AERONOMY AND THE MAGNETOSPHERE

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Summary

Neutral atmospheric structure is usually defined in terms of temperature. The part of the atmosphere in which we live, the troposphere, is a region of decreasing temperatures with altitude. Above it there is a region where temperatures increase with altitude, the stratosphere. These two regions are not subjects of this article. Instead the article is concerned with the regions, both neutral and charged, that lie above the top of the stratosphere. Although this region is a vacuum by the standards of someone on the earth's surface, there is still a significant neutral gas up to heights of over 500 km. Two more atmospheric regions occur here: the mesosphere, a region where temperatures decrease with altitude between about 45 and 90 km; and the thermosphere, a region of increasing or constant temperatures with altitude that occurs above the mesosphere. As well as the neutral gas, the atmosphere also has an ionized component.

The sun's corona and chromosphere emit radiation in the Extreme Ultraviolet (EUV) and Far Ultraviolet (FUV). When this radiation reaches the earth it ionizes the upper parts of the atmosphere. Ionization by this electromagnetic radiation creates concentrations of ions and electrons down to about 50–60 km. At these altitudes the atmosphere is only very lightly ionized: ion densities are only about 10^3 to 10^4 cm^{-2},

whereas neutral densities are about 10^{15} cm^{-2}. The relative importance of ions (and electrons) increases as altitude increases. Even so, at about 300 km, where the greatest ion densities in the atmosphere are found, ions still make up only about 0.001 of the total gas density or less. As altitudes increase further, ions (and electrons) become relatively more important, so that, above about 1000 to 2000 km, most of the gas consists of ions (and electrons) and the gas is now almost entirely plasma. The part of this region that corotates with the earth is called the "plasmasphere." Higher still the nature of the gas is strongly controlled by the earth's magnetic field in a region that is called the "magnetosphere." At the top of this region, control passes from the earth's magnetic field to the solar wind and the magnetic field embedded in it. Each of these regions has its own characteristics: these are outlined in the body of this article.

1. The Neutral Atmosphere

The atmosphere is usually defined in terms of the way temperatures vary with altitude (see Figure 1). Temperatures decrease from the ground up to heights of about 10 km; this region is called the "troposphere." Above 10 km, temperatures start to increase again as a result of residual heat from the absorption of UV radiation by ozone. The region in which these increasing temperatures occur is called the "stratosphere" (10–50 km). Ozone heating becomes less significant at greater heights and the region above 50 km, where temperatures decrease, becomes increasingly dominated by cooling as a result of CO_2 radiation processes: this region is called the "mesosphere" (50–90 km). The last atmospheric region that is defined by its neutral temperature structure is called the "thermosphere" (90–500 km). The last two regions are the subjects of this section.

Significant ionization occurs down to about 70 km, one height that might provide a boundary for the discussion in this article. Another useful lower boundary is the region below which significant interaction with tropospheric weather occurs. The stratopause provides a convenient boundary using this definition. Henceforth, this article will only discuss the major features of the regions above this altitude.

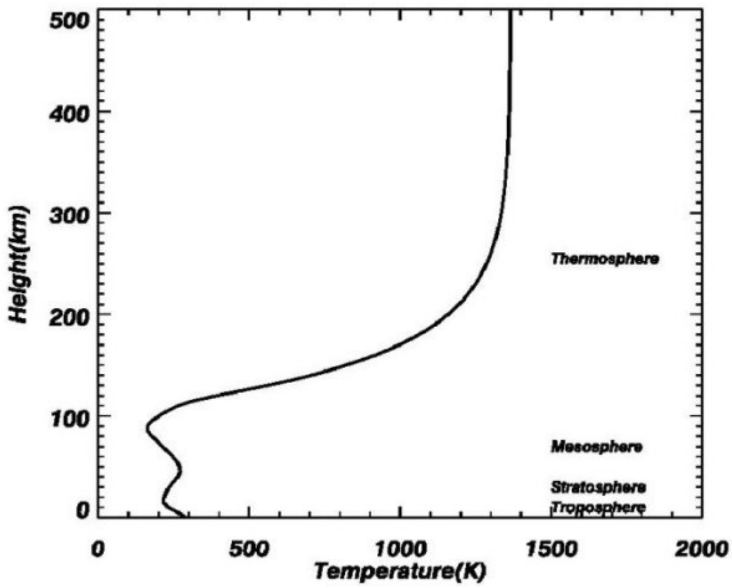

Figure 1.This figure plots the temperature variation of the neutral atmosphere with height above the earth. Several regions of temperature decrease and increase occur in the earth's atmosphere, corresponding to the various layers in the atmosphere. The thermosphere is the region above about 90 km, where temperatures increase from the lows at the mesopause (150 K to 220 K depending on the season) up to about 250 km, above which they are constant with altitude. The temperatures on this plot are appropriate for a daytime, summer, and solar maximum case. This graph was drawn using calculations from MSIS 90.

1.1. The Mesosphere

The mesosphere is characterized primarily by its thermal structure. In this region temperatures decrease from the stratopause to the mesopause. Warm temperatures occur at its lower boundary as a result of the heat produced by the absorption of ultraviolet radiation by ozone. Cold temperatures occur at the top of the mesosphere because there is no strong *in situ* heat source in that region and there is a strong cooling mechanism, radiative cooling by CO_2.

Seasonal variations of temperature at the stratopause are similar to those near the ground; temperatures at this height are hotter in summer than they are in winter. But temperatures at the mesopause are significantly hotter in winter than in summer. In fact, the coldest temperatures reported in the atmosphere have occurred in the summer mesopause region. This apparent incongruity can be explained by the circulation pattern in the mesosphere. As solar energy heats the ozone layer more in summer than in winter, and cooling is similar in each season, there is a net upwelling in summer and a net downwelling in winter. The resulting circulation cell is closed by a summer-to-winter flow in the upper mesosphere. Internal gravity wave breaking makes important, even dominant contributions to this circulation pattern, as the momentum deposited by this breaking near the mesopause helps to drive the summer-to-winter flow and thus also increases upwelling in summer and downwelling in winter. This breaking is discussed further in the next paragraph. The circulation pattern that these processes create affects the morphology of the temperature field near the mesopause through compression and expansion. As air upwells it expands and cools; as it downwells it compresses and heats. Thus, the warm winter mesopause results from compressional heating and the cold summer mesopause results from cooling by expansion.

Internal gravity wave breaking plays an important role in determining mesospheric circulation, as was mentioned in the previous paragraph. Internal gravity waves are generated by a variety of mechanisms in the troposphere and propagate up into the mesosphere. In the mesosphere, they typically have periods from about five minutes (the Brunt-Vaisala period), to a few hours, horizontal wavelengths from tens to hundreds of kilometers, and vertical wavelengths of greater than a few kilometers. They break in the mesosphere for two reasons. First, temperatures decrease with height in the mesosphere, creating an unstable region where convective instabilities can occur. Second, the amplitude of upwardly propagating waves increases exponentially in the absence of dissipation. Thus, the mesosphere is a region where waves, and particularly internal gravity waves, tend to become unstable and dissipate their momentum. In the lower parts of the mesosphere this momentum acts as a drag on the mean wind, reducing the amplitude of the westward jets in the summer mesosphere and the eastward jets in the winter hemisphere. But the jets themselves affect the deposition of momentum by gravity waves. Westward jets in summer preferentially decrease the westward momentum of the upwardly propagating gravity waves. Similarly, eastward jets in winter preferentially decrease the eastward momentum associated with the upwardly propagating waves. Thus, above the jets, internal gravity waves deposit eastward momentum in summer and westward momentum in winter, causing a reversal of the zonal circulation and (as a result of changes in the Coriolis force driven by this circulation) an increase in the summer-to-winter flow.

Another characteristic of the mesopause is the importance of waves to the dynamics and thermodynamics of the region. The influence of internal gravity waves in determining the general circulation of the mesosphere was discussed in the preceding paragraph. They are also an influential component of local wind fields. Other waves are also important thermodynamically and dynamically. Large diurnal tides occur near the equator, and the semi-diurnal tide is important throughout the region. Longer period oscillations are also seen. Although planetary waves are mainly observed in the lower parts of the region in winter, some planetary wave modes may be present anywhere in the mesosphere at any time. At an even longer period, the quasi-biennial oscillation is important in the mesosphere. But other dynamical processes that are important at greater altitudes, such as solar cycle effects and geomagnetic storm effects, are of much less importance in the mesosphere.

The major gases in the mesosphere are thoroughly mixed by the turbulence generated by the dissipation of gravity waves and tides. Thus the proportion of N_2 in the atmosphere is 78% at the mesopause, the same as it is at the ground. However, important chemical reactions involving minor species occur throughout the mesosphere. For example, water vapor, which itself varies greatly with time and location in the mesosphere, can undergo photodissociation, especially near the mesopause. This dissociation creates significant amounts of hydroxide (OH) in the upper parts of the mesosphere. In turn, this OH reacts with ozone altering its distribution. Other trace species play important roles in the mesosphere, but, in general, their chemistry is relatively complicated and is thus beyond the scope of this brief survey.

1.2. The Thermosphere

The thermosphere is a region that is characterized by very hot temperatures, rising, occasionally, to over 2000 K in the upper thermosphere. This maximum temperature is highly variable, especially over a solar cycle. At solar minimum, normal daytime temperatures are about 600–700 K, whereas normal, daytime, solar maximum temperatures are about 1400 K. These high temperatures occur as a result of two factors. First, energetic EUV radiation, produced in the sun's corona, ionizes some of the neutral gas in the upper thermosphere. Some of the electrons produced by this ionization are thermally excited. These collide with ions, which, in turn, collide with and heat the neutral gas. The second factor is that there is no strong internal cooling mechanism in the thermosphere. Therefore the hot gas, which was created as part of the ionization process, is mainly cooled by the relatively inefficient process of downward heat conduction.

Another major feature of the thermosphere is diffusive separation. In the atmosphere below about 100 km air is mixed uniformly by turbulence, so the proportion of the major gases N_2 and O_2 is approximately the same at 100 km as it is at the ground. The region below 100 km is sometimes called the “homosphere” for this reason. Above about 100 km molecular diffusion and ion drag prevent the formation of turbulence, so this mixing cannot occur. Instead, each species decreases vertically at a rate that is dependent on the mass of the species. Therefore, although N_2 is the major gas at 100 km, while the lighter gas O is dominant at 300 km. Each species is roughly in diffusive equilibrium above the turbopause, hence the term “diffusive separation.” This region is sometimes called the “heterosphere.”

The thermosphere changes in a number of ways as a result of changes in external forcing. There are long-term changes that should be regarded as climatological. Several different forcing mechanisms lead to this climatology. The sun’s output in the EUV and UV wavelengths varies in 11-year cycles, called “solar cycles.” At solar maximum, high levels of this radiation can cause temperatures in the upper thermosphere to reach 1400 K, as discussed earlier. At solar minimum, the much-reduced levels of radiation lead to maximum temperatures of only 600 K in the upper thermosphere. These temperature changes over a solar cycle also lead to large variations in density at a particular altitude. The difference in density over a solar cycle at 350 km, for example, can be more than an order of magnitude. Winds in the upper thermosphere are less affected by these solar cycle variations. This lack of major wind variations occurs because, although the increased temperatures at solar maximum create greater pressure gradients to drive the winds, these gradients are balanced by increases in ion drag that result from the enhanced electron densities. Solar output also varies on a 27-day cycle (the solar rotation period). Upper thermospheric temperatures, winds, and composition are also affected by this cycle.

Seasonal variations occur as well. Temperatures are much hotter in summer than in winter and this, in turn, drives summer-to-winter winds. Because winds blow away from the summer hemisphere, there is more upwelling in this season and consequently the upper thermosphere is richer in molecular species than the winter is. The relative preponderance of atomic species in winter causes the “winter anomaly” in electron densities that will be discussed in the next section.

Another feature that has been observed in the thermosphere is the semiannual oscillation. Temperatures, winds, and composition have been observed to oscillate every six months. Although this effect has been known since the beginning of the space age, explanations for it have proved more elusive. It has been suggested that upwardly propagating waves cause the effect, but it has proved difficult to model these effects to produce reasonable amplitudes for this phenomenon. A recent explanation is called the “thermospheric spoon.” This explanation involves the differences in circulation in the equinoxes, when upwelling occurs primarily at the equator, and the solstices when it occurs at high latitudes in the summer hemisphere. The consequent differences in the thermal, dynamical, and compositional structure appear to be sufficient to explain the semiannual oscillation.

Although they are not, in a sense, climatological, there are also regular daily variations in the thermosphere. The sun heats the thermosphere from dawn to dusk, which would lead to the highest temperatures in the day occurring at the later time if other processes were not occurring. Solar heating also causes upwelling. This upwelling, in turn, results in expansion of the gas and hence cooling. By about 1500 hours this cooling, and the decreased heat input that occurs as the sun gets closer to the horizon, lead to net thermospheric cooling. Thus maximum temperatures occur at about 1500 hours. Similarly thermospheric compressional heating resulting from downward winds at night acts, in part, to counterbalance the effects of downward heat conduction. Winds blow away from the temperature maximum, so they are predominantly westward during the daylight hours and eastward at night. Composition is also affected by this diurnal cycle. Upwelling in the daytime leads to increases in the molecular species, whereas downwelling at night causes enhanced densities of the lighter atomic species.

One interesting consequence of the diurnal temperature cycle is a super-rotation of the neutral winds. To the first order, the ionosphere occurs at one altitude during the daylight hours and at another, higher altitude at night. Therefore there is a height region in which the neutral winds are retarded by ion drag during the day, but not at night. As the winds are westward during the day and eastward at night this leads to an eastward super-rotation. In addition the upper thermosphere is very viscous and vertical gradients in the horizontal wind are hard to maintain. Thus this super-rotation is also driven at most altitudes in this region.

"Weather" effects also occur in the thermosphere. In this case weather refers to non-regular, relatively short-period changes in the region. Certainly upwardly propagating waves affect the region, but their influence throughout the thermosphere has not been fully investigated. More dramatic changes occur during geomagnetic storms. These will be described in a later section.

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Biographical Sketches

Dr. Alan Burns, is a scientist at the National Center for Atmospheric Research (NCAR). Dr. Burns received his Ph.D. from the University of Canterbury in New Zealand in 1986. He was a Postdoctoral Research Fellow at the Space Physics Research Laboratory (SPRL) at the University of Michigan for two years and a Research Scientist there from 1988 to 2000. He has been at NCAR since then. In addition, he is an Associate Editor of the *Journal of Atmospheric and Solar-Terrestrial Physics*. His research interests center on the thermosphere and ionosphere and their coupling with other regions. He uses both modeling and data analysis techniques to study how various thermospheric parameters, and particularly neutral composition, respond to external forcing.

Dr. Timothy Killeen, is the Director of NCAR. Dr. Killeen received his Ph.D. from the University College London (UCL) in 1975. He was Research Assistant at UCL from 1975 to 1978, a Postdoctoral Research Fellow in SPRL at the University of Michigan from 1978 to 1979, a Research Scientist at SPRL from 1979 to 1987, an Associate Professor in the Department of Atmospheric Oceanic and Space Science (AOSS) from 1987 to 1990, and a Full Professor between 1990 and 2000. He was the Director of SPRL from 1993 to 1998 and an Associate Vice President for Research at the University of Michigan from 1997 to 2000. He was appointed Director of NCAR in 2000. In addition, he is Editor-in-Chief of the *Journal of Atmospheric and Solar-Terrestrial Physics.* His research interests center on experimental and theoretical studies of the thermosphere, ionosphere and upper mesosphere. A particular interest of his has involved studying the dynamics of the high latitude thermosphere.