Investigations with Remote Sensing Data using MODIS, AIRS, & AMSU (4 Labs)

Paul Menzel
UW
and colleagues at CIMSS

Access to text, visualization tools, and data

For Remote Sensing Applications Text
ftp://ftp.ssec.wisc.edu/pub/menzel/AppMetSat09.pdf

For HYDRA
http://www.ssec.wisc.edu/hydra/

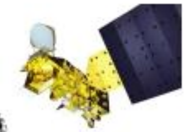
For MODIS data and quick browse images
http://rapidfire.sci.gsfc.nasa.gov/realtime

For MODIS data
http://ladsweb.nascom.nasa.gov/

For AIRS and AMSU data
http://daac.gsfc.nasa.gov/

HYperspectral viewer for Development of Research Applications - HYDRA

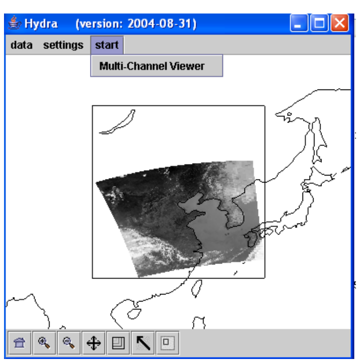
MSG,
GOES



MODIS,
AIRS, IASI,
AMSU,
CALIPSO

Freely available software
For researchers and educators
Computer platform independent
Extendable to more sensors and applications
Based in VisAD
(Visualization for Algorithm Development)
Uses Jython (Java implementation of Python)
runs on most machines
512MB main memory & 32MB graphics card suggested
on-going development effort

Rink et al, BAMS 2007



Developed at CIMSS by
Tom Rink
Tom Whittaker
Kevin Baggett

With guidance from
Paolo Antonelli
Liam Gumley
Paul Menzel
Allen Huang

CIMSS

http://www.ssec.wisc.edu/hydra/

Intro to VIS-IR Radiation Lab

Lecture - Labs in Italy

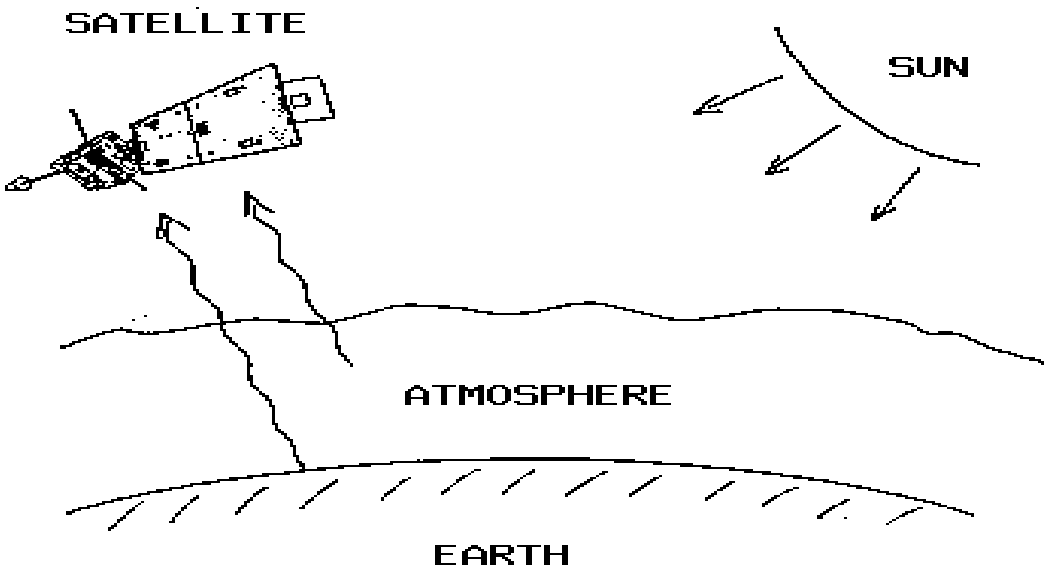
18 – 25 May 2012 in Bologna

28 May - 1 June 2012 in Potenza

Paul Menzel
UW/CIMSS/AOS

Relevant Material in Applications of Meteorological Satellites

Satellite remote sensing of the Earth-atmosphere

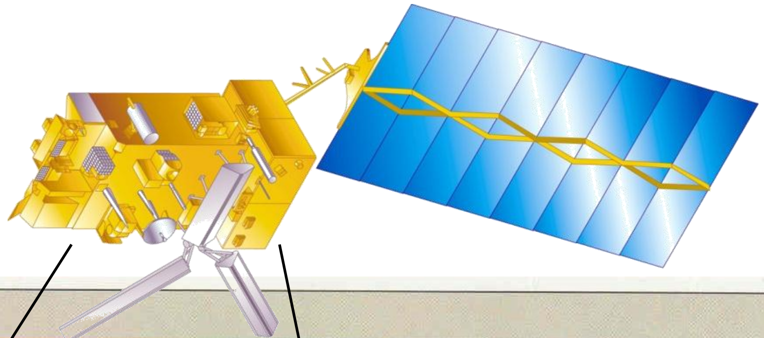


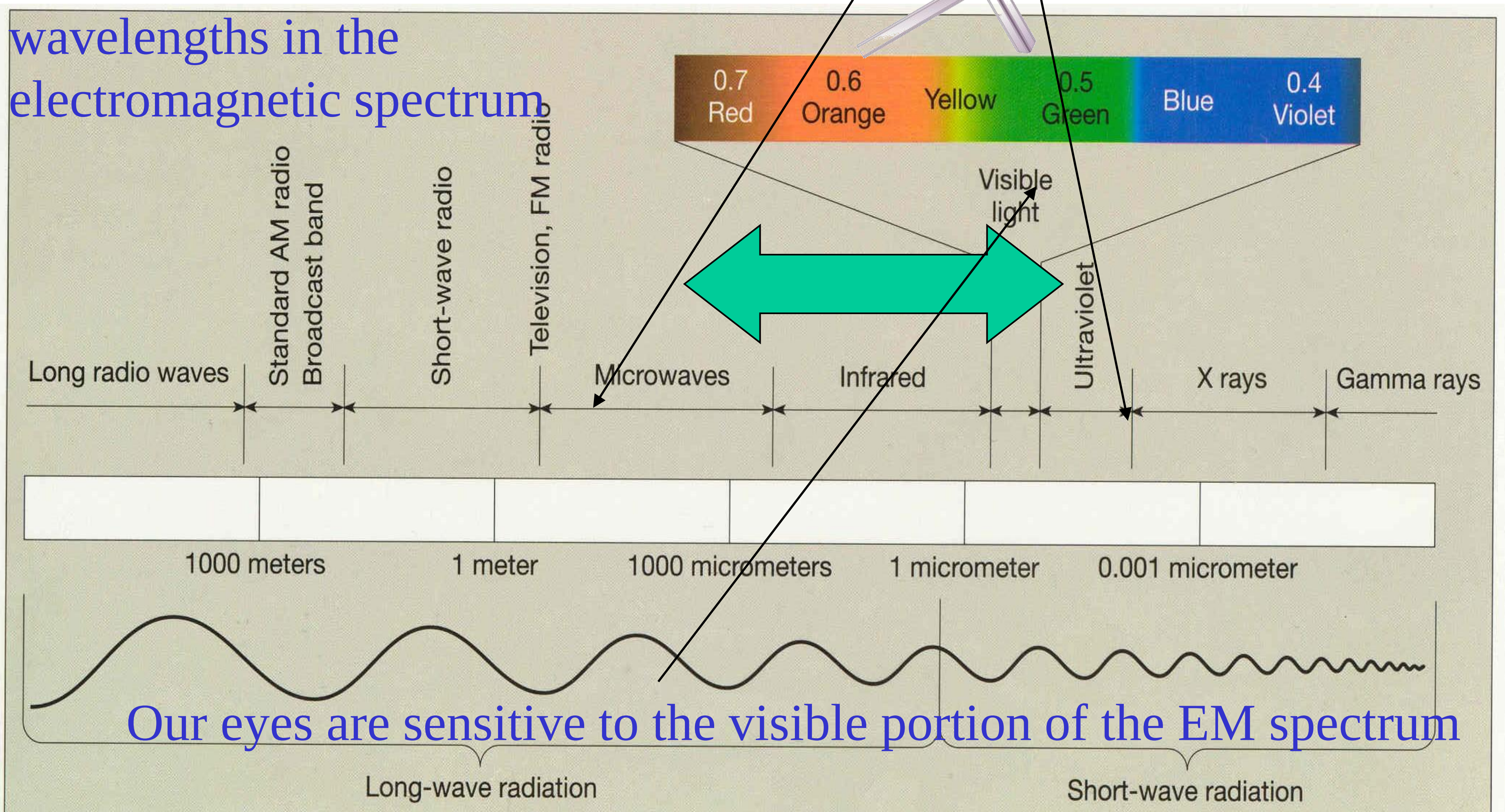
Observations depend on

telescope characteristics (resolving power, diffraction)
detector characteristics (field of view, signal to noise)
communications bandwidth (bit depth)
spectral intervals (window, absorption band)
time of day (daylight visible)
atmospheric state (T, Q, clouds)
earth surface (Ts, vegetation cover)

Electromagnetic spectrum

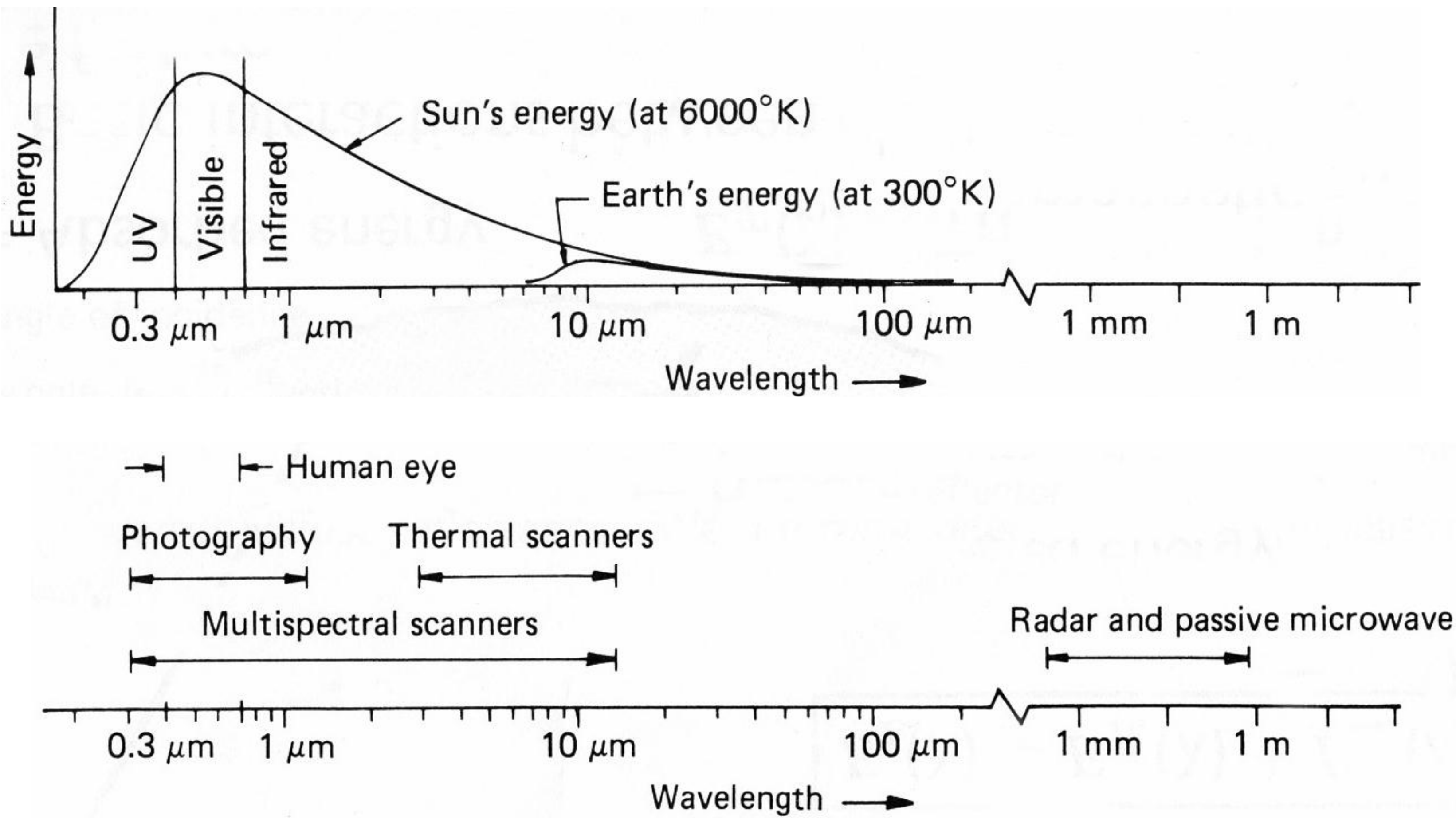
Remote sensing uses radiant energy that is reflected and emitted from Earth at various wavelengths in the electromagnetic spectrum





Our eyes are sensitive to the visible portion of the EM spectrum

Spectral Characteristics of Energy Sources and Sensing Systems



Definitions of Radiation

QUANTITY	SYMBOL	UNITS
Energy	dQ	Joules
Flux	dQ/dt	Joules/sec = Watts
Irradiance	dQ/dt/dA	Watts/meter^2
Monochromatic Irradiance	dQ/dt/dA/dλ	W/m^2/micron
	or	
	dQ/dt/dA/dν	W/m^2/cm^{-1}
Radiance	dQ/dt/dA/dλ/dΩ	W/m^2/micron/ster
	or	
	dQ/dt/dA/dν/dΩ	W/m^2/cm^{-1}/ster

Using wavelengths

Planck's Law $B(\lambda,T) = c_1 / \lambda^5 / [e^{c_2/\lambda T} - 1]$ $(mW/m^2/ster/cm)$

where λ = wavelengths in cm

T = temperature of emitting surface (deg K)

$c_1 = 1.191044 \times 10^{-5}$ $(mW/m^2/ster/cm^{-4})$

$c_2 = 1.438769$ (cm deg K)

Wien's Law $dB(\lambda_{max},T) / d\lambda = 0$ where $\lambda(max) = .2897/T$

indicates peak of Planck function curve shifts to shorter wavelengths (greater wavenumbers) with temperature increase. Note $B(\lambda_{max},T) \sim T^5$.

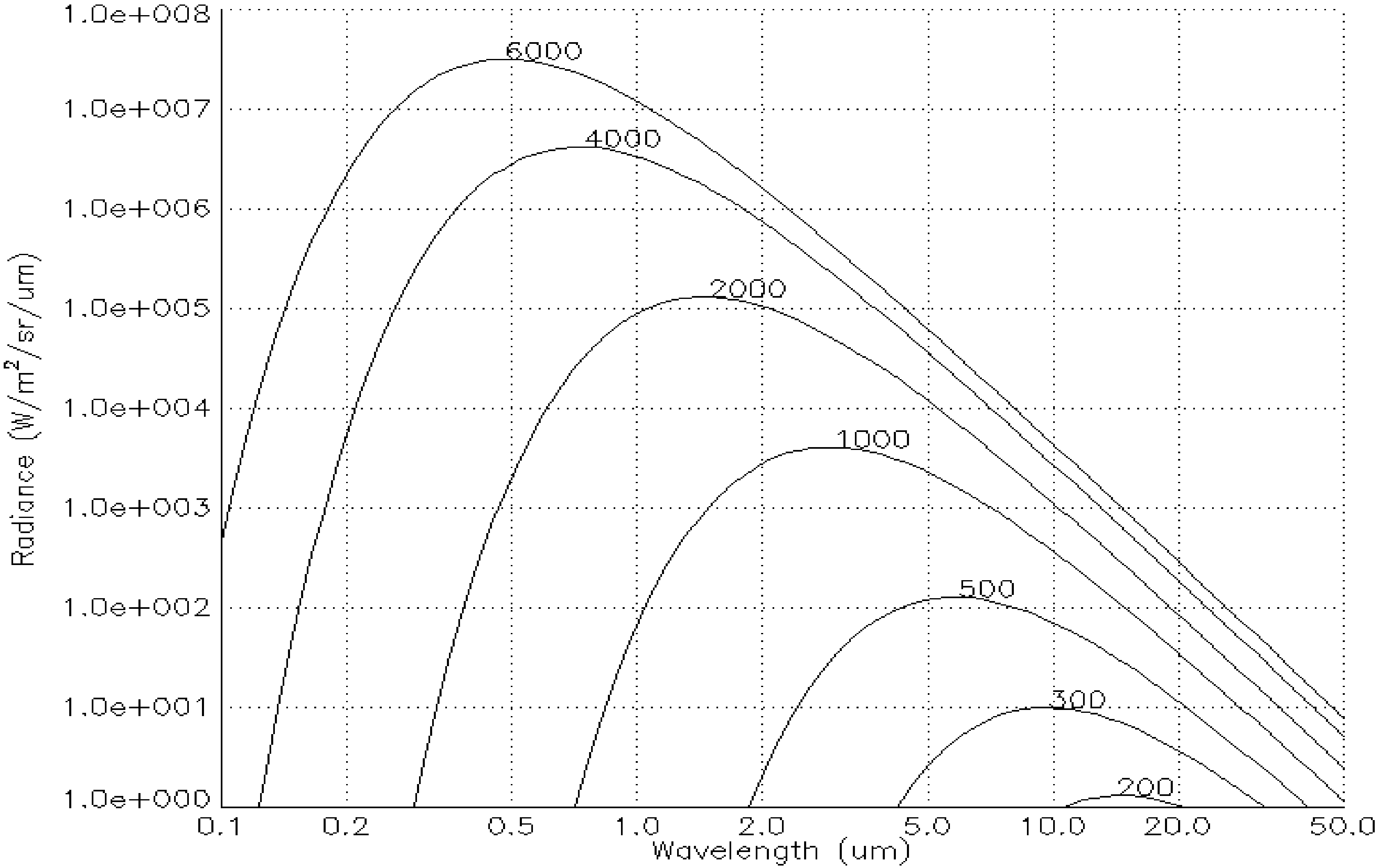
Stefan-Boltzmann Law $E = \pi \int_0^\infty B(\lambda,T)\, d\lambda = \sigma T^4$, where σ = 5.67 x 10-8 W/m2/deg4.

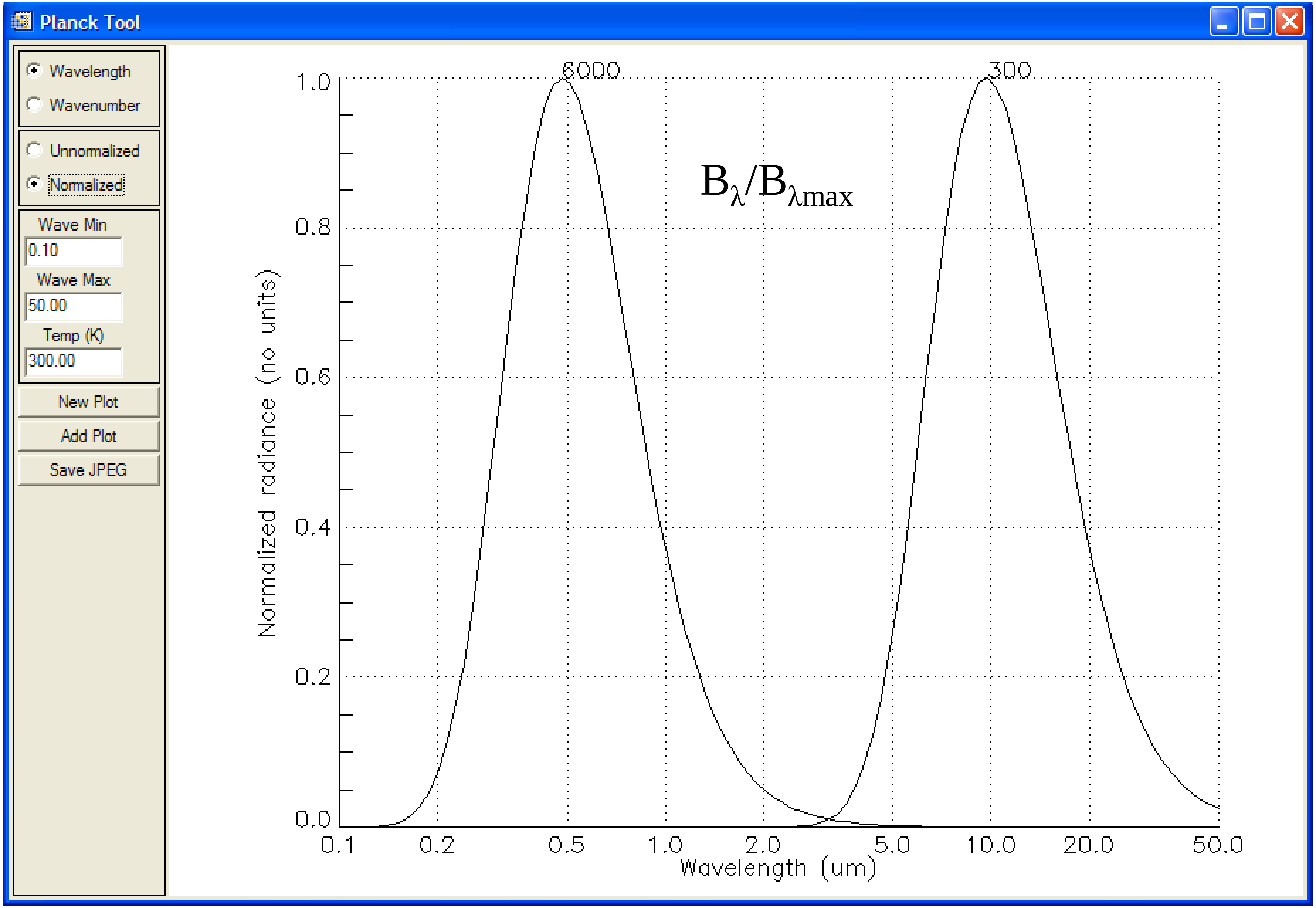
states that irradiance of a black body (area under Planck curve) is proportional to T^4 .

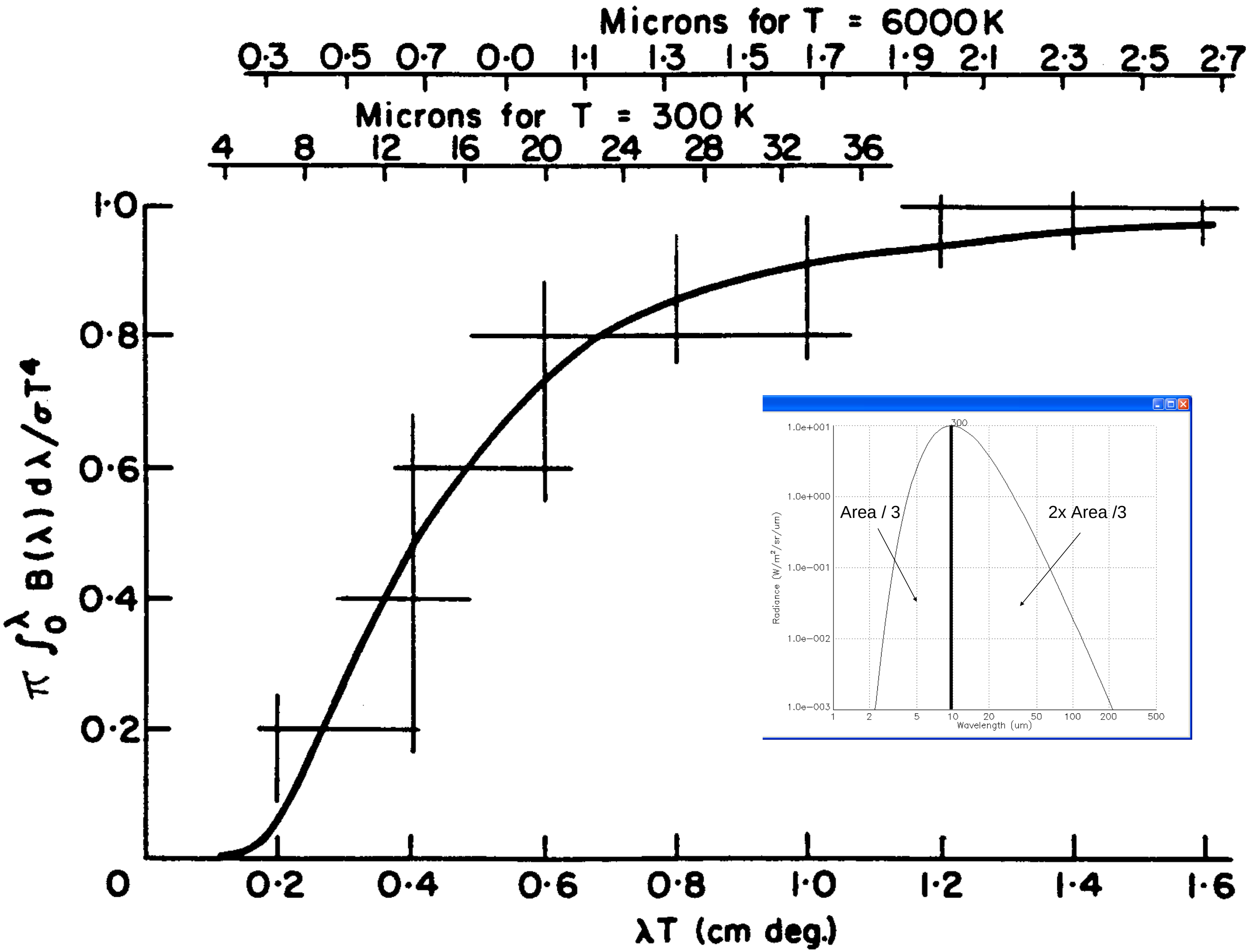
Brightness Temperature

$$T = c_2 / [\lambda \ln(\frac{c_1}{\lambda^5 B_\lambda} + 1)]$$ is determined by inverting Planck function

Spectral Distribution of Energy Radiated from Blackbodies at Various Temperatures




Planck Tool
Wavelength
Wavenumber
Unnormalized
Normalized
Wave Min
0.10
Wave Max
50.00
Temp (K)
300.00
New Plot
Add Plot
Save JPEG
6000
300
$B_\lambda / B_{\lambda max}$
Normalized radiance (no units)
1.0
0.8
0.6
0.4
0.2
0.0
0.1
0.2
0.5
1.0
2.0
5.0
10.0
20.0
50.0
Wavelength (um)



Microns for T = 6000 K
0·3 0·5 0·7 0·0 1·1 1·3 1·5 1·7 1·9 2·1 2·3 2·5 2·7
Microns for T = 300 K
4 8 12 16 20 24 28 32 36
$\pi \int_0^\lambda B(\lambda)\,d\lambda / \sigma T^4$
1·0 0·8 0·6 0·4 0·2 0
0·2 0·4 0·6 0·8 1·0 1·2 1·4 1·6
λT (cm deg.)
Area / 3
2x Area /3
300
Radiance (W/m^2/sr/um)
1.0e+001 1.0e+000 1.0e−001 1.0e−002 1.0e−003
1 2 5 10 20 50 100 200 500
Wavelength (um)

Using wavenumbers

Planck's Law $B(\nu,T) = c_1\nu^3 / [e^{c_2\nu/T} - 1]$ $(mW/m^2/ster/cm^{-1})$

where ν = # wavelengths in one centimeter (cm-1)

T = temperature of emitting surface (deg K)

c_1 = 1.191044 x 10-5 $(mW/m^2/ster/cm^{-4})$

c_2 = 1.438769 (cm deg K)

Wien's Law $dB(\nu_{max},T) / d\nu = 0$ where $\nu max) = 1.95T$

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states that irradiance of a black body (area under Planck curve) is proportional to T^4 .

Brightness Temperature

$$T = c_2\nu/[\ln(\frac{c_1\nu^3}{B_\nu} + 1)]$$ is determined by inverting Planck function

Using wavenumbers

$$B(\nu,T) = c_1\nu^3 / [e^{c_2\nu/T} - 1]$$

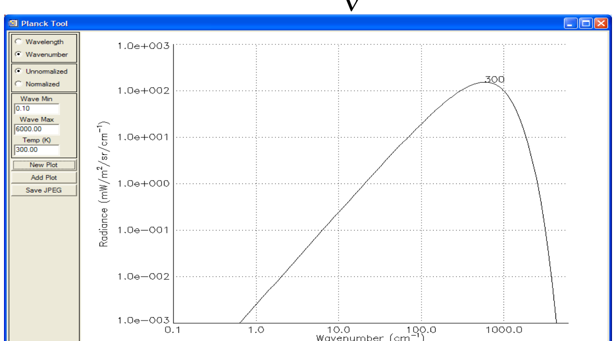
$(mW/m^2/ster/cm^{-1})$

ν(max in cm-1) = 1.95T

$B(\nu_{max},T) \sim T^{**}3.$

$$E = \pi \int_0^\infty B(\nu,T)\, d\nu = \sigma T^4,$$

$$T = c_2\nu/[\ln(\frac{c_1\nu^3}{B_\nu} + 1)]$$



Using wavelengths

$$B(\lambda,T) = c_1 / \{ \lambda^5 [e^{c_2/\lambda T} - 1] \}$$

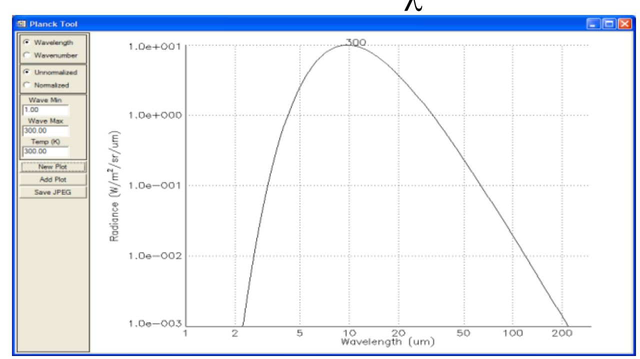
$(mW/m^2/ster/\mu m)$

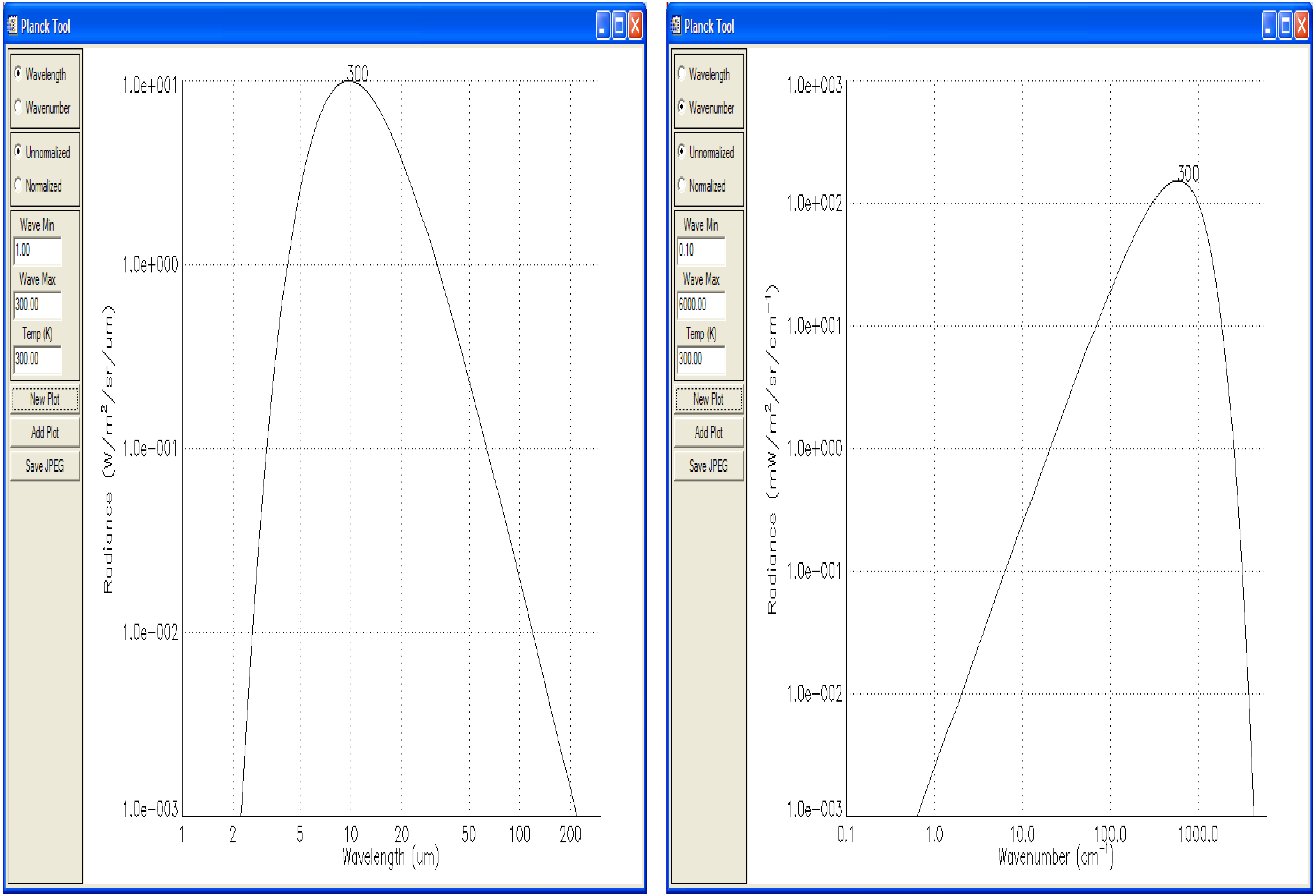
λ(max in cm)T = 0.2897

$B(\lambda_{max},T) \sim T^{**}5.$

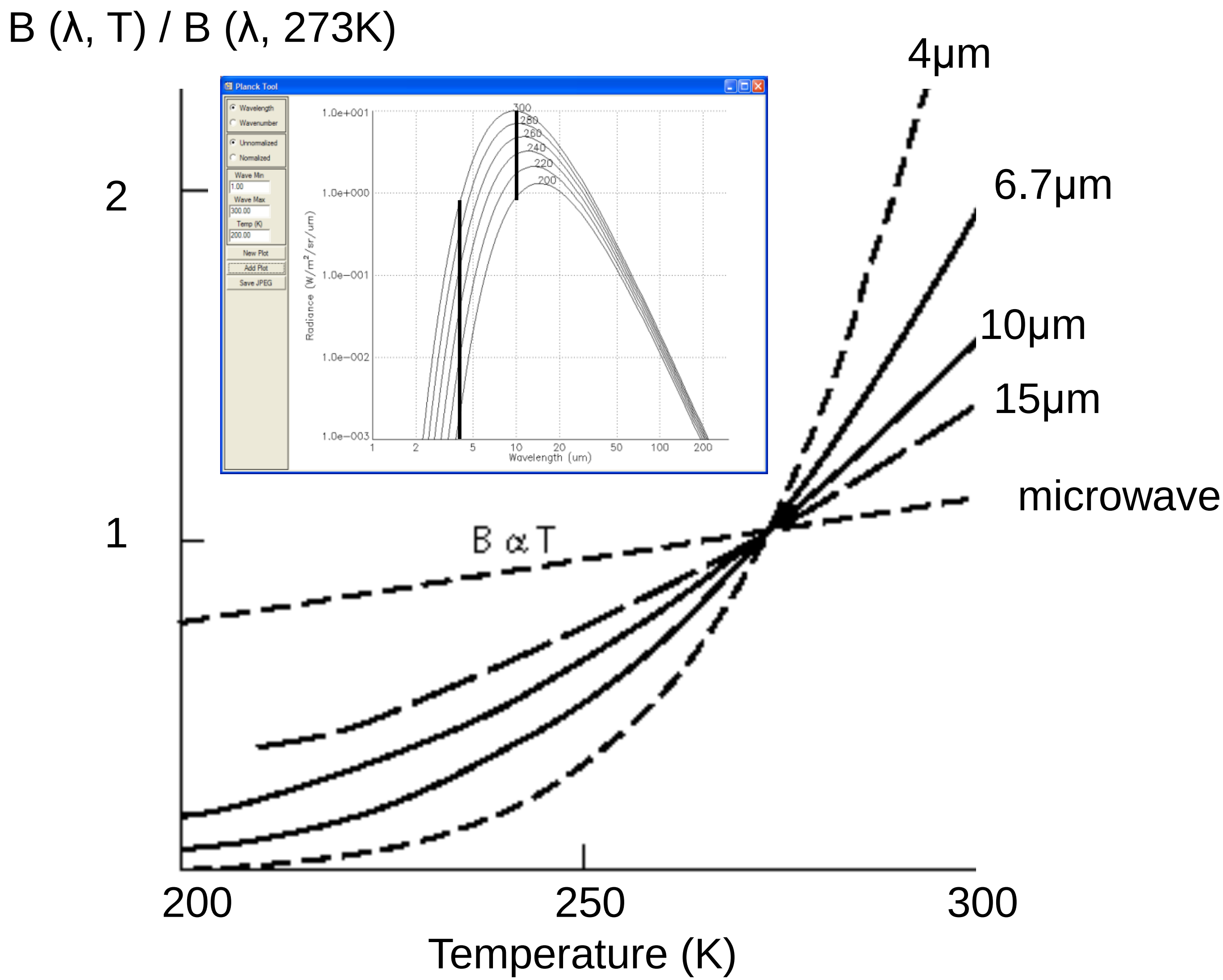
$$E = \pi \int_0^\infty B(\lambda,T)\, d\lambda = \sigma T^4,$$

$$T = c_2/[\lambda \ln(\frac{c_1}{\lambda^5 B_\lambda} + 1)]$$




Planck Tool
Wavelength
Wavenumber
Unnormalized
Normalized
Wave Min
1.00
Wave Max
300.00
Temp (K)
300.00
New Plot
Add Plot
Save JPEG
300
Radiance (W/m2/sr/um)
Wavelength (um)
Planck Tool
Wavelength
Wavenumber
Unnormalized
Normalized
Wave Min
0.10
Wave Max
6000.00
Temp (K)
300.00
New Plot
Add Plot
Save JPEG
300
Radiance (mW/m2/sr/cm-1)
Wavenumber (cm-1)

Temperature Sensitivity of B(λ,T) for typical earth temperatures



(Approximation of) B as function of α and T

$\Delta B/B=\alpha\ \Delta T/T$

Integrating the Temperature Sensitivity Equation Between T_{ref} and T (B_{ref} and B):

$B=B_{ref}(T/T_{ref})^{\alpha}$

Where $\alpha=c_2\nu/T_{ref}$ (in wavenumber space)

$$B = B_{ref} (T/T_{ref})^{\alpha}$$

$$\Downarrow$$

$$B = (B_{ref} / T_{ref}^{\alpha}) \; T^{\alpha}$$

$$\Downarrow$$

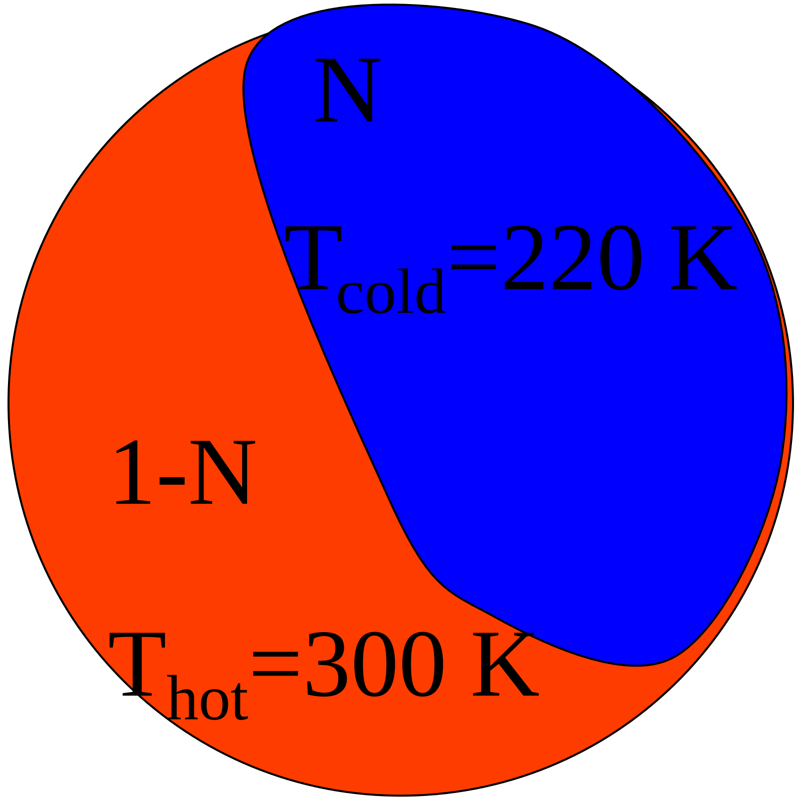
$$B \propto T^{\alpha}$$

The temperature sensitivity indicates the power to which the Planck radiance depends on temperature, since B proportional to T^{α} satisfies the equation. For infrared wavelengths,

$\alpha = c_2\nu/T = c_2/\lambda T.$

Wavenumber	Typical Scene Temperature	Temperature Sensitivity
900	300	4.32
2500	300	11.99

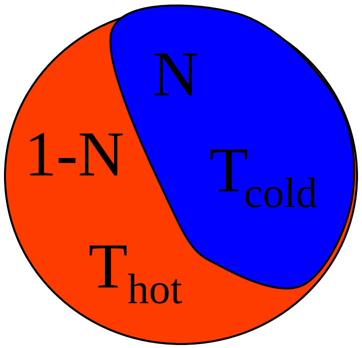
Non-Homogeneous FOV



$$B = N*B(T_{cold}) + (1-N)*B(T_{hot})$$

$$BT = N*T_{cold} + (1-N)*T_{hot}$$

For NON-UNIFORM FOVs:



$B_{obs}=NB_{cold}+(1-N)B_{hot}$

$B_{obs}=N\ B_{ref}(T_{cold}/T_{ref})^{\alpha}+(1-N)\ B_{ref}(T_{hot}/T_{ref})^{\alpha}$

$B_{obs}=B_{ref}(1/T_{ref})^{\alpha}\ (N\ T_{cold}^{\alpha}+(1-N)T_{hot}^{\alpha})$

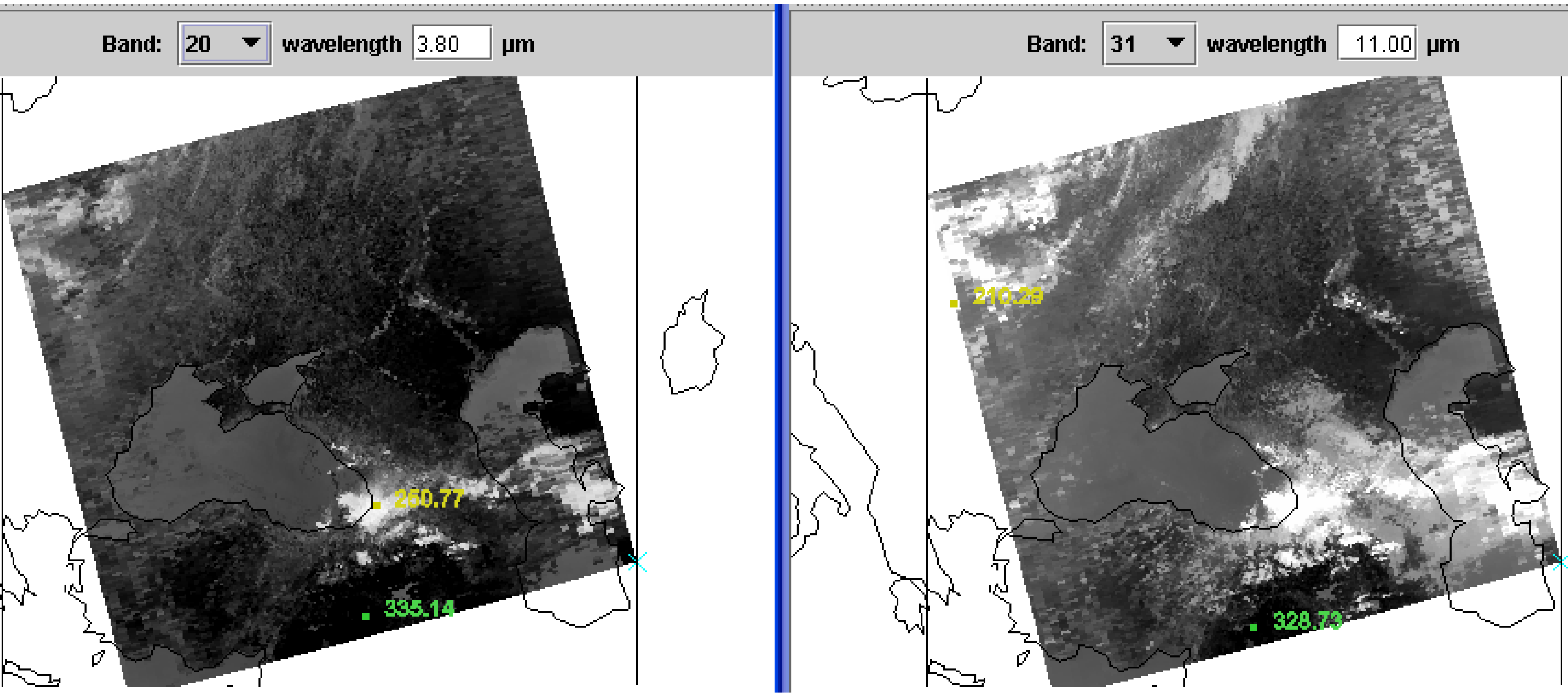
For N=.5

$B_{obs}/B_{ref}=.5\ (1/T_{ref})^{\alpha}\ (T_{cold}^{\alpha}+T_{hot}^{\alpha})$

$B_{obs}/B_{ref}=.5\ (1/T_{ref}T_{cold})^{\alpha}\ (1+(T_{hot}/T_{cold})^{\alpha})$

The greater α the more predominant the hot term

At 4 μm (α=12) the hot term more dominating than at 11 μm (α=4)



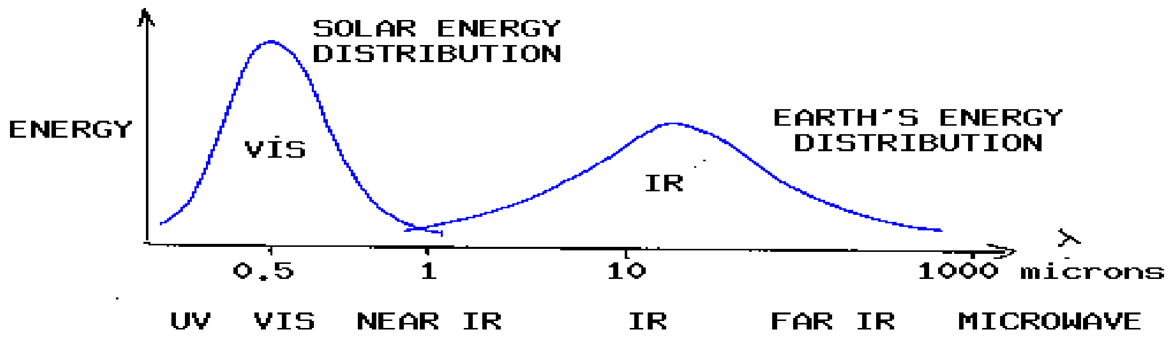
Cloud edges and broken clouds appear different in 11 and 4 um images.

T(11)**4=(1-N)*Tclr**4+N*Tcld**4~(1-N)*300**4+N*200**4
T(4)**12=(1-N)*Tclr**12+N*Tcld**12~(1-N)*300**12+N*200**12

Cold part of pixel has more influence for B(11) than B(4)

Relevant Material in Applications of Meteorological Satellites

Solar (visible) and Earth emitted (infrared) energy

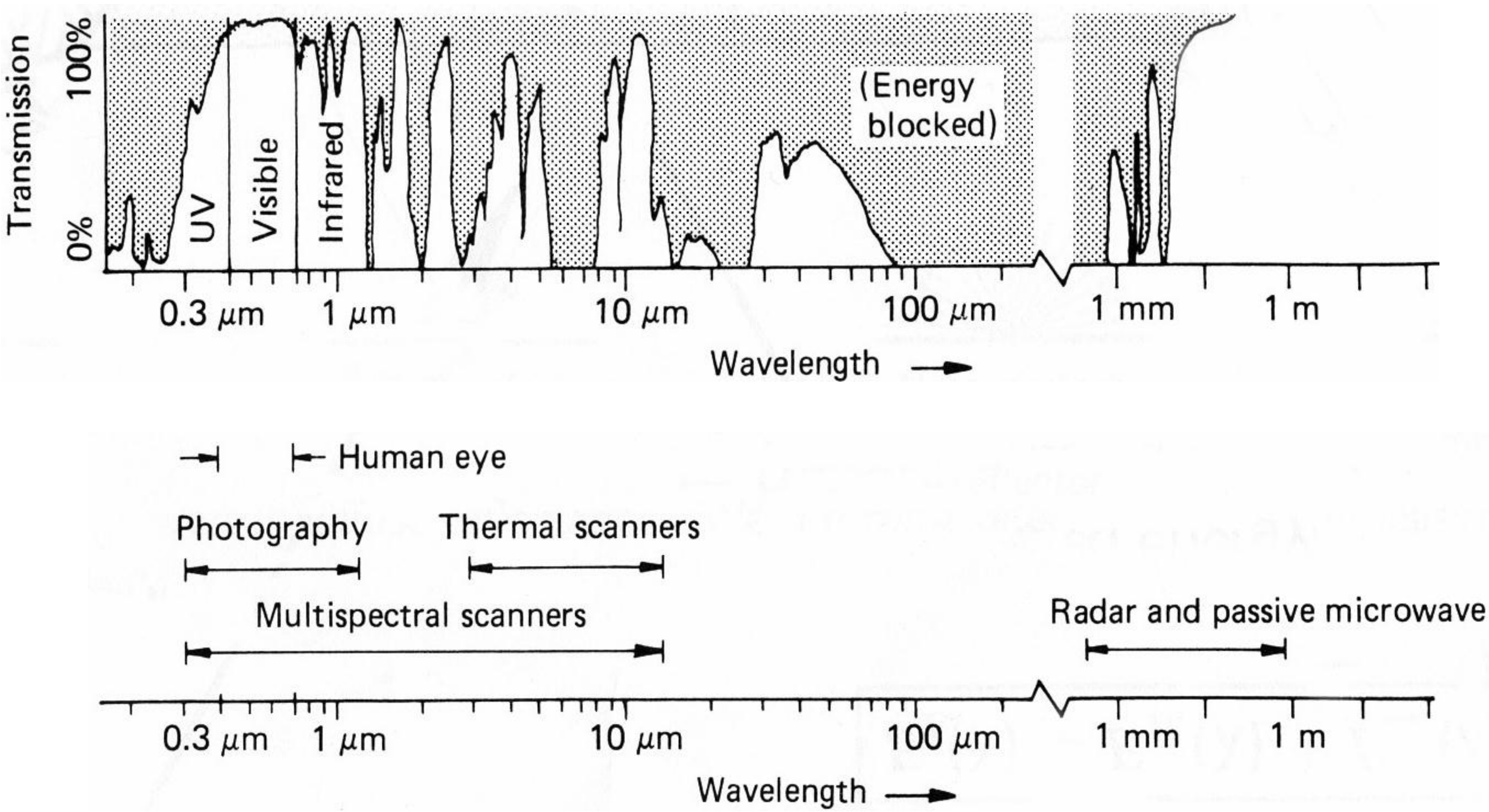


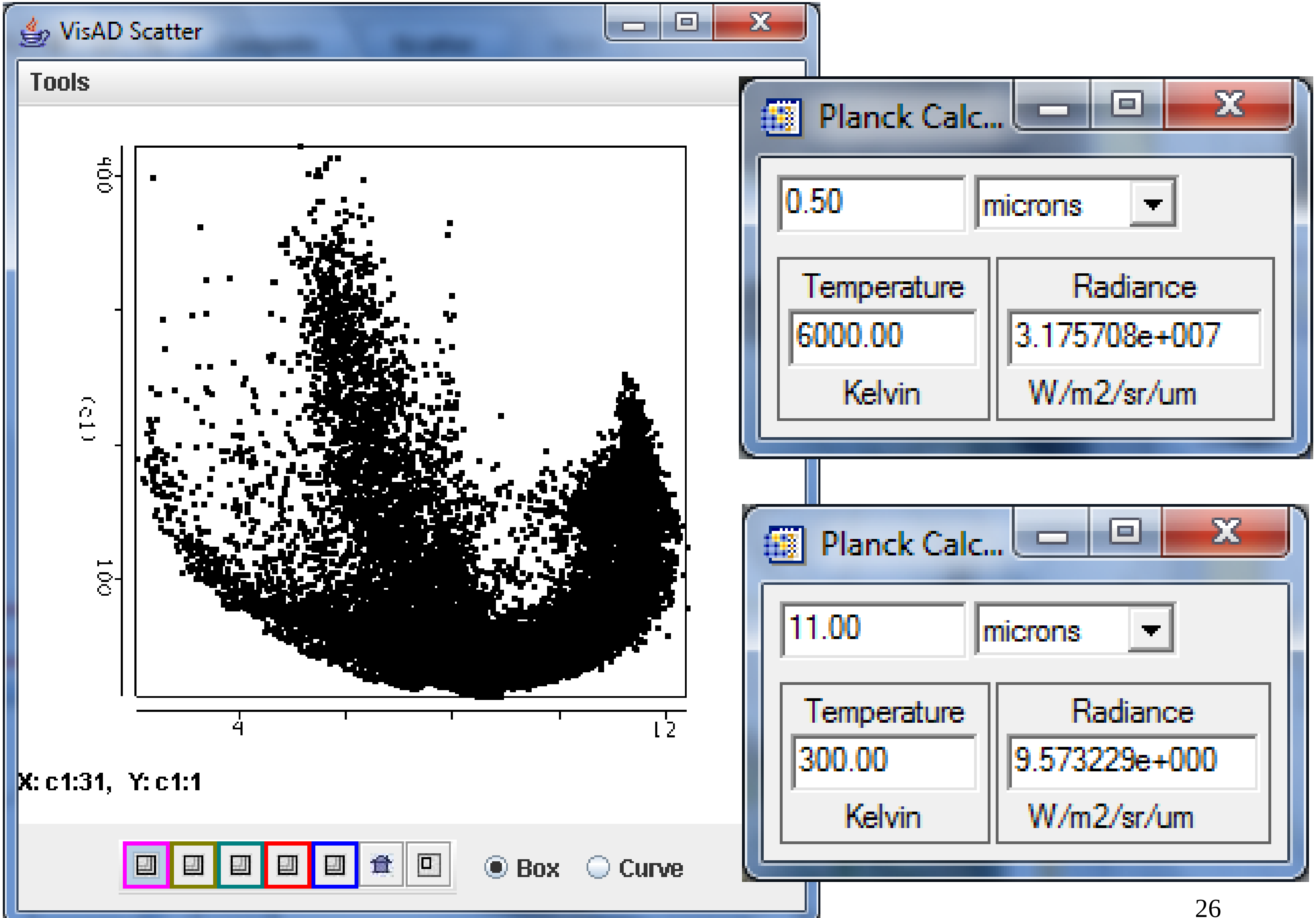
Incoming solar radiation (mostly visible) drives the earth-atmosphere (which emits infrared).

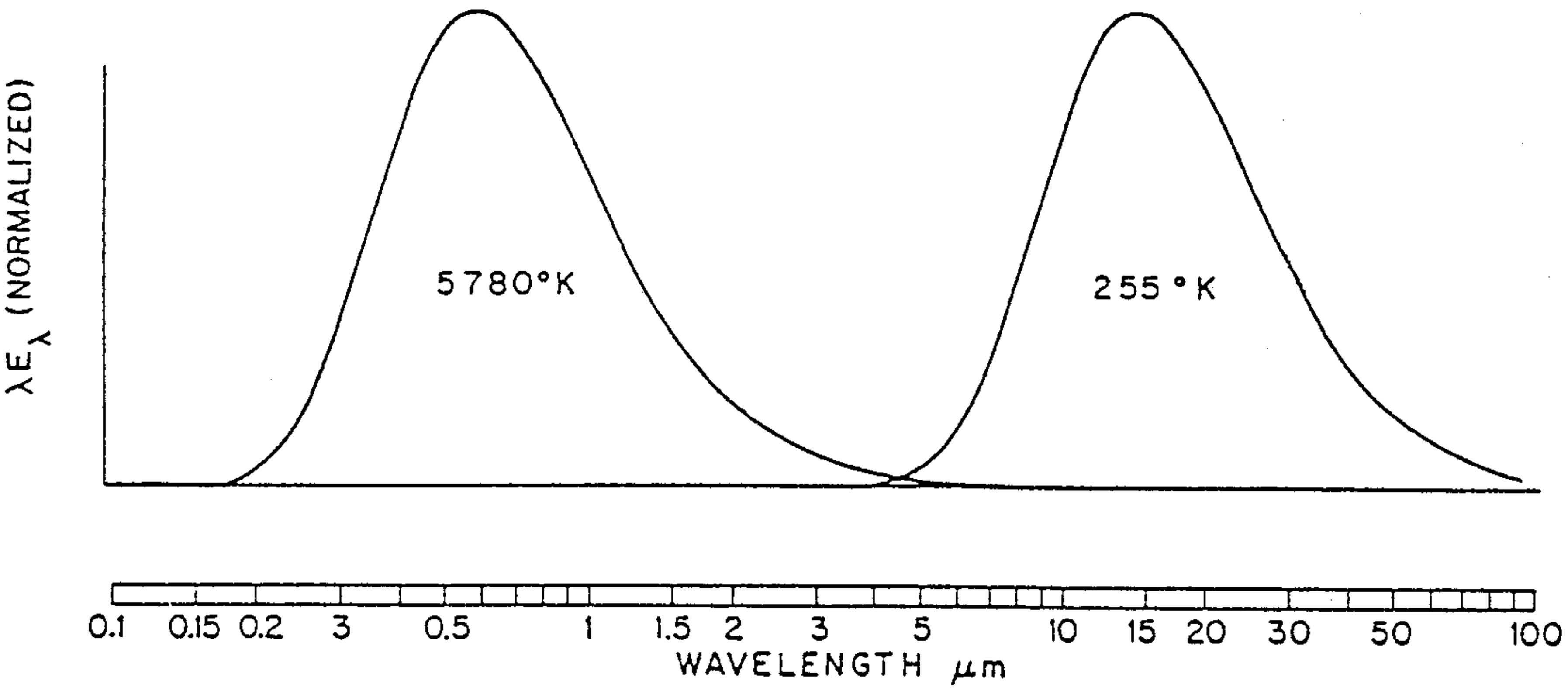
Over the annual cycle, the incoming solar energy that makes it to the earth surface (about 50 %) is balanced by the outgoing thermal infrared energy emitted through the atmosphere.

The atmosphere transmits, absorbs (by H2O, O2, O3, dust) reflects (by clouds), and scatters (by aerosols) incoming visible; the earth surface absorbs and reflects the transmitted visible. Atmospheric H2O, CO2, and O3 selectively transmit or absorb the outgoing infrared radiation. The outgoing microwave is primarily affected by H2O and O2.

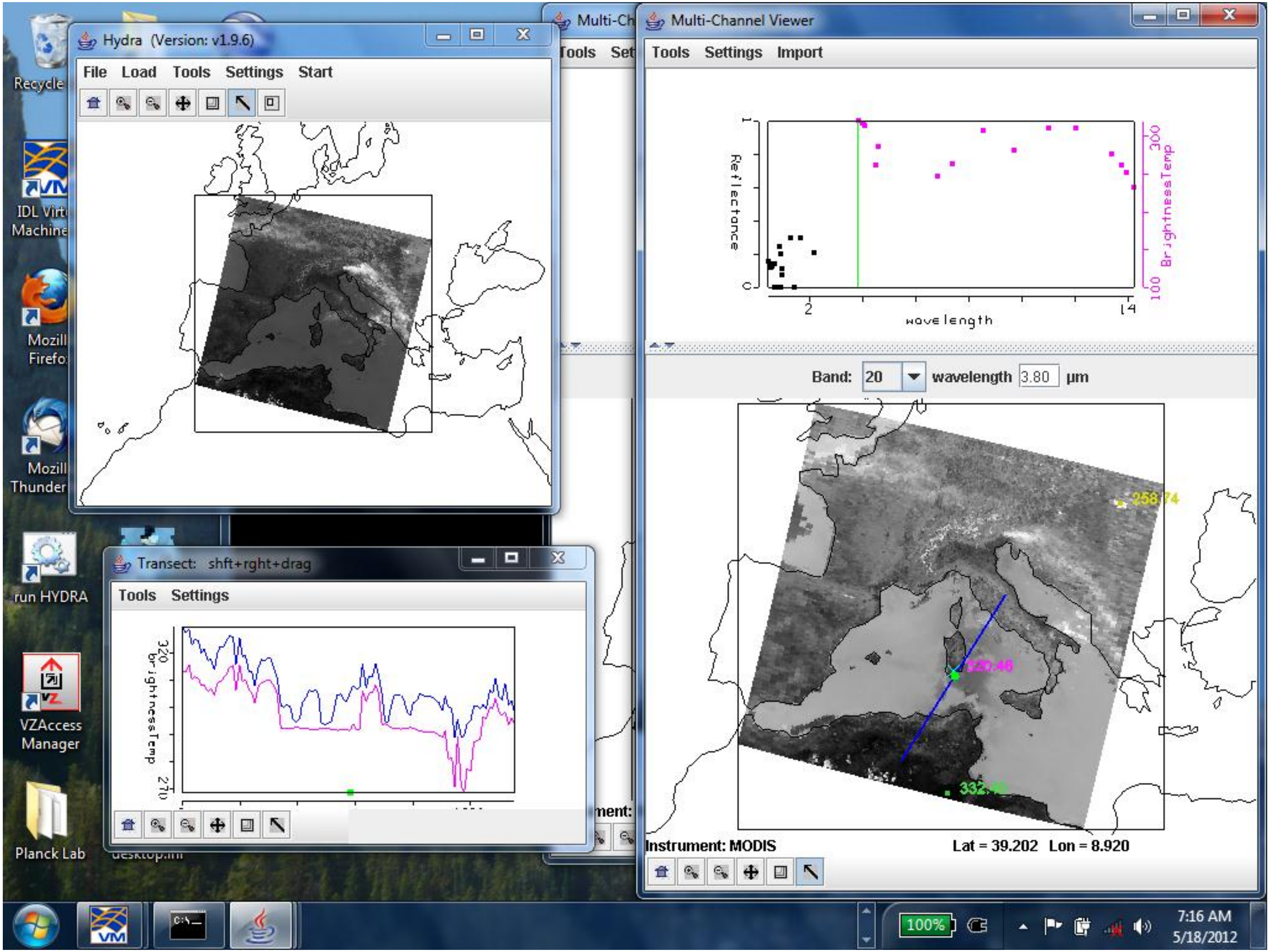
Spectral Characteristics of Atmospheric Transmission and Sensing Systems



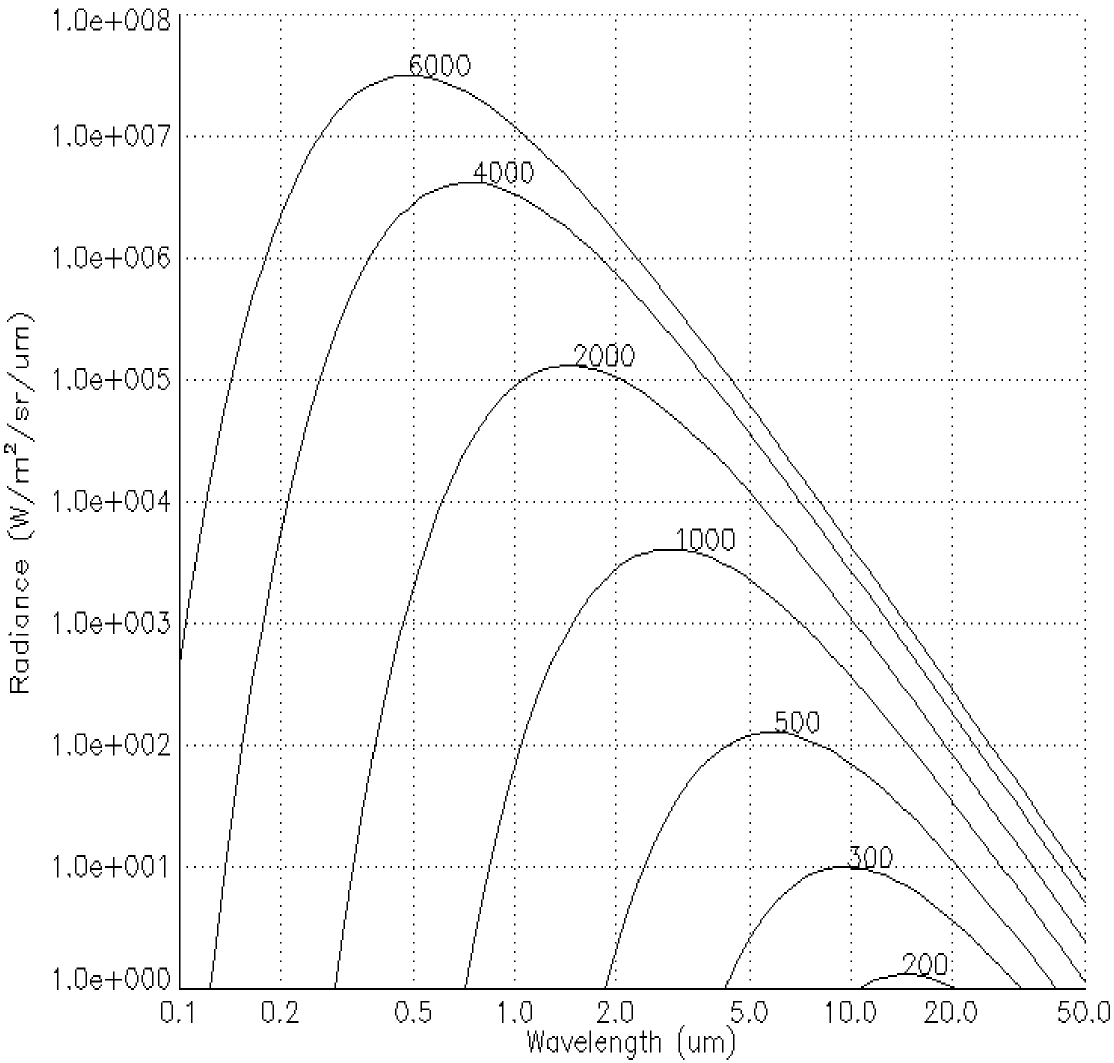

VisAD Scatter
Tools
400
(c1)
100
4
12
X: c1:31, Y: c1:1
Box
Curve
Planck Calc...
0.50
microns
Temperature
6000.00
Kelvin
Radiance
3.175708e+007
W/m2/sr/um
Planck Calc...
11.00
microns
Temperature
300.00
Kelvin
Radiance
9.573229e+000
W/m2/sr/um



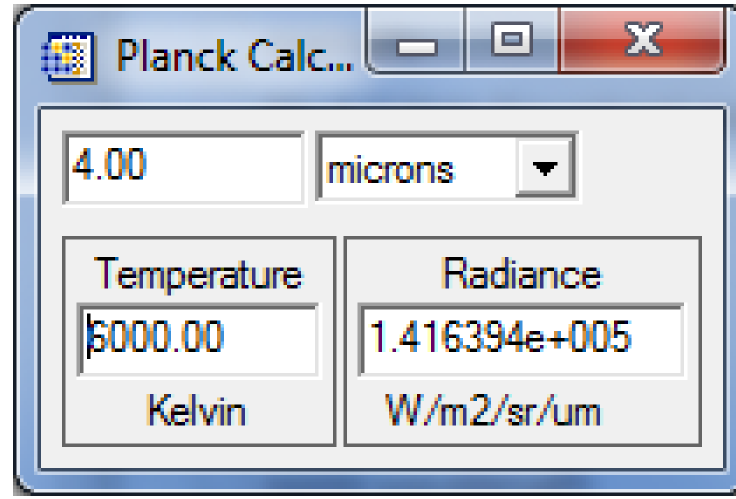
Normalized black body spectra representative of the sun (left) and earth (right), plotted on a logarithmic wavelength scale. The ordinate is multiplied by wavelength so that the area under the curves is proportional to irradiance.



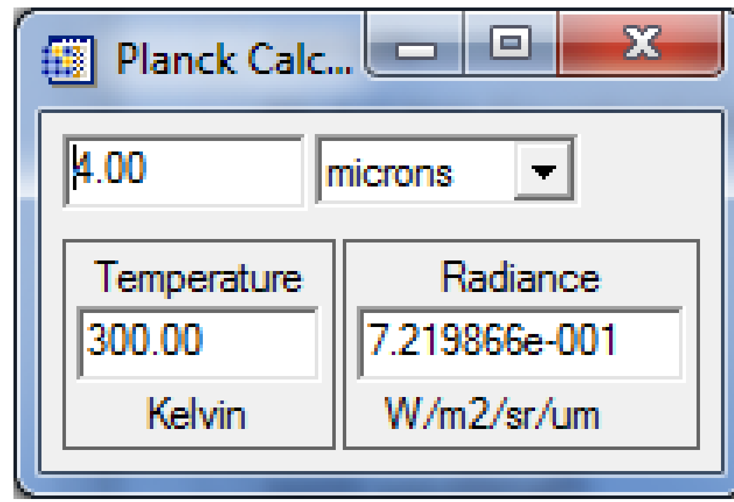
Hydra (Version: v1.9.6)
File Load Tools Settings Start
Multi-Channel Viewer
Tools Settings Import
Reflectance
BrightnessTemp
wavelength
Band: 20 wavelength 3.80 µm
Instrument: MODIS
Lat = 39.202 Lon = 8.920
Transect: shft+rght+drag
Tools Settings
brightnessTemp
Recycle
IDL Virt Machine
run HYDRA
VZAccess Manager
Planck Lab
7:16 AM
5/18/2012



1.0e+008
1.0e+007
1.0e+006
1.0e+005
1.0e+004
1.0e+003
1.0e+002
1.0e+001
1.0e+000
Radiance (W/m²/sr/um)
6000
4000
2000
1000
500
300
200
0.1
0.2
0.5
1.0
2.0
5.0
10.0
20.0
50.0
Wavelength (um)



Planck Calc...
4.00
microns
Temperature
6000.00
Kelvin
Radiance
1.416394e+005
W/m2/sr/um

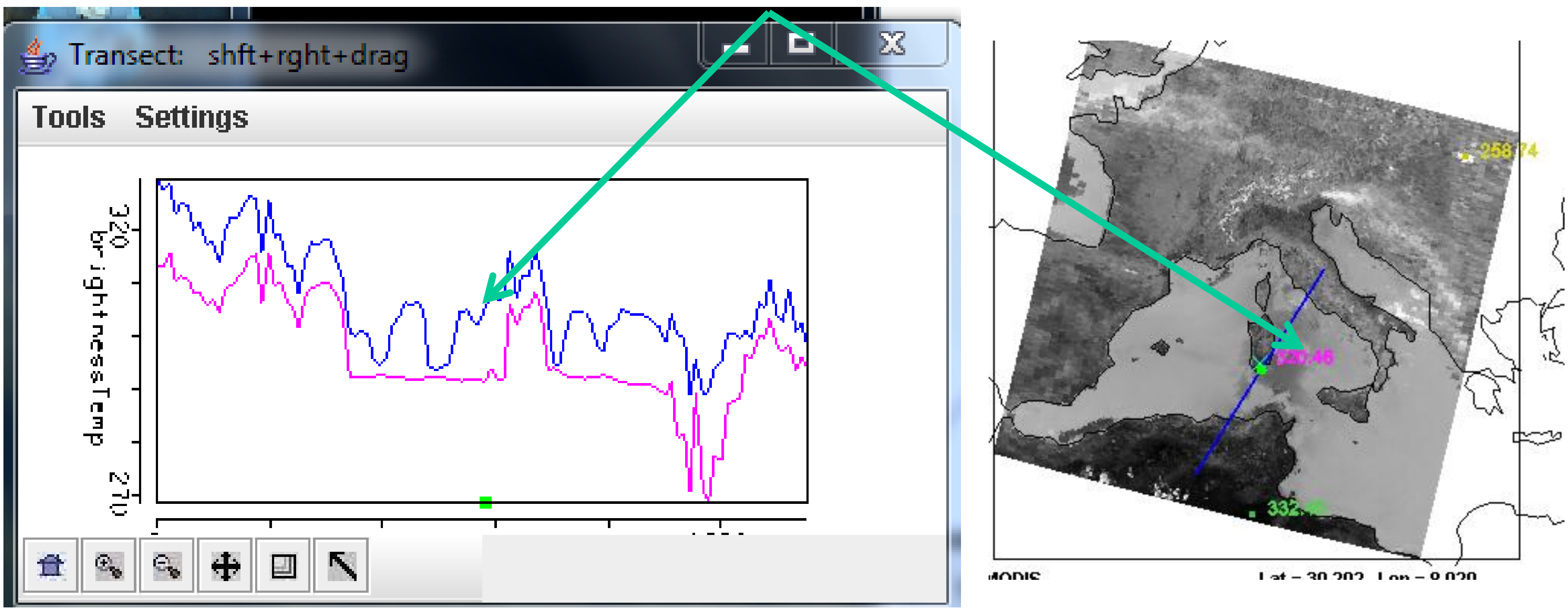


Planck Calc...
4.00
microns
Temperature
300.00
Kelvin
Radiance
7.219866e-001
W/m2/sr/um

BT11=290K and BT4=310K.
What fraction of R4 is due to reflected solar radiance?

R4 = R4 refl + R4 emiss
BT4 emiss = BT11
R4 ~ T**12

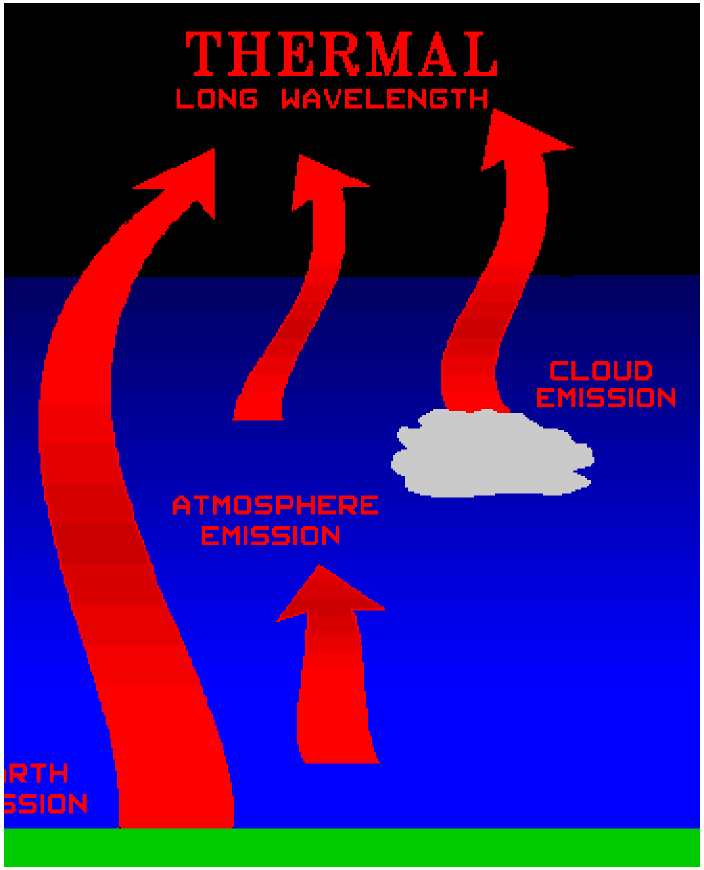
Fraction = [310**12 – 290**12]/ 310**12 ~ .55



Radiative Transfer Equation

in the IR

Infrared
(Emissive Bands)



Emission, Absorption, Reflection, and Scattering

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

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

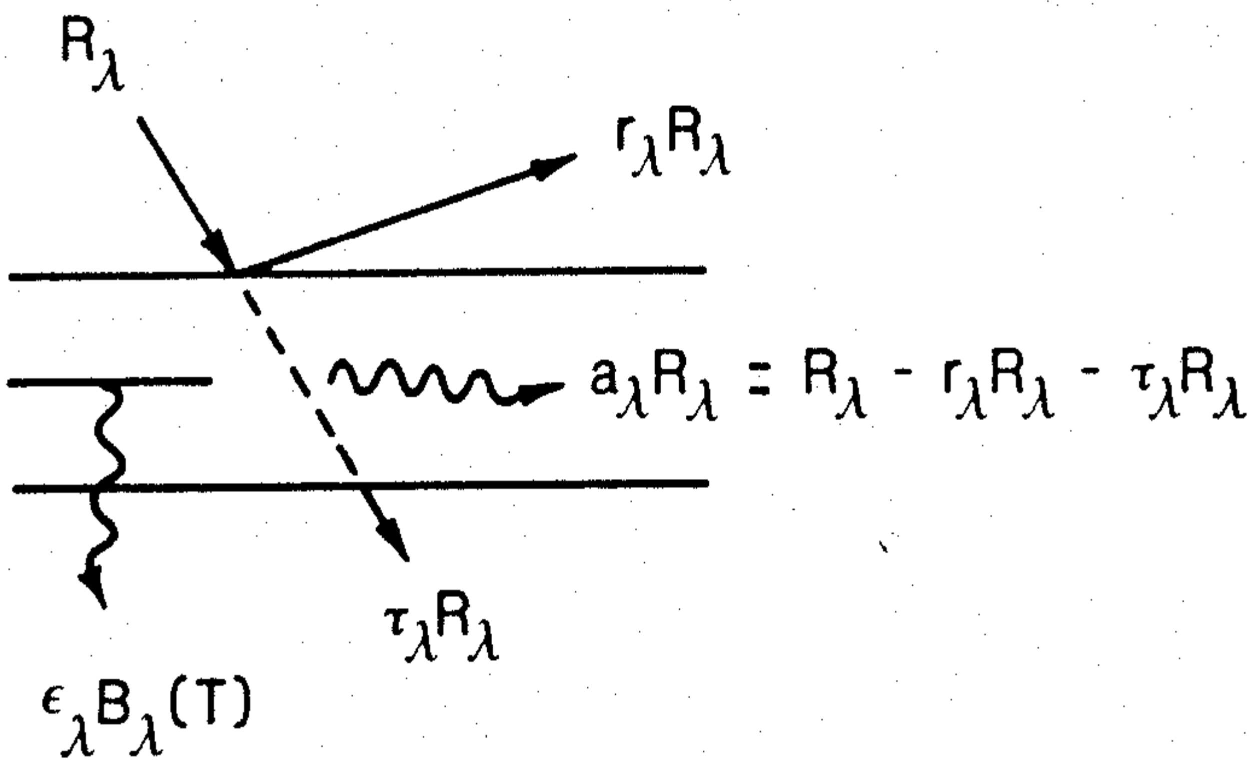
Thus, materials which are strong absorbers at a given wavelength are also strong emitters at that wavelength; similarly weak absorbers are weak emitters.

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

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

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).



"ENERGY CONSERVATION"

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

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

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

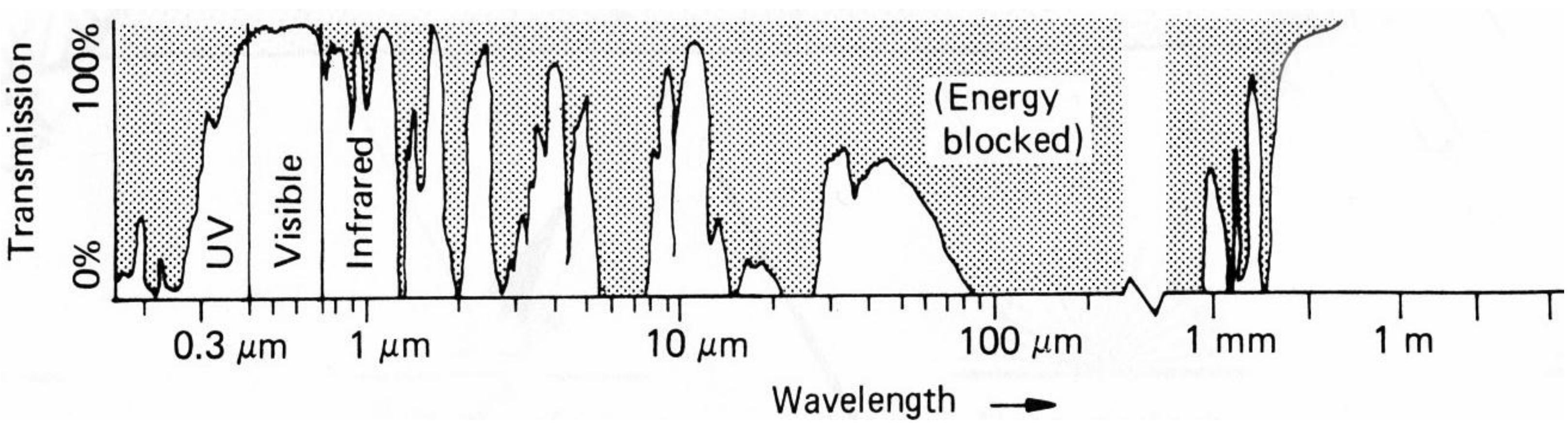
$k_\lambda u$ is a measure of the cumulative depletion that the beam of radiation has experienced as a result of its passage through the layer and is often called the optical depth σ_λ.

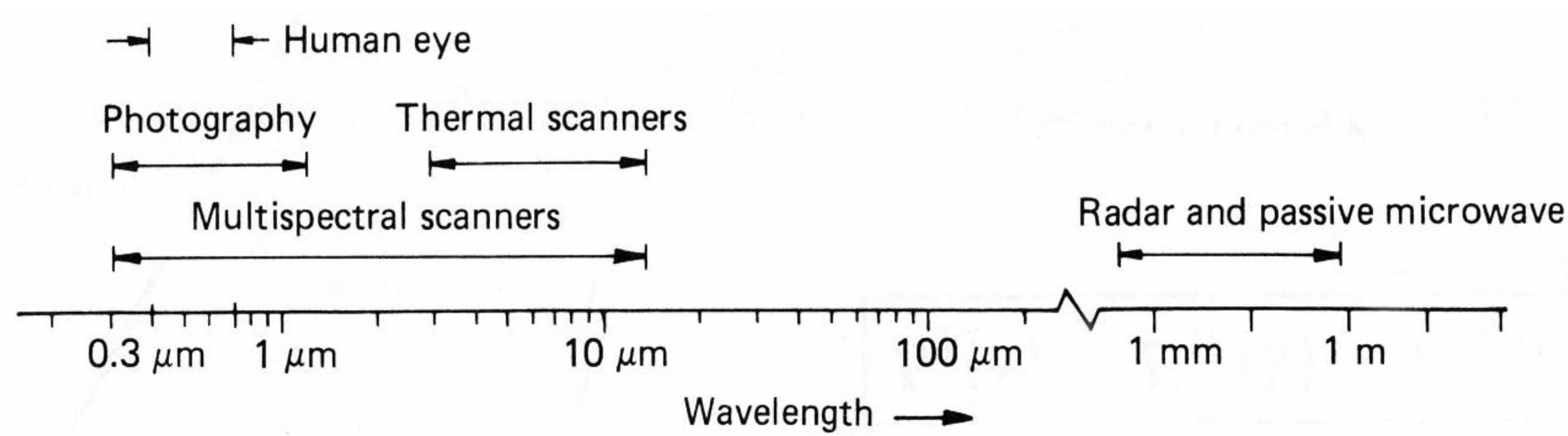
Realizing that the hydrostatic equation implies $g \rho \, dz = - q \, dp$

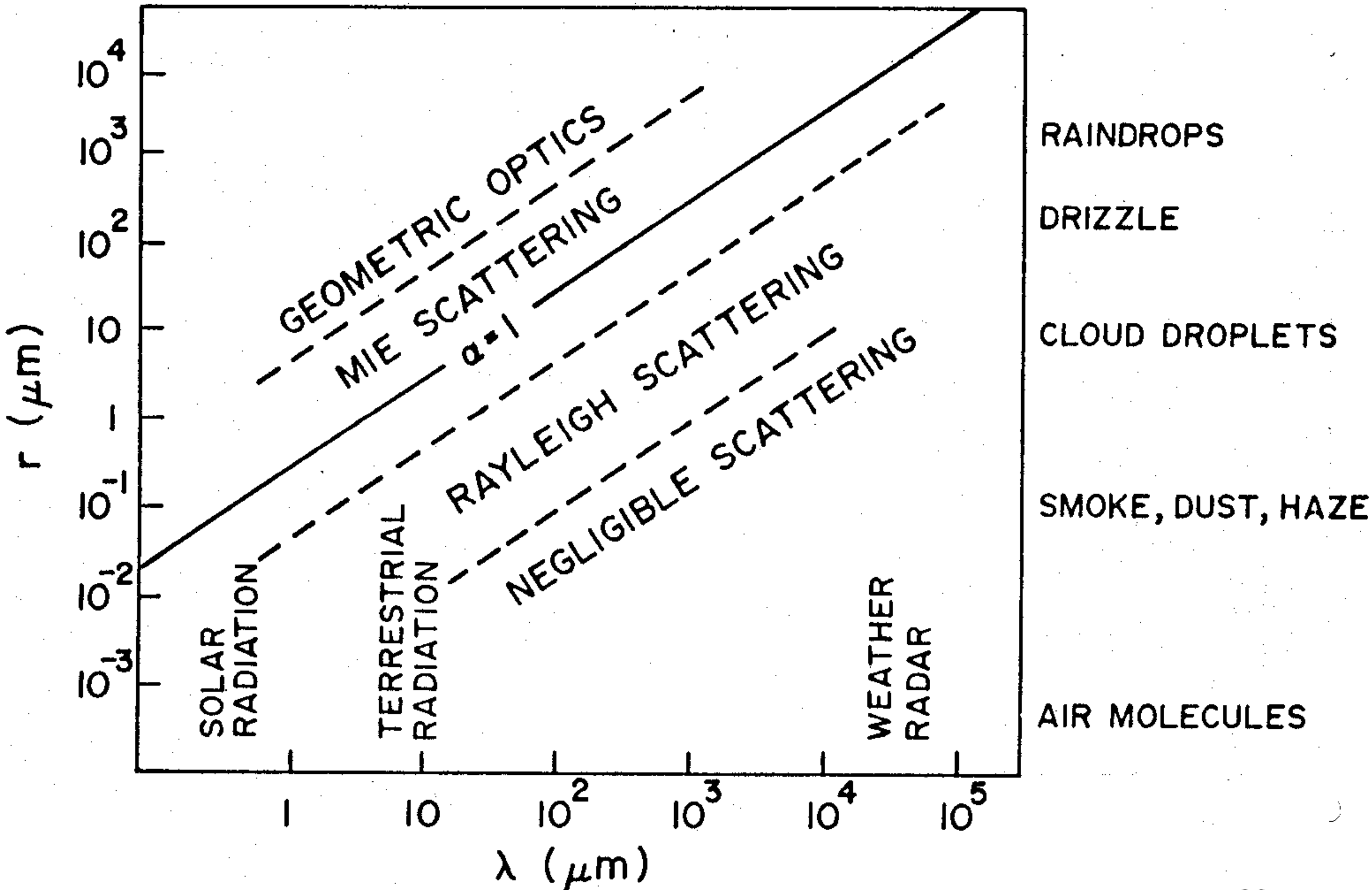
where q is the mixing ratio and ρ is the density of the atmosphere, then

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

Spectral Characteristics of Atmospheric Transmission and Sensing Systems







GEOMETRIC OPTICS
MIE SCATTERING
α=1
RAYLEIGH SCATTERING
NEGLIGIBLE SCATTERING
SOLAR RADIATION
TERRESTRIAL RADIATION
WEATHER RADAR
RAINDROPS
DRIZZLE
CLOUD DROPLETS
SMOKE, DUST, HAZE
AIR MOLECULES
r (μm)
λ (μm)

Aerosol Size Distribution

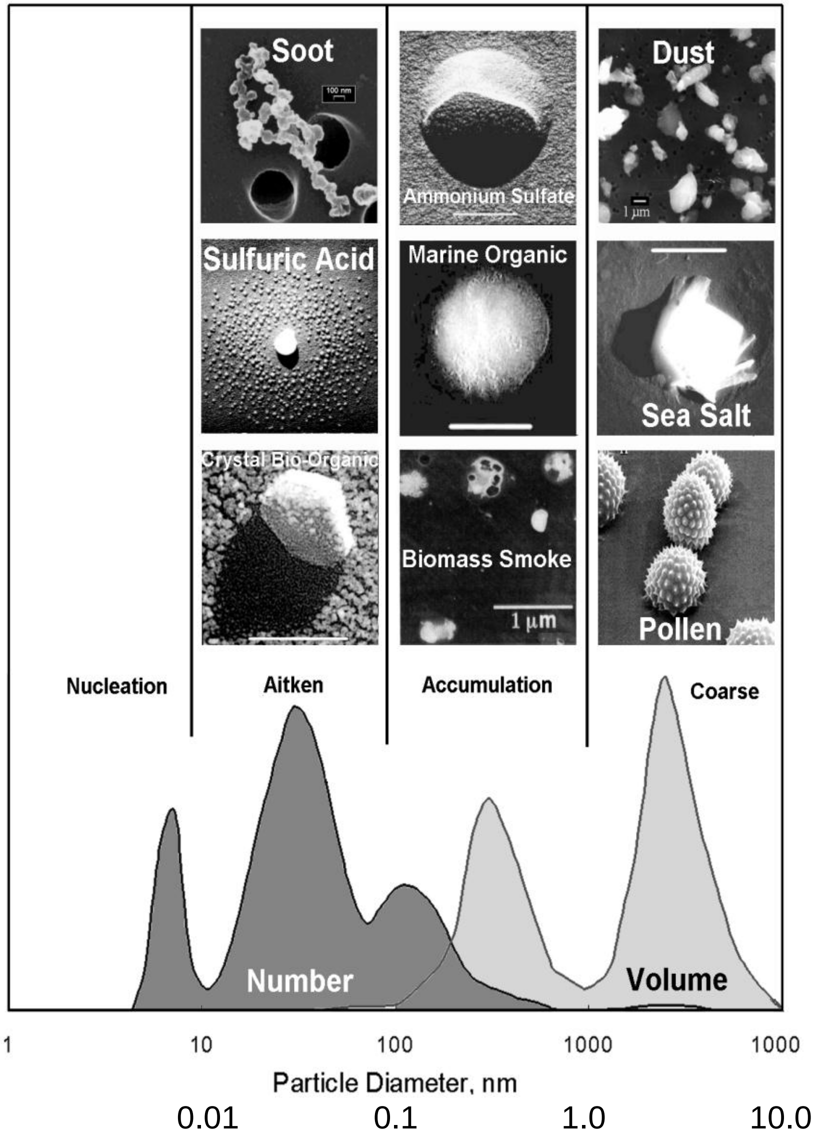
There are 3 modes :

- « **nucleation** »: radius is between 0.002 and 0.05 µm. They result from combustion processes, photo-chemical reactions, etc.

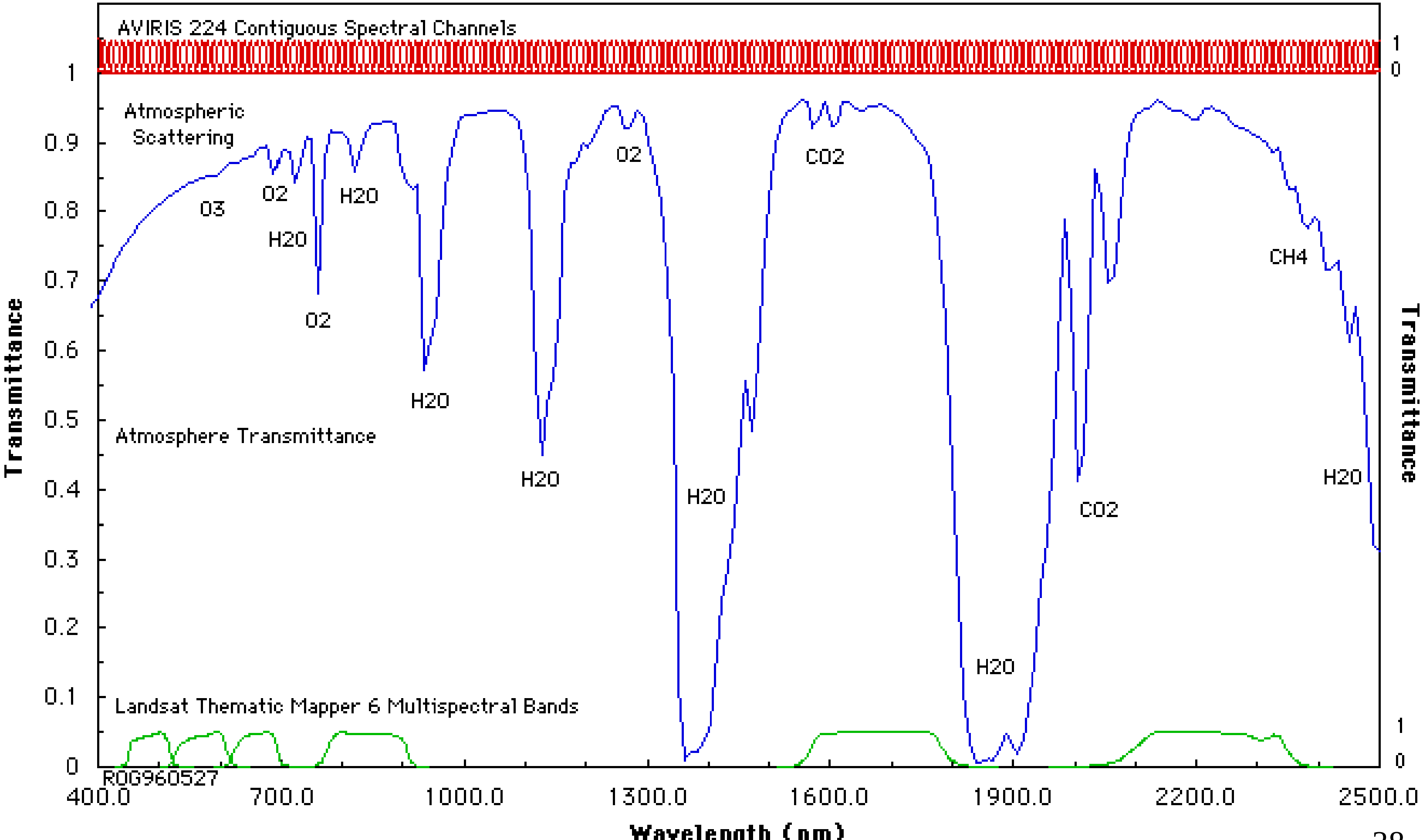
- **« accumulation »:** radius is between 0.05 µm and 0.5 µm**.** Coagulation processes.

- « **coarse** »: larger than 1 µm**.** From mechanical processes like aeolian erosion.

« fine » particles (nucleation and accumulation) result from anthropogenic activities, coarse particles come from natural processes.

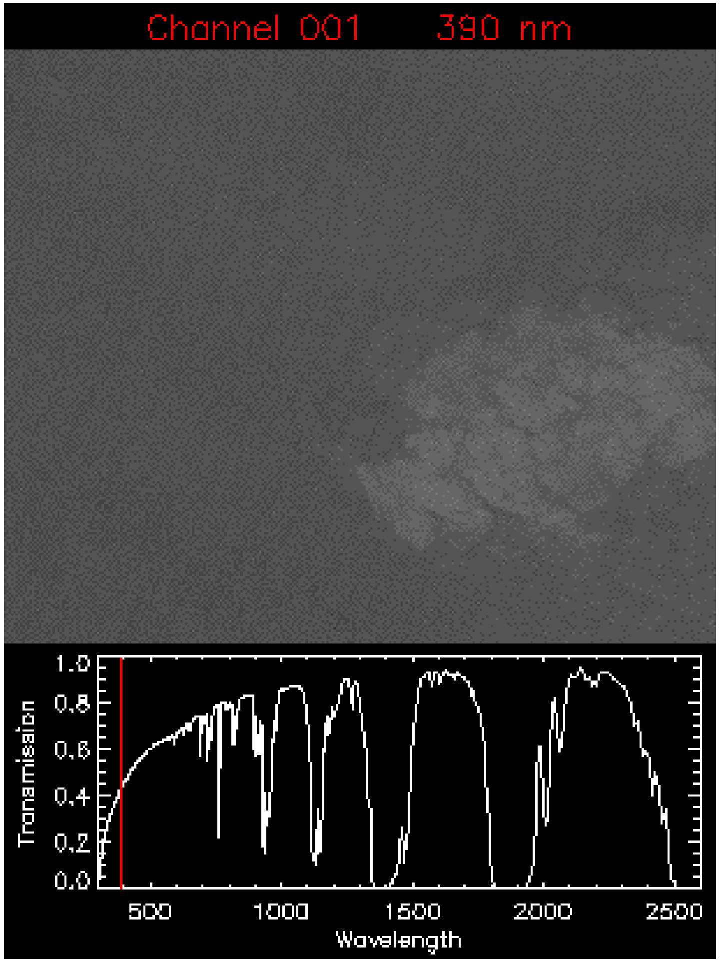


Measurements in the Solar Reflected Spectrum across the region covered by AVIRIS



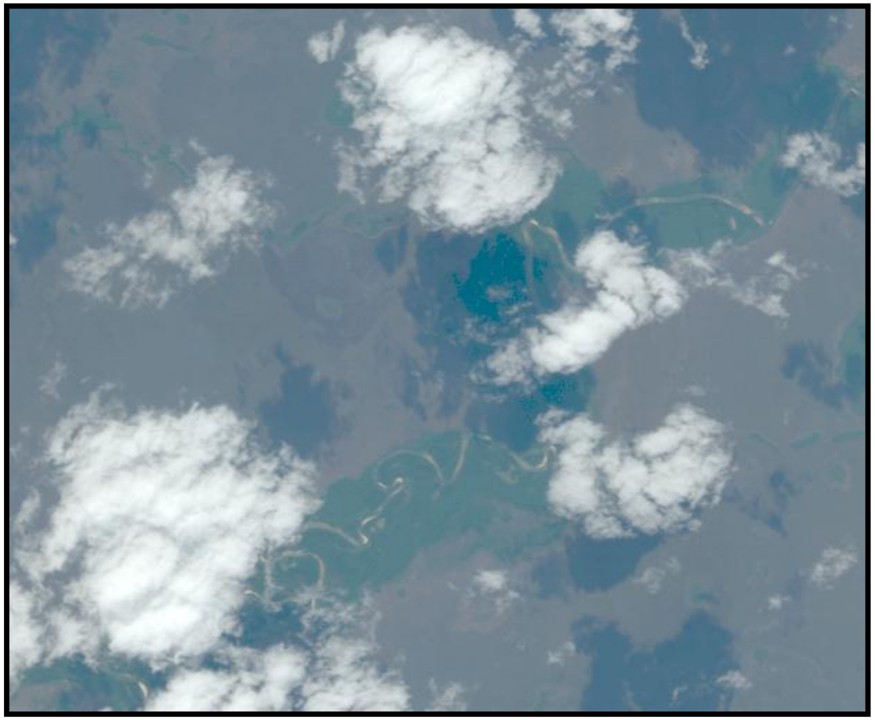
AVIRIS Movie #1

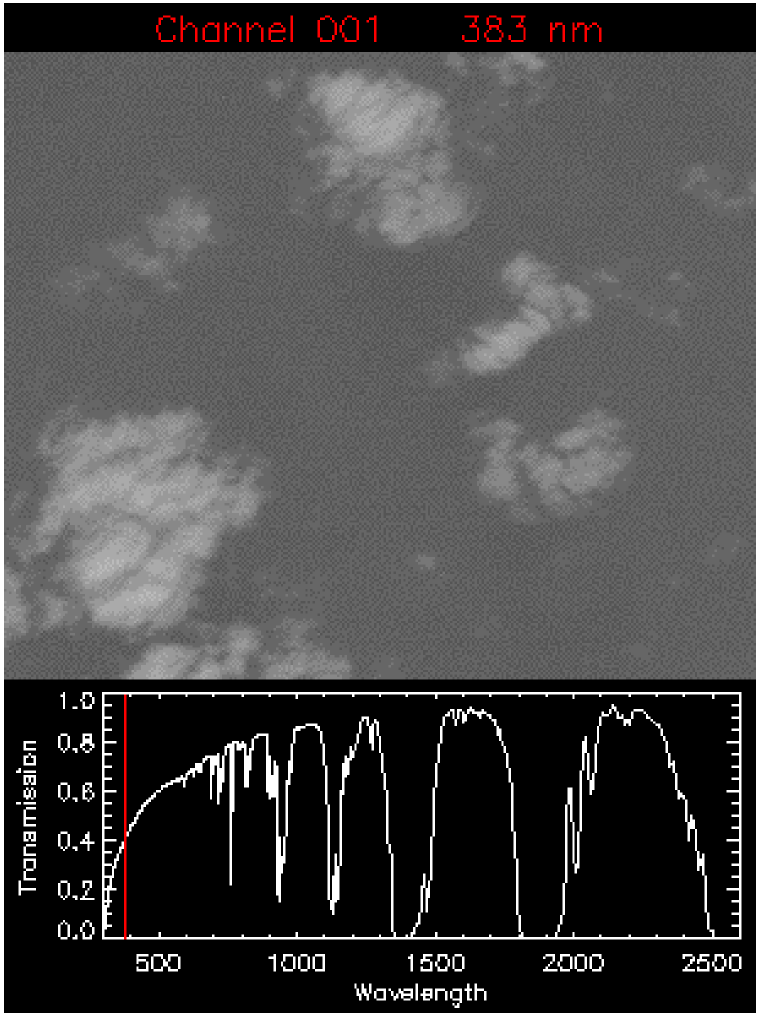
AVIRIS Image - Linden CA 20-Aug-1992
224 Spectral Bands: 0.4 - 2.5 μm
Pixel: 20m x 20m   Scene: 10km x 10km



AVIRIS Movie #2

AVIRIS Image - Porto Nacional, Brazil
20-Aug-1995
224 Spectral Bands: 0.4 - 2.5 μm
Pixel: 20m x 20m   Scene: 10km x 10km





Intro to Land-Ocean-Atmosphere Remote Sensing Lab

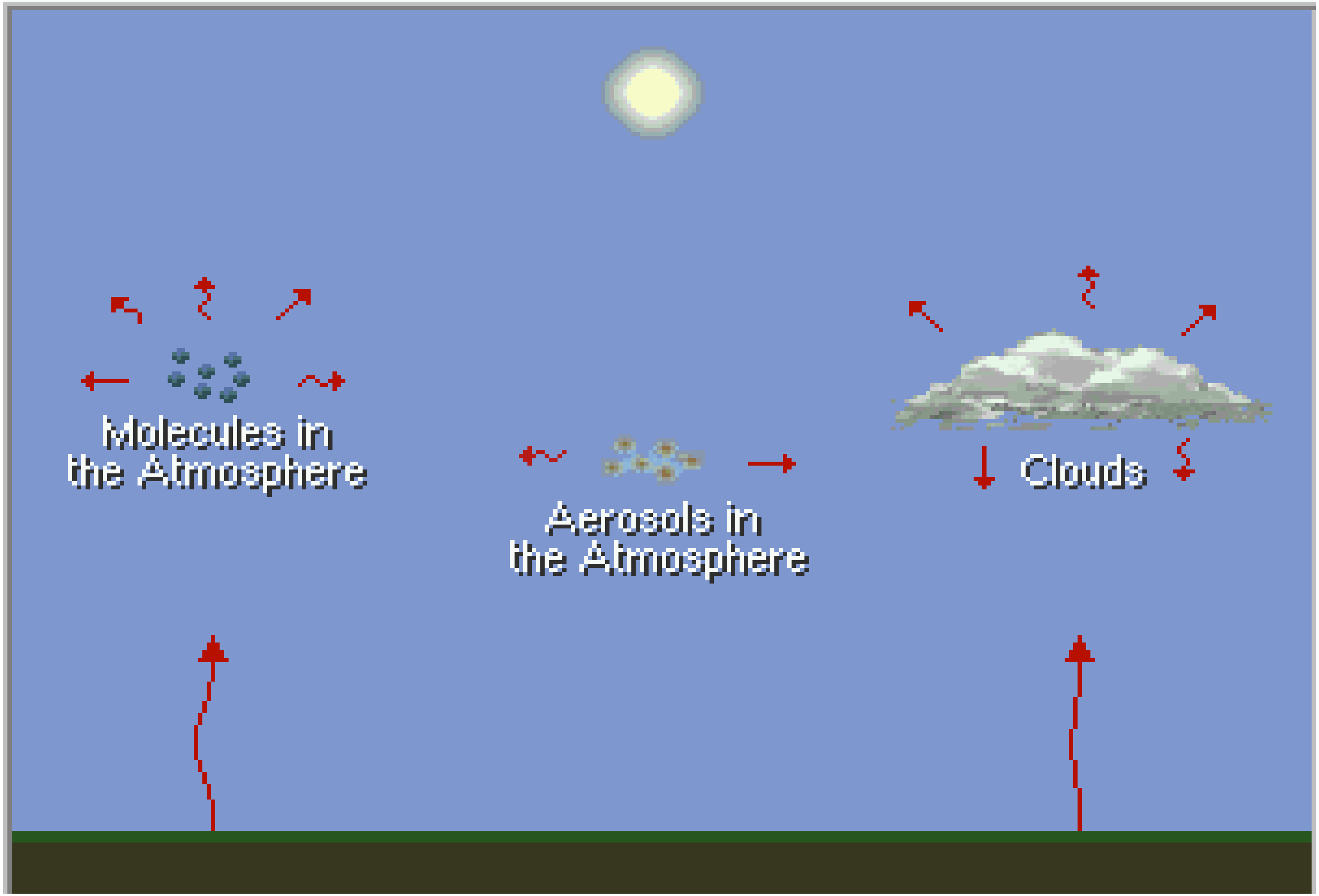
Lecture - Labs in Bologna & Potenza

May-June 2012

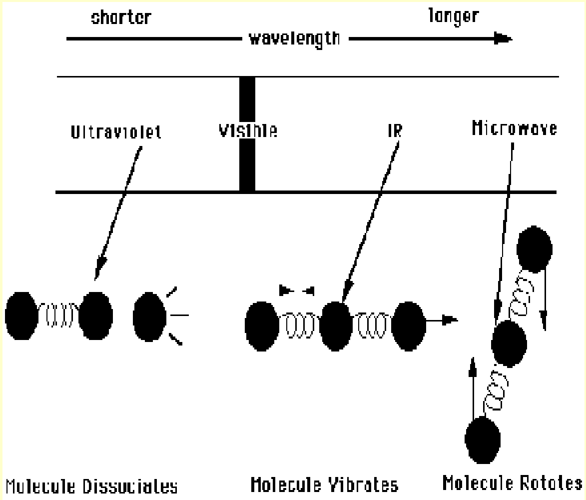
Paul Menzel
UW/CIMSS/AOS

Relevant Material in Applications of Meteorological Satellites

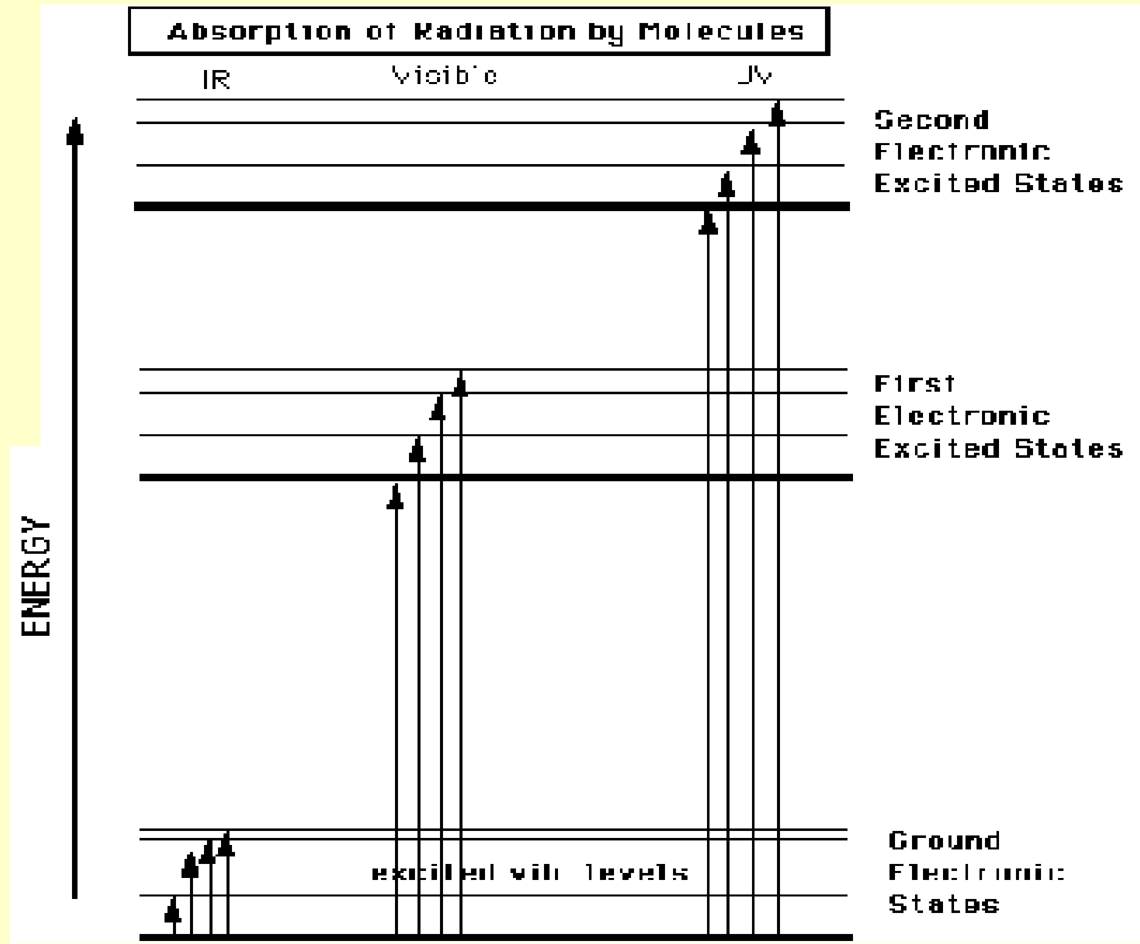
Re-emission of Infrared Radiation



Molecular Responses to Radiation

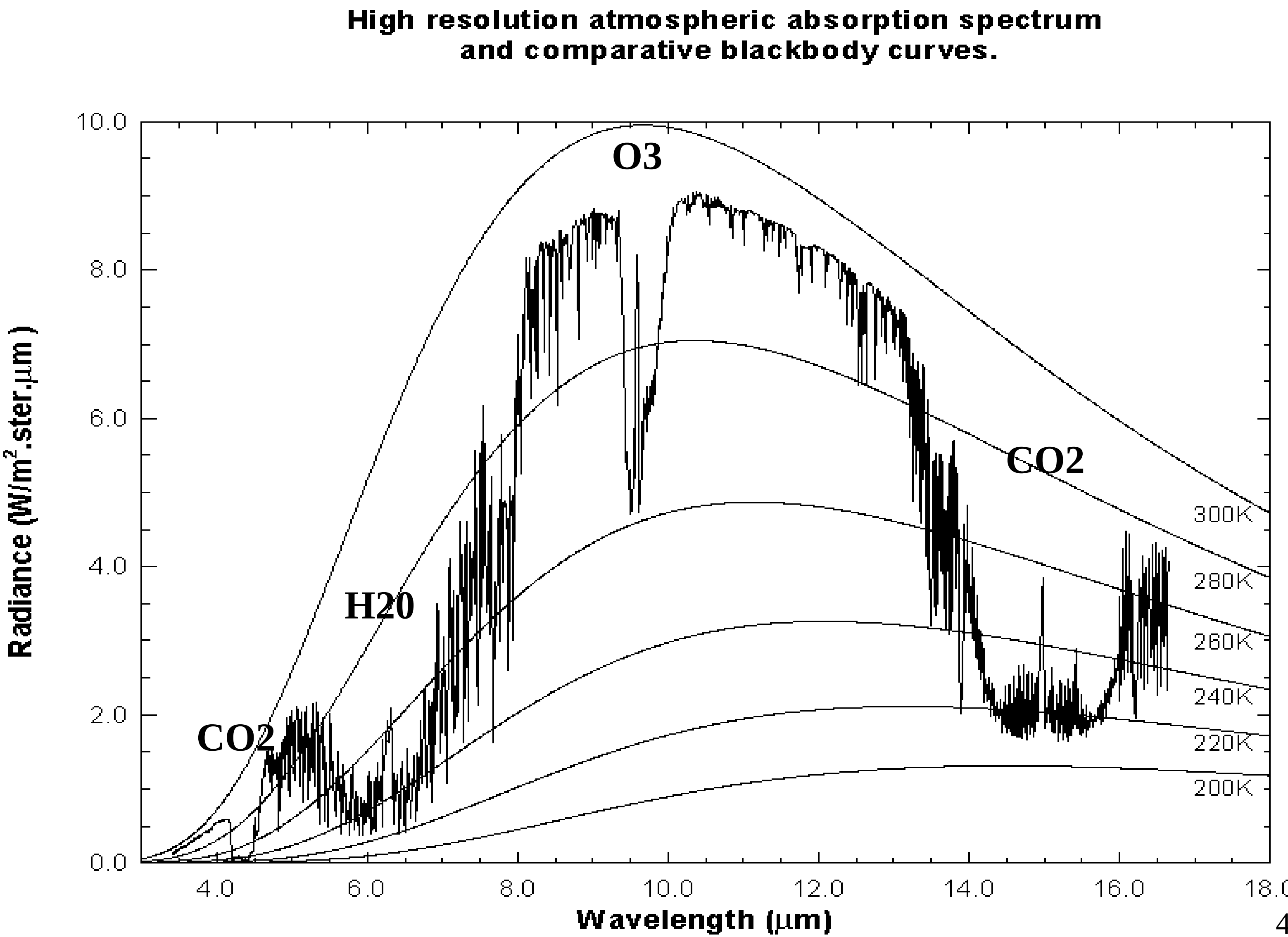


Molecular absorption of IR by vibrational and rotational excitation

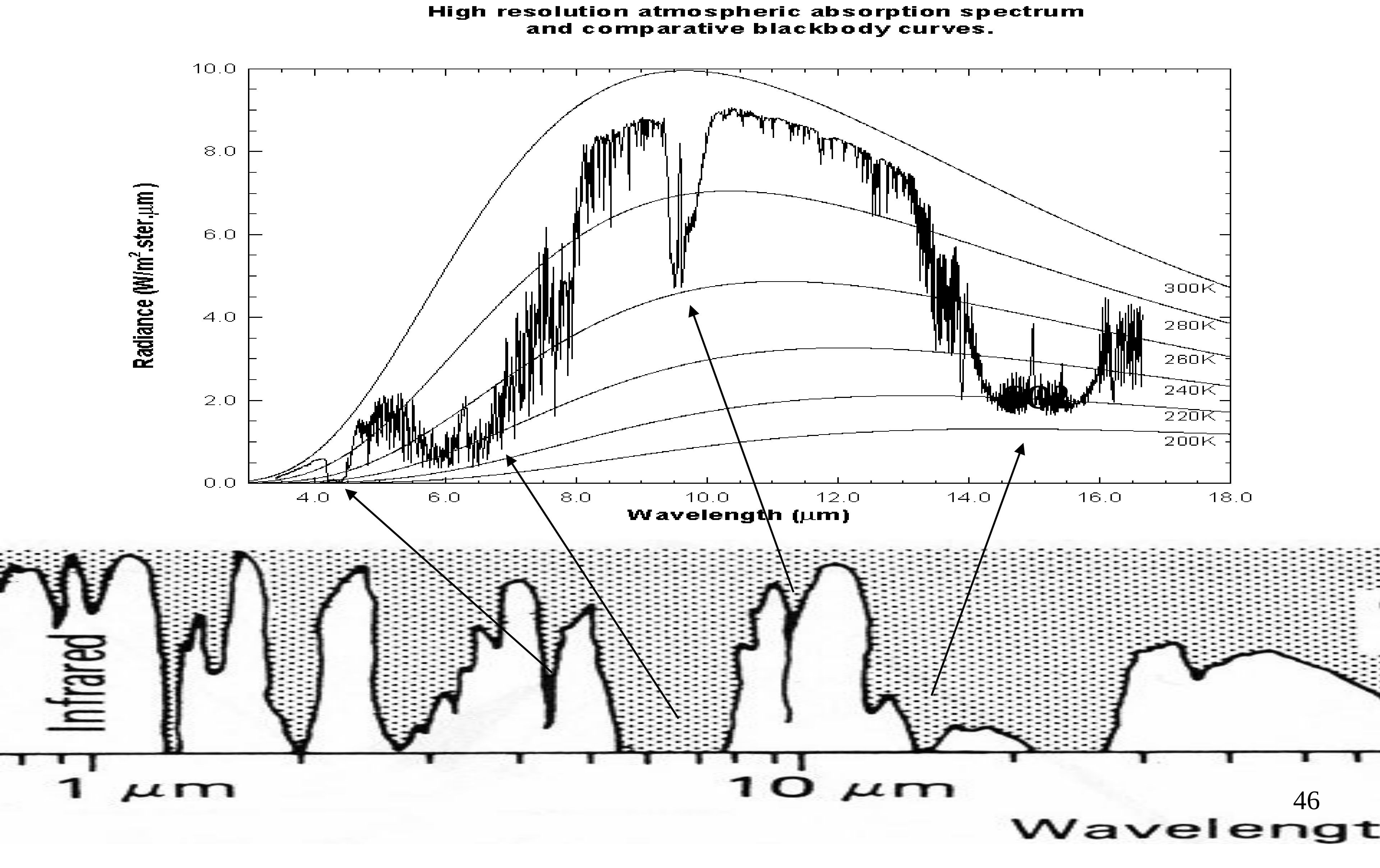


CO2, H2O, and O3

Earth emitted spectra overlaid on Planck function envelopes



Earth emitted spectra overlaid on Planck function envelopes



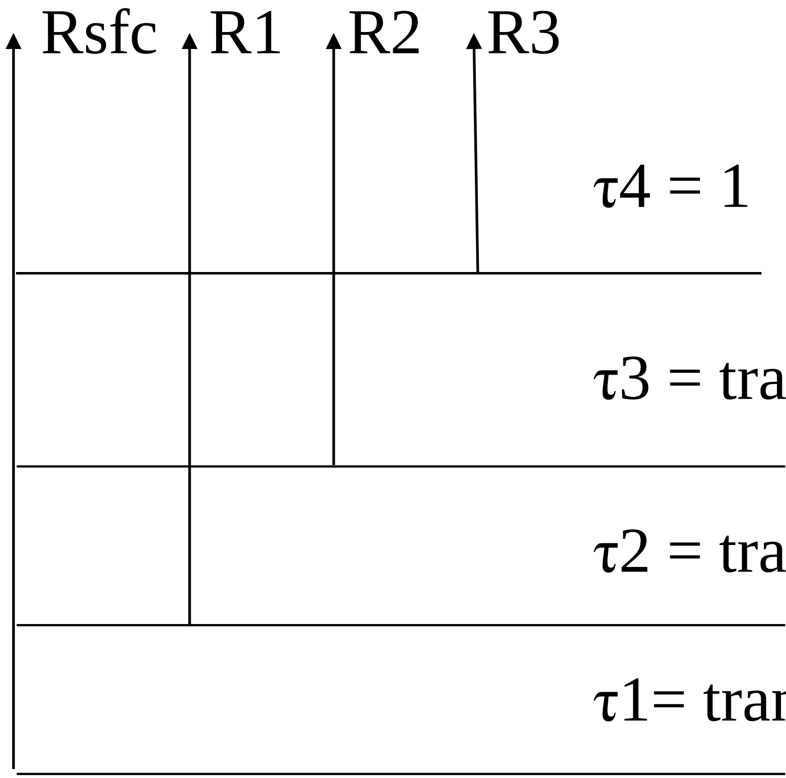
Radiative Transfer Equation

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.

Satellite observation comes from the sfc and the layers in the atm



$\tau 3$ = transmittance of upper layer of atm

$\tau 2$ = transmittance of middle layer of atm

$\tau 1$= transmittance of lower layer of atm

εsfc for earth surface

recalling that $\varepsilon i = 1 - \tau i$ for each layer, then

$$R_{obs} = \varepsilon sfc\ B_{sfc}\ \tau 1\ \tau 2\ \tau 3 + (1-\tau 1)\ B1\ \tau 2\ \tau 3 + (1-\tau 2)\ B2\ \tau 3 + (1-\tau 3)\ B3$$

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

The emission of an infinitesimal layer of the atmosphere at pressure p is equal to the absorption (1 - transmission). So,

$$\varepsilon_\lambda(layer)\, \tau_\lambda(layer\ to\ top) = [1 - \tau_\lambda(layer)]\, \tau_\lambda(layer\ to\ top)$$

Since transmission is multiplicative

$$\tau_\lambda(layer\ to\ top) - \tau_\lambda(layer)\, \tau_\lambda(layer\ to\ top) = -\Delta\tau_\lambda(layer\ to\ top)$$

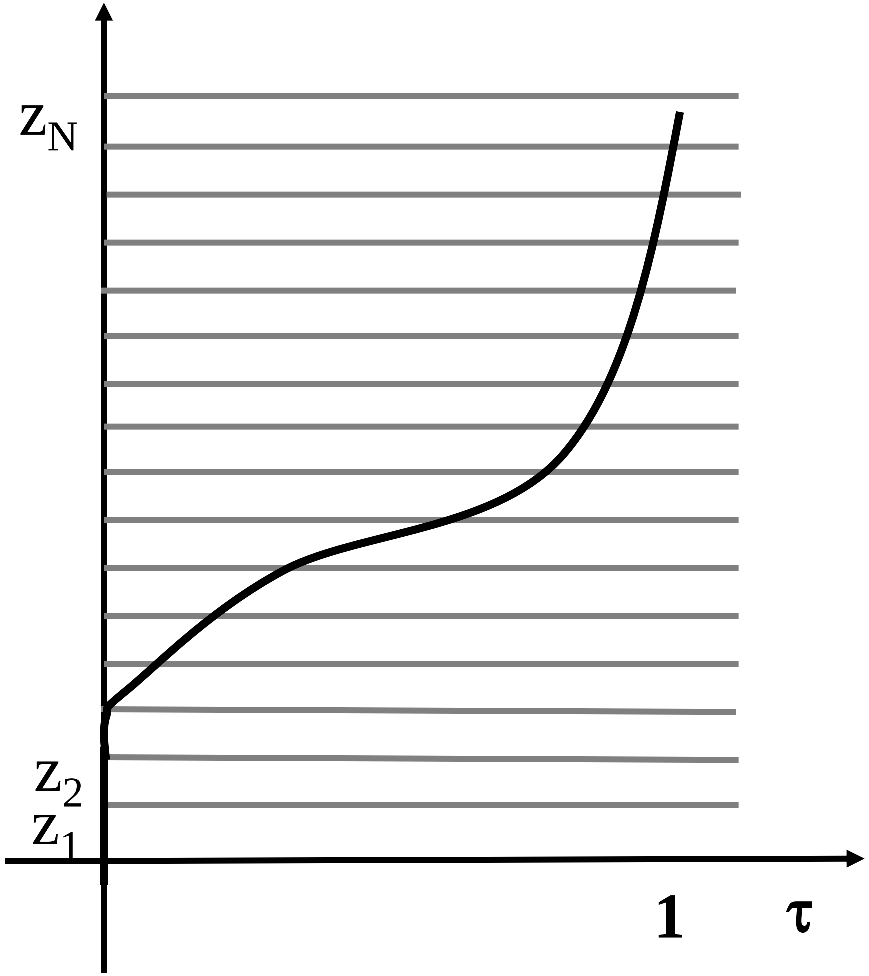
So we can write

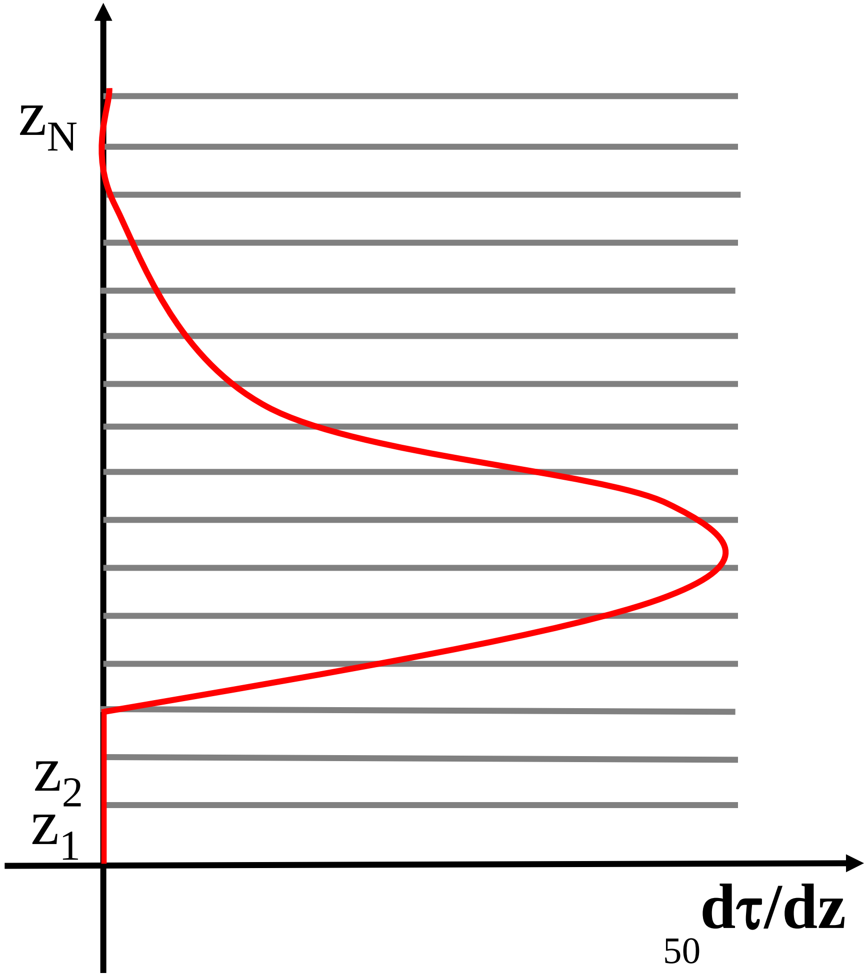
$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T(p_s))\, \tau_\lambda(p_s) - \sum_{p} B_\lambda(T(p))\, \Delta\tau_\lambda(p) .$$

which when written in integral form reads

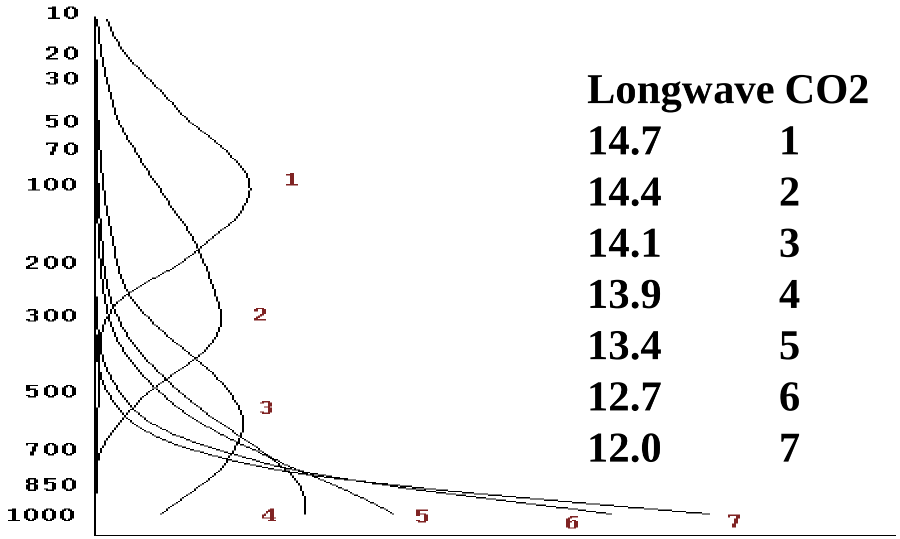
$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T(p_s))\, \tau_\lambda(p_s) - \int_{o}^{p_s} B_\lambda(T(p))\, [\, d\tau_\lambda(p) / dp \,]\, dp .$$

Weighting Functions



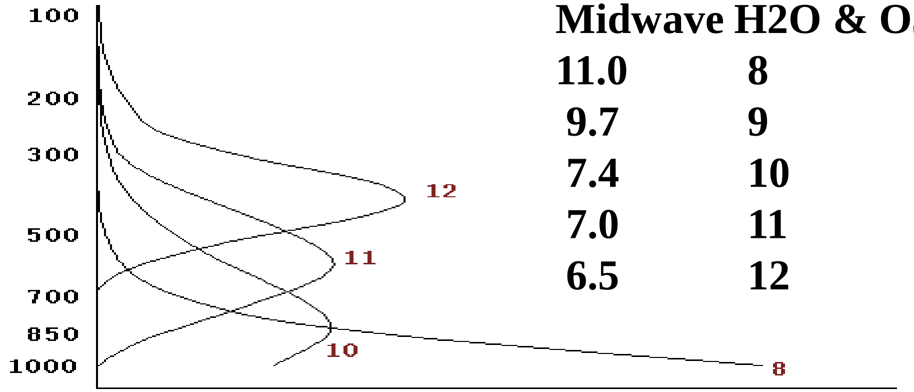


Weighting Functions



Longwave CO2

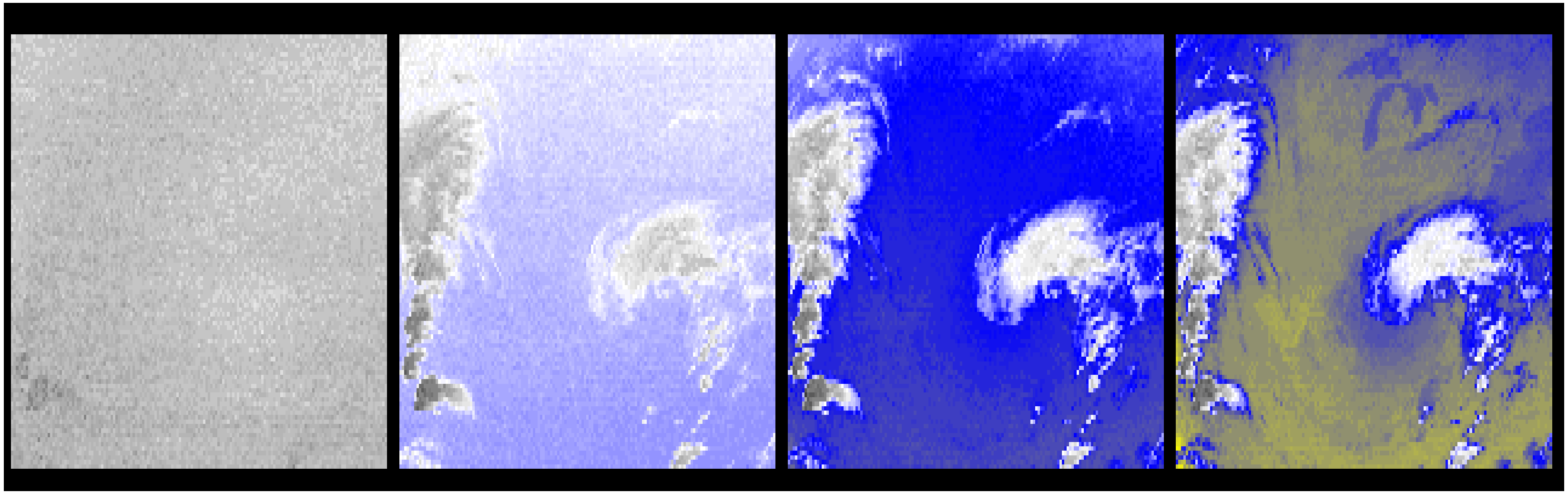
14.7	1	680	CO2, strat temp
14.4	2	696	CO2, strat temp
14.1	3	711	CO2, upper trop temp
13.9	4	733	CO2, mid trop temp
13.4	5	748	CO2, lower trop temp
12.7	6	790	H2O, lower trop moisture
12.0	7	832	H2O, dirty window



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

CO2 channels see to different levels in the atmosphere



14.2 um 13.9 um 13.6 um 13.3 um

Characteristics of RTE

* Radiance arises from deep and overlapping layers

* The radiance observations are not independent

* There is no unique relation between the spectrum of the outgoing radiance and T(p) or Q(p)

* T(p) is buried in an exponent in the denominator in the integral

* Q(p) is implicit in the transmittance

* Boundary conditions are necessary for a solution; the better the first guess the better the final solution

Profile Retrieval from Sounder Radiances

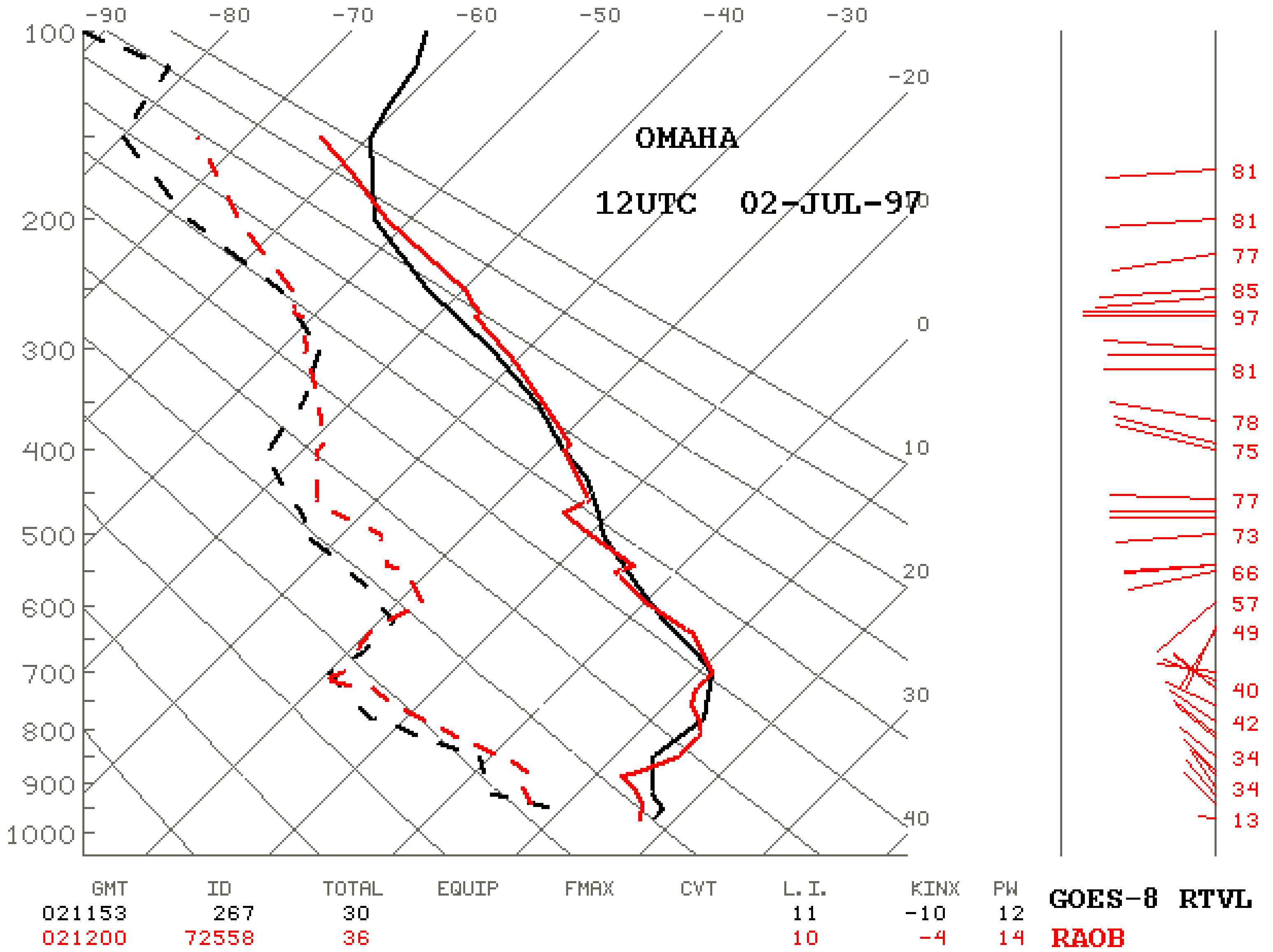
$$I_\lambda = \varepsilon_\lambda^{sfc} B_\lambda(T(p_s)) \tau_\lambda(p_s) - \int_o^{p_s} B_\lambda(T(p)) F_\lambda(p) [d\tau_\lambda(p) / dp] dp .$$

I1, I2, I3, ... , In are measured with the sounding radiometer
P(sfc) and T(sfc) come from ground based conventional observations
$\tau_\lambda(p)$ are calculated with physics models (using for CO2 and O3)
$\varepsilon_\lambda^{sfc}$ is estimated from a priori information (or regression guess)

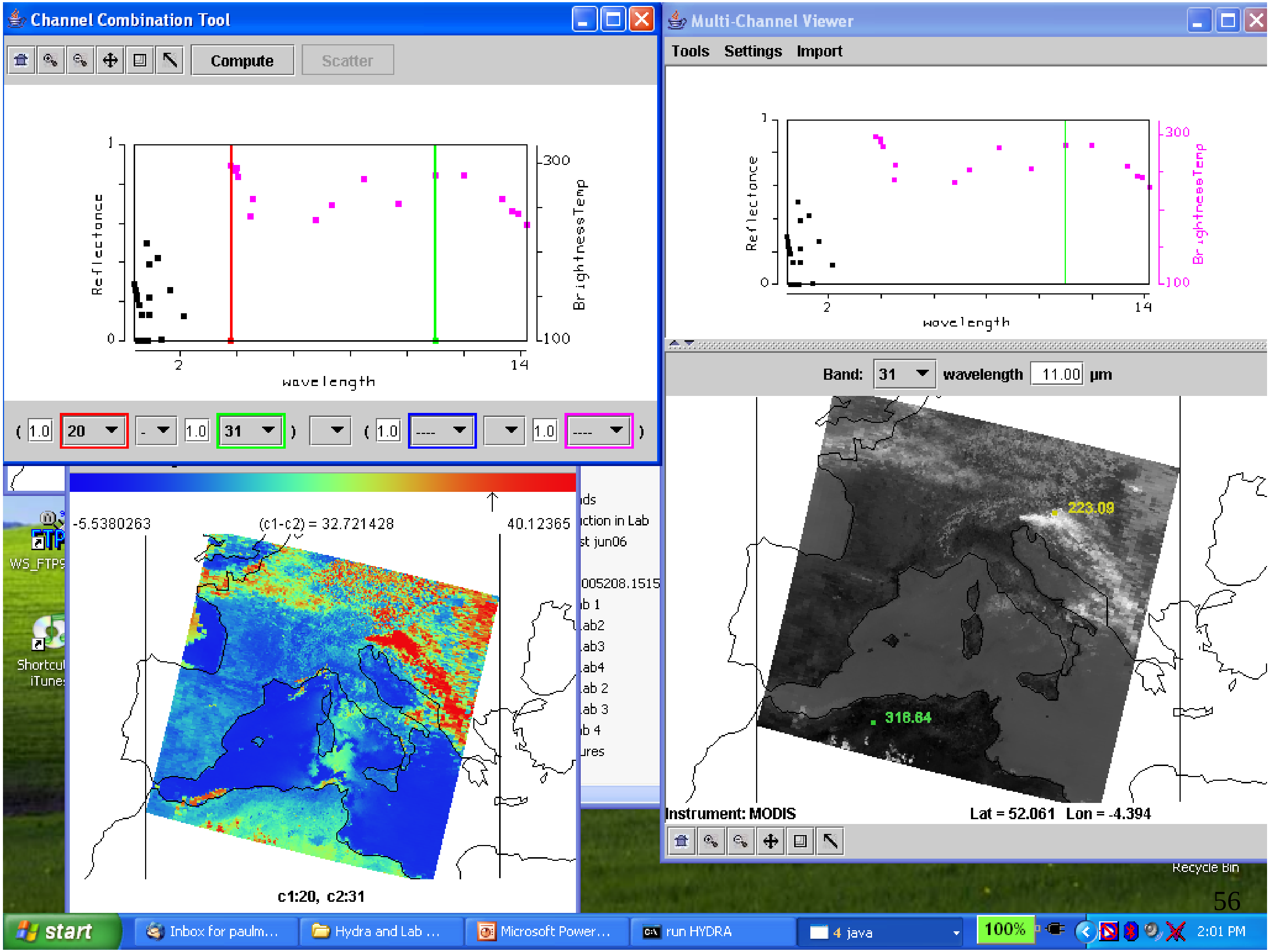
First guess solution is inferred from (1) in situ radiosonde reports,
(2) model prediction, or (3) blending of (1) and (2)

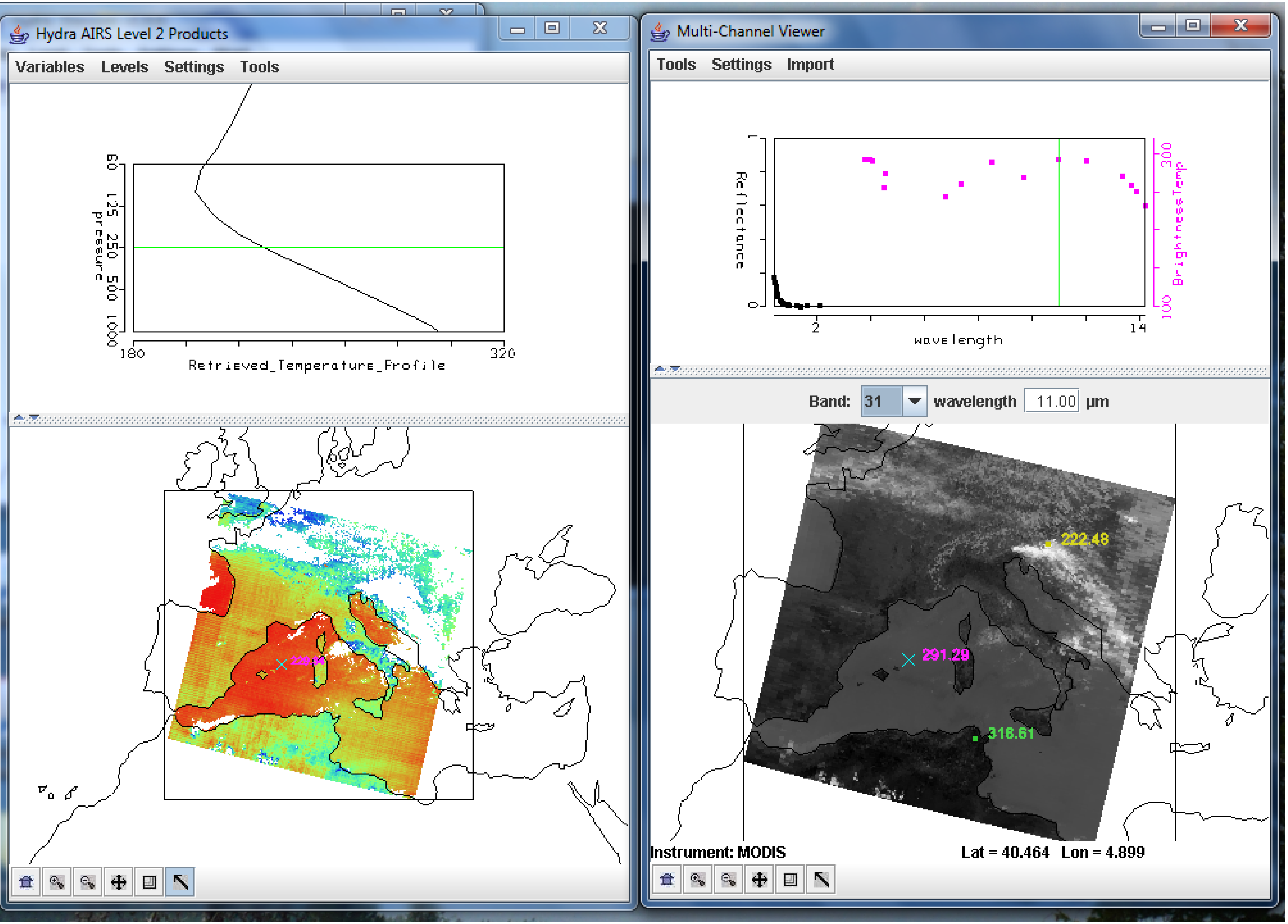
Profile retrieval from perturbing guess to match measured sounder radiances

Example Sounding



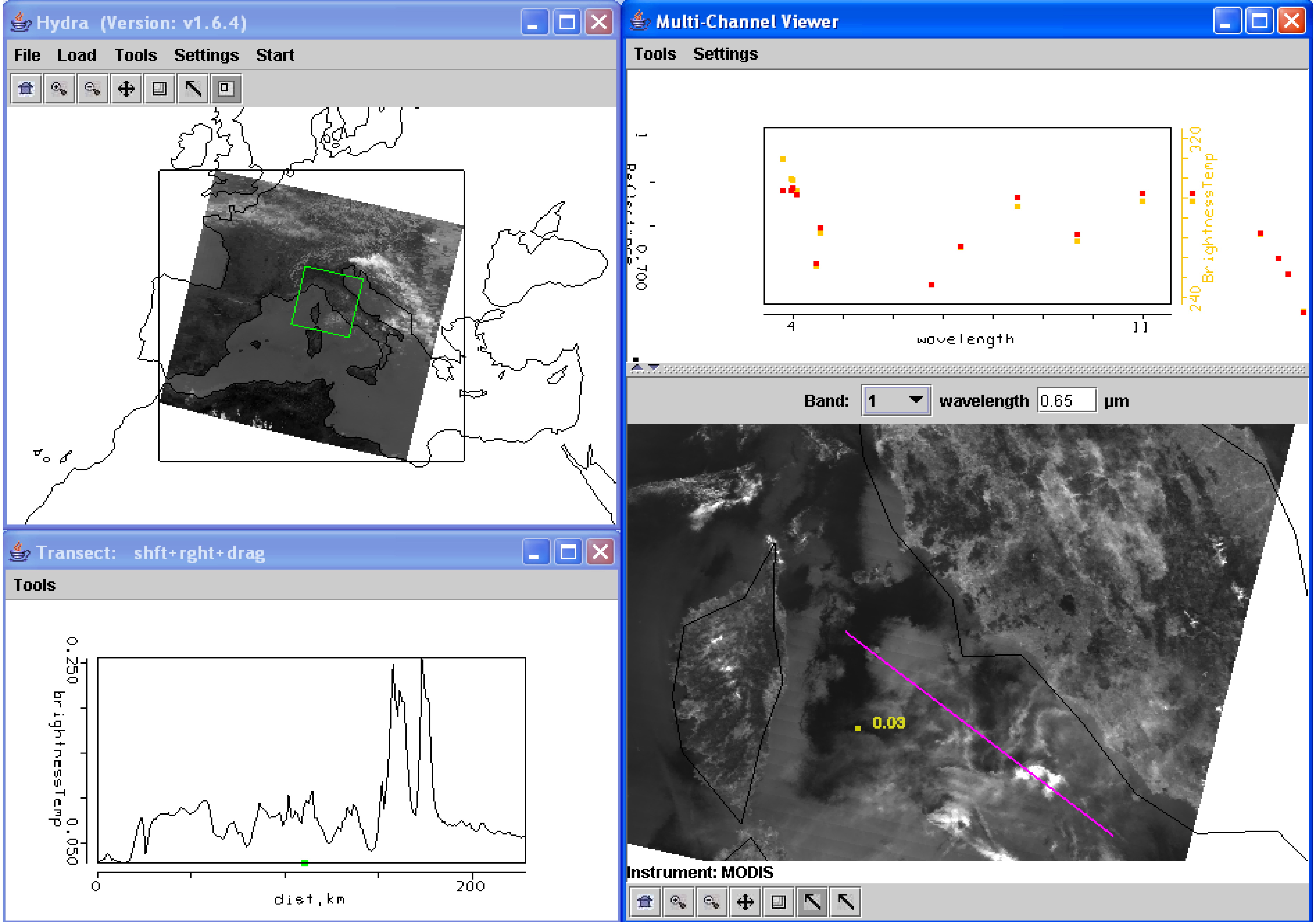
Viewing remote sensing data with HYDRA



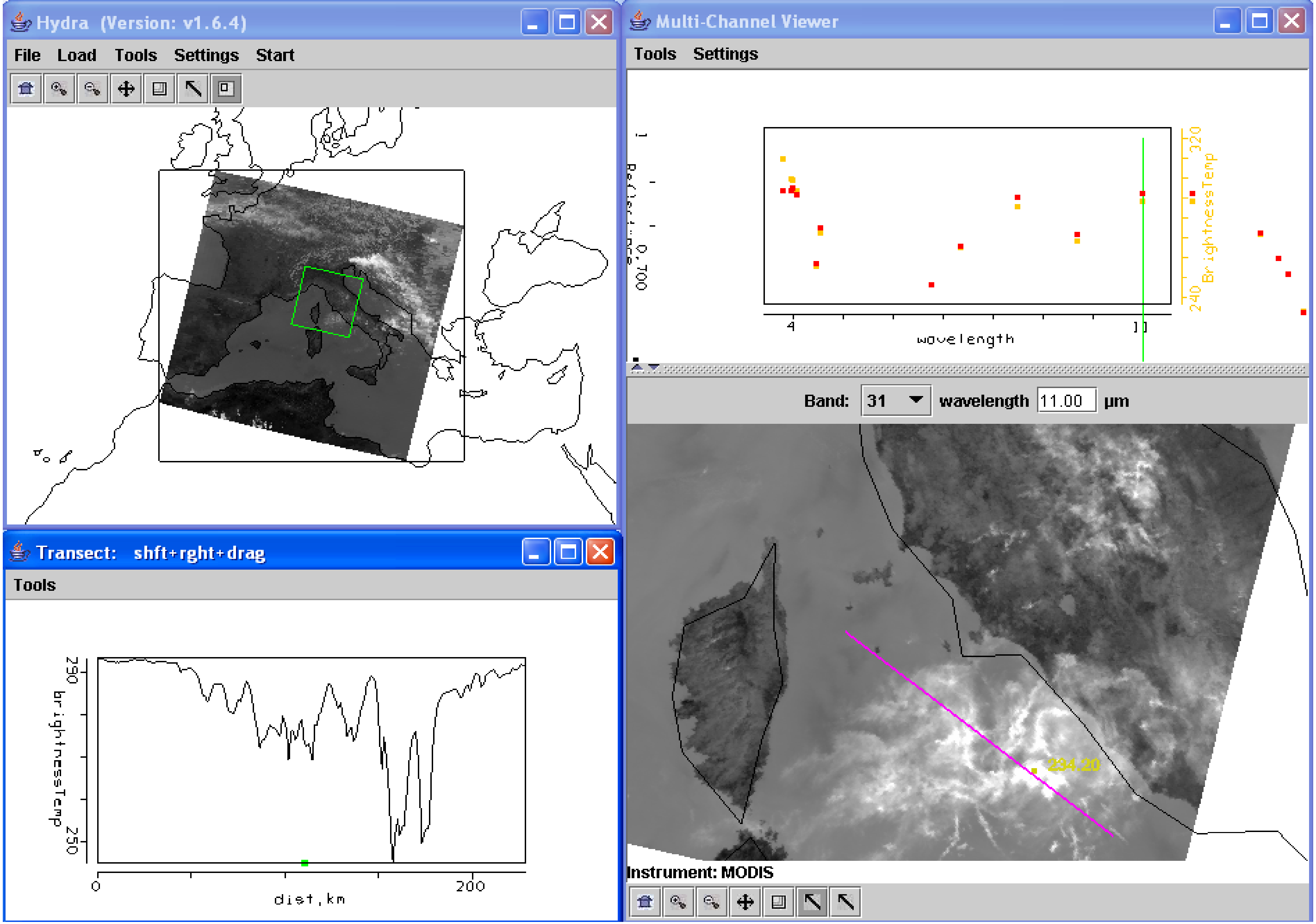


Hydra AIRS Level 2 Products
Variables
Levels
Settings
Tools
pressure
60
125
250
500
1000
180
320
Retrieved_Temperature_Profile
Multi-Channel Viewer
Tools
Settings
Import
Reflectance
BrightnessTemp
300
100
0
2
14
wavelength
Band:
31
wavelength
11.00
µm
222.48
291.29
318.61
Instrument: MODIS
Lat = 40.464
Lon = 4.899

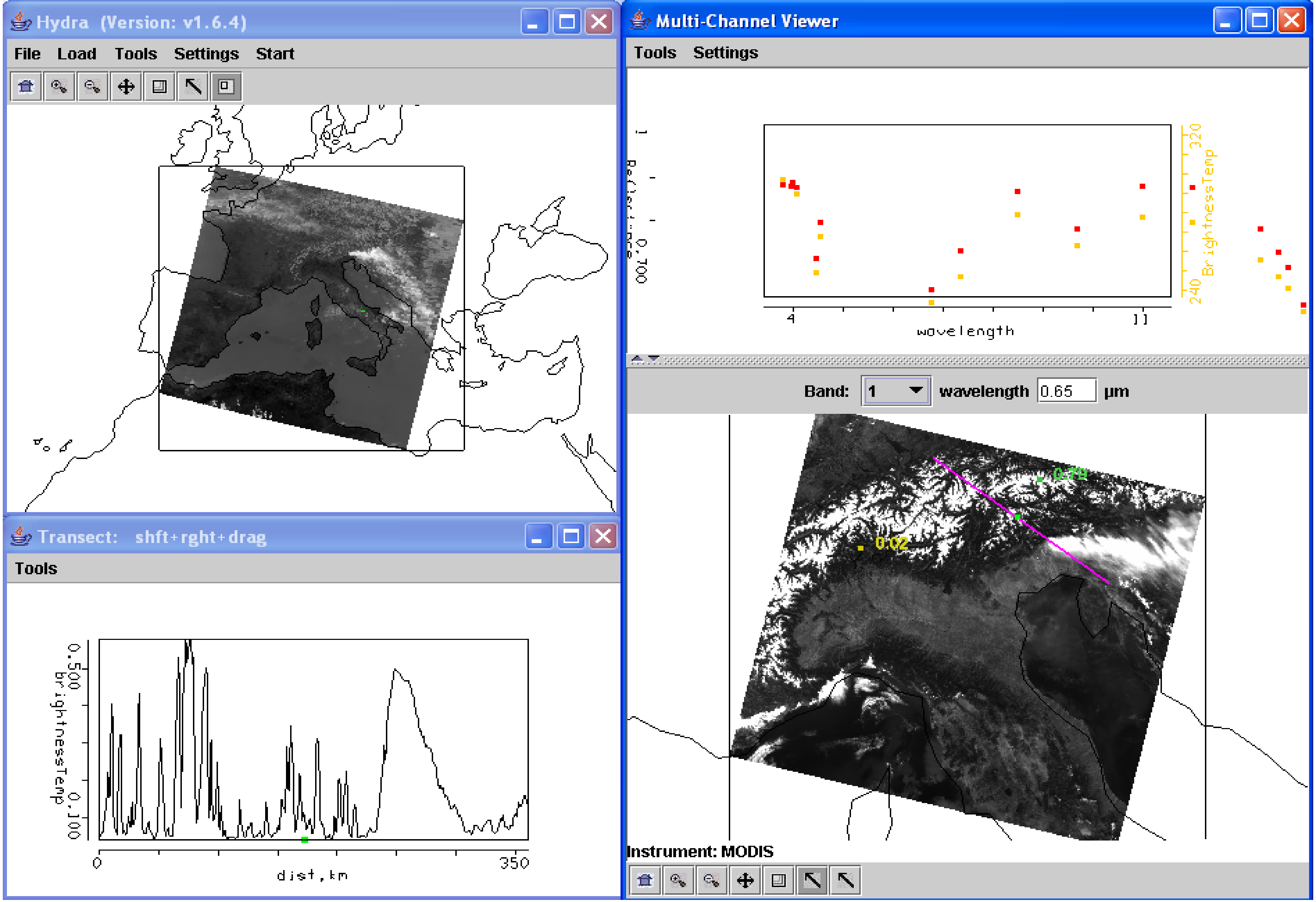
Relevant Material in Applications of Meteorological Satellites



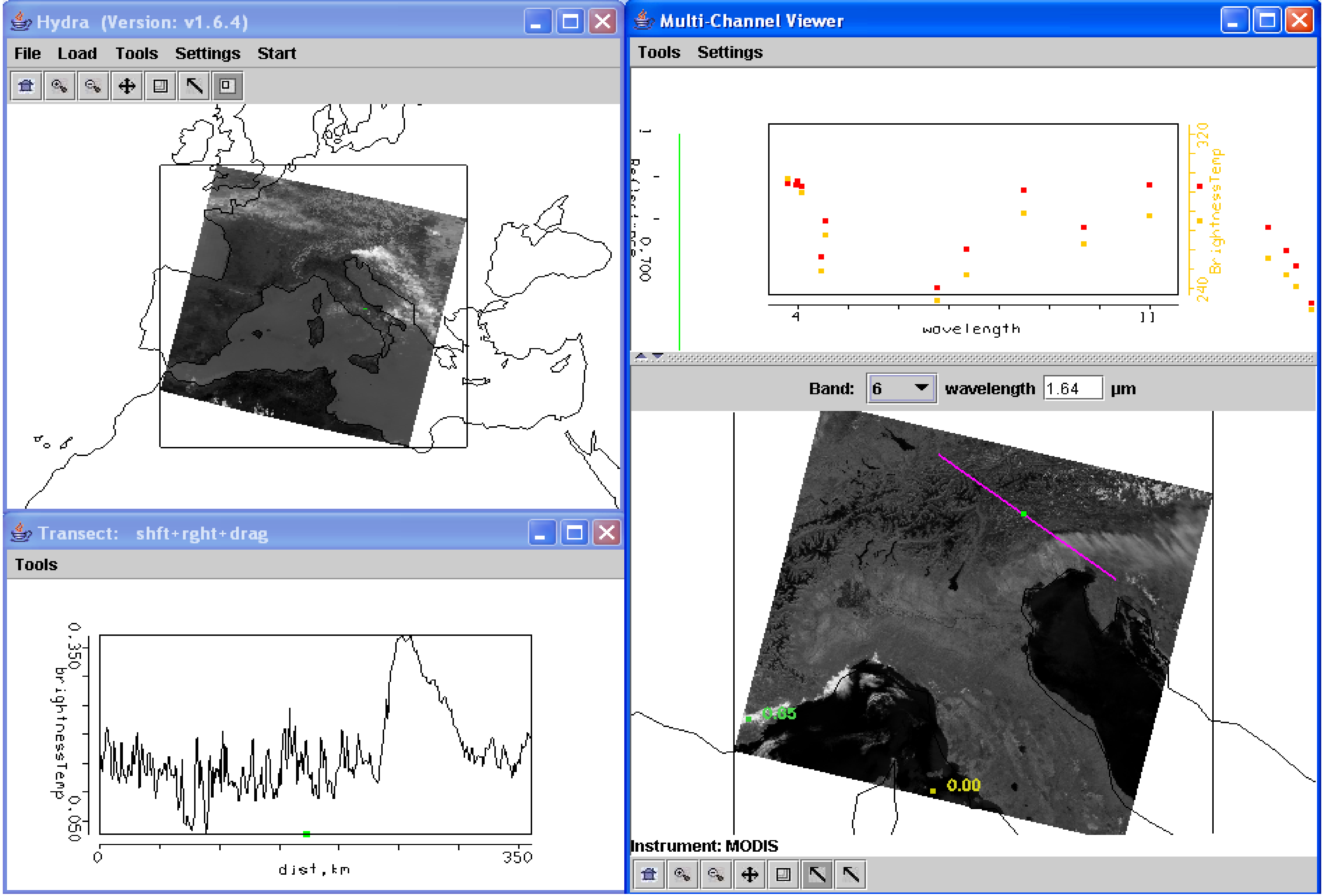
High clouds reflect more than surface at 0.65 µm



High clouds, cooler than surface, create lower 11 µm BTs

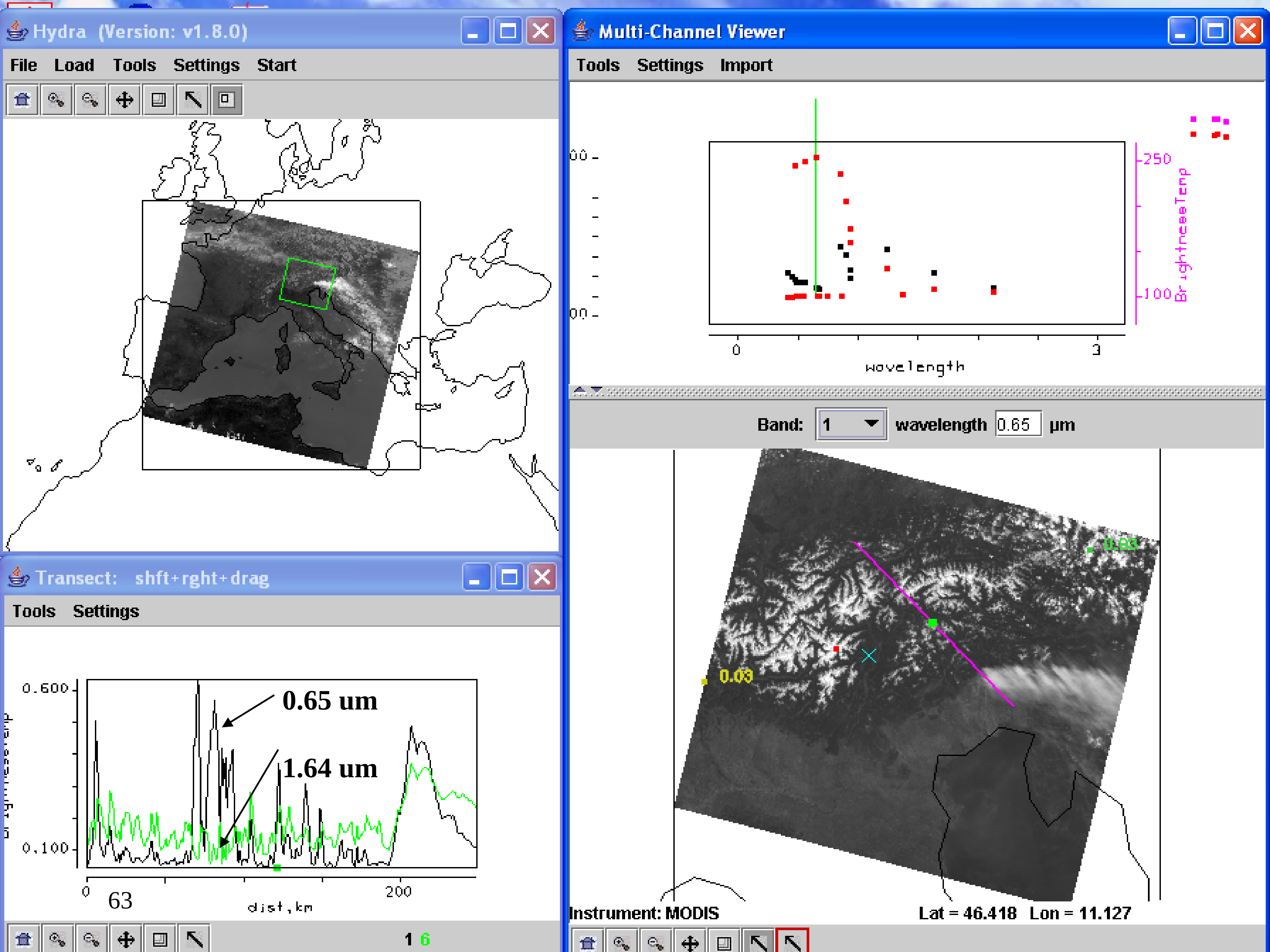


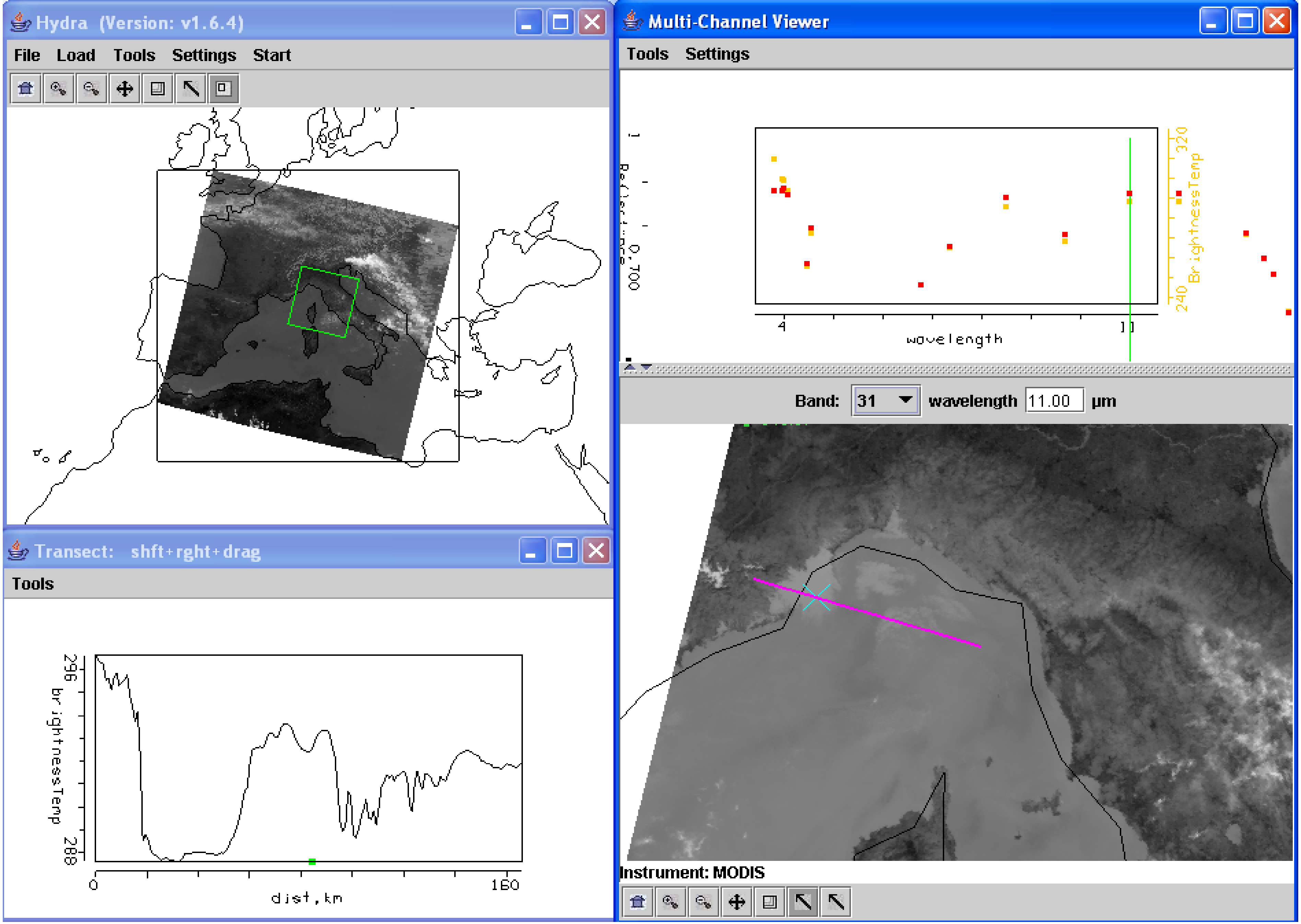
High clouds and snow both reflect a lot at 0.65 µm



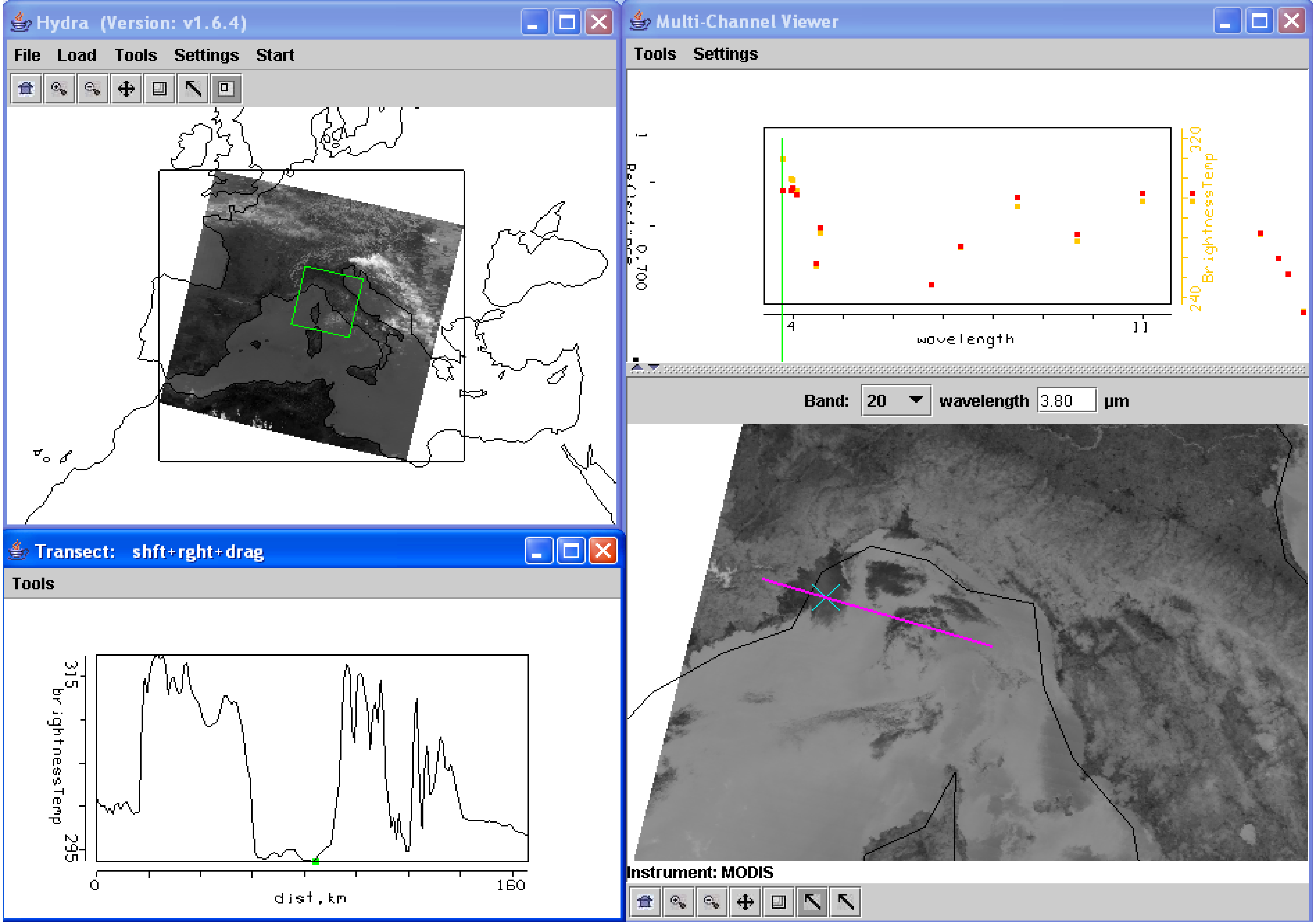
62

High clouds reflect but snow doesn't at 1.64 µm

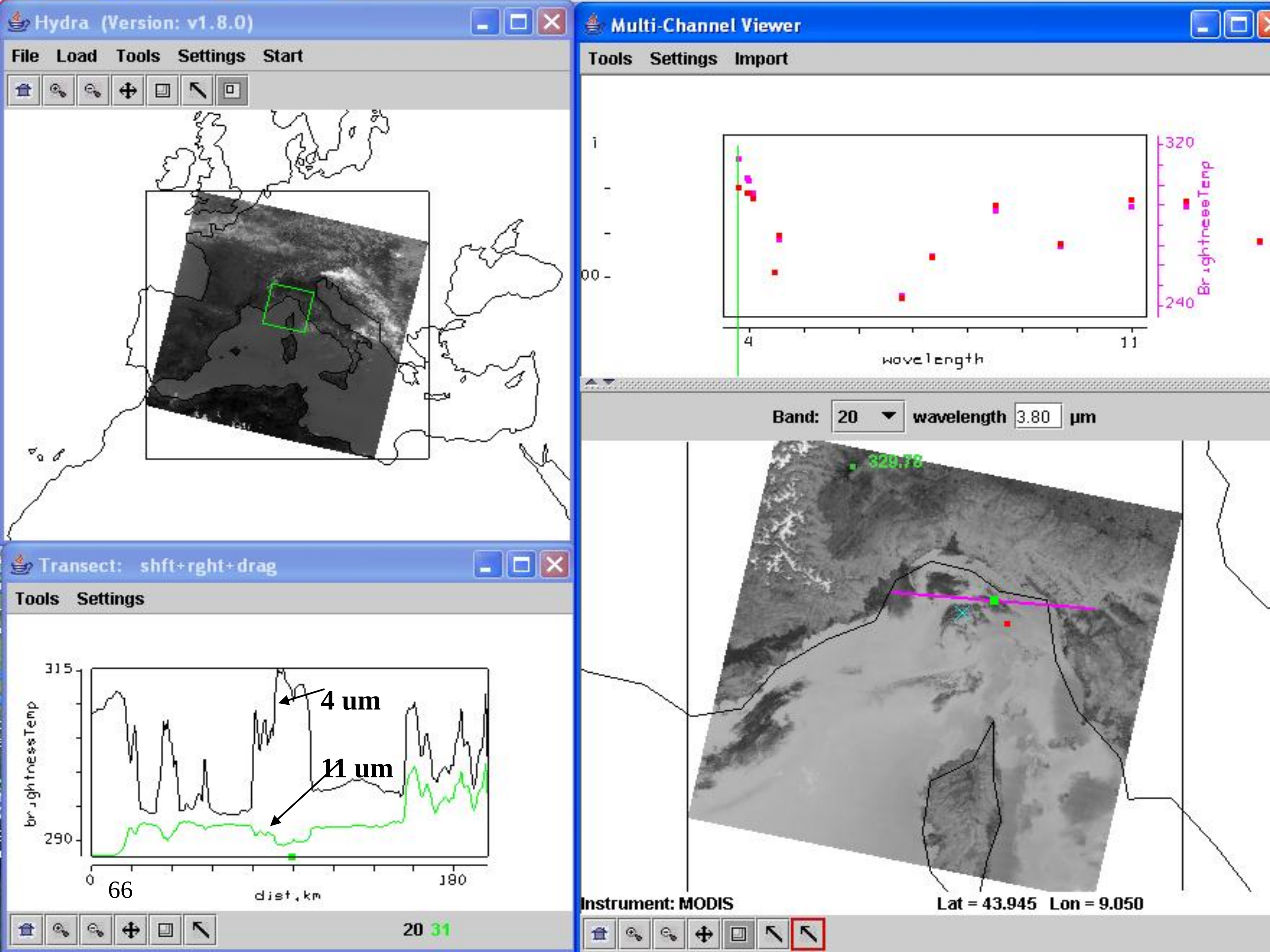

Hydra (Version: v1.8.0)
File Load Tools Settings Start
Multi-Channel Viewer
Tools Settings Import
250
100
BrightnessTemp
0
3
wavelength
Band: 1 wavelength 0.65 µm
0.03
0.03
Instrument: MODIS
Lat = 46.418 Lon = 11.127
Transect: shft+rght+drag
Tools Settings
0.600
0.100
0.65 um
1.64 um
0
200
dist,km
1 6

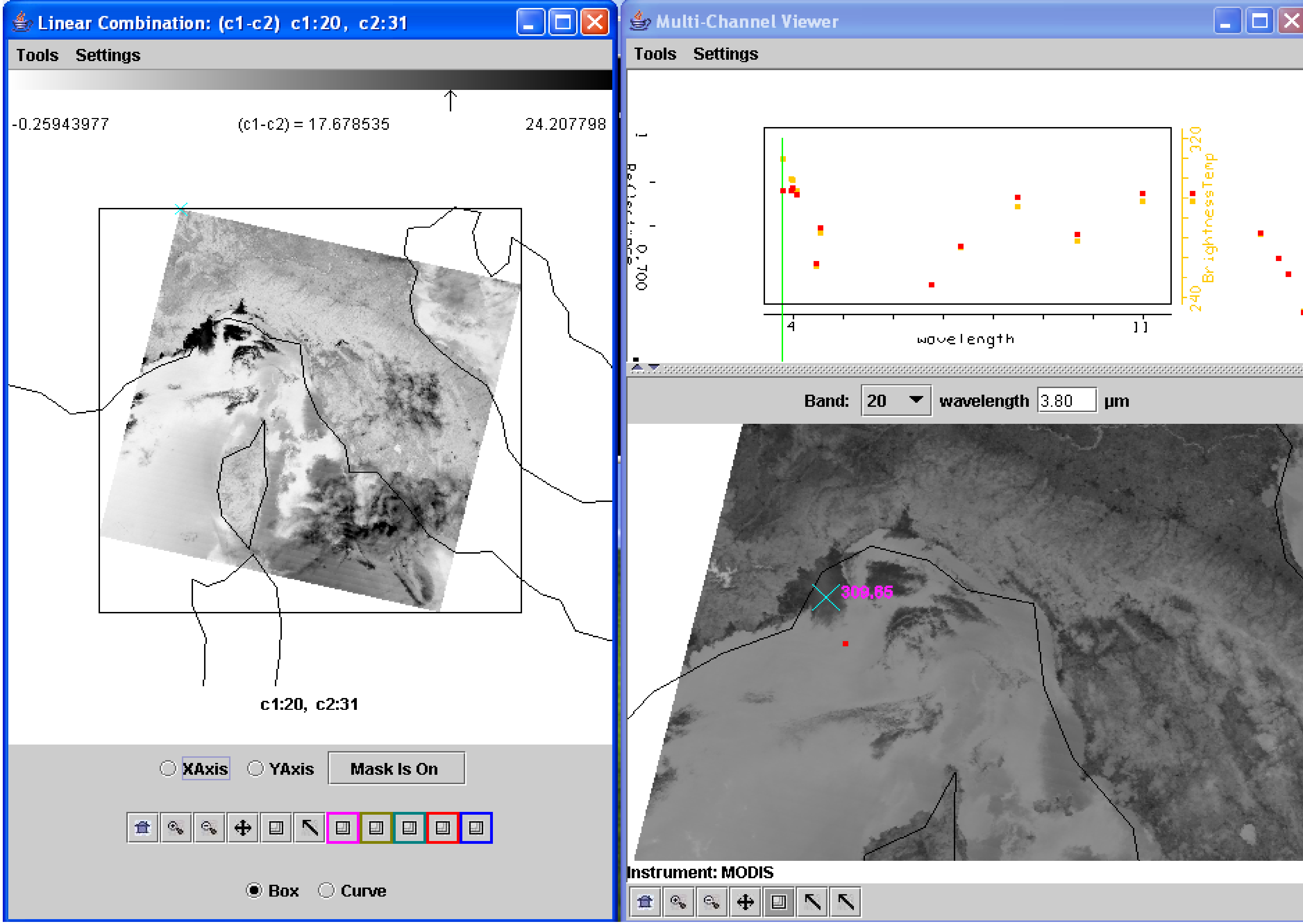


Low clouds, cooler than surface, create lower 11 µm BTs



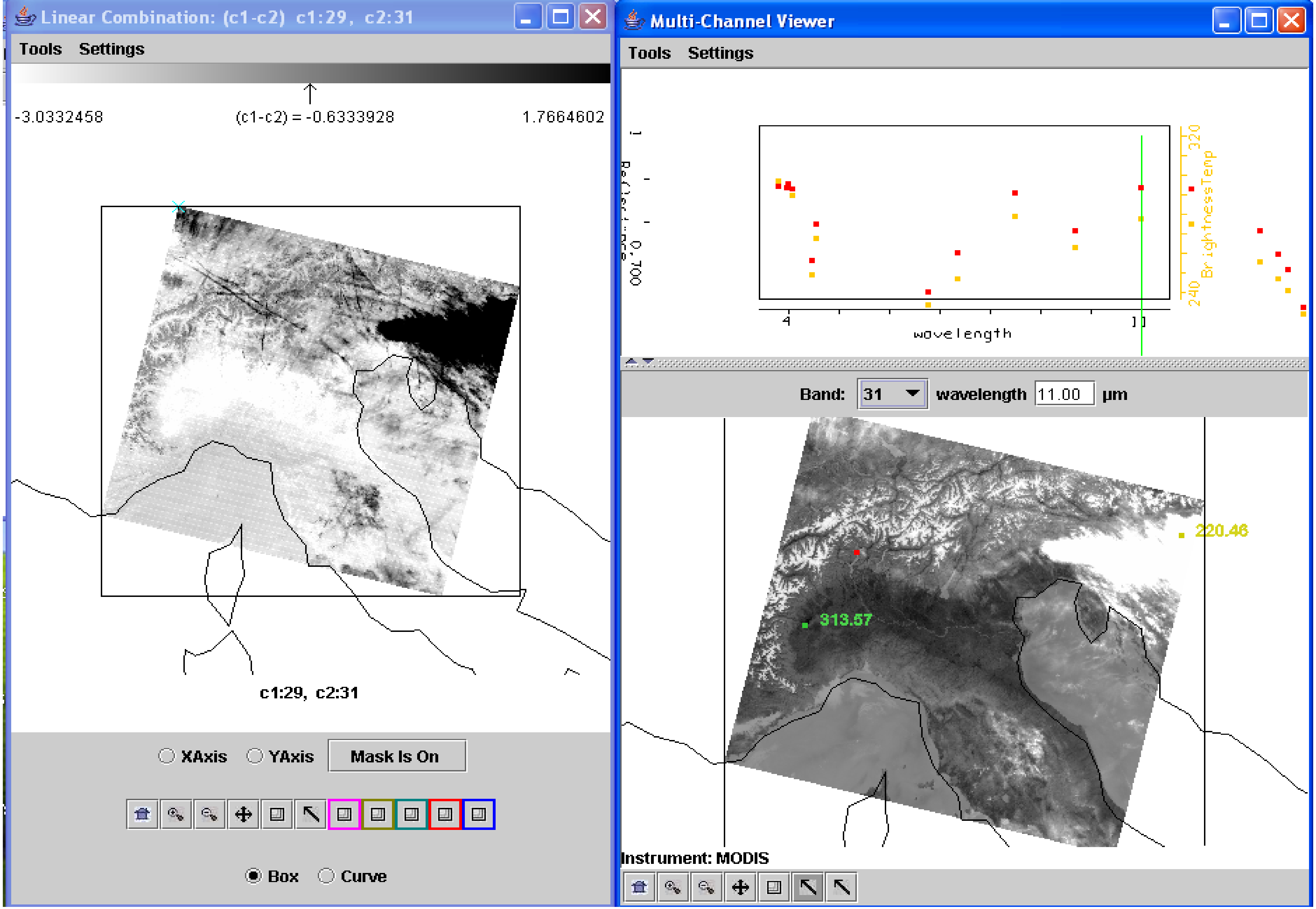
Low clouds reflecting create larger 4 µm brightness temperatures


Hydra (Version: v1.8.0)
File Load Tools Settings Start
Multi-Channel Viewer
Tools Settings Import
wavelength
BrightnessTemp
Band: 20 wavelength 3.80 µm
Instrument: MODIS
Lat = 43.945 Lon = 9.050
Transect: shft+rght+drag
Tools Settings
brightnessTemp
dist,km
4 um
11 um
20 31



67

Detecting low clouds in 4-11 µm brightness temperature differences



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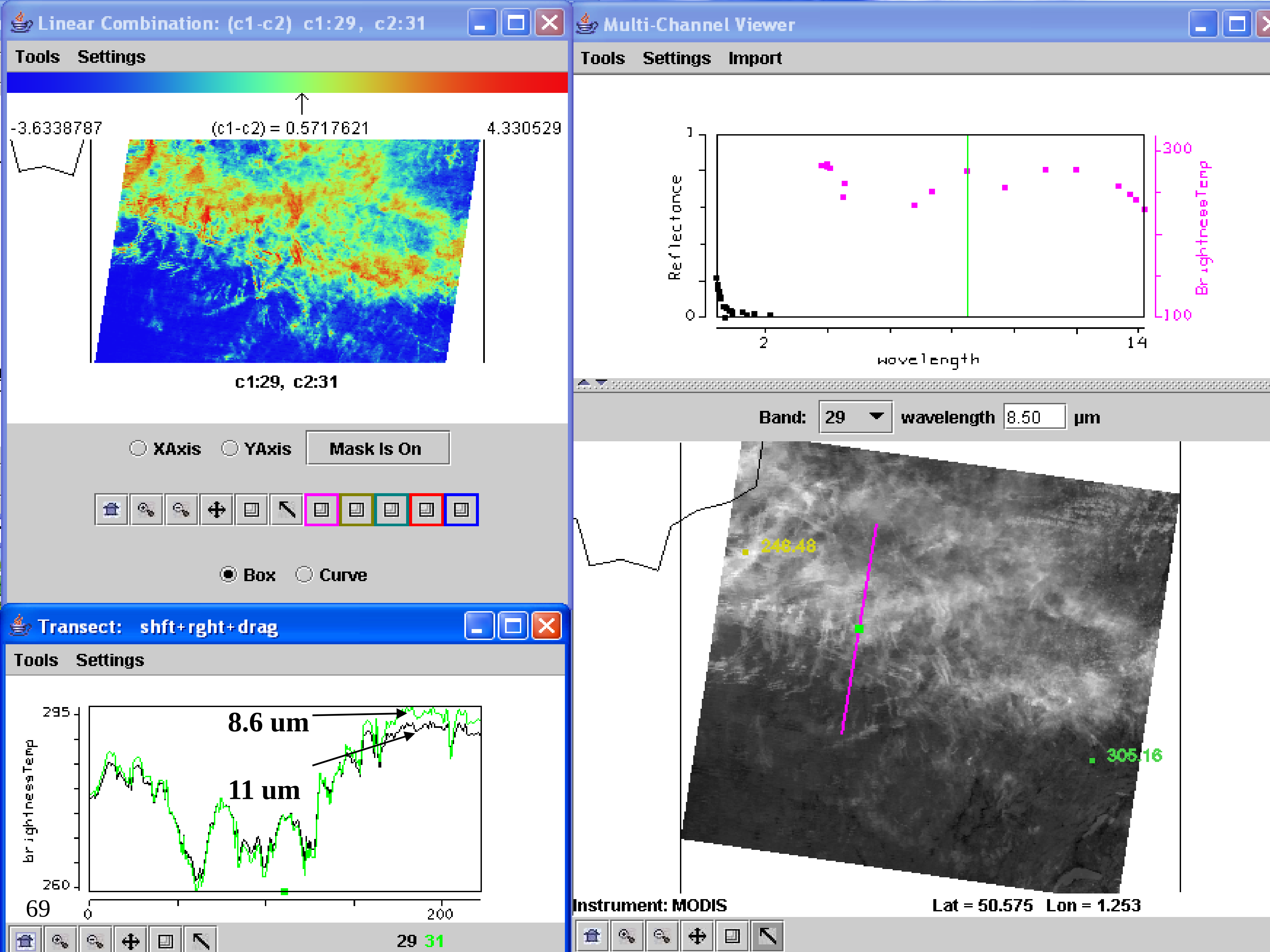
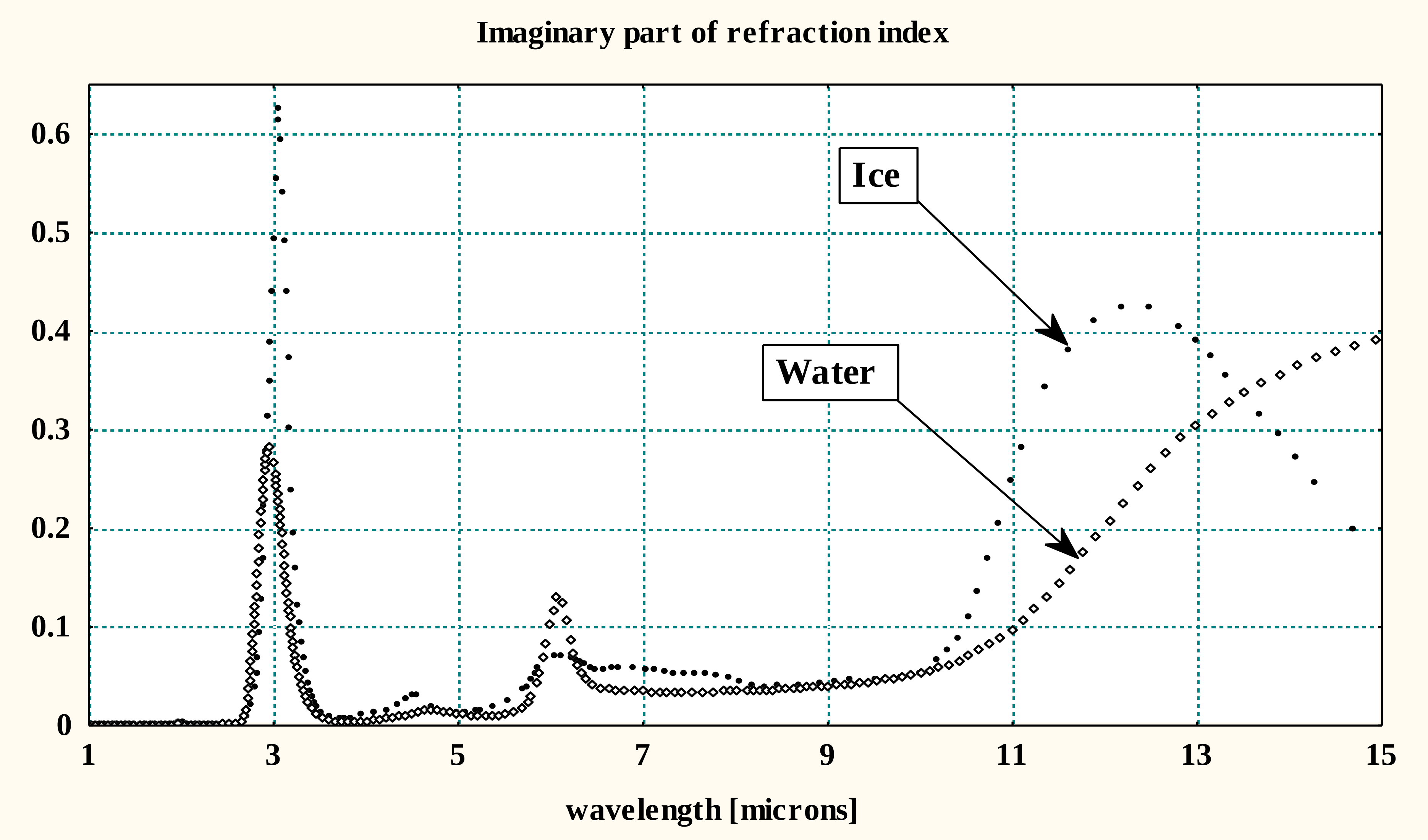
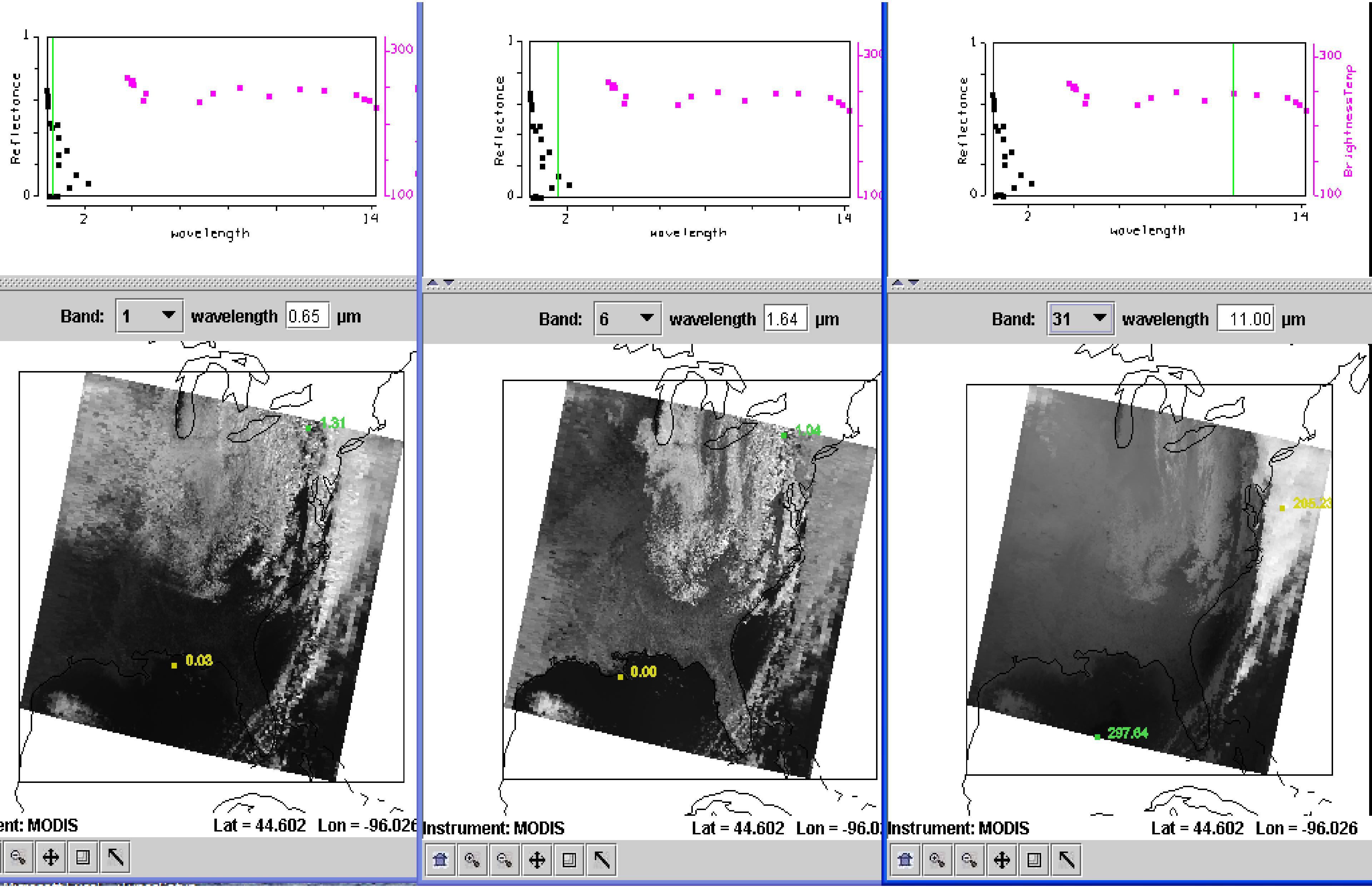
Detecting ice clouds in 8.6-11 μm brightness temperature differences


Linear Combination: (c1-c2) c1:29, c2:31
Tools Settings
-3.6338787
(c1-c2) = 0.5717621
4.330529
c1:29, c2:31
XAxis
YAxis
Mask Is On
Box
Curve
Multi-Channel Viewer
Tools Settings Import
Reflectance
BrightnessTemp
wavelength
1
0
300
100
2
14
Band: 29
wavelength 8.50 µm
248.48
305.16
Instrument: MODIS
Lat = 50.575 Lon = 1.253
Transect: shft+rght+drag
Tools Settings
brightnessTemp
295
260
8.6 um
11 um
0
200
29 31

Optical properties of cloud particles: imaginary part of refraction index

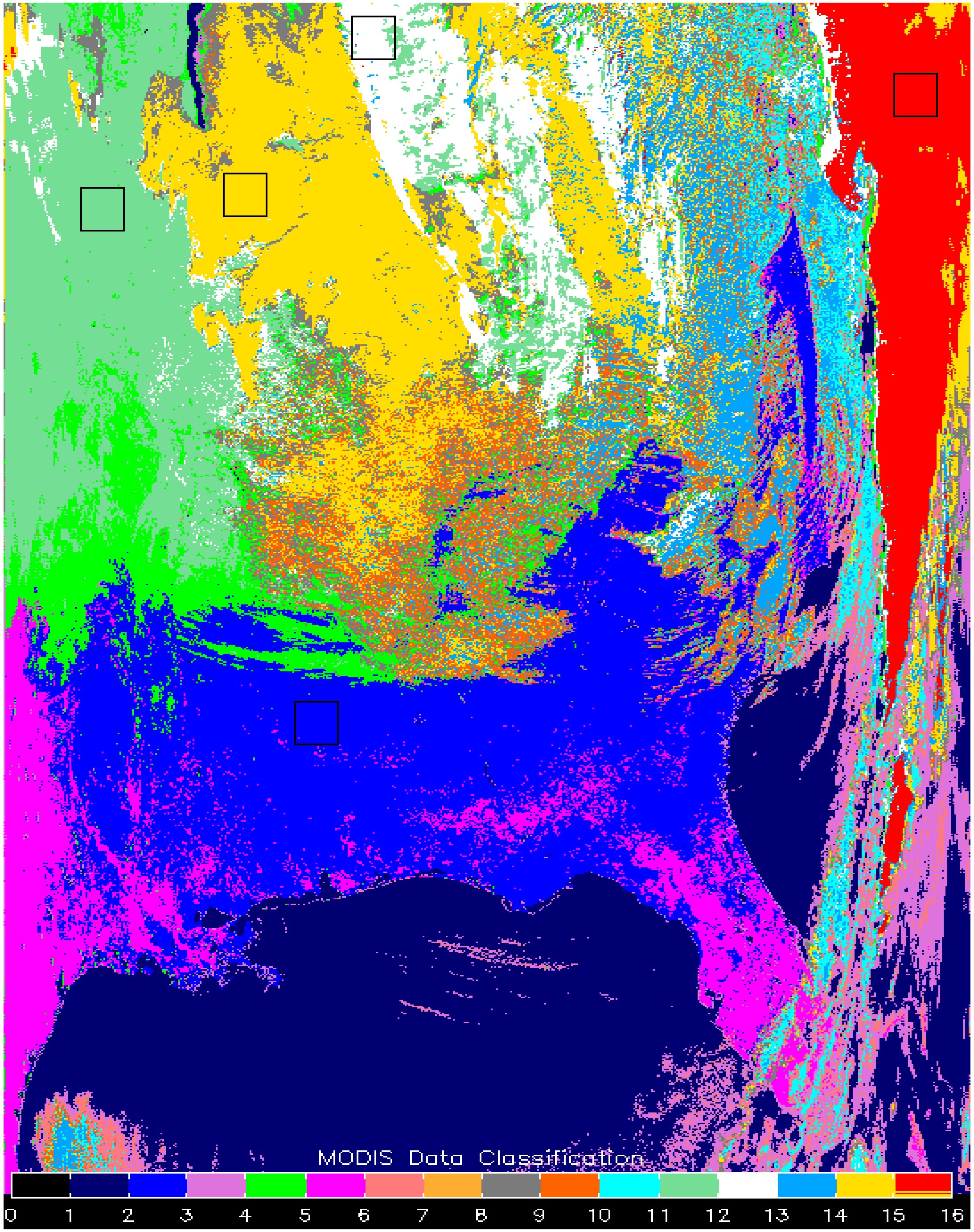


BT[8.6] – BT[11] will be positive for transmissive ice clouds



Reflectance
wavelength
BrightnessTemp
Band: 1
wavelength 0.65 µm
0.03
1.31
ent: MODIS
Lat = 44.602 Lon = -96.026
Band: 6
wavelength 1.64 µm
0.00
1.04
Instrument: MODIS
Lat = 44.602 Lon = -96.0
Band: 31
wavelength 11.00 µm
297.64
Instrument: MODIS
Lat = 44.602 Lon = -96.026

MODIS identifies cloud classes



Hi cld

Mid cld

Lo cld

Snow

clr

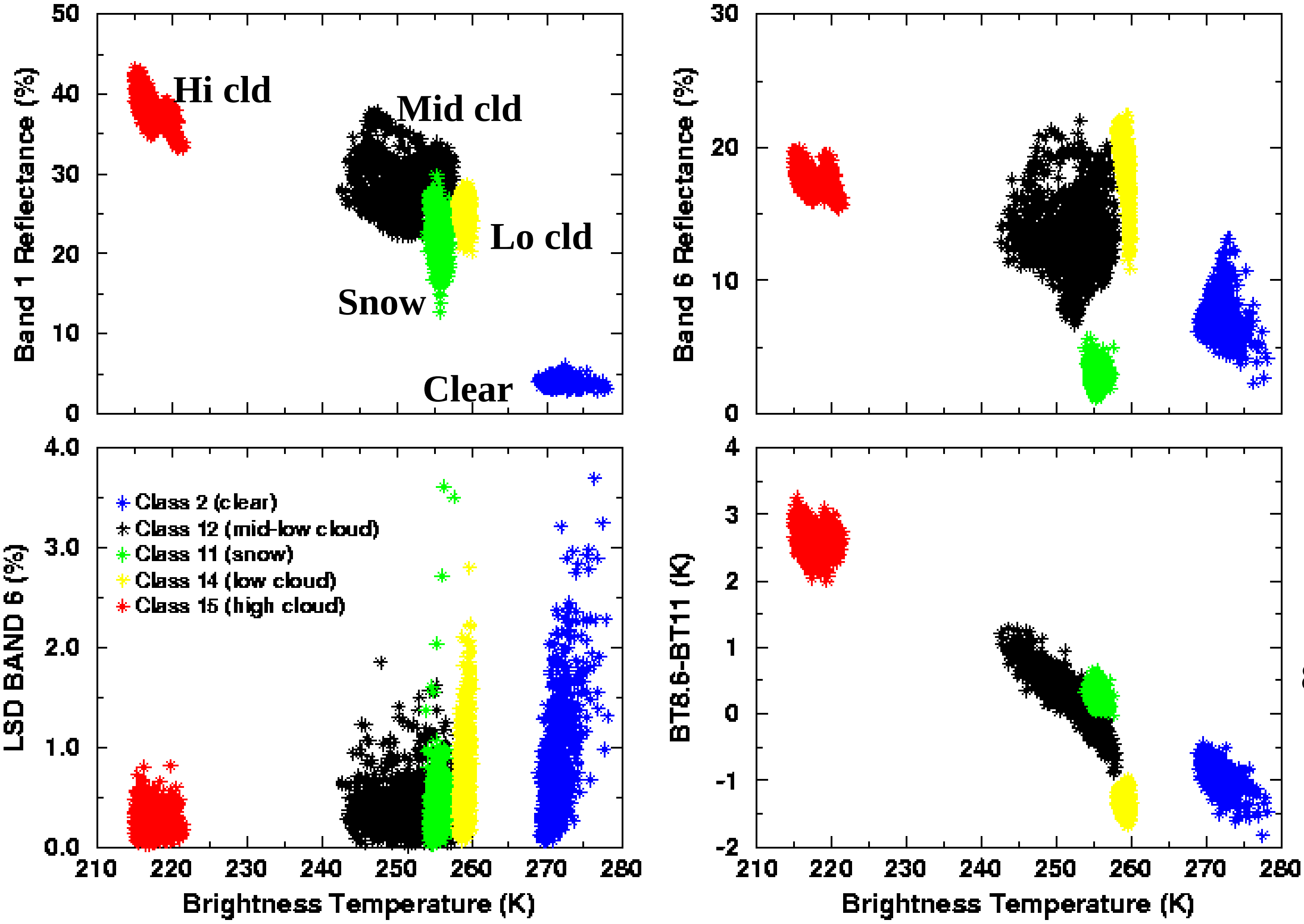
Clouds separate into classes when multispectral radiance information is viewed

vis

1.6 um

LSD
1.6 um

8.6-11 um

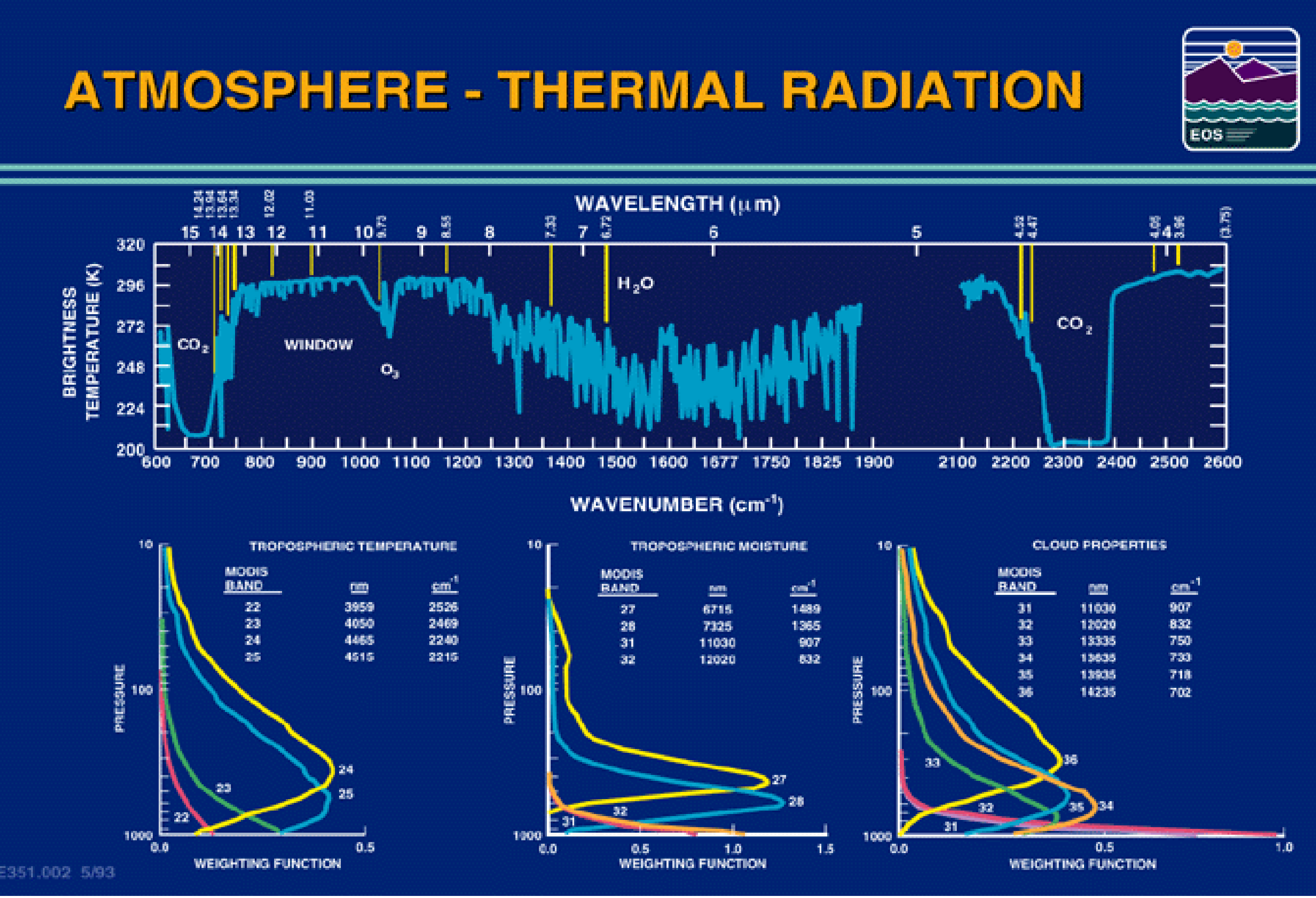


11 um

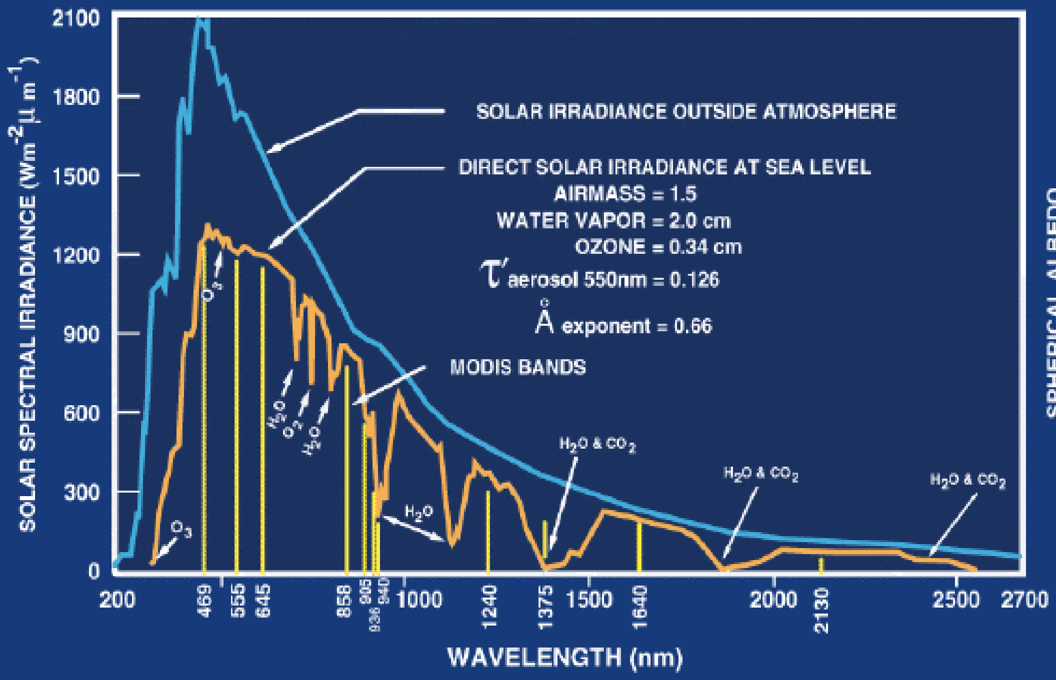
11 um

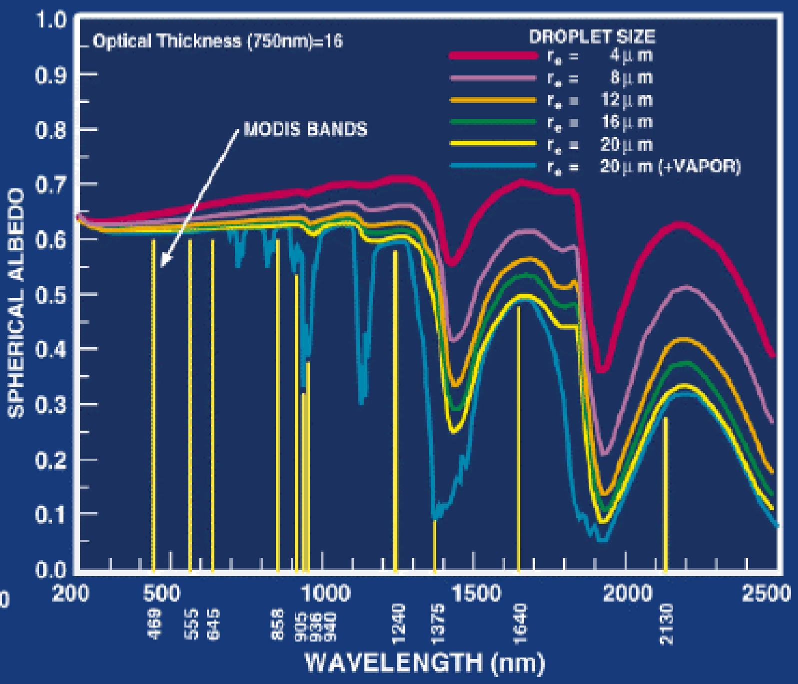
Cloud Mask Tests

• BT11	clouds over ocean
• BT13.9	high clouds
• BT6.7	high clouds
• BT3.9-BT11	broken or scattered clouds
• BT11-BT12	high clouds in tropics
• BT8.6-BT11	ice clouds
• BT6.7-BT11 or BT13.9-BT11	clouds in polar regions
• BT11+aPW(BT11-BT12)	clouds over ocean
• r0.65	clouds over land
• r0.85	clouds over ocean
• r1.38	thin cirrus
• r1.6	clouds over snow, ice cloud
• r0.85/r0.65 or NDVI	clouds over vegetation
• σ(BT11)	clouds over ocean

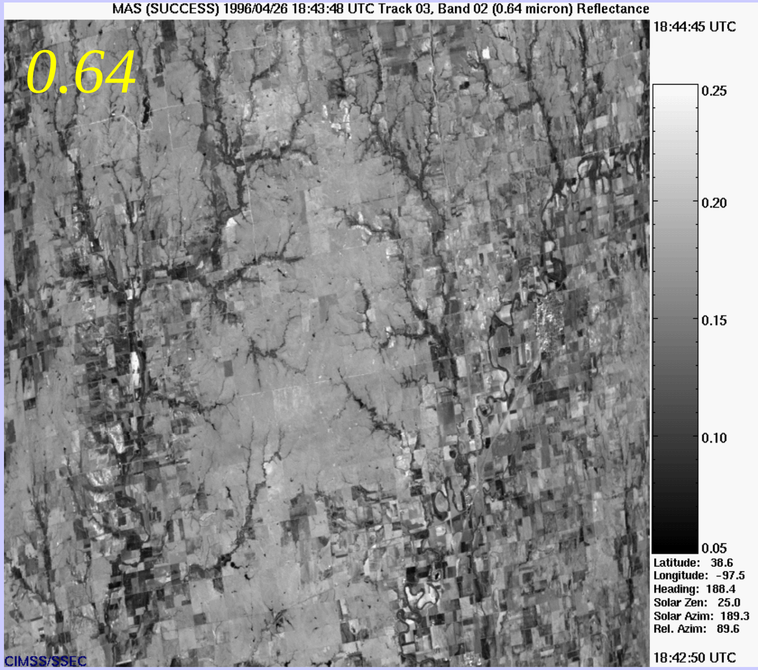

ATMOSPHERE - THERMAL RADIATION
EOS
WAVELENGTH (μm)
15 14 13 12 11 10 9 8 7 6 5 4
14.24 13.94 13.64 13.34 12.02 11.03 9.73 8.55 7.33 6.72 4.52 4.47 4.08 3.96 (3.75)
BRIGHTNESS TEMPERATURE (K)
320 296 272 248 224 200
CO2
WINDOW
O3
H2O
CO2
600 700 800 900 1000 1100 1200 1300 1400 1500 1600 1677 1750 1825 1900 2100 2200 2300 2400 2500 2600
WAVENUMBER (cm-1)
TROPOSPHERIC TEMPERATURE
MODIS BAND nm cm-1
22 3959 2526
23 4050 2469
24 4465 2240
25 4515 2215
PRESSURE
10 100 1000
WEIGHTING FUNCTION
0.0 0.5
TROPOSPHERIC MOISTURE
MODIS BAND nm cm-1
27 6715 1489
28 7325 1365
31 11030 907
32 12020 832
0.0 0.5 1.0 1.5
CLOUD PROPERTIES
MODIS BAND nm cm-1
31 11030 907
32 12020 832
33 13335 750
34 13635 733
35 13935 718
36 14235 702
0.0 0.5 1.0
E351.002 5/93

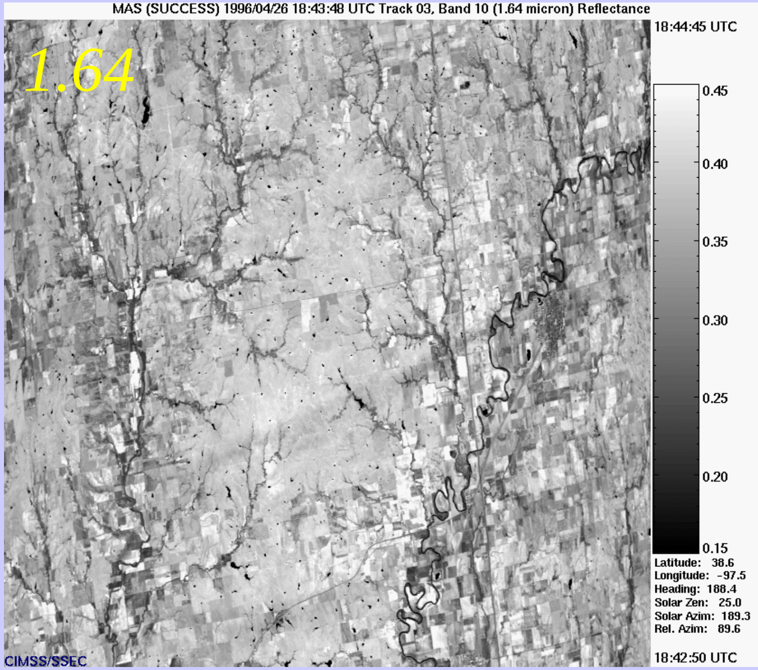
ATMOSPHERE-SOLAR RADIATION

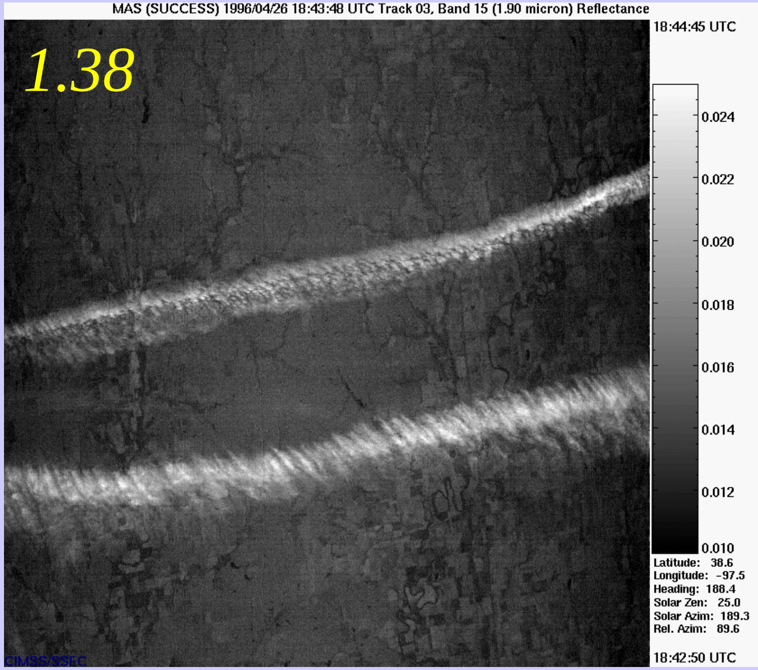


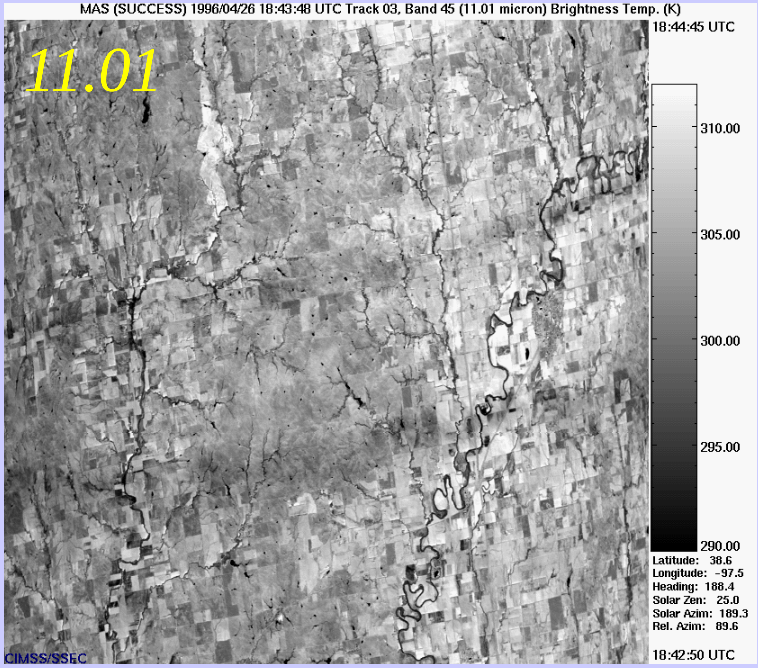


C351.008 5/93

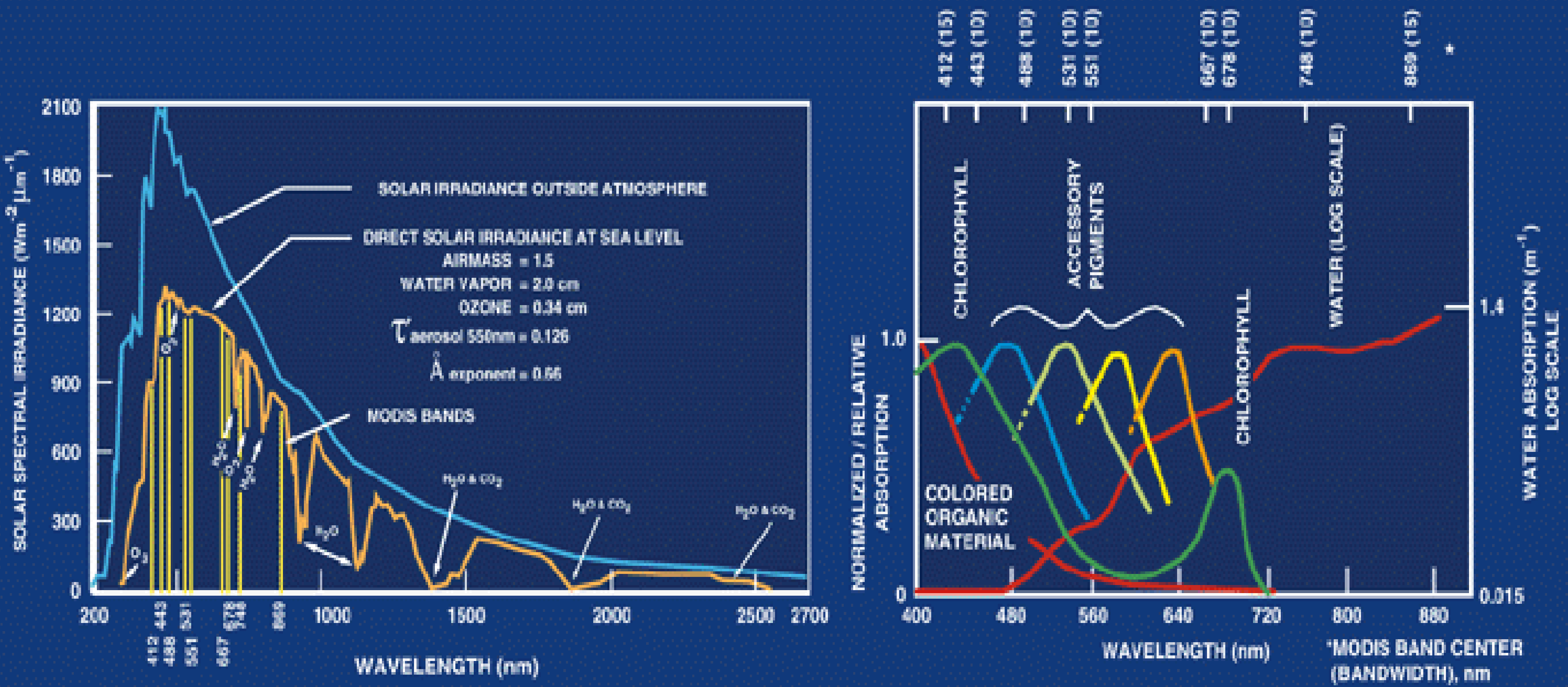

MAS (SUCCESS) 1996/04/26 18:43:48 UTC Track 03, Band 02 (0.64 micron) Reflectance
0.64
18:44:45 UTC
0.25
0.20
0.15
0.10
0.05
Latitude: 38.6
Longitude: -97.5
Heading: 188.4
Solar Zen: 25.0
Solar Azim: 189.3
Rel. Azim: 89.6
18:42:50 UTC
CIMSS/SSEC


MAS (SUCCESS) 1996/04/26 18:43:48 UTC Track 03, Band 10 (1.64 micron) Reflectance
1.64
18:44:45 UTC
0.45
0.40
0.35
0.30
0.25
0.20
0.15
Latitude: 38.6
Longitude: -97.5
Heading: 188.4
Solar Zen: 25.0
Solar Azim: 189.3
Rel. Azim: 89.6
18:42:50 UTC
CIMSS/SSEC

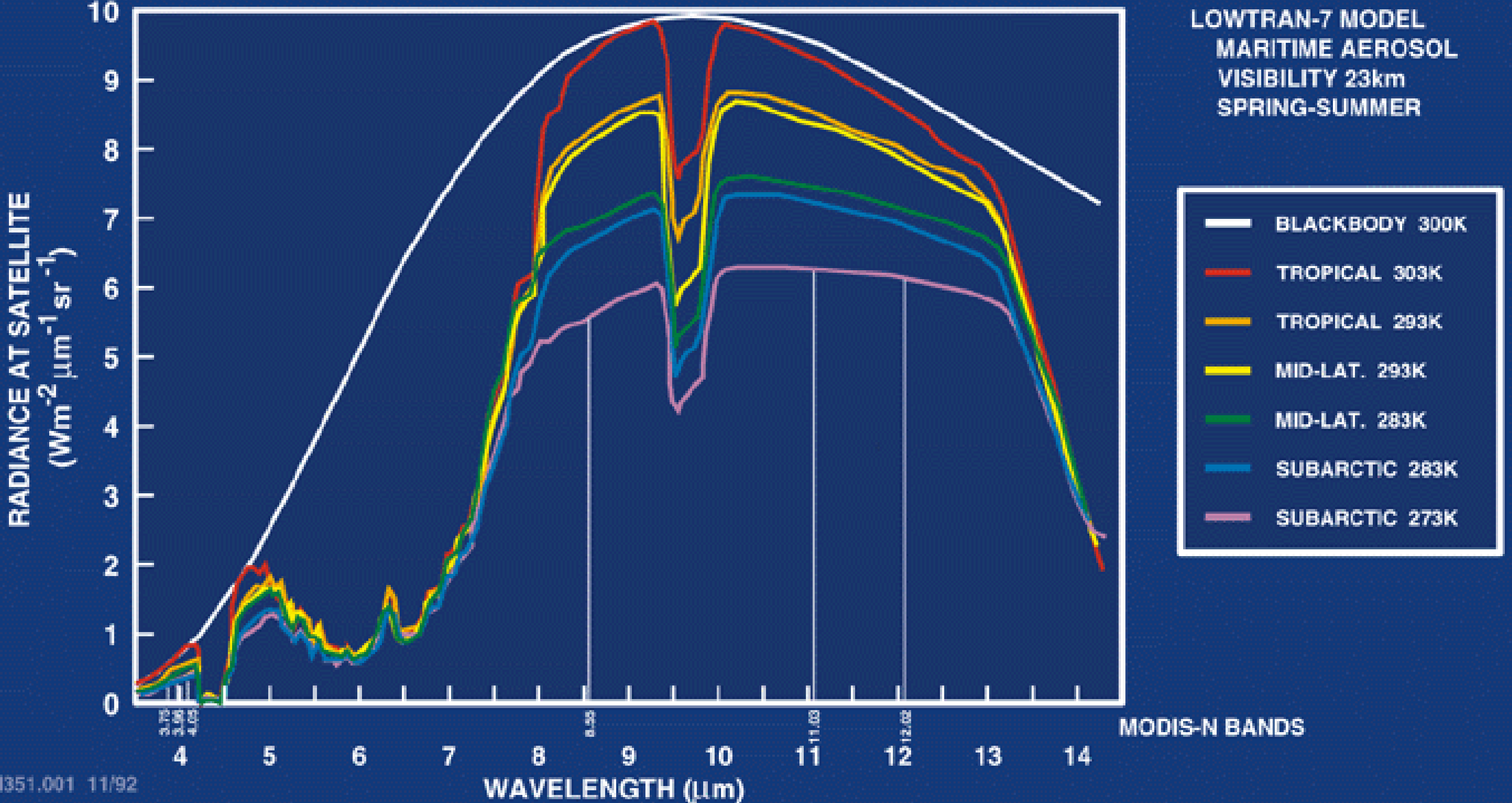

MAS (SUCCESS) 1996/04/26 18:43:48 UTC Track 03, Band 15 (1.90 micron) Reflectance
1.38
18:44:45 UTC
0.024
0.022
0.020
0.018
0.016
0.014
0.012
0.010
Latitude: 38.6
Longitude: -97.5
Heading: 188.4
Solar Zen: 25.0
Solar Azim: 189.3
Rel. Azim: 89.6
18:42:50 UTC
CIMSS/SSEC


MAS (SUCCESS) 1996/04/26 18:43:48 UTC Track 03, Band 45 (11.01 micron) Brightness Temp. (K)
11.01
18:44:45 UTC
310.00
305.00
300.00
295.00
290.00
Latitude: 38.6
Longitude: -97.5
Heading: 188.4
Solar Zen: 25.0
Solar Azim: 189.3
Rel. Azim: 89.6
18:42:50 UTC
CIMSS/SSEC

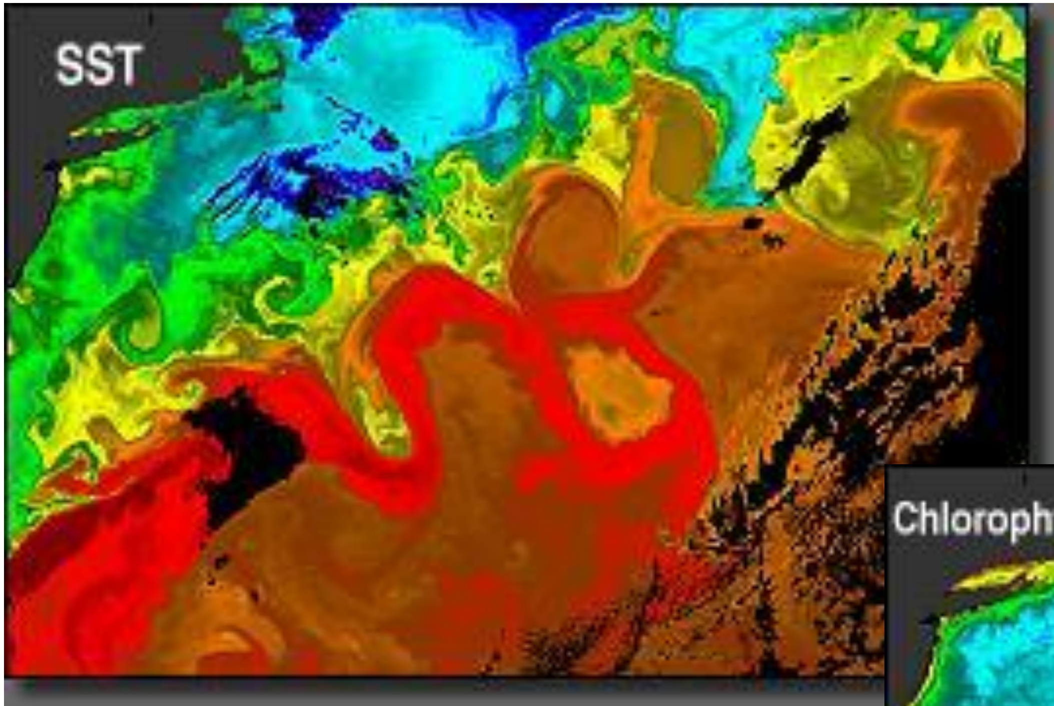
OCEAN-SOLAR RADIATION

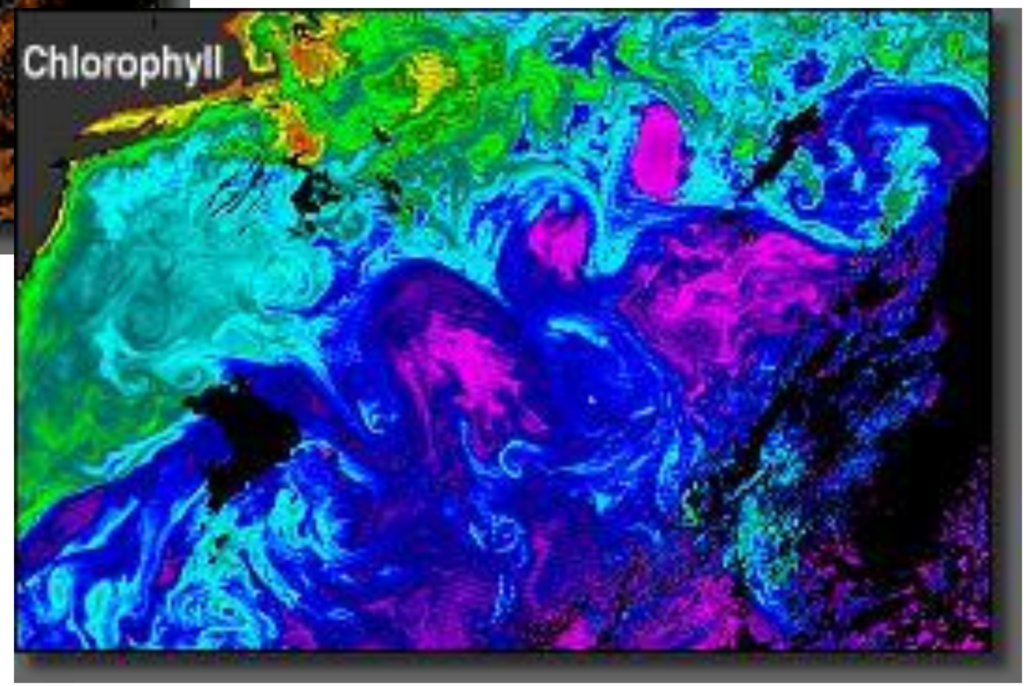


MODIS SEA SURFACE TEMPERATURE



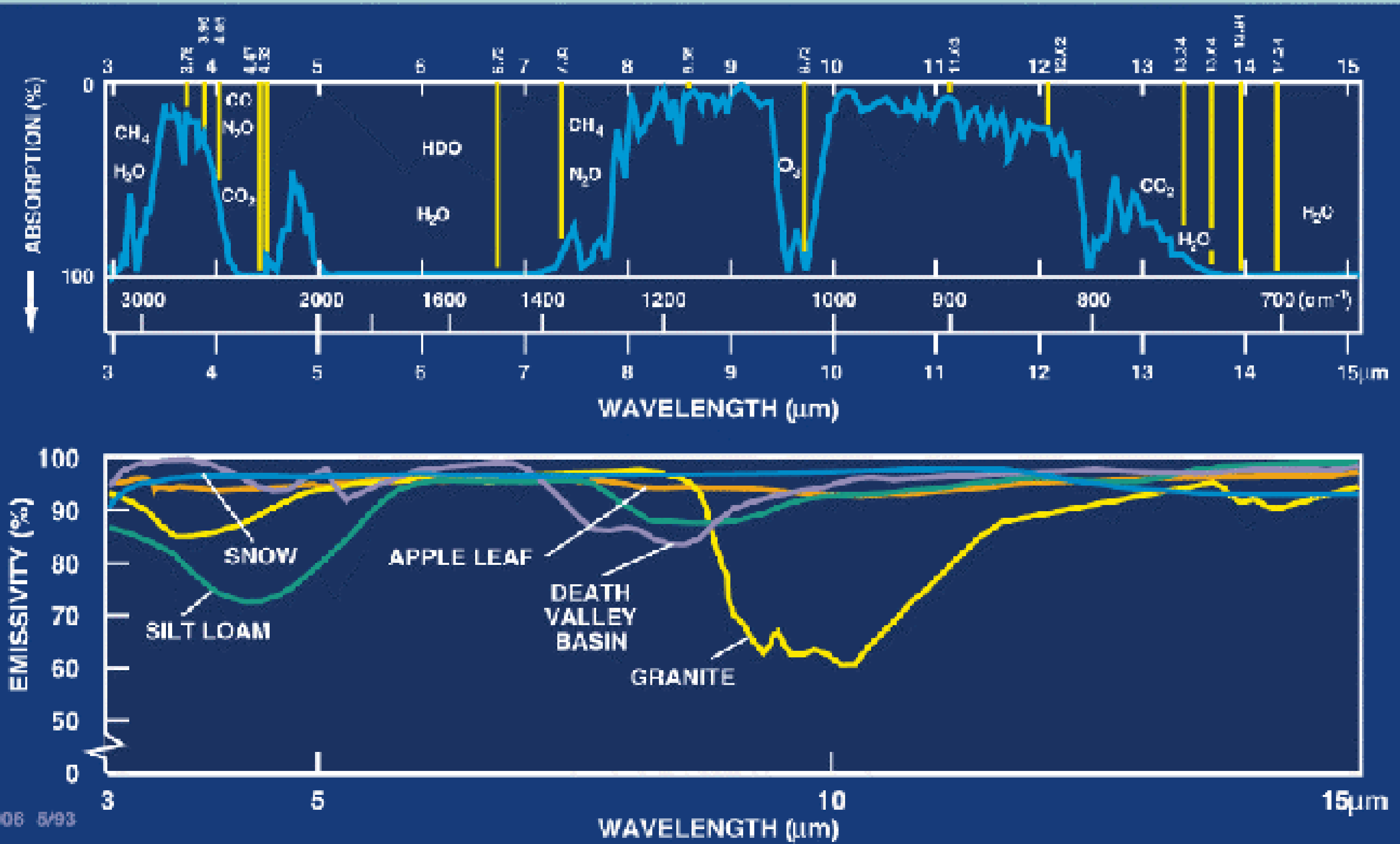
H351.001 11/92



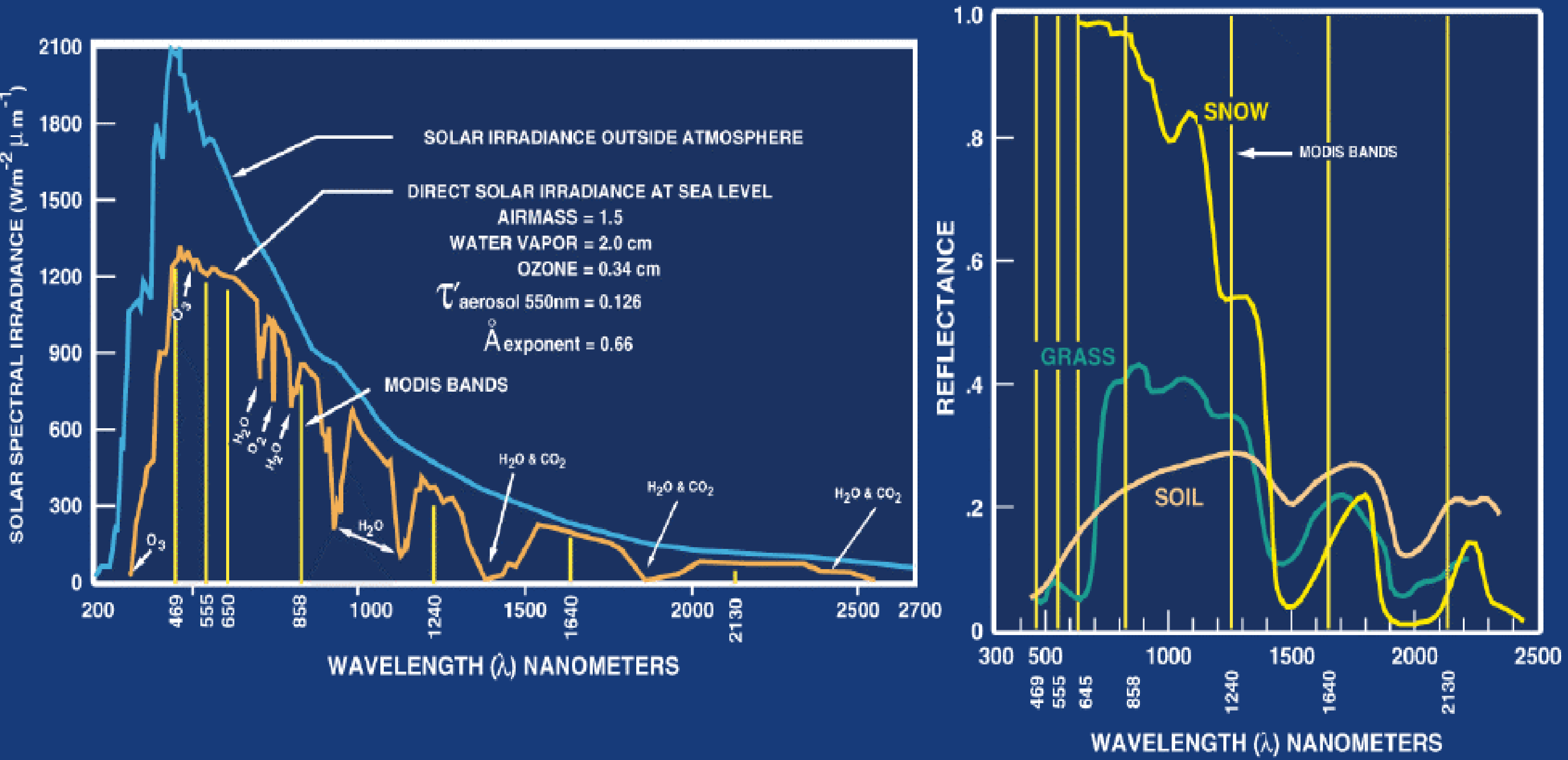


The warm heart of the Gulf Stream is readily apparent in the top SST image. As the current flows toward the northeast it begins to meander and pinch off eddies that transport warm water northward and cold water southward. The current also divides the local ocean into a low-biomass region to the south and a higher-biomass region to the north. The data were collected by MODIS aboard Aqua on April 18, 2005.

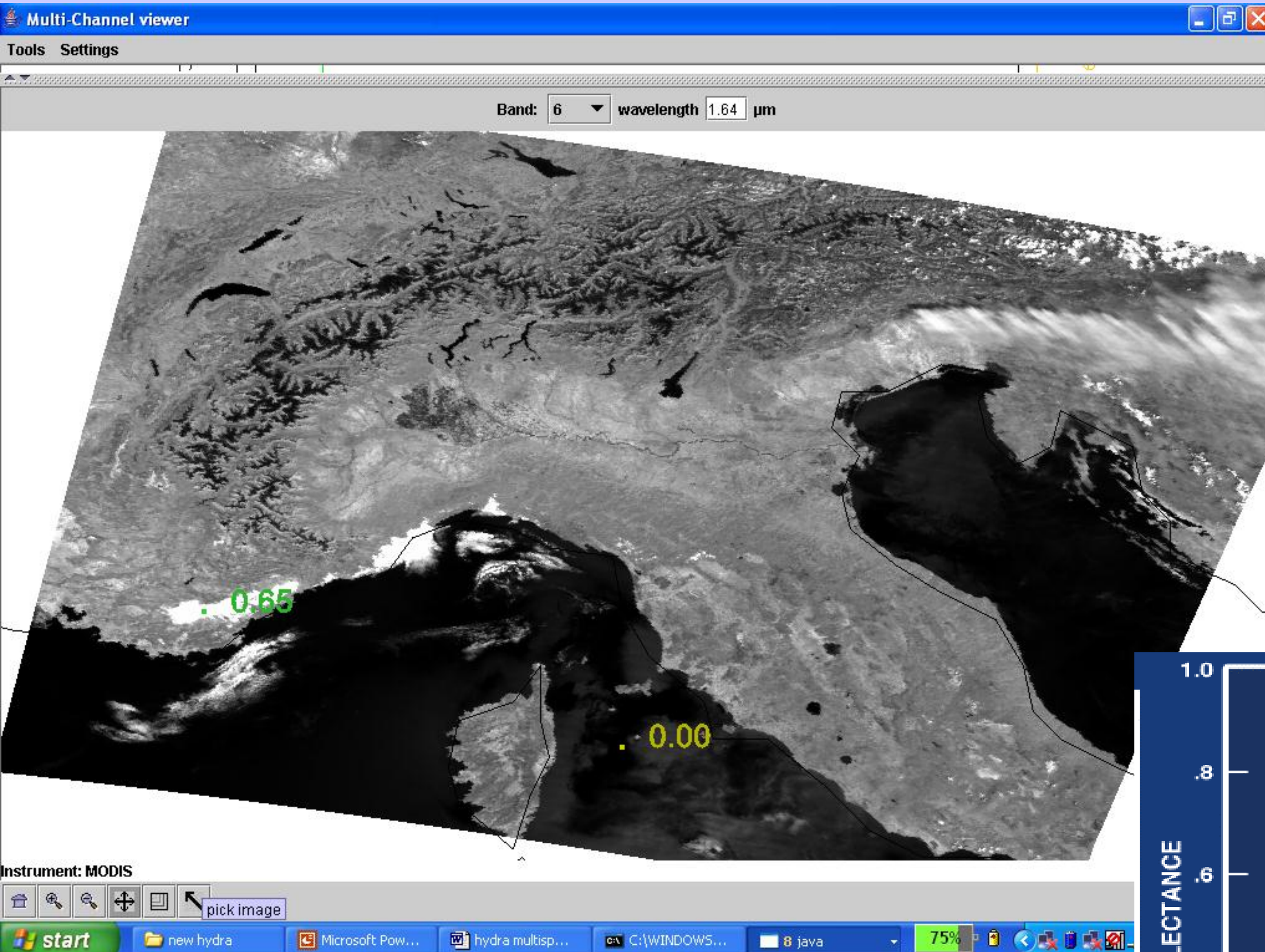
LAND - THERMAL RADIATION

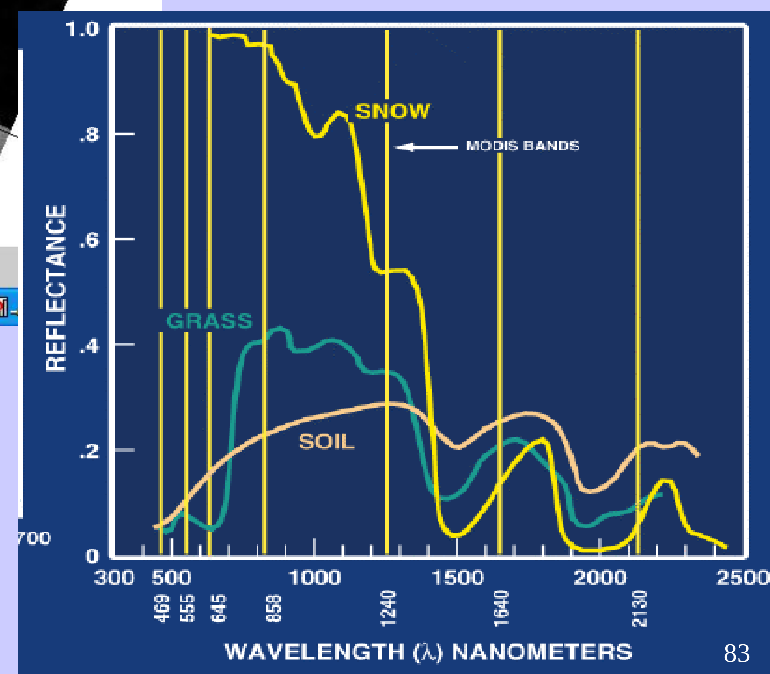


LAND-SOLAR RADIATION

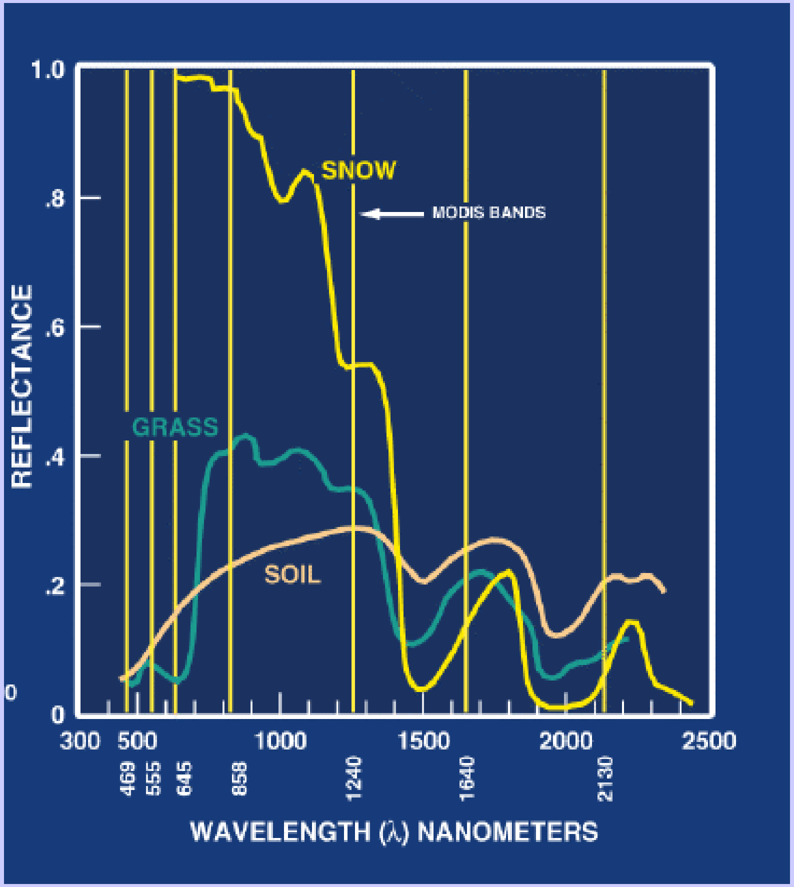


Example with MODIS





low refl at 1.6 um from snow in mountains

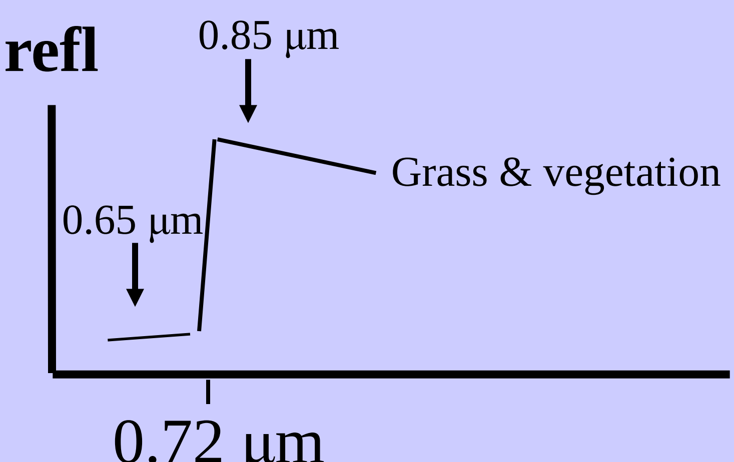


Investigating with Multi-spectral Combinations

Given the spectral response of a surface or atmospheric feature

Select a part of the spectrum where the reflectance or absorption changes with wavelength

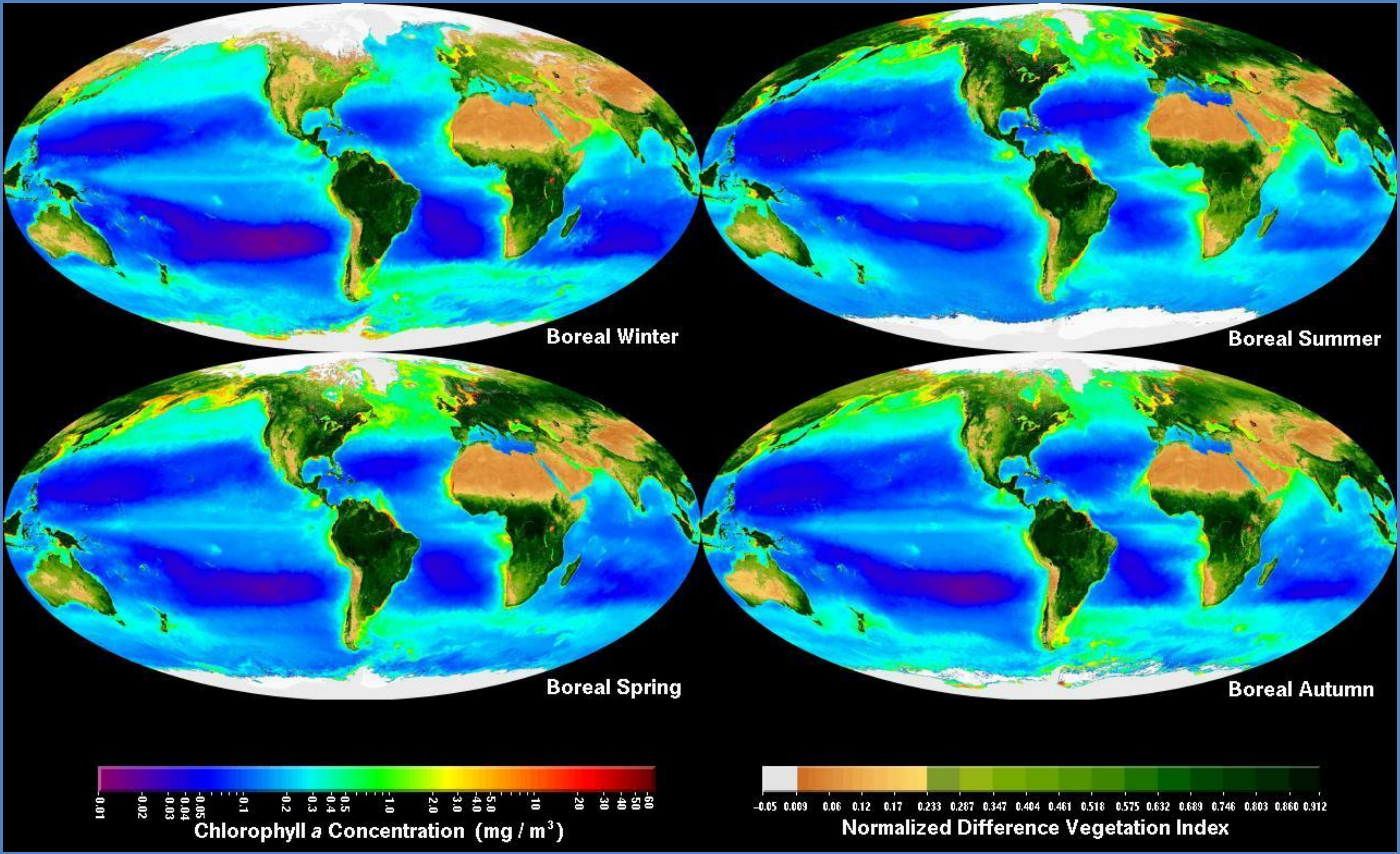
e.g. reflection from grass

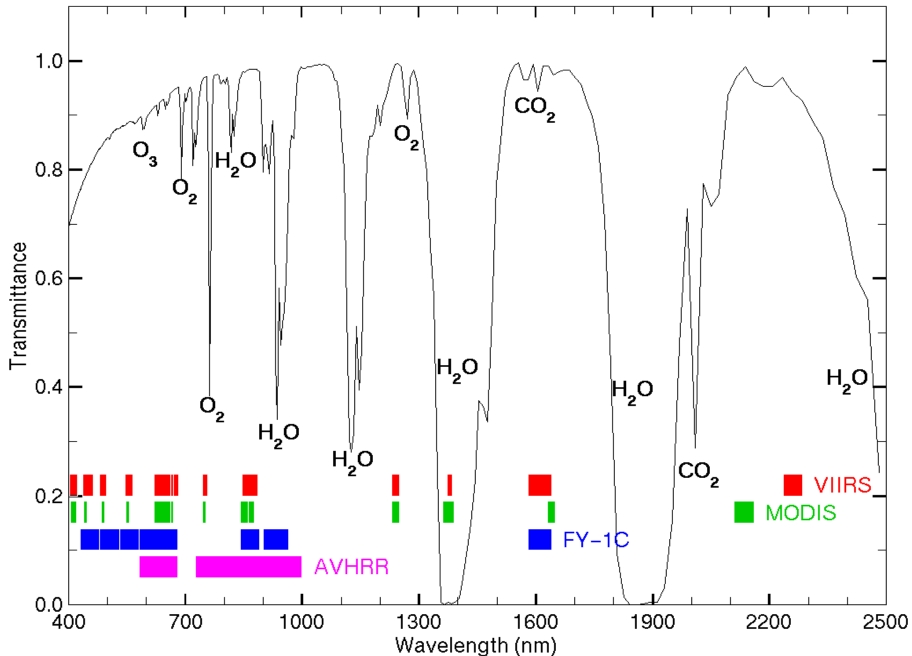


If 0.65 µm and 0.85 µm channels see the same reflectance than surface viewed is not grass;
if 0.85 µm sees considerably higher reflectance than 0.65 µm then surface might be grass

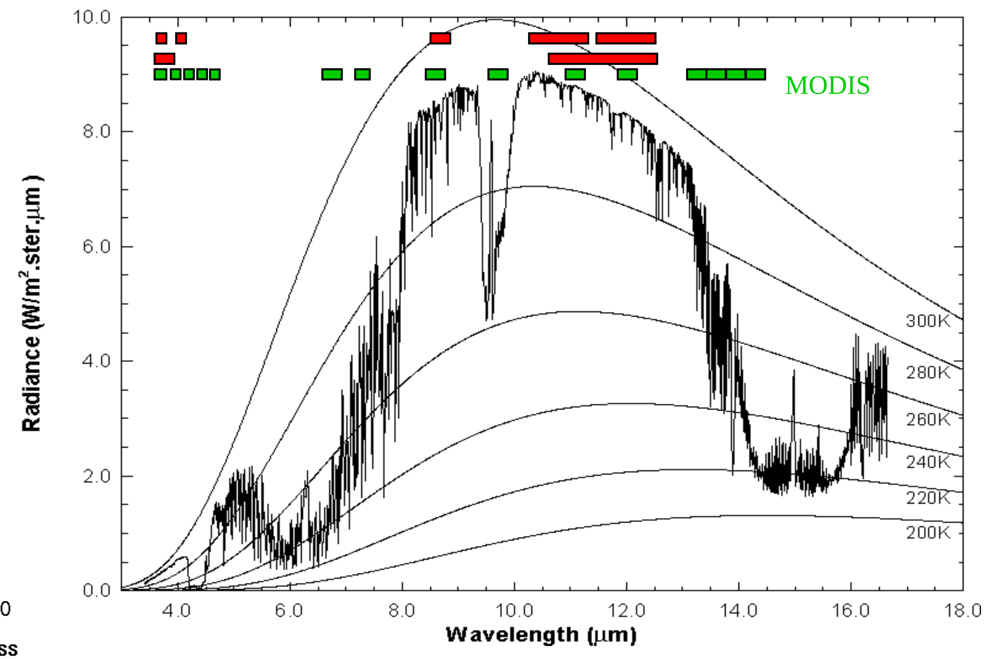
Seasonal Biosphere

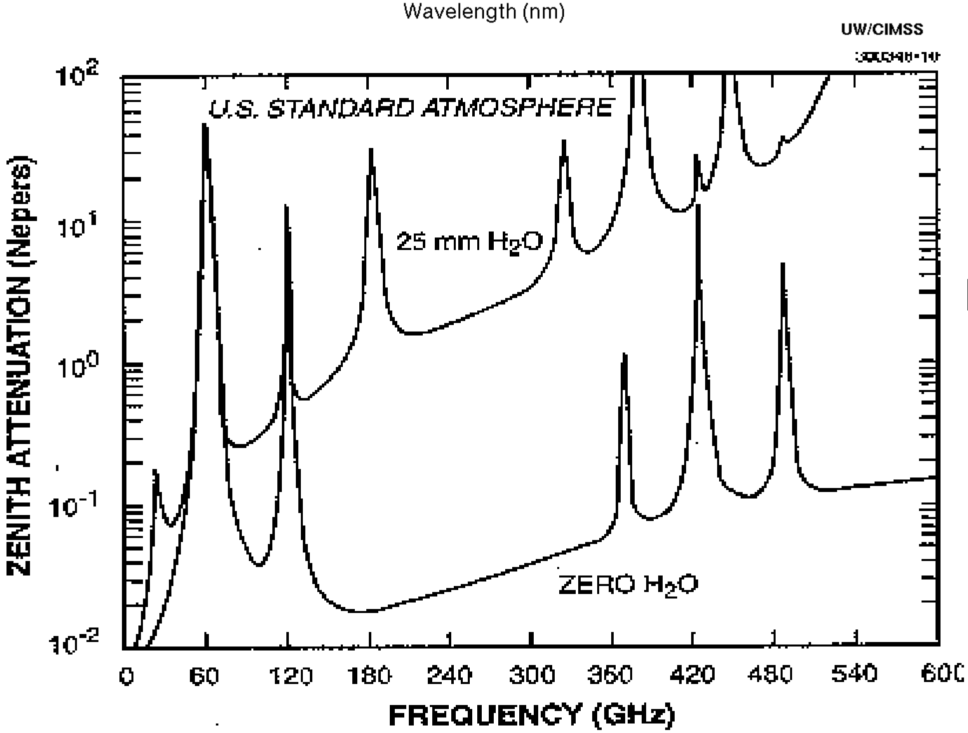
Ocean Chlorophyll-a & Terrestrial NDVI

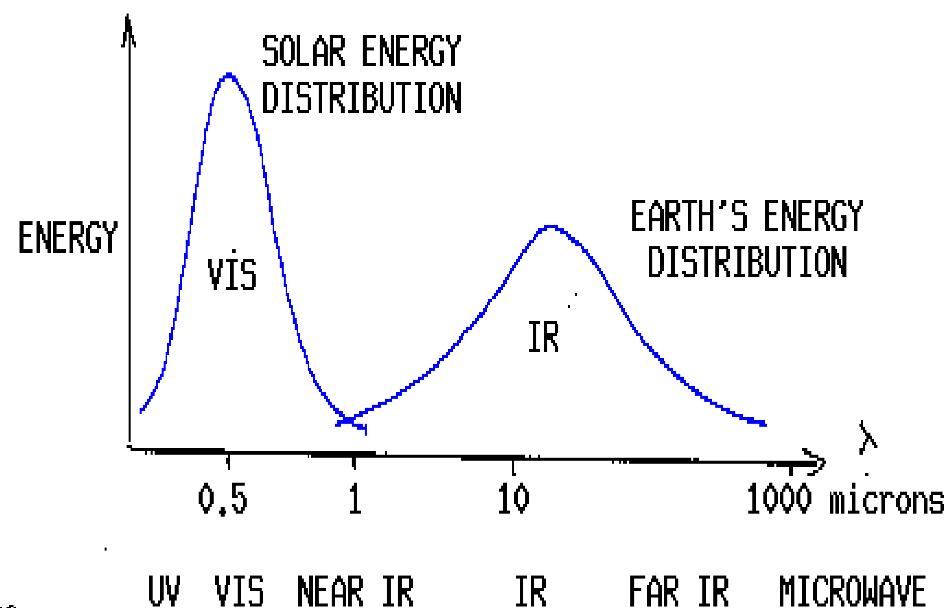




High resolution atmospheric absorption spectrum
and comparative blackbody curves.







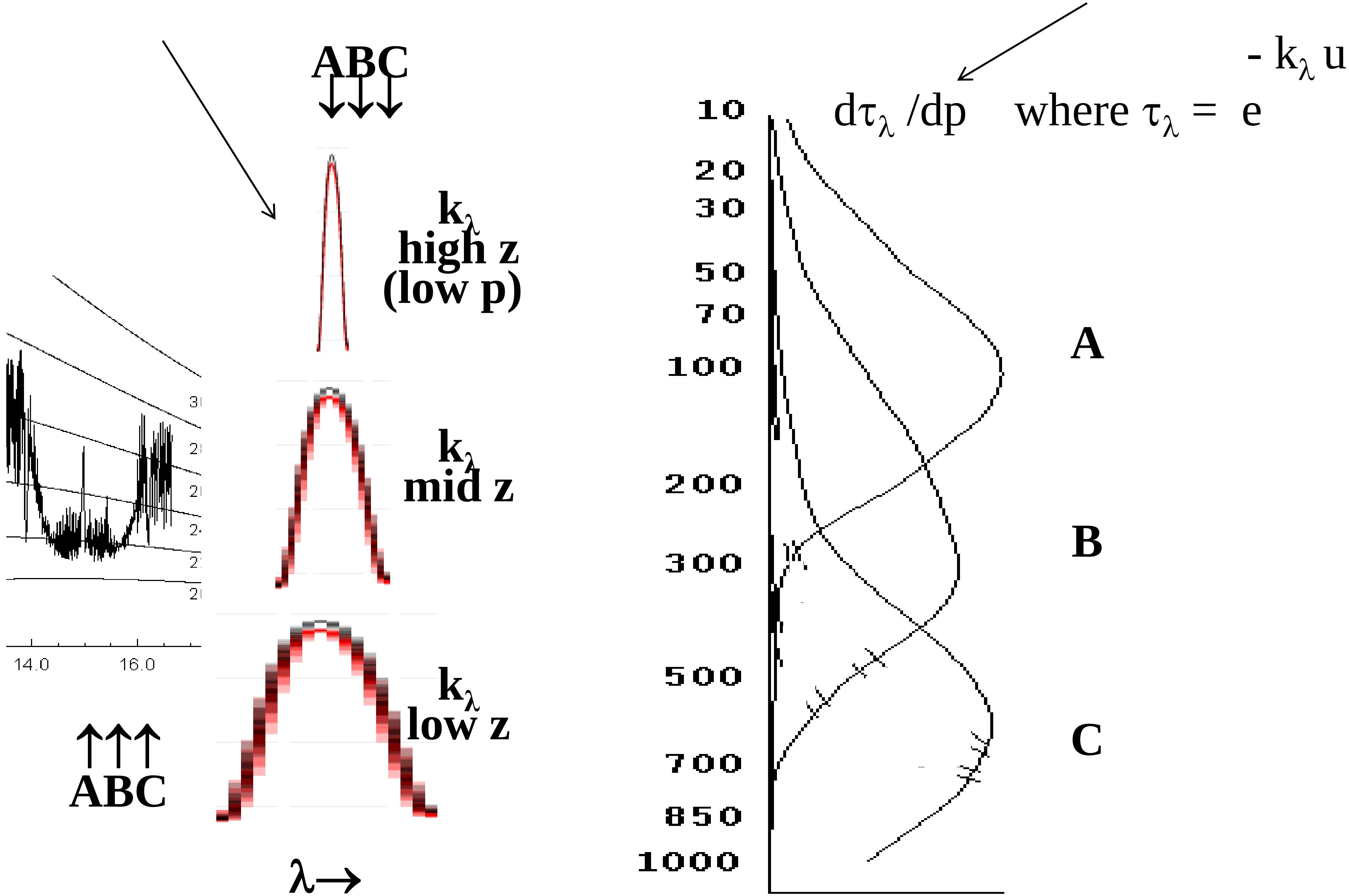
Intro to Lab on High Spectral Resolution IR Measurements

Lecture - Labs in Bologna & Potenza

May-June 2012

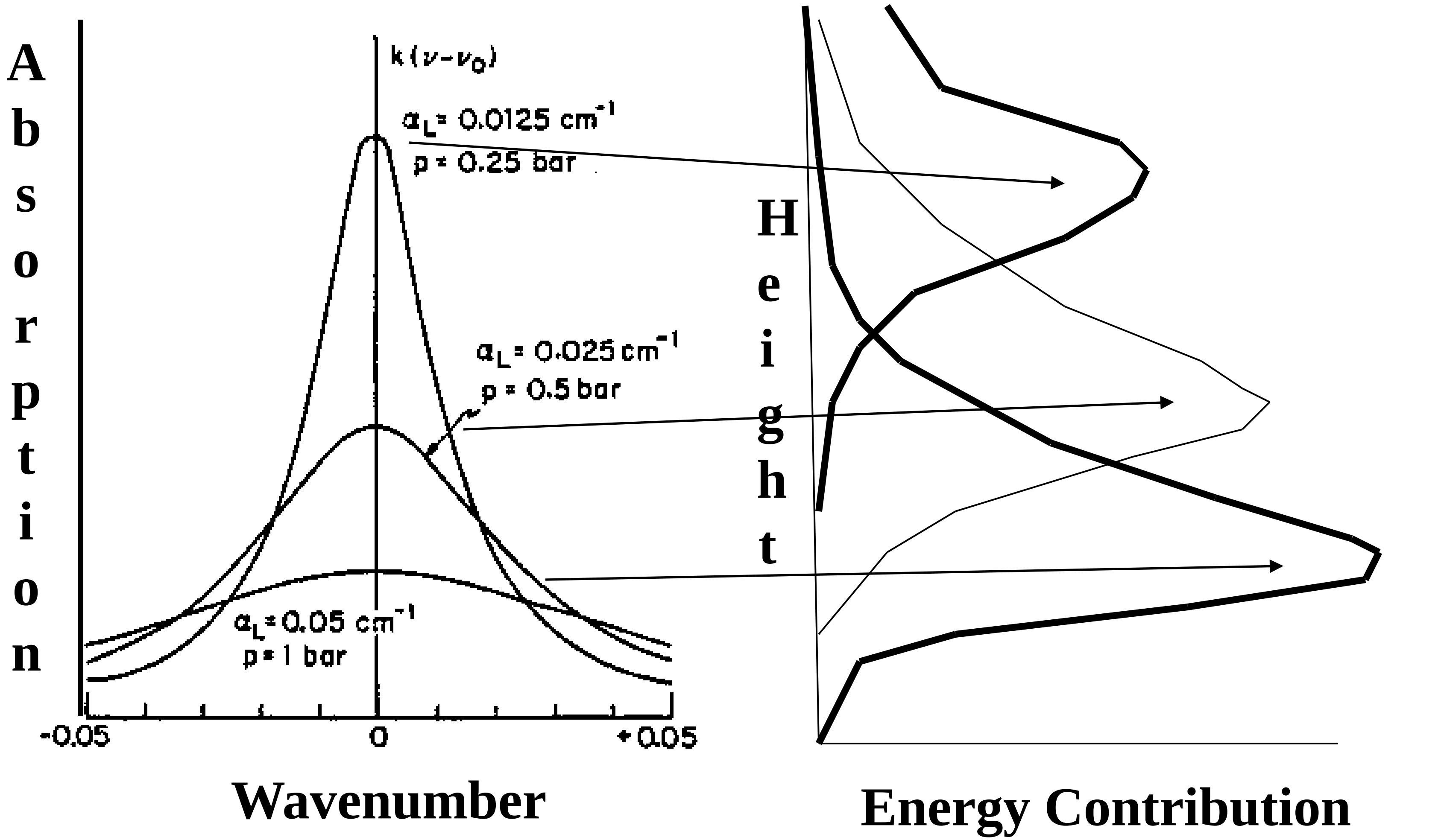
Paul Menzel
UW/CIMSS/AOS

line broadening with pressure helps to explain weighting functions

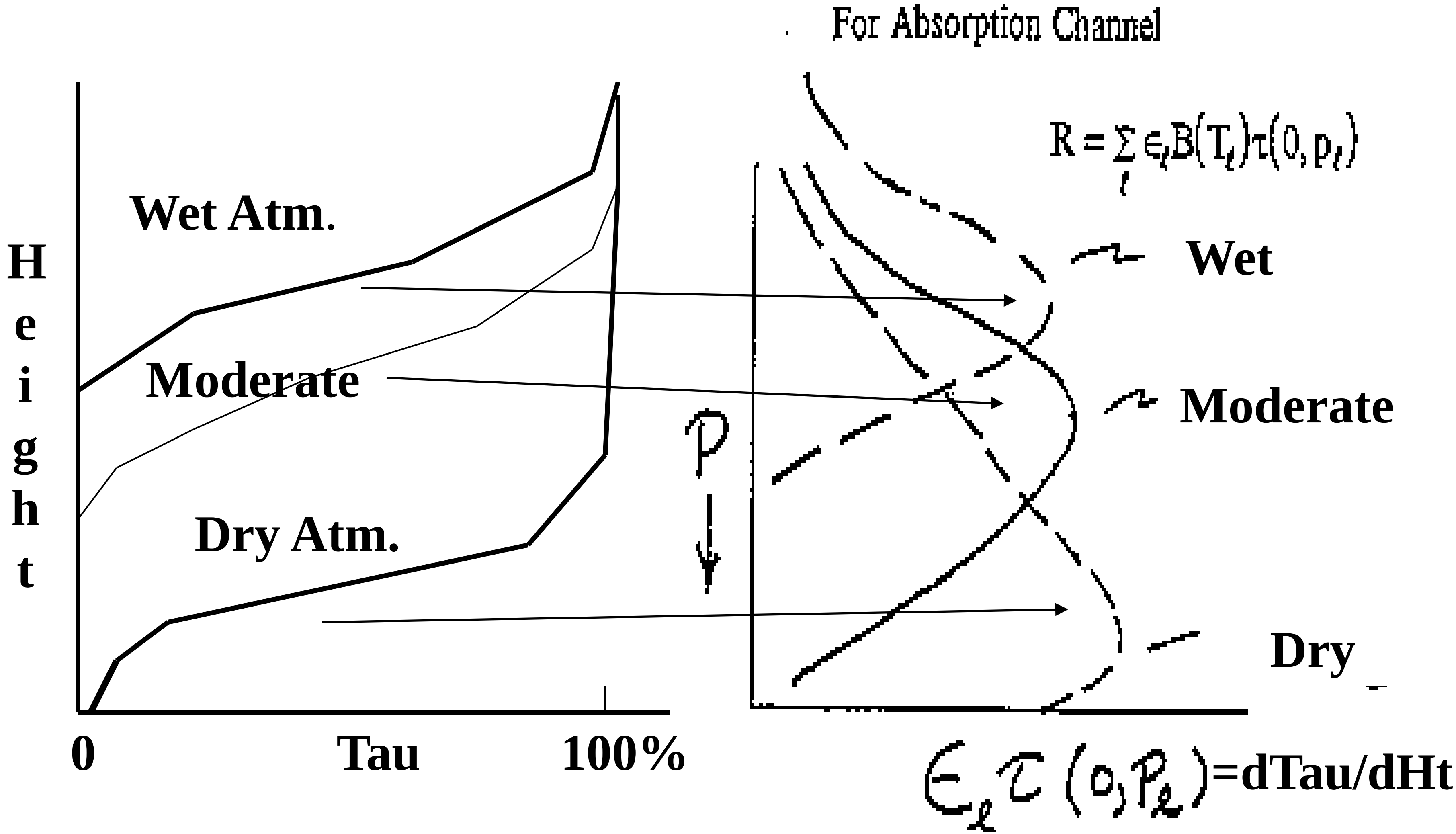


line broadening with pressure helps to explain weighting functions

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

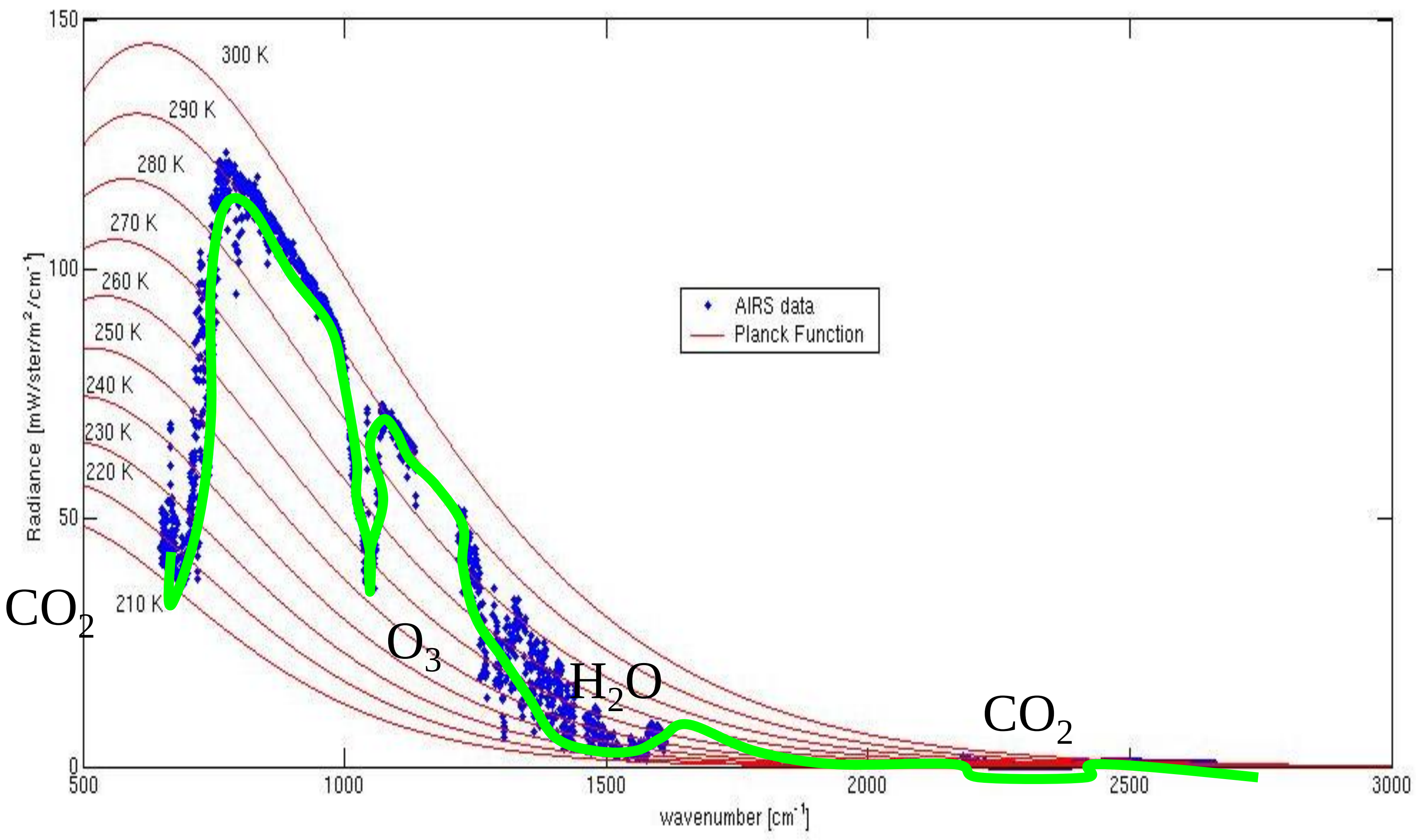


For a given water vapor spectral channel the weighting function depends on the amount of water vapor in the atmospheric column



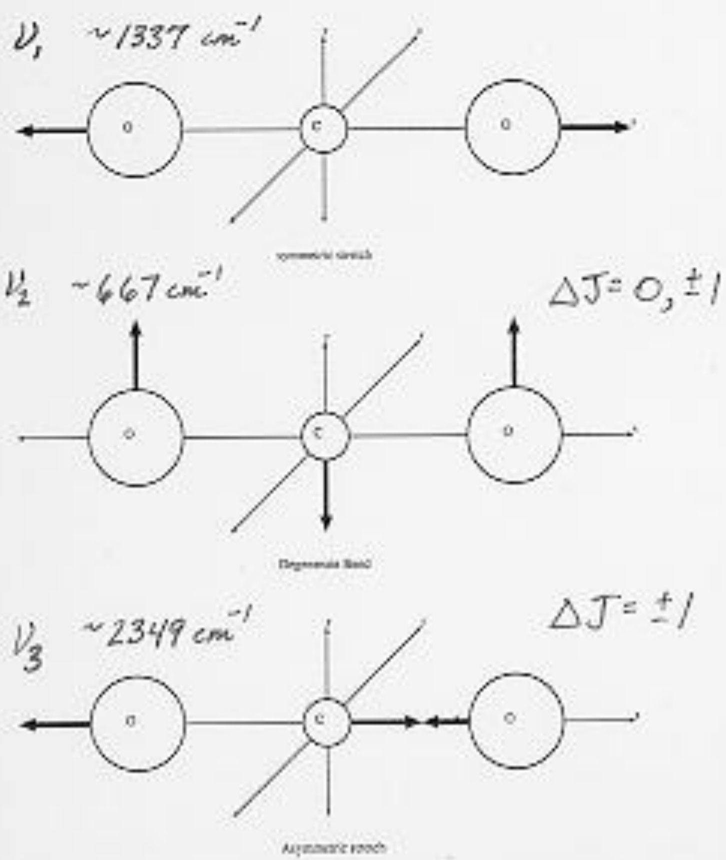
CO2 is about the same everywhere, the weighting function for a given CO2 spectral channel is the same everywhere

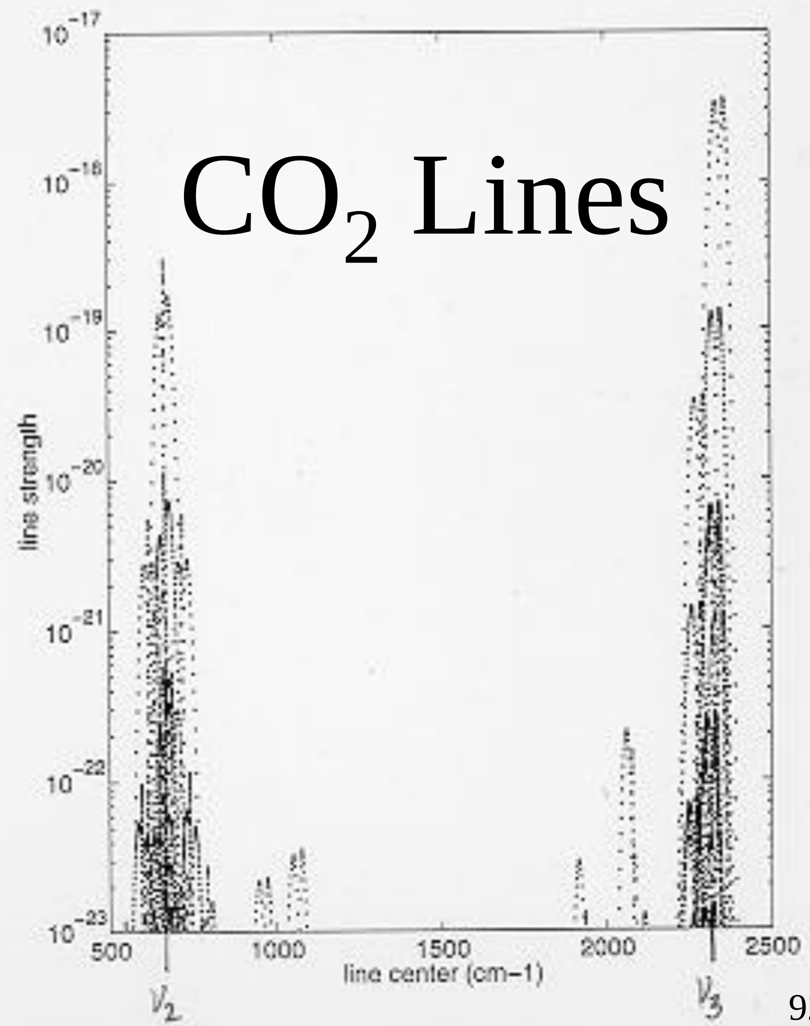
Vibrational Bands



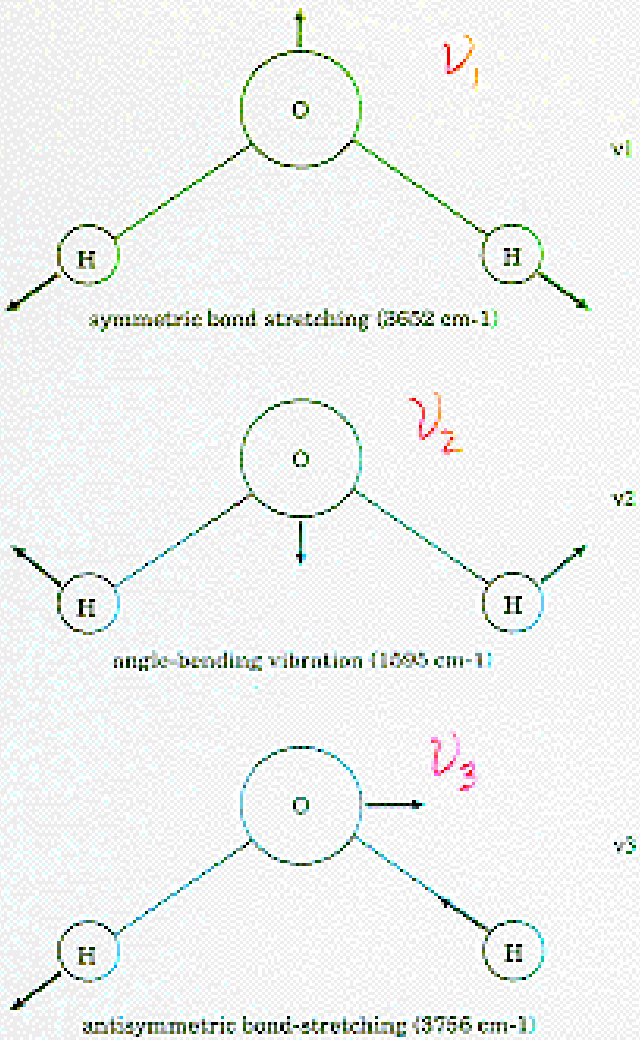
CO_2 Vibration – Rotation Spectra

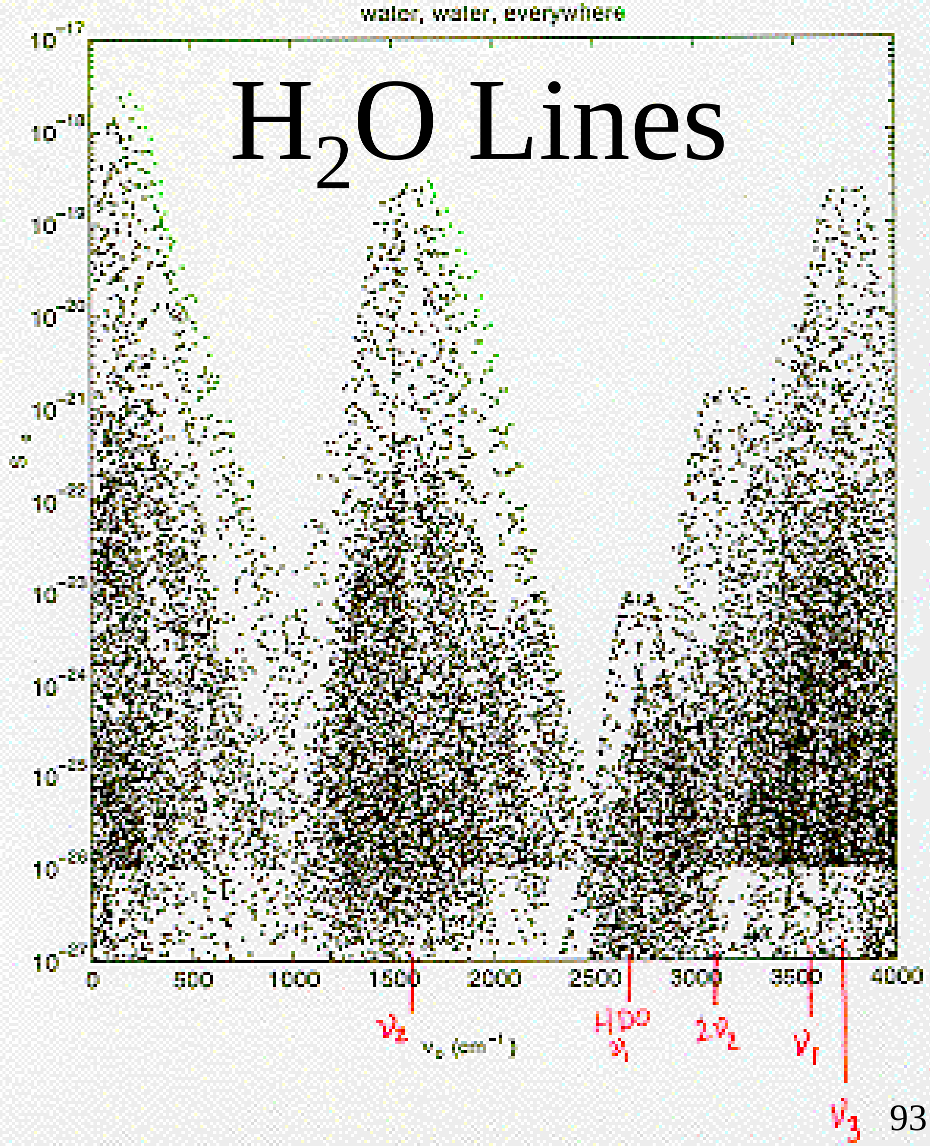
$$E(\upsilon, J) = \underbrace{h\nu(\upsilon + \frac{1}{2}) - xh\nu(\upsilon + \frac{1}{2})^2 + \ldots}_{\text{vibration}} + \underbrace{B_\nu[J(J+1) - \ell^2] - D_\nu[J(J+1) - \ell^2]^2 + \ldots}_{\text{rotation}}$$

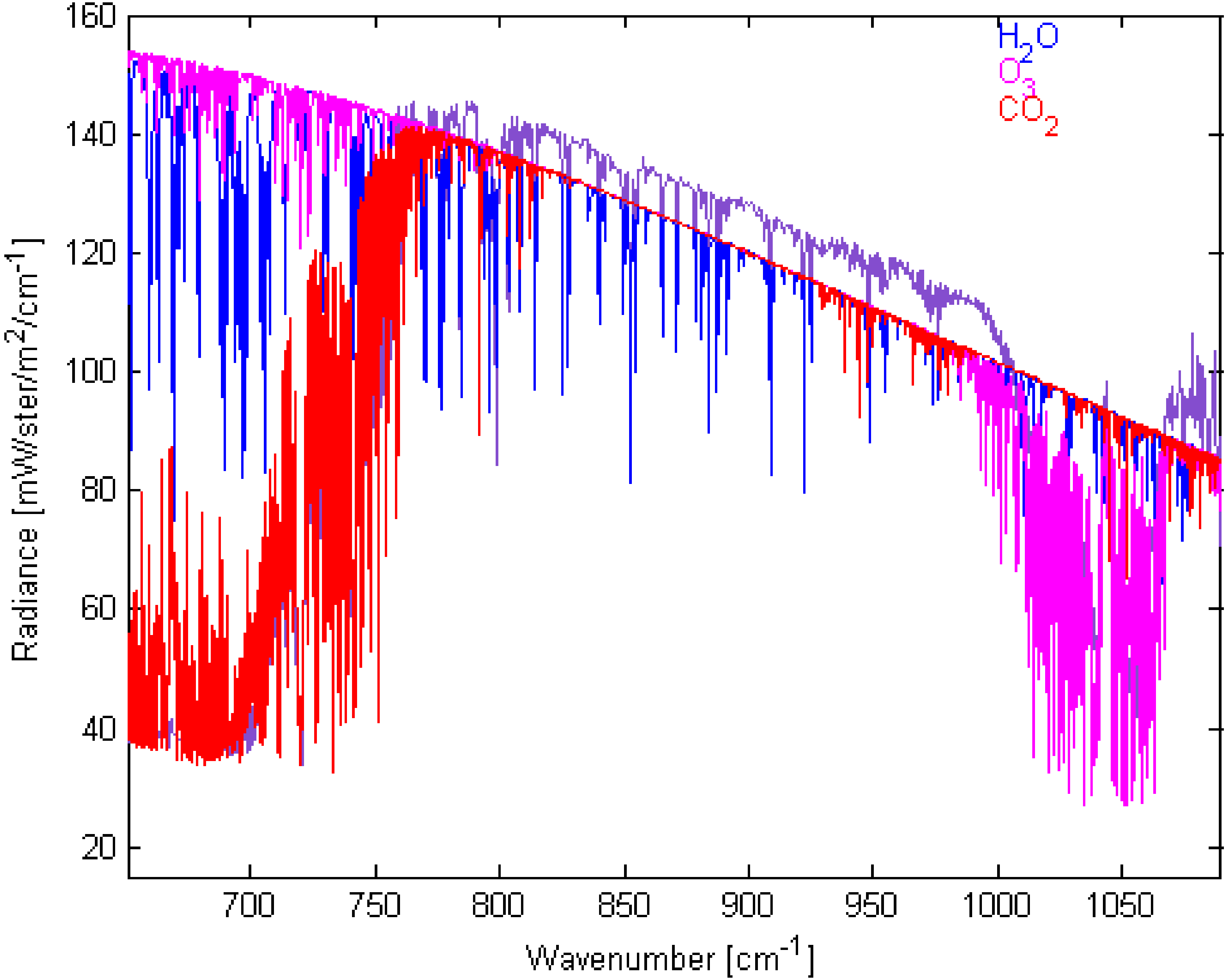




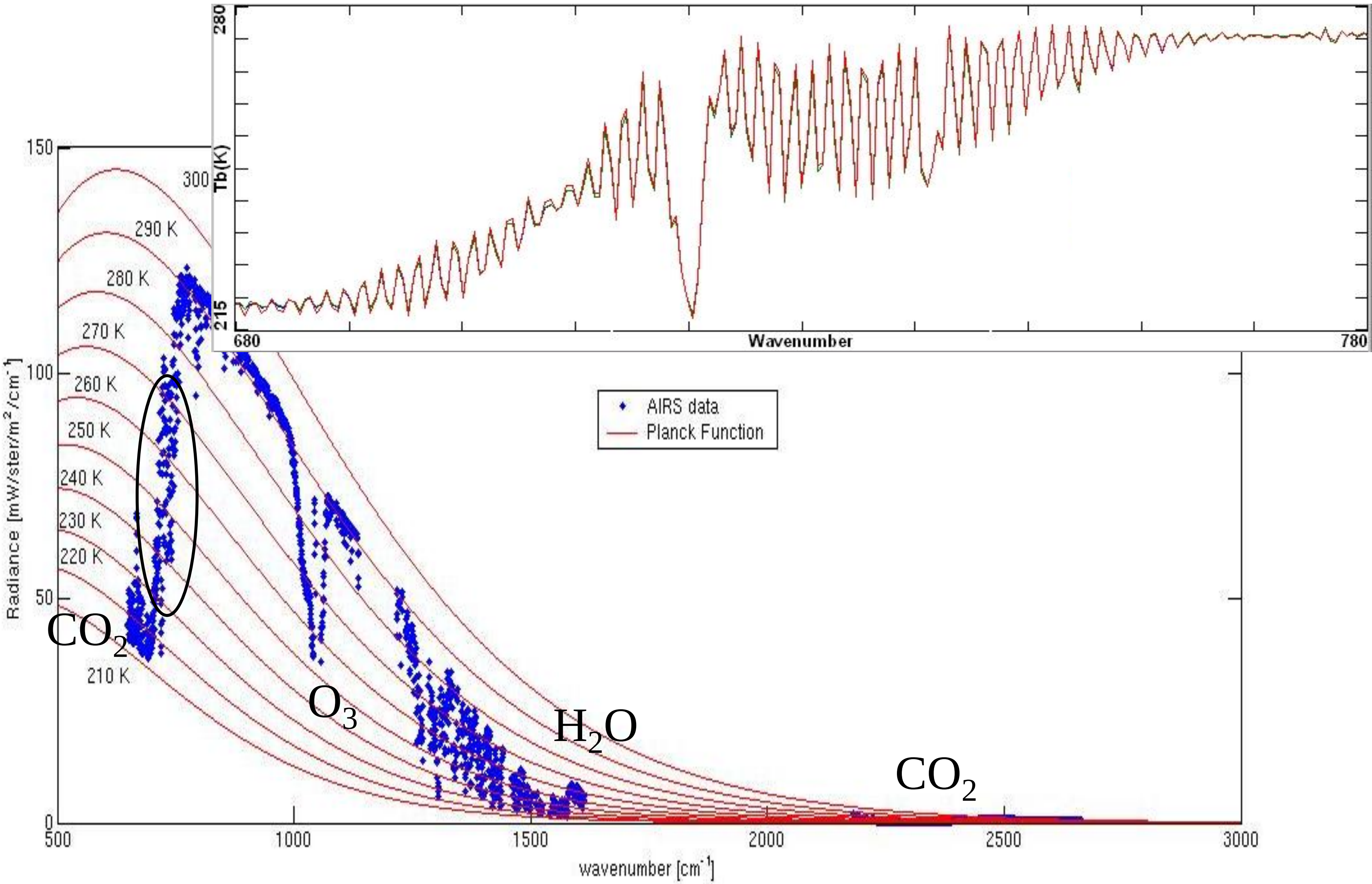
H_2O Vibration – Rotation Spectra



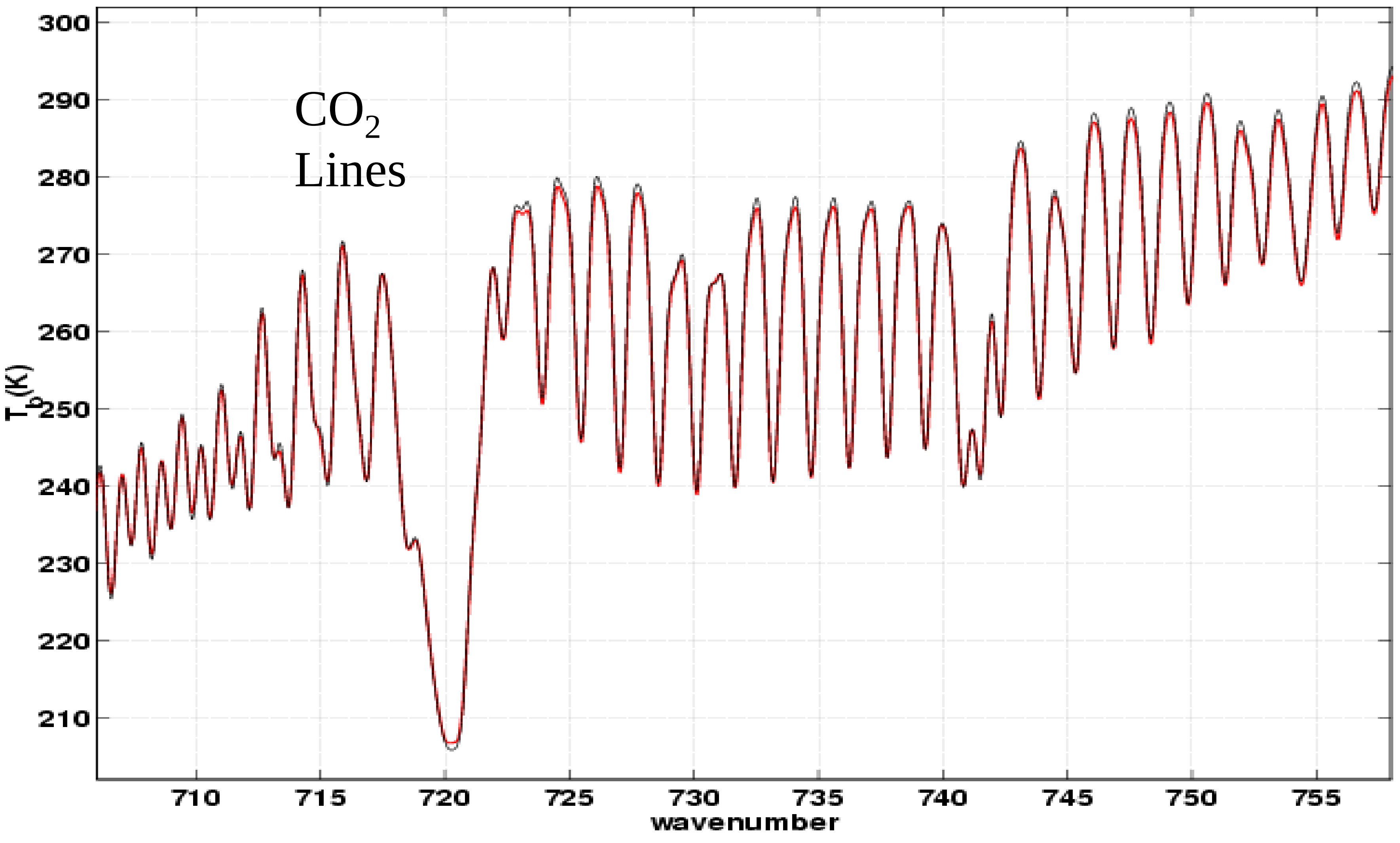



H2O
O3
CO2
Radiance [mW/ster/m2/cm-1]
Wavenumber [cm-1]
160
140
120
100
80
60
40
20
700
750
800
850
900
950
1000
1050

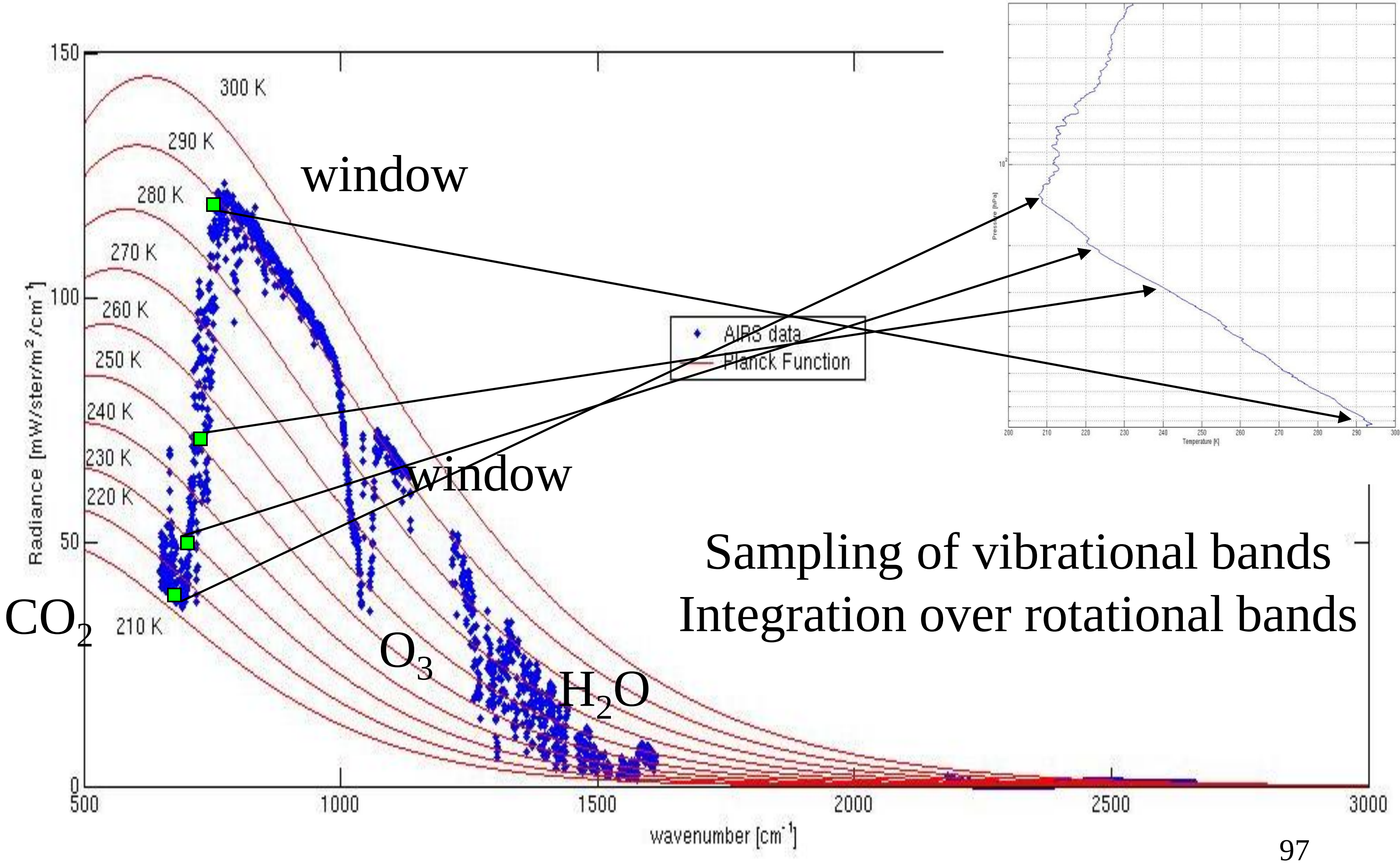
Rotational Lines



Earth emitted spectrum in CO2 sensitive 705 to 760 cm-1

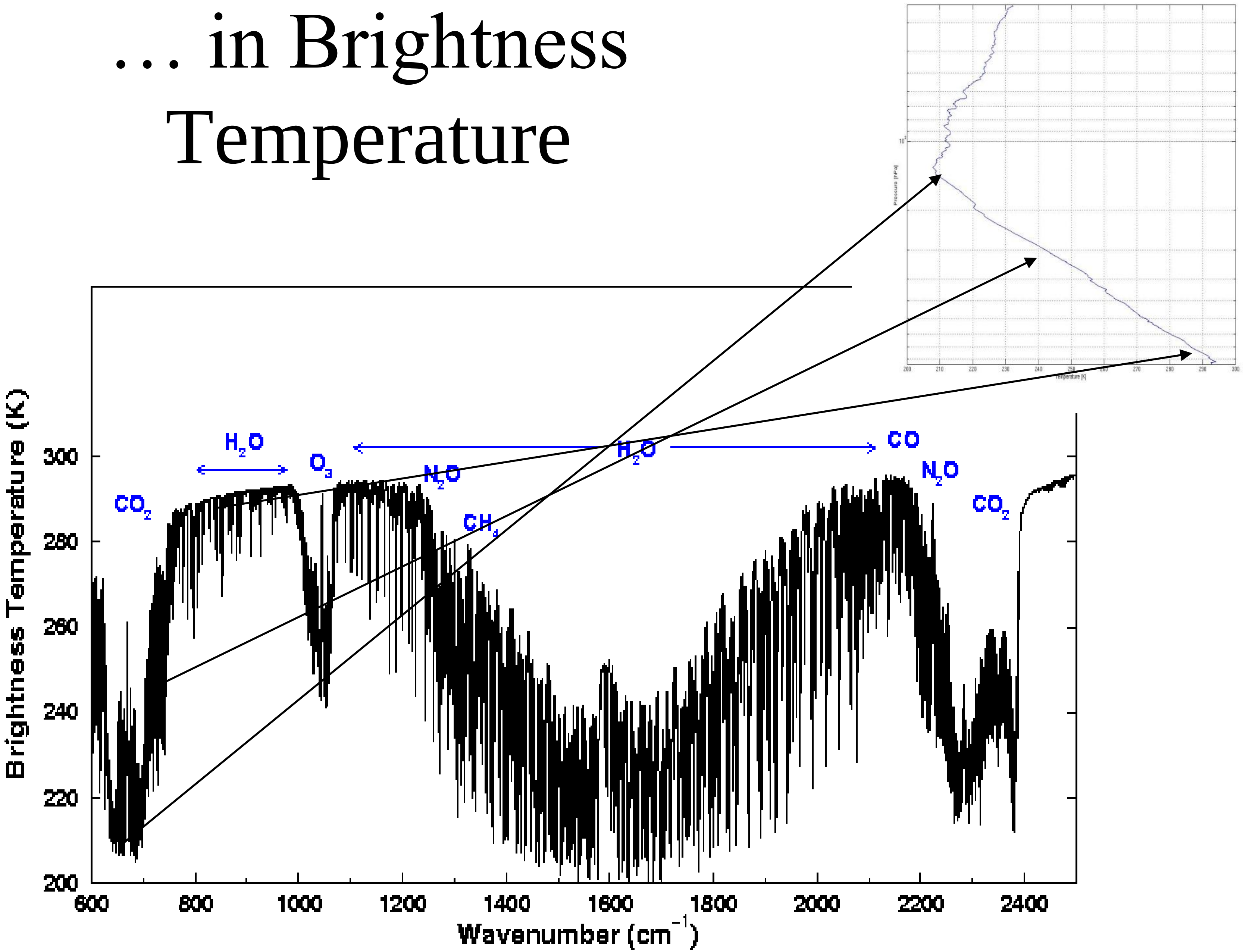


Broad Band

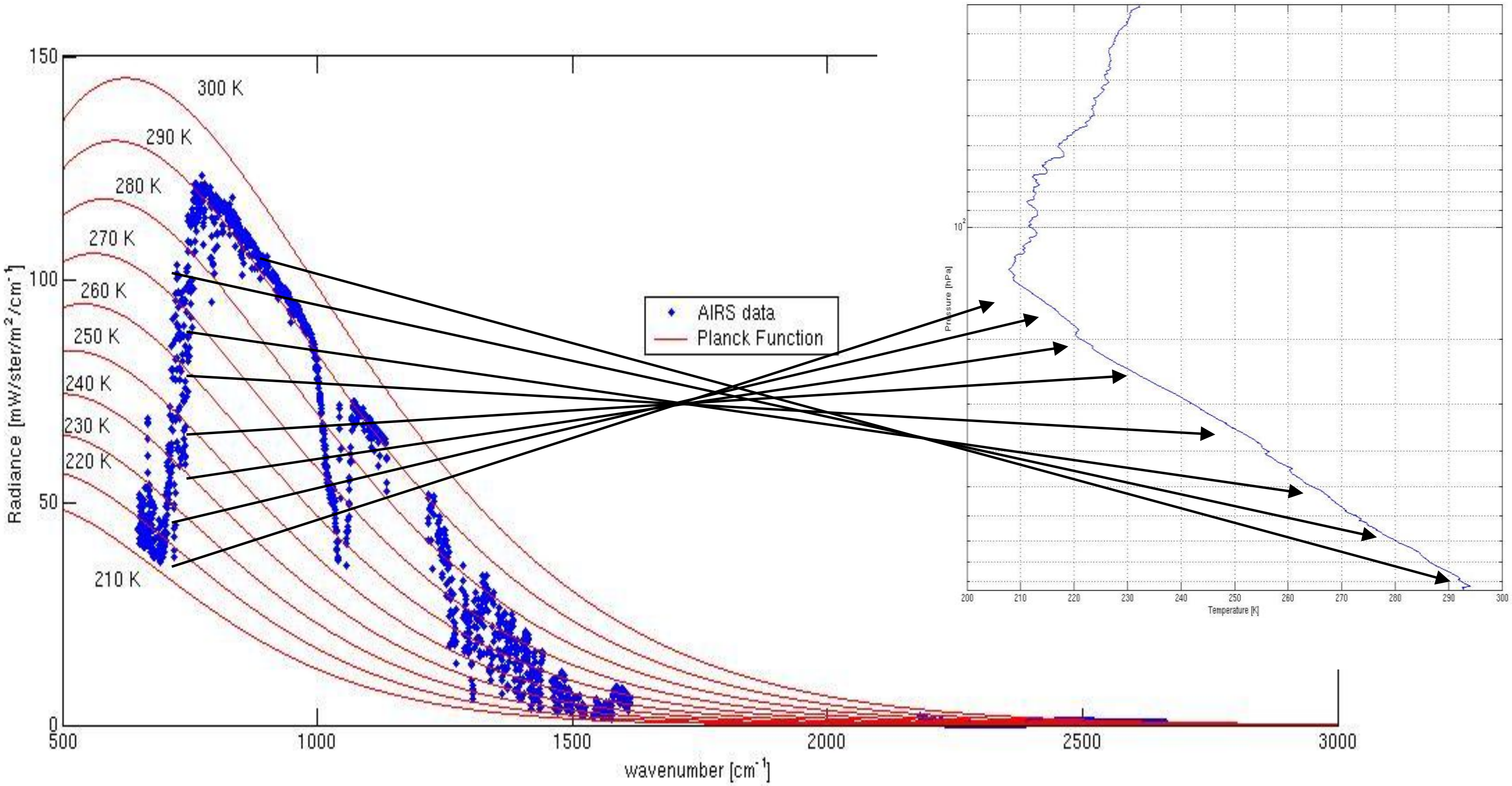


Sampling of vibrational bands
Integration over rotational bands

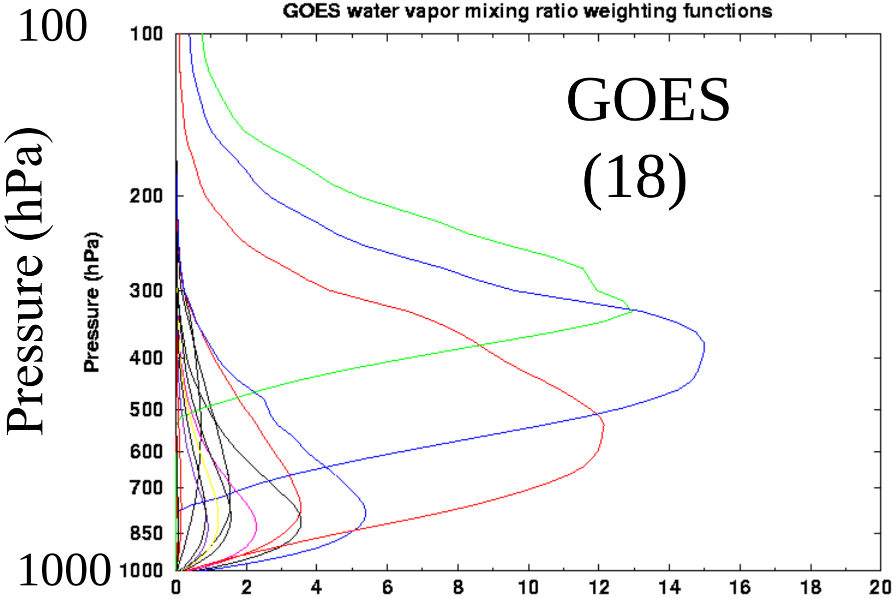
… in Brightness Temperature

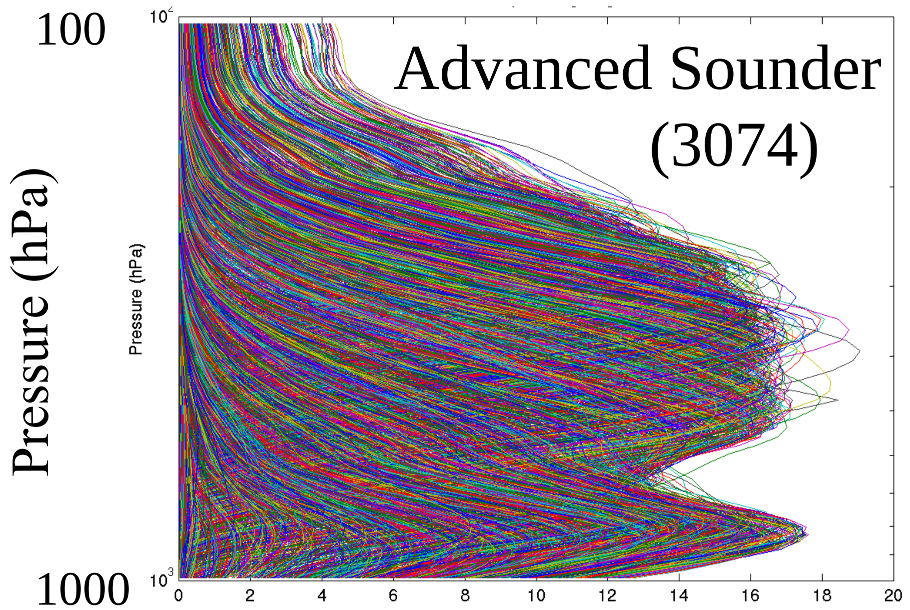


High Spectral Resolution



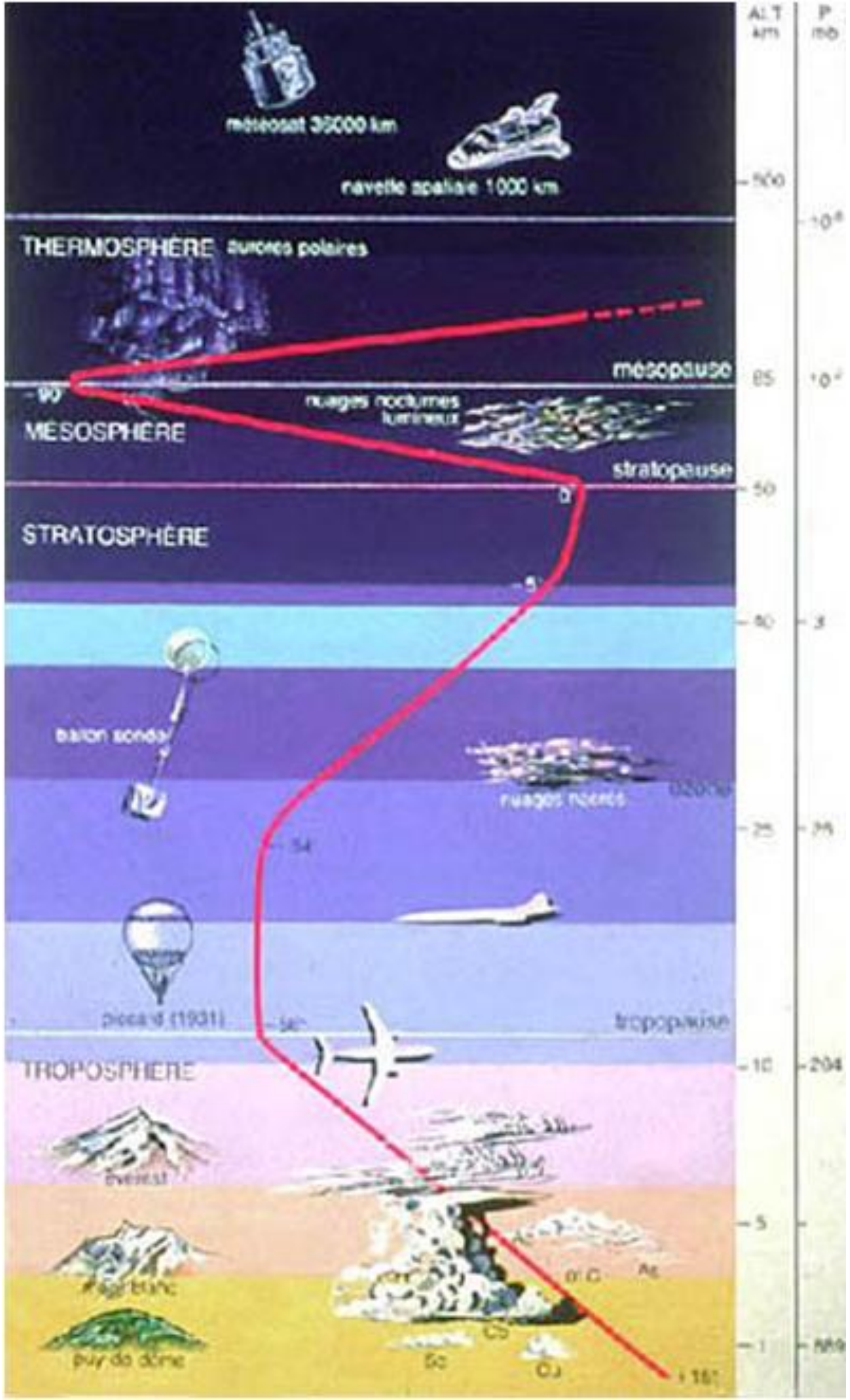
Sampling over rotational bands

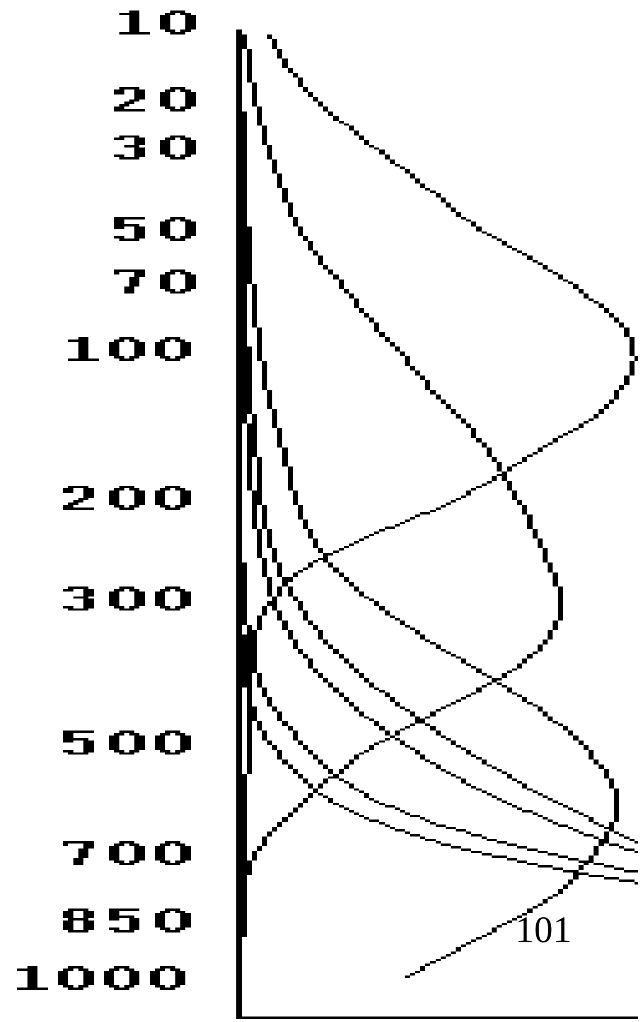


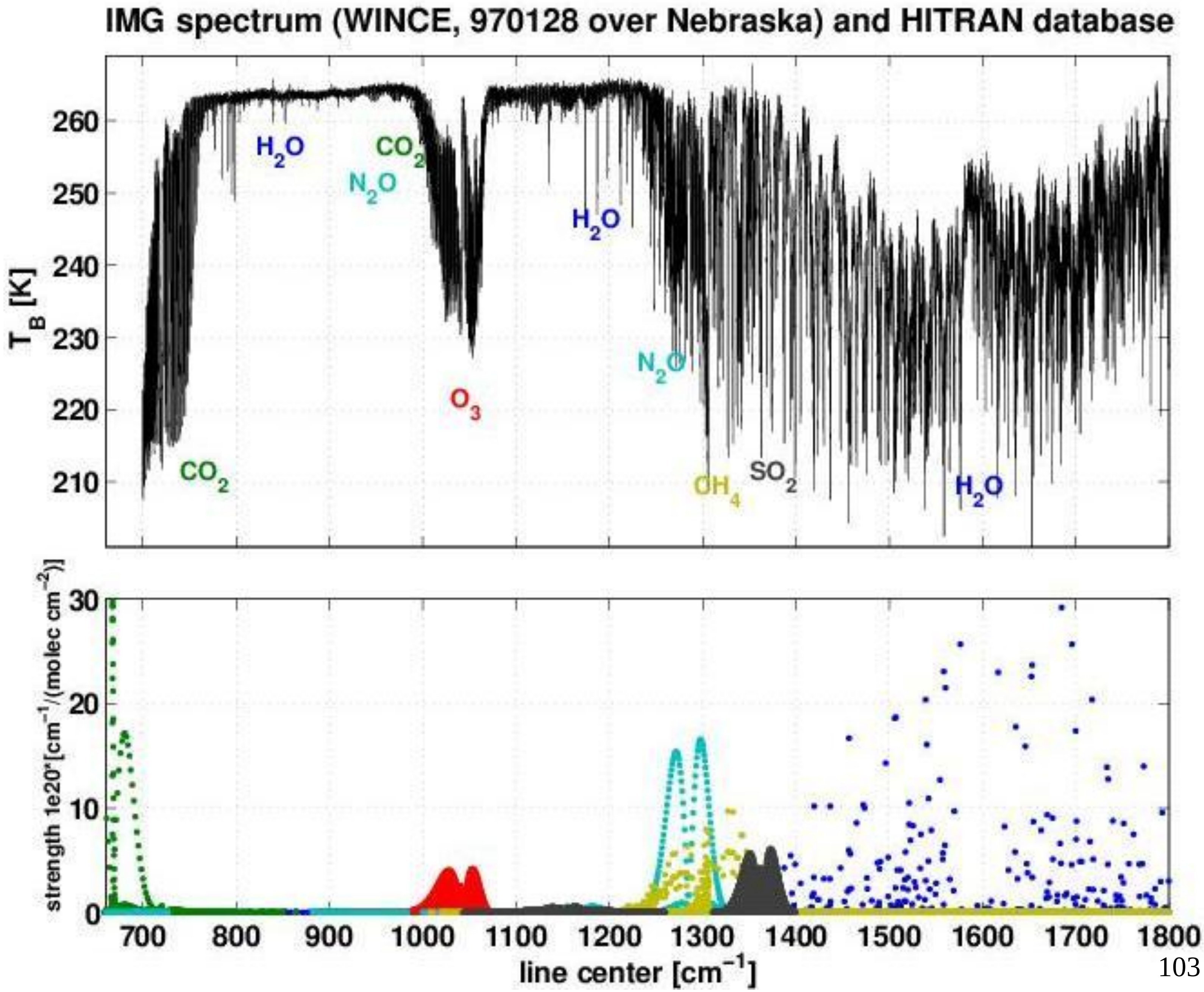


Moisture Weighting Functions

High spectral resolution advanced sounder will have more and sharper weighting functions compared to current GOES sounder. Retrievals will have better vertical resolution.

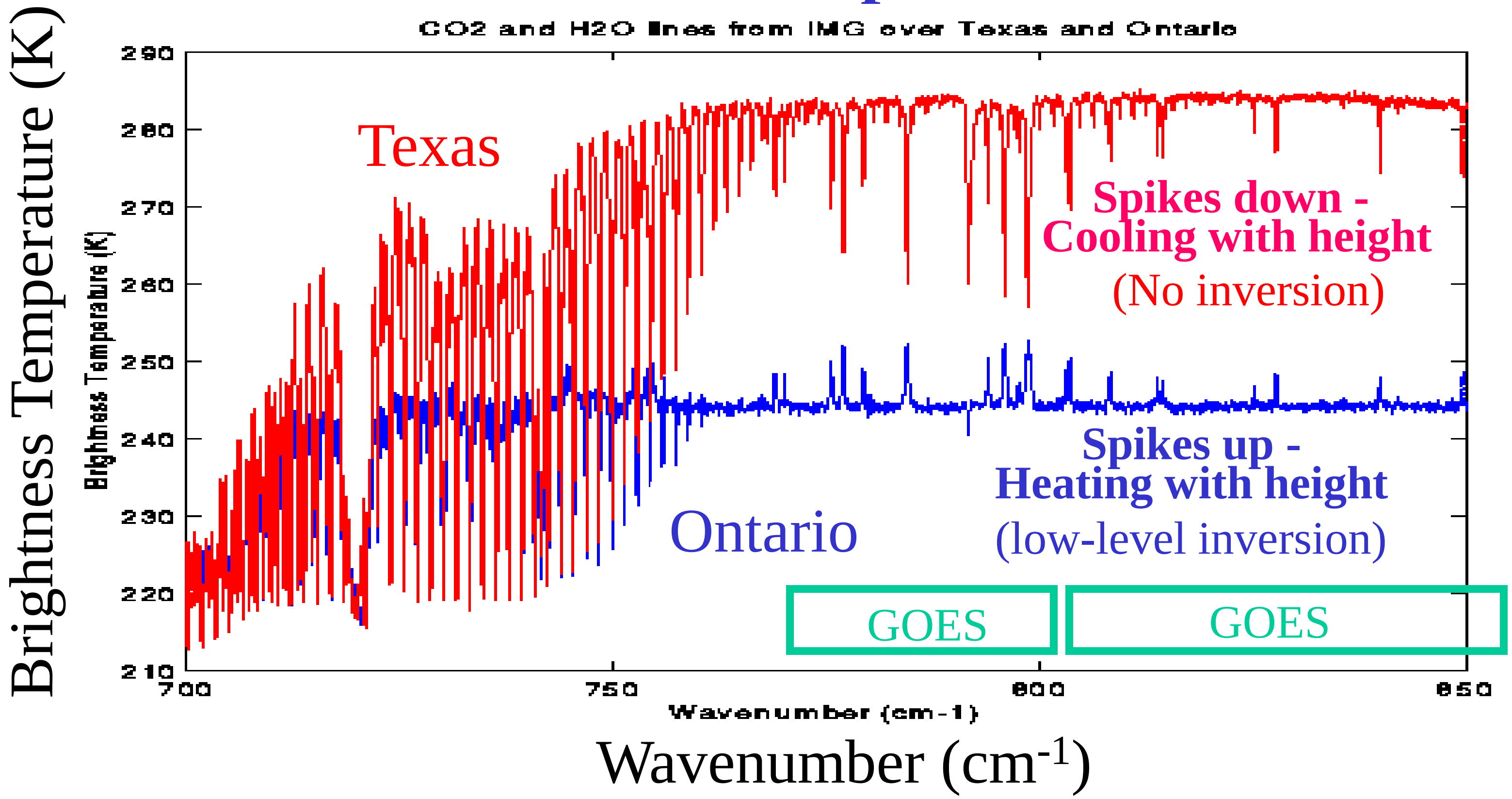

ALT km
P mb
navette spatiale 1000 km
THERMOSPHÈRE
mésopause
MÉSOSPHÈRE
stratopause
STRATOSPHÈRE
ballon sonde
tropopause
TROPOSPHÈRE
puy de dôme


10
20
30
50
70
100
200
300
500
700
850
1000



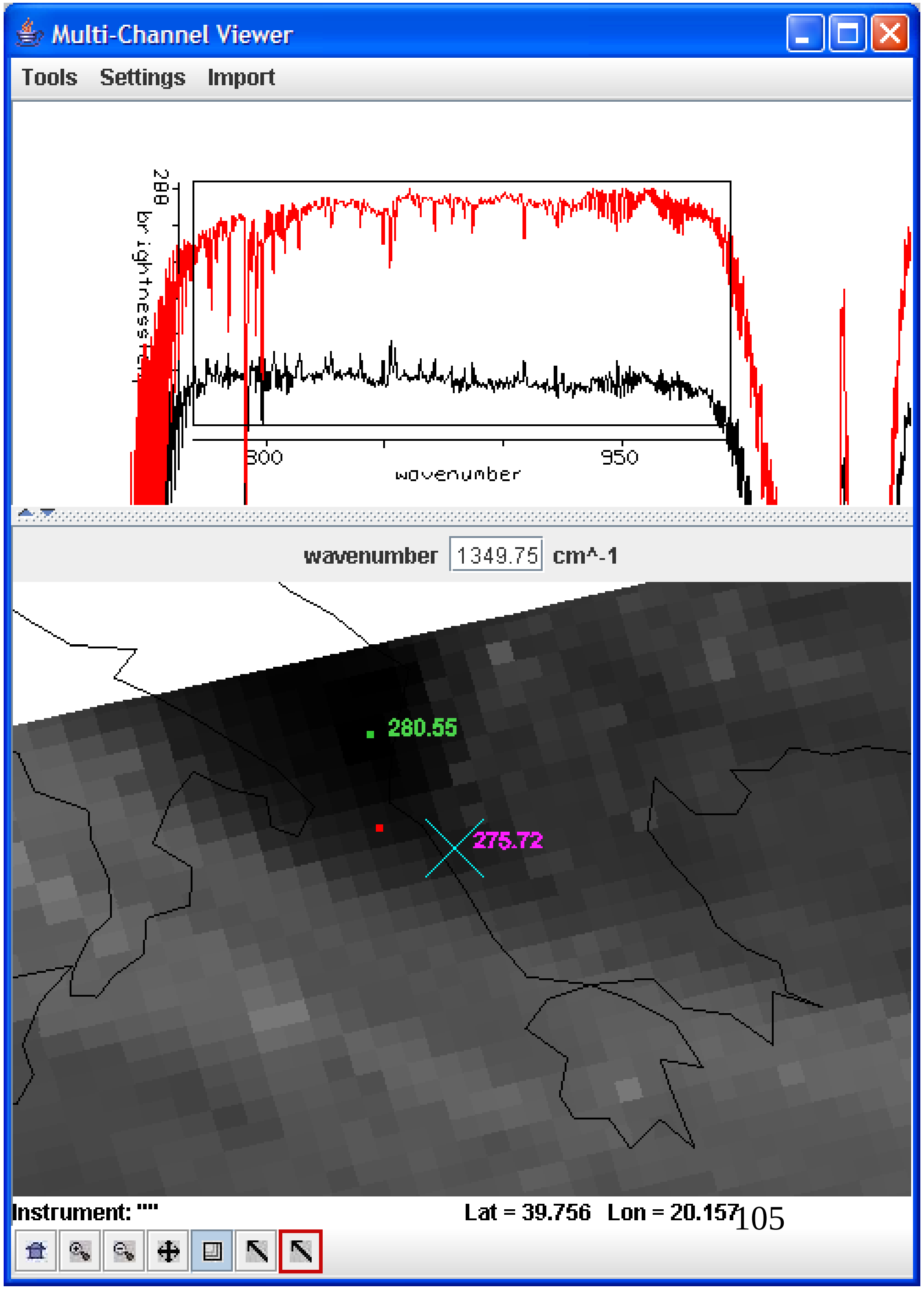
IMG spectrum (WINCE, 970128 over Nebraska) and HITRAN database
T_B [K]
260
250
240
230
220
210
H2O
CO2
N2O
H2O
N2O
O3
CO2
CH4
SO2
H2O
strength 1e20*[cm^-1/(molec cm^-2)]
30
20
10
0
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
1800
line center [cm^-1]

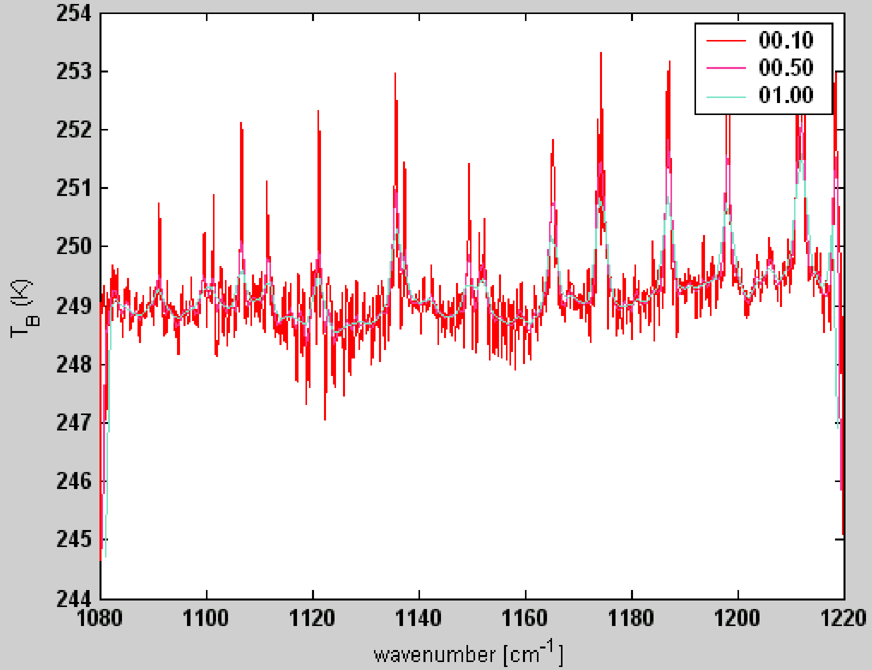
Resolving absorption features in atmospheric windows enables detection of temperature inversions

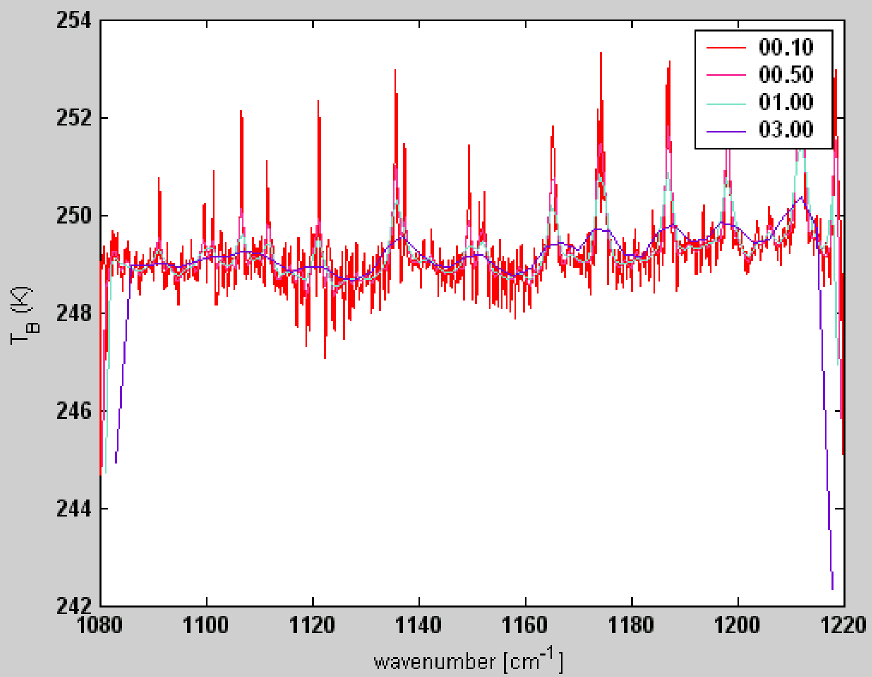


Detection of inversions is critical for severe weather forecasting. Combined with improved low-level moisture depiction, key ingredients for night-time severe storm development can be monitored.

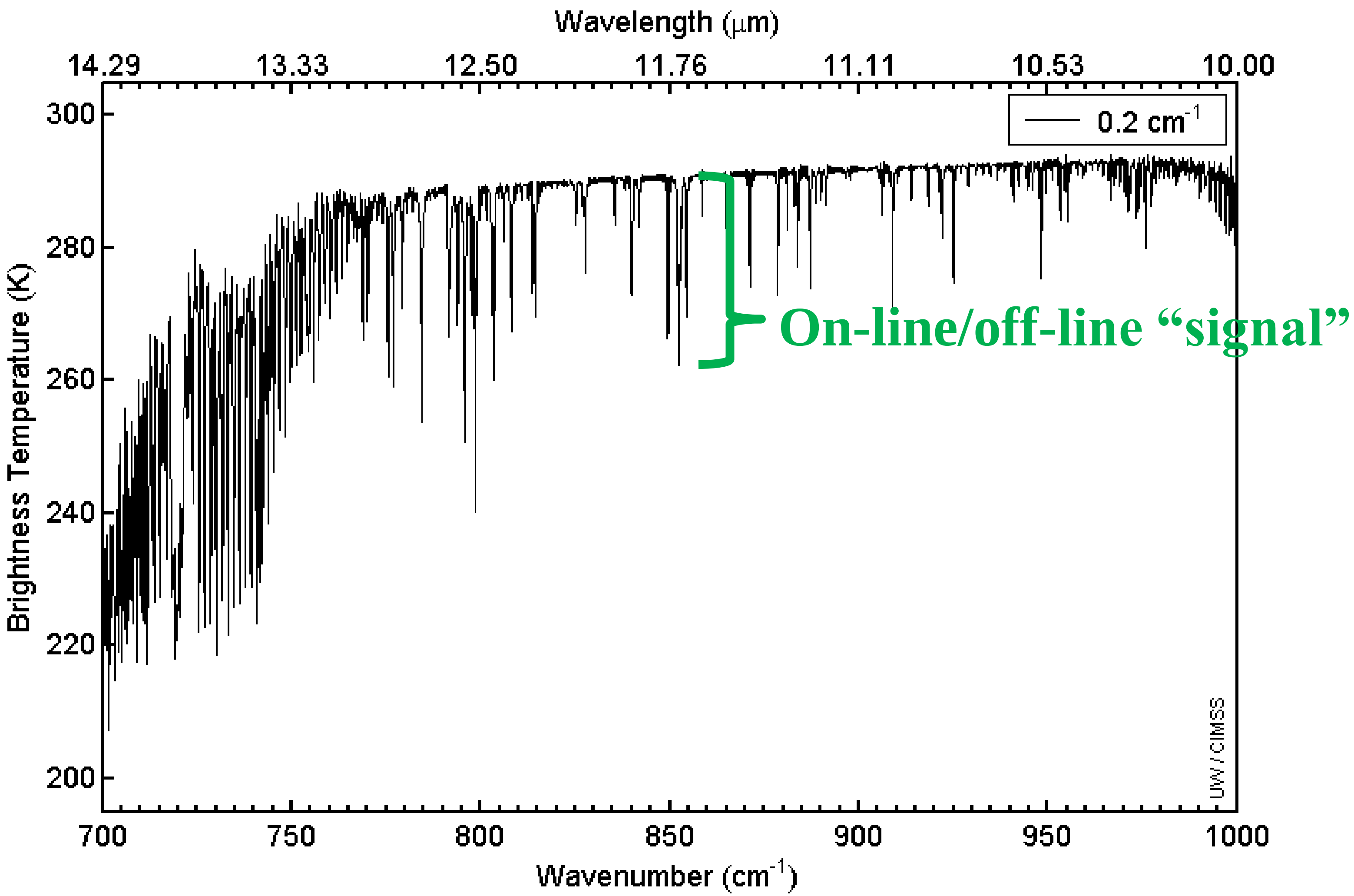
IASI detection of temperature inversion (black spectrum) vs clear ocean (red spectrum)



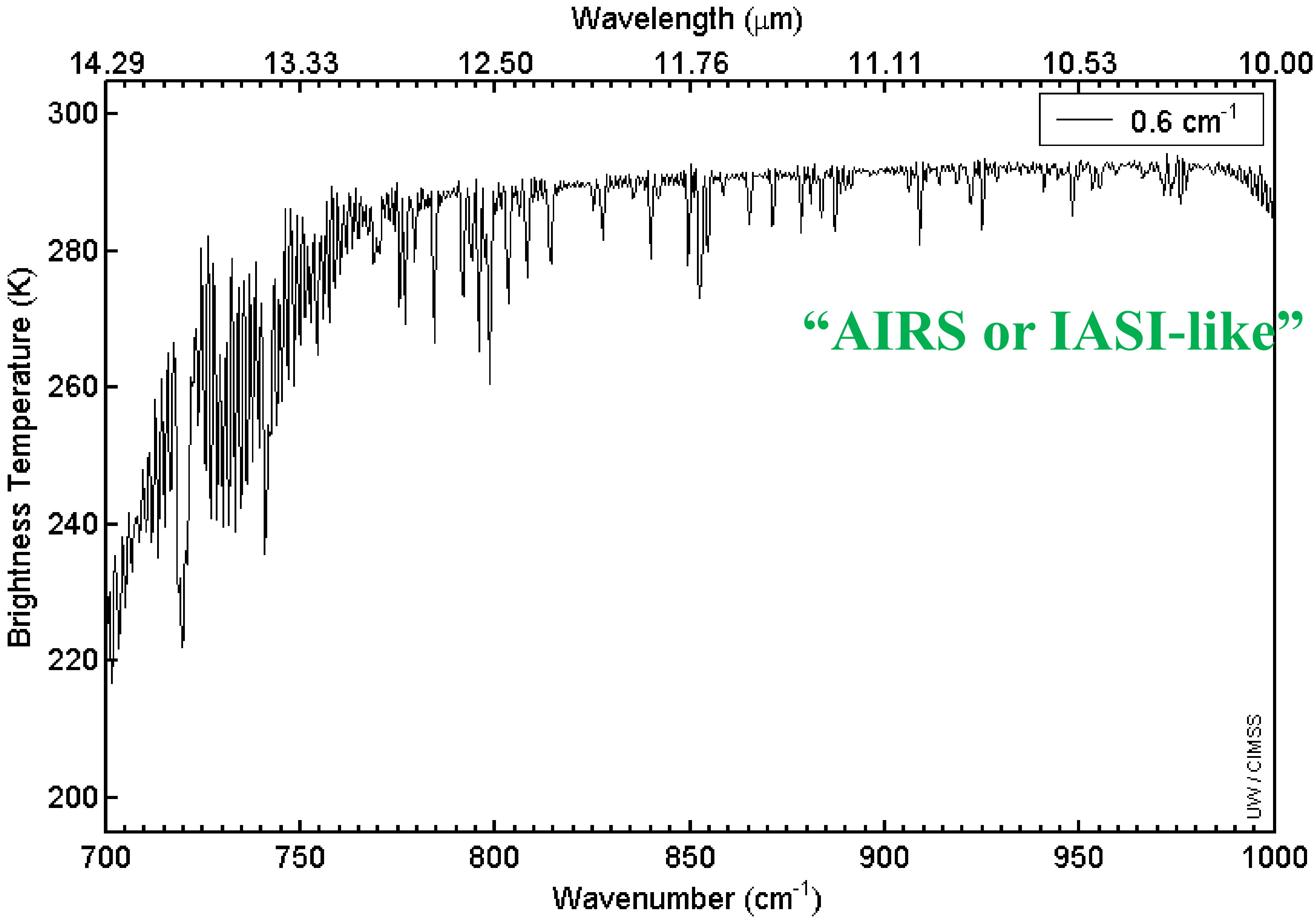




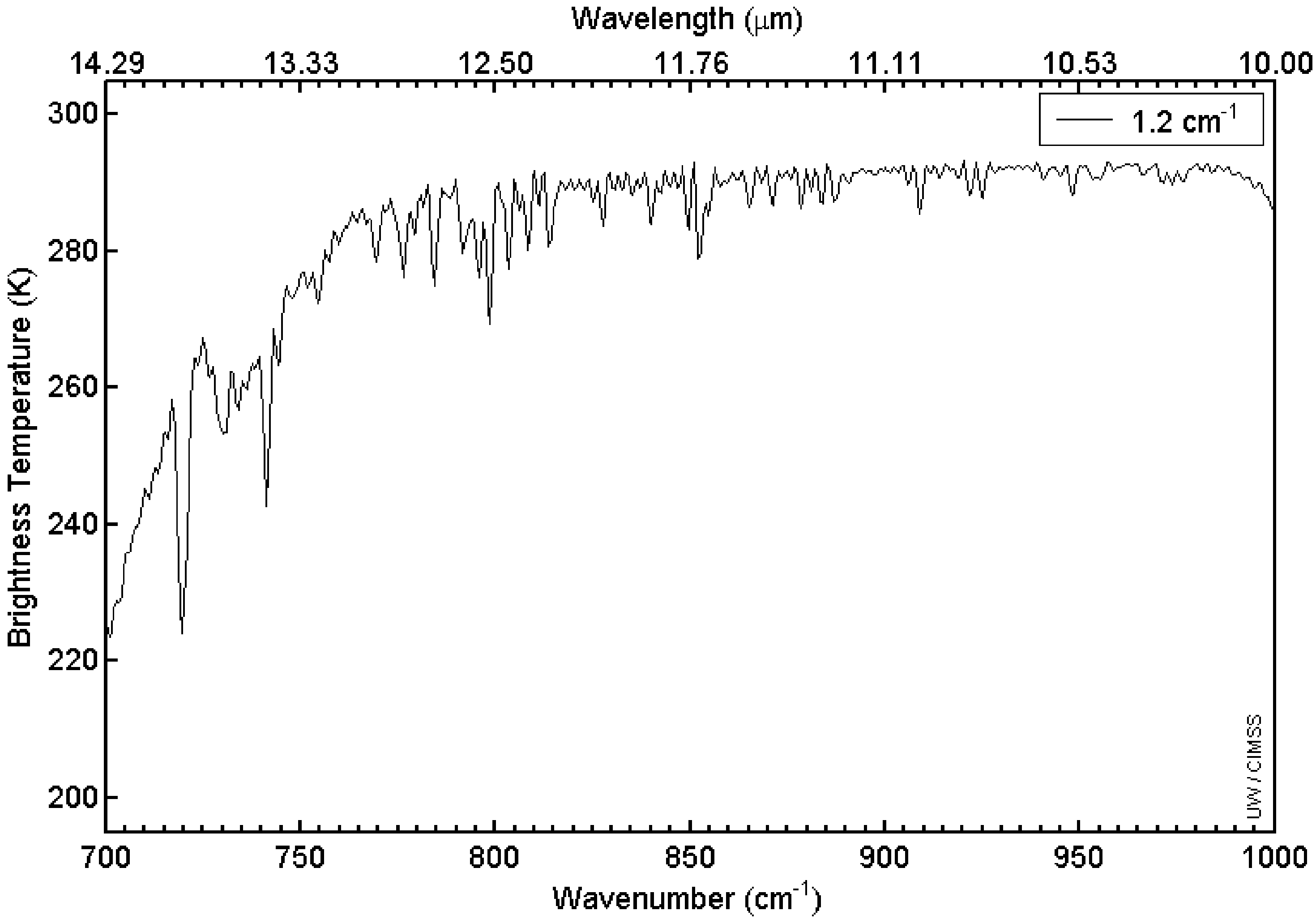
Ability to detect inversions disappears with broadband observations (> 3 cm-1)



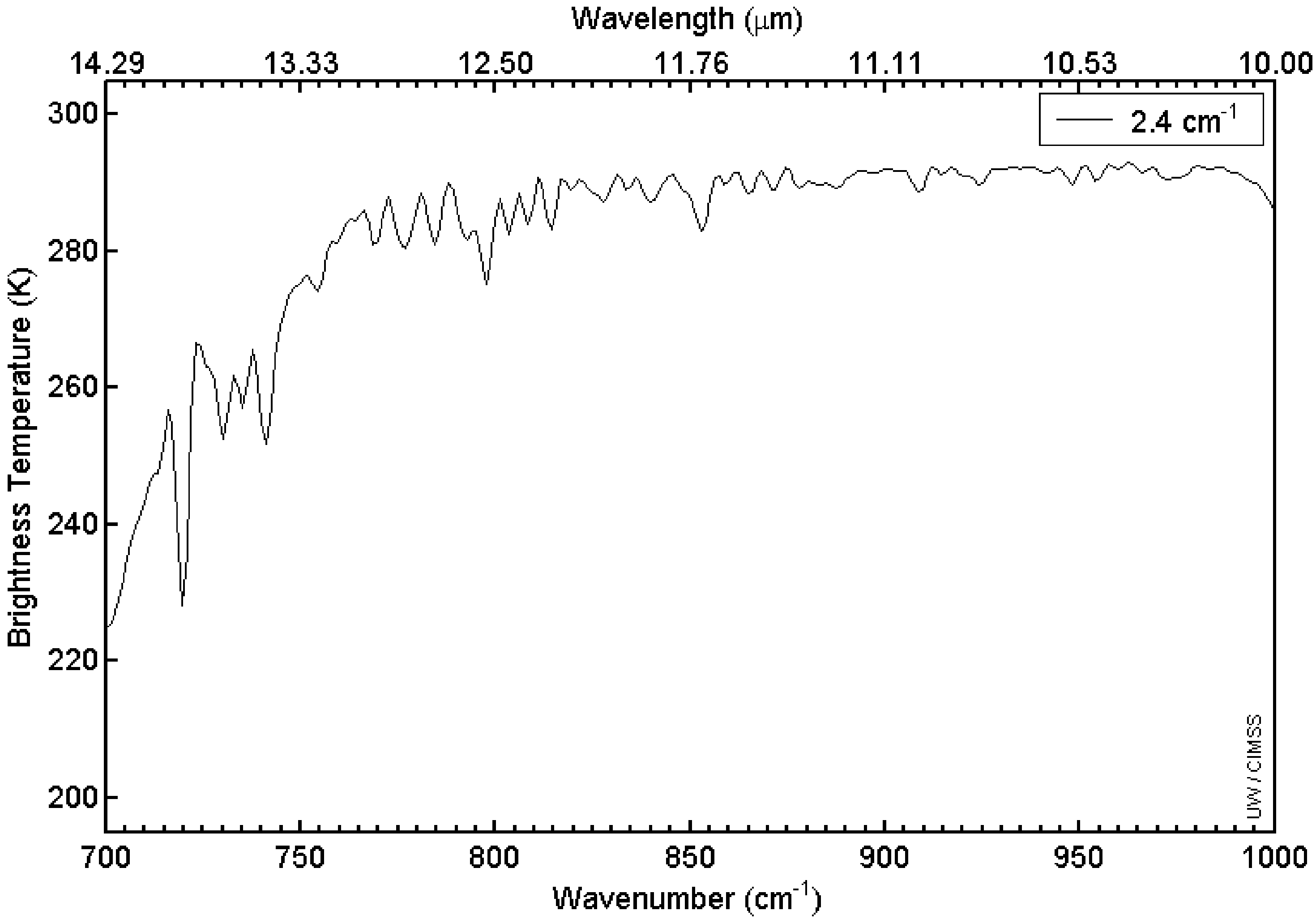
Longwave window region



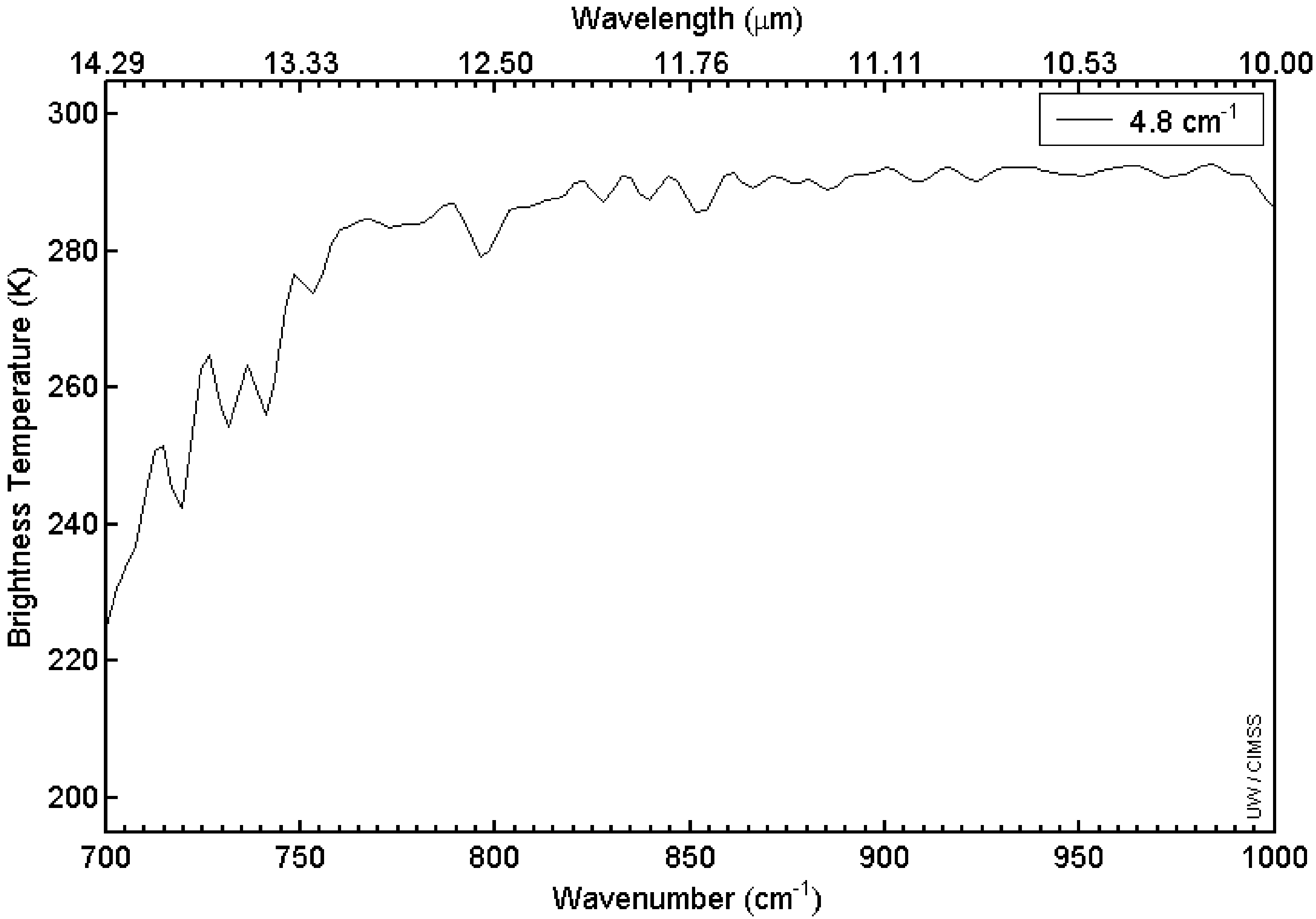
Longwave window region



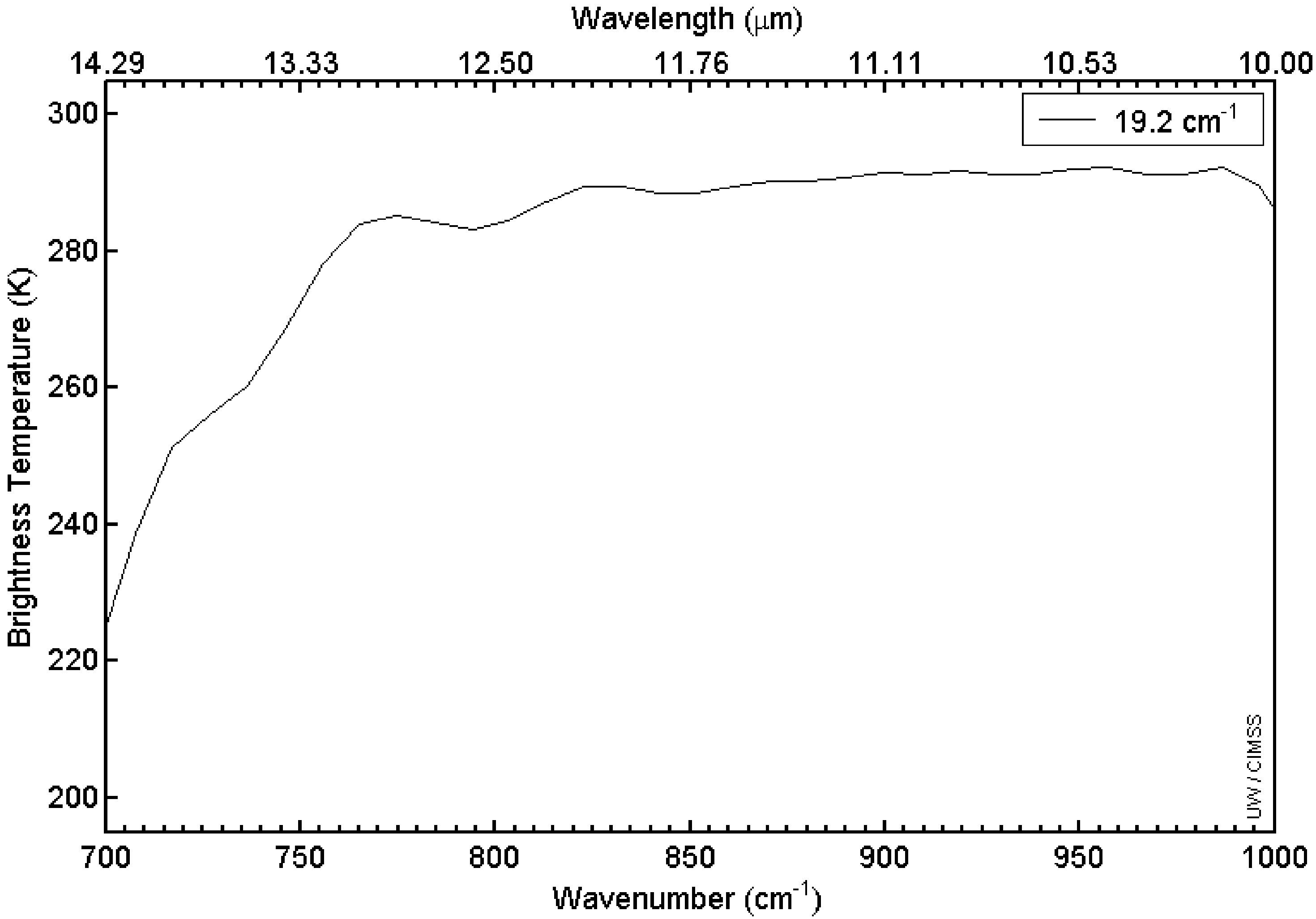
Longwave window region



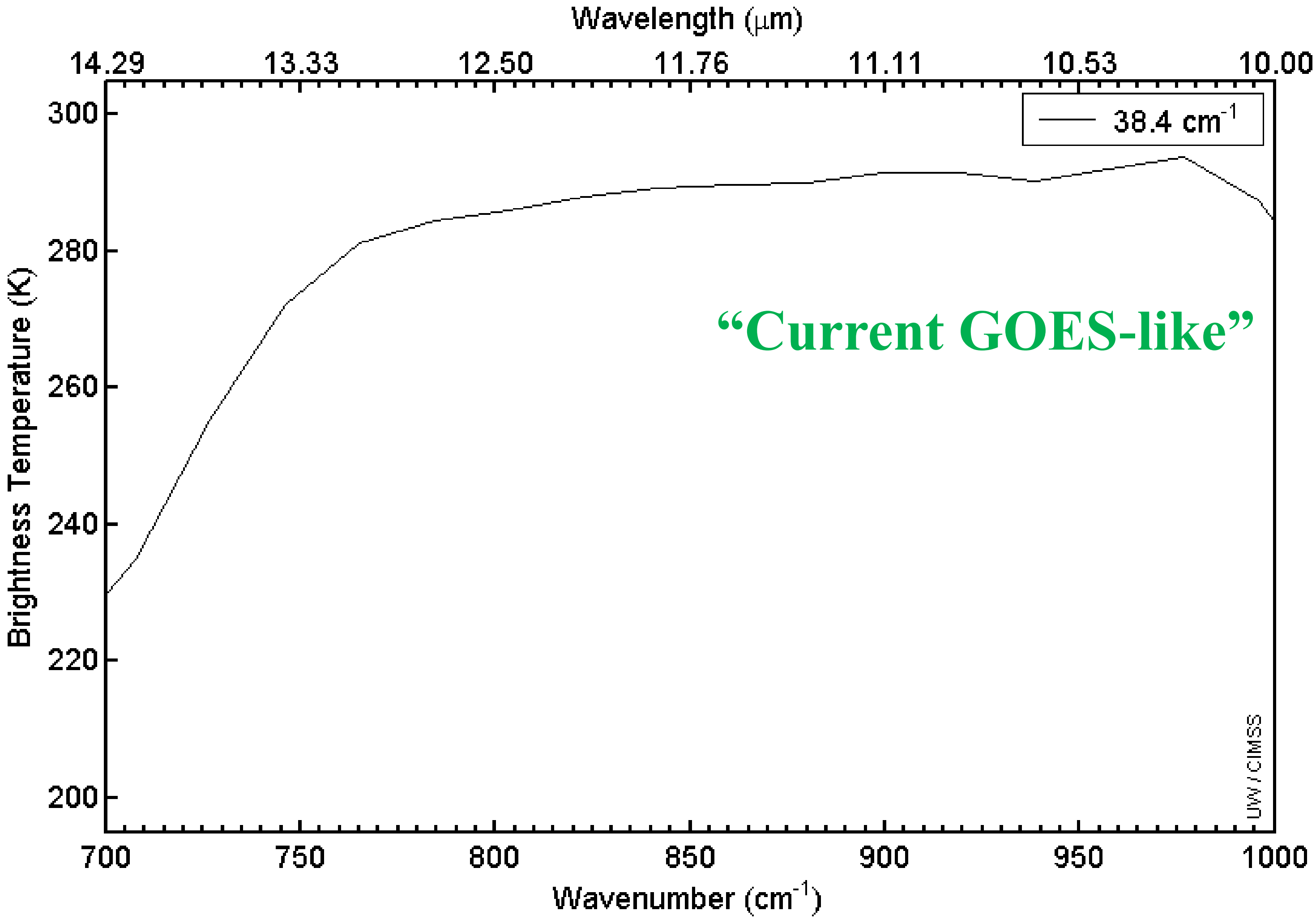
Longwave window region



Longwave window region



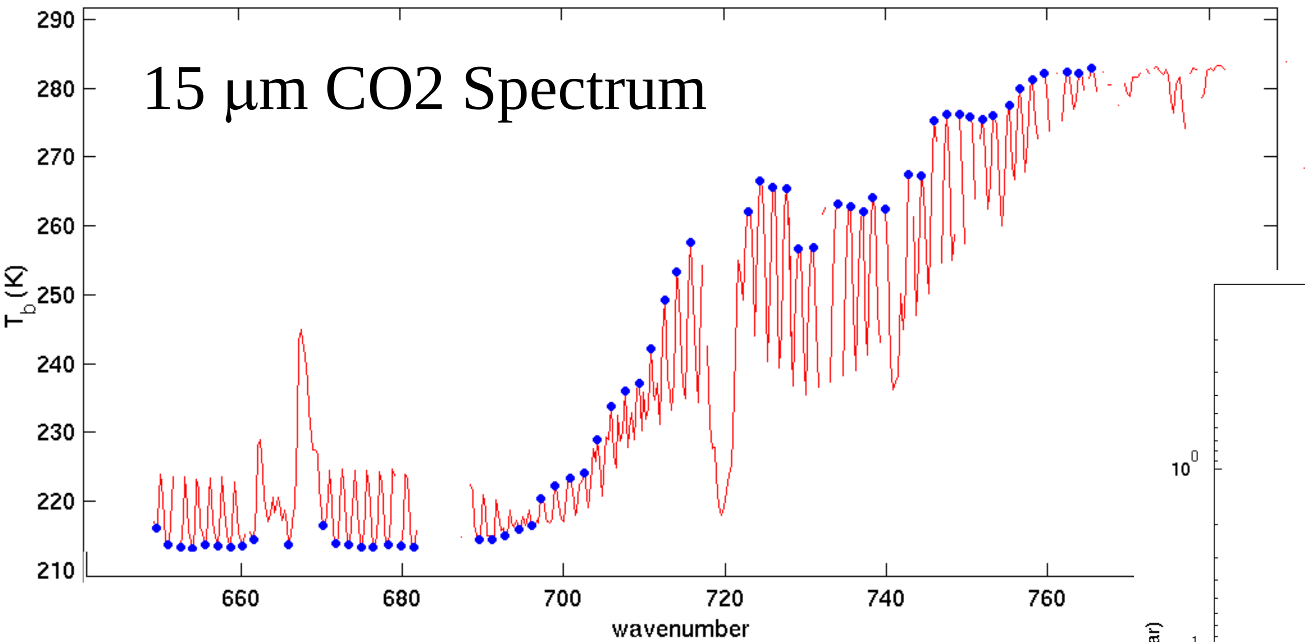
Longwave window region



Longwave window region

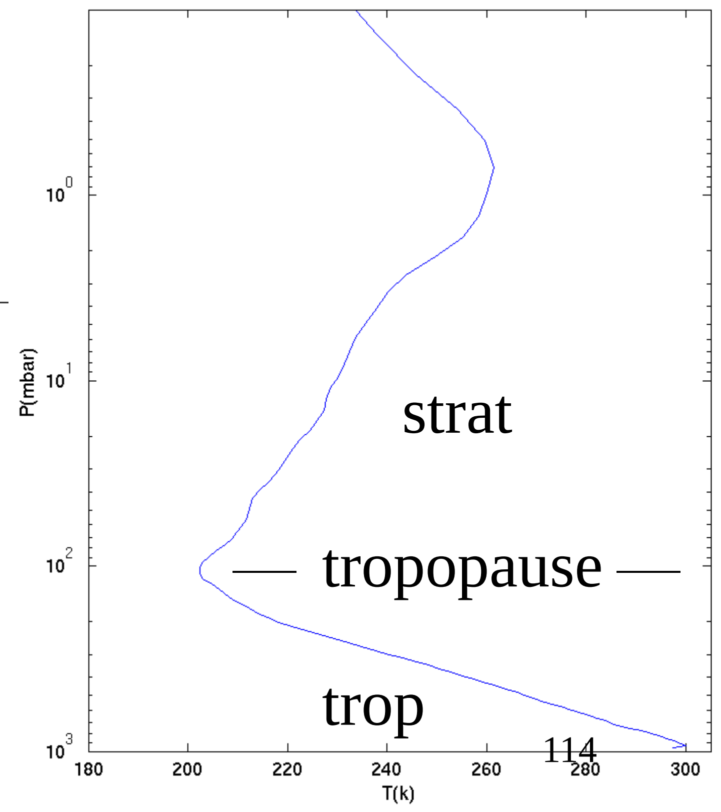
Twisted Ribbon formed by CO_2 spectrum:

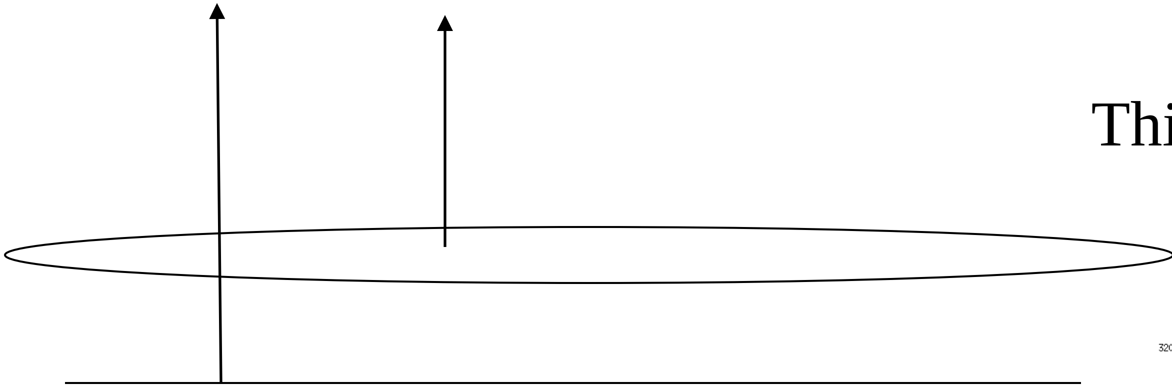
Tropopause inversion causes On-line & off-line patterns to cross



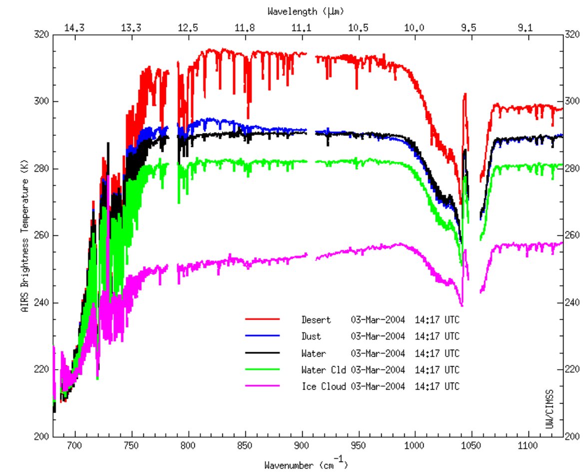
Blue between-line T_b
warmer for tropospheric channels,
colder for stratospheric channels

Signature not available at low resolution



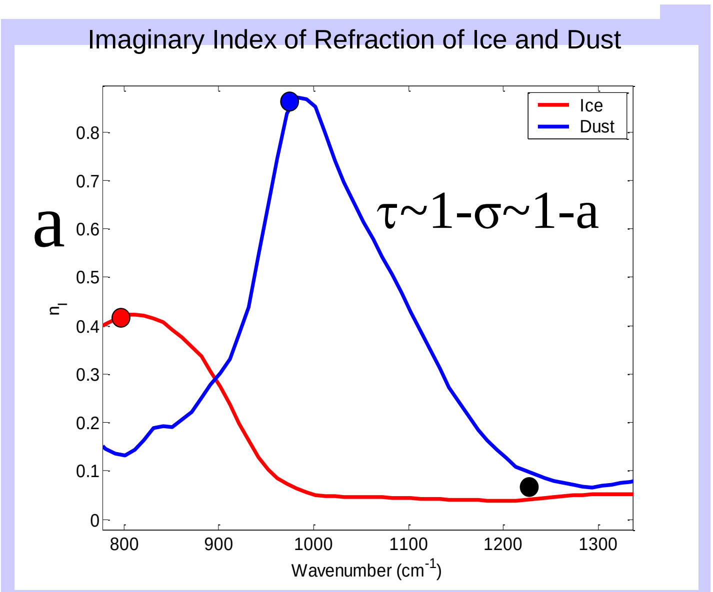


Thin ice cloud over ocean



$R = \boldsymbol{\varepsilon}_s B_s (1-\boldsymbol{\sigma}_c) + \boldsymbol{\sigma}_c B_c$ using $e^{-\sigma} = 1 - \sigma$

So difference of thin ice cloud over ocean
from clear sky over ocean is given by

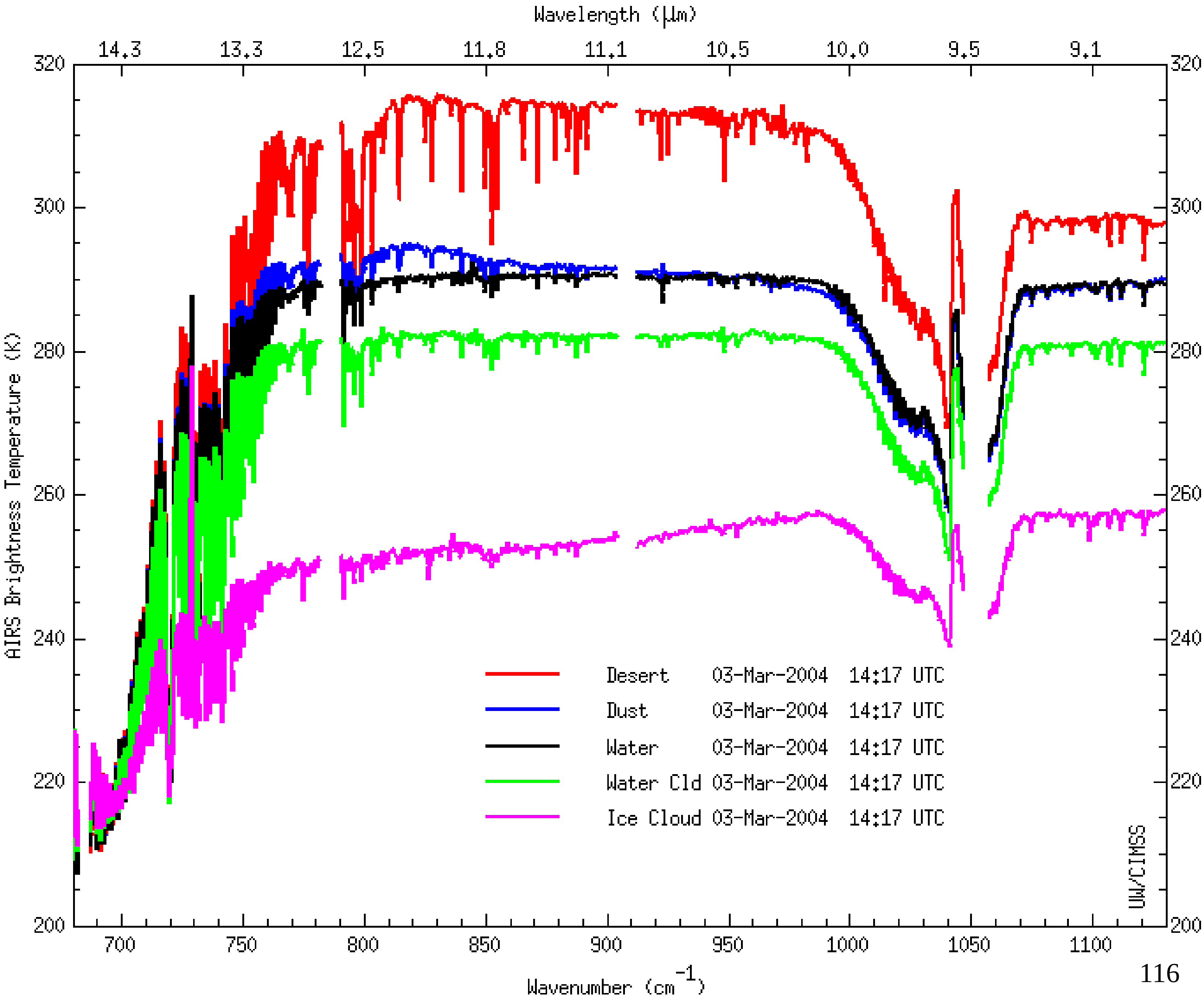


$\Delta R = - \boldsymbol{\varepsilon}_s \boldsymbol{\sigma}_c B_s + \boldsymbol{\sigma}_c B_c$

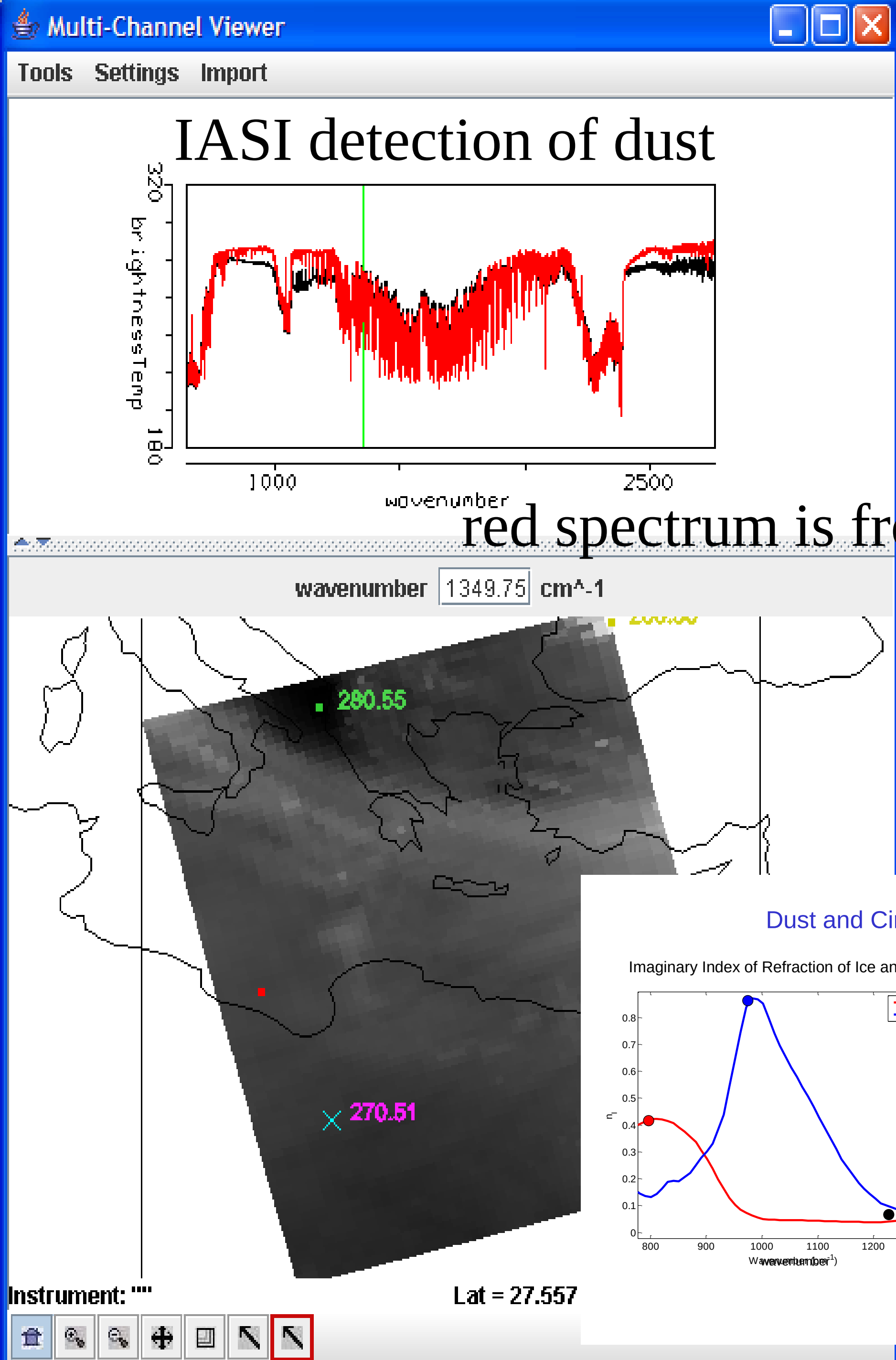
For $B_s > B_c$ and $\boldsymbol{\varepsilon}_s \sim 1$

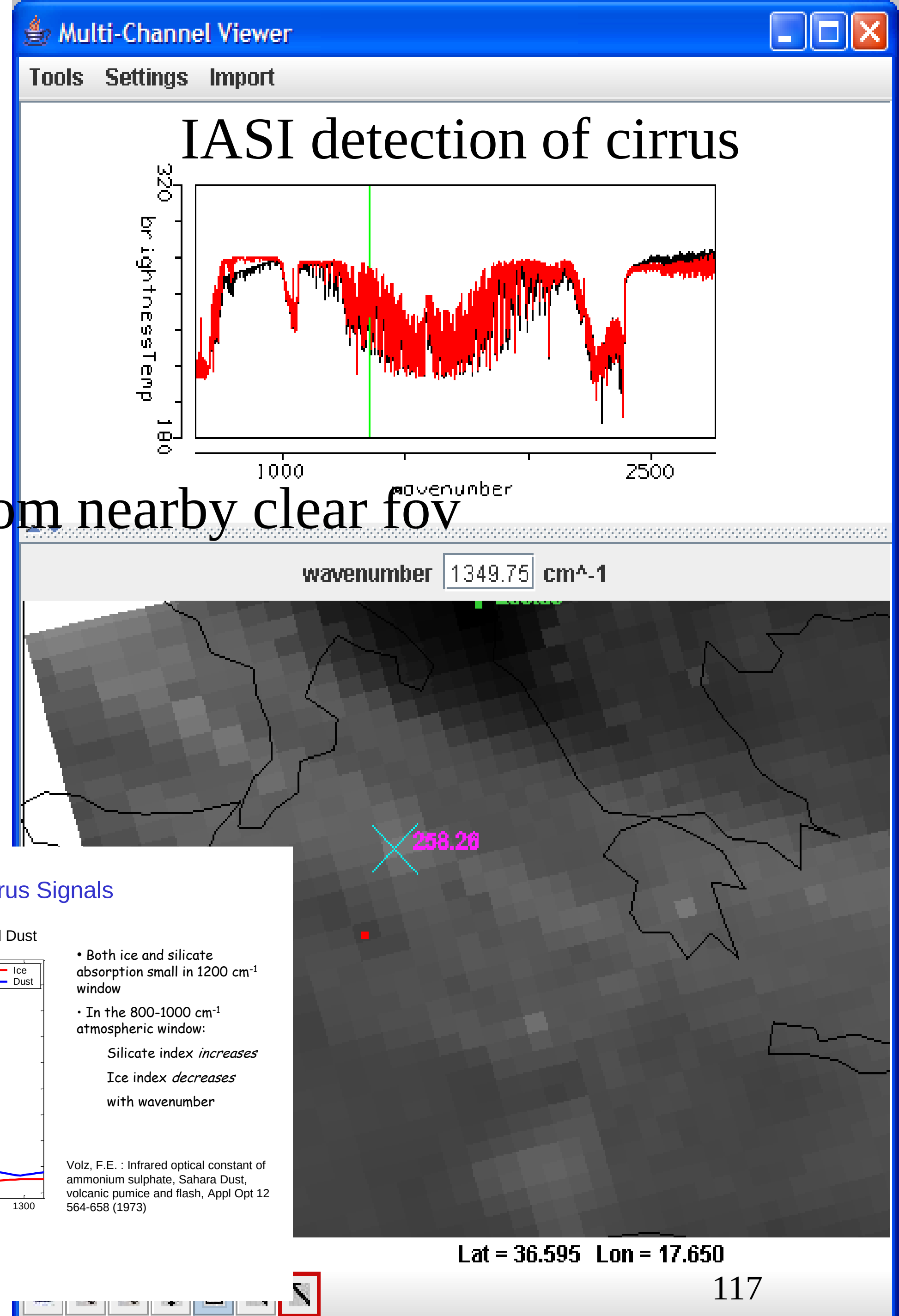
$\Delta R = - \boldsymbol{\sigma}_c B_s + \boldsymbol{\sigma}_c B_c = \boldsymbol{\sigma}_c [B_c - B_s]$

As $\boldsymbol{\sigma}_c$ increases (decreases) then ΔR becomes more negative (positive)

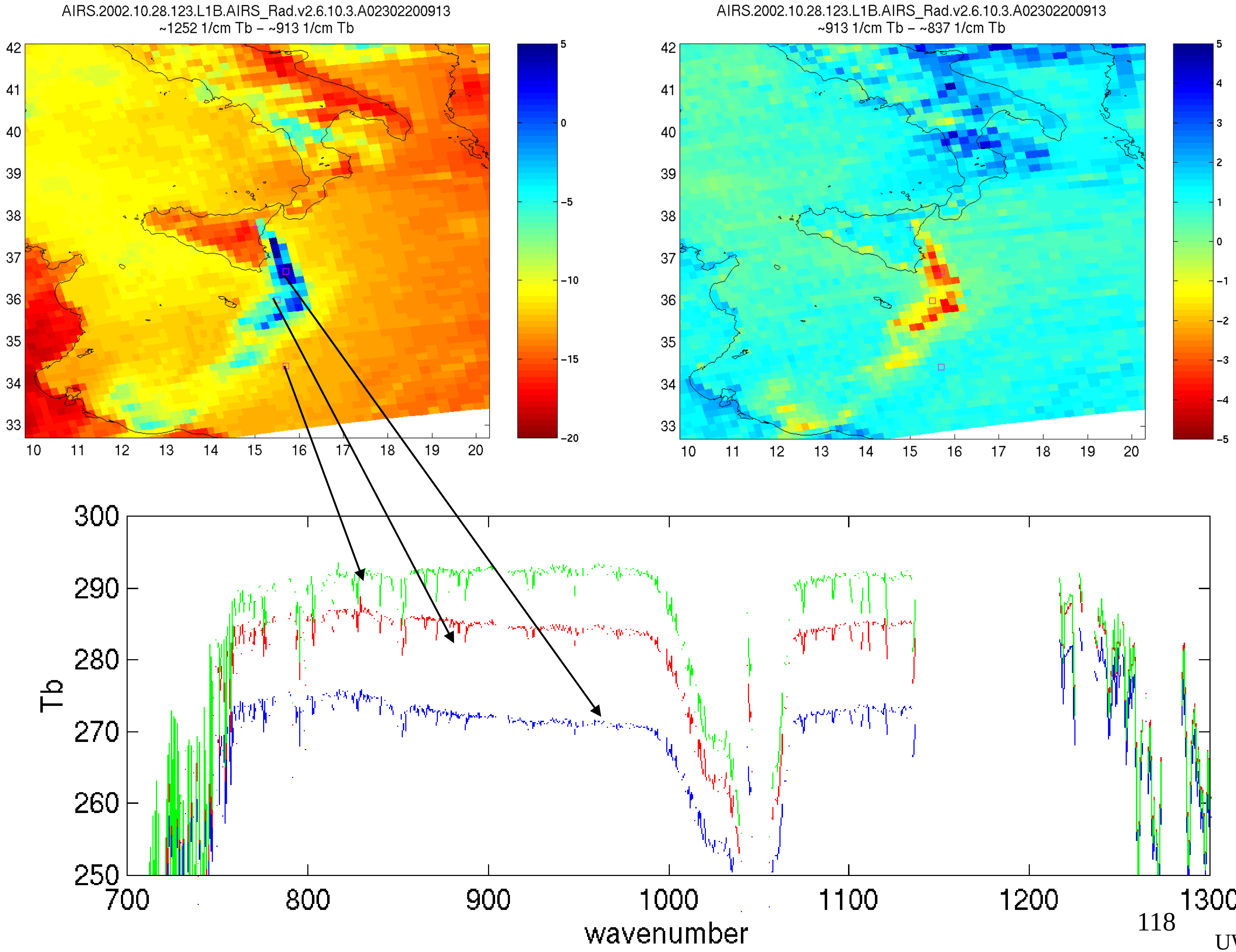


Wavelength (µm)
14.3
13.3
12.5
11.8
11.1
10.5
10.0
9.5
9.1
320
300
280
260
240
220
200
AIRS Brightness Temperature (K)
Desert 03-Mar-2004 14:17 UTC
Dust 03-Mar-2004 14:17 UTC
Water 03-Mar-2004 14:17 UTC
Water Cld 03-Mar-2004 14:17 UTC
Ice Cloud 03-Mar-2004 14:17 UTC
UW/CIMSS
700
750
800
850
900
950
1000
1050
1100
Wavenumber (cm^{-1})



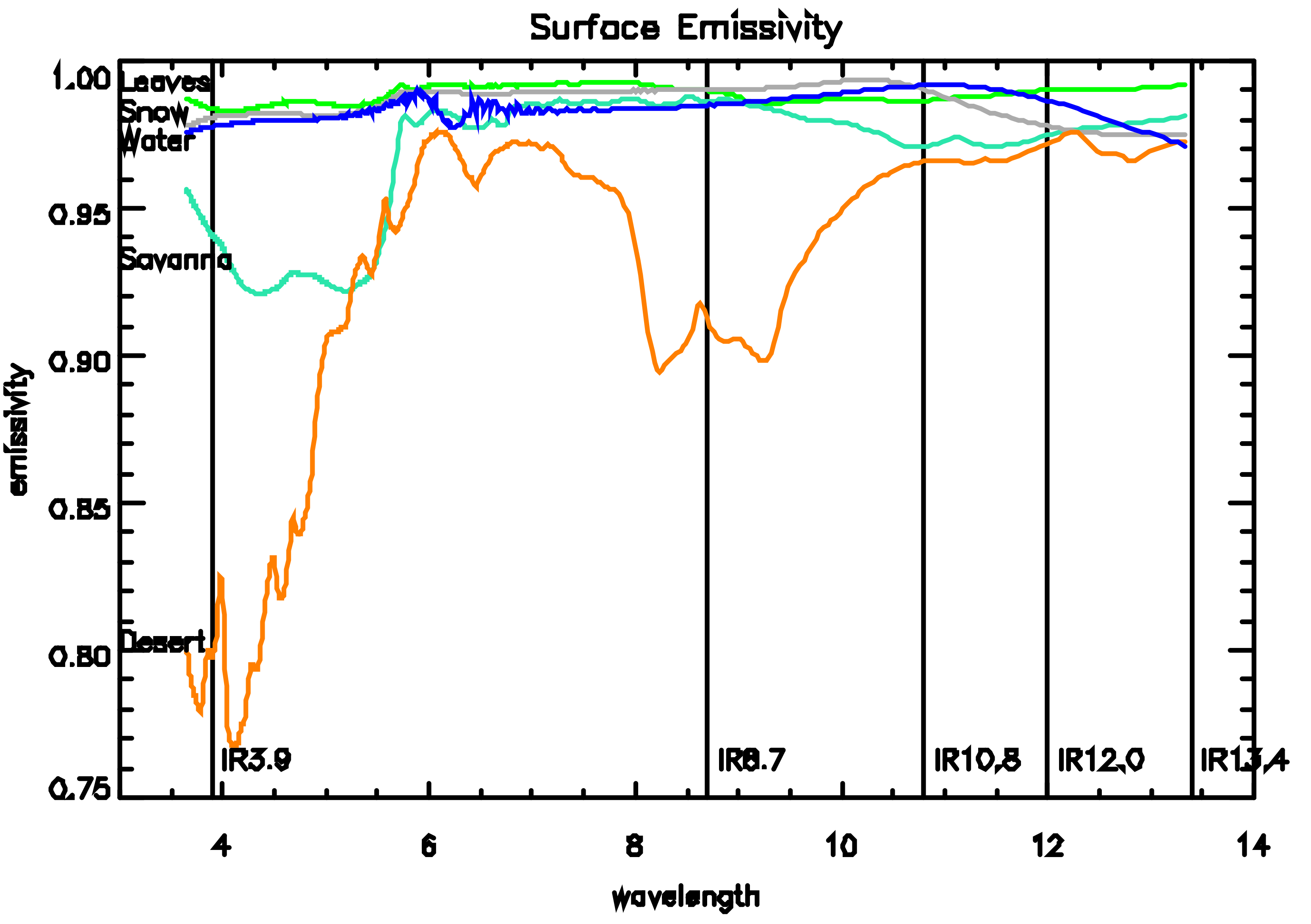


red spectrum is from nearby clear fov


AIRS.2002.10.28.123.L1B.AIRS_Rad.v2.6.10.3.A02302200913
~1252 1/cm Tb – ~913 1/cm Tb
AIRS.2002.10.28.123.L1B.AIRS_Rad.v2.6.10.3.A02302200913
~913 1/cm Tb – ~837 1/cm Tb
Tb
wavenumber

UW
CIMSS

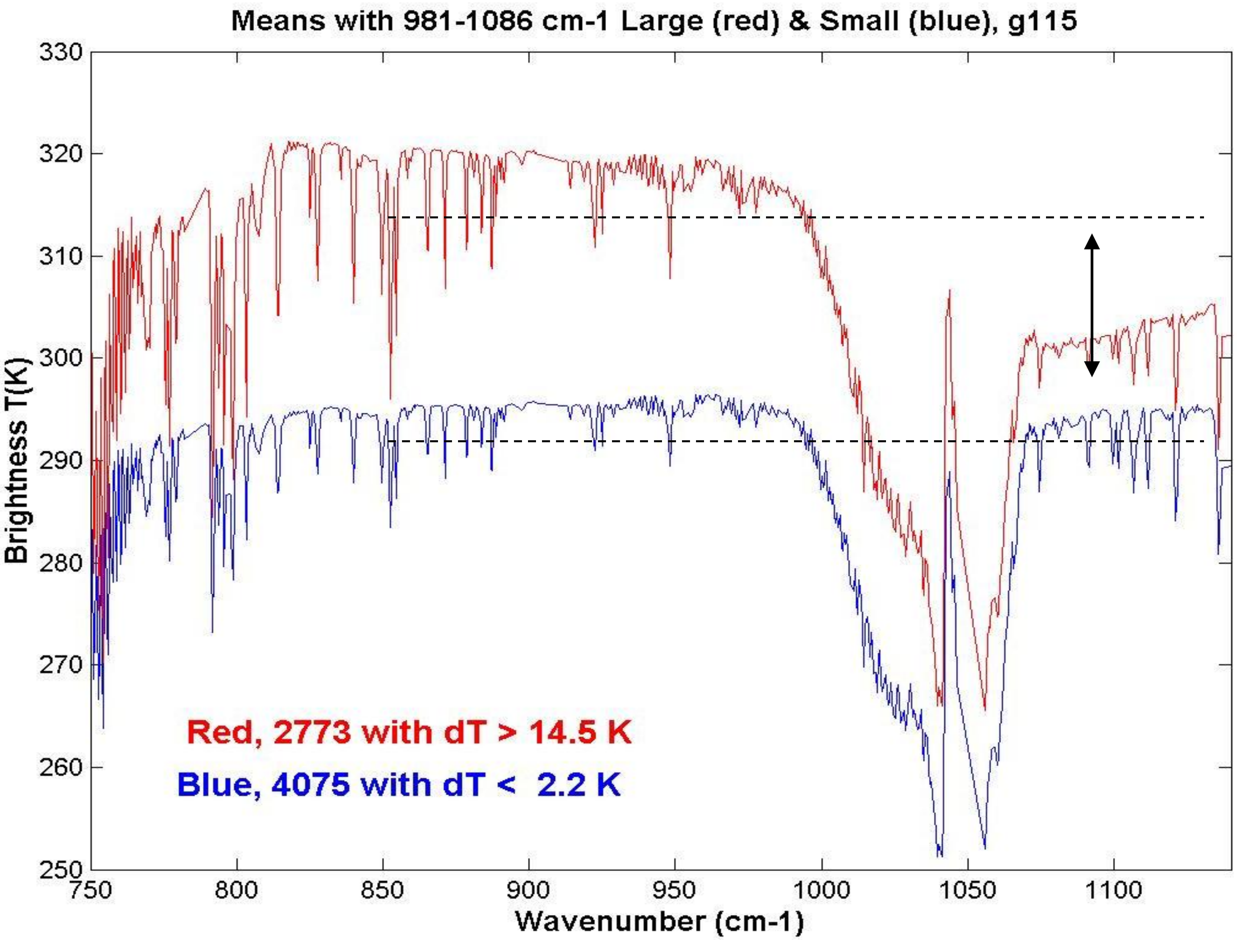
2500 1000 715 cm-1



Inferring surface properties with AIRS high spectral resolution data

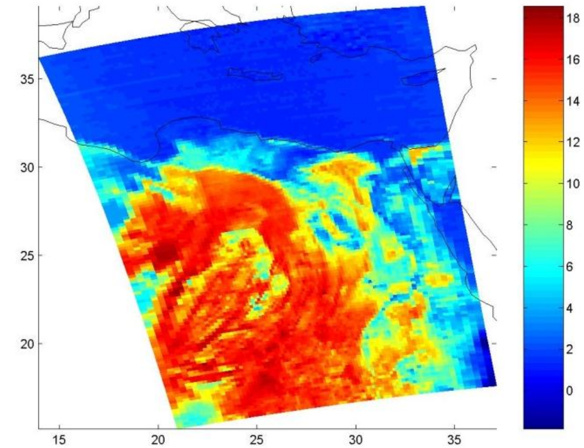
Barren region detection if T1086 < T981

Barren vs Water/Vegetated

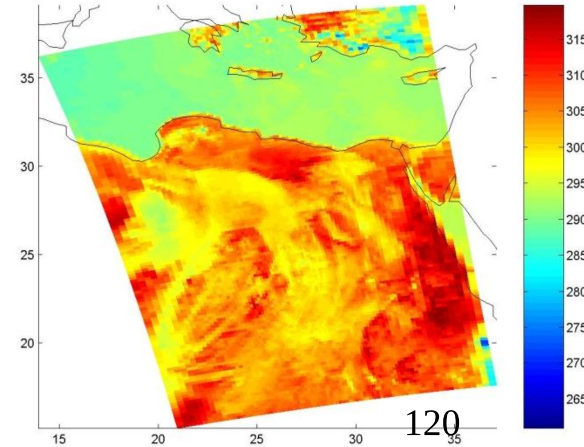


AIRS data from 14 June 2002

T(981 cm^{-1})-T(1086 cm^{-1})

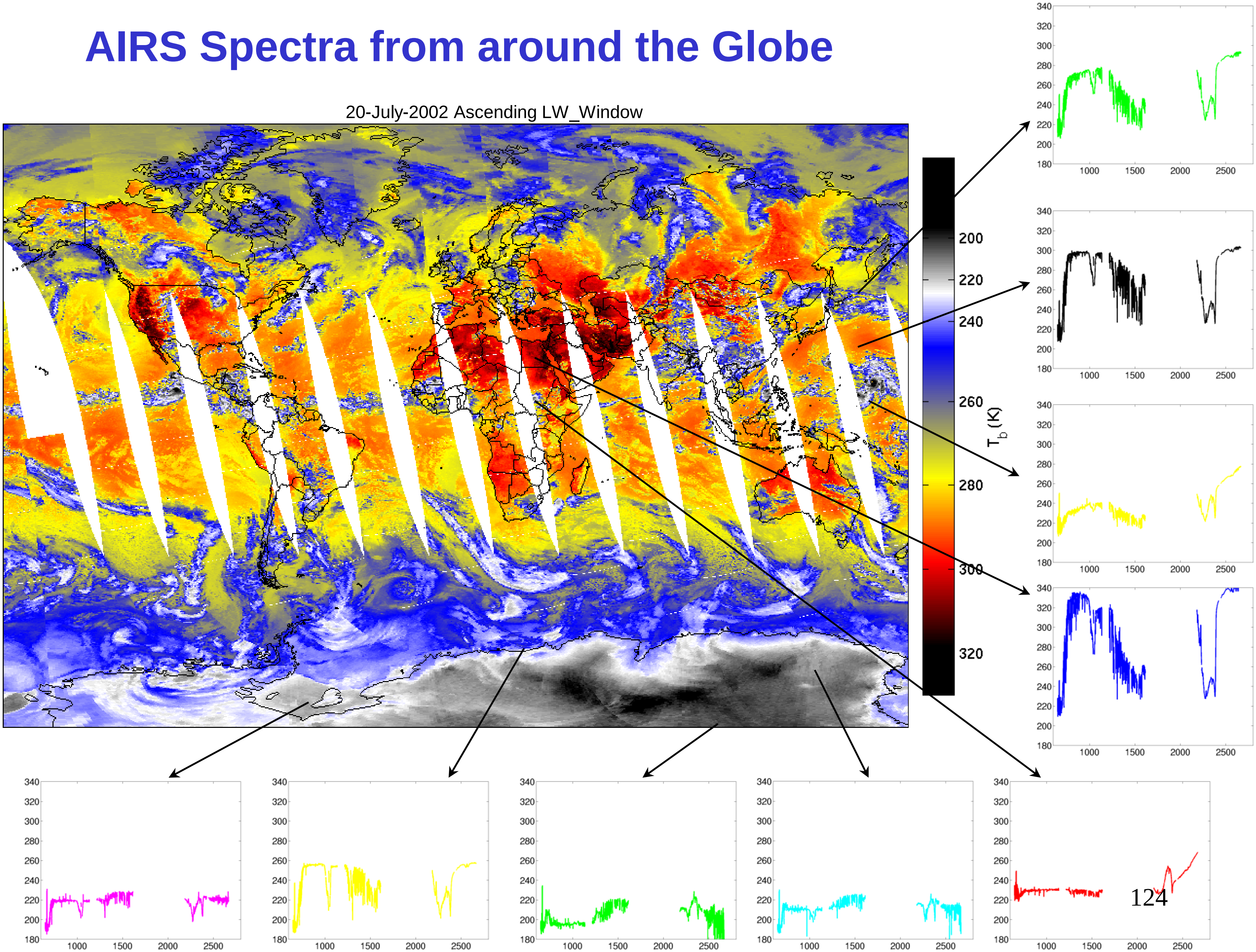


T(1086 cm^{-1})



from Tobin et al.

AIRS Spectra from around the Globe



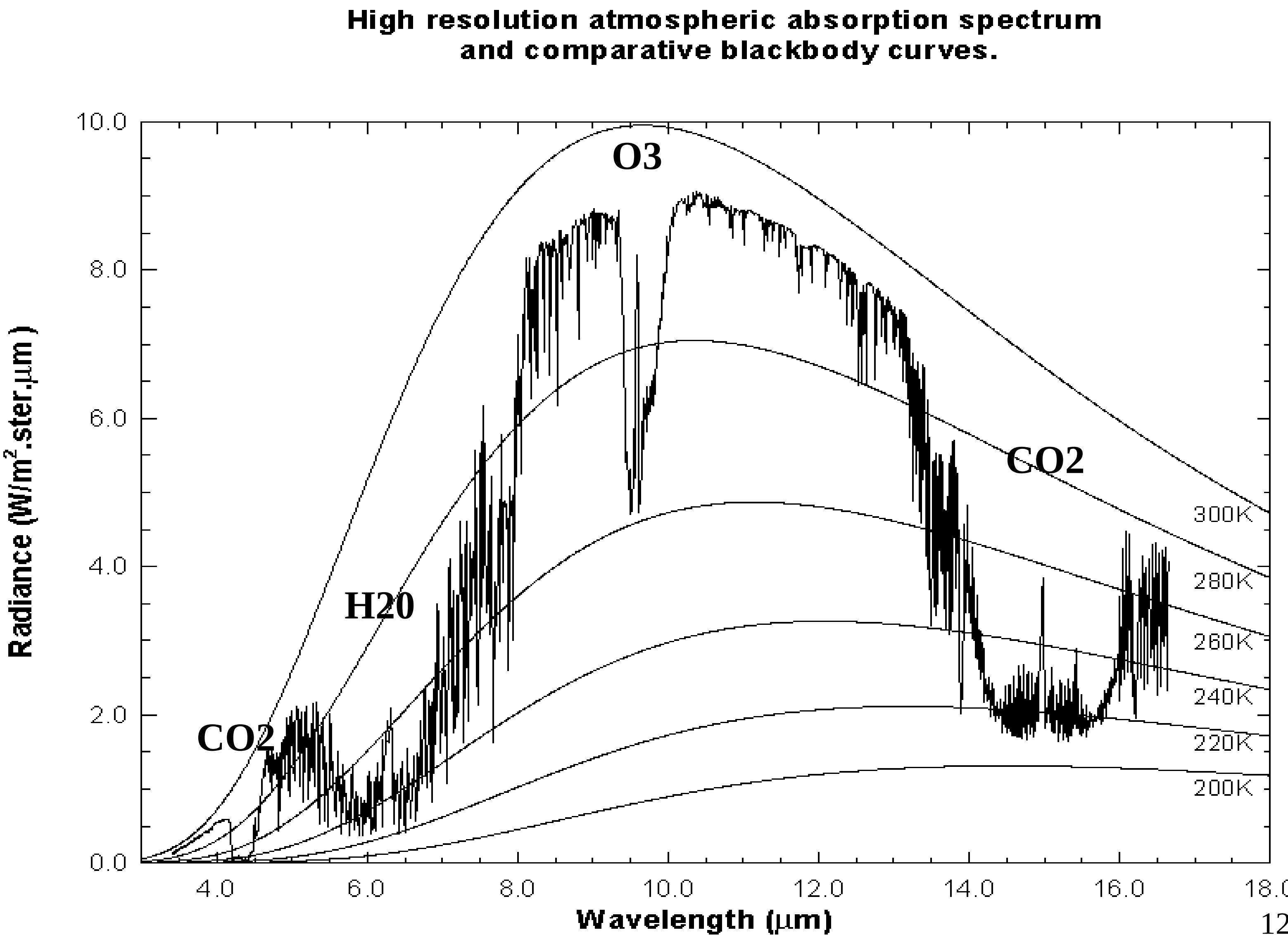
Intro to Lab on Split Window Moisture

Lecture - Labs in Bologna & Potenza

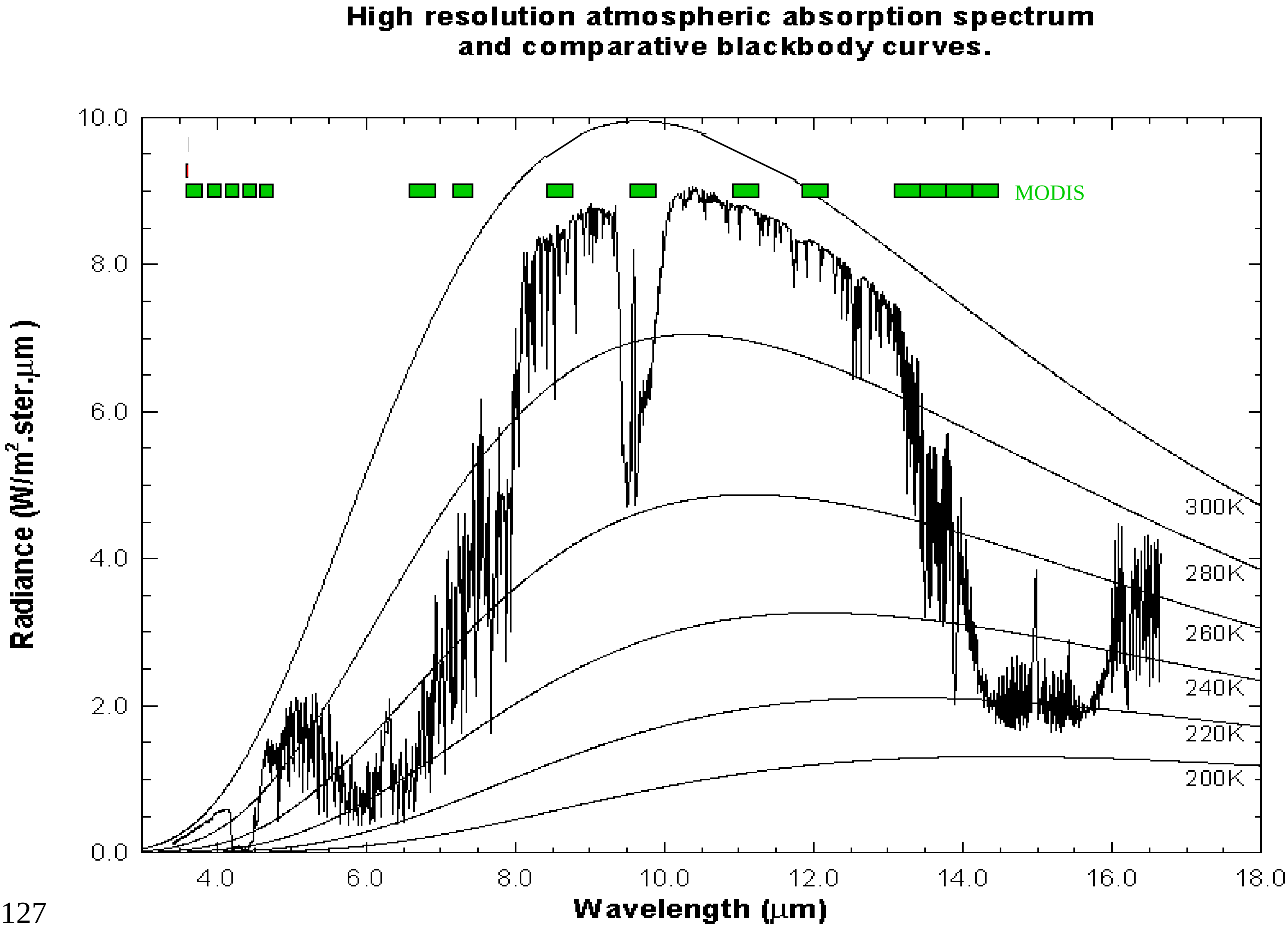
May-June 2012

Paul Menzel
UW/CIMSS/AOS

Earth emitted spectra overlaid on Planck function envelopes



MODIS IR Spectral Bands



First order estimation of SST correcting for low level moisture

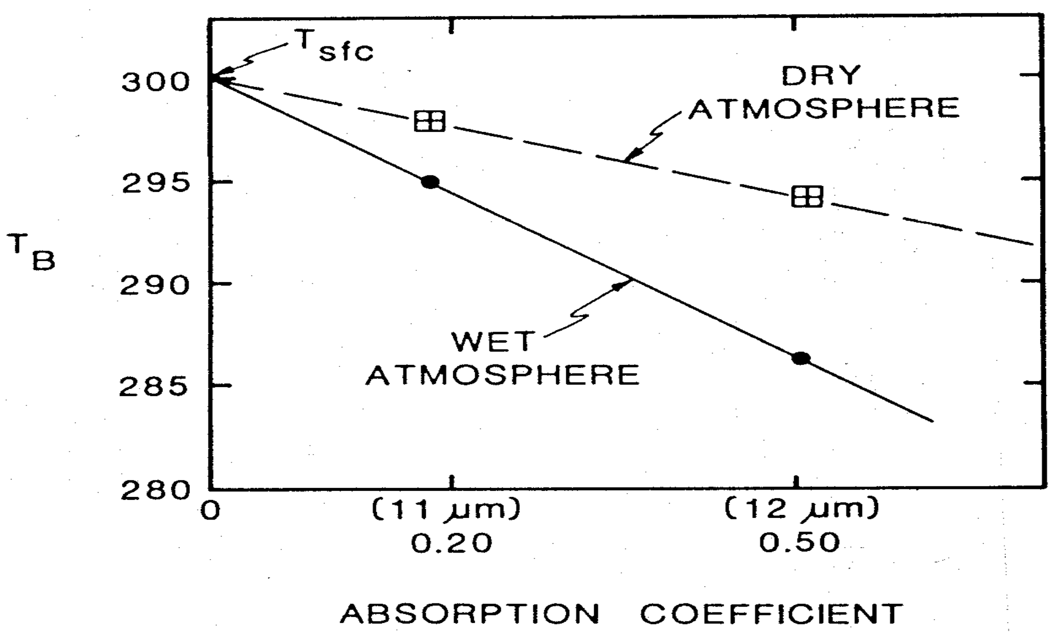
Moisture attenuation in atmospheric windows varies linearly with optical depth.

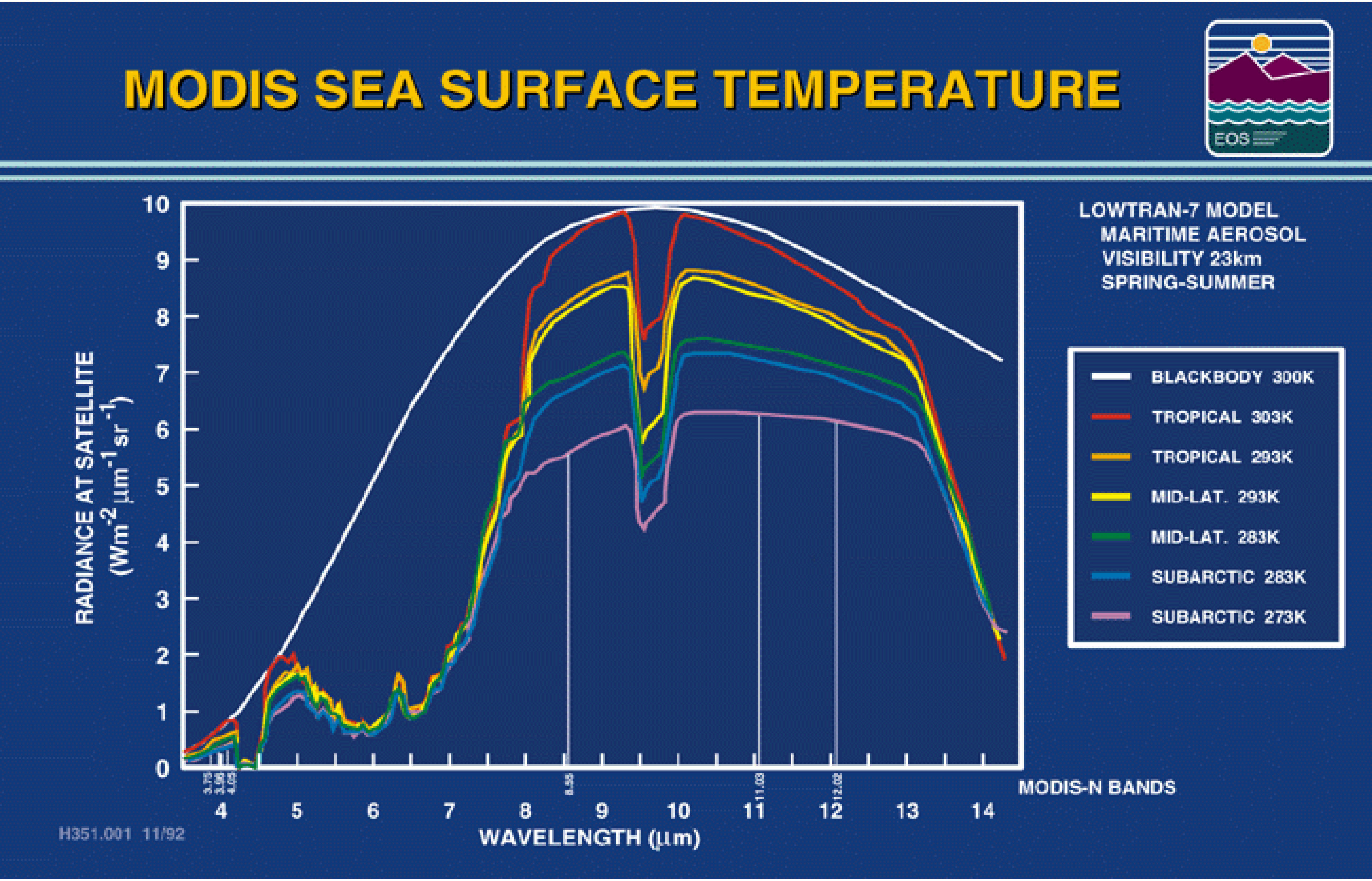
$$\tau_\lambda = e^{-k_\lambda u} = 1 - k_\lambda u$$

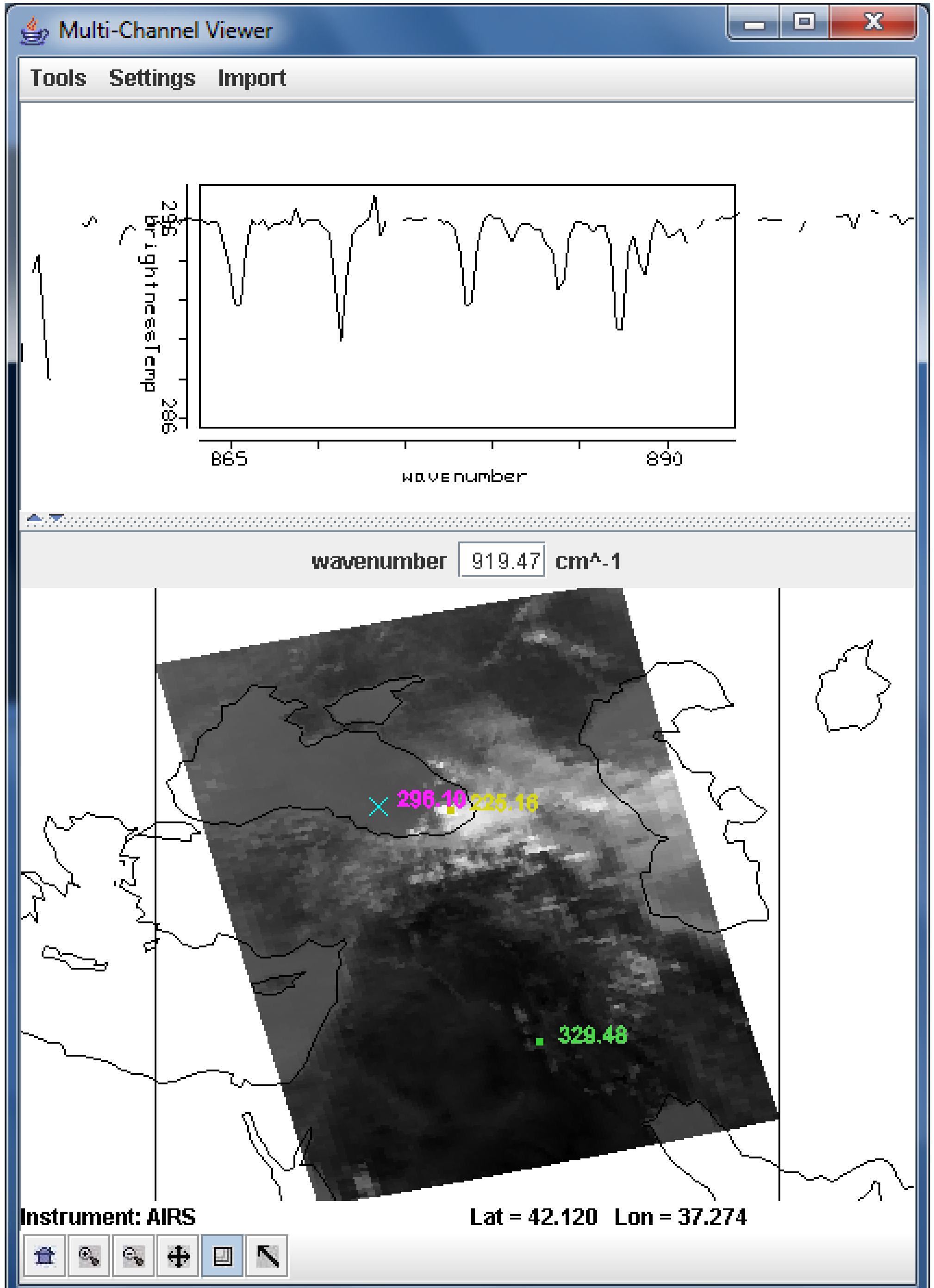
For same atmosphere, deviation of brightness temperature from surface temperature is a linear function of absorbing power. Thus moisture corrected SST can inferred by using split window measurements and extrapolating to zero k_λ

$$T_s = T_{bw1} + [\, k_{w1} / (k_{w2} - k_{w1}) \,] \, [T_{bw1} - T_{bw2}] \ .$$

Moisture content of atmosphere inferred from slope of linear relation.

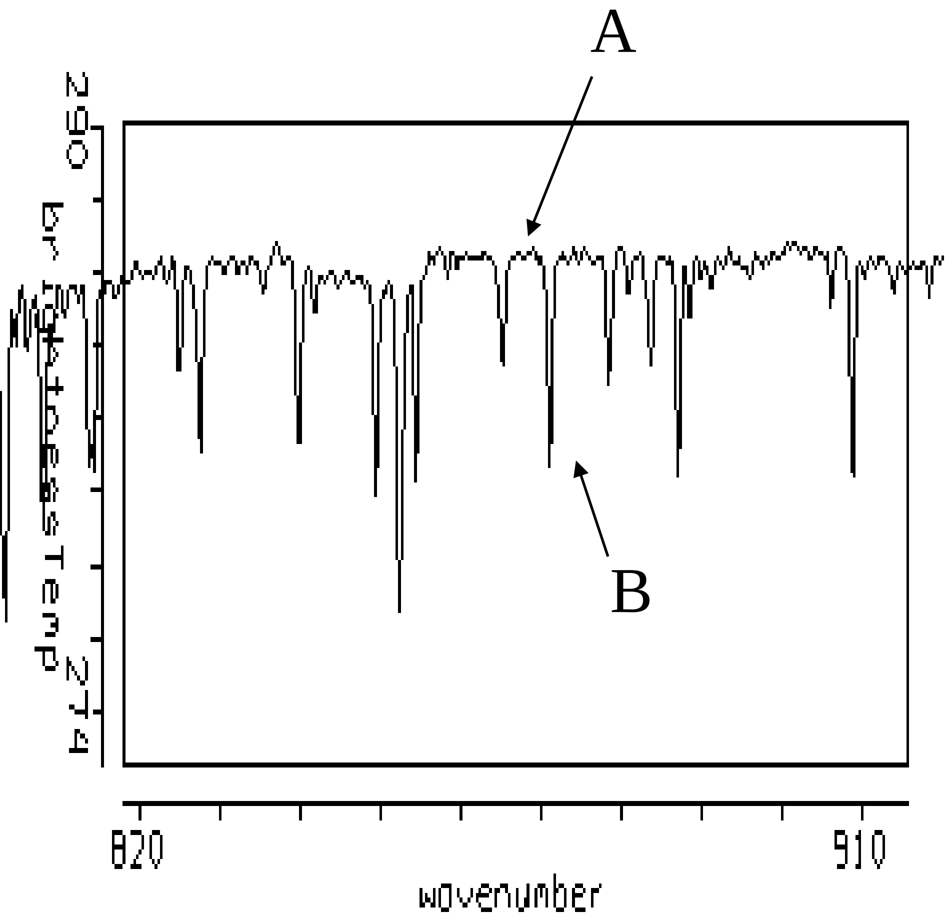



MODIS SEA SURFACE TEMPERATURE
EOS
LOWTRAN-7 MODEL
MARITIME AEROSOL
VISIBILITY 23km
SPRING-SUMMER
BLACKBODY 300K
TROPICAL 303K
TROPICAL 293K
MID-LAT. 293K
MID-LAT. 283K
SUBARCTIC 283K
SUBARCTIC 273K
RADIANCE AT SATELLITE
(Wm-2 μm-1 sr-1)
0
1
2
3
4
5
6
7
8
9
10
4
5
6
7
8
9
10
11
12
13
14
WAVELENGTH (μm)
MODIS-N BANDS
H351.001 11/92

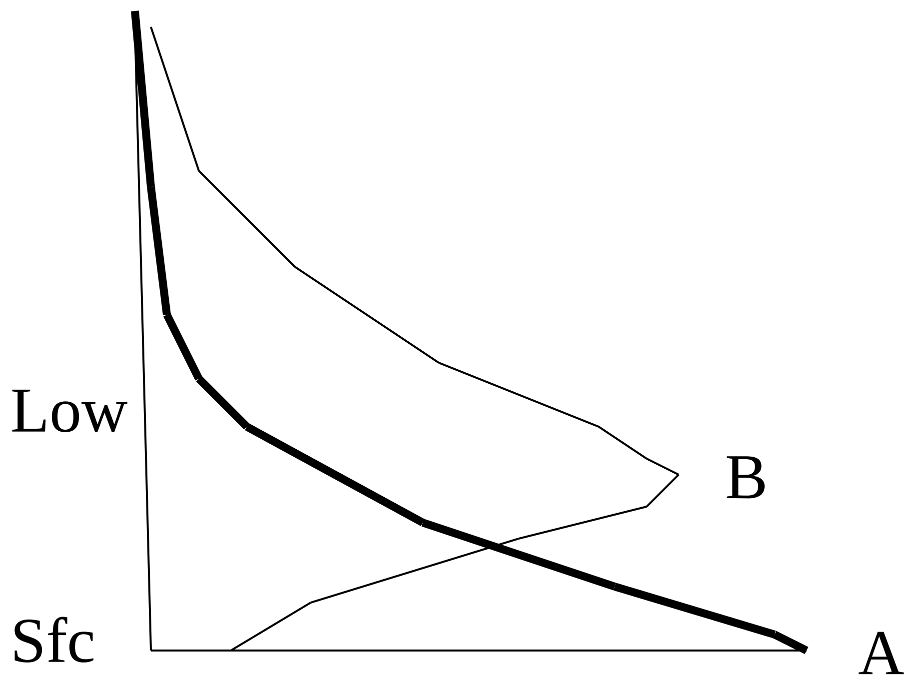


Multi-Channel Viewer
Tools Settings Import
296
BrightnessTemp
286
865
890
wavenumber
wavenumber 919.47 cm^-1
296.10
225.16
329.48
Instrument: AIRS
Lat = 42.120 Lon = 37.274

In the IRW - A is off H2O line and B is on H2O line



IRW spectrum



Weighting Function

Radiation is governed by Planck's Law

$$B(\lambda,T) = c_1 / \{ \lambda^5 [e^{c_2/\lambda T} - 1] \}$$

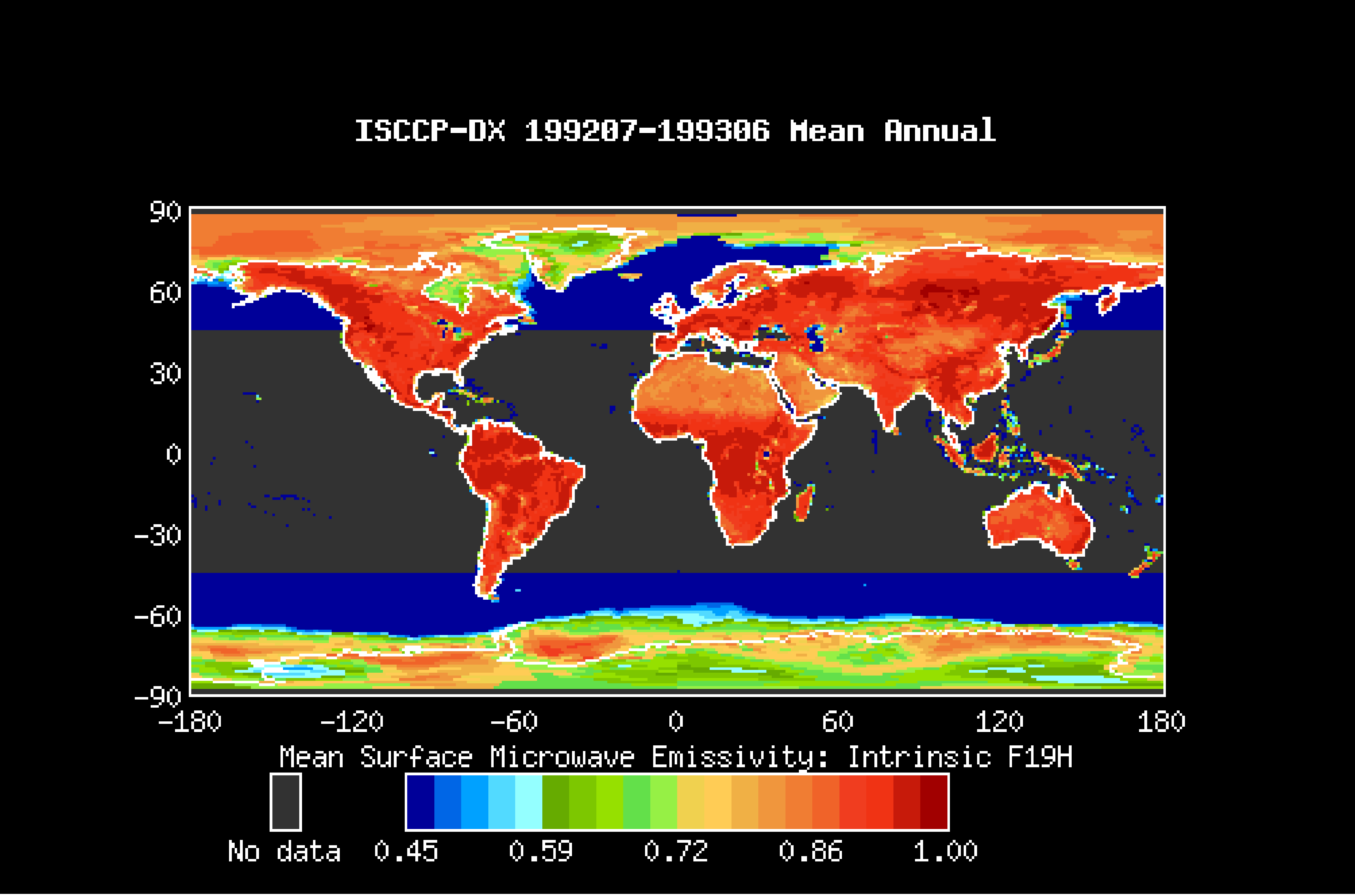
In microwave region $c_2/\lambda T << 1$ so that

$$e^{c_2/\lambda T} = 1 + c_2/\lambda T + \text{second order}$$

And classical Rayleigh Jeans radiation equation emerges

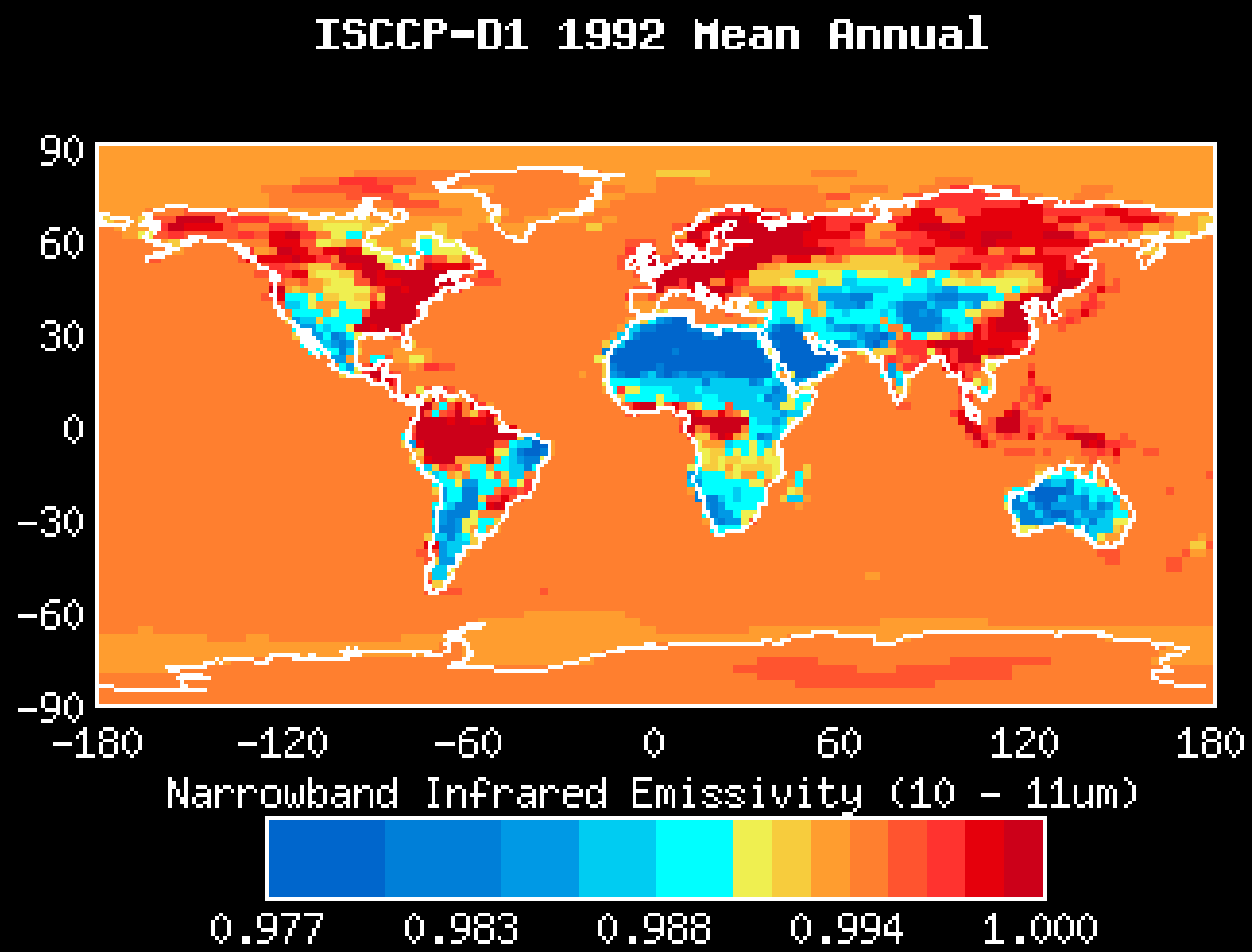
$$B_\lambda(T) \approx [c_1 / c_2] [T / \lambda^4]$$

Radiance is linear function of brightness temperature.



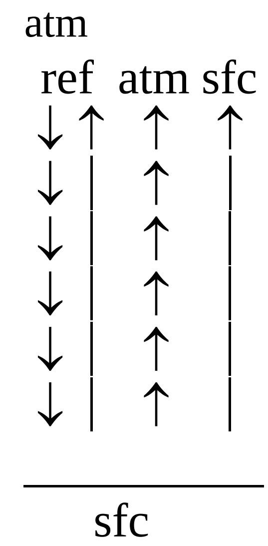
ISCCP-DX 199207-199306 Mean Annual
90
60
30
0
-30
-60
-90
-180
-120
-60
0
60
120
180
Mean Surface Microwave Emissivity: Intrinsic F19H
No data
0.45
0.59
0.72
0.86
1.00

19H Ghz


ISCCP-D1 1992 Mean Annual
90
60
30
0
-30
-60
-90
-180
-120
-60
0
60
120
180
Narrowband Infrared Emissivity (10 - 11um)
0.977
0.983
0.988
0.994
1.000

10 to 11 um

Microwave Form of RTE



$$I^{sfc}_{\lambda} = \varepsilon_\lambda B_\lambda(T_s)\, \tau_\lambda(p_s) + (1-\varepsilon_\lambda)\, \tau_\lambda(p_s) \int_{o}^{p_s} B_\lambda(T(p)) \frac{\partial \tau'_\lambda(p)}{\partial \ln p}\, d \ln p$$

$$I_\lambda = \varepsilon_\lambda B_\lambda(T_s)\, \tau_\lambda(p_s) + (1-\varepsilon_\lambda)\, \tau_\lambda(p_s) \int_{o}^{p_s} B_\lambda(T(p)) \frac{\partial \tau'_\lambda(p)}{\partial \ln p}\, d \ln p$$

$$+ \int_{p_s}^{o} B_\lambda(T(p)) \frac{\partial \tau_\lambda(p)}{\partial \ln p}\, d \ln p$$

In the microwave region $c_2 / \lambda T << 1$, so the Planck radiance is linearly proportional to the brightness temperature

$$B_\lambda(T) \approx [c_1 / c_2]\, [T / \lambda^4]$$

So

$$T_{b\lambda} = \varepsilon_\lambda T_s(p_s)\, \tau_\lambda(p_s) + \int_{p_s}^{o} T(p)\, F_\lambda(p) \frac{\partial \tau_\lambda(p)}{\partial \ln p}\, d \ln p$$

where

$$F_\lambda(p) = \left\{ 1 + (1 - \varepsilon_\lambda) \left[\frac{\tau_\lambda(p_s)}{\tau_\lambda(p)}\right]^2 \right\} .$$

Transmittance

$\tau(a,b) = \tau(b,a)$
$\tau(a,c) = \tau(a,b) * \tau(b,c)$

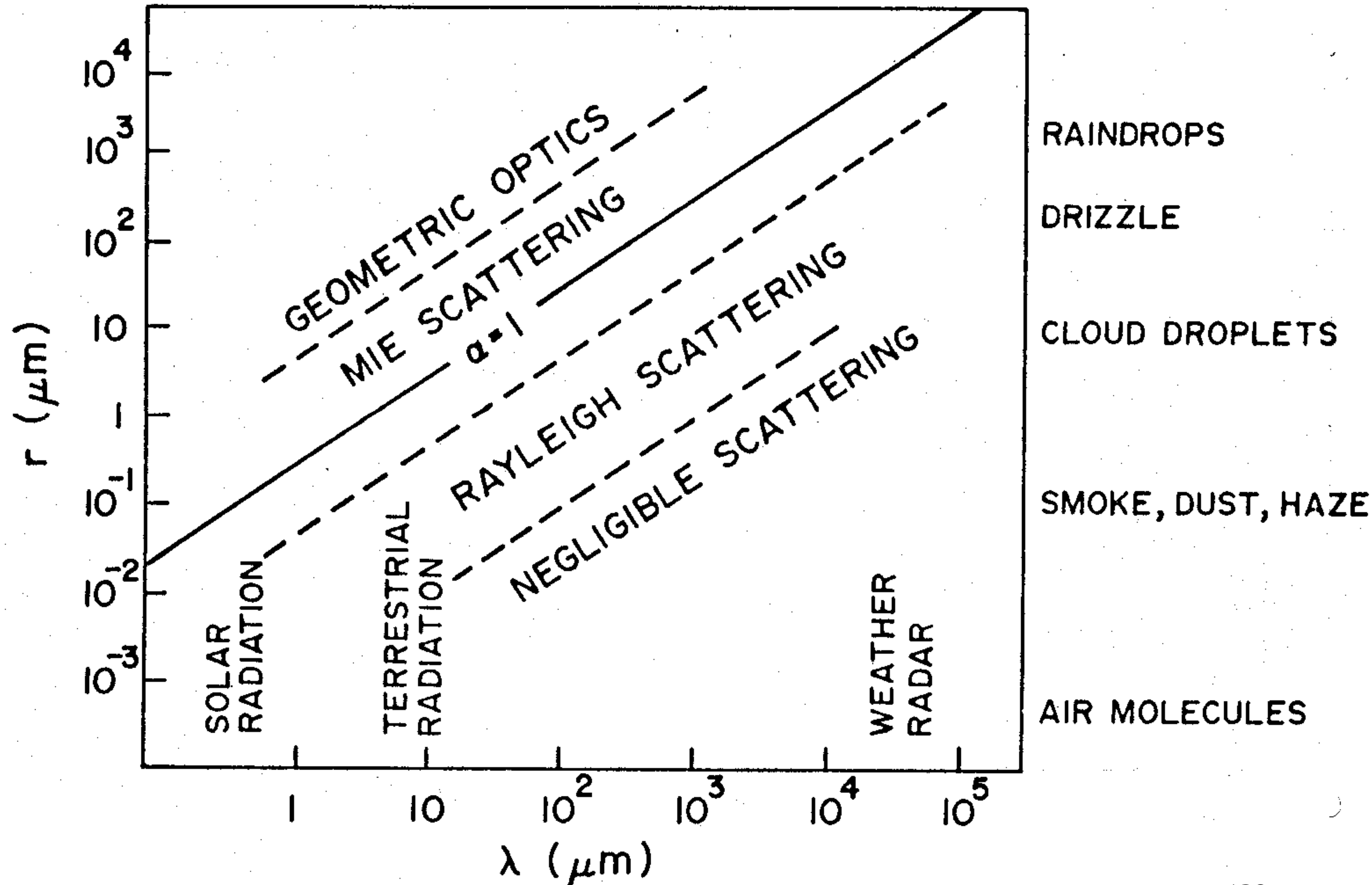
Thus downwelling in terms of upwelling can be written

$\tau'(p,ps) = \tau(ps,p) = \tau(ps,0) / \tau(p,0)$

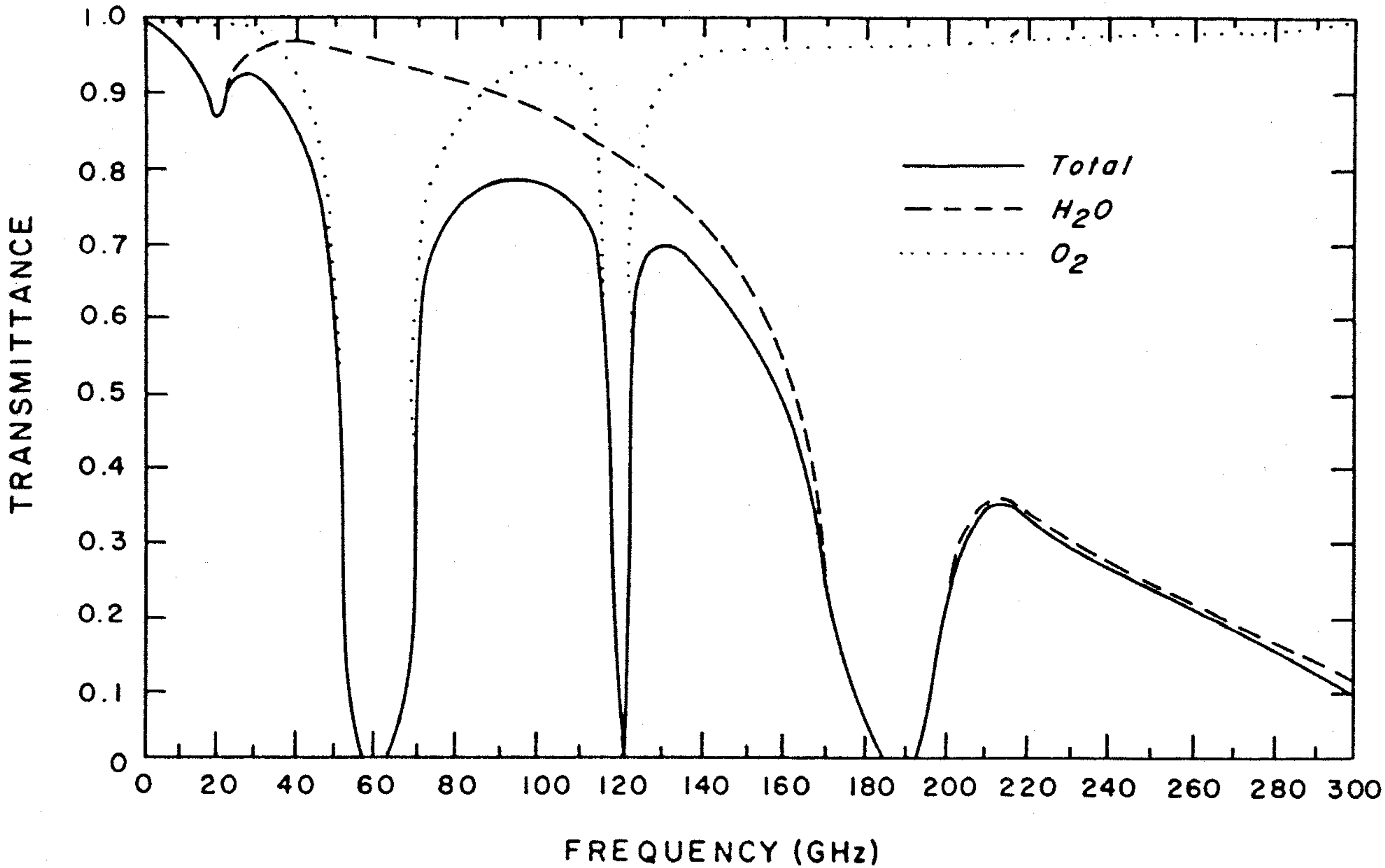
and

$d\tau'(p,ps) = - d\tau(p,0) * \tau(ps,0) / [\tau(p,0)]^2$

WAVELENGTH			FREQUENCY		WAVENUMBER
cm	μm	Å	Hz	GHz	cm^{-1}
10^{-5} Near Ultraviolet (UV)	0.1	1,000	$3x10^{15}$		
$4x10^{-5}$ Visible	0.4	4,000	$7.5x10^{14}$		
$7.5x10^{-5}$ Near Infrared (IR)	0.75	7,500	$4x10^{14}$		13,333
$2x10^{-3}$ Far Infrared (IR)	20	$2x10^{5}$	$1.5x10^{13}$		500
0.1 Microwave (MW)	10^{3}		$3x10^{11}$	300	10

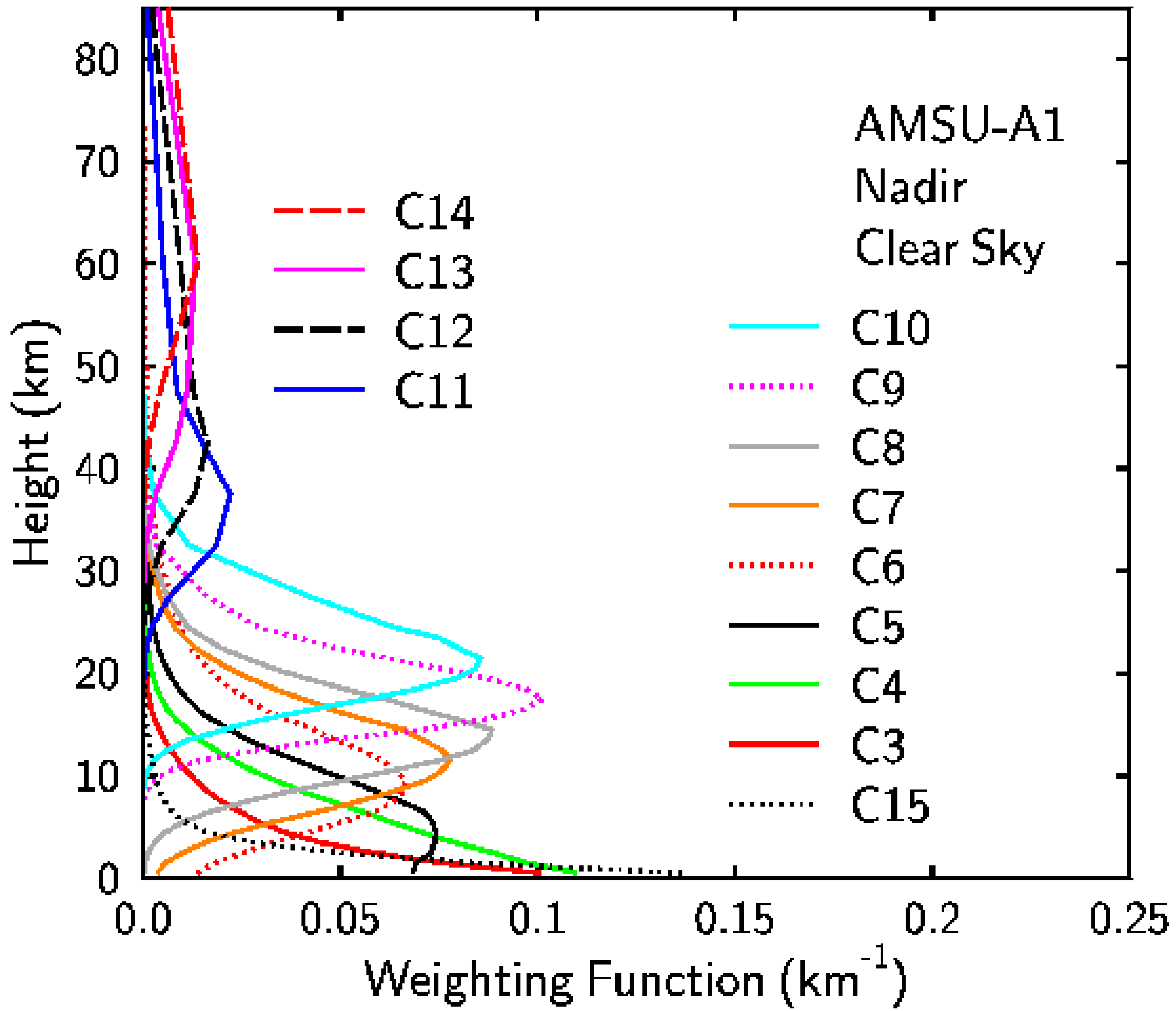


GEOMETRIC OPTICS
MIE SCATTERING
α = 1
RAYLEIGH SCATTERING
NEGLIGIBLE SCATTERING
SOLAR RADIATION
TERRESTRIAL RADIATION
WEATHER RADAR
RAINDROPS
DRIZZLE
CLOUD DROPLETS
SMOKE, DUST, HAZE
AIR MOLECULES
r (μm)
λ (μm)

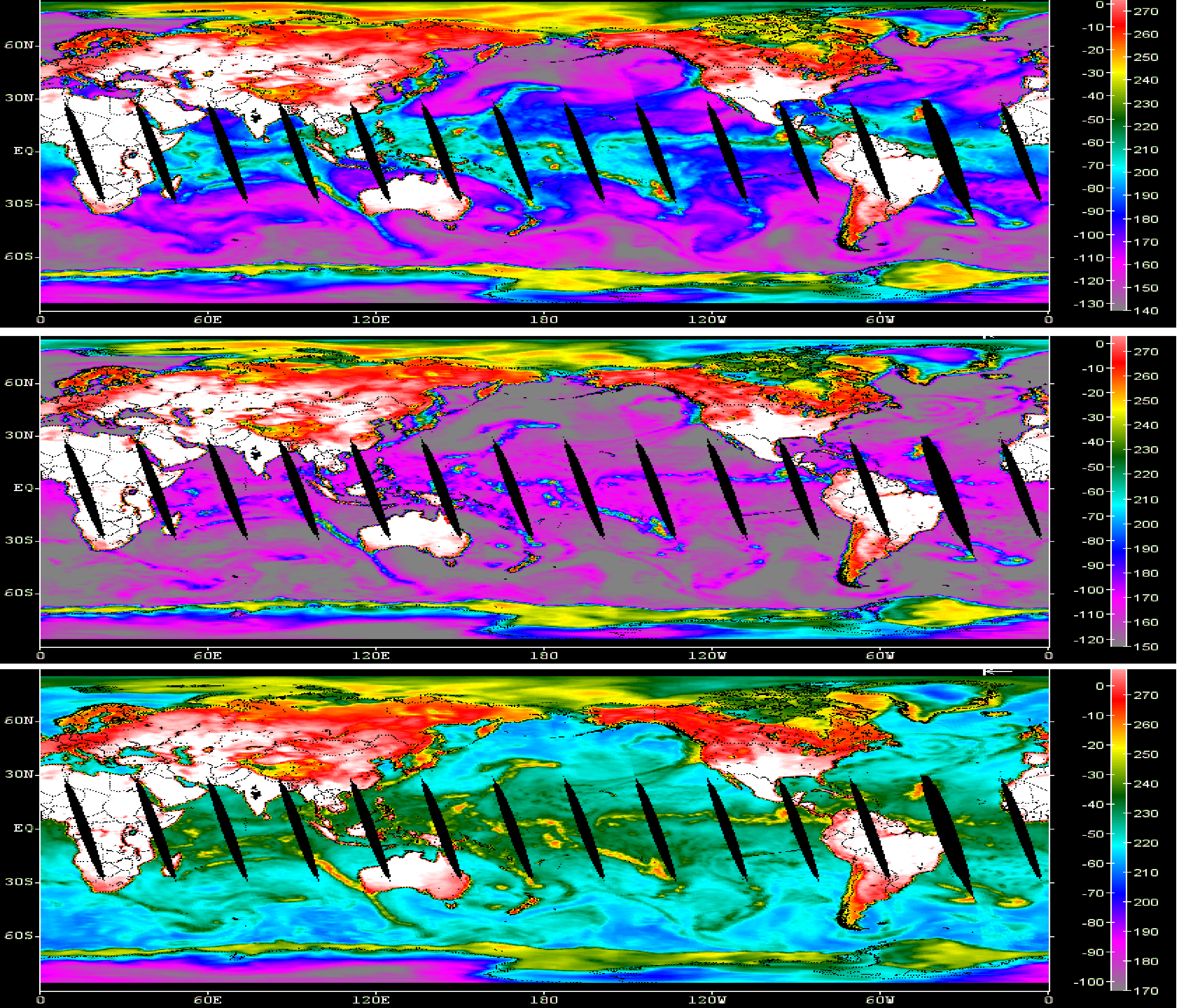

1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
TRANSMITTANCE
0 20 40 60 80 100 120 140 160 180 200 220 240 260 280 300
FREQUENCY (GHz)
Total
H2O
O2

Microwave spectral bands

23.8 GHz	dirty window H_2O absorption
31.4 GHz	window
60 GHz	O_2 sounding
120 GHz	O_2 sounding
183 GHz	H_2O sounding



23.8, 31.4, 50.3, 52.8, 53.6, 54.4, 54.9, 55.5, 57.3 (6 chs), 89.0 GHz

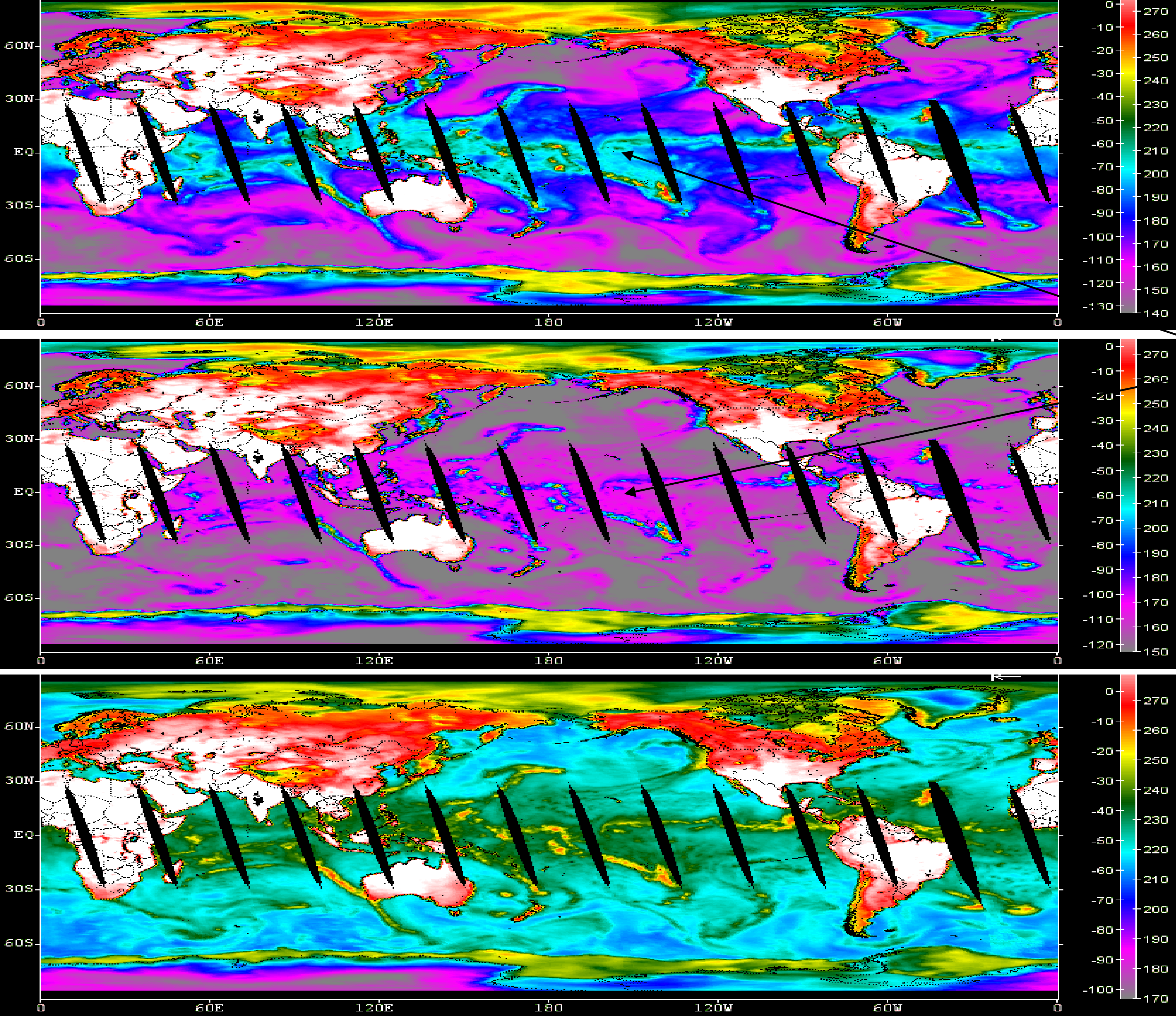


AMSU

23.8

31.4

50.3 GHz



AMSU

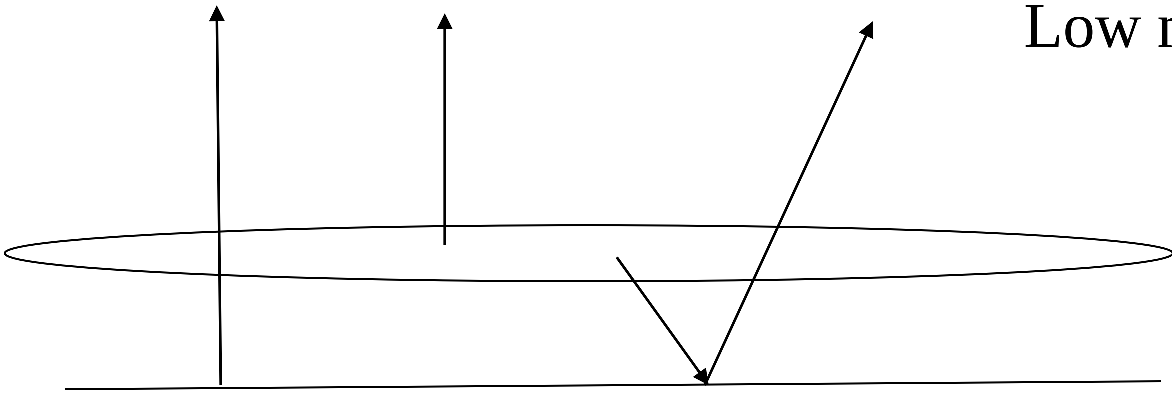
23.8 dirty window

atm Q warms BT

31.4 window

50.3 GHz

Low mist over ocean (MW)



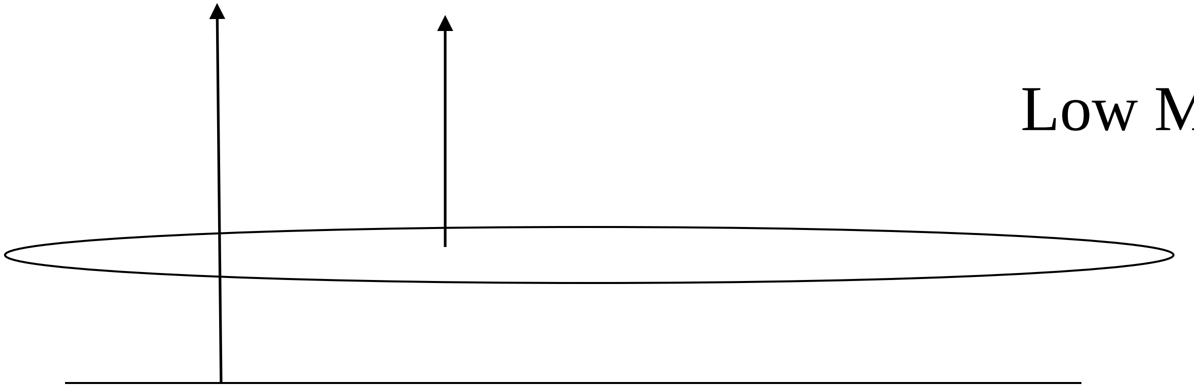
$T_b = \varepsilon_s T_s \tau_m + \varepsilon_m T_m + \varepsilon_m r_s \tau_m T_m$

$T_b = \varepsilon_s T_s (1-\sigma_m) + \sigma_m T_m + \sigma_m (1-\varepsilon_s)(1-\sigma_m) T_m$ using $e^{-\sigma} = 1 - \sigma$

So temperature difference of low moist over ocean from clear sky over ocean is given by

$\Delta T_b = -\varepsilon_s \sigma_m T_s + \sigma_m T_m + \sigma_m (1-\varepsilon_s)(1-\sigma_m) T_m$

For $\varepsilon_s \sim 0.5$ and $T_s \sim T_m$ this is always positive for $0 < \sigma_m < 1$



Low Mist over ocean (IRW)

$R = \varepsilon_s B_s (1-\sigma_m) + \sigma_m B_m$ using $e^{-\sigma} = 1 - \sigma$ and $\tau \sim 1-\sigma \sim 1-a$

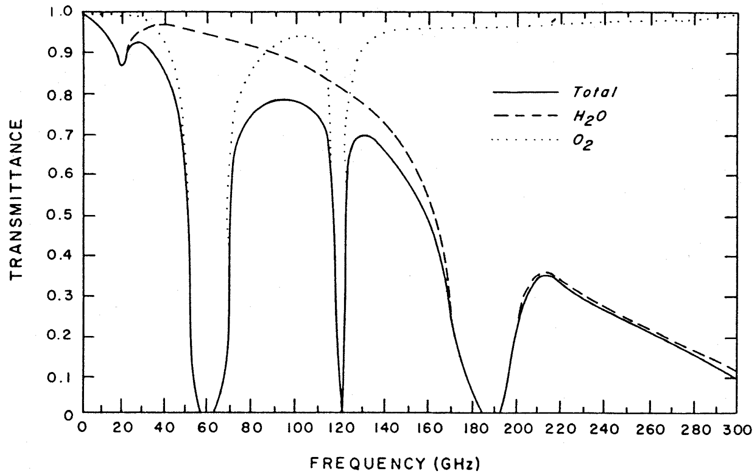
So difference of low mist over ocean
from clear sky over ocean is given by

$\Delta R = - \varepsilon_s \sigma_m B_s + \sigma_m B_m$

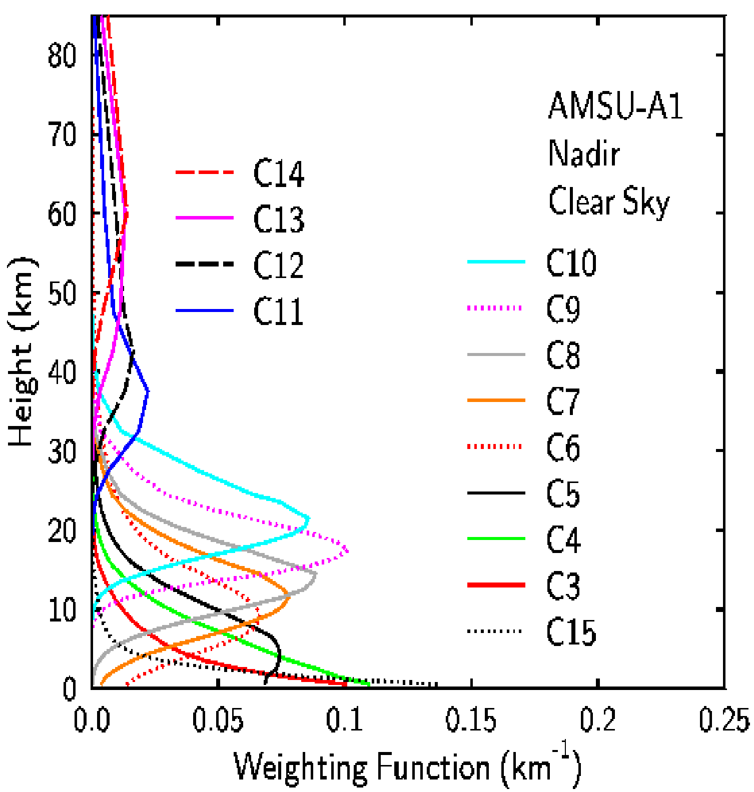
For $\varepsilon_s \sim 1$

$\Delta R = - \sigma_m B_s + \sigma_m B_m = \sigma_m [B_m - B_s]$

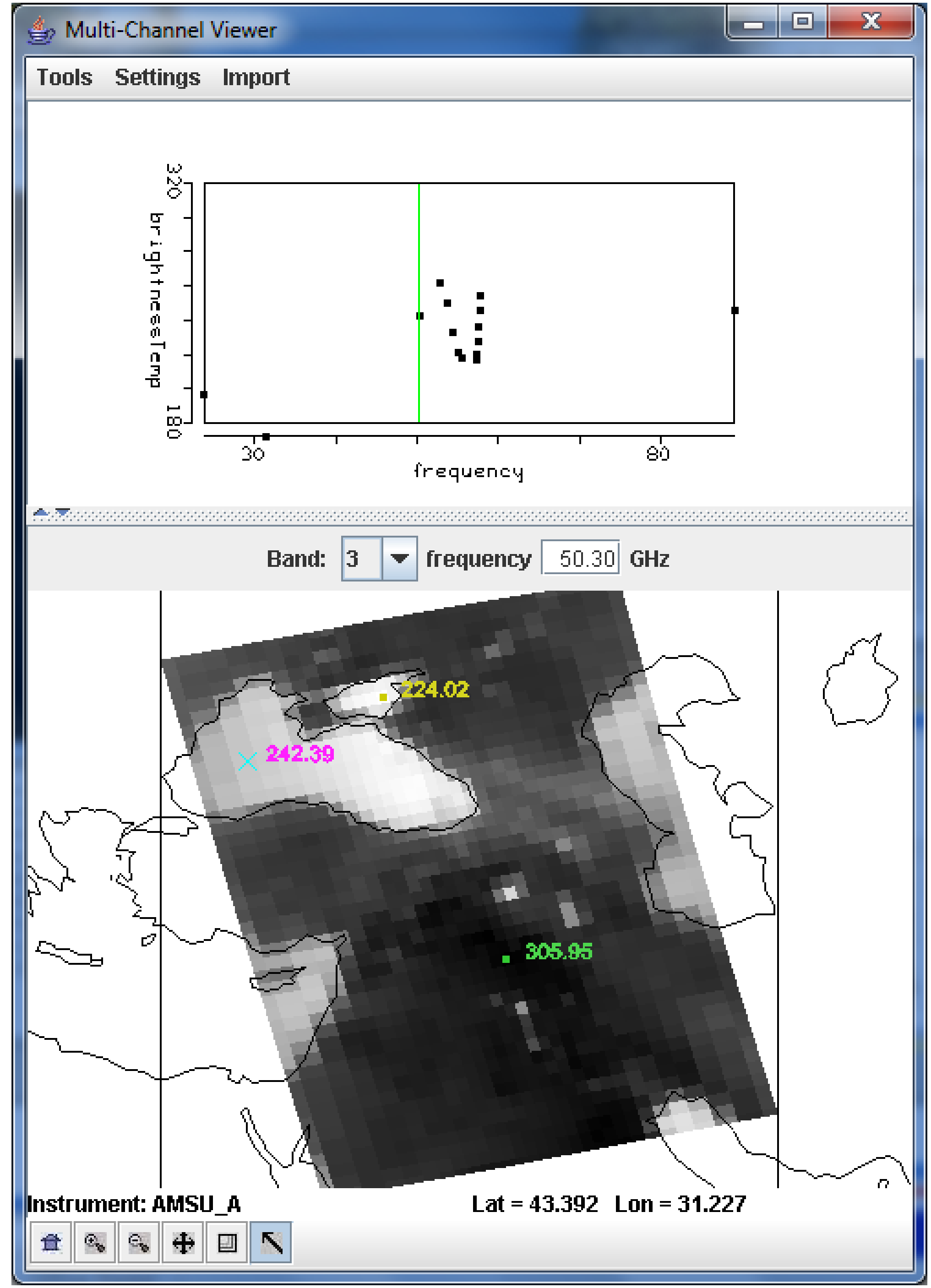
So if $[B_m - B_s] < 0$ then as σ_m increases ΔR becomes more negative



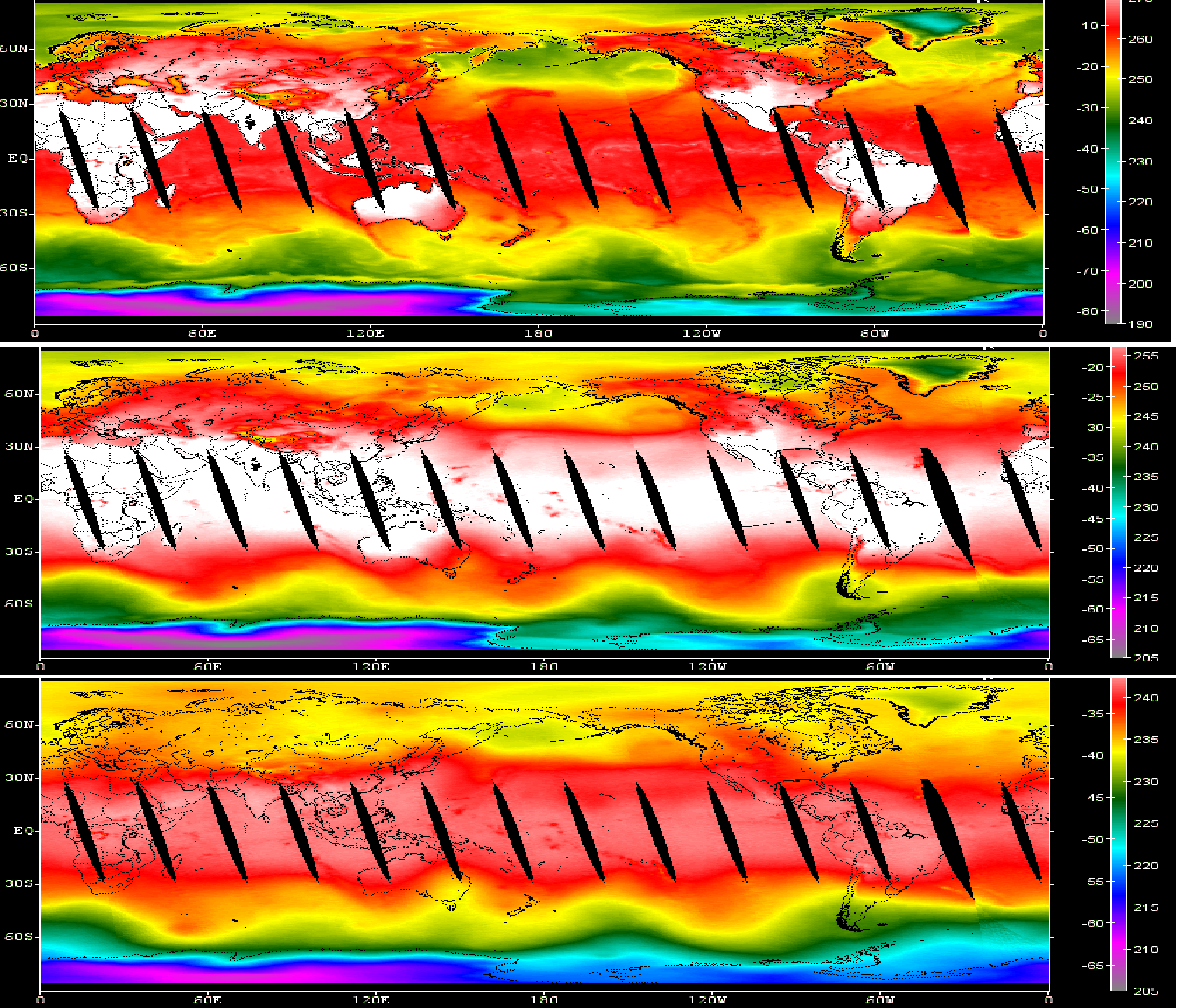
TRANSMITTANCE
FREQUENCY (GHz)
Total
H_2O
O_2



AMSU-A1
Nadir
Clear Sky
C14
C13
C12
C11
C10
C9
C8
C7
C6
C5
C4
C3
C15
Height (km)
Weighting Function (km^{-1})



Multi-Channel Viewer
Tools
Settings
Import
brightnessTemp
frequency
Band: 3
frequency 50.30 GHz
224.02
242.39
305.95
Instrument: AMSU_A
Lat = 43.392 Lon = 31.227

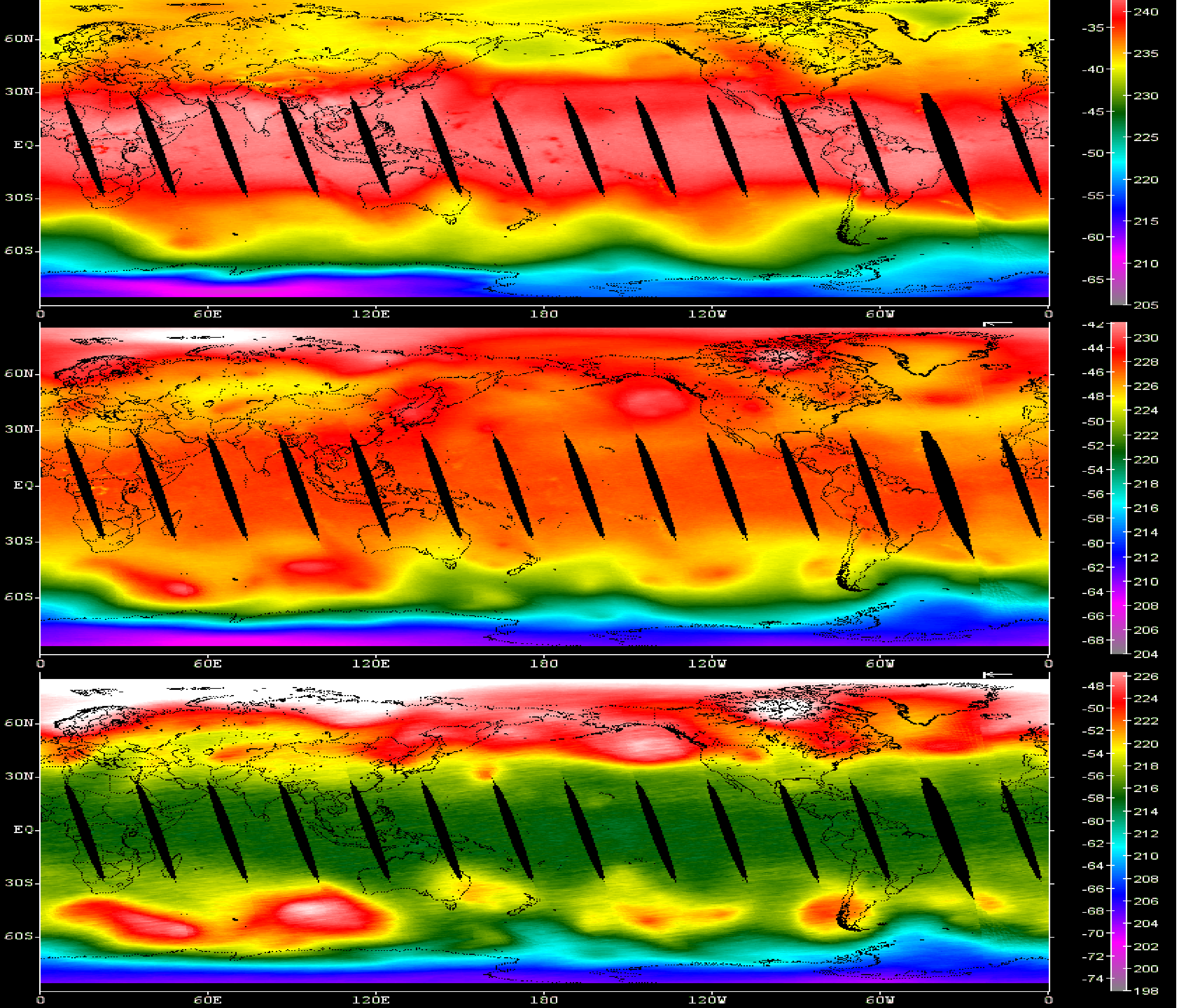


AMSU

52.8

53.6

54.4 GHz

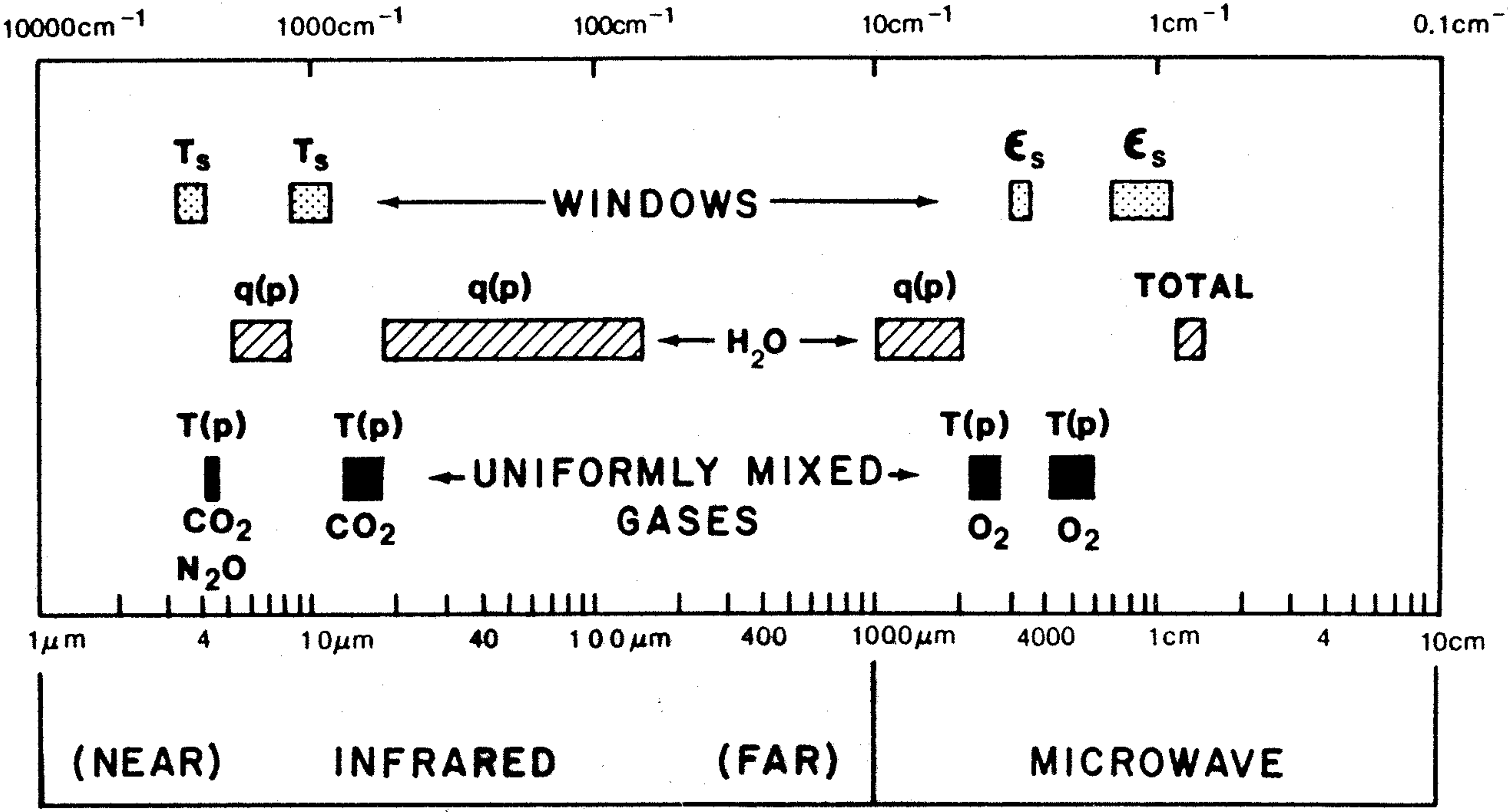


AMSU

54.4

54.9

55.5 GHz



Spectral regions used for remote sensing of the earth atmosphere and surface from satellites. ε indicates emissivity, q denotes water vapour, and T represents temperature.