The atmosphere: A general introduction

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Facts about the atmosphere

- The atmosphere is 'kept in place' on Earth by gravity

- The Earth-Atmosphere system is 'spinning' around the rotation axis of the Earth and is simultaneously moving in an eliptical orbit around the Sun with a period of approx. 365 days

- The 'spinnig' of the Earth and the associated variation in incoming solar radiation gives rise to **diurnal** variations in the state of the atmosphere (and its underlying surface)

- The movement of the Earth around the Sun gives rise to **seasonal** variation in the state of the atmosphere (and its underlying surface), since the rotation axis of the Earth tilts with respect to the orbital plane

- The angular velocity of the Earth is

$$\Omega = \frac{2\pi \cdot (n_d + 1)}{n_d \cdot 24 \cdot 60^2} \, rad \, sek^{-1} \approx 7.29 {\cdot} 10^{-5} rad \, sek^{-1} \tag{1}$$

  for n_d=365 days

- In the past variations in orbital parameters (including tilt of the Earth's axis of rotation with respect to the orbital plane) have given rise to alternating **glacial** and **interglacial** climates on time scales > 10000 years

- **Composition of the atmosphere**
- **Trace gases**
- **Layering of and vertical temperature profile in the atmosphere**

Gas	Symbol	Concentration (ppmv)	Variation in time
Nitrogen	N_2	780 840	Constant
Oxygen	O_2	209 476	Constant
Argon	Ar	9 340	Increasing
Carbondioxid	CO_2	357	Increasing
Neon	Ne	18,18	Constant
Helium	He	5,24	Constant
Methan	CH_3	1,70	Increasing
Krypton	Kr	1,14	Constant
Hydrogen	H_2	0,56	Constant
Nitrous oxide	N_2O	0,31	Increasing
Carbon monoxide	CO	0,15	Variable
Xenon	Xe	0,087	Constant
Ozon	O_3	0,020	Variable
Nitrogen oxid	NO	0,0005	Variable
Nitrogen dioxid	NO_2	0,0005	Variable
Sulfur dioxide	SO_2	0,0004	Variable
Freon 12	CCL_2F_2	0,0003	Increasing
Freon 11	CCL_3F	0,0002	Increasing
Ammonia	NH_3	0,0002	Variable

Constituents of dry air in the Earth's atmosphere

Trace gas	Mixing ratio	Distribution	Sources	Sinks
Methane, CH_4	1.7 ppmv rising	Uniform (8 yr)	Farming, cumbust.	OH, soil, stratosph.
Hydrogen,	0.5 ppmv	Uniform (2 yr)	Oxidation cumbust.	OH, soil uptake
Carbon, CO Monoxid	0.2 ppmv .05 ppmv	NH SH	Oxidation cumbust.	OH, soil, stratosph.
Ozone, O_3	15-50 ppbv	Low to high Eq. to Pole	Stratosph. photoche.	Dry dep. photoche.
Nitrous oxide, N_2O	0.3 ppmv rising	Uniform (110 yr)	Soil, oce. Antropog.	Stratosph.
Nitrogen oxides, NO, NO_2	30 pptv 300 pptv 5 ppb	Marine air Cont. air Cont. rual	Cumbust. Soils lightening	Oxidation to HNO_3 and O_3
Amonia, NH_3	100 pptv 5 ppbv	Marine air Cont. air	Farming, oceans	Dry/wett deposit. NH_4 aerosols
Carbonyl Sulfide, OCS	500 pptv	Uniform residence time(7 yr)	Soil, oce., Oxid. of CS_2,DMS	Vegeta-tion, Stratosph.
Hydrogen sulfide, H_2S	5-90 pptv	Marine air Cont. rual Cont. urban	Soil, vegeta., volcanoes	Reaction with OH
Dimethyl sulfide, CH_3SCH_3	5-70 pptv (DMS)	Marine air Cont. rual Cont. urban	Ocean, soils, vegeta.	Reaction OH, NO_3

Trace gases in the Earth's atmosphere

Trace gas	Mixing ratio	Distribution	Sources $(Tg\,yr^{-1})$	Sinks $(Tg\,yr^{-1})$
Sulfur dioxide, SO_2	20-90 pptv 5 ppbv	Marine air Contin. air	Fossil fuel Volcanoes Oxidation of sulfides	Dry/wett deposition Oxidation Aerosoles
Isoprene, C_3H_8	0.6-2.5 ppbv	Contin. Surf. air	Emission decedious trees	Reaction with OH
Terpenes, $C_{10}H_{16}$	0.03-2ppbv	Contin. Surf. air	Emmision coniferous trees	Reaction with OH

Trace gases in the Earth's atmosphere - continued
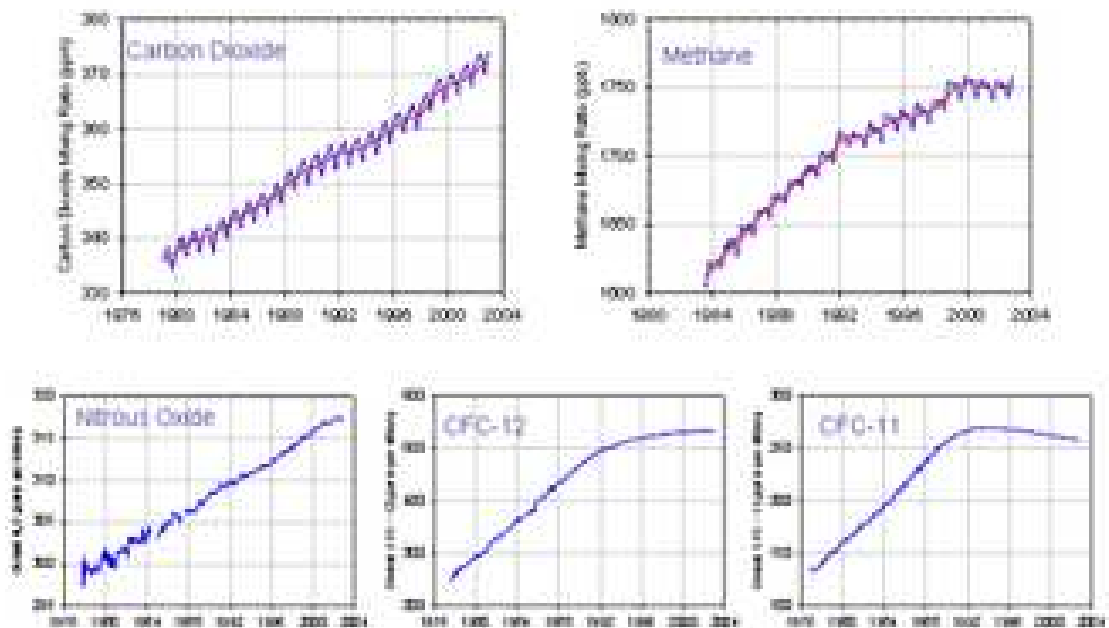
Global trends in CO$_2$, CH$_4$, N$_2$ O, CFC-12 and CFC-11 in recent years until January 2003 (time inerval: 4 years)

Ozone depletion in the stratosphere

Due to photo-chemical reactions on the surface of polar stratospheric clouds (either water, ice of nitric acid depending on temperature).

Formation of polar stratospheric clouds occurs at very low temperatures (¡ 80 degree Celsius).

These cold temperatures occur every winter within the antarctic strotospheric polar vortex and is a result of of radiative cooling of relatively undisturbed air inside the vortex.

Less regular and smaller cold spots also occur in the polar winter stratosphere.

Chemical reactions in the ozone depletion process requires short wave solar radiation.

Therefore the ozone hole is a spring phenomenon. In this period both solar radiation and cold temperatures are present.

Antropogenically created dhemical compounds are active in the depletion.
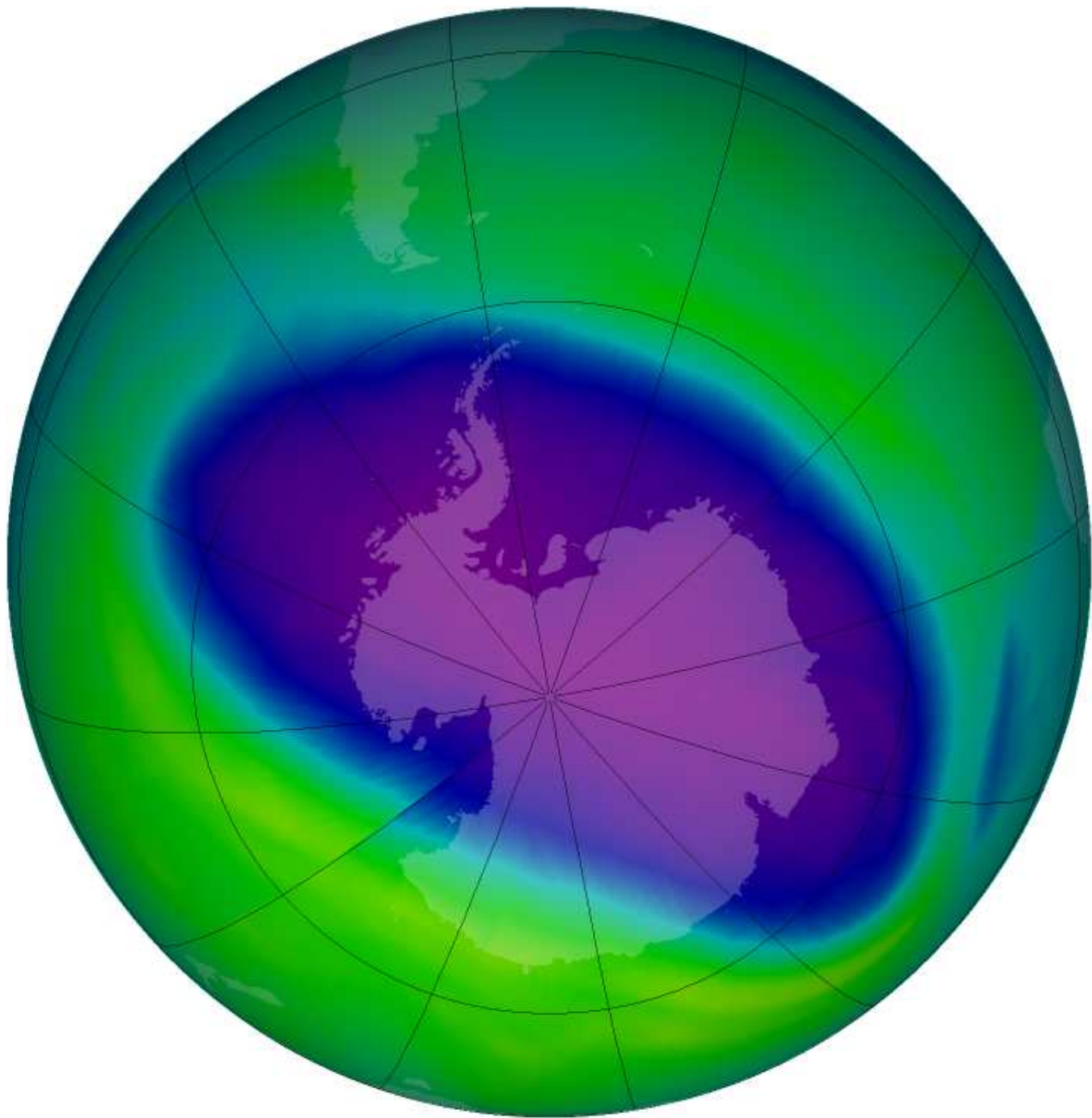
Ozone hole over Antarctica 1 October 2006. The region inside the ring of light blue has Dobson units below 220.
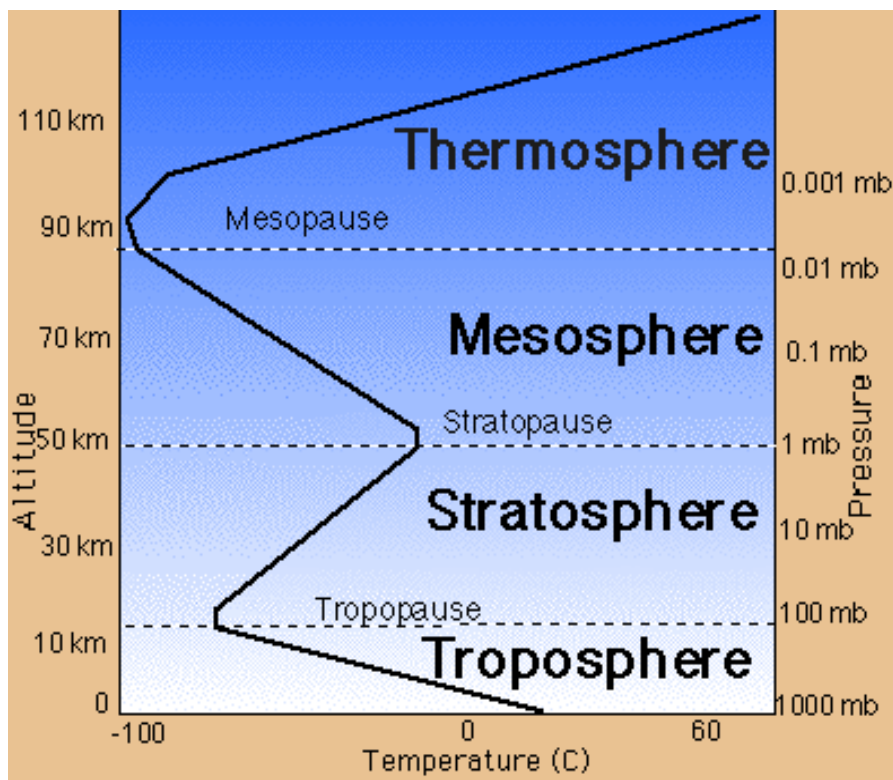
The layer structure and vertical temperature variation of the US standard atmosphere

It has bee estimated that Dobson units below 220 are due to antropogenic effects. The stratosphere and thermosphere are layers with high static stability bounded by the troposphere and mesosphere with much lower static stability.

This layering results mainly from absorption of solar radiation at the surface of the Eath and in ozone in the stratosphere

Boundaries of the atmosphere

No clear **upper** boundary

Radiative equlibrium between incoming short wave (solar) radiation and outgoing long wave (terrestial) radiation

Distinct **lover** boundary (sea, sea ice, complex land surface with/without vegetation, snow, ice etc.)

A large number of exchange processes occur across this boundary (momemtum, heat, trace gases, aerosols, biologic and antropogenic compounds)

Secondary boundaries

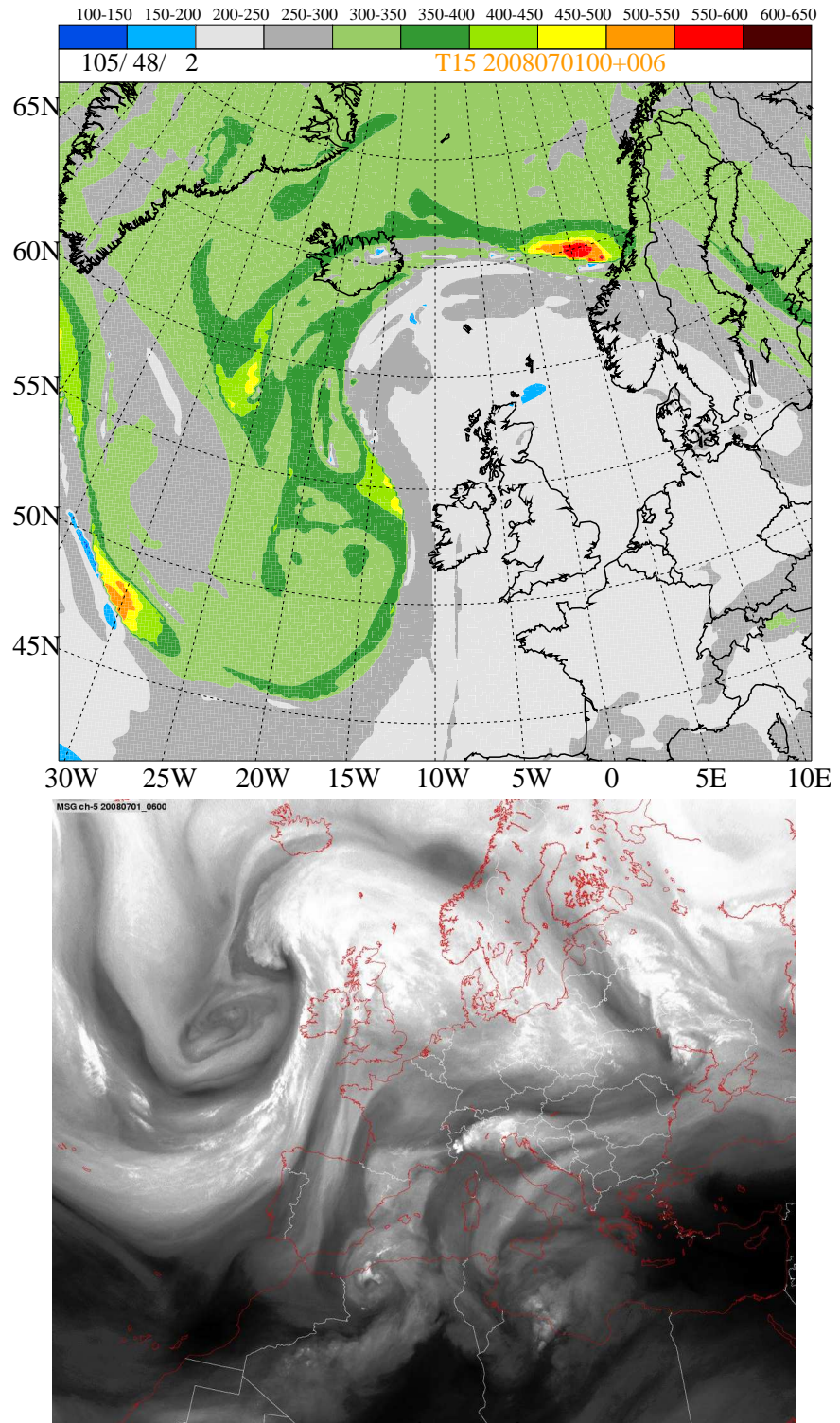- Tropopause

- Stratopause

- Mesopause

Exchange processes across secondary boundaries

Important is mass ecchange between the troposphere and the stratosphere.

This exchange is typically associated with

- deep convection

- tropopause folding associated with upper tropospheric jet streams and superposed jet streaks.

Interaction between dynamics in the troposphere and stratosphere (ongoing research field).

Tropopause height (hPa) and MSG ch5
image 1 July 06 UTC 2008

Top: Tropopause height (hPa) and bot-
tom: Meteosat Second Generation water
vapor image from 1 July 06 UTc 2008

The equation of state of air

To a good approximation the ideal gas law can be applied, i.e.

$$p = R\rho T, \tag{2}$$

with the gas constant $R = R^*/m$, where R^* is the universal gas constant and m the mean molecular weight of the constituents of the atmosphere.

Since $R = R^*/m_d \cdot (m_d/m)$ the equation of state can be rewritten

$$p = R_d\rho T_v, \tag{3}$$

where $R_d = R^*/m_d$ and $T_v = T \cdot m_d/m$ is the virtual temperature. Since $m_d \geq m$ it follows that $T_v \geq T$. The relation between T and T_v can be written

$$T_v = T \cdot m_d/m = T \cdot (1 + \epsilon q), \tag{4}$$

where q and m_d are specific humidity and mean molecular weight of dry air, respectively and $\epsilon = 0.61$.

Approximate equation of state for sea water

$$\alpha \approx \alpha_0 \left[1 + A_T - A_P - A_S + A_{TT} + A_{TP} \right], \quad (5)$$

where

$A_T = \beta_T (T - T_0)$

$A_P = \beta_p (p - p_0)$

$A_S = \beta_S (S - S_0)$

$A_{TT} = \frac{\beta_T^*}{2} (T - T_0)^2$

$A_{TP} = \beta_T \gamma^* p (T - T_0)$

In (4) T is absolute temperature, p is pressure and S salinity of seawater.

Index "0" refers to reference values (T_0=283 K, $\alpha_0 = 9.730 \cdot 10^{-4} \, \text{m}^3 \, \text{kg}^{-1}$, p_0=0).

The "beta"-coefficients are constants ($\beta_T = 1.67 \cdot 10^{-4} \, \text{K}^{-1}$, $\beta_T^* = 1.00 \cdot 10^{-5} \, \text{K}^{-2}$, $\beta_S = 0.78 \cdot 10^{-3} \, psu^{-1}$ and $\beta_p = 4.39 \cdot 10^{-10} \, \text{m} \, \text{s}^2 \, \text{kg}^{-1}$).

Finally, $\gamma^* = 1.1 \cdot 10^{-8} \, \text{Pa}^{-1}$.

'psu' is a practical salinity unit, approximately one part per thousand.

Conservation equations for

- Momentum

- Energy

- Mass

The momentum equation

$$\frac{D\vec{U}}{Dt} = -\frac{1}{\rho}\nabla p - 2\vec{\Omega} \times \vec{U} + \vec{g} + \vec{F}, \qquad (6)$$

where $\vec{U}$ is the velocity relative to the rotating Earth.

The forces on the right hand side are:

- The pressure gradient force

$$-\frac{1}{\rho}\nabla p = -\frac{1}{\rho}\left(\frac{\partial p}{\partial x}\vec{i} + \frac{\partial p}{\partial y}\vec{j} + \frac{\partial p}{\partial z}\vec{k}\right), \qquad (7)$$

  where $\vec{i}$, $\vec{j}$ and $\vec{k}$ are unit vectors along the orthogonal coordinate axes.

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  The Coriolis force

$$-2\vec{\Omega} \times \vec{U} = 2\Omega(-w\cos\phi + v\sin\phi)\vec{i} +$$
$$+ 2\Omega((-u\sin\phi)\vec{j} + (u\cos\phi)\vec{k}) \qquad (8)$$

  where ϕ is latitude and

$$\vec{\Omega} = \Omega\cos\phi\vec{j} + \Omega\sin\phi\vec{k}. \qquad (9)$$

- The gravity force

$$\vec{g} = \Omega^2 \vec{R} + \vec{g_*} = -g\vec{k}, \qquad (10)$$

where $\Omega^2 \vec{R}$ is the centrifugal force and $\vec{g_*}$ is the gravitational force. $\vec{R}$ is the distance vector along the outward direction perpendicular to the axis of rotation

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The molecular frictional force

$$\vec{F} \approx \nu \left(\nabla^2 u \cdot \vec{i} + \nabla^2 v \cdot \vec{j} + \nabla^2 w \cdot \vec{k} \right), \quad (11)$$

where ν is the kinematic viscosity.

The momentum equation in spherical coordinates

In spherical coordinates (λ, ϕ and z) the velocity components in latitudinal(λ), meridional (ϕ) and vertical direction (z) are

$u = r \cos \phi \frac{D\lambda}{Dt}$, $v = r \frac{D\phi}{Dt}$ and $w = \frac{Dz}{Dt}$

Since $\vec{U} = u\vec{i} + v\vec{j} + w\vec{k}$ it follows that

$$\frac{D\vec{U}}{Dt} = \frac{Du}{Dt}\vec{i} + \frac{Dv}{Dt}\vec{j} + \frac{Dw}{Dt}\vec{k} + u\frac{D\vec{i}}{Dt} + v\frac{D\vec{j}}{Dt} + w\frac{D\vec{k}}{Dt} \tag{12}$$

Geometrical considerstions leads to

$$\frac{D\vec{i}}{Dt} = \frac{u}{a \cos \phi} \left(\sin \phi \vec{j} - \cos \phi \vec{k} \right) \tag{13}$$

$$\frac{D\vec{j}}{Dt} = -\frac{u \tan \phi}{a}\vec{i} - \frac{v}{a}\vec{k} \tag{14}$$

$$\frac{D\vec{k}}{Dt} = \frac{u}{a}\vec{i} + \frac{v}{a}\vec{j} \tag{15}$$

In spherical coordinates the momentum equations become

$$\frac{Du}{Dt} + S_x = -\frac{1}{\rho}\frac{\partial p}{\partial x} + 2\Omega v \sin \phi - 2\Omega w \cos \phi + F_x$$

$$\frac{Dv}{Dt} + S_y = -\frac{1}{\rho}\frac{\partial p}{\partial y} - 2\Omega u \sin \phi + F_y$$

$$\frac{Dw}{Dt} + S_z = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g + 2\Omega u \cos \phi + F_z \qquad (16)$$

where $S_x = -\left(uv \tan \phi - uw\right)/a$,
$S_y = \left(uu \tan \phi + vw\right)/a$ and $S_z = -\left(u^2 + v\right)/a$.

The thermodynamic energy equation for a dry atmosphere

$$c_p \frac{DT}{Dt} - \alpha \frac{Dp}{Dt} = J, \qquad (17)$$

where c_p is the specific heat at constant pressure, J the diabatic heating rate per unit mass (due to radiation, latent heat release and conduction), α is specific volume ($\mathrm{m^3\,kg^{-1}}$), p pressure (in Pa) and T temperature (in K).

The consevation equation for mass (or mass continuity equation)

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho \vec{U}) = 0 \qquad (18)$$

Generalization If ζ is the amount of some property of the air per unit volume (i.e. the **concentration** of the property), then the continuity equation for ζ can be written

$$\frac{\partial \zeta}{\partial t} + \nabla \cdot (\zeta \vec{U}) = Q_v[\zeta], \qquad (19)$$

where $Q_v[\zeta]$ is the net effect per unit volume of all non-conservative processes.

Usually a **tracer** in the atmosphere is measured per unit mass of air. The continuity equation for a tracer mixing ratio ψ then becomes

$$\frac{D\psi}{Dt} = Q_m[\psi], \qquad (20)$$

where $Q_m[\psi]$ is the source/sink term.

Radiation

Absorption and emission of radiation in the atmosphere and at its underlying surface is the most important diabatic process on Earth. Solar radiation into the Earth-atmosphere system is responsible for atmospheric circulations.

Black body radiation (Planck's law)

at temperature T in ,K and wavelength λ

$$E_\lambda = \frac{c_1}{\lambda^5[\exp(c_2/(\lambda T)) - 1]},\qquad (21)$$

where $c_1 = 3.74 \cdot 10^{-16}\,\mathrm{W\,m^2}$ and $c_2 = 1.44 \cdot 10^{-2}\,\mathrm{m\,K}$.

The Wien displacement law

$$\lambda_m = \frac{2897}{T},\qquad (22)$$

where λ_m (in micrometer) is the peak emission wavelength for a black body at temperature T in Kelvin.

The Stefan-Boltzmann law

Irradiance from a black body at temperature T is

$$E = \sigma T^4, \tag{23}$$

where $\sigma = 5.67 \cdot 10^{-6} \, \mathrm{W \, m^{-2}}$ is the Stefan-Boltzmann constant.
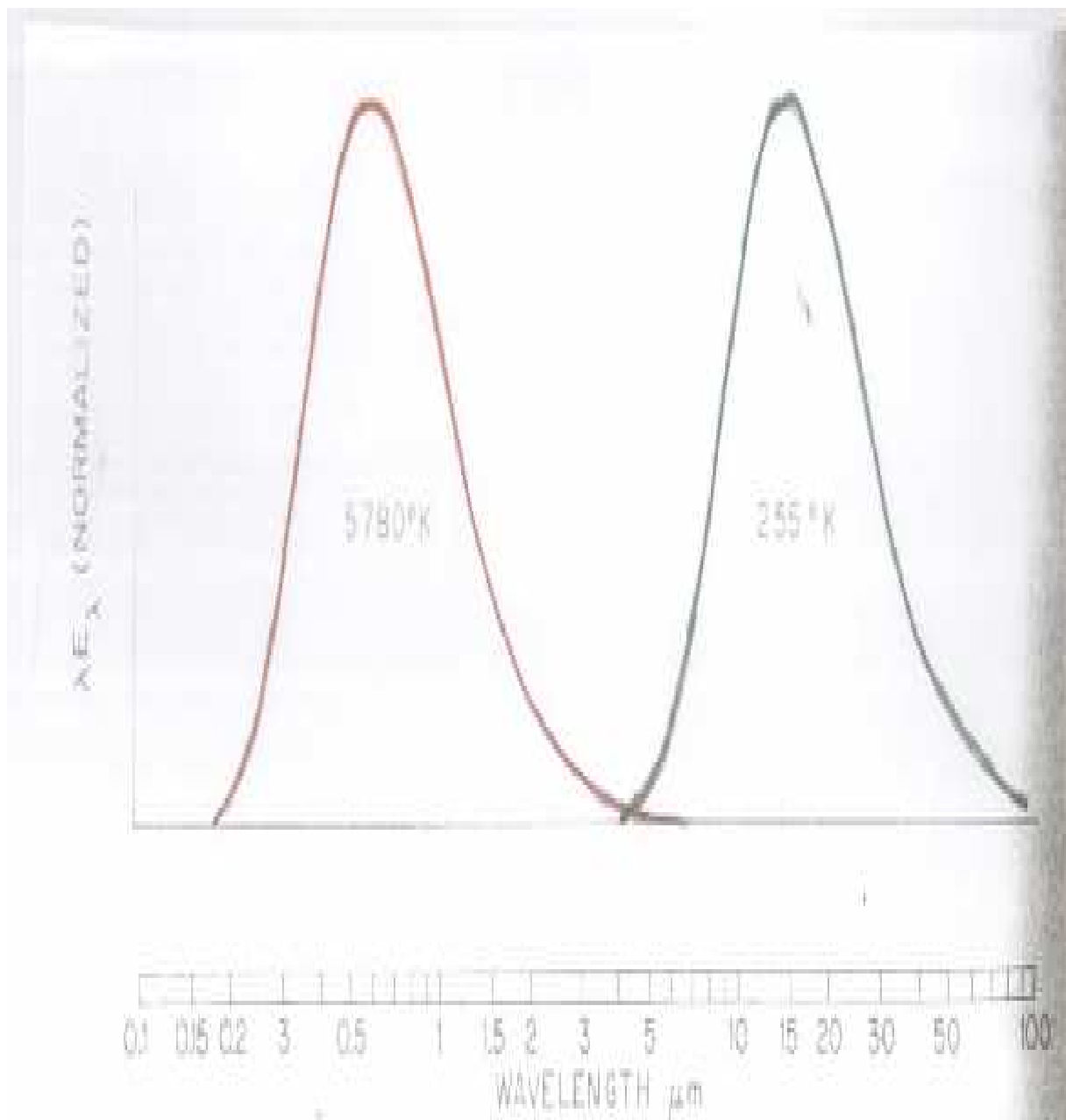
Fig. 6.4 Normalized blackbody spectra representative of the sun (left) and earth (right), plotted on a logarithmic wavelength scale. The ordinate is multiplied by wavelength in order to make area under the curves proportional to irradiance [Adapted from R. M. Goody, "Atmospheric Radiation." Oxford Univ Press (1964), p. 4]

Normalized black body irradiance spectra for the Sun and the Earth

Radiative equilibrium

Radiative equilibrium at the outer edge of the Earth's atmosphere (at radius R_E) yields

$$(1 - A)S_0 \pi R_E{}^2 = E_A \cdot 4\pi R_E{}^2. \qquad (24)$$

Therefore

$$E_A = \frac{(1 - A)S_0}{4} = \sigma T_E{}^4. \qquad (25)$$

If the planetary albedo is $A = 0.3$,
the solar radiation at the top of the atmosphere is $S_0 = 1380\,\mathrm{W\,m^{-2}}$,
then the equvivalent equilibrium temperature of the Earth becomes $T_E \approx 255.5\,\mathrm{K}$.

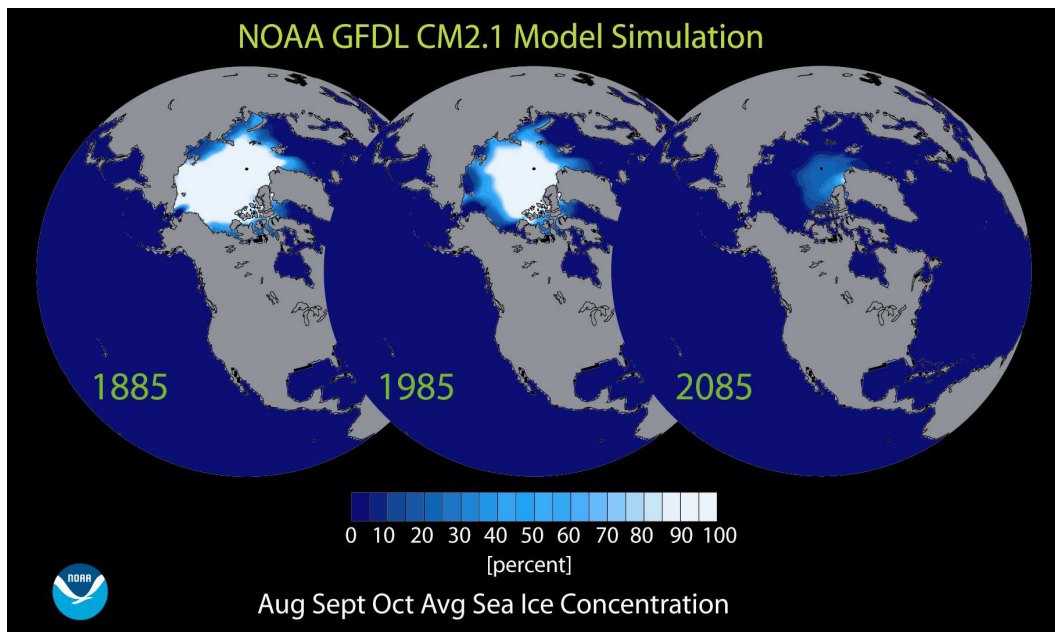A small change δA in planetary albedo results in a change δT_E in the equilibrium temperature, which is approximately

$$\delta T_E \approx -\frac{S_0}{16\sigma T_E{}^3} \cdot \delta A, \qquad (26)$$

or at $T_E = 255.5\,\mathrm{K}$

$$\delta T_E \approx -92 \cdot \delta A, \qquad (27)$$

indicating that climate on Earth is quit sensitive to small changes in global albedo.

Climate scenario simulation of average ice cover (percent) in the Polar sea in the period August to October
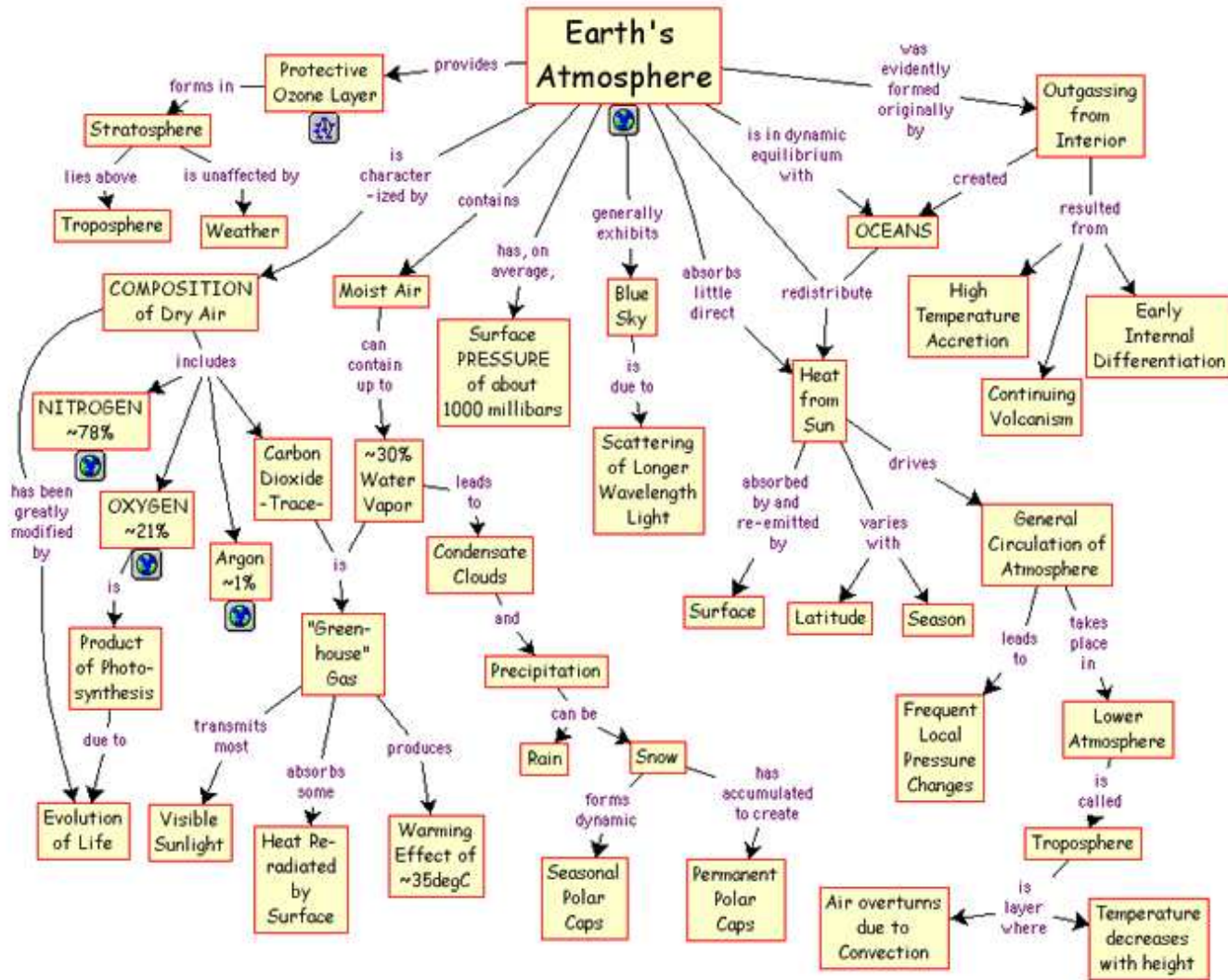
Diagram giving a brief and incomplete overwiev of composition and properties of and processes occurring in the atmosphere