Web page: http://geophysics.tau.ac.il/personal/nili/GCirc/GCirc-course.htm

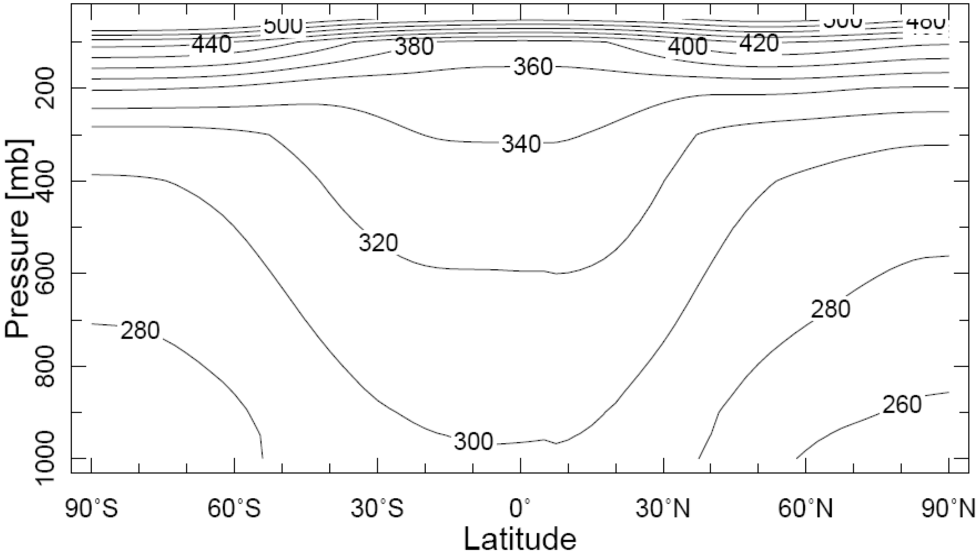
References:

Main text:

- Geoffrey Vallis (2006) "Atmospheric and Oceanic Fluid Dynamics." Cambridge press. Part III- Large scale atmospheric circulation (chapters 11-12). Parts rely on the theory developed in parts I, II.

Other texts and supplemental reading (partial list):

- Richard Lindzen (1990) "Dynamics in atmospheric physics." Cambridge press. In particular the chapter on the Hadley cell (7). Chapter 5 on the observed circulation is also good.
- Holton, (1992, 2004) "An introduction to dynamic meteorology." Elsevier, Academic Press. Chapter on the general circulation (10).
- Dennis Hartmann (1994), "Global Physical Climatology".
- John Marshall and Alan Plumb, (2007), "Atmosphere, Ocean and Climate dynamics". An introductory level text book with nice rotating tank experiments and a good basic physical understanding.
- Tapio Schneider (2006) The general circulation of the atmosphere, Annu. Rev. Earth. Planet. Sci. A review article which can be found online at http://www.gps.caltech.edu/~tapio/pubs.html
- Edward N. Lorenz (1967) "The nature and Theory of the general circulation of the atmosphere." WMO monograph 218. Includes a historical overview of general circulation theories.
- Isaac Held (2000) The general circulation of the atmosphere. Woodshole summer school class notes, found at http://www.gfdl.noaa.gov/~ih/. General circulation theory.
- Peixoto and Oort (1992) "Physics of Climate." AIP. Introduction to the equations and general circulation "observations".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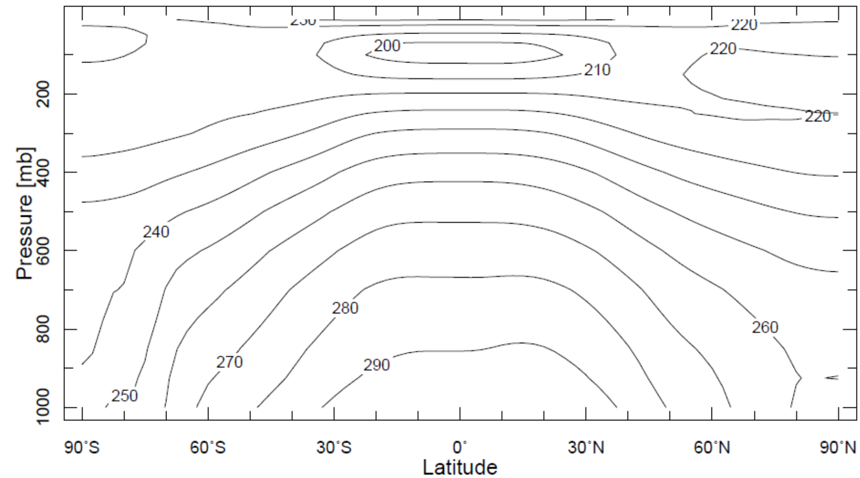
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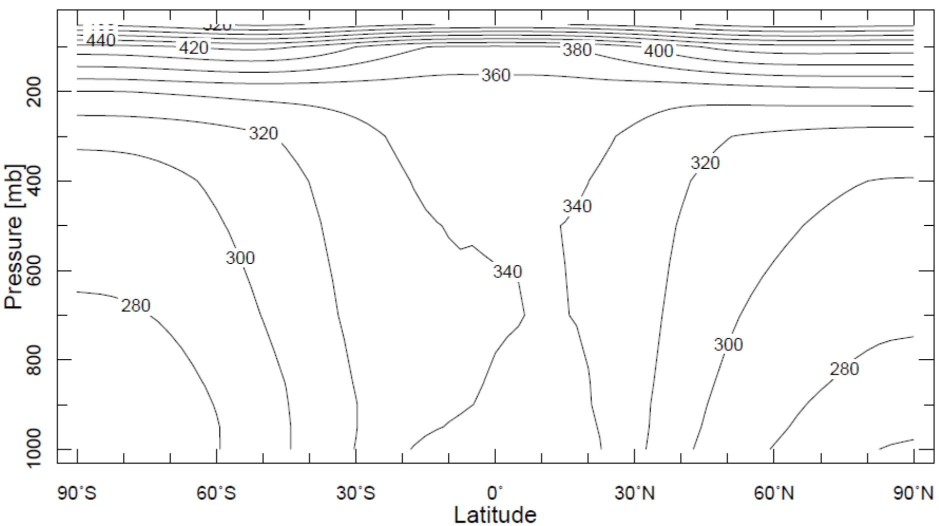
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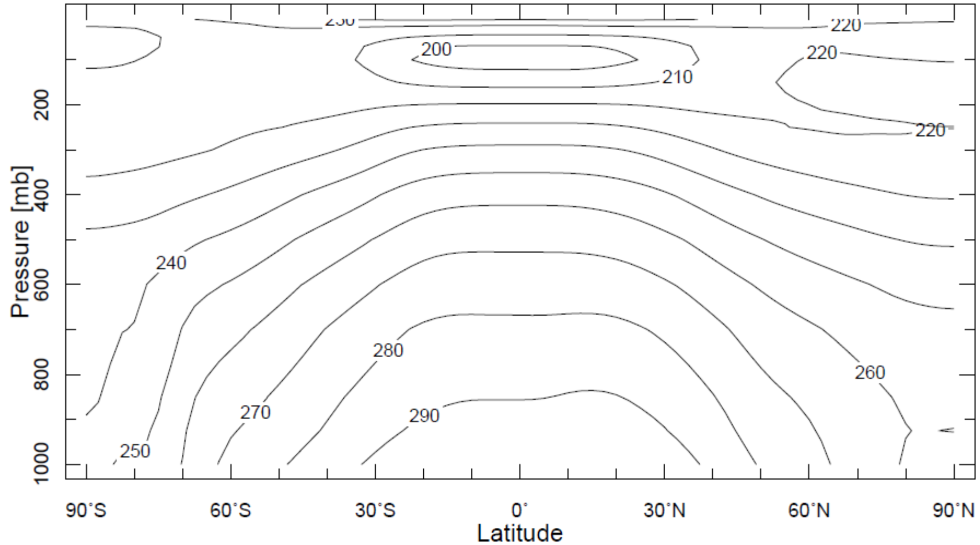

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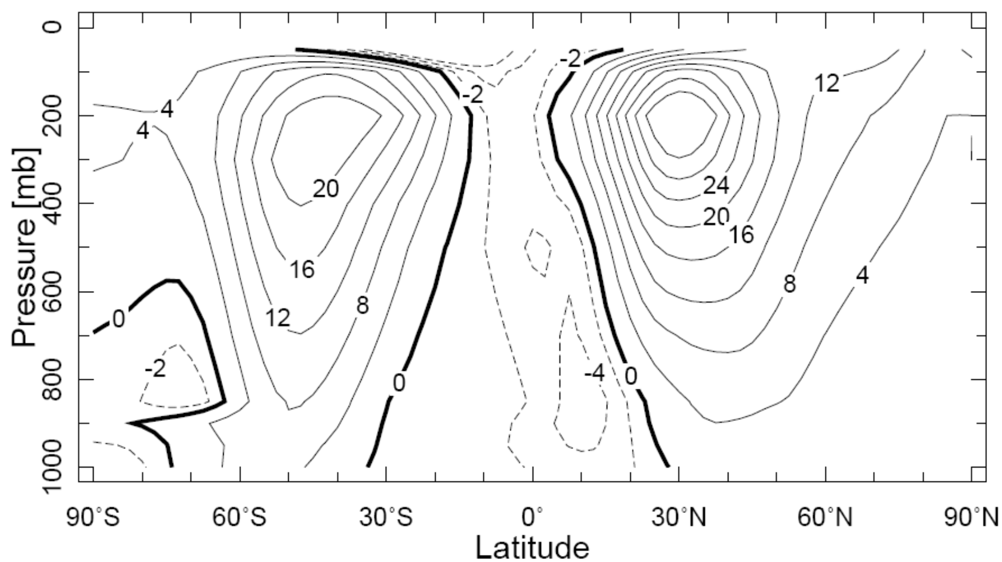

4
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Pressure [mb]
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Latitude

Vallis (2006)

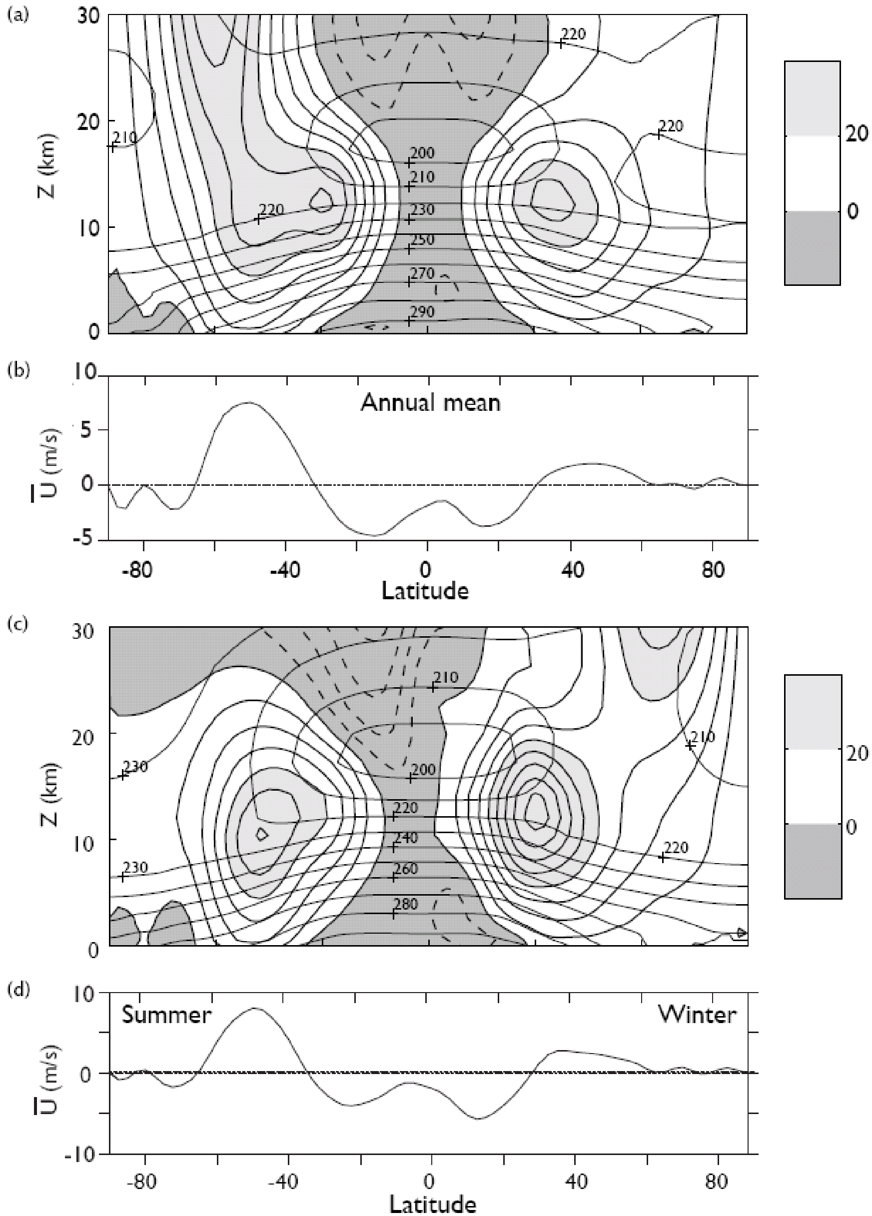
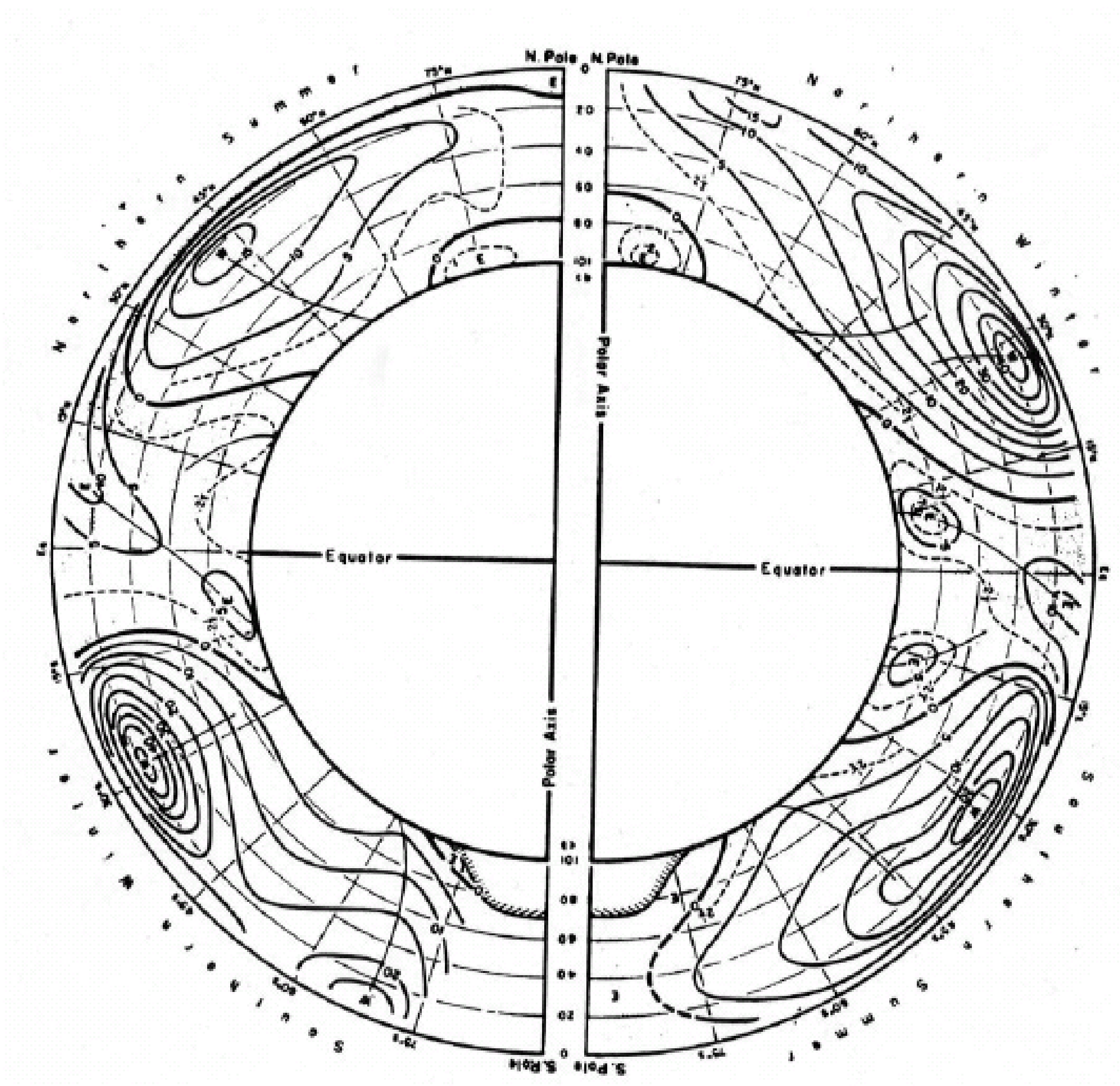


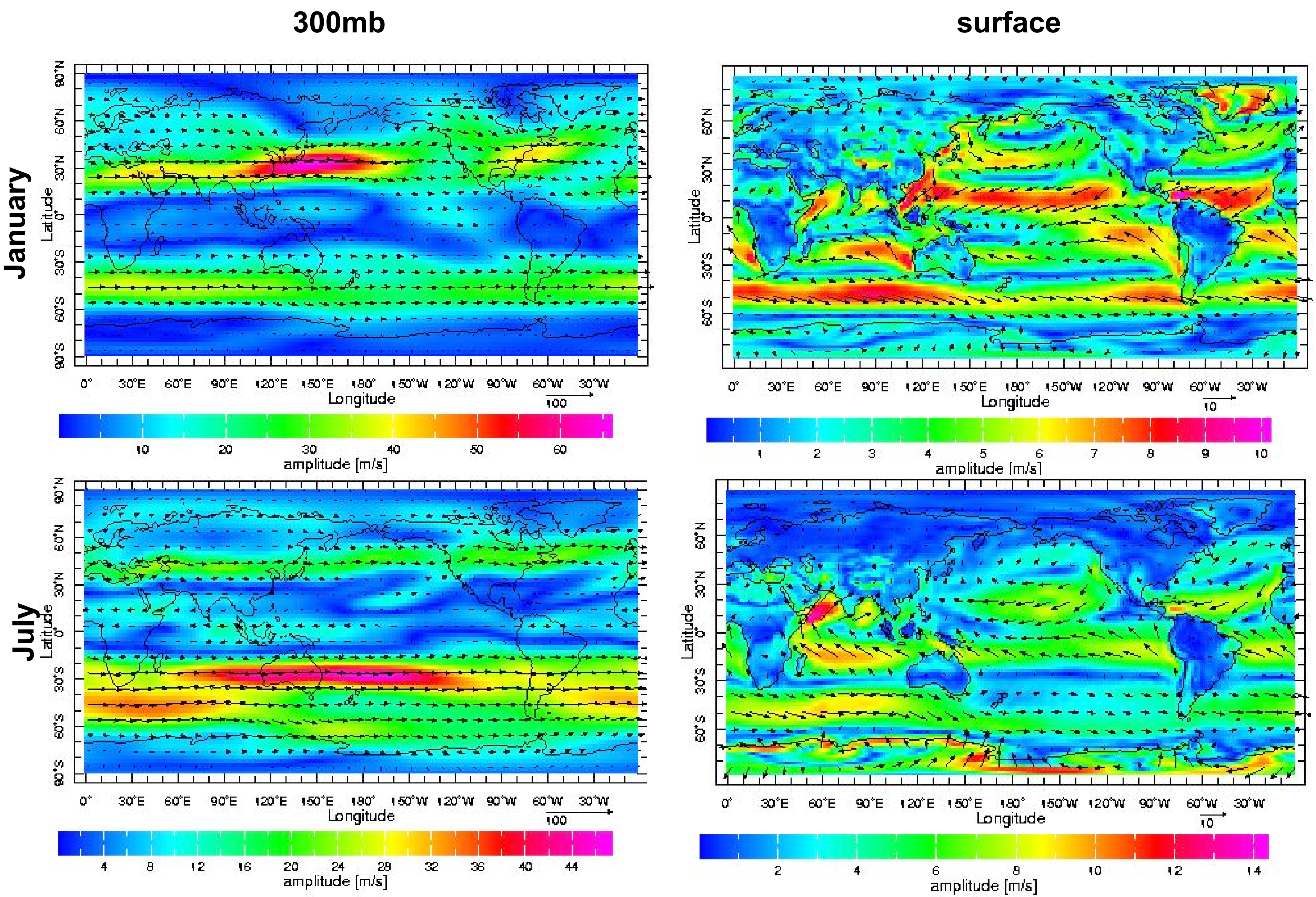
Fig. 11.2 (a) Annual mean, zonally averaged zonal wind (heavy contours and shading) and the zonally averaged temperature (lighter contours). (b) Annual mean, zonally averaged zonal winds at the surface. (c) and (d) Same as (a) and (b), except for northern hemisphere winter (DJF). The wind contours are at intervals of $5\,\mathrm{m\,s^{-1}}$ with shading for eastward winds above $20\,\mathrm{m\,s^{-1}}$ and for all westward winds, and the

Thermal wind balance

Vallis (2006)

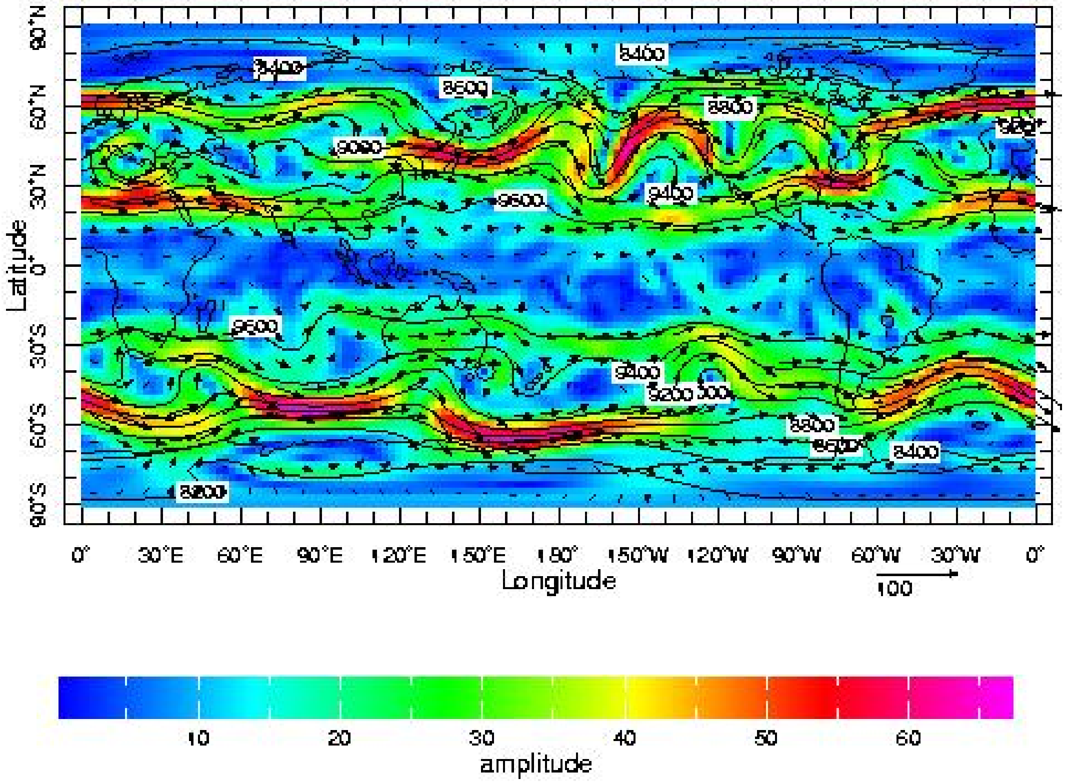


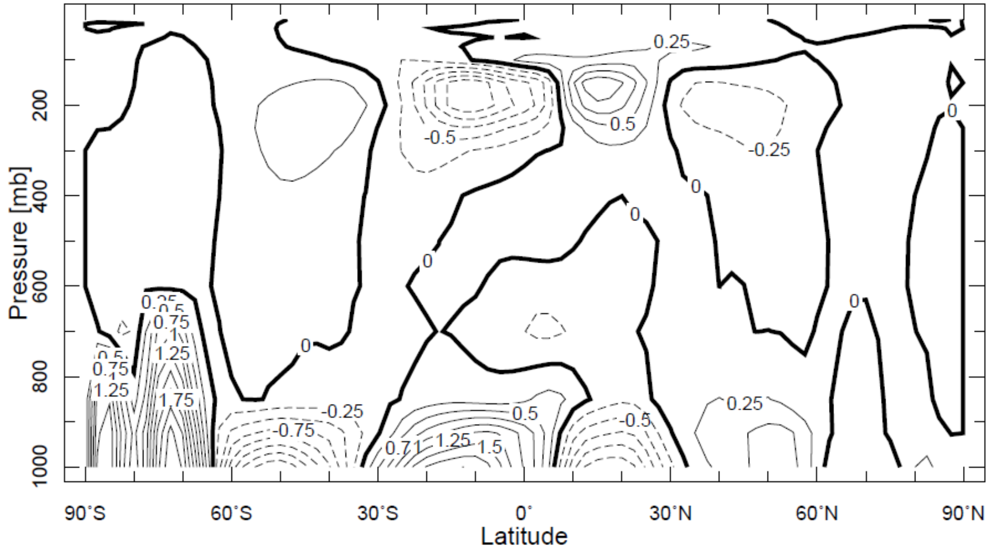
Palmen and Newton (1969)

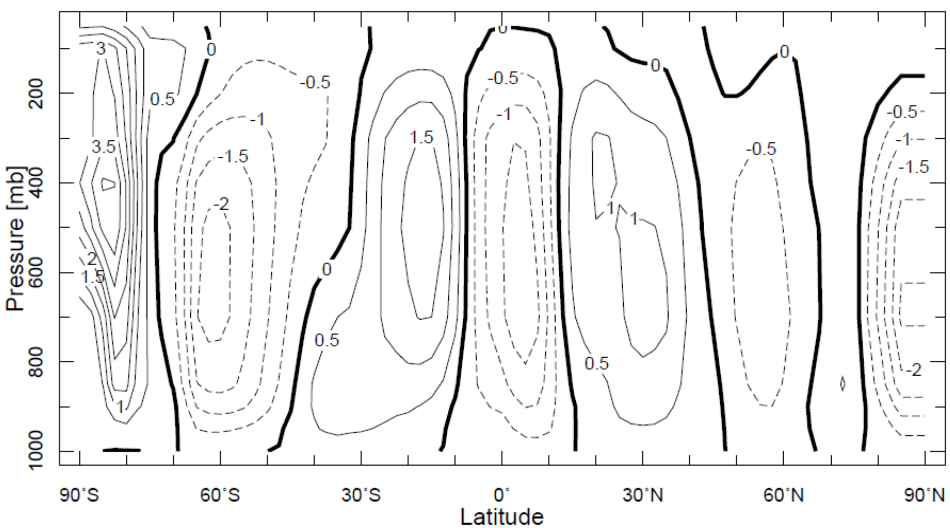

300mb
surface
January
July
Latitude
Longitude
90°N 60°N 30°N 0° 30°S 60°S 90°S
0° 30°E 60°E 90°E 120°E 150°E 180° 150°W 120°W 90°W 60°W 30°W
100
10
10 20 30 40 50 60
1 2 3 4 5 6 7 8 9 10
4 8 12 16 20 24 28 32 36 40 44
2 4 6 8 10 12 14
amplitude [m/s]

Midlatitude weather systems

NOAA NCEP-NCAR CDAS-1 DAILY
300 mb height (m) and winds (m/s) 1 Apr 1997



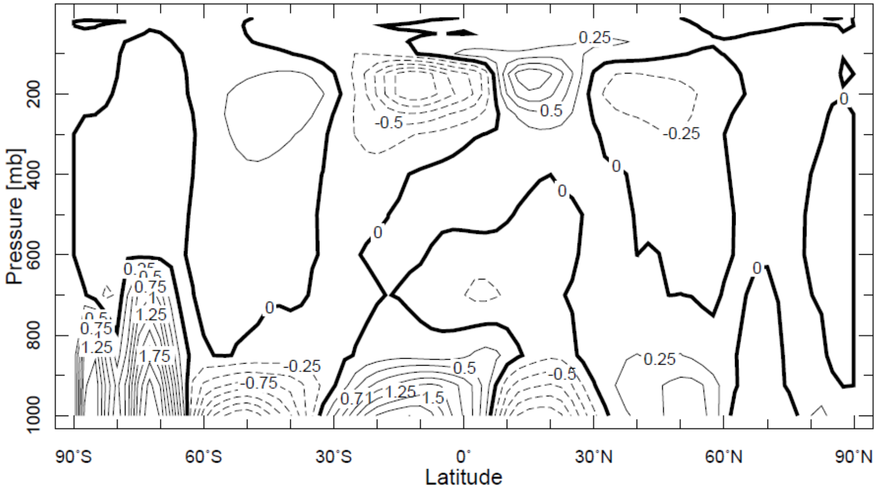




ממוצע זונאלי ושנתי של המהירות המרידיונאלית. יחידות: m/sec.

זכרו: מהירות חיובית היא זרימה צפונה ושלילית זרימה דרומה.

ממוצע זונאלי ושנתי של מהירות הלחץ האנכית ($\omega = \frac{dp}{dt}$. 10^{-4}mb/sec



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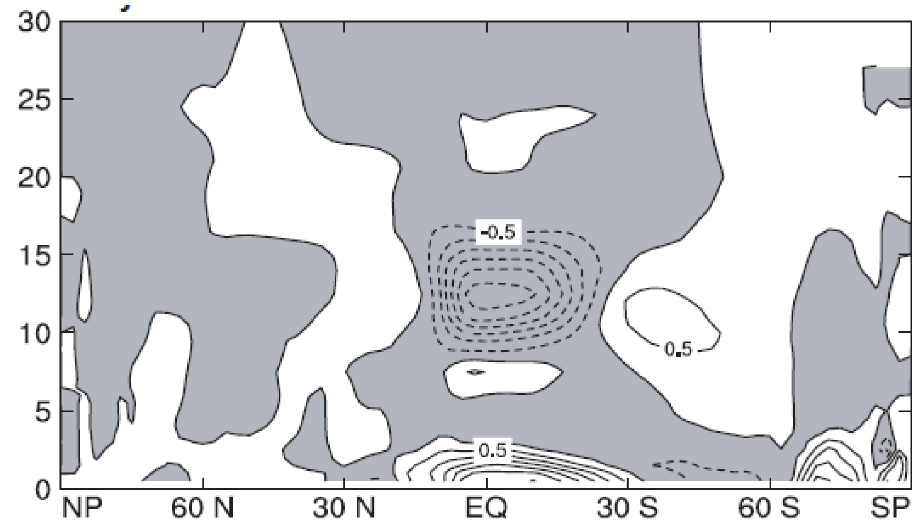
90°N
Latitude



30
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0.5
NP
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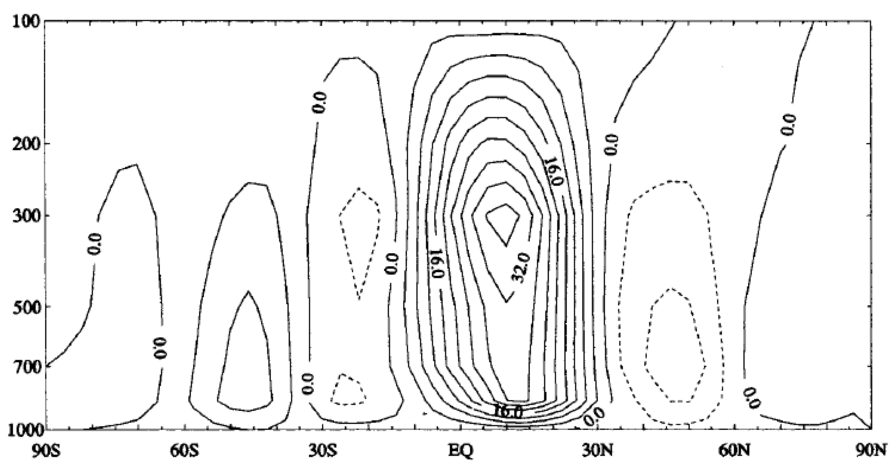
30 N
EQ
30 S
60 S
SP



100
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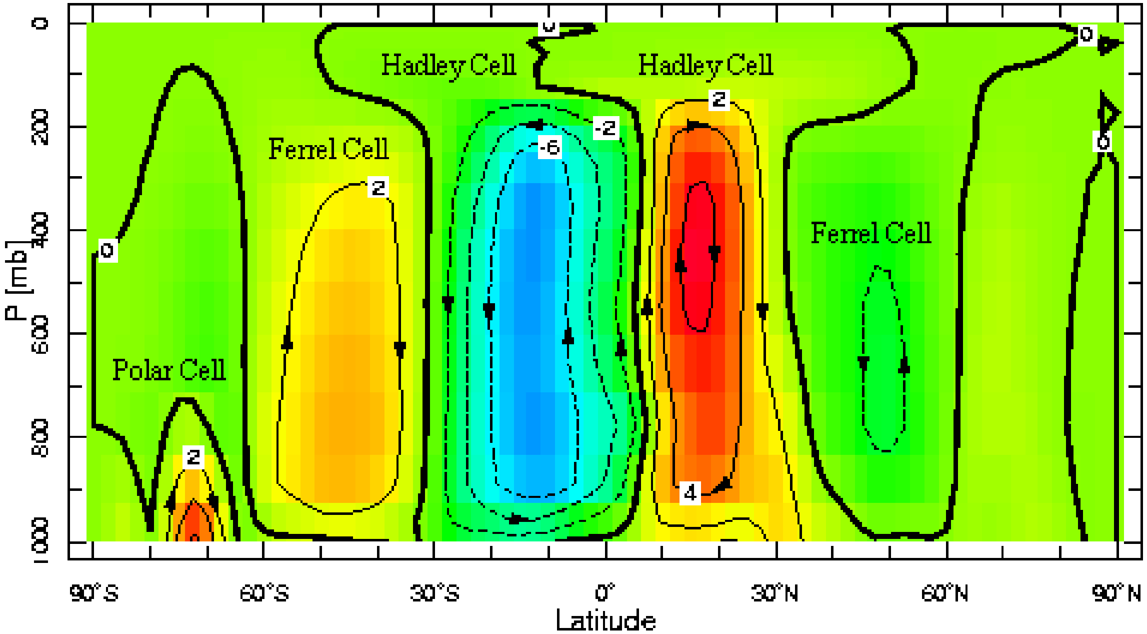
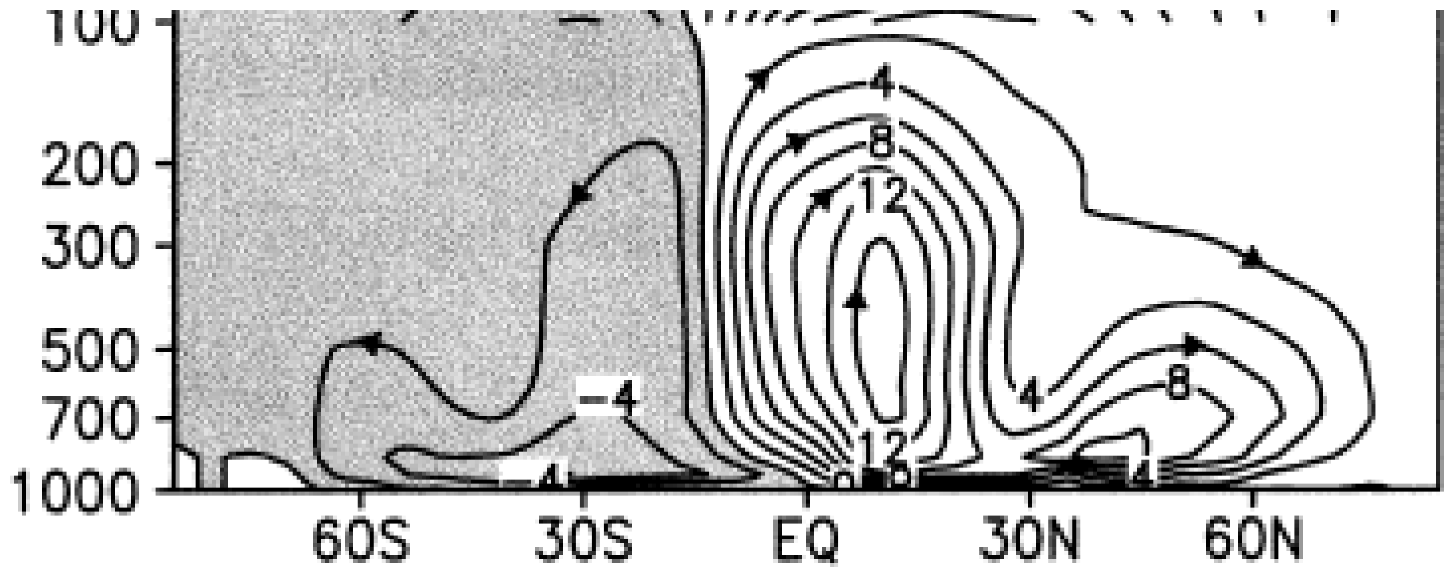
90N

The Zonally Averaged vertical-meridional Circulation



The annually-averaged atmospheric circulation in the latitude pressure plane (the meridional plan). The arrows depict the direction of air movement in the meridional plane. The contour interval is 2x10 10 Kg/sec - this is the amount of mass that is circulating between every two contours. The total amount of mass circulating around each "cell" is given by the largest value in that cell. Data based on the NCEP-NCAR reanalysis project 1958-1998.



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4
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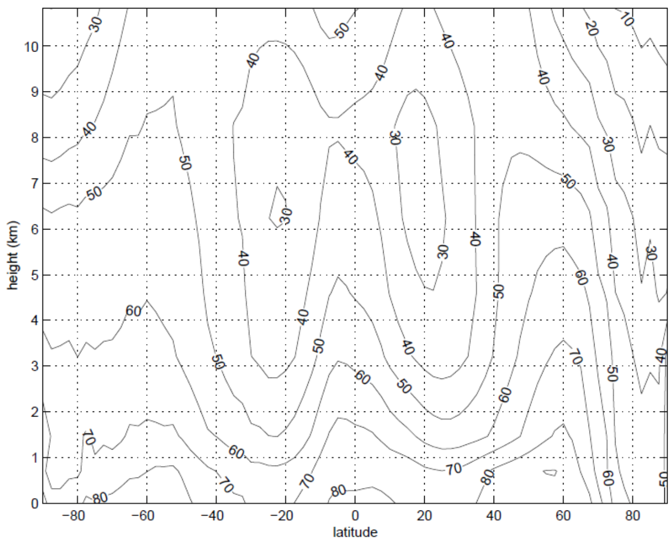
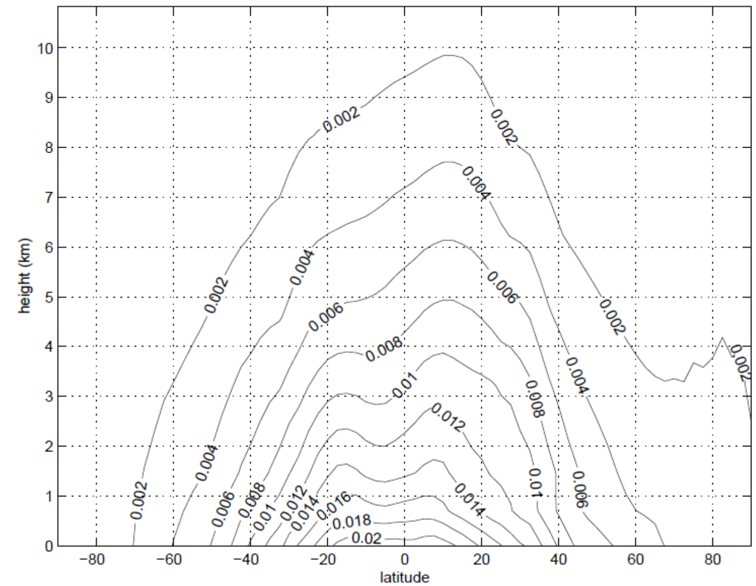
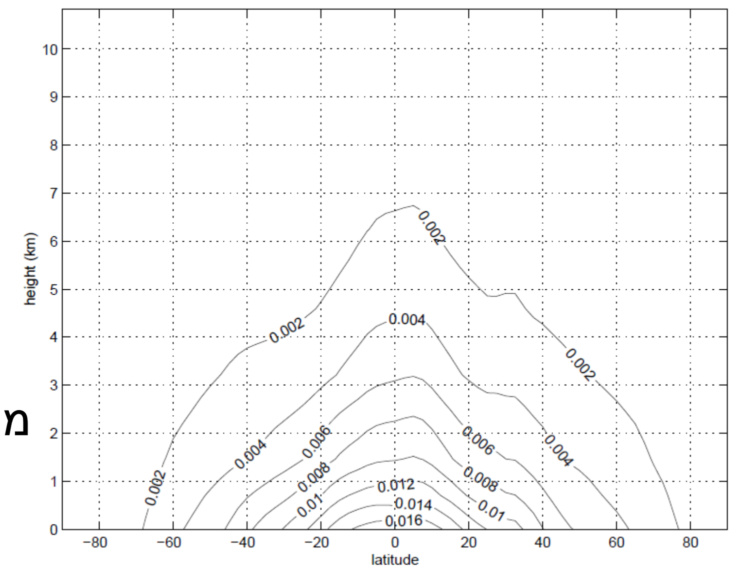
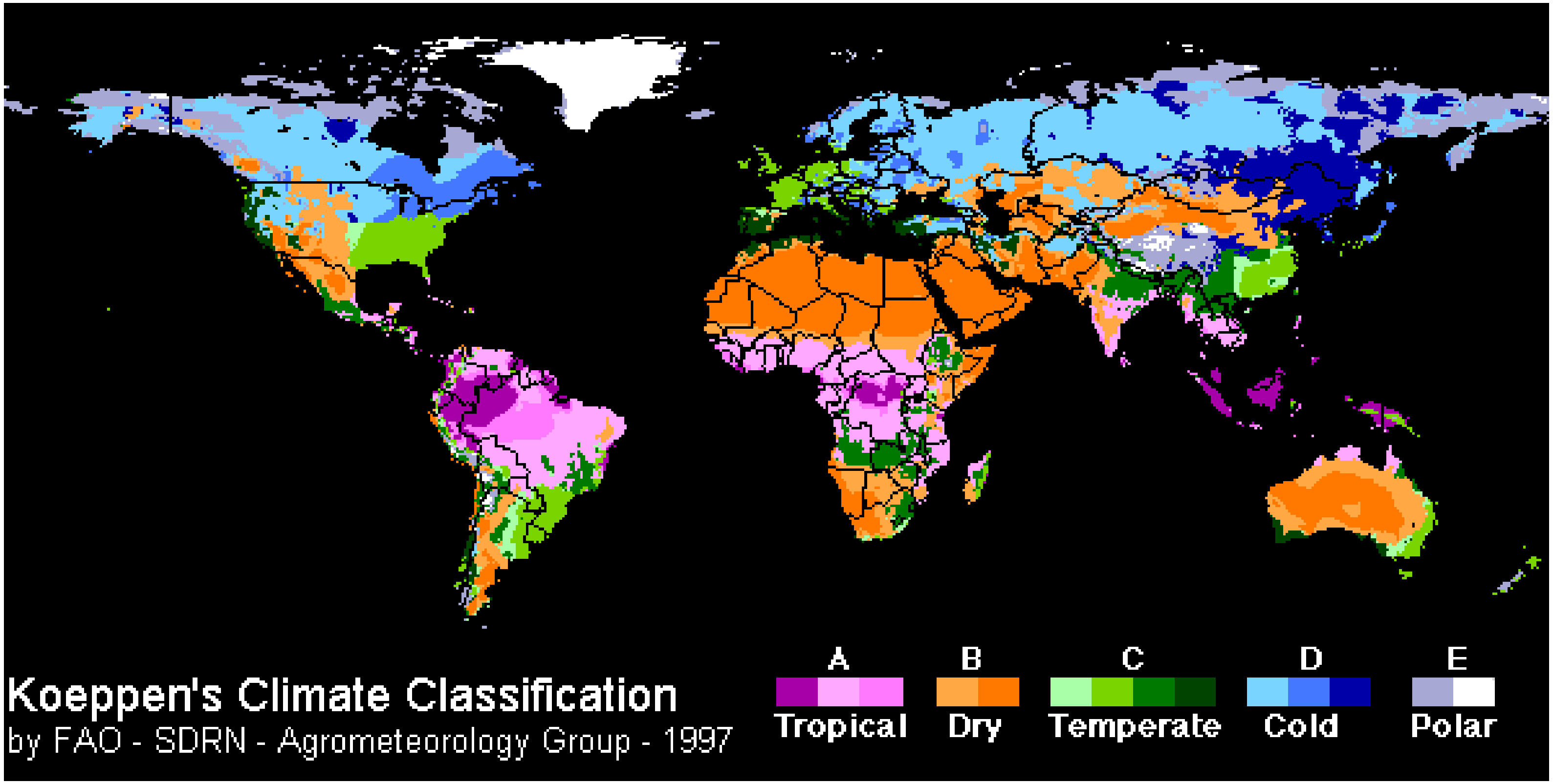




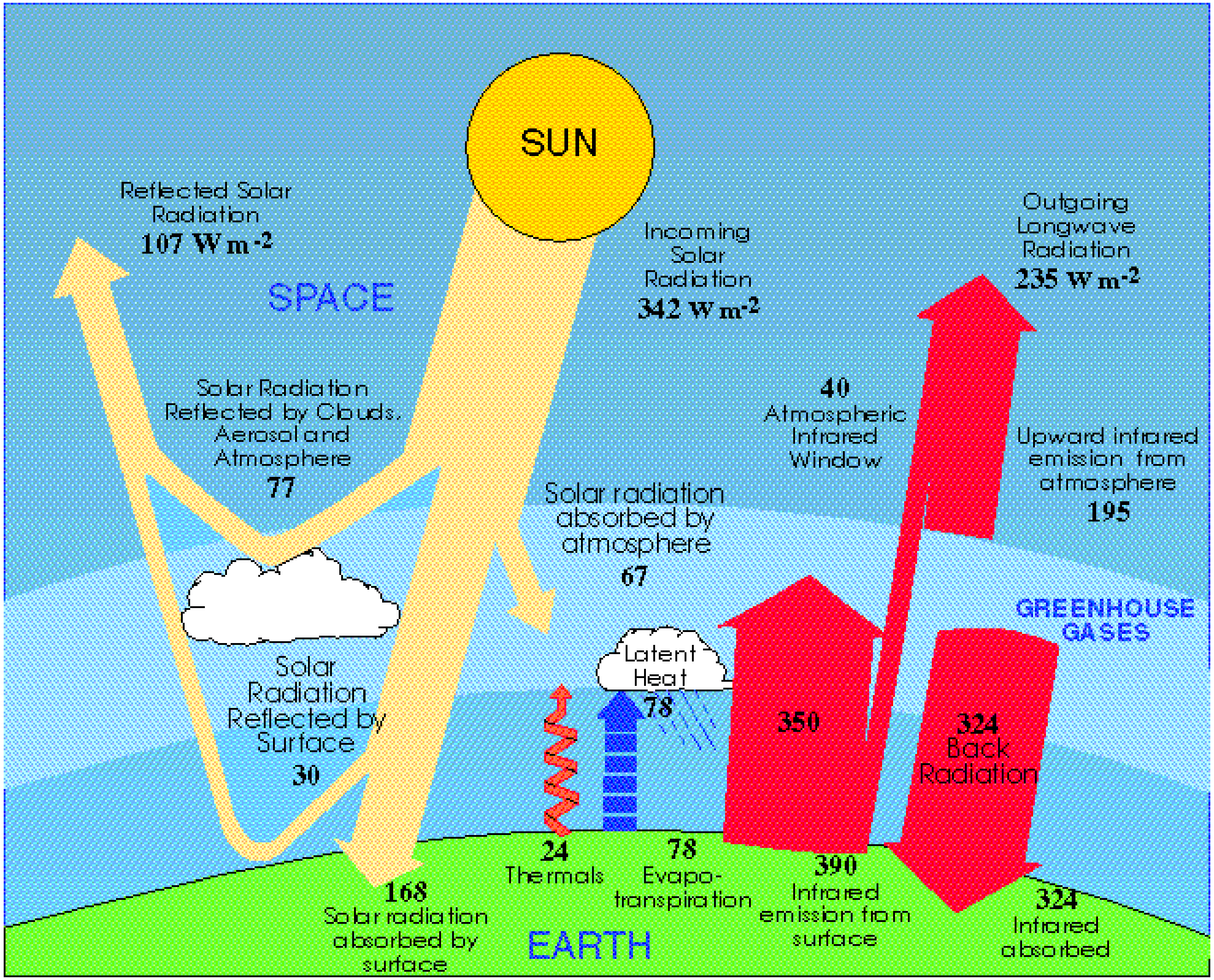


ממוצע זונאלי ושנתי של הלחות הסגולית. חסר יחידות.
ממוצע זונאלי ושנתי של הלחות הסגולית ברוויה. חסר יחידות.
ממוצע זונאלי ושנתי של הלחות היחסית. יחידות: %.

Climate Zones according to Koeppen



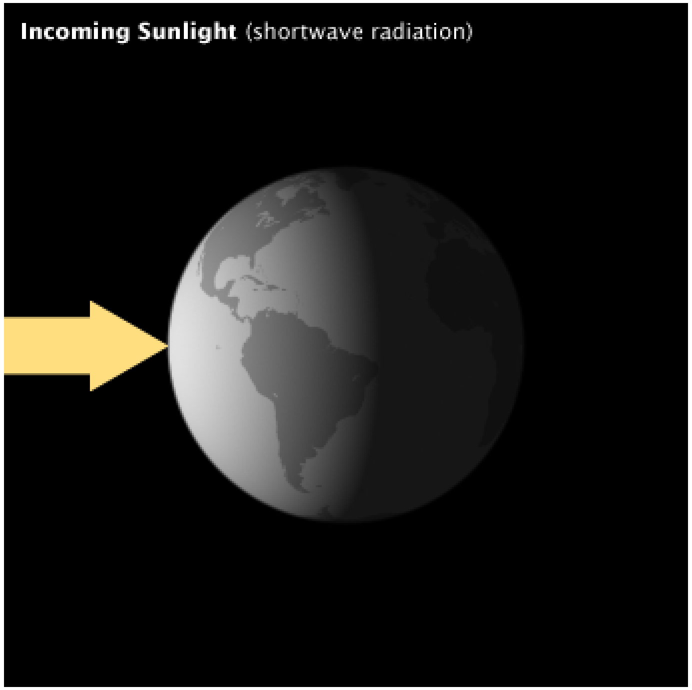
- For more information about this map see:
- http://www.blueplanetbiomes.org/climate.htm

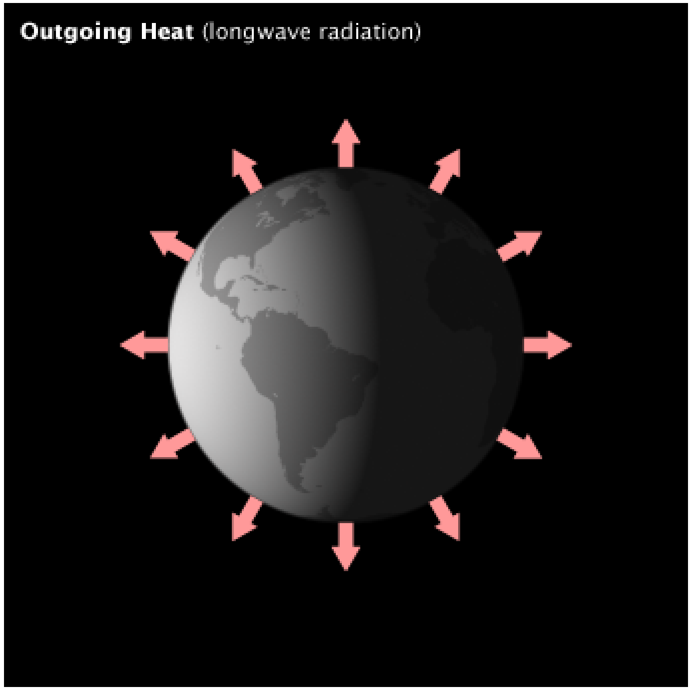

SUN
Reflected Solar Radiation
107 W m-2
SPACE
Incoming Solar Radiation
342 W m-2
Outgoing Longwave Radiation
235 W m-2
Solar Radiation Reflected by Clouds, Aerosol and Atmosphere
77
40
Atmospheric Infrared Window
Upward infrared emission from atmosphere
195
Solar radiation absorbed by atmosphere
67
GREENHOUSE GASES
Latent Heat
78
Solar Radiation Reflected by Surface
30
350
324
Back Radiation
24
Thermals
78
Evapo-transpiration
390
Infrared emission from surface
324
Infrared absorbed
168
Solar radiation absorbed by surface
EARTH

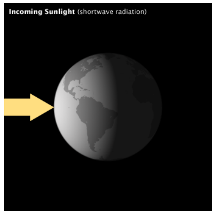
I assume you already know that :

Energy balance - what comes in must go out

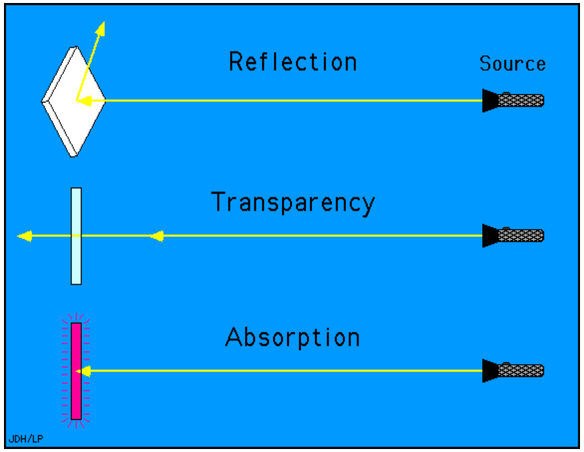
Earth is heated by short wave (energetic) solar radiation and cooled by emitting less energetic, long wave infra-red (IR) radiation.

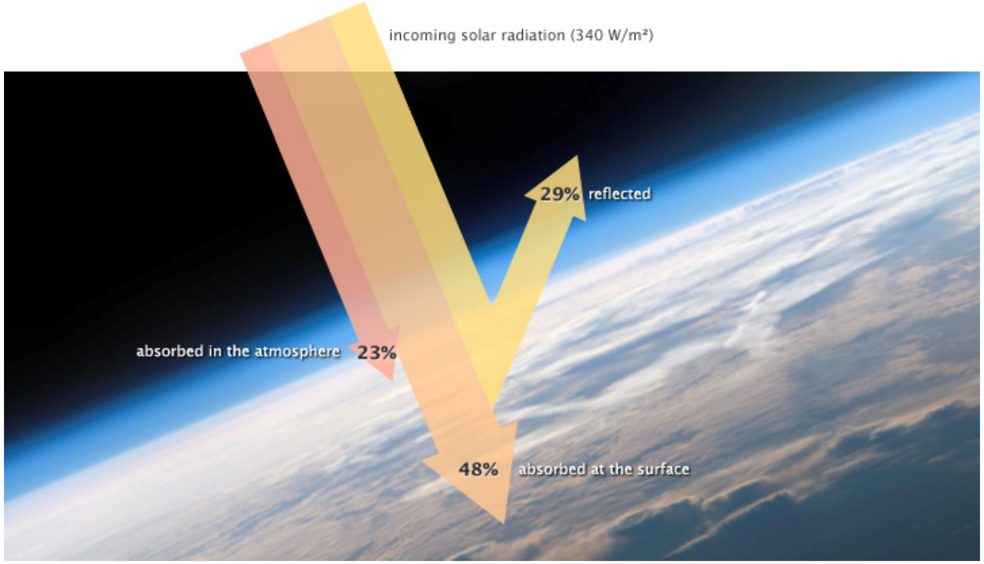






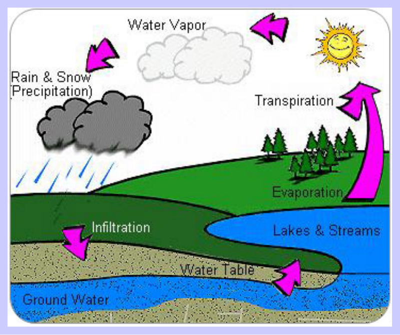
The incoming solar radiation is partly reflected back to space (30%), partly absorbed by the atmosphere, and partly absorbed by the earth (50%).



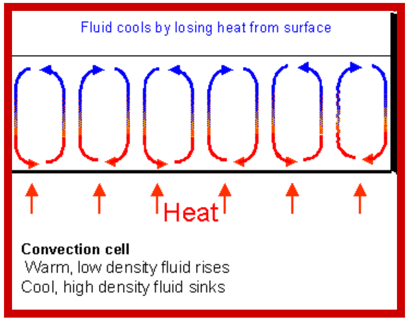


The earth's surface, which absorbs much of the solar radiation, emits heat in 3 ways:

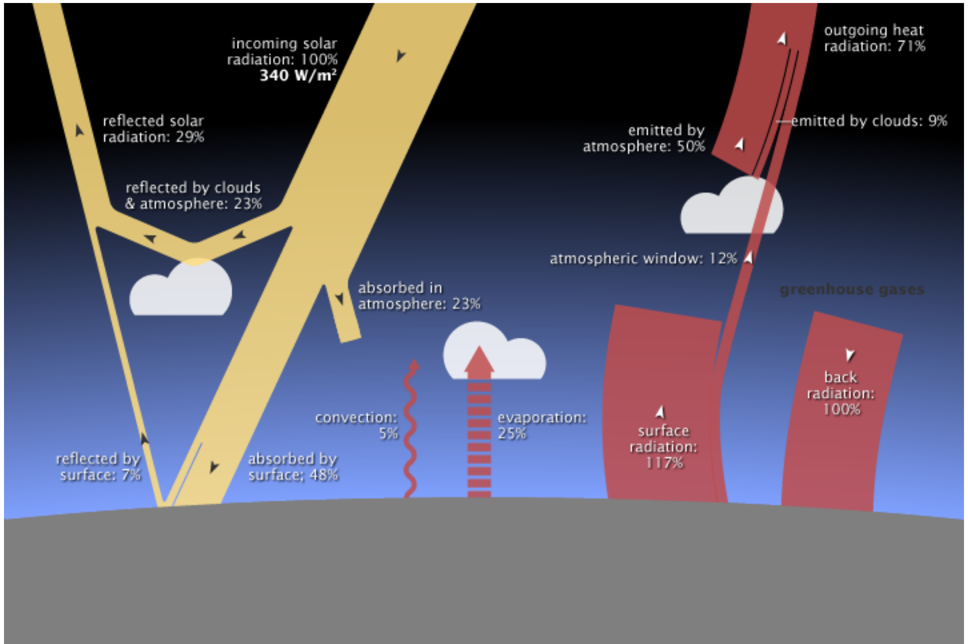
- Latent heat – water cools the surface when it evaporates into the air, and releases heat when it condenses into cloud and rain drops



- Sensible heat – hot buoyant air convection

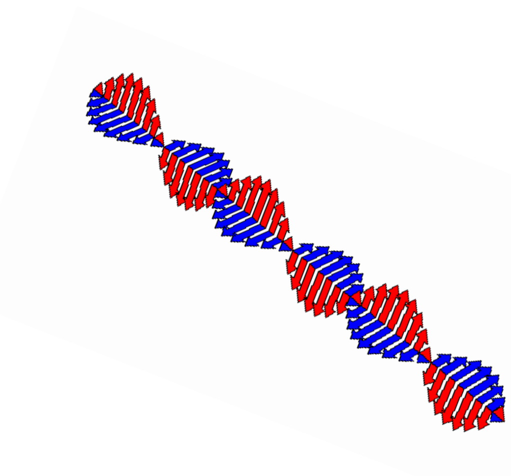


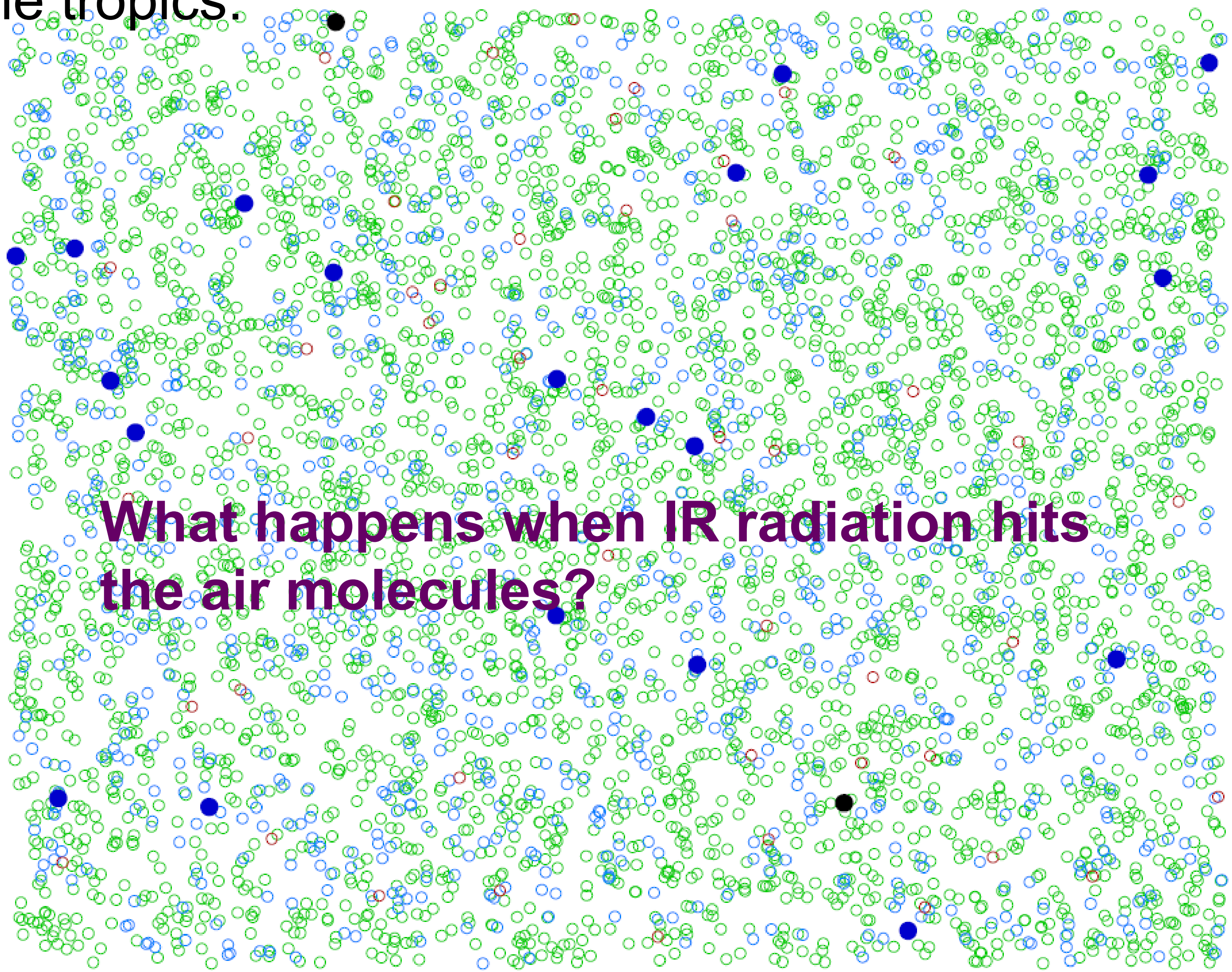
- IR radiation



on which we will concentrate next…

The atmosphere is a mixture of molecular Nitrogen and Oxygen, with a bit of Argon, and trace amounts of other gases, like CO_2. Water varies between ~0.4% in very dry air, to ~4% in the tropics.

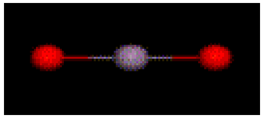


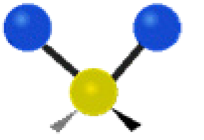


What happens when IR radiation hits the air molecules?

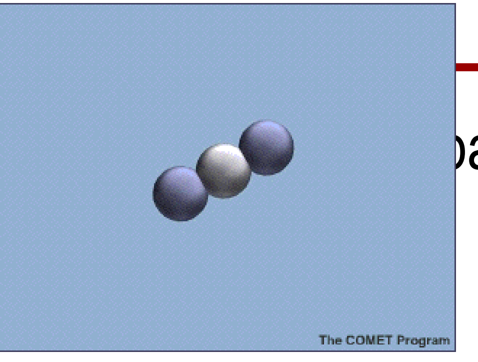
That depends on the type of molecule and radiation

Molecules have energy, stored in the bonds between atoms, and in the relative motions between the different molecule atoms.



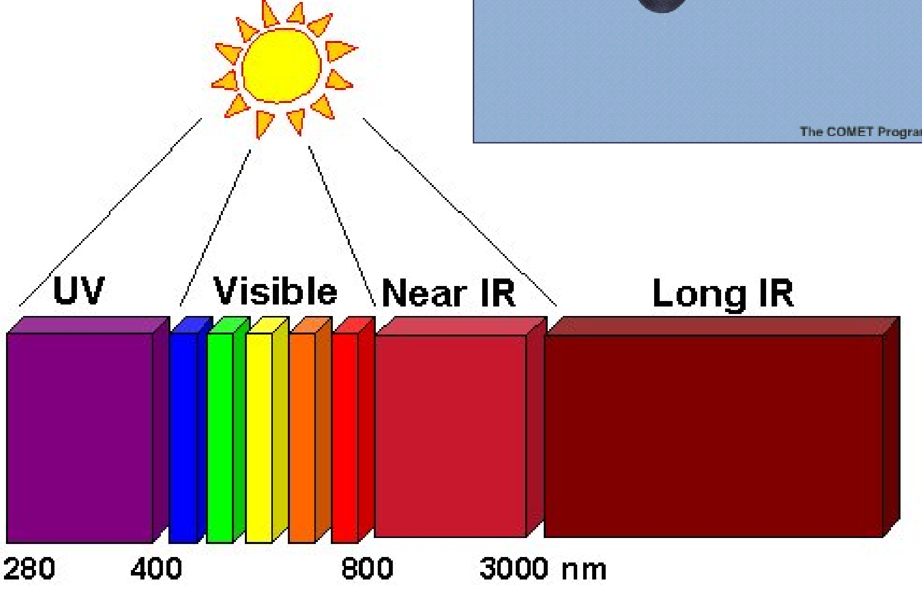


When radiation hits a molecule, it will pass its energy on to the molecule, if the molecule can "contain" this energy.



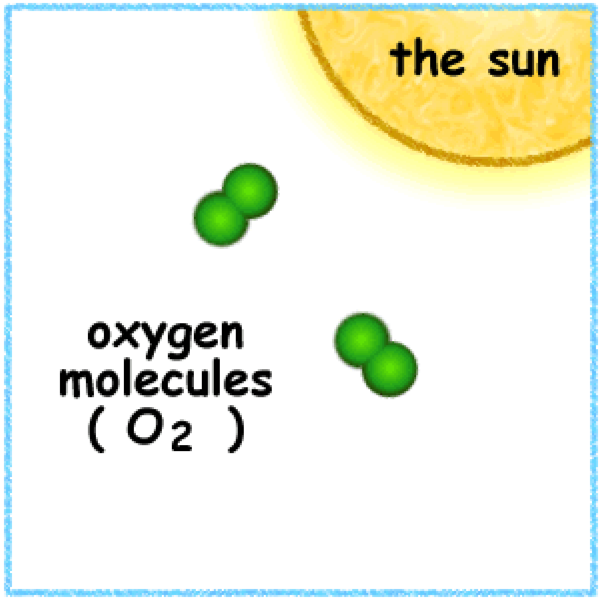
Thus, each mode of energy absorption occurs a … band of the solar spectrum.

Radiation energy spectrum



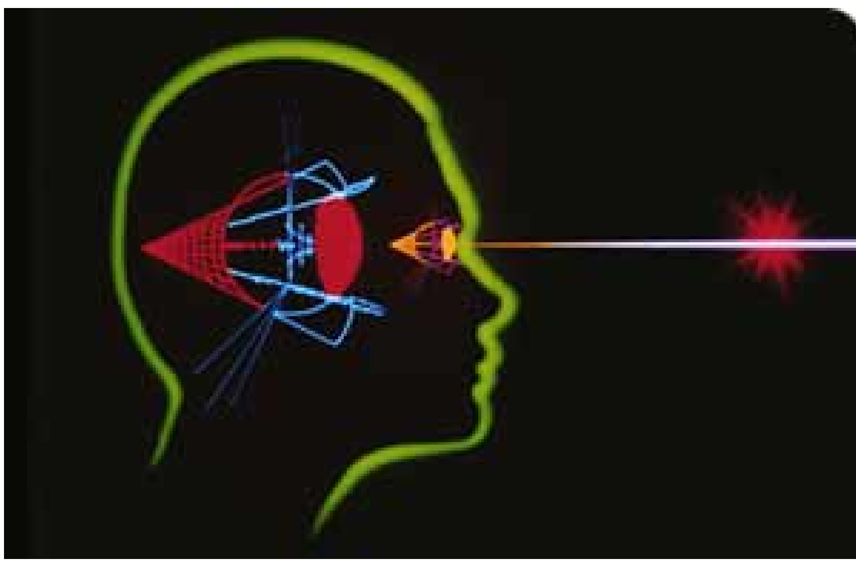
Energetic UV radiation can break apart the molecules.

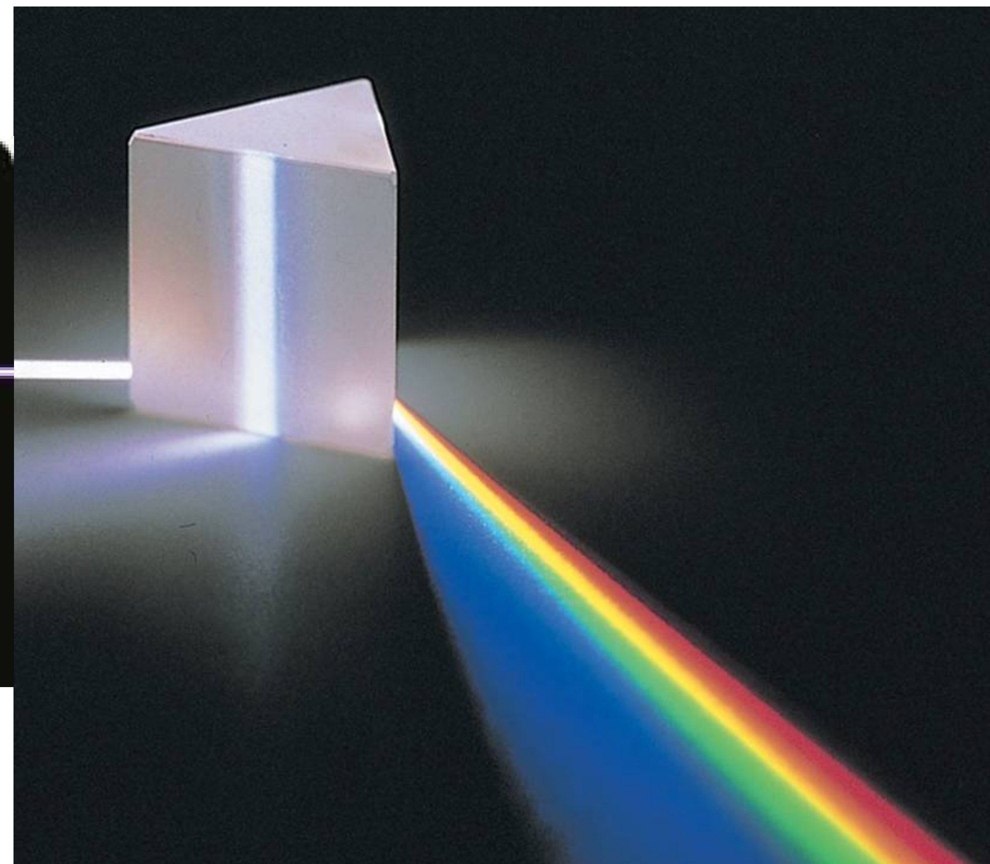
For example: breakup of O_2, to form O_3



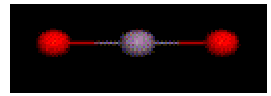
Visible light does not interact with air molecules.

Our eyes developed to make use of this, to see things on earth.

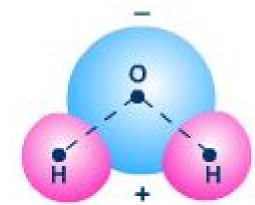




IR radiation gets absorbed by exciting rotational and vibrational modes of specific molecules

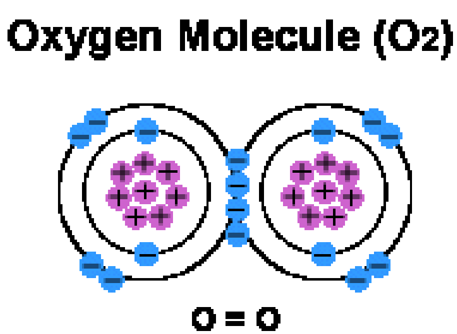


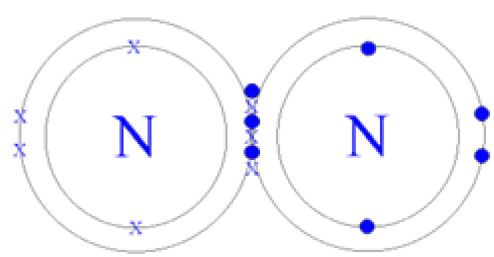
Absorption of electromagnetic radiation requires an electric dipole.



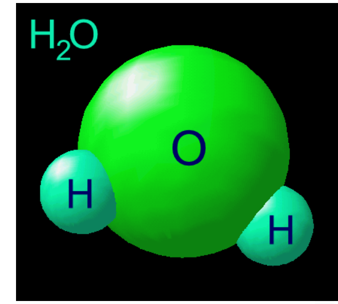
Electric dipoles require some asymmetry in the molecule structure.

→ O_2, N_2 don't interact with IR radiation.

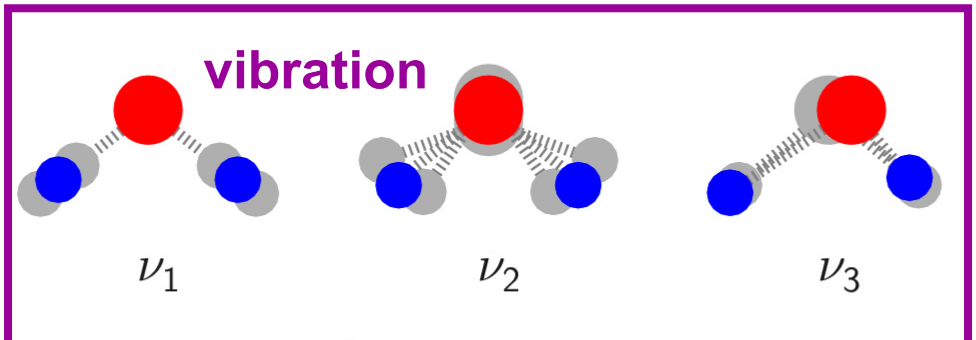


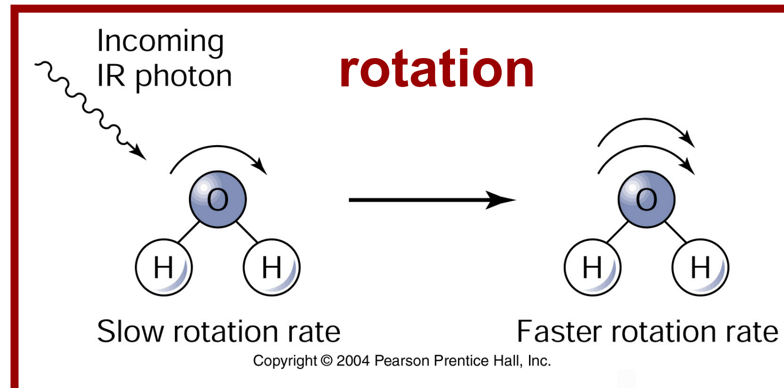


H_2O molecules have a triangular structure, with a dipole.

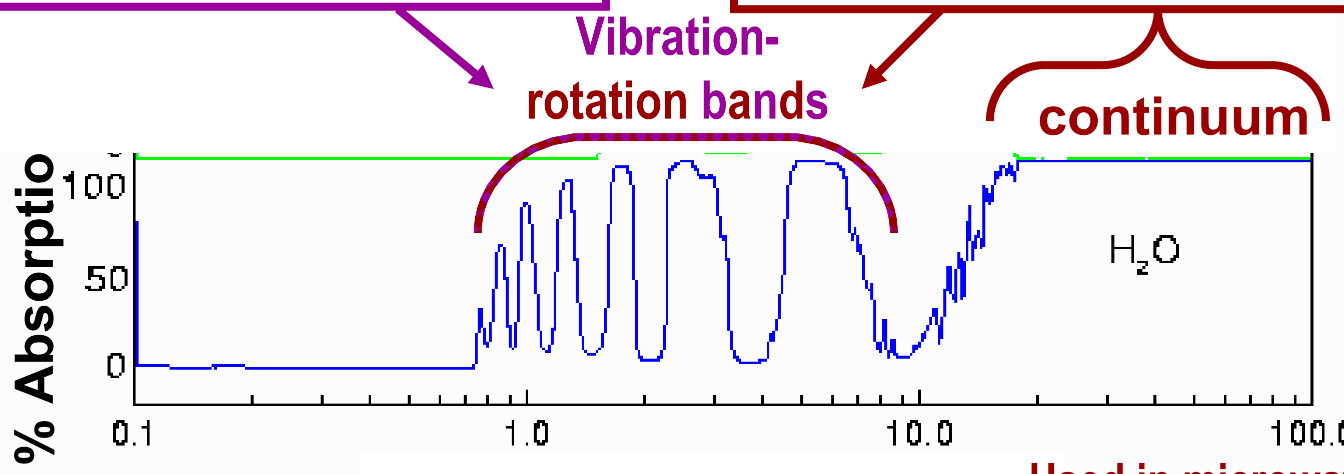


IR is absorbed by exciting **vibrational** and **rotational** energy.



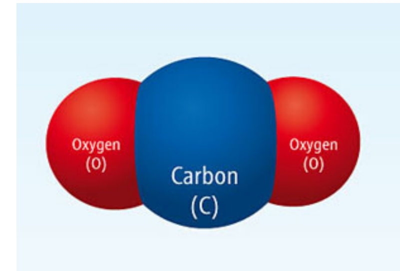


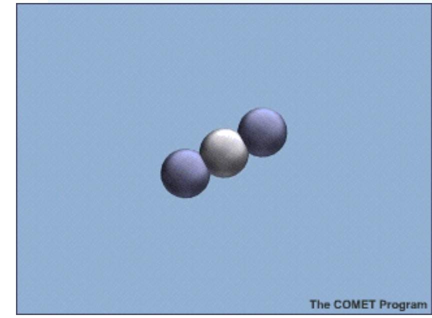
H_2O absorption spectrum



Used in microwave ovens

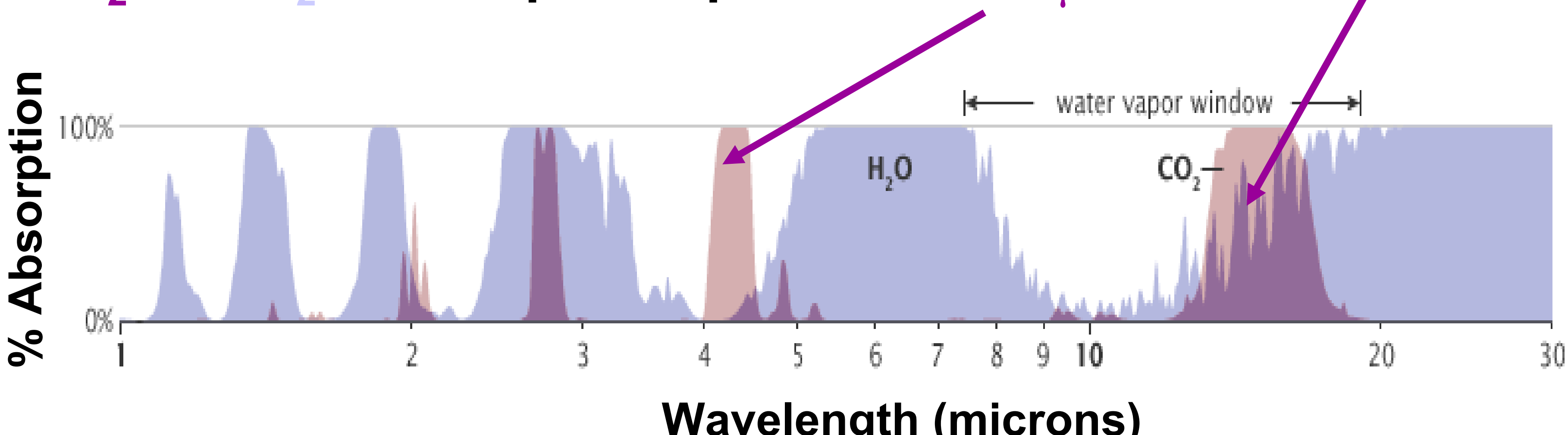
CO_2 molecules have a linear symmetric structure, with no dipole, but when they vibrate, the symmetry breaks, a dipole forms and IR can be absorbed



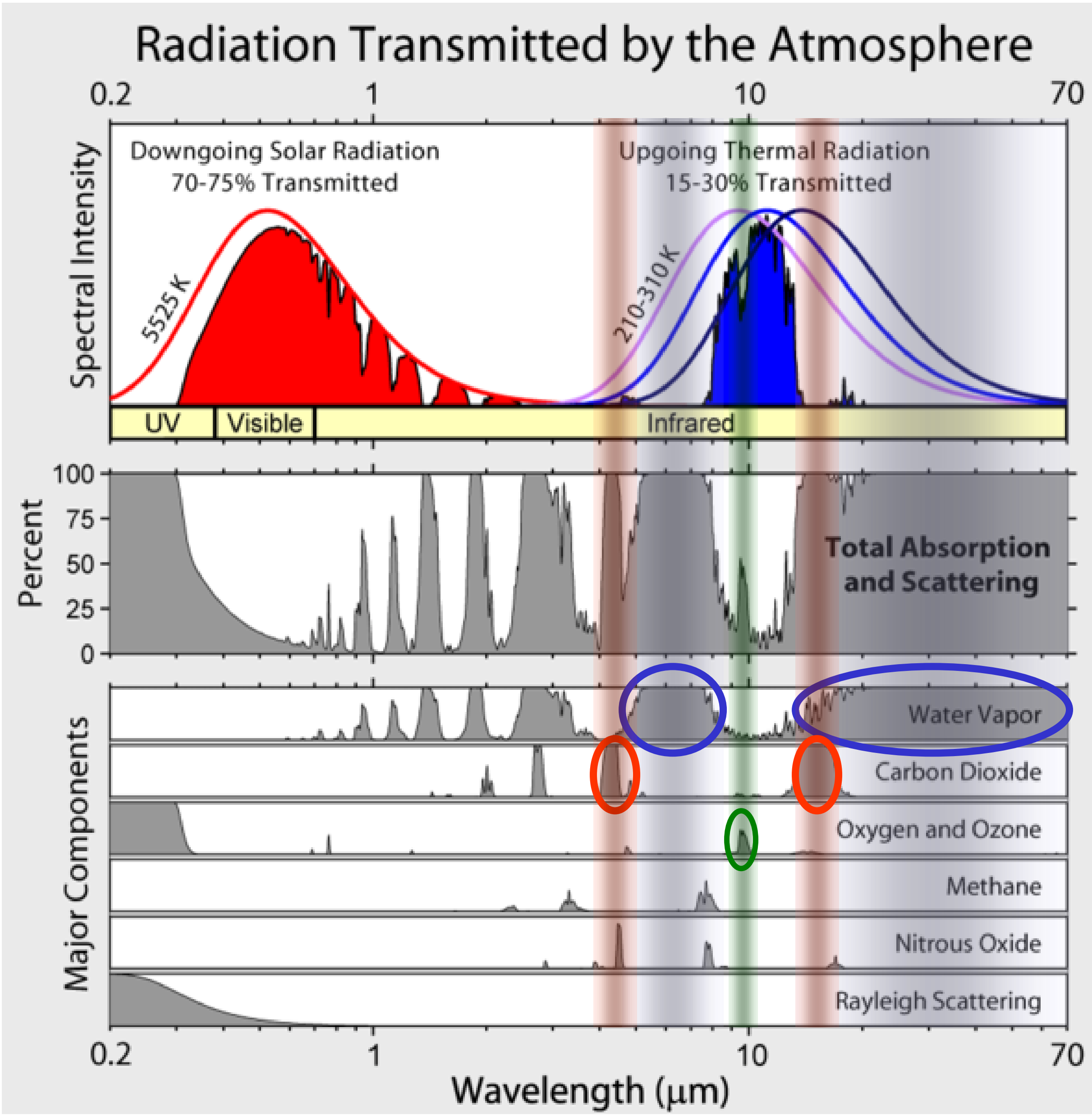


Some of the CO2 vibration bands lie in the water vapor "windows"

CO_2 and H_2O absorption spectrum



Other gases also absorb IR: O_3, CH_4, NO



Most important are the absorption peaks which sit on the Earth's peak emission spectrum:

H_2O - 6.3μm

H_2O > 12μm

CO_2 – 15μm

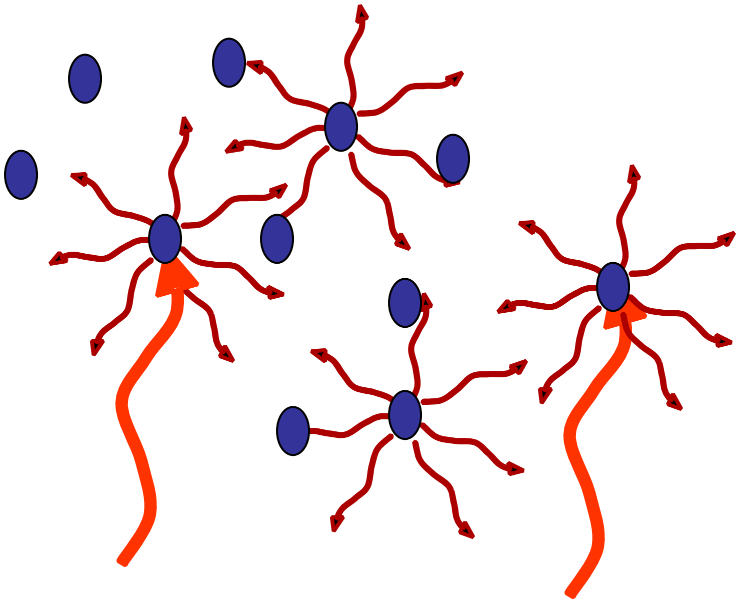
O_3 – 9.6μm

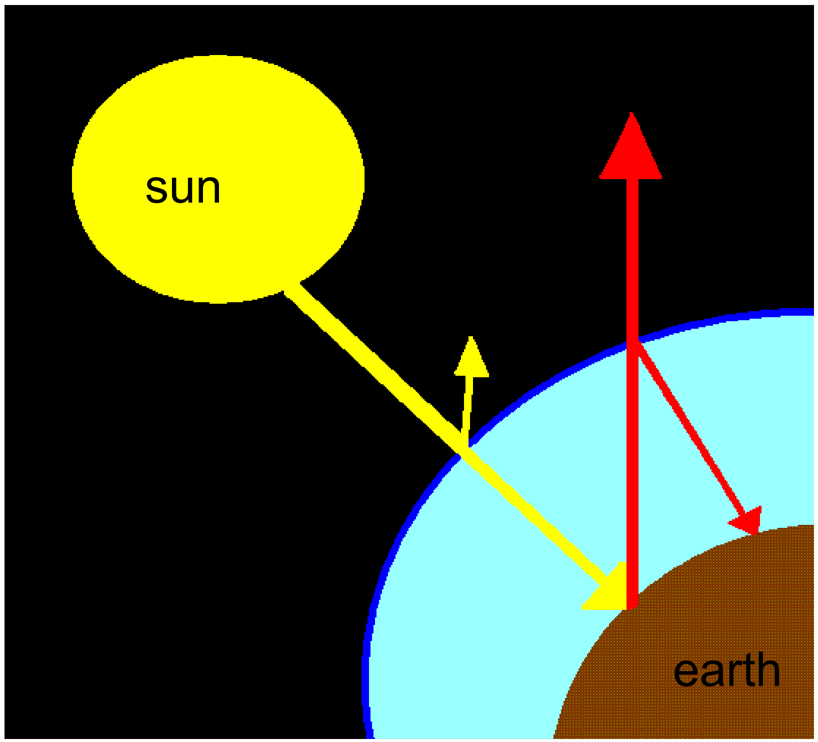
The IR radiation which is absorbed in the air molecules is emitted back out in all directions (including the surface).

Part of the energy absorbed causes the molecules to move faster → temperature rises.

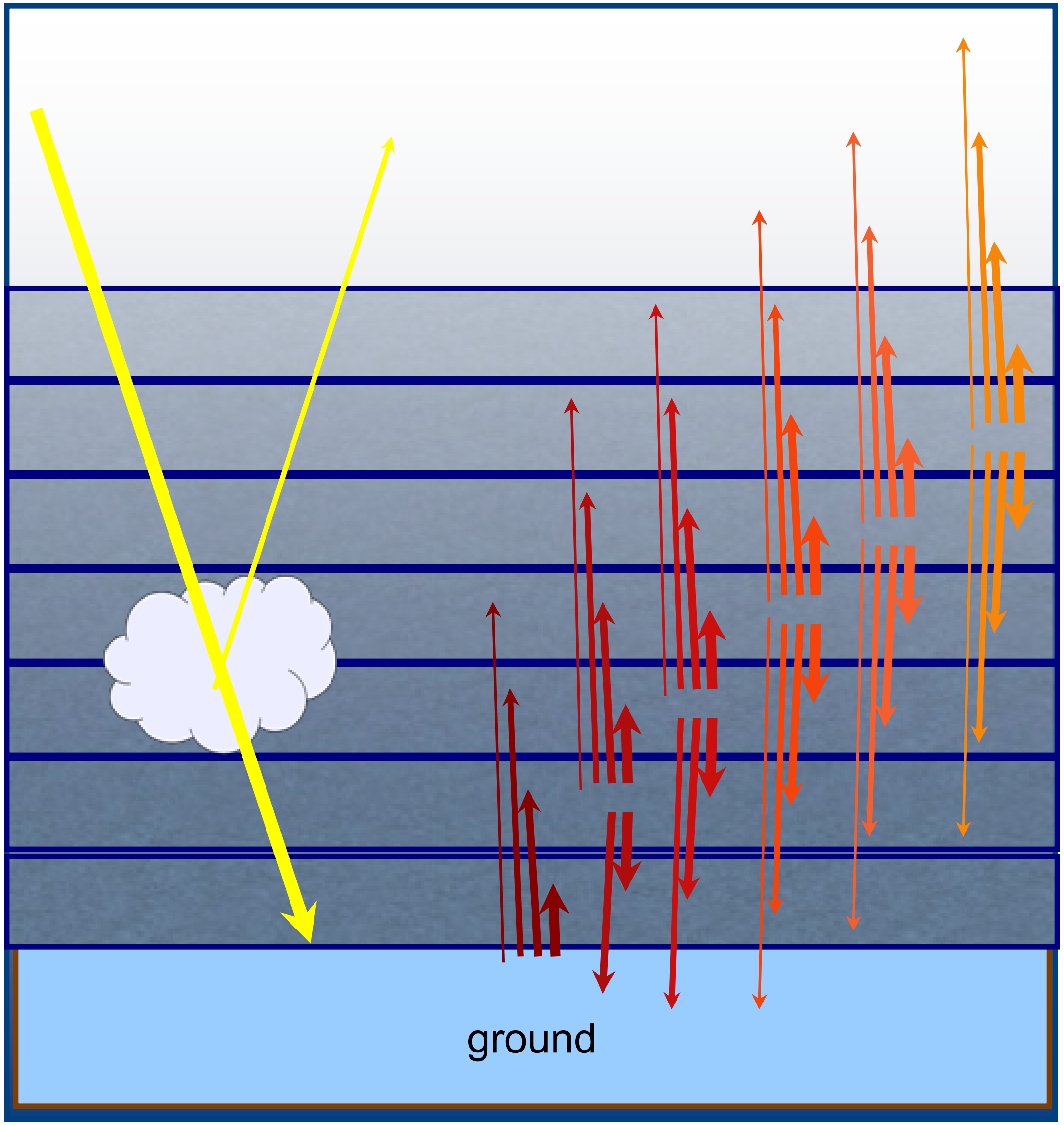
When temperature rises, more radiation is emitted (black body)

Temperature will rise until equilibrium is reached.





Equations…



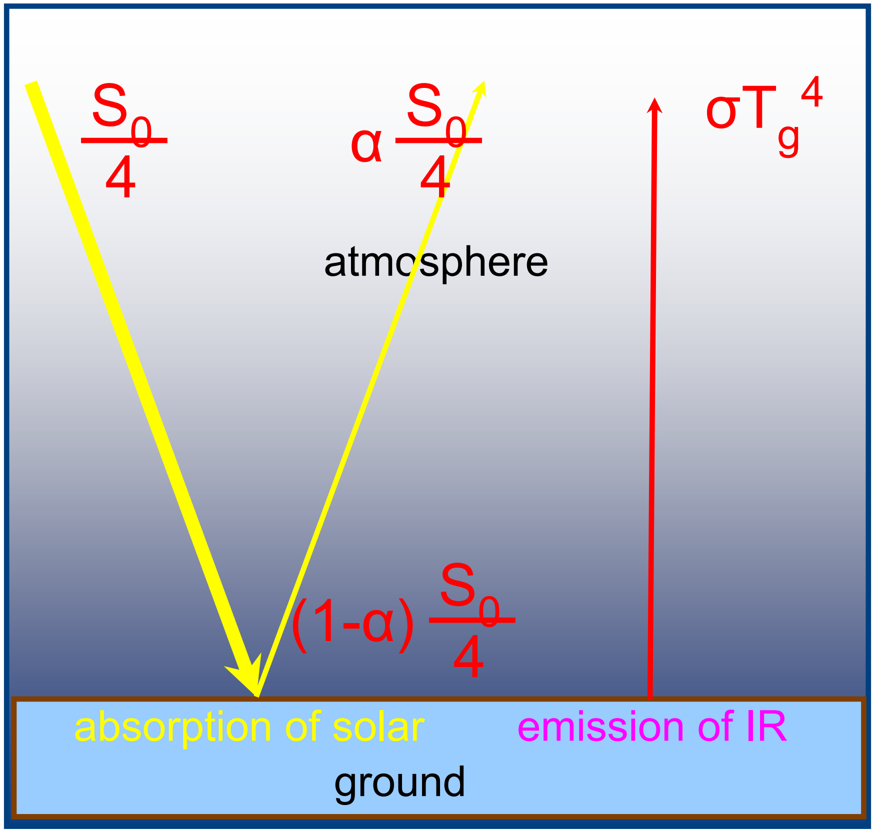
The terrestrial IR radiation gets gradually absorbed as it propagates upwards in the atmosphere.

The IR emitted from each layer, also gets gradually absorbed in the layers above and below.

The full problem consists of summing up the contributions and effects of all layers.

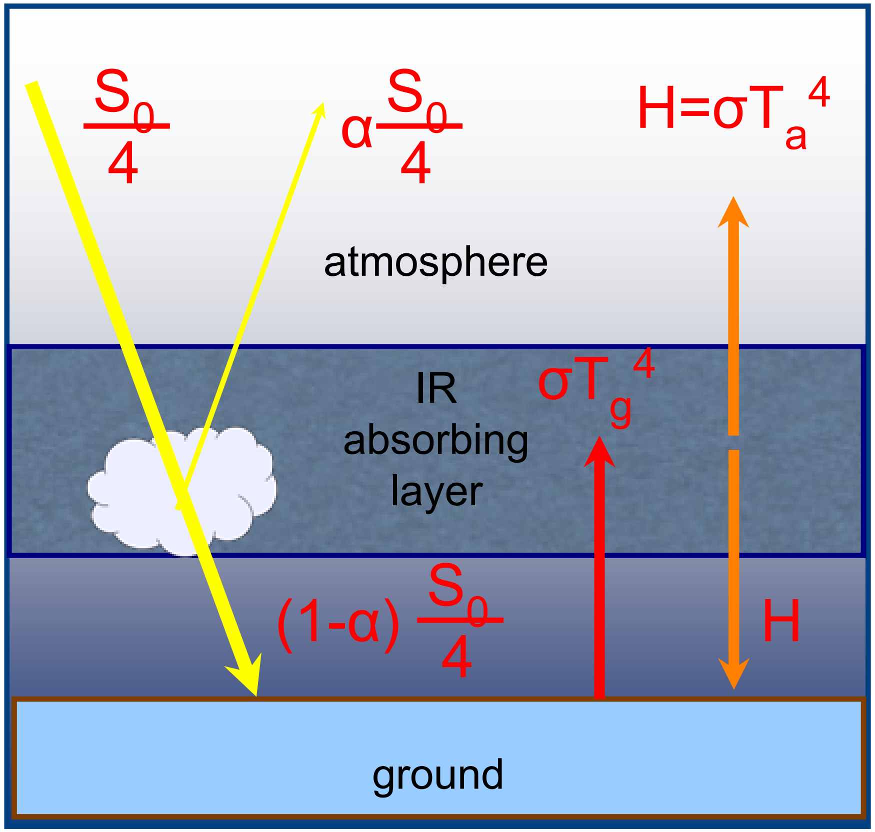
Effect of Atmospheric Absorption: 1 sheet

w/o absorption



$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4}$$

w. absorption



$$\sigma T_a^4 = (1-\alpha)\frac{S_0}{4}$$

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4} + \sigma T_a^4$$

Effect of Atmospheric Absorption: 1 sheet

w/o absorption

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4} = 239\ \mathrm{Wm}^{-2}$$

$$T_g = \sqrt[4]{(1-\alpha)\frac{S_0}{4\sigma}}$$

W/O absorption: **T_g=255 K or -18°C**

The Greenhouse Effect

w. absorption

$$\sigma T_a^4 = (1-\alpha)\frac{S_0}{4}$$

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{4} + \sigma T_a^4$$

$$\sigma T_g^4 = (1-\alpha)\frac{S_0}{2}$$

$$T_g = \sqrt[4]{(1-\alpha)\frac{S_0}{2\sigma}}$$

W. absorption **T_g=303.5 K or +30.5°C**

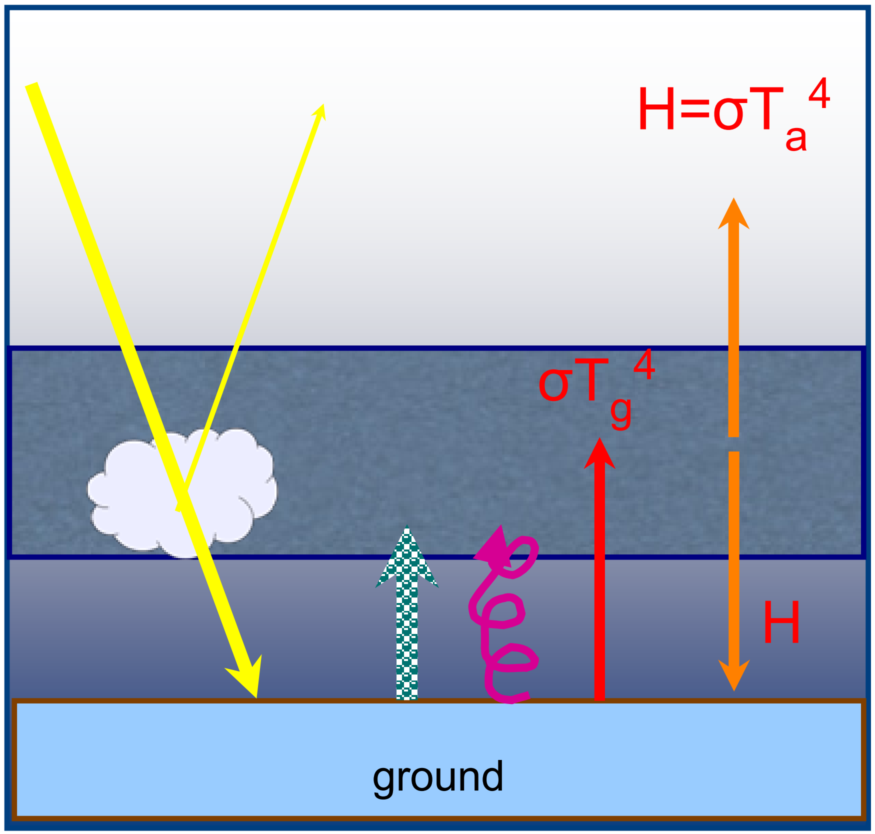
Too warm

2 layers

N layers

Surface and surface air temperature

Since radiation makes the surface hotter than the air just above it, the surface looses heat not only through radiation, but also through latent and sensible heat transports (both upwards and polewards).



These fluxes do not depend so strongly on temperature – a linear, rather than "to the 4th" dependence on T:

$$\sigma T_g^4 + A T_g + B = (1-\alpha)\frac{S_0}{4} + \sigma T_a^4$$

This yields colder T_g than with pure radiative surface cooling

Radiative convective equilibrium…
Manabe and Strickler, 1964

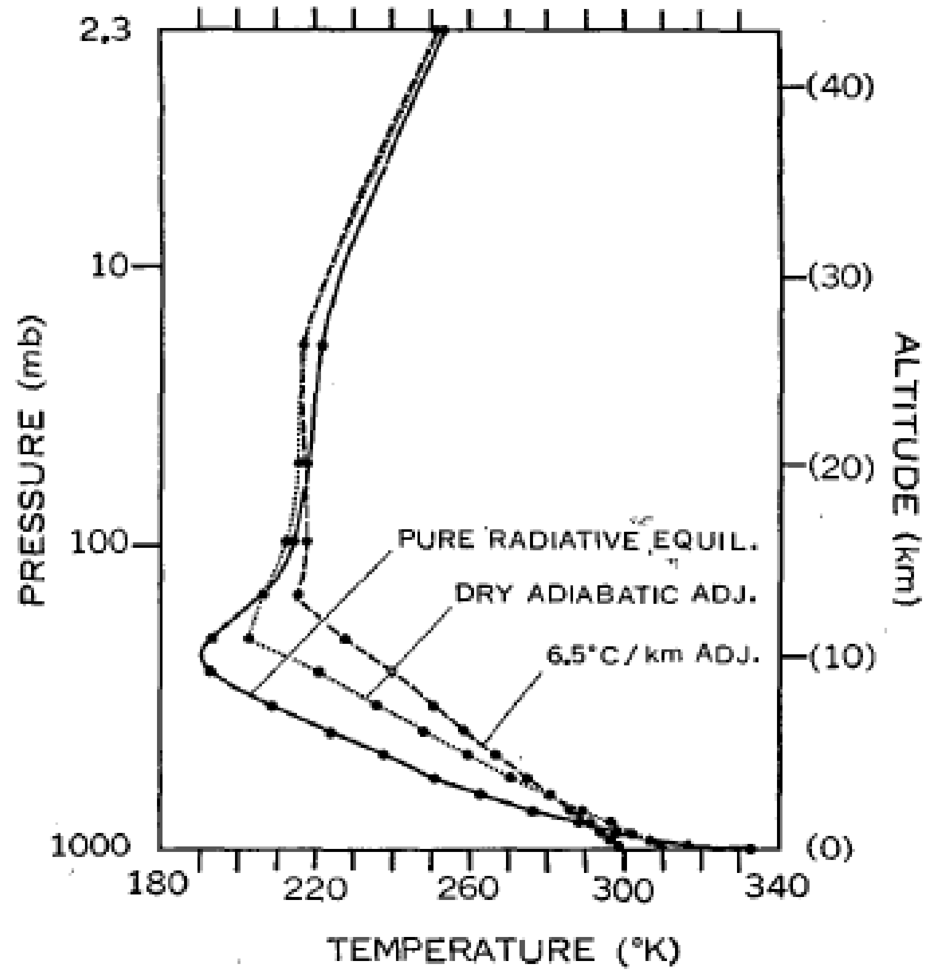
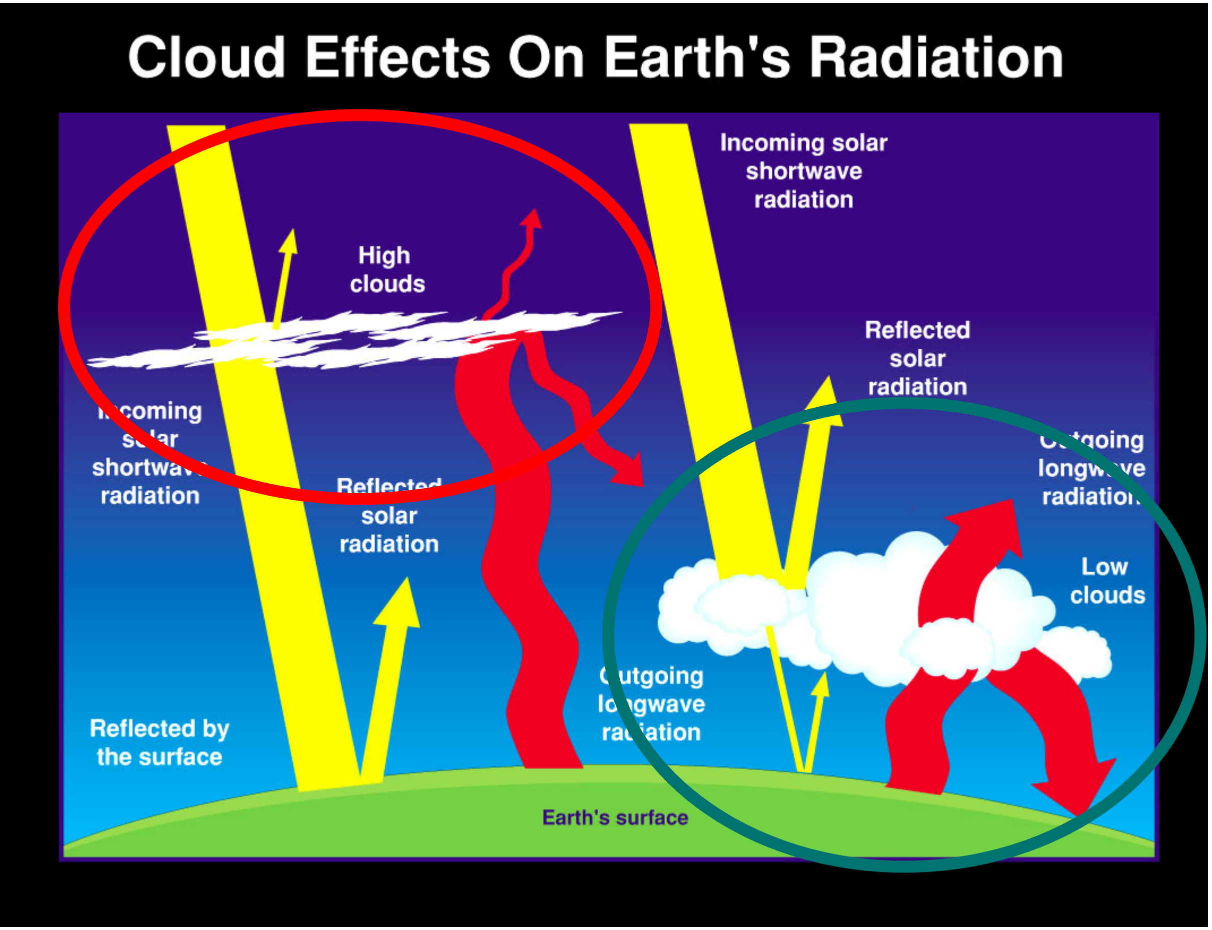


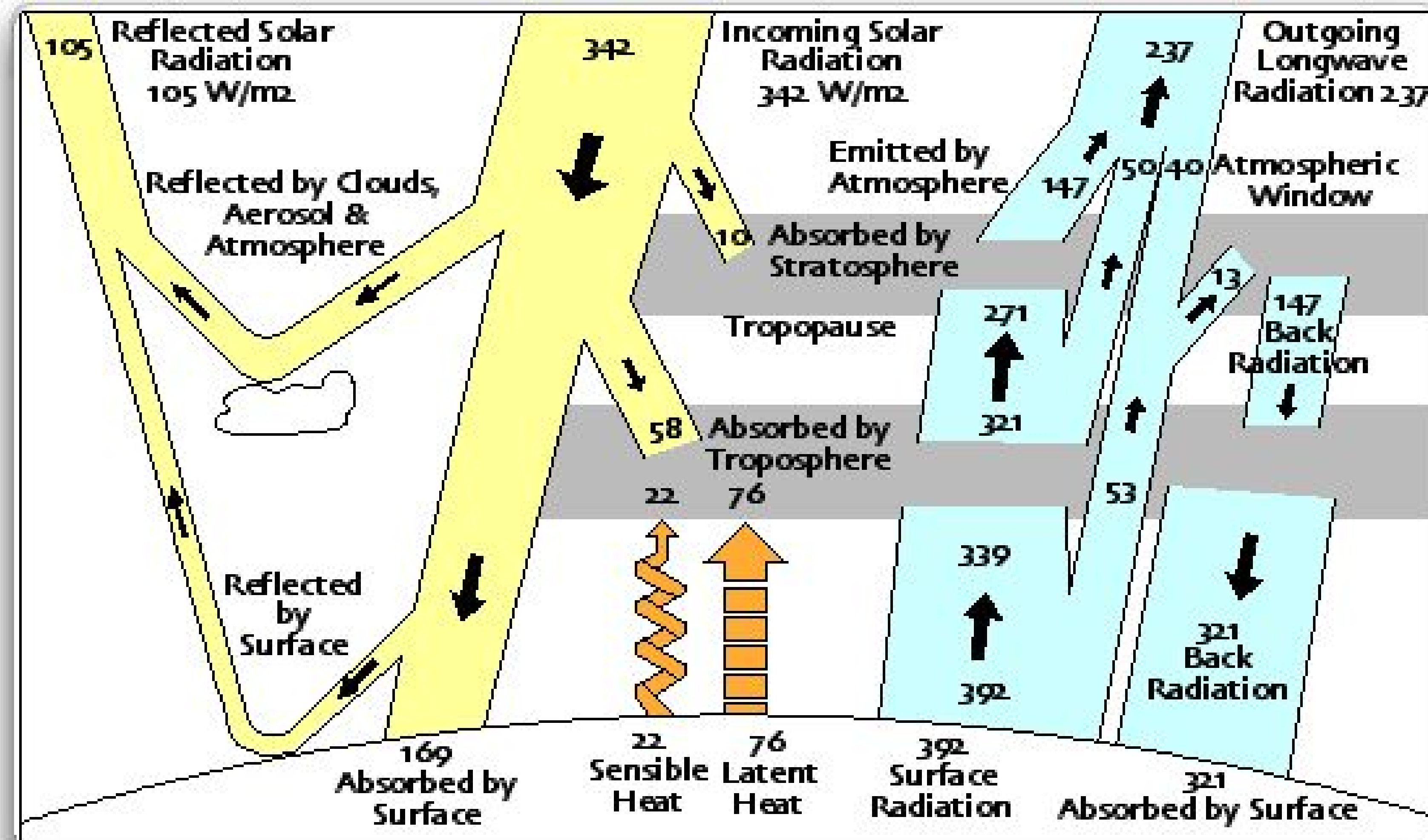
FIG. 4. The dashed, dotted, and solid lines show the thermal equilibrium with a critical lapse rate of 6.5 deg km^{-1}, a dry-adiabatic critical lapse rate (10 deg km^{-1}), and pure radiative equilibrium.

Clouds complicate the picture:

Thin high cirrus clouds reflect little solar radiation, and absorb IR efficiently. They then emit IR at a cold temperature (air is colder higher up) → **net warming**



Low thick clouds reflect solar radiation efficiently. They absorb IR, but they emit at a relatively warm temperature (near the ground) → **net cooling**



Overall, we get a surface temperature of $T_g=(390/\sigma)^{1/4}=288°K=15°C$

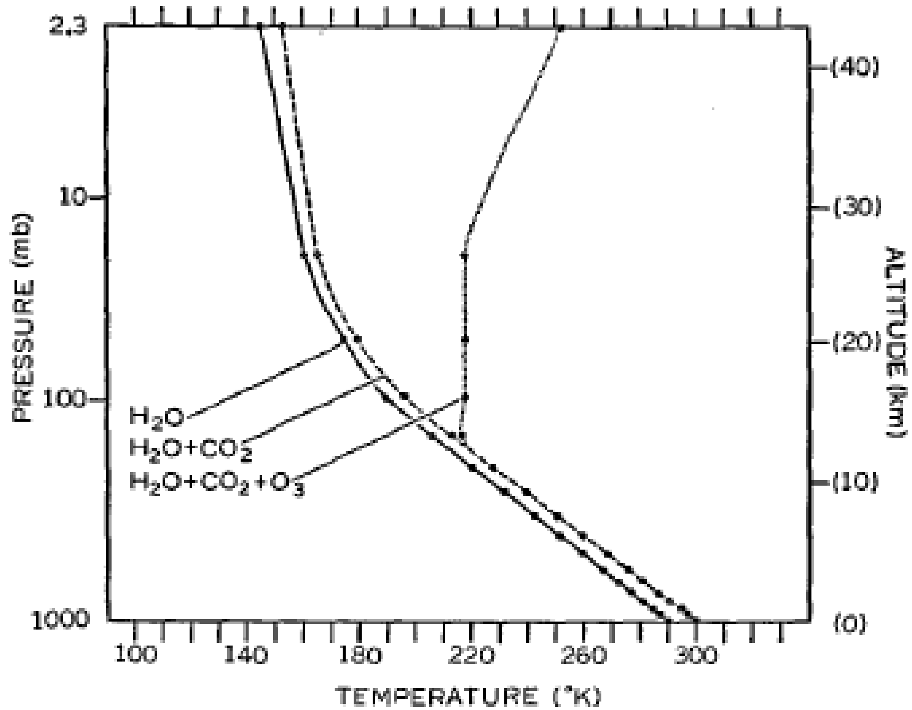
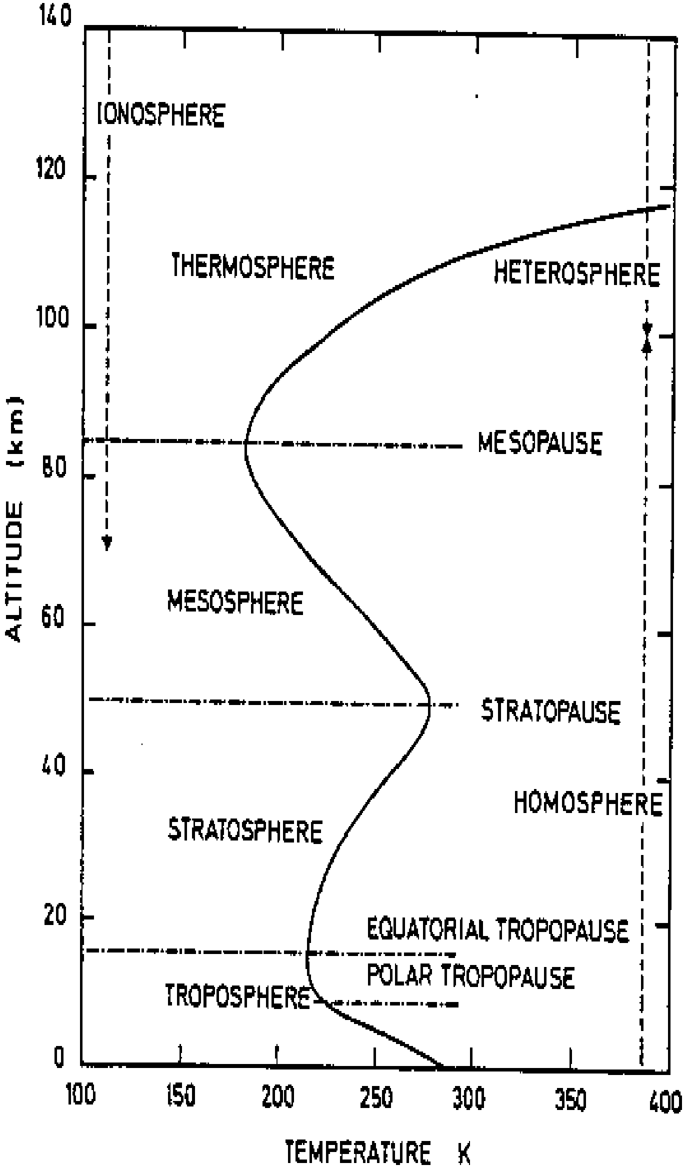


FIG. 6C. Thermal equilibrium of various atmospheres which have a critical lapse rate of 6.5 deg km^{-1}. Vertical distributions of gaseous absorbers at 35N, April, were used. $S_c=2$ ly min^{-1}, $\cos\bar{\zeta}=0.5$, $r=0.5$, no clouds.



Mean Atmospheric Temperature Profile
(Adapted From Brasseur and Solomon)

consideration.

LW Cooling **SW Heating**

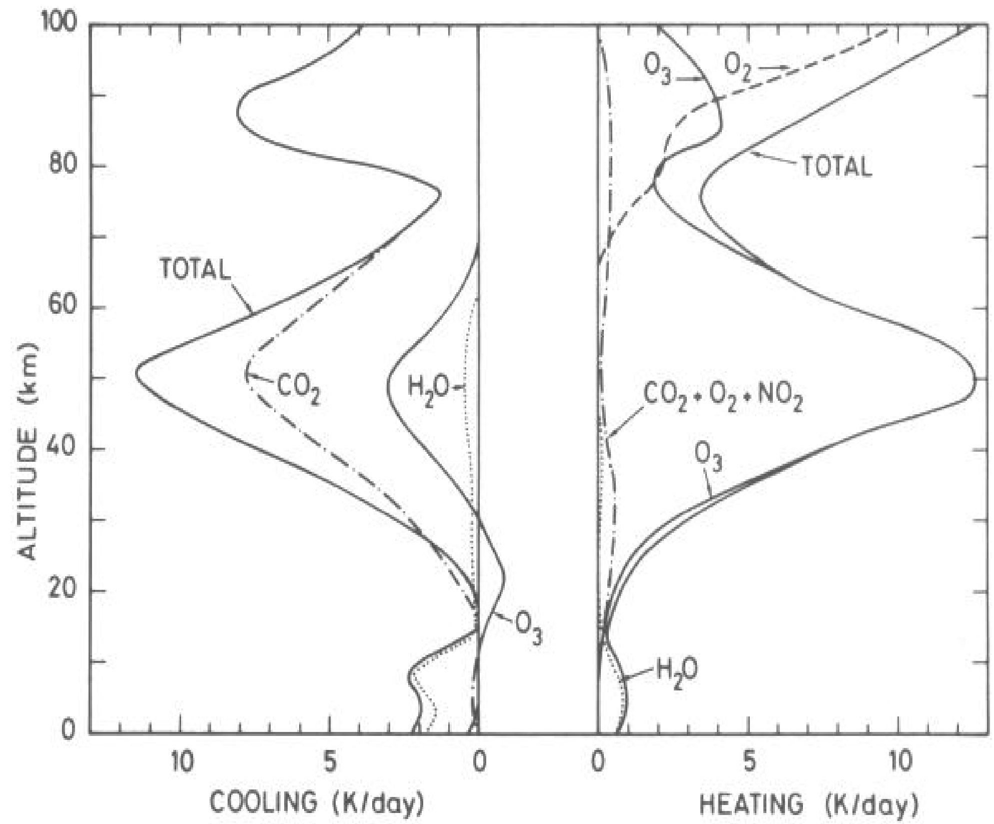
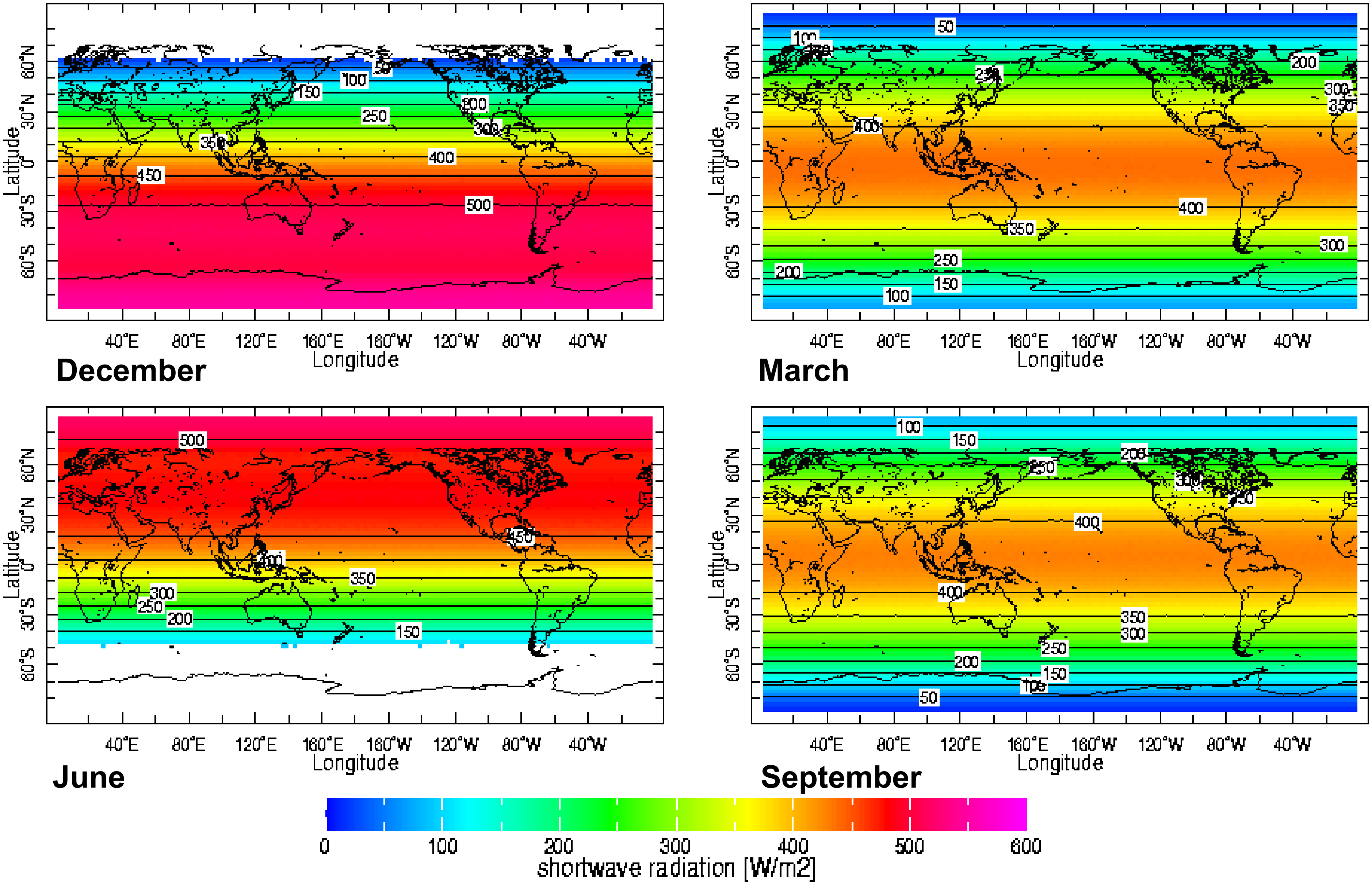
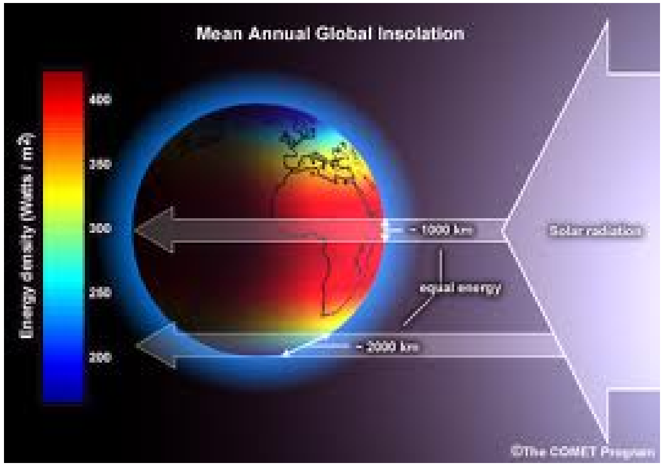


Fig. 4.19b. Vertical distribution of solar short wave heating rates by O_3, NO_2, H_2O, CO_2, and of terrestrial long wave cooling rates by CO_2, O_3, H_2O. From London (1980).

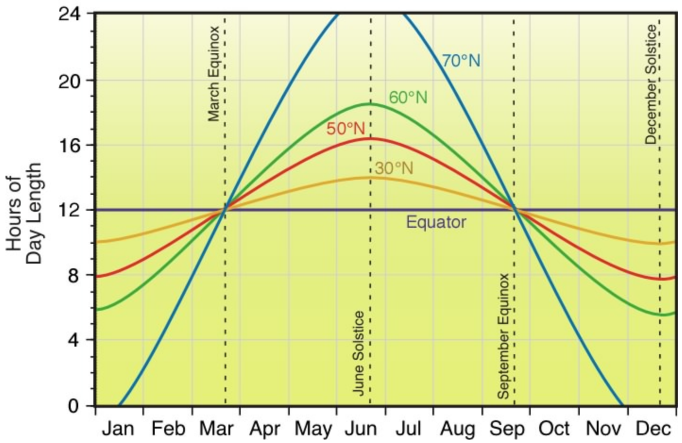
Incoming Solar (Shortwave) at TOA



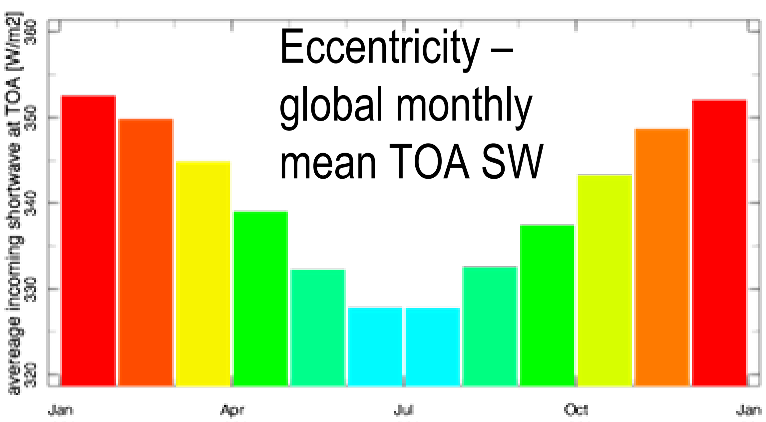
Daily mean incoming solar radiation at top of the atmosphere (W/m^2)



declination

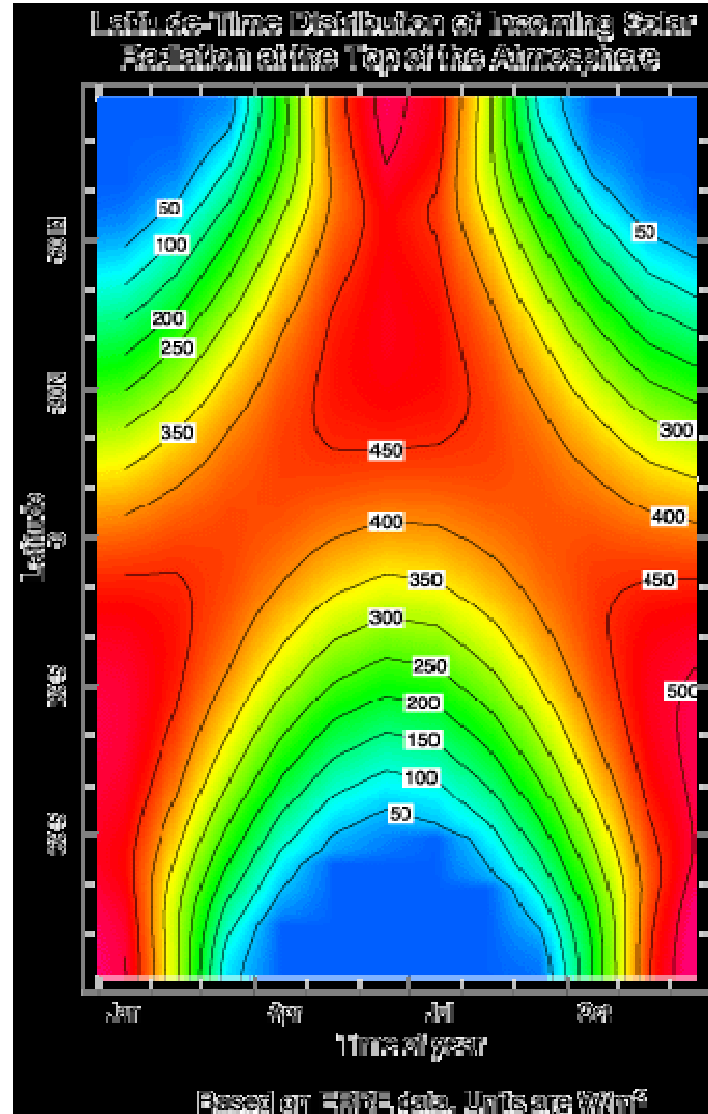


Length of day



Eccentricity – global monthly mean TOA SW

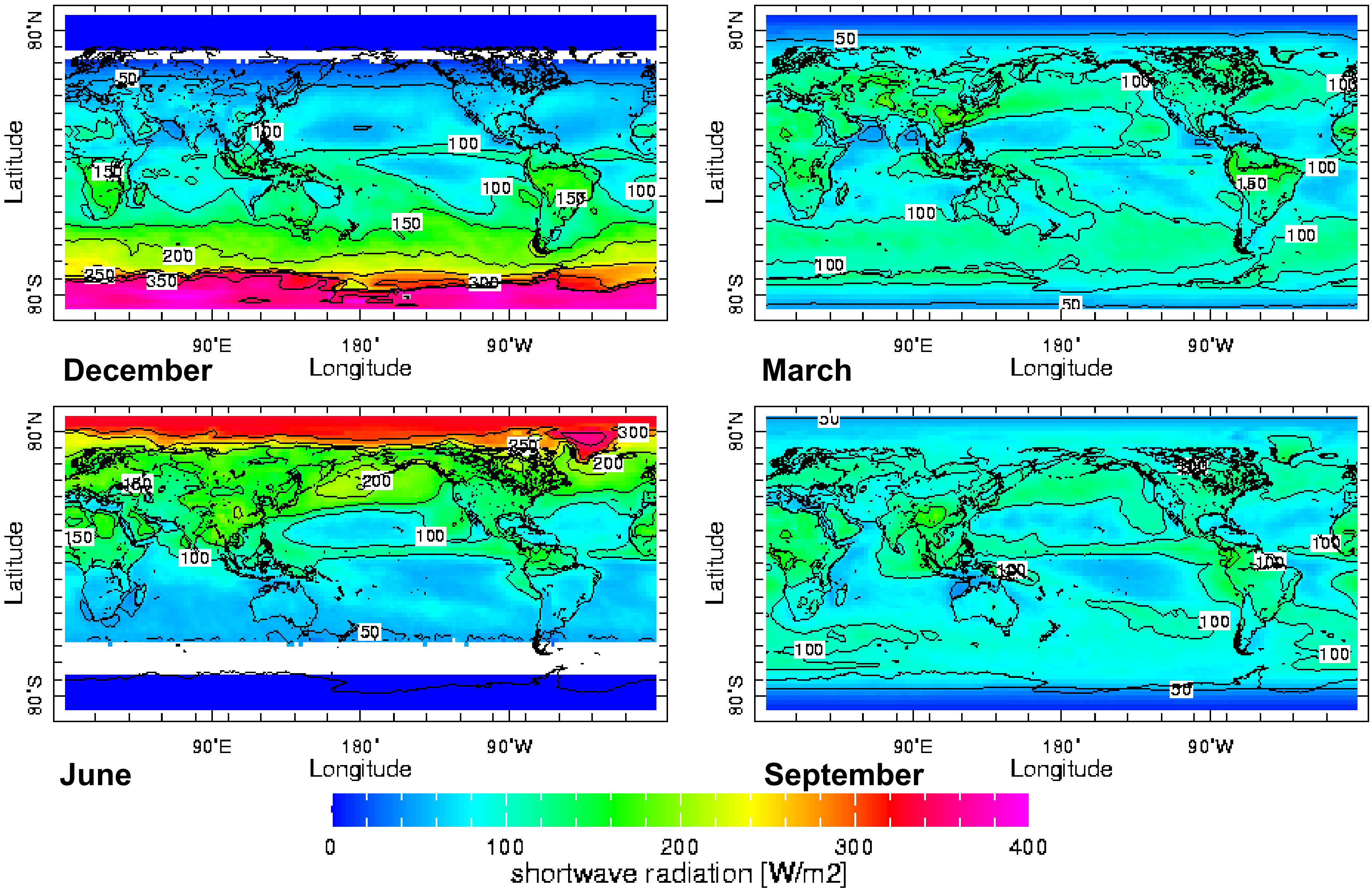
Latitude



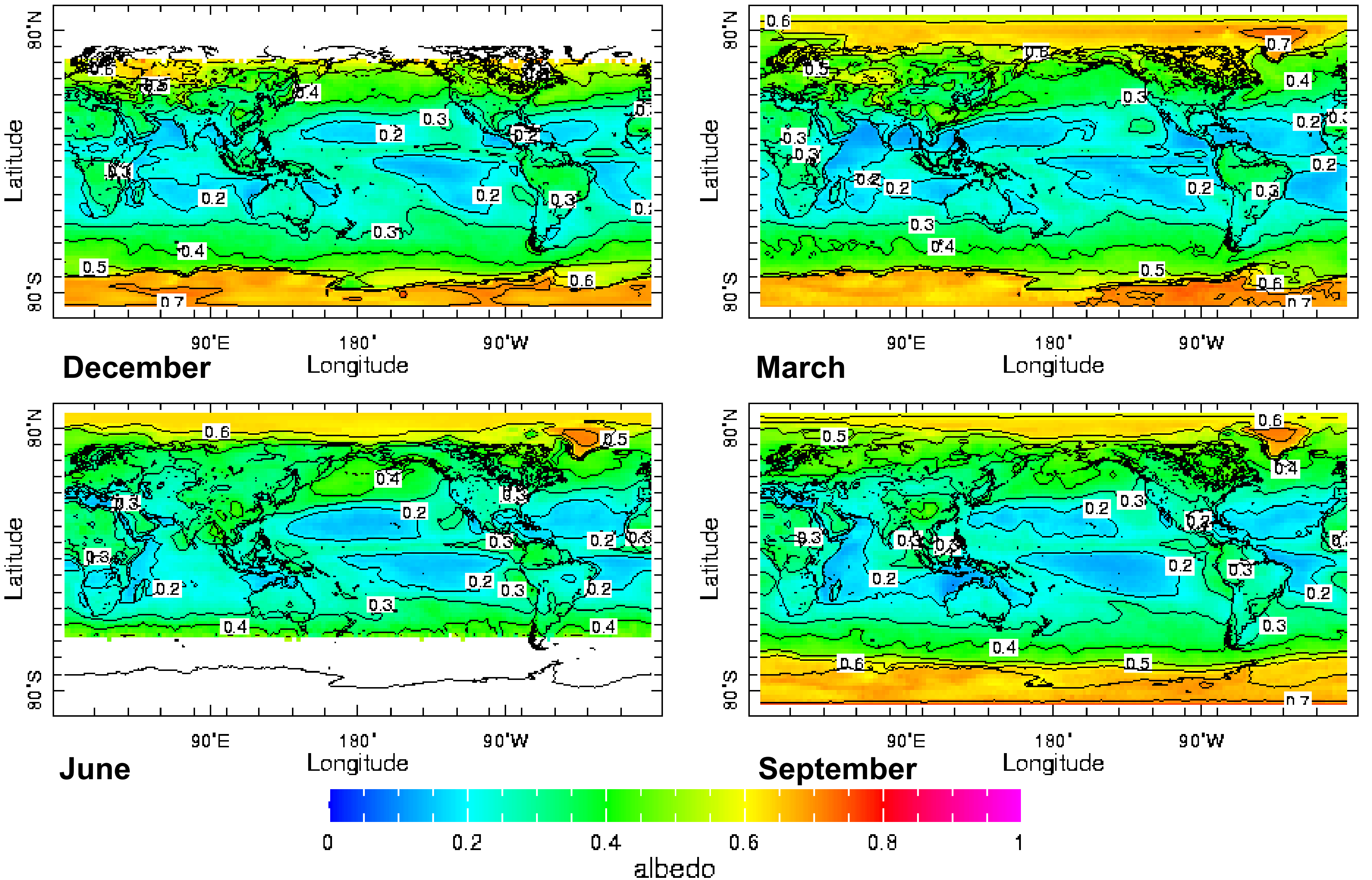
Based on ERBE data

Time of year

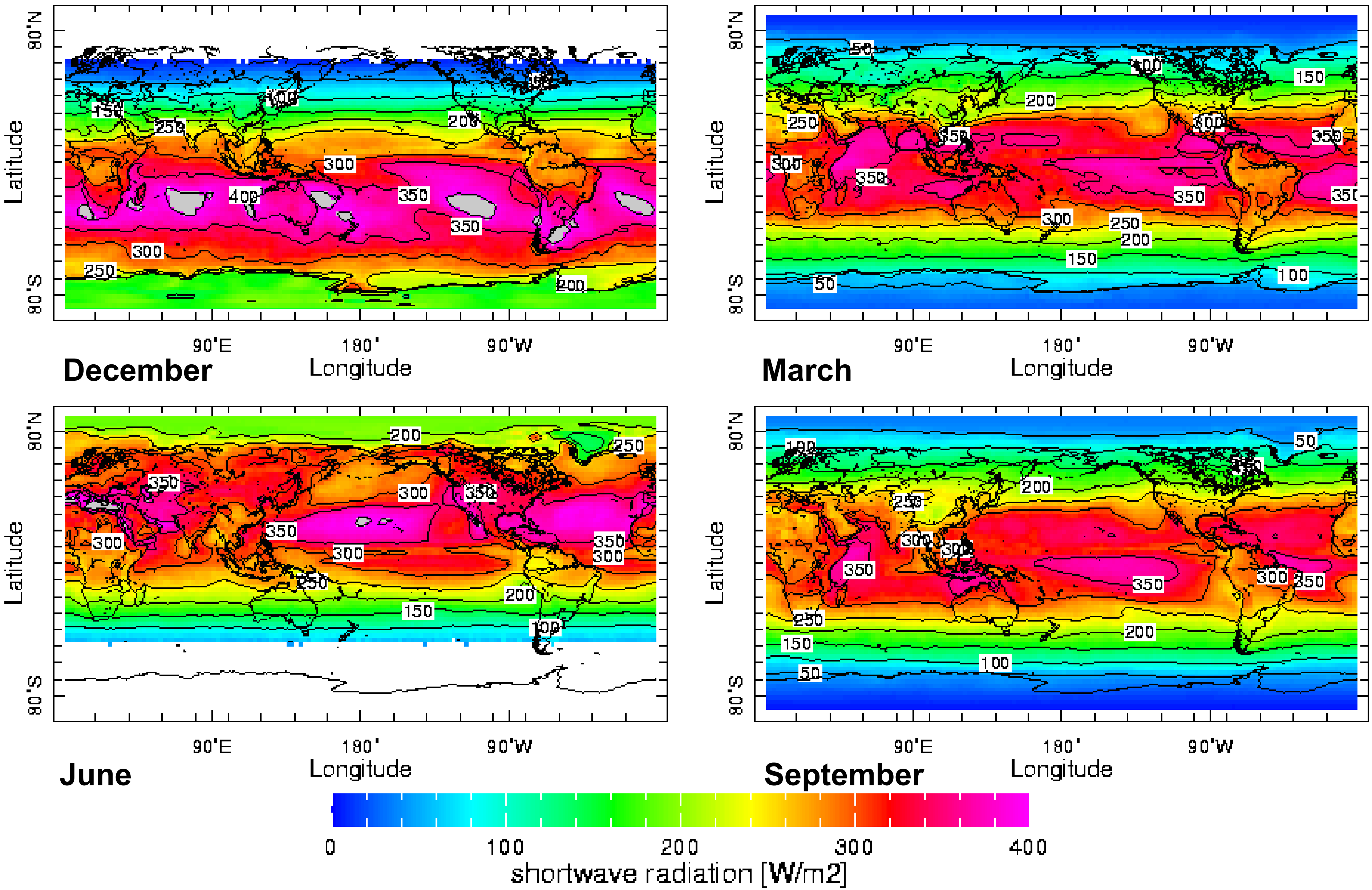
Reflected Solar at TOA



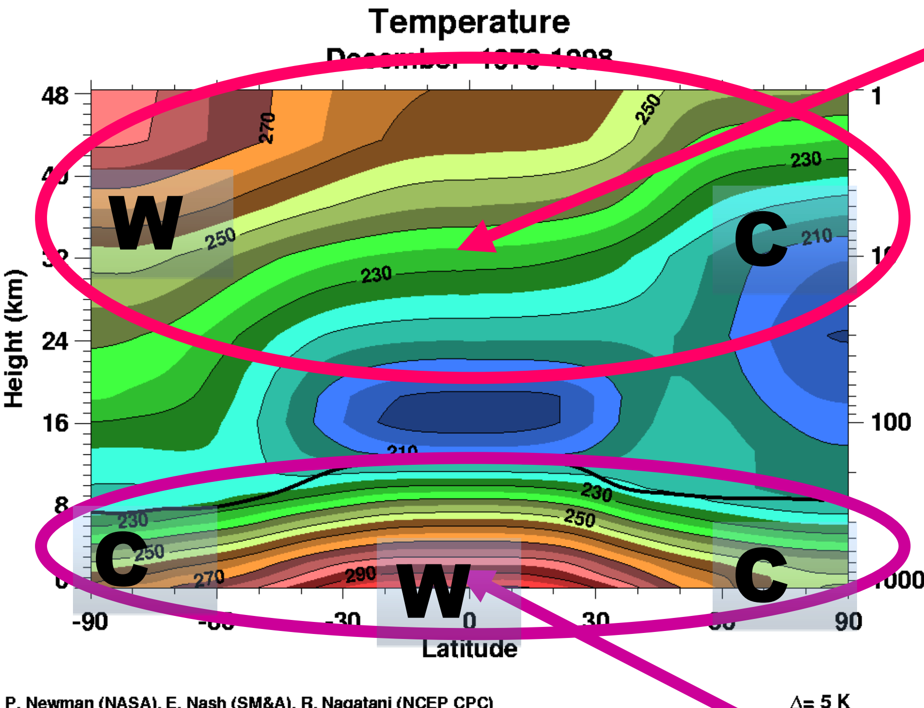
Planetary Albedo



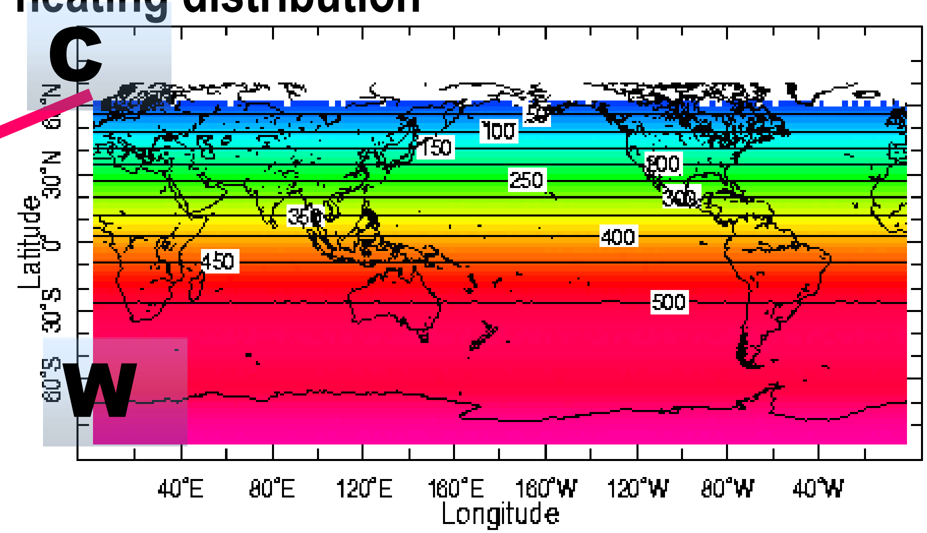
Net Shortwave (Solar) Radiation



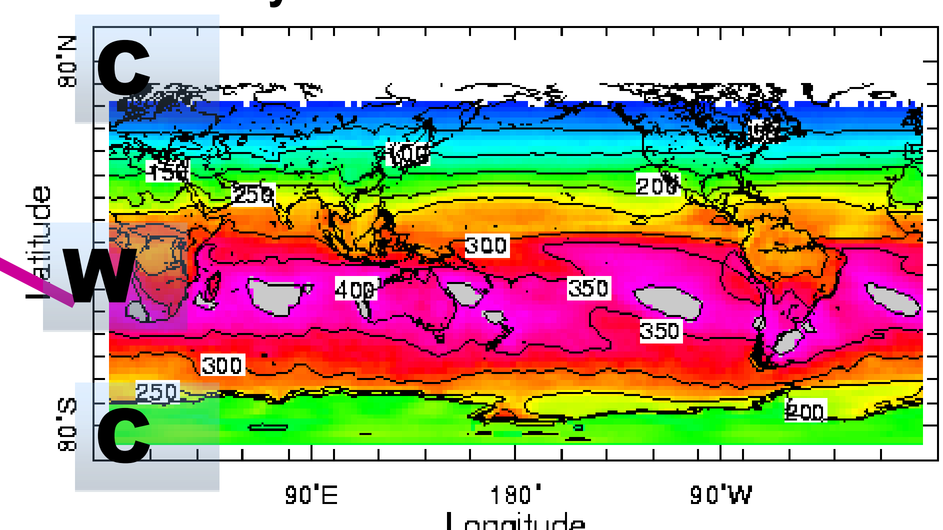
Latitude-height zonal mean temperature structure



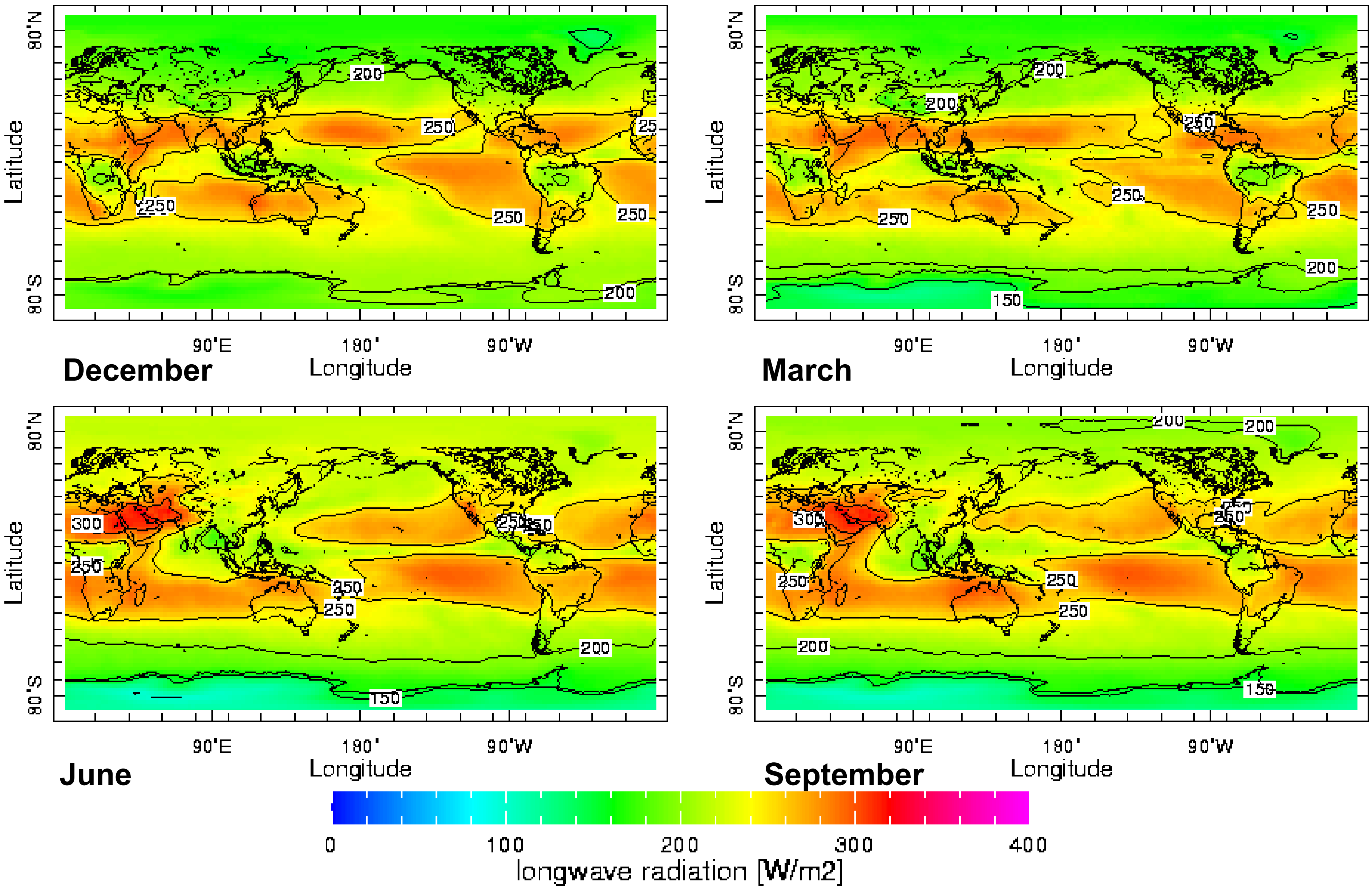
Daily incoming radiation at top of atmosphere during December – ozone heating distribution



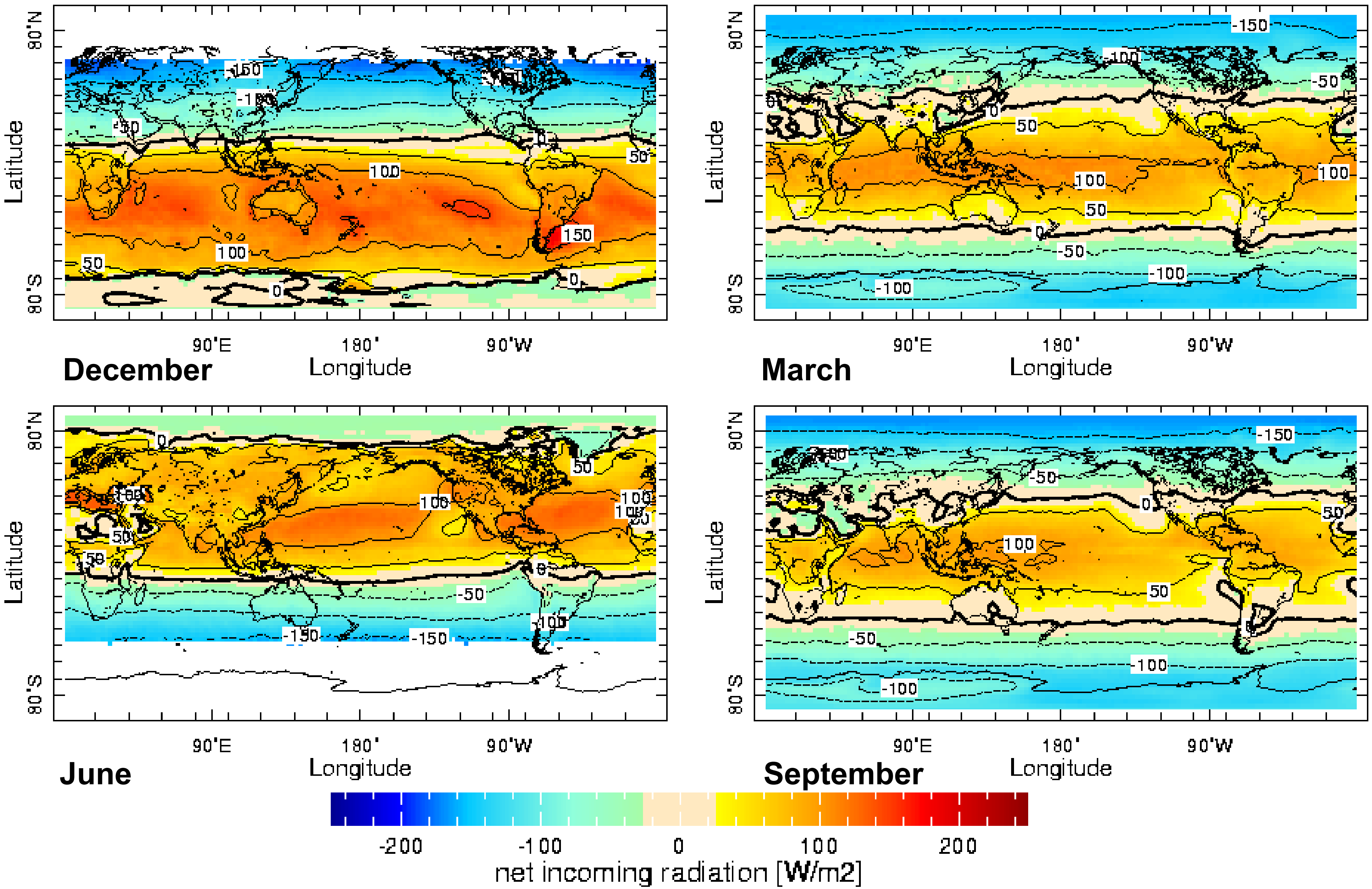
Net daily solar radiation at surface



Outgoing Longwave (IR) at TOA

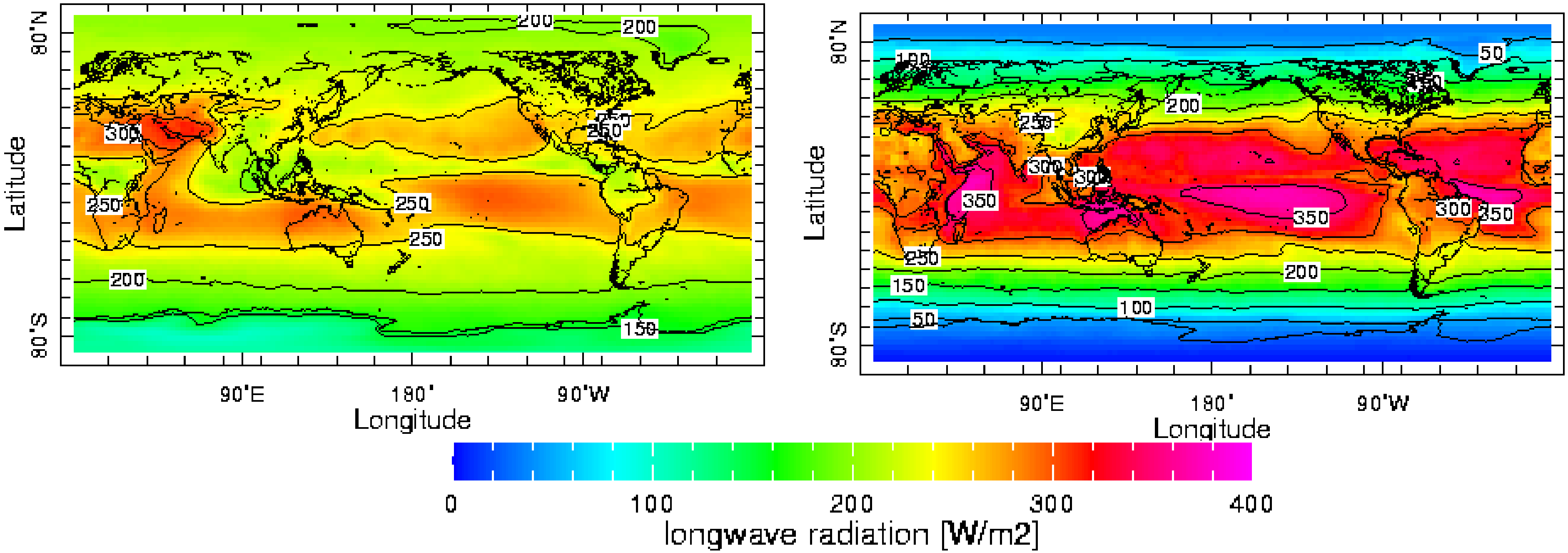


Net Incoming Radiation (LW and SW)

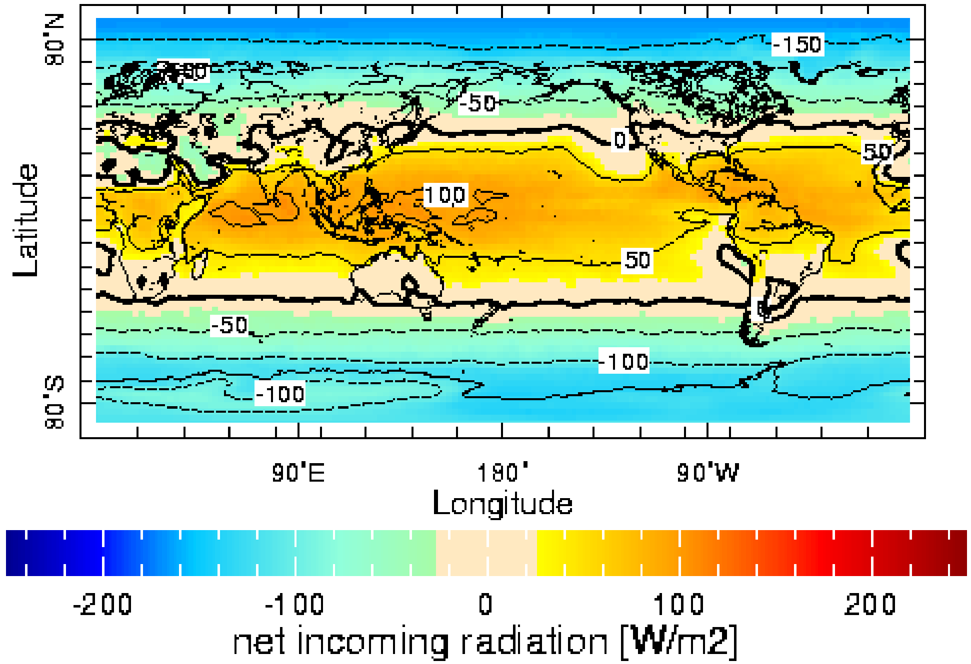


Outgoing longwave radiation (OLR)

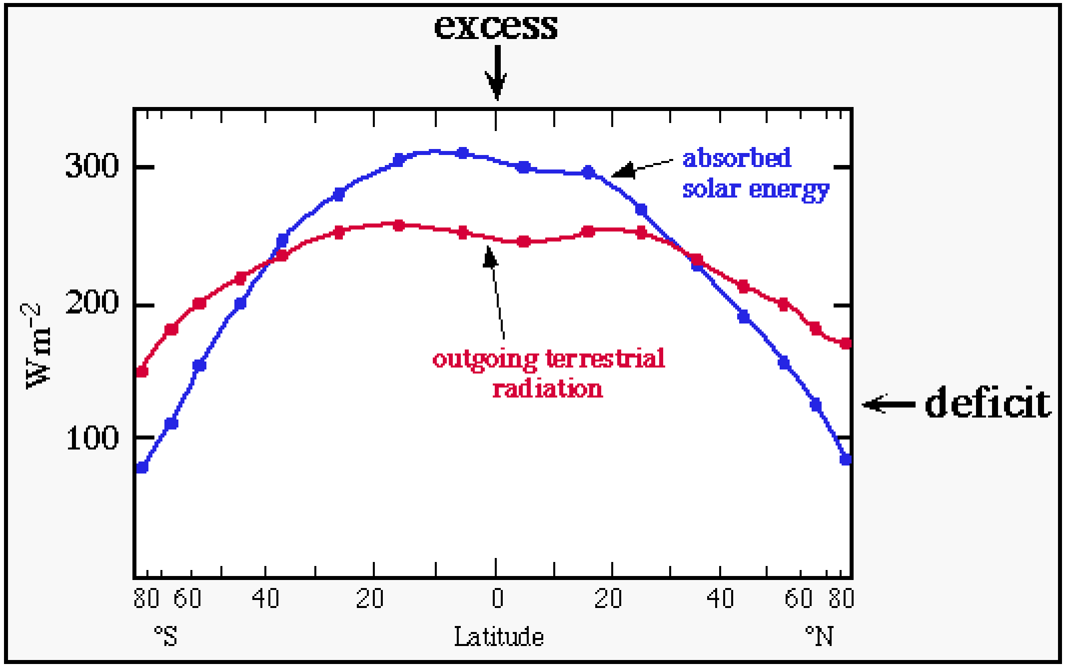
Net Shortwave (Solar) Radiation

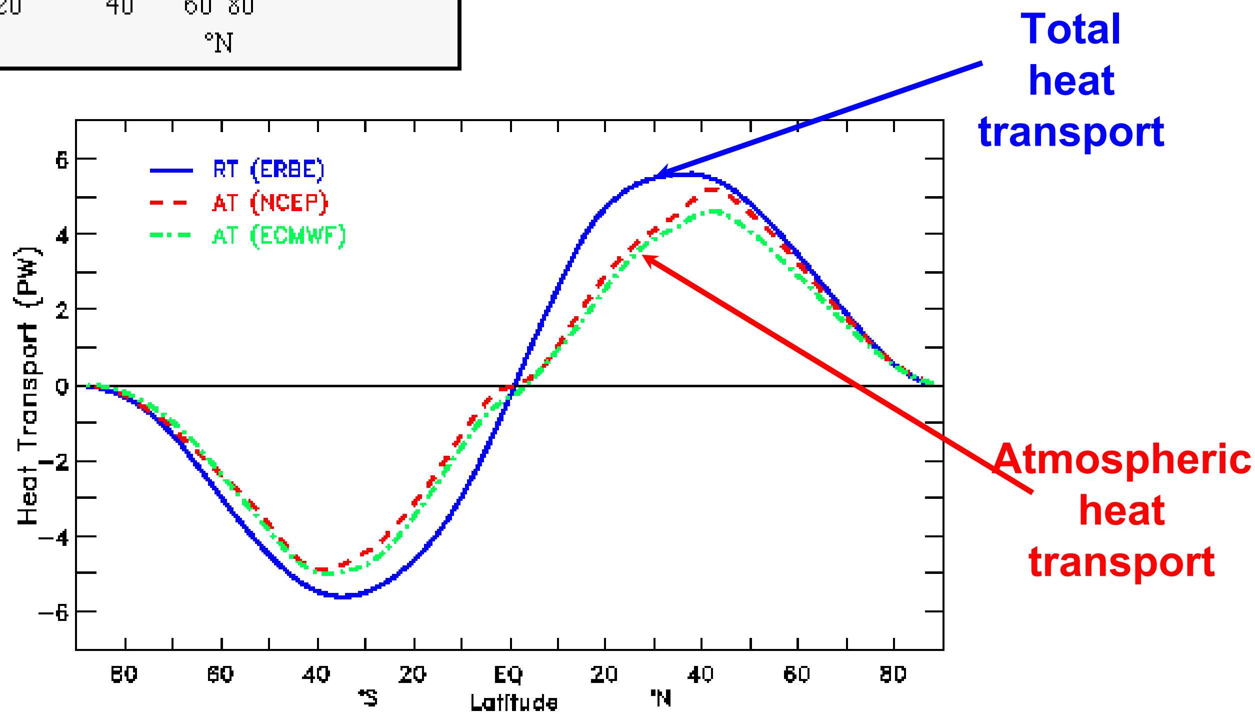


Net incoming Radiation (SW-LW)



Latitudinal Radiation Imbalance





adapted from Trenberth and Caron (2001)

PW=10^{15} W

Meridional heat transport

Trenberth and Caron (2001)

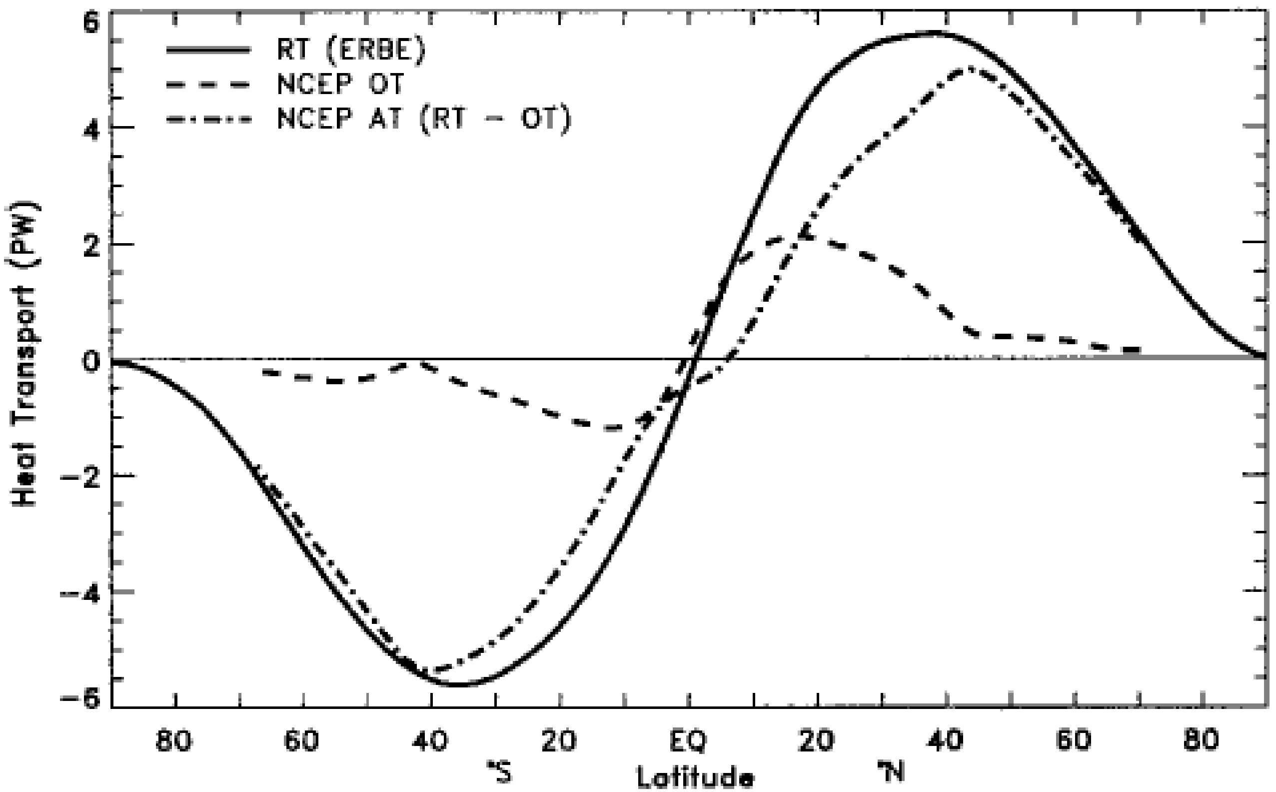


FIG. 7. The required total heat transport from the TOA radiation RT is compared with the derived estimate of the adjusted ocean heat transport OT (dashed) and implied atmospheric transport AT from NCEP reanalyses (PW).

PW=10^{15} W