Quick Review of Remote Sensing Basic Theory

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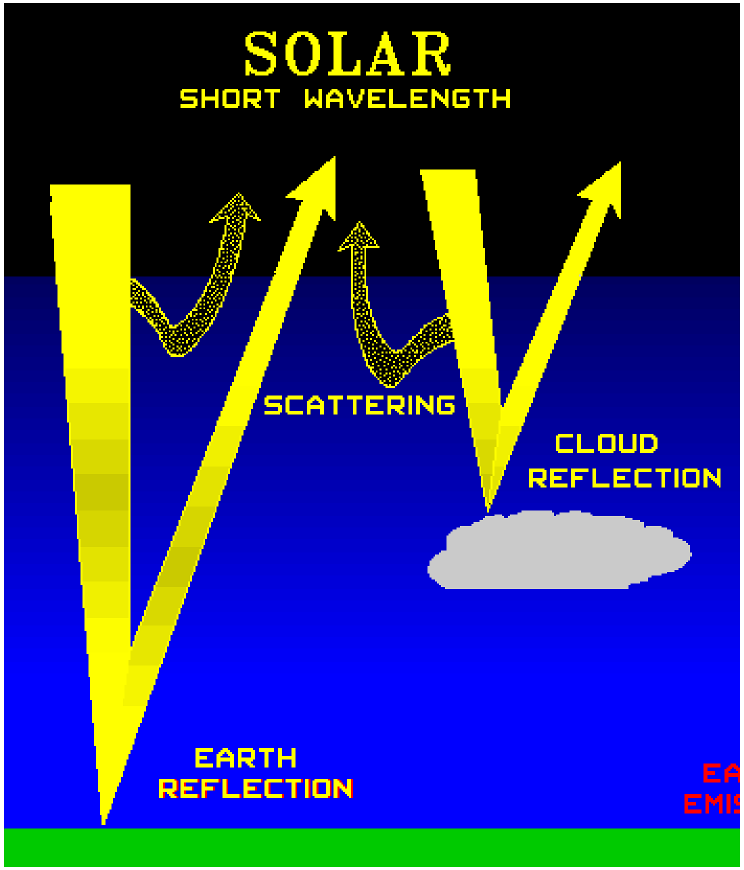
Outline

- Bit Depth
- Radiative Transfer Equation in the IR

Visible
(Reflective Bands)



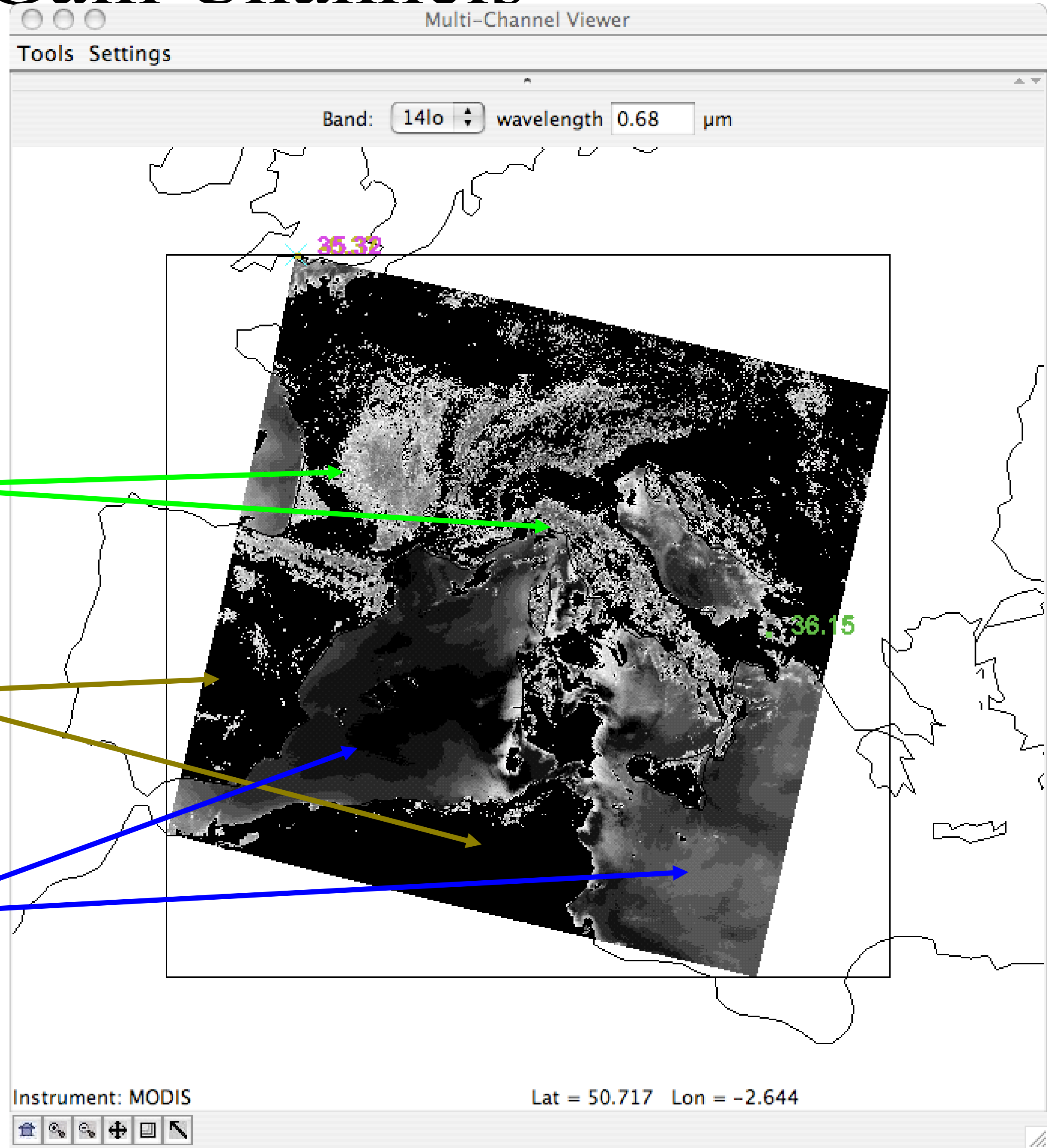
Low Gain Channels

Band 14 low
0.68 µm

Vegetated areas
Are visible

Saturation over
Barren Soil

Visible details
over water



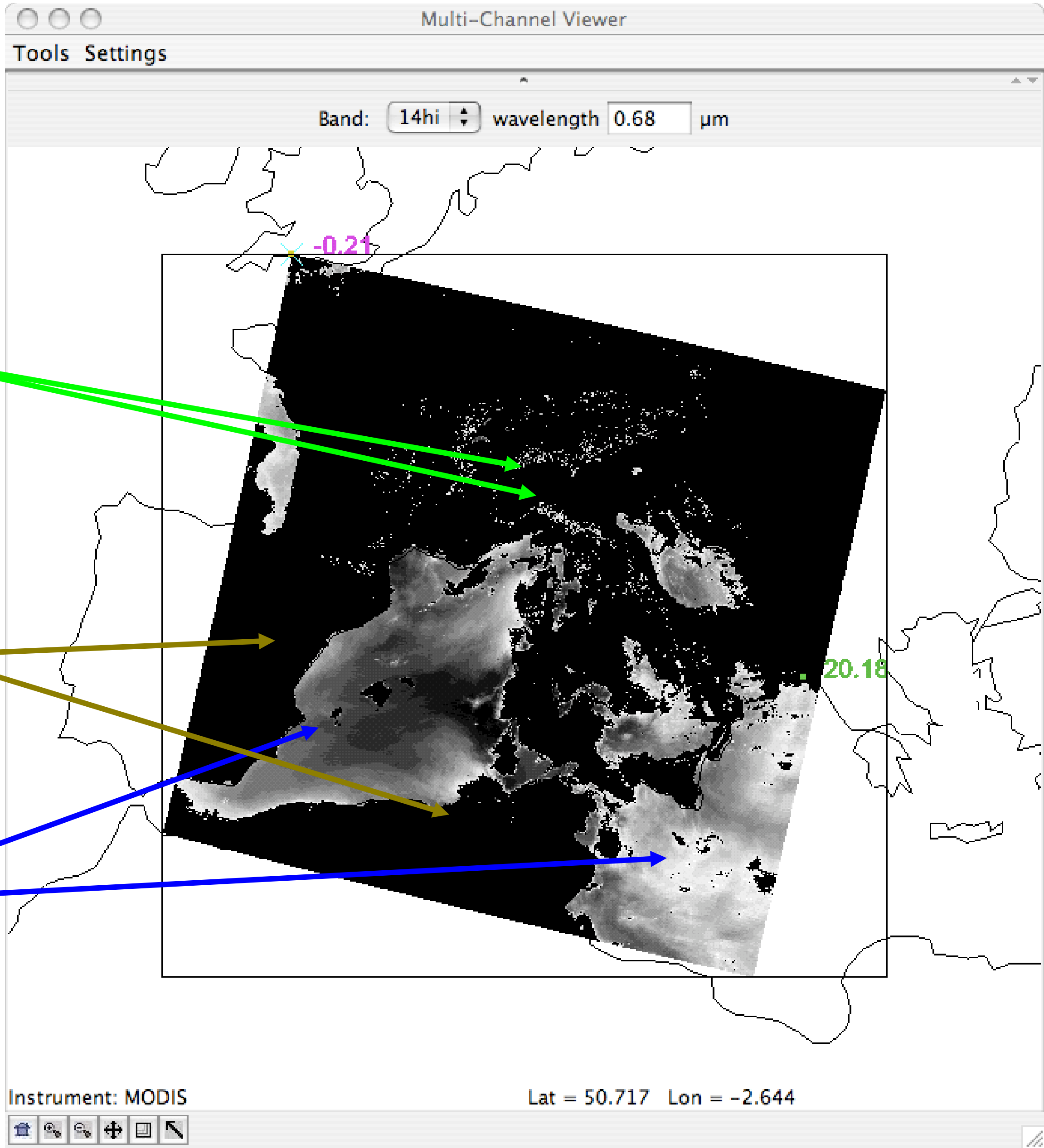
High Gain Channels

Band 14 hi
0.68 µm

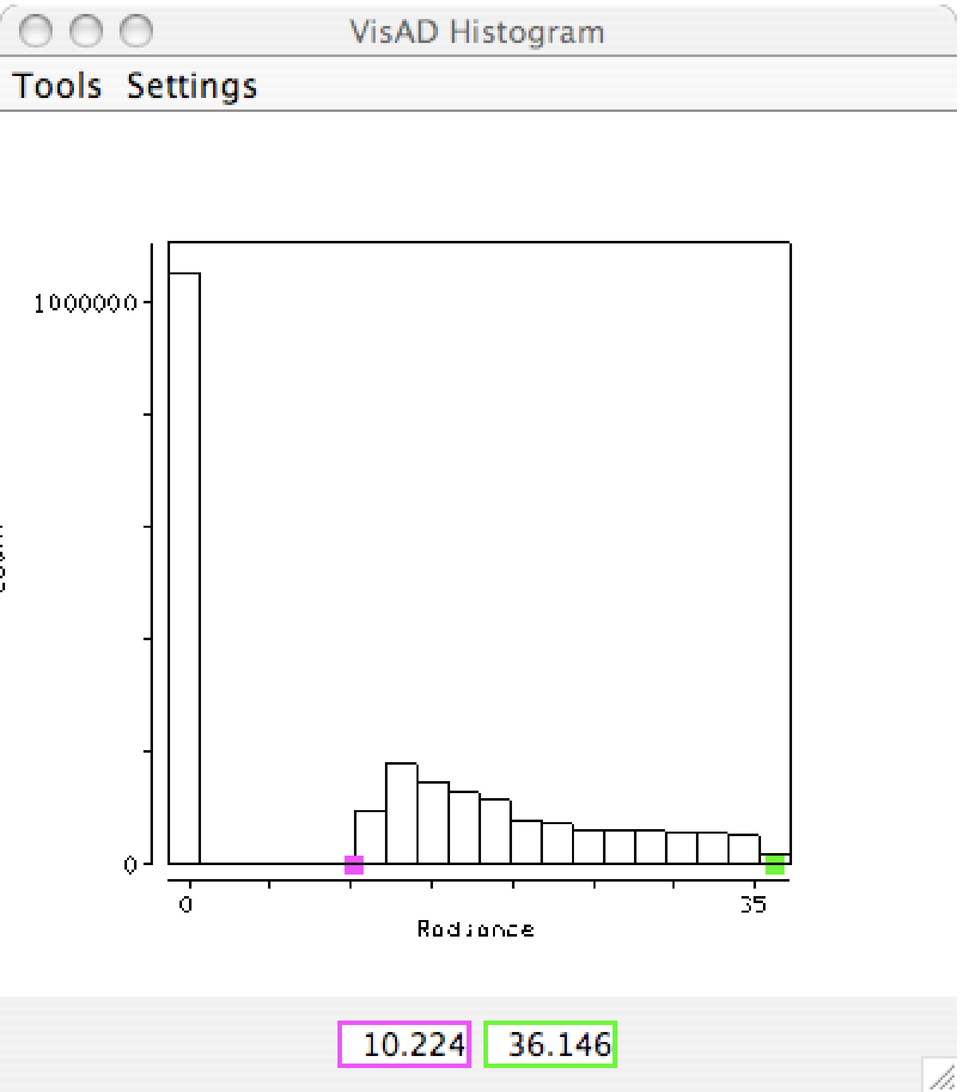
Saturation over
Vegetated areas
little barely visible

Saturation over
Barren Soil

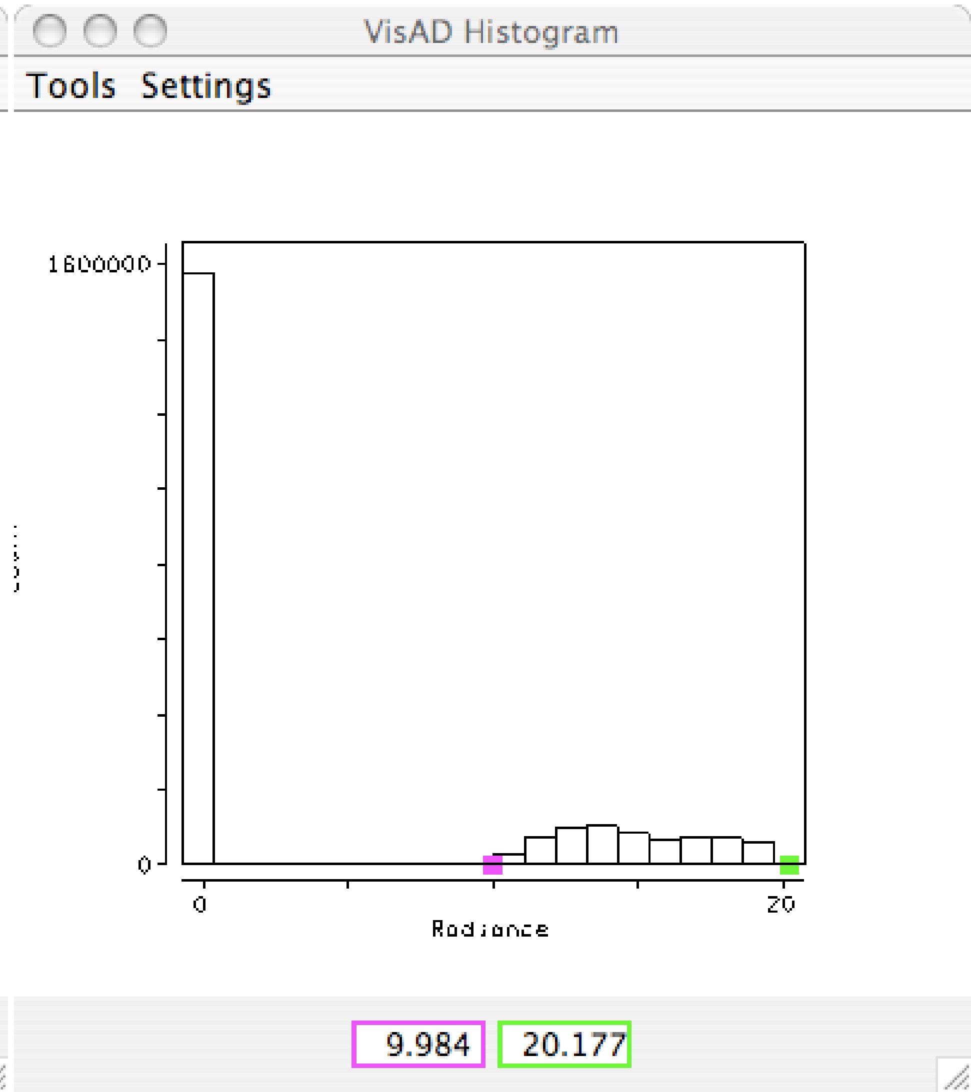
Visible details
over water



Range for Band 14 low 0.68 µm



Range for Band 14 high 0.68 µm



Bit Depth and Value Range

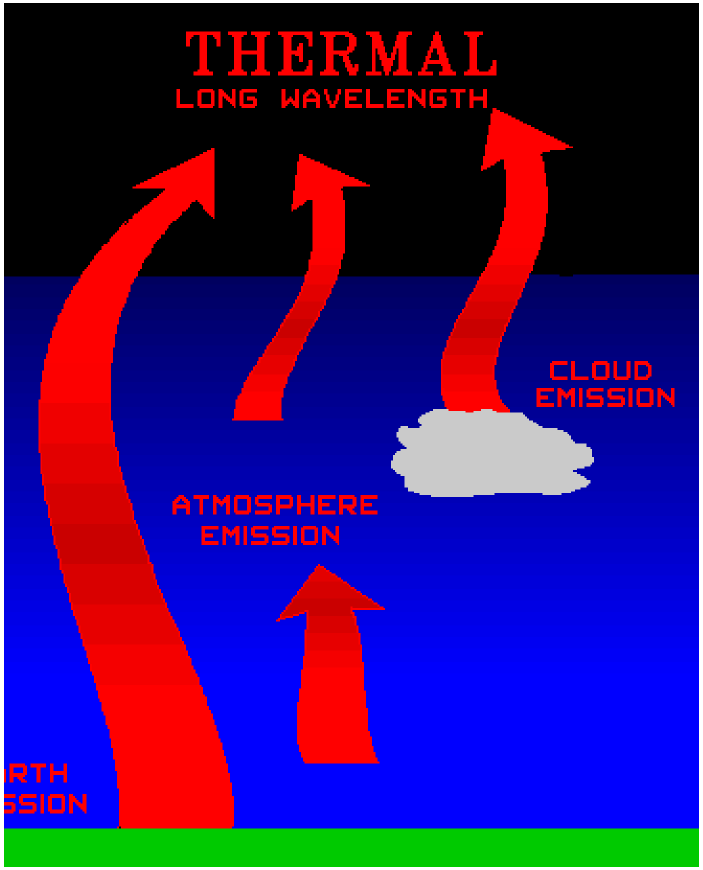
- With 12 bits 2^{12} integer numbers can be represented
  - Given ΔR, the range of radiances we want to observe, the smallest observable variation is $\Delta R / 2^{12}$
  - Given dR smallest observable variation, the range of observable radiances is $dR * 2^{12}$

For this reason Band 14low (larger range) is used for cloud detection and Band 14hi (smaller range) is used for ocean products

Infrared
(Emissive Bands)

Radiative Transfer Equation

in the IR



Relevant Material in Applications of Meteorological Satellites

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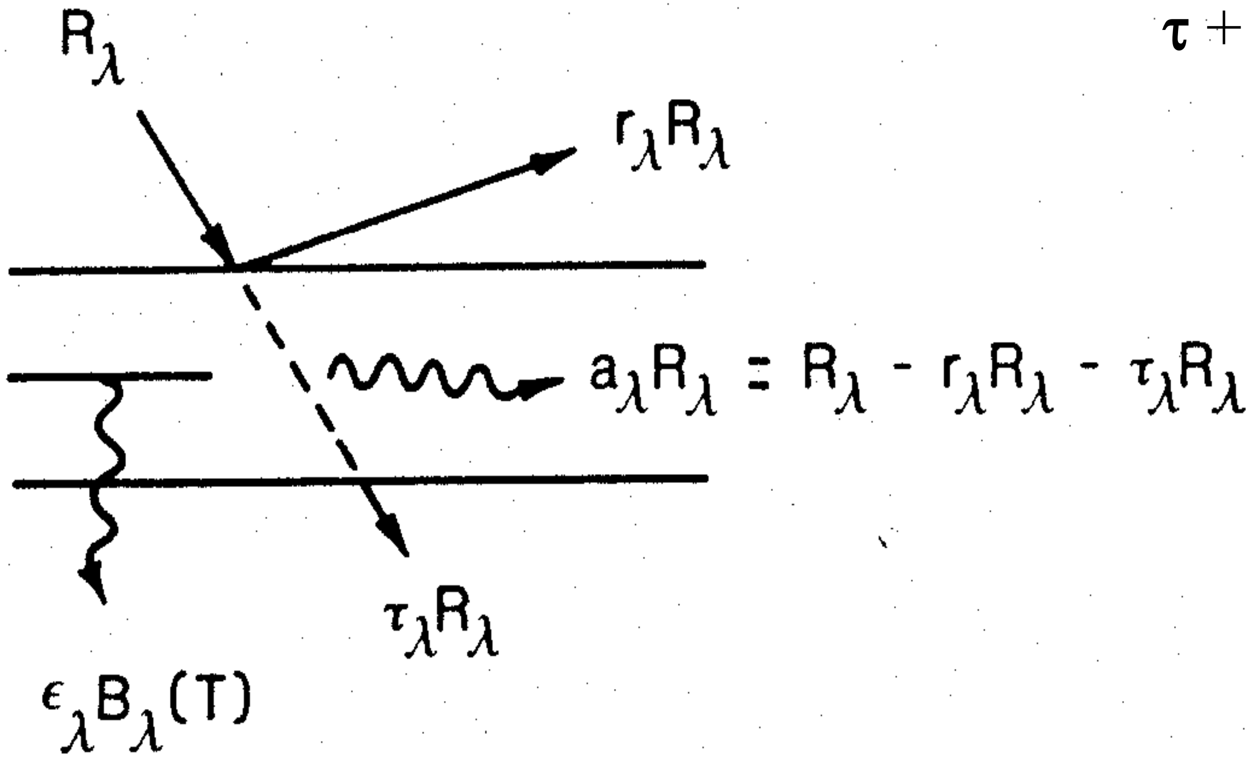
Energy conservation

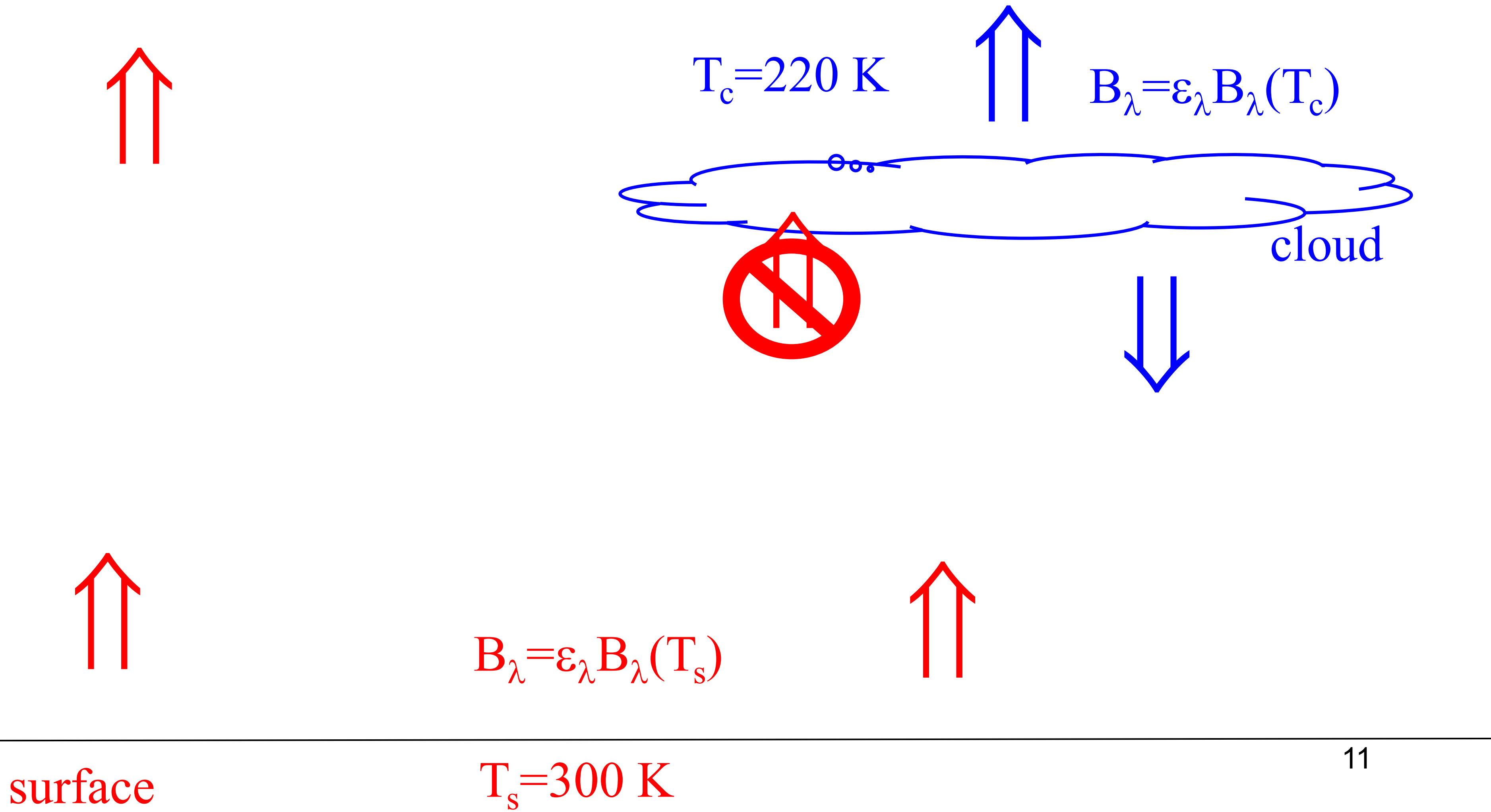
$\boldsymbol{\tau} + a + \cancel{r} = 1$



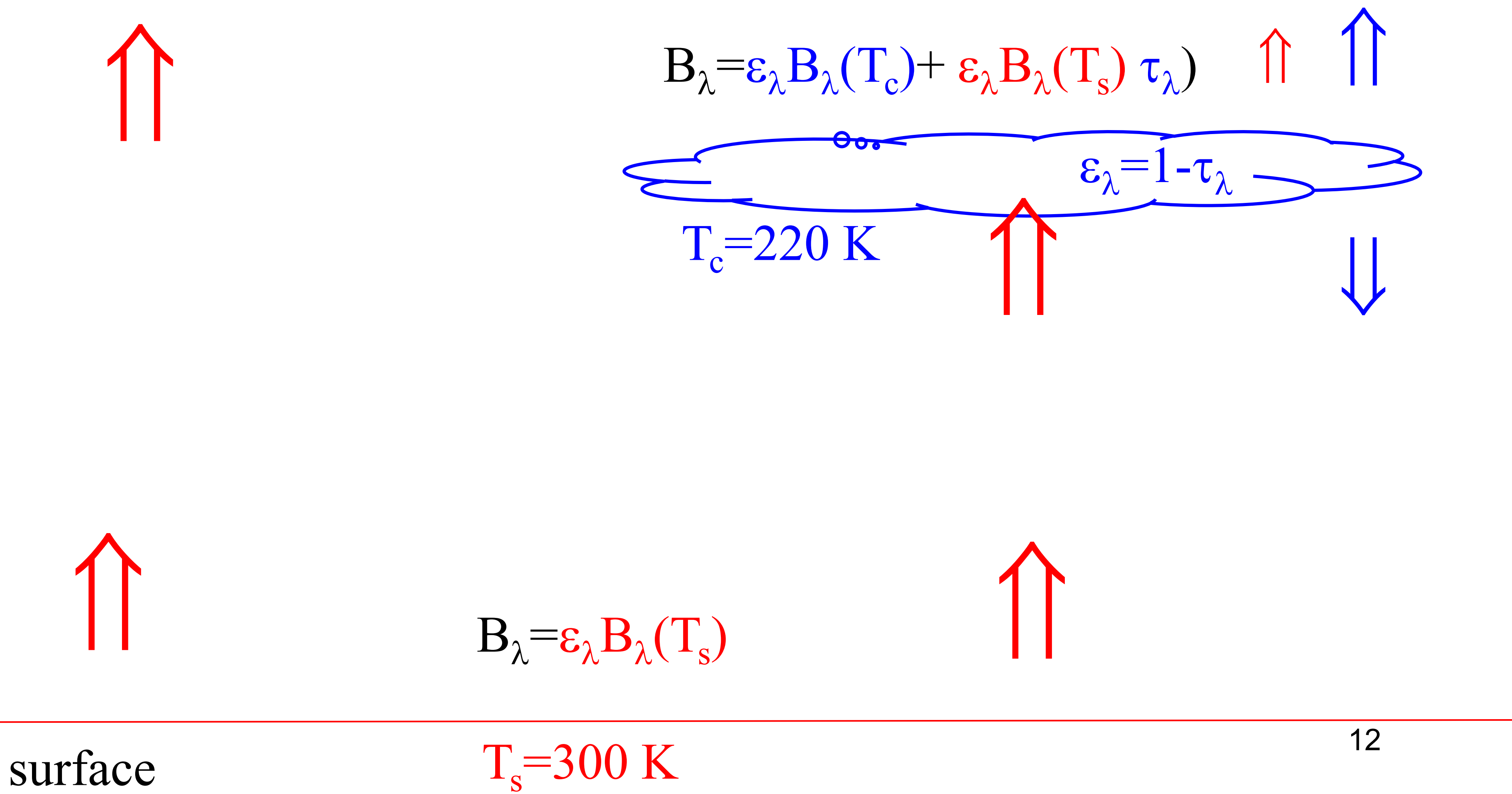
'ENERGY CONSERVATION'

$\boldsymbol{\tau} + a + r = 1$

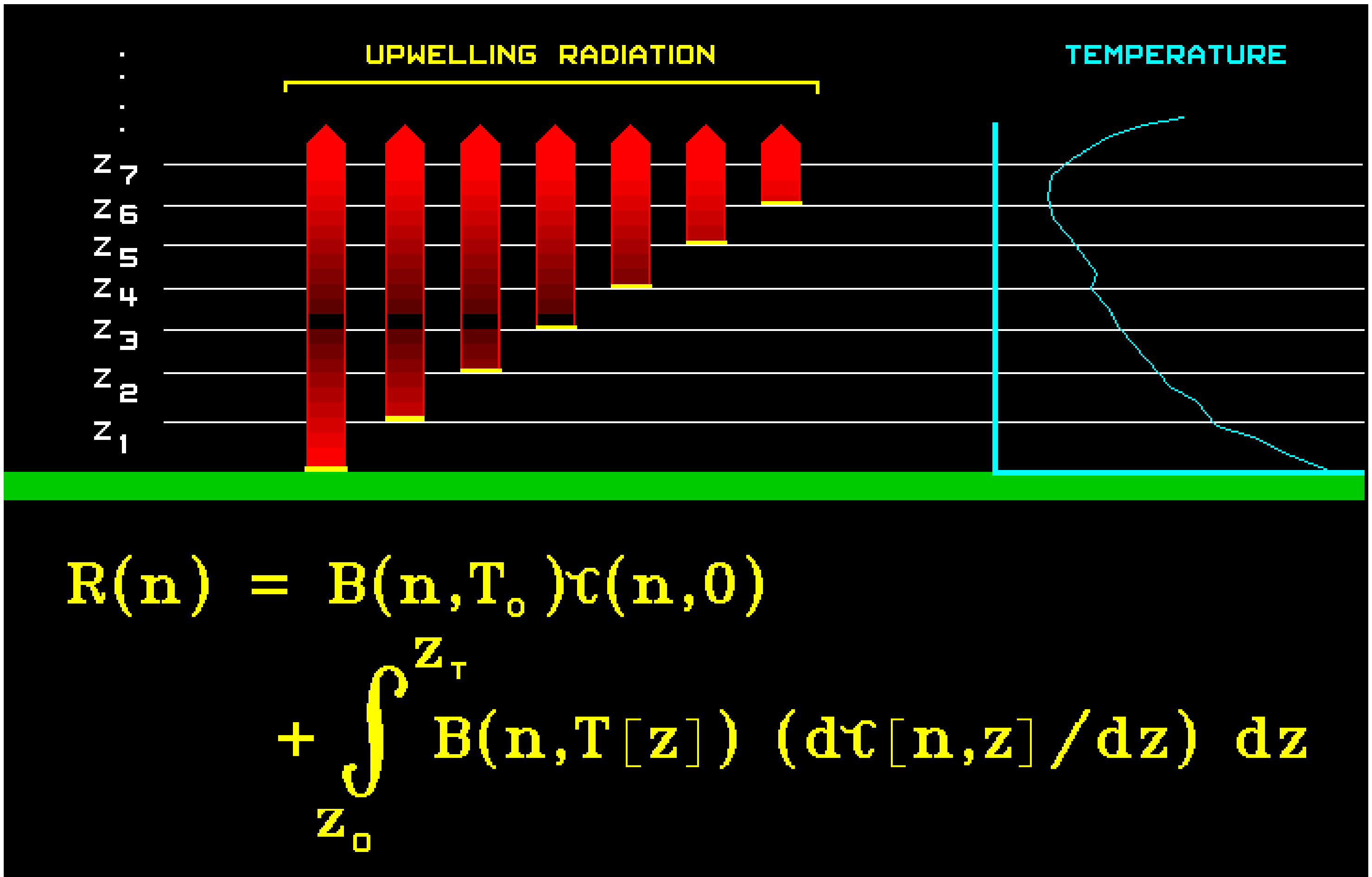
Simple case with no atmosphere and opaque cloud



Simple case with no atmosphere and semi-trasparent cloud

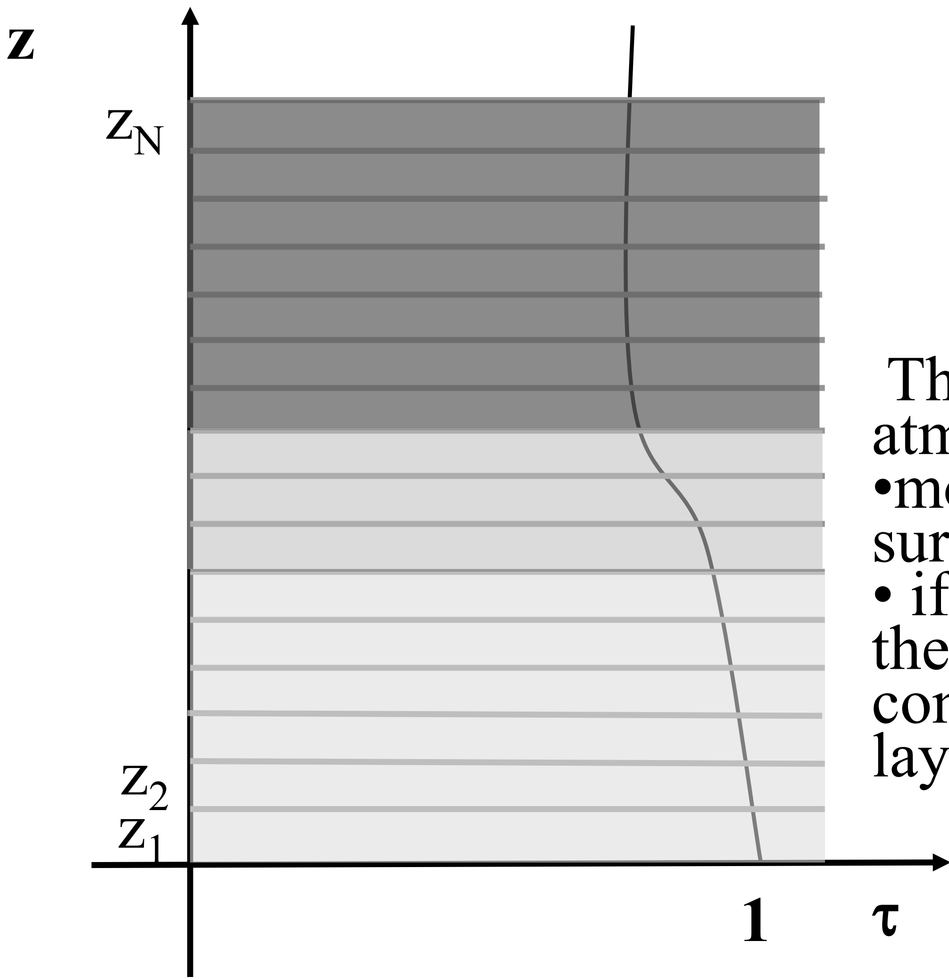


Radiative Transfer Equation



Transmittance for Window Channels

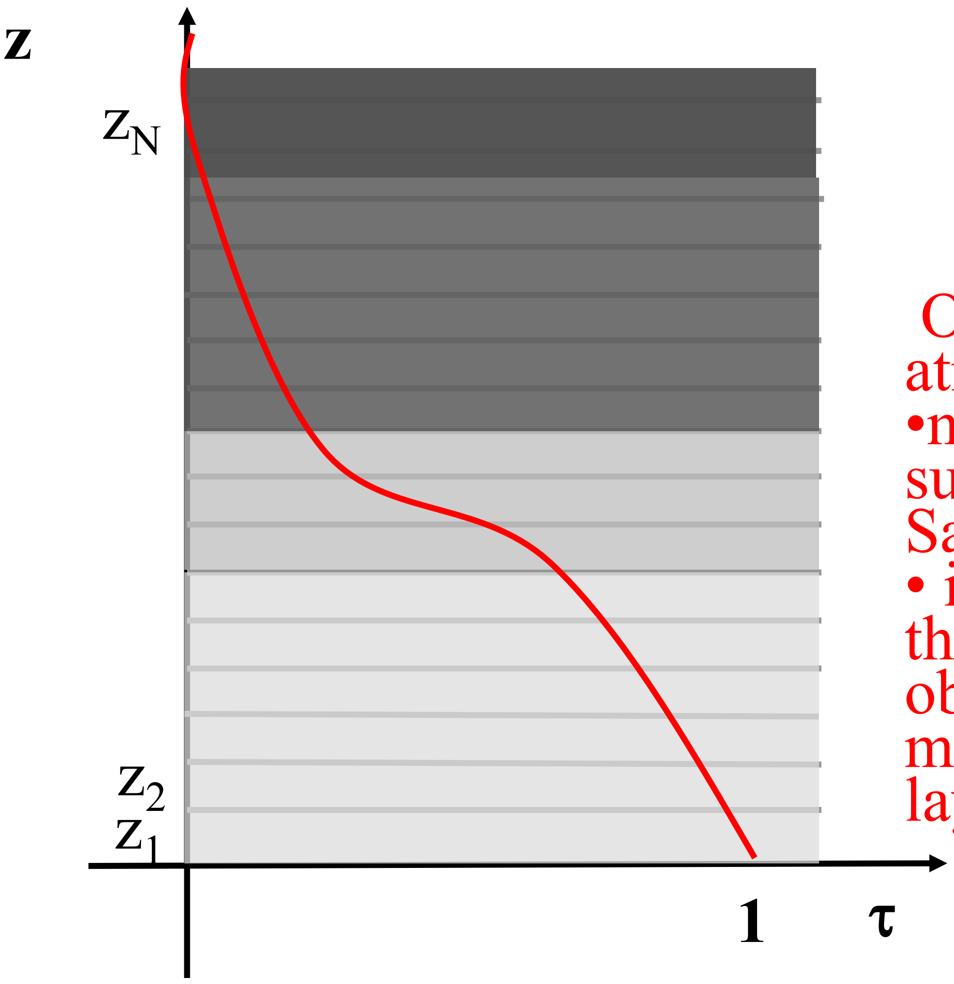
$$\boldsymbol{\tau} + a + \cancel{r} = 1$$



τ close to 1
a close to 0

The molecular species in the atmosphere are not very active:
- most of the photons emitted by the surface make it to the Satellite
- if a is close to 0 in the atmosphere then ε is close to 0, not much contribution from the atmospheric layers

Trasmittance for Absorption Channels

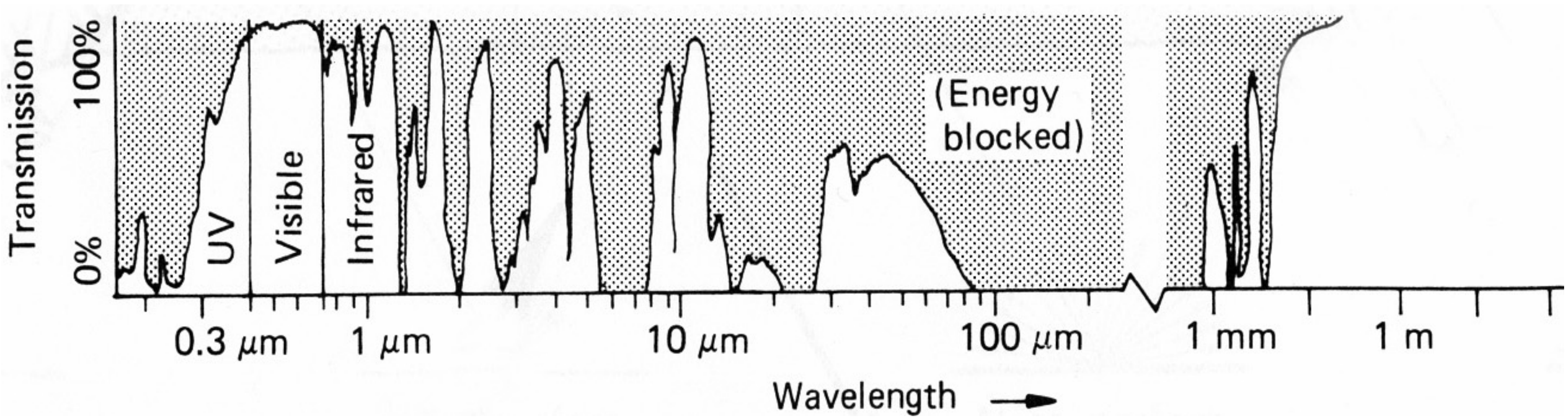


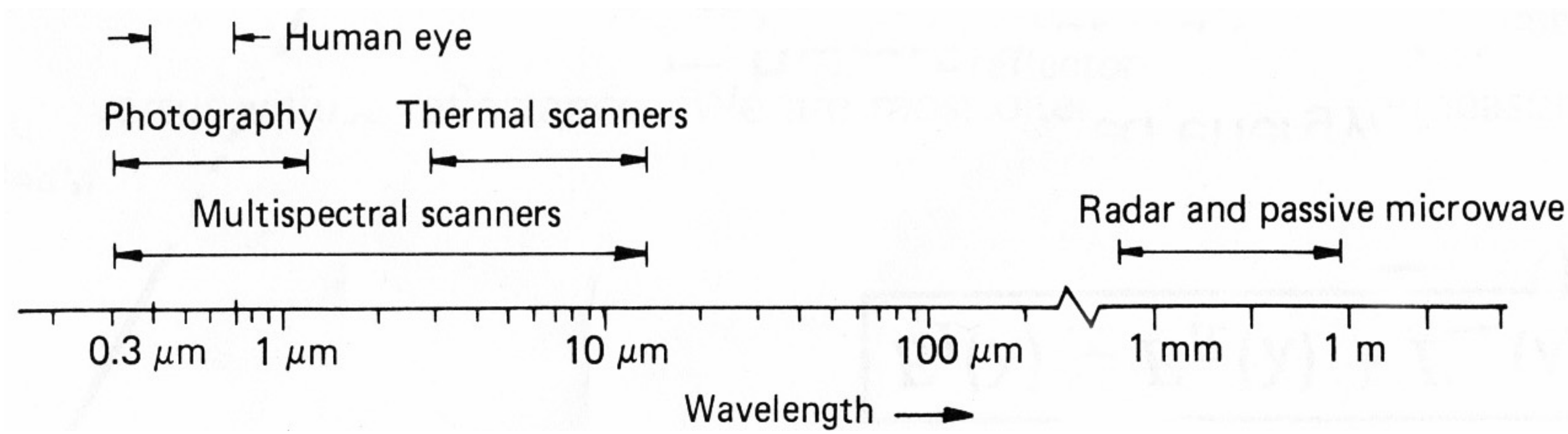
Absorption Channel:
τ close to 0
a close to 1

One or more molecular species in the atmosphere is/are very active:

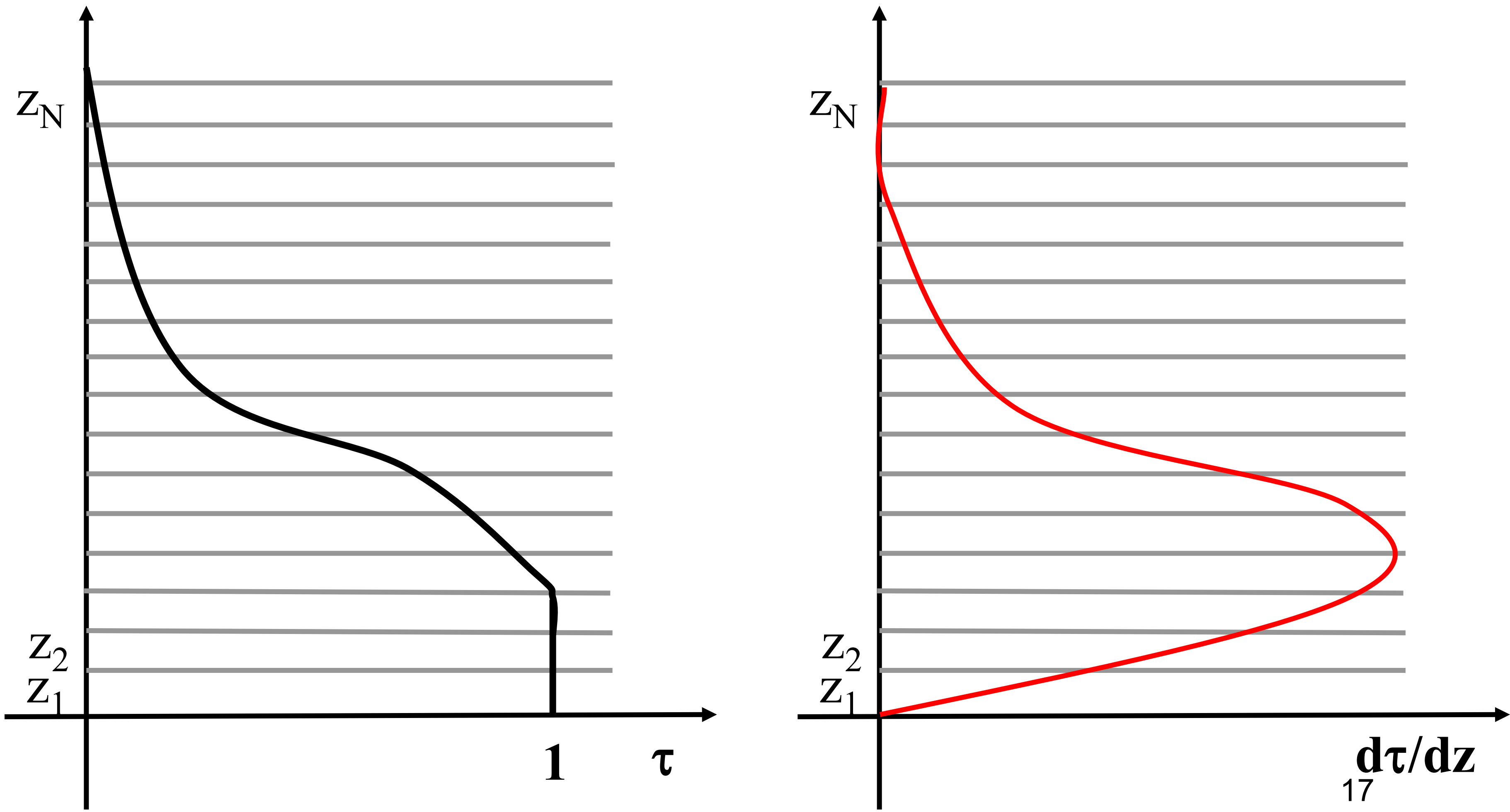
- most of the photons emitted by the surface will not make it to the Satellite (they will be absorbed)
- if a is close to 1 in the atmosphere then ε is close to 1, most of the observed energy comes from one or more of the uppermost atmospheric layers

Spectral Characteristics of Atmospheric Transmission and Sensing Systems

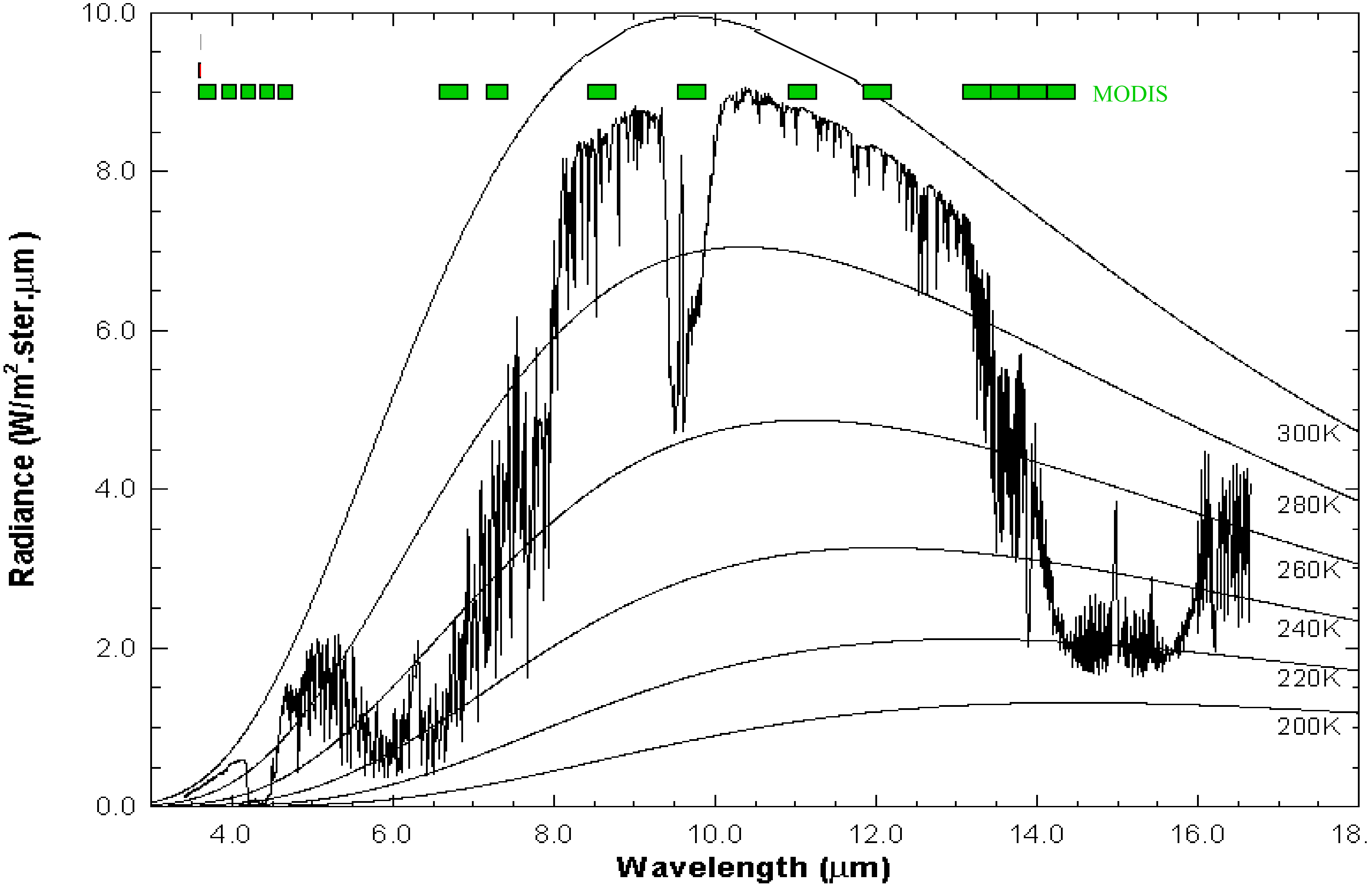




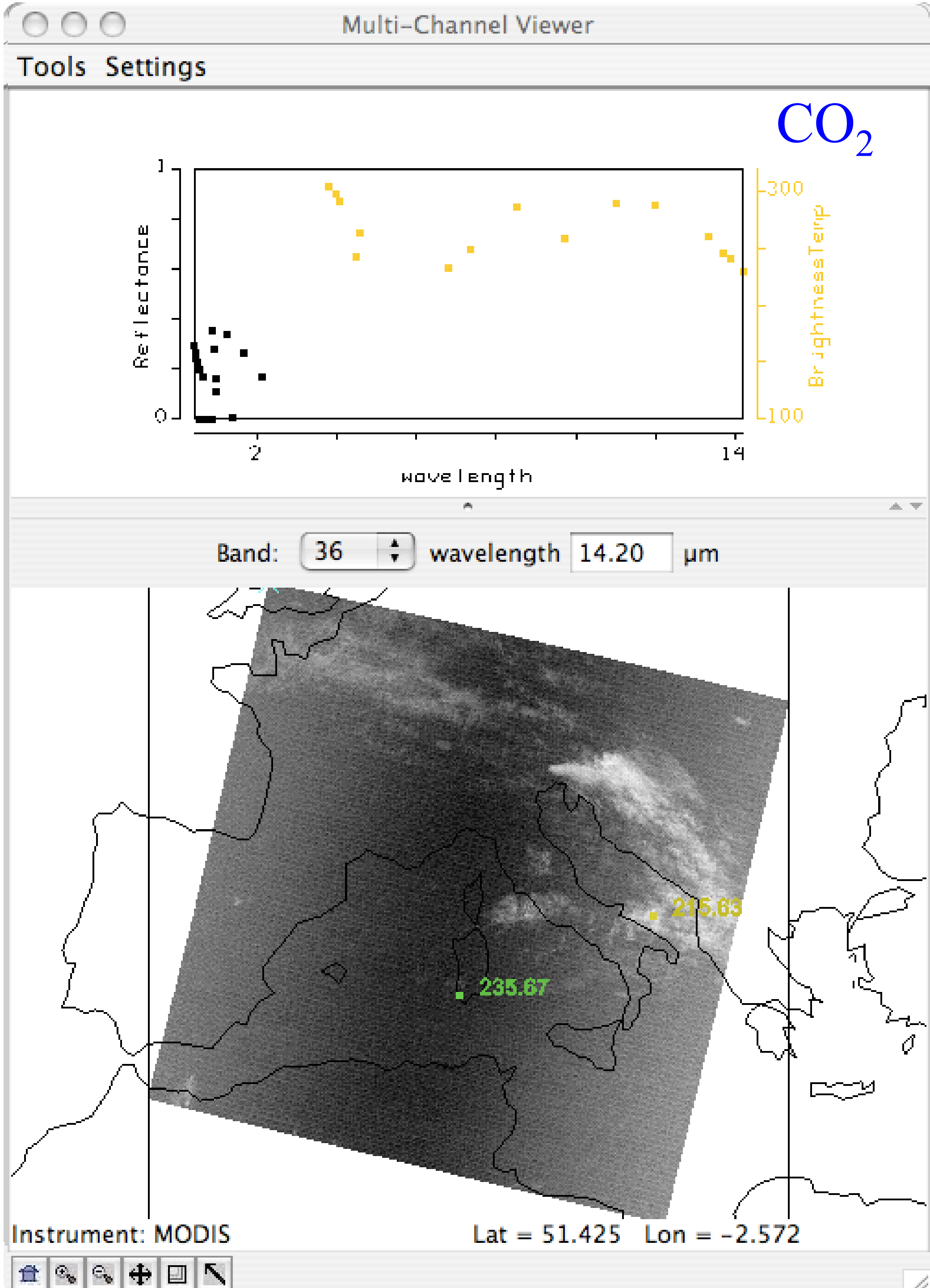
Weighting Functions



High resolution atmospheric absorption spectrum and comparative blackbody curves.



MODIS absorption bands



Emission, Absorption

Blackbody radiation B_λ represents the upper limit to the amount of radiation that a real substance may emit at a given temperature for a given wavelength.

Emissivity ε_λ is defined as the fraction of emitted radiation R_λ to Blackbody radiation,

$$\varepsilon_\lambda = R_\lambda / B_\lambda \, .$$

In a medium at thermal equilibrium, what is absorbed is emitted (what goes in comes out) so

$$a_\lambda = \varepsilon_\lambda \, .$$

Thus, materials which are strong absorbers at a given wavelength are also strong emitters at that wavelength; similarly weak absorbers are weak emitters.

Transmittance

Transmission through an absorbing medium for a given wavelength is governed by the number of intervening absorbing molecules (path length u) and their absorbing power (k_λ) at that wavelength. Beer's law indicates that transmittance decays exponentially with increasing path length

$$\tau_\lambda (z \to \infty) = e^{- k_\lambda u(z)}$$

where the path length is given by $u(z) = \int_z^\infty \rho \, dz$.

$k_\lambda u$ is a measure of the cumulative depletion that the beam of radiation has experienced as a result of its passage through the layer and is often called the optical depth σ_λ.

Realizing that the hydrostatic equation implies $g \rho \, dz = - q \, dp$

where q is the mixing ratio and ρ is the density of the atmosphere, then

$$u(p) = \int_o^p q \, g^{-1} \, dp \qquad \text{and} \qquad \tau_\lambda (p \to o) = e^{- k_\lambda u(p)} \; .$$

Emission, Absorption, Reflection, and Scattering

If a_λ, r_λ, and τ_λ represent the fractional absorption, reflectance, and transmittance, respectively, then conservation of energy says

$$a_\lambda + r_\lambda + \tau_\lambda = 1 \quad .$$

For a blackbody $a_\lambda = 1$, it follows that $r_\lambda = 0$ and $\tau_\lambda = 0$ for blackbody radiation. Also, for a perfect window $\tau_\lambda = 1$, $a_\lambda = 0$ and $r_\lambda = 0$. For any opaque surface $\tau_\lambda = 0$, so radiation is either absorbed or reflected $a_\lambda + r_\lambda = 1$.

At any wavelength, strong reflectors are weak absorbers (i.e., snow at visible wavelengths), and weak reflectors are strong absorbers (i.e., asphalt at visible wavelengths).

Radiative Transfer Equation

The radiance leaving the earth-atmosphere system sensed by a satellite borne radiometer is the sum of radiation emissions from the earth-surface and each atmospheric level that are transmitted to the top of the atmosphere. Considering the earth's surface to be a blackbody emitter (emissivity equal to unity), the upwelling radiance intensity, I_λ, for a cloudless atmosphere is given by the expression

$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T_{sfc}) \tau_\lambda(sfc - top) + \sum_{layers} \varepsilon_\lambda^{layer} B_\lambda(T_{layer}) \tau_\lambda(layer - top)$$

where the first term is the surface contribution and the second term is the atmospheric contribution to the radiance to space.

In standard notation,

$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T(p_s))\, \tau_\lambda(p_s) + \sum_p \varepsilon_\lambda(\Delta p)\, B_\lambda(T(p))\, \tau_\lambda(p)$$

The emissivity of an infinitesimal layer of the atmosphere at pressure p is equal to the absorptance (one minus the transmittance of the layer). Consequently,

$$\varepsilon_\lambda(\Delta p)\, \tau_\lambda(p) = [1 - \tau_\lambda(\Delta p)]\, \tau_\lambda(p)$$

Since transmittance is an exponential function of depth of absorbing constituent,

$$\tau_\lambda(\Delta p)\, \tau_\lambda(p) = \exp\left[- \int_p^{p+\Delta p} k_\lambda\, q\, g^{-1}\, dp\right] * \exp\left[- \int_o^{p} k_\lambda\, q\, g^{-1}\, dp\right] = \tau_\lambda(p + \Delta p)$$

Therefore

$$\varepsilon_\lambda(\Delta p)\, \tau_\lambda(p) = \tau_\lambda(p) - \tau_\lambda(p + \Delta p) = - \Delta\tau_\lambda(p) .$$

So we can write

$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T(p_s))\, \tau_\lambda(p_s) - \sum_p B_\lambda(T(p))\, \Delta\tau_\lambda(p) .$$

which when written in integral form reads

$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T(p_s))\, \tau_\lambda(p_s) - \int_o^{p_s} B_\lambda(T(p))\, [\, d\tau_\lambda(p) / dp \,]\, dp .$$

When reflection from the earth surface is also considered, the Radiative Transfer Equation for infrared radiation can be written

$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T_s)\, \tau_\lambda(p_s) + \int_{p_s}^{o} B_\lambda(T(p))\, F_\lambda(p)\, [d\tau_\lambda(p)/ dp]\, dp$$

where

$$F_\lambda(p) = \{ 1 + (1 - \varepsilon_\lambda)\, [\tau_\lambda(p_s) / \tau_\lambda(p)]^2 \}$$

The first term is the spectral radiance emitted by the surface and attenuated by the atmosphere, often called the boundary term and the second term is the spectral radiance emitted to space by the atmosphere directly or by reflection from the earth surface.

The atmospheric contribution is the weighted sum of the Planck radiance contribution from each layer, where the weighting function is [$d\tau_\lambda(p) / dp$]. This weighting function is an indication of where in the atmosphere the majority of the radiation for a given spectral band comes from.

Mathematical Derivation of the Radiative Transfer Equation

Schwarzchild's equation

At wavelengths of terrestrial radiation, absorption and emission are equally important and must be considered simultaneously. Absorption of terrestrial radiation along an upward path through the atmosphere is described by the relation

$$-dL_\lambda^{abs} = L_\lambda k_\lambda \rho \sec \varphi \, dz .$$

Making use of Kirchhoff's law it is possible to write an analogous expression for the emission,

$$dL_\lambda^{em} = B_\lambda d\varepsilon_\lambda = B_\lambda da_\lambda = B_\lambda k_\lambda \rho \sec \varphi \, dz ,$$

where B_λ is the blackbody monochromatic radiance specified by Planck's law. Together

$$dL_\lambda = -(L_\lambda - B_\lambda) k_\lambda \rho \sec \varphi \, dz .$$

This expression, known as Schwarzchild's equation, is the basis for computations of the transfer of infrared radiation.

Schwarzschild to RTE

$$dL_\lambda = -(L_\lambda - B_\lambda)\, k_\lambda\, \rho\, dz$$

but

$$d\tau_\lambda = \tau_\lambda\, k\, \rho\, dz \quad \text{since} \quad \tau_\lambda = \exp\left[-k_\lambda \int_z^\infty \rho\, dz\right].$$

so

$$\tau_\lambda\, dL_\lambda = -(L_\lambda - B_\lambda)\, d\tau_\lambda$$

$$\tau_\lambda\, dL_\lambda + L_\lambda\, d\tau_\lambda = B_\lambda d\tau_\lambda$$

$$d\,(L_\lambda\, \tau_\lambda) = B_\lambda d\tau_\lambda$$

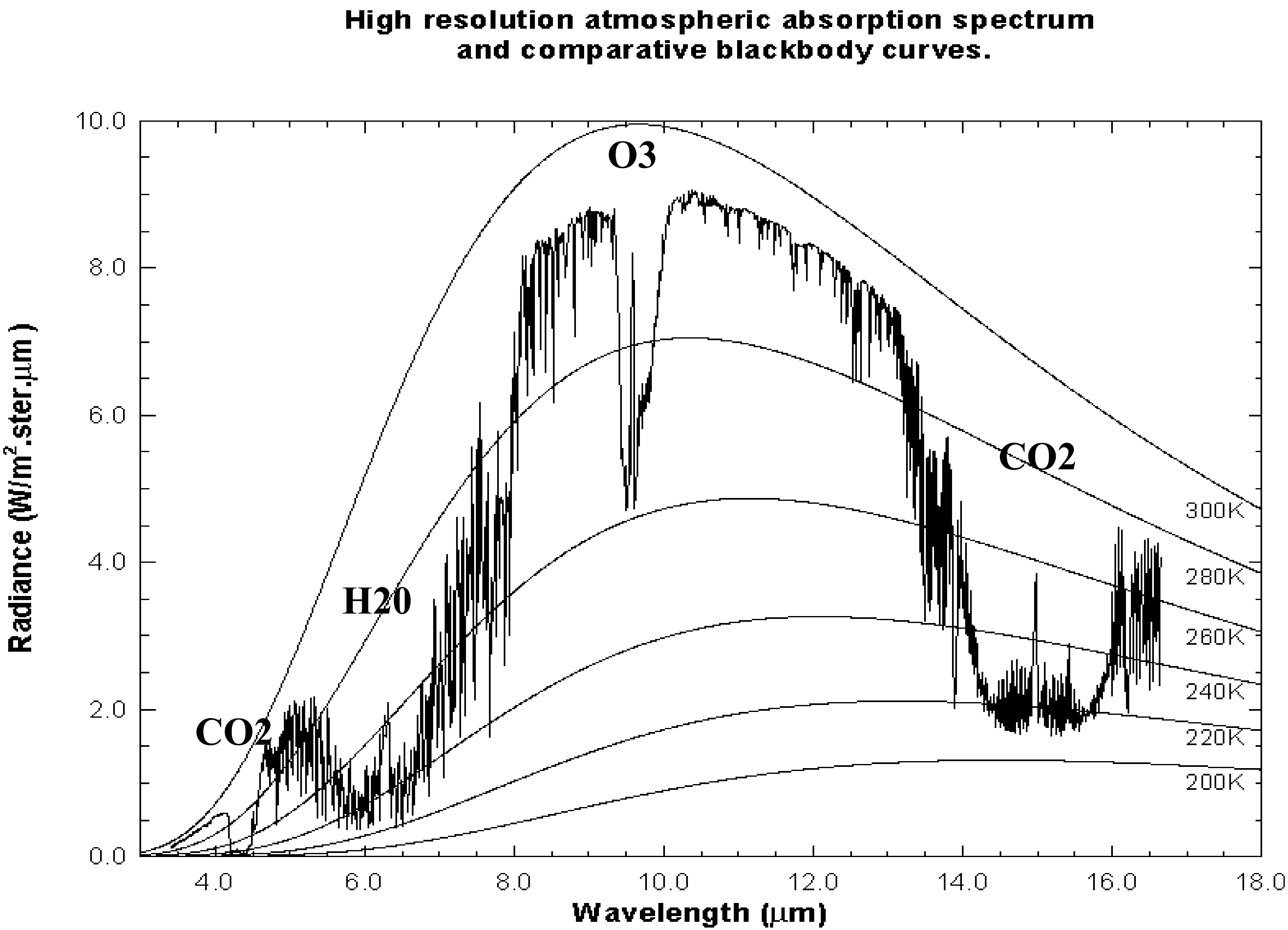
Integrate from 0 to ∞

$$L_\lambda(\infty)\, \tau_\lambda(\infty) - L_\lambda(0)\, \tau_\lambda(0) = \int_0^\infty B_\lambda\, [d\tau_\lambda / dz]\, dz.$$

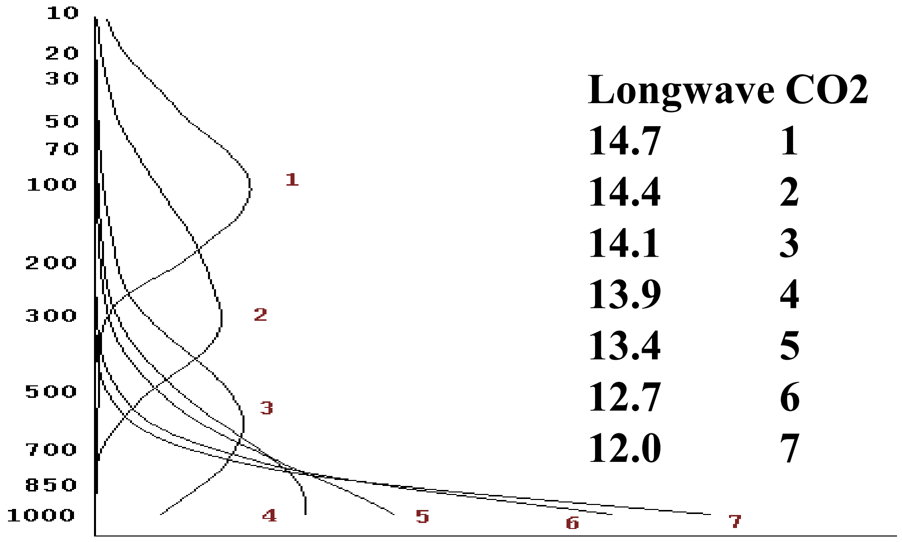
and

$$L_\lambda(\text{sat}) = L_\lambda(\text{sfc})\, \tau_\lambda(\text{sfc}) + \int_0^\infty B_\lambda\, [d\tau_\lambda / dz]\, dz.$$

Earth emitted spectra overlaid on Planck function envelopes

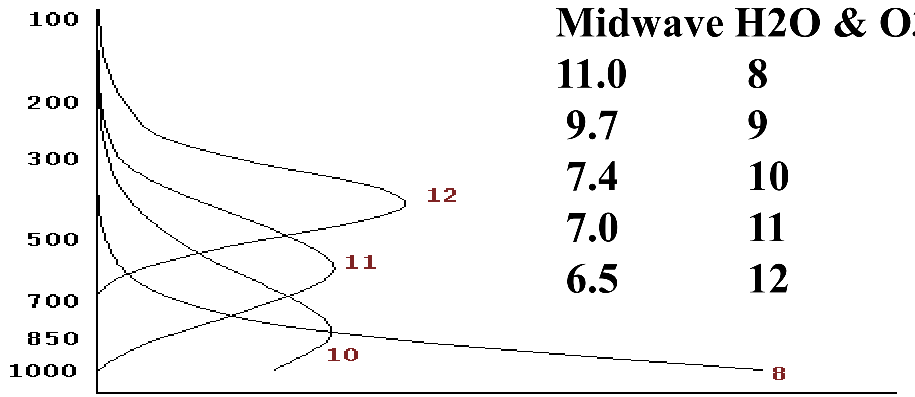


Weighting Functions



Longwave CO2

14.7	1	680	CO2, strat temp
14.4	2	696	CO2, strat temp
14.1	3	711	CO2, upper trop temp
13.9	4	733	CO2, mid trop temp
13.4	5	748	CO2, lower trop temp
12.7	6	790	H2O, lower trop moisture
12.0	7	832	H2O, dirty window



Midwave H2O & O3

11.0	8	907	window
9.7	9	1030	O3, strat ozone
7.4	10	1345	H2O, lower mid trop moisture
7.0	11	1425	H2O, mid trop moisture
6.5	12	1535	H2O, upper trop moisture

RTE in Cloudy Conditions

$$I_\lambda = \eta\, I^{cd}_\lambda + (1 - \eta)\, I^{c}_\lambda \quad \text{where cd = cloud, c = clear, } \eta = \text{cloud fraction}$$

$$I^{c}_\lambda = B_\lambda(T_s)\, \tau_\lambda(p_s) + \int_{p_s}^{o} B_\lambda(T(p))\, d\tau_\lambda \,.$$

$$I^{cd}_\lambda = (1-\varepsilon_\lambda)\, B_\lambda(T_s)\, \tau_\lambda(p_s) + (1-\varepsilon_\lambda) \int_{p_s}^{p_c} B_\lambda(T(p))\, d\tau_\lambda$$

$$+ \varepsilon_\lambda\, B_\lambda(T(p_c))\, \tau_\lambda(p_c) + \int_{p_c}^{o} B_\lambda(T(p))\, d\tau_\lambda$$

ε_λ is emittance of cloud. First two terms are from below cloud, third term is cloud contribution, and fourth term is from above cloud. After rearranging

$$I_\lambda - I_\lambda^{c} = \eta\varepsilon_\lambda \int_{p_s}^{p_c} \tau(p)\, \frac{dB_\lambda}{dp}\, dp \,.$$

Techniques for dealing with clouds fall into three categories: (a) searching for cloudless fields of view, (b) specifying cloud top pressure and sounding down to cloud level as in the cloudless case, and (c) employing adjacent fields of view to determine clear sky signal from partly cloudy observations.

Cloud Properties

RTE for cloudy conditions indicates dependence of cloud forcing (observed minus clear sky radiance) on cloud amount ($\eta\varepsilon_\lambda$) and cloud top pressure (p_c)

$$(I_\lambda - I_\lambda^{clr}) = \eta\varepsilon_\lambda \int_{p_s}^{p_c} \tau_\lambda \, dB_\lambda \, .$$

Higher colder cloud or greater cloud amount produces greater cloud forcing; dense low cloud can be confused for high thin cloud. Two unknowns require two equations.

p_c can be inferred from radiance measurements in two spectral bands where cloud emissivity is the same. $\eta\varepsilon_\lambda$ is derived from the infrared window, once p_c is known. This is the essence of the CO2 slicing technique.

Conclusion

- Bit Depth: given the range of observable values characterizes the minimal detectable variation in radiance and/or reflectance;
- Radiative Transfer Equation (IR): models the propagation of terrestrial emitted energy through the atmosphere