precipitation pattern is obtained when using the electric field by Foster et al. in conjunction with the precipitation model by Fuller-Rowell and Evans, or the model by Heppner and Maynard with that by Hardy et al. The electric fields are assumed constant with height, and furthermore the feedback of moving ions, which in reality cause a continuous change of the electric field, is ignored. All the CTIM simulations presented in this thesis use the electric field model by Foster et al. and the particle precipitation model by Fuller-Rowell and Evans.

The above mentioned electric field models describe the convection field at high latitudes only. Furthermore, the dynamo action of thermospheric winds (see Chapter 1, section 1.2.4) generates an electric field at mid- to low latitudes [Evans, 1978; Richmond, 1979b]. Unless the dynamo-mechanism is modelled self-consistently, this field needs to be implemented externally in parameterized form. An example for an empirical model of the dynamo electric field is that of Richmond et al. [1980]. In all simulations carried out here, this model is used to define the low-latitude dynamo electric field. The field model by Richmond et al. is also used in the current version of the self-consistent plasmasphere code (see III.2), causing zonal drift of the magnetic flux tubes at low latitudes.

III. 3. 9. NUMERICAL METHODS

In order to solve numerically the above non-linear equations of momentum [(III.18), (III.19)], energy (III.22) and continuity (III.26) on the global grid used in the model, numerical techniques are used which calculate the differentials and time integration. The techniques used are briefly summarized here. A more comprehensive review is given by [Fuller-Rowell, 1981].
i) DIFFERENTIAL TECHNIQUES

The first- and second-order derivatives are evaluated numerically by finite-difference two-point centred approximations. The first-order derivative of a variable $f$ is given by

$$
\left( \frac{\partial f_j}{\partial \alpha_x} \right)_\Delta = \frac{f_{j+1} - f_{j-1}}{2\Delta}
$$

(III.50)

where $f_{j-1}, f_j, f_{j+1}$ are values of the variable $f$ on the grid points along a mesh $1 \leq j \leq N$ and $\Delta$ is the mesh step length. In the model, values of $f_j$ are located on the same pressure level along lines of constant latitude (for longitude derivatives) or lines of constant longitude (for latitude derivatives).

For vertical ($p$-) derivatives the points are located along the vertical (pressure-) axis at fixed latitude and longitude. The second-order derivative is expressed by

$$
\left( \frac{\partial^2 f_j}{\partial \alpha_x^2} \right)_\Delta = \frac{f_{j+1} - 2f_j + f_{j-1}}{\Delta^2}
$$

(III.51)

The accuracy of these finite differential approximations may be estimated by evaluating the difference between using this scheme and analytically differentiating a Fourier mode,

$$
A = A_0 e^{ika_x}
$$

(III.52)

where $A_0$ is the amplitude, $a_x$ the dependent variable and $k$ the wavenumber. Applying (III.50) and (III.51) to (III.52) one can derive that

$$
\left( \frac{\partial^2 A_j}{\partial \alpha_x^2} \right)_\Delta = \left[ 1 - \frac{k^2 \Delta^2}{6} + O(k^4 \Delta^4) \right] \cdot \frac{\partial A}{\partial \alpha_x}
$$

(III.53)

and

90
\[
\left( \frac{\partial^2 A}{\partial x^2} \right)_\Delta = \left[ 1 - \frac{k^2}{12} + O(k^4) \right] \cdot \frac{\partial A}{\partial x^2} \quad (\text{III.54})
\]

The accuracy therefore depends on the grid resolution, \( \Delta \), and the Fourier wavenumber, \( k \). The latter determines the degree of change of the parameter along the distance, \( \Delta \). If the parameter \( A \) varies considerably over a distance of the order of \( \Delta \), its Fourier transform will consist of modes with large wavenumbers \( k \), implying a larger error in the above numerical techniques. It is of interest here to estimate the accuracy of the numerical techniques in longitudinal direction for diurnal and semidiurnal tidal variations. The longitudinal resolution is 18°, giving a total of 20 local-time gridpoints per day, and therefore one may write for the diurnal and semidiurnal wavenumbers \( k_{\text{diu}} \) and \( k_{\text{sem}} \), respectively

\[
k_{\text{diu}} = \frac{2 \pi}{20 \Delta} = \frac{\pi}{10 \Delta} \quad (\text{III.55})
\]

\[
k_{\text{sem}} = \frac{2 \pi}{10 \Delta} = \frac{\pi}{5 \Delta} \quad (\text{III.56})
\]

which, inserted into (\text{III.53}) and (\text{III.54}) give the following numerical errors:

<table>
<thead>
<tr>
<th></th>
<th>first derivative</th>
<th>second derivative</th>
</tr>
</thead>
<tbody>
<tr>
<td>diurnal oscillation</td>
<td>&lt; 2 %</td>
<td>&lt; 1 %</td>
</tr>
<tr>
<td>semidiurnal oscillation</td>
<td>&lt; 7 %</td>
<td>&lt; 3.5 %</td>
</tr>
</tbody>
</table>

\textbf{Table (III.6)}: \textit{Numerical errors introduced by the numerical differentiation in CTIM.}

Values in \textbf{Table (III.6)} shows that the errors are all smaller than 7\% in longitudinal direction. For semidiurnal winds of 40 m/s amplitude the largest error is therefore around 3 m/s. In order to estimate numerical errors for a particular variable it needs to be Fourier-decomposed and the wavenumber of relevant components inserted into (\text{III.53}) and (\text{III.54}) along with the resolution \( \Delta \) in the direction of the derivative. In order to maintain numerical errors of the first derivative
below 10% the resolution needs to satisfy the following condition, where $k$ is the wavenumber of the feature to be resolved,

$$\Delta < \frac{0.8}{k}$$  \hspace{1cm} (III.57)

which was derived from the above expressions.

\textbf{ii) INTEGRATION METHOD}

While the fundamental differential equations used in the model describe the changes of velocity, energy and mass, they give no information about the absolute value of a parameter at a particular time. In order to obtain this, it is necessary to integrate them numerically. In practice, the model starts a run at time $t = 0$ with initial values of all relevant parameters which in most cases are given by output from a previous run. In order to calculate parameters at a time step $t$ the model uses values from time step $t-1$ and carries out a numerical integration which is described here briefly.

The momentum-[(III.19), (III.18)] and energy (III.22) equations are time-dependent differential equations of the form

$$\frac{\partial A(t)}{\partial t} = L \cdot A(t)$$  \hspace{1cm} (III.58)

where $L$ is a spatial differential operator with first- and second-order derivatives acting on the parameter $A$. One may integrate (III.58) over a small time step, $\Delta t$, and obtains for $A$ at time $t+\Delta t$

$$A(t + \Delta t) = A(t) + \int_{t}^{t+\Delta t} L \cdot A(t') \, dt'$$  \hspace{1cm} (III.59)

with $t \leq t' \leq t+\Delta t$. Since $A(t')$ is not known for all $t'$ the integral on the right side needs to be approximated. The simplest approximation to the integral in (III.59) is given by the Euler method

$$A(t + \Delta t) = A(t) + L \cdot A(t) \cdot \Delta t$$  \hspace{1cm} (III.60)

However, this method proves to be unstable for differential equations with first-order spatial derivatives [Fuller-Rowell, 1981], such as the one-dimensional advective equation
The M odel

Chapter III

\[
\frac{\partial u}{\partial t} + v \frac{\partial u}{\partial x} = 0 \tag{III.61}
\]

where \(u\) and \(v\) are components of horizontal velocities. In contrast, the Euler method is stable for equations with second-order spatial differentials, such as the diffusion equation

\[
\frac{\partial u}{\partial t} - K \frac{\partial^2 u}{\partial x^2} = 0 \tag{III.62}
\]

for time steps smaller than

\[\Delta t \leq 0.5 \cdot \frac{\Delta^2}{K} \tag{III.63}\]

where \(\Delta\) is the mesh length in direction of \(x\) and \(K\), in this particular case, the diffusion coefficient. The reason for the Eulerian approach to be unstable in equations with first-order spatial derivatives is that the time derivative is not time centred. Time centering techniques (such as the Leapfrog method or the Lax scheme) cannot be used here since they cause instabilities in the second-order spatial derivatives. Furthermore, although not being an important issue with today's computing capacities, time-centering techniques require the storage of parameters from two previous time steps. Fuller-Rowell [1981] derived a scheme which overcomes the problems of time-centering for the first spatial derivative, instabilities for the second derivative and storage of parameters from more than one previous time step. The integration method used in the model is based on the Euler-method (III.60), but also considers a numerical diffusion term which is added to the solution

\[
A(t + \Delta t) = A(t) + L \cdot A(t) \cdot \Delta t - \frac{1}{2} \Delta t (\vec{V} \cdot \vec{V}) (\vec{V} \cdot \vec{V}) A(t) \tag{III.64}
\]

With this simple yet powerful numerical integration scheme only the values from one previous time step need to be stored.

III. 4. THE IONOSPHERE MODEL

All numerical simulations presented in this thesis are carried out with the Coupled Thermosphere-Ionosphere Model (CTIM, as described in III.2). However, the thesis will focus on the neutral gas response to tidal forcing, given that the full neutral-ion coupling is at present not treated self-
consistently, particularly at low latitudes. Therefore, only the main properties of the ionosphere code are outlined in this section. More comprehensive descriptions can be found in [Fuller-Rowell et al., 1996; Fuller-Rowell et al., 1987 and Quegan et al., 1982].

The high- to mid latitude ionospheric convection model used was developed independently by Quegan et al. [1982] and later coupled to the thermosphere code [Fuller-Rowell et al., 1987]. The two models use different coordinate systems. While the thermosphere performs calculations on an Eulerian frame, the ionosphere code uses a Lagrangian system which has been modified by Fuller-Rowell et al. [1988] to be more compatible with the thermospheric frame. The Lagrangian frame of reference is fixed relative to the Sun-Earth axis and therefore fixed relative to the high-latitude electric convection potential, which simplifies calculations considerably. The height range of the ionospheric model is 100 to around 10000 km. Since the thermosphere model has a height range of 80 to around 400 km, the neutral gas parameters fed into the ionospheric code between 400 and 10000 km height are not calculated self-consistently, but upper boundary values from the thermosphere model are used. This, however, is a valid approximation since the ionosphere at those heights is hardly influenced by the neutral gases.

The main constituents considered in the ionosphere model are the atomic ions H\(^+\) and O\(^+\) and, between 100 and 400 km, N\(_2\)\(^+\), O\(_2\)\(^+\), NO\(^+\) as well as N\(^+\). The high-latitude particle precipitation and electric convection field are the same as those used in the thermosphere model and described in section III.3.8.

### III. 4.1. ION CONTINUITY EQUATION

In the ionosphere above around 150 km altitude, ion velocities may to a good approximation be divided into two perpendicular components, one being parallel to the magnetic field lines and the other perpendicular to the magnetic and electric field. Denoting the ion velocity of species \(i\) with \(U_i\) and the parallel and perpendicular components with \(U_i^\parallel\) and \(U_i^\perp\), respectively, one may thus write

\[
\vec{U}_i = \vec{U}_i^\parallel + \vec{U}_i^\perp
\]  

(III.65)

where the perpendicular component is given by
The ion continuity equation is therefore with (III.65) and (III.66) written as

\[
\frac{dn_i}{dt} = P_i - L_i - n_i \nabla \cdot \vec{U}_i - \nabla \cdot (n_i \vec{U}_i) \quad (III.67)
\]

where \( n_i \) is the number density of species \( i \), \( P_i \) and \( L_i \) are production- and loss rates, respectively, and the time-derivative is given by

\[
\frac{d}{dt} = \frac{\partial}{\partial t} + \vec{U}_i \cdot \vec{\nabla} \quad (III.68)
\]

### III. 4. 2. EQUATION OF ION MOMENTUM

The ion velocity component parallel to the magnetic field lines arises from diffusion of ions under the influence of gravity, the partial pressure gradient, thermal diffusion and ion-neutral collisional drag. The momentum equation along magnetic field lines is, following [Conrad and Schunk, 1979, and St.-Maurice and Schunk, 1977], given by

\[
U_i \parallel = h_{ij} U_j \parallel + h_{in} V \parallel - D_i \left( \frac{1}{n_i} \frac{\partial n_i}{\partial s} - \frac{m_i g \parallel}{kT_i} + \frac{1}{T_i} \frac{\partial}{\partial s} \left( T_i + T_e \right) \right)
+ D_i \left( \frac{T_e}{T_i} \frac{1}{n_e} \frac{\partial n_e}{\partial s} + \frac{1}{T_i} \left[ \beta_{ij} \frac{\partial T_i}{\partial s} - \beta_{ij}^* \frac{\partial T_e}{\partial s} \right] \right) \quad (III.69)
\]

with \( h_{ij} \) and \( h_{in} \) being the drag coefficients between the ion species \( i \) and \( j \) and the neutral particles \( n, V \parallel \) is the neutral velocity along the field lines, \( D_i \) is the diffusion coefficient for \( i \), \( m \) the ion mass, \( g \parallel \) is the field-aligned gravity component, \( T_i \) and \( T_e \) the ion- and electron temperatures, respectively and \( \beta_{ij} \) and \( \beta_{ij}^* \) are dimensionless coefficients of thermal diffusion. The variable \( s \) denotes the angular (arc-) length along field lines.

The ionosphere model solves equations (III.67) and (III.69) as a pair of coupled first-order differential equations, using numerical techniques described in section III.3.9 and thus obtains the densities and field-aligned velocities for \( O^+ \) and \( H^+ \).
III. 4. 3. ION CHEMISTRY

While the ionosphere code calculates properties for the main constituents $O^+$ and $H^+$ over its entire height range (around 100-10000 km), molecular ions densities are calculated up to around 400 km, thus enabling fully self-consistent ion-ion and ion-neutral chemistry. The main reactions and rates are given in [Fuller-Rowell, 1993] and listed in Table (III.7).

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Rate constant</th>
</tr>
</thead>
<tbody>
<tr>
<td>$NO^- + e^- \rightarrow N(^3D) + O$ (80%)</td>
<td>$af_1 = 4.2 \cdot 10^{-13} (300/T_e)^{0.85} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$O_2^+ + e^- \rightarrow O + O$</td>
<td>$af_2 = 1.6 \cdot 10^{-13} (300/T_e)^{0.55} \text{ m}^3\text{s}^{-1}$ ... $T_e &gt; 1200^\circ\text{K}$</td>
</tr>
<tr>
<td>$N_2^- + e^- \rightarrow N(^4S) + N(^4S)$ (10%)</td>
<td>$af_3 = 1.8 \cdot 10^{-13} (T_e/300)^{-0.39} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$O^+ + N_2 \rightarrow NO^+ + N(^4S)$</td>
<td>$k_3 = 1.533 \cdot 10^{-18} - 5.92 \cdot 10^{-19} (T_e/300)$ + $8.6 \cdot 10^{-20} (T_e/300)^2 \text{ m}^3\text{s}^{-1}$ ... $300^\circ\text{K} &lt; T_e &lt; 1700^\circ\text{K}$</td>
</tr>
<tr>
<td>$N_2^+ + O \rightarrow N(^2D) + NO^+$</td>
<td>$k_4 = 1.4 \cdot 10^{-16} (300/T_R)^{0.44} \text{ m}^3\text{s}^{-1}$ ... $T_R &lt; 1500^\circ\text{K}$</td>
</tr>
<tr>
<td>$N^+ + O_2 \rightarrow NO^+ + O$</td>
<td>$k_5 = 4.0 \cdot 10^{-16} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$O_2^+ + N(^4S) \rightarrow NO^+ + O$</td>
<td>$k_6 = 1.2 \cdot 10^{-16} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$O_2^+ + NO \rightarrow NO^+ + O_2$</td>
<td>$k_7 = 4.4 \cdot 10^{-16} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$O^+ + O_2 \rightarrow O_2^+ + O$</td>
<td>$k_8 = 2.82 \cdot 10^{-17} - 7.74 \cdot 10^{-18} (T_e/300)$ + $1.073 \cdot 10^{-18} (T_e/300)^2$ - $5.17 \cdot 10^{-20} (T_e/300)^3$ + $9.65 \cdot 10^{-22} (T_e/300)^4 \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$N^+ + O_2 \rightarrow O_2^+ + N(^4S)$</td>
<td>$k_{10} = 4.0 \cdot 10^{-16} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$N^+ + O \rightarrow O^+ + N$</td>
<td>$k_{15} = 1.0 \cdot 10^{-18} \text{ m}^3\text{s}^{-1}$</td>
</tr>
<tr>
<td>$NO + h\nu_{Lyman \alpha} \rightarrow NO^+ + e^-$</td>
<td>see [Fuller-Rowell, 1993]</td>
</tr>
</tbody>
</table>

Table (III.7) Ion-neutral reactions in the model. Rate constant names are identical to those used in the code. $T_i$ is the ion temperature, $T_e$ the electron temperature, $T_n$ the neutral temperature, $T_1 = 0.667 T_i + 0.333 T_n$, $T_2 = 0.6363 T_i + 0.3637 T_n$ and $T_R = (T_1 + T_n)/2$. 

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All calculations assume chemical equilibrium, and vertical profiles of $N(\text{^2 D})$, $N(\text{^4 S})$ and NO assume vertical diffusion and chemistry to be balanced at each location.

III. 5. THE NEW LOWER BOUNDARY

The thermosphere model's lower boundary is set at an average height of 80 km. This implies that the chemistry of the stratosphere and mesosphere cannot be calculated. Therefore, tides generated in these regions (see Figure I.1) which propagate upwards into the thermosphere need to be implemented through external forcing. In a previous attempt [Parish et al., 1990], this forcing was implemented at pressure level 4 (around 96 km height) and proven successful for many studies above 130 km height. Below this height, however, an inconsistency in the implementation caused results to be unreliable. Future ambitions to couple CTIM to a middle atmosphere model not only emphasize the need to model more accurately tides in the 80-130 km height regime but also to make the model's lower boundary more flexible. Therefore, tidal forcing in CTIM has been moved from level 4 to the lower boundary and, in addition, the method of forcing has been improved to avoid the earlier inconsistencies. In Chapter VII, further expressions are presented which were used there to implement planetary wave forcing at CTIM's lower boundary.

When implementing tidal forcing externally, it is necessary to simultaneously force geopotential height, winds and temperature to oscillate and ensure that they form a self-consistent set of parameters. Once the forcing has occurred at the lower boundary, their self-consistency is automatically maintained at higher levels.

In practice, any set of realistic geopotential height-, temperature- and wind values with tidal oscillations can be used at CTIM's lower boundary if self-consistency of the parameters is ensured. This can be obtained either numerically or analytically. The approach presented here is analytical and the parameters are linked through equations from Classical Tidal Theory which was outlined in Chapter II. While equations for the global structure of geopotential height and winds were already derived there, the equation for temperature is derived in III.5.3. In Chapter VII, an example is shown of a study where CTIM's lower boundary is forced not through the analytical equations derived here, but using a set of self-consistent wind-, temperature- and pressure oscillations which were provided from another numerical model. When coupling the CTIM to a middle atmosphere model in future, the analytical approach would be replaced in a similar manner by an externally
given set of parameters.

III. 5.1. GEOPOTENTIAL HEIGHT

In order to simulate pressure oscillations at CTIM's lower boundary, the geopotential height of that pressure level is perturbed diurnally and semidiurnally. As derived in Chapter II, section II.3, the total geopotential perturbation $\Phi(\theta, t)$ due to tides is at a fixed altitude given by the real component of

$$
\Phi(\theta, t) = \sum_s \Phi_s = g \sum_s \sum_n \left[ Z_{s,n}^0 \cdot \Theta_{s,n}(\theta) \cdot e^{i(\omega_s t + \varphi_s + \phi_{s,n})} \right]
$$

where, as previously, $\omega_s$ is the tidal frequency, $t$ is the universal time, $\theta$ the longitude, $\varphi$ the latitude and $g$ is the gravity acceleration, taken as $g=9.50 \text{ ms}^{-2}$. $\Theta_{s,n}(\theta)$ are the normalized Hough functions for modes $(s,n)$ (see also Figure II.2). In the model they are used in parameterized form which is based on the equations given by Chapman and Lindzen [1970]. $Z_{s,n}^0$ denote the amplitudes (in dimension of height) which are specified separately for each $(s,n)$ mode. The phase term, $\phi_{s,n}$, is specified for each mode and the value inserted determines the local time of maximum of the mode $(s,n)$. In (III.70), the summation over $s$ allows the use of diurnal ($s=1$) and semidiurnal ($s=2$) modes and the summation over $n$ allows the use of a variety of latitudinal Hough modes.

In the model, therefore, the geopotential height is perturbed around its mean value of 80 km by the value of $\Delta z(\theta, t)$, given by

$$
\Delta z(\theta, t) = \sum_s \sum_n \left[ Z_{s,n}^0 \cdot \Theta_{s,n}(\theta) \cdot \cos (\omega_s t + \varphi_s + \phi_{s,n}) \right]
$$

The values used for the amplitudes, $Z_{s,n}^0$, at the lower boundary typically lie around 50-400 m, corresponding to pressure oscillations by up to around 0.06 Pa, or 4%. The inserted amplitude and phase values for each mode are either given by simulations from other middle atmosphere models [eg. Hagan et al., 1993; Forbes and Vial, 1989] or extracted from measurements, as described in more detail in Chapter V. While the Hough modes, $\Theta_{s,n}(\theta)$, determine the latitudinal structure of the oscillations, the cosine-term on the right side of equation (III.71) defines its zonal structure.
Typical global perturbation profiles are shown in Chapter IV. It is important to note that the expression given in (III.71) is for one specific height only, at the lower boundary. Although tidal theory describes a vertical dependence of the perturbations, as described in II.4.4.ii) for internal waves, this vertical structure does not need to be implemented here since the model calculates the vertical progression self-consistently. Therefore, the factor $e^{i\omega t}$ found in II.4.4.ii) is not used here.

III.5.2. HORIZONTAL WINDS

The horizontal wind equations used at CTIM's new lower boundary are based on equations (II.23) and (II.24) in Chapter II. Taking the real parts of the complex values given there, the southward ($u$) and eastward ($v$) winds are specified at the lower boundary as

\[
\begin{align*}
\nu(\theta, t) &= \frac{g}{R_E} \sum_s \sum_n \left( \frac{1}{\omega_s \left(1 - \frac{f^2}{\omega_s^2}\right)} \cdot Z_{s,n}^0 \right. \\
& \quad \left. \cdot \left[ \frac{sf}{\omega_s \cos \theta} \cdot \Theta_{s,n}(\theta) - \frac{\partial \Theta_{s,n}(\theta)}{\partial \theta} \right] \cdot \sin \left( \omega_s t + s \varphi + \phi_{s,n} \right) \right) \\
\end{align*}
\]

\[
\begin{align*}
\nu(\theta, t) &= \frac{g}{R_E} \sum_s \sum_n \left( \frac{1}{\omega_s \left(1 - \frac{f^2}{\omega_s^2}\right)} \cdot Z_{s,n}^0 \right. \\
& \quad \left. \cdot \left[ \frac{f}{\omega_s} \cdot \frac{\partial \Theta_{s,n}(\theta)}{\partial \theta} - \frac{s}{\cos \theta} \cdot \Theta_{s,n}(\theta) \right] \cdot \cos \left( \omega_s t + s \varphi + \phi_{s,n} \right) \right) \\
\end{align*}
\]

where, as previously, $f=2\Omega \sin \theta$ is the Coriolis term at latitude $\theta$ and $R_E$ the Earth's radius. Typical diurnal and semidiurnal wind amplitudes at 80 km range between 10-50 m/s. The background wind is globally set to zero.

Since Hough mode values $\Theta_{s,n}(\theta)$ are given in parameterized form in the model, the derivative terms $\frac{\partial \Theta_{s,n}(\theta)}{\partial \theta}$ need to be treated with caution. In some instances, the derivatives of the (1,1) mode Hough functions showed "spikes" which lead to unstable behaviour of the model. These were
found at latitudes where the meridional gradient of the Hough function was small. Since derivatives are taken numerically, they rely on the precision of the parameter values on CTIM's grid points. In regions where these parameters change very little, gradients can fluctuate between small positive and small negative values, an error which at first instance is negligible. If however this value is used in other calculations, such as here for the winds, the effect of these marginal fluctuations can be significant. Therefore, the horizontal velocities in (III.72) and (III.73) are smoothed numerically, using a technique which calculates the Fast Fourier Transforms (FFT's) of the parameterized Hough modes. The routine used here is based on that given by Press et al. [1992].

III. 5.3. TEMPERATURE

An expression for temperature perturbations will be derived in the following and is based on the results and assumptions from Classical Tidal Theory. The horizontal winds, geopotential height perturbations described above and the temperature perturbations derived here form a self-consistent set of parameters.

In Chapter II, equation (II.29), the dimensionless vertical variable \( x \) was defined as \( x = -\ln(p/p_0) \). This gives:

\[
\frac{\partial x}{\partial z} = -\frac{\partial p}{p} \tag{III.74}
\]

One may use (III.74) to substitute \( \partial p \) in the hydrostatic equation (III.1) and furthermore the ideal gas law (II.1) to substitute \( p \) and obtains

\[
\frac{\partial z}{\partial x} = \frac{R_{\text{air}} T}{g} \tag{III.75}
\]

where, as previously, \( R_{\text{air}} \) is the gas constant for dry air \( (R_{\text{air}} = 287.05 \text{ J K}^{-1} \text{ (kg)}^{-1}) \) and \( z \) is the geographic height coordinate. In particular, \( z \) may be regarded as the height of a constant pressure level. If \( \Delta z \) is a periodic (tidal-) perturbation of the pressure level height, \( z \), then (III.75) may be rewritten as

\[
\frac{\partial (\Delta z)}{\partial x} = \frac{R_{\text{air}} \Delta T}{g} \tag{III.76}
\]
The tidal waves considered here are propagating (internal) waves only, and therefore the vertical dependence of $\Delta z$ is given by $\Delta z \propto e^{i\omega x}$, as derived in II.4.ii). One may thus write for the left term in (III.76):

$$\frac{\partial (\Delta z)}{\partial x} = i \alpha (\Delta z) \quad (III.77)$$

where

$$\alpha = \sqrt{\frac{k H}{h_n} - \frac{1}{4}} \quad (III.78)$$

with $k = \frac{R_{air}}{c_p} = \frac{2}{7}$, $H$ being the scale height and $h_n$, the equivalent depth. Rewriting (III.76) gives

$$\Delta T(\theta, t) = \frac{g}{R_{air}} i \alpha \cdot \Delta z(\theta, t) \quad (III.79)$$

which is a relationship between temperature ($\Delta T$) - and geopotential height ($\Delta z$) perturbations. Substituting $\Delta z(\theta, t)$ from (III.70) finally gives

$$\Delta T(\theta, t) = - \frac{g}{R_{air}} \sum \sum \left[ \alpha \cdot Z_{r,n}^0 \cdot \Theta_{r,n}(\theta) \cdot \sin(\omega t + s\varphi + \phi_{r,n}) \right] \quad (III.80)$$

which is the equation used at the lower boundary. The equivalent depth values, $h_n$, used in the code are taken from Chapman and Lindzen [1970] and shown in Table II.1.

It is shown in Chapter VI that temperature perturbations on a pressure level are generated by horizontal wind gradients which cause up-and downwelling relative to pressure levels, resulting in adiabatic heating and cooling. The expression for temperature perturbations at the lower boundary as derived above intrinsically considers this effect. Although wind gradients are not explicitly used in (III.80), the temperature expression has in common with the horizontal wind expressions (III.72) and (III.73) the geopotential height perturbation term, $\Delta z (\theta, t)$, which couples the parameters. The expression for the geopotential height perturbation term (III.71) is determined by the derivations in Chapter II which use the coupled energy- and momentum equations of the atmospheric gas, implying that all of the physical principles governing the atmosphere under the
idealized conditions outlined in II.2 are automatically considered in the expressions at the model's lower boundary.
CHAPTER IV. PROPAGATING HOUGH MODES

ABSTRACT

This and following chapters analyse output from the Coupled Thermosphere Ionosphere Model (CTIM) in order to validate the code and give a deeper understanding of the effects of upwards propagating tides on the dynamics, energy and composition of the thermosphere. A series of test runs are presented which differ in the applied external tidal forcing. The effects of individual tidal Hough modes on dynamics and energy of the thermosphere are examined by analyzing the terms in the equations of momentum and energy. A distinction is made between the tidal oscillations generated in-situ, which are found to be mostly non-propagating (or evanescent), and those forced externally at the lower boundary. In the real atmosphere, the latter are those propagating into the thermosphere from the lower- and middle atmosphere. In-situ winds at low- to mid latitudes are found to be forced diurnally by the pressure gradients and Coriolis forces, with ion drag playing a role only during daytime at low-to mid latitudes and partly also at nighttime for high latitudes, where fast moving auroral ions drag the neutrals and generate mainly semidiurnal tidal perturbations. Horizontal- and vertical advection as well as vertical viscosity also play an important role. In-situ temperature oscillations are found to be caused at high latitudes by Joule heating and at low latitudes by UV- and EUV absorption. Temperature oscillations under tidal forcing are generated by adiabatic heating- and cooling which result from up- and down welling relative to pressure levels. Variations of density and vertical winds are, through changed solar absorption and vertical advection, found to affect the temperatures as well in regions where the upwards propagating tides dissipate and induced changes of background winds and temperature. The influence of tides on the background atmosphere does not occur isotropically in both horizontal directions. In the case of the (1,1) mode a westward jet is generated at around 120 km height while in meridional direction the momentum is released mainly in the vicinity of 160 km. The non-linearities are in both cases caused mainly by the vertical and horizontal advection as well as vertical viscosity. CTIM successfully simulates the effect of mode coupling, with higher order Hough modes being generated in regions of dissipation.
IV. 1. THE FORCING FUNCTIONS

The tidal forcing implemented at CTIM’s lower boundary, as described in Chapters II and III, allows the choice of five Hough functions, the diurnal (1,1) mode and the semidiurnal (2,2), (2,3), (2,4) and (2,5) modes, which can be applied individually as well as in any combination in order to produce a global oscillation profile. Other higher order modes have not been implemented since CTIM’s current vertical resolution of one scale height is insufficient for their shorter vertical wavelengths. Since only the propagating Hough modes are implemented externally, the lower boundary forcing approaches zero towards the poles and occurs mainly within ±60° latitude. Any of the non-propagating tidal modes occurring in the thermosphere, are generated in-situ in the model, particularly at high latitudes under the auroral forcing, and need no consideration at its lower boundary.

IV. 1.1. THE DIURNAL (1,1) MODE

The (1,1) mode is the first symmetric propagating diurnal Hough mode. With its vertical wavelength of around 25 km (see Table II.1) it is the most important diurnal mode propagating into the thermosphere. Other propagating diurnal modes have smaller vertical wavelengths and tend to undergo destructive interference within the region where they are generated, the stratospheric ozone layer. In contrast, the (1,1) mode propagating into the thermosphere is generated by water vapour absorption in the troposphere within a layer which is thin enough for destructive interferences not to take place [Forbes, 1995]. The latitudinal profile of geopotential height amplitudes produced by the (1,1) Hough mode is given in Figure II.2 and leads to a global oscillation profile at pressure level 1 as shown in Figure IV.1.a (top). In the global profile of Figure IV.1.a, an amplitude of 100 m and local time of maximum at 12 h were chosen. Since the model uses a pressure coordinate system, oscillations of pressure are expressed in terms of oscillations of the geopotential height of a pressure level. In the remainder of this chapter, therefore, the terms geopotential height- and pressure oscillations refer to the same physical process.

Main characteristics of the (1,1) mode are the equatorial symmetry, with geopotential height (or pressure-) peaks occurring on the geographic equator and weaker local extrema at ±30° latitude.
Propagating Hough Modes

Chapter IV

The strongest pressure gradients imposed by the (1,1) mode therefore also occur at low latitudes, near ±20°. Figures IV.1.a (bottom) and b show the resulting temperature- and horizontal wind oscillations, respectively, for the same pressure level, as calculated at CTIM's lower boundary, using the relations from Classical Tidal Theory given in Chapter III, section III.3.5.

Temperature oscillations (Figure IV.1.a, bottom) are applied to a globally constant background value of 192°K and the plot thus suggests that a geopotential height amplitude of 100 m produces diurnal temperature variations of around 5K. The latitudinal temperature oscillation pattern is very similar to the geopotential height oscillation pattern, but shifted backwards in phase by 6 hours. The temperature oscillations are a result of adiabatic heating and cooling which is caused by the up- and down welling relative to the fixed pressure level due to horizontal wind gradients (see also Chapter VI).

The horizontal wind patterns (Figure IV.1.b) are produced by the pressure- and Coriolis forces. The zonal wind extrema occur near ±22°latitude which is where the Coriolis- and pressure forces balance each other, giving a geostrophic wind flow along the isobars. Equatorward of these extrema, the Coriolis force decreases and winds are determined mainly by the pressure gradients. Although the Coriolis term is important above ±22°latitude, the meridional pressure gradients are too weak to drive significant winds. Since also the zonal pressure gradient is weak in those regions, both the diurnal meridional- and zonal winds virtually disappear poleward of around ±50°latitude. The meridional wind pattern is antisymmetric to the equator since geostrophic winds have opposite directions in the different hemispheres, being clockwise around pressure peaks in the northern- and anticlockwise in the southern hemisphere. As mentioned earlier, the vertical wavelength of the (1,1) mode is around 25 km in the 80-100 km height regime.

IV. 1.2. THE SEMIDIURNAL MODES

Semidiurnal modes are the most important tides propagating into the lower thermosphere since their vertical wavelengths are larger than those of any of the diurnal modes, implying that the latter are damped more efficiently (see also Chapter V). Absorption by water vapour in the troposphere and ozone in the stratosphere both generate, in addition to diurnal tides, primarily the (2,2) mode. Higher order semidiurnal modes are insignificant below around 60 km altitude [Forbes, 1982b]. As the (2,2) mode propagates upwards, its exponential amplitude growth (due to the density
decrease with height) is damped by *mode coupling*, an interaction with the background wind which generates higher order modes (see also section IV.6.2), as well as by other friction forces. It was suggested by *Forbes* [1982b] that the global semidiurnal tidal profile in the 70-90 km height region consists of almost equal contributions from the (2,2), (2,3) and (2,4) modes. As pointed out in Chapter II, section II.4.4.i, sufficiently negative vertical gradients of scale height, \( \frac{dH}{dx} \), may lead to the (2,2) mode becoming evanescent over a limited height range near the mesopause. The occurrence of a region of evanescence leads to considerable damping of the (2,2) mode as it propagates upwards into the thermosphere.

The geopotential height-, temperature- and horizontal wind perturbations generated by the first four semidiurnal propagating modes are shown in *Figures IV.2.-5.*. All patterns are based on the same geopotential height amplitude of 200 m, with the local time of maxima set to 12.0 h. As for the diurnal (1,1) mode, temperature profiles are a result of mainly adiabatic heating and wind profiles follow from pressure gradients and the Coriolis force. The plots show that the complexity of patterns increases with the mode order. Furthermore, the latitudinal extent of the forcing becomes broader for higher order modes, with strong wind patterns under the (2,5) mode being generated at up to around \( \pm 70^\circ \) latitude, in contrast to the diurnal (2,2) mode where forcing is limited to a narrower band around the equator. The vertical wavelengths of semidiurnal modes become smaller for higher orders and values for the (2,2), (2,3), (2,4) and (2,5) modes expected from tidal theory are around 1600 km, 99 km, 46 km and 38 km, respectively. These values are larger than that of the (1,1) mode, which is, as mentioned earlier, why the semidiurnal modes propagate to higher altitudes in the atmosphere. It is important to note, as pointed out already in Chapter II, that the above stated values for vertical wavelengths are based on equation (II.48) (see Chapter II), which assumes idealized conditions. In the real atmosphere the wavelengths become smaller with altitude as a result of damping processes. In particular, the (2,2) mode wavelength is in the thermosphere much smaller than stated, with a values of around 80 km. What remains unchanged, though, is the aspect that higher order modes have shorter vertical wavelengths.

The phase relationships between pressure-, temperature- and wind oscillations are as for the diurnal mode. Phase shifts relative to the pressure oscillations are -3 h and \( \pm 6 \) h for temperatures and zonal winds, respectively. Meridional wind phases are shifted by -3 h in the northern- and +3 h in the southern hemisphere. These values are given by the analytical expressions in III.3.5 and valid in the CTIM at its lower boundary only, while at higher altitudes they change considerably,
as shown in the simulations presented later. Temperature amplitudes increase with mode order from less than 1 °K for the (2,2) mode to around 6 °K for the (2,5) mode. So even the largest of these amplitudes are only around 3% of the lower boundary background temperature in CTIM. Horizontal winds equally increase with mode order, from around 5 m/s for the (2,2) mode to 23 m/s for the (2,5) mode, when using the same geopotential height amplitude of 200 m. These values merely indicate that the larger pressure gradients introduced by a more complex latitudinal pressure perturbation profile lead to a stronger temperature- and wind response, but they do not reflect the actual relative intensity of modes in the atmosphere at 80 km. A realistic balance of the modes is achieved by selecting different amplitudes for each of them, as done explicitly in Chapter V.

While only the basic structure and properties of diurnal and semidiurnal Hough modes used at CTIM’s lower boundary have been described so far, their effects on thermospheric dynamics and energy are investigated in the remainder of this chapter. The discussions address not only the perturbations caused by the lower boundary forcing but also the influences found in the background atmosphere.

IV. 2. THE MODEL RUNS AND PLOTS

The main purpose of the study presented below is to examine the atmosphere’s response to forcing with each of the Hough modes, and therefore the choice of input amplitudes and phases at the lower boundary was empirical, although within realistic boundaries. The runs are thus not intended to represent a measured situation; they are illustrative only. The aim is to understand the differences between the individual tidal modes. Model runs with realistic tidal forcing are presented in Chapter V and compared with results from other models as well as measurements.

All the simulations presented here are for medium solar activity conditions (F10.7=100) at low magnetic activity (Kp=2+). In order to study the upwards propagating modes in as many different conditions as possible, the series of simulations presented in Table IV.1 were carried out. Each has been given a name listed in the left column for easier reference in the following discussions.

All plots shown here are snapshots at 12 h Universal Time (UT). In order to better distinguish the externally forced tides from in-situ generated oscillations the following technique was used. All runs were carried out using one common initial input file which was adapted from a non-tidal run
by globally averaging parameter values on each pressure level, thus giving a globally uniform height profile only. In addition to the runs with tidal forcing a further simulation (NT) was carried out, using the same input file but leaving the tidal forcing switched off. Many of the plots presented here were taken as differences between the conditions with external forcing and the non-tidal simulation (NT). In that way, any in-situ forced oscillations could be eliminated and the tides propagating upwards as a result of lower boundary forcing could more easily be identified.

As shown in Table IV.1, all simulations are carried out for equinox conditions. It was initially intended to present results also for solstice conditions. The motivation for doing so was that the background atmosphere has a different latitudinal structure during solstice, giving also a different background wind structure which might ultimately influence the upwards propagating tides. It was found, though, that the oscillation profiles differed only marginally in both seasons when applying identical lower boundary forcing, and therefore only the equinox simulations are presented here.
### Table IV.1: Model runs carried out to study the influence of individual tidal modes in the thermosphere. All runs are for medium solar activity (F10.7=100) and low magnetic activity (Kp=2+). *Names are used in the text to refer to individual runs.*

<table>
<thead>
<tr>
<th>Name</th>
<th>Season</th>
<th>Forcing</th>
<th>Amplitude</th>
<th>Phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>NT</td>
<td>Equinox</td>
<td>none</td>
<td>/</td>
<td>/</td>
</tr>
<tr>
<td>11E</td>
<td>Equinox</td>
<td>(1,1)</td>
<td>100 m</td>
<td>12 h</td>
</tr>
<tr>
<td>22E</td>
<td>Equinox</td>
<td>(2,2)</td>
<td>200 m</td>
<td>12 h</td>
</tr>
<tr>
<td>23E</td>
<td>Equinox</td>
<td>(2,3)</td>
<td>200 m</td>
<td>12 h</td>
</tr>
<tr>
<td>24E</td>
<td>Equinox</td>
<td>(2,4)</td>
<td>200 m</td>
<td>12 h</td>
</tr>
<tr>
<td>25E</td>
<td>Equinox</td>
<td>(2,5)</td>
<td>200 m</td>
<td>12 h</td>
</tr>
<tr>
<td>2+4E</td>
<td>Equinox</td>
<td>(2,2)+(2,4)</td>
<td>200 m/200 m</td>
<td>12 h/ 12 h</td>
</tr>
</tbody>
</table>

#### IV. 3. THE THERMOSPHERIC RESPONSE TO LOWER BOUNDARY FORCING

This paragraph is divided into three main sections, of which the first describes the thermospheric oscillations which are generated in-situ, without any lower boundary forcing. The second and third section then discuss the thermosphere’s response to diurnal and semidiurnal forcing, respectively.

#### IV. 3.1. IN SITU OSCILLATIONS

The CTIM simulates self-consistently the physical processes of the atmosphere above 80 km. An important fraction of diurnal and semidiurnal oscillations found there are generated within that region (see Figure 1.1) and are thus intrinsically included in the model, without any lower boundary forcing. This section will examine the profiles of these in-situ generated oscillations in order to later distinguish them from those propagating upwards through the lower boundary.
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i) TEMPERATURE

The main energy source of the thermosphere is solar radiation which is absorbed primarily by atomic and molecular oxygen. Since the solar irradiance varies diurnally, this energy input is mostly diurnal but also leads to semi-diurnal oscillations being forced in-situ as sub-harmonics. In the auroral region, semi-diurnal oscillations are stronger due to the two-cell shape of the auroral ion flow pattern which is through ion-neutral coupling transferred onto the neutral atmosphere (see discussion below).

Figures IV.6 and IV.7 show diurnal and semi-diurnal amplitudes of temperature and winds in the thermosphere, as calculated by CTIM without lower boundary forcing (run NT). Figure IV.6 (top) shows that temperature varies diurnally by up to around 100° K above 200 km altitude but by less than 10° K below 150 km. Atomic oxygen is the main neutral constituent in the thermosphere above 150 km height and thus causes most of the observed diurnal heating through solar UV and XUV absorption. Small amplitudes below 150 km suggest that no considerable solar absorption takes place there.

Semi-diurnal amplitudes of temperature in Figure IV.7 (top) are generally smaller than the diurnal ones. A region of enhanced amplitudes is found at 190 km height between around ±50° latitude, with two additional peaks at 60°S and 70°N. The low- to mid latitude enhanced semi-diurnal amplitudes have peak values of around 20°K and in the model are found to be generated by the solar extreme ultraviolet (EUV) heating term (not shown here) which has its peak semi-diurnal amplitudes at the location of the enhanced temperature amplitudes in Figure IV.7. In the real atmosphere a similar amplitude peak is often found and associated with energy damping of upwards propagating tides. The oscillations generated in-situ in Run NT were generally found not to propagate vertically but to be constrained to the region of forcing. This suggests that most of the in-situ generated tides of the thermosphere are evanescent (see section II.4.i)

In order to closer investigate the origin of these in-situ temperature oscillations, the main energy terms in this simulation are shown for pressure level 10 (around 190 km altitude) in Figure IV.8 and sums of these curves are displayed in Figure IV.9 (solid curves with circles). Equatorward of around 60° latitude, the principal heating comes directly from the solar EUV and thus during daytime only, causing primarily diurnal oscillations. This is confirmed by the upper- and middle
plot of Figure IV.8, where the solar heating curve (filled circles) is given as the sum of EUV heating and infrared cooling. In what follows, the EUV heating curve always includes the infrared cooling component since in CTIM they are calculated together on the basis of the same parameterization. The other main cooling mechanism in the model is vertical heat conduction, which is calculated self-consistently. The plots for low- to mid latitudes show that solar heating occurs between 6 and 18 h local time and downwards heat conduction reacts with a time delay of around 3 h. The only other important factor contributing to the energy at those latitudes is the vertical energy advection, or energy transport by vertical winds.

The situation is however different at 80° latitudes (Figures IV.8 and 9, bottom). Here, the solar heating is much less important since the simulation is for equinox condition, with peak solar irradiance at the equator. The strongest energy source at 80° latitude is Joule heating from the auroral oval, a ring shaped area of enhanced electron densities around 5-10° width in latitude and in this particular simulation lying between roughly 80°N latitude in the early evening and about 60°N in the early morning hours. The auroral oval plays an important role on the night side in that it provides, through Joule heating, a source of energy in regions which experience no direct solar radiation. Figure IV.8 (bottom) shows that Joule heating occurs between around 13 h and 20 h local time, thus 2 h longer into the evening compared with the solar heating, an aspect which is described in more detail later. In addition, horizontal and vertical energy advection are important energy sources on the night side and during the early morning hours, implying that energy is transported into those regions from elsewhere by horizontal and vertical winds. Figures IV.8 and 9 (bottom) therefore suggest that high latitude energy input during equinox depends more on the "indirect" energy sources such as advective redistribution, and less on direct solar absorption. Since the energy input is weaker at the high latitudes during equinox it occurs more evenly distributed in time, it has a weaker diurnal profile (Figure IV.9, bottom), as opposed to the strongly diurnal heating at lower latitudes.

At the chosen latitude of 80°N the auroral oval is encountered only during early evening hours. When choosing a different location the timing and duration of exposure to Joule heating will vary, depending on the exact shape of the auroral oval which varies with the shape of the high-latitude electric convection field as well as the intensity of magnetospheric particle precipitation. At different latitudes, therefore, the night side Joule heating occurs at different hours and not necessarily in the early evening, as in Figure IV.8 (bottom). The peaks of semidiurnal amplitudes
at around 70° N latitude in Figure IV.7 (top) indicate that locations at that latitude spend more time than other locations under the influence of the night side auroral oval and the associated Joule heating. A more detailed modelling study investigating the neutral atmosphere’s response to auroral electrojet forcing was presented by Larsen and Mikkelsen [1987].

ii) WINDS

Figure IV.6 (middle, bottom) shows the amplitudes of diurnal zonal- and meridional winds, and one prominent feature are the relatively small values (under 20 m/s) below 150 km altitude. As is the case for the temperatures, the wind profiles above 150 km are roughly symmetric around the equator, showing little structure. Diurnal amplitudes are largest at high latitudes with values of up to 250 m/s but at mid-latitudes only lie around 90 m/s for zonal- and below 50 m/s for meridional winds. The latter decrease further towards the equator where they reach a minimum of around 10 m/s. In order to understand better the cause of these wind structures, horizontal momentum terms at level 10 (around 190 km) are displayed in Figures IV.9-11 for the meridional and zonal components at latitudes 20° N, 50° N and 80° N. The dashed and solid curves in Figure IV.9. are the total eastward- and southward momentum changes, respectively, per unit mass per time interval of 60 seconds and Figures IV.10, 11 are the breakdowns of these total changes into individual components. The plots in Figure IV.9 show that differences are smaller between the low- and mid latitudes (top and middle) than between the mid- and high latitudes (middle and bottom). The low- and mid latitude momentum behaviour is primarily diurnal with peak eastward forcing around 16 h local time and peak southward forcing 4 h later, at 20 h local time.

The zonal pressure forcing is similar at low- and mid latitudes and primarily westward between around 5 h and 13 h local time and eastward at other times. This is consistent with the fact that the day side heating causes a pressure gradient towards the night side. In the meridional direction the pressure gradient is much smaller at low latitudes and increases poleward, being generally northward on the northern hemisphere day side- and southward on its night side. This again is consistent with properties of the day- to night pressure gradient. Figure IV.11(bottom) shows that the horizontal pressure gradient is however mainly semidiurnal at high latitudes. This can be explained in similar manner as the semidiurnal heating found near the poles (lower plot in Figure IV.9) which was discussed in the previous section.
At low- to mid latitudes the ion drag is clearly important only during daytime, being eastward during most of the period after 6 h and westward in the late afternoon hours. In the meridional direction the northern hemisphere ion drag is almost entirely southward. The reason for this daytime occurrence is that ion drag depends on the ion concentration. With most ionization occurring on the day side due to solar radiation, this enhancement of ion drag is caused by the increased day side charged particle densities. Similarly, the ion drag term depends on altitude and is generally negligible in the lower thermosphere at low-to mid latitude since charged particle densities are relatively small there. An exception to this is the region near the equatorial electrojet. The plots in Figure IV.11 show that at low-to mid latitude the horizontal pressure gradient is larger than the ion drag term and opposite in sign, implying that the ions are pulled by neutral particles across the weak electric field lines, thus slowing down the neutrals to some degree.

The high latitude ion forcing term is considerably different from that found at the low-to mid latitudes in that it is stronger than the pressure forcing term during the early evening hours, particularly in the zonal direction. The strong westward forcing occurs simultaneously in time with the ion heating found in Figure IV.8 (bottom), and is caused by fast ions in the auroral oval. In contrast to the situation at lower latitudes, therefore, ions “pull” the neutral particles at high latitudes. Since the high latitude ions flow zonally, ion drag in the meridional direction is very weak. As described earlier, the chosen latitude of 80°N encounters the auroral band of ion winds only during early evening hours. Again, choosing a different location will change the degree of exposure to large ion winds. At a different latitude, therefore, the forcing of neutrals by ions occurs at different hours.

Coriolis forcing increases with latitude and appears more important in the meridional direction. At mid latitudes it balances roughly the pressure forcing, especially during the nighttime where the absence of any other significant forcing gives a quasi geostrophic wind flow parallel to pressure gradients. At high latitudes Coriolis forcing becomes more irregular.

A number of other momentum terms become important at high latitudes, in particular horizontal advection and vertical viscous drag. In the meridional direction (Figure IV.10, bottom) one prominent feature is the strong northward advection in the early morning hours, around 6-7 h local time. The reason for this can be found in the high latitude wind field. At 80°N latitude the neutral winds flow towards the night side at dawn, giving them a south- to westward direction shortly
before 6 h local time and north- to westward direction after that. As a result, the meridional orientation of the wind changes at dawn from southward to northward, giving a strong northward advection term. Vertical viscosity is generally of secondary importance and appears to counteract partly the pressure gradient.

In summary, in-situ wind patterns found near 200 km height in low- to mid latitude regions are generally forced diurnally by the pressure gradient and Coriolis force, with ion drag playing a role only during daytime. At high latitude, neutral winds are forced not only by the pressure gradient but in the early evening or night by fast moving ions. Horizontal advection and vertical viscosity also play an important role and the total forcing at high latitudes is mainly semidiurnal.

IV. 3. 2. DIURNAL FORCING

Diurnal forcing is applied to the lower boundary of CTIM through the (1,1) Hough mode which is the most important diurnal mode propagating into the thermosphere from below. A run was carried out (1 IE), using a forcing amplitude of 100 m geopotential height and a maximum at 12 h local time. The resulting lower boundary profiles are shown in Figure IV.1. Effects of this forcing on the thermosphere below 200 km are shown in Figure IV.12 for temperature and horizontal winds. The plots show diurnal amplitude changes and were produced as described in section IV.2, by subtracting the diurnal amplitudes under in-situ forcing from those under (1,1) mode forcing.

i) TEMPERATURE

The change of diurnal temperature amplitude shows a distinct pattern which is symmetric to the equator and strongest between around ±30° latitude (Figure IV.12, top). Of the three main maxima in the pattern, the strongest lies above the equator at around 130-140 km height. In this region the diurnal amplitudes have increased by about 60°K, which is significant when compared to the in-situ amplitudes in that region which lie below 10°K (Figure IV.6, top). Two weaker maxima are located at around ±18° latitude and extend to almost 150 km altitude. The diurnal amplitude increases by around 40° K which again is significant compared with the in-situ forcing found there. Outside the region of this pattern, poleward of around ±40° latitude, the diurnal amplitudes do not change under the influence of the lower boundary diurnal forcing. The “V” shape
of the response pattern in Figure IV.12 (top) suggests that oscillations propagate upwards and simultaneously from lower to higher latitudes. This is examined further in IV.6.1.

The marked curve in Figure IV.13 (top) shows the total energy flux per 60 sec time step at 20°N and level 7 (around 122 km altitude), where the response to diurnal forcing is strongest (Figure IV.12). A breakdown of this curve into individual terms is shown in Figure IV.13 (bottom). For comparison, the same parameters, as produced by run NT, are shown for the same location in Figure IV.14. One sees that the total energy flux is smaller by an average factor of 5 and mainly semidiurnal in the case of in-situ forcing (Figure IV.14, top). Comparison of Figures IV.13 and 14 (bottom) shows that these differences are caused by stronger horizontal and vertical energy advection terms as well as increased vertical heat conduction. As shown in equation III.22, the vertical energy advection term is given by

\[ w \frac{\partial (e+gZ_p)}{\partial p} \]

and thus depends on the vertical wind in the pressure frame, \( w \). As discussed further in Chapter VI, \( w \) is linked to the horizontal wind gradient. The (1,1) forcing at the lower boundary produces additional horizontal wind gradients which generate up- and downwelling relative to pressure levels and, in terms of energy, adiabatic heating and cooling on pressure levels. This is confirmed in Figure IV.13 (bottom) through the vertical energy advection term which forms the dominant heat flux at that location. Therefore, the observed temperature changes are caused mainly by adiabatic heating and cooling.

ii) WINDS

Figure IV.12 (middle and bottom) shows that meridional winds respond to diurnal forcing stronger than the zonal winds. The peak zonal wind amplitude change lies around 20 m/s, which is about half the increase observed for meridional winds. The pattern in Figure IV.12 (middle) consists mainly of two symmetric maxima located at ±26° latitude and around 115 km height. Higher up, there are two thin regions of amplitude enhancement at midlatitude which disappear above 220 km height, but these hardly influence the total winds since in situ zonal wind amplitudes at mid latitudes are much stronger. Regions of enhanced meridional wind amplitudes form a pattern which is more complex than that of zonal winds and temperatures, with essentially two "V" shaped patterns which again suggest simultaneous upward and poleward propagation of the oscillations. The meridional winds respond between 80 and 220 km height, while the temperature response is limited to altitudes between around 100 and 160 km.
Meridional momentum terms at 20°N latitude and pressure level 7 are plotted in Figure IV.13 (middle) and show that the observed wind changes at this location are associated primarily with stronger horizontal pressure gradients, followed by larger horizontal advection and vertical viscous drag. The peak of the horizontal pressure gradient lies at 18 h local time but is generally dependent on the choice of phase for the lower boundary forcing. As a result of stronger horizontal winds the ion drag has also increased, but overall is not an important factor at that latitude. The relative importance of terms is very similar for the zonal momentum components, but the zonal pressure gradient is smaller than the meridional one, thus giving the different responses of zonal- and meridional winds.

The diurnal phases are plotted in Figure IV.15 for temperatures (solid), zonal- (dashed) and meridional (dotted) winds. The decreasing phase values with height indicate westward propagation of the diurnal tide, in agreement with expectations. The plotted values are the local times of maxima and demonstrate for all three parameters that phase changes occur only below around 160 km in the thermosphere. Above that altitude the phases are almost constant with height, implying that there is no vertical flux of energy through oscillations. The implication of this is that the diurnal wave at higher altitudes is non-propagating only. This is confirmed by the Hough mode decomposition carried out in section IV.6.1. The plot also shows that the diurnal phase changes considerably with height at a rate of around 14 h over a 20 km range which is roughly the same for all three parameters. At higher altitudes this rate falls gradually, and above 160 km suddenly approaches zero.

Finally, one can estimate the vertical wavelength of the (1,1) mode from Figure IV.15 and obtains a value of around 10 km near the lower boundary which increases to around 35 km near the height of dissipation.

IV. 3.3. SEMIDIURNAL FORCING

While the diurnal forcing at CTIM's lower boundary uses only the (1,1) mode, a selection of four different modes is available for semidiurnal forcing. As outlined earlier, their main differences are the vertical wavelength and the latitudinal structure. The latter is reflected in the global profiles of Figures IV.2-5 and the latitudinal profiles of Figure II.2. This section will compare the different responses of thermospheric temperatures and winds to forcing with various semidiurnal modes.
i) TEMPERATURE

The changes of semidiurnal temperature amplitudes for different mode forcing are shown in upper plots of Figures IV.16 - 19 and were taken from runs 22E - 25E, respectively (see Table IV.1). Using the same geopotential height amplitude of 200 m, the response is stronger for the higher order modes, increasing from around 7°K with (2,2) mode- to 30° K with (2,5) mode forcing. As seen previously, these plots show the change of semidiurnal amplitudes introduced by forcing with the relevant modes, compared with in-situ forcing. All of the increases found are significant, when compared with in-situ amplitudes (Figure IV.7). For the (2,2) and (2,4) modes the response patterns are symmetric to the equator, while the (2,3) and (2,5) modes both have antisymmetric response patterns. The patterns become more complex for higher modes as well as extending to higher latitudes.

The semidiurnal temperatures are only changed within around ±30° latitude by (2,2) mode forcing, but this band extends to roughly ±40°, ±50° and ±60° latitude when forcing with the (2,3), (2,4) and (2,5) modes, respectively. The (2,2) mode pattern extends to altitudes of around 200 km, while all other modes are confined to below around 160 km. This is associated with the vertical wavelength (Table II.1) which is largest for the (2,2) mode, thus giving least effective damping, as shown further in Chapter V. One sees in Figure IV.16 (top) the periodic change of temperature amplitude when moving upwards near the equator, with similar but weaker vertical cells of oscillations appearing at around ±20°, ±40° and ±50° latitude. At around 105 km there is a peak increase of temperature amplitude above the equator, followed by a peak decrease at 140 km and again a peak increase at 185 km. These extrema, due to the falling density, become stronger with height but then disappear above 200 km, indicating the onset of energy damping. In regions of amplitude decrease the 12 h temperature amplitude has been reduced by the lower boundary forcing through destructive interference with the in-situ oscillations. Similar vertical periodic structures are found in the temperature responses of higher modes in Figures IV.17 - 19.

From the plots one can estimate the vertical wavelengths of these oscillations. For the (2,2), (2,4) and (2,5) modes the values are around 80 km, 50 km and 45 km, respectively. The wavelength is more difficult to obtain for the (2,3) mode (Figure 17) since the vertical wave structure is less distinct. These figures show the decrease of vertical wavelength with latitudinal wave number, as expected from theory (see Table II.1).
In what follows, the factors causing the observed temperature responses to tidal forcing are examined in more detail for the (2,2) mode forcing. After that, the relative importance of the various energy sources under forcing with the other modes are also discussed. Results obtained for the (2,2) mode forcing case can then be applied with some modifications to the other modes as well without the need to carry out the entire (2,2) mode analysis again.

In Figure IV.20 the total energy change and a breakdown into individual terms are shown, respectively, for 20°N and pressure level 9 (around 163 km) in run 22E. As previously, the location was chosen to lie within a region of strong temperature response (see Figure IV.16). Compared with the total energy flux under in-situ forcing (dashed curve in Figure IV.20, top) the changes at first appear minor only. Comparing Figures IV.6 and 7 (top) one sees that the semidiurnal amplitudes at the chosen level are small in relation to the diurnal ones, which is confirmed by the mainly diurnal behaviour of the curves in Figure IV.20 (top). When examining the differences in semidiurnal amplitudes of these curves one finds, however, that they fully account for the observed temperature pattern in Figure IV.16.

In order to find which of the energy terms are causing the temperature behaviour each term was Fourier-decomposed individually for every model run. The semidiurnal amplitudes of each term under forcing were subtracted from those under in-situ forcing and the resulting profiles compared with the temperature response pattern of Figure IV.16. This method was necessary since a comparison of individual terms from plots such as Figure IV.20 (bottom) proved problematic due to the dominance of the diurnal behaviour. According to this plot the main terms at this latitude and height are solar heating with infrared cooling, vertical heat conduction and vertical energy advection. The analysis of semidiurnal amplitudes of these three terms found that vertical heat conduction only made a minor contribution to the main temperature pattern of Figure IV.16. The terms causing this pattern were however found to be the solar EUV heating and the vertical energy advection, of which the sum is shown in Figure IV.21 (top).

The change of solar heating can be linked to the redistribution of mass by the tidal motion which leads to different density distributions and different rates of absorption. This is confirmed by the middle plot in Figure IV.21 which shows the change of semidiurnal density amplitudes.

One interesting aspect is that the neutral density maximum at 200 km (middle plot of Figure
IV.21) is smaller than the other maximum at 105 km, while in the temperature plot (Figure IV.16) the upper maximum is larger. This is caused by vertical energy advection, or transport of energy by vertical winds. In the bottom plot of Figure IV.21 the semidiurnal amplitude changes of vertical wind are displayed. The plot shows that below 180 km no distinct structure is present, implying that the vertical winds play only a minor role in that region, which is confirmed by their small values of below 0.3 m/s. However, a strong minimum is seen at 200 km height which coincides with the temperature extremum in Figure IV.16. This minimum expresses that the semidiurnal amplitudes of vertical winds have at that location been reduced by the lower boundary tidal forcing. Therefore, transport of energy by vertical winds has been reduced as well, implying that the energy produced in that region by the increased solar absorption is transported away less effectively than at other locations. The (opposite) density- and vertical wind changes thereby both have the effect of increasing the semidiurnal temperature amplitudes at 200 km.

ii) WINDS

The changes of semidiurnal wind amplitudes in the middle- and lower plot of Figures IV.16 can be attributed mostly to the Coriolis force and horizontal pressure gradient which are displayed in Figures IV.23 and 24 for southward- and eastward direction, respectively. In both directions the Coriolis term is strongest between around ±30° and ±60° latitude and 100-130 km altitude. The horizontal pressure gradient term is strongest within ±40° latitude and reaches to altitudes of around 200 km. When plotting the individual momentum terms of run 22E at pressure level 10 (around 190 km) they appear very similar to those in Figures IV.10 and 11 and are thus not shown here separately. It is evident from these plots that the major mechanisms damping the propagation of tides at that height are ion drag and, to a lesser extent, vertical viscous drag. Horizontal viscous drag was found to be very minor only in that height regime and thus unimportant here for the dissipation of tides.

In Figures IV.17-19 the wind response to forcing with the (2,3), (2,4) and (2,5) modes, respectively, is shown. The complexity of the response pattern increases with mode order as well as broadening more towards the poles. While the zonal wind response remains roughly the same for all modes the meridional winds experience a stronger response under higher mode forcing. The cause of this behaviour is illustrated in Figures IV.25 and 26 which show the semidiurnal amplitudes of the main southward and eastward momentum terms, respectively, at pressure level.
Comparison of the top left plots in Figure IV.25 and 26 show that in the southward direction the horizontal pressure gradient becomes generally larger with mode order, while in eastward direction there is no clear trend to a general increase or decrease. While for the (2,2) forcing mode the pressure gradient is similar in both directions, the southward component is generally stronger by up to a factor of 2 in the southward direction under the higher order modes. The reason for this is that the zonal wave frequency remains the same for all modes, since they are all semi-diurnal, while in the latitudinal direction the variability of the profiles increases, thus causing the larger meridional gradients. In both plots the change of the latitudinal profile shape is also seen. At high latitudes the situation is largely unaffected by the forcing mode.

The Coriolis term (top right plots in Figures IV.25 and 26) appears stronger for the (2,4) and (2,5) modes, in particular above +60° latitude, but other than that no consistent change of overall strength is noticeable. The extrema move poleward with higher mode order and change only little in strength. At low latitudes the Coriolis term is mostly unaffected by the mode of forcing.

The vertical viscosity term (bottom left plots in Figures IV.25 and 26) shows a very regular latitudinal pattern in both directions and has semi-diurnal amplitudes of the order of around 5-10% of the pressure gradient term. Similarly to the pressure gradient, the vertical viscosity is highly variable in meridional direction and increases with mode order in terms of intensity and variability, while in eastward direction it becomes larger only between around 30° and 70° latitude in both hemispheres.

The ion drag term (bottom right plots in Figures IV.25 and 26) is least affected by the choice of forcing mode and at low- to mid latitudes has a small semi-diurnal component only, which is plausible since in those regions the ion drag is present only during daytime and thus mainly diurnal.

In summary, therefore, the differences in wind response to forcing with various Hough modes is caused primarily by the stronger meridional pressure gradients for higher order modes. Although vertical viscosity increases as well it remains small compared with the pressure gradient and is unable to neutralize the effect. Also, the ion damping is very similar in magnitude under all the modes and thus does not contribute towards damping the higher modes more effectively. The
changes of latitudinal structure are mainly controlled by the pressure gradient and Coriolis term.

A comparison of semidiurnal temperature- and wind phases at 20°N under different semidiurnal forcing is given in Figure IV.27. As for the diurnal phase in Figure IV.15, the semidiurnal phase decreases with height, implying westward propagation of the tides. The plots show that phases of temperature and zonal wind are constant above an altitude of around 160 km, while meridional wind phase changes with height up to around 240 km altitude. From Figure IV.27 one can see that the change of phase over a given height range is stronger for the higher order Hough modes. In the case of the zonal winds the phase over a height extent of 20 km changes by around 1.2 hours (72 minutes) under (2,2) forcing and by around 4.5 hours under (2,4) forcing. The rate is roughly constant up to an altitude of around 120 km and then reverses before approaching zero above 160 km. The constant phase change below 120 km indicates that any damping processes are roughly constant there. However, damping becomes stronger above 120 km and slows down the phase progression over a height range of less than 20 km. The reversal of vertical phase progression which is seen in Figure IV.27 indicates that the tide is eastward propagating over a short height range at that latitude before dying away. The same phenomenon is also seen in Figure IV.15 for the diurnal tide under (1,1) mode forcing. The rates of phase change with height are much lower for semidiurnal tides than for the diurnal tide in Figure IV.15, implying that the diurnal tide propagates upwards faster than the semidiurnal tide.

IV. 4. INTERACTIONS BETWEEN SEMIDIURNAL HOUGH MODES

This and the following section will investigate two types of interactions which may occur in the thermosphere as a result of tidal oscillations. While this section looks at the interaction between semidiurnal tidal oscillations, the other examines the interactions between oscillations and the background atmosphere.

In order to determine whether semidiurnal tidal modes interact non-linearly as they propagate upwards, three different model simulations were carried out. Of these, two were forced with one Hough mode each and the third was forced with both of these Hough modes simultaneously. If the non-linear interaction does take place between the tides, the sum of the responses under individual mode forcing (first two runs) will not be the same as the total response found when forcing with both modes simultaneously.
The response under simultaneous forcing with the (2,2) and (2,4) modes (run 2+4E, see Table IV.1) is compared with the sum of responses under the individual modes from Figures IV.16 and IV.18. In Figures IV.28 and 29 these two cases are shown for temperature, zonal- and meridional winds. Comparison of the plots shows that they are different for all three parameters. The overall response is slightly weaker when forced with the two modes simultaneously, and thus non-linear interactions take place and lead to an overall damping of the perturbations. This effect is found to be most important below 140 km altitude for winds and below 180 km for temperature. It was found earlier that these are also the regions where most of the tidal dissipation takes place, and therefore these results confirm each other since non-linear processes lead to damping of the oscillations.

IV. 5. INTERACTIONS OF TIDES WITH THE BACKGROUND ATMOSPHERE

Results from the previous sections have suggested that non-linear interactions play an important role in the thermosphere. It is therefore necessary to assume that the oscillations and the background atmosphere are no longer separable, as was assumed in Classical Tidal Theory. In other words, the perturbations can no longer be studied without considering the background atmosphere. The non-linear interactions between oscillations and the background atmosphere influence zonally averaged parameters as well as the oscillations themselves. This is also examined in Chapter V.

IV. 5.1. TEMPERATURE

The zonally averaged profile of temperature under in-situ forcing (run NT) is shown in Figure IV.30. The plot shows that temperature values increase with height from around 190° to 750° K with little latitudinal structure. Above around 160 km temperature values are larger over the equator since the solar irradiance is stronger there.

In the upper plots of Figures IV.31 and IV.32, the changes of zonally averaged temperatures are displayed under forcing with the (1,1) and (2,2) modes. In order to identify better the changes taking place, the plots were, as previously, produced by subtracting from the zonally averaged values under external forcing those generated under in-situ forcing (of Figure IV.30). In both plots a regular pattern can be seen within the band between around 40°N and 40°S which reaches up to
an altitude of around 200 km. It consists mainly of three vertical bands of cells, with other much weaker bands being partly visible at higher latitudes as well. The strength of the pattern increases with height, reaches a maximum at around 160 km altitude and then falls off again. This shows that zonally averaged temperatures are influenced strongest at an altitude of around 160 km, suggesting that this is the height at which the non-linear interactions, and thus the tidal damping, are strongest. This finding is in agreement with earlier observations of the tidal amplitudes.

However, a number of unexplained features can be seen in the plots, all of which could have a common cause. In the upper plot of Figure IV.31 the overall trend is clearly a decrease of background temperature under tidal forcing, compared with the in-situ values. This is against expectations since the forced lower boundary forms an additional source of energy and should lead to an overall increase of temperature. The same type of background temperature decrease was also found under forcing with the (2,3), (2,4) and (2,5) Hough modes (not shown). Under the (1,1) forcing of run 11E the zonally averaged temperature falls by up to around 80° K, under (2,4) forcing (run 24E) the decrease is by around 20° K and under (2,2) forcing (run 22E) only by around 2° K. Immediately above the lower boundary one can see that temperature is generally increased by the forcing, but this trend disappears above 90 km height. When comparing the response also under the remaining semidiurnal modes (not shown here) it becomes apparent that for higher order modes the zonally averaged temperature is reduced even more.

This point is discussed further in chapter VIII in the context of possible limitations of the current CTIM model which need to be addressed in the future. Essentially, two possible explanations can be brought forward for the observation. One possibility is that the modelled effect is physically correct and forms a new discovery. This is unlikely from the present knowledge but cannot be entirely ruled out at this stage. The other and more likely option is that CTIM is in its present version unable to simulate correctly the non-linear interactions between tides and background atmosphere in terms of energy, although in terms of dynamics the responses match expectations better, as will be shown shortly. The discussion in Chapter VIII essentially proposes that the current vertical resolution of CTIM may be insufficient to model the non-linear processes. In order to simulate these correctly, high frequency oscillations need to be considered which have smaller vertical wavelengths than the standard Hough modes used here and, more importantly, than the vertical resolution of CTIM. This theory is supported by the fact that results from the (2,2) mode simulation in Figure IV.32 match best the expectations, while at the same time the vertical
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wavelength of this mode is larger than that of all the other modes.

IV. 5. 2. WINDS

The impact of (1,1), (2,2) and (2,4) mode forcing on zonally averaged winds is illustrated in the middle- and lower plots of Figures IV.31 and 32 for the eastward- and southward directions, respectively. The equivalent parameters under in-situ forcing are displayed in Figure IV.30. There, the meridional winds are close to zero, which is consistent with the lack of latitudinal structure in the temperature profile in the upper plots of Figure IV.30. The averaged zonal winds are generally small as well, except at high latitudes where a strong westward and weaker eastward flow are found in the northern- and southern hemisphere, respectively, at around 10° distance from the poles. These zonal winds are driven by the fast ions in the auroral oval (see IV.3.1).

The main effect of the diurnal (1,1) mode forcing on zonal winds is the generation of a low-latitude westward jet of around 40 m/s speed between around 100 and 120 km altitude, as shown in the middle plot of Figure IV.31. Two weaker westward jets are seen at roughly 50° latitude and 160 km altitude in both hemispheres. Westward jets are also generated by the semidiurnal modes, as shown for the (2,2) mode in the middle plot of Figure IV.32. The altitude at which these jets occur varies with mode, ranging from near 140 km under (2,2) forcing to 110 km under (2,5) forcing (not shown here). These altitudes coincide with the maxima of semidiurnal amplitudes found for the respective modes in Figures IV.16 to IV.19, suggesting that they are generated by the dissipation of the tides. The impact of tides on the background atmosphere is thus strongest in regions where the tides are damped. This is plausible since the damping extracts momentum and energy from the oscillations which, as a result of conservation of these quantities, must be released to the background atmosphere.

Meridional winds are influenced in a different manner from the zonal components, with more complex patterns of wind change occurring at low latitudes, within 30° of the equator on both hemispheres. These patterns are similar to those found for temperature and they vary for the different modes only in terms of intensity and not in their height- or latitude ranges. The cell structure of the meridional wind response suggests that the dissipation of tides generates regions of equatorward flow within around ±15° latitude at some heights which alternate vertically with regions of flow away from the equator.
For closer examination of the causes of these wind responses, zonally averaged momentum terms are shown in Figure IV.33 for the zonal and meridional components at pressure level 6 (~115 km). From these plots one sees that the zonally averaged zonal wind response at low- to mid latitudes is influenced mainly by the terms of vertical viscosity, vertical advection, horizontal advection and Coriolis force (upper plot in Figure IV.33). At high latitude the response is determined mostly by horizontal advection. The zonally averaged pressure gradient is very small in comparison. The meridional wind response is caused by the same terms, but here the Coriolis force and horizontal pressure gradient in the bottom plot of Figure IV.33 are both larger than the other terms. However, they balance each other at that height, suggesting a generally geostrophic flow. The total meridional response at that height is much weaker than the zonal response. The bottom plot of Figure IV.32 shows that the situation reverses at around 160 km height where the wind changes are stronger in the meridional direction.

In conclusion, therefore, the findings from an analysis of zonally averaged winds suggest that the influence of tides on the background atmosphere does not occur isotropically in both horizontal directions. In the case of the (1,1) mode a westward jet is generated at around 120 km height while in meridional direction the momentum is released mainly in the vicinity of 160 km. The non-lineairities are in both cases caused mainly by the vertical and horizontal advection as well as vertical viscosity.

IV. 6. MODE COUPLING

Mode coupling describes the generation of higher order Hough modes through non-linear interactions with the background atmosphere and may thus also be linked to the dissipation of tides in the thermosphere. As a result of higher order propagating modes being generated, the overall vertical wavelength of the upwards propagating tides is reduced. Due to the differences in latitudinal profile of the various modes, mode coupling is linked to a broadening of the latitudinal range of the tide as it propagates upwards, or essentially propagation of the tide to higher latitudes. Forbes [1995] suggested that the (2,4) mode at 90 km height was generated in equal proportion by thermal heating and mode coupling, the (2,5) mode was mainly unaffected by mode coupling and the (2,3) mode was damped by the process. It is of interest, therefore, whether mode coupling is seen also in the CTIM simulations.
IV. 6.1. THE (1,1) MODE

The mode coupling associated with the upwards propagating (1,1) mode has been studied by Forbes and Hagan [1987] and Vial [1986]. These authors found that dissipation of the (1,1) mode in the vicinity of 90 km altitude leads to broadening of its latitudinal structure. Forbes and Hagan attributed this to the generation of the non-propagating (1,-2) Hough mode by non-linear interactions.

The thermospheric response to (1,1) forcing is shown in Figure IV.12 and was described in more detailed in section IV.3.2. For all three parameters the response pattern broadens with altitude from the region within around 30° of the equator in both hemispheres at 80 km to around ±60° at 200 km. This indicates that the mode broadening is reproduced by CTIM, but more detailed analysis is necessary. Therefore, the diurnal southward winds under (1,1) forcing have been decomposed into the (1,1) and (1,-2) Hough modes and the profiles are compared to those obtained under in-situ forcing (run NT). Theoretical profiles of southward winds for the two Hough modes are shown in Figure IV.34. The plots show that the (1,1) mode winds are strongest at low latitude, while the (1,-2) mode winds increase towards the poles. A further important difference between the modes is that the (1,-2) mode is non-propagating and thus exists only in the region of forcing (see also Chapter II).

A decomposition of southward winds is shown in Figure IV.35 for in-situ forcing (run NT) and under (1,1) mode forcing (run 1IE). The (1,1) mode is of minor importance only under in-situ forcing where the (1,-2) mode is dominant throughout the thermosphere, as a result of high-latitude momentum forcing. The (1,-2) mode amplitudes increase with height to values of up to around 200 m/s at 250 km. When applying lower boundary forcing the balance between the modes is altered in the lower thermosphere. Up to around 120 km altitude the (1,1) mode is dominant and then decreases as a result of dissipation. Up to around 150 km the intensity of the (1,-2) mode is unaffected by the forcing but above that altitude is stronger by up to 30 m/s as a result of lower boundary (1,1) mode forcing. The plots therefore suggest that the (1,1) mode dissipation generates non-propagating (1,-2) oscillations. The altitude of (1,1) mode dissipation lies around 120 km and is thus higher than given by most other perturbation models (Forbes and Vial, 1989; Hagan and Forbes, 1993). The reason for this is most likely to lie in the vertical profile of the eddy diffusion coefficient used in CTIM. Many authors (eg. Forbes and Hagan, 1988) have emphasized the
importance of this coefficient to the behaviour of the (1,1) mode in the lower thermosphere. More detailed comparisons with other perturbation models are presented and discussed in Chapter V.

IV. 6.2. THE SEMIDIURNAL MODES

The atmospheric response to semidiurnal tidal forcing is shown in Figures IV.16-IV.19 and was described in section IV.3.3. From these plots the mode broadening effect is not as obvious as for the diurnal mode and therefore a decomposition has been carried out also for semidiurnal southward winds. The vertical profiles of the (2,2), (2,4) and (2,5) modes are shown in Figure IV.36 under in-situ and under (2,2) mode forcing. The (2,3) mode is not displayed since it found to be of minor importance only in the two runs used here. The plots indicate that in-situ forcing leads to a roughly equal proportion of the (2,2) and (2,4) modes. The (2,5) mode is dominant above around 110 km altitude and considerably larger than the other two. It peaks near 200 km height with around 30 m/s amplitude. When considering that peak semidiurnal southward wind amplitudes in the thermosphere can exceed 150 m/s (Figure IV.7) it is clear that the three modes shown in the bottom plot of Figure IV.36 do not fully account for the total profiles. The strongest contribution to semidiurnal oscillations at altitudes above 200 km comes from non-propagating modes, as shown for the diurnal case in the previous section. Non-propagating semidiurnal modes are however not discussed here.

Under (2,2) mode forcing (upper lot in Figure IV.36) the profiles of the (2,2) and (2,4) modes change below 200 km height and are unaffected by the lower boundary forcing above that altitude. The (2,2) mode is strongest up to around 140 km and then falls below the level of the (2,4) mode. The former peaks near 120 km height and the latter near 170 km. When comparing the two plots in Figure IV.36 one can see that the in-situ (2,4) mode is unaffected by the (2,2) mode up to around 115 km altitude. The (2,2) mode then dissipates near 120 km and as a result of that, the (2,4) mode becomes stronger. These results suggest that mode coupling does occur in these model runs and is directly related to the tidal dissipation. Finally, a comparison of plots in Figure IV.36 shows that the (2,5) mode is almost identical under in-situ and (2,2) mode forcing. This result is in agreement with findings by Forbes [1995] who suggested that mode coupling does not affect the (2,5) Hough mode.

In summary, CTIM does simulate the effect of mode coupling in accordance with findings by other
authors. When tidal modes dissipate they generate a higher order Hough mode which then propagates to higher altitudes until it also dissipates. As a result, the vertical wavelength of a tide is reduced. This effect might play a role in the earlier findings which suggested that CTIM's vertical resolution might pose a problem to some of the results. This issue is discussed further in Chapter VIII.
Figure IV.1.a: Geopotential height- and temperature perturbations (in [m] and [°K], respectively) of the (1,1) Hough mode with a geopotential forcing amplitude of 100 m and phase set to 12.0 h local time.
Figure IV.1.b: Same as Figure IV.1 a, but for eastward- and southward winds, in m/s.
Figure IV.2.a: Geopotential height- and temperature perturbations (in [m] and [°K], respectively) of the (2,2) Hough mode with a geopotential forcing amplitude of 200 m and phase set to 12.0 h local time.
Eastward wind (2,2) mode

Southward wind (2,2) mode

Figure IV.2.b: Same as Figure IV.2.a, but for eastward- and southward winds, in m/s.
Figure IV.3.a: Geopotential height- and temperature perturbations (in m and °K, respectively) of the (2,3) Hough mode with a geopotential forcing amplitude of 200 m and phase set to 12.0 h local time.
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Eastward wind (2,3) mode

Southward wind (2,3) mode

Figure IV.3.b: Same as Figure IV 3a, but for eastward- and southward winds, in m/s.
Figure IV.4.a: Geopotential height- and temperature perturbations (in [m] and [°K], respectively) of the (2,4) Hough mode with a geopotential forcing amplitude of 200 m and phase set to 12.0 h local time.
Eastward wind (2,4) mode

Southward wind (2,4) mode

Figure IV.4.b: Same as Figure IV.4.a, but for eastward- and southward winds, in m/s.
Figure IV.5.a: Geopotential height- and temperature perturbations (in [m] and [°K], respectively) of the (2,5) Hough mode with a geopotential forcing amplitude of 200 m and phase set to 12.0 h local time.
Figure IV.5.b: Same as Figure IV.5.a, but for eastward- and southward winds, in m/s.
Figure IV.6: Diurnal in-situ amplitudes of temperature, zonal- and meridional winds, as simulated in Run NT. Winds are given in m/s and temperature in °K.
Figure IV.7: Semidiurnal in-situ amplitudes of temperature, zonal- and meridional winds, as simulated in Run NT. Winds are given in m/s and temperature in °K.
Figure IV.8: Energy terms at level 10 (~190 km) for in-situ forcing (run NT). Values are for 60 sec intervals per unit mass in m²s⁻². Curves are solar heating (●), particle precipitation energy (△), Joule heating (▼), horizontal energy advection (×), vertical energy advection (♦) and vertical heat conduction (+).
Figure IV.9: Total momentum and energy changes per 60 sec interval per unit mass at level 10 (~190 km) for in-situ forcing (run NT). Momentum values are for eastward- (solid) and southward (dashed) components (in m s\(^{-2}\)), energy values (solid with dots, in m\(^2\) s\(^{-2}\)) are summed over zonal and meridional contributions.
Figure IV.10: Southward momentum terms at level 10 (~190 km) for in-situ forcing (run NT). Values are for 60 sec intervals per unit mass in m/s². Curves are horizontal advection (●), Coriolis force (X), horizontal Geopotential height gradient (▲), ion drag (□) and vertical viscous drag (+).
Figure IV.11: Same as Figure IV.10, but for eastward momentum terms.
Figure IV.12: Change of diurnal amplitudes when forcing with the (1,1) mode. Winds are given in m/s and temperature in °K.
Figure IV.13: Momentum and energy changes per 60 sec interval per unit mass (in m s\(^{-2}\) and m\(^2\) s\(^{-2}\), respectively) at level 7 (~122 km altitude) for (1,1) mode forcing (Run 11E). For caption of curves in upper plot, see Figure IV.9, for middle plot see Figure IV.10 and for lower plot see Figure IV.8.
Figure IV.14: Same as Figure IV.13, but for Run NT.
Diurnal phase
lat = +20

Figure IV.15: Diurnal phases of temperature (solid), eastward-(dashed) and southward (dotted) wind, as simulated by CTIM under (1,1) mode forcing (Run 11E).
Figure IV.16: Semidiurnal amplitude changes induced by (2,2) mode forcing for temperature, zonal- and meridional winds. Winds are given in m/s and temperature in °K.
Figure IV.17: Same as Figure IV.16, but for (2,3) mode forcing.
Figure IV.18: Same as Figure IV.16, but for (2,4) mode forcing.
**Figure IV.19:** Same as Figure IV.16, but for (2,5) mode forcing.
Figure IV.20: Energy changes per 60 sec interval per unit mass at level 9 (~163 km) for (2,2) mode forcing (run 22E), in units of $m^2 s^{-2}$. The upper plot also contains values from run NT (dashed). For caption of curves in lower plot see Figure IV.8.
Figure IV.21: Changes of semidiurnal amplitudes through (2,2) forcing (run 22E) for the solar heating and vertical advection energy terms (top), the neutral density (middle) and vertical wind (bottom). Values of the upper plot are in units of m^2 s^{-2}, of the middle plot in log_{10} (kg m^{-3}) and the bottom plot in m/s.
Figure IV.22: Semidiurnal amplitudes of the horizontal and vertical energy advection terms (top, middle) and vertical heat conduction (bottom). Plots show the amplitudes under semidiurnal forcing with the (2,2), (2,3), (2,4) and (2,5) modes (runs 22E, 23E, 24E and 25E, respectively) at pressure level 8 (around 135-140 km, depending on mode). The terms are for 60 sec intervals in units of m²s⁻².
Figure IV.23: Changes of 12 h amplitudes through (2,2) forcing (run 22E) for the southward components of Coriolis force (top) and horizontal geopotential gradient (bottom). Values are for 60 sec intervals in units of ms^{-2}. 
Figure IV.24: Same as Figure IV.23, but for eastward components.
Figure IV.25: Semidiurnal amplitudes of the southward terms of horizontal geopotential height gradient (top left), Coriolis force (top right), vertical viscosity (bottom left) and ion drag (bottom right). Plots show the 12 h amplitudes under semidiurnal forcing with the (2,2), (2,3), (2,4) and (2,5) modes (runs 22E, 23E, 24E and 25E, respectively) at pressure level 7 (around 120-126 km altitude, depending on mode). The terms are for 60 sec intervals in units of ms\(^{-2}\).
**Figure IV.26**: Same as Figure IV.25, but for eastward terms.
Figure IV.27: Semidiurnal phases (in h LT) of temperature, eastward- and southward wind at latitude 20.0N, under forcing with the (2,2) mode (solid), (2,3) mode (dashed), (2,4) mode (dotted) and (2,5) mode (dashed-dotted).
Figure IV.28: Changes of semidiurnal temperature, zonal- and meridional wind amplitudes induced by external forcing with the summed (2,2) and (2,4) modes (run 2-4E). Winds are given in m/s and temperature in °K.
Figure IV.29: Same as Figure IV.28, but with (2,2) and (2,4) mode forcing carried out in two separate runs. The plots are the sums of Figures IV.16 and IV.18.
Figure IV.30: Zonally averaged temperature [K], eastward- and southward wind [m/s] under in-situ forcing (run NT), at equinox.
Figure IV.31: Changes of zonally averaged temperature [K], eastward- and southward wind [m/s] caused by (1,1) mode forcing. The plots are Run 11E-Run NT.
Figure IV.32: Same as Figure IV.31, but generated by (2,2) mode forcing.
Figure IV.33: Zonally averaged momentum term changes caused by (2,4) mode forcing. Plots show the terms of hor.advection, vert.advection, vert.viscosity, coriolis force and horiz.pressure gradient at pressure level 6 (~115 km). Values are the differences between terms under (2,4) forcing and under in-situ forcing. The sum of all curves is given as thick black line. Values are the momentum changes per 60 sec per mass, in units of ms⁻².
Figure IV.34: Latitudinal profiles of diurnal southward wind amplitudes for the propagating (1,1) Hough mode (top) and the non-propagating (1,-2) mode (bottom). All values are normalized to 1.
Figure IV.35: Vertical profiles of diurnal (1,1) and (1,2) Hough modes, as generated in-situ (bottom) and under (1,1) mode forcing (top). The decomposition was carried out for southward wind.
Figure IV.36: Vertical profiles of semidiurnal (2,2) , (2,4) and (2,5) Hough modes, as generated in-situ (bottom) and under (2,2) mode forcing (top). The decomposition was carried out for southward wind.
CHAPTER V. VALIDATION

ABSTRACT

Two comparative studies are presented in this chapter between CTIM, other numerical models and measurements. They are preceded by a general discussion of factors influencing the simulated tidal profiles, thus adding to the findings of Chapter IV. It is found that wind amplitude profiles are sensitive to the background temperature in the atmosphere, to the molecular viscosity coefficient and, at mid-to high latitudes, to the background wind field as well as the relative phase between forcing from below and the solar- and auroral in-situ momentum sources. Temperature amplitude profiles are found to be influenced by the same factors as the winds and, additionally, by the coefficients of heat conduction and by the background winds also at low latitudes. The height of momentum dissipation varies with tidal amplitude, while the height of energy dissipation hardly does. The tides' vertical wavelength, however, is important for the height of both the momentum- and energy dissipation. Comparisons with the TIGCM model generally show good agreement, with discrepancies being explained primarily by the differences in background temperature at low- to mid latitudes and differences in phase with the auroral forcing at mid- to high latitudes. Comparisons with the GSWM, HWM and MSISE90 show reasonable agreement of the phases but generally lower amplitudes in CTIM. These cannot be accounted for comprehensively by the discussed influencing factors. Comparisons with measurements at low- and mid-latitude from the LTCS-9 campaign show good agreement of meridional wind amplitudes, some similarities in the temperatures and less agreement in the zonal winds. Comparisons with high latitude EISCAT measurements for seasonally averaged conditions at equinox show reasonable agreement. Finally, a series of plots are presented which compare tidal amplitudes and phases from CTIM at the mid-latitude site of Millstone Hill (42°N) to measurements and output from the above models as well as the TMTM model. Most of the discrepancies are attributed to ambiguity in the lower boundary forcing profile of CTIM, due to the lack of global tidal measurements.

V. 1. INTRODUCTION

This and following chapters present a number of studies carried out with the new CTIM model code described in Chapter III. While the analyses presented in Chapter IV are mainly numerical
Validation

Chapter V

experiments, this chapter focuses on comparisons with measured data and output from other numerical models. These comparisons are essential for validating the code. However, the results presented here are not only intended for validation but also complement those discussed in Chapter IV to give a broader understanding of the factors influencing the tidal propagation into the thermosphere. A discussion of these factors is presented in the first section of this chapter and then applied to the comparisons and, where necessary, explored further. Chapters IV and V thus together form a comprehensive study into the nature of tides.

Tidal simulations of CTIM are first compared with output from two numerical and two empirical models. The numerical models used are the National Center for Atmospheric Research Thermosphere Ionosphere General Circulation Model (TIGCM and TIEGCM) [Roble et al., 1988; Richmond et al., 1992] and the Global Scale Wave Model (GSWM) [Hagan et al., 1995]. A number of runs from both models are available on the CEDAR (Coupling, Energetics and Dynamics of Atmospheric Regions) database, but since these are for different conditions they cannot be compared to the same CTIM run. For this reason it was necessary to carry out two separate simulations with CTIM which differ in their external tidal forcing and solar activity. Settings in the CTIM runs were chosen to match as closely as possible the settings used by the other models, in particular regarding the specification of lower boundary tides. The TIGCM run uses output from the model by Forbes and Vial [1989] for tidal input at its lower boundary, while the GSWM model simulates self-consistently the formation and upwards propagation of tides, thus requiring no external forcing.

The tidal winds are also compared with the empirical Hedin Wind Model (HWM) [Hedin et al., 1993] and temperatures with the empirical MSISE90 [Hedin, 1991] atmosphere model. Since semidiurnal amplitudes of those models at 80 km altitude were found to be roughly similar to those produced by the GSWM they are compared to the same CTIM simulation as the GSWM run.

In V.5, a combined comparison for the site of the Millstone Hill observatory (42.6°N, 71.5°W) is presented between CTIM, measurements, the TIEGCM, GSWM and TMTM models. Of these, the TIEGCM is an updated version of the TIGCM (see also V.3.1) and the TMTM is the Tuned Mechanistic Tidal Model, which has been developed recently only and is still largely undocumented.
V. 2. TIDAL DAMPING IN THE THERMOSPHERE

Comparisons of two or more large numerical models, as carried out here, are often made difficult by the fact that not all information about other models which is necessary for an in-depth interpretation is available to an external user, but often known in detail only to the developer. Furthermore, a statement as to which of the models simulates situations more accurately is difficult if the runs are not compared to measurements at different globally distributed locations. Therefore, the factors influencing the tidal propagation are at first discussed in general terms in this and the following paragraph, and later applied to the comparisons. This general discussion is important not only for the comparisons with models but also for those with measurements. The relevant momentum- and energy drag terms used in CTIM are reviewed to discuss what model properties could be responsible for any observed discrepancies.

Upwards propagating tides are identified by a horizontal- as well as a vertical structure. A common method of studying tides is thus to examine the vertical profiles of horizontal amplitudes, which is done later in this chapter. As expressed by equations (II.25) - (II.27) in Chapter II, a parameter in the presence of a propagating tide has at any fixed moment in time a periodic vertical structure. This is illustrated in Figure V.1 for an idealized situation where no damping of the tide occurs and thus the horizontal amplitude increases exponentially with altitude (dashed line).
Parameter value →

Figure V.1: Vertical profile of any atmospheric parameter (solid line), such as wind and temperature, in the presence of a propagating tide, at a fixed moment in time. The horizontal tidal amplitudes rise exponentially (dashed line) due to the decrease of density with height.

The vertical distances between extrema in Figure V.1 are half the tide’s vertical wavelength. The occurrence of damping prevents amplitudes from growing continuously with height, and in the real atmosphere they reach a maximum value at an altitude, the dissipation height, above which the vertical and horizontal oscillations dissipate and release their energy- and momentum to the background atmosphere. Two categories of factors are found to influence profiles such as that of Figure V.1. One category includes all those parameters which influence the undamped tidal profile and the other category are those which affect the damping processes.

For quick reference, the findings of the following sections are listed in simplified form in Table V.1. Various background atmosphere- and wave parameters are listed along with terms describing their importance for semidiurnal tidal profiles of winds and temperature. A distinction is made between three latitudinal regimes of around 0° to 30°, 30° to 60° and 60° to 90°, denoted as “low”, “mid” and “high”, respectively. The terms “primary”, “secondary” and “negligible” have been used to express the degree of influence of a listed parameter on the semidiurnal tidal profiles.
at various latitudes. The table is based on the findings described in more detail below, and thus on
simulations from the CTIM.

<table>
<thead>
<tr>
<th></th>
<th>low latitude</th>
<th>mid-latitude</th>
<th>high latitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>background wind</td>
<td>negligible</td>
<td>negligible</td>
<td>primary</td>
</tr>
<tr>
<td>background temperature</td>
<td>primary</td>
<td>primary</td>
<td>secondary</td>
</tr>
<tr>
<td>tidal mode</td>
<td>primary</td>
<td>primary</td>
<td>primary</td>
</tr>
<tr>
<td>tidal amplitude</td>
<td>primary</td>
<td>primary</td>
<td>primary</td>
</tr>
<tr>
<td>tidal phase</td>
<td>primary</td>
<td>primary</td>
<td>primary</td>
</tr>
<tr>
<td>molecular viscosity coefficient</td>
<td>primary</td>
<td>secondary</td>
<td>secondary</td>
</tr>
<tr>
<td>molecular heat conduction coeff.</td>
<td>secondary</td>
<td>secondary</td>
<td>secondary</td>
</tr>
<tr>
<td>ion density</td>
<td>negligible</td>
<td>secondary</td>
<td>primary</td>
</tr>
</tbody>
</table>

Table V.1: The degree of influence of various parameters on the vertical profiles of semidiurnal tidal amplitudes between 80 and around 200 km altitude. The findings were obtained from theoretical discussions as well as CTIM simulations.

V. 2. 1. THE UNDAMPED TIDAL PROFILE

The profile in Figure V.1 is influenced by the value of scale height and the altitude at which the tide is generated. It was shown in Chapter II that the vertical wavelength of a tide is expressed in terms of scale height, which itself depends on temperature. In a pressure coordinate frame, the profile of Figure V.1 is to first a approximation invariant with changes of scale height or temperature, but when plotted in a height coordinate frame the curve is “stretched” or “compressed” vertically when increasing- or decreasing the background temperature. For easier
comparison with measured data all plots in this chapter use the height coordinate frame, and therefore differences of background temperature need to be considered. It will be shown later that these also influence the momentum- and energy damping processes.

The other aspect which can, in theory, influence a tidal profile such as that in Figure V.1 is the height at which a tide is generated. In CTIM the tides are generated at 80 km, in the TIGCM (see V.3.1) at 97 km and in the other models used here as well as in the real atmosphere this forcing occurs below 50 km altitude. Comparisons of the CTIM with other models or data require the knowledge of forcing amplitudes and -phases at its lower boundary height (80 km). However, another factor plays a role and potentially might influence the vertical amplitude profiles. In the CTIM and TIGCM, the external tides are generated by oscillating the lower boundary pressure levels. A vertical structure is not defined externally but follows from the background atmosphere properties. In the real atmosphere as well as in models at sufficient distance from the forcing level a tidal oscillation “knows” how far it has progressed from the region of excitation. This can be understood from the finding in Classical Tidal Theory (see Chapter II) that the horizontal- and vertical profiles of tides are coupled. This information cannot be specified when forcing tides externally in the manner used in CTIM and the TIGCM. A tide forced below CTIM’s lower boundary will thus contain information about its further vertical progression which cannot be specified in the CTIM, and similarly a tide forced at 80 km in CTIM contains information at 97 km height which cannot be specified in the TIGCM’s forcing. In Figure V.2 the situation which might arise is illustrated for a simplified case of an undamped tide.
Here, the dashed curve represents a tide which is identical in terms of vertical wavelength and amplitude to the original tide (solid curve), but forced at a vertical distance of $\Delta H$ above the other. A result of this shift, the amplitude maxima do not occur at the same heights. At some heights the two curves cross and give the same parameter values. When specifying forcing in horizontal direction only, it is thus ambiguous whether the vertical progression will follow the solid or the dashed curve. In principle, there are then an unlimited number of possible scenarios. In order to avoid the situation illustrated in Figure V.2, one would need to define a “vertical phase” of the tide at CTIM’s lower boundary. No numerical experiments have yet been carried out to demonstrate the effect under realistic atmospheric conditions, and therefore it is at present a theoretical possibility which might play a role in the CTIM/TIGCM comparisons. A collaborative study with M. Hagan is planned for the near future to carry out such a joint numerical experiment with the CTIM and GSWM models. When ignoring any in-situ forcing with which an upwards propagating tide could interact, the vertical profile shift of Figure V.2 is independent of the choice of horizontal phase since the amplitude at any height is independent of phase as well. In the real atmosphere the horizontal phase does play a role, as outlined below, since interference between the upwards propagating- and in-situ solar- and auroral tides occurs, affecting the total amplitude at
each height, but nevertheless it is expected not to play an important role in this effect.

V. 2. 2. MOMENTUM DRAG

The second category of factors influencing the vertical propagation profiles of tides is linked to the damping processes which occur in the “real” atmosphere. The main damping terms for tidal momentum were found in Chapter IV to be vertical viscous drag at low-to mid latitudes and ion drag at mid- to high latitudes. Vertical viscous drag is in a pressure coordinate frame given by

\[ g \left( \frac{\partial}{\partial p} \left[ \frac{\mu}{H} p \frac{\partial}{\partial p} U \right] \right) \]

where \( \mu \) and \( g \) are the molecular viscosity coefficient and gravitational acceleration, respectively, \( p \) is the pressure, \( H \) the scale height and \( U \) is the neutral velocity vector. The above expression shows that vertical viscous drag is influenced by the values and vertical gradients of molecular viscosity, \( \mu \), the wind vector, \( U \), and the scale height, \( H \). As a result of this, vertical viscous drag is stronger, the more the winds change with height. The vertical gradient of \( U \) is influenced by the vertical wavelength of the oscillation as well as its amplitude. It was already shown in Chapter IV that higher order Hough modes have shorter vertical wavelengths and are therefore damped more effectively than the lower order modes. The scale height dependency implies that vertical viscous drag also increases with falling background temperatures.

The influence of a tide’s amplitude on its height of dissipation is not trivial and needs to be considered in some detail. In principle, two contributing factors have opposite effects. When the tidal amplitudes are larger, the vertical changes of winds are stronger, and thus the viscous drag. This would suggest stronger damping and dissipation at a lower height. However, stronger amplitudes also imply that the kinetic energy of the wave is larger, giving it a higher altitude of dissipation. In order to investigate which of these two factors dominates under realistic conditions in the thermosphere, a numerical experiment was carried out with CTIM. Two test runs were made in which (2,4) mode forcing was applied at an amplitude of 200 m and 400 m geopotential height, with all other parameters being identical. When plotted in a height coordinate system the resulting vertical wind- and temperature amplitudes profiles were shifted vertically as a result of different forcing by up to 30 km. In the pressure coordinate system, amplitude maxima occurred on the same pressure levels in both runs, but numerically fitting the profile suggested a downwards shift by
around a half pressure level under stronger tidal forcing. The runs therefore suggest that larger amplitudes tend to lower the height of dissipation, assuming the same vertical wavelength.

The second important momentum damping is caused by ion drag which is given by

$$
\frac{1}{\rho} \cdot \sigma (E + U \times B) \times B
$$

where $\sigma$, $E$, and $B$ are the conductivity tensor, the external electric polarisation- and convection field vector and the Earth’s magnetic field vector, respectively. At mid- to high latitudes the ions are influenced by the electric field and, through collisions with the neutral particles, effectively dampen oscillations in the neutral motion. The ion drag term thus depends on the shape of the external high-latitude polarisation field as well as the conductivity tensor, which itself depends on the charged particle densities. These are sensitive to neutral composition, in particular the ratio of atomic oxygen to molecular nitrogen, to winds as well as season, solar- and magnetic activity. As described further down, seasonal changes of ion density in the lower thermosphere were found in CTIM to have a negligible influence on the upwards propagating tides at high latitude, but they do influence strongly the in-situ generated oscillations.

Earlier findings showed that the background temperatures in the atmosphere have an impact on vertical viscous drag. It is useful to examine whether the background winds also have a similar influence, other than that via temperatures. Chapman and Lindzen [1970] found that the effect of background winds on semidiurnal solar tides was insignificant, whereas it played an important role for the lunar tides. Following Booker and Bretherton [1967], the importance of background wind on the tidal propagation can be estimated from the ratio of the wind speed to the horizontal phase speed of the tide. If the phase speed is much larger than the background wind speed the tide is hardly influenced by the wind. For westward propagating waves the zonal phase speed, $C_{ph}$, is given by

$$
C_{ph} = -\frac{\Omega R_E \cos \theta}{s T_p}
$$

where $\Omega$, $R_E$, $\theta$, $s$ and $T_p$ are the Earth’s angular velocity, its radius, the latitude, wave number of the tide and its period (in days), respectively. In Table V.2 average zonal phase speeds are given for semidiurnal solar tides at different latitudes. The values are based on CTIM simulations for the
thermosphere below around 200 km.

<table>
<thead>
<tr>
<th>Latitude</th>
<th>0°</th>
<th>18°</th>
<th>42°</th>
<th>68°</th>
</tr>
</thead>
<tbody>
<tr>
<td>$C_{ph}$ [m/s]</td>
<td>464</td>
<td>441</td>
<td>345</td>
<td>174</td>
</tr>
</tbody>
</table>

**Table V.2: Zonal phase speeds of westward propagating semidiurnal tides at various latitudes**

The table shows that phase speeds at low-to mid latitudes (on either hemisphere) are considerably larger than the zonally averaged winds found in the thermosphere (see also Chapter IV), while at high latitude they can be similar. Therefore, low- to mid latitude tides are unlikely to be influenced by the background winds. This was also confirmed by runs carried out for studies in Chapter IV.

Tidal amplitudes there are shown for equinox conditions, but applying the same lower boundary tidal forcing under solstice conditions was there found to produce almost identical tidal profiles at low- to mid latitudes in spite of the fact that a stronger north-south temperature gradient is present at solstice and generates larger zonal winds. At high latitudes, the effects on upwards propagating tides were minor as well since geomagnetic conditions were assumed identical. Zonally averaged $E$-layer electron densities were found near 70°N to be enhanced in December only to $6.90 \times 10^{10}$ m$^{-3}$ from the June value of $6.75 \times 10^{10}$ m$^{-3}$. At the same time the tidal amplitudes were almost identical below 150 km. Only at higher altitudes did the charged particle densities change more with season and give stronger in-situ forced semidiurnal winds. In summary, therefore, a coupling between background wind and tides does exist at high latitudes but is most likely to be accompanied by changes in the auroral forcing, either through magnetic- and solar activity or in models through different parameterizations of the auroral precipitation patterns and high-latitude convection field.

*Volland* [1988] discussed the interaction of tides and background winds further by addressing the problem of large-scale flow friction. Large-scale flow in the atmosphere, such as that generated by tidal oscillations, interacts with small-scale eddies which are mainly short-periodic gravity waves. These extract momentum from the large-scale flow and convert it into internal energy. Although details of this process are not yet well understood it is often parameterized through a *Rayleigh friction* term

$$F_R = -\Omega \times B \times U$$
where the Rayleigh friction coefficient, $v_r$, is an empirical number which varies with height, latitude, season and period of the wave. It is generally of the same order of magnitude as the Newtonian cooling coefficient. In the mesosphere the Rayleigh friction coefficient increases with height as a result of breaking small-scale internal gravity waves [Volland, 1988]. Rayleigh friction is a function of the mean background wind, $U$, and therefore does provide a means by which the background winds influence tides. Rayleigh friction is not considered in CTIM, which is why tidal profiles were found not to depend on the background winds. Above the turbopause (near 100 km) small-scale eddies are not found any more and therefore Rayleigh friction is unimportant for the height regime of interest in simulations discussed here. A Rayleigh friction term is used in the GSWM (see V.3.1) and other middle atmosphere models, and studies by M. Hagan found that the choice of background winds in the GSWM had considerable influence on the resulting tidal profiles [M. Hagan, private comm., 1995]. The Rayleigh friction parameterization used in that model [Hagan et al., 1993] however gives a peak value for the coefficient at 60-70 km altitude. Above 80 km the term is very small and thus insignificant for the propagation of tides in the regions where they are compared to the CTIM. Therefore, the low- to mid latitude background wind fields of the GSWM as well as the other models used here do not need to be considered when interpreting the comparisons. It is sufficient to take into account only the background temperature.

V. 2.3. ENERGY DRAG

In the previous section the various factors were discussed which influence the tidal momentum drag. This section will discuss the various contributions to tidal energy damping. At heights where no damping occurs the wind- and temperature amplitudes are directly correlated through relations described in Classical Tidal Theory (Chapter II), and the issues discussed in V.2.1 are equally valid for the temperature case. Temperature amplitude profiles can thus be offset vertically in a height frame of reference if the scale height is different or when changing the altitude of tidal forcing.

When temperature is locally enhanced, as through tidal oscillations, various energy redistribution processes begin. If these are effective enough to redistribute an important fraction of tidal energy on the time scale of the oscillations, the tides are damped. The CTIM runs described in Chapter IV found the main energy redistribution in the thermosphere to occur through heat conduction and advection. A further process found in the real atmosphere is referred to as Newtonian cooling and describes radiative cooling at various wavelengths. In the upper thermosphere, Newtonian cooling
is found at 63\textmu m from the fine structure of atomic oxygen (O) but plays a minor role compared with the downward molecular heat conduction. In the middle thermosphere, near 180 km altitude, 5.3\textmu m cooling from nitric oxide (NO) occurs but is also thought to be of secondary importance [Roble, 1995]. Below 140 km the main radiative cooling occurs through 15\textmu m emissions from carbon dioxide (CO$_2$). At present, this cooling process is not considered in CTIM. However, Hagan et al. [1993] found CO$_2$ cooling to be important only for waves with a vertical wavelength larger than 40 km. As found in Chapter IV, the tidal modes dissipating below 140 km are higher order modes only with vertical wavelengths below that value. Therefore, the overall effect of CO$_2$ cooling on energy damping of the tidal oscillations discussed here is thought to be negligible. Various authors such as Chapman and Lindzen [1970] and Lindzen and Goody [1965] also found that Newtonian cooling was of minor importance to solar semidiurnal propagating tides.

The vertical molecular and turbulent heat conduction term in a pressure coordinate frame is given by

$$ g \frac{\partial}{\partial p} \frac{K_m + K_T}{H} p \frac{\partial}{\partial p} T $$

where $K_m$ and $K_T$ are the coefficients of molecular- and turbulent heat conduction, respectively, $T$ is the temperature, $H$ the scale height, $p$ the pressure and $g$ the gravity acceleration. Above the turbopause, $K_T$ is negligible compared with $K_m$ and only molecular heat conduction plays a role [Roble, 1995]. The above expression shows that heat conduction depends on the vertical profiles of molecular- and eddy conduction coefficients as well as scale height. The dependency on scale height leads to a similar conclusion as in the case of vertical viscous drag, with the heat conduction increasing for smaller background temperatures. One sees from the above expression that heat conduction is a function of the vertical temperature gradient, as viscosity was earlier found to be a function of the vertical wind gradient.

Horizontal energy advection was found in Chapter IV to be important for redistributing tidal energy at low latitudes. On a pressure level the term is given by

$$ U \cdot \nabla_p \left[ c_p T + \frac{1}{2} U^2 + \Phi \right] $$

where $\nabla_p$ is the horizontal 2-dimensional gradient operator at constant pressure. It is important to
Validation

note the dependency of this term on horizontal background winds. While these were found in the previous paragraph not to influence the tidal momentum profiles they do influence the temperature profiles. Stronger background winds ensure more effective horizontal energy transport.

Vertical advection of energy is given by

\[ w \cdot \frac{\partial}{\partial p} \left( c_p T + \frac{1}{2} U^2 + \Phi \right) \]

where \( w \) and \( \Phi \) are the vertical wind relative to a fixed pressure level and the geopotential, respectively. The term shows that vertical advection increases with tidal amplitude through the vertical wind, \( w \), and the vertical geopotential gradient. The vertical heat conduction term does not have the same strong dependency on vertical gradients of wind as was found in the case of vertical viscous drag (see V.2.2). It is therefore of interest to examine whether doubling the amplitude of tidal forcing at CTIM's lower boundary would have a different effect on the altitude of energy dissipation than it has on the altitude of momentum dissipation. The latter had been found to be reduced by a half-scale height under the larger amplitudes. A comparison of temperature amplitudes in the same pair of runs used in V.2.2 reveals that temperature amplitude patterns in a pressure coordinate system are almost identical in both runs and differ only in the amplitude values. The altitude of energy dissipation thus depends less on the amplitude of a tide than was found in the case for momentum dissipation. However, the value of the peak amplitude itself does increase with the forcing amplitude. As shown also in Figure V.13, the two are not proportional, and doubling the forcing amplitude will lead to a temperature amplitude rise of around 10-40\%, depending on latitude. This is plausible from results in Chapter IV, that linearity breaks down where dissipation occurs.

V. 3. COMPARISONS OF CTIM WITH OTHER MODELS

V. 3.1. THE MODELS

A range of available numerical models for tidal simulations was described in Chapter I. The models referred to in the following have been described extensively in the literature, and therefore only their basic properties are reviewed.
The NCAR model is in many respects similar to CTIM and therefore particularly suitable for comparisons. The code is available in different versions, the earliest being the Thermosphere General Circulations Model (TGCM) [Dickinson et al., 1981; 1984] which does not calculate self-consistently the ionosphere and uses parameterized charged particle properties. The TIGCM [Roble et al., 1988] is an extension of the TGCM in two aspects, using a self-consistent ionosphere and simulating upwards propagating tides through forcing of its lower boundary at 97 km [Fesen et al., 1986]. Later, the TIGCM code was extended by Richmond et al. [1992] to include self-consistent low latitude electric dynamo field calculations and is in that version referred to as the TIEGCM. This version is used in Figures V.26 - V.31 for comparisons. The most recent version of the code (TIME-GCM) is a downwards extension from the previous lower boundary at 97 km to 30 km altitude, with full treatment of the stratosphere and mesosphere chemistry and dynamics, as described by [Roble and Ridley, 1994]. The version used primarily in the following (TIGCM) is similar to the CTIM in that it calculates the parameters on a pressure coordinate system, uses similar basic equations of momentum, energy and transport and has external tidal forcing implemented in the same manner. Main differences to the CTIM model are the grid resolution (5° latitude by 5° longitude horizontally, half-scale heights vertically), the altitude range (lower boundary height at 97 km) and coefficients describing chemical and diffusive processes.

The tidal model by Forbes and Vial [1989] differs in many respects from the CTIM and TIGCM models. The most important differences are its height range (ground level to 110 km) and the methods of calculation. The model neglects eddy- and molecular dissipation and thereby solves a single second-order partial differential equation in height and latitude for the perturbation geopotential, assuming only linear Rayleigh-friction and a given background atmosphere. This aspect along with only small vertical overlap of the Forbes and Vial model with CTIM make a direct comparison between the two models difficult. The Forbes/Vial model is therefore in this context used only to provide the TIGCM with a set of tidal amplitude- and phase values at its lower boundary.

The GSWM [Hagan et al., 1995] simulates solar diurnal and semidiurnal tides using realistic ozone and water vapour forcing. It is more suitable for direct comparisons with CTIM than the Forbes/Vial model since it not only reaches a higher altitude (125 km) but also considers the full dissipation processes occurring in the lower thermosphere. With this height range the region of overlap with CTIM extends over roughly 7 scale heights (around 45 km). The GSWM is essentially
an updated version of the model by Forbes [1982a] and solves numerically the four coupled partial differential equations for three velocity components and temperature, as opposed to only one equation in the model by Forbes and Vial. Both the original model by Forbes and the updated GSWM require specification of a zonally averaged background atmosphere which varies only with latitude. The background temperature and density in the run presented here are taken from the empirical MSISE90 model [Hedin, 1991], using a solar flux of 120 and magnetic index of Ap=4. The background winds between 20-80 km were calculated from the temperature profiles, assuming geostrophic balance. Above 80 km they were taken from the empirical Portnyagin and Solov'eva [1992 a,b] model.

The HWM [Hedin et al., 1993] is an empirical model of the horizontal neutral wind in the upper thermosphere based on wind data obtained from the AE-E and DE-2 satellites. A limited set of vector spherical harmonics is used to describe the zonal and meridional wind components. The model is available in several versions which differ mostly in their height ranges. The first edition of the model, released in 1987, (HWM87) gave the winds above 220 km only. With the inclusion of wind data from ground-based incoherent scatter radar and Fabry-Perot optical interferometers, HWM87 was extended down to 100 km, giving the HWM90. The HWM90 thermospheric wind model was then revised in the lower thermosphere and extended into the mesosphere and lower atmosphere to provide HWM93, which is a single analytic model for calculating zonal and meridional wind profiles representative of the climatological average for various geophysical conditions. Gradient winds from CIRA-86 and rocket soundings, incoherent scatter radar, MF radar, and meteor radar provide the data base and are supplemented by previous data driven models. Low-order spherical harmonics and Fourier series are used to describe the major variations throughout the atmosphere including latitude, annual, semiannual, and longitude. The HWM93 model represents a smoothed compromise between the data sources. Although agreement between various data sources is generally good, some systematic differences have been noted, particularly near the mesopause [Hedin et al., 1993].

The MSISE90 model [Hedin, 1991] was developed using satellite-borne mass spectrometer and ground-based incoherent scatter radar measurements in the thermosphere. In the lower and middle atmosphere, mesospheric rocket data as well as the tabulations of monthly average temperatures, densities and pressures from the Middle Atmosphere Program (MAP) Handbook 16 [Barnett and Corney, 1985] were used. The MSISE90 and HWM93 models do not normally use data from the
same instruments and therefore are not expected to be entirely consistent with each other. Nevertheless they are here compared to the same CTIM run.

The Tuned Mechanistic Tidal Model (TMTM) model is used only in the comparisons presented in V.5 (Figures V.26 - V.31) and thus described further there.

V. 3.2. THE CTIM RUNS

As mentioned previously, two separate simulations were necessary for the following comparisons. They differ in the tidal forcing and the level of solar activity, the latter being F10.7=75 for the TIGCM comparison and F10.7=120 in the other run. Both are carried out for equinox conditions at magnetic activity of Kp=2+

For comparisons with the TIGCM model the lower boundary forcing in CTIM at 80 km had to be derived from the values used by the TIGCM at 100 km, which are themselves based on values suggested by the model of Forbes and Vial [1989]. The aim was to obtain in the CTIM run at 100 km altitude tidal amplitudes and phases which are similar to those used in the TIGCM model at its lower boundary. To achieve this, the amplification of tidal amplitudes between 80 and 100 km as well as the shift in phase had to be taken into account. It was shown in Chapter IV that tidal amplitudes above 80 km behave differently for the various Hough modes. The simulations with individual tidal modes presented there show that amplitudes of the semidiurnal (2,2), (2,3), (2,4) and (2,5) modes are amplified between 80 and 100 km by factors of around 1.5, 1.5, 1.4 and 1.0, respectively. However, some wave broadening towards the poles also takes place at those altitudes and therefore the latitudinal structure of a mode at 100 km is not the same as that of the same mode at 80 km. It was found that using the above amplification factors in deriving the lower boundary Hough mode amplitudes did not give a satisfactory agreement between the TIGCM and CTIM models at 100 km. The amplitudes at CTIM’s lower boundary were therefore optimized through a method of trial and error to find the best fit. This required a considerable number of CTIM model runs. An alternative method to this would have been to use at CTIM’s lower boundary tidal amplitudes and phases suggested by the Forbes and Vial model for 80 km altitude. However, the reason for not doing so was that this did not lead to an acceptable agreement between the TIGCM and CTIM values at 100 km. Since the aim was to compare these two models with each other rather than each of them individually with the Forbes and Vial model it was decided that finding the best
possible agreement between the CTIM and TIGCM models at 100 km was most important.

In order to match the phases at 100 km the local times of maxima used by the TIGCM run at 100 km were shifted backwards by the amount of phase change which, as found in Chapter IV, is different for each of the Hough modes. The phase changes with height for semidiurnal modes are illustrated in Chapter IV, Fig. IV.27, and the exact values of change when moving from 80 km to 100 km were found to be for the (2,2), (2,3), (2,4) and (2,5) modes -0.8 h, -2.0 h, -3.1 h and -4.0, respectively.

The values of geopotential height amplitudes and phases at 80 km used for the TIGCM comparisons are given in the first row of Table V.3. This CTIM model run is referred to below as Run 1.

<table>
<thead>
<tr>
<th>Modes</th>
<th>(1,1)</th>
<th>(2,2)</th>
<th>(2,3)</th>
<th>(2,4)</th>
<th>(2,5)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TIGCM</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>comparison</td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>(Run 1)</td>
<td>150 m/4.2 h</td>
<td>200 m/2.5 h</td>
<td>10 m/8.2 h</td>
<td>400 m/7.5 h</td>
<td>10 m/7.7 h</td>
</tr>
<tr>
<td>GSWM/</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MSISE90/</td>
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<tr>
<td>HWM</td>
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<tr>
<td>comparison</td>
<td></td>
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</tr>
<tr>
<td>(Run 2)</td>
<td>150 m/8.0 h</td>
<td>62 m/2.9 h</td>
<td>21 m/1.6 h</td>
<td>50.4 m/7.9 h</td>
<td>27 m/2.3 h</td>
</tr>
</tbody>
</table>

Table V.3: Geopotential height amplitudes and local times of maxima used at CTIM’s lower boundary for two comparative model runs. The run for comparison with TIGCM used a solar activity level of \(F_10.7=75\), while the other used a value of \(F_10.7=120\).

The tidal forcing values used at 100 km by the TIGCM run for the (1,1), (2,2), (2,3), (2,4) and (2,5) modes are 400 m (4.0 h), 150 m (1.7h), 10 m (6.2 h), 280 m (4.4 h) and 10 m (3.7 h), respectively, where the local times of maxima are given in brackets. The latitudinal profile of semidiurnal amplitudes produced by CTIM at 100 km, using values from Table V.3 at the lower boundary, does not match entirely that given in the TIGCM run, due to the difficulties outlined earlier. However, the relative proportion of each mode is very similar, with the (2,4) mode being the strongest, roughly twice the amplitude of the (2,2) mode, and the antisymmetric semidiurnal modes being only of minor importance. One problem lay in matching the large (1,1) amplitude used
in the TIGCM run. With an amplification factor of 1.8 between 80 and 100 km the amplitude used in CTIM should be chosen at 222 m. However, this large a value leads to unresolved stability problems in the model and therefore the value was reduced by 35% to 150 m. However, from the results in Chapter IV, section IV.4.1 this should have no influence on the semidiurnal tides. The problem with the upper diurnal amplitude limit is addressed further in Chapter VIII and attributed to the limited vertical resolution in CTIM. In the following, therefore, comparisons are carried out for the semidiurnal tide only.

The second CTIM run (which is referred to as Run 2 below) uses lower boundary tidal amplitudes modelled by the GSWM at 80 km altitude and decomposed by M.Hagan [private comm., 1996]. Since values from this decomposition were provided for the altitudes 78 km and 82 km, the amplitudes and phases used in CTIM at 80 km, given in the second row of Table V.3., are linear interpolations between these two heights. The GSWM run used here is for the month of April. A March run is not yet available but the differences are not expected to be significant [M.Hagan, private comm., 1996].

One sees that the tidal forcing amplitudes in Run 1 and Run 2 differ substantially. The overall forcing is considerably stronger in Run 1, with the semidiurnal symmetric modes being dominant. In Run 2 the symmetric modes are still dominant, but antisymmetric modes are relatively stronger than in Run 1. The local times of maxima of the total perturbation profile at 80 km between the simulations differ by less than 2 hours in the southern hemisphere, but in the northern hemisphere the profile in Run 1 leads by around 4 hours local time poleward of 30° latitude. These examples show that substantial ambiguity lies in the specification of lower boundary tides, emphasizing the need for accurate tidal analysis at the input height before carrying out any comparative studies.

Three northern hemisphere latitudes were chosen for amplitude- and phase height profiles, at 18.0°N, 42.0°N and 68.0°N. These coincide roughly with the latitudes of the incoherent scatter radars at Arecibo, Puerto Rico (18.3°N, 66.8°W) and Millstone Hill (42.6°N, 71.5°W) as well as the three instrument sites of the EISCAT consortium at Tromsø, Norway (69.3°N, 19.2°E), Kiruna, Sweden (69.9°N, 20.4°E) and Sodankylä, Finland (67.4°N, 26.6°E).
V. 3.3. COMPARISONS OF WINDS

Vertical profiles of semidiurnal southward wind amplitudes and local times of maxima are shown in Figures V.3 and V.4 for Run 1 and in Figures V.5 and V.6 for Run 2. In the following, the comparisons are described for the low-, mid- and high latitudes. The results are discussed in section V.3.5.

i) LOW LATITUDE

The upper image in Figure V.3 shows a comparison of semidiurnal amplitudes at 18.0°N, as simulated by CTIM (solid) and the TIGCM (dashed). The CTIM wind amplitude has stronger variation with height, reaching from roughly 5 m/s to 35 m/s, than the TIGCM amplitude which varies only by less than 4 m/s around an average of 20 m/s. Amplitude maxima are found near 100 and 120 km altitude in the CTIM values, while TIGCM amplitudes peak near 110 and 160 km. Above around 240 km the amplitude reaches a constant value in both models which is roughly 18 m/s in the TIGCM- and 6 m/s in the CTIM output. In the zonal direction (not shown here) the variability of amplitude with height is similar in both models, but shifted to lower values in the TIGCM run, where values go from 2 m/s at 100 km to 13 m/s near 125 km and back to around 2 m/s at 200 km. Zonal wind amplitudes in CTIM range from 6 m/s near 110 km to 26 m/s at 120 km and around 16 m/s at 140 km. The CTIM zonal winds reach a constant amplitude of 17 m/s above 240 km, while the TIGCM wind amplitudes increase up to an altitude of around 340 km until they reach their saturation value of around 8 m/s which is, in contrast to the case of meridional wind, smaller than the value in CTIM. A number of local maxima can be found between 80 and 100 km in the CTIM zonal wind amplitudes and lie below the height range of the TIGCM.

The local times of semidiurnal southward wind maxima are shown in Figure V.4 for CTIM and the TIGCM. Both models have almost identical phase values below 150 km and above 320 km. Between those heights the TIGCM wind maxima occur later by up to 2 hours between 180 and 220 km altitude. The models reach their constant southward wind phase value at a different altitude. The CTIM phase is shifted backwards in time more quickly above 150 km than is the case for the TIGCM and therefore reaches constant local time of maximum of around 0:30 h (or 12:30 h) already at around 200 km altitude, while the same happens in the TIGCM only 100 km above that.
Southward wind amplitudes produced by CTIM in Run 2 are shown in Figure V.5 along with values from the GSWM and HWM (see V.3.1 for description of these models). The plot shows that amplitudes are similar at 80 km in all three models but then increase much stronger with height in the GSWM and HWM than in CTIM. Between around 100 and 125 km the CTIM amplitudes reach values of 10 m/s and those of the GSWM and HWM approach 70 and 25 m/s, respectively. The CTIM amplitudes peak near 105 km, 120 km and 155 km with values of around 5 m/s, 10 m/s and 15 m/s, respectively. The GSWM amplitudes continuously rise with altitude and the HWM values have one maximum near 115 km of around 25 m/s which is more distinct than the relatively weak maxima in the CTIM output. Between 120 and 145 km altitude the CTIM and HWM values are roughly the same. In the zonal direction (not shown here) the CTIM wind amplitudes peak at around 5 m/s at roughly 105 km altitude and between 120 and 200 km rise continuously from almost 0 to 15 m/s. The GSWM zonal winds also reach a peak at 105 km altitude with around twice the value of the CTIM peak, then fall off and above 110 km rise again to a peak of 10 m/s at 120 km. The agreement between CTIM and the GSWM is thus much better below 115 km than for meridional winds. Zonal wind amplitudes produced by the HWM are similar to the GSWM values up to 105 km, but then increase much stronger than the GSWM and CTIM winds to a value of around 50 m/s near 120 km. They remain large also above 120 km and do not agree as well with the CTIM values as meridional winds at that height.

The phase plot in Figure V.6 shows that CTIM local times of maxima are up to 2 hours ahead of the GSWM values, but they change with altitude at nearly the same rate. In contrast, the HWM phase at 80 km lies around 2 hours behind that of the other models and then changes with height at a slower rate. Above 130 km the vertical phase change is roughly similar in the CTIM and HWM, although the HWM southward winds peak around 4 hours before the CTIM values. This height regime of similar vertical phase change coincides roughly with the region of similar amplitudes described earlier.

ii) MID-LATITUDE

At 42.0°N the meridional winds produced by the CTIM and TIGCM (Figure V.3, middle plot) agree better than at low latitudes. Although the values at a fixed height differ by up to 20 m/s, the overall vertical patterns are very similar, though shifted upwards for the TIGCM values. One difference between the vertical patterns is that the CTIM winds make two oscillations, thus giving
them two local maxima (at 110 and 160 km), while the TIGCM winds make only one oscillation with one main wind amplitude maximum at 120 km. The saturation wind amplitudes are more similar than at low latitude with values of around 12 m/s and 16 m/s for the CTIM and TIGCM winds, respectively, above 300 km. Zonal wind amplitudes (not shown) behave similar to the meridional winds, with the only difference being that the saturated amplitudes are identical in both models at around 4 m/s above 320 km altitude.

The phases at mid latitude (middle plot in Figure V.4) differ more between the two models than at low latitude. Below around 120 km they have similar values, but then a reversal of the phase propagation is seen in the CTIM winds, producing a phase difference of around 7 hours between the models. This difference decreases with altitude and for the saturated phases above around 240 km remains at a 2 hour phase lead of the CTIM southward winds.

The meridional wind amplitudes produced in CTIM's Run 2 at 42.0°N (middle plot in Figure V.5) are, as expected from the already different amplitudes at the lower boundary (see Table V.3), smaller than in Run 1 and have less variability with height. The amplitudes are also lower than those of the GSWM and HWM. The reason for 12 h amplitudes being lower than in the GSWM is that CTIM does not use semidiurnal forcing modes higher than the (2,5) mode, while the GSWM gives also the (2,6) and higher modes at 80 km. A similar feature in the vertical profiles from all three models is the local amplitude peak near 110 km. Largest amplitude values are predicted by the GSWM (at 65 m/s), followed by the HWM (at 35 m/s) and CTIM (10 m/s). In the zonal direction the CTIM predictions are again smaller than those of the other two models. A local peak is again found near 100 km, but here the HWM peak is higher by around 10 km. The zonal wind amplitudes are again generally largest in the GSWM, but the vertical amplitude growth occurs at a similar rate in the HWM and GSWM.

The vertical phase profiles (middle plot in Figure V.6) show the largest vertical phase velocity in the GSWM values and the smallest in the HWM data. The CTIM phase velocity lies between the two. The phase reversal seen in CTIM values near 140 km altitude is not reproduced by the HWM and GSWM. A similar phase reversal was however seen in the TIGCM values (middle plot in Figure V.4). One interesting feature in Figure V.6 is the almost constant phase between 90 and 100 km altitude. In weaker form this feature is also seen at the low latitude location. It is neither present in any of the other models or in CTIM Run 1. Possible causes for this are discussed in
section V.3.5.

iii) HIGH LATITUDE

At high latitudes the meridional wind amplitudes and phases are very similar below 160 km in CTIM and the TIGCM (lower plots in Figures V.3 and V.4). The shape of the vertical amplitude pattern is almost identical in both models, with a wind amplitude of around 40 m/s near 130 km altitude. Above 160 km the CTIM amplitudes continue to fall towards the saturation value of 25 m/s, while the TIGCM amplitudes reach a minimum at 180 km and then rise again towards 32 m/s above 300 km. The southward wind maxima above 200 km occur around 1 hour earlier in CTIM. The zonal winds (not shown) also follow a very similar vertical pattern, but the amplitude peak near 110 km altitude is smaller in CTIM at 33 m/s, as opposed to 42 m/s in TIGCM. The TIGCM zonal winds after a minimum at 150 km rise with altitude towards a saturation amplitude of 42 m/s, while at the same time the CTIM amplitudes continue to fall after the peak at 110 km and reach their saturation amplitude of 10 m/s above 200 km.

Although CTIM meridional wind amplitudes become larger when moving poleward they are smaller by up to 50 m/s below 120 km altitude when compared to the GSWM values and by up to 30 m/s when compared to the HWM. A wind amplitude maximum is found in the HWM output at 120 km altitude, but neither CTIM nor the GSWM show this feature. Below 110 km the CTIM amplitudes are almost zero and then rise continuously with altitude, reaching values of 65 m/s at 200 km. In spite of these differences in absolute value between CTIM and the other two models the rate of amplitude growth with altitude is more similar in the three models than at lower latitudes. The same applies to zonal winds which are also smaller in CTIM and largest in the GSWM.

The meridional wind phase change with altitude at high latitudes (lower plot in Figure V.6) is generally strongest in the GSWM. The rate above 100 km is roughly similar in CTIM and the HWM, and the phase values of both models are almost identical between 110 and 140 km altitude. Below that, the CTIM meridional wind phase is characterized by a phase reversal between 90 and 100 km, resulting in a shift of the southward wind maxima from 11 hours near 90 km towards later times, by around 2.5 hours. Above this phase reversal the waves are again westward travelling. A weaker phase reversal can be found in the GSWM values near 85 km, but not in the HWM.
V. 3. 4. COMPARISONS OF TEMPERATURE

Vertical profiles of semidiurnal temperature amplitudes and local times of maxima are shown in Figures V.7 and V.8 for Run 1 and in Figures V.9 and V.10 for Run 2. In the following, these are compared for the low-, mid- and high latitudes. The results are discussed in section V.3.5.

i) LOW LATITUDE

Semidiurnal temperature amplitudes at 18.0°N are shown in the upper plots of Figure V.7 and V.9. Compared with the TIGCM, the CTIM temperatures below 180 km altitude are more variable and oscillate around an average of about 12°K with a vertical wavelength of around 40 km. In the height regime below 180 km three local maxima are found in the CTIM temperature amplitudes, while only one occurs near 110 km altitude in the TIGCM run. A similar behaviour has already been seen in the low-latitude wind amplitudes (Figure V.3). As is the case for the winds, the temperatures reach their constant saturation value at a lower height in the CTIM run than in the TIGCM values. Both saturation amplitudes are roughly similar. Temperature phases agree very well below 120 km and at 160 km are around 3 hours behind in the CTIM run. The saturation phases differ by less than an hour at the low latitudes. Compared with the GSWM and MSISE90 the CTIM generally underestimates the semidiurnal temperatures below 130 km by up to 5°K and 15°K, respectively. Above that height they agree well with the MSISE90 values. The CTIM temperature amplitudes furthermore have three maxima near 100, 115 and 140 km altitude which are only partly found in the other models. A maximum near 115 km is also found in the GSWM values, but is around 3 times larger than that in CTIM. As for the case of the winds (Figure V.6), temperature phases agree well in CTIM and the GSWM in terms of vertical slope. Below 100 km, the CTIM phase is more variable than the GSWM phase, but the average values are very similar.

ii) MID-LATITUDE

At 42.0°N the temperature amplitudes generally show a similar vertical shape to that of the TIGCM (middle plot in Figure V.7), but while one main maximum is found in the TIGCM values near 120 km there are two maxima in CTIM, one at 120 km and the larger near 160 km. The amplitude values are larger by up to 5°K in the TIGCM below 140 km and smaller above 200 km.
by around 6°K. The phases agree reasonably well below 200 km but above that the temperature maxima in CTIM are around 2 hours ahead of those in the TIGCM. In Run 2 the CTIM amplitudes below 130 km are again (as for low latitude) smaller in CTIM than in the GSWM and MSISE90, but agree better above that with MSISE90 values with an offset of less than 10°K. A large discrepancy is found near 120 km altitude with the GSWM which suggests a peak amplitude of 55°K, whereas in CTIM it lies near 5°K. The MSISE90 value at that height lies around 15°K. The phases generally agree less than at low latitude. Above 90 km they generally have the same vertical phase speed, but are offset by 5 h and 2.5 h.

### iii) HIGH LATITUDE

At 68.0°N the CTIM temperature amplitudes are larger than the TIGCM values (lower plot in Figure V.7), but their vertical shapes are similar with two local maxima occurring below 200 km. The saturation amplitudes above 280 km are the same in both models. As for the mid-latitude case the temperature phases (lower plot in Figure V.8) agree well below 160 km and above 240 km height are earlier by around 5 h in the TIGCM. The amplitude profile from Run 2 (lower plot in Figure V.9), as for the low- and mid latitudes, has lower values by up to 20°K compared with the GSWM and phases generally agree well in terms of vertical phase speed, although there is an offset of around 3 h local time. Compared with MSISE90 the CTIM values are smaller by around 5°K below 110 km and above that exceed the MSISE90 values by up to around 10°K, although the general profile shapes are similar. Above 150 km the phases (lower plot in Figure V.10) in CTIM and MSISE90 are very similar, while at lower altitudes their profiles are considerably different with discrepancies of up to 6 h.

### V. 3. 5. DISCUSSION OF MODEL COMPARISONS

The comparisons of semidiurnal wind- and temperature amplitudes, as described in the previous sections, are a mixture of partial agreement and have some disagreements between CTIM and the other four models. Overall, CTIM amplitudes compare better to the TIGCM than to the GSWM or HWM/MSISE90 which generally tend to show larger values below 120 km. The CTIM phases often agree better with the GSWM and HWM/MSISE90 than the amplitudes. This paragraph will discuss results in more detail, using the findings of discussions in section V.2.
i) CTIM RUN 1 AND TIGCM

Discussions in this section will show that most of the discrepancies between the CTIM and TIGCM models can be explained by differences in background temperature, low-latitude forcing amplitudes, high-latitude neutral winds and the auroral forcing patterns.

At low- and mid latitudes in Figure V.3 the CTIM profiles are shifted to lower altitudes, compared with the TIGCM profiles, and vertical wavelengths of the oscillations appear larger in the TIGCM run. The apparent difference in vertical wavelength can however be explained by the different scale heights in these runs. A comparison of zonally averaged background temperatures in CTIM and the TIGCM (not shown) reveals that the vertical profiles in the models above 105 km are almost identical in terms of shape, but shifted to lower temperatures by around 90-130°K in the CTIM run, depending on latitude. As a result of this, the scale heights are smaller in CTIM by around 3-5 km between 105 and 200 km altitude, while being similar at 100 km. This, as will be shown, explains most of the differences at 18.0°N in Figure V.3. The upper of the two wind amplitude maxima at 18.0°N in CTIM lies 4 scale heights above 100 km. Assuming a mean difference in scale height of around 3 km in that regime the total scale height-induced offset of that second maximum to the corresponding peak in the TIGCM is around 12 km. The lower two maxima are shifted by around 12 km height as well. Adding these values gives roughly the difference in height of the two upper maxima in the models. Therefore, the height differences of the upper maxima can be attributed to the scale height differences. It thus only remains to be explained what causes the shift of the lower two maxima. This may be due also to scale height differences, but in the lower thermosphere the temperatures in the two models were too similar for a comprehensive explanation via the scale heights. Other factors which may play a role are given in V.2.2.

This rationale assumed intrinsically the finding of V.2.2, that applying tidal forcing at different altitudes also vertically shifts the tides' height of dissipation. One may therefore assume that the oscillations produced by both models have a similar vertical wavelength. Although the vertical shifts in Figure V.3 are sufficiently accounted for by the two described effects, a third is likely to have some importance as well, and that is the difference in amplitudes of the (2,4) mode at that latitude which was inevitable in the attempt to fit tidal forcing amplitudes in CTIM and the TIGCM. Following discussions in V.2.2, this should further lower the height of dissipation in the
CTIM run. The wind amplitude comparisons at 42.0°N (middle plot in Figure V.3) can be explained by the same reasoning as for 18.0°N, with the only difference being that forcing amplitudes are similar in both models and the third factor lowering dissipation heights is thus not important here. Low- to mid-latitude temperatures were found to show more variability with altitude in the CTIM, an effect which is most likely to be caused by the larger horizontal wind gradients in CTIM which cause stronger horizontal energy advection.

The wind amplitudes at 68.0°N (bottom plot in Figure V.3) agree better than at lower latitudes, particularly below 160 km altitude. At high latitudes, in-situ forcing through auroral forcing becomes important, but the amplitude peaks at around 120 km are caused by the upwards propagating (2,4) mode which is the strongest mode used in these runs. This is also confirmed when comparing the lower plot in Figure V.3 to that in Figure V.5, where the (2,4) mode used was considerably weaker and the same amplitude maximum is not found. It remains to be explained why the differences in scale height in the models do not show up as strongly here as at low- to mid-latitudes. A comparison of zonally averaged winds at mid- to high latitudes in both models (not shown) reveals stronger westward flow in CTIM between 120 and 180 km height, with westward speeds of around 15 m/s and 30 m/s in the TIGCM and CTIM, respectively. As shown in V.2.2, ion drag is sensitive to neutral background winds and in the case presented here may compensate for the different scale heights in the models. Zonal wind amplitudes (not shown) show stronger discrepancies in CTIM and the TIGCM, and the temperature amplitudes (Figure V.7) similarly suggest that differences in the background atmospheres of the models are responsible for relative properties of tidal amplitudes. Fesen et al. [1993b] discussed the TIGCM model run used here and found that the semidiurnal amplitudes of the meridional wind were sensitive to the phase relationship between auroral- and other tidal sources. The finding is confirmed by discussions in Chapter IV and the shape of the auroral precipitation pattern is thus likely to influence TIGCM profiles in Figures V.3, V.4, V.7 and V.8 as well.

ii) CTIM RUN 2 AND GSWM, HWM/MSISE90

While in the previous section the discrepancies between CTIM and the TIGCM could be attributed to differences in scale height, low-latitude forcing amplitude, high-latitude background wind and auroral properties the following discussion will show that these are not sufficient to account for differences between CTIM and the GSWM, HWM and MSISE90. Discussions and test runs in
Validation

what follows investigate the influence of parameters such as the coefficients of molecular energy conduction and viscous drag. Although these factors are found to play a role, the observed model discrepancies cannot comprehensively be explained.

Wind- and temperature amplitudes are generally underestimated by the CTIM, but phases change at roughly the same rate with height. Also, comparisons showed that some height profile patterns found in the HWM are successfully reproduced by CTIM. Some of the discrepancies found between the CTIM and HWM as well as MSISE90 can be attributed to the differences in tidal forcing. The CTIM Run 2 used tidal forcing amplitudes which were given by the GSWM, but they were not adjusted to match HWM and MSISE90 values. One may therefore not expect the same quality of match between the CTIM and those two models. This is also confirmed by the phase plots which generally show better agreement between CTIM and the GSWM. As outlined in V.3.1, the GSWM uses zonally averaged temperature profile from the MSISE90 model. In Figure V.11 these are compared to the zonally averaged temperatures in CTIM’s Run 2. At all latitudes the quality of agreement is very similar, with temperatures being overestimated by CTIM below 110 km and underestimated above that height. Discrepancies however never exceed 60°K and are thus smaller than those found previously between CTIM Run 1 and the TIGCM. Therefore, differences of scale height are expected to be of less importance in the Run 2 comparisons. Another factor found to play a role in the Run 1 comparison was the low-latitude forcing amplitude which was higher in CTIM. The same is not found here, and amplitudes generally agree well at 80 km, especially between the CTIM and GSWM. A third factor which was found to play a role in the Run 1 comparison, the high-latitude auroral- and wind patterns, is likely to be important here as well. At mid- to high latitudes below 200 km height the zonally averaged winds produced by the HWM are generally smaller than CTIM values. This is likely to cause important differences between these two models at those latitudes. The CTIM and GSWM do not use the same parameterization of the auroral region and therefore differences found at mid- to high latitudes are likely to be caused partly by this discrepancy. In summary, though, the discrepancies between CTIM Run 2 and the models cannot be explained entirely by the same arguments used in the Run 1 comparison, particularly at low- to mid latitude.

The main problem which needs to be addressed is why the CTIM amplitudes do not grow at the same rate with height as those given by the GSWM, HWM and MSISE90. Some measurements presented later in this chapter suggest that the amplitude increase rates suggested by those models
are not unrealistic and for this reason the issue is discussed further in the following. Generally, the amplitude growth with height is caused by the decrease of density with height in the atmosphere. The density was in CTIM found to decrease exponentially with altitude and this therefore does not appear to play a role in the present discussion.

In order to investigate various other possible factors which influence the momentum- and energy drag terms discussed in V.2, a set of three test runs have been carried out with the CTIM. They investigate the influence of forcing amplitude and the vertical profiles of molecular energy conduction and viscosity on the resulting tidal amplitudes. In the first of these runs, forcing amplitudes were doubled at CTIM’s lower boundary for each of the semidiurnal Hough modes, thus maintaining their relative balance. The motivation for this was to test the sensitivity of the resulting amplitudes on the strength of forcing. A second run used the original forcing amplitudes, but with a modified coefficient of molecular heat conduction, $K_m$. As outlined in Chapter III (section III.3.4), values for $K_m$ are consistent with those by Bauer [1973], with initial conditions taken from the Jacchia [1971] model atmosphere for 1000°C exosphere temperature. In order to investigate the sensitivity to this profile all $K_m$ values were reduced by 50% on each of the pressure levels, thus maintaining the shape of the vertical $K_m$ profile. In the third test run, the original forcing amplitudes and $K_m$ values were used, but the viscosity coefficient, $\mu_m$, was altered in the following way. In its original form the coefficient is calculated from the expression by Dalgarno and Smith [1962] which is given by

$$\mu_m = 4.5 \cdot 10^{-5} \cdot \left( \frac{T}{1000} \right)^{0.71} [kg \ m^{-1} \ s^{-1}]$$

and uses a standard vertical profile of temperature, $T$ (in °K), which is invariant throughout a simulation. In order to make the approach more self-consistent the coefficient was recalculated every 60 minutes in the third test run with the new globally averaged temperature profile, resulting in a change of the vertical gradients of $\mu_m$. This approach is more self-consistent but ultimately has a minor influence only on the results. Furthermore, therefore, the above expression was multiplied by 0.67, or the coefficient reduced by 1/3, in order to reduce the vertical viscous drag and thereby the tidal damping in CTIM. The typical vertical $\mu_m$ profile which resulted from this is shown in Figure V.12 (dashed) along with the original values (solid). The plot shows that the strongest influence of this change is found at pressure levels 2-7, covering the height ranges of between 85 and 125 km which is where most discrepancies were found in CTIM Run 2. Semidiurnal
amplitudes of meridional wind and temperature resulting from these parameter changes are shown in Figure V.13 at 42.0°N. The solid lines are amplitudes from Run 2, dotted curves are values obtained from doubling the forcing amplitudes, the dashed curves show values when halving $K_m$ and the dashed-dotted lines are from the run with modified $\mu_m$. One sees from the plots that none of these three test runs give large enough amplitude growth with height to match the GSWM better as well as HWM and MSISE90. However, they show a series of important features which are described in the following.

Stronger forcing at 80 km (dotted lines) does lead to enhanced amplitudes below around 140 km. Surprisingly, though, amplitudes above that height are reduced by the stronger forcing. This is most likely to be caused by destructive wave interference with in-situ generated waves. As pointed out previously (section V.2.2), the upwardly propagating tides interact with those generated in-situ in the thermosphere. The overall response thus depends on the amplitudes and relative phases of these two components. In-situ forcing was in the simulations here found to be strongest at mid-to high latitudes and it is thus important to consider these interactions at those latitudes. This is confirmed by the CTIM simulations. Comparing the semidiurnal phases in the two model runs (not shown here), they are, as expected, identical at the lower boundary as well as at high altitudes, but between 85 and around 200 km altitude they differ noticeably. At 68.0°N the local times of maxima of semidiurnal southward wind change from 6.6 hours local time (LT) near 105 km altitude to 2.8 hours LT near 190 km in the original simulation, but when using doubled forcing amplitudes the respective values are from 8.6 h to 3.8 h LT. A similar phase shift was found at 42.0°N (not shown here), whereas at 18.0°N (not shown) the phase change was also found, but the high-altitude amplitude decrease was not. While the phase shift in the simulations here has lead to an overall decrease of tidal amplitudes at mid- to high latitudes it can just as well have the opposite effect in other simulations with different phase values at the level of forcing.

Halving the $K_m$ values at all heights (dashed lines in Figure V.13) causes no significant wind amplitude changes at mid latitudes but almost doubles the high-latitude values (not shown) between around 100 and 130 km height. Temperatures are generally affected more than the winds and at 42.0°N one sees that halving $K_m$ has almost the same effect on temperatures below 130 km as doubling the forcing amplitudes.

Using the modified profiles of vertical viscosity (dashed-dotted lines in Figure V.13) generally
Validation

changes the amplitude profiles and in most cases causes larger values. Some of the local maxima in the profiles are enhanced by the new viscosity profile, but others are shifted in height. At 42.0°N the effects of changing \( \mu \) on wind amplitudes are not significant but they are stronger at other latitudes (not shown). No consistent pattern of behaviour can be read from these plots, but they show that vertical amplitude profiles are sensitive to the choice of both \( K_m \) and \( \mu_m \).

In summary, the presented GSWM, HWM- and MSISE90 amplitudes could not be comprehensively matched by the CTIM runs. The various test runs presented have demonstrated the sensitivity of profiles to various externally specified coefficients as well as to the choice of forcing amplitudes. The strength of forcing at 80 km was found to influence the vertical phase progression and at mid- to high latitudes this considerably changes the resulting profiles because of interference between the upwardly propagating- and the in-situ forced tides.

Other evidence suggests that interference among the externally forced upwards propagating Hough modes may also play a role for the overall profile. In Run 2, forcing amplitudes and phases were given by the GSWM values at 80 km. As shown in Table V.3, the forcing consisted of the semidiurnal Hough modes (2,2) to (2,5) which all have amplitudes of similar order, as opposed to the forcing of Run 1 which was dominated by the (2,4) mode, followed by the (2,2) mode and almost negligible contributions from the remaining modes. It was shown in Chapter IV that vertical phase speeds were different for the various Hough modes, and forcing with several modes at similar amplitudes is thus likely to cause destructive interferences at some altitude, depending on the relative phases at the level of forcing. This would be consistent with the finding that amplitudes generally increase stronger with height in Run 1. Although phases generally agreed well in the CTIM and GSWM, small differences could already lead to considerable effects on the total amplitudes, and therefore this potentially offers another explanation for the discrepancies.

V. 4. COMPARISONS OF CTIM WITH DATA

In the remainder of this chapter comparisons are shown between the CTIM and measured data. Three northern hemisphere sites have been chosen for the comparisons, at Arecibo, Puerto Rico (18°N, 67°W), Millstone Hill, USA (42.6°N, 71.5°W) and the European Incoherent Scatter scientific radar system (EISCAT) high-latitude sites at Tromsø, Norway (69.6°N,19.2°W), Sodankylä, Finland (67.4°N, 26.6°E) and Kiruna, Sweden (67.9°N, 20.4°E). Since EISCAT
measurements are combined from all three high-latitude sites these are referred to as one site, with an average latitude of 69°N. The nearest latitudes on the CTIM grid are at 18.0°N, 42.0°N and 68.0°N and will thus be used in these comparisons. Since tides are global phenomena the aim here was to present data at these latitudes which have been measured over the same periods or at similar season. The first of these sets consists of measurements carried out during the ninth Lower Thermosphere Coupling Study (LTCS 9), between 20-30 January, 1993, while the other set are data averaged for equinox conditions. In V.5, a further comparison is shown between the CTIM, measurements and models. There, the CTIM simulations presented in the following are used.

As in any numerical model, the values produced by calculations have any scientific value only at some distance from boundary regions. In the case of CTIM, data should be located at least 2 grid points above the lower boundary, or 2 scale heights. Considering that CTIM’s lower boundary lies at 80 km altitude, any comparisons of tidal amplitudes with other models or data are thus sensible only above 90 km. The ideal height regime for comparing the properties of upwards propagating tides is therefore between around 90 and 200 km; above that regime most tidal oscillations are formed in-situ and hardly depend on the lower boundary forcing. Numerous tidal measurements have been made below 110 km using Medium- and High Frequency (MF/HF) radars as well as Very High Frequency (VHF) meteor radars. Examples of such studies can be found in Manson et al. [1991], Roper et al., [1993] and Deng et al., [1997]. However, the vertical range overlap between such measurements and the CTIM is too small to allow detailed vertical structure comparisons. The measurement of tidal oscillations of the neutral atmosphere above 110 km relies on indirect measurement techniques. Incoherent scatter radars (see also section I.4.3.ii ) are used to obtain the ion drift velocities, and neutral winds are derived from these. Examples for tidal measurements above 110 km using incoherent scatter radars are given by Virdi and Williams [1993], Brekke et al. [1994] and Kunitake and Schlegel [1991]. At high latitudes the level of ionization during nighttime is high enough to allow continuous measurements, but at low-to mid latitudes nighttime ion concentration is too low to allow reliable measurements. Therefore, diurnal tides are more difficult to measure at the low- to mid-latitude sites. Examples of low- to mid latitude semidiurnal tide measurements are presented by Salah et al. [1991, 1994] and, most recently, by Zhou et al., [1997].
V. 4.1. THE DIFFICULTIES WITH DATA COMPARISONS

When comparing tidal measurements to simulations, a number of difficulties arise which strongly influence the quality of agreement. One category of difficulties addresses the quality of the data and another consists of the problems associated with finding the best external forcing profile at CTIM's lower boundary. Where relevant, these difficulties are also mentioned in the data comparisons below, but in the following discussion they are summarized and described in more detail.

As outlined briefly in the previous section, obtaining tidal profiles in the neutral atmosphere above 100 km relies on an indirect measurement technique in which the ion drifts are measured. The correlation between neutral winds and ion drift is controlled mainly by two parameters, the electric field and the ion-neutral collision frequency. In measurements such as those carried out at EISCAT in the CP-2 mode (see V.4.3), the electric field is measured simultaneously, but this is not always the case, as in the CP-1 mode used previously at EISCAT, and an error may thus be associated with the electric field assumed. Secondly, the ion-neutral collision frequency has a degree of uncertainty since it relies on the knowledge of the neutral number densities. One common method of evaluating the ion-neutral frequency is through analytical relations such as that by Schunk and Walker [1973] which is also used in the CTIM (see Chapter III, section III.3.7, equation III.47). The number densities of $O$, $O_2$ and $N_2$ are for the measurements often assumed from empirical models, such as MSISE90 [Hedin, 1991]. Once the neutral velocities have been calculated from the ion drifts, they are binned together into intervals of typically 1 hour. A curve is then fitted to these data points, and amplitudes are extracted by Fourier-decomposing the fitted curve. Producing this curve involves some ambiguity as well, depending on the degree of fluctuation of data points during the measurement period. It is evident, therefore, that the tidal amplitudes derived from ion drift measurements can contain errors.

Furthermore, measured tides are frequently found to show strong variability which is often attributed to interactions of the tides with gravity waves. Such variations were also found in the LTCS-9 measurements presented below [L.Goncharenko, private comm., 1997], and the comparisons here use amplitude values which were averaged over the entire measurement period of 10 days. The short-term variations are not considered in the forcing at CTIM's lower boundary. Since interactions such as those between tides and gravity waves are non-linear, averaging a profile
which has fluctuations (the measurements) will not produce the same result as using a constant, averaged profile for the forcing (in CTIM). Some features may therefore be lost as a result of this averaging. This is even more important in the case of the equinox comparisons presented in V.4.3, which were averaged at 42°N over several equinoctial LTCS campaigns and at 69°N over days close to equinox. The ambiguity associated with averaging over periods in different years under different solar- and geomagnetic conditions is even larger.

It was found in the comparisons presented below that the quality of agreement between CTIM and measurements was very sensitive to the lower boundary forcing profile. As was described earlier in the CTIM-TIGCM comparisons, one difficulty generally lies in finding an optimum forcing profile if a global amplitude profile is known for a height other than that of CTIM’s lower boundary. In the case of the TIGCM, a global amplitude profile was given at 97 km, and this difference of height to CTIM’s lower boundary, at 80 km, made it difficult to “interpolate down” the amplitudes at 97 km. The main reason for this was the wave broadening, or propagation to higher latitudes, which gave each of the Hough modes a different latitudinal structure when they propagated from 80 km to 97 km height. Even when carrying out a Hough mode decomposition of the TIGCM profile at 97 km and interpolating down the amplitudes of each mode to 80 km, the profile produced by CTIM at 97 km would not entirely match that of the TIGCM. In principle, a similar problem is faced with the data, but another difficulty arises which further complicates the matter. Although at a different height from CTIM’s lower boundary, the TIGCM did produce a global profile of tidal amplitudes. Measurements of tidal amplitudes are to-date not available globally, but for a limited number of sites only, most of which are generally located in the northern hemisphere since more facilities are operated there. The global tidal profile produced during LTCS-9 (see V.4.2.i) is an example for one of the most comprehensive simultaneous tidal measurements currently available. The amplitudes of winds were during the campaign measured for 90 km altitude at 12 different latitudes, mostly in the northern hemisphere. As a result, considerable flexibility lay in fitting a global latitudinal amplitude curve through these points. Therefore, a large series of test simulations were carried out with CTIM, all of which produced an acceptable match to the measured profile at 90 km altitude, but varied considerably in their vertical behaviour at various latitudes. From these, the best fitting simulation is presented below. In practise, this process is time-consuming since each simulation requires around 24 hours of run time, corresponding to around 6 days of real time, to stabilize sufficiently. When comparing tidal simulations to measurements, therefore, any discrepancies do not necessarily express limitations
of the model but may just as well be linked to any of the described factors.

V. 4. 2. THE LTCS 9 CAMPAIGN

In the following, measurements carried out during 20-30 January, 1993 at the low- and mid-latitude sites of Arecibo and Millstone Hill, respectively, are compared to a CTIM run simulating tides under similar conditions. The data presented for the Arecibo site were provided by Q. Zhou [private comm., 1997] and are described in more detail by Zhou et al., [1997]. The Millstone Hill data were provided by L. Goncharenko [private comm., 1997], and the measurement techniques used for these data are described by Salah et al. [1991] and Salah [1994].

i) THE CTIM RUN

One of the important challenges in obtaining good agreement between measured and simulated tidal profiles is to specify the correct lower boundary forcing in CTIM. There is a large degree of flexibility in doing so since not only does the absolute amplitude at any particular location matter, but so also does the balance of Hough modes and their relative phases, and thus the global tidal profile. This already proved a difficult task in the earlier comparisons with the TIGCM model, and is even more difficult here since accurate Hough mode decomposition is only possible if amplitudes are known at many globally distributed latitudes for the same altitude. Whether any semidiurnal amplitude at a fixed latitude is produced by the (2,2), (2,3), (2,4) or (2,5) mode, or any combination of these, has a considerable influence on how the wave propagates vertically, and the main difficulty in carrying out the CTIM comparisons with data lies in finding the best combination of Hough modes. A global distribution profile of semidiurnal amplitudes at 90 km altitude for the LTCS 9 period was derived by Deng et al. [1997] from measurements at 12 different latitudes, of which only 4 lay in the southern hemisphere. This gives uncertainty regarding the asymmetric Hough modes, and furthermore the lack of measurements above 60°N gives ambiguity about the contribution of higher order modes, such as the (2,4) and (2,5) modes, which have strong wind components also above 60°N. However, their amplitude profile is well supported by measurements between around 20° and 60°N. Following Deng et al., semidiurnal amplitudes of zonal wind during LTCS 9 at 90 km are around 2 m/s near 20°N, rise to values of around 18 m/s near 42°N and then fall again to around 5 m/s near 50°N. The amplitudes of meridional wind at 90 km have a similar global shape, but values at mid latitudes are larger by around 10 m/s than the zonal
components. It was found that the height profiles of amplitudes were best matched by CTIM if this latitudinal profile was maintained in the model, but with on average 60% stronger amplitude values. The lower boundary forcing in CTIM was specified as follows.

<table>
<thead>
<tr>
<th>Mode</th>
<th>(1,1)</th>
<th>(2,2)</th>
<th>(2,3)</th>
<th>(2,4)</th>
<th>(2,5)</th>
</tr>
</thead>
<tbody>
<tr>
<td>amplitude [m]</td>
<td>150</td>
<td>200</td>
<td>135</td>
<td>240</td>
<td>200</td>
</tr>
<tr>
<td>phase [h LT]</td>
<td>5.5</td>
<td>6.7</td>
<td>7.3</td>
<td>7.8</td>
<td>10.0</td>
</tr>
</tbody>
</table>

Table V.4: Amplitudes and local times of maxima of tidal forcing at CTIM’s lower boundary for comparisons with measurements from the LTCS 9 campaign. The amplitudes are perturbations of geopotential height of pressure level 1 (1.0376 Pa).

The simulation was carried out for late January, using a constant solar flux index of F10.7=106 and a magnetic activity index of 2°. These were derived by averaging the solar- and magnetic flux values found during the measurement period [Deng et al., 1997] which fluctuated over F10.7=101-110 and Kp=0°-5°, respectively. Since the amplitude data used here were also averaged over the period this was assumed to be the most appropriate choice. Some ambiguity lies in the phases of Table V.4 which were chosen primarily to obtain the best possible amplitude fit. The main interest of the following comparisons lies therefore in the vertical amplitude shapes as well as the phase progression.

**ii) WIND- AND TEMPERATURE COMPARISONS**

The data and model output for the LTCS 9 comparisons are given in Figures V.14 - V.20. These are subdivided into the zonal- and meridional semidiurnal wind amplitudes (Figures V.14, V.15), the semidiurnal temperature amplitudes (Figure V.16) and plots of the respective phases (Figures V.17-V.19). Figure V.20 shows a comparison of background winds- and temperature at the Millstone Hill site. Background parameters were not measured successfully during LTCS 9 at Arecibo [Q.Zhou, private comm., 1997] and are thus not given here.

The figures show partial agreement between the CTIM simulation and measurements. The zonal winds at Arecibo (Figure V.14, top) are underestimated by CTIM, but vertical amplitude shapes
Validation

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are similar. A maximum of amplitudes is found near 110 km altitude in the data as well as model output, with peak values of around 33 m/s and 13 m/s, respectively. The meridional winds at Arecibo agree better in terms of value (Figure V.15, top) which are of similar order. Below 110 km the CTIM meridional wind amplitudes are around 10 m/s larger than the measured values, and between around 110 and 140 km they are smaller by a similar amount and less. The fact that modelled wind amplitudes are smaller in the zonal direction may be linked to the fundamental properties of Hough modes. The lower boundary forcing values of Table V.4 show that the (2,4) mode is the dominant forcing mode in this CTIM simulation. It was found in Chapter IV that the overall wind oscillations produced by the (2,4) Hough modes were smaller in zonal- than in meridional direction (Figure IV.18). The reason for this is that the (2,4) Hough mode produces smaller zonal- than meridional wind amplitudes between -30° and +30° latitude (see Figure IV.4). The effect is not as strong here as in Figure IV.18 since other Hough modes are used as well in the forcing which produce more isotropic wind amplitudes than the (2,4) mode. At 42°N the zonal wind amplitudes in CTIM are more like the meridional ones. Figure V.15 shows that the CTIM wind amplitudes agree reasonably well with measurements in the meridional direction also at 42°N where the general shape of the measured vertical amplitude curve is reproduced by the CTIM. The measurements show a maximum near 115 km (with a secondary peak near 105 km lying within the error bars and thus representing no clear feature), while in the CTIM output one maximum is also found, but just below 100 km. A meridional wind minimum found in the measurements is successfully reproduced by the model and lies only around 5 km higher than in the data. The zonal wind amplitudes at 42°N (Figure V.14, bottom) are very different in the model and measurements.

Generally, the different qualities of agreement in the meridional- and zonal directions can either be caused by the non-isotropic background atmosphere or by non-isotropic upwards propagating oscillations. From Figures V.14 and V.15 it is not evident which of these are relevant to the discrepancies found. In Figure V.20 the background winds- and temperature are shown for the same measurements and simulation. The agreement of winds is very good in terms of profile shape. Both in the meridional- and the zonal direction the main features of the measured vertical profiles are reproduced by CTIM. In both directions more detailed features in the measurements are not reproduced, such as the eastward wind minimum near 130 km and the southward wind minimum near 120 km. The CTIM values at some heights lie within the error bars of measurements, but generally the eastward winds are weaker in CTIM and southward winds are stronger, thus giving a background wind field which is more southward oriented. The background temperatures are
above 110 km lower in CTIM by up to around 70°K, but the vertical gradients above 120 km are the same. The temperature- and associated scale height differences account for most of the height differences of the features found in the comparisons, as in the CTIM/TIGCM comparisons described in V.3.5.i). The wind field differences are however, with the findings of V.2.2, unlikely to cause the observed zonal wind amplitude differences at 42°N. The phase plots in Figures V.17 and V.18 surprisingly show better agreement of the vertical phase change in zonal- than in meridional direction at 42°N. In the meridional direction the CTIM rate of phase change with height is generally larger, and in the vicinity of 120 km height a sudden reversal of the phase propagation indicates strong damping. The reversal is also seen in measurements, but is considerably weaker (Figure V.18, bottom). The different vertical phase speeds indicate either that the vertical wavelengths of the oscillations are smaller in the CTIM run than in the measurements, or that stronger damping occurs in CTIM. In the zonal direction the rate of phase change with height is however very similar in the data and simulation, suggesting either that the vertical wavelengths are different, but this discrepancy is compensated for by other damping rates, or that these two factors roughly agree in the model and real atmosphere. Considering that the discrepancies are largest in the zonal direction at mid-latitude one factor which is likely to play an important role is the ion drag term which is controlled by the relative phase of external tides and auroral forcing. The eastward wind phases at 42°N are offset by around 3 h LT in the model and data, and this may play a crucial role for the overall profile. Uncertainties remain, though, and cannot fully be explained from the available information. Overall, the CTIM simulation did manage to reproduce some of the measured semidiurnal wind patterns and discrepancies are caused mainly by uncertainties in CTIM's tidal forcing amplitudes, -modes and -phases. As mentioned earlier, phenomena such as small-scale oscillations are not modelled in CTIM and may also be important.

Temperature amplitudes (Figure V.16) are generally underestimated in CTIM by around 10-30°K at 18°N. At 42°N the discrepancies are larger and of a similar nature as that found for the zonal winds. However, the low- latitude vertical patterns agree well in shape, with a main maximum found near 120 km both in the model and data. The discrepancy in amplitude at 42°N is, as in the case of zonal winds, accompanied by much better agreement of the phases (Figure V.19) which propagate with height at roughly an equal rate, although being offset by around 2 h LT. Amplitude discrepancies might be explained in a similar manner as for the zonal winds, considering that the phase difference between external tidal forcing and solar- as well as auroral heating matters for the resulting profile. At low latitude the phases differ substantially, being offset by up to 6 hours below
120 km and propagating upwards at a slower rate in the CTIM. This however appears not to have a large influence on the amplitude behaviour which agree much better. As outlined earlier, the auroral forcing plays a very minor role to semidiurnal tidal profiles at low latitudes, and the phase sensitivity is thus not as important as at mid- and high latitudes.

Generally, it must be emphasized that the reasonable match between the model and data in many amplitude features tells us that the physical processes appear to be reproduced fairly accurately by the model. The sometimes large discrepancies either in absolute value only or vertical shape (especially zonal winds and temperature at mid-latitude) may be due either to the difficulty in finding an accurate forcing for CTIM or they may be due also to errors in the data. Although the CTIM amplitudes often lie outside the error bars shown here, the tides measured at Millstone Hill in particular fluctuated considerably over the measurement period (L. Goncharenko, private comm., 1997). Many features are expected to be lost by averaging parameters such as solar- and magnetic activity, and furthermore applying one uniform profile of tidal forcing at CTIM’s lower boundary for the entire period. The discrepancies found therefore do not necessarily express limitations of the model but rather that the simulations are incomplete with respect to the strongly fluctuating background conditions which for simplicity were averaged. Future studies could address the extent of discrepancy caused by such averaging. A further reason for discrepancies is likely to lie in the fact that many small-scale phenomena such as gravity waves are not simulated by the model but play an important role in the real atmosphere. The interaction of gravity waves and tides is assumed to be significant but is still poorly understood [Forbes, 1995].

V. 4. 3. EQUINOX MEASUREMENTS

A second CTIM simulation has been carried out for equinox conditions and will be compared with measurements at mid- and high latitudes. In contrast to data presented in the previous section, the values used here were not measured during one particular campaign but are seasonal averages.

i) THE CTIM RUN

The mid-latitude data from Millstone Hill were averaged over the equinox LTCS campaigns 1, 5, 7, 12, 13 and 14 which were carried out between 1987 and 1996. The average levels of solar- and magnetic activity over these measurements are F10.7=121 and Kp=2.4, or Kp=2+.

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campaigns were at spring equinoxes and the other three at autumn equinoxes. Spring- to autumn differences were not larger than the campaign- to campaign differences (L. Goncharenko, private comm., 1997). The peak amplitudes of zonal winds at spring lay within the range of 80-95 m/s, between altitudes 109-118 km, with phases varying by more than one hour, particularly above 115 km. The eastward wind amplitude peaks at fall were in the range of 70-80 m/s with heights of maximum between 115-125 km and phases varying by 1-2 hours. The meridional winds were more variable, with amplitude maxima reaching from 35 m/s to 90 m/s and phase differences of up to 3 hours.

High-latitude data are taken from Virdi and Williams [1993] and consist of EISCAT measurements carried out between 1987 and 1990. These measurements were made using the Common Programme Two (CP-2) mode, as opposed to the CP-1 mode which was used for similar measurements in Williams and Virdi [1989]. In the latter, the Tromsø antenna was fixed pointing along the magnetic field line. Estimates of ion velocity were then based on tristatic measurements at four heights in the E-region which were carried out by the Kiruna- and Sodankylä antennae by scanning up and down the Tromsø beam in a 10-min cycle. The main disadvantages of this technique were that velocities in the E-region were not measured simultaneously with electric field measurements in the F-region, that the observation times of each antenna were too short to allow reliable nighttime measurements and, thirdly, that ion velocities were given at four E-region heights only which were about 10 km apart, which is not enough for more detailed tidal analyses. These limitations were overcome in the CP-2 mode where the Tromsø antenna scanned through four different pointing directions in a 6-min cycle and the two remote antennae tracked the Tromsø beam at 279 km altitude to provide continuous tristatic measurements of the electric field. This is essential for determining the exact E-region velocities since the high-latitude electric field was found to often vary considerably over periods shorter than 6 min. More detailed descriptions and discussions of the CP-2 technique are given by Virdi and Williams [1993]. The data used here were averaged over the measurement days listed in Table V.5.

<table>
<thead>
<tr>
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</thead>
<tbody>
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<td>April 11,12,13,14,15</td>
<td>February 14,15</td>
<td>October 24,25</td>
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</table>

Table V.5: Measurement days used for the equinox-averaged data from EISCAT (see Virdi and Williams [1993]).
In order to obtain the corresponding background conditions for the model simulation the numbers of days separating dates in Table V.5 from the corresponding spring- or autumn equinoxes were evaluated, averaged and then added to the autumn equinox date. CTIM does at present not distinguish between the equinoxes and therefore the spring equinox date could just as well have been used as initial reference date. The described averaging lead to a simulation date of day 284, or October 11. The averaged solar activity was F10.7=155 and the magnetic power index was Kp=2°. Since solar- as well as magnetic activity varied considerably during the measurement period this averaging is expected to introduce inaccuracies. The simulation was primarily fitted to the EISCAT measurements in terms of solar- and magnetic activity as well as simulation day. However, it was found that the used values matched well also to the averaged Millstone Hill data set, and therefore the same simulation is compared to both data sets. Unlike the LTCS 9 comparisons carried out in the previous section, no global profile of tidal amplitudes was available for the present comparisons, leading to even more ambiguity in the tidal forcing specification than for the LTCS 9 comparisons. However, the fact that the averages at Millstone Hill and EISCAT were taken over different periods in time (1987-1996 and 1987-1990, respectively) suggests that the amplitude profiles are not necessarily matched by a single fit. It was thus decided to give priority to the EISCAT data and try to obtain the best possible fit to those. The lower boundary tidal forcing used in for the model run here are given in Table V.6.

<table>
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<th>(2,4)</th>
<th>(2,5)</th>
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<td>0</td>
<td>600</td>
<td>200</td>
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<tr>
<td>phase [h LT]</td>
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<td>n.a.</td>
<td>n.a.</td>
<td>4.4</td>
<td>7.7</td>
</tr>
</tbody>
</table>

Table V.6: Amplitudes and local times of maxima of tidal forcing used in CTIM for equinox comparisons with EISCAT and Millstone Hill data. The amplitudes are perturbations of the geopotential height of pressure level 1 (1.0376 Pa).

The table shows that the diurnal- as well as low-order semidiurnal modes were set to zero. The reason for this is that insufficient information was available to specify these.

ii) WIND COMPARISONS

Measured and simulated wind amplitudes are shown in Figures V.21 and V.22, while the phases are given in Figures V.23 and V.24. It is evident from the amplitude plots that the agreement
between the model and measurements is generally better at high latitudes. In the zonal direction (Figure V.21) the amplitude maximum between 100 and 110 km altitude is reproduced by the simulation, with peak amplitudes of around 10 m/s less than in the measurements. The minimum in the data near 125 km altitude is also successfully reproduced by the simulation and agrees well with the data in terms of value. At 100 km CTIM overestimates zonal wind amplitudes by around 15 m/s. The meridional wind amplitudes (Figure V.22) at the same latitude agree less well. In the data a maximum is found near 105 km altitude, in the simulation it lies at 120 km. At that height there is an amplitude minimum in the data which in the simulation lies around 35 km higher. However, these extrema in both cases agree well in terms of value. When looking at the high-latitude phases (Figures V.23 and V.24) one sees that in zonal direction they are behind in the simulation by about three hours, while in meridional direction they lead by the same amount over measured phases. In eastward direction the vertical rate of phase change is reproduced well by CTIM and only above 120 km the discrepancies become larger. In the southward direction the measured phases are more variable with height and reverse twice, between around 115 and 120 km as well as 125 and 130 km. A sharp phase change is seen above 130 km in the data, suggesting sudden onset of damping. The measured phases do not reproduce any of these features but change with height at an overall similar rate as the measured phases. The good match of amplitudes in the zonal direction suggest that the forcing is roughly correct, and the discrepancies in the southward direction may be due to the background atmospheric properties. Figure V.25 shows that background winds in CTIM at 42°N are too strongly westward and generally too weak in meridional direction. It is evident from the phase plots that the southward wind amplitude changes with height are influenced in the real atmosphere by aspects not reproduced in the CTIM. These may, as outlined on previous occasions, include interactions with gravity waves which are still poorly understood and not modelled by the CTIM. Discrepancies of phases may just as well play a role, and from the plots presented here no consistent evidence is found identifying the central factor causing the meridional discrepancies.

At mid-latitudes the measured amplitude behaviour is generally not reproduced well by the CTIM, and there is mainly a lack of features in the simulated profiles. A strong amplitude peak is found in measured meridional wind amplitudes near 130 km height which is not found in the CTIM amplitudes. In the zonal direction the measured peak near 125 km is not found in CTIM, but a peak with similar amplitude value is found there around 20 km below. The phases at mid latitude agree well between 100 and 110 km in both directions, but generally the CTIM values change faster with
height. This is most likely to be caused by the strong abundance of high order Hough modes in the CTIM run (see Table V.6), while in the real atmosphere the lower order modes appear more important. The mid-latitude amplitude differences may thus also be caused to large extent by the different vertical wavelengths of measured and simulated oscillations. The result illustrates very well the importance of finding an accurate global shape for the lower boundary forcing through the correct choice of Hough modes and their relative balance.

V. 5. A COMBINED MODEL-DATA COMPARISON AT 42°N

In Figures V.26 - V.31 a series of comparisons of semidiurnal tides are shown for the Millstone Hill site at 42°N between the CTIM, measurements (denoted MHO), the GSWM and TIEGCM models (see V.3.1) as well as another model, the Tuned Mechanistic Tidal Model (TMTM). The latter is an empirical model, using combined data from the UARS satellite measurements and other numerical tidal models throughout the thermosphere. At the time of writing, though, the model is in an early stage of development and details have not yet been published. An introduction into the general properties of an earlier version of this model is given in Khattatov et al., [1997]. The plots were produced by J.Salah and L.Goncharenko (private comm., 1997). The CTIM- and the MHO values are the same as those presented in V.4.2 and V.4.3, and the purpose of showing Figures V.26 - V.31 is to give an impression how other tidal models perform under similar conditions. A detailed interpretation is thus not given here; the CTIM results are discussed in V.4.2 and V.4.3. The plots show amplitudes and phases for the 2 equinoxes and 2 solstices, but CTIM values have not been produced for the summer solstice. The equinox plots both use the same CTIM values.

One sees from the plots that CTIM amplitudes and phases are generally within the range of amplitudes produced by the other models. For the phases, there is often an offset, but the vertical phase speed is generally reproduced very well. The offset of phases may play an important role in the discrepancies of amplitudes, as mentioned earlier on numerous occasions. One sees that the CTIM wind amplitudes are often smaller than the data (MHO) and output from the GSWM model, which is an issue already discussed above. Curiously, the TIEGCM has the similar problem, and its amplitudes show weak vertical variability only. The northward- and eastward winds during winter (LTCS-9) are similar in many aspects in CTIM and the TMTM. Generally, it must be emphasized that the tidal profiles in CTIM very much depend on the lower boundary forcing profiles.
V. 6. CONCLUDING REMARKS

Previous attempts to fit tidal simulations of CTIM to data were presented by Fuller-Rowell et al. [1991] and Parish [1989]. The studies by Fuller-Rowell et al. are comparable to those carried out here since data from the first LTCS campaign at EISCAT, Millstone Hill and Arecibo were used. In those comparisons the CTIM largely underestimated measured amplitudes and the vertical profile shapes were in most cases not reproduced well. The quality of agreement is thus considerably better in the simulations carried out here, and this may be due largely to the inconsistencies of the method of external tidal forcing which was used then.
FIGURES CHAPTER V
Figure V.3: Semidiurnal amplitudes of meridional winds, as simulated by CTIM (solid) and the TIGCM (dashed) at equinox conditions.
Validation

\[ \text{lat} = +18^\circ \]

\[ \text{lat} = +42^\circ \]

\[ \text{lat} = +68^\circ \]

Figure V.4: Local times of maxima of semidiurnal southward winds, as simulated by CTIM (solid) and the TIGCM (dashed) at equinox conditions.
Figure V.5: Semidiurnal amplitudes of meridional winds, as simulated by CTIM (solid), the GSWM (dashed) and the HWM (dashed-dotted) at equinox conditions.
Veil ionion Chapter

\[ \text{lat} = +18^\circ \]

\[ \text{lat} = +42^\circ \]

\[ \text{lat} = +68^\circ \]

Figure V.6: Local times of maxima of semidiurnal southward winds, as simulated by CTIM (solid), the GSWM (dashed) and the HWM (dashed-dotted), at equinox conditions.
Figure V.7: Same as Figure V.3, but for temperatures
Figure V.8: Same as Figure V.4, but for temperatures.
Figure V.9: Semidiurnal amplitudes of temperatures, as simulated by CTIM (solid), the GSWM (dashed) and MSISE90 (dashed-dotted) at equinox conditions.
Validation

Figure V.10: Local times of maxima of temperatures, as modelled by CTIM (solid), the GSWM (dashed) and the MSISE90 model (dashed-dotted) at equinox conditions.
Figure V.11: Zonally averaged temperatures for March at $F10.7 = 120$, $Ap = 4$, as modelled by CHIM (solid) and MSISE90 (dashed).
Figure V.12: Vertical profile of the molecular viscosity coefficient used originally in CTIM (solid) and in its modified form (dashed). The original value is based on the expression by Dalgarno and Smith [1962], using a standard global vertical temperature profile. In its modified version the coefficient is recalculated every 60 minutes, using the actual globally averaged temperature profile, and furthermore the expression by Dalgarno and Smith is multiplied by 0.67.
Figure V.13: Semidiurnal amplitudes of temperature and meridional winds at 42.0°N, as simulated by CTIM in four different runs: 1) Run 2 (solid); 2) same as 1), but with doubled semidiurnal forcing at 80 km (dotted); 3) same as 1), but using a halved molecular heat conduction coefficient (dashed); 4) same as 1), but using the modified molecular viscosity coefficient of Figure V.12 (dashed-dotted).
Figure V.14: Semidiurnal amplitudes of zonal wind, as measured during the LTCS-9 campaign on January 20-30, 1993 at Arecibo (top) and Millstone Hill (bottom). Dashed lines are amplitudes simulated by the CTIM for similar conditions.
Figure V.15: Same as Figure V.14, but for meridional winds.
Figure V.16: Same as Figure V.14, but for temperatures.
Figure V.17: Local times of semidiurnal eastward wind maxima, as measured at Arecibo (top) and Millstone Hill (bottom) during January 20-30, 1993. Dashed lines are the phases simulated by the CTIM for similar conditions.
Figure V.18: Same as Figure V.17, but for southward winds.
Figure V.19: Same as Figure V.17, but for temperatures.
Figure V.20: Background values of temperature, eastward- and southward winds during January 20-30, 1993, as measured at Millstone Hill. Dashed lines are simulations by the CTIM under similar conditions.
Zonal 12h at 42 N

Zonal 12h at 69 N

Figure V.21: Seasonally averaged semi-diurnal zonal winds for equinox, as measured at Millstone Hill (top) and Eiscat (bottom). Millstone Hill measurements are averages over the equinox campaigns LTCS 1, 5, 7, 12, 13 and 14. Eiscat values are averaged over measurements carried out between 1987 and 1990. Dashed lines are the amplitudes simulated by CTIM for similar conditions. Millstone Hill data were provided by L.Goncharenko [private communication, 1997] and Eiscat data are taken from [Viridi and Williams, 1993].
Figure V.22: Same as Figure V.21, but for meridional winds.
Figure V.23: Same as Figure V.21, but showing local times of maxima of semidiurnal eastward winds.
Figure V.24: Same as Figure V.23, but for southward winds.
Figure V.25: Background values of eastward- and southward winds, as measured at Millstone Hill during the same period as data in Figures V.21-24. Dashed curves are background winds simulated by CTIM under similar conditions.
Figure A.26: Semidiurnal amplitudes of temperature at 5°N as measured by
the Millstone Hill Observatory (MHO) and simulated by the CTM-CSEM.

The CTM model compared to MHO observations.

Top left: Spring (March) 1957.
Top right: Summer (June) 1957.
Bottom left: Autumn (September) 1957.
Bottom right: Winter (December) 1957.
Figure V.27: Semidiurnal phases of temperature at 42°EN, as measured by the Millstone Hill Observatory (MHO) and simulated by the CTIM, GSWM, TIEGCM and TMTM for solstices and equinoxes. [J. Salah and L. Goncharenko, private comm., 1997].
Figure V.28: Same as Figure V.26 but for eastward winds.
Figure V.26: Same as Figure V.27, but for eastward winds.

Time of Max East Wind

- Spring
- Summer
- Fall
- Winter

Altitude (km)

EST (hr)
Figure V.30: Same as Figure V.26, but for northward winds.
Figure V.31: Same as Figure V.27, but for northward winds.
CHAPTER VI. TIDES AND COMPOSITION

ABSTRACT

This chapter investigates mechanisms by which tidal oscillations in the lower thermosphere can change its composition. In order to investigate whether diffusive- or chemical redistribution of constituents occur as a result of upwards propagating tides, a distinction is made between vertical motion of pressure levels and vertical motion relative to pressure levels. It is shown that only the latter disturbs diffusive equilibrium, while the former changes composition at a fixed height but not on a pressure level, causing no departure from diffusive equilibrium. An estimate of diffusion times shows that molecular diffusion in the lower thermosphere cannot restore diffusive equilibrium on the time scale of diurnal and semidiurnal tides. Chemical lifetimes are shown to be short enough only below 90 km to cause chemical redistribution on the time scale of tides, once the chemical balance is disturbed. It is shown that diffusive balance is affected also by background winds in regions of tidal dissipation if the horizontal winds change sufficiently over a vertical height regime of around one scale height. Large-scale transport of constituents by winds may also be affected in regions where the tides influence background winds. Theoretical predictions are accompanied by simulations with the CTIM. These CTIM simulations also illustrate the nature of tidal temperature oscillations. The main processes generating temperature oscillations are found to be adiabatic heating and cooling which occur as a result of up- and down welling relative to levels of fixed pressure.

VI. 1. COMPOSITION OF THE UPPER ATMOSPHERE

As outlined in Chapter I, the Earth's atmosphere is divided with respect to composition into two regions which are separated by the turbopause at around 100 km altitude. Below the turbopause, all atmospheric gases are mixed, while above it the eddy (turbulent-) mixing disappears and constituents are separated. Each constituent will approach hydrostatic equilibrium, where the partial pressure of a species is balanced by its own gravitational weight. The vertical distribution of each constituent is then in diffusive equilibrium, implying an exponential fall-off of the particle densities with the individual scale heights, \( H_i = \frac{RT}{M_i g} \), which depend on the individual molecular masses, \( M_i \), where \( i \) are species such as O, O\(_2\) and N\(_2\). Figure VI.1 shows the globally averaged vertical
distribution of \( O, O_2 \) and \( N_2 \), as modelled by CTIM. The plot shows that atomic oxygen, \( O \), (solid curve) is the main neutral constituent above around 180 km altitude, while below that molecular nitrogen, \( N_2 \), (dotted) has the largest concentration. As was also outlined in Chapter I, the stronger abundance of molecular nitrogen compared with molecular oxygen at lower heights is caused by the fact that molecular oxygen is dissociated into atomic oxygen, while molecular nitrogen, due to its much smaller photo-dissociation cross-section, is affected to a lesser extent by the UV and XUV radiation. One sees from the plot that constituents are already partly separated in the vicinity of the turbopause and down to CTIM's lower boundary height at 80 km.

When trying to identify what mechanisms upset the diffusive- and chemical balance of the thermosphere and ionosphere, *Rishbeth et al.* [1987] suggested vertical flow, changes in photochemistry, vertical components of centrifugal- and Coriolis forces and electrostatic forces between charged particles which change the scale height of ions. This chapter discusses disturbances caused by tidally induced vertical flow, or up- and down welling. In section VI.5 a discussion is presented of how horizontal winds may also play a role.

One common way of analysing composition changes is not only to look at the individual number densities of constituents but to examine their number density ratios, such as \( O/O_2 \) and \( O/N_2 \). From Figure VI.1 one can see that ratios \( O/O_2 \) and \( O/N_2 \) increase with altitude. The hydrostatic distribution of constituents above the turbopause implies that constituent ratios are constant on any level of fixed pressure as long as diffusive- and chemical equilibrium holds. In order to examine the influence of tides on composition, one therefore needs to determine how far equilibrium can be disturbed and how quickly the balance can be restored. Generally, this can happen through molecular diffusion as well as chemical reactions. The following discussions will first show how tidally induced vertical flow causes departures from diffusive- and chemical equilibrium and then investigate the processes which attempt to restore balance.

**VI.2. THE TWO VERTICAL WIND COMPONENTS**

In order to investigate the influence of vertical flow on composition, it is necessary to distinguish between two components of vertical wind, the *barometric velocity*, \( v_b \), and *divergence velocity* \( v_d \), where the latter, as outlined previously in III.3.2, represents motion relative to a fixed pressure-level and the former is the motion of the pressure level itself [*Rishbeth et al.*, 1969]. The total vertical
wind, \( v \), in the height coordinate system is then given by their sum, \( v = v_b + v_d \).

While the barometric wind can be caused by thermal expansion and contraction, the divergence wind is related to the vertical wind in the pressure frame, \( w = dp/dt \) (see III.3.2, equation III.16), by

\[
v_d = -\frac{w}{\rho g}
\]

One can see from the above relation and the continuity equation in the pressure coordinate frame that the vertical divergence wind and the divergence of the horizontal wind field are directly related to each other. The continuity equation in the pressure coordinate frame was given in Chapter III (equation III.26) as

\[
\frac{\partial w}{\partial p} + \nabla_p \cdot \mathbf{V}_p = 0
\]

where \( \mathbf{V}_p \) and \( \nabla_p \) are the horizontal wind vector and gradient operator, respectively, on a level of constant pressure. Since horizontal wind gradients are strongly influenced by tidal oscillations in the atmosphere, the vertical divergence wind is similarly affected. The effect of this on composition is as follows.

The two vertical wind components described above influence composition in different ways. Barometric winds are caused by thermal expansion- and contraction, and the temperature variations which accompany barometric winds will change the scale heights of each constituent by the same proportion. This will change number densities in the height coordinate system since at any fixed altitude the pressure is different and therefore also the composition. However, the constituent ratios are maintained on a level of fixed pressure, even when this level is at a different altitude, and barometric winds will thus not disturb diffusive- and chemical balance.

In contrast, the vertical divergence winds will move parcels of gas relative to pressure levels and cause departures from diffusive and chemical equilibrium. This is due to the fact that constituent ratios change with height, as shown in Figure VI.1, and a vertically displaced parcel of air will be surrounded by gas which has a different constituent ratio. Molecular diffusion and chemical reactions act to try to restore equilibrium, and an effective constituent flow may take place as a result of the vertical divergence wind, depending on the time scales involved. If time scales of
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processes restoring the balance are much longer than those of the perturbation, composition is effectively not changed when averaged over the period of the tidal oscillation. If horizontal winds are sufficiently strong they may play a role as well, which is discussed in section VI.5.

VI.3. DIFFUSION TIMES IN THE THERMOSPHERE

When the applied divergence winds are periodic in time, which is the case when they are caused by tides, vertical displacement of air parcels occurs periodically. It is of interest, thus, to examine the relative time scales of the oscillations which disturb diffusive equilibrium and molecular diffusion which tries to restore the balance.

The time necessary to reach diffusive equilibrium, \( \tau_{eq} \), can be estimated by

\[
\tau_{eq} = \frac{H^2}{D}
\]

[Rishbeth & Garriott, 1969], where \( H \) is the scale height and \( D \) the diffusion coefficient. Using the molecular diffusion rates in CTIM which are taken from Colegrove et al. [1966], values of \( \tau_{eq} \) can be calculated, as shown in Table VI.1 for globally averaged conditions.

<table>
<thead>
<tr>
<th>Height [km]</th>
<th>105</th>
<th>115</th>
<th>125</th>
<th>140</th>
<th>160</th>
<th>180</th>
<th>200</th>
<th>&gt; 200</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \tau_{eq} ) [h]</td>
<td>140</td>
<td>55</td>
<td>23</td>
<td>10</td>
<td>5</td>
<td>2</td>
<td>1</td>
<td>&lt; 1</td>
</tr>
</tbody>
</table>

Table VI.1: Time scales (in hours) for molecular diffusion in the globally averaged upper atmosphere. Values are based on the molecular diffusion coefficients by Colegrove et al. [1966].

The values shown were produced by averaging the individual \( \tau_{eq} \) values for the three molecular diffusion rates \( D \) [O-O\(_2\)], \( D \) [O-N\(_2\)] and \( D \) [O\(_2\)-N\(_2\)] (see Table III.1) between the major constituents. Between 80 and 105 km altitude, the eddy diffusion is important and the coefficient used in CTIM (see Chapter III, Table III.2) gives time scales, \( \tau_{eq} \) to reach equilibrium of around 65-72 hours, depending on the height. These values as well as those in Table VI.1 show that time scales necessary to reach diffusive equilibrium are larger than the periods of diurnal and semidiurnal tides for heights below around 120 and 135 km, respectively. This implies that below
these altitudes diffusion is unable to restore equilibrium fast enough if disturbed by upwards propagating tides. Diffusive redistribution of constituents is therefore below 120 km expected not to play an important role on the time scale of tides. The situation changes when diffusion times become shorter. During a perturbation, diffusion between the displaced air parcel and the background atmosphere then takes place and as a result of this, composition changes also when averaged over the period of the oscillation. After an oscillation cycle, the original constituent ratios are not restored.

Departures from diffusive equilibrium in the thermosphere are also known from other studies to lead to significant redistribution of the constituents. Mayr et al. [1973] showed that wind-induced diffusion in the thermosphere leads to transport of atomic oxygen towards earlier local times. They found this redistribution of O to be one of the mechanisms causing the thermospheric phase anomaly in which temperature maxima occur between 1 and 3 hours after the O density maxima. This effect is successfully reproduced in CTIM above 200 km altitude, where temperature maxima occur at around 15.6 h local time, while atomic oxygen densities peak near noon, and above 300 km towards 13.2 h local time.

VI. 4. CHEMICAL LIFETIMES IN THE THERMOSPHERE

The tidally induced changes of constituent ratio upset chemical equilibrium in the same way as diffusive balance is disturbed, and chemical reactions can thus similarly lead to composition changes in the background atmosphere. It depends on the chemical lifetimes of constituents in comparison with the tidal oscillation time scale whether the periodic changes of equilibrium lead to their long-term redistribution. The neutral constituent chemical reactions considered in the CTIM as well as their reaction rates are given in Table III.3. From these reaction rates and typical concentrations, chemical lifetimes of constituents associated with each reaction can be derived and are summarized in Table VI.2. The letters in the left column of the table denote the reactions $O_2+h v\rightarrow O+O$ (A); $O+O+M\rightarrow O_2+M$ (B); $O+OH\rightarrow O_2+H$ (C); $O+HO_2\rightarrow O_2+OH$ (D); $O+O\rightarrow O_2+h v$ (E). Fields are marked "n.a." where the reaction rate is set to zero in CTIM.
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Table VI.2: Chemical lifetimes (in hours) of constituents in the thermosphere involved in various chemical reactions, based on reaction rates from Table III.3 and average constituent densities modelled by CTIM. The letters in the left column of the table denote the reactions $O_2 + hv \rightleftharpoons O + O$ (A), $O + O + M \rightleftharpoons O_2 + M$ (B); $O + OH \rightleftharpoons O_2 + H$ (C); $O + HO_2 \rightleftharpoons O_2 + OH$ (D); $O + O - O_2 + hv$ (E). Fields are marked "n.a." where the reaction rate is set to zero in CTIM.

The table illustrates that all chemical reactions except for (C) and (D) give constituent lifetimes throughout the thermosphere which are much larger than the time scale of tides. The dominant photochemical process affecting major species thermospheric composition above around 100 km is the photo-dissociation of $O_2$ (A) [Roble et al., 1987]. Photo-chemical lifetimes of $O_2$ associated with this reaction are between 100 and 200 km altitude longer than 4 days (see first row in Table VI.2), so the photochemistry will not cause noticeable composition changes in that height regime on the time scale of diurnal and semidiurnal perturbations. Below 90 km, reactions of atomic oxygen with OH and HO$_2$ are likely to play a role in connection with tides. Other than that, chemistry plays no role in restoring balance disturbed by tides below around 200 km altitude.

VI. 5. THE ROLE OF HORIZONTAL WINDS

Horizontal winds are known to play an important role in the transport of constituents in the lower and middle atmosphere [Brasseur and Solomon, 1986]. It is therefore of interest to investigate whether they play a role in the upper atmosphere as well, especially in connection with the tides. In the following, the term horizontal wind refers to motion along isobars.

Generally, transport by winds occurs if the lifetime of a constituent is comparable to or larger than the time scale associated with the background winds [Brasseur and Solomon, 1986]. One may define a transport time scale, $\tau_j$, of winds as the time needed for transport in direction $j$ over a
distance \( d_j \), assuming the wind speed of \( v_j \):

\[
\tau_j = \frac{d_j}{v_j}
\]

*Brasseur and Solomon* chose characteristic distances of 1000 km in the zonal- and meridional directions and 5 km in the vertical direction. In the present context it is more appropriate to chose a characteristic distance of 1 scale height in the vertical direction, ranging from around 6 km at 80 km height to around 20 km at 200 km height.

It was found in Chapter IV that in-situ zonal background winds below 200 km reach values of around 70 m/s at 60° latitude, 50 m/s at 40° latitude and less than 30 m/s at lower latitudes, on either the hemispheres (see Figure IV.30). The meridional in-situ winds were found to be of less than 10 m/s in the same regions. This gives characteristic wind transport times of at least 4 hours in zonal direction and at least 1 day in the meridional direction. With the photochemical lifetimes being longer than around 4 days between 100 and 200 km (see Table VI.3), the background winds therefore do influence the composition structure at those altitudes.

Between 80 and 90 km, the chemical lifetime of atomic oxygen is in the order of a few hours or less, while wind transport times are of more than 2 days in CTIM. Therefore, atomic oxygen is not influenced by background winds at those heights in CTIM. The lifetimes of molecular oxygen and nitrogen are however much longer and thus their distribution does depend on the background winds also the low altitudes in CTIM.

It must be emphasized that the background winds currently produced by CTIM below 90 km are smaller than would be expected in the real atmosphere. The reason for this is that the background atmosphere at the lower boundary is assumed horizontally stratified, implying, amongst other, that the latitudinal temperature gradient found at solstice in the real atmosphere is not considered in the model. This gradient produces zonal background winds of the order of tens of metres per second which give transport times in the order of one day or less.

Background winds can affect composition in the following two ways if they are accompanied by tidal oscillations.
Wherever the horizontal background winds are important for the global distribution of constituents (see earlier discussion of this), the tides can alter the composition through changes of the background wind structure. It was found in Chapter IV that tides release momentum to the background atmosphere when they dissipate, and therefore the horizontal background wind pattern will be affected by the tides. Changes of the background vertical wind above 160 km are later shown to explain features in Figure VI.6.

The second type of wind influence is more complex than the first. It was described earlier that tidal oscillations cause periodic vertical displacements of air parcels. If horizontal winds are strong enough, giving characteristic wind transport times of less than a few hours, these vertically displaced air parcels are transported horizontally away from the original region of tidal perturbation. The periodic displacement of an air parcel can be seen as taking place within a horizontal layer which is centred vertically around the average height of the air parcel and has a depth of twice the maximum distance by which the parcel is perturbed. If the horizontal wind is strong, but roughly constant throughout this layer, any horizontal transport will not matter.

However, an additional mixing of constituents occurs if horizontal winds change vertically within the layer. The principle is illustrated in Figure VI.2. The blue and read filled circles represent air parcels which have been displaced vertically through tidally induced up- and down welling. At time \( t_A \) the parcels have a distance of \( A \), at time \( t_B \) their horizontal distance is \( B \). If horizontal winds are not only strong enough but furthermore different for the red and blue parcels, the situation arises which is illustrated in the plot. One sees that \( B < A \) as a result of the stronger horizontal winds on the blue parcels. As a result, constituents are mixed.

One can see from these discussions that it is difficult to identify this effect in model simulations since it occurs only in combination with up- and down welling, which on its own already causes departures from diffusive balance. Since one of the conditions for the effect to happen are strong vertical gradients of horizontal background winds, it is furthermore most likely to occur in regions of tidal dissipation, where the release of tidal momentum causes changes in the background winds over short vertical distances. The typical thickness of the layers described earlier depends on the vertical divergence wind. CTIM simulations suggest that these winds are less than 0.01 m/s between 80 and 150 km altitude and less than 0.5 m/s between 150 and 200 km. For semidiurnal tides, the period of up welling lasts 3 hours, giving a layer thickness in the order of 200 m below
150 km altitude and up to 10 km above that. For diurnal tides the values are in the order of roughly 15 km and 40 km for the altitude ranges below and above 130 km, respectively. Tidal oscillations were found to change the background winds by up to around 20 m/s over one scale height under (1,1) mode forcing (using 100 m lower boundary amplitude; see also Figure IV.31), by up to around 10 m/s over one scale height under (2,4) and (2,5) mode forcing and less than 5 m/s under (2,2) mode forcing (using 200 m forcing amplitude at the lower boundary; see also Figure IV.32).

These numbers suggest that the described horizontal wind effect under semidiurnal forcing does not occur below 150 km since the horizontal winds hardly change over an altitude of a few hundred metres. Above 150 km it is more likely to play a role, but only when the tides are of higher semidiurnal mode order, such as the (2,4) or (2,5) modes. Under diurnal forcing the effect may occur above 100 km already.

In summary, the horizontal winds do play a role in conjunction with tides in disturbing the diffusive balance. This influence is however limited to regions where the tides dissipate, and is stronger at altitudes above 150 km. The effect is difficult to identify in simulations or measurements since it occurs only with the up-and down welling which by itself also disturbs the diffusive balance.

VI. 6. TIDES AND COMPOSITION IN CTIM

In order to investigate whether CTIM reproduces the above theoretical predictions, two runs were carried out for equinox conditions, using F10.7 solar flux of 85 and Kp=2. At the lower boundary, semidiurnal tidal (2,4) forcing was applied in both runs with an amplitude of 200 m geopotential height, setting the phase of local time maximum to 12.0 h. In order to investigate the influence of horizontal wind gradients on composition changes, however, in one of the runs horizontal momentum gradients were artificially set to 0, which resulted in the horizontal wind gradients being zero.

Results of these two simulations are given in Figure VI.3. Both plots show the height of pressure level 7 (2.57·10^3 Pa) at 18°N (dashed curves) and the heights of the levels of constant O/O_2 constituent ratios (solid). The upper plot in Figure VI.3 shows the two parameters under normal conditions, while the lower plot shows the same parameters under zero horizontal wind gradients. Two different constituent ratios were used in the plots, with values of 2.1 (upper plot) and 3.0
In the upper plot of Figure VI.3 both curves show primarily semidiurnal behaviour and are shifted in phase by around 4 hours. The fact that a phase shift is found between these curves tells that the \( \text{O}_2/\text{O} \) ratio is not constant on a fixed pressure level. This demonstrates not only that tides disturb the diffusive balance but also that molecular diffusion and chemical reactions are unable to restore equilibrium on the tidal timescale. The situation is different in the lower plot of Figure VI.3. Here, both curves have the same phases, thus showing that the \( \text{O}_2/\text{O} \) ratio is constant on the pressure level. With the earlier findings about diffusion- and chemical life times one may conclude that no departure from diffusive equilibrium occurs when the horizontal velocity divergence is ignored since no vertical divergence winds are generated.

In Figure VI.4 the heights of pressure levels 6 and 8 are shown with those of \( \text{O}_2/\text{O} = 0.85 \) and \( \text{O}_2/\text{O} = 4.2 \), respectively. Two important features can be seen in the plots. At pressure level 6 (\( 6.99 \times 10^3 \) Pa) the average phase difference between the curves is larger by around 1 hour, compared with pressure level 7 (Figure VI.3, top). At pressure level 8 (\( 9.46 \times 10^4 \) Pa) the situation differs considerably in that the \( \text{O}_2/\text{O} \) ratio height behaves rather diurnally, while the pressure level height is predominantly semidiurnal. This suggests that photochemistry starts to become important near 140 km height, overshadowing any diffusion-driven composition changes. At level 6 the phase shift is larger between the curves than at level 7, and values from Table VI.1 suggest that the shorter diffusion times at level 7 are likely to cause some redistribution of constituents, giving a marginally more balanced situation. Plots in Figures VI.3 and VI.4 thus illustrate how strongly composition is affected by the tides below around 140 km altitude through up-and down welling of air parcels relative to the pressure levels.

In Figure VI.5 a global profile is shown for the 12 h amplitudes of the \( \text{O}_2/\text{O} \) ratio under (2,4) mode forcing. The amplitudes are given as fractions of the background \( \text{O}_2/\text{O} \) ratio (in %). The clear pattern at low- to mid latitude again shows the influence of lower boundary tidal forcing on the aspect ratio, while at high latitudes the semidiurnal composition changes are caused by auroral effects above 120 km. Average amplitudes at low-to mid latitudes are around 10% of the background value, and only below 90 km are they considerably larger, approaching 100%. However, the absolute (as opposed to the relative-) values of amplitude do not show the same strong peaks below 90 km, but they decrease from 90 km towards the lower boundary. The reason
for these strong features at 85 km in Figure VI.5 is the background O/O₂ ratio, which is smaller by a factor of around 50 at pressure level 2 (85 km), compared with pressure level 3 (91 km). The strong features in the plot are thus rather a result of the lower boundary conditions and need no further consideration.

Figure VI.6 shows the change of background O/O₂ ratio caused by the tidal forcing. Values in the plots are differences between background O/O₂ in the simulation with (2,4) tidal forcing and a simulation using no lower boundary forcing (run 24E- run NT, see Table IV.1), in percent. The lower plot shows a vertical “slice” at 0° latitude from the upper plot (solid blue curve). Furthermore, it shows the changes of background O density (dashed red) and O₂ density (dotted green) for the same latitude, thus allowing to identify which of the two cause the O/O₂ ratio behaviour at different heights.

Apart from high-latitude features which are not linked to the tidal forcing and will not be discussed further in the present context, a distinct pattern of local extrema is found near and on the equator. These are linked to the tidal forcing, and the question is thus which of the processes discussed in VI.3, VI.4 and VI.5 causes them. From the lower plot in Figure VI.6 one sees that below around 100 km the O/O₂ ratio follows primarily variations of O, while the O₂ density is relatively invariant. Above 140 km the opposite is seen. There, the O density varies to a lesser extent with height and the periodic behaviour of the O/O₂ ratio is caused primarily by the O₂ density changes. Between these heights, the two constituents both vary with altitude, but generally the O₂ density changes more under tidal forcing above around 100 km altitude. Below 100 km the dominant chemical reactions (see Table VI.2) remove O and produce O₂, while above 100 km the opposite is the case. The reason why below 100 km the changes of O appear more strongly in Figure VI.6 than those of O₂ is that the overall O density in that height regime is considerably smaller than the O₂ density (see Figure VI.1), so the changes in number density induced by the tides are larger in relation to the overall O density. The opposite is the case above around 100 km, where the O density is larger than the O₂ density. The composition changes below 100 km are thus primarily related to the chemistry which gives shorter lifetimes of O. Although the O₂ and N₂ densities there are with the longer lifetimes likely to be affected by background winds, it was found that tidally induced changes in background wind at those heights were very minor only and unable to explain features in Figure VI.6. At higher altitudes, the winds do however play an important role along with molecular diffusion. Background winds above 100 km were found to vary with height by up
to around 10 m/s over one scale height and this along with the shorter diffusion times above around 130 km (see Table VI.1) explains the behaviour in Figure VI.6.

The large region at low- to mid-latitude above 160 km in which the O/O₂ ratio decreases is a result of stronger up welling in that region. When plotting the change of background vertical winds (not shown) one sees a region of stronger upward winds by around 25% which coincides with the region of lower O/O₂ ratio in Figure VI.6 above 160 km. As mentioned earlier, the O/O₂ ratio increases with height, and the up welling therefore reduces the ratio.

VI. 7. TIDALLY INDUCED TEMPERATURE VARIATIONS

The CTIM simulations presented earlier not only allow an insight into the mechanisms causing composition changes, but they similarly illustrate the nature of tidal temperature oscillations outside the region of tidal (radiative-) forcing. This can be examined by plotting temperatures from the two tidal model simulations used earlier. The upper plot in Figure VI.7 shows temperature at 18°N on pressure level 5 (solid) and at a fixed height of 106.1 km (dashed). The curves are almost identical and show predominantly semidiurnal behaviour of the temperature. The lower plot in Figure VI.7 shows the equivalent parameters produced when horizontal gradients are set to 0. One sees a considerable difference between the curves in the lower plot. The temperature on a fixed pressure level there has a stronger diurnal component, while values at a fixed height are predominantly semidiurnal.

This shows that the solar heating with its primarily diurnal behaviour dominates the temperature behaviour on a pressure level when ignoring divergence winds. In the upper plot, the divergence winds are accompanied by up- and down welling, and this causes adiabatic heating- and cooling. The strongly semidiurnal temperatures on a pressure level are therefore caused by the divergence winds. When the divergence winds are set to zero (lower plot in Figure VI.7) this adiabatic heating does not occur on a pressure level and temperature is thus controlled by other sources.

Nevertheless, temperatures are semidiurnal at fixed altitude in the lower plot since oscillations of the pressure level height are accompanied by heating- and cooling through compression and expansion. This compressional heating and cooling also occurs in the upper plot of Figure VI.7, but is overshadowed by the adiabatic processes. If the two heating and cooling processes were
similar in strength, the curves in the upper plot would differ more.

The finding that temperature variations in the upper plot of Figure VI.7 are caused primarily by
the adiabatic heating and cooling is confirmed when analysing the difference of the curves in the
upper plot of Figure VI.3. This difference is largest when the heights of constant pressure level and
O/O₂ ratio differ most. It can therefore be regarded as a measure of the degree of departure from
diffusive equilibrium, or similarly as a measure of the strength of up- and down welling relative to
a fixed pressure level. When plotting this difference versus local time (not shown here) and
comparing the profile to the temperature curves in the upper plot of Figure VI.7 one sees that
temperature extrema coincide with extrema of the difference curve. This confirms the strong
correlation between the temperature behaviour and up- or down welling, through adiabatic heating-
and cooling.

The solar heating through absorption increases with height and eventually becomes stronger than
the adiabatic heating from tides, resulting in a primarily diurnal temperature behaviour at higher
altitudes. Above 200 km, most of the tides propagating upwards from the middle atmosphere will
have dissipated, thus further reducing the semidiurnal component in temperatures, especially at
low-to mid latitudes.
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Figure VI.1: Neutral composition of the atmosphere, under average solar- and geomagnetic conditions. The plot shows globally averaged number densities in units of \( \log_{10} [m^{-3}] \) for atomic oxygen (solid), molecular oxygen (dashed) and molecular nitrogen (dotted).
Figure VI.2: Illustration of the role of horizontal winds in causing composition changes. The blue and red filled circles represent air parcels which have been displaced vertically through tidally induced up- and downwelling. At time $t_A$ the parcels have a distance of $A$, at time $t_B$ their horizontal distance is $B$. If time scales for horizontal wind transport are smaller than the perturbation period and furthermore the horizontal winds are different for the red and blue parcels, the distances $A$ and $B$ are different, thus causing departure from diffusive equilibrium.
Figure VI.3: Height of pressure level 7 (dashed) in a regular CTIM simulation (top) and when setting horizontal gradients in the model artificially to zero (bottom). Also shown are heights of levels with constant O/O\textsubscript{2} ratios (solid). The ratios have values of 2.1 (top) and 3.0 (bottom).
**Figure VI.4:** Heights of pressure level 6 (top) and 8 (bottom) from the regular CTIM run used in Figure VI.3. Solid curves are the heights of constant $O/O_2=0.85$ (top) and $O/O_2=4.2$ (bottom) ratios.
Figure VI.5: Semidiurnal amplitudes of the $O/O_2$ ratio, as modeled by CTIM with (2,4) mode forcing. The values shown are 12 h amplitudes in percent of the mean background aspect ratio.
Figure VI.6: The change (in %) of background \( O/O_2 \) ratio caused by lower boundary (2,4) forcing (top). The bottom plot shows the change of \( O/O_2 \) ratio at latitude 0° (solid) as well as the changes of \( O \) density (dashed) and \( O_2 \) density (dotted). Positive values denote increase of the ratio or densities due to the tides.
Figure VI.7: Temperature on pressure level 5 (solid) and at fixed heights (dashed) of 106.1 km (top) and 107.2 km (bottom). Values in the bottom plot are taken from the CTIM simulation in which horizontal wind gradients were set to 0.
CHAPTER VII. PLANETARY WAVES

ABSTRACT

This chapter presents a study in which the CTIM is used to simulate the upwards propagating 2-day wave. Motivation for the study is to investigate reasons for the occurrence of 2-day wave signatures in the ionospheric \( F \)-layer, while neutral atmosphere measurements have found 2-day wave signatures up to altitudes of around 120 km only. Three model simulations are presented, in which the first uses planetary wave forcing of lower boundary parameters, the second adds to that lower boundary tidal forcing and the third uses only tidal forcing, but with the tidal amplitudes being modulated with a quasi-2-day period. The first two simulations produce very similar results and both show the sensitivity of profiles to eastward background winds which make the wave evanescent and prevent further vertical propagation. Replacing the planetary wave forcing with modulated tides (third simulation) produces 2-day wave oscillations at altitudes of which the amplitude maxima lie up to 50 km above those produced in the first two simulations. Furthermore, the modulated tides are found to propagate undisturbed through regions of eastward winds. From the results in the neutral parameters it is predicted that modulated tides provide a means by which planetary wave signatures, through chemistry, are transferred into the ionosphere near 200 km height. An updated version of CTIM will in future studies be used to evaluate the relative importance of dynamo coupling in the \( E \)-region for explaining the ionospheric 2-day wave signatures. Results also encourage to examine in the future measurements of neutral parameters above around 200 km for planetary wave signatures.

VII. 1. BACKGROUND

Planetary waves are global atmospheric oscillations with periods of more than one day. Of these, one well known example is the “quasi-2-day wave” which is investigated in more detail in what follows.

Features having a quasi-2-day oscillation period where first found in meteor wind measurements from Sheffield, UK, by Müller [1972]. It was not immediately evident what the global structure of this wave was, and by combining measurements from two different locations in France and Russia,
Glass et al. [1975] were the first to propose that the oscillations had a zonal wavenumber of 3. Muller and Nelson [1978] confirmed this finding and added that the 2-day wave was westward travelling. Furthermore, Clark [1975] found a seasonal dependence in the amplitude, being strongest in the middle to late summer. More recent observations of the 2-day wave are by Plumb et al. [1987], Tsuda et al. [1988], Williams and Avery [1992] and Clark et al. [1993].

Significant advances were made with the theoretical studies by Salby [1981 a,b,c] who presented results from numerical calculations between ground level and 85 km altitude. He identified the 2-day wave as being the [3,-3] Hough mode, using the same nomenclature as in Chapter II. Using the nomenclature of Longuet-Higgins [1968], it is also in the literature sometimes referred to as the [3,0] mode. The theoretical studies by Salby were revolutionary in that they explored the properties of the 2-day wave in unprecedented detail. Nowadays, his results still largely agree with measurements and are thus assumed to be an accurate description of this wave. His model simulated the wave properties in an idealized background atmosphere which ignored damping forces, such as viscosity. Some key findings were that the exact frequency of the 2-day wave as well as its vertical structure were dependent on the background atmosphere wind field. Specifically, the wave is sensitive to the zonal winds, being more (less) evanescent when eastward winds are strong (weak) compared with the wave’s zonal phase speed. This, he argued, was due to the stronger zonal winds causing a reduced refractive index in the atmosphere. The average period of the 2-day wave was determined to be 54 hours, which is why the wave is commonly also referred to as the “quasi-2-day wave”. Studies by Salby suggested that the 2-day wave is a resonant oscillation of the atmosphere and thus has no clear source. Plumb [1983] investigated this further and proposed that the 2-day wave was generated by instabilities which were caused by a shear of the westward wind jet found in the mesosphere at solstice. This, he found, also explained the annual variability of the wave. Pfister [1985] extended the 1-dimensional calculations by Plumb into 2 dimensions and found 2-day oscillations which were generated by the instabilities, but he found these waves at mid-to high latitudes to be trapped vertically, thus not accounting for observations of the 2-day wave in the upper mesosphere and lower thermosphere. At present, no conclusive evidence has been found for the source of the 2-day wave and the idea proposed by Salby best explains observations.

VII. 2. THE QUASI 2-DAY WAVE IN THE UPPER ATMOSPHERE

Observations and studies of the 2-day wave listed in the previous section, including those by Salby,
Planetary waves described the wave as being confined vertically to below around 120 km height. To-date, no 2-day signatures have been found in neutral atmosphere measurements at higher altitudes. It came as a surprise, therefore, that more recent measurements [Pancheva, 1988; Chen, 1992; Apostolov et al., 1994, 1995] found westward travelling quasi-2-day wave signatures in ionospheric parameters, such as the critical F-layer frequency, $f_{\text{F2}}$. These observations also found modulations of the ionospheric 2-day wave amplitudes by the 11-year solar cycle as well as geomagnetic activity variations. In many cases, the ionospheric 2-day wave observations could be linked to simultaneous upper mesosphere- lower thermosphere measurements of 2-day wave signatures. The key question arising from these observations is how the wave can penetrate into the ionosphere, while at the same time appearing to be confined in the neutral atmosphere to below around 120 km. The purpose of modelling studies with the CTIM presented below is to investigate further this issue.

Three possible scenarios can be proposed as explanations. The first is that the [3,-3] Hough mode does penetrate higher into the thermosphere than previously found. The second possible explanation is that low-latitude planetary waves in the lower thermosphere through dynamo coupling modulate the electric field, thus transferring the signatures into the ionosphere at a lower altitude than in the previously listed option. Chen [1992] carried out numerical simulations suggesting that dynamo coupling was responsible for observations of the 2-day wave in the equatorial ionization anomaly. The third possibility is that tides, which are known to propagate to altitudes near 200 km, are modulated by the planetary waves at lower altitudes and carry the signatures up, while the original waves dissipates below around 120 km. The modulation of tides by planetary waves was studied in detail by Teitelbaum and Vial [1991] and measured by Mitchell et al., [1996], Ruster [1994] and Huuskonen et al., [1991].

This third option has in common with the first that planetary wave features are carried higher into the neutral atmosphere than previously thought, but the first option describes a situation in which the planetary waves directly propagate higher, while in the third option the coupling between lower- and higher altitudes is linked to modulation of tides, giving a “hidden” planetary wave at some altitudes.

In what follows, results from a modelling study carried out by Hagan et al. [1993] are described in more detail since they investigated the propagation of the 2-day wave into the lower thermosphere, while at the same time presenting some key properties of the wave. Most of the
lower- and middle atmosphere wave properties found in this study largely agree with findings by \textit{Salby} [1981 \textit{a,b,c}], but the effects of lower thermosphere damping were also included. In the CTIM study presented later, 2-day wave signatures given by the third simulation in \textit{Hagan et al.} are used at CTIM's lower boundary as the external forcing profile, and some of the results produced with the CTIM are compared to those by \textit{Hagan et al.}.

The modelling study by \textit{Hagan et al.} used the Global Scale Wave Model (GSWM) which was described in Chapter V, section V.3.1. Excitation of planetary waves in the GSWM simulations was carried out by implementing an artificial heating source at the bottom boundary (ground level) with the latitudinal structure of a [3,-3] mode. Three simulations were carried out, using different background wind fields. The first used a standard wind field which above 20 km was calculated from the MSISE90 [Hedin, 1991] temperature profile, assuming geostrophic balance. These winds were for the second simulation replaced in the 25-75 km height regime with parameterized winds by \textit{Smith} [1985], while using the original winds at other heights. In the third simulation the \textit{Smith} winds were kept, and furthermore winds between 80 km and the upper boundary, at 150 km, were replaced by values from the empirical model by \textit{Portnyagin and Solv'yeva} [1992 \textit{a, b}], giving a more realistic profile for the lower thermosphere. Thus, the effects of changing mesospheric- and lower thermospheric winds on the 2-day wave could be examined separately.

Key results were that changing mesospheric wind fields lead to wave structure changes below 80 km, but not above, and furthermore to a wave period change from 2.3 days (in the first simulation) to 1.95 days. Introducing the more realistic winds above 80 km lead to changes of the 2-day wave also above 80 km. Essentially, the simulations confirmed the importance of the mean zonal wind which was first found by \textit{Salby} [1981 \textit{a, b, c}]. The wind model by \textit{Portnyagin and Solv'yeva} introduced an additional eastward jet at mid-latitudes in the thermosphere, and this jet was found to suppress the 2-day wave propagation into the thermosphere at those latitudes, again because of refractive index changes caused by the wind, in accordance with findings by \textit{Salby}. Changing the wind field above 80 km hardly affected the wave's period. These studies by \textit{Hagan et al.} furthermore investigated effects of molecular- and eddy diffusion on the 2-day wave. The introduction of molecular diffusion was found to weaken the 2-day wave in the lower thermosphere by around 30\%, while eddy diffusion was found to increase considerably the wave response in the 75-95 km height regime.
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The simulations suggest that the 2-day wave dissipates below around 100-120 km, and found only negligible 2-day wave features in the neutral atmosphere above that height. These results suggest that the wave itself does not penetrate sufficiently into the upper atmosphere to explain ionospheric observations.

VII. 3. PLANETARY WAVE SIMULATIONS WITH CTIM

Of the three earlier proposed mechanisms causing planetary wave signatures in the ionosphere (see VII.2), only the first and the third can be investigated with the current version of CTIM since at present, the low-latitude dynamo coupling is not modelled self-consistently. Results from simulations by Hagan et al. [1993], as described in the previous section, and findings with the CTIM presented below both suggest that the first mechanisms is unlikely to be important, and therefore the discussions below will focus on the third. Simulation settings are described in the following along with the methods used to force CTIM's lower boundary with planetary wave perturbations. The study below is the first of its kind with an upper atmosphere model.

VII. 3.1. THE MODEL RUNS

In order to investigate the mechanisms by which quasi-2-day wave signatures propagate into the upper atmosphere, three simulations were carried out with CTIM for January conditions, using a solar flux index of F10.7=106 and magnetic index of Kp=2. The simulations differed only in their lower boundary forcing. The first run, (A), used planetary wave forcing without tides, the second, (B), used both planetary wave- and tidal forcing, while the third simulation, (C), used only tidal forcing, but with modulated amplitudes.

The tidal forcing used in simulation (B) consisted of the (2,2) Hough mode at 200 m amplitude, with the phase set to 6.7 h local time. The choice of phase was in agreement with the January simulation presented in Chapter V (see Table V.4) and the amplitude value was chosen empirically. It was decided to use the (2,2) mode since this propagates to greater heights than the higher order Hough modes (see Chapter IV).
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VII. 3.2. PLANETARY WAVE FORCING IN CTIM

Planetary wave forcing at CTIM's lower boundary used in Runs (A) and (B) consisted of oscillating the winds, geopotential height and temperature with a period of 55.2 hours. In principle, such forcing can be specified in CTIM by using the $[3,-3]$ Hough mode in a similar manner as for the tides, but in these simulations an alternative approach was successfully used. As pointed out previously on numerous occasions, the lower boundary can be forced by any profile of winds, geopotential height and temperature, as long as the set of parameters is self-consistent. A complete set of self-consistent quasi-2-day wave wind-, geopotential height- and temperature amplitudes as well as phases was provided for 80 km altitude by M. Hagan [private comm., 1997] and used at CTIM's lower boundary for the external forcing. The forcing profile was extracted from the third simulation presented in Hagan et al. [1993] (see earlier description), which had used the realistic lower thermosphere winds by Portnyagin and Solv'yeva. The set of amplitudes is shown in Figure VII.1, and one interesting feature is that peak amplitudes occur in the southern hemisphere. The simulations by Hagan et al. were carried out for January, and this feature thus agrees with the seasonal behaviour found by other authors (see VII.1), giving stronger 2-day wave amplitudes in the summer hemisphere. This characteristic would not have been implemented if using a theoretical $[3,-3]$ mode profile and shows the limitations of the Hough mode approach for planetary waves at that height.

The phases used are not shown here, but presented in Hagan et al. [1993]. Phases of planetary waves are given not in terms of local times of maxima, but longitude. The reason for this is that planetary waves, as outlined in VII.1, are not forced by the Sun. The scenario is thus different than for solar tides. The phases of tides are constant with local time, implying that the Earth, in a simplified picture, "rotates under the oscillation", while the oscillation itself at any location remains fixed relative to the Sun. Planetary waves, in contrast, "rotate with the Earth", and their phases are therefore defined as longitudes of maxima. This needs to be considered when implementing a planetary wave at CTIM's lower boundary.

The zonal wavenumber of the 2-day wave is 3 (see VII.1), and therefore the zonal structure of 2-day wave perturbations is given by a sinusoid with a spatial period of 120° longitude. The 2-day wave perturbation, $\Delta_i(\theta,\varphi)$, of parameter $i$ at latitude $\theta$ and longitude $\varphi$ (in degrees) added to the
background field at CTIM's lower boundary was thus in Runs (A) and (B) expressed by the relation

\[
\Delta_i(\theta, \varphi) = A_i(\theta) \cdot \cos\left(\frac{2\pi}{120} \varphi + \phi_i(\theta)\right) \cdot \cos\left(\frac{2\pi}{T_{2\text{-day}}} t\right)
\]

Here, the \(A_i(\theta)\) are the 2-day wave amplitudes of Figure VII.1 for each parameter \(i\) at latitude \(\theta\). The time dependence is introduced by multiplying this global pattern at each location with a time-dependent sinusoid (second cosine term on the right side) which has the quasi-2-day wave period, \(T_{2\text{-day}}\), and changes with universal time, \(t\). The 2-day wave phase, \(\phi_i(\theta)\), (in degrees longitude) is implemented in the longitudinal cosine term, but the phase of the time-dependent sinusoid is arbitrary. Due to the earlier argument about the 2-day wave not being forced by the Sun any time-phase can be used, and for simplicity it was set to 0, implying that the simulations were started at 12.0 h universal time with a 2-day wave maximum in the time-domain.

The global 2-day wave profile implemented at CTIM's lower boundary can thus be pictured in the following way. At each location on the globe there is an oscillation of winds, temperature and geopotential height with time. The amplitude of this oscillation at each location depends on its latitude and longitude. The latitudinal dependence is given by the profiles in Figure VII.1 for each of these parameters. In longitude, the amplitude varies with a period of 120°, implying that locations at the same latitude which lie 120° apart in longitude have the same amplitudes. The strongest amplitudes are found at certain longitudes which are defined as the phases and implemented externally. These depend on latitude, implying that the strongest amplitudes do not necessarily occur at the same longitudes for two locations which have different latitudes.

In Run (C), a different method of planetary wave forcing was used. Here, the lower boundary was perturbed with tidal oscillations (see Chapters III-V), and the tidal amplitudes were modulated with a quasi-2-day period. The latitudinal amplitude profiles of Figure VII.1 could however not be used, for the following reason. The geopotential height amplitude term, \(Z_{s,m}^0\), as defined in Classical Tidal Theory (see Chapter III, equation III.71), is independent of longitude and latitude and gives the strength of a Hough mode \((s,m)\) in the global tidal profile. Any global dependence of the tidal profile is thus expressed by the Hough modes, and amplitude terms, \(Z_{s,m}^0\), are constant parameters only. Implementing a latitude- and longitude dependent tidal amplitude term thus conflicts with the tidal theory expressions used at the lower boundary and would lead to an inconsistency. Therefore, tidal amplitudes have been modulated with a quasi-2-day wave amplitude which varies only with
universal time, but not with location. The expression for the (2,2) mode amplitude in Run (C) is

\[ Z_{2,2}^0 = 400 \ m \cdot \left[ \frac{3}{2} + \cos \left( \frac{2\pi}{T_{\text{2-day}}} t \right) \right] \]

and results in a 2-day wave amplitude of 800 m geopotential height at the equator, where the (2,2)
mode is strongest (see Figure II.2). This amplitude, as well as the term "3/2" used in the above
expression, were chosen empirically. The resultant 2-day wind- and temperature oscillations were
calculated automatically by the analytical expressions for tides. Due to the unavoidable differences
in the planetary wave forcing profiles of Run (C), compared with Runs (A) and (B), the results
cannot be compared quantitatively, but only qualitatively.

VII. 4. RESULTS OF SIMULATIONS AND KEY FINDINGS

Results of the CTIM simulations are displayed in Figures VII.2 - VII.5. Global profiles of quasi-
2-day amplitudes are presented only for Runs (A) and (C) since results from Run (B) were found
to match closely those of Run (A). Due to the large changes of electron density with altitude in the
thermosphere it was found most appropriate to show in the contour plots electron density
amplitudes in proportion to the background electron density, in percent.

One sees that wind- and temperature profiles differ substantially in Run (C) from Runs (A) and
(B). Complex amplitude patterns are seen for the meridional winds in Run (A) (Figure VII.2, top).
The latitudinal structure between 80 and 160 km shows generally stronger amplitudes in the
southern hemisphere, which is due to the specific lower boundary forcing profile used (see bottom
right plot in Figure VII.1). Furthermore, a vertical structure can be seen, with essentially a pair of
strong maxima at 140 km height and weaker ones at 115 km. Amplitudes are smaller than those
implemented at the lower boundary, with peak values of less than 20 m/s, indicating that some of
the momentum from the lower boundary is not transferred upwards. The amplitude pattern is
constrained to the region between roughly 40°S and 30°N, and reaches to an altitude of around 160
km.

In simulations by Hagan et al. [1993] which have the identical 2-day wave profile at 80 km,
meridional wind amplitudes decreased to less than 5 m/s above 100 km, and are thus much lower.
Hagan et al. suggested that their low wind amplitudes above 100 km were a result of the
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background wind structure, with an eastward jet of up to 40 m/s above 80 km preventing an amplitude growth in the southern (summer-) hemisphere. The low- to mid latitude background wind structure in CTIM's Run (A) (not shown) is different from that, with a westward wind in both hemispheres up to an altitude of around 140 km which then becomes eastward to the north of around 30°S. The westward winds between 80 and 140 km in CTIM do not inhibit amplitude growth, as does the eastward wind in the GSWM simulations. It is of interest to note that the height of the main meridional wind maxima in Figure VII.2 coincides with the height at which the background winds become eastward throughout the relevant latitude range. This agrees with predictions by Salby and Hagan et al., emphasizing the importance of the background wind structure, particularly in the case of eastward winds. Adding the tidal forcing to the planetary waves, as was done in Run (B) (not shown), changes the zonal background wind between around 30°S and 30°N, reducing westward background winds (not shown) by up to 4 m/s, but the important aspect is that this tidal forcing does not reverse them to eastward. Therefore, the amplitude response patterns of Run (B) were so similar to those of Run (A). In future studies it will be of interest to apply different tidal forcing to produce a reversal of the zonal wind field, which would then affect the 2-day wave.

Meridional quasi-2-day wave amplitudes produced by Run (C) (Figure VII.4, top) differ substantially from the others in that patterns are symmetric to the equator and strongest in the height regime between around 120 and 200 km. The exact shape of the pattern can however not be compared to that of the other simulations since a different latitude structure was used at the lower boundary in Run (C). Two aspects are important, though; the first is that 2-day wave amplitudes are not seen below around 120 km and the other is that amplitude peaks occur at 160 and 190 km, which is higher by up to 50 km than in the other simulations.

The first of these findings shows that the 2-day wave implemented by modulating tidal amplitudes remains "hidden" throughout the 80-120 km height regime. In Run (C), zonal background wind patterns (not shown) were found in this height regime to be eastward in the southern- and westward in the northern hemisphere, which is different from what was found in Runs (A) and (B). This shows not only that the different methods of planetary wave forcing affected the zonal background winds but, more importantly, it showed that the "hidden" planetary wave of Run (C) propagated through regions of eastward winds apparently undisturbed. This suggests that planetary waves travelling upwards in the form of modulated tides are not as sensitive to the background wind field
as the “ordinary” planetary waves of Runs (A) and (B).

The second finding shows that planetary wave signatures propagate to larger heights if carried upwards by the tides. This is directly relevant to the original question, how planetary waves propagate into the upper thermosphere and ionosphere. This issue is discussed in more detail below. This finding is also interesting in that it would suggest the occurrence of planetary wave signatures also in the neutral atmosphere above around 160 km. Although the modulation of tides has been measured (Mitchell et al., [1996]; Ruster, [1994]; Huuskonen et al., [1991]), planetary wave signatures have been found only in the lower thermosphere measurements. The simulation thus motivates to search for planetary wave signatures in thermosphere measurements at higher altitudes.

Zonal wind response patterns in Runs (A), (B) and (C) are consistent with the earlier discussions of meridional wind amplitudes. The zonal wind response is generally smaller than that of meridional wind, which in Runs (A) and (B) is consistent with the smaller zonal wind forcing amplitudes at 80 km (see Figure VII.1). In Run (C), the zonal wind response is also weaker, in accordance with results for tidal amplitudes, as shown in Chapter IV (section IV.3.3.ii). The main reason for the smaller amplitudes in zonal direction is the stronger damping through ion drag in the zonal direction at mid- to high latitudes. The latitudinal profile of zonal wind 2-day wave amplitudes is different from that of meridional winds, concentrating at mid latitudes and being smaller at low latitudes, in accordance with findings by Salby and Hagan et al.

The 2-day wave amplitudes of temperature are shown in Figures VII.3 and VII.5 for Runs (A) and (C), respectively. As for the winds, the temperature response is more localized in terms of latitude and height in Run (C), leading to the same conclusions stated earlier. A further observable difference between Runs (A) and (C) is that temperature response signatures in Run (A) do not disappear entirely above 160 km height, and particularly in the northern hemisphere a band of enhanced 2-day-wave amplitudes is seen between around 20°N and 40°N, extending upwards with peak values of around 25°K. This is not significant when compared to the background temperature at those heights which lies above 550°K, giving a 2-day wave amplitude of less than 5% of the background value. Nevertheless, the fact that this feature does not appear in Run (C) suggests that some of the energy perturbations are not damped as effectively in Runs (A) and (B) as in (C). The phenomenon appears only in the temperature and not in horizontal winds.
Electron density 2-day wave amplitudes are shown in Figures VII.3 and VII.5 for Runs (A) and (C), respectively. One sees that no clear statement can be made from the calculated electron densities as to whether 2-day wave signatures are carried into the ionosphere by tides. The contour plots demonstrate the difficulty with the model version used here. At 34°S and 34°N two discontinuities are seen in the relative amplitudes at low- and higher latitudes. In this low-latitude regime the empirical model by Chiu [1975] is used to provide ionospheric background data to the thermosphere. However, the feedback of the thermosphere onto the ionosphere is not given, and thus electron density values at those latitudes are of no use in the present discussions. Poleward of these latitudes, however, the thermosphere-ionosphere interaction is simulated self-consistently. Unfortunately, the strongest meridional wind- and temperature response to planetary wave forcing in Run (C) was found at low latitudes, and 2-day wave effects at mid- to high latitudes are therefore very weak in the ionosphere as well. Some 2-day wave signatures are seen in the electron densities from Run (A) poleward of 34°N since the neutral atmosphere extends to higher latitudes as well (see Figure VII.3).

Even though electron density amplitudes in these simulations do not show clear evidence of modulated tides carrying planetary waves into the ionosphere, the response patterns of the neutral parameters do suggest this process to take place. Wind- and temperature amplitudes were found to peak at higher altitudes by around 20-50 km when using modulated tides. These signatures were also accompanied by periodic up- and down welling (not shown), and thus changes in the O/N₂ ratio. This, in turn, has an influence on ionospheric chemistry. Atomic oxygen (O) absorbs XUV radiation in the upper atmosphere and is ionized. The sink of ionized atomic oxygen in the ionosphere is its reaction with molecular nitrogen (N₂). Therefore, the charged particle density in the ionosphere is sensitive to the O/N₂ ratio. The model simulations here show that modulated tides carry 2-day wave signatures to an altitude in the thermosphere where changes of neutral composition can affect the charged particle densities. Future simulations with an improved version of the model need to examine this further.

The CTIM needs two major improvements in order to be able to confirm the theoretical predictions, an improved plasmasphere code for self-consistent charged particle properties at low latitudes and, in connection with this, the full dynamo calculations to account for the entire ion-neutral coupling. These two modifications have recently been added to the model by the collaborating group at the Space Environment Center in Boulder, Colorado, but the new code is still in an experimental stage.
The proposed planetary wave simulations are expected to be possible within the next year.

VII. 5. CONCLUDING COMMENTS

The planetary wave simulations presented in this chapter have produced a series of important results which were outlined in the previous section. The purpose of these experiments was not to carry out a detailed quantitative comparison with measurements or other models, but to test the upper atmosphere's response to different methods of external tidal forcing.

What remains to be shown is whether this coupling mechanism is more important, or whether the low- to mid latitude dynamo coupling which occurs at lower altitudes, typically below 120 km, is similarly significant. While the former relies on modulated tidal amplitudes, the latter occurs in regions where the "ordinary" planetary wave signatures have not dissipated entirely. It is possible that the modulated tides are the principle means for carrying planetary wave signatures to higher latitudes, where the dynamo field is less important, and more detailed modelling studies are needed to investigate this. The improved lower boundary of CTIM has thus proven to offer a wide spectrum of potential planetary wave research for the future.
FIGURES CHAPTER VII
Figure VII.1: Quasi-2-day wave amplitudes at 80 km used at CTIM’s lower boundary. Values are based on simulations by [Hagan et al., 1993].
Planetary waves  Chapter VII

Figure VII.2: Amplitudes of quasi-2-day meridional- and zonal winds, as modelled by CTIM in Run A, using planetary wave forcing at the lower boundary and no tides. The simulation was carried out for January conditions.

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Figure VII.3: *Same as Figure VII.2, but for temperatures and electron densities. Electron density values are given as ratios of amplitude to background value (in %).*
Figure VII.4: Same as Figure VII.2, but for Run C, using modulated tidal amplitudes at the lower boundary.
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Temperature 2-day amplitude
Run C

Latitude
Altitude [km]

320
280
240
200
160
120
80

0
30
60
-90
-60
-30
0

[K]
30+
27 to 30
25 to 27
22 to 25
19 to 22
16 to 19
14 to 16
11 to 14
8 to 11
5 to 8
3 to 5
0 to 3

Figure VII.5: Same as Figure VII.3, but for Run C, using modulated tidal amplitudes at the lower boundary.

e^ density 2-day amplitude
Run C

Latitude
Altitude [km]

320
280
240
200
160
120
80

0
30
60
-90
-60
-30
0

[%]
15+
14 to 15
12 to 14
11 to 12
10 to 11
8 to 10
7 to 8
5 to 7
4 to 5
3 to 4
1 to 3
0 to 1
CHAPTER VIII. SUMMARY AND FUTURE WORK

VIII. 1. SUMMARY

VIII. 1. 1. THE NEW LOWER BOUNDARY

This thesis has investigated the effects of upwardly propagating tides and planetary waves on the terrestrial thermosphere above 80 km altitude. The main tool in these studies was the Coupled Thermosphere-Ionosphere Model (CTIM) which simulates self-consistently the time-dependent dynamics, energy and composition of the upper atmosphere in three dimensions. Since an important fraction of the tidal oscillations found in the thermosphere originate from the troposphere and stratosphere, both of which lie outside the vertical range of the model, the upwardly propagating tides from these regions need to be specified externally in CTIM. The model’s previously fixed lower boundary, at 80 km, was replaced with a flexible layer at which a global profile of pressure-, wind- and temperature oscillations can be specified. One key issue in doing this was to ensure that these parameters are perturbed self-consistently. In principle, any global profile of parameter oscillations can be used at CTIM’s new lower boundary as long as their self-consistency is ensured. Two alternative approaches to lower boundary forcing were therefore presented.

The first method of tidal forcing consists in specifying a latitudinal profile of geopotential height perturbations, from which the accompanying temperature- and wind oscillations at the lower boundary are calculated, using analytical expressions from Classical Tidal Theory. The latitudinal geopotential height perturbation profile is given by Hough functions, which are the eigenfunctions of Laplace’s Tidal Equation. According to tidal theory, any global tidal profile can be decomposed into an infinite series of propagating- and non-propagating Hough functions and CTIM’s new lower boundary allows the use of the diurnal (1,1) mode as well as the semidiurnal (2,2), (2,3), (2,4) and (2,5) modes. The desired global oscillation profile at 80 km is thus specified by choosing an appropriate combination of individual mode amplitudes and phases.

In the other method of lower boundary forcing, a complete set of geopotential height, wind and temperature perturbations are used at the lower boundary, as provided ideally by another numerical model, thus ensuring their self-consistency without using the Classical Tidal Theory relations at the
lower boundary. This approach was found to be more realistic than the other for planetary wave forcing, since the approximations on which tidal theory relies are less acceptable for planetary waves. In particular, the fact that planetary waves interacted strongly with zonal background winds in the stratosphere and mesosphere was found to produce latitudinal amplitude profiles at the lower boundary height which could not satisfactorily be approximated by Hough functions.

**VIII. 1.2. TIDES IN THE THERMOSPHERE**

In Chapters IV, V and VI the characteristics of upwardly propagating tides in the thermosphere were examined in depth, together with their influence on the thermosphere’s momentum, energy and composition. It was found that tidal wind oscillations at low latitudes are driven mainly by pressure gradients and Coriolis forces, with ion drag playing an important role towards the poles where charged particle concentrations and velocities are larger. In the auroral oval, above around 120 km altitude, fast moving ions were found to provide an important source of mainly semidiurnal momentum and energy to the neutral atmosphere. Of the processes redistributing momentum internally, the horizontal and vertical advection, or transport by background winds, were found to be important at all latitudes. Momentum damping was found to occur mainly through vertical viscous drag at low- to mid-latitudes. Ion drag was found to be equally important at mid-latitudes and at the high latitudes generally stronger than the viscosity. Therefore, ion drag plays a double role, damping the tidal momentum at some locations and acting as momentum source at others, particularly in the auroral oval where ion velocities exceed the neutral wind speeds.

Similarly, a variety of external energy sources, internal redistribution and damping processes can be identified which play a role with tides. Divergence of the horizontal wind field leads to vertical winds relative to levels of constant pressure, causing adiabatic heating and cooling which generate oscillations of temperature. Above around 150 km altitude photodissociation of atomic oxygen becomes sufficiently effective to give photochemical lifetimes comparable to the oscillation periods of tides. Therefore, a further factor causing temperature oscillations at those altitudes are the tidally induced changes of density which lead to changes in photochemistry and thus energy. At high latitudes, a primarily semidiurnal energy source is the auroral oval, through Joule heating. The most important energy redistribution and damping processes were found to be horizontal and vertical advection and vertical heat conduction.
The described damping and redistribution processes for energy and momentum prevent tidal amplitudes from growing exponentially with height in the lower thermosphere, which one would otherwise expect from the exponential decrease of density. The tidal amplitudes in the lower thermosphere reach maxima and then fall off rapidly. At this altitude, the dissipation height, the tides release their energy and momentum to the background atmosphere, thus causing stronger background winds as well as heating the atmosphere.

The effectiveness of the various damping mechanisms, and thus the dissipation height, depends on horizontal and vertical gradients of winds, temperature and pressure. Higher order Hough modes have more complex latitudinal structures and smaller vertical wavelengths than the lower order modes. This causes stronger damping of the higher modes, and dissipation at lower altitudes. Typical heights of momentum dissipation were found to lie near 120 km for the semidiurnal \((2,4)\) and \((2,5)\) modes and near 180 km for the \((2,2)\) mode. The diurnal \((1,1)\) mode was found to dissipate near 140 km. Energy dissipation was generally found to occur around 20 km higher than that of momentum. In regions where the tides dissipate, their impact on the background atmosphere was found to be significant.

Background winds were found at low- to mid latitude not to influence noticeably the wind amplitudes in the lower thermosphere, but to some degree the temperature amplitudes. In the upper mesosphere, various authors have found that short-scale perturbations such as gravity waves do interact with the background winds as well as the tides. This interaction is still poorly understood and generally parameterized in middle atmosphere models through a Rayleigh friction term, which is not considered in the CTIM. Background temperature was found to influence the tides, though, with a cooler atmosphere causing stronger viscous damping.

Tides were found to upset diffusive equilibrium in the lower thermosphere by the periodic up- and down-welling which is caused by divergence of the horizontal wind field. Molecular diffusion times below around 150 km are too long to restore balance on the time scale of tidal oscillations. Below around 90 km, chemical lifetimes of atomic oxygen were found to be comparable with tidal periods, indicating that chemistry causes redistribution of constituents when the chemical balance is upset by the tides. The background winds are found below 100 km altitude to affect the distribution of molecular oxygen and nitrogen and above 100 km that of atomic oxygen as well. Where strong vertical gradients of horizontal winds occur, the winds are found to cause additional mixing of
constituents when accompanied by tidal up- and down-welling. In regions of tidal dissipation the changes to the background wind field can affect the global distribution of constituents.

VIII. 1. 3. COMPARISONS OF CTIM WITH DATA AND MODELS

Simulations from the CTIM were compared to various other middle- and upper atmosphere models as well as tidal measurements. While comparisons with the NCAR-TIGCM model [Roble et al., 1988] showed reasonable agreement, other models such as the GSWM [Hagan et al., 1993] and the empirical Hedin Wind Model (HWM) [Hedin et al., 1993] and MSISE90 [Hedin, 1991] generally gave larger amplitudes in the 100-130 km height regime than the CTIM, although the phases often matched fairly well in terms of their vertical gradients. These discrepancies could not be entirely explained, but ideas have been brought forward for further studies to investigate this. Comparisons with data were carried out for the ninth Lower Thermosphere Coupling Study (LTCS) campaign of January 1993, and the agreement was found to be good for meridional winds and less for temperatures and zonal winds. The discrepancies were primarily attributed to uncertainties in the global lower boundary forcing profile applied in CTIM. Since measurements during that campaign were available mostly for the northern hemisphere at low- to mid latitudes, with high-latitudes as well as large parts of the southern hemisphere not being considered, no unique global tidal profile was suggested by the data to be used as lower boundary forcing in CTIM. Generally, the issue of finding the best forcing profile in CTIM was found to be equally important and difficult to solve, given that local properties of tides are very much linked to their global profile, but quasi-global measurements are very rare. A number of factors have also been presented which give the measurements a degree of uncertainty. On the whole, though, comparisons have shown that the CTIM does reproduce some of the characteristics of tides in the thermosphere. Unresolved difficulties as well as suggestions for further development of the code are discussed in VIII.2.

VIII. 1. 4. PLANETARY WAVES

It was shown in Chapter VII that CTIM is capable of simulating upwardly propagating planetary waves as well. Basic properties of planetary waves which have in the past been suggested by other authors (eg. Salby, 1981a, b, c), could successfully be reproduced by CTIM. As an example, it was found that eastward winds prevented planetary waves from propagating vertically. The scientific aim of the CTIM studies was to investigate whether tides modulated by planetary waves would be
capable of carrying planetary wave signatures to altitudes of above 200 km, where planetary wave patterns had been measured in the ionosphere, conflicting with other observations which suggested that planetary waves in the neutral atmosphere dissipated below 120 km altitude. The simulations with CTIM found that modulated tides were not as sensitive to the zonal background winds as planetary waves, and produced strong planetary wave signatures between around 160 and 200 km height. Due to limitations of the current CTIM version, which is incapable of modelling the self-consistent neutral-ion coupling at low latitudes, planetary wave signatures were not convincingly reproduced in electron densities. Nevertheless, it was argued that the modelled signatures in the neutral atmosphere did suggest that modulated tides could produce ionospheric planetary waves as a result of changes in the ionospheric chemistry which were caused by up- and downwelling in the thermosphere near 200 km.

VIII. 2. FUTURE WORK

The studies presented in this thesis have shown many qualities of the CTIM in reproducing much of the thermospheric behaviour expected from theory and measurements. Below, a number of extensions to the present code are proposed which are intended to overcome some of the model's limitations found in some of the studies. These show that the improvements of the code which were carried out and presented in Chapter III, section III.5., formed an essential step towards more comprehensive modelling not only of the thermospheric- and ionospheric environment but also towards an understanding of coupling processes between the middle- and upper atmosphere.

VIII. 2.1. VERTICAL RESOLUTION

On several occasions it was found that CTIM's current vertical resolution of one scale height was insufficient to reproduce some of the tidal features. In the simulations presented throughout the thesis, semidiurnal Hough modes (2,2) up to (2,5) were used since higher order modes have vertical wavelengths of one scale height or less and cannot be resolved with the current code. This limits the flexibility of the latitudinal forcing profiles at the lower boundary. Particularly when trying to resolve accurately tidal profiles poleward of around ±60° latitude, higher order modes are important since they produce the strongest features nearer to the poles than the currently used modes. Also, the (1,1) mode which has a small vertical wavelength could be better resolved with smaller vertical grid spacing than it is currently.
Another difficulty found in some of the simulations may be linked to the limited vertical resolution. In Chapters IV and V it was found that the background atmospheric temperature under some circumstances was reduced throughout most of the thermosphere as a result of applying tidal forcing, rather than increased, which would be expected from the fact that lower boundary forcing provides an additional source of energy. This unexpected effect was found to be strongest when forcing with higher order Hough modes, such as the (2,4) and (2,5) mode. With (2,2) mode forcing the temperature increased slightly. This behaviour suggests the possibility of a link to the vertical wavelengths of tides. The release of energy to the background atmosphere occurs when a tide dissipates. In Chapter IV it was found that dissipation is also accompanied by the creation of higher order Hough modes, through a process commonly referred to as mode coupling. When mode coupling occurs as a result of (2,2) mode forcing, the (2,4) mode is generated. When the (2,4) mode dissipates, though, a higher order mode is generated which has a vertical wavelength smaller than the vertical resolution of CTIM. It is possible that the lack of vertical resolution causes inaccurate modelling of the energy dissipation process and thus the observed temperature effect, but numerical experiments are necessary to investigate this further.

Improving the vertical resolution will also provide better resolution of processes near the lower boundary. At present, 4 pressure levels (including the lower boundary) lie below the turbopause, thus not allowing comprehensive simulation of the turbulent mixing processes occurring there. With a finer vertical grid, that region below 100 km would be simulated in more detail, possibly influencing also the composition at higher altitudes. This will be of use also in studying tidally induced composition changes in the lower thermosphere, such as in Chapter VI, where it was found that chemical reactions of atomic oxygen with constituents such as OH and HO₂ played an important role under tidal forcing. However, this is at present insufficiently simulated since this chemistry has the relevant time scales up to an altitude of 90 km only, covering merely 3 pressure levels.

Tidal studies in Chapter V found that the height of tidal momentum dissipation was dependent on tidal amplitudes. Doubling the forcing amplitudes at the lower boundary was found to reduce the height of dissipation by around half a pressure level. With the current resolution, this shift could not be resolved accurately but was derived from fitting a curve to the vertical profiles of tidal amplitudes. On a finer grid, though, the different heights of dissipation would be resolved on different pressure levels. These examples shows that the quality of the tidal modelling studies
would be improved noticeably by a finer vertical grid.

As a result of these concerns, a modification of CTIM’s vertical resolution has been undertaken recently. However, a number of difficulties prevented the task from being completed at the time of writing. The code has been changed to allow specification of the vertical resolution to any fraction of one scale height. At present, this improved version of CTIM is in an experimental stage in which a number of numerical difficulties need to be resolved. The main concerns were found to be in the vertical as well as the horizontal gradients of winds and temperature, which produced unpredictable singularities. This is attributed to the fact that neighbouring grid points in both the horizontal and vertical planes have more similar values of parameters, and so the gradients are more susceptible to numerical errors with the finite difference technique of evaluation. This requires a better matching of smoothing to the problem — if smoothed too strongly, small-scale features disappear, whereas too weak smoothing will not remove the numerical artefacts. It is expected, though, that the new version of CTIM will be completed in due course and solve any problems associated with the limited vertical resolution.

VIII. 2. 2. DYNAMO FIELD

One important, and in terms of numerical modelling still relatively unexplored, area of study is the influence of upwardly propagating tides and other oscillations on the ionosphere. As described in Chapter I, neutral winds drag charged particles across the magnetic field lines in the low- to mid latitude $E$-region, causing a build-up of polarization charges and thus an electric field. This process is referred to as the dynamo effect, and the electric field created is the so-called dynamo electric field. The electric field is mapped along the conducting magnetic field lines into the $F$-region where it influences the dynamics of charged particles. At present, this electric dynamo field is implemented at low latitudes in CTIM in parameterized form (see Chapter III and Figure I.3), based on the model by Richmond et al. [1980]. Therefore, the tidal signatures in the neutral winds at $E$-region heights are not mapped into the ionosphere, and the external tidal forcing in CTIM affects primarily the neutral atmosphere.

This was also found to be an important constraint in the planetary wave study presented in Chapter VII. Chen [1992] found 2-day wave features in the equatorial ionization (or Appleton-) anomaly, a region on the magnetic equator which is characterized by lower charged particle densities than
surrounding regions. The anomaly is generally attributed to an effect commonly referred to as the “equatorial fountain” [Rishbeth and Garriott, 1969], where charged particles are transported to higher altitudes through upwelling above the magnetic equator. The effect can only be simulated when the dynamo field calculations are carried out self-consistently. It is therefore essential to carry out the relevant extension to the CTIM code in order to reproduce findings such as those by Chen.

At present, a new version of CTIP (see Chapter III, section III.2) is in an experimental stage at the Space Environment Center in Boulder, Colorado [G.Millward, private comm., 1997] and in first experiments has proven successful in reproducing the equatorial ionization anomaly. The new code is based on studies by Richmond et al. [1992], who carried out a similar improvement in the NCAR-TIGCM model (see Chapter V, section V.3.1). The promising initial results suggest that this improvement of the CTIP code will be available also for the tidal and planetary wave studies within roughly a year. It will thus allow an unprecedented variety of modelling research to be carried out. The calculations of the dynamo electric field rely on the neutral wind profiles, and therefore the improved lower boundary in CTIM and CTIP formed an essential step for the dynamo calculations.

VIII. 2.3. GRAVITY WAVES

Gravity waves are small-scale oscillations which originate often from orographic texture changes, such as mountains, which cause periodic upwelling, with gravity as the main restoring force. Gravity waves are known to strongly influence the background atmosphere, particularly in the upper mesosphere and lower thermosphere [Walterscheid, 1995]. A typical example is the mesopause temperature anomaly. The summer hemisphere at the mesopause is considerably cooler than the winter hemisphere. In fact, the summer mesopause with temperatures as low as 140 °K represents the coolest region in the Earth’s atmosphere [Rishbeth and Garriott, 1969]. The general belief is that gravity wave drag in the summer hemisphere near the mesopause damps considerably the westward winds which follow from Coriolis forces acting on the meridional winds, thus enabling transport of energy to the winter hemisphere [Garcia and Solomon, 1985] and setting up a large-scale circulation.

Lindzen [1970] examined gravity waves breaking in the mesosphere and predicted that this process would considerably influence the zonally averaged structure of the mesosphere and lower
thermosphere. A parameterization for both the diffusion and momentum deposition due to breaking gravity waves was presented by Lindzen [1981]. This parameterization was used in simulations by Garcia and Salomon, [1985] who confirmed the strong influence of gravity waves on the background atmosphere. It is important to note that this parameterization did not represent the gravity waves themselves, but rather their influence on the background atmosphere. With gravity wave periods being smaller than those of tidal oscillations, interactions between gravity waves and tides have an important influence on the tidal response profiles as well. Few successful attempts have been made to-date to simulate or understand in detail these interactions, mainly because of the difficulty of parameterizing the gravity waves themselves [Forbes, 1995].

To account for the effects of gravity waves on the lower thermosphere, an important future study would be to implement into CTIM a parameterized scheme such as that of Lindzen [1981] for investigating the influence of this on the tides and background atmosphere. Such simulations could be compared to results such as those by Walterscheid [1981], Fritts and Vincent [1987] and Wang and Fritts [1991]. The former two are modelling studies investigating the tides-gravity wave interactions. Forbes et al. [1991] numerically simulated the effects of gravity wave stresses on the diurnal tide, using the parameterization by Lindzen. These modelling studies are limited to heights below 140 km, and to-date no simulations have been carried out for higher altitudes. The relatively simple parameterization by Lindzen would require few systematic changes to the current CTIM code, and the novelty of such a study for the thermospheric environment is a strong motivation for exploring this in the near future.

VIII. 2. 4. LOWER THERMOSPHERE CHEMISTRY

It was shown in Chapter VI that chemical reactions of O with OH and HO₂ in the thermosphere below 90 km play a role in the tidally induced composition changes of the thermosphere. It was, however, also pointed out in VIII.2.1 that the current vertical grid resolution does not allow any detailed study with these reactions since they are of importance only on the lower 3 pressure levels of CTIM. Introducing a finer vertical grid, therefore, would allow an improvement in these calculations and make them more reliable.

Furthermore, a number of chemical reactions occur in the lower thermosphere which influence the three major constituents O, O₂ and N₂, as well as a further constituent which is currently not
considered, nitric oxide (NO). Around 170 km altitude, NO is responsible for radiative cooling of up to around 50°K per day [Roble, 1995]. In the lower thermosphere, photochemical lifetimes of NO are similar to the transport time scales of horizontal winds (see Chapter VI, section VI.5), implying that NO transport occurs through the lower thermosphere winds [Brasseur and Solomon, 1986]. Simulations by Brasseur and Salomon have shown a strong sensitivity of NO densities in the lower thermosphere to the dynamics of the region. Therefore, levels of NO are also expected to be affected by the tides, and this provides another way in which tides can influence the thermosphere, also at higher altitudes. Implementing the self-consistent odd nitrogen chemistry in CTIM would also, as mentioned earlier, affect the O, O₂ and N₂ densities in the lower thermosphere, and it would be of great interest to investigate the effects of this on the thermosphere and possibly the ionosphere. Attempts at implementing these chemical reaction schemes into CTIM are currently being made by T.J. Fuller-Rowell at the Space Environment Center in Boulder, Colorado, but it is unclear at this stage when these will be available for use.

Further cooling in the lower thermosphere is provided by carbon dioxide (CO₂), which has its strongest effect at around 120 km height with a rate of around 70°K per day [Roble, 1995]. A self-consistent treatment of CO₂ densities would require a chemical reaction scheme to be implemented which involves the species CH₄, CO and OH. The CO₂ is produced by three-body reactions of CO with O and by reactions of CO with OH. The CO is produced by reactions of CH₄ with O, O(¹D) or OH. It is evident from this that an implementation of carbon dioxide chemistry involves also a number of other species which are currently not considered in CTIM. Although CO₂ acts as a coolant in the thermosphere, some of the chemical reactions, such as the oxidation of CO, involve production of energy, and the extended chemistry might have an influence on the overall vertical energy distribution of the lower thermosphere.

The chemistry described above has an influence on lower thermosphere properties and should be included in CTIM. Historically, it was not implemented since the lower thermosphere region with the old lower boundary was not assumed to be realistic in terms of dynamics, and the main application of the model lay in the thermosphere and ionosphere above 200 km altitude. With the new lower boundary and an interest in carrying out lower thermosphere studies as well, the shortcomings of the current model version in terms of chemistry should be addressed. A further reason for doing so is the long-term aim of coupling CTIM to a middle atmosphere model (see following section).
VIII. 2.5. LOWER BOUNDARY EXTENSION

Full and self-consistent coupling between the upper and the middle atmosphere can be achieved by coupling CTIM to a middle atmosphere model. Technically, this task would correspond to that undertaken in 1987, of coupling the UCL Thermosphere model by Fuller-Rowell and Rees [1980] to the Sheffield University high latitude ionosphere model by Quegan et al. [1982], forming the CTIM model [Fuller-Rowell et al., 1987]. Before coupling these models, initial studies had been carried out by swapping output parameters between the separate codes. A similar procedure could be followed when testing any coupling of CTIM to a middle atmosphere model.

A necessary condition for carrying out such a task is that the models overlap sufficiently in terms of height range that values of the middle atmosphere model are reliable at the altitude of CTIM's lower boundary (80 km). In practice, unfortunately, many available middle atmosphere models only reach to altitudes of around 70-90 km, which is insufficient for any coupling. One possible scenario might therefore be to extend CTIM's current lower boundary down by around 4 pressure levels, or roughly 25 km. Then one could attempt coupling studies with models such as the UGAMP middle atmosphere model. Such a downwards extension would however ideally need to be accompanied by an extension of the chemical scheme in CTIM (see VIII.2.4).

A simpler approach could be used in the first instance, without changing the current lower boundary, if considering only the upwardly directed influence, and not CTIM's feedback into the mesosphere. In Chapter VII, a first example was presented of using a complete set of parameters produced by a lower- and middle atmosphere model at the lower boundary of CTIM. In that example only the perturbations of geopotential height, winds and temperature were implemented from the GSWM model [Hagan et al., 1993], but similarly, the background profiles of these and other parameters can also be used. This would allow for considering effects such as the meridional temperature gradient of the mesopause at solstice, which includes the temperature anomaly referred to in VIII.2.3. The GSWM model and the model by Forbes and Vial [1989] are ideal in terms of height overlap, but they are both perturbation models, implying that the background atmospheres are parameterized and only the perturbations are calculated self-consistently. This may be sufficient to improve the current lower boundary profiles and involves no further development of the code. Such studies can therefore be carried out easily in the near future.
It was found in Chapter V that one crucial aspect of comparing modelled tides to measurements was to find an accurate lower boundary forcing profile in CTIM. The quality of this profile depends strongly on the global coverage of measurements, and ideally instruments at the low-, mid- and high latitudes in both hemispheres should be used to carry out tidal measurements at overlapping altitude ranges simultaneously. One of the principal aims of the Lower Thermosphere Coupling Study (LTCS) campaigns is to perform such measurements in order to improve understanding of global perturbations such as tides and planetary waves. Between 1987 and early 1997, 15 such LTCS campaigns have been carried out, or on average around 1.5 per year. This allows one to study tides and planetary waves globally for particular periods of time, but climatological studies are not possible. Long-term measurements at single locations show evidence of semiannual variations of tidal amplitudes in the mesosphere and lower thermosphere, but global long-term measurements are needed to investigate these in detail with numerical models such as CTIM. Global tidal measurements such as those in the LTCS campaigns should therefore be carried out more frequently, ideally at intervals of around one month.

The thesis has also shown that the investigation of the propagation of tides into the thermosphere requires measured tidal amplitudes above 100 km altitude. At present, radar measurements of ion drifts give a height coverage of up to around 130 km, implying that much of the region between 140 and around 200 km altitude is currently not measured, least of all in global campaigns. That height regime is therefore still poorly known in terms of tidal oscillations of the neutral atmosphere. Research in the future should try to address this issue, possibly by developing new measurement techniques which would allow a height coverage which the current techniques are not capable of providing.
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REFERENCES


Bauer, S.J. (1973) Physics of planetary ionospheres, Springer Verlag


Booker, J.R., and F.P. Bretherton (1967), The critical layer for internal gravity waves in a shear flow, J. Fluid Mech., 27, 513-519


References


Evans, J.V.(1978) Incoherent scatter contributions to studies of the dynamics of the lower thermosphere, Rev. Geophys. Space Phys., 16, 195

References


Fritts, D.C., and R.A. Vincent (1987) Mesospheric momentum flux studies at Adelaide, Australia:
Observations and a gravity wave-tidal interaction model, *J. Atmos. Sci.*, 44, 605-619


Glass, M., J.L. Fellous, M. Massebeuf, A. Spizzichino, I.A. Lysenko, and Y.I. Portnyagin (1975) Comparison and interpretation of the results of simultaneous wind measurements in the lower
thermosphere at Garchy (France) and Obninsk (USSR) by meteor radar technique, 
*J. Atmos. Terr. Phys.*, 37, 1077-1087


Harris, M.F. (1959) Diurnal and semidiurnal variations of wind, pressure and temperature in the troposphere and stratosphere over the Azores, *J. Atmos. Sci.*, 19, 136-149


299
References


Lindzen, R.S., and D.J. McKenzie (1967) Tidal theory with Newtonian cooling, Pageoph, 66, 90-96

300
References


Lindzen, R.S. (1970) Internal gravity waves in the atmosphere with realistic dissipation and temperature, I, Mathematical development and propagation of was into the thermosphere, *Geophys. Fl. Dyn.*, 1, 303-355


Miers, B.T. (1965) Wind oscillations between 30 and 60 km over White Sands Missile Range, New Mexico, *J. Atmos. Sci.*, 22, 382-387


References


302
References


Richmond, A.D. (1979a) Large amplitude gravity wave energy production and dissipation in the thermosphere, J. Geophys. Res., 84, 1880-1890


303


References


305
References


Wilkes, M.V. (1949) Oscillations of the Earth’s atmosphere, *Cambridge University Press*


Young, C., and E.S.Epstein (1962) Atomic oxygen in the polar winter mesosphere, *J.Atmospheric Sci.*, **19**, 435,