

Thermosphere

The **thermosphere** is the layer of the Earth's atmosphere directly above the mesosphere and below the exosphere. Within this layer of the atmosphere, ultraviolet radiation causes photoionization/photodissociation of molecules, creating ions in the ionosphere. Taking its name from the Greek θερμός (pronounced *thermos*) meaning heat, the thermosphere begins about 85 kilometres (53 mi) above the Earth.^[1] At these high altitudes, the residual atmospheric gases sort into strata according to molecular mass (see turbosphere). Thermospheric temperatures increase with altitude due to absorption of highly energetic solar radiation. Temperatures are highly dependent on solar activity, and can rise to 2,000 °C (3,630 °F). Radiation causes the atmosphere particles in this layer to become electrically charged (see ionosphere), enabling radio waves to be refracted and thus be received beyond the horizon. In the exosphere, beginning at 500 to 1,000 kilometres (310 to 620 mi) above the Earth's surface, the atmosphere turns into space, although by the criteria set for the definition of the Kármán line, the thermosphere itself is part of space.

The highly diluted gas in this layer can reach 2,500 °C (4,530 °F) during the day. Even though the temperature is so high, one would not feel warm in the thermosphere, because it is so near vacuum that there is not enough contact with the few atoms of gas to transfer much heat. A normal thermometer might indicate significantly below 0 °C (32 °F), at least at night, because the energy lost by thermal radiation would exceed the energy acquired from the atmospheric gas by direct contact. In the anacoustic zone above 160 kilometres (99 mi), the density is so low that molecular interactions are too infrequent to permit the transmission of sound.

The dynamics of the thermosphere are dominated by atmospheric tides, which are driven by the very significant diurnal heating. Atmospheric waves dissipate above this level because of collisions between the neutral gas and the ionospheric plasma.

The International Space Station orbits the Earth within the middle of the thermosphere, between 330 and 435 kilometres (205 and 270 mi).

Contents

Neutral gas constituents

Energy input

- Energy budget
- Solar XUV radiation
- Solar wind
- Atmospheric waves

Dynamics

Thermospheric storms

See also

References

Neutral gas constituents

It is convenient to separate the atmospheric regions according to the two temperature minima at about 12 km altitude (the tropopause) and at about 85 km (the mesopause) (Figure 1). The thermosphere (or the upper atmosphere) is the height region above 85 km, while the region between the tropopause and the mesopause is the middle atmosphere (stratosphere and mesosphere) where absorption of solar UV radiation generates the temperature maximum near 45 km altitude and causes the ozone layer.

The density of the Earth's atmosphere decreases nearly exponentially with altitude. The total mass of the atmosphere is $M = \rho_A H \approx 1 \text{ kg/cm}^2$ within a column of one square centimeter above the ground (with $\rho_A = 1.29 \text{ kg/m}^3$ the atmospheric density on the ground at $z = 0 \text{ m}$ altitude, and $H \approx 8 \text{ km}$ the average atmospheric scale height). 80% of that mass is concentrated within the troposphere. The mass of the thermosphere above about 85 km is only 0.002% of the total mass. Therefore, no significant energetic feedback from the thermosphere to the lower atmospheric regions can be expected.

Turbulence causes the air within the lower atmospheric regions below the turbopause at about 110 km to be a mixture of gases that does not change its composition. Its mean molecular weight is 29 g/mol with molecular oxygen (O_2) and nitrogen (N_2) as the two dominant constituents. Above the turbopause, however, diffusive separation of the various constituents is significant, so that each constituent follows its own barometric height structure with a scale height inversely proportional to its molecular weight. The lighter constituents atomic oxygen (O), helium (He), and hydrogen (H) successively dominate above about 200 km altitude and vary with geographic location, time, and solar activity. The ratio N_2/O which is a measure of the electron density at the ionospheric F region is highly affected by these variations.^[2] These changes follow from the diffusion of the minor constituents through the major gas component during dynamic processes.

The thermosphere contains an appreciable concentration of elemental sodium located in a 10-km thick band that occurs at the edge of the mesosphere, 80 to 100 km above Earth's surface. The sodium has an average concentration of 400,000 atoms per cubic centimeter. This band is regularly replenished by sodium sublimating from incoming meteors. Astronomers have begun utilizing this sodium band to create "guide stars" as part of the optical correction process in producing ultra-sharp ground-based observations.^[3]

Energy input

Energy budget

The thermospheric temperature can be determined from density observations as well as from direct satellite measurements. The temperature vs. altitude z in Fig. 1 can be simulated by the so-called Bates profile:^[4]

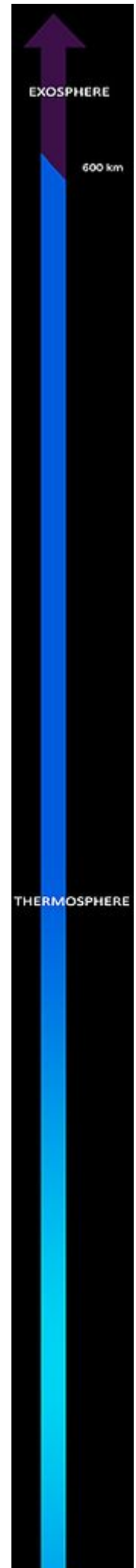
$$(1) \quad T = T_\infty - (T_\infty - T_0) \exp\{-s(z - z_0)\}$$

with T_∞ the exospheric temperature above about 400 km altitude, $T_0 = 355 \text{ K}$, and $z_0 = 120 \text{ km}$ reference temperature and height, and s an empirical parameter depending on T_∞ and decreasing with T_∞ . That formula is derived from a simple equation of heat conduction. One estimates a total heat input of $q_0 \approx 0.8$ to 1.6 mW/m^2 above $z_0 = 120 \text{ km}$ altitude. In order to obtain equilibrium conditions, that heat input q_0 above z_0 is lost to the lower atmospheric regions by heat conduction.

The exospheric temperature T_∞ is a fair measurement of the solar XUV radiation. Since solar radio emission F at 10.7 cm wavelength is a good indicator of solar activity, one can apply the empirical formula for quiet magnetospheric conditions:^[5]

$$(2) \quad T_\infty \approx 500 + 3.4F_0$$

with T_∞ in K, F_0 in $10^{-2} \text{ W m}^{-2} \text{ Hz}^{-1}$ (the Covington index) a value of F averaged over several solar cycles. The Covington index varies typically between 70 and 250 during a solar cycle, and never drops below about 50. Thus, T_∞ varies between about 740 and 1350 K. During very quiet magnetospheric conditions, the still continuously flowing magnetospheric energy input contributes by about 250 K to the residual temperature of 500 K in eq.(2). The rest of 250 K in eq.(2) can be attributed to atmospheric waves generated within the troposphere and dissipated within the lower thermosphere.



Solar XUV radiation

The solar X-ray and extreme ultraviolet radiation (XUV) at wavelengths < 170 nm is almost completely absorbed within the thermosphere. This radiation causes the various ionospheric layers as well as a temperature increase at these heights (Figure 1). While the solar visible light (380 to 780 nm) is nearly constant with a variability of not more than about 0.1% of the solar constant,^[6] the solar XUV radiation is highly variable in time and space. For instance, X-ray bursts associated with solar flares can dramatically increase their intensity over preflare levels by many orders of magnitude over a time span of tens of minutes. In the extreme ultraviolet, the Lyman α line at 121.6 nm represents an important source of ionization and dissociation at ionospheric D layer heights.^[7] During quiet periods of solar activity, it alone contains more energy than the rest of the XUV spectrum. Quasi-periodic changes of the order of 100% or greater, with periods with period of 27 days and 11 years, belong to the prominent variations of solar XUV radiation. However, irregular fluctuations over all time scales are present all the time.^[8] During low solar activity, about half of the total energy input into the thermosphere is thought to be solar XUV radiation. Evidently, that solar XUV energy input occurs only during daytime conditions, maximizing at the equator during equinox.



Earth atmosphere diagram showing the exosphere and other layers. The layers are to scale. From Earth's surface to the top of the stratosphere (50 kilometres (31 mi)) is just under 1% of Earth's radius.

Solar wind

A second source of energy input into the thermosphere is solar wind energy which is transferred to the magnetosphere by mechanisms that are not well understood. One possible way to transfer energy is via a hydrodynamic dynamo process. Solar wind particles penetrate into the polar regions of the magnetosphere where the geomagnetic field lines are essentially vertically directed. An electric field is generated, directed from dawn to dusk. Along the last closed geomagnetic field lines with their footpoints within the auroral zones, field aligned electric currents can flow into the ionospheric dynamo region where they are closed by electric Pedersen and Hall currents. Ohmic losses of the Pedersen currents heat the lower thermosphere (see e.g., Magnetospheric electric convection field). In addition, penetration of high energetic particles from the magnetosphere into the auroral regions enhance drastically the electric conductivity, further increasing the electric currents and thus Joule heating. During quiet magnetospheric activity, the magnetosphere contributes perhaps by a quarter to the thermosphere's energy budget.^[9] This is about 250 K of the exospheric temperature in eq.(2). During very large activity, however, this heat input can increase substantially, by a factor of four or more. That solar wind input occurs mainly in the auroral regions during both day and night.

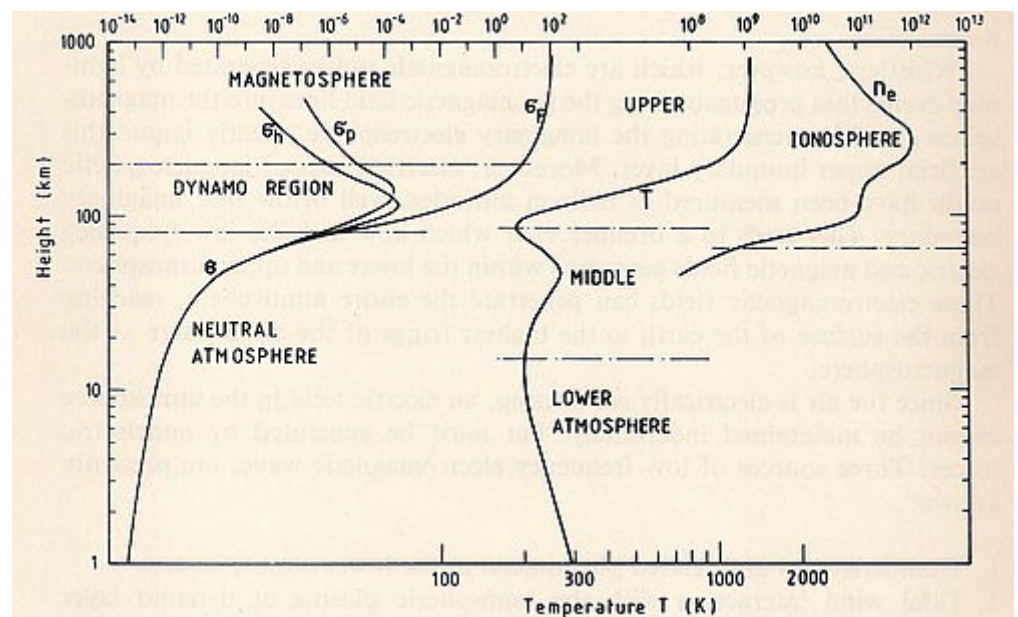


Figure 1. Nomenclature of atmospheric regions based on the profiles of electric conductivity (left), temperature (middle), and electron number density in m^{-3} (right)

Atmospheric waves

Two kinds of large-scale atmospheric waves within the lower atmosphere exist: internal waves with finite vertical wavelengths which can transport wave energy upward; and external waves with infinitely large wavelengths which cannot transport wave energy.^[10] Atmospheric gravity waves and most of the atmospheric tides generated within the troposphere belong to the internal waves. Their density amplitudes increase exponentially with height, so that at the mesopause these waves become turbulent and their energy is dissipated (similar to breaking of ocean waves at the coast), thus contributing to the heating of the thermosphere by about 250 K in eq.(2). On the other hand, the fundamental diurnal tide labelled (1, -2) which is most efficiently excited by solar irradiance is an external wave and plays only a marginal role within lower and middle atmosphere. However, at thermospheric altitudes, it becomes the predominant wave. It drives the electric Sq-current within the ionospheric dynamo region between about 100 and 200 km height.

Heating, predominately by tidal waves, occurs mainly at lower and middle latitudes. The variability of this heating depends on the meteorological conditions within troposphere and middle atmosphere, and may not exceed about 50%.

Dynamics

Within the thermosphere above about 150 km height, all atmospheric waves successively become external waves, and no significant vertical wave structure is visible. The atmospheric wave modes degenerate to the spherical functions P_n^m with m a meridional wave number and n the zonal wave number ($m = 0$: zonal mean flow; $m = 1$: diurnal tides; $m = 2$: semidiurnal tides; etc.). The thermosphere becomes a damped oscillator system with low-pass filter characteristics. This means that smaller-scale waves (greater numbers of (n,m)) and higher frequencies are suppressed in favor of large-scale waves and lower frequencies. If one considers

very quiet magnetospheric disturbances and a constant mean exospheric temperature (averaged over the sphere), the observed temporal and spatial distribution of the exospheric temperature distribution can be described by a sum of spheric functions.^[11]

$$(3) \quad T(\varphi, \lambda, t) = T_{\infty} \{1 + \Delta T_2^0 P_2^0(\varphi) + \Delta T_1^0 P_1^0(\varphi) \cos[\omega_a(t - t_a)] + \Delta T_1^1 P_1^1(\varphi) \cos(\tau - \tau_d) + \dots\}$$

Here, it is φ latitude, λ longitude, and t time, ω_a the angular frequency of one year, ω_d the angular frequency of one solar day, and $\tau = \omega_d t + \lambda$ the local time. $t_a = \text{June 21}$ is the date of northern summer solstice, and $\tau_d = 15:00$ is the local time of maximum diurnal temperature.

The first term in (3) on the right is the global mean of the exospheric temperature (of the order of 1000 K). The second term [with $P_2^0 = 0.5(3 \sin^2(\varphi) - 1)$] represents the heat surplus at lower latitudes and a corresponding heat deficit at higher latitudes (Fig. 2a). A thermal wind system develops with wind toward the poles in the upper level and wind away from the poles in the lower level. The coefficient $\Delta T_2^0 \approx 0.004$ is small because Joule heating in the aurora regions compensates that heat surplus even during quiet magnetospheric conditions. During disturbed conditions, however, that term becomes dominant, changing sign so that now heat

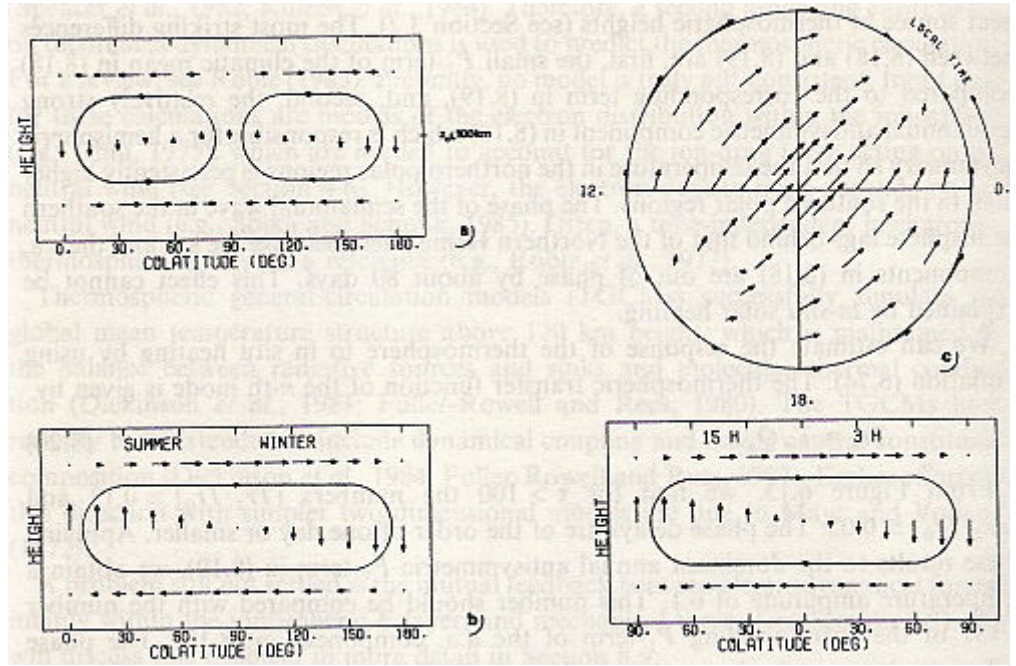


Figure 2. Schematic meridian-height cross-section of circulation of (a) symmetric wind component (P_2^0), (b) of antisymmetric wind component (P_1^0), and (d) of symmetric diurnal wind component (P_1^1) at 3 h and 15 h local time. Upper right pannel (c) shows the horizontal wind vectors of the diurnal component in the northern hemisphere depending on local time.

surplus is transported from the poles to the equator. The third term (with $P_1^0 = \sin \varphi$) represents heat surplus on the summer hemisphere and is responsible for the transport of excess heat from the summer into the winter hemisphere (Fig. 2b). Its relative amplitude is of the order $\Delta T_1^0 \approx 0.13$. The fourth term (with $P_1^1(\varphi) = \cos \varphi$) is the dominant diurnal wave (the tidal mode (1,−2)). It is responsible for the transport of excess heat from the daytime hemisphere into the nighttime hemisphere (Fig. 2d). Its relative amplitude is $\Delta T_1^1 \approx 0.15$, thus on the order of 150 K. Additional terms (e.g., semiannual, semidiurnal terms and higher order terms) must be added to eq.(3). However, they are of minor importance. Corresponding sums can be developed for density, pressure, and the various gas constituents.^{[5][12]}

Thermospheric storms

In contrast to solar XUV radiation, magnetospheric disturbances, indicated on the ground by geomagnetic variations, show an unpredictable impulsive character; from short periodic disturbances of the order of hours to long-standing giant storms of several days' duration. The reaction of the thermosphere to a large magnetospheric storm is called thermospheric storm. Since the heat input into the thermosphere occurs at high latitudes (mainly into the auroral regions), the heat transport represented by the term P_2^0 in eq. (3) is reversed. In addition, due to the impulsive form of the disturbance, higher-order terms are generated which, however, possess short decay times and thus quickly disappear. The sum of these modes determines the "travel time" of the disturbance to the lower latitudes, and thus the response time of the thermosphere with respect to the magnetospheric disturbance. Important for the development of an ionospheric storm is the increase of the ratio N_2/O during a thermospheric storm at middle and higher latitude.^[13] An increase of N_2 increases the loss process of the ionospheric plasma and causes therefore a decrease of the electron density within the ionospheric F-layer (negative ionospheric storm).

See also

-  [Atmospheric sciences portal](#)

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