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***Ionosphere-Thermosphere Basics - I  
Neutral Atmosphere Vertical  
Structure***

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# **Ionosphere-Thermosphere Basics - I**

## **Neutral Atmosphere *Vertical Structure***

*Jeffrey M. Forbes, University of Colorado*

<http://spot.colorado.edu/~forbes/Home.html>

### **Lecture Topics**

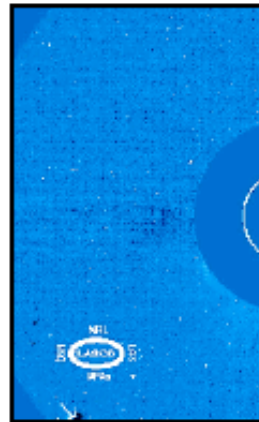
- Thermal Structure and Thermal Balance
  - ➔ Energy Sources and Sinks
  - ➔ Molecular Conduction and Thermal Gradient
- Hydrostatic Balance
  - ➔ Thermosphere Composition
  - ➔ Mean Molecular Weight vs. Height
  - ➔ Homosphere-Heterosphere
  - ➔ Deviations from Hydrostatic Balance: Diffusive and Chemical Time Constants
- Thermosphere Variability & Static Diffusion Empirical Models

**ASEN-**  
The Space Weat

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**COURSE IN**

The various co  
Magnetosphere, Rad  
the interactions bet  
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which aerospace vel  
planetary environme  
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Coronal Mass Ejec  
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# Lectures

Introduction

The Sun

Solar Wind

Geomagnetism

Orbital Debris

Radiation Belts

Magnetospheres

Radiation Effects on Systems

Upper Atmospheres

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orado, Boulder

**NING GOALS**

(Sun, Solar Wind,  
Atmosphere), and  
rbance events, are  
nvironments within  
yes on comparative  
Jupiter and Saturn)

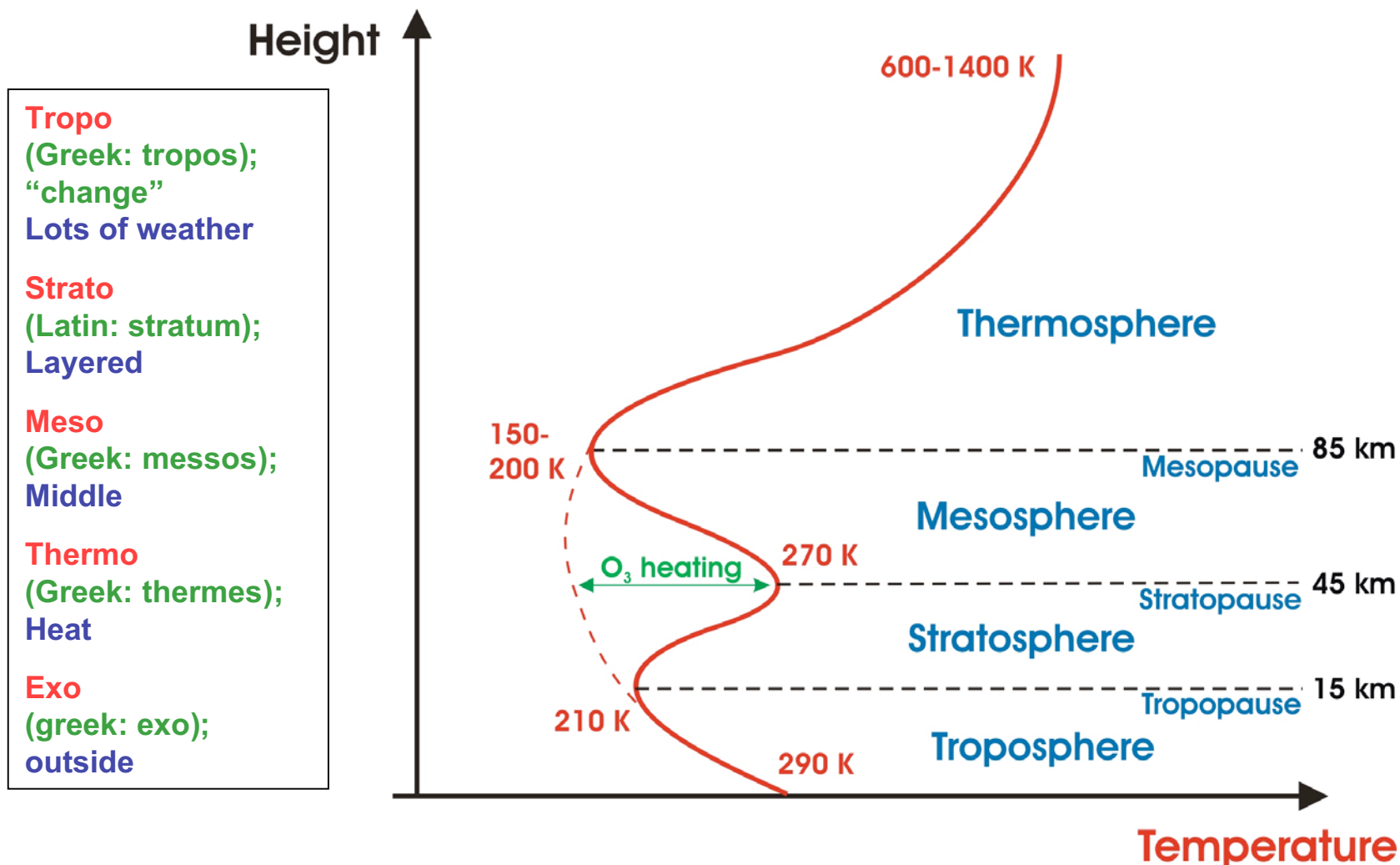


charge (Arcing) is one of  
ts on spacecraft in the

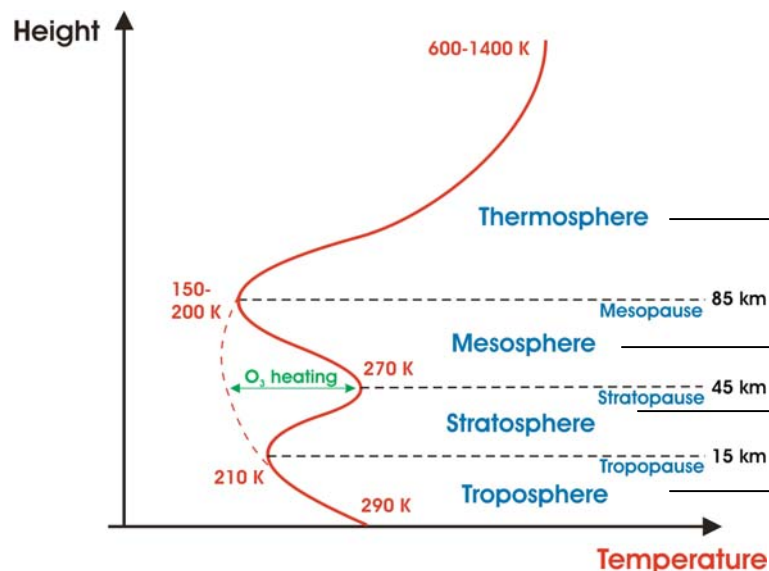
please visit: <http://sisko.Colorado.EDU/FORBES/asen5335/>

# *Thermal Structure & Thermal Balance*

# Global Mean Temperature Structure



# Global Mean Energy Sources and Sinks



## Troposphere:

- **Energy sources:**
  - Planetary surface absorption (IR, visible), convection & conduction to atmosphere
  - Atmospheric absorption of terrestrial and solar IR
  - Latent heat release by H<sub>2</sub>O
- **Energy sinks:**
  - IR radiation
  - Evaporation of H<sub>2</sub>O

## Thermosphere:

- **Energy sources:**
  - Absorption of EUV (200-1000Å; photoionizing O, O<sub>2</sub>, N<sub>2</sub>) and UV (1200-2000 Å), photodissociating O<sub>2</sub>, leading to chemical reactions and particle collisions, liberating energy
  - Joule heating by auroral electrical currents
  - Particle precipitation from the magnetosphere
  - Dissipation of upward propagating waves (tides, planetary waves, gravity waves)
- **Energy sinks:**
  - Thermal conduction into the mesosphere, where energy is radiated by CO<sub>2</sub>, O<sub>3</sub> and H<sub>2</sub>O
  - IR cooling by CO<sub>2</sub>, NO, O

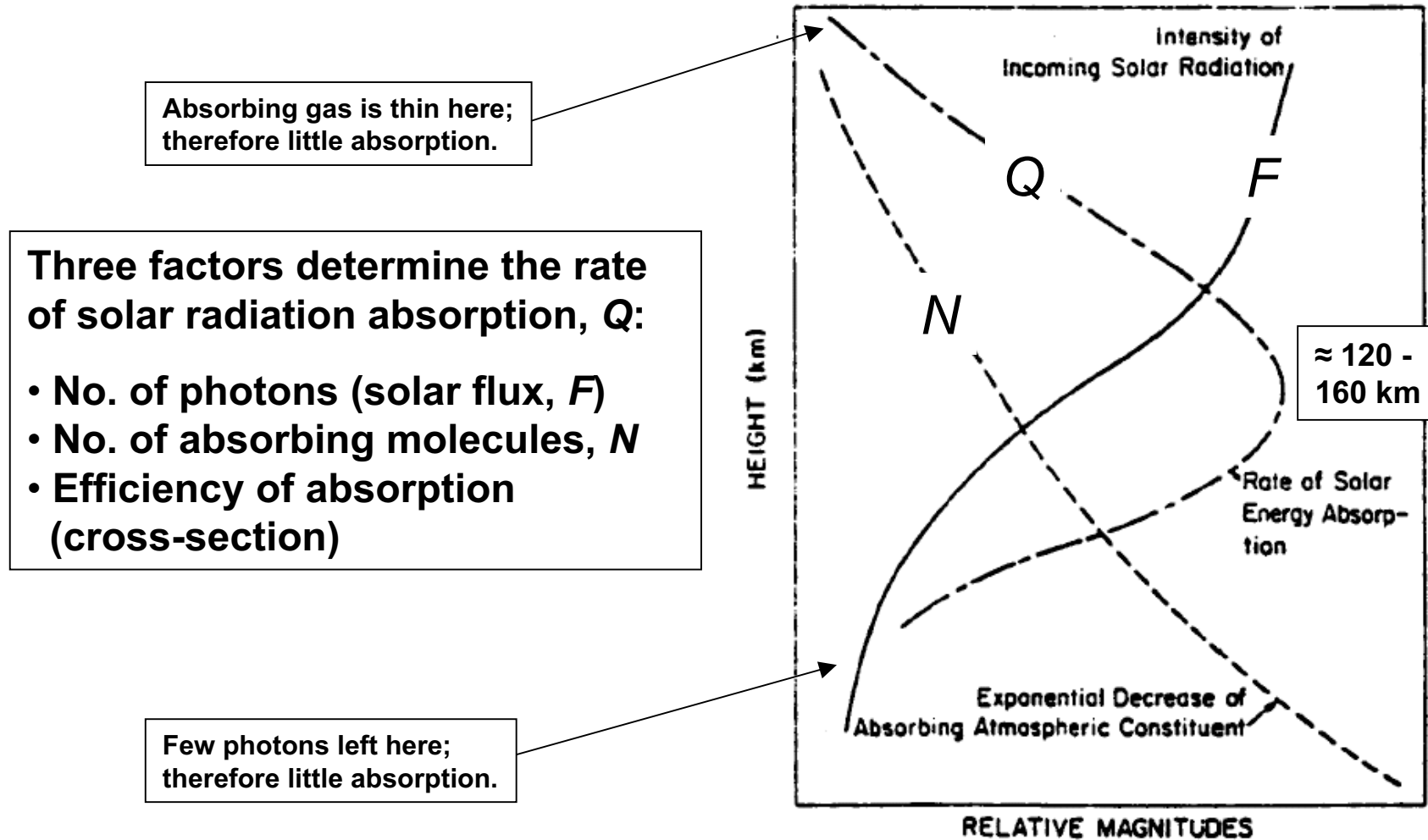
## Mesosphere:

- **Energy sources:**
  - Some UV absorption by O<sub>3</sub> (lower heights)
  - Heat transport down from thermosphere (minor, upper heights only)
  - Chemical heating
- **Energy sinks:**
  - IR radiation by CO<sub>2</sub>, H<sub>2</sub>O, OH

## Stratosphere:

- **Energy sources:**
  - Strong absorption of UV by ozone (causing stratopause temperature peak)
- **Energy sinks:**
  - IR radiation by O<sub>3</sub>, CO<sub>2</sub>, H<sub>2</sub>O

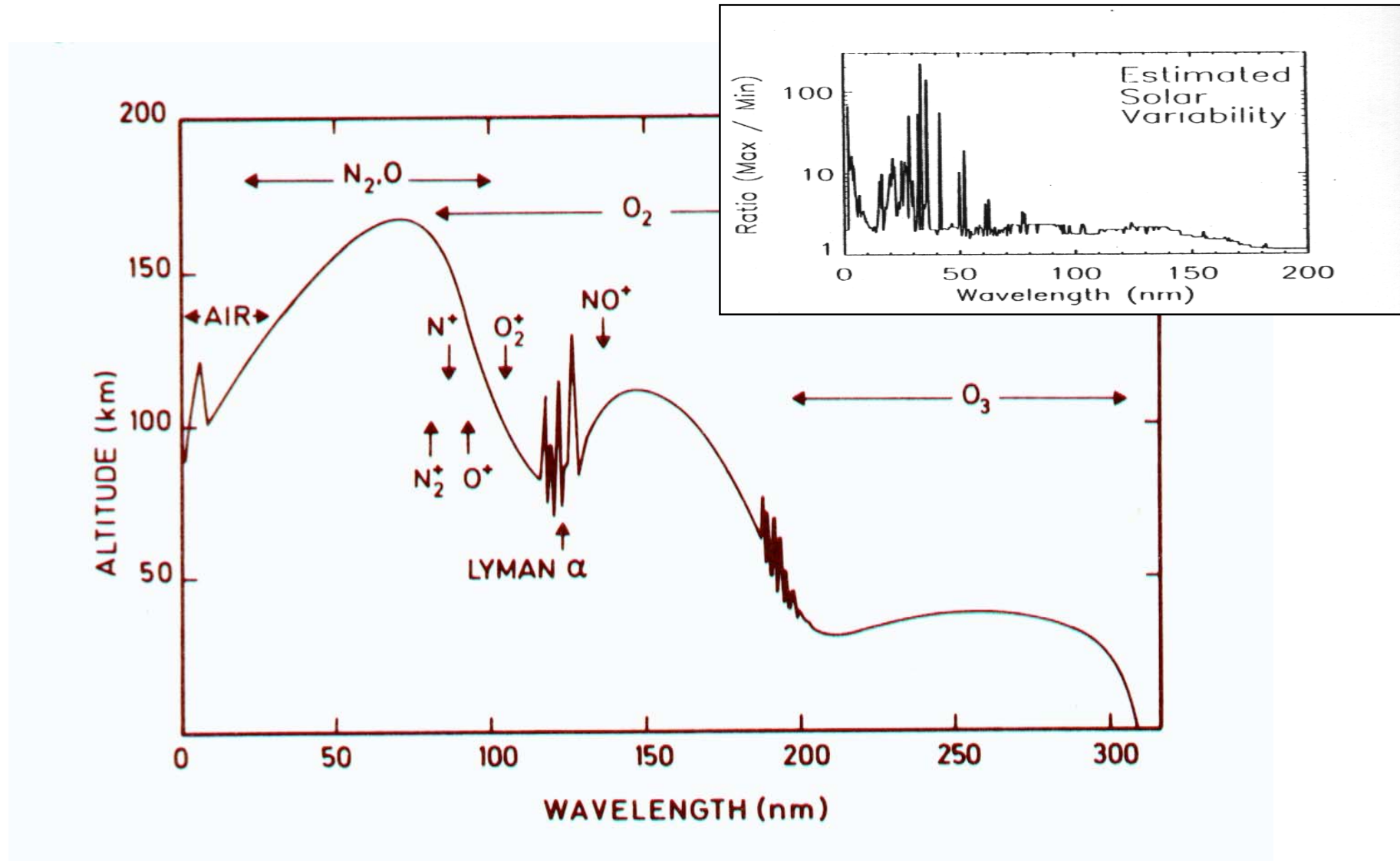
# Solar Radiation Absorption in the Thermosphere



Three factors determine the rate of solar radiation absorption,  $Q$ :

- No. of photons (solar flux,  $F$ )
- No. of absorbing molecules,  $N$
- Efficiency of absorption (cross-section)

# Altitude of Maximum Solar Radiation Absorption





# Molecular Conduction Determines Thermosphere Temperature Profile Shape

Thermal conduction (molecular and turbulent) removes heat from the thermosphere to the mesosphere (here collision frequencies are high enough that polyatomic molecules  $\text{CO}_2$ ,  $\text{O}_3$ ,  $\text{H}_2\text{O}$  can radiate energy away in infrared).

$$\text{Let } \Phi = \text{heat flux due to conduction} = k \frac{dT}{dh}$$

As a first approximation, heat input is balanced by loss due to conduction:

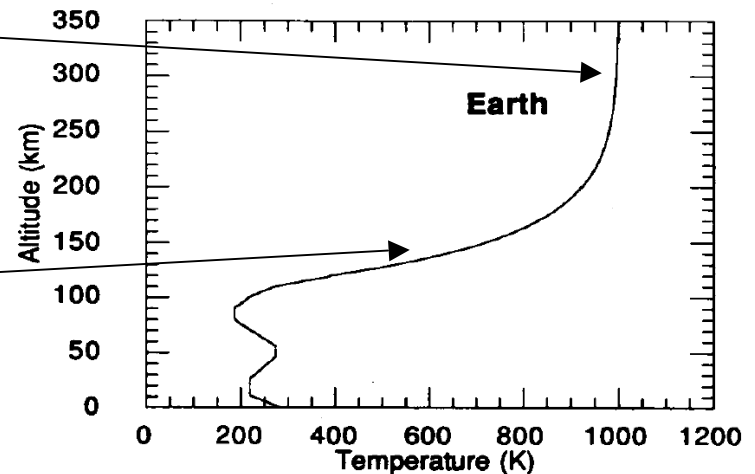
$$Q \approx \frac{d\Phi}{dh} \implies \Phi \approx \int Q dh$$

Therefore 
$$\left. \frac{dT}{dh} \right|_z \approx \frac{1}{k} \int_z^\infty Q dh$$

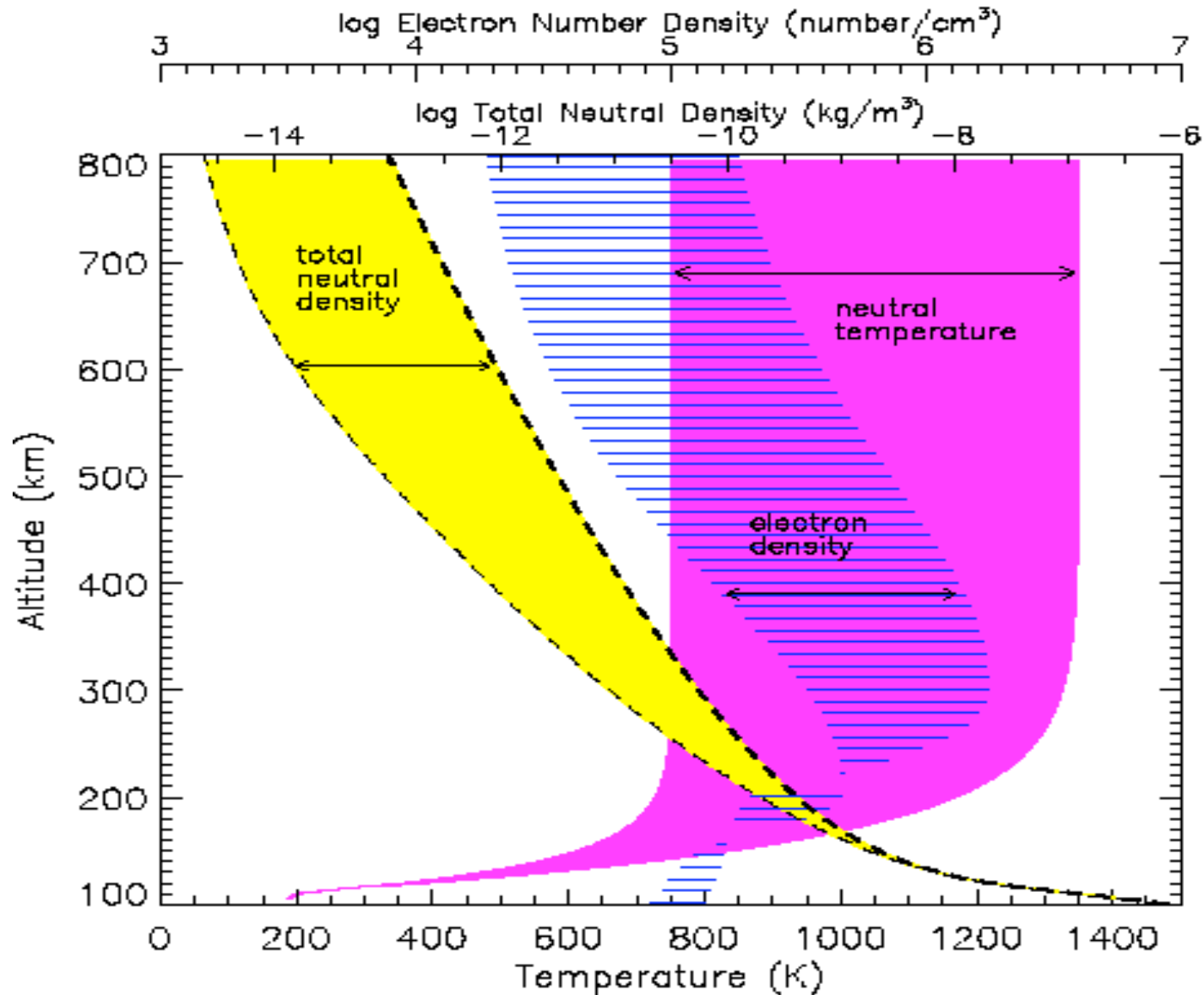
$\frac{dT}{dh}$  must always be sufficiently large to conduct away  $dh$  heat deposited at higher levels. Therefore

•  $\frac{dT}{dh} \rightarrow 0$  above 200 km since  $Q \rightarrow 0$

• also  $\frac{dT}{dh}$  is maximum around 120 - 130 km



# Solar Radiation and Thermosphere Variability

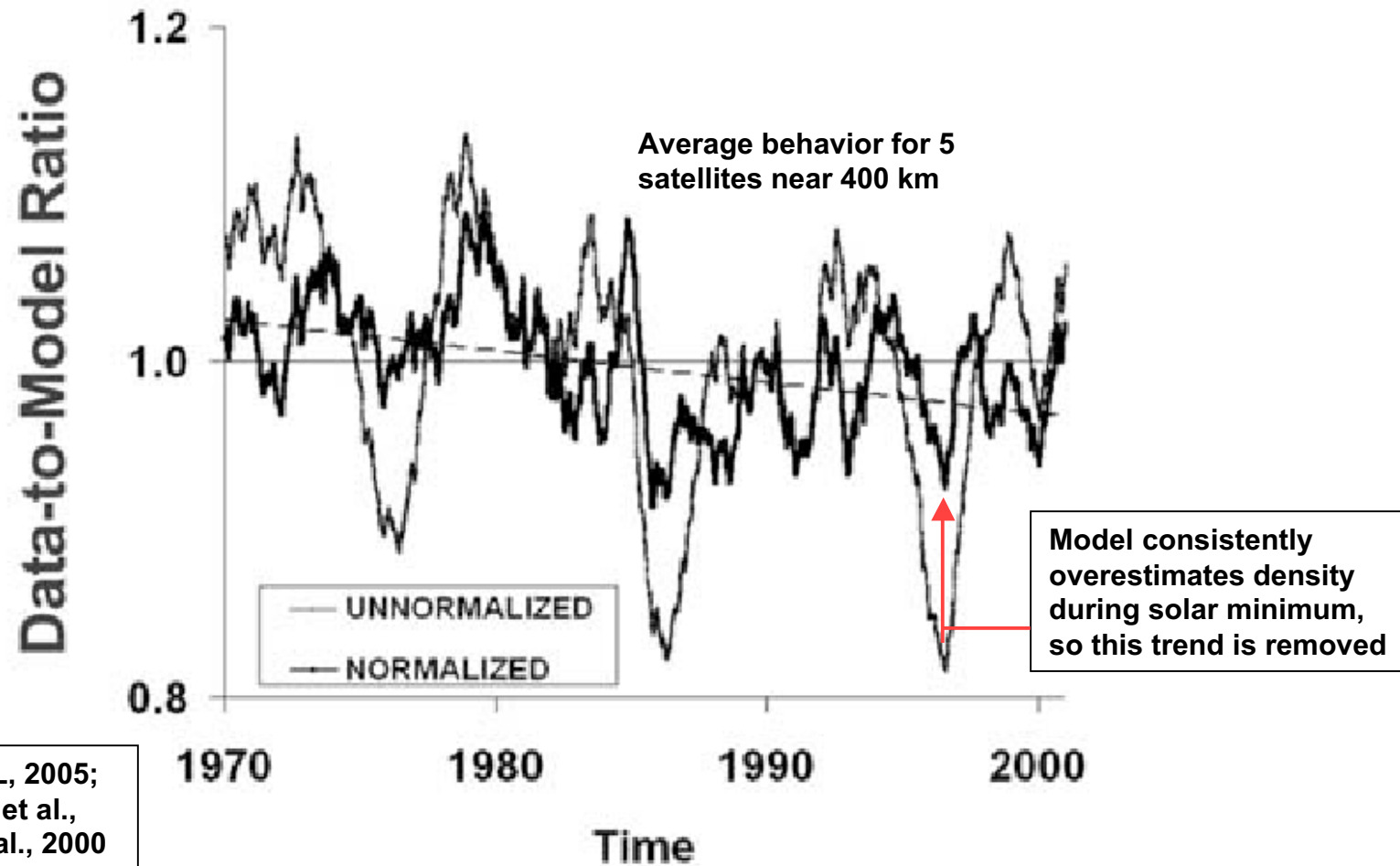


# Radiative Cooling by CO<sub>2</sub>, NO, O is also important to the thermal budget

- Q<sub>T</sub> total neutral heating rate
- K<sub>m</sub> molecular thermal conduction
- K<sub>e</sub> eddy thermal conduction
- NO cooling at 5.3 mm
- CO<sub>2</sub> total CO<sub>2</sub> rate
- O(<sup>3</sup>P) fine structure cooling by O
- IR total of O, NO, CO<sub>2</sub>
  
- NO cooling rate can increase by two orders of magnitude during a storm.

Roble, 1995

## Long-Term Decrease in Thermosphere Density - CO2 Cooling Effect?



# *Hydrostatic Balance & Thermosphere Composition*

# HYDROSTATIC EQUILIBRIUM

If .....

$n$  = # molecules per unit volume, and

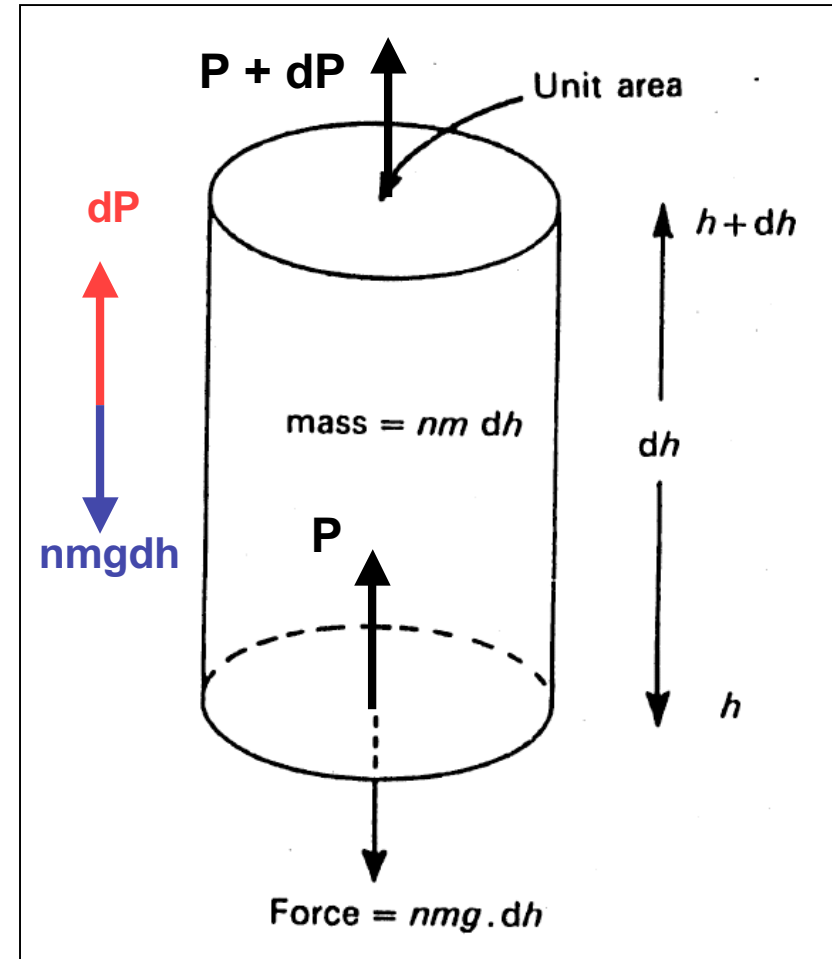
$m$  = mass of each particle, then

$nm \, dh$  = total mass contained in a cylinder of air (of unit cross-sectional area)

and the force due to gravity on the cylindrical mass =  $nmg \, dh$

The difference in pressure between the lower and upper faces of the cylinder balances the above force in an equilibrium situation:

$$(P + dP) - P = -nmgdh$$



$$\Rightarrow \frac{dP}{dh} = -nmg$$

$$R^* = 8.314 \text{ Jmol}^{-1}\text{K}^{-1}$$

$$M = N_a m = 28.97 \text{ gmol}^{-1}$$

for air

$$R^* = kN_a$$

Assuming the ideal gas law holds,

$$n = \frac{P}{kT} \qquad R = \frac{R^*}{M}$$

Then the previous expression may be written:

where H is called the scale height  
and

$$\frac{1}{P} \frac{dP}{dh} = -\frac{1}{H}$$

$$H = \frac{kT}{mg} = \frac{RT}{g} \qquad g = g(0) \frac{R_E^2}{(R_E + h)^2}$$



**This is the so-called hydrostatic law or barometric law.**

Integrating,  $P = P_0 e^{-z}$  where  $z = \int_0^h \frac{dh}{H}$

and  $z$  is referred to as the "reduced height" and the subscript zero refers to a reference height at  $h = 0$ .

Similarly,  $n = n_0 \left( \frac{T_0}{T} \right) e^{-z}$

**For an isothermal atmosphere, then,**

$$P = P_0 e^{-h/H}$$

$$n = n_0 e^{-h/H}$$

$$\rho = \rho_0 e^{-h/H}$$

**Strictly speaking, since  $m$  varies from constituent to constituent (i.e., H, H<sub>e</sub>, O, O<sub>2</sub>, N<sub>2</sub>, ....), the above relations apply to individual constituents, i.e.,**

$$P = P_{i0} e^{-h/H_i}$$

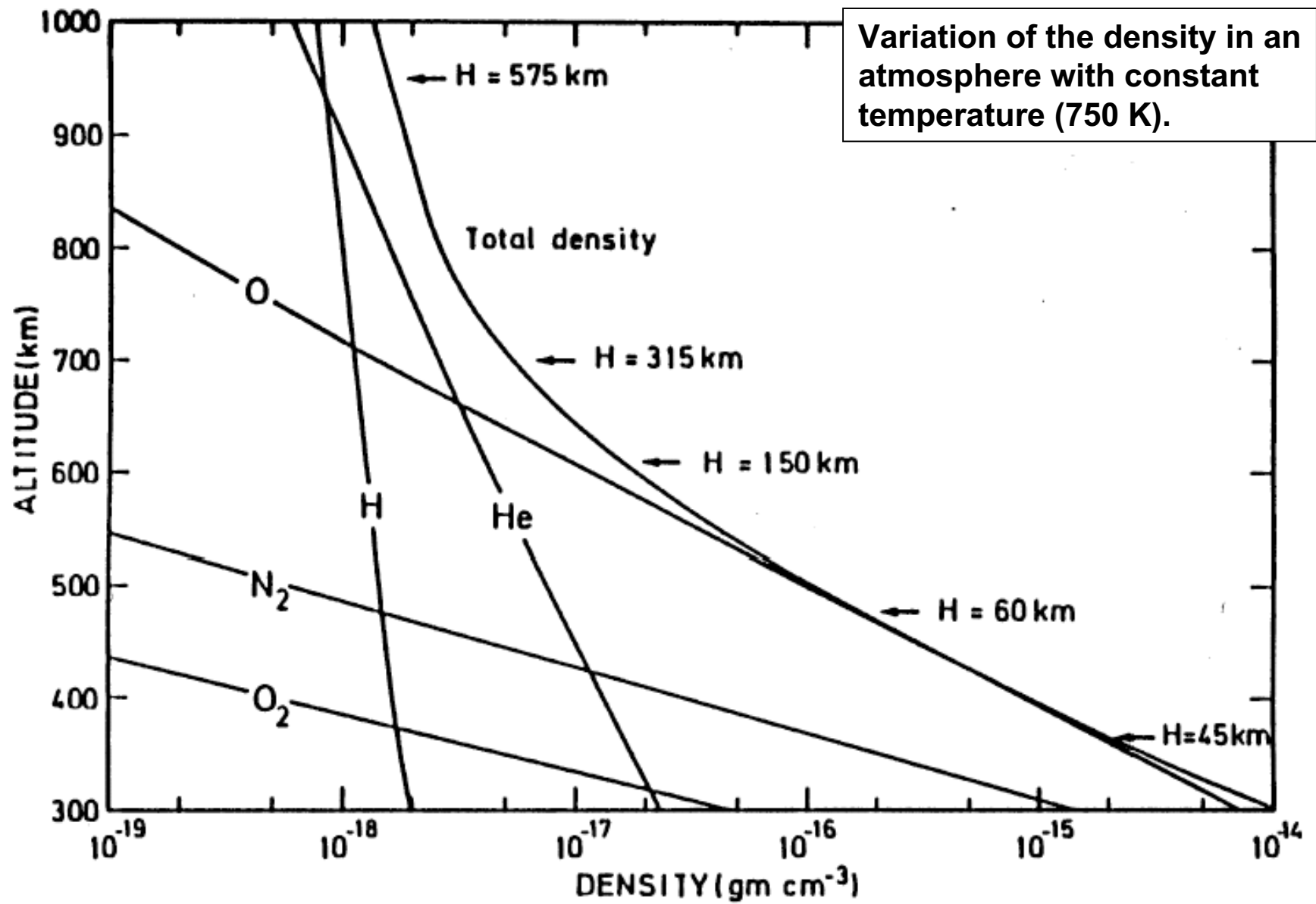
$$n = n_{i0} e^{-h/H_i}$$

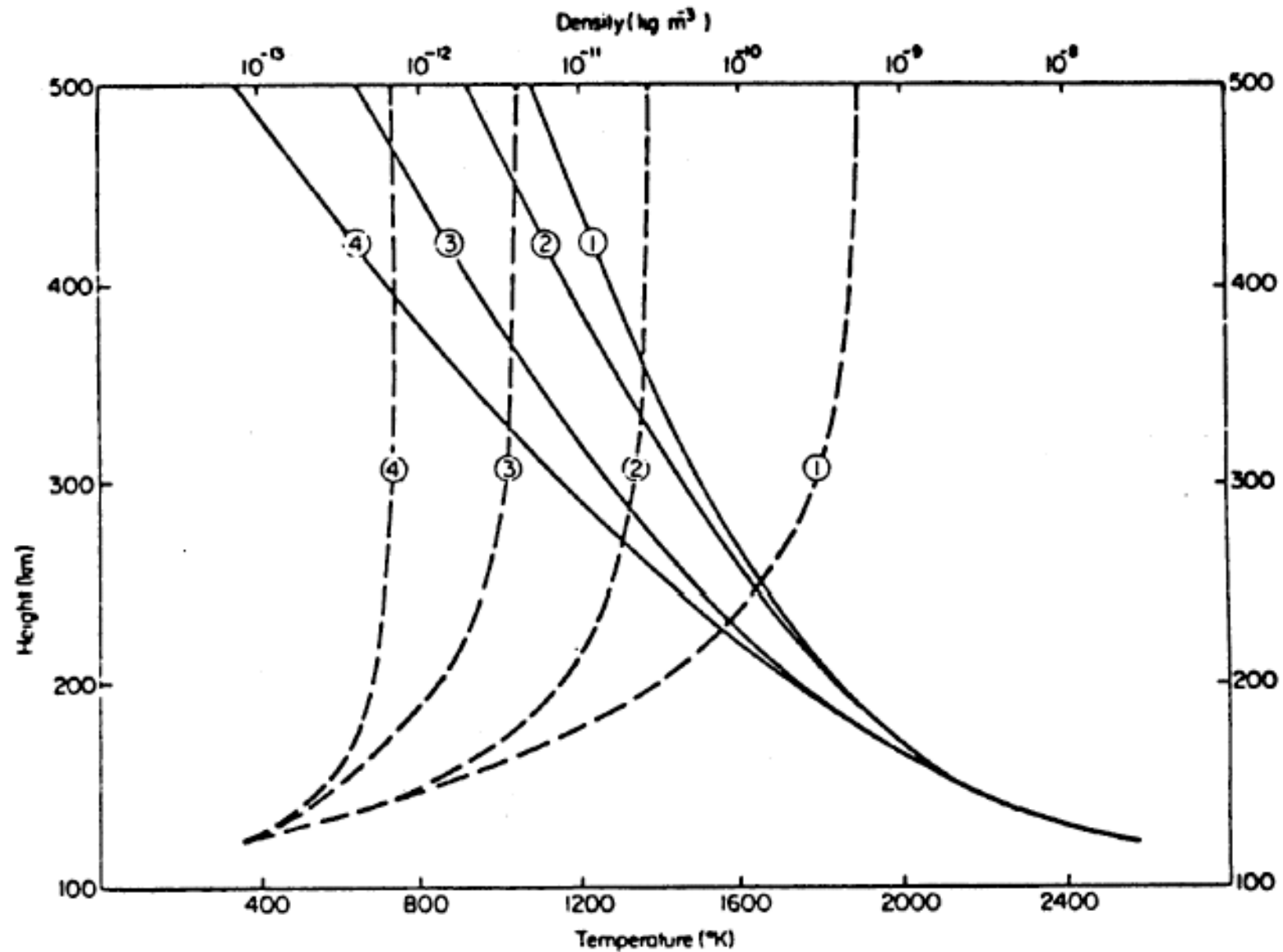
$$\rho = \rho_{i0} e^{-h/H_i}$$

where  $P_i$  is the partial pressure and

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

**Thus, each individual constituent has the tendency to distribute vertically according to its own individual scale height (see following figure). The process which makes this possible is molecular diffusion.**





Vertical distribution of density and temperature for high solar activity (F10.7 = 250) at noon (1) and midnight (2), and for low solar activity (F10.7 = 75) at noon (3) and midnight (4) according to the COSPAR International Reference Atmosphere (CIRA) 1965.

Now, the efficiency of molecular diffusion increases according to the mean free path of atmospheric particles, and hence inversely with atmospheric density. At sufficiently low altitudes in the atmosphere, molecular diffusion is not able to compete with the various mixing processes in the atmosphere (turbulent diffusion, wave and general dynamical transport, etc.).

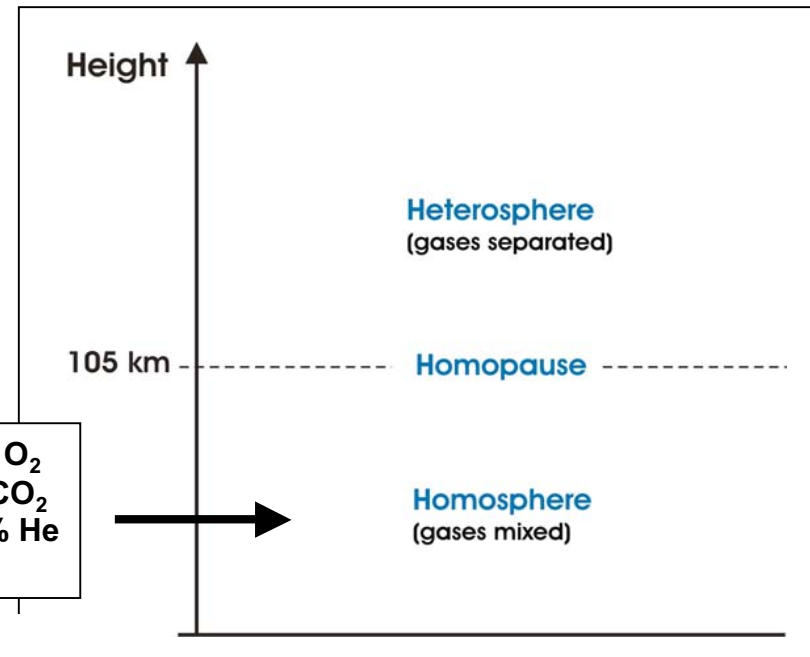
The atmosphere, in fact, remains well-mixed below about 100 km. This regime is called the homosphere and is characterized by a constant mean molecular weight as a function of height:

$$\bar{m} = \frac{\sum m_i n_i}{\sum n_i} = 28.97 \text{ g mole}^{-1}$$

A mean scale height is thus defined:

$$\bar{H} = \frac{kT}{mg}$$

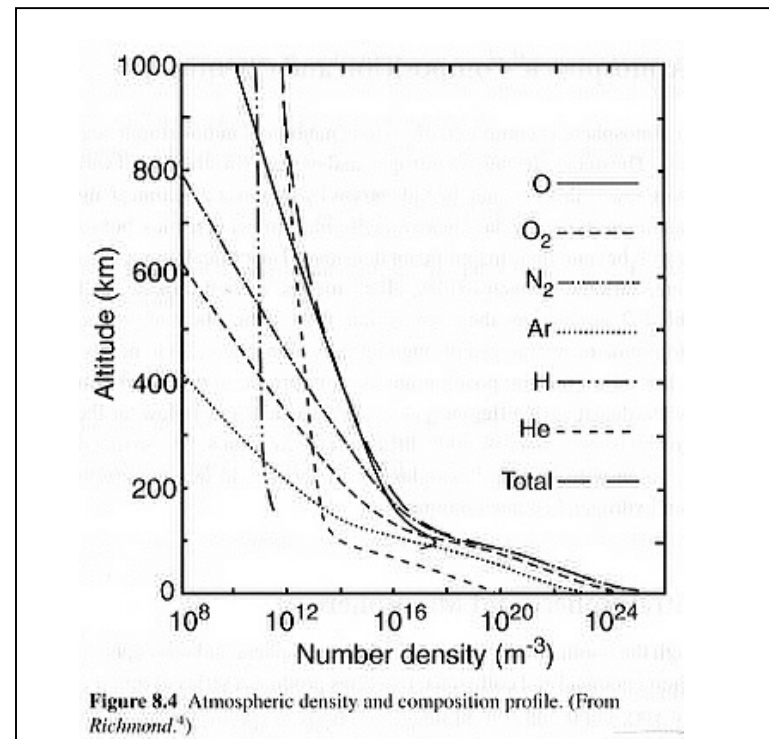
78.1% N <sub>2</sub>	20.9% O <sub>2</sub>
.9% Ar	.03% CO <sub>2</sub>
.002% Ne	.0005% He
Variable H <sub>2</sub> O	



..... and all constituents possess the same scale height and number density (and pressure) distributions with height:

$$n_i = \bar{n} = n_0 \left( \frac{T_0}{T} \right) e^{-z/H}$$

- It is not until about 100 km (the exact height is species dependent, due to the dependence of molecular diffusion velocity on mean molecular weight) that molecular diffusion begins to take over, and each species separates according to its individual scale height.
- This separation occurs at the homopause, sometimes called the turbopause. Above the homopause is the heterosphere; homosphere below.



# Deviations from Hydrostatic Equilibrium: Diffusive and Chemical Time Constants

**Distributions of atmospheric constituents in the atmosphere are not just affected by gravity, but also by chemical processes and transport by winds and diffusion.**

***The continuity equation expresses the balance between these competing processes.***

**The equation of continuity for the number density of the  $i^{\text{th}}$  constituent  $N_i$  is**

$$\frac{\partial N_i}{\partial t} = P_i - L_i - \nabla \cdot \vec{\Phi}_i$$

$\vec{\Phi}_i = \text{flux } N_i \vec{V}_i \text{ due to mass transport}$

with velocity  $\vec{V}_i$

**Generally, vertical variations are most important for many atmospheric problems:**

$$\nabla \cdot \vec{\Phi}_i \approx \frac{\partial}{\partial z} N_i w_i$$



From gas kinetic theory (i.e., Chapman & Cowling 1961):

$$\Phi_i = N_i w_i \approx -N_i D_i \left[ \frac{1}{N_i} \frac{dN_i}{dz} + \frac{1}{H_i} + \frac{(1 + \alpha_i)}{T} \frac{dT}{dz} \right] \quad \text{molecular diffusion}$$

$$- N_i K \left[ \frac{1}{N_i} \frac{dN_i}{dz} + \frac{1}{H} + \frac{1}{T} \frac{dT}{dz} \right] \quad \text{eddy diffusion}$$

Note:  
Magnitude  
 $N_i w_i \sim N_i D_i / H_i$

$$\frac{1}{P_i} \frac{dP_i}{dz}$$

Note: for  $w_i = 0$  we have  $\frac{1}{P_i} \frac{dP_i}{dz} = -\frac{1}{H_i}$  or  $\frac{1}{P_i} \frac{dP_i}{dz} = -\frac{1}{H}$

$$H_i = \frac{R^* T}{m_i g}$$

$$H = \frac{R^* T}{\bar{m} g}$$

$$D_i = \frac{kT}{m \nu}$$

$\nu =$  collision frequency

Note:  $D_i$  increases exponentially with height

**We can define a time constant for the diffusive process.**

$$\frac{\partial N_i}{\partial t} \sim \frac{\partial}{\partial z} N_i w_i \sim \frac{N_i D_i}{H_i^2}$$

$$\frac{1}{N_i} \frac{\partial}{\partial t} N_i \sim \frac{D_i}{H_i^2} \Rightarrow N_i \sim N_{i0} e^{-t/\tau_D}$$

**Suppose diffusion is the only process acting.**

$$\text{where } \tau_D \sim \frac{H_i^2}{D_i} \quad (\text{or } \tau_M \sim \frac{H^2}{K})$$

**The relative importance of various process can be determined by comparing their time constants. *The smaller time constant indicates the dominant process.***

**For instance, we may define the homopause or turbopause according to the relative lifetimes of a constituent with respect to molecular diffusion or mixing:**

$$\tau_D \ll \tau_M \quad z > 105 \text{ km} \quad \Rightarrow \quad \text{“heterosphere”}$$

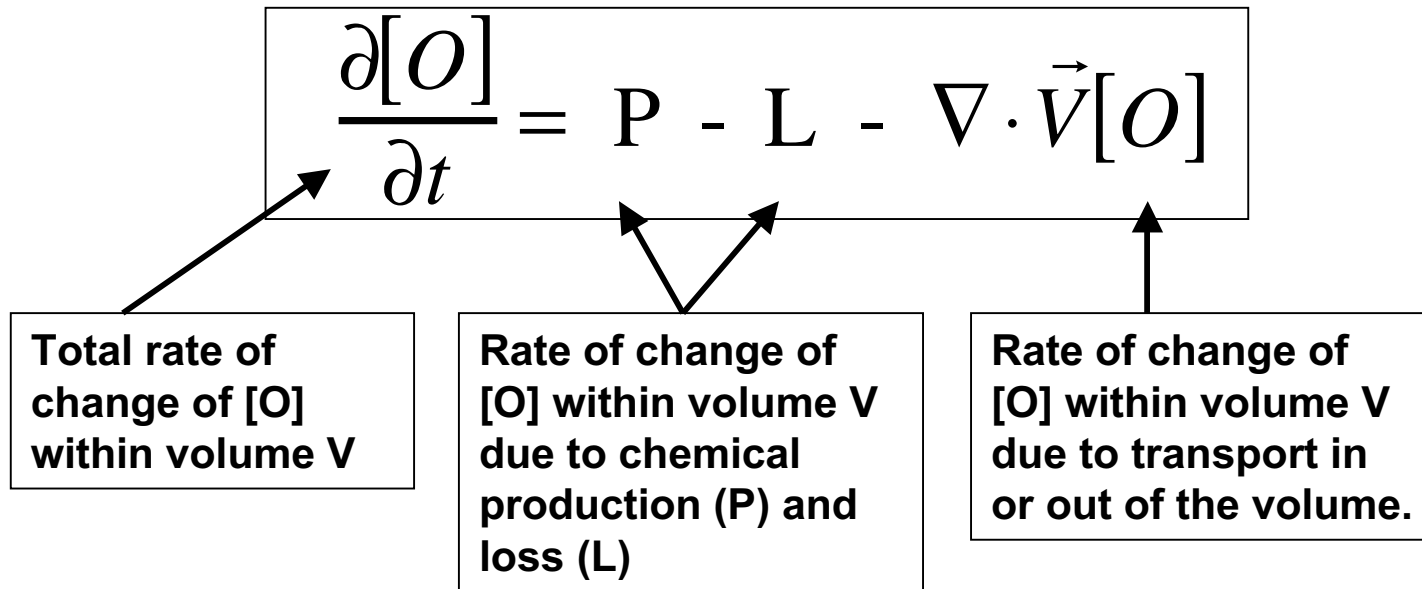
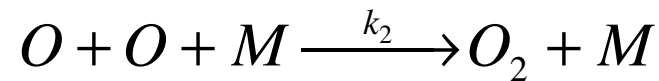
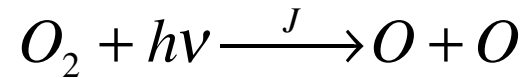
$$\tau_D \gg \tau_M \quad z < 105 \text{ km} \quad \Rightarrow \quad \text{“homosphere”}$$

$$\tau_D \sim \tau_M \quad z \sim 105 \text{ km} \quad \Rightarrow \quad \text{“turbopause”}$$

$$\tau_D \sim \frac{H_i^2}{D_i} \quad \text{and} \quad \tau_M \sim \frac{H^2}{K}$$

**Are the time constants relevant for re-establishing hydrostatic equilibrium if the species are driven from hydrostatic equilibrium by some process (i.e., vertical motion due to sudden heating; introduction of a reactive chemical into the atmosphere; sudden lifting of the ionosphere by electric fields).**

# Atomic Oxygen



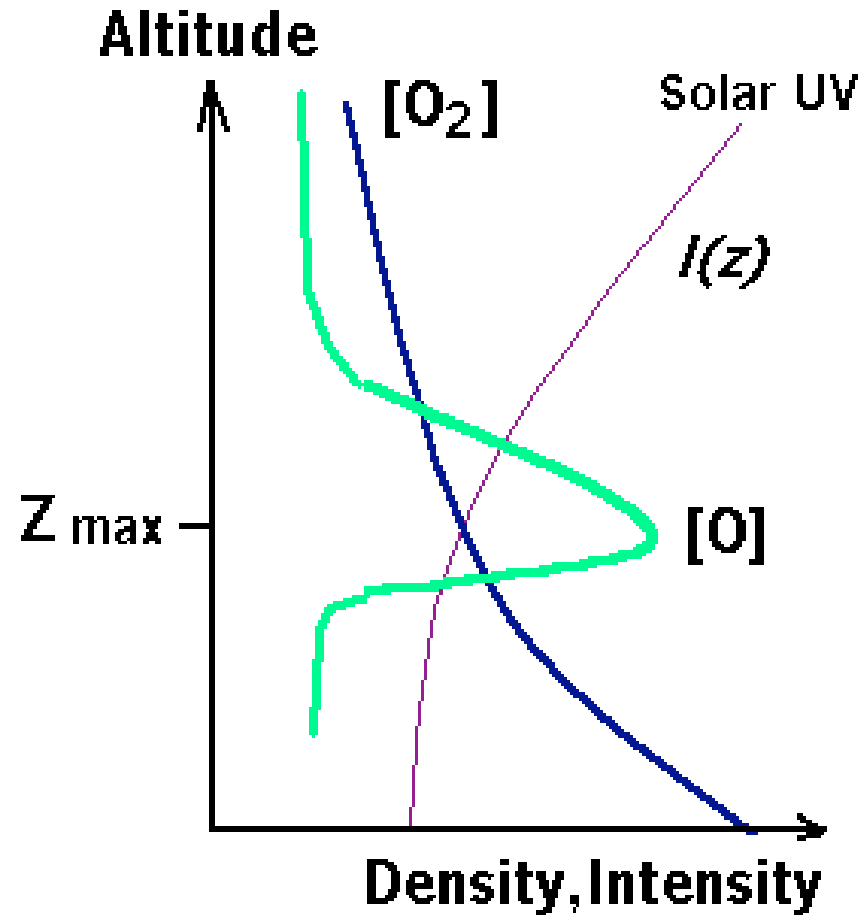
Photochemical equilibrium implies a balance between production and loss, i.e.,  $P = L$ .

Note: The **law of mass action** says that the rate of a chemical reaction is proportional to the product of the densities of species taking part in the reaction.

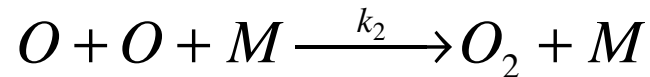
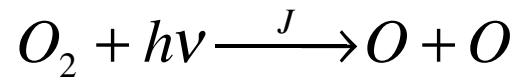
$$2J[O_2] = k_2[O]^2[M]$$

$$\therefore [O] = \left\{ \frac{2J[O_2]}{k_2[M]} \right\}^{1/2}$$

Thus, given  $[O_2]$ ,  $J$ ,  $k$  and  $[M]$ , one can calculate the atomic oxygen profile in Earth's atmosphere:



However, this profile does not agree with measurements, particularly the height of the peak.



So we ask,  
after photo-dissociation, what  
is the most likely fate for O?

Near 95 km,  
 $\tau_L \sim 100$  d

$$\frac{\partial [O]}{\partial t} \sim -k_2 [O]^2 [M]$$

$$\frac{1}{[O]} \frac{\partial}{\partial t} [O] \sim -k_2 [O][M] \Rightarrow [O] \sim [O]_0 e^{-t/\tau_L}$$

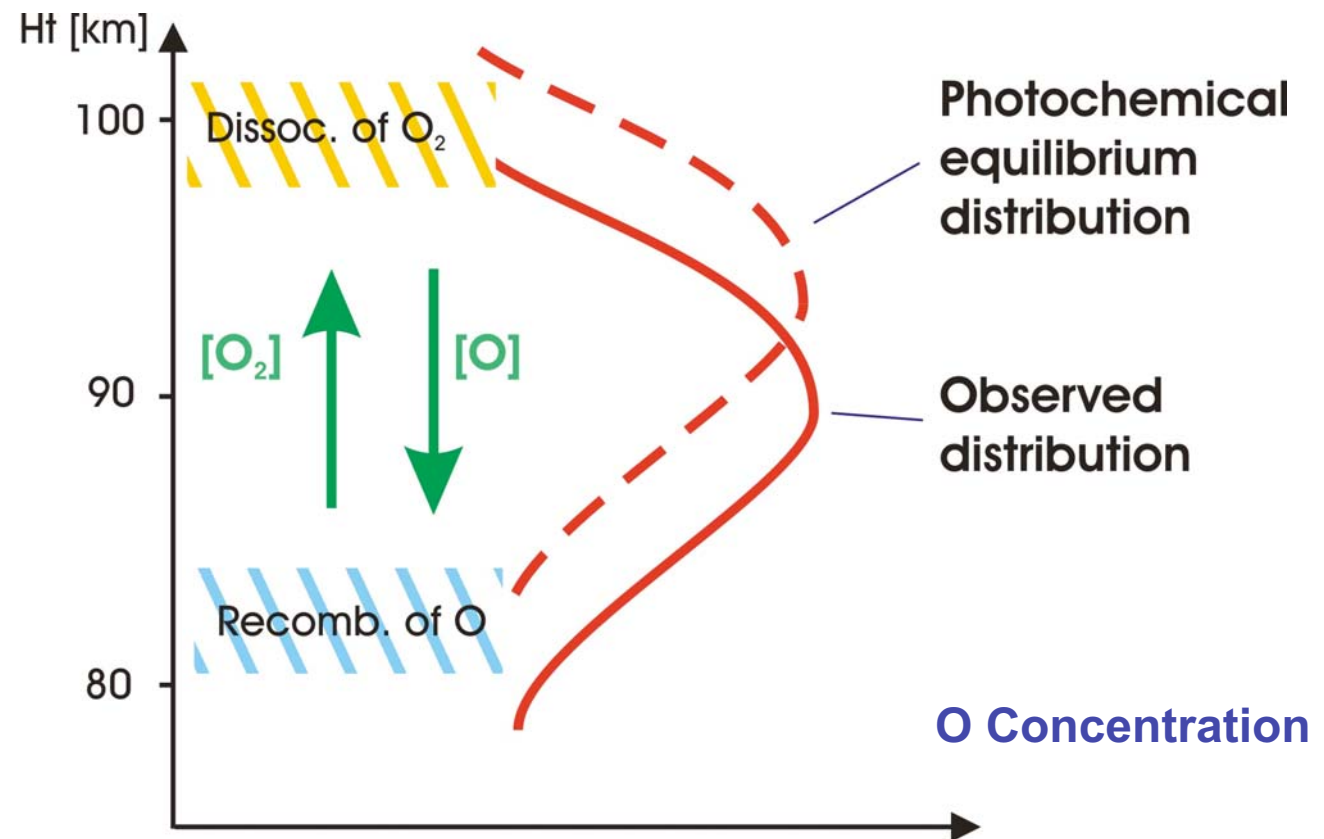
where  $\tau_L \sim \frac{1}{k_2 [O][M]}$

$$\tau_M \sim \frac{H^2}{K}$$

Near 95 km,  $K \sim 100 \text{ m}^2\text{s}^{-1}$  (no  
systematic height dependence),  
and  $H \sim 8$  km, so  $t_M \sim 6$  d

At 80-100 km, the time constant for mixing is more efficient than recombination, so *mixing due to turbulence and other dynamical processes must be taken into account* (i.e., photochemical equilibrium does not hold).

Mixing transports O down to lower (denser) levels where recombination proceeds rapidly (the "sink" for O).



After the O recombines to produce O<sub>2</sub>, the O<sub>2</sub> is transported upward by dynamical mixing to be photodissociated once again (the "source" for O).

**Thus, “hydrostatic equilibrium” for a particular atmospheric constituent implies**

- **Steady-state conditions**
- **Chemical production and loss have long time constants compared to diffusion**
- **There is no net diffusion between the constituent and the other major species**

**Planetary escape by atomic hydrogen is another case where the mean vertical profile of a thermosphere constituent deviates from hydrostatic balance. This is because there is a net escape flux of H from the atmosphere**



# Thermosphere Variability & Empirical Models of the Thermosphere

## ***Our Early Knowledge of the Thermosphere was Derived from the Observed Effects of Atmospheric Drag on Satellite orbits***

Through Kepler's laws, one can derive the rate of change of orbital period ( T ) in terms of the atmospheric density:

$$\frac{dT}{dt} = -\frac{3}{2} B \rho_P \int \frac{\rho}{\rho_P} ds$$

where **B** = **B-factor (ballistic coefficient)** =  $\frac{C_D A}{m}$

$\rho_P$  = density at perigee

$\rho$  = density

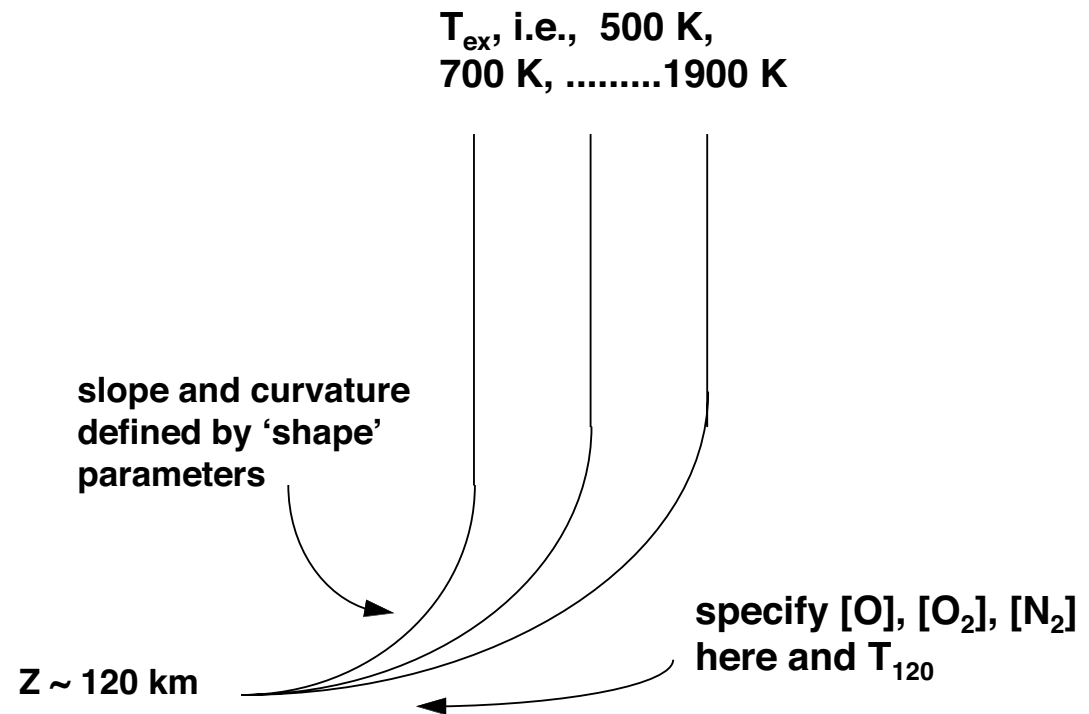
**s** = satellite path

From radar tracking, one can derive the atmospheric density (the more accurate the tracking, the shorter time required to determine density). Typical resolution is about 1 day below 200 km and 5 days at 500 km. (Much better with laser beacon, etc.)

The above procedure requires some knowledge about the variation of density with height.

Given the equations defining the hydrostatic law, the chemical composition at some height, and an expression for the temperature profile shape, a density profile can be retrieved from the satellite observations.

In practical applications of this method, if a reasonable initial first guess of the vertical structure is provided, an iterative procedure leads to an accurate determination that is independent of the initial guess.



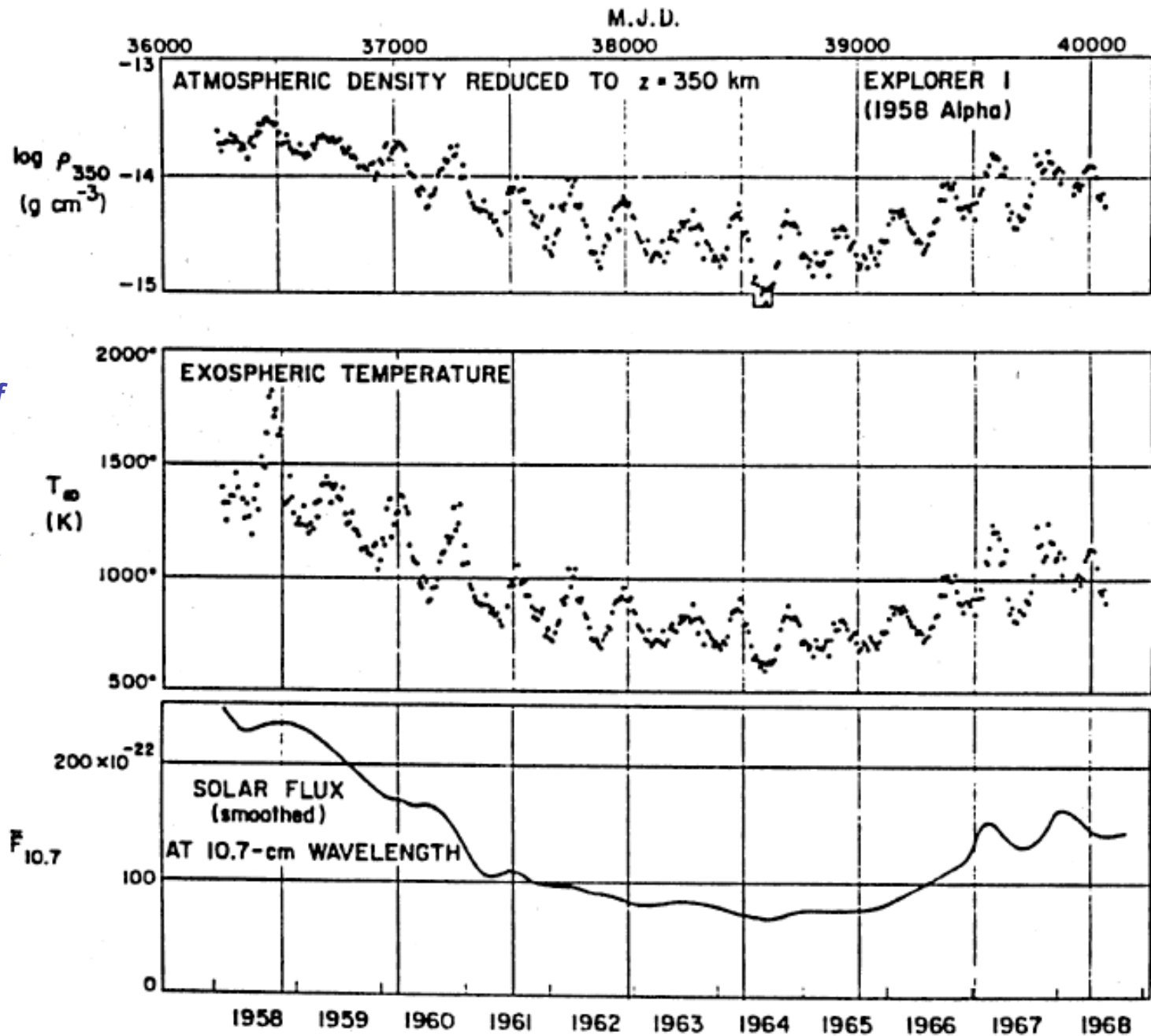
***A set of "exospheric temperatures" emerges from this process.***

**Analyses of many satellite orbits led to the so-called "static diffusion models" developed by Jacchia and which form the basis of many operational drag models. These are based on a parametric dependence of density on exospheric temperature.**

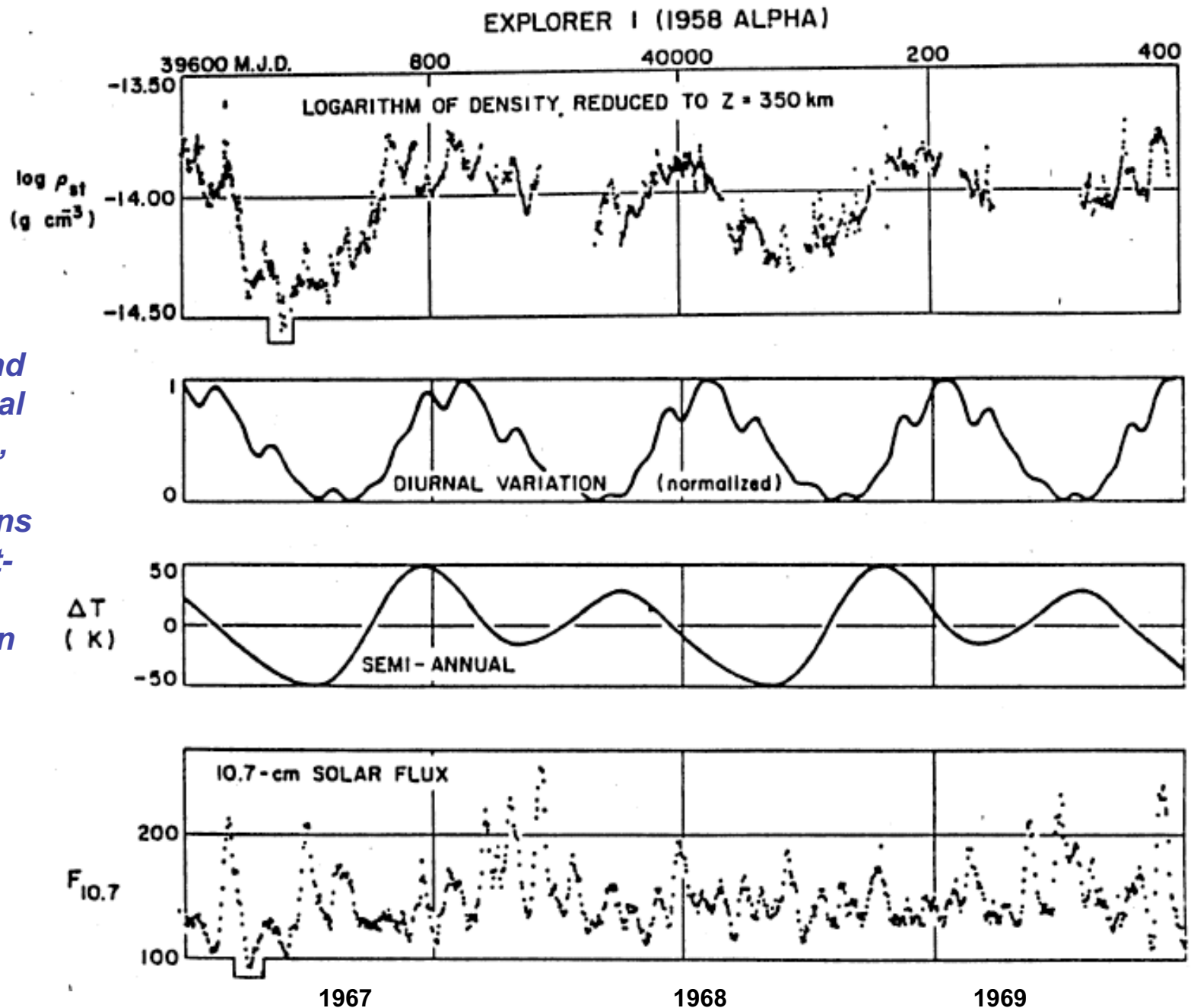
**The derived exospheric temperatures (and the densities) reveal many of the variations typical of the thermosphere: annual, semiannual, solar activity, magnetic activity, diurnal, etc. (see following figures).**

**DESPITE THE IMMENSE SUCCESS OF THESE MODELS AT THE TIME, THEY SUFFER FROM SOME FUNDAMENTAL LIMITATIONS:**

- The derived temperature is more of a 'virtual' temperature than 'real' (kinetic) temperature**
- Rocket measurements of O, O<sub>2</sub> at the lower boundary are difficult to interpret (i.e., O recombines into O<sub>2</sub> against walls of measuring device, meaning that O can be underestimated and O<sub>2</sub> overestimated).**
- Wind-induced diffusion is also important, i.e., for [O]; The 'static diffusion' or 'hydrostatic' restriction is not amenable to addressing vertical transport (i.e., upwelling) or horizontal transport.**



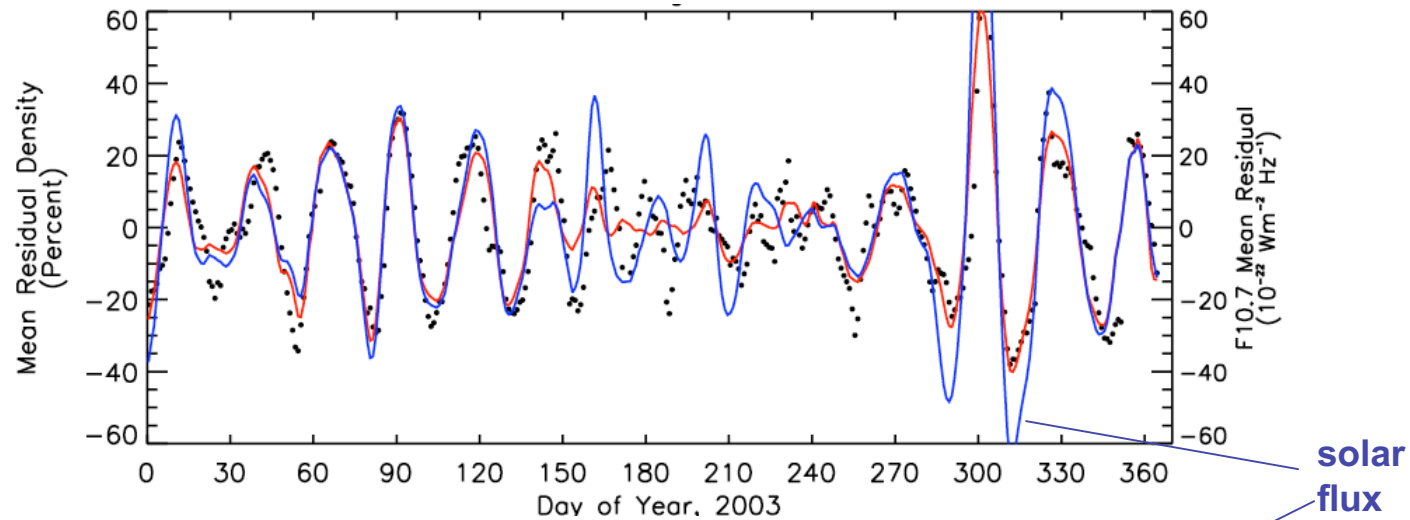
*Correlation of density and temperature with long-term changes in solar activity.*



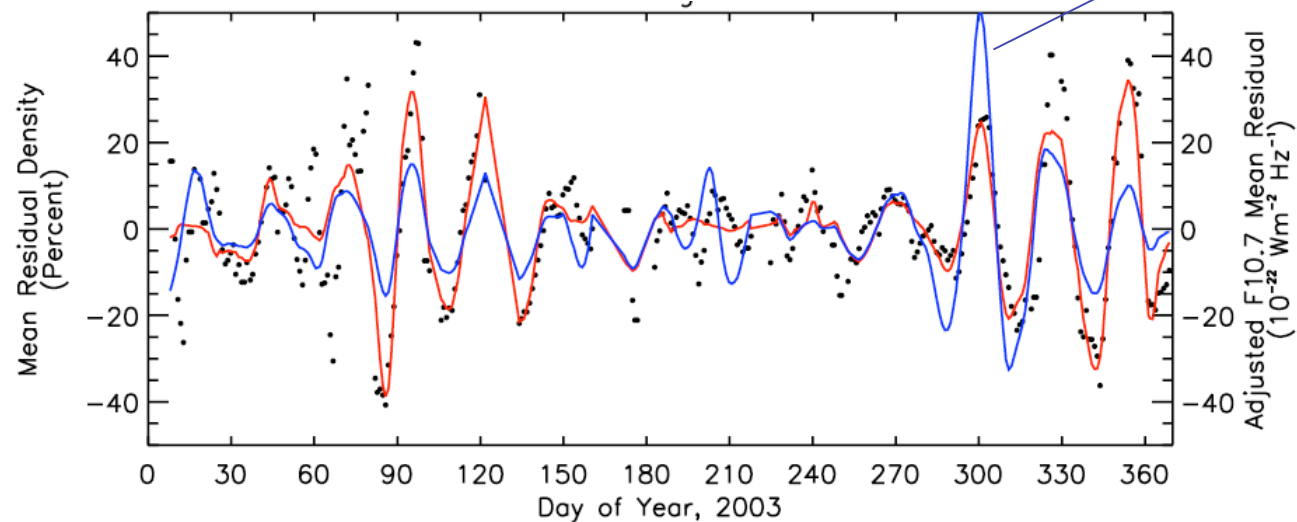
*Diurnal and semiannual variations, and correlations with short-term changes in solar activity.*

# Contemporaneous Solar Rotation Effects at Mars and Earth

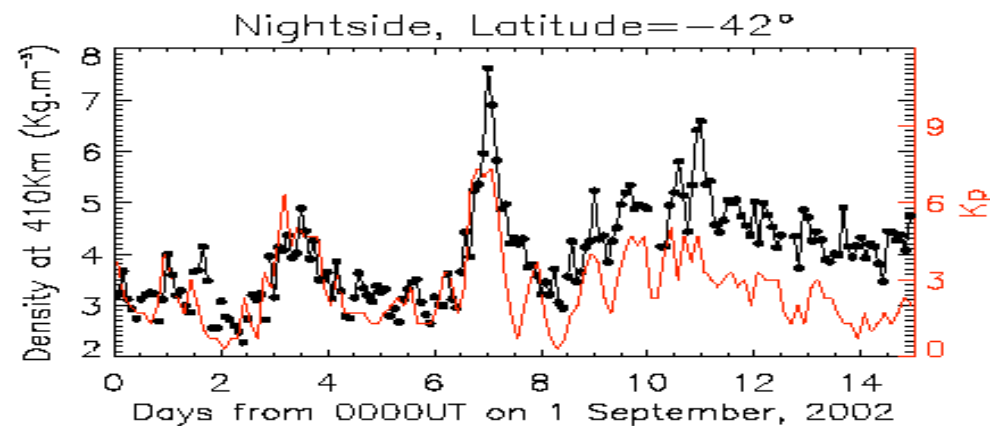
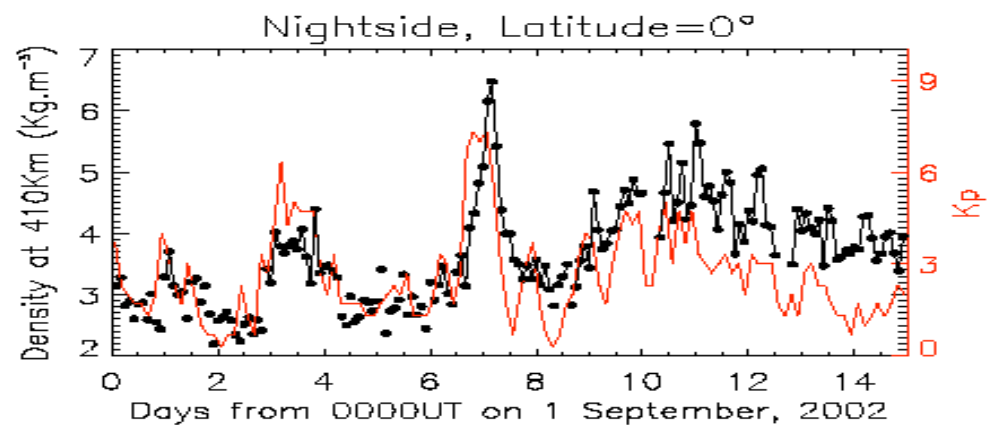
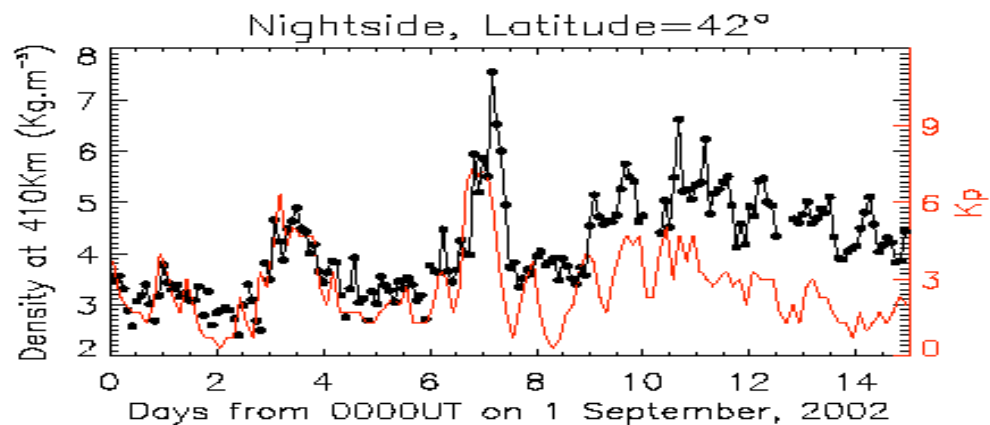
## CHAMP 5d Orbit-Integrated Densities - 410 km



## Mars Global Surveyor Orbital Decay (5d) - 390 km



**Densities at  
410 km  
derived from  
the STAR  
Accelerometer  
on the CHAMP  
Satellite**





**IN THE 1970'S, TWO DATA SETS BECAME AVAILABLE THAT ADDRESS THE PREVIOUSLY-MENTIONED LIMITATIONS FOR DRAG-BASED MODELS:**

- **determinations of  $T_{\text{ex}}$  from incoherent scatter radar measurements.**
- **satellite mass spectrometer measurements of O, O<sub>2</sub>, N<sub>2</sub>, He, H, etc., and also measurements of  $T_{\text{ex}}$ . (satellites like OGO-6, AE, and many others).**

**HENCE LEADING TO THE**

**Mass Spectrometer Incoherent Scatter (MSIS) models of A. Hedin (NASA/GSFC).**

**However, the MSIS models are not optimized with respect to satellite drag, and so have not been widely adopted for ephemeris computations in lieu of the Jacchia models. In principle, though, getting closer to the correct physics should lead to improved orbital predictions.**

## *Some of the other models developed during the past 30 years*

### EMPIRICAL

**U.S. Standard 1962**

**U.S. Standard Supplements, 1966**

**MSIS86, MSIS90, MSISE90**

**CIRA-1986**

**Jacchia 1964**

**CIRA-1961, 1965**

**Jacchia-1971, 1977**

**NRLMSIS2000**

### NUMERICAL/THEORETICAL MODELS

**University College London Thermosphere-Ionosphere Model**

**(now Coupled Thermosphere-Ionosphere-Plasmasphere  
Model (CTIP) at CU/CIRES)**

**NCAR Thermosphere-Ionosphere GCM**

**TGCM**

**TIGCM**

**TIE-GCM**

**TIME-GCM**

# Additional Slides

# Planetary Escape

# Atmospheric Escape

## Maxwell-Boltzmann Distribution

For an ideal gas consisting of perfectly elastic spheres in random thermal motion, and under equilibrium conditions, the number of molecules  $dN$  out of a total  $N$  having a speed between  $c$  and  $(c+dc)$  is given by the *Maxwell-Boltzmann distribution*:

$$\frac{dN}{N} = 4\pi \left( \frac{m}{2\pi kT} \right)^{3/2} e^{-\frac{mc^2}{2kT}} c^2 dc$$

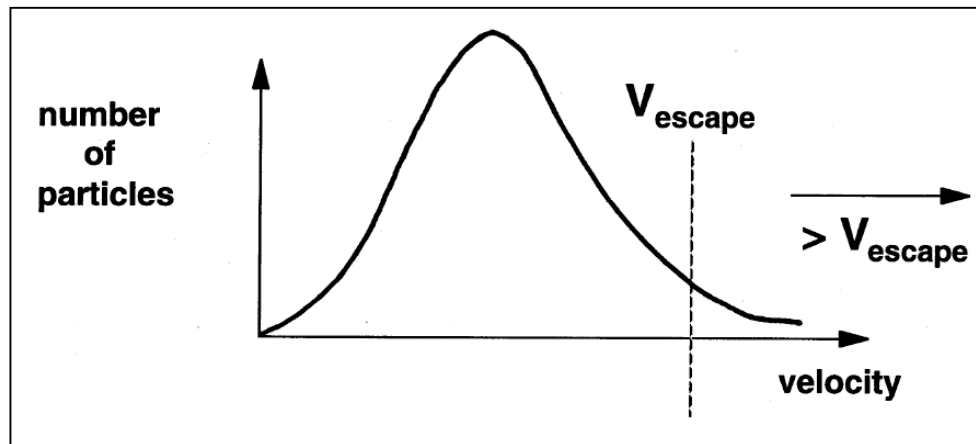
The above provides a definition of *kinetic temperature* valid whenever a gas is in *thermal equilibrium*.

## Critical Level

Above a certain level the mean free path of the particles becomes so long (i.e., on the order of a scale height) that a molecule traveling upward can go all the way out to space without colliding at all. The level at which this occurs is called the *critical level*.

## Exosphere and Exobase

*The region above the critical level is called the exosphere. The critical level is sometimes called the exobase. The exobase is between about 500-800 on Earth (depending on solar activity), and at about 200 km on Venus.*



**Above the exobase there exists a substantial fraction of the particles with velocities greater than the escape velocity.**

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.

*This produces a net upward flux of atmospheric gases, in practice only relevant for atomic hydrogen, that introduces an additional term in the “hydrostatic equation”.*

## Jeans Escape

Assuming that particle velocities are thermal, the escape flux is given by:

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

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

where

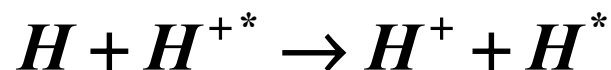
$$U = \left( \frac{2kT}{M_{particle}} \right)^{1/2} \quad \text{and} \quad \lambda = \frac{u_{esc}^2}{U^2}$$

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

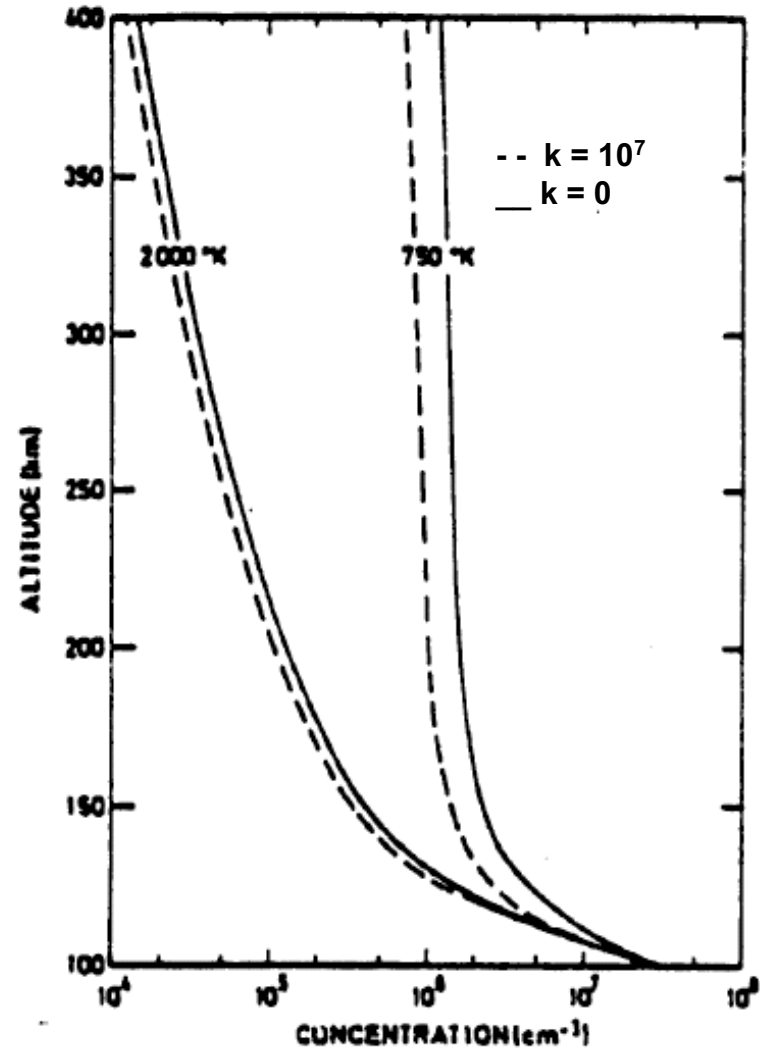
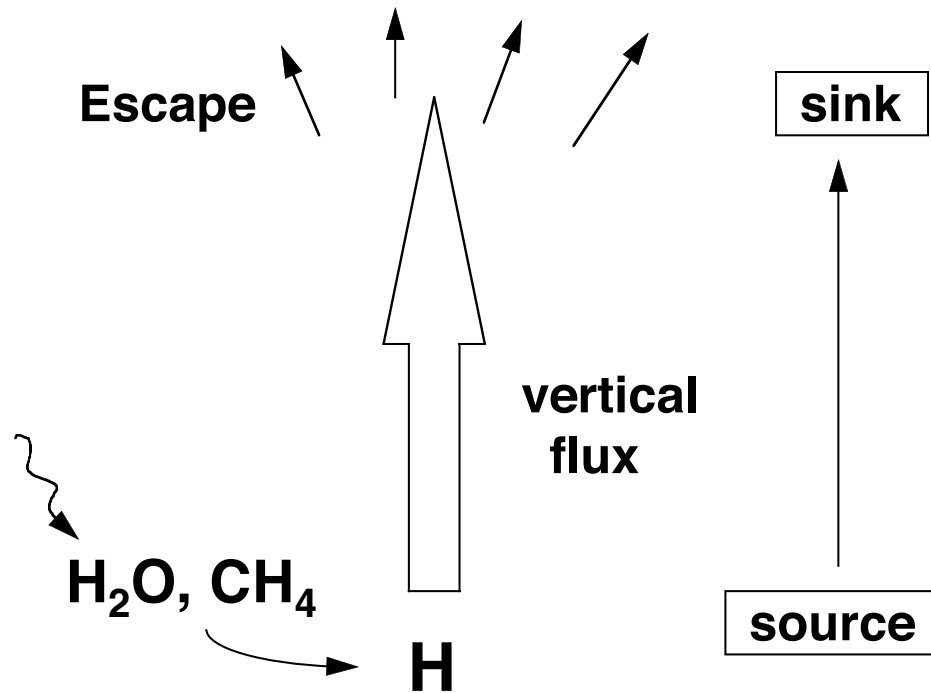
$U$  is the most probable velocity of a Maxwellian distribution of thermal velocity.

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:



In Earth's atmosphere, escape has an important effect on variations in Hydrogen densities.



The thermal or Jeans escape flux increases with temperature, causing the hydrogen density to decrease with increasing solar activity. ↗