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The Neutral Atmosphere

I. Mueller-Wodarg
(University College, London)

strada costiera, 11 - 34014 trieste italy - tel. +39 0402240111 fax +39 040224163 - sci_info@ictp.trieste.it - www.ictp.trieste.it

THE NEUTRAL ATMOSPHERE

Ingo Mueller-Wodarg

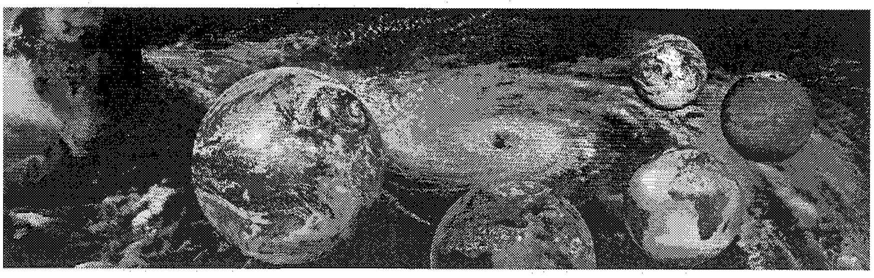
Atmospheric Physics Laboratory
University College London
67-73 Riding House Street
London W1P 7PP, U.K.
i.mueller-wodarg@ucl.ac.uk

THE BASICS:

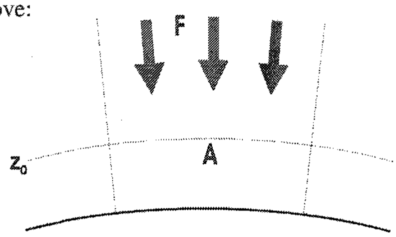
PRESSURE, SCALE HEIGHT

What is an atmosphere?

An atmosphere is a gas layer surrounding a planet. Our solar system has planets with permanent (non-transient) atmospheres (Earth, Mars, Venus, Giant planets, Titan) as well as bodies with noon-permanent (transient) atmospheres (Moon, Mercury, Jovian moons, ..).



The **pressure** is defined as the weight (per surface area) of the atmosphere above:



$$p(z_0) = \frac{F}{A} = \int_{z=z_0}^{\infty} \rho(z) \cdot g(z)\, dz$$

and the change or pressure with height is given by:

$$dp = -\rho(z) \cdot g(z)\, dz$$

Pressure decreases with altitude since the weight of the atmosphere above becomes smaller for increasing altitude.

From the ideal gas law:

$$p = n\,k\,T$$

and:

$$\frac{dp}{dz} = -\rho \cdot g = -\overline{m} \cdot n \cdot g$$

$\overline{m}$ is the mean molecular mass (in units of mass)

we get:

$$\frac{dp}{dz} = -\frac{\overline{m} \cdot g}{kT} \cdot p = -\frac{1}{H} \cdot p$$

where:

$$\boxed{H = \frac{kT}{\overline{m}g}}$$

is the **scale height**

giving:

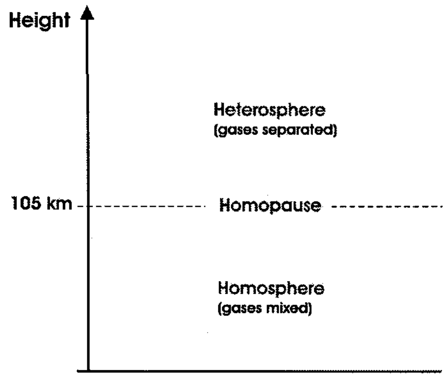
$$\boxed{p(z) = p(z_0) \cdot e^{-\frac{z-z_0}{H(z)}}}$$

So, when moving up in altitude by one scale height, pressure decreases by a factor of 1/e ~ 0.37.

The same applies to number densities:

$$n = \frac{p}{kT} \quad \text{, so:} \quad n = n_0 \cdot \left(\frac{T_0}{T}\right) \cdot e^{-\int_{z_0}^{z} \frac{dz}{H}}$$

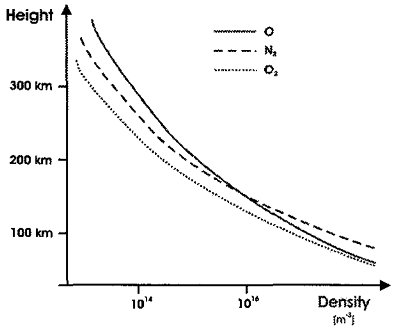
In practise, the atmosphere consists of many gases which below the homopause (near 105 km on Earth) are well mixed due to small scale turbulence and larger scale winds:



In the heterosphere, molecular diffusion is more effective than turbulent mixing, so gases behave more independently. They distribute vertically according to their **individual scale heights**,

$$H_i = \frac{kT}{m_i g}$$

So, the lighter a gas, the larger its scale height and the slower its densities fall with altitude. Therefore, lighter gases become relatively more abundant at higher altitudes.



For T = 800 K:

$H_{O2} \sim 24$ km

$H_O \sim 48$ km

$H_{N2} \sim 27$ km

Thermosphere Composition

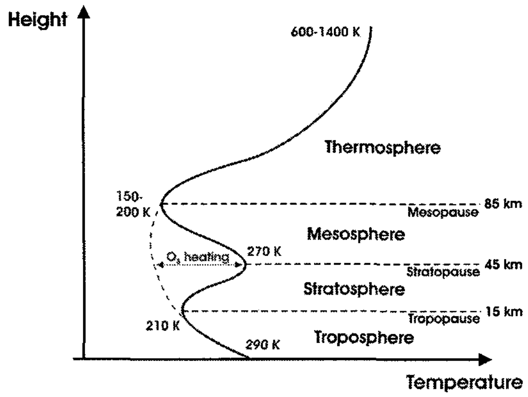
Height	O [m^{-3}]	O_2 [m^{-3}]	N_2 [m^{-3}]
100 km	$4.55 \cdot 10^{17}$ (3.5 %)	$2.38 \cdot 10^{18}$ (18.3 %)	$1.02 \cdot 10^{19}$ (78.2 %)
200 km	$3.71 \cdot 10^{15}$ (57.6 %)	$1.64 \cdot 10^{14}$ (2.5 %)	$2.57 \cdot 10^{15}$ (39.9 %)
300 km	$4.39 \cdot 10^{14}$ (86.8 %)	$2.44 \cdot 10^{12}$ (0.5 %)	$6.42 \cdot 10^{13}$ (12.7 %)

(MSISE 90 Model, March, lat=0, lon=0, 15:00 LT, F10.7=100, Ap=6, 1987)

Note the change in relative abundances with height, N_2 being dominant at 100 km (78 %), O being dominant near and above 200 km (>58 %), and O_2, being the heaviest of the 3 molecules, falling off strongest, being a minor constituent only above 200 km.

Thermal structure

The Earth's atmosphere is thermally subdivided vertically into different regions:

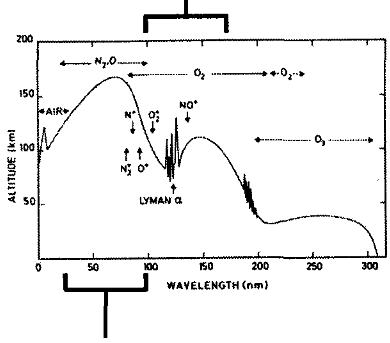


This structure is a result of different gases absorbing different wavelengths at different altitudes:

Spectral regions of photochemical importance in the atmosphere

Wavelength	Atmospheric absorbers
121.6 nm	Solar Lyman α line, absorbed by O_2 in the mesosphere; no absorption by O_3
100 to 175 nm	O_2 Schumann Runge continuum. Absorption by O_2 in the thermosphere. Can be neglected in the mesosphere and stratosphere.
175 to 200 nm	O_2 Schumann Runge bands. Absorption by O_2 in the mesosphere and upper stratosphere. Effect of O_3 can be neglected in the mesosphere, but is important in the stratosphere.
200 to 242 nm	O_2 Herzberg continuum. Absorption by O_2 in the stratosphere and weak absorption in the mesosphere. Absorption by the O_3 Hartley band is also important; both must be considered.
242 to 310 nm	O_3 Hartley band. Absorption by O_3 in the stratosphere leading to the formation of $O(^1D)$.
310 to 400 nm	O_3 Huggins bands. Absorption by O_3 in the stratosphere and troposphere leads to the formation of $O(^3P)$.
400 to 850 nm	O_3 Chappuis bands. Absorption by O_3 in the troposphere induces photodissociation even at the surface.

Dissociation of O_2 in Thermosphere



Ionization of O, O_2, N_2

Troposphere:
- Energy sources:
  - Planetary surface absorption (IR, visible), conduction to atmosphere
  - Atmospheric absorption or terrestrial and solar IR
  - Latent heat release by H_2O
- Energy sinks:
  - IR radiation
  - Evaporation of H_2O

Stratosphere:
- Energy sources:
  - Strong absorption of UV by ozone (causing stratopause temperature peak)
- Energy sinks:
  - IR radiation by O_3, CO_2, H_2O

Mesosphere:
- Energy sources:
  - Some UV absorption by O_3 (lower heights)
  - Heat transport down from thermosphere (minor, top heights only)
- Energy sinks:
  - IR radiation by CO_2 H_2O, OH

Thermosphere:

- Energy sources:

  - Absorption of EUV (200-1000Å; photoionizing O, O_2, N_2) and UV (1200-2000 Å), photodissociating O_2), leading to chemical reactions and particle collisions, liberating energy
  - Dissipation of upward propagating waves (tides, planetary waves, gravity waves)
  - Joule heating by auroral electrical currents
  - Particle precipitation from the magnetosphere
  - Dynamics: advection, adiabatic heating

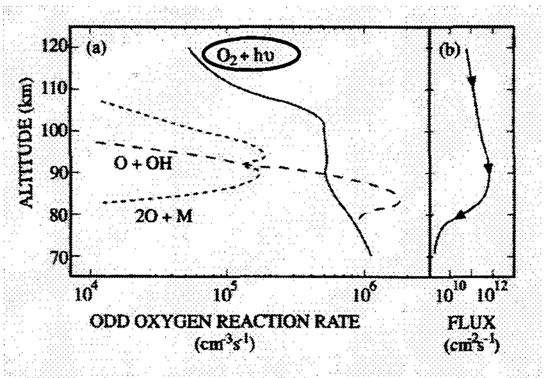
- Energy sinks:

  - Thermal conduction into the mesosphere, where energy is radiated by CO_2, O_3 and H_2O
  - IR cooling by NO and CO_2 (after geomagnetic storms)
  - Dynamics: advection, adiabatic cooling

Photochemistry in the Thermosphere

a) Photodissociation:

$$O_2 + h\nu \quad \rightarrow \quad O(^3D) + O(^1D)$$

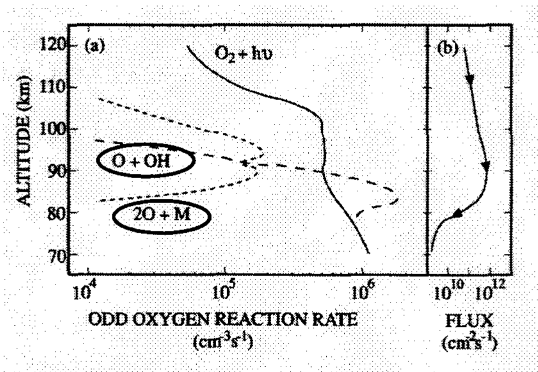


b) Recombination:

$$O+O+M \rightarrow O_2+M \qquad \left[\approx 3\times10^{-34}\exp\left(\frac{710}{T}\right) cm^6/\sec\right]$$

Where M is a molecule; O_2 or N_2. Three-body recombination is necessary to get rid of excess energy of metastable O_2^*. Note the temperature dependence of the reaction rate.

$$O+OH \rightarrow O_2+H$$



FUNDAMENTAL EQUATIONS

Energy equation

Advection and adiabatic terms **EUV heating, IR cooling**

$$\frac{\partial \varepsilon}{\partial t} + \vec{V}_p \vec{\nabla}_p(\varepsilon + gh_p) + w\frac{\partial(\varepsilon + gh_p)}{\partial p} = Q_{EUV} + Q_{IR}$$

Molecular & turbulent heat conduction

$$+ g\frac{\partial}{\partial p}\left(\frac{K_M + K_T}{H}p\frac{\partial T}{\partial p}\right) + \frac{1}{\rho}(K_M + K_T)\nabla_p^2 T$$

Viscous heating

$$+ g\frac{\partial}{\partial p}\left(u_\theta \mu \frac{\partial u_\theta}{\partial p} + u_\varphi \mu \frac{\partial u_\varphi}{\partial p}\right)$$

where $\varepsilon = c_p T + \frac{1}{2}\left(u_\theta^2 + u_\varphi^2\right)$ is the sum of internal and kinetic energies per unit mass.

Continuity equation

The concentration of gas particles is determined by the balance between sources and sinks. Gas concentrations are affected by **chemical** sources and sinks, **transport** (**advection**) by winds and **diffusion** of gas particles (their microscopic motion).

$$\frac{\partial Y_i}{\partial t} + u\frac{\partial Y_i}{\partial x} + v\frac{\partial Y_i}{\partial y} + w\frac{\partial Y_i}{\partial z} + g\frac{\partial}{\partial p}\left(\rho Y_i(w_i^D + w_i^K)\right) + J_i$$

Advection **Molecular & Eddy Diffusion** **Net chemical source rate**

where the molecular diffusion velocities w_i^D satisfy the equation:

$$\frac{\partial Y_i}{\partial z} - \left(1 - \frac{m_i}{m} - \frac{H}{m}\frac{\partial m}{\partial z}\right)\frac{Y_i}{H} = -\sum_{j \neq i}\frac{mY_iY_j}{m_j D_{ij}}\left(w_i^D - w_j^D\right)$$

and Eddy diffusion velocities w_i^K are given by $w_i^K = -K\frac{\partial \ln(Y_i)}{\partial z}$

K is the **Eddy diffusion coefficient** and D_{ij} is the **molecular diffusion coefficient** between gases i and j.

In this, the concentration of gas constituents may be expressed in 3 ways:

i) **number density** (particles per volume), $\boldsymbol{n_i}$

ii) **mole fraction** (number mixing ratio),

$$X_i = \frac{n_i}{n_{tot}} \quad \text{where} \quad n_{tot} = \sum_i n_i$$

iii) **mass fraction** (mass mixing ratio),

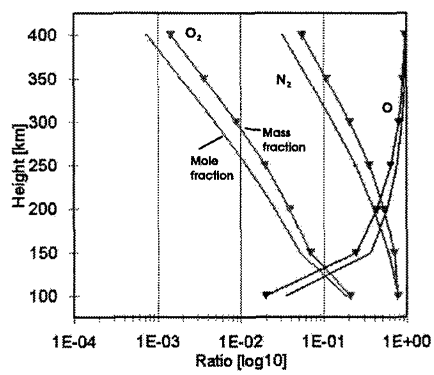
$$Y_i = \frac{n_i m_i}{n_{tot} m_{tot}} \quad \text{where} \quad m_{tot} = \sum_i \frac{n_i m_i}{n_{tot}}$$

Note that:

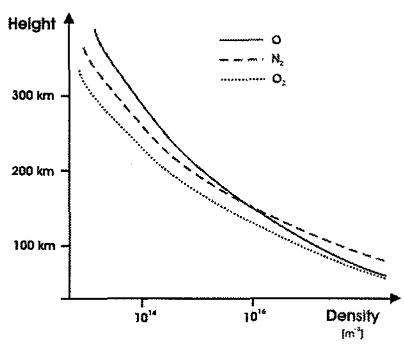
$$\sum_i X_i = 1 \quad \text{and} \quad \sum_i \Psi_i = 1$$

Therefore, the continuity equation can be expressed in 3 ways.

Mass and mole fractions



Number densities



Time scales

In order to understand changes in composition and estimate which processes are dominant, we introduce the concept of **time scale**, which quantifies how long changes in composition take for each type of process.

i) Dynamical time scale:

Estimates time scale of gas transport by winds:

$$\boxed{\tau_{dyn} \approx \frac{dist}{V}}$$

where ***dist*** is a typical distance and ***V*** the velocity.

Horizontal: ***dist*** = a few degrees lat/lon (depending on problem)
Vertical: ***dist*** = one scale height

ii) Chemical time scale:

a) Photochemical production/loss:

$$N_i + h\nu \xrightarrow{J} N_i$$

J is rate coefficient for photoproduction or loss (in [1/sec])

$$\frac{\partial N_i}{\partial t} = -J \cdot N_i \quad \Rightarrow \quad N_i = N_{i0} \cdot e^{-Jt}$$

So, the **photochemical lifetime** is defined as:

$$\boxed{\tau_p = \frac{1}{J}}$$

b) Chemical production/loss:

$$N_i + N_j \xrightarrow{K} \ldots$$

K is rate coefficient for chemical production or loss (in [cm^3/sec])

$$\frac{\partial N_i}{\partial t} = -K \cdot N_i \cdot N_j \quad \Rightarrow \quad N_i = N_{i0} \cdot e^{-KN_j t}$$

So, the **chemical lifetime** is defined as:

$$\boxed{\tau_C = \frac{1}{KN_j}}$$

c) Molecular diffusion time constant:

$$\boxed{\tau_D = \frac{H^2}{D_{i,j}}}$$

$D_{i,j}$ is the molecular diffusion coefficient between species i and j (in [m^2/sec]); H is the average scale height.

τ_D decreases roughly exponentially with height.

d) Turbulent mixing time constant:

$$\boxed{\tau_K = \frac{H^2}{K_D}}$$

K_D is the turbulent (Eddy) diffusion coefficient (in [m^2/sec])

τ_K is roughly invariant with height.

Time scales can be used very effectively to work out the importance of various processes without the need for sophisticated calculations.

For example...

If	**Then ....**
$\tau_D \ll \tau_{dyn}$	Molecular diffusion is more effective than winds in changing composition $\Rightarrow$ diffusive balance holds, winds don't matter
$\tau_K \ll \tau_C$	Turbulent mixing is more effective than chemical changes, so the gas distribution is strongly affected by turbulence
$\tau_C \approx \tau_{dyn}$	Chemical changes and winds are equally important in changing the composition.

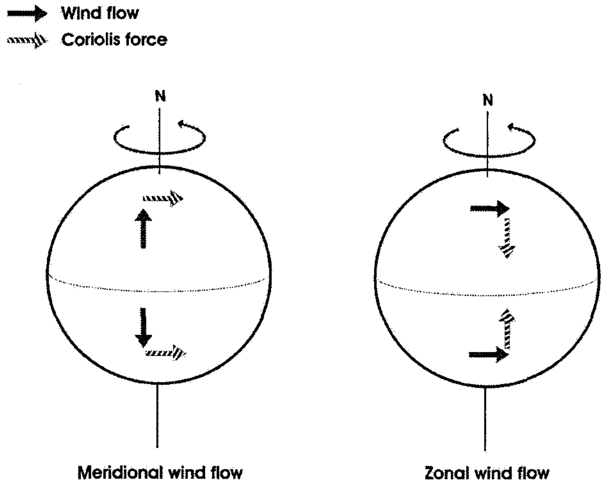
Momentum equation

$$\frac{d\vec{U}}{dt} = \overbrace{\vec{g}}^{\text{Gravity}} - \underbrace{\frac{1}{\rho}\vec{\nabla}p}_{\text{Pressure gradient}} - \overbrace{2\vec{\Omega}\times\vec{U}}^{\text{Coriolis}} + \underbrace{\frac{1}{\rho}\vec{\nabla}\left(\mu\vec{\nabla}\vec{U}\right)}_{\text{Viscosity}} - \overbrace{\nu_{ni}\left(\vec{U}-\vec{V}\right)}^{\text{Ion drag term}}$$

Pressure gradients	are driven by temperature differences
Advection	is transport of momentum by winds
Coriolis force	is caused by the Earth's rotation
Viscosity	is due to gas particle collisions
Ion drag	is transfer of momentum from ions to neutrals

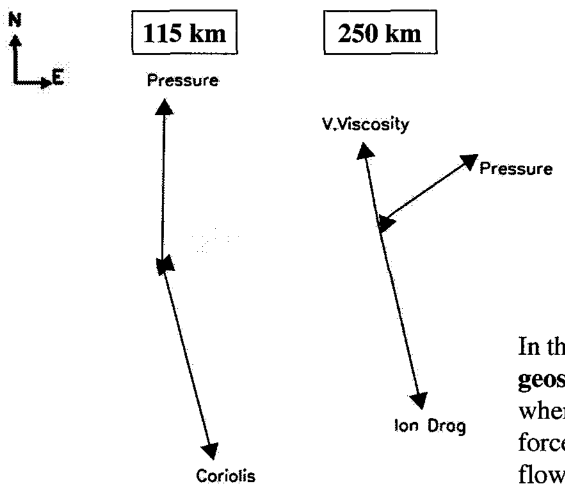
Pressure gradients and **Ion drag** are **external forces**, the rest are **internal**

Coriolis force acts **perpendicular** to the **wind vector**. It deflects poleward winds towards the east and eastward winds equatorward. So, winds are driven clockwise (anticlockwise) in the northern (southern) hemisphere around pressure minima.



Momentum balance

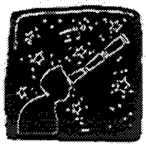
March, 30N, 54E, 15:40 LT



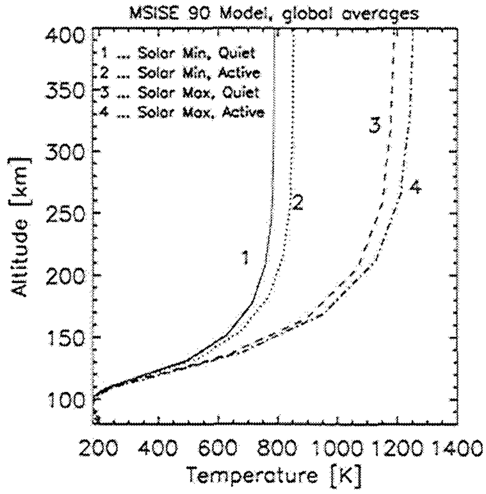
Note the differences in momentum balance at different altitudes!

In the lower thermosphere, **geostrophic balance** applies, where Pressure and Coriolis forces almost balance. Winds flow roughly **perpendicular** to isobars.

THERMAL STRUCTURE, COMPOSITION
AND DYNAMICS
OF THE THERMOSPHERE

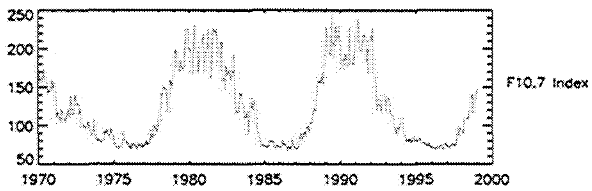


The Solar Cycle behaviour
of Thermospheric Temperature
and Composition

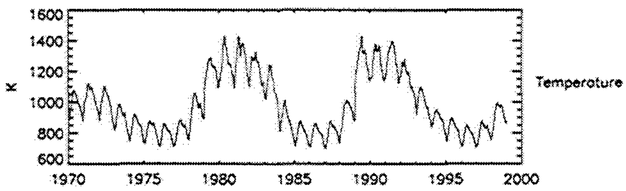


Solar flux intensity is quantified by the **F10.7 index**, which gives the flux of solar radiation at 10.7 cm wavelength. Although this wavelength is of no importance to the upper atmosphere, its flux correlates reasonably well with UV and EUV fluxes.

The solar flux intensity varies with an 11 year cycle



Exospheric temperatures vary strongly with solar activity as well as season.

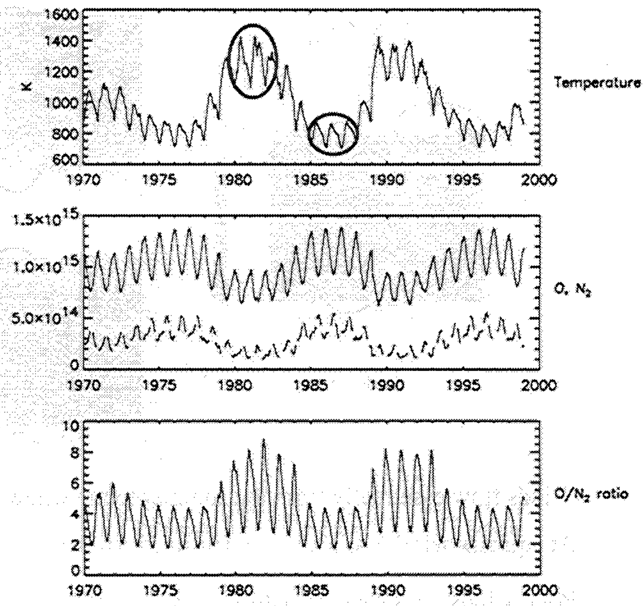


300 km, 50°N, local noon, F10.7=100

Seasonal temperature changes are stronger during solar max than solar minimum.

Number densities of gases vary with solar cycle and season

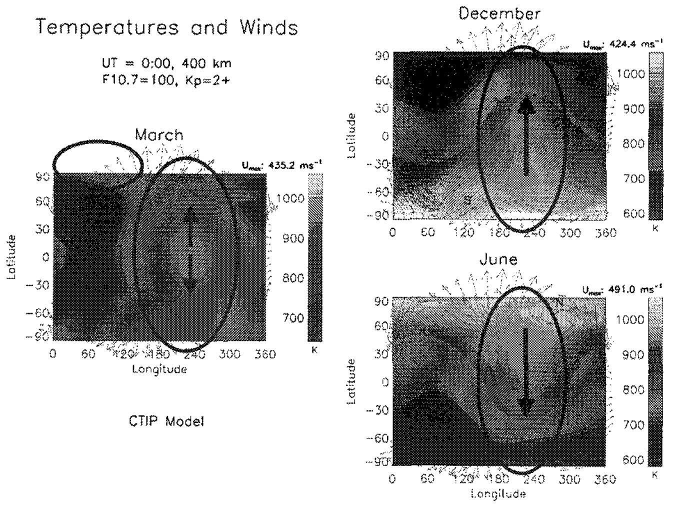
During active solar conditions there is proportionally more O than N_2 in the thermosphere, due to stronger photodissociation.



The Seasonal behaviour

of Thermospheric Temperature, Winds

and Composition



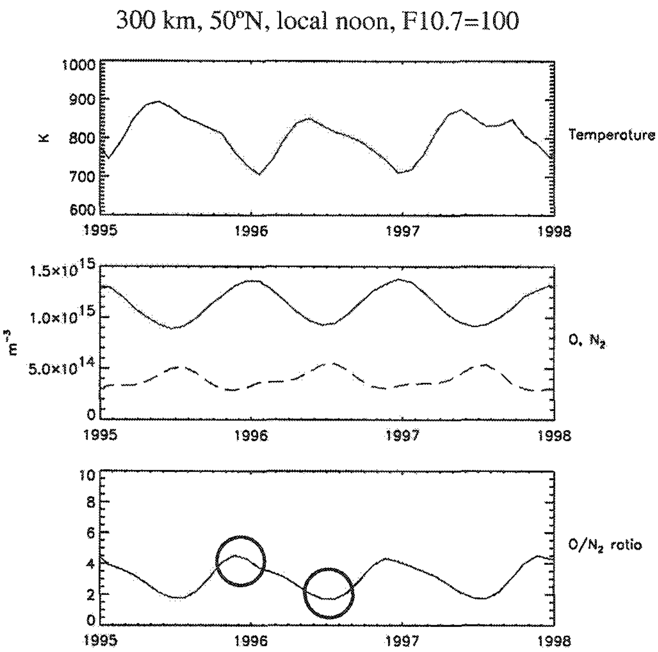
Winds flow essentially from the summer to the winter hemisphere.

At equinox winds are quasi-symmetric, from the equator towards the poles.

Polar winds are strongly controlled by ion drag

Exospheric noon-time temperatures peak near solstice

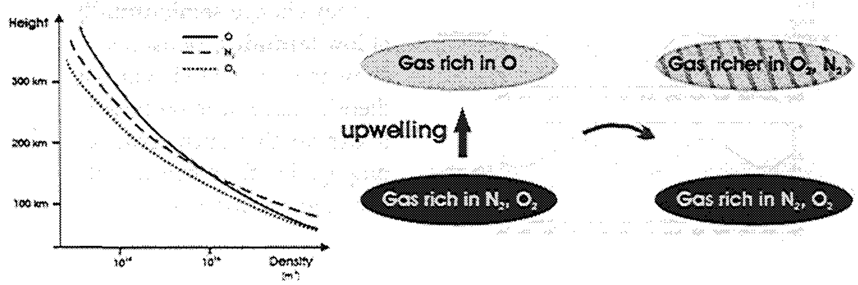
In spite of increased dissociation at summer solstice, the abundance of O in proportion to N_2 is at its lowest then. This is due to vertical transport of gases: summer upwelling transports N_2 rich gases from lower to higher altitudes .



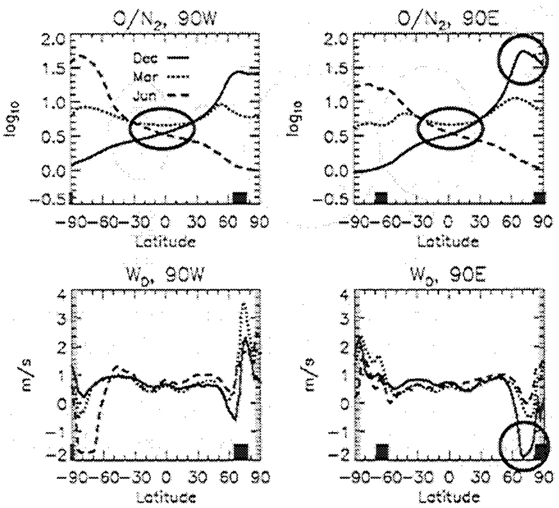
The total vertical wind may be expressed in terms of 2 components, the **barometric** wind and the **divergence** wind. The former is due to thermal expansion/contraction of the atmosphere, the latter is caused by diverging horizontal winds and the conservation of mass:

$$U_z = W_B + W_D = \left(\frac{\partial h}{\partial t}\right)_p + \frac{1}{n}\int_z^{\infty}\left(\frac{\partial(nU_x)}{\partial x}+\frac{\partial(nU_y)}{\partial y}\right)\cdot H\,dz$$

Upward vertical *divergence* winds transport gases from lower to higher altitudes. Gases at lower heights are richer in molecular constituents, so the upward winds cause gases higher up to be relatively more molecular.
So, **upward** winds cause a **decrease** in the O/N_2 ratio.

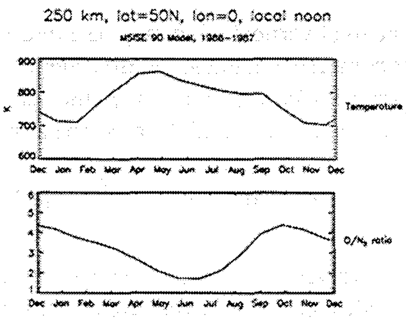


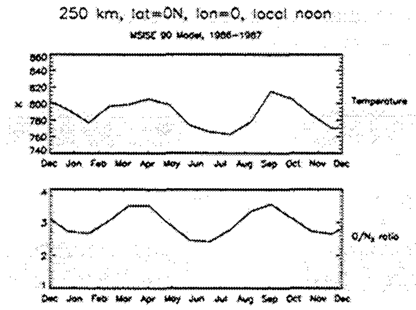
The principle is illustrated in these plots. Note the changes in composition near the auroral ovals (black boxes), where strong vertical winds are found.



Note the different behaviour at low latitudes: the March values are larger than both December and June ones.

While temperatures and composition show an **annual** change at **mid-latitudes**...

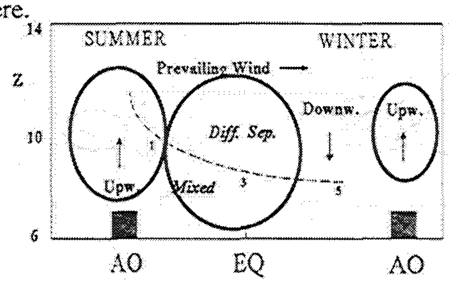




... they change **semiannually** at **low latitudes.** Densities show the same trend. Various theories have been proposed to explain this, such as wave propagation from below and geomagnetic forcing.

Inter-hemispheric winds and composition changes

This figure illustrates the interplay between winds and composition in the thermosphere.

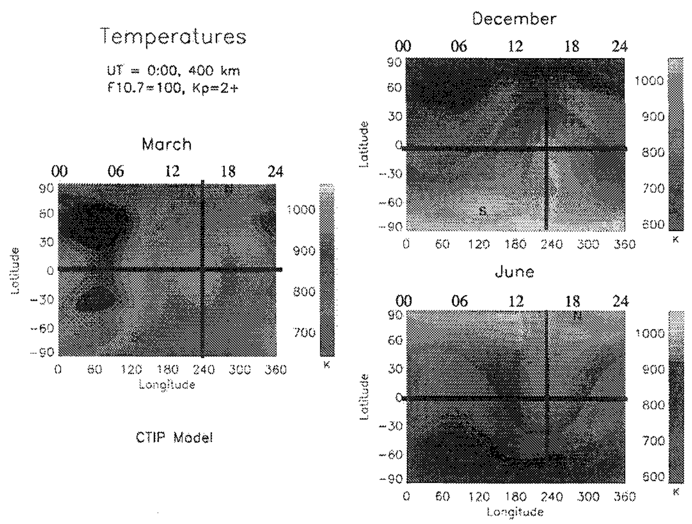


Upwelling occurs in the summer hemisphere and over the auroral ovals ("AO"). The summer upwelling upsets diffusive equilibrium,

Molecular-rich gases are transported by horizontal winds towards the winter hemisphere, where diffusive balance is progressively restored, from top (where diffusion is faster) to bottom.

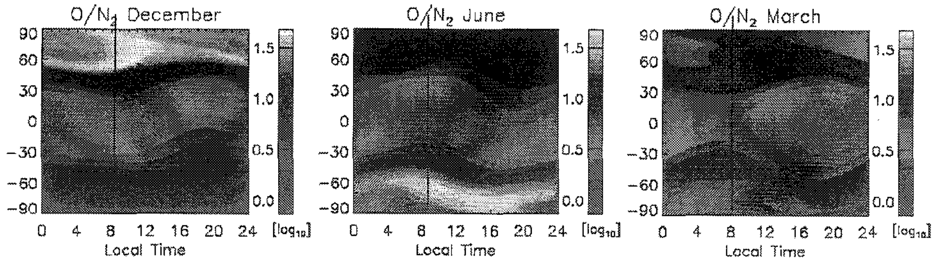
An additional circulation cell exists in the winter hemisphere through upwelling over the aurora, but doesn't affect much the equatorial regions.

The Diurnal behaviour of Thermospheric Temperature, Winds and Composition

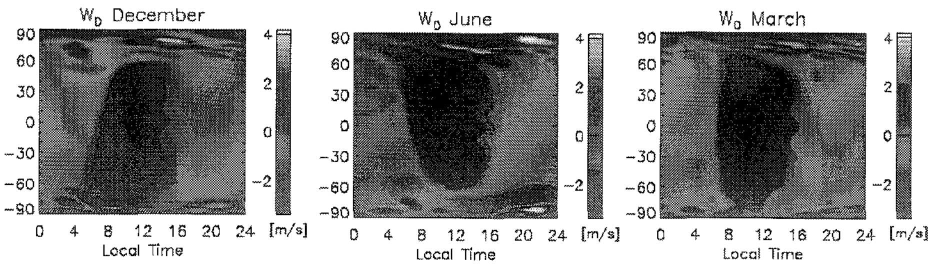


Exospheric temperatures peak near **15:30 h** local time.

Day-night temperature differences at low latitudes reach around **200 K**.

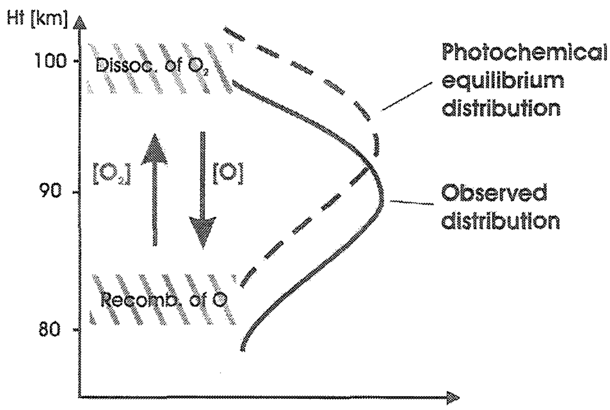


The O/N_2 ratio peaks near noon and is smallest near 02:00 local time.



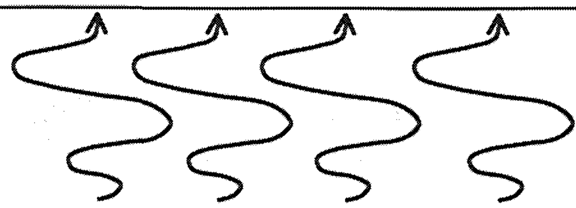
Diurnally, composition is almost anti-correlated with vertical winds and thus determined to large part by photodissociation. Note that vertical winds however determine the *seasonal* behaviour of composition!

The impact of winds on the vertical distribution of [O]



At 80-100 km $\tau_K \ll \tau_{recombination}$ $\Rightarrow$ eddy mixing important

Mixing transports O down to lower heights, where recombination is more rapid



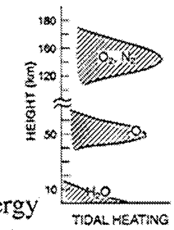
COUPLING FROM BELOW:

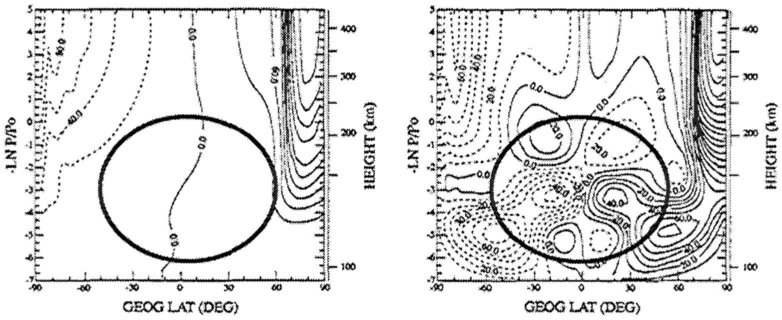
TIDES AND PLANETARY WAVES

Winds in the lower thermosphere (80-200 km) are strongly affected by global waves propagating upward from the lower atmosphere. These are:

a) Tides:

- Periods: 24*h*, 12*h*, 8*h*
- Driven by solar heating (*thermal tides*) and, to less extent, by the Moon's gravitation field (*lunar tides*)
- Follow the Sun, ie. westward propagating
- Distinct latitudinal structure described, for an idealized atmosphere, by *Hough modes.*
- Peak amplitudes in lower thermosphere ~50 m/s and ~10 K
- Dissipate in 100-160 km height regime, depositing large amounts of momentum (westward zonal acceleration) and energy (temperature increase) into the background atmosphere.
- Show diurnal and seasonal variability





Zonally averaged meridional winds at 70°W and 18:00 UT for quiet-time conditions with (left) and without (right) tidal oscillations. Contours are positive southward.

Note how the tides dominate the low- to mid latitude thermosphere!

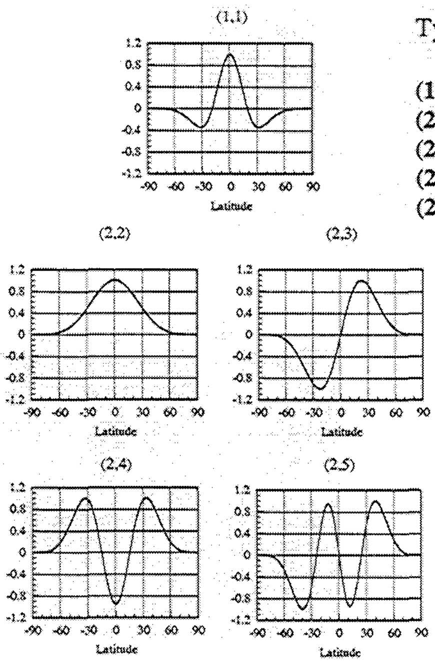
Classical Tidal Theory

Classical tidal theory analytically describes the global structure of tidal oscillations. The analytical solutions strictly apply only to an idealized atmosphere, ignoring non-linear processes, but are an effective tool for describing also the real atmosphere. **Above around 100 km, predictions from Classical Tidal theory become increasingly unrealistic.**

Global oscillations can be decomposed into **Hough modes** (similar to Fourier decomposition), which allow a physical interpretation of the tidal structure. The modes are classified as

- **migrating**: the tides propagate vertically and horizontally
- **non-migrating**: they remain constrained to where they are generated

Hough modes are named as (n, m), where n is their longitudinal wavenumber and m the latitudinal wavenumber.

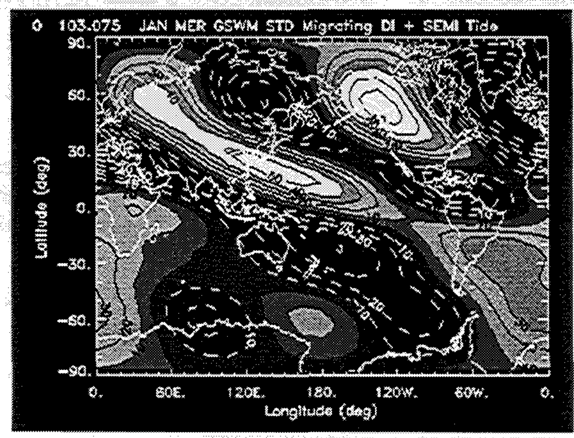


Typical migrating tidal modes are:

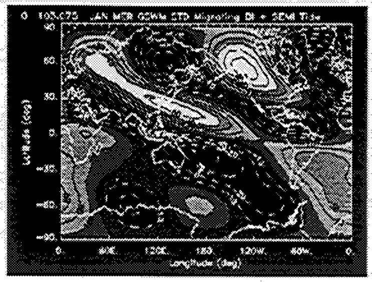
(1,1): diurnal, symmetric
(2,2): semidiurnal, 1st symmetric
(2,3): semidiurnal, 1st asymmetric
(2,4): semidiurnal, 2nd symmetric
(2,3): semidiurnal, 2nd asymmetric

These figures show common Hough modes. The lines show the amplitudes of geopotential height of a surface of constant pressure. Classical tidal theory gives analytical equations for deriving wind and temperature amplitudes from those of geopotential height.

This figure shows a global profile of meridional winds versus latitude and local time (longitude) near 100 km altitude for January conditions, as simulated by the GSWM model



This global profile can be decomposed into diurnal and semidiurnal migrating tides:



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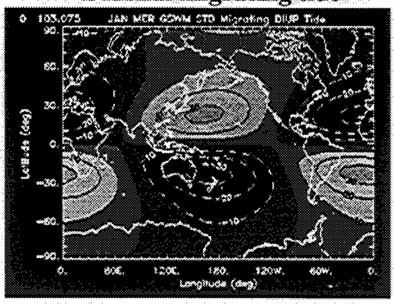
Diurnal migrating tide



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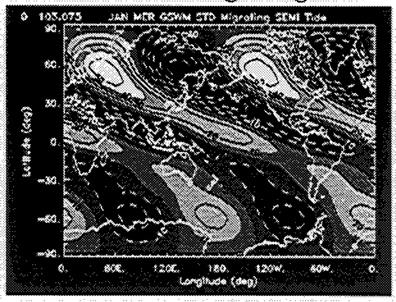
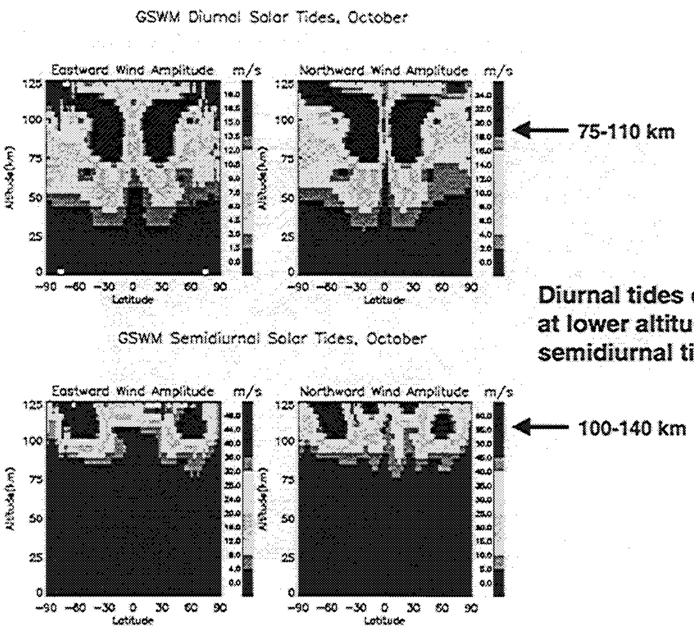
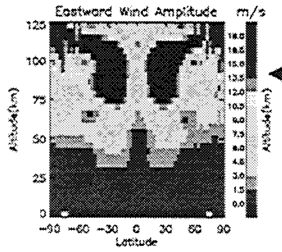
Semidiurnal migrating tide



The change of tidal amplitudes with height

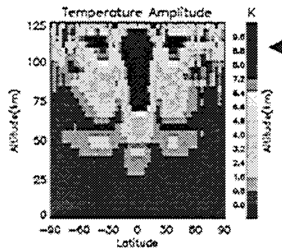


Diurnal tides dissipate at lower altitudes than semidiurnal tides.



75-110 km

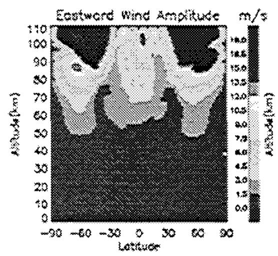
Semidiurnal wind amplitudes peak at lower altitudes than Temperature amplitudes.



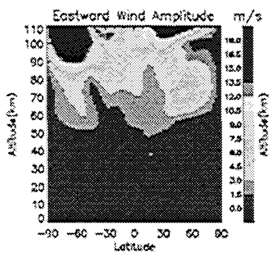
75-125 km

Tidal amplitudes show variations from day to day and with season:

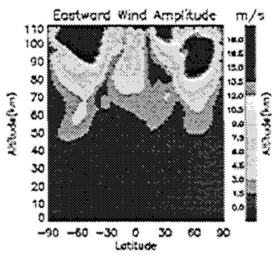
March



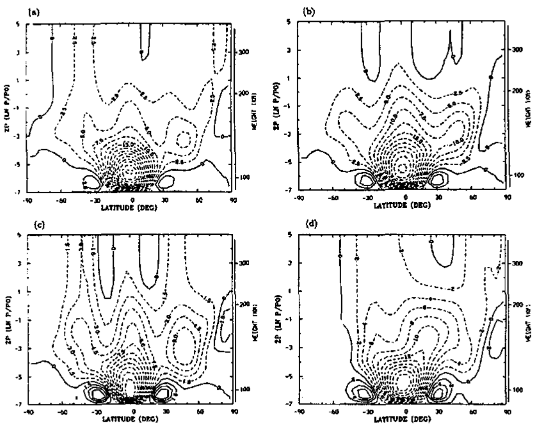
June



September

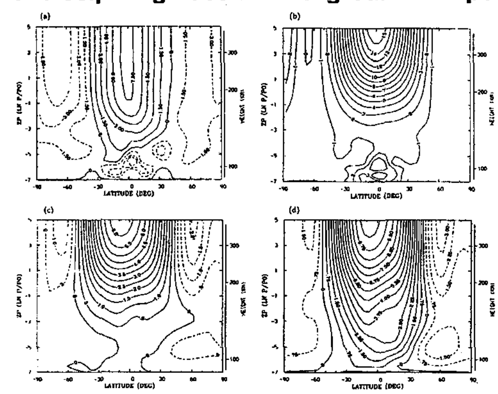


Effects of dissipating tides on background winds



When tides dissipate, they release momentum and energy into the background atmosphere, accelerating the mean winds. The plots illustrate that the dissipating diurnal (1,1) mode generates a westward acceleration near 100 km altitude. The response is sensitive to season and the phase of the upward propagating tide. Plots (a)-(c) are for March conditions and have tidal phases 12 h , 0 h and 18 h LT, plot (d) is for December conditions.

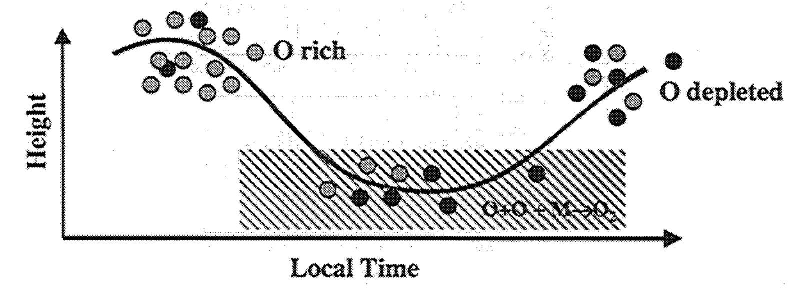
Effects of dissipating tides on background composition



Dissipating tides also affect temperature and composition. Figure (a) shows temperature increases above 120 km (in [K]), (b) shows the change of N_2 due to the diurnal and semidiurnal tides, (c) are the N_2 changes due to the semidiurnal tide only and (d) shows changes in O_2 due to the semidiurnal tide (in [%]). Tides have no significant effect on thermospheric mean temperature. The semidiurnal tide affects mostly upper thermosphere composition, the diurnal one primarily causes changes near 100 km.

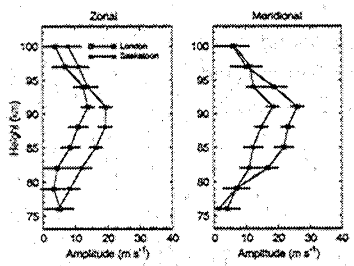
Depletion of O due to the diurnal tide in the lower thermosphere

- The diurnal tide displaces gases, including atomix oxygen (O), vertically during the day
- As the O is transported to lower altitudes, where 3-body recombination times are fast, some of the O is transformed into O_2.
- The net effect of this is a reduction in O.
- The stronger the diurnal tide, the more efficient this process

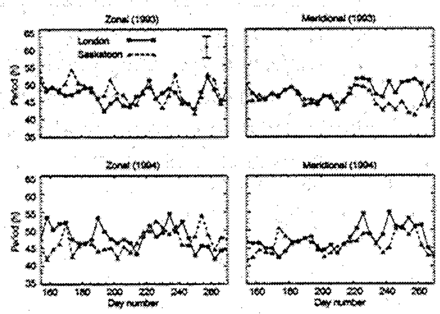


b) Planetary waves:

- Periods: > 1 day, typically 2 days, 5 days, 16 days, …
- Periods are not exact, they vary with time
- Probably Resonant oscillations of the atmosphere; no clear source
- Don't propagate horizontally; phases are locked in longitude
- Strongly sensitive to the background atmosphere, particularly the zonal winds when those are comparable to the wave's phase speed
- Dissipate typically below 100 km altitude in the neutral atmosphere

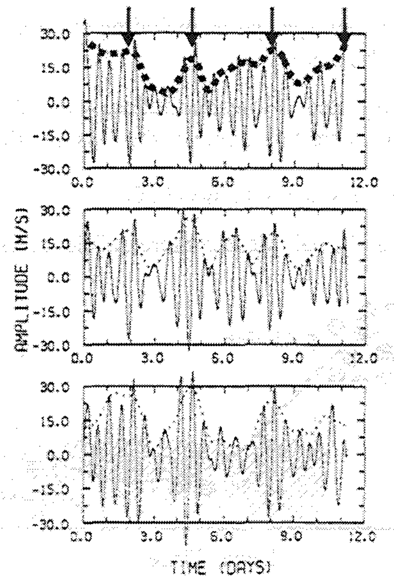


Measured amplitudes of the "Quasi-2-day-wave" (Q2DW)



The variability of Q2DW period

Planetary waves interact non-linearly with tides, generating **modulated tidal amplitudes** which are detectable in the thermosphere.



c) Gravity waves:

Gravity waves are short period (minutes-hours) waves generated locally by topography, localized heating events, wind shears or other small scale processes. Those gravity waves generated in the lower atmosphere propagate upward and, by dissipating or breaking, considerably affect the structure of mesospheric winds. Their effect on the thermosphere is less significant in comparison.

Horizontally propagating gravity waves are found frequently in the thermosphere, in particular during and after geomagnetic events or solar eclipses. They reach velocities of around 300 m/s.

ATMOSPHERIC ESCAPE

Atmospheric Escape

In the upper regions of the atmosphere, gravity is weak enough for gas particles to escape. They gain their necessary kinetic energy from processes like:

- Thermal escape ("Jeans escape")
- Nonthermal processes (mostly involving ions):
    - Charge exchange
    - Dissociative recombination
    - Impact and photo-dissociation
    - Ion-neutral reactions
    - Sputtering
    - Solar wind pickup
    - Ion escape
    - Electric fields

Light gases in the Earth's upper atmosphere, such as H, D, He are affected by these escape mechanisms.

Jeans escape

Atmospheric particles escape if their velocities exceed the *escape velocity*, which is determined by the gravitational field:

$$u_{esc} = \left(\frac{2GM}{r}\right)^{1/2}$$

where G, M and r are the Gravity constant, planet mass and radius. For Earth, u_{esc}=11.2 km/s. Assuming that particle velocities are thermal, the flux of escaping particles is given by:

$$F_{Jeans}(r_c) = \frac{N(r_c)\cdot U}{2\pi^{1/2}} e^{-\lambda}(\lambda+1) \qquad [cm^{-2}\sec^{-1}]$$

r_c ..... Exobase radius
N ..... Number density

where

$$U = \left(\frac{2kT}{M_{particle}}\right)^{1/2} \quad \text{and} \quad \lambda = \frac{u_{esc}^2}{U^2}$$

$M_{particle}$ ...Particle mass
T .... Temperature
k .... Botzmann const.

U is the most probable velocity of a Maxwellian distribution of thermal velocity.

So, Jeans escape increases with temperature. A rise in temperature should thus lead to a considerable decrease of N(H) at the exobase.

This decrease in N(H) with temperature is observed, but smaller than expected from Jeans escape. Thus, H escape may be due to other processes less sensitive to temperature changes, such as charge exchange:

$$H + H^{+*} \rightarrow H^{+} + H^{*}$$

AND FINALLY....

Summary of basic Thermosphere characteristics:

- Solar heating is main energy source
- Main cooling is through molecular conduction
- Main gases: O, O_2, N_2, He (high altitudes only)
- Strong variability of temperature, winds and composition with solar cycle, season, local time
- The seasonal composition changes are controlled primarily by global winds, the diurnal ones by photochemistry
- At low latitudes effects of upward propagating tides, planetary waves and gravity waves are important
- At high latitudes, heating from the magnetosphere occurs in the form of Joule heating and precipitating particles

Further Reading

Key text books:

- Chapman, S. C., and R. S. Lindzen, *Atmospheric Tides*, D. Reidel, Dordrecht, 1970
- Banks, P. M., and G. Kockarts, *Aeronomy*, Academic Press, New York, 1973
- Kato, S., *Dynamics of the Upper Atmosphere*, D. Reidel, Boston, 1980
- Chamberlain, J. W., and D. M. Hunten, *Theory of Planetary Atmospheres*, Academic Press, New York, 1987
- Rees, M. H., *Physics and Chemistry of the Upper Atmosphere*, Cambridge University Press, Cambridge, U.K., 1989
- Schunk, R. W., and A. F. Nagy, *Ionospheres*, Cambridge University Press, Cambridge, U.K., 2000