General Meteorology

II Earth‘s atmosphere and sun

II.1 Chemical composition of dry atmosphere

II.2 Physical units for trace gas amounts

II.3 Layering of the atmosphere

II.4 Hydrostatic Equation

II.5 Temperature profile of the atmosphere

II.6 Moisture – water vapour in the air

II.7 Earth´s radiation budget

UNIVERSITÄT BREMEN

II.1 Chemical composition of dry atmosphere

Molecule	partial volume/ partial pressure	Molar mass [10^{-3} kg/mole]	Boiling point [°C]
Nitrogen N_2	0.7809	28.016	-196
Oxygen O_2	0.2095	32.000	-183
Argon Ar	0.0093	39.944	-186
Carbon dioxide CO_2	0.0003	44.010	-78
Major components (air)	1.0000	28.966	(-193)
Neon Ne	18×10^{-6}	20.182	-246
Helium He	5.2×10^{-6}	4.003	-269
Methane CH_4	$1\text{-}2 \times 10^{-6}$	16.03	-
Crypton Kr	1.0×10^{-6}	83.8	-153
Hydrogen H_2	0.5×10^{-5}	2.016	-253
Nitrous oxide N_2O	$0.2\text{-}0.6 \times 10^{-6}$	44.016	-89
Xenon Xe	0.08×10^{-6}	131.30	-107
Carbon monoxide CO	$0.01\text{-}0.2 \times 10^{-6}$	28.01	-191
Ozone O_3 (*)	0.01×10^{-6}	48.00	-112

(*) surface value, stratosphere: $2\text{-}8 \times 10^{-6}$

→ other minor components (note: chemical active species are in minority!!)
CH_2O (formaldehyde), NO_2 (nitric oxide), NH_3 (ammonia), SO_2(sulfur dioxide), I_2 (jodine), Cl_2 (chlorine), Rn (radon)

→ water vapour: very variable 0-4%

UNIVERSITÄT BREMEN

II.2 Physical units for trace gas amounts

- ideal gas law $\boldsymbol{p \cdot V = n \cdot R \cdot T}$ (Eq. II.1)
  - ➔ p pressure
  - ➔ V volume
  - ➔ n amount in units of mole
  - ➔ R =8.31 J/(mole·K), gas constant
  - ➔ T temperature in K

- trace gas amount:
  - ➔ 1 mole is the amount of a substance which has the same number of molecules as 12g of ^{12}C isotope
  - ➔ 1 mole contains 6.022 x 10^{23} molecules
  - ➔ N_A=6.022 x 10^{23} molecules/mole → Avogadro number

- molecular density:

→molar density $$\rho_{mol} = {}^{n}\!/_{V} = {}^{p}\!/_{R \cdot T}$$ units: mole/m³ (Eq. II.2)

→number density $$\rho = {}^{n \cdot N_A}\!/_{V} = {}^{p \cdot N_A}\!/_{R \cdot T} = {}^{p}\!/_{k \cdot T}$$ units: molec./m³ (Eq. II.3)

❖ k=1.38x10^{-23} J/K Boltzmann constant

→mass density $$\rho_m = {}^{n \cdot m}\!/_{V} = {}^{m \cdot p}\!/_{R \cdot T}$$ units: kg/m³ (Eq. II.4)

❖ m: molar mass in g/mole (air ~29 g/mole)

Trace gas amounts

- volume mixing ratio ξ_i of *i*-th trace gas: $$\xi_i = \frac{\rho_i}{\rho} = \frac{p_i}{p} = \frac{\rho_{i,mol}}{\rho_{mol}} = \frac{V_i}{V}$$

    ➔ P_i, V_i: partial pressure and partial volume of i-th trace gas

    ➔ p air pressure; ρ, ρ_{mol} air density (molar or number density)

    ➔ volume mixing ratio units: volume parts per million=ppmv=10^{-6} or ppbv=10^{-9}

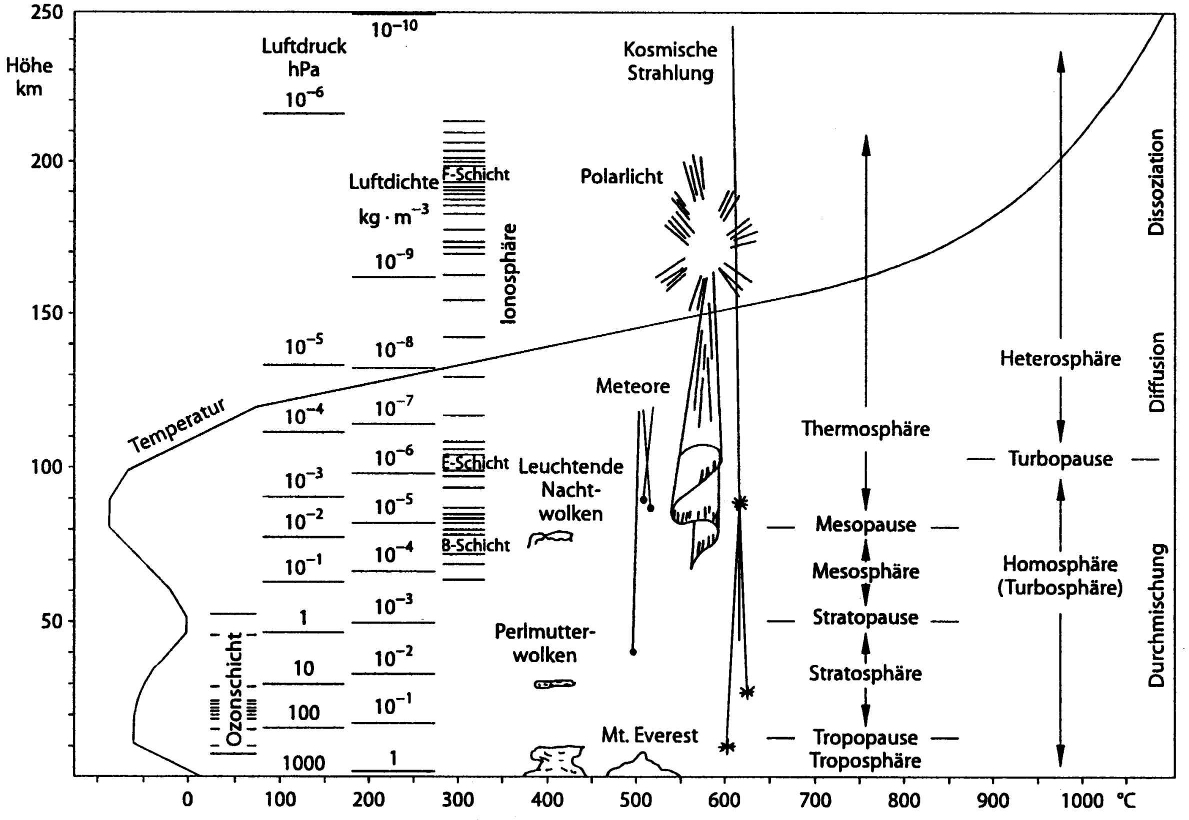
$$\rho_m = \sum_i \rho_{i,m} = \rho_{N2,m} + \rho_{O2,m} + \cdots \qquad p = \sum_i p_i = p_{N2} + p_{O2} + \cdots$$

$$\rho = \sum_i \rho_i = \rho_{N2} + \rho_{O2} + \cdots \qquad \rho_{mol} = \sum_i \rho_{i,mol} = \rho_{N2,mol} + \rho_{O2,mol} + \cdots$$

- other units used are ppmm (mass parts per million) or ppbm according to partial mass density

    ➔ Note: if neither volume or mass units are given, then most likely:
    1 ppm=1 ppmv or 1 ppb =1 ppbv

II.3 Layering of the atmosphere



dissociation

molecular diffusion

mixing

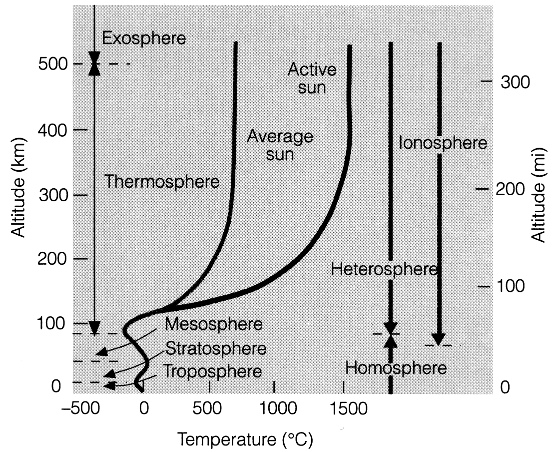
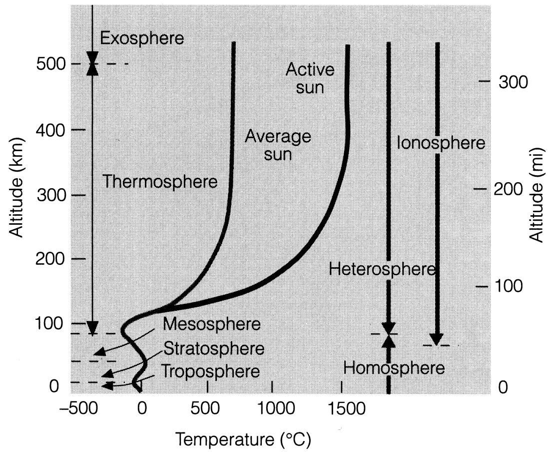


Abb. 2.20. Vertikale Verteilung von Temperatur (*ausgezogene Kurve*), Luftdruck und Luftdichte (Zahlenangaben für ausgesuchte Punkte) sowie Zuordnung der geläufigen Schichtbezeichnungen und einiger bekannter Phänomene (nach Liljequist 1974)

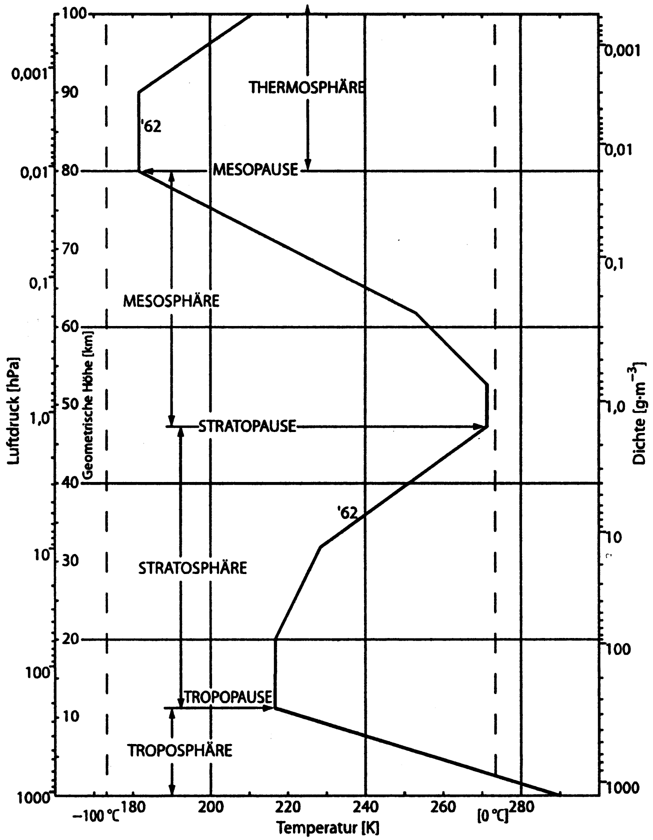
glossary: Luftdruck=pressure, Luftdichte =air density, Ozonschicht=ozone layer, Höhe=height, Ozonschicht=ozone layer, B- E-, F-Schicht=B-,E-,F-layer (ionic layers), kosmische Strahlung=(galactic) cosmic ray, Polarlicht=aurora, Meteore=meteorites, leuchtende Nachtwolken=noctilucent clouds, Perlmutterwolken=polar stratospheric clouds (mother of pearl clouds)

UNIVERSITÄT BREMEN

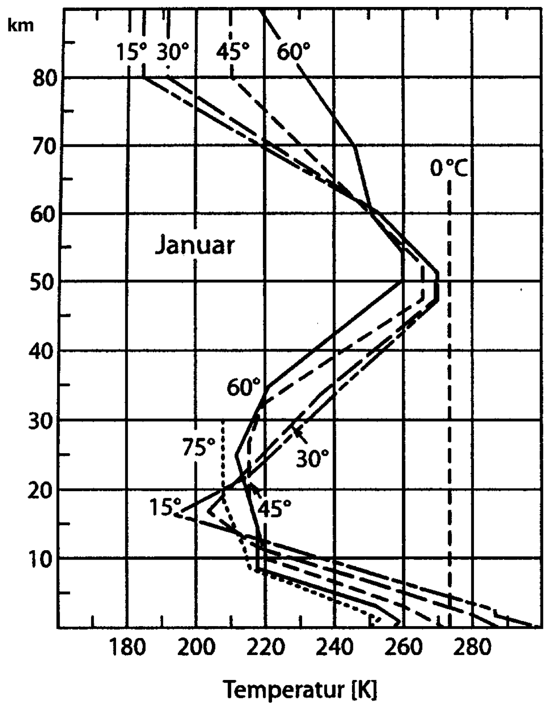
Criterium	term	altitude
life forms	biosphere	0-20 km
composition	homosphere	0-100 km
	homopause	100-120 km
	heterosphere	>120 km
temperature	troposphere	0-12 km
	tropopause	~12 km
	stratosphere	12-50 km
middle atmosphere	stratopause	~50 km
	mesosphere	50-85 km
upper atmosphere	mesopause	~85 km
	thermosphere	85-500 km
	exosphere	>500 km
radio physics	ionosphere	50 –600 km
	magnetosphere	>300 km

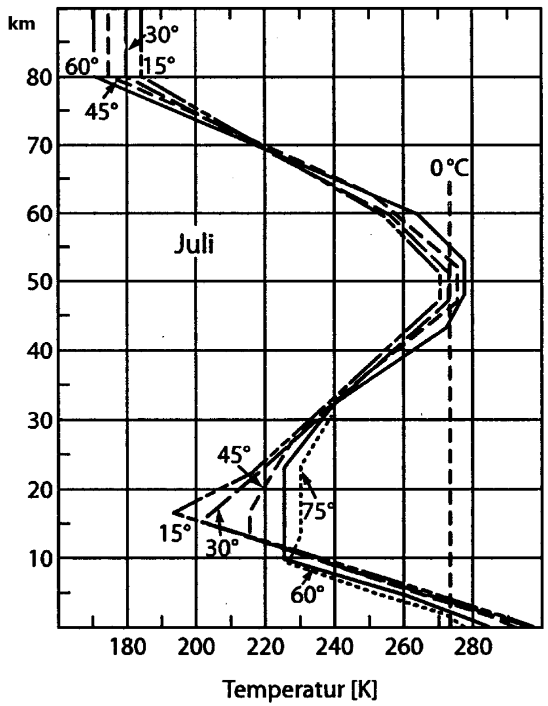


atmospheric layering and temperature

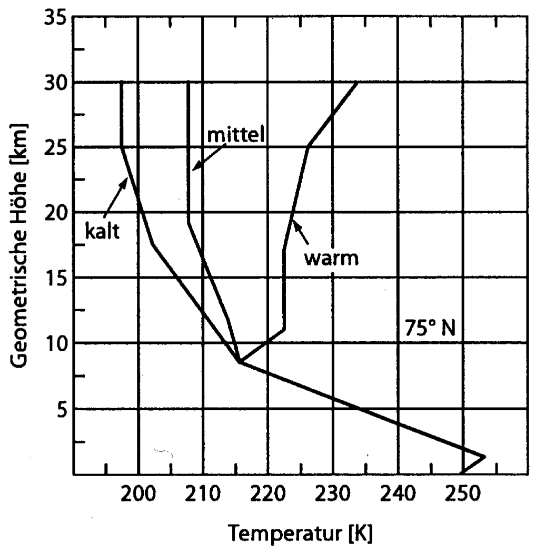


WMO standard atmosphere 1962 (global)





seasonal variation in NH
January & July (Valley 1965)

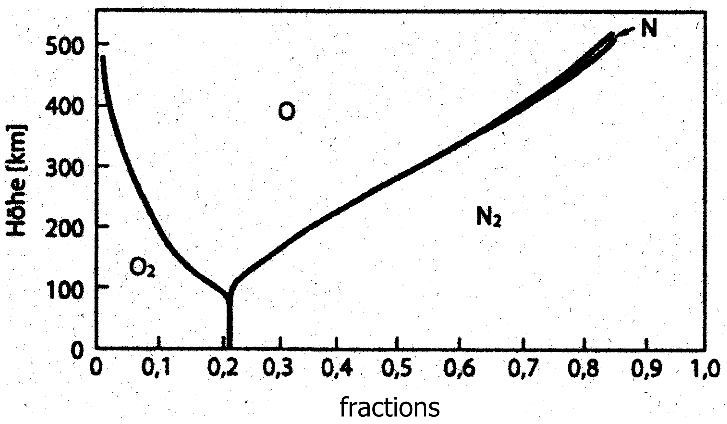


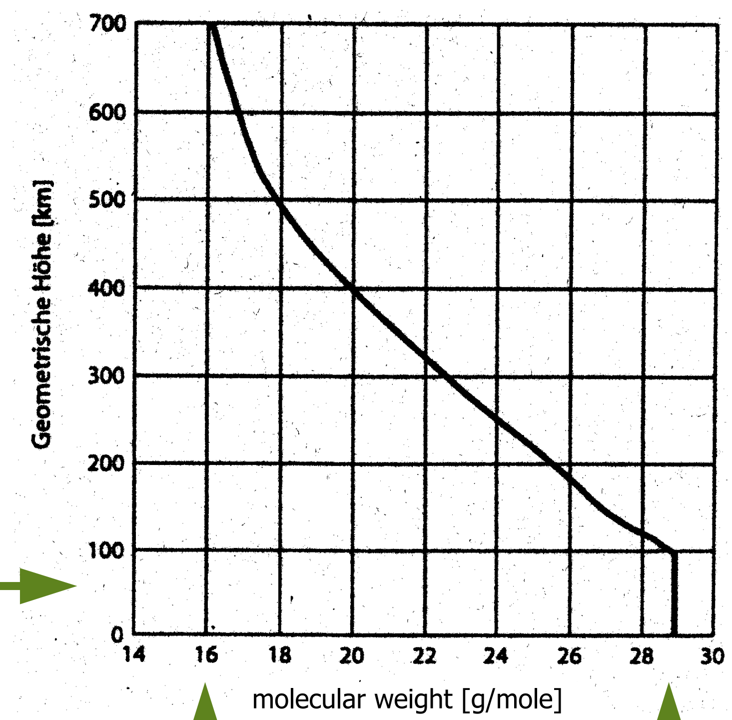
stratospheric warming in winter
75°N (Valley 1965)

atmospheric mass:
- troposphere 90%
- stratosphere 9.5%
- mesosphere 0.5%

UNIVERSITÄT BREMEN

- other criteria:
  - ➔ aerodynamical state (planetary boundary layer)
    - ❍ Prandtl layer 0-50 m
    - ❍ Ekman layer 50-1000 m
    - ❍ free atmosphere (above boundary layer > 1000 m)





homosphere

O

O_2, N_2

The sun and ionosphere

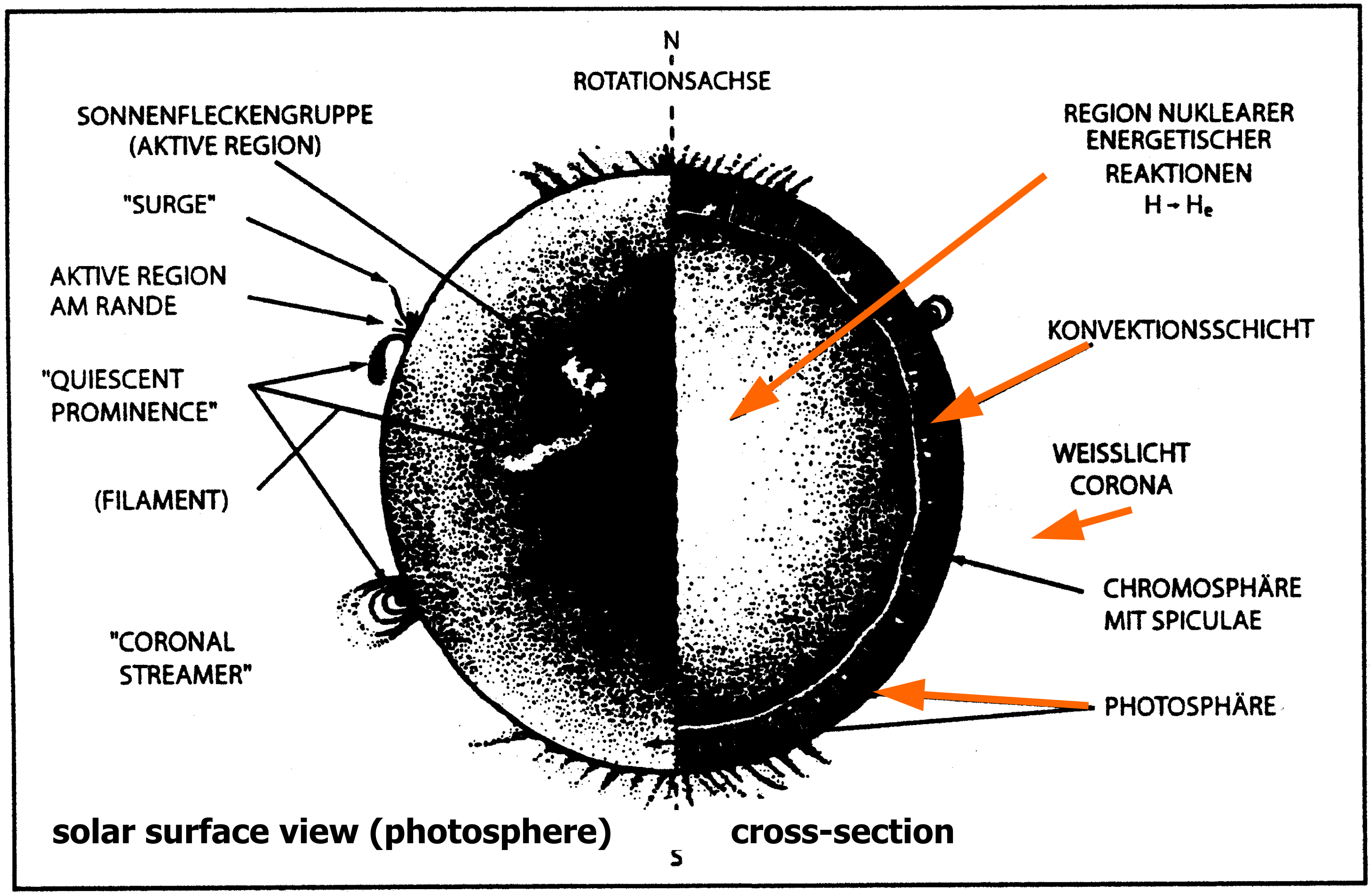


Abb. 3.3. Erscheinungen auf der Sonne und ihre Bezeichnungen (nach Valley 1965)

glossary: Sonnenfleckengruppe=sun spot group, Rotationsachse=rotation axis

UNIVERSITÄT BREMEN

The sun and ionosphere

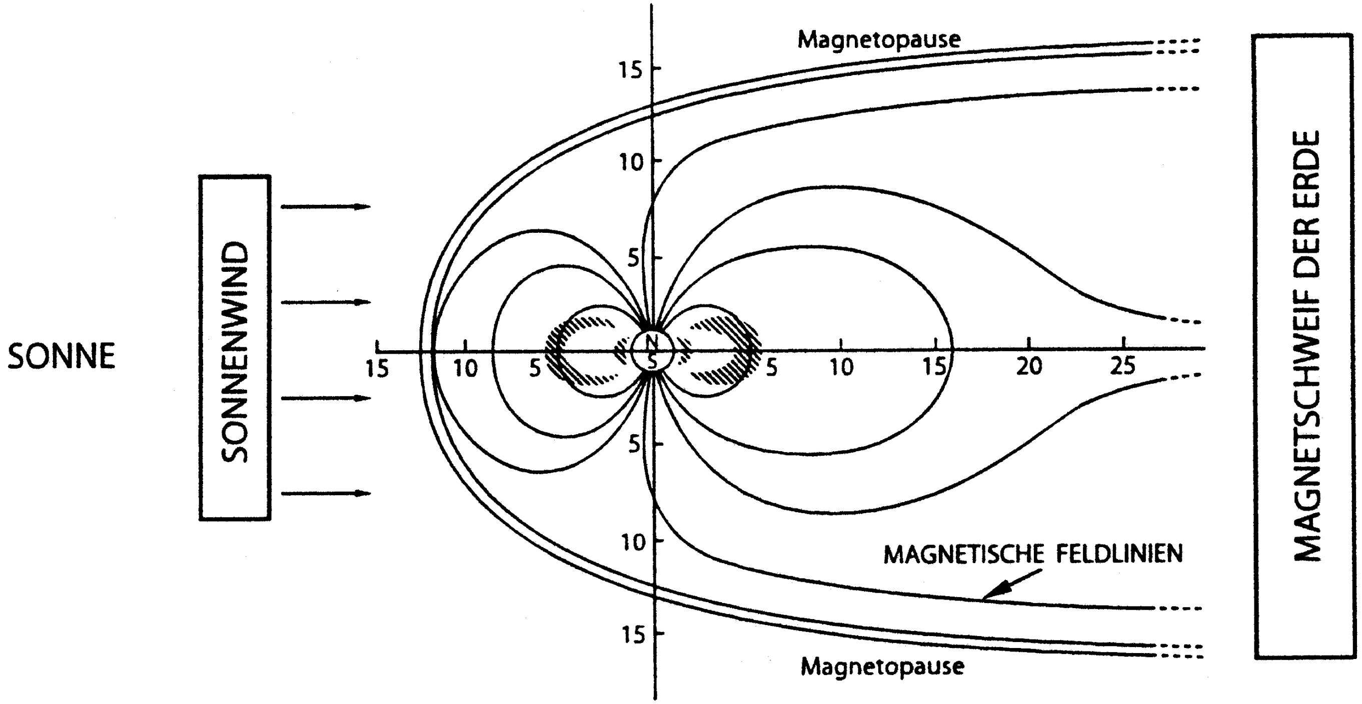


Abb. 3.4. Das Magnetfeld der Erde im Sonnenwind. Die *Doppellinie* markiert die Magnetopause. Die ungefähre Lage der Van-Allen-Strahlungsgürtel ist *schraffiert* gekennzeichnet. Die Anströmrichtung des Sonnenwindes in bezug auf die erdmagnetische Achse variiert mit der Jahreszeit bzw. im Gefolge von säkularen Magnetpolwanderungen (nach Dobson 1968)

glossary: Sonne=sun, magnetische Feldlinien=magnetic field lines, Sonnenwind=solar wind, Magnetschweif der Erde=magnetic tail, shaded area=Van Allen belt (ion plasma)

The sun and ionosphere

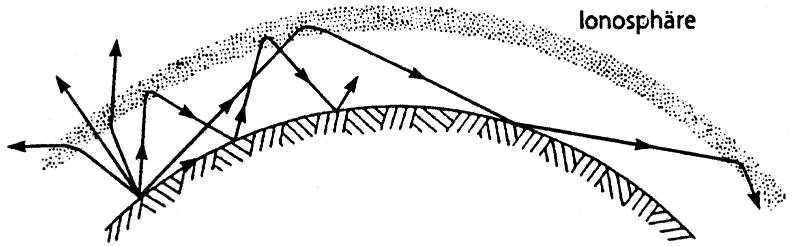


Abb. 3.8. Radiowellenausbreitung und Ionosphäre (nach Liljequist 1974): In den elektrisch leitenden Schichten der Ionosphäre werden die Radiowellen abgelenkt, bei entsprechend schrägem Einfall zur Erdoberfläche zurückgebrochen. Diese „Totalreflexion" ermöglicht die Radiowellenausbreitung über weite Distanzen. Dadurch ist z.B. der Kurzwellen-Radioempfang rund um den Globus möglich. Vor dem Satellitenzeitalter war das ein wichtiges Element der Radio-Telekommunikation

Abb. 3.9. Schematische Lage der irdischen Ionosphärenschichten, relativ zur Sonnenposition, zur Zeit der Tag- und Nachtgleiche (nach Liljequist 1974). Die Darstellung ist nicht maßstabsgerecht!

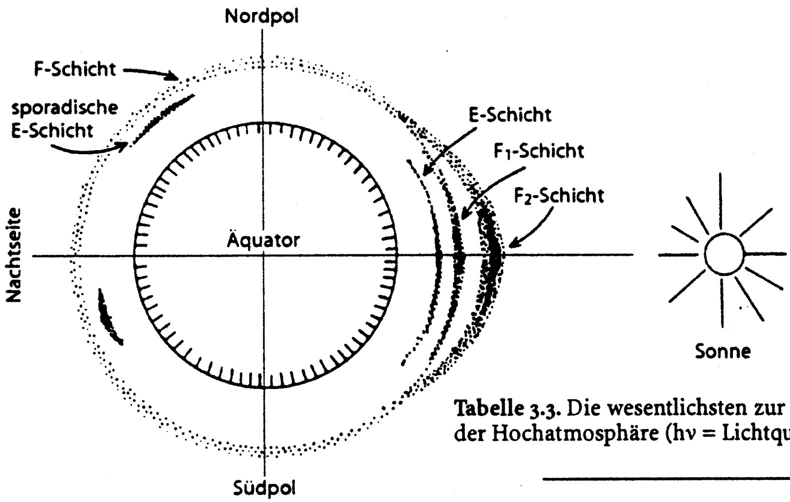


Tabelle 3.3. Die wesentlichsten zur Bildung der Ionosphäre beitragenden photochemischen Prozesse der Hochatmosphäre (hν = Lichtquant, ν = Frequenz der Strahlung, h = Planck-Konstante)

Photochemical process	wavelength [Å]		layer
$NO + h\nu \longrightarrow NO^+ + e$	1215,7	(Lyman-a)	D
$O_2 + h\nu \longrightarrow O_2^+ + e$	1024,7	(Lyman-ß)	E
$O_2 + h\nu \longrightarrow O_2^+ + e$	1012 – 910		D
$O + h\nu \longrightarrow O^+ + e$	910 – 795		F_1, F_2
$N_2 + h\nu \longrightarrow N_2^+ + e$	795 – 755		E
$O_2 + h\nu \longrightarrow O_2^+ + e$	744 – 661		E
$N_2 + h\nu \longrightarrow N_2^+ + e$	661 – 585		F

glossary: Nachtseite= night side, sporadische E-Schicht=sporadic E-layer, photochemischer process=photochemical reaction, Wellenlänge =wavelength

The sun and ionosphere

Abb. 3.10. Vertikalverteilung der Konzentration von Luftmolekülen insgesamt [Teilchenzahl pro cm^{-3}] (*rechte Kurve*) und freien Elektronen [Anzahl pro cm^{-3}] (*linke Kurve*) (nach Wallace u. Hobbs 1977)

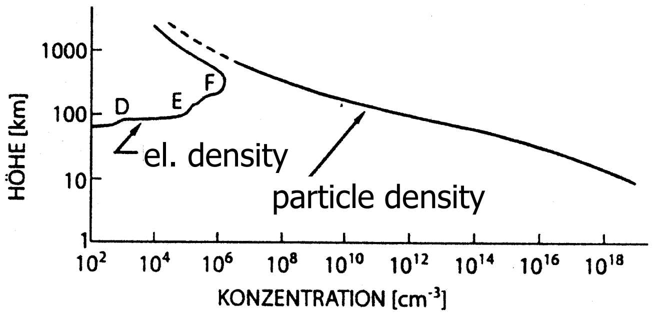
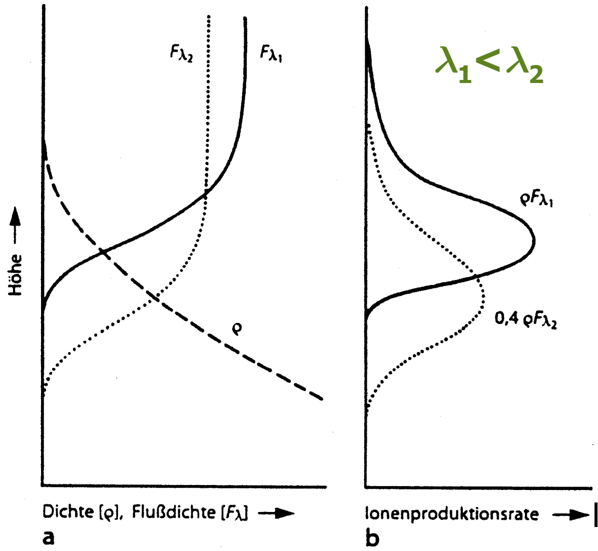


Abb. 3.11. Schematische Illustration der Schichtenbildung in der Ionosphäre als Folge gegenläufigen Verhaltens mit der Höhe von spektraler solarer Strahlungsflußdichte F_λ und Dichte ϱ des bei der entsprechenden Wellenlänge ionisierbaren Gases (nach Fleagle u. Businger 1980)



$$P \propto \rho \cdot F_\lambda$$

P : ion production rate

ρ : density of ionisable trace gases

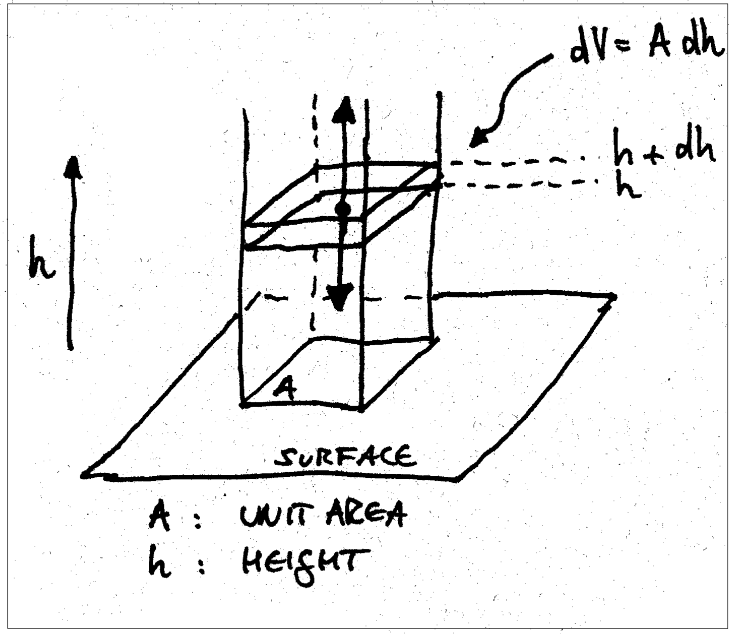
F_λ: solar flux density at wavelength λ

Tabelle 3.4. Übersicht über den Schichtaufbau der Ionosphäre und die in den einzelnen Schichten vorherrschenden physikalischen Bedingungen

layer	altitude [km]	radiation	ions	el.density [cm^{-3}]	max e- [km]
D	50–85	Lyman-α, Röntgen, Kosmische	NO^+, O_2^+	≈ 10^3	80
E	85–150	UV, weiche Röntgen ≈ 100 Å	O_2^+, NO^+, O^+	$1{,}5 \cdot 10^5$	105
F	> 150	UV 300–1 000 Å	untere Region: O_2^+, NO^+;	$2 \cdot 10^5$ (F_1)	170
			mittl. Region: O^+;	10^6 (F_2)	320
			obere Region: He^+, H		

glossary: Röntgen=x-ray, weiche Röntgen= soft x-ray, Dichte=density, Flußdichte=flux density, Ionenproduktionsrate=ion production rate

II.4 Hydrostatic equation



• FORCE : $F = m \cdot g = \rho_m \cdot V \cdot g$

m MASS
g gravity acceleration
V VOLUME

• PRESSURE = FORCE/UNIT AREA

$$p = F/A \Rightarrow dF = A \cdot dp$$

• DIFFERENCE OF FORCE APPLIED AT ALTITUDE h+dh AND ALTITUDE h :

$$\Delta F = F(h+dh) - F(h) = (p-dp)A - p \cdot A$$

$$= -A\,dp = m(dV) \cdot g = \rho_m g \cdot A\,dh$$

MASS IN PARTIAL VOLUME $m(dV) = \rho_m \cdot dV = \rho_m \cdot A\,dh$

- HYDROSTATIC EQUATION:

$$dp = -\rho_m \cdot g \cdot dh \qquad \text{(Eq. II.5)}$$

- PRESSURE AT ALTITUDE h IS $p(h)$

$$\int_{p(h)}^{p(\infty)} dp' = -\int_h^{\infty} g \cdot \rho_m(h')\, dh'$$

integration on the left from $p(h)$ to $p(\infty)$ and from height h to top-of-athmosphere ($h = \infty$)

$$\xRightarrow{p(\infty)=0} \quad p(h) = \int_h^{\infty} g \cdot \rho_m(h')\,dh' \underset{g = \text{const}}{\simeq} g \cdot \underbrace{\int_h^{\infty} \rho_m(h')\,dh'} \qquad \text{(Eq. II.6)}$$

partial mass density of air column above height h in units of kg/m^2

that means that the pressure at altitude h is proportional to the weight of air column above that altitude

UNIVERSITÄT BREMEN

hydrostatic equation (cont'd)

- PUT IDEAL GAS LAW INTO HYDROSTATIC EQUATION, i.e.

Eq. II-4: $\rho_m = m_L \cdot p / R \cdot T$

$$\Rightarrow dp = -\frac{m_L \cdot p}{R \cdot T}\, g\, dh$$

m_L MOLAR MASS OF AIR $= 28.97\,\frac{g}{mole}$

$g = 9.81\ m/s^2$

$R = 8.31\ J/(mole \cdot K)$

$R_L = R/m_L = 287\ J/(kg \cdot K)$

AIR GAS CONSTANT

$$\Rightarrow \frac{dp}{p} = -\frac{g}{R_L \cdot T(h)} \cdot dh \qquad \text{(Eq. II.7)}$$

Now we have separated variables, pressure terms to the left and height dependent terms to the right

- ISOTHERMAL ATMOSPHERE: $T(h) = \overline{\overline{T}} = \text{const.}$

$$\int_{p(h_1)}^{p(h_2)} \frac{dp'}{p'} = -\frac{g}{R_L \cdot \overline{\overline{T}}} \int_{h_1}^{h_2} dh'$$

$$\Rightarrow \ln p(h_2) - \ln p(h_1) = \frac{-g}{R_L \cdot \overline{\overline{T}}}(h_2 - h_1$$

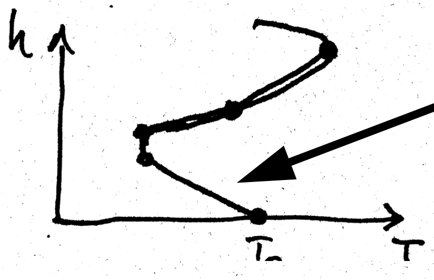
$$\Rightarrow \overline{\overline{T}} = -g/R_L \cdot \frac{h_2 - h_1}{\ln p(h_2) - \ln p(h_1)} \qquad \text{(Eq. II.8)}$$

This is an important relationship that will allow us to estimate temperature from geopotential height maps at two pressure levels

NOTE: THE DIFFERENCE OF THE TWO ALTITUDES BELONGING TO TWO PRESSURE LEVELS, RESPECTIVELY IS PROPORTIONAL TO THE MEAN TEMPERATURE BETWEEN THE TWO PRESSURE LEVELS

UNIVERSITÄT BREMEN

- LINEAR INCREASE (DECREASE) OF TEMPERATURE WITH HEIGHT. GENERALLY TEMPERATURE DOES NOT CHANGE LINEARLY, BUT TEMPERATURE PROFILE CAN BE APPROXIMATED BY PIECEWISE LINEAR CHANGES AND/OR ISOTHERMAL LAYERS



$$T = T_0 - \gamma \cdot h$$

TROPOSPHERE : $\gamma = 6.5 K/km$ LAPSE RATE

SURFACE : $p_0 = p(h=0) = 1013 hPa$

$T_0 = T(h=0) = 288 K$

$$\int_{p_0}^{p(h)} \frac{dp'}{p'} = -\frac{g}{R_L} \int_0^h \frac{dh'}{T_0 - \gamma \cdot h'}$$

$$\Rightarrow \ln p(h) - \ln p_0 = +\frac{g}{R_L \cdot \gamma} \ln (T_0 - \gamma h') \Big|_0^h$$

$$= +\frac{g}{R_L \cdot \gamma} \left[\ln (T_0 - \gamma \cdot h) - \ln T_0 \right]$$

This way the relationship between atmospheric pressure and temperature can be well approximated at any altitude

$$\Rightarrow \ln \frac{p(h)}{p_0} = \frac{g}{R_L \cdot \gamma} \ln \left(\frac{T_0 - \gamma \cdot h}{T_0} \right)$$

$$\Rightarrow p(h) = p_0 \left[\frac{T_0 - \gamma \cdot h}{T_0} \right]^{g/R_L \cdot \gamma} \qquad \text{(Eq. II.9)}$$

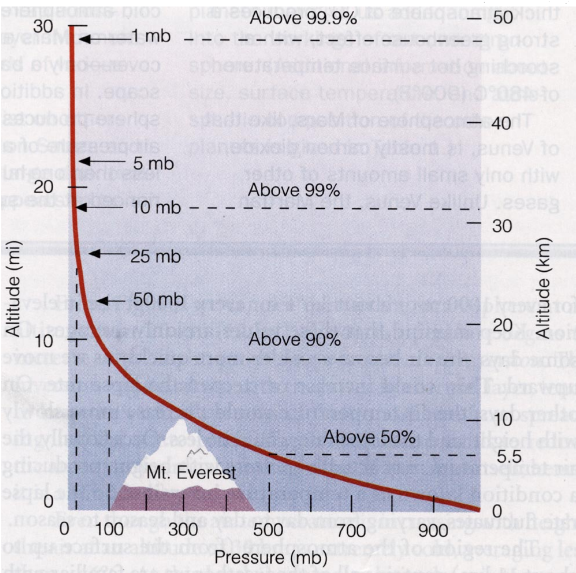


FIGURE 1.8
Atmospheric pressure decreases rapidly with height. Climbing to an altitude of only 5.5 km, where the pressure is 500 mb, would put you above one-half of the atmosphere's molecules.

Ahrens 1999

- DEFINITION OF SCALE HEIGHT

FROM Eq. II-7 : $\frac{dp}{p} = -\frac{g}{R_L \cdot \bar{T}}\, dh$ (ISOTHERMAL LAYER);

FOLLOWS IF INTEGRATED FROM p_0 TO $p(h)$

$$p(h) = p_0 \cdot \exp\left(\frac{-g}{R_L \cdot \bar{T}} \cdot h\right)$$

$$\equiv p_0 \cdot \exp(-h/H)$$

$H = \frac{R_L \cdot \bar{T}}{g}$ IS CALLED SCALE HEIGHT OF ATMOSPHERE

$$\left.\begin{aligned} R_L &= 287\ J/(kg \cdot K) \\ g &= 9.81\ m/s^2 \\ \bar{T} &\approx 273K \end{aligned}\right\} \Rightarrow H = 8\ \text{km}$$

➔if our atmosphere is compressed to normal sea level pressure the atmosphere would be 8 km thick

➔at about 5.5 km altitude the pressure is half the value at sea level (~500 hPa)

➔air number density and pressure decrease exponentially with altitude

UNIVERSITÄT BREMEN

Ω is called the air vertical (total) column density usually given in molec./cm^2

• DEFINITION OF TOTAL COLUMN AIR DENSITY

AIR COLUMN ABOVE UNIT AREA A

$$\rho(h) = \rho_0 \cdot \exp(-h/H_0)$$

$$\text{TOTAL COLUMN } \Omega = \int_{h=0}^{\infty} \rho(h')\,dh' = -\rho_0 \cdot H \exp\left(-\frac{h}{H}\right)\Big|_0^{\infty}$$

$$= -\rho_0 \cdot H\,(0-1) = \rho_0 \cdot H$$

$$\Rightarrow \Omega = \rho_0 \cdot H = \frac{p_0}{k \cdot T_0} H \quad \text{(Eq. II.10)}$$

$$\rho_0 = \frac{p_0}{k \cdot T_0} \quad (\text{Eq. II.3})$$

Ahrens 1999

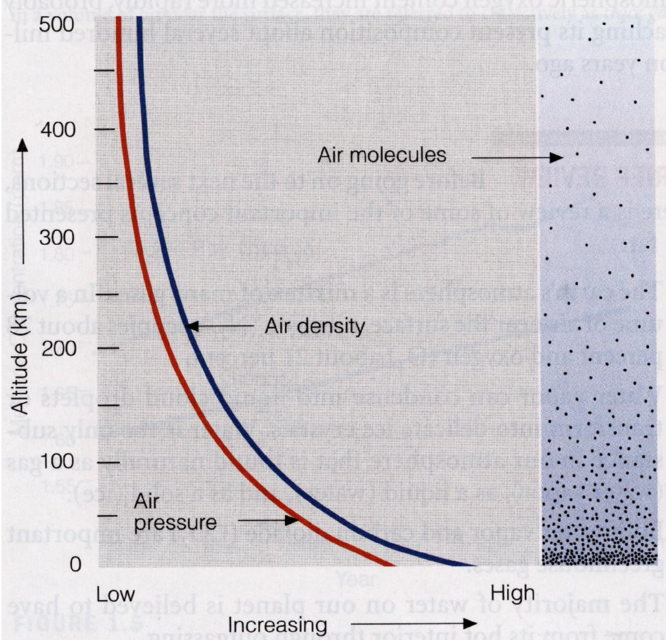


FIGURE 1.7
Both air pressure and air density decrease with increasing altitude.

IN AN ANALOGOUS MANNER TOTAL COLUMNS OF INDIVIDUAL TRACE GASES CAN BE DEFINED

$$\Omega_i = \rho_0 \cdot H_i$$

AGAIN: $H = \sum_i H_i = H_{N_2} + H_{O_2} + \cdots$

$$\Omega = \sum_i \Omega_i = \Omega_{N_2} + \Omega_{O_2} + \cdots$$

EXAMPLE: SCALE HEIGHT OF OZONE = 0.3cm

$$\Omega_{O_3} = \rho_0 \cdot H_{O_3} = \frac{p_0}{k \cdot T_0} H_{O_3}$$

$$= \frac{1013\,hPa}{1.38\times10^{-23}\frac{J}{K}\cdot 273K} \cdot 3\times10^{-3}\ \mathrm{m}$$

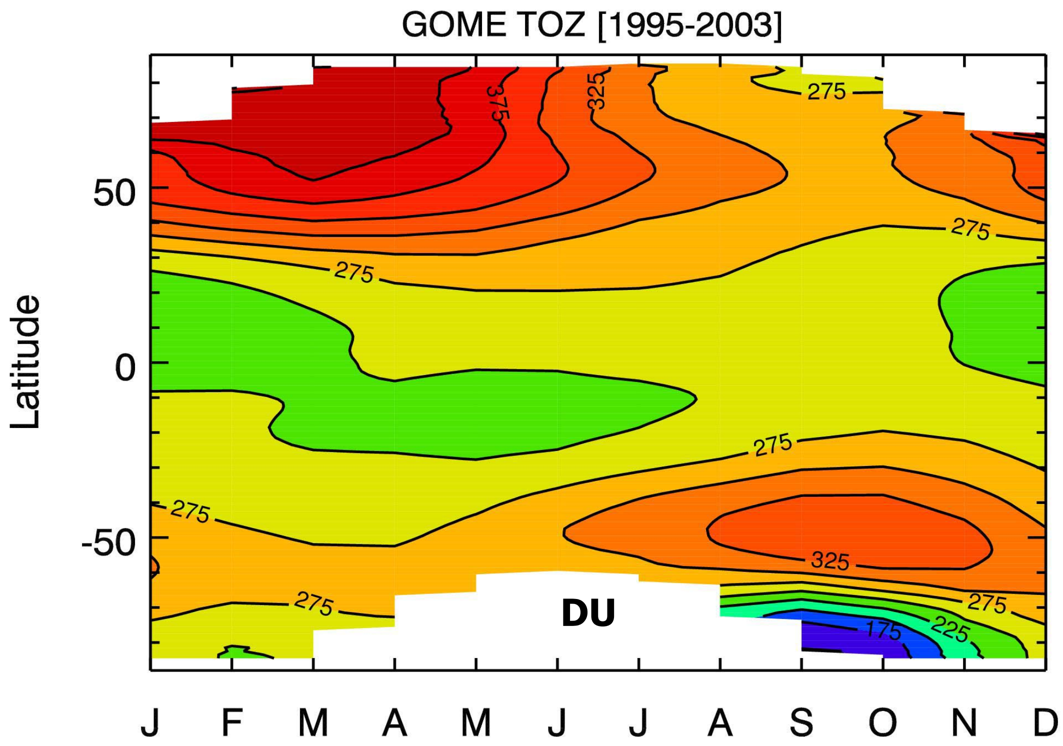
$$= 8.07\times10^{22}\ \frac{\text{MOLEC.}}{m^2} = 300\ \text{DU}$$

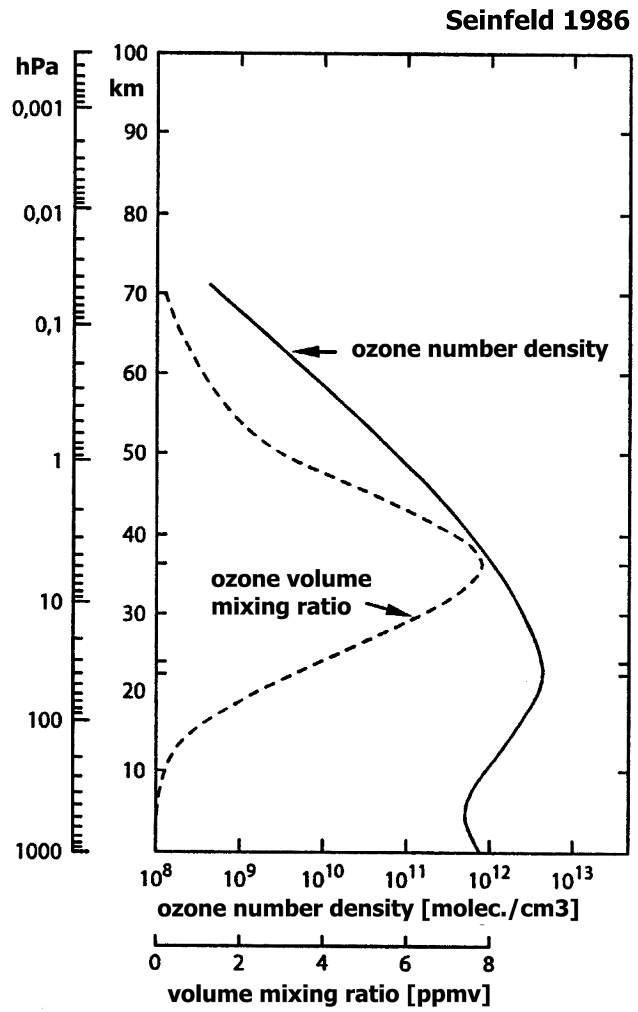
DOBSON UNIT

$1\,\text{DU} = 2.689\times10^{20}\ \frac{\text{MOLEC.}}{m^2}$

➔global ozone total column average about 300 DU

➔ozone „layer" is extremely thin (3 mm) but still very important since O_3 is a very strong absorber (per molecule) of UV radiation





➔global ozone total column average about 300 DU

➔ozone „layer" is extremely thin (3 mm) but still very important since O_3 is a very strong absorber (per molecule) of UV radiation

geopotential height

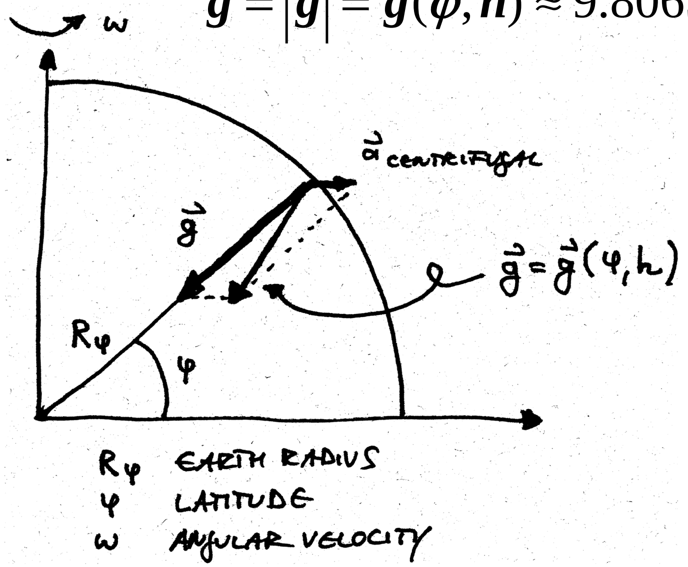
- geopotential ϕ at any point of the atmosphere is defined as the work to be done to bring a 1 kg mass against gravity to that point from sea level

$$[\phi] = \frac{J}{kg} = m \cdot \frac{m}{s^2}$$

energy per unit mass=path·acceleration

$$d\phi = g \cdot dh \longrightarrow \phi(h) = \int_{h'=0}^{h} g(h')dh' \quad \text{(Eq. II.11)}$$

- g is not constant and its value depends on latitude and height



$$g = |\vec{g}| = g(\varphi, h) \approx 9.8065 \cdot [1 - 2.6373 \times 10^{-3} \cdot \cos(2\varphi) - 5.9 \times 10^{-6} \cdot \cos^2(2\varphi)] \times \times [1 - 3.14 \times 10^{-7} \cdot h]$$

Boguer formula with h in meter and g in m/s²

g [m/s²]	φ=0°	φ= 45°	φ= 90°
h=0 km	9.780	9.807	9.832
h=30 km	9.689	9.715	9.742

UNIVERSITÄT BREMEN

geopotential height

- Definition of geopotential height z (GPH):

$$z = \frac{1}{g} \int_{h'=0}^{h} g(\varphi, h')dh' \equiv \frac{\phi(h)}{g_o} \qquad \text{(Eq. II.12)}$$

$$g_o \equiv 9.8 \ m/s^2$$

$$d\phi = g \cdot dh = g_o \cdot dz$$

ϕ=45°:	h=5 km	→	z=4.996 km
	h=50 km	→	z=49.607 km
	h=500 km	→	z=463.597 km

→ at lower altitudes geopotential height z is almost identical to geometric height h.
→ at surfaces of equal geopotential heights the gravity force remains the same.
→ this is important for energy considerations in large-scale motion (see atmospheric dynamics)

- hypsometric equation is an important tool for weather charts:
  ➔ starting from Eq. II.8

$$\bar{T} = -\frac{g}{R_L} \frac{h_2 - h_1}{\ln p(h_2) - \ln p(h_1)} =$$

$$\Delta h = h_2 - h_1 \qquad = \frac{+g\,\Delta h}{R_L\left[\ln p(h_1) - \ln p(h_2)\right]} =$$

$$\Delta\phi = g \cdot \Delta h = g_o \cdot \Delta z$$

$$= \frac{g_o}{R_L} \frac{z_2 - z_1}{\ln p(z_1) - \ln p(z_2)}$$

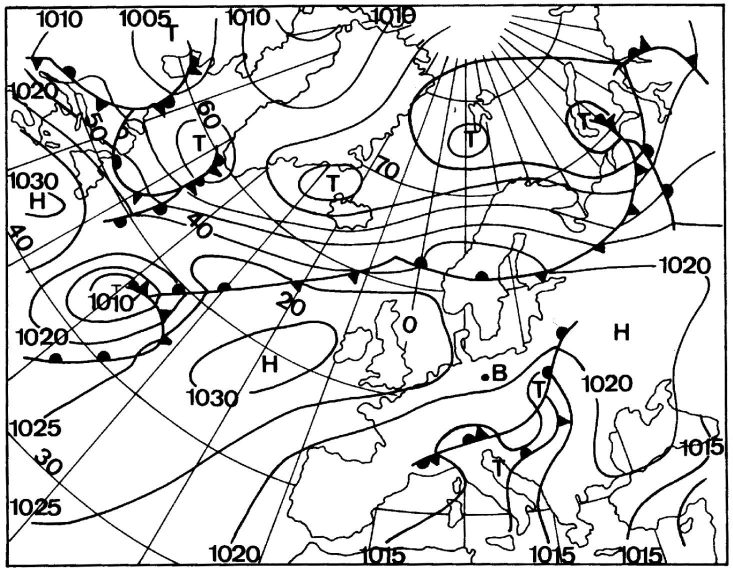
OR

$$\bar{T} = \frac{g_o}{R_L\left(\ln p_1 - \ln p_2\right)} \cdot (z_2 - z_1)$$

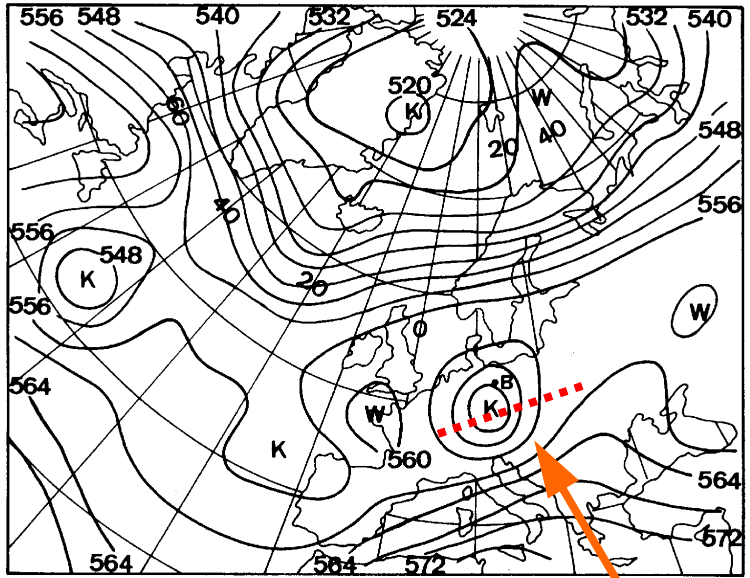
This is the hypsometric equation

➔ difference in geoptential heights (z_2-z_1) from two pressure levels (p_1 and p_2) is called relative topography and is proportional to the mean temperature between these two layers

relative topography: cold droplet



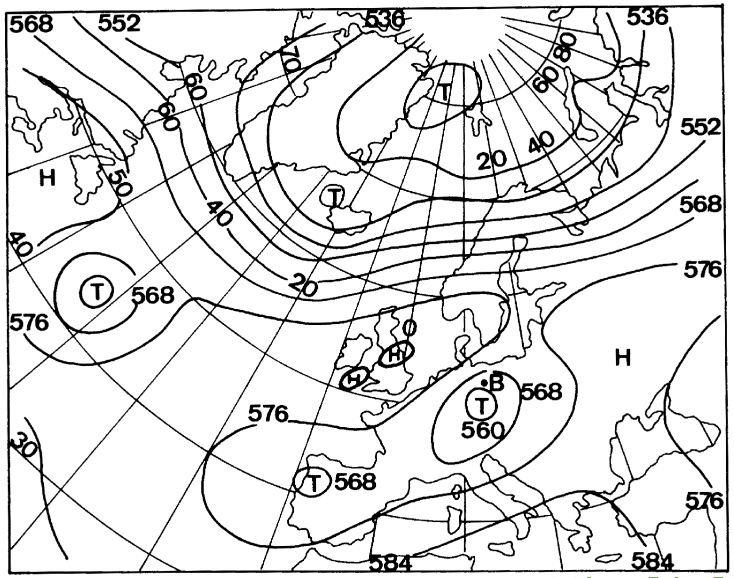
a surface pressure [hPa]



c GPH difference between 500 and 1000hPa [dm]

T low
H high

K cold
W warm



b GPH at 500hPa[dm]

Abb. 7.19 a – c. Karten vom 7. Juni 1983. **a** Bodenwetterkarte; **b** 500-hPa-Höhenkarte; **c** relative Topographie 500/1000 hPa

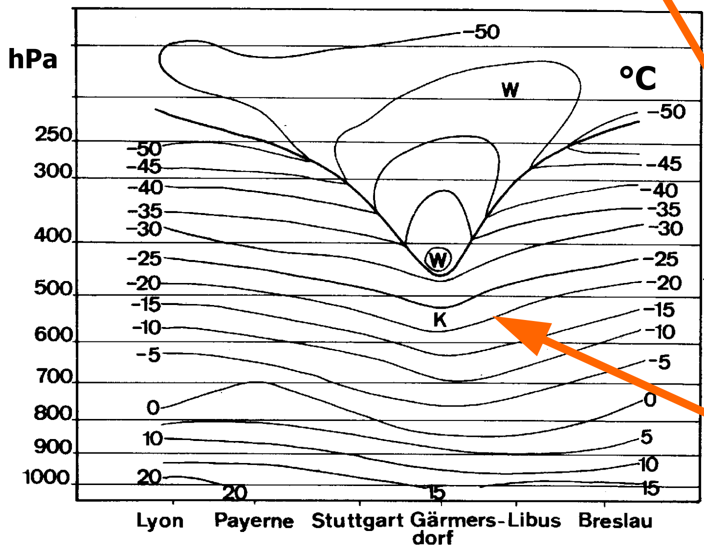


Abb. 7.20. Thermische Struktur eines Kaltlufttropfens im Vertikalschnitt

cold droplet

Malberg 1997

relative topography: cold droplet

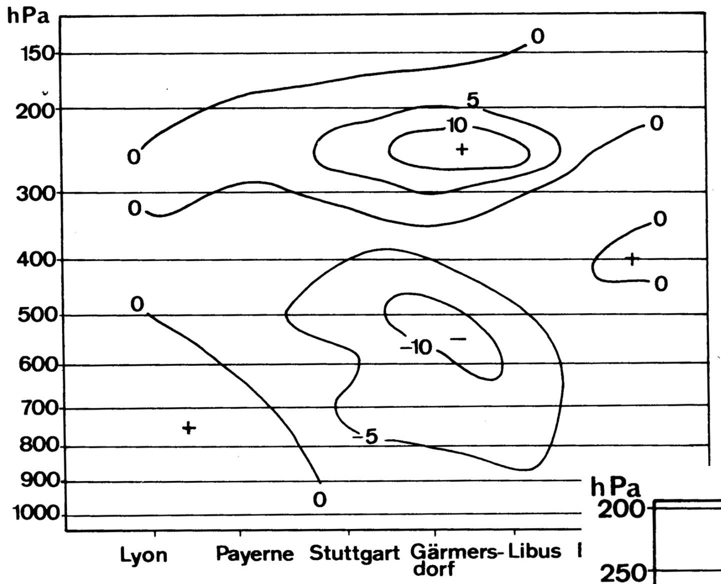


Abb. 7.21. Temperaturdifferenz eines Kaltlufttropfens

temperature anomaly in °C

dry condition (clouds dissolving)

high convection (clouds, thunderstorm)

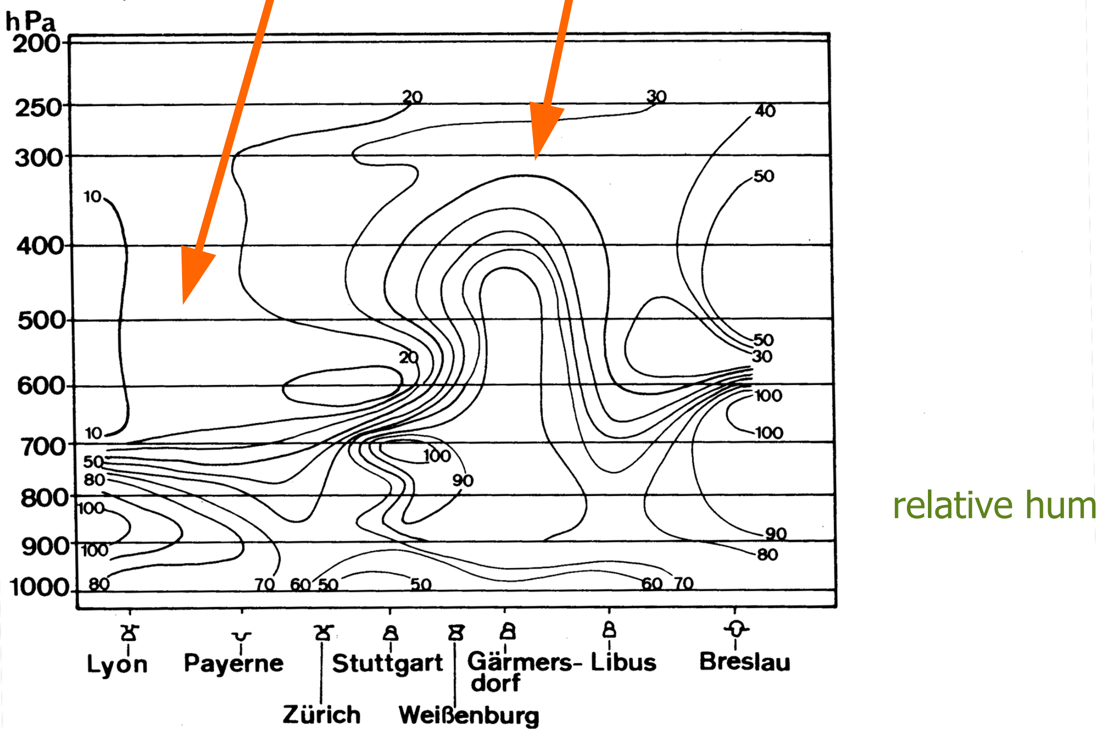


Abb. 7.22. Vertikalschnitt der relativen Feuchte durch einen Kaltlufttropfen (7. Juni 1973)

relative humidity in %

Malberg 1997