Atmospheric Sciences 321

Science of Climate

Lecture 3: Global Energy Balance

Community Business

- Check the assignments
- You should have read chapters 1 and 2 of book
- First HW due next Wednesday in class
- Questions?
- Will review a bit today.

Review Last Time

- Conservation of energy and hydrostatic relation indicate a relationship between temperature and altitude we call the dry **adiabatic lapse rate.**

$$c_p\, dT = -\alpha\rho g\, dz\,, \quad \text{or} \quad c_p\, dT = -g\, dz\,, \quad \text{or} \quad \boxed{\frac{dT}{dz} = -\frac{g}{c_p}}$$

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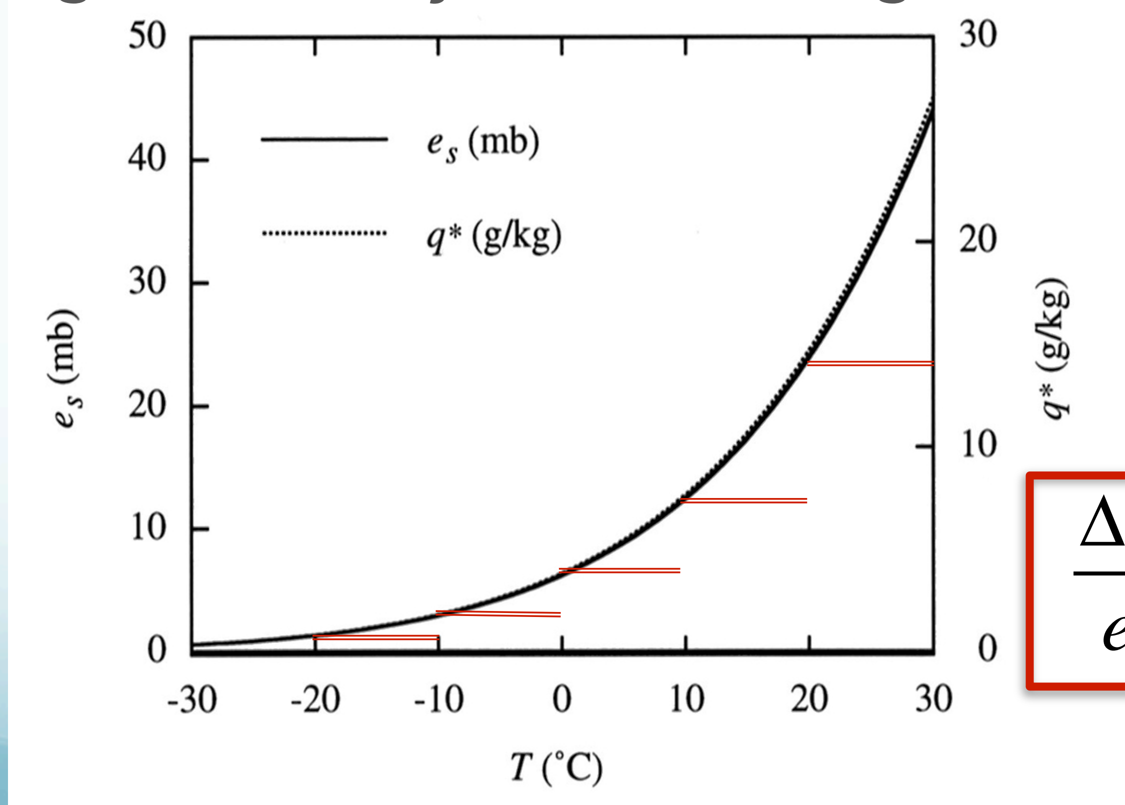
Review Last Time - Water

- Saturation water vapor pressure increases rapidly with temperature, about 7% for 1°K warming, or doubling about every 10°K warming.



$$\boxed{\frac{\Delta e_s}{e_s} \approx 20\,\frac{\Delta T}{T}}$$

Scaling Clausius-Clapeyron

- The saturation vapor pressure e_s depends only on Temperature!

$$\frac{de_s}{dT} = \frac{L}{T(\alpha_v - \alpha_l)}$$

- We can manipulate this a little, to show.

$$\frac{\Delta q^*}{q^*} = \frac{\Delta e_s}{e_s} = \left(\frac{L}{R_v T}\right)\frac{\Delta T}{T} = r\frac{\Delta T}{T} \qquad q \simeq 0.622\,\frac{e}{p}$$

- Where r turns out to be about 20, so for a 1% change in temperature, we get a 20% change in saturation vapor pressure and specific humidity q, or 7% per degree C or K, a strong factor in climate change.

Check Linearization

- Let's do some math

$$\frac{\Delta q^*}{q^*} = \frac{\Delta e_s}{e_s} = \left(\frac{L}{R_v}\right)\frac{\Delta T}{T^2} = r\frac{\Delta T}{T}$$

- What is the error involved in ignoring the change in the denominator T?

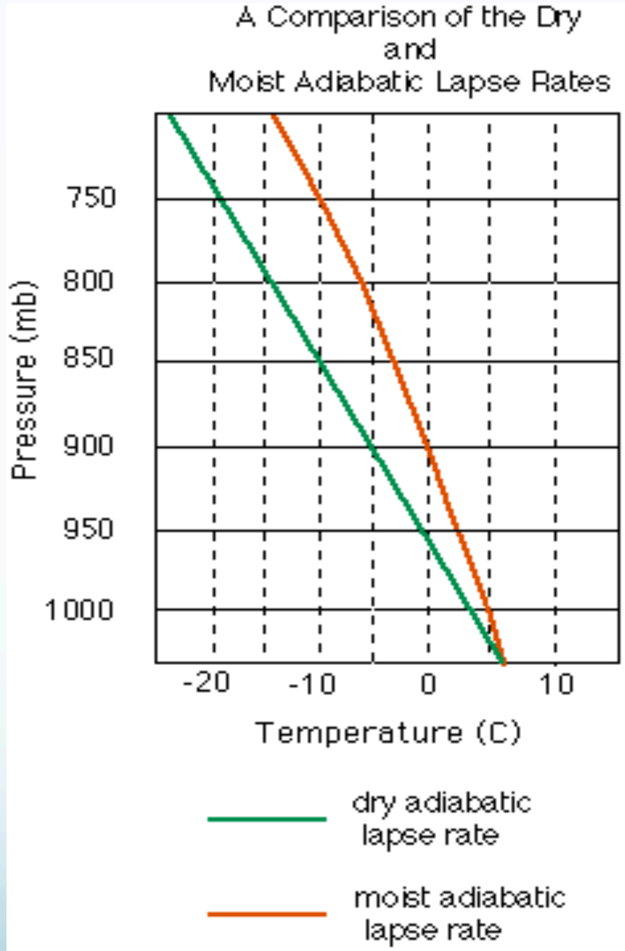
$$\frac{\Delta T}{T^2} = \frac{\Delta T}{(T+\Delta T)^2} = \frac{\Delta T}{\left(T_o^2 + 2T_o\Delta T + \Delta T^2\right)}$$

- Suppose T=300 and DT=3 --> 2% error

$$\frac{\Delta T}{T^2} = \frac{\Delta T}{(T+\Delta T)^2} = \frac{\Delta T}{9\times 10^4(1.0+0.02+0.0001)} \simeq \frac{\Delta T}{T_o^2}$$

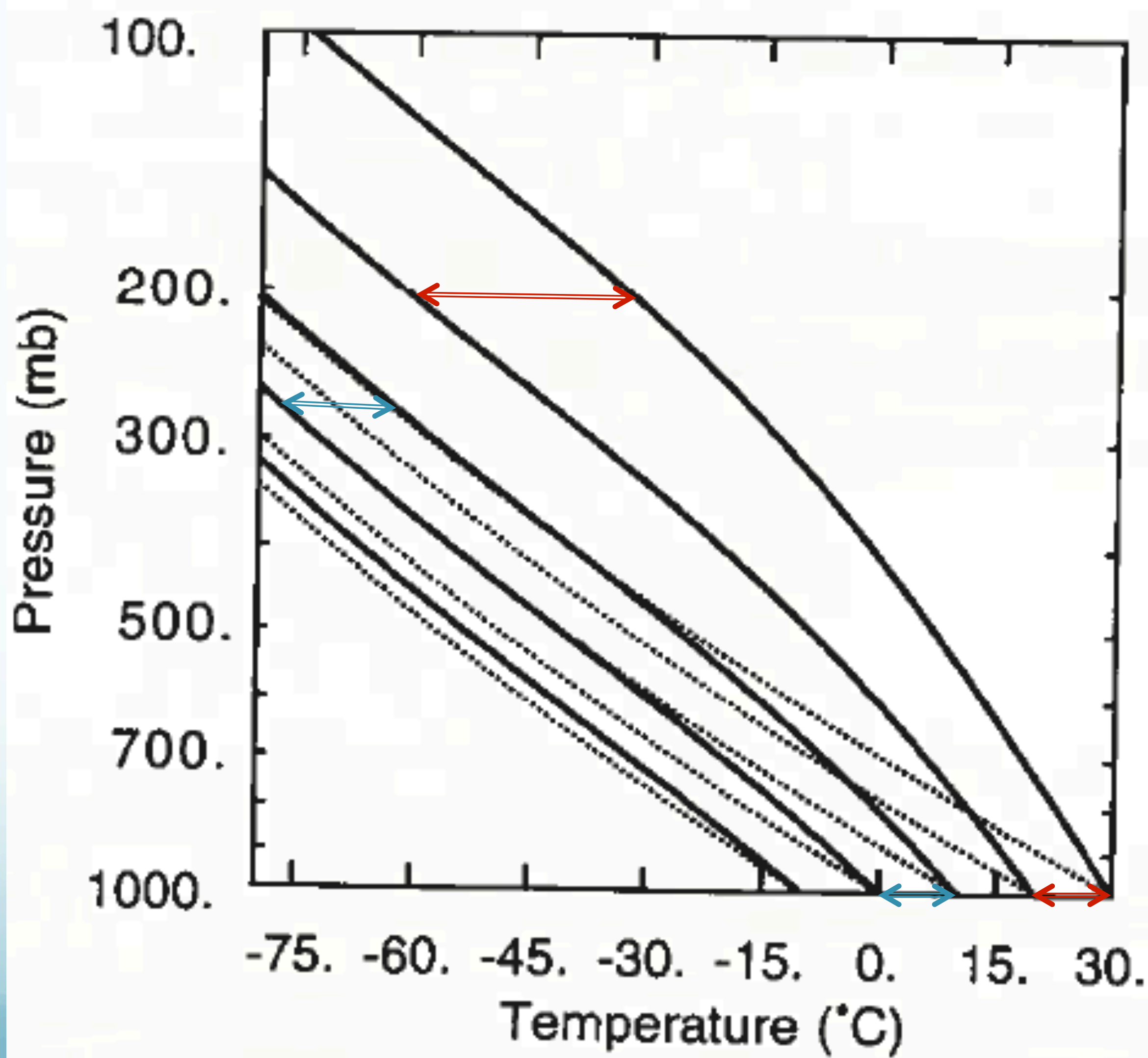
Review Last Time - 2

- Latent heat in the air is important. If you raise a saturated parcel of air and assume that the latent heat stays in the parcel as it is lifted up, then the lapse rate of temperature is a lot less than for dry thermodynamics.
- The warmer the surface air, the greater the difference between the dry and moist lapse rates.



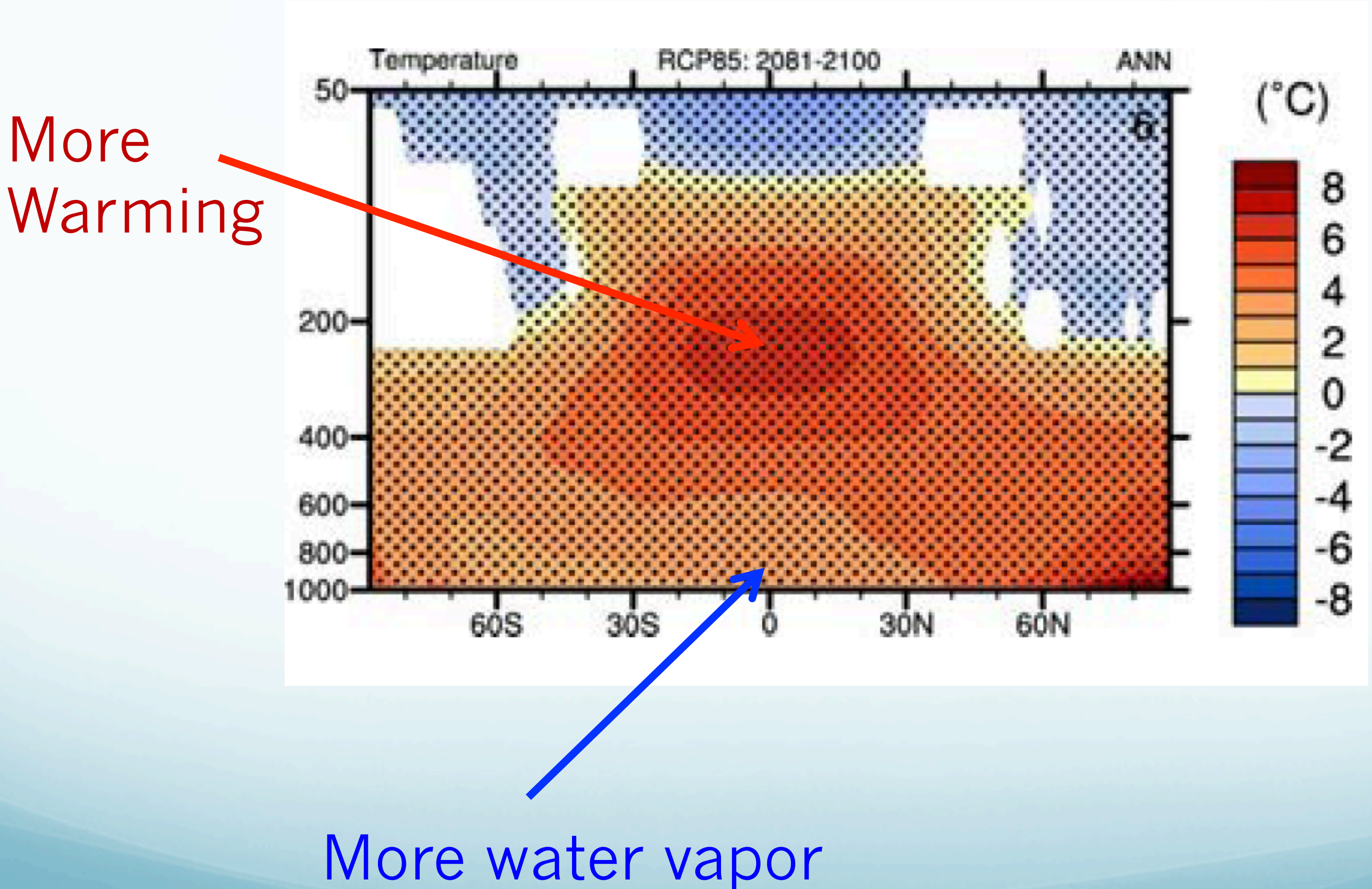
Moist Adiabatic Lapse Rates are Temperature Dependent

- Dashed -Dry
- Solid – Moist
- As surface temperature goes up, the moist lapse rate decreases,
- and the warming amplification with altitude increases



Zonal Structure of Change

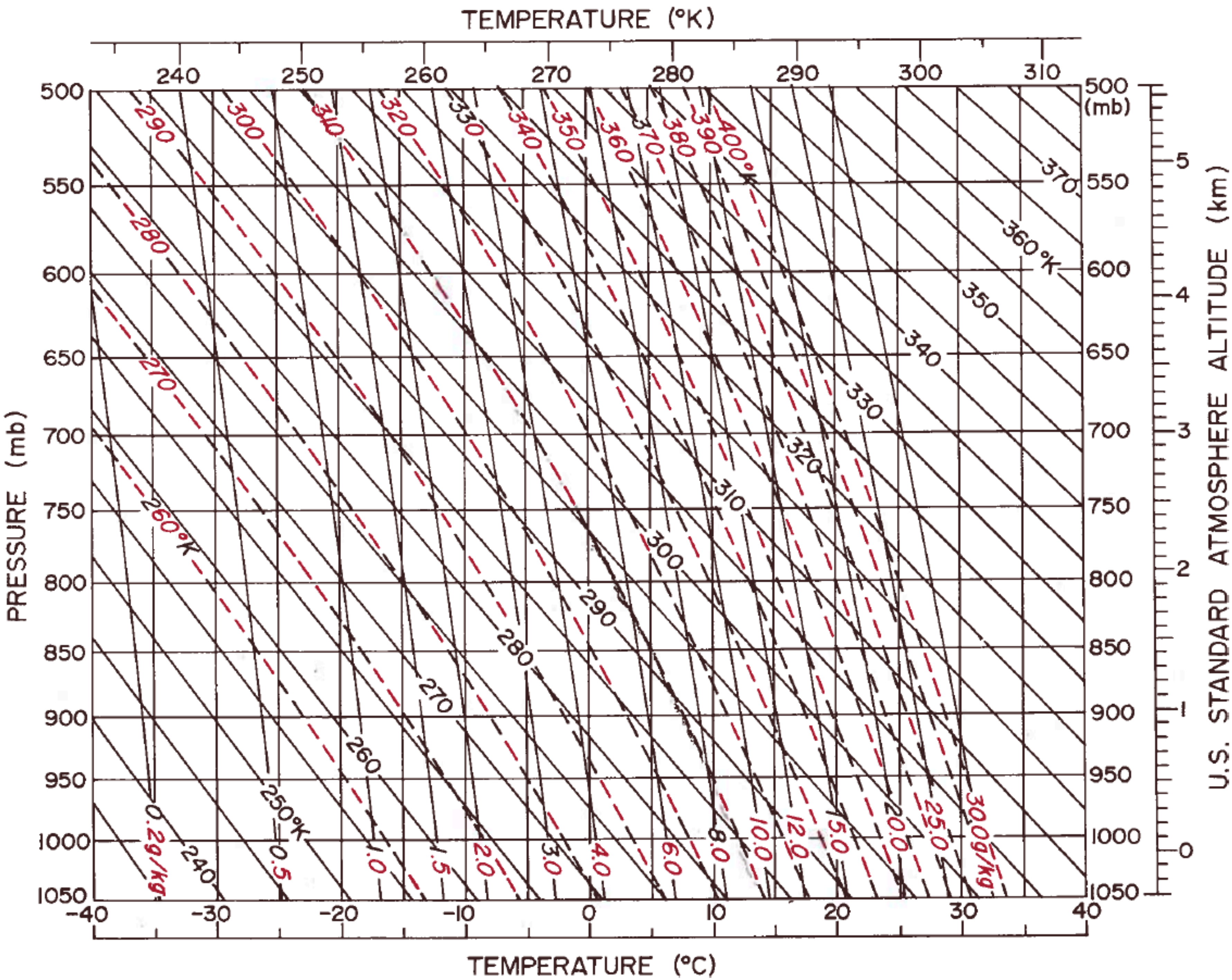
In Lecture 1 we showed that the warming increases with altitude in the Tropics. This is because of water vapor increasing near the surface, following the 7% per C° rule



Pseudoadiabatic Chart

- We will use coordinates of $p^{R/cp}$ versus T
- In these coordinates, lines of constant potential temperature are straight and are called **adiabats** – they are black and marked in °K
- Lines of constant equivalent potential temperature are called **pseudoadiabats** and are red dashed, also marked in °K
- Lines of constant saturation mass mixing ratio, w_s are plotted in solid red. (how slope these?)

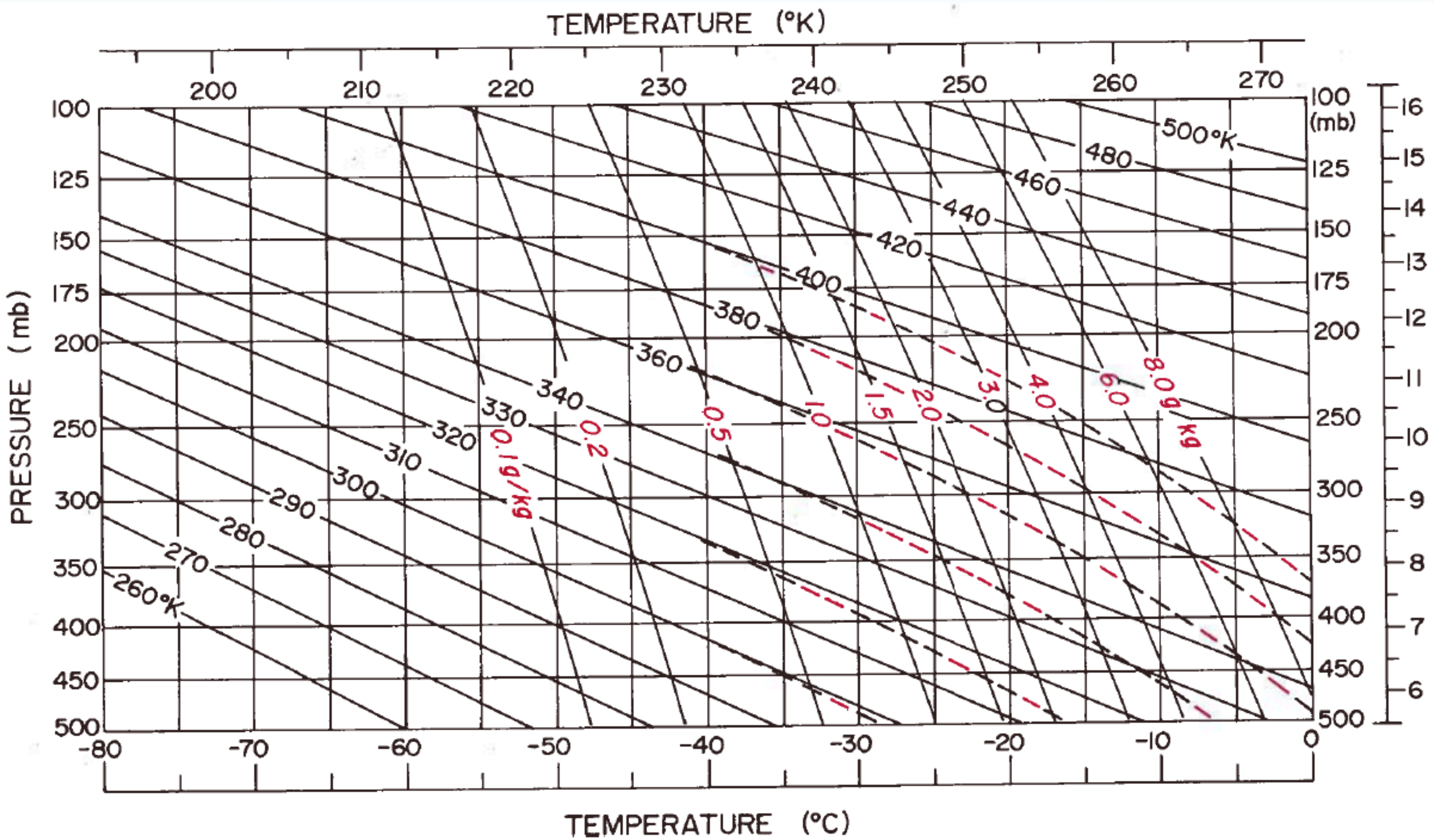
$$\boxed{w_s = \frac{R_d}{R_v}\left(\frac{e_s}{p-e_s}\right) \approx \frac{R_d}{R_v}\frac{e_s}{p}}$$



TEMPERATURE (°K)
TEMPERATURE (°C)
PRESSURE (mb)
U.S. STANDARD ATMOSPHERE ALTITUDE (km)

Pseudoadiabatic Chart

(higher up)



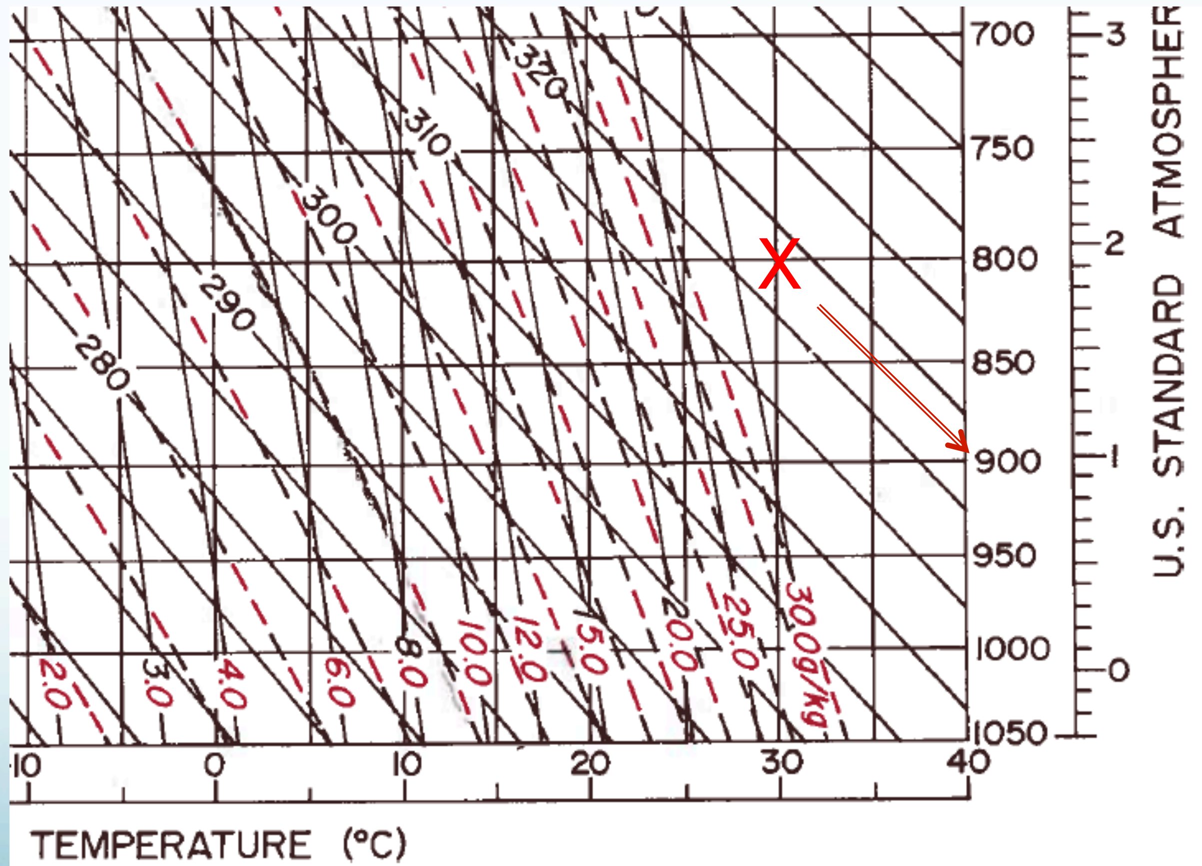
Potential Temperature

- The plane is pressurized to 800mb, the temperature in side the cabin is 30°C, the captain dives down to 1000mb. If no heat is added or subtracted. How much does the temperature increase?
- The potential temperature, using 1000mb as the reference pressure is

$$\theta = T\left(\frac{p_0}{p}\right)^{\frac{R}{c_p}} = \text{constant} = 303K\left(\frac{1000}{800}\right)^{\frac{287}{1004}} = 323K = 50^{\circ}C$$

Same problem with chart

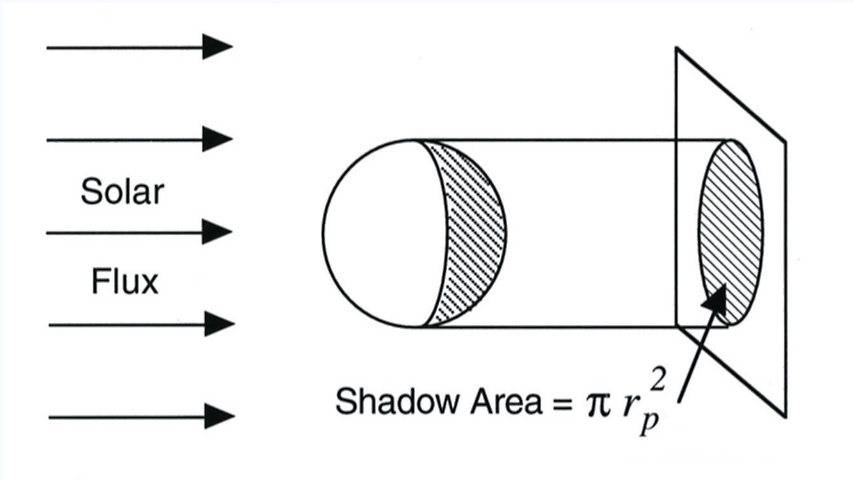
- Start at 800mb , 30˚C. Read potential temperature, assume reference pressure for chart is 1000mb
- Get 326K
- Close to 323K
- Reference p 1013.25mb most likely



Global Energy Balance

- Started to talk about solar constant and emission temperature of planet.
- Shadow Area = Emission Area/4, since Earth is a sphere, more or less

Emission Temperature of Planet



- Set absorbed solar radiation equal to longwave emission
- Solar absorption and longwave emission areas are different
- Shadow Area is solar absorption area, a circle

Emission Temperature

- Absorbed Solar radiation = Solar Irradiance x absorptivity x shadow area = Watts

$$S_0 (1-\alpha_p)\, \pi r_p^2 = Absorbed\ Solar\ Radiation$$

- Emitted Terrestrial Radiation = emissivity x blackbody emission x emission area = Watts

$$\sigma T_e^4\ 4\pi r_p^2 = Emitted\ Terrestrial\ Radiation$$

- In = Out

$$S_0 (1-\alpha_p)\, \pi r_p^2 = \sigma T_e^4\ 4\pi r_p^2$$

$$\boxed{T_e = \sqrt[4]{\frac{S_0 (1-\alpha_p)}{4\sigma}}}$$

Earth's Emission Temperature

- S_0 is about 1360 Wm^{-2}
- Planetary albedo is about 0.3

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

$$T_e = \sqrt[4]{\frac{1360\,Wm^{-2}(1-0.3)}{4\times 5.67\times 10^{-8}\,Wm^{-2}K^{-4}}} \simeq 255K = -18^{\circ}C = 0^{\circ}F$$

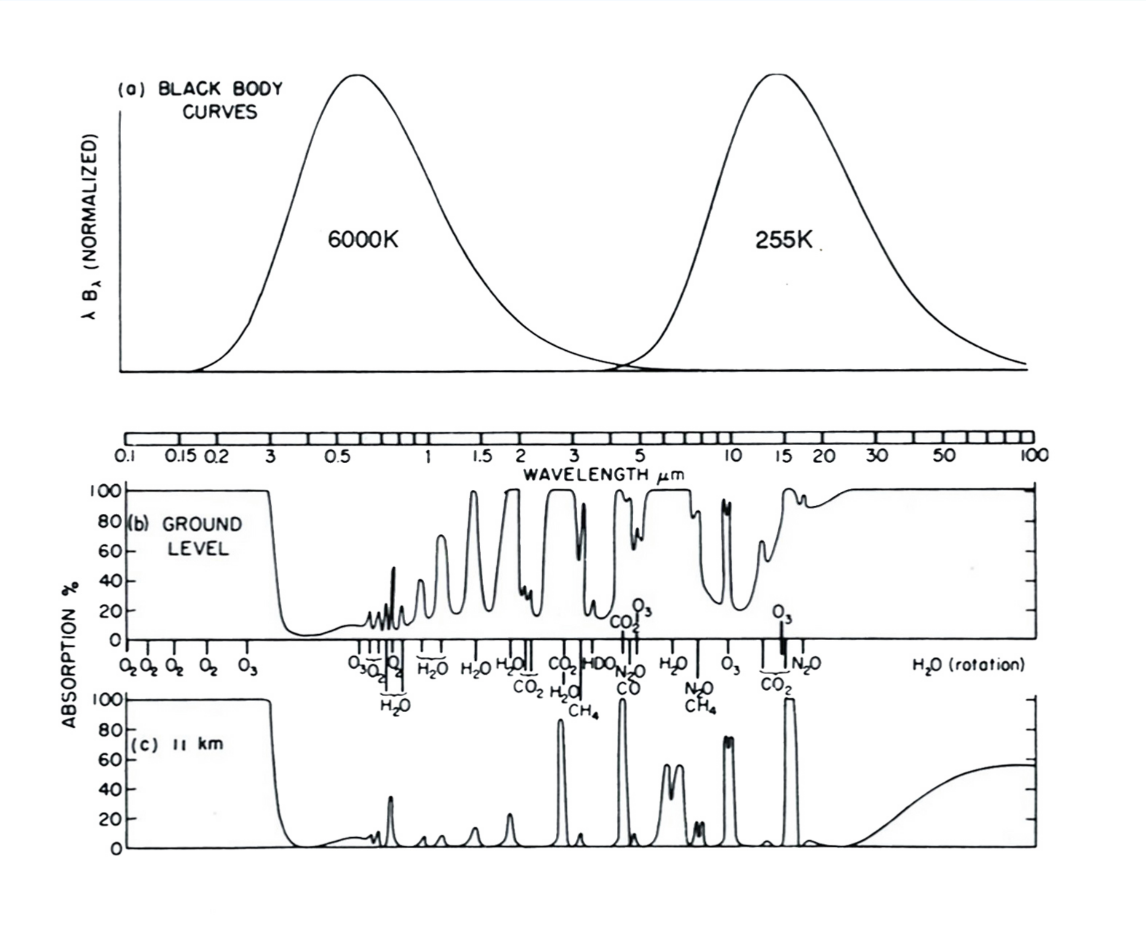
- Not a good estimate of Earth's surface temperature, which is about 288K
- What's missing?

Emission Temperature

- Effective emission temperature of Earth is about 255K, whereas surface temperature of Earth is about 288K.
- If lapse rate is 6.5 K/km, then, emission is coming from an effective altitude of
- (288K – 255K)/(6.5 K/km) = 5 km altitude

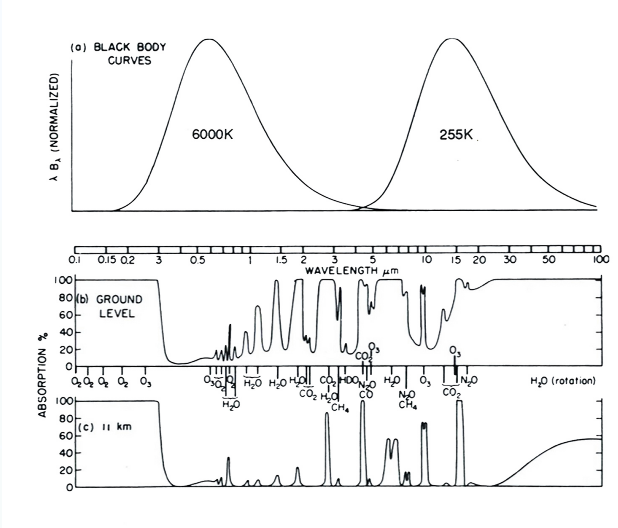
Solar and Terrestrial Radiation Transmission in the Atmosphere

- Solar and Terrestrial are separate and different

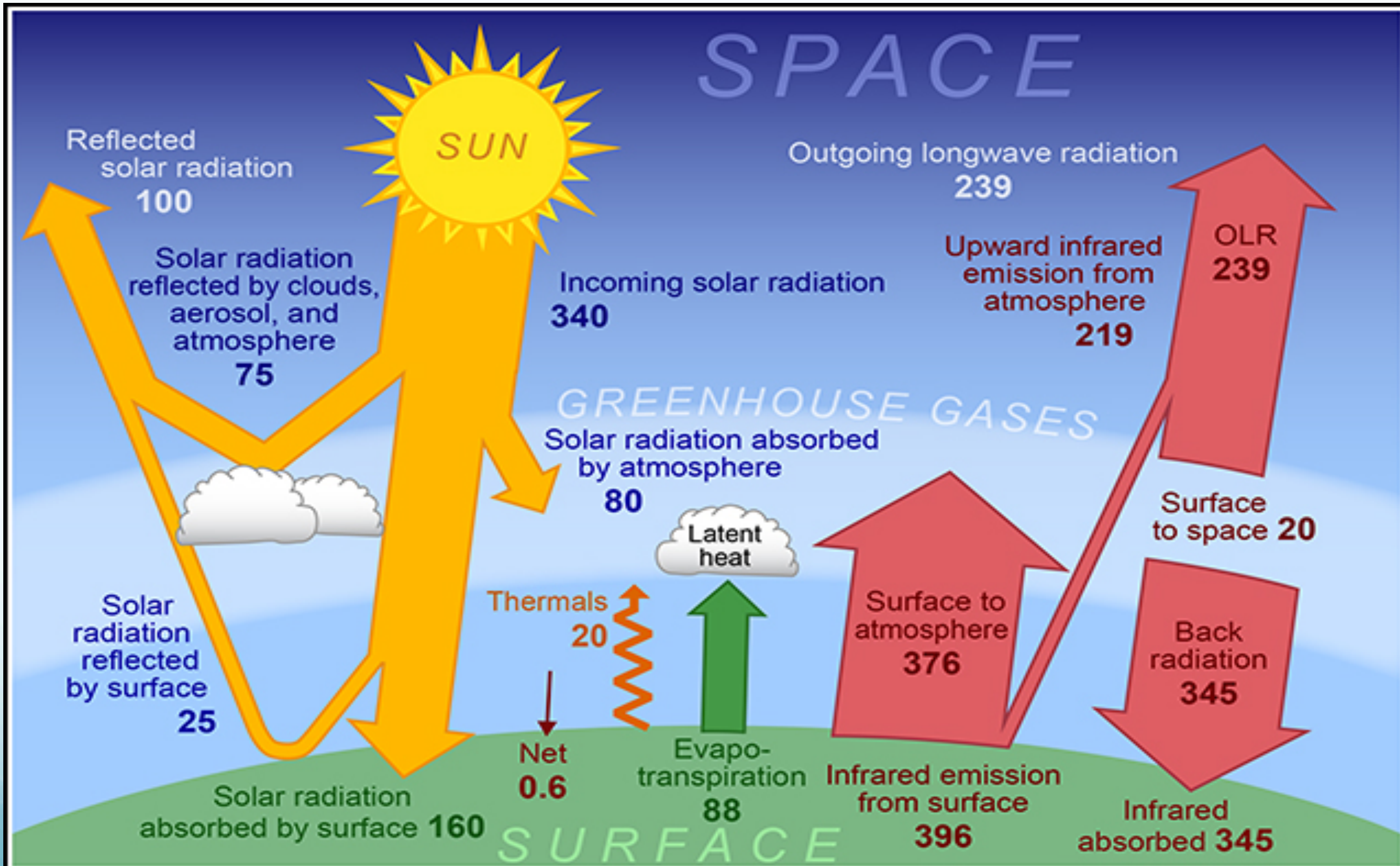


Atmosphere & Radiation

- Atmosphere is
  - Fairly transparent to solar radiation, especially in visible wavelengths
  - Fairly opaque to terrestrial radiation (thermal infrared) Only the atmospheric window between 8 and 12 microns is fairly open
  - Water is the most important absorber, both in solar and terrestrial. CO_2 and Ozone come second and third.
  - Ultraviolet is screened out in the stratosphere, or above.



Global Vertical Energy Balance



Energy Balances

- Top of Atmosphere

Incoming solar – reflected solar - emitted terrestrial radiation = storage
340 Wm^{-2} - 100 Wm^{-2} - 239 Wm^{-2} = 0.6 Wm^{-2}

- Atmosphere

Absorbed solar + Thermals + Evaporation + surface longwave – OLR = 0
80 Wm^{-2} + 20 Wm^{-2} + 88 Wm^{-2} +51 Wm^{-2} – 239 Wm^{-2} = 0

- Surface

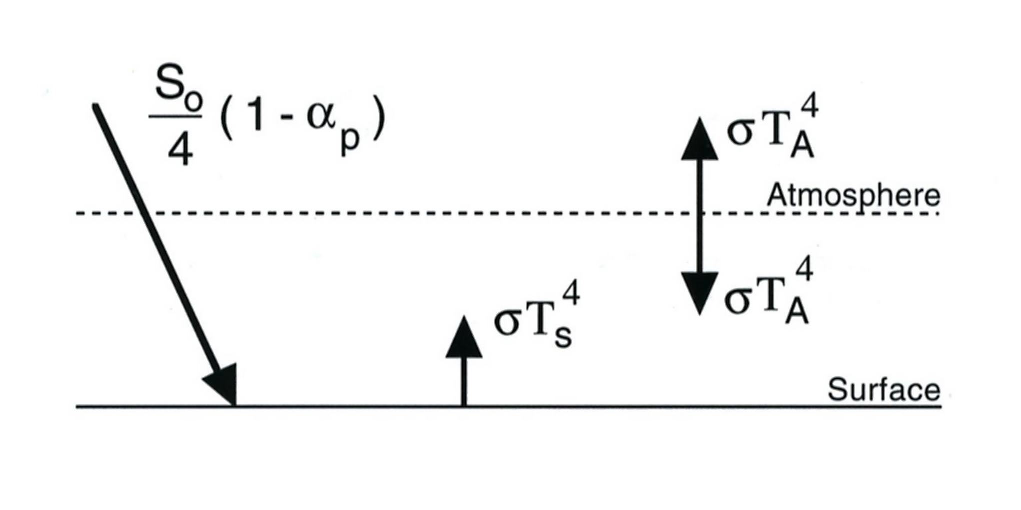
Absorbed solar – Thermals – latent heating – surface longwave = Storage
160 Wm^{-2} – 20 Wm^{-2} – 88 Wm^{-2} – 51 Wm^{-2} = 0.6 Wm^{-2}

Magnitude of Greenhouse Effect for Earth

- Outgoing Longwave Radiation (OLR) is 239 Wm^{-2},
  Surface longwave emission is 396 Wm^{-2}
  Magnitude of TOA greenhouse effect = 396 – 239 = 157Wm^{-2} heating

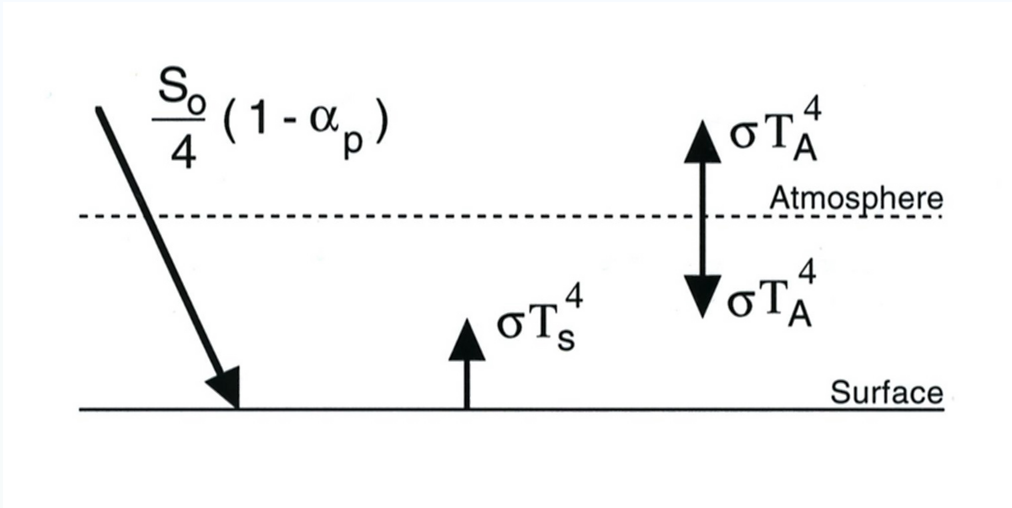
- Solar heating of surface is 160Wm^{-2}
  Downward longwave heating of surface by atmosphere is 345Wm^{-2}
  Magnitude of surface greenhouse effect = twice the solar heating of the surface.

Greenhouse Effect
Simplest possible model



- The atmosphere is nearly transparent to solar radiation (UV, Visible and Near IR).
- The atmosphere is nearly opaque to terrestrial radiation (Thermal Infrared)
- The surface absorbs all the solar radiation and emits like a black body

Greenhouse Effect



- TOA $$\frac{S_0}{4}(1-\alpha_p) = \sigma T_A^4 = \sigma T_e^4$$
- Atmosphere $$\sigma T_s^4 = 2\sigma T_A^4$$
- Surface $$\sigma T_A^4 + \frac{S_0}{4}(1-\alpha_p) = \sigma T_s^4$$

Magnitude of Greenhouse effect in simplest model

- TOA Definition
  surface emission – OLR =

$$\sigma T_s^4 = 2\sigma T_e^4,\ so\ G = 2\sigma T_e^4 - \sigma T_e^4 = \sigma T_e^4 = 239 Wm^{-2}$$

  Too big, real one 157 Wm^{-2}

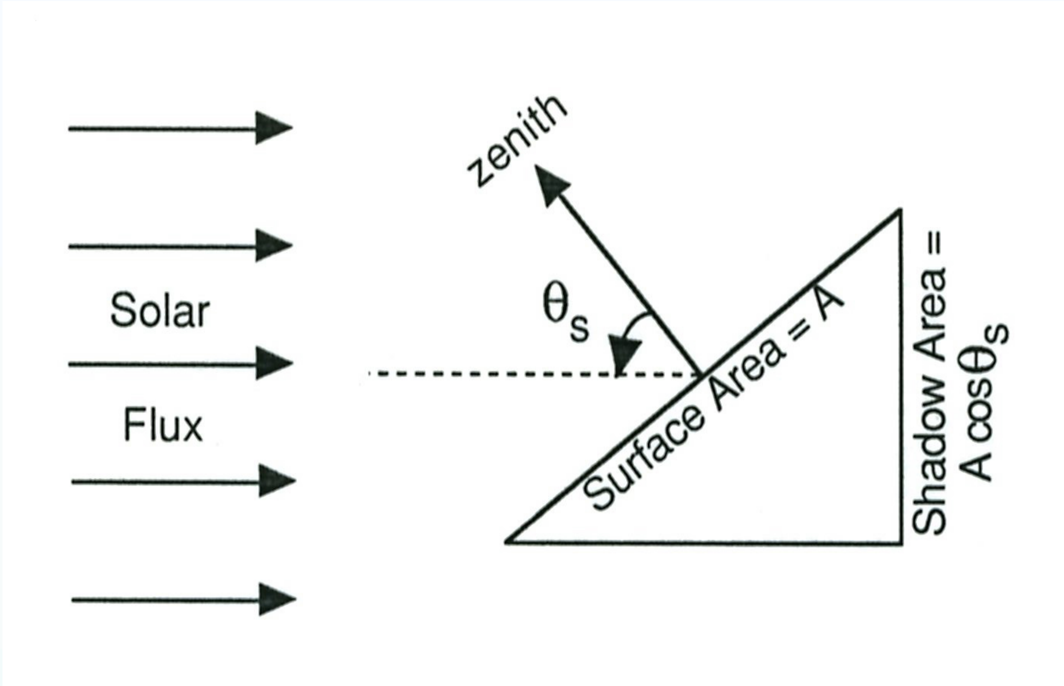
- Surface Definition
  Downward Longwave/Surface Solar

$$\sigma T_e^4 / \sigma T_e^4 = 1$$

  Too small, ratio should be about 2

- So model not perfect, duh . . . see CHAPTER 3

Solar Zenith Angle



- A parallel wall of irradiance arrives at Earth from the Sun. The irradiance per unit area depends on the zenith angle.

Solar Zenith Angle

- Insolation at top of atmosphere – TOA

$$Q = S_0 \left(\frac{\bar{d}}{d} \right)^2 \cos\theta_s$$

- Zenith Angle Formula

$$\cos\theta_s = \sin\phi \sin\delta + \cos\phi \cos\delta \cos h$$

- Sunrise/set hour angle

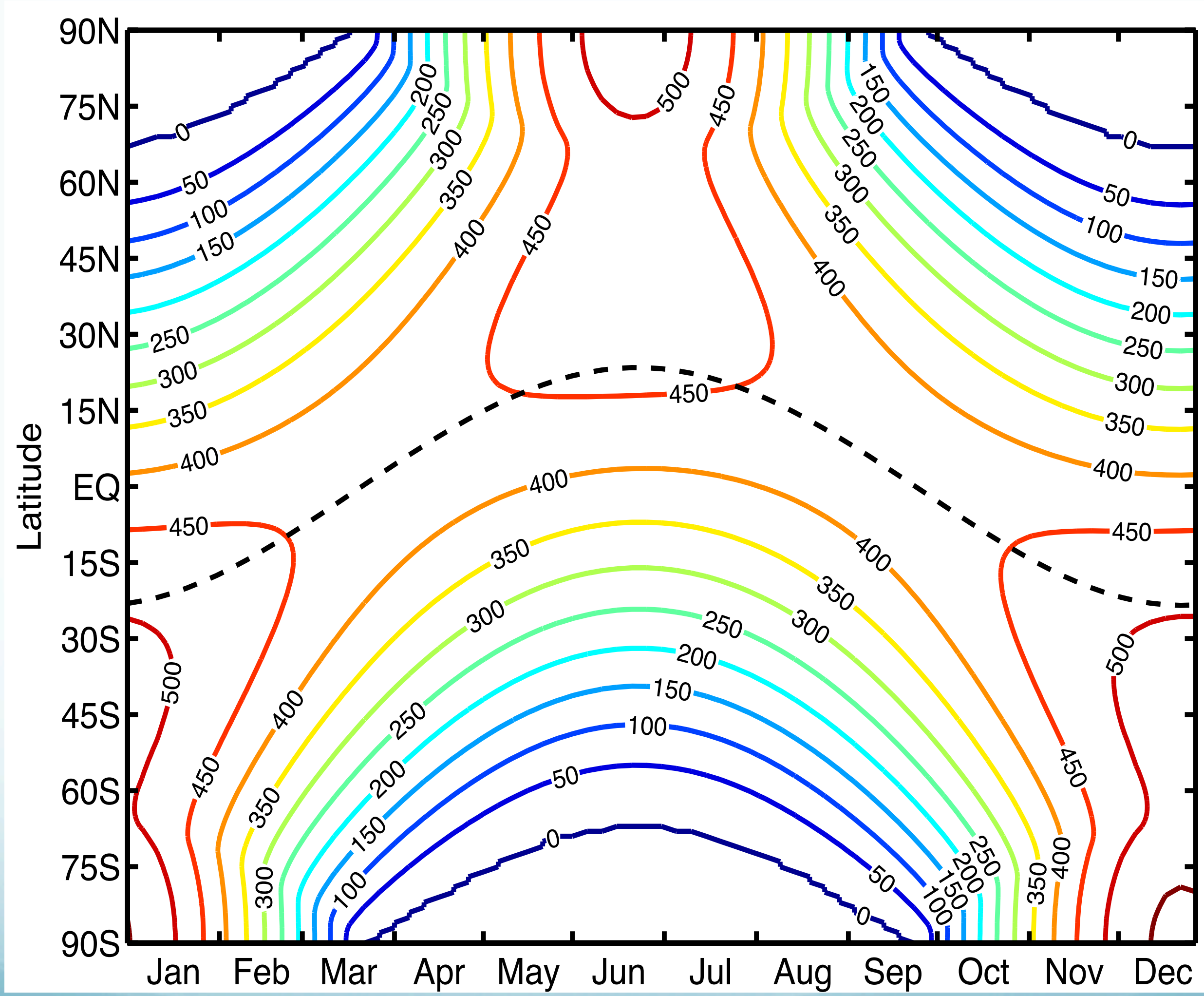
$$\cos h_0 = -\tan\phi \tan\delta$$

Daily Insolation

$$\bar{Q}^{day} = \frac{S_0}{\pi}\left(\frac{\bar{d}}{d}\right)^2 \left[h_0 \sin\phi \sin\delta + \cos\phi \cos\delta \sin h_0\right]$$

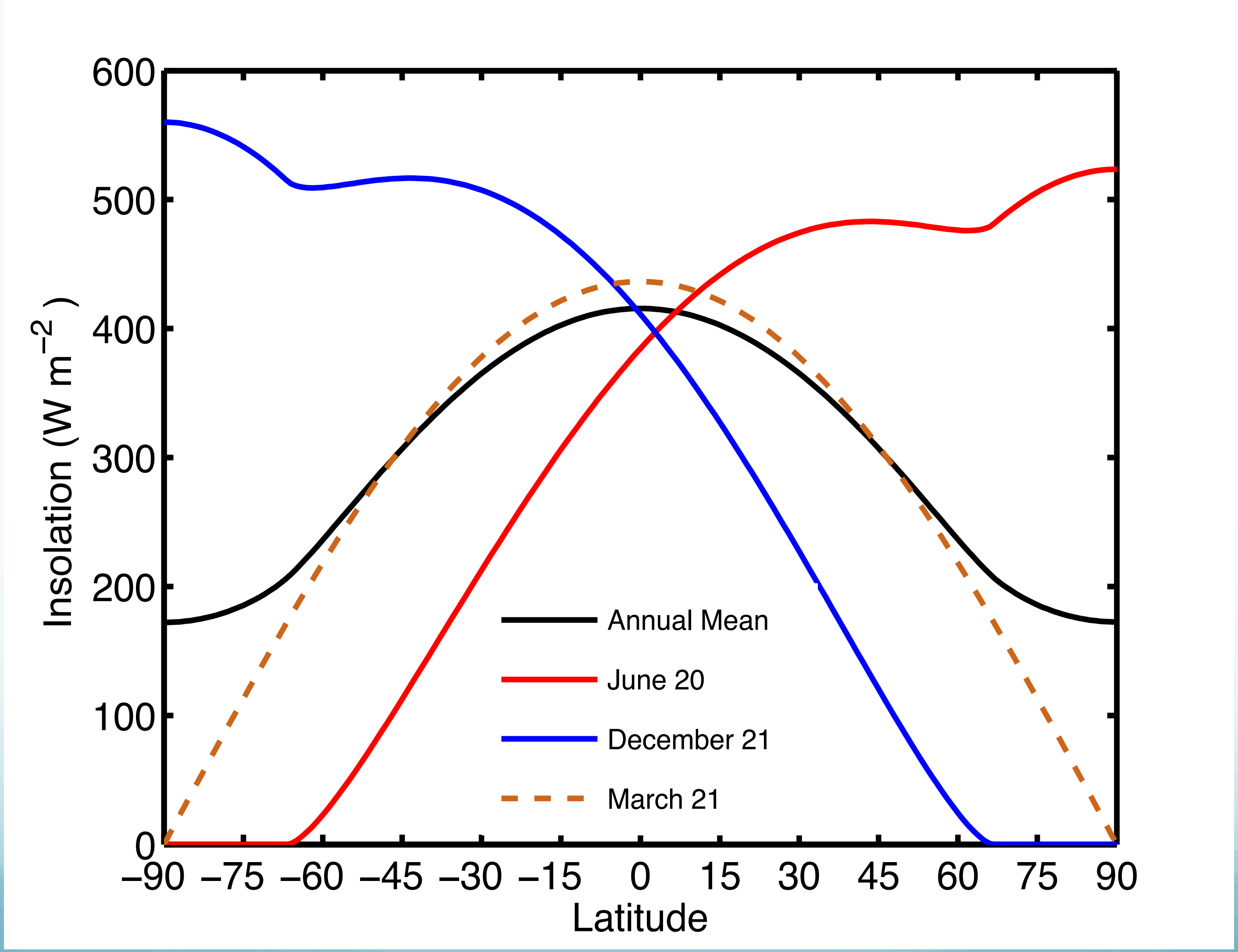
See book for derivation

Daily Insolation Plot



Watts per square meter

Latitude Distribution of Daily Insolation



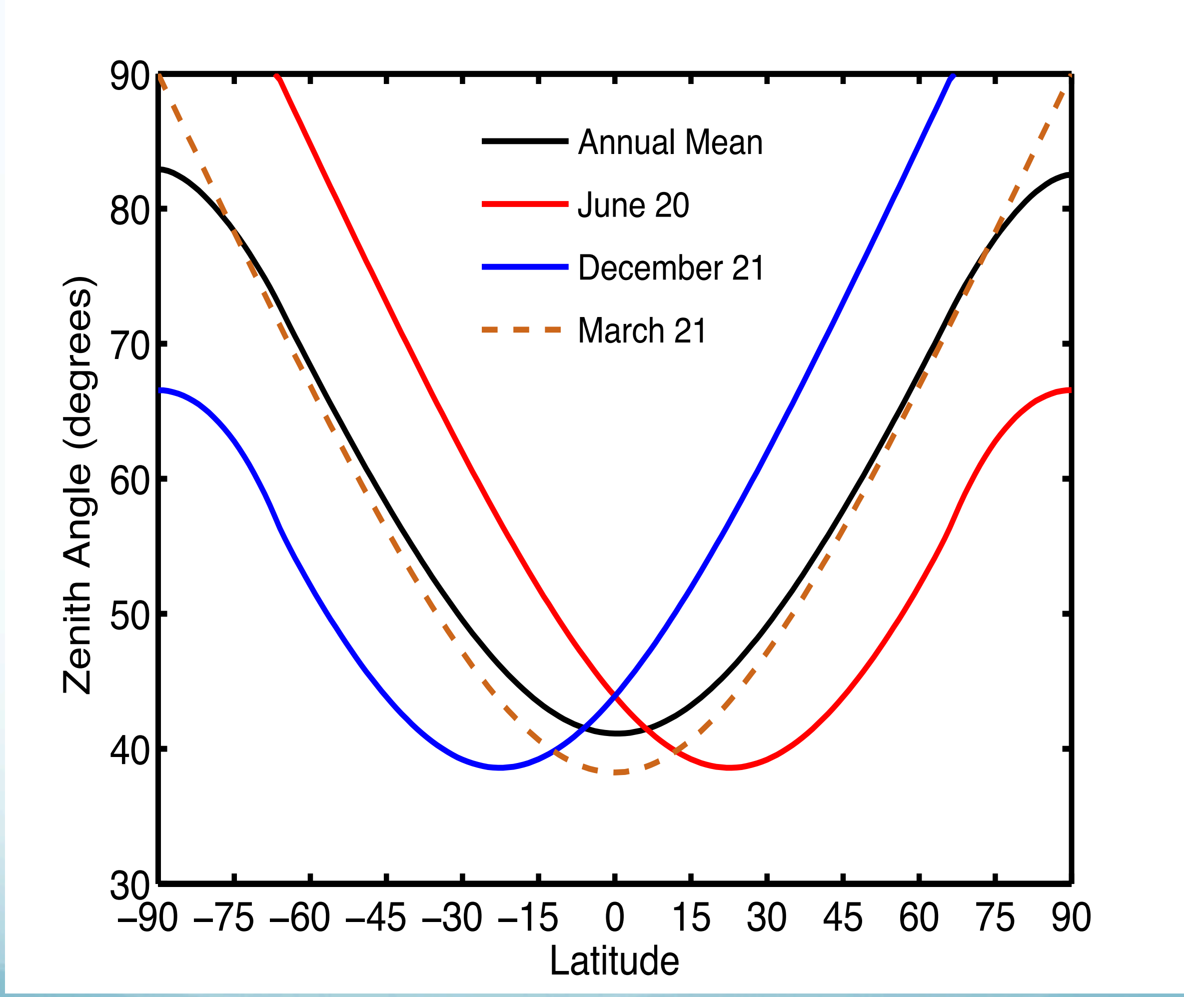
Insolation-Weighted Zenith Angle

- IF you wanted to calculate an average zenith angle, you should weight the average by insolation.
- Of course this would still give the wrong albedo, because the reflection of surfaces depend on zenith angle in complicated ways.

$$\overline{\cos\theta_s}^{day} = \frac{\int_{-h_0}^{h_0} Q \cos\theta_s \, dh}{\int_{-h_0}^{h_0} Q \, dh}$$

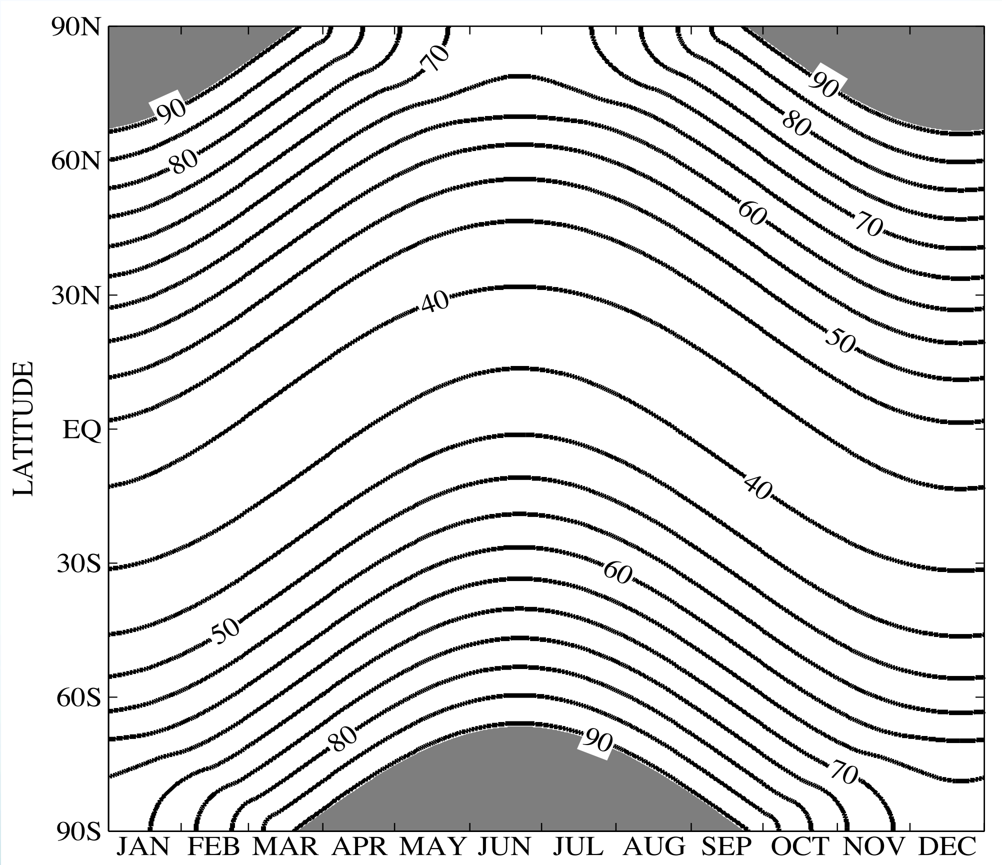
$$Q = S_0 \left(\frac{\bar{d}}{d} \right)^2 \cos\theta_s$$

Solar Zenith Angle Insolation-Weighted over diurnal cycle



Insolation-Weighted Zenith Angle

- is here



TOA Energy Balance

- Net Radiation = Absorbed Solar – OLR

$$R_{TOA} = Q_{abs} - OLR$$

$$Q_{abs} = S_{TOA}(1-\alpha)$$

- Albedo alpha is important

Albedo

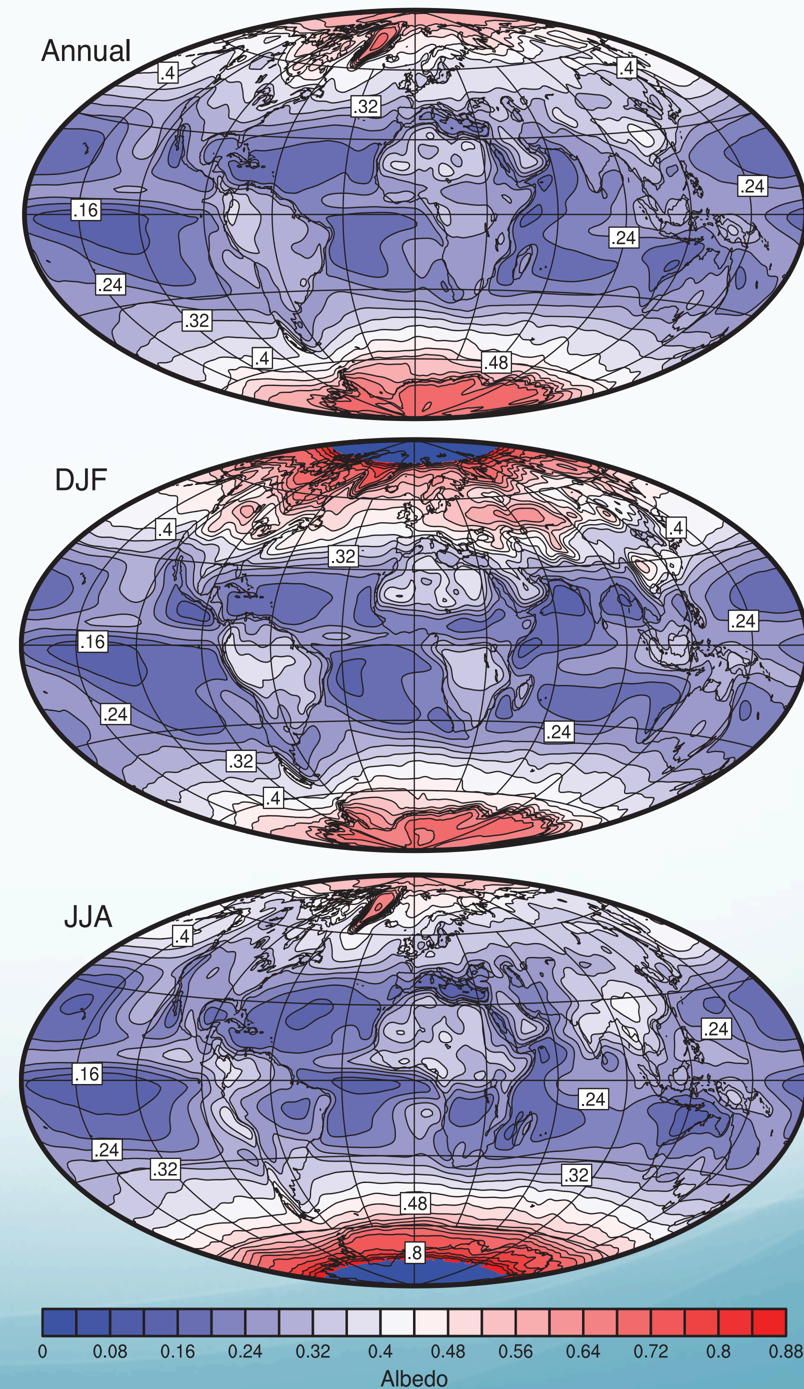
Albedo measured by satellite at top of atmosphere.

Includes effects of surface, clear atmosphere, and clouds

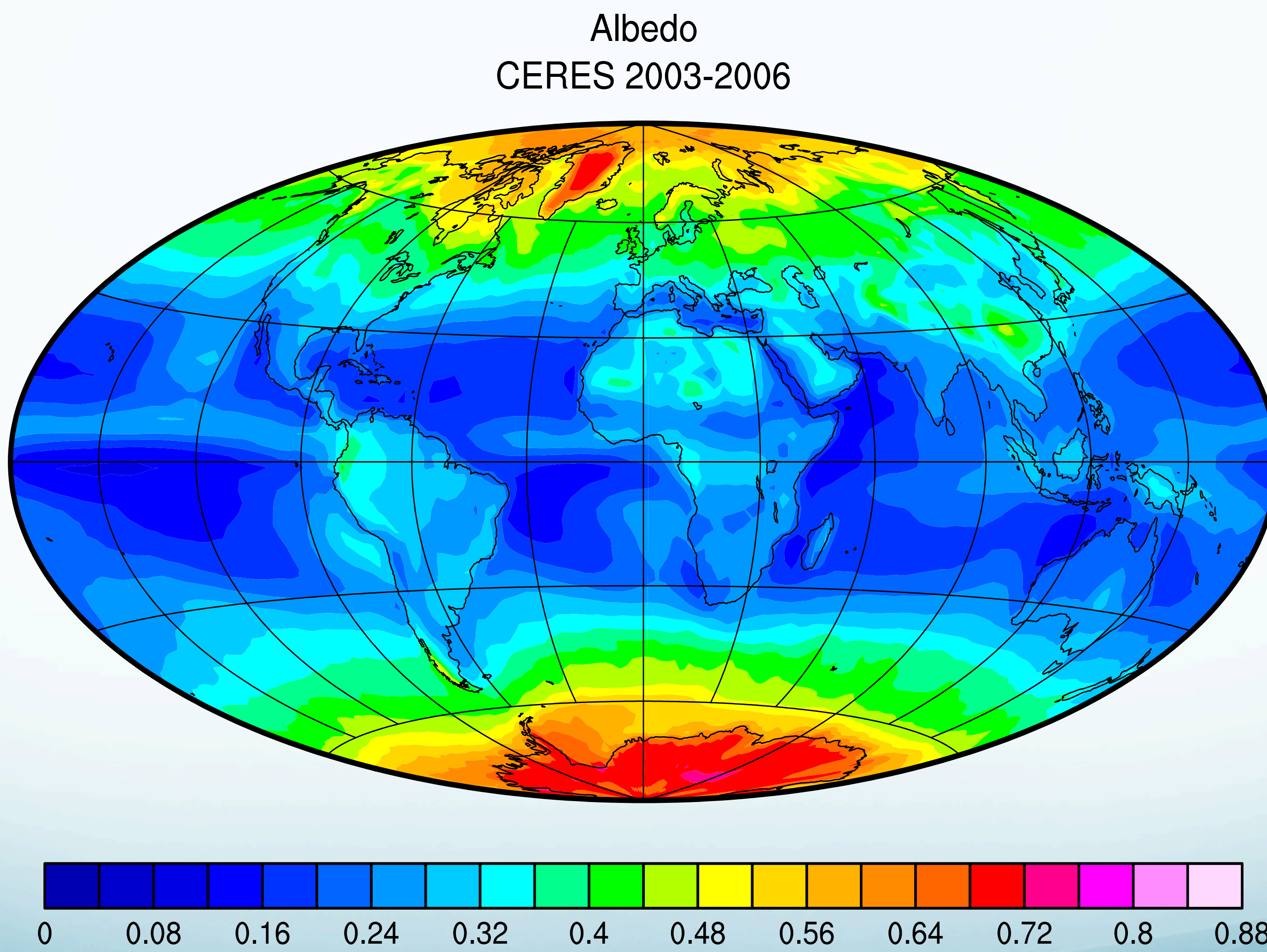
Higher over land than ocean, because of surface albedo.

Higher in high latitudes because of zenith angle, clouds and surface ice.

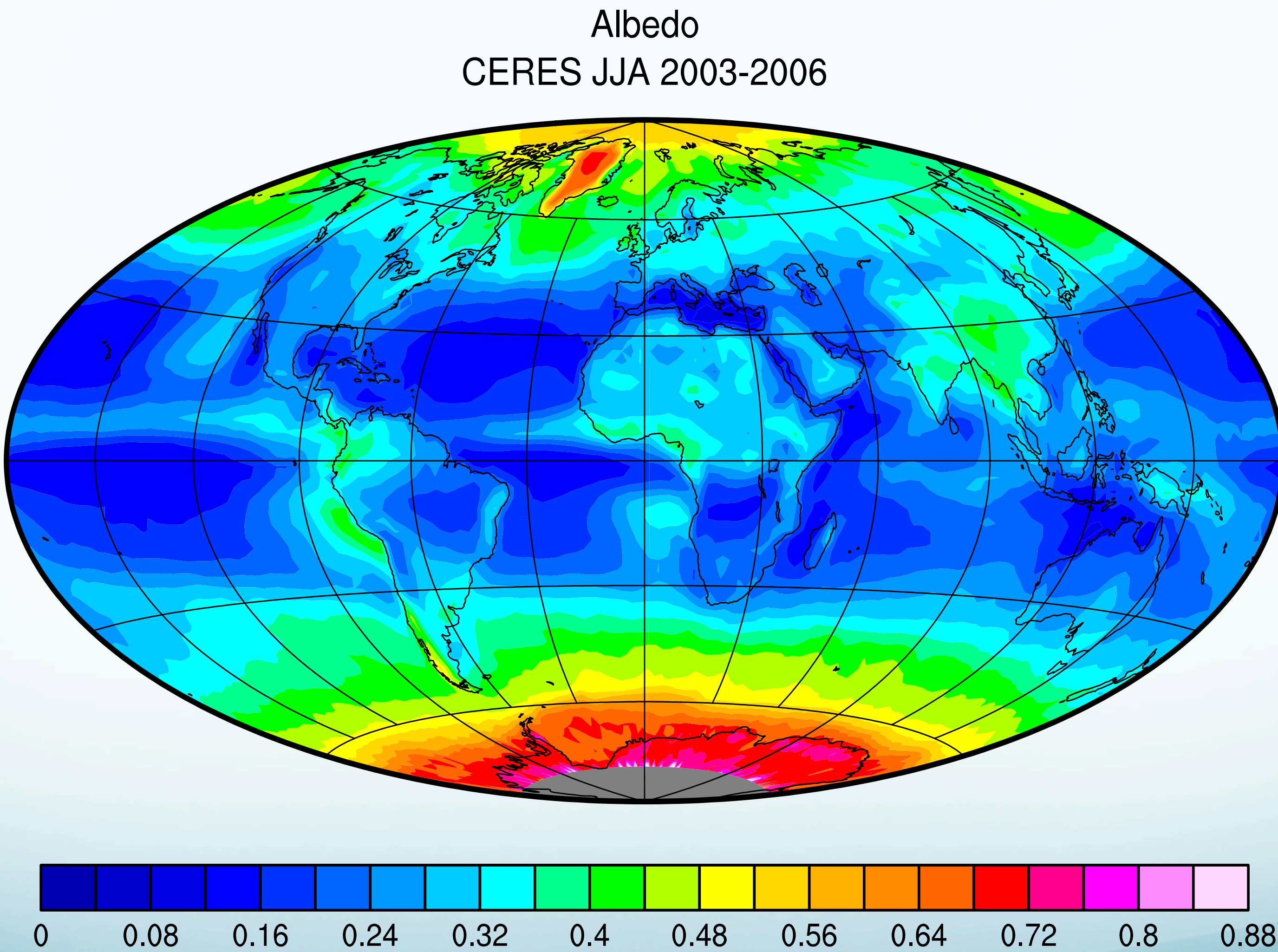
CERES data, 2000-2013.



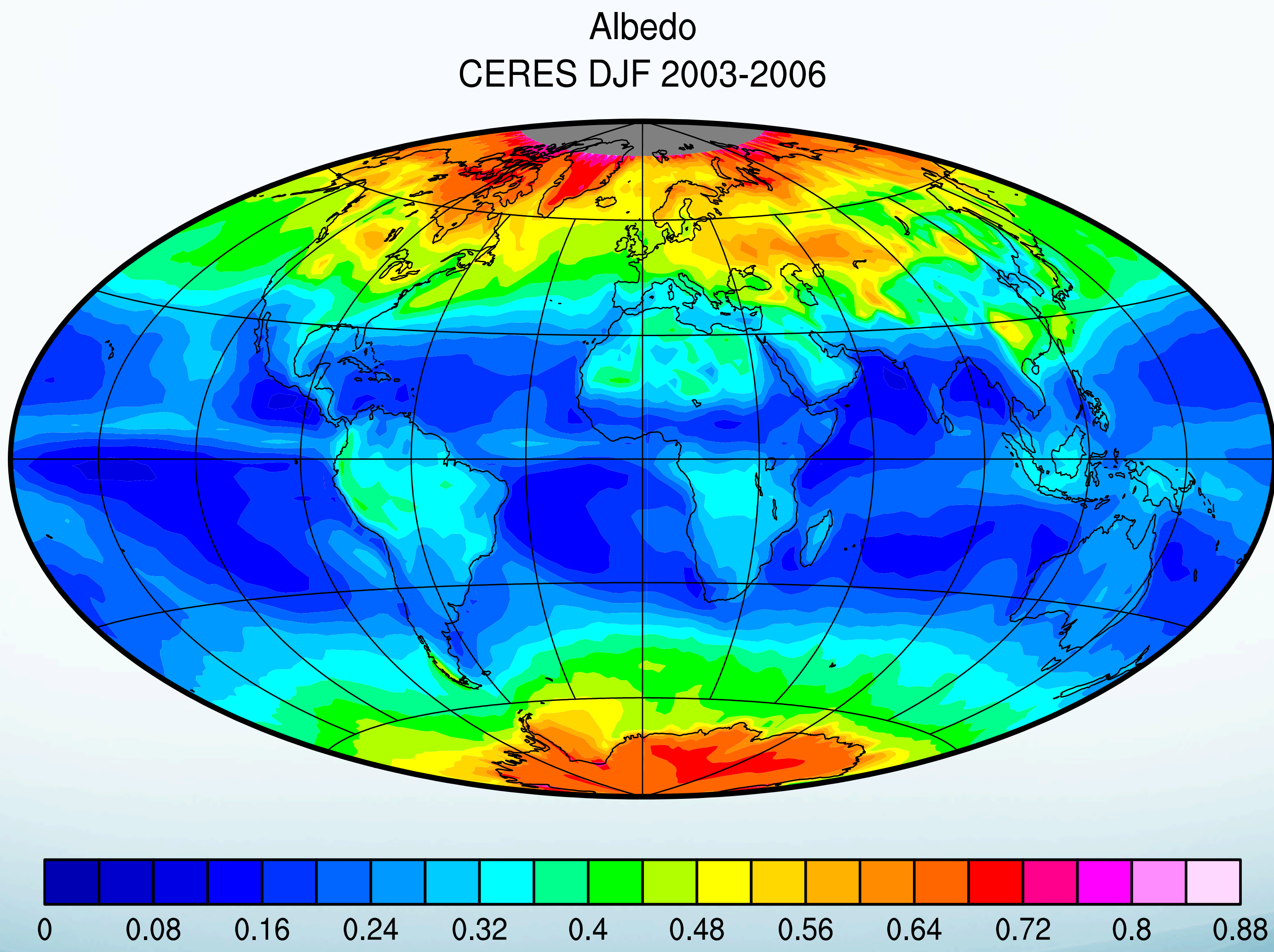
TOA Albedo – annual mean



TOA Albedo – JJA



TOA Albedo – DJF



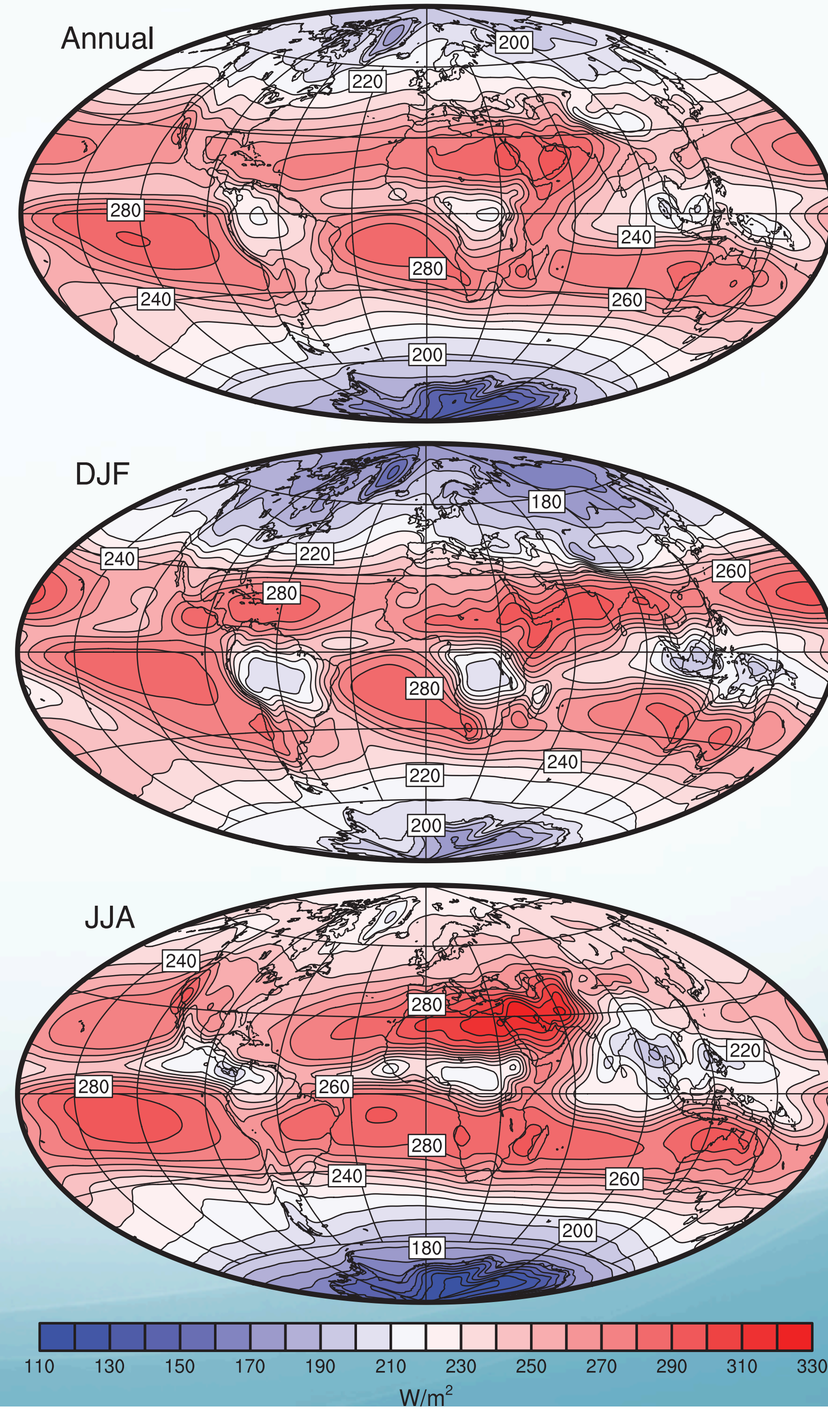
Outgoing Longwave Radiation (OLR)

Greatest in tropical latitudes that have not much cloud, since warm temperatures emit to space there.

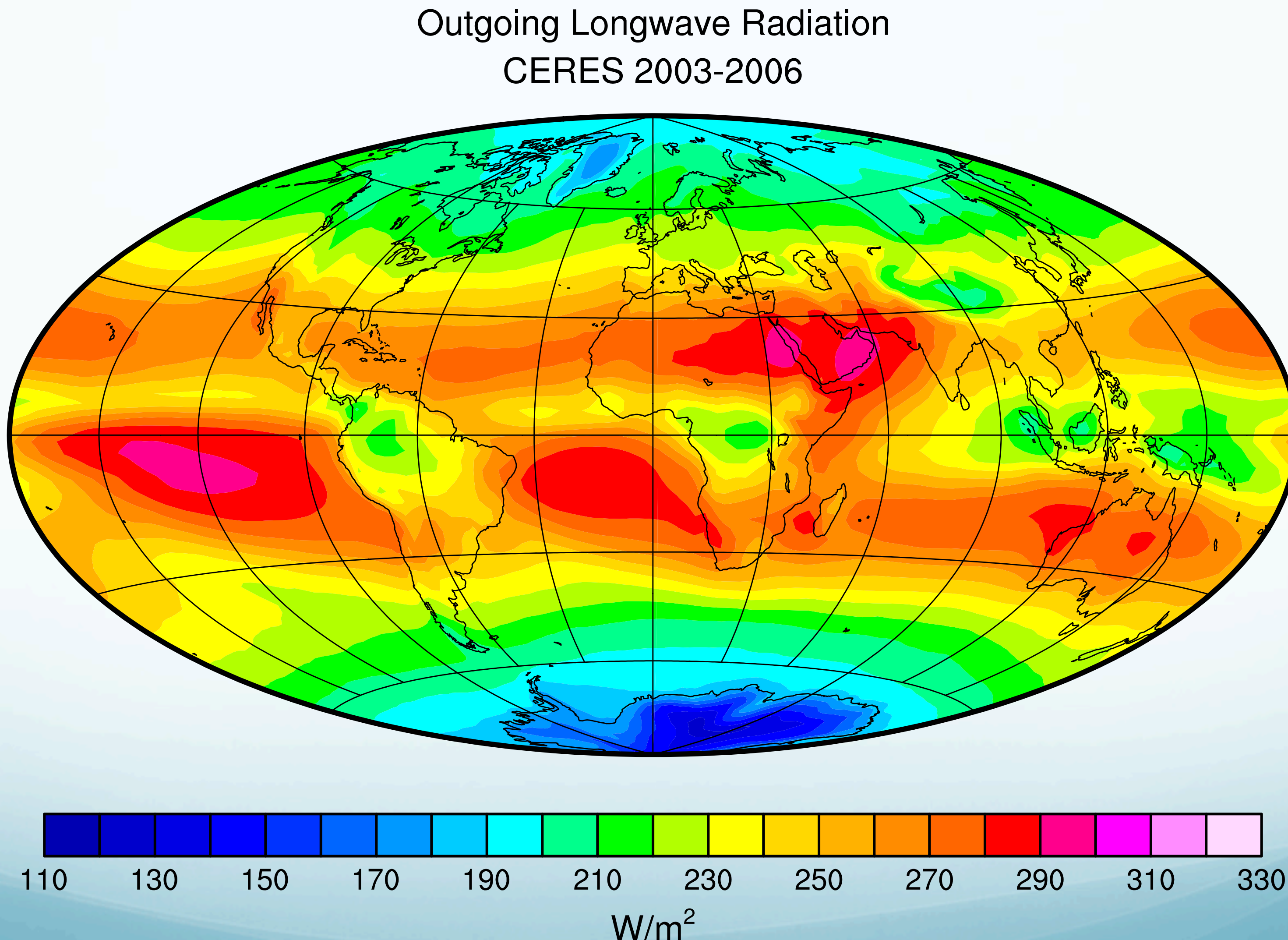
Lower over tropical cloudy regions since clouds are opaque to thermal infrared and their tops are cold.

Lower in high latitudes because it is colder there.

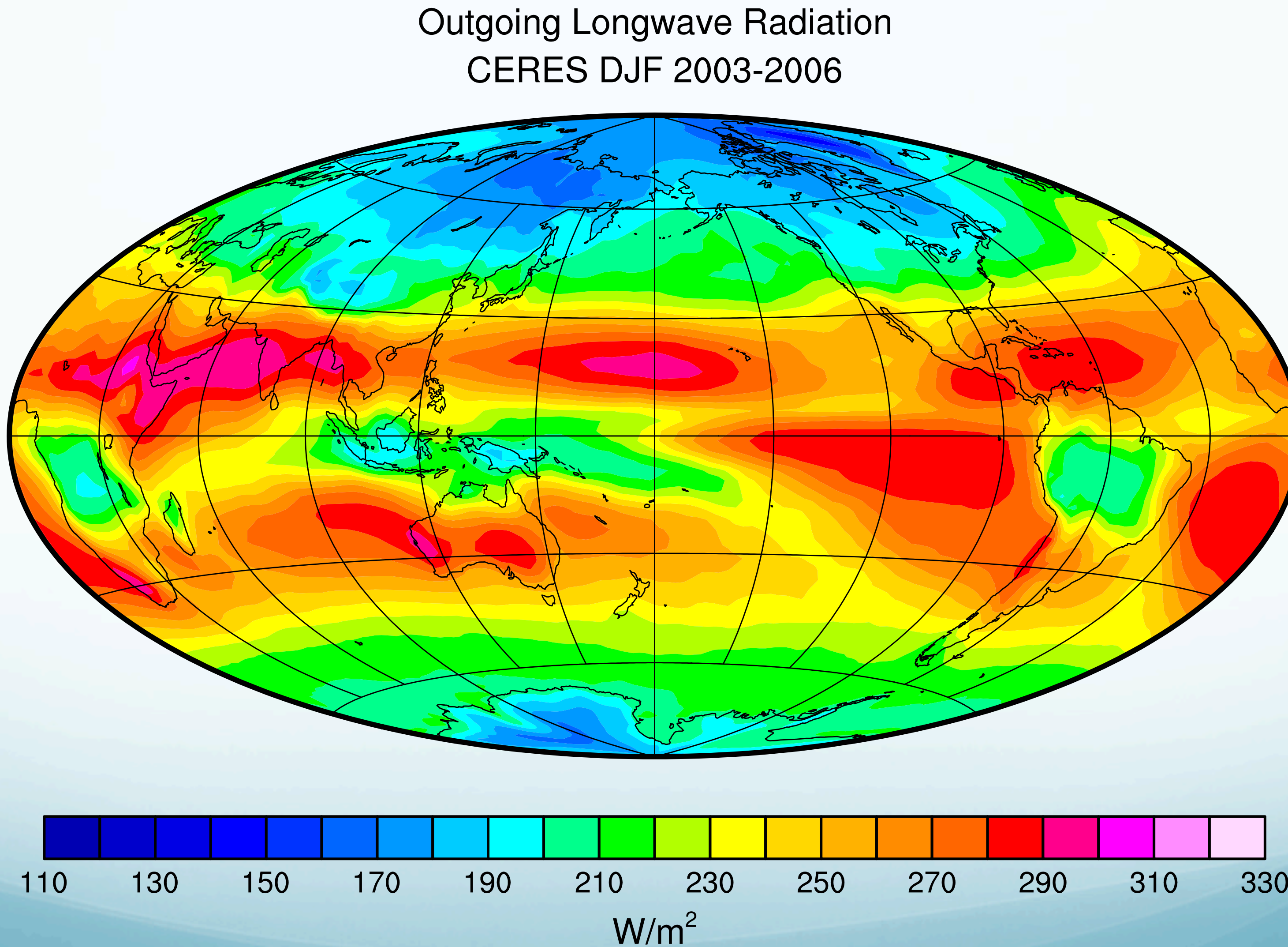
Super high over tropical deserts in summer because the surface is hot and there are few clouds or water vapor to trap IR.



OLR = Outgoing Longwave Radiation

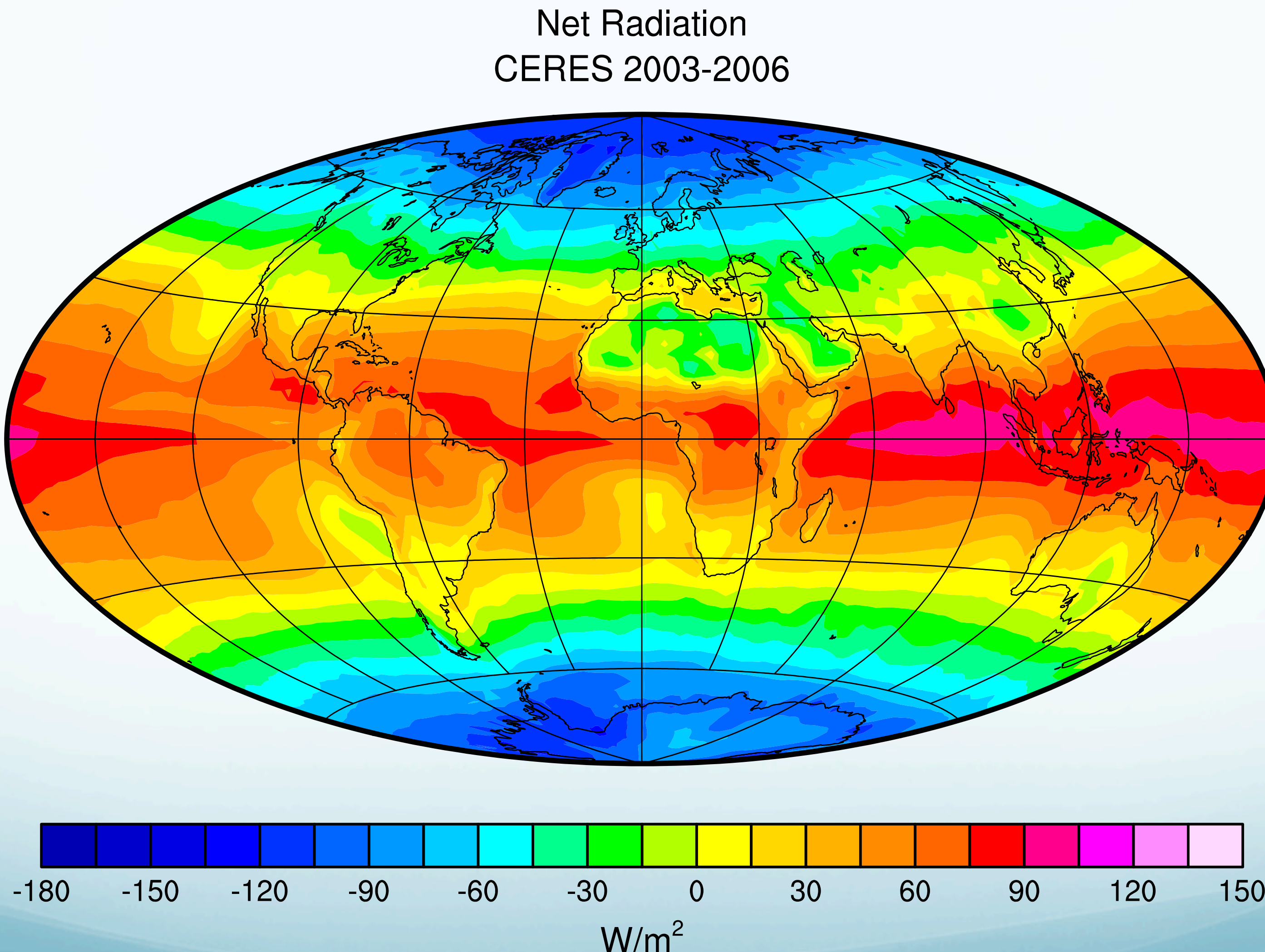


OLR for DJF



Outgoing Longwave Radiation
CERES DJF 2003-2006
110
130
150
170
190
210
230
250
270
290
310
330
W/m2

Net Radiation – Annual Mean

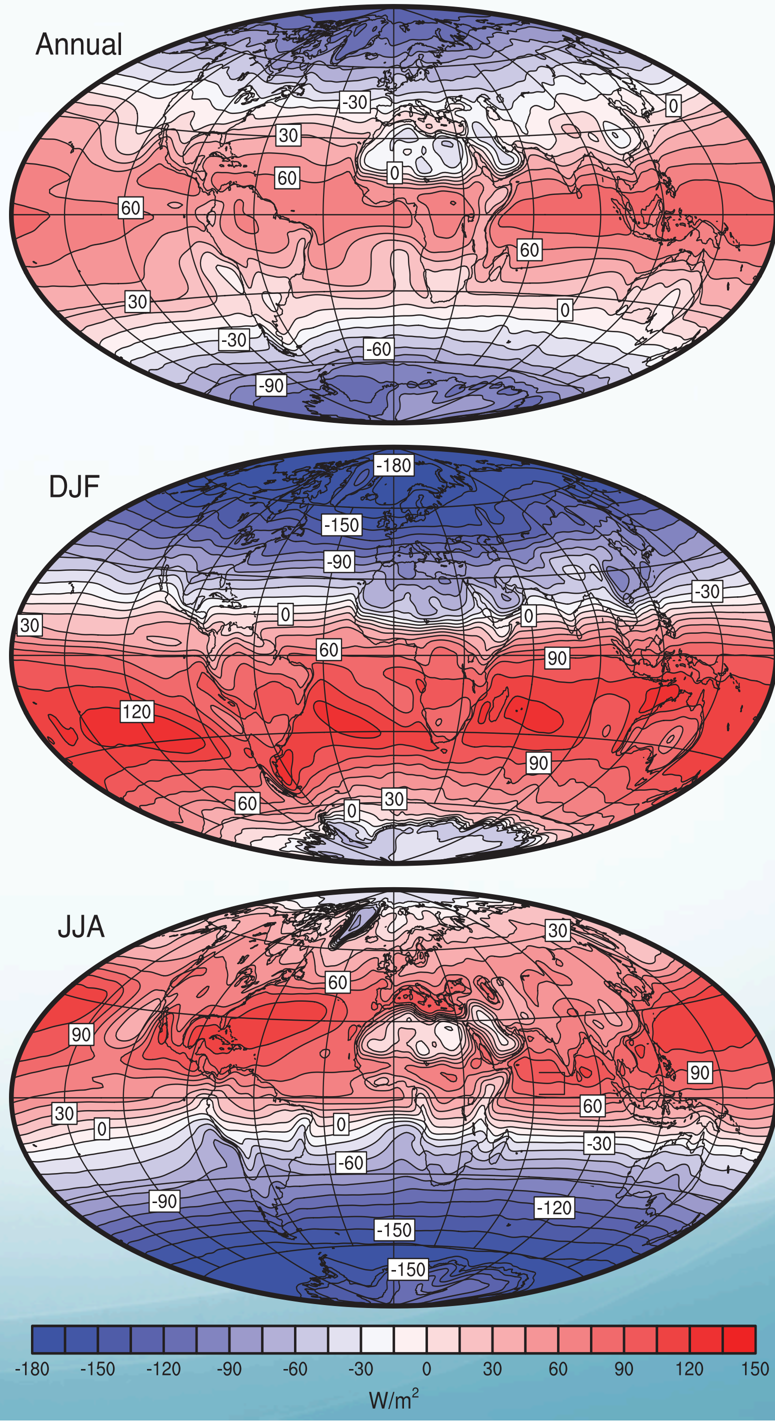


Net Incoming Radiation Absorbed Solar minus OLR

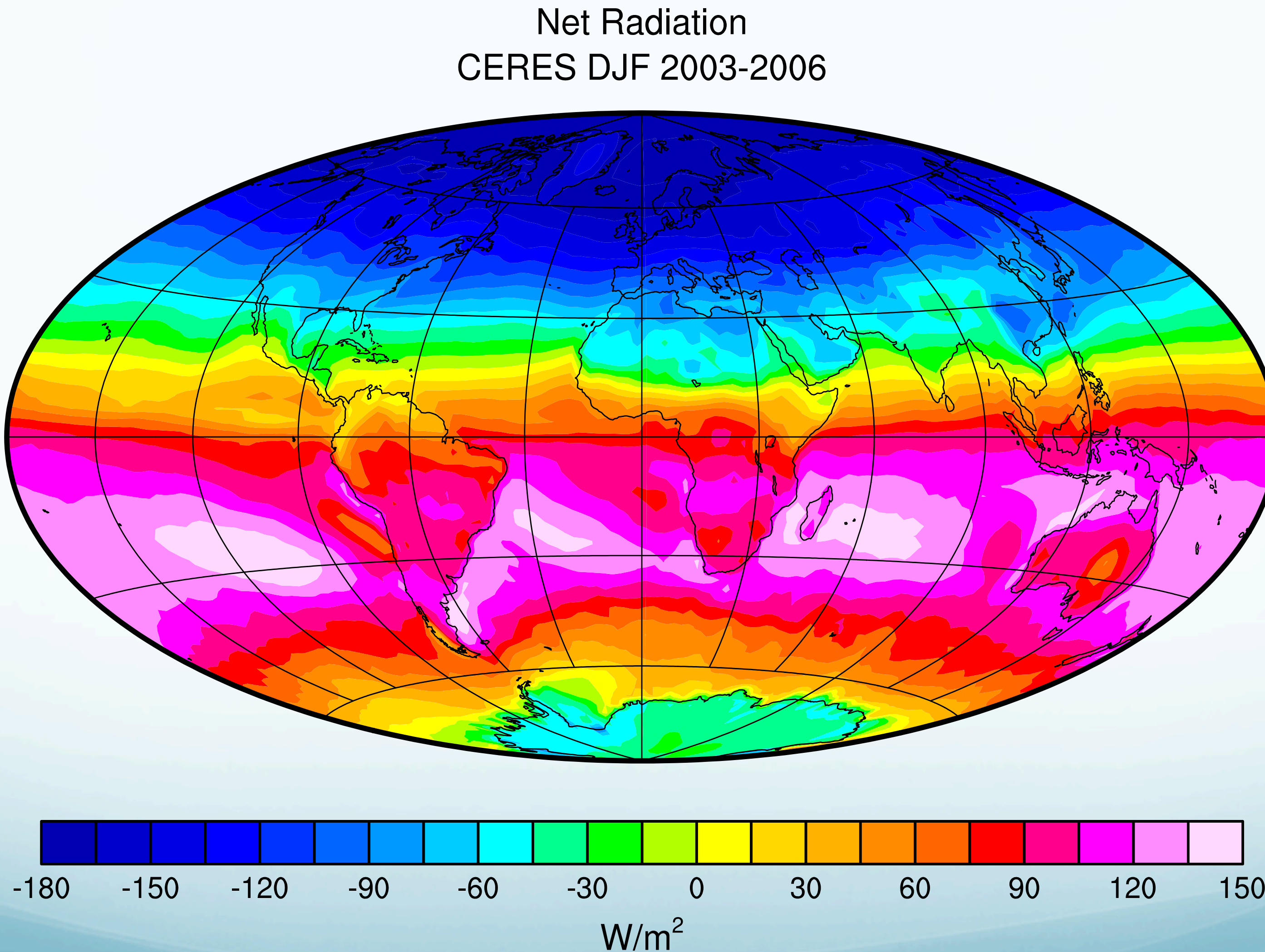
More near equator than
in high latitudes

Much more in summer hemisphere
than winter hemisphere

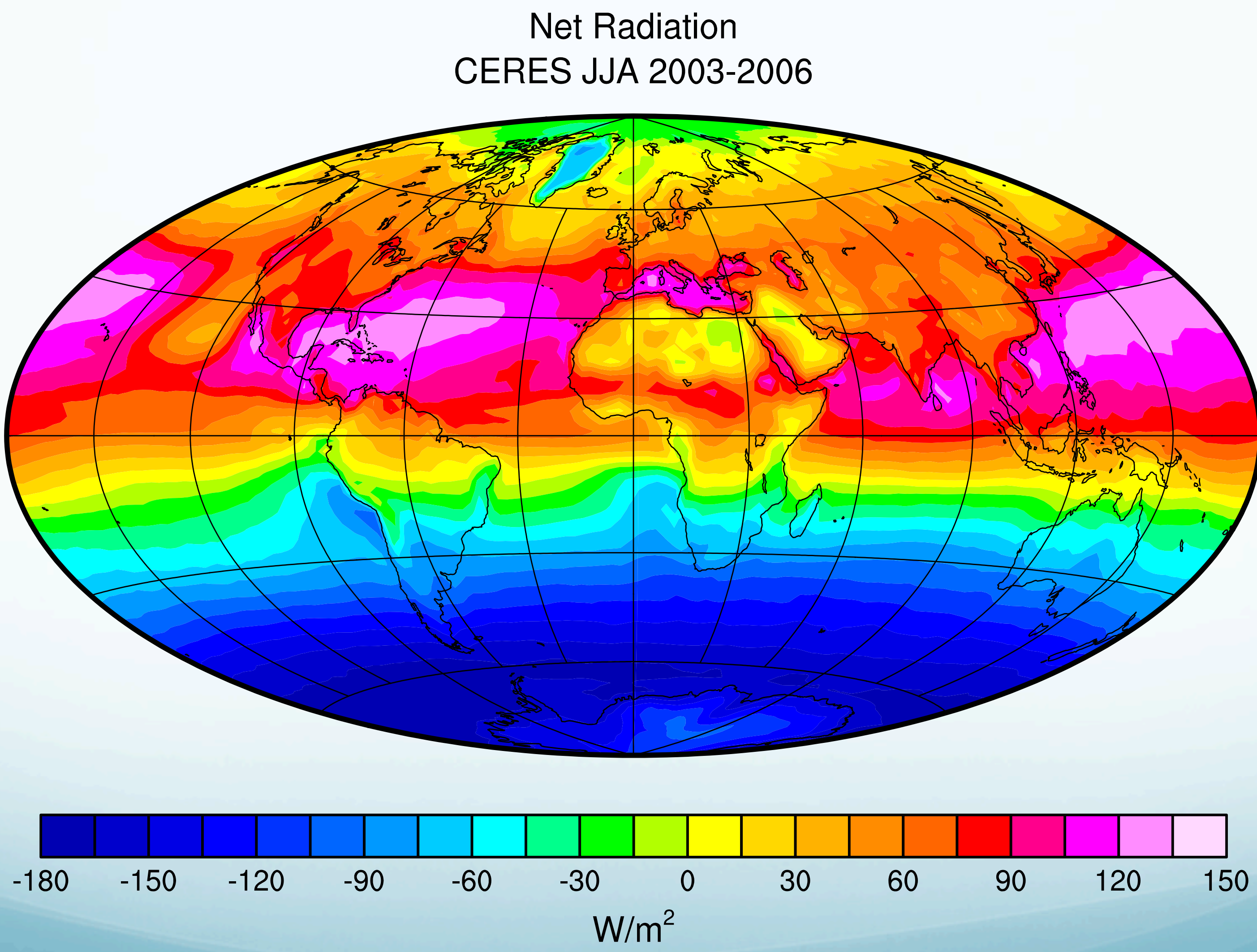
Note annual mean is negative
over the Sahara/Arabian Desert.
How does this support dryness there?



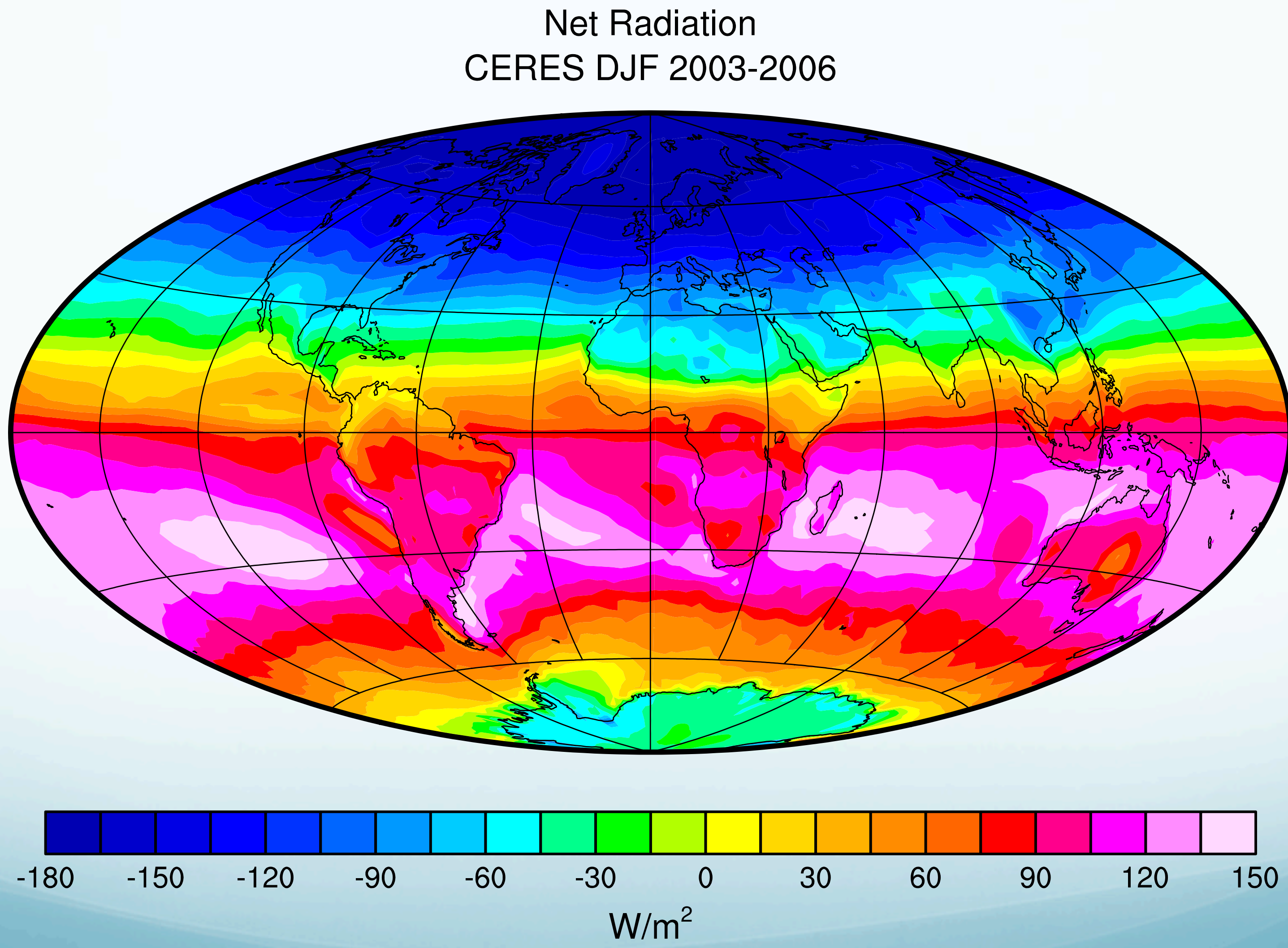
Net Radiation Annual



Net Radiation – JJA

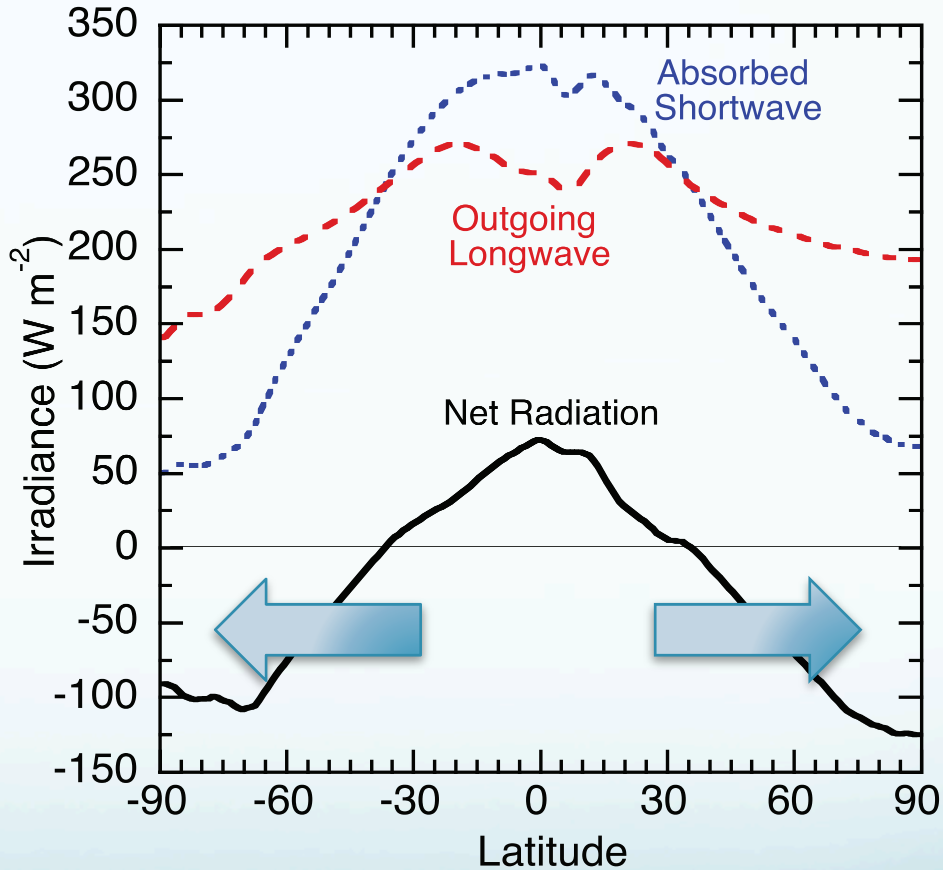


Net Radiation – DJF



Zonal Average Top-of-Atmosphere (TOA) – Annual mean

Need to move Energy Poleward – in both hemispheres



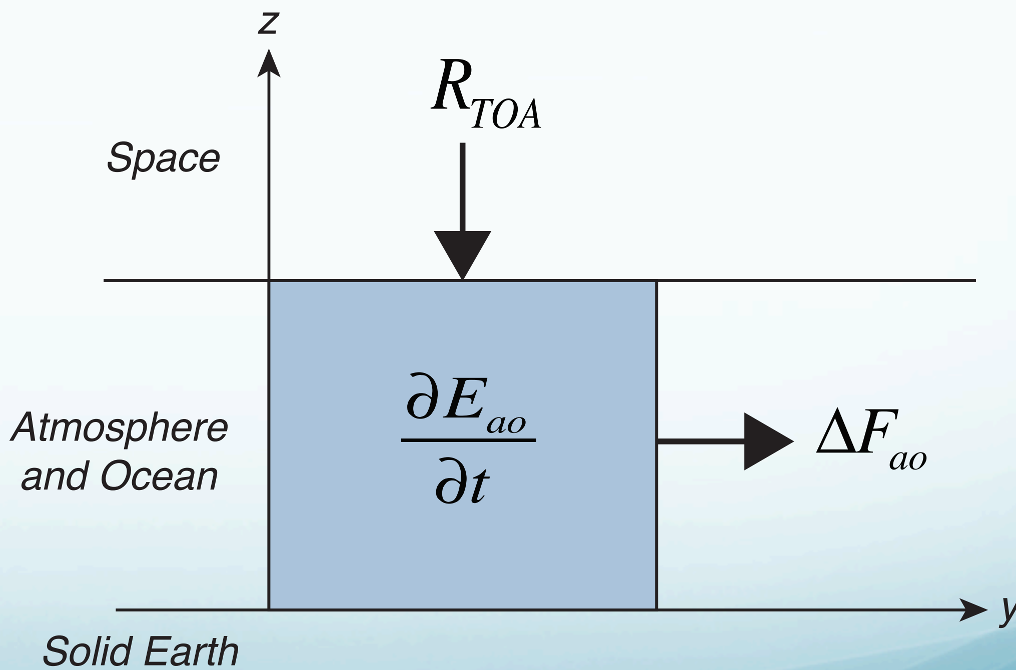
- Net Radiation has a latitude gradient - duh

Computing Poleward Flux

- We can compute the poleward energy flux in the atmosphere and ocean necessary to balance the TOA exchanges: First Law

$$\frac{\partial E_{ao}}{\partial t} = R_{TOA} - \Delta F_{ao}$$

$$R_{TOA} = \Delta F_{ao}$$



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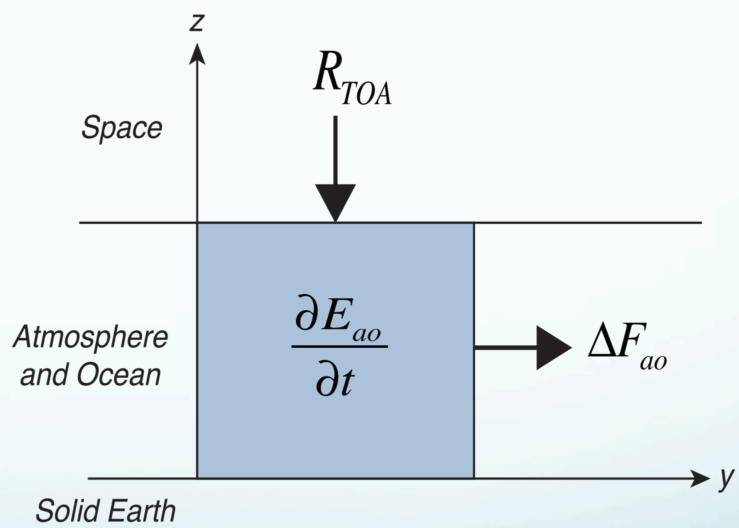


$$F(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_{0}^{2\pi} R_{TOA} a^2 \cos\phi \, d\lambda \, d\phi$$

Poleward Energy Flux

- Compute from TOA Balance by integrating over area of polar cap

$$F(\phi) = \int_{-\frac{\pi}{2}}^{\phi} \int_{0}^{2\pi} R_{TOA} a^2 \cos\phi \, d\lambda \, d\phi$$

- Use Atmospheric Observations to compute Atmospheric Flux, then Ocean flux is residual

$$F_{Total}(\phi) = F_{Atmosphere}(\phi) + F_{Ocean}(\phi)$$

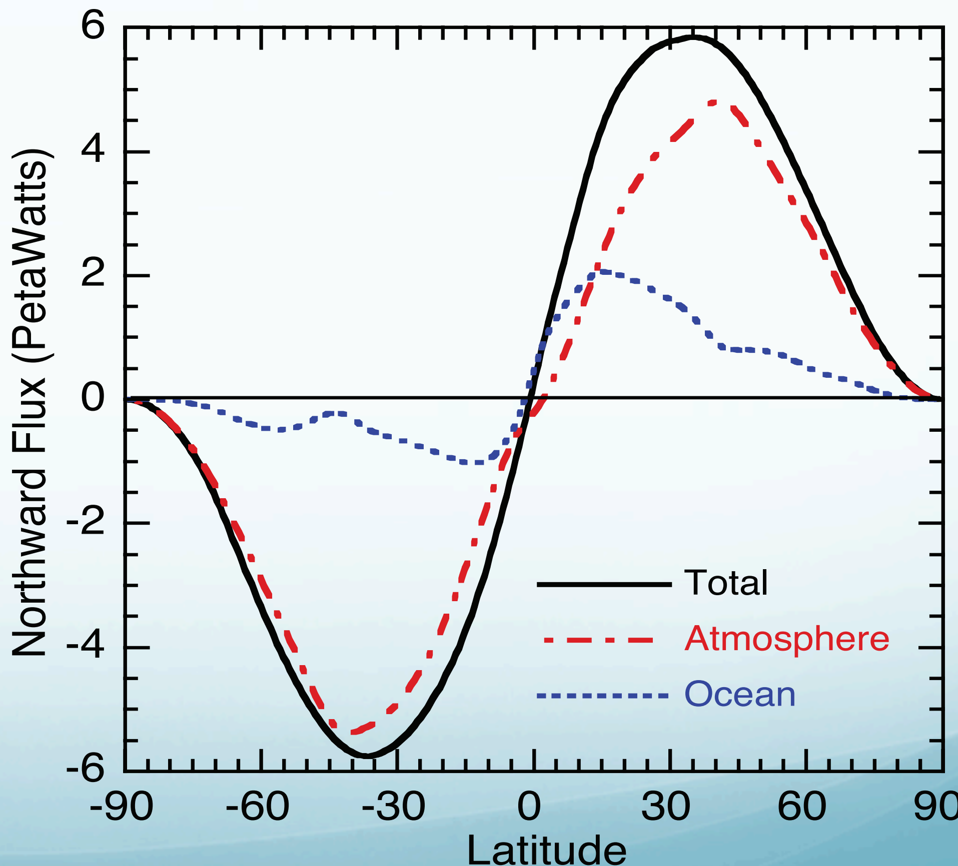
Poleward Heat Flux

- Contributions of atmosphere and ocean are both important, but are different functions of latitude.

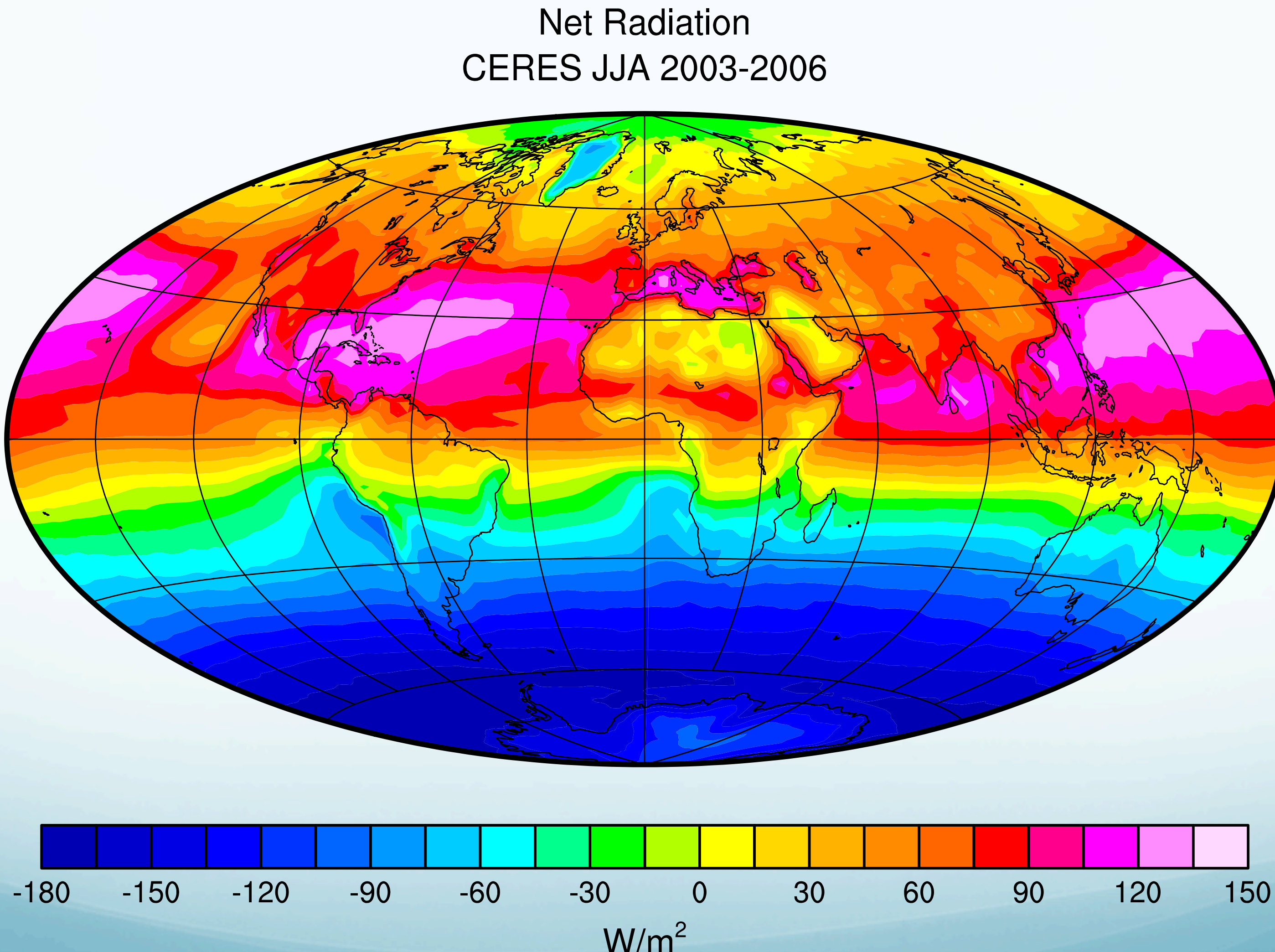
- Total from TOA net radiation satellite data.

- Atmosphere from atmosphere measurements

- Ocean flux is residual in energy budget



Thanks!


Net Radiation
CERES JJA 2003-2006
-180
-150
-120
-90
-60
-30
0
30
60
90
120
150
W/m2