The neutral upper atmosphere
thermosphere and exosphere

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Earth's atmosphere layers

Extreme UltraViolet

Schumann-Runge Bands & Continuum

O
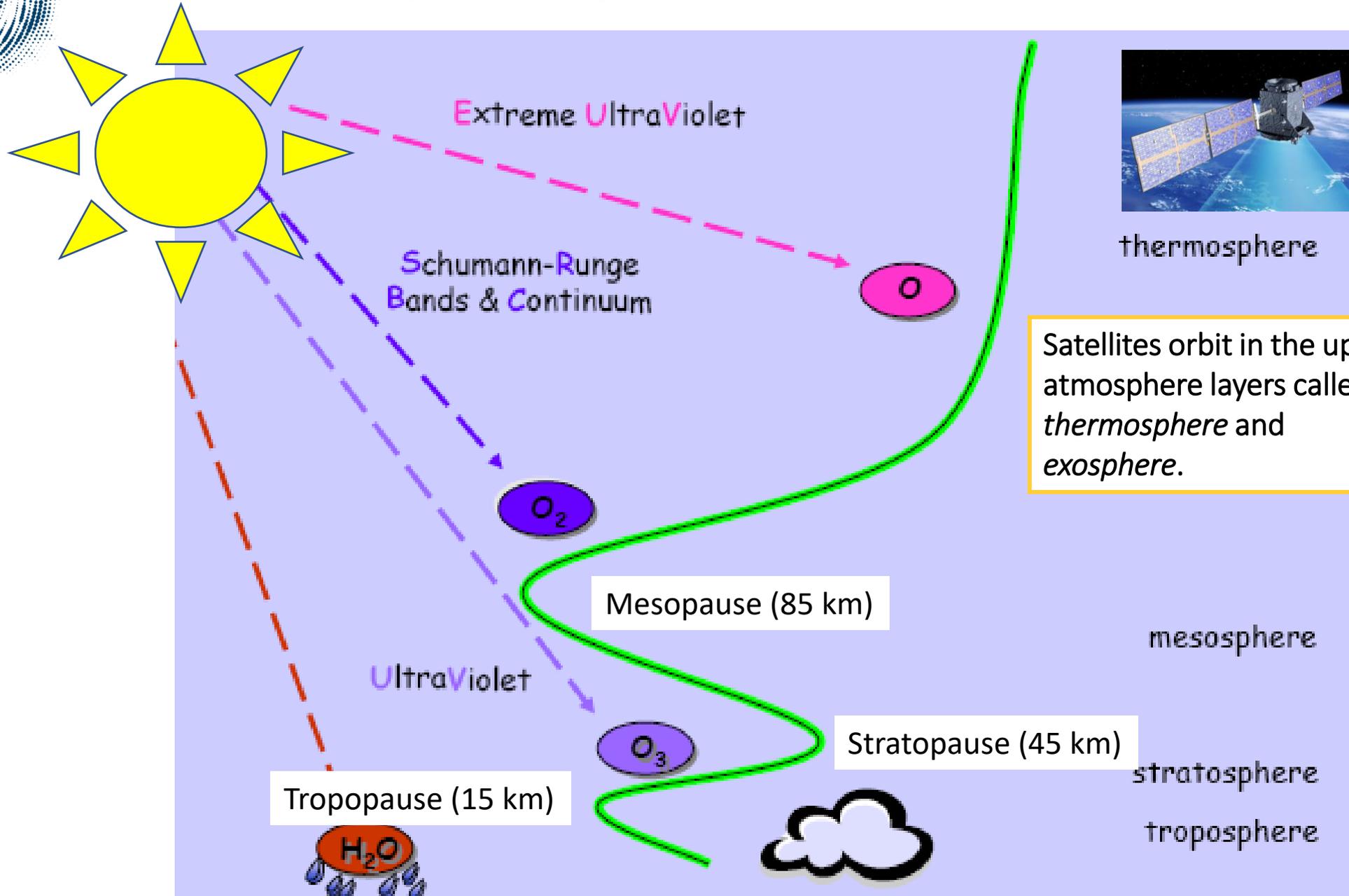
thermosphere

Satellites orbit in the upper atmosphere layers called *thermosphere* and *exosphere*.

O_2

Mesopause (85 km)

mesosphere

UltraViolet

O_3

Stratopause (45 km)

stratosphere

Tropopause (15 km)

H_2O

troposphere

Exosphere — 10,000 km

690 km

Thermosphere — Shuttle — Aurora — 100 km (Kármán line)

85 km — Mesosphere — Meteors

50 km — Stratosphere — Weather balloon

6 – 20 km — Troposphere — Mount Everest
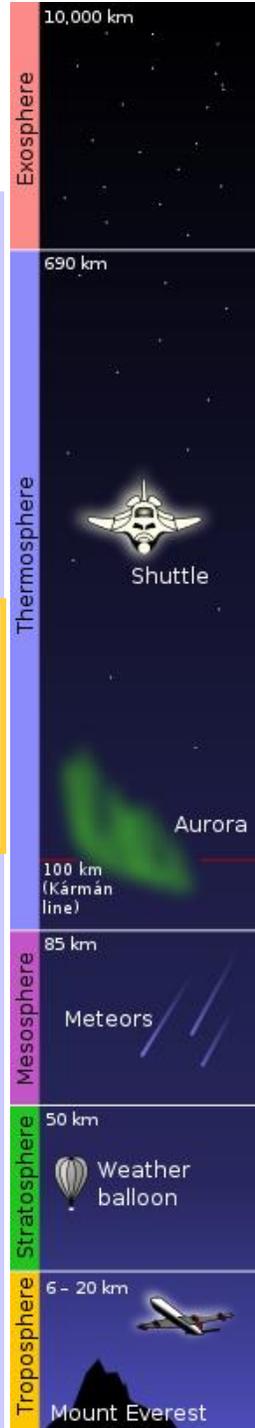
Global Mean Energy Sources and Sinks

Thermosphere:
- Energy sources:
  - Absorption of EUV (20-100 nm; photoionizing O, O_2, N_2) and UV (120-200 nm; photodissociating O_2), 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_2, O_3 and H_2O
  - IR cooling by CO_2 NO, O

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

Troposphere:
- **Energy sources**:
  - Planetary surface absorption (IR, visible), convection & conduction to atmosphere
  - Atmospheric absorption of 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

Solar Radiation Absorption in the Thermosphere

Absorbing gas is thin here; therefore little absorption.

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)**

Few photons left here; therefore little absorption.

≈ 120 - 160 km

Intensity of Incoming Solar Radiation

Q

F

N

HEIGHT (km)

Rate of Solar Energy Absorption

Exponential Decrease of Absorbing Atmospheric Constituent

RELATIVE MAGNITUDES
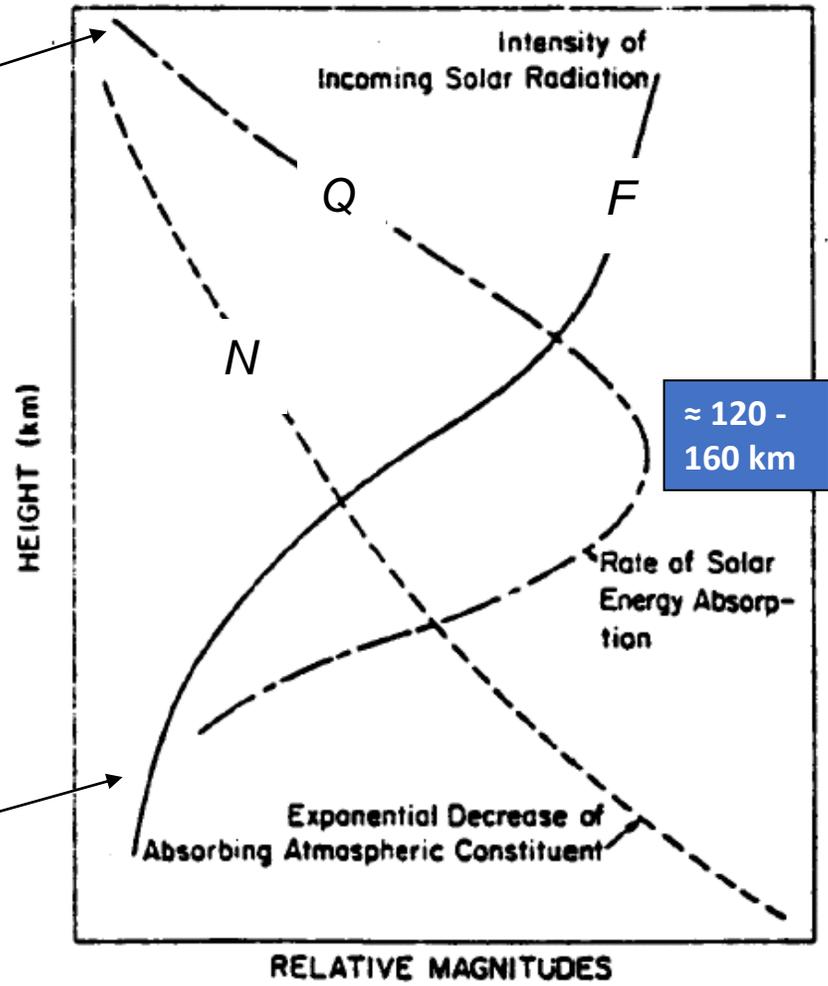
Hydrostatic equilibrium

n = # molecules per unit volume

m = mass of each particle

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

The force due to gravity g on the cylindrical mass is: $nmg\ dh$

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

$(P + dP) – P = -nmg\ dh$
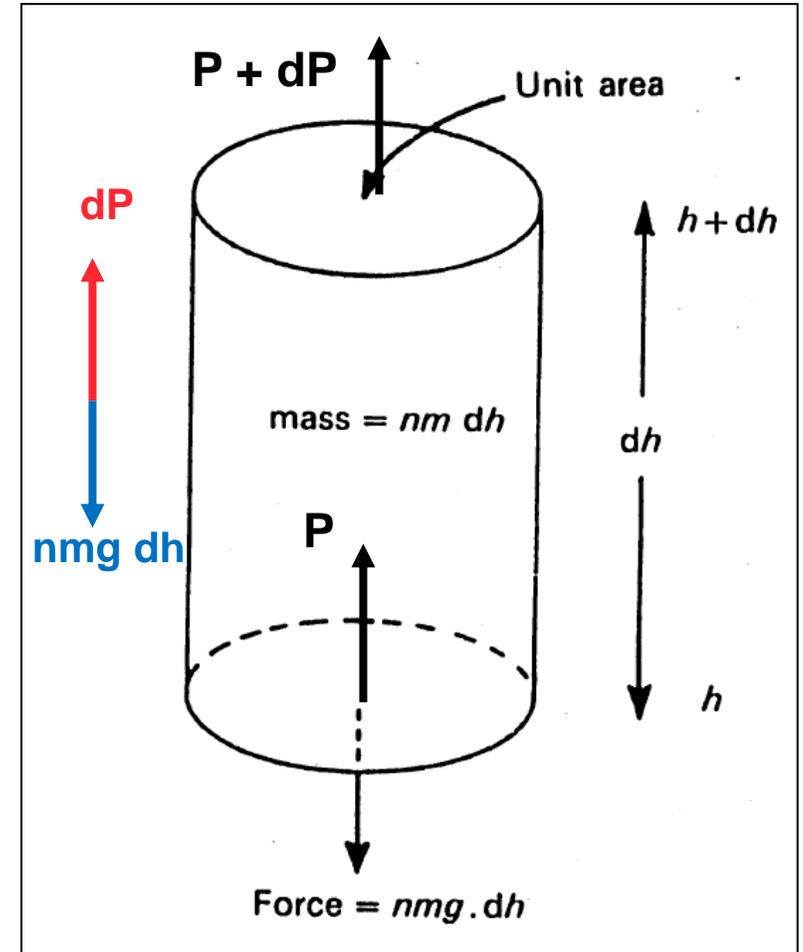
Hydrostatic equilibrium

Since *m* varies from constituent to constituent (i.e., H, He, O, O_2, N_2, ...), the equations apply to individual constituents *i*:

$$P = P_{io}\, e^{-h/H_i} \qquad n = n_{io}\, e^{-h/H_i} \qquad \rho = \rho_{io}\, e^{-h/H_i}$$

Derivation given at the end of the presentation in A(ppendix)1

where P_i is the partial pressure and
$$H_i = \frac{kT}{m_i g}$$
where *H* is called the scale height (general way to describe how a value fades away, it is the distance for density to drop by 1/e = 0.37)

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.

Hydrostatic equilibrium

The total density and composition as a function of altitude

Total density

Partial densities

Variation of density for an atmosphere with constant temperature of 750 K;
The distribution is a function of temperature, i.e. solar heating
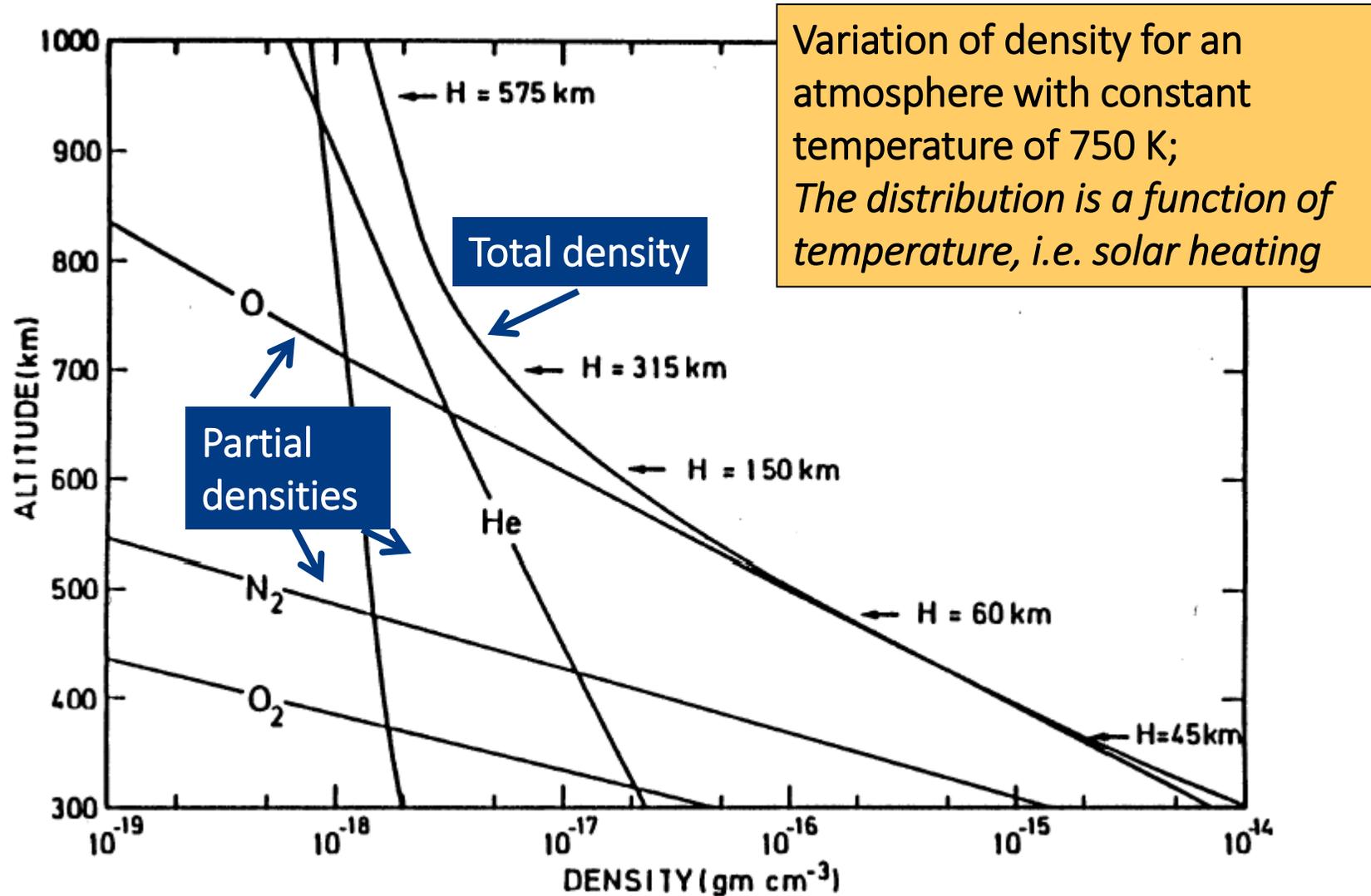
Hydrostatic equilibrium

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.

- 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, or turbopause. Above the homopause is the heterosphere; homosphere below.

Upper atmosphere heating

Solar radiation – EUV emissions

Correction due to eccentricity: ±3%

$$E = E_s \times (\overline{R} / r)^2 \times \cos \chi$$

Solar constant   Solar zenith angle

Earth rotation: ('day-night' effect):

Seasonal and latitudinal effects

Received energy at a location is variable:

Solar declination

equator

(22 December)
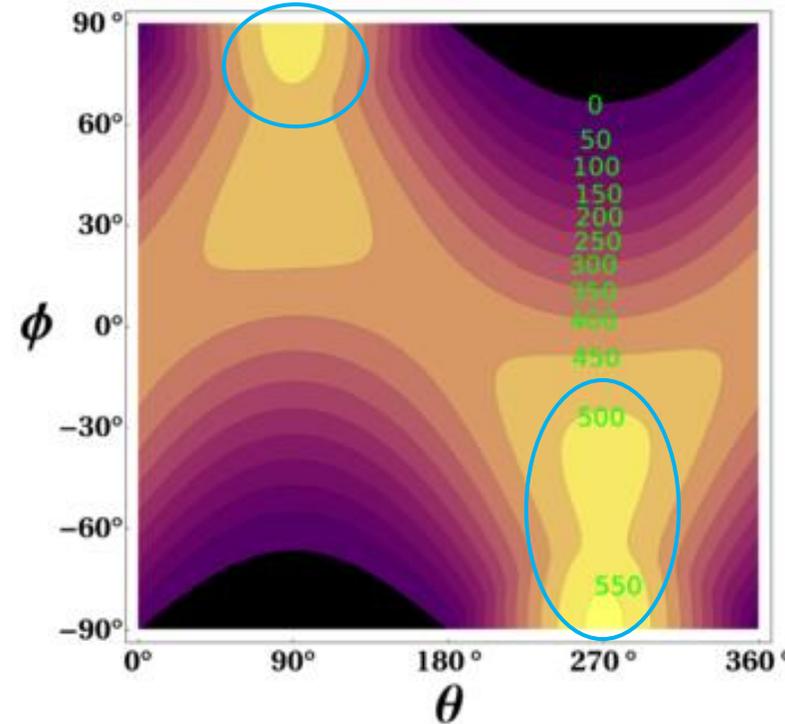
Energy deposition is unequal: *Perihelion is early January*

Upper atmosphere heating

Variability in solar UV/EUV emissions: the approximately 11 year solar cycle

81-day mean and daily F10.7 solar radio flux

Variability over a solar cycle as a function of wavelength
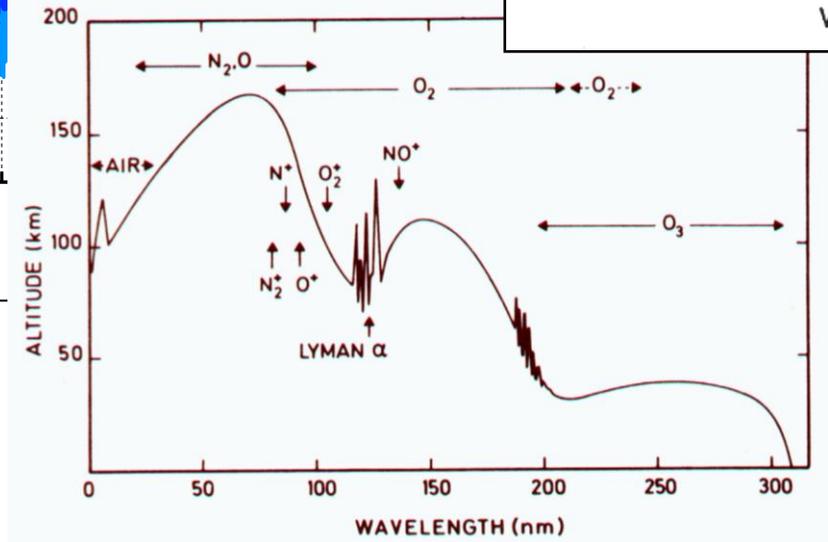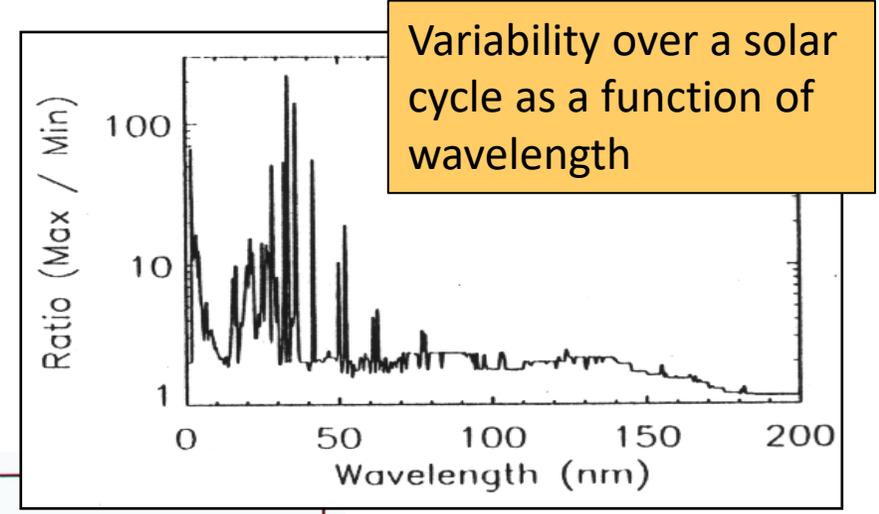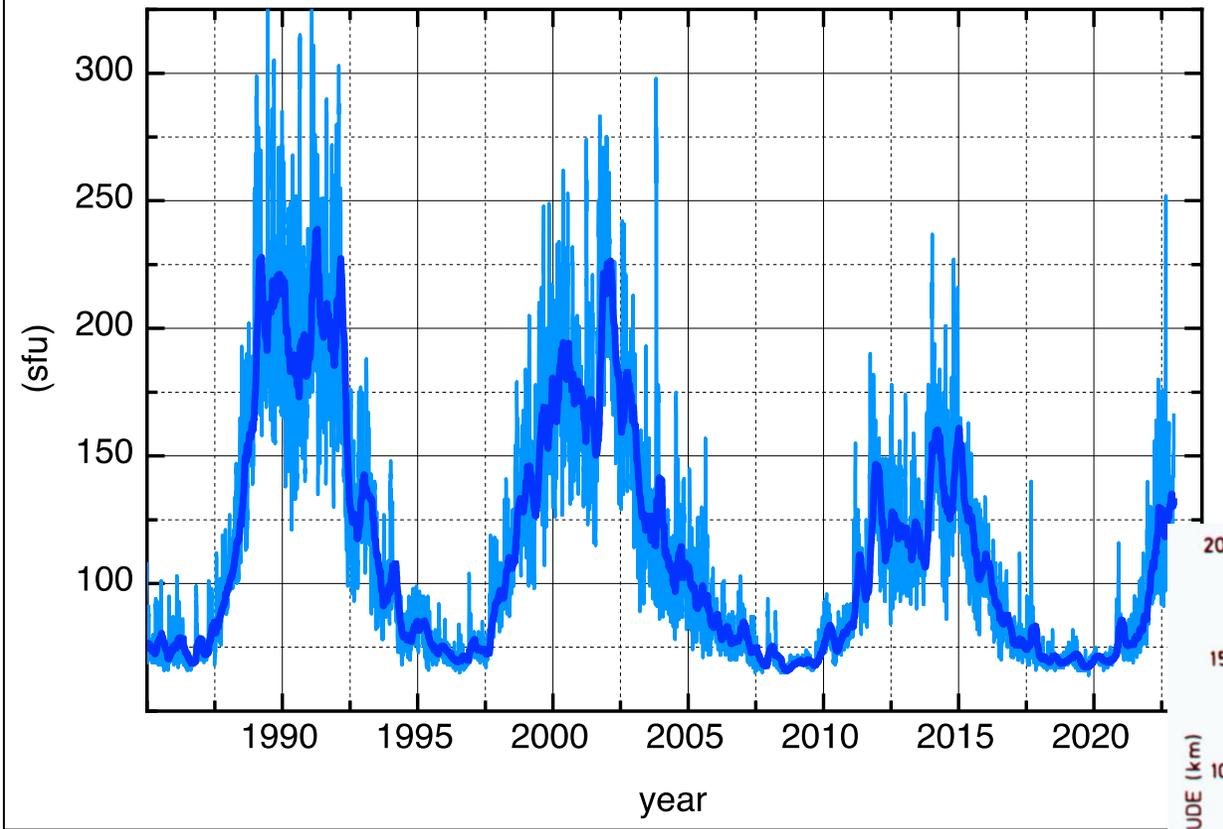
Altitude of maximum solar radiation absorption

Upper atmosphere heating

Variability in EUV (30 nm) over a solar cycle: examples

2009 (weak activity: cycle minimum)
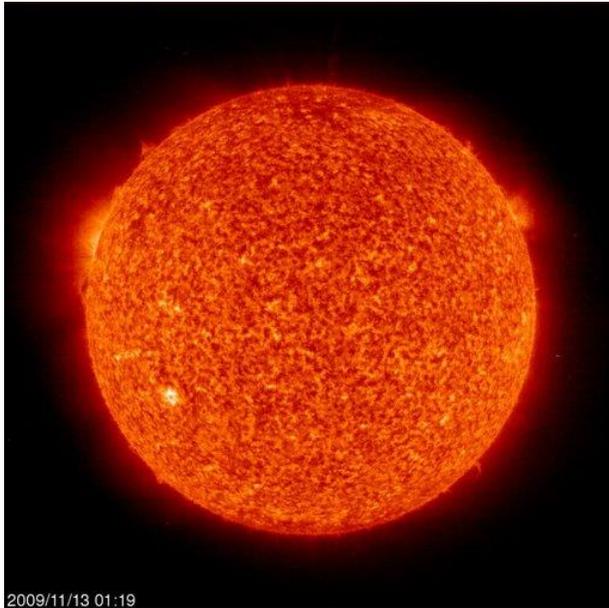
2001 (high activity: cycle maximum)
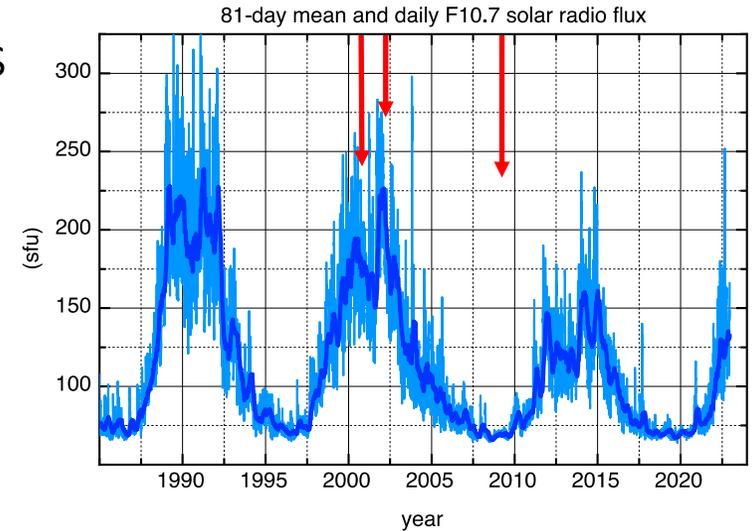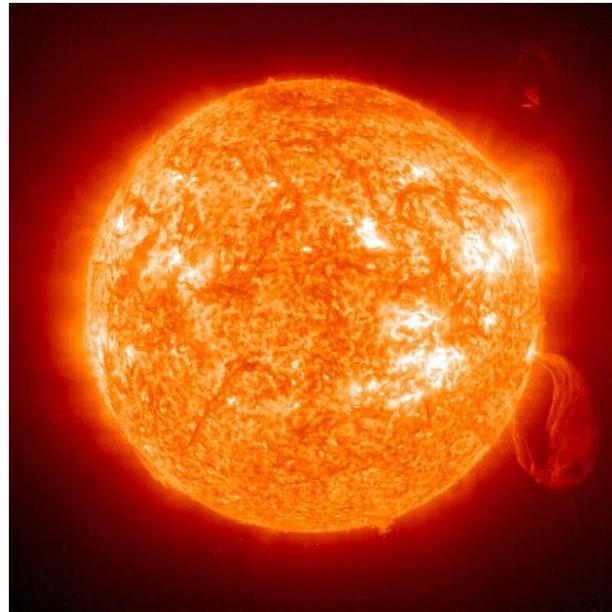
2002 (high activity+Coronal Mass (CME))

81-day mean and daily F10.7 solar radio flux

SOHO images : Courtesy ESA/NASA
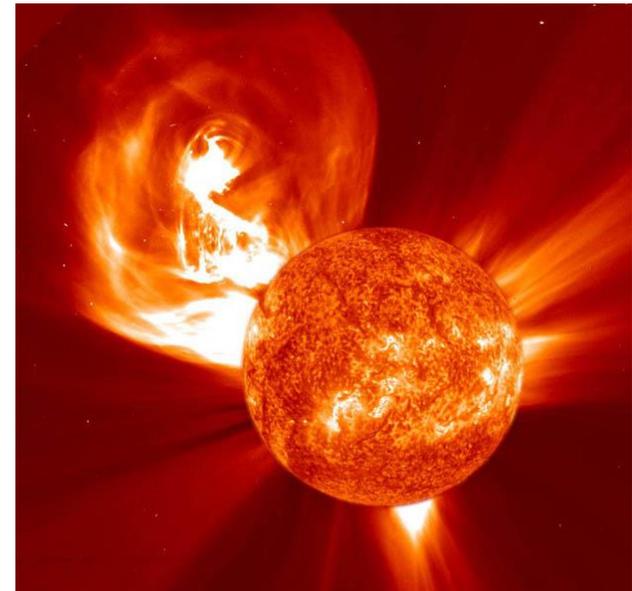
Upper atmosphere heating

Solar wind *particles* – Magnetosphere coupling, Joule heating

Coupling between interplanetary medium and the magnetosphere depends on:

- Orientation of the interplanetary magnetic field (Bz in particular)

- Density and speed of the solar wind

Upper atmosphere heating is due to:
'Solar activity'   = EUV
'Geomagnetic activity' = Solar wind
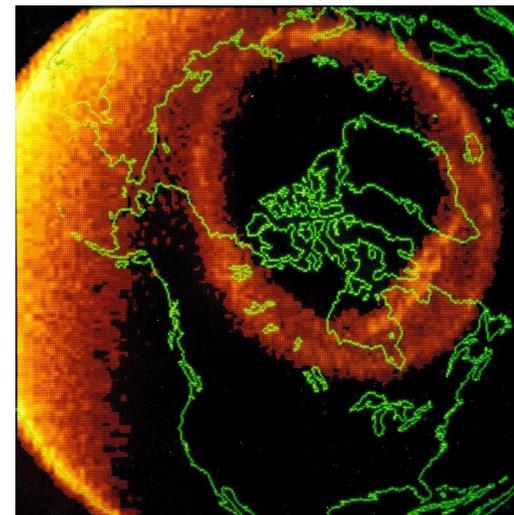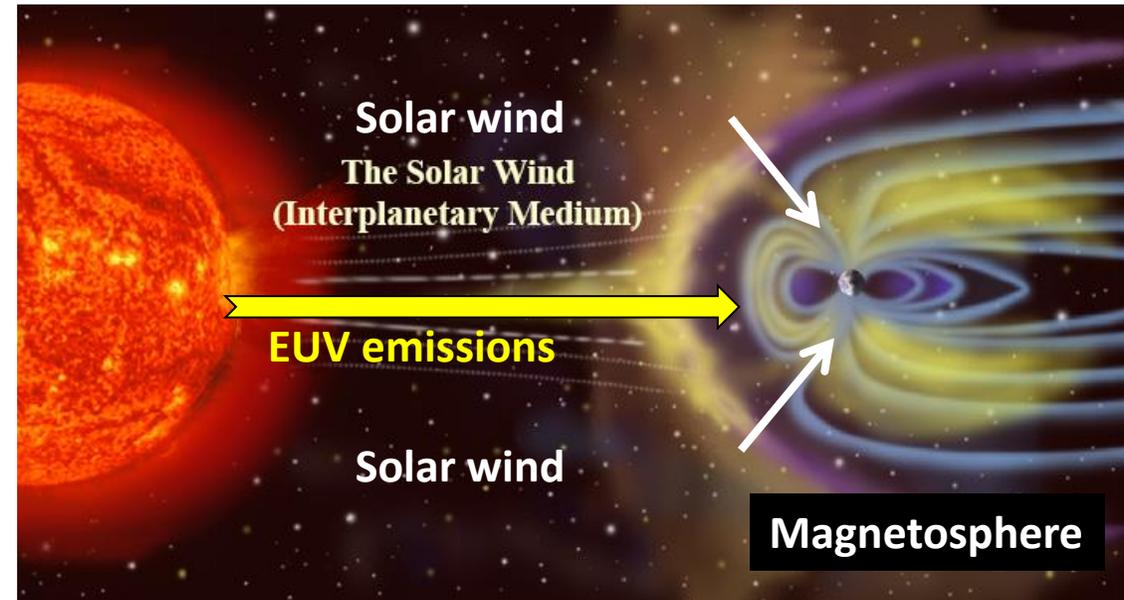
Ergo:
Aurora visible at high latitude

Upper atmosphere heating

Total heating = solar + geomagnetic activity

Daily Average Power Values for Solar Cycles 21-23

Courtesy: D. Knipp
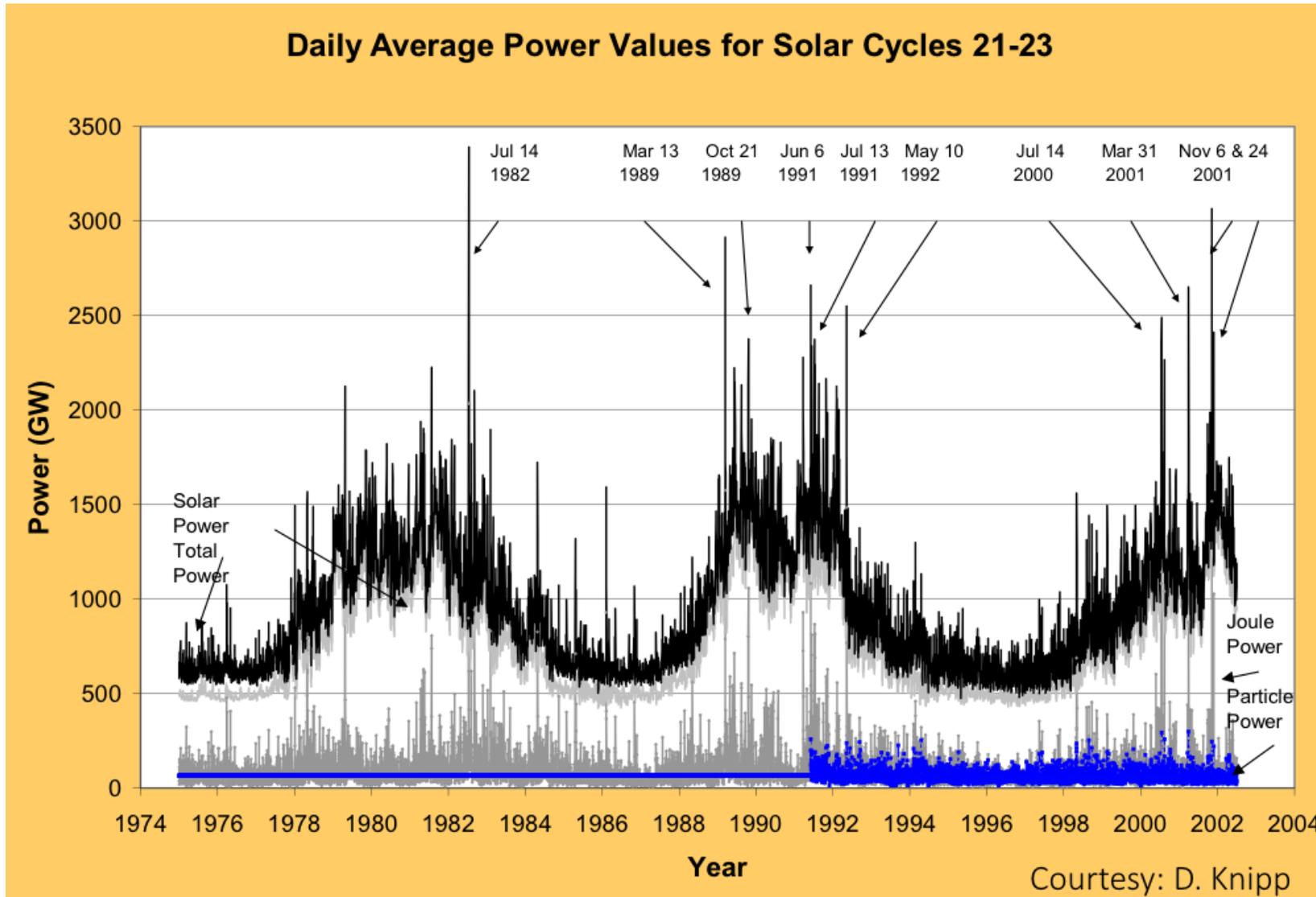
Thermosphere models

1 – **Physical**, or first principles models: three-dimensional fluid equations are solved with a time-step of typically 1 minute

Example: TIE-GCM (Thermosphere-Ionosphere-Electrodynamics General Circulation Model) - a global first-principles model with 2.5^o latitude by longitude grid and 29 (or more) pressure level from 97km to ~600km

(and: WAM, WACCM-X, GITM, Gaia, CTIPe)

Inputs to TIE-GCM:
• Convection and auroral precipitation patterns from *AMIE* (Assimilative Mapping of Ionospheric Electrodynamics), Weimer, Heelis
• Solar EUV fluxes (NB: lower-resolution spectrum scaled by F10.7)
• Semidiurnal and diurnal tides from the *GSWM* (Global Scale Wave Model)

advantage: realistic physics, variability qualitatively correct
disadvantage: complex and costly calculation, for expert users, often no Helium
uncertainty: ?

Thermosphere models

2 - **Tabular** (interpolation)

Example: MCM (Mowa Climatological Model) - model of the whole atmosphere by means of blending averaged data tables of the Unified Model (UM) from the Met Office for the atmosphere (0 to 120 km) and DTM2020 (120-1500 km)*

(and: GRAM, MarsGRAM, MCD & VCD database)*

Typical input:
• date, position
• solar flux, solar activity conditions, geomagnetic activity index (e.g. Kp)

advantage :   fast and easy calculation, simple algorithm
disadvantage :   truncated resolution, interpolations errors, limited by number of scenarios
uncertainty: ?

* Combination of tabular + empirical

Thermosphere models

3 - **Empirical**

Example: DTM (drag Temperature Model) - thermosphere model from 120-1500 km reproducing a limited number of known variations by fitting to observations, using a simple and fast algorithm.

(and: Jacchia, NRLMSIS, JB2008, Stewart Mars model, Hedin Venus model)

input:
• date, position
• solar activity proxy or proxies (e.g. F10.7), geomagnetic activity index (e.g. Kp)

advantage :   fast and easy calculation, simple algorithm, robust
disadvantage :   low resolution, simple algorithm, highly dependent on quality of data
uncertainty: 8-25% (1σ, for Earth)

NB: despite its weaknesses, this kind of model is used in orbit computation

Thermosphere models

1, 3 – plus data assimilation

Example empirical model: HASDM

Example physical models: WAM, Gaia, Aeneas

HASDM input:
- date, position
- solar activity proxy or proxies (F10.7), geomagnetic activity index (ap, Dst)

advantage: idem physical/empirical models + high accuracy

disadvantage: idem physical/empirical models + data management and quality control

uncertainty: 3-10% for HASDM; physical models ?

NB: (no) data availability is a show stopper

Thermosphere models

1, 3 – plus data assimilation

Example empirical model: HASDM - comparison with GOCE densities at 255 km

Daily-mean densities

GOCE
HASDM: mean O/C = 1.00 / StD = 3 %
JB2008: mean O/C = 1.12 / StD = 10%

DA: higher accuracy

18
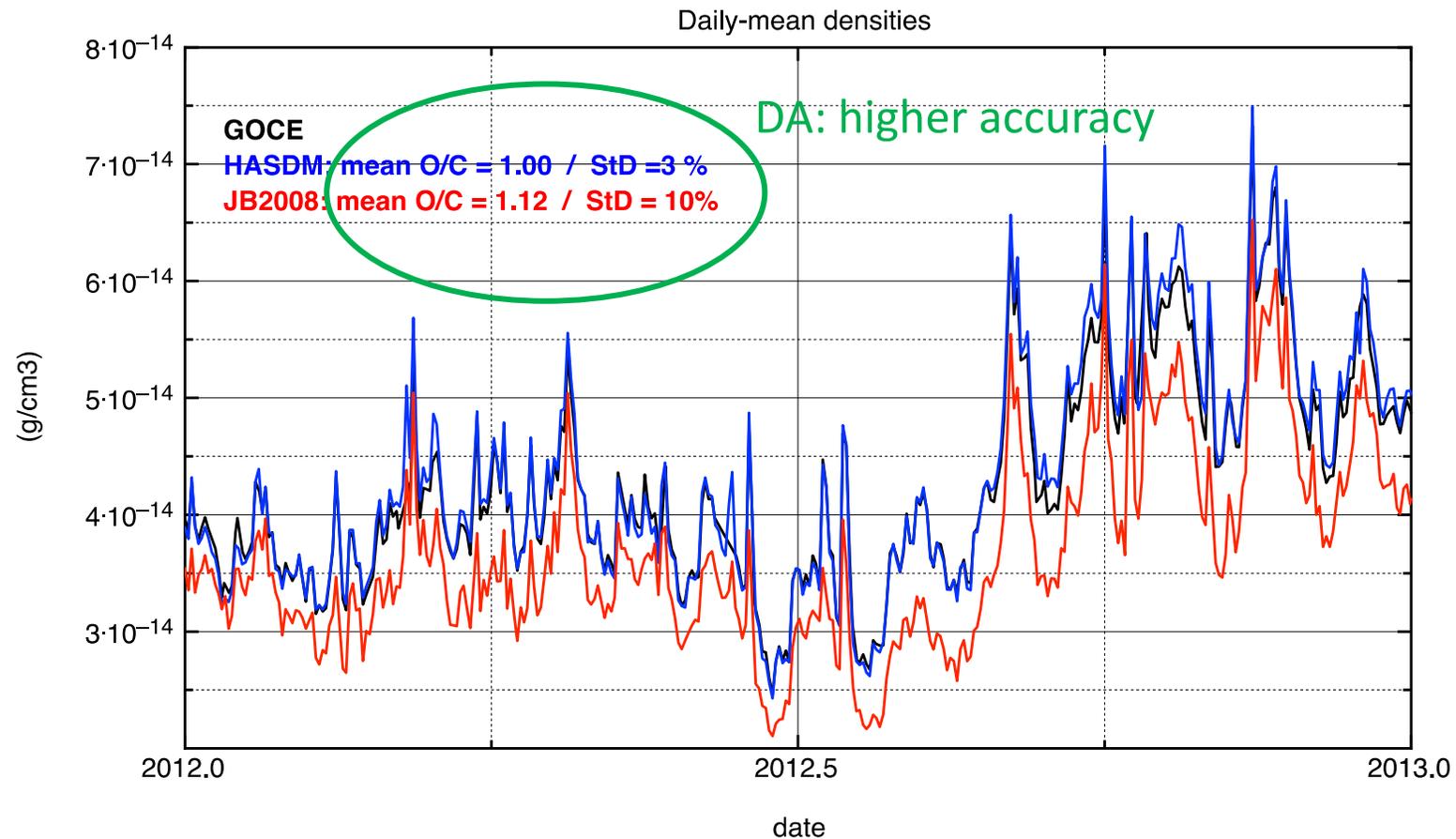
Thermosphere density: variability

Thermosphere density is function of location:

- Altitude
- Latitude, longitude
- Local solar time
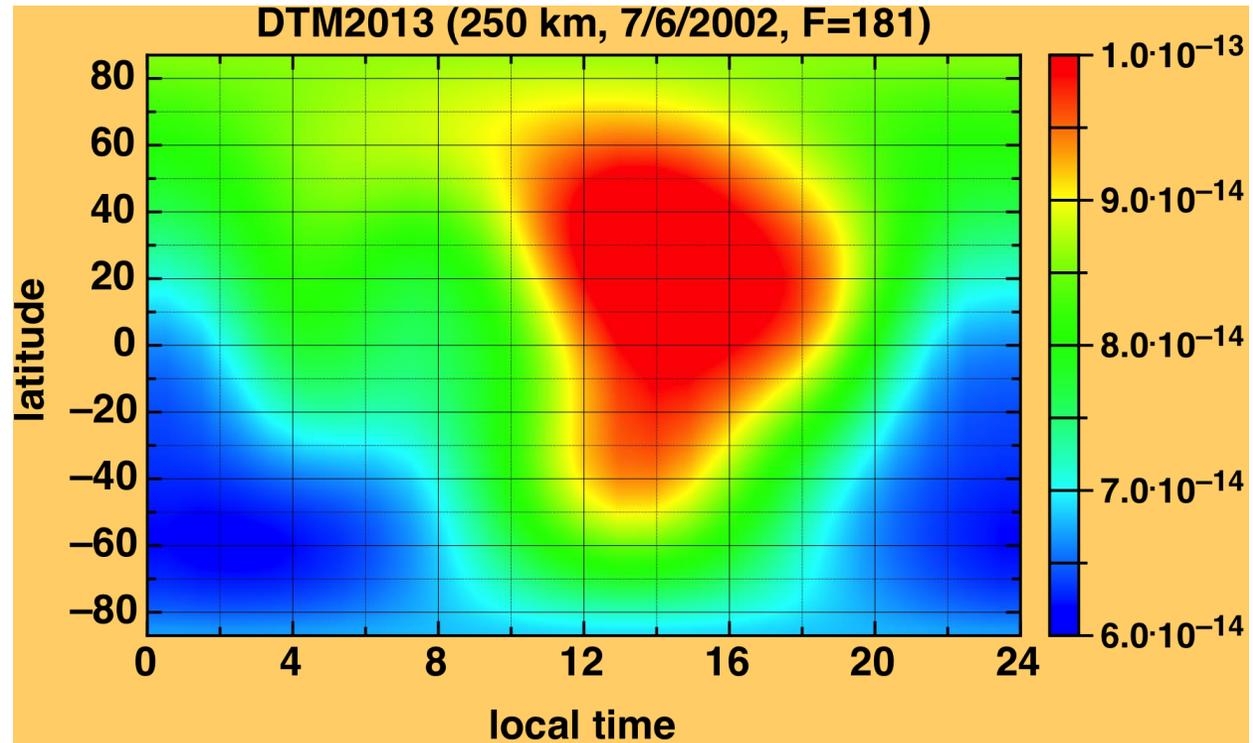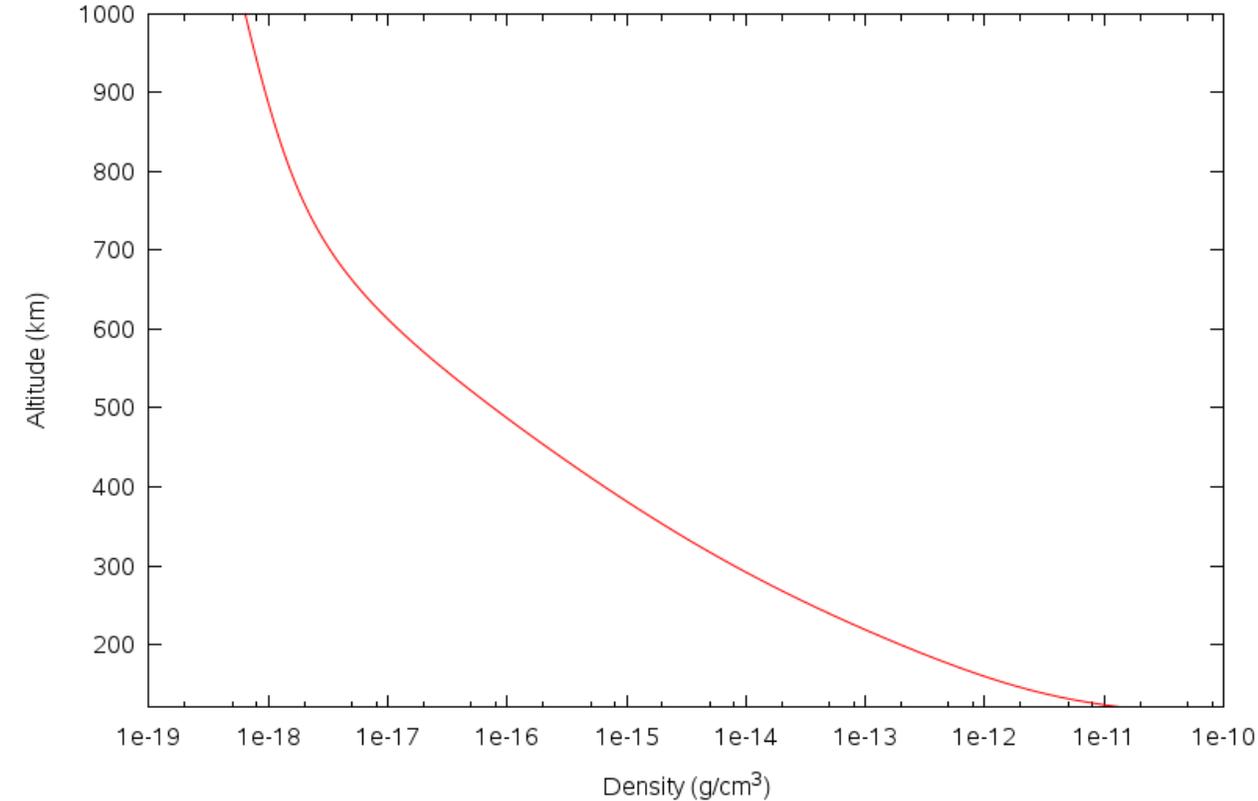
DTM2013 (250 km, 7/6/2002, F=181)

Thermosphere density: variability

Thermosphere density is function of:

- Season

Earth — Low density of incident rays (northern winter)

Night — Equator — Day

High density of incident rays (southern summer)

Sun

NRLMSISE-00: 500 km, F10.7=150, Ap=4 - NH Summer (18 July)

NRLMSISE-00: 500 km, F10.7=150, Ap=4 - SH Summer (1 January)

Temperature (K)

Local solar time

Densities SWARM-A: 22 July 2020

Solar local time: 12h

(g/cm3)

Densities SWARM-A: 1 January 2020

Solar local time: 18h

(g/cm3)
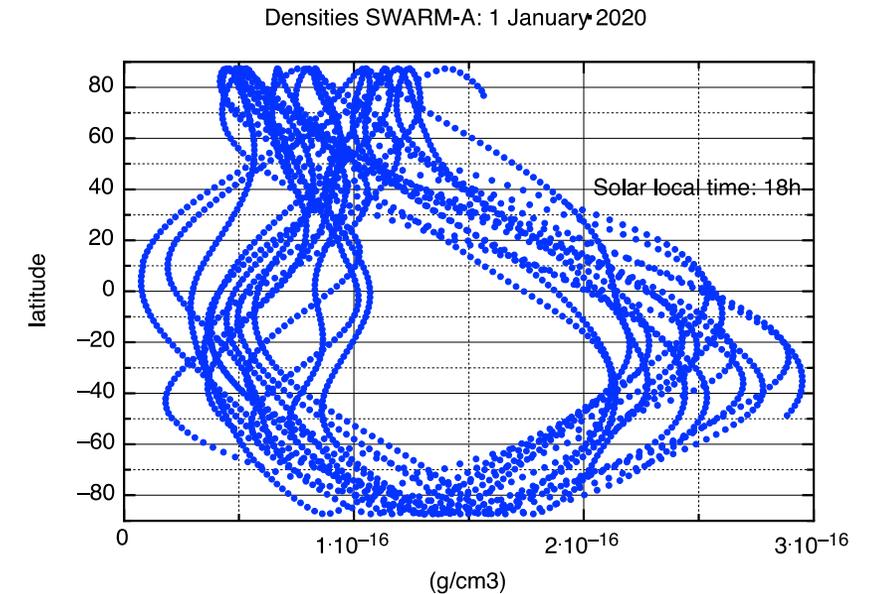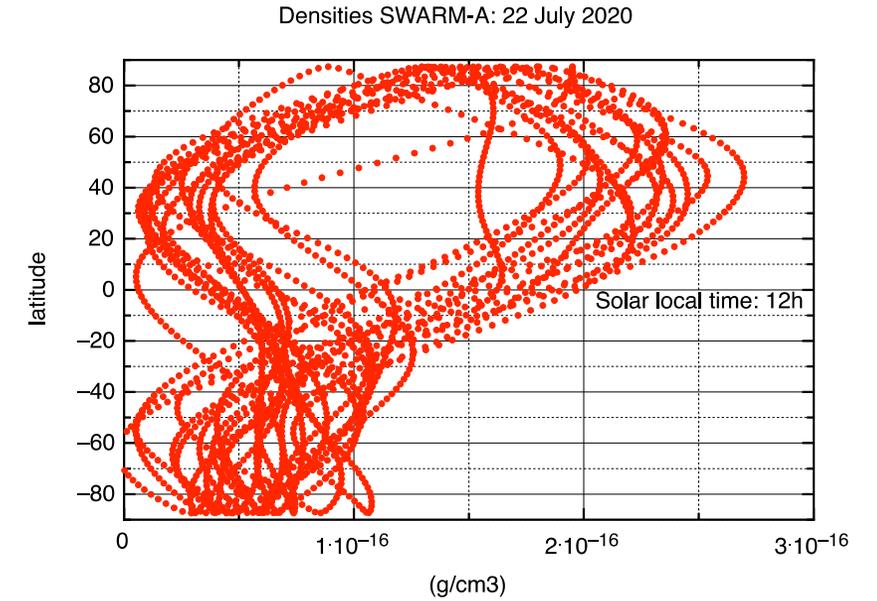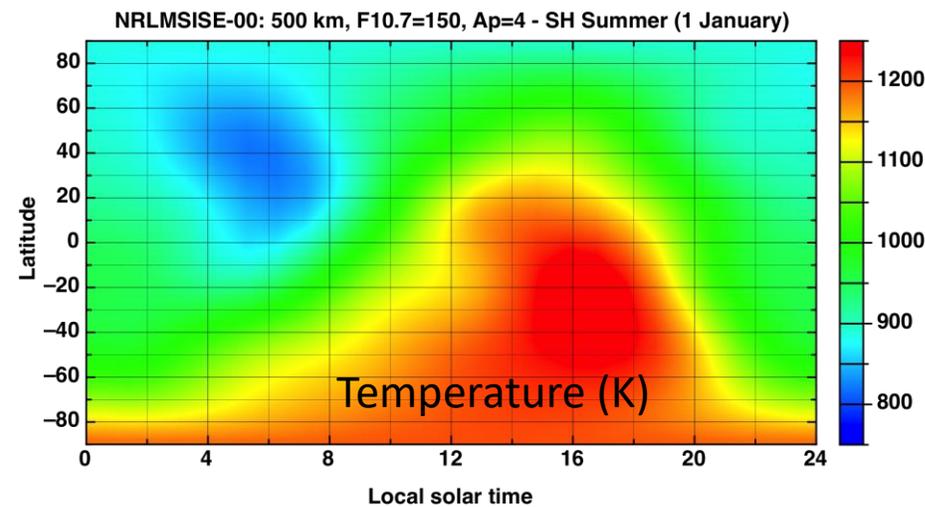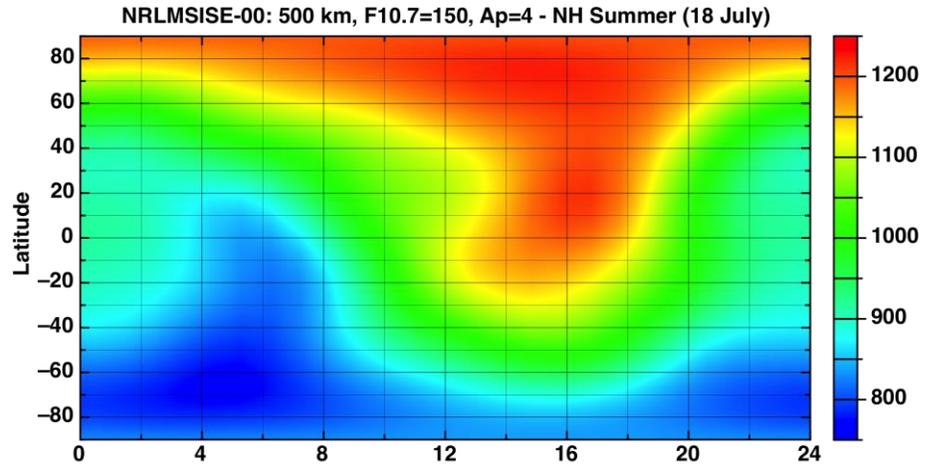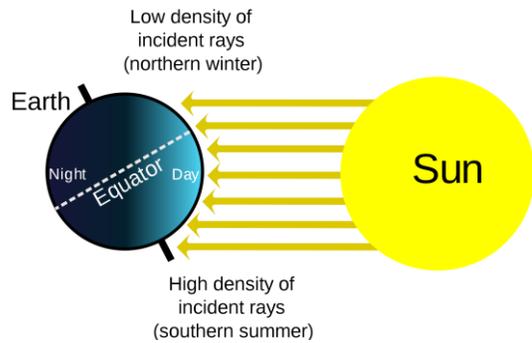
Thermosphere density: variability

Thermosphere density is function of:

- Solar and geomagnetic activity > *proxies*

(measurement that mimics variations of another observable)

81-day mean and daily F10.7 solar radio flux
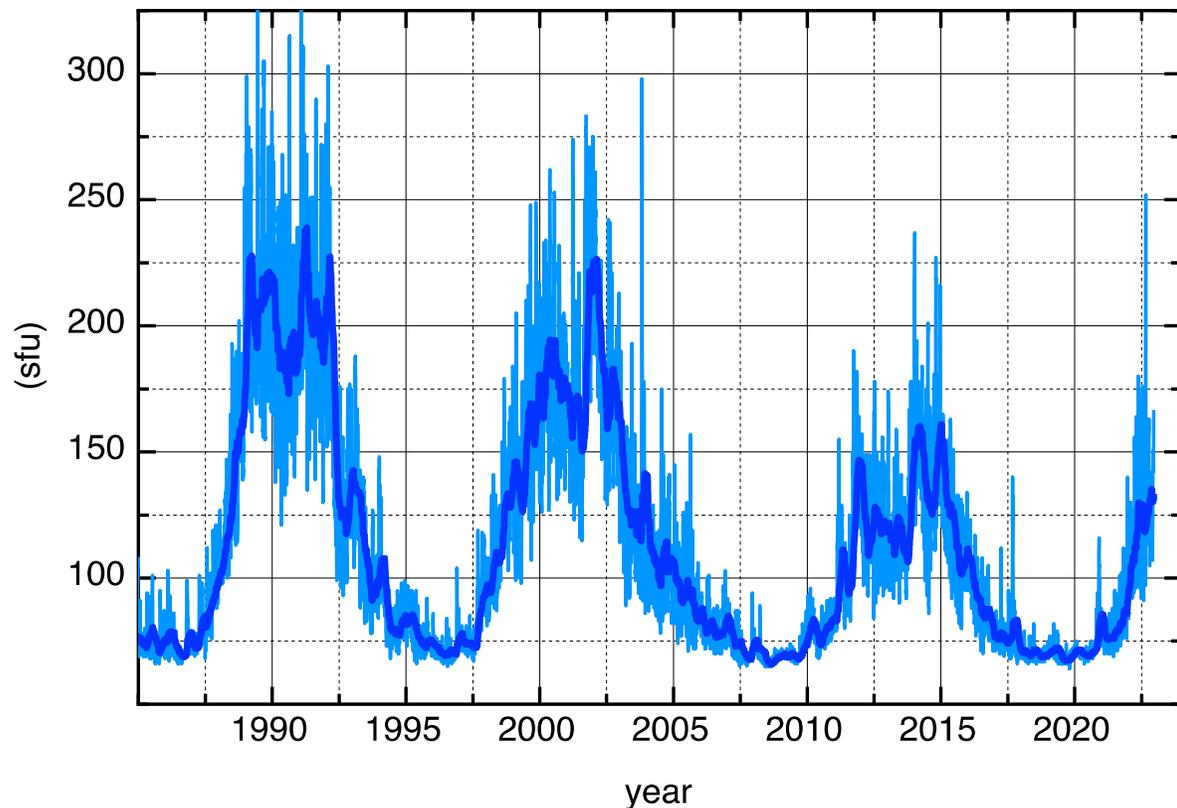
Proxy for Joule heating by solar wind (particles): daily mean Kp for 2003
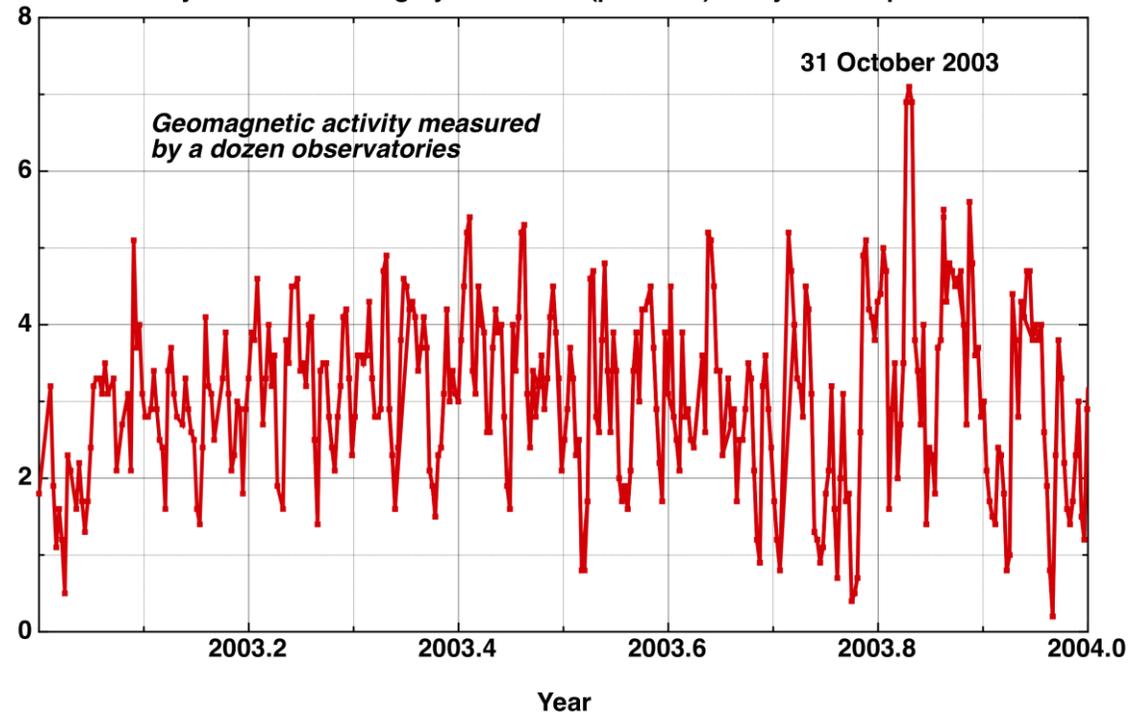
Thermosphere density: variability

Maximum versus minimum density as a function of altitude (*model*):

Thermosphere density: variability

Slow and fast temporal variations:
- Solar cycle (≈11 years)
- Season (6 months & 12 months)
- Active regions (months)
- Solar rotation (≈27 days)
- Corotating Interaction Regions (9 & 13.5 days)
- Day to day variations
- Solar/geomagnetic storms (hours – days)
- Solar flares (hours)

Daily-mean densities CHAMP (red) and GRACE (blue) vs 81-day mean F10.7 (yellow)

Decaying orbit altitude

Same orbit altitude

Thermosphere density: variability

Slow and fast temporal variations:
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Daily-mean density Stella

Thermosphere density: variability

Slow and fast temporal variations:

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Daily-mean densities CHAMP (red) and GRACE (blue) vs mean (yellow dotted) and daily F10.7

Thermosphere density: variability

Slow and fast temporal variations:
- Solar cycle (≈11 years)
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- Solar flares (hours)

(date)	(8 3-hourly Kp)								(8 3-hourly ap)								(F10.7)
2023 05 01	2.667	1.333	2.333	1.333	1.333	2.000	2.333	3.333	12	5	9	5	5	7	9	18 9	147.9
2023 05 02	3.000	3.000	2.000	2.667	1.000	0.667	1.333	1.000	15	15	7	12	4	3	5	4 8	156.8
2023 05 03	0.000	0.000	0.333	0.333	0.667	0.667	0.667	1.667	0	0	2	2	3	3	3	6 2	156.2

➢ Due to daily changes in **geomagnetic** and **solar** activity

Daily-mean densities GOCE

(g/cm3) vs date
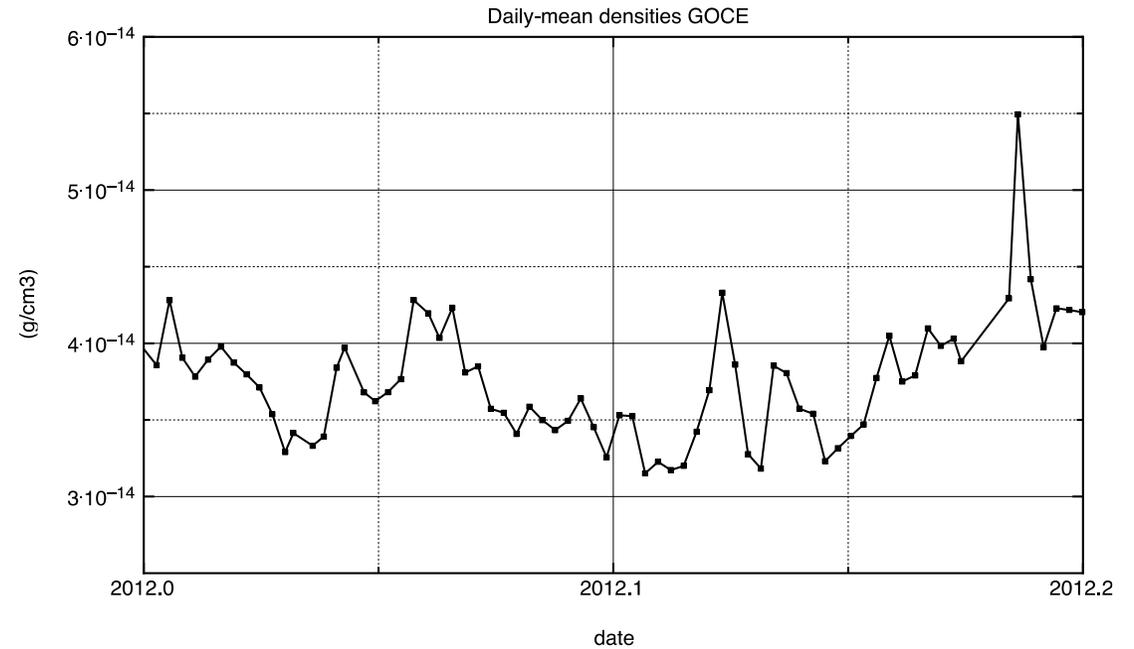
➢ *And due to a superposition of tidal components*

Explained at the end of the presentation in A(ppendix)2

Thermosphere density: variability

Slow and fast temporal variations:
- Solar cycle (≈11 years)
- Season (6 months & 12 months)
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- Day to day variations
- Solar/geomagnetic storms (hours – days)
- Solar flares (hours)

Geomagnetic storm: 19-21 November 2003

Global increase in density + large-scale waves
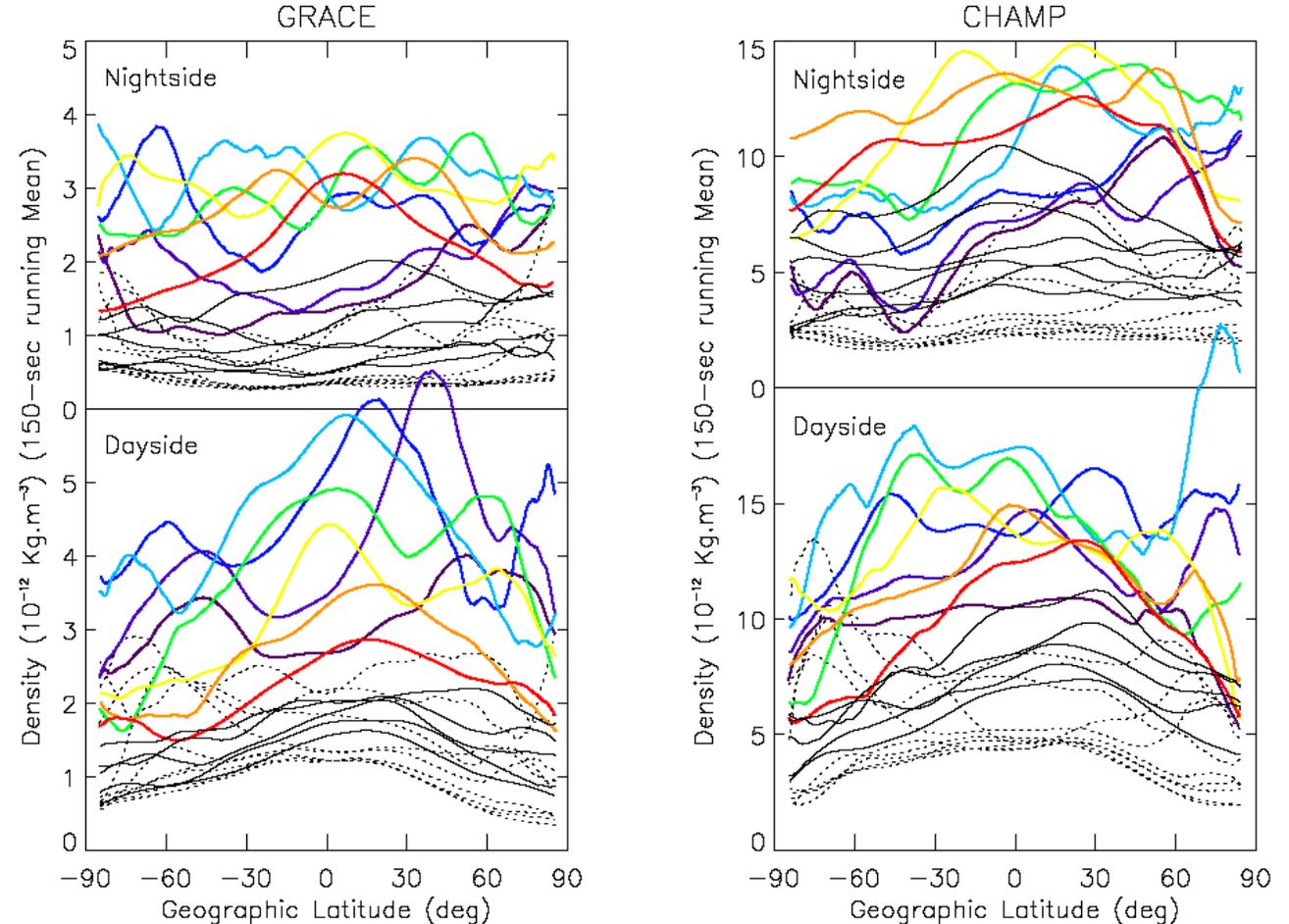
(150-sec (~1200 km) running means applied to raw data)

Thermosphere density: variability

Residual = observation − trend

Relative variation = residual / trend

We can localize and quantify the wave-like perturbations by computing relative density variations

First, the trend

By detrending the data, we extract specific scales.

CHAMP night side density profiles: observations (black), 15-point trend (red), 59-point trend (blue)

59-point trend: blue

CHAMP night side relative variations profiles: 15p-residuals (black), 59p-residuals (red), 15p-59p residuals (blue)

59-point detrended CHAMP residuals: 160-2400 km scales

Relative density variation

Thermosphere density: variability

Density variability:
CHAMP, Kp < 3

Thermosphere density: variability

Relvar_LatvsSLT_160–2400km_kpmin4_0510_RMS.dat

Density variability:
CHAMP, Kp > 4+

CHAMP 2005-2010

Mean altitude:
340 km

485 km

GRACE 2005-2010

Density variability:
GRACE, Kp > 4+

Relvar_LatvsSLT_80–2400km_kpmin4_0510_RMS.dat

Thermosphere density: variability

Slow and fast temporal variations:
- Solar cycle (≈11 years)
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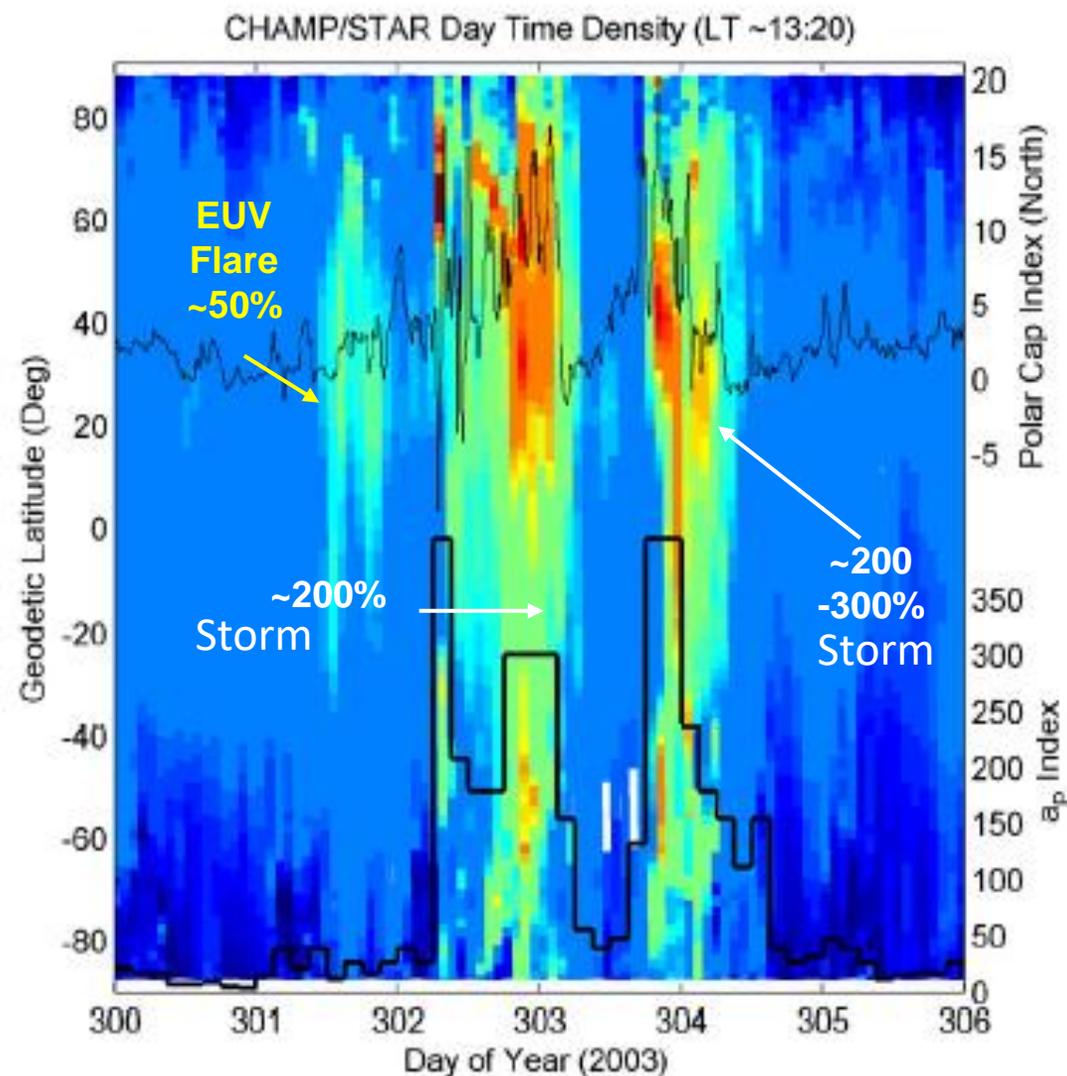
* Thermosphere Density Response to the October 29-31 2003 Storms from CHAMP Accelerometer Measurements (*Sutton et al., JGR, 2005*)

Summary of basic thermosphere characteristics

• Solar heating is the main energy source

• Main cooling is through molecular conduction; secondary, radiative cooling by O, CO_2, NO

• Main gases: O, O_2, N_2, He (high altitudes only)

• Strong variability of temperature, winds and composition with solar cycle, season, local time, geomagnetic activity

• 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 most important (but still rather small)

• At high latitudes, heating from the magnetosphere occurs in the form of Joule heating and precipitating particles (*NB: can be equal to solar heating for severe storms*)

Thank you for your attention

Questions?

Hydrostatic equilibrium

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

g = gravity: 9.81 m s^{-2} (at the surface)
k = Boltzmann constant: 1.38 J K^{-1} mol^{-1}
R = Molar gas constant: 8.314 J K^{-1}

Assuming the ideal gas law holds:

$$\boxed{P = nkT}$$

Then the previous expression may be written:

$$\boxed{\frac{1}{P}\frac{dP}{dh} = -\frac{1}{H}}$$ *integrating* →

where *H* is called the scale height
(general way to describe how a value fades away,
it is the distance for density to drop by 1/e = 0.37)

$$\boxed{H = \frac{kT}{mg} = \frac{RT}{g}}$$

A1

Hydrostatic equilibrium

Leads to the so-called hydrostatic law or barometric law:

$$P = P_0 e^{-z}$$

$$z = \int_0^h \frac{dh}{H}$$

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

Similarly,

$$n = n_o \left(\frac{T_O}{T}\right) e^{-z}$$

For an isothermal atmosphere, then,

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

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

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

A1

Thermosphere density: variability

Slow and fast temporal variations:
- Solar cycle (≈11 years)
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➢ Due to a superposition of tidal components
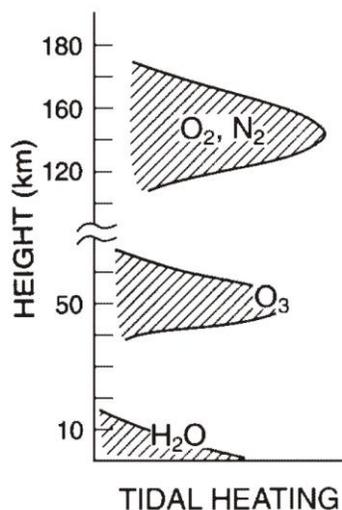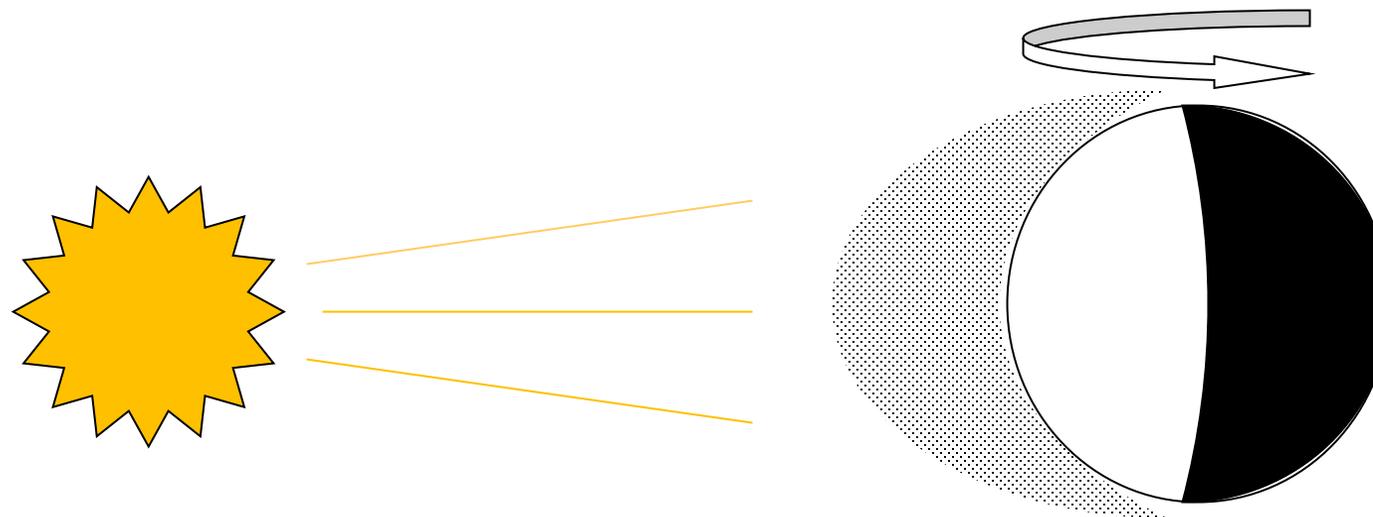
Solar thermal tides are excited in a planetary atmosphere through the periodic (local time, longitude) absorption of solar radiation.

In general, tides are capable of propagating vertically to higher, less dense, regions of the atmosphere; the oscillations grow exponentially with height.

The tides are dissipated by molecular diffusion above 100 km, their exponential growth with height ceases, and they deposit mean momentum and energy into the thermosphere.

HEIGHT (km)

180

160

O_2, N_2

120

50

O_3

10

H_2O

TIDAL HEATING

A2

Thermosphere density: variability

- Solar cycle (≈11 years)
- Season (6 months & 12 months)
- Active regions (months)
- Solar rotation (≈27 days)
- Corotating Interaction Regions (9 & 13.5 days)
- Day to day variations
- Solar/geomagnetic storms (hours – days)
- Solar flares (hours)

➢ **Due to a superposition of tidal components**

To be complete:
 if the excitation also depends on longitude, the spectrum of tides that is produced is more generally expressed as a linear superposition of waves of various frequencies (n) and zonal wavenumbers (s):

$$\sum_{s=-k}^{s=+k} \sum_{n=1}^{N} A_n \cos(n\Omega t + s\lambda - \phi)$$

λ= longitude

Ω= 2π/24
(rotating planet)

The waves with **s = n** are referred to as migrating tides because they migrate with respect to the Sun to a planetary-fixed observer. (*n=1: diurnal / n=2: semidiurnal*)

The waves with **s ≠ n** are referred to as non-migrating tides because they do not migrate with respect to the Sun to a planetary-fixed observer
They can migrate *westward* (slower or faster than the Sun), *eastward*, or they can be *standing waves* for s=0.

NB: non-migrating tides are very small in the thermosphere above 200 km, a few percent

A2

References

Empirical models

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