Summary
Remote Sensing Seminar

Lectures in Krakow
May 2006

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NOAA/NESDIS/ORA

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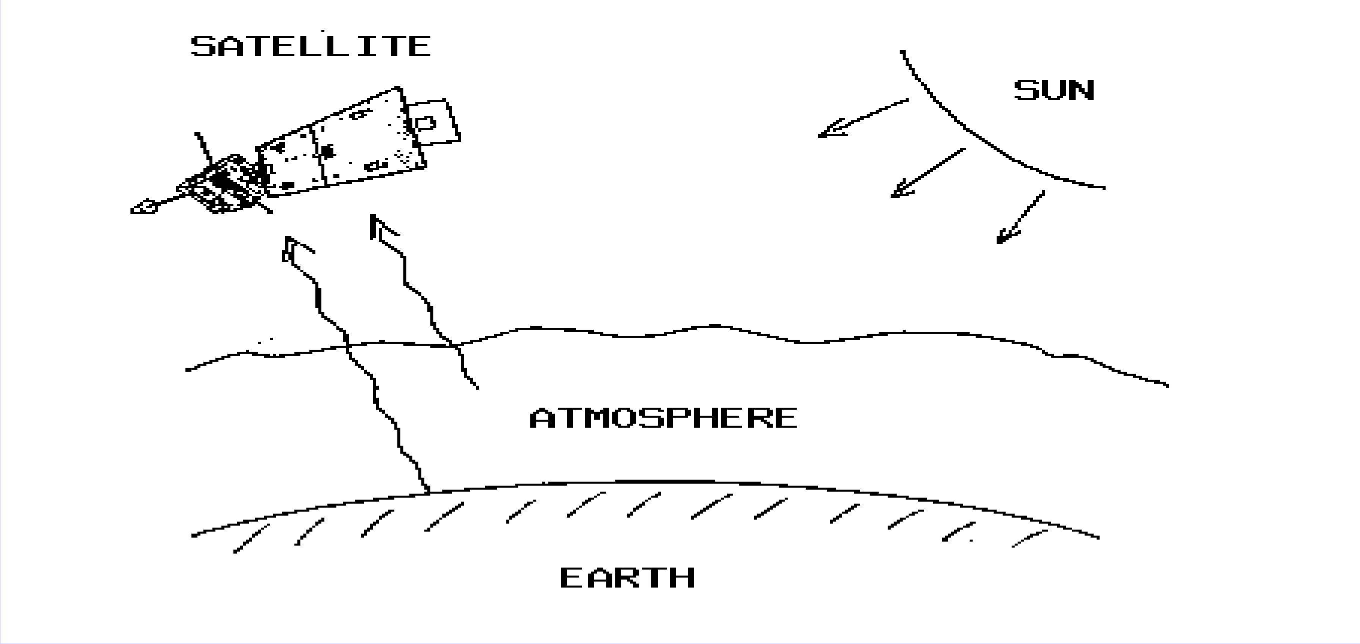
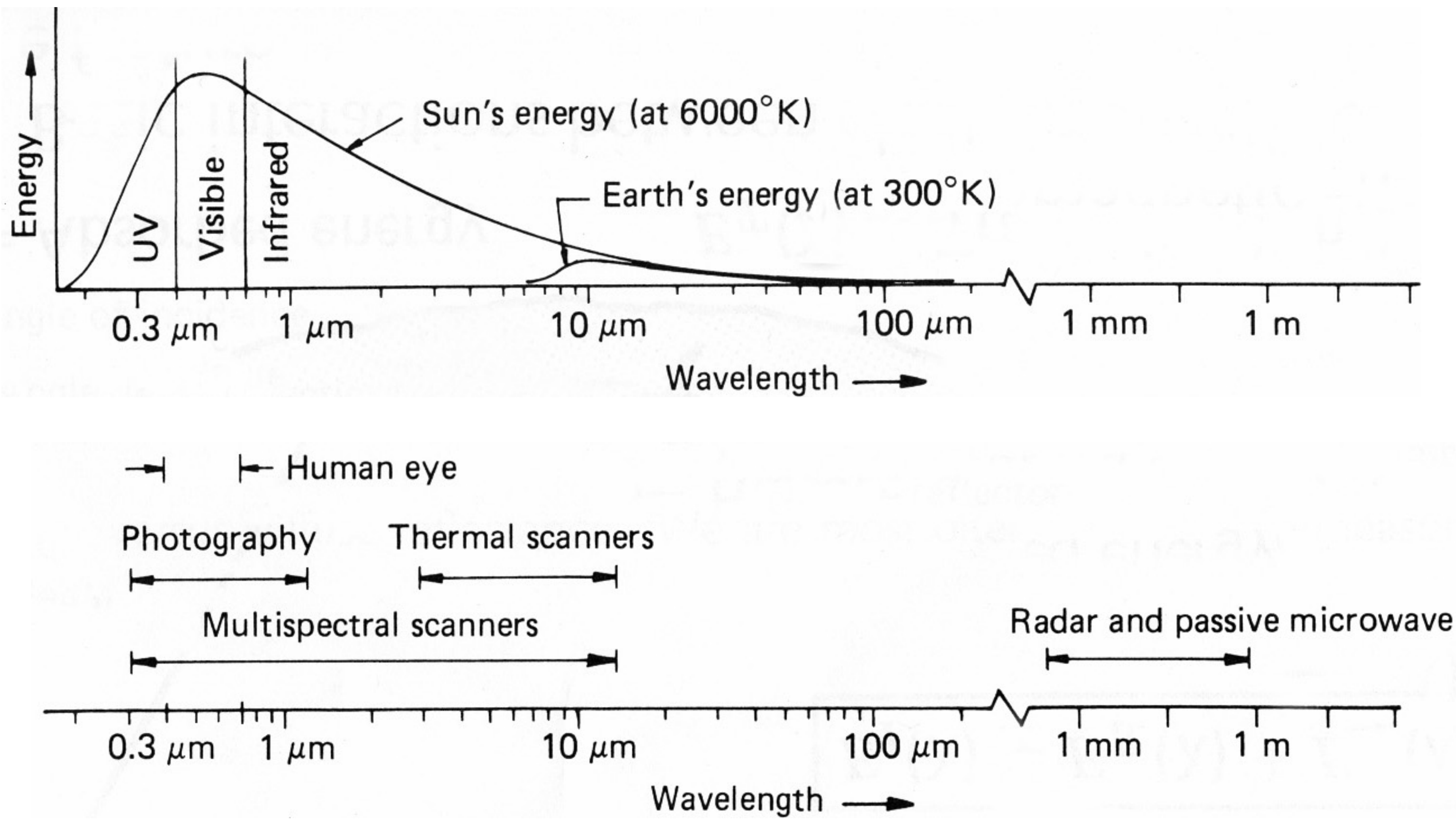
Satellite remote sensing of the Earth-atmosphere



Observations depend on

- telescope characteristics (resolving power, diffraction)
- detector characteristics (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)

Spectral Characteristics of Energy Sources and Sensing Systems



Terminology of radiant energy

Energy from the Earth Atmosphere

over time is

Flux

which strikes the detector area

Irradiance

at a given wavelength interval

Monochromatic Irradiance

over a solid angle on the Earth

Radiance observed by satellite radiometer

is described by

The Planck function

can be inverted to

Brightness temperature

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 wavenumbers

Planck's Law

$$B(\nu,T) = c_1\nu^3 / [e^{c_2\nu/T} - 1] \quad (mW/m^2/ster/cm^{-1})$$

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 $\nu max) = 1.95T$

indicates peak of Planck function curve shifts to shorter wavelengths (greater wavenumbers) with temperature increase.

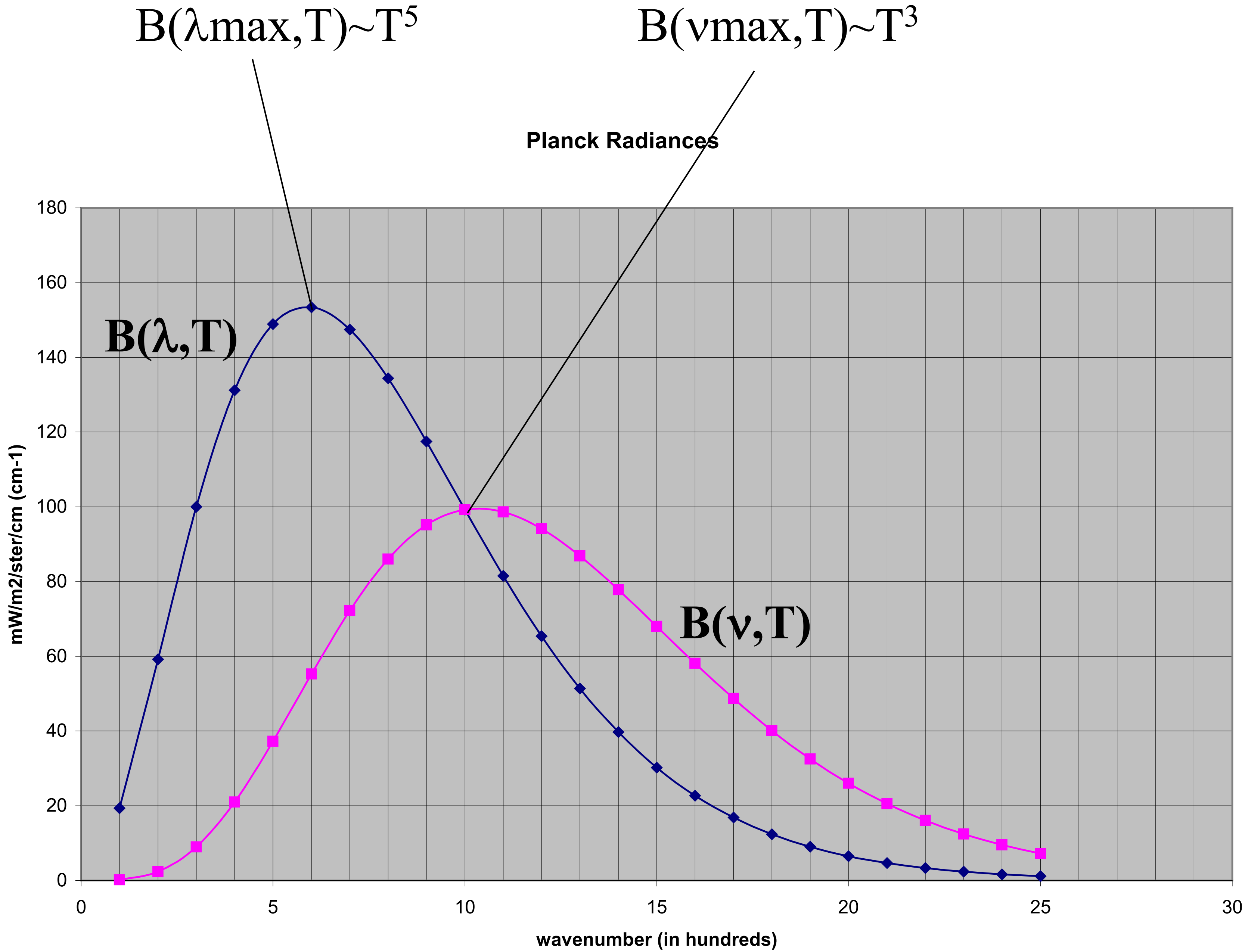
Stefan-Boltzmann Law

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

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



B(λ,T) versus B(ν,T)

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

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

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

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

Using wavelengths

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

(mW/m^2/ster/μm)

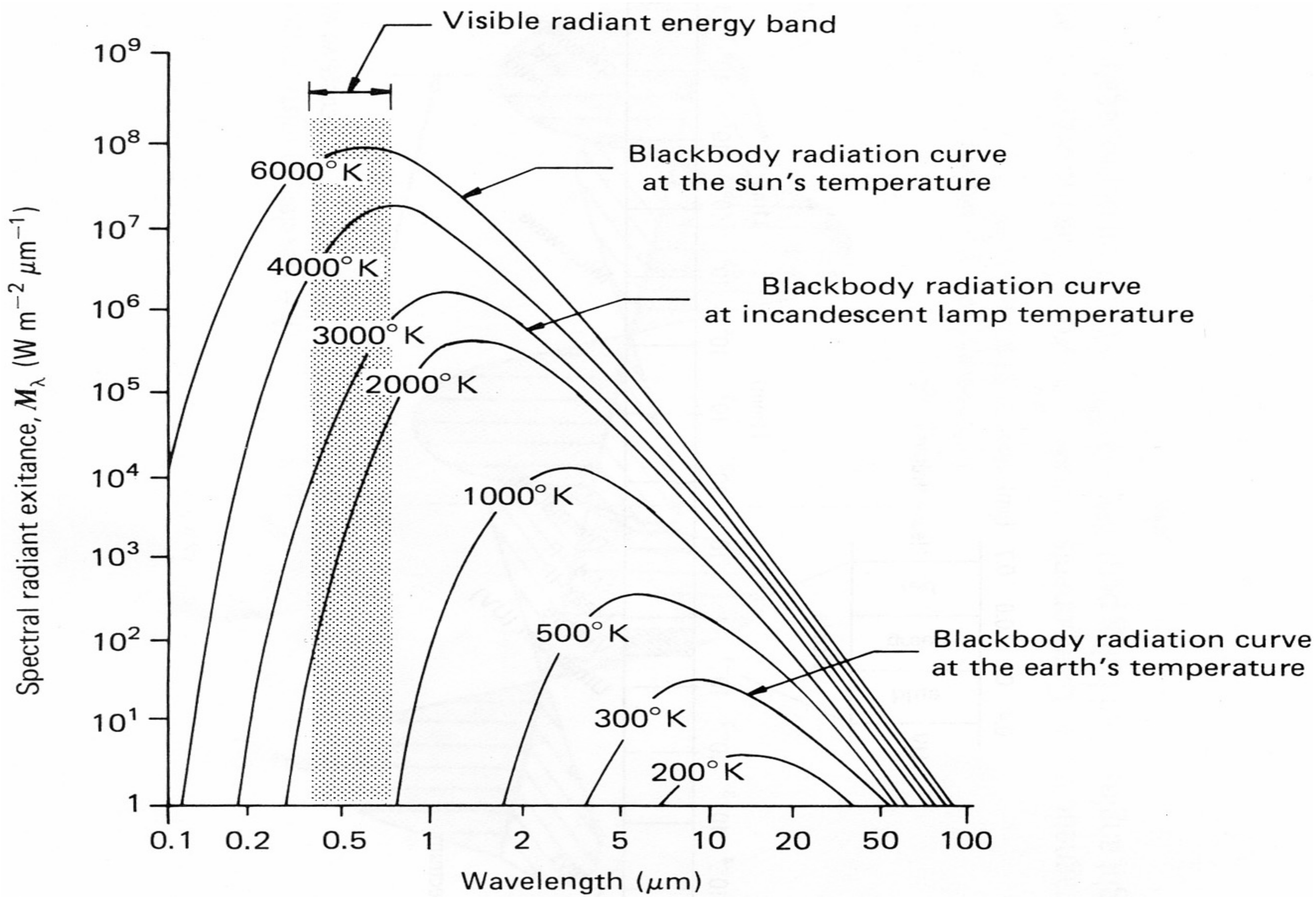
λ(max in cm)T = 0.2897

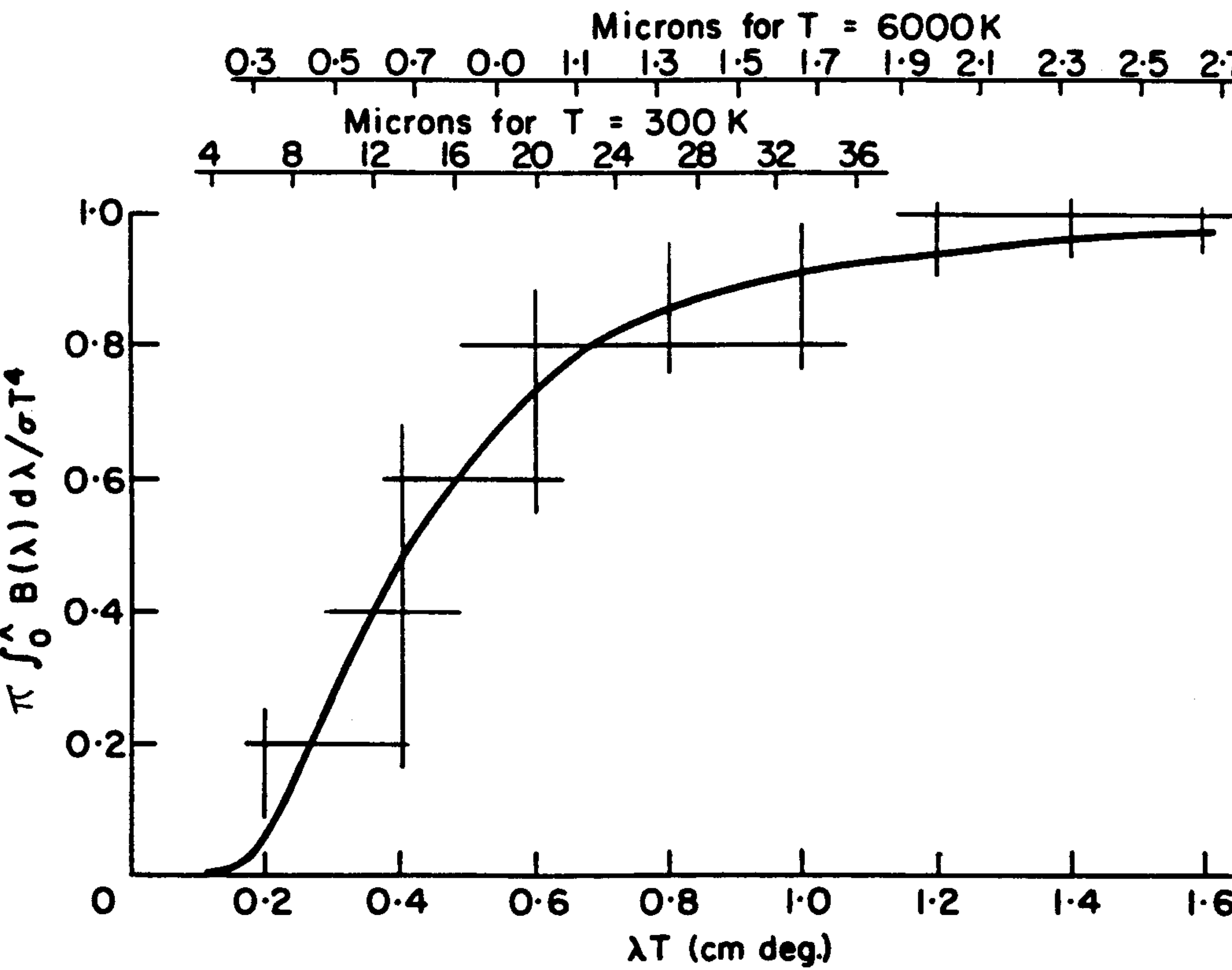
$B(\lambda_{max},T) \sim T**5.$

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

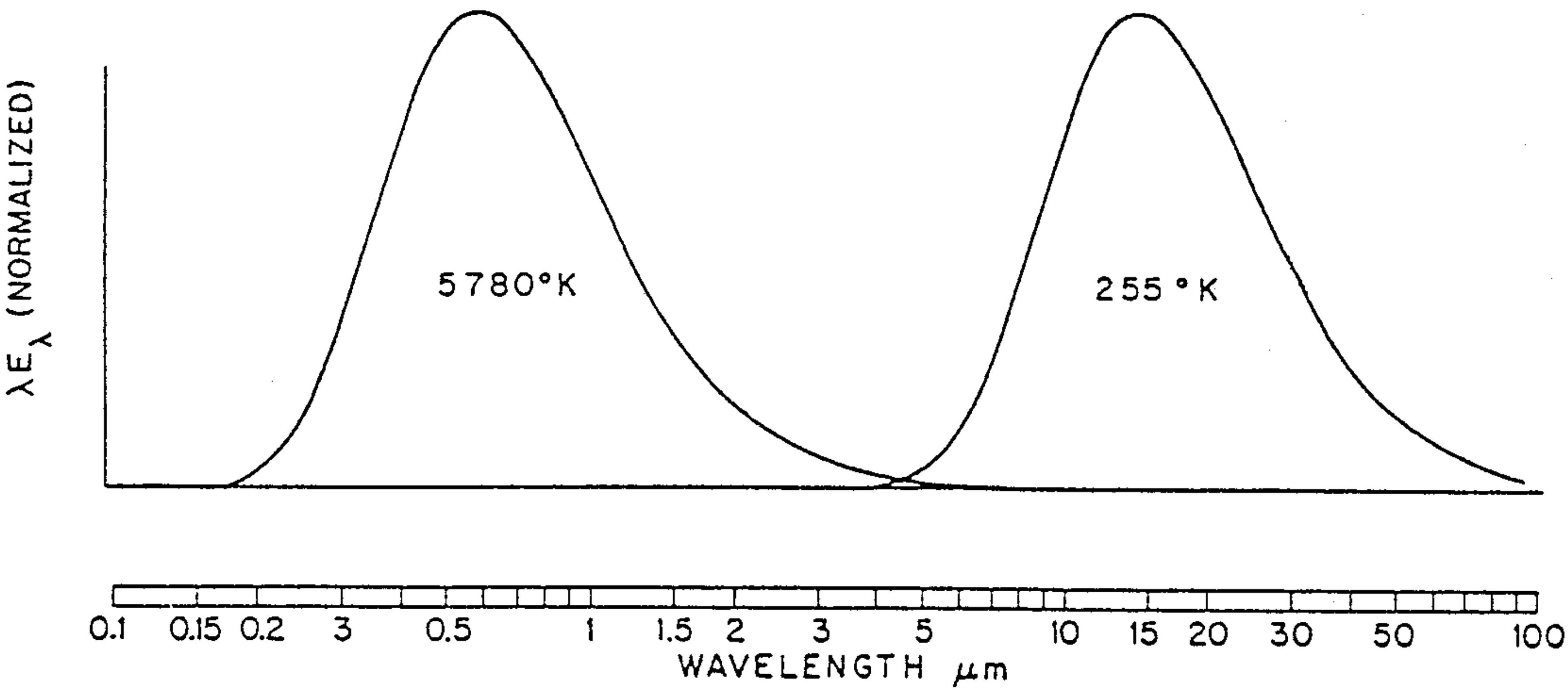
$$T = c_2 / [\lambda \ln(\frac{c_1}{\lambda^5 B_\lambda} + 1)]$$

Spectral Distribution of Energy Radiated from Blackbodies at Various Temperatures



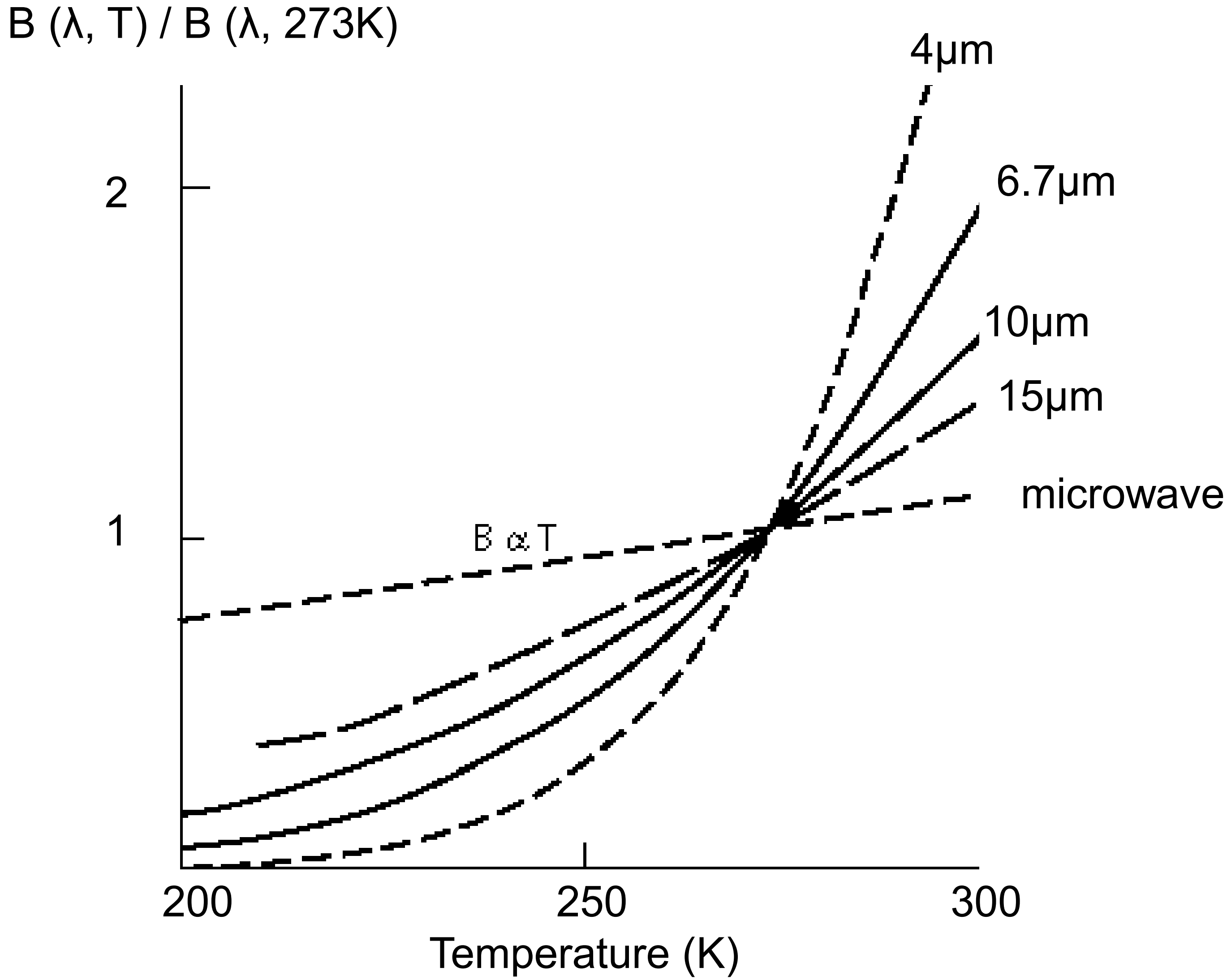


Microns for T = 6000K
0·3 0·5 0·7 0·0 1·1 1·3 1·5 1·7 1·9 2·1 2·3 2·5 2·7
Microns for T = 300 K
4 8 12 16 20 24 28 32 36
1·0
0·8
0·6
0·4
0·2
0
$\pi \int_0^\lambda B(\lambda)\, d\lambda / \sigma T^4$
0·2 0·4 0·6 0·8 1·0 1·2 1·4 1·6
λT (cm deg.)

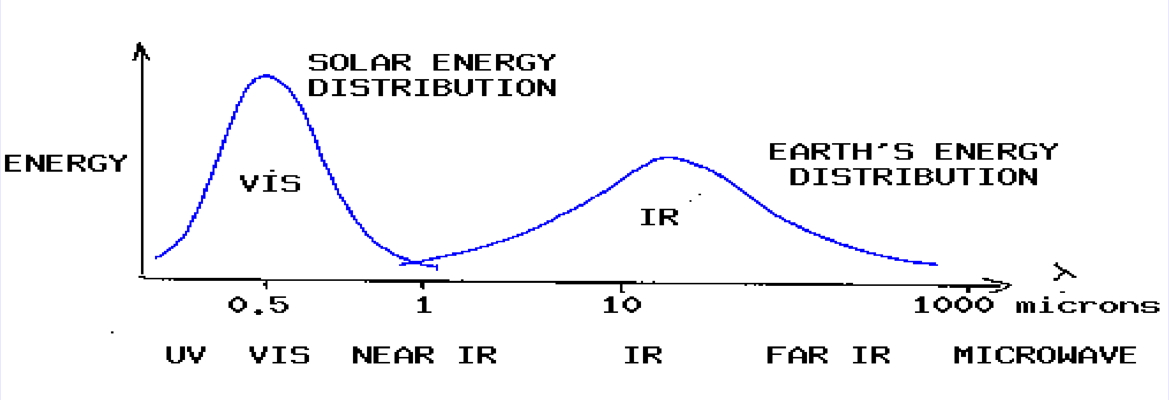


Normalized black body spectra representative of the sun (left) and earth (right), plotted on a logarithmic wavelength scale. The ordinate is multiplied by wavelength so that the area under the curves is proportional to irradiance.

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



Solar (visible) and Earth emitted (infrared) energy

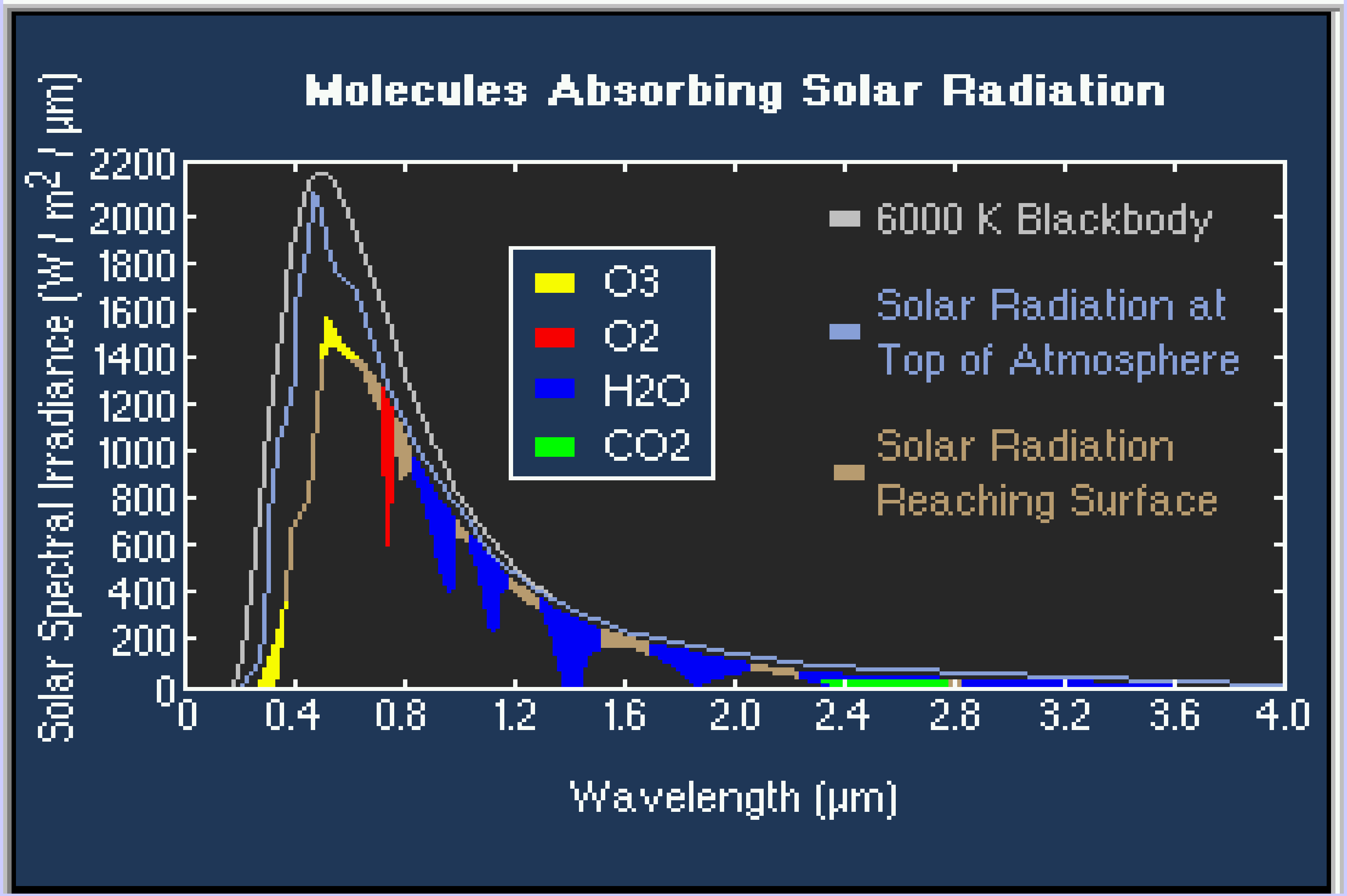


Incoming solar radiation (mostly visible) drives the earth-atmosphere (which emits infrared).

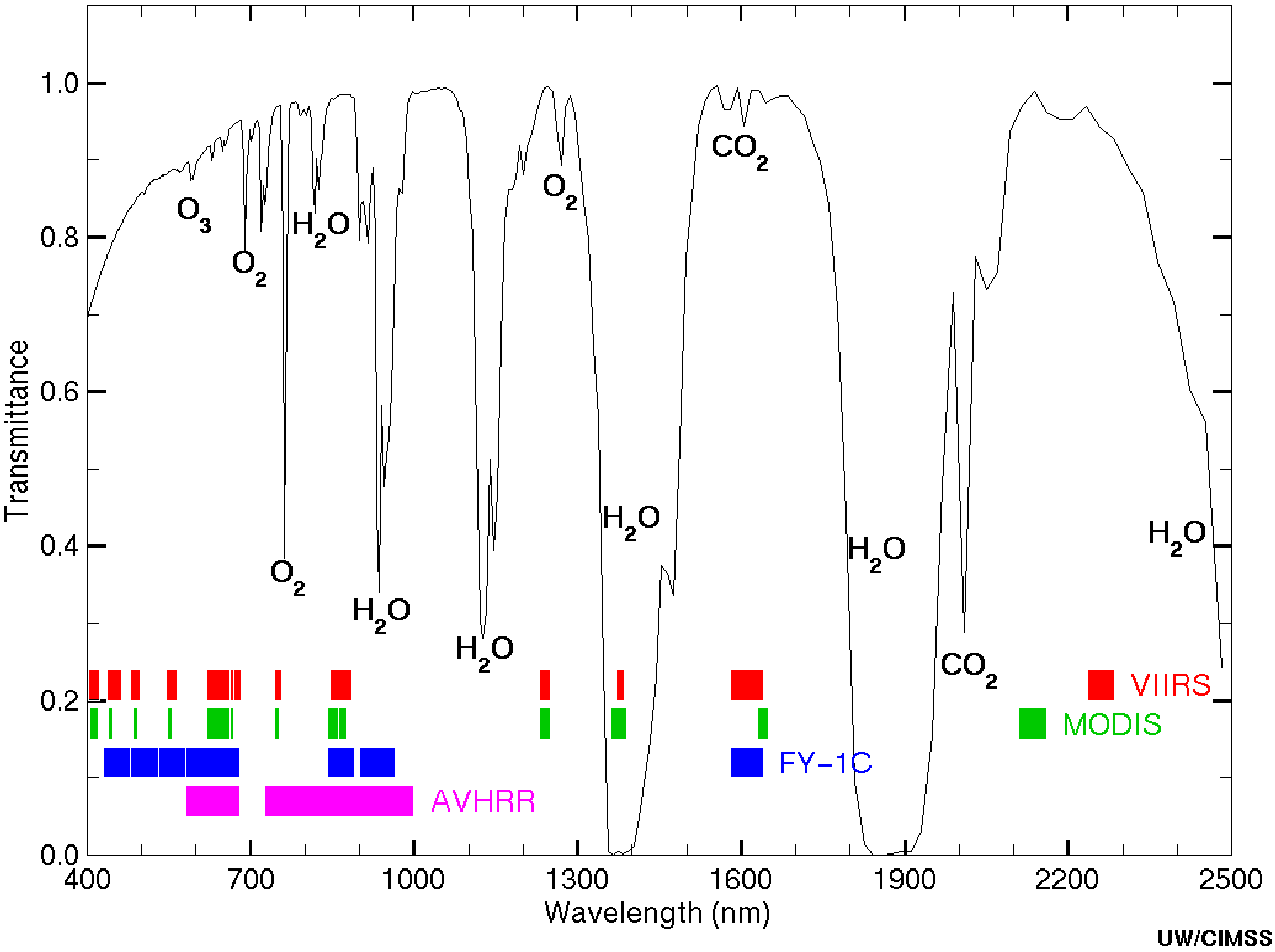
Over the annual cycle, the incoming solar energy that makes it to the earth surface (about 50 %) is balanced by the outgoing thermal infrared energy emitted through the atmosphere.

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.

Solar Spectrum

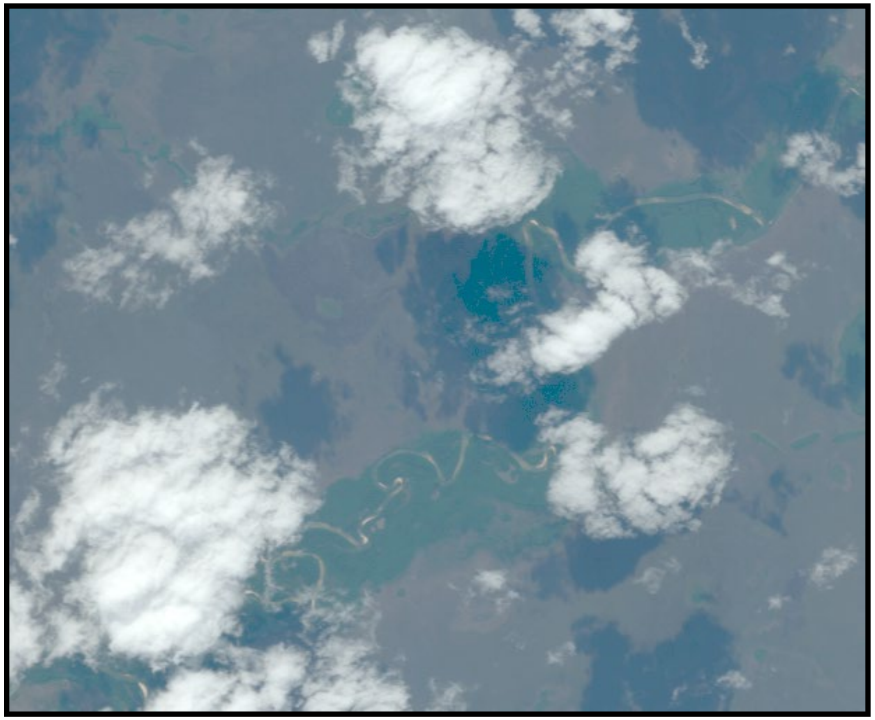


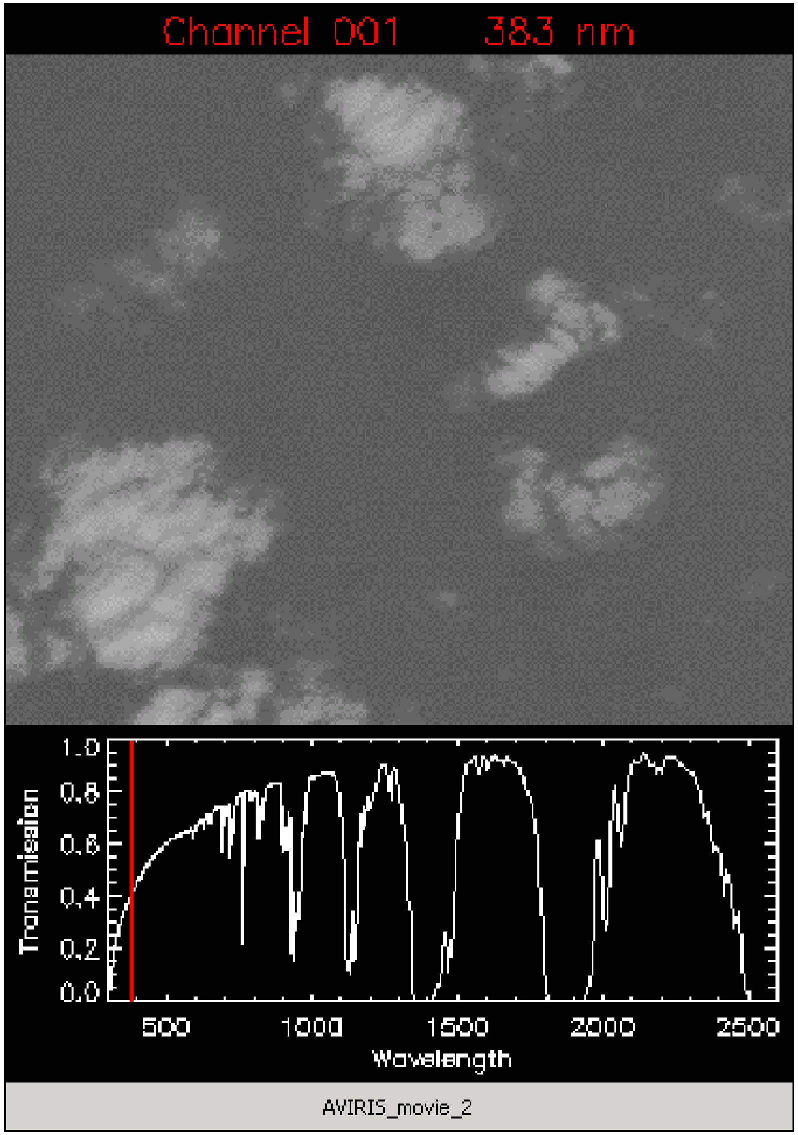
VIIRS, MODIS, FY-1C, AVHRR

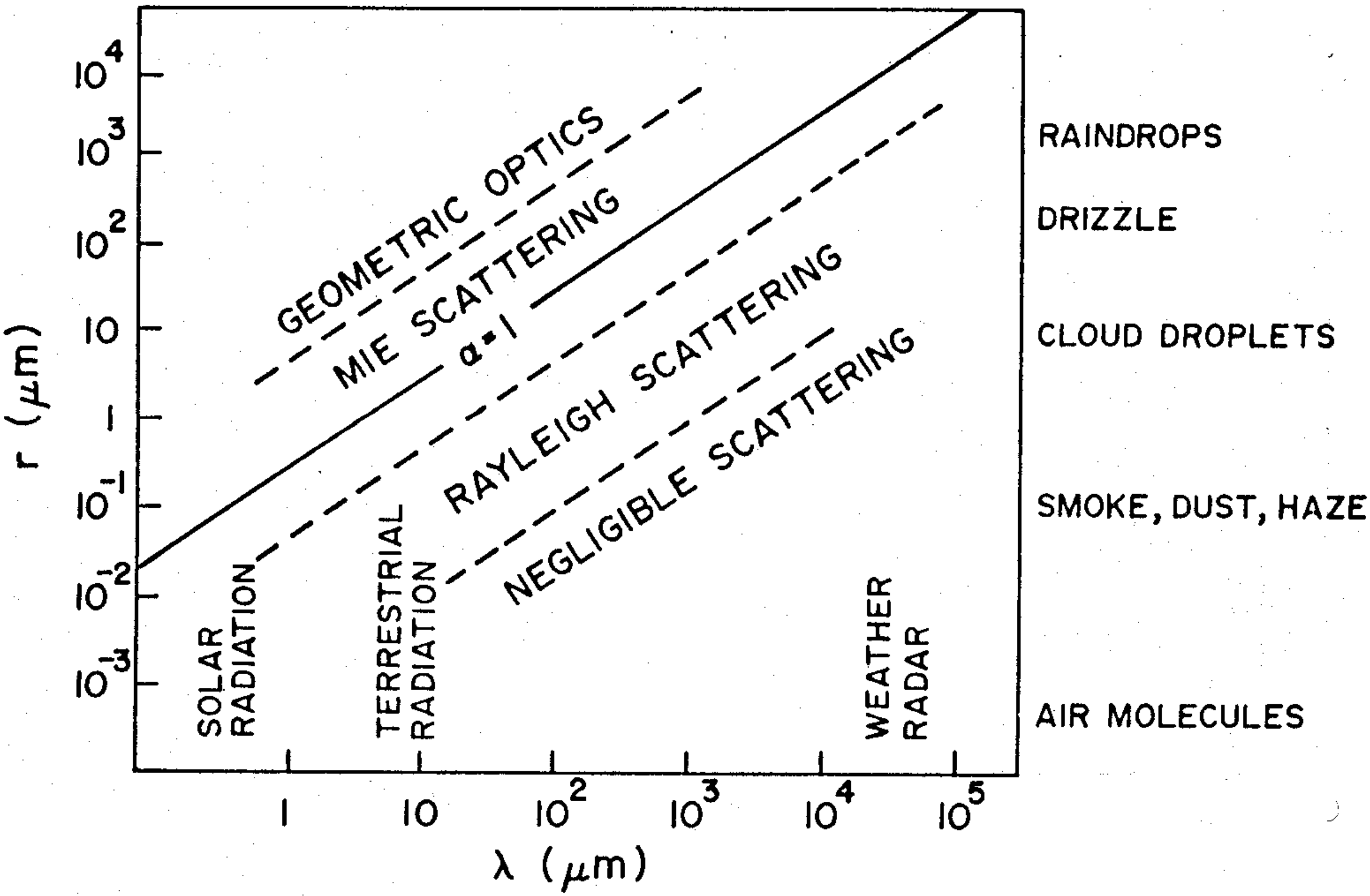


AVIRIS Movie #2

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






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)
10^4
10^3
10^2
10
1
10^-1
10^-2
10^-3
1
10
10^2
10^3
10^4
10^5

Aerosol Size Distribution

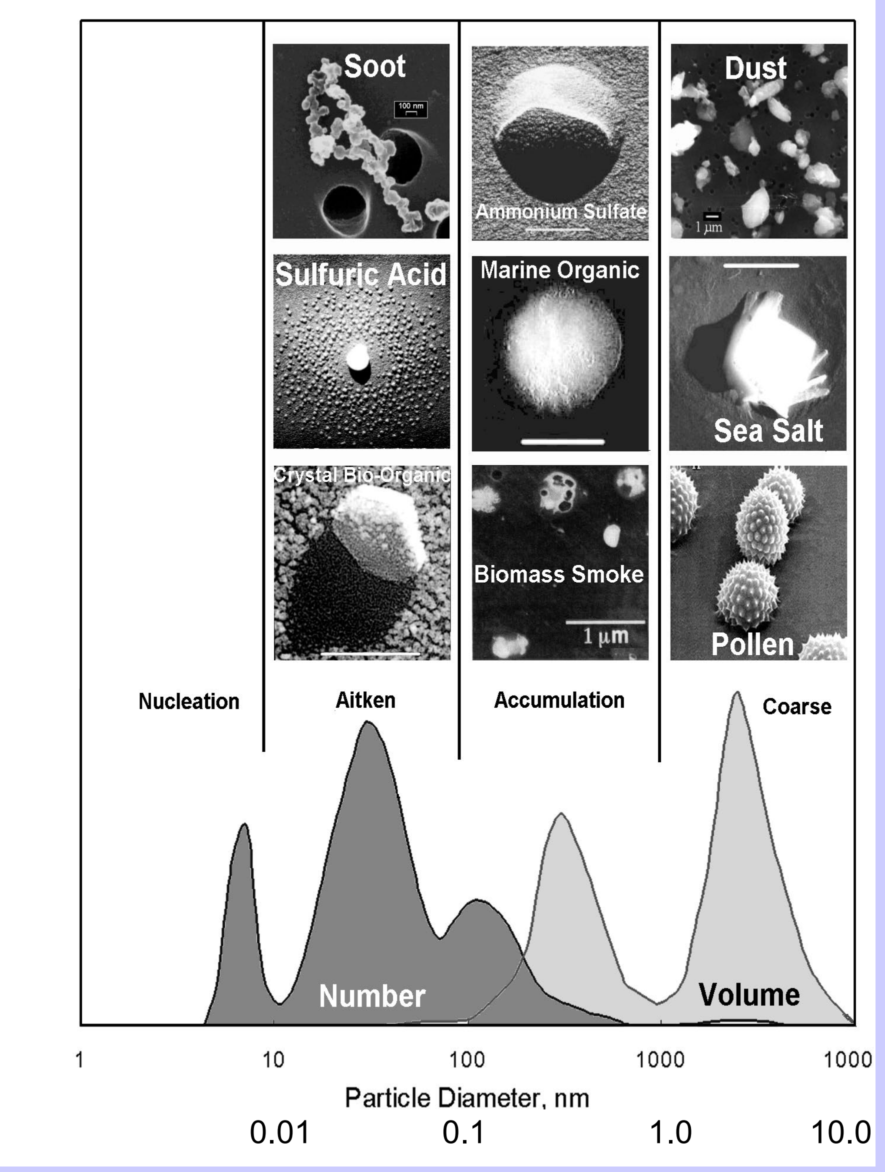
There are 3 modes :

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

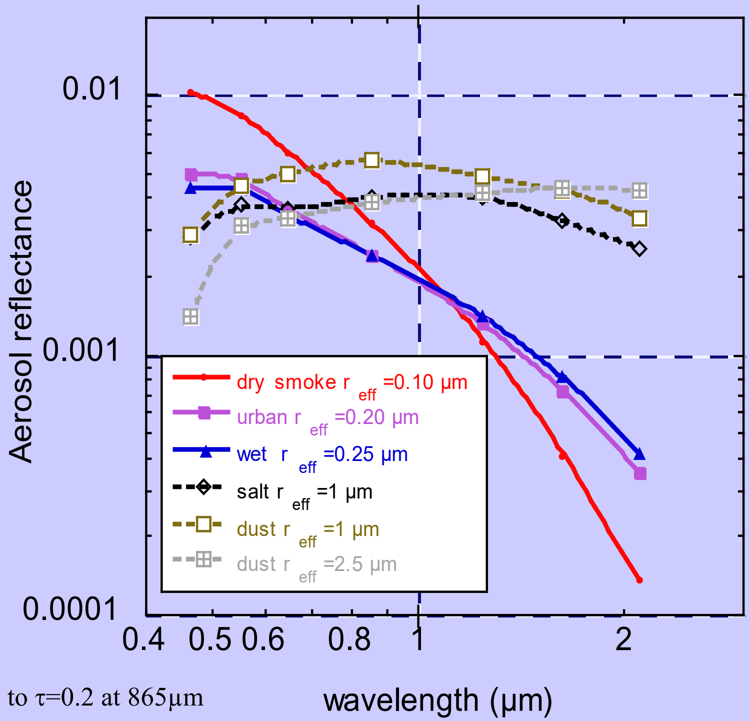
- « **accumulation »:** radius is between 0.05 µm and 0.5 µm**.** Coagulation processes.

- « **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.



Aerosols over Ocean

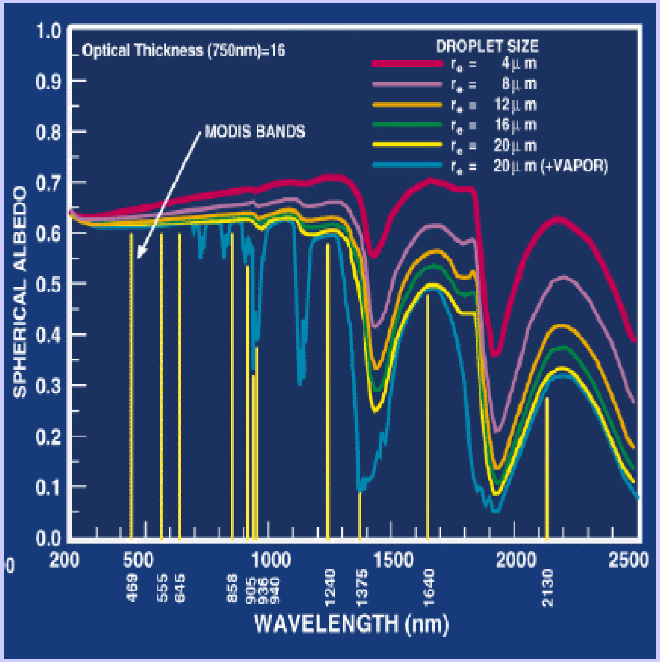


Normalized to τ=0.2 at 865µm

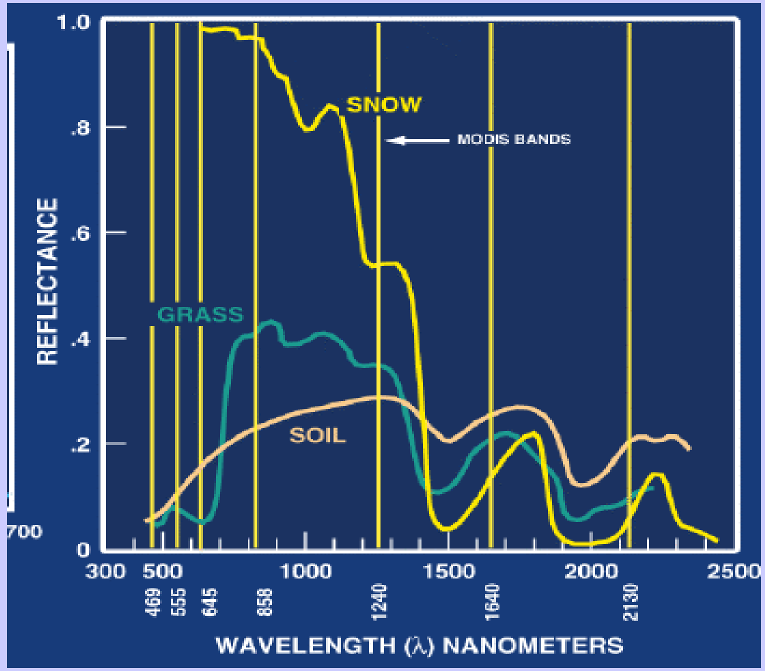
- Radiance data in 6 bands (550-2130nm).
- Spectral radiances (LUT) to derive the aerosol size distribution
- Two modes (accumulation 0.10-0.25µm; coarse1.0-2.5µm); ratio is a free parameter
- Radiance at 865µm to derive τ

Ocean products :

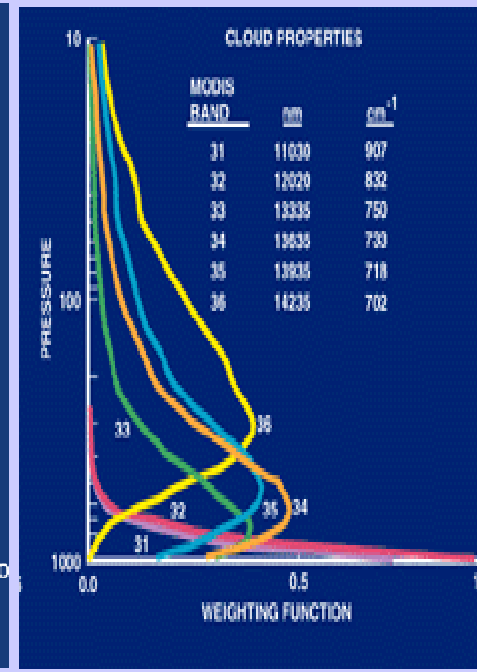
- The total Spectral Optical thickness
- The effective radius
- The optical thickness of small & large modes/ratio between the 2 modes



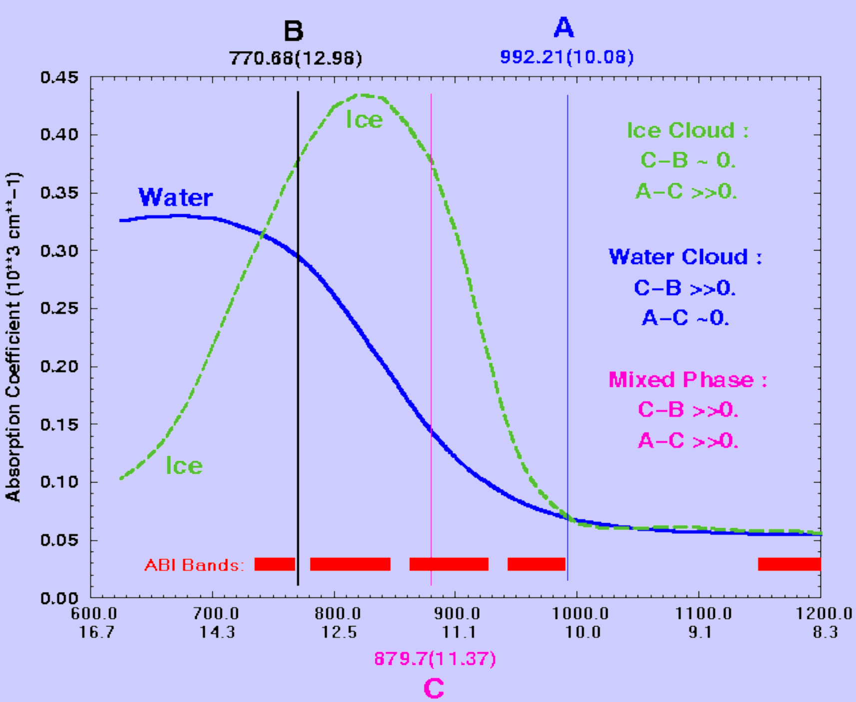
Optical Thickness (750nm)=16
DROPLET SIZE
r_e = 4 μ m
r_e = 8 μ m
r_e = 12 μ m
r_e = 16 μ m
r_e = 20 μ m
r_e = 20 μ m (+VAPOR)
MODIS BANDS
SPHERICAL ALBEDO
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0.0
200
500
1000
1500
2000
2500
469
555
645
858
905
936
940
1240
1375
1640
2130
WAVELENGTH (nm)



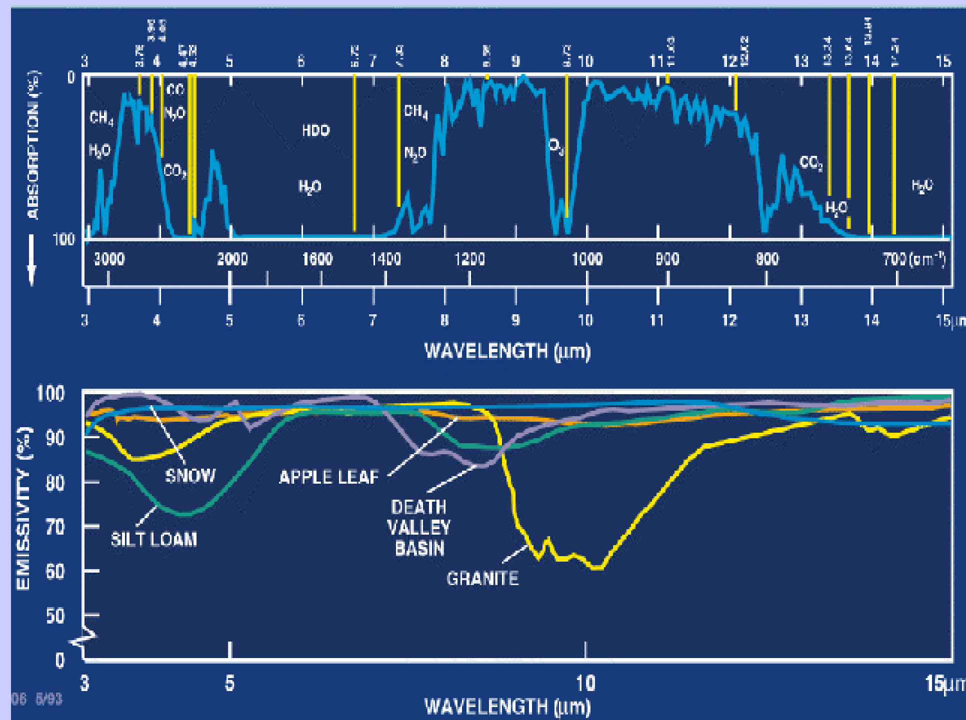
SNOW
MODIS BANDS
GRASS
SOIL
REFLECTANCE
1.0
.8
.6
.4
.2
0
300
500
1000
1500
2000
2500
469
555
645
858
1240
1640
2130
WAVELENGTH (λ) NANOMETERS



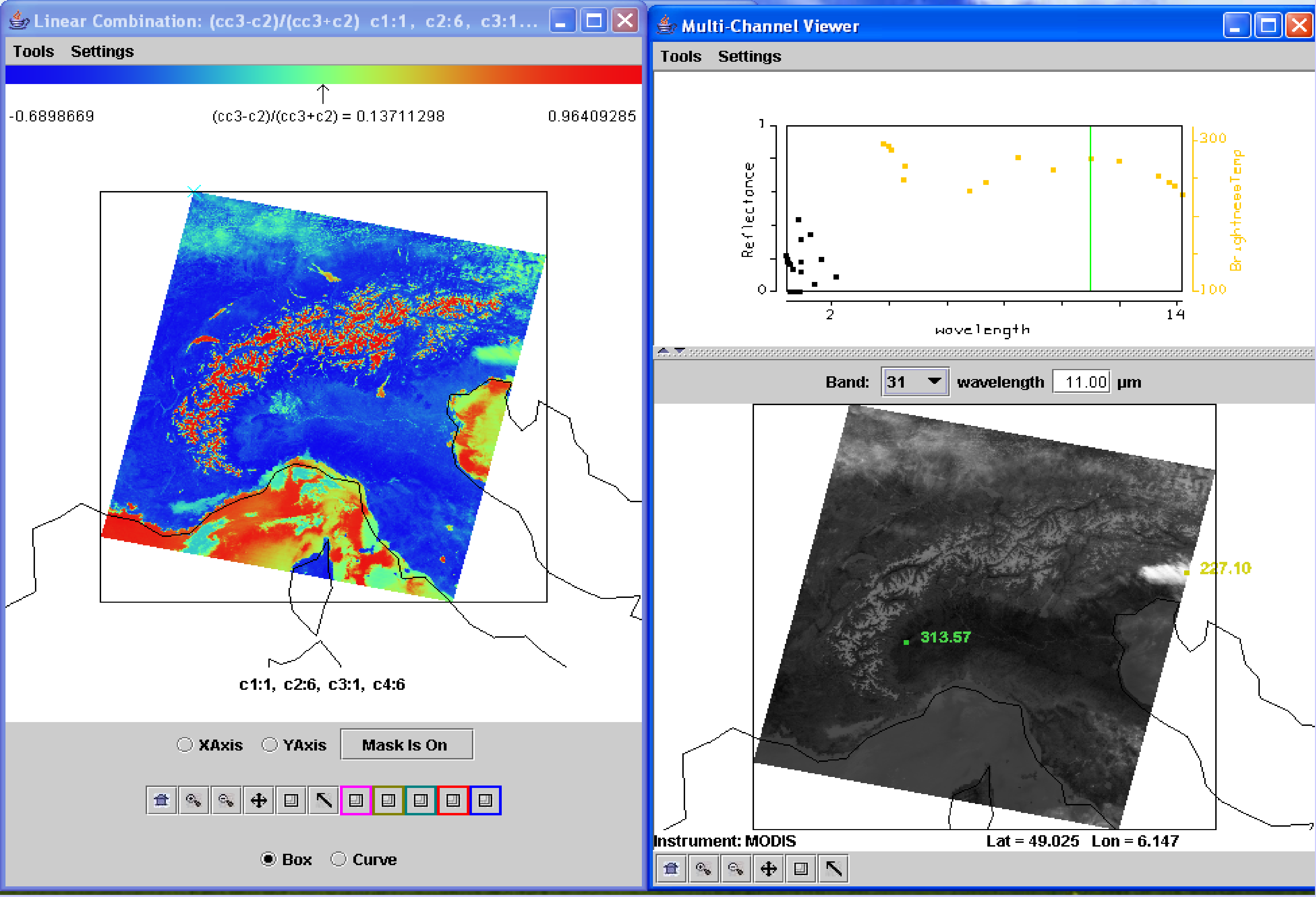
CLOUD PROPERTIES
MODIS BAND
nm
cm⁻¹
31 11030 907
32 12020 832
33 13335 750
34 13635 733
35 13935 718
36 14235 702
PRESSURE
10
100
1000
33
36
32
35
34
31
0.0
0.5
1.0
WEIGHTING FUNCTION



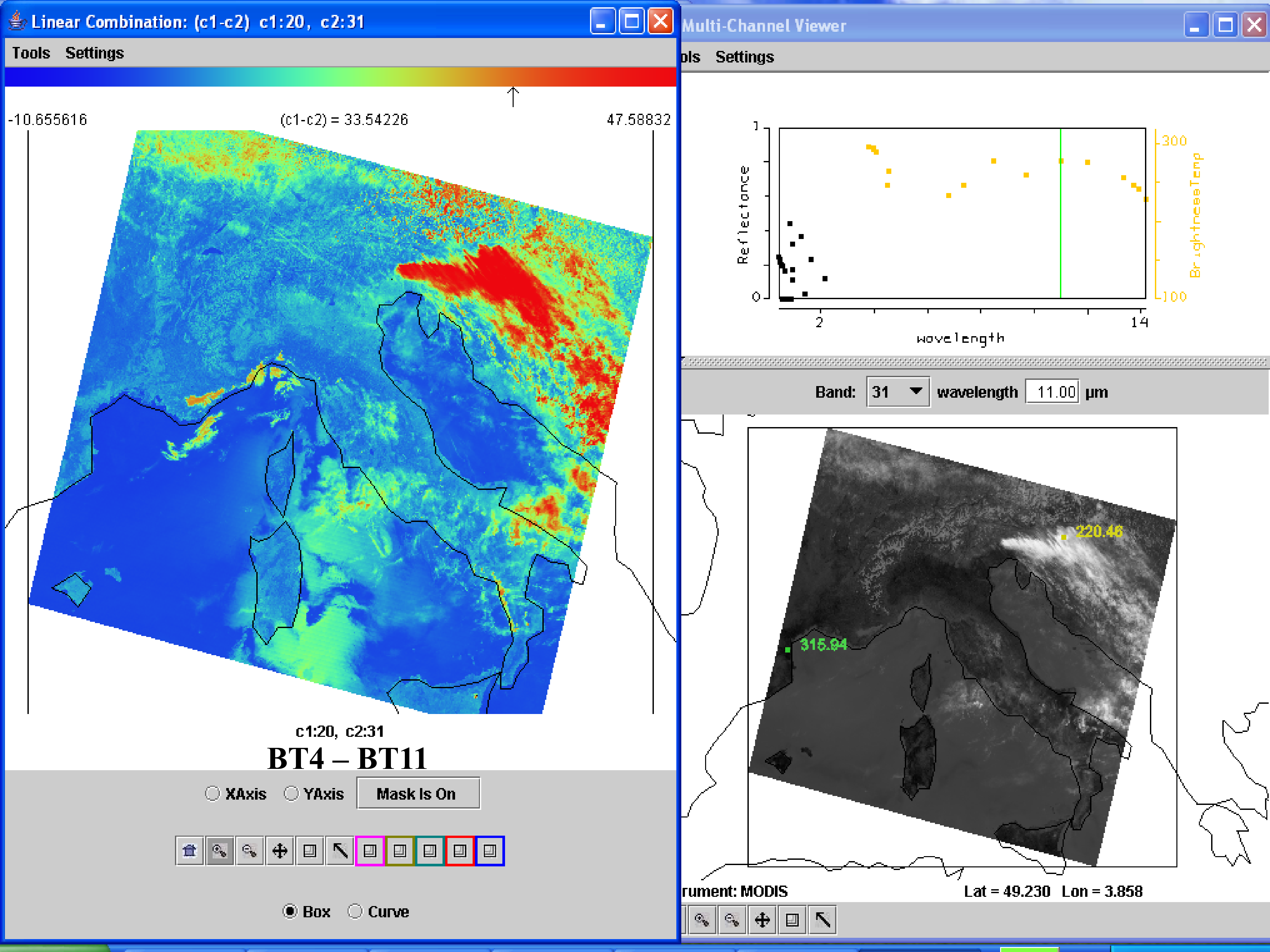
B
770.68(12.98)
A
992.21(10.08)
Ice
Water
Ice
Ice Cloud :
C–B ~ 0.
A–C >>0.
Water Cloud :
C–B >>0.
A–C ~0.
Mixed Phase :
C–B >>0.
A–C >>0.
ABI Bands:
Absorption Coefficient (10**3 cm**-1)
0.45
0.40
0.35
0.30
0.25
0.20
0.15
0.10
0.05
0.00
600.0 16.7
700.0 14.3
800.0 12.5
900.0 11.1
1000.0 10.0
1100.0 9.1
1200.0 8.3
879.7(11.37)
C



ABSORPTION (%)
0
100
CH4
H2O
CO
N2O
CO2
HDO
H2O
CH4
N2O
O3
CO2
H2O
H2O
3000
2000
1600
1400
1200
1000
900
800
700 (cm⁻¹)
3
4
5
6
7
8
9
10
11
12
13
14
15μm
WAVELENGTH (μm)
EMISSIVITY (%)
100
90
80
70
60
50
0
SNOW
APPLE LEAF
DEATH VALLEY BASIN
SILT LOAM
GRANITE
3
5
10
15μm
WAVELENGTH (μm)
06 5/93



NDSI = [r0.6-r1.6]/[r0.6+r1.6] is near one in snow in Alps



Linear Combination: (c1-c2) c1:20, c2:31
Tools Settings
-10.655616
(c1-c2) = 33.54226
47.58832
c1:20, c2:31
BT4 – BT11
XAxis
YAxis
Mask Is On
Box
Curve
Multi-Channel Viewer
ols Settings
Reflectance
BrightnessTemp
wavelength
300
100
2
14
Band: 31 wavelength 11.00 µm
220.46
315.94
rument: MODIS
Lat = 49.230 Lon = 3.858

Selective Absorption
Atmosphere transmits visible and traps infrared

Incoming solar   Outgoing IR

$\downarrow E$ $\uparrow (1-a_l)\, Y_{sfc}$ $\uparrow Y_a$

top of the atmosphere

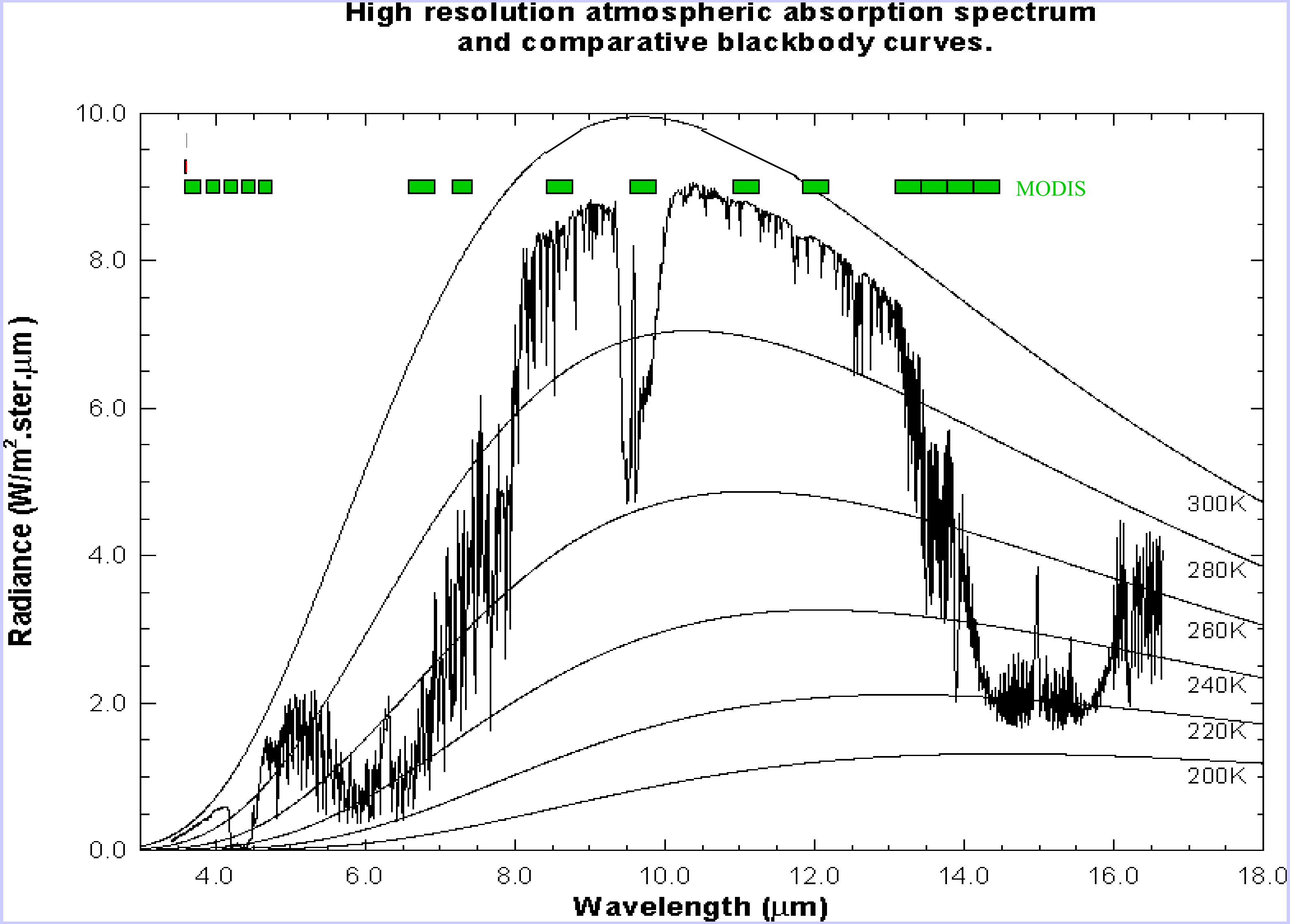
$\downarrow (1-a_s)\, E$ $\uparrow Y_{sfc}$ $\downarrow Y_a$

earth surface.

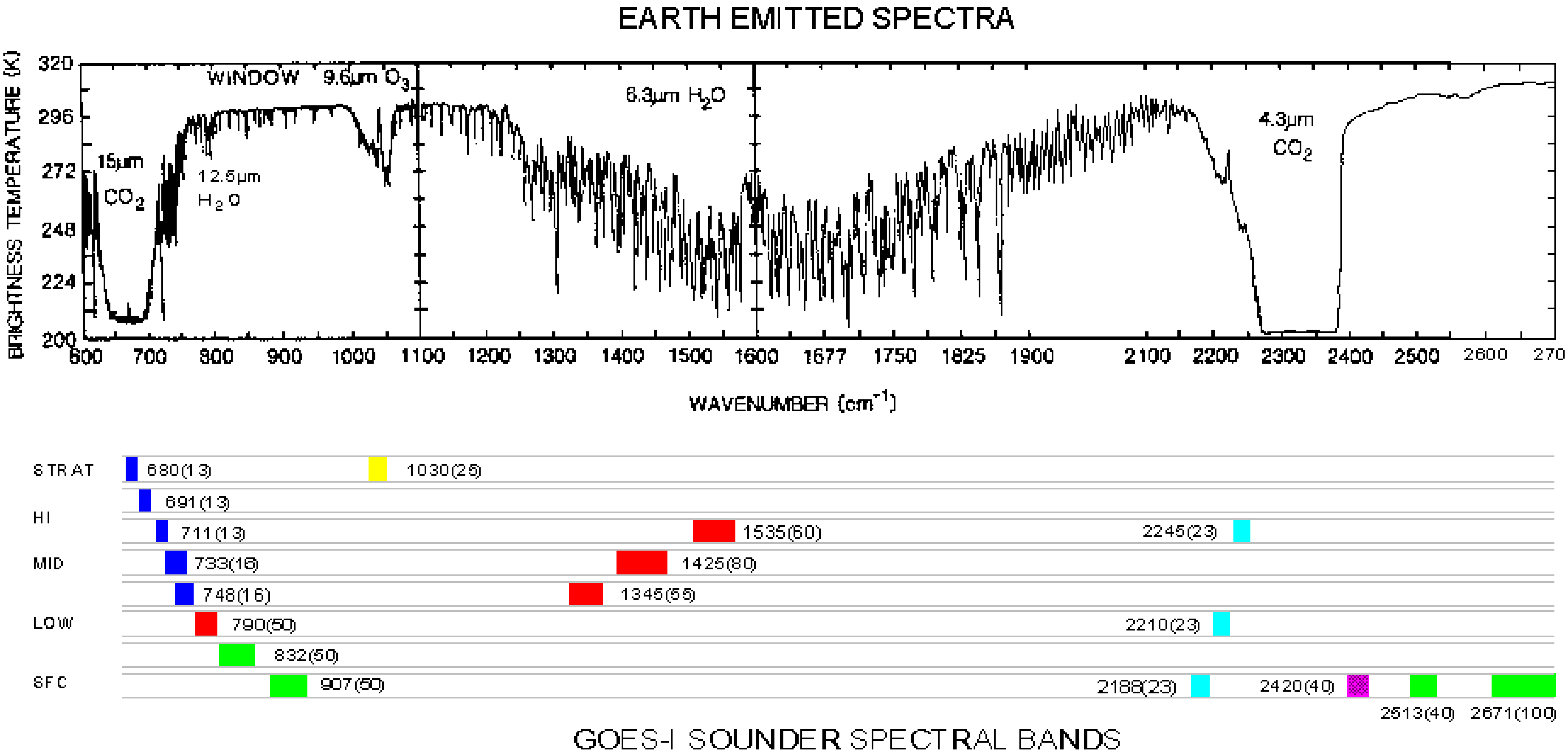
$$Y_{sfc} = \frac{(2-a_S)}{(2-a_L)} E = \sigma T_{sfc}^4$$

thus if $a_s < a_L$ then $Y_{sfc} > E$

MODIS IR Spectral Bands

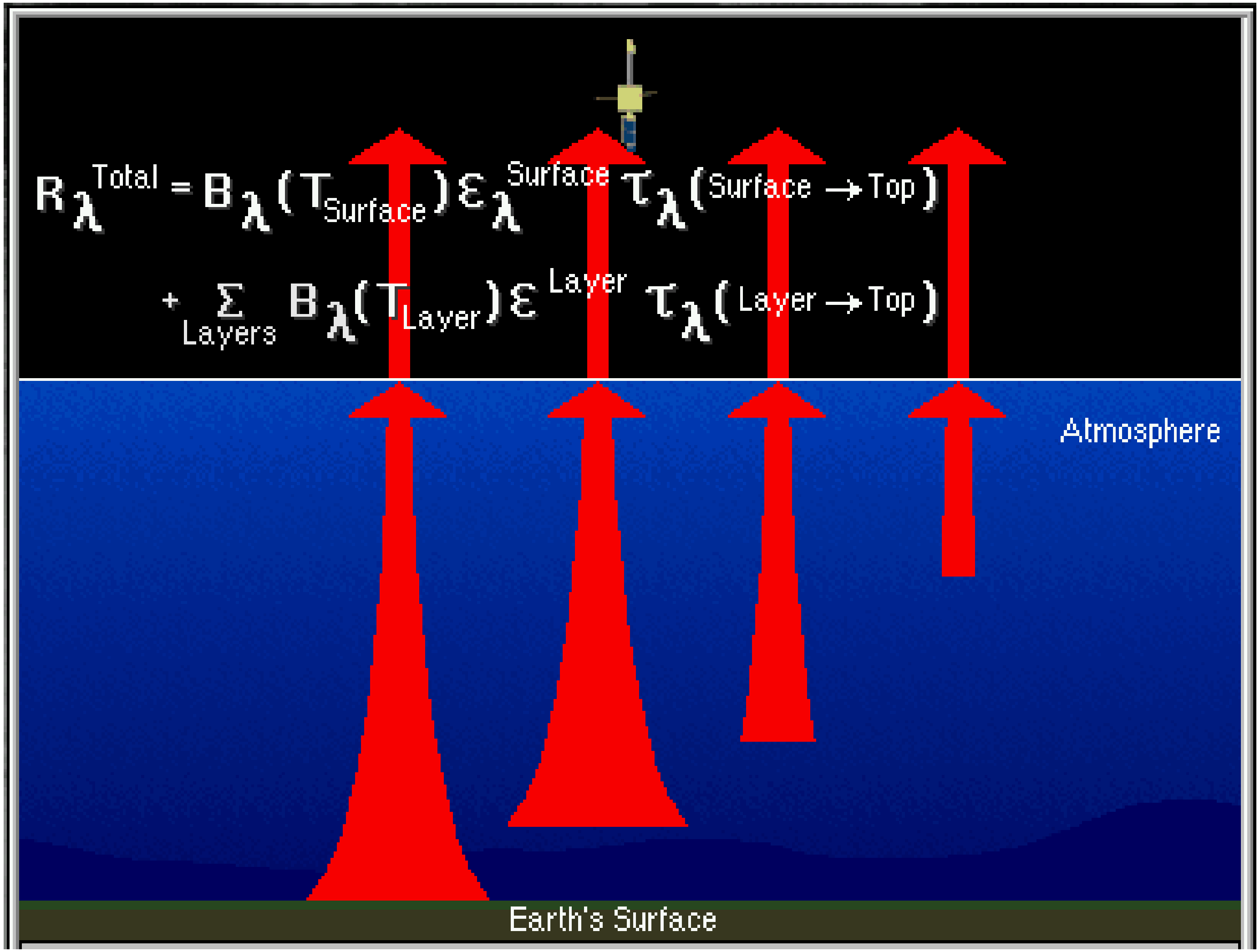


GOES Sounder Spectral Bands: 14.7 to 3.7 um and vis

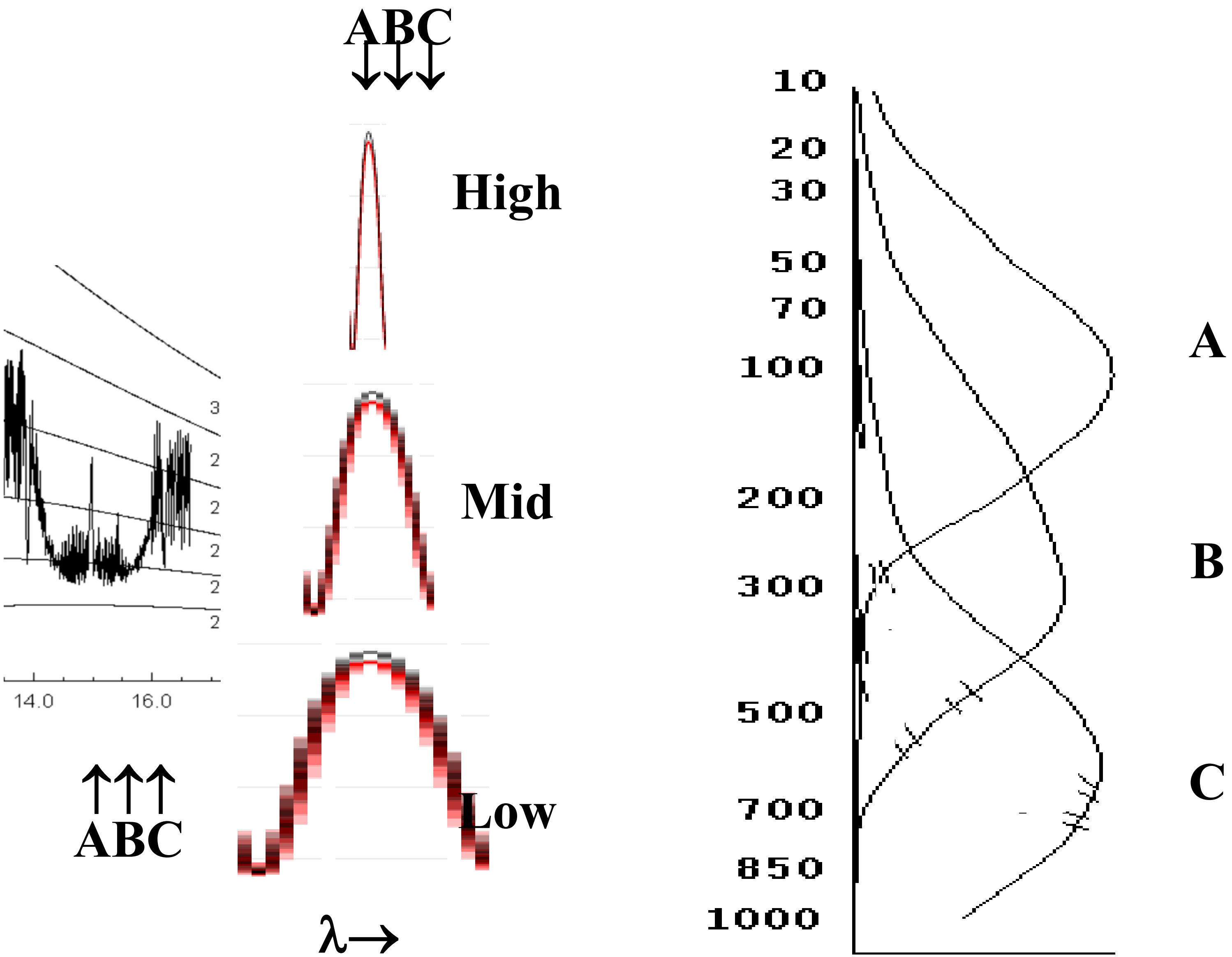


COOPERATIVE INSTITUTE FOR METEOROLOGICAL SATELLITE STUDIES

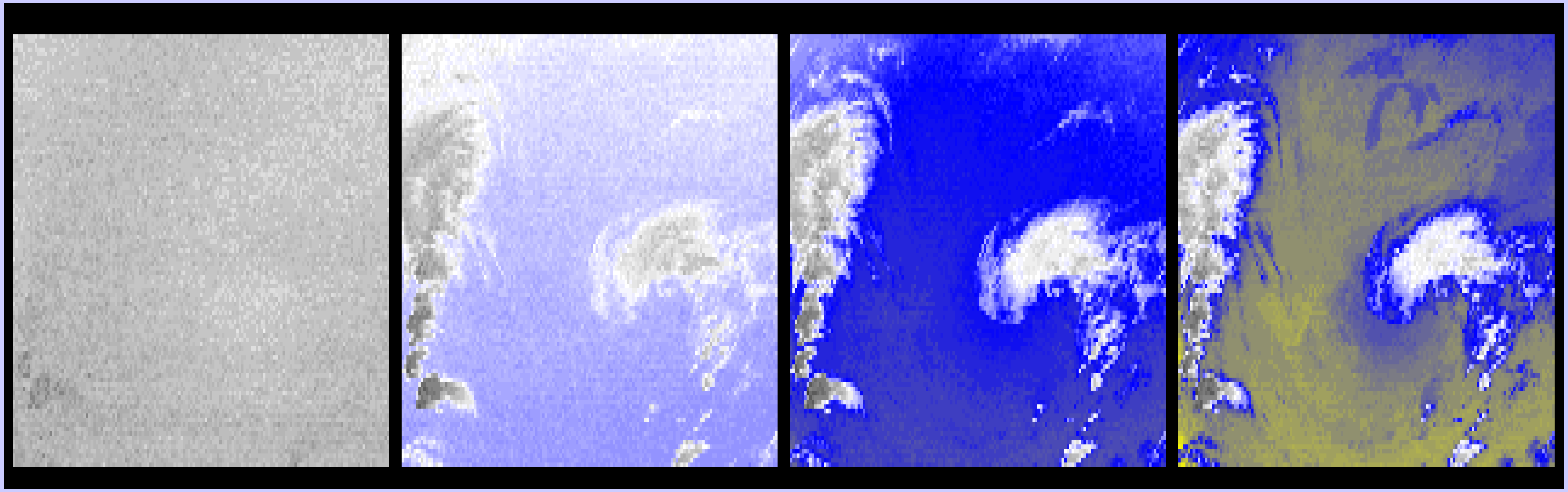
Radiative Transfer through the Atmosphere



line broadening with pressure helps to explain weighting functions



CO2 channels see to different levels in the atmosphere



14.2 um 13.9 um 13.6 um 13.3 um

Radiative Transfer Equation

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

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

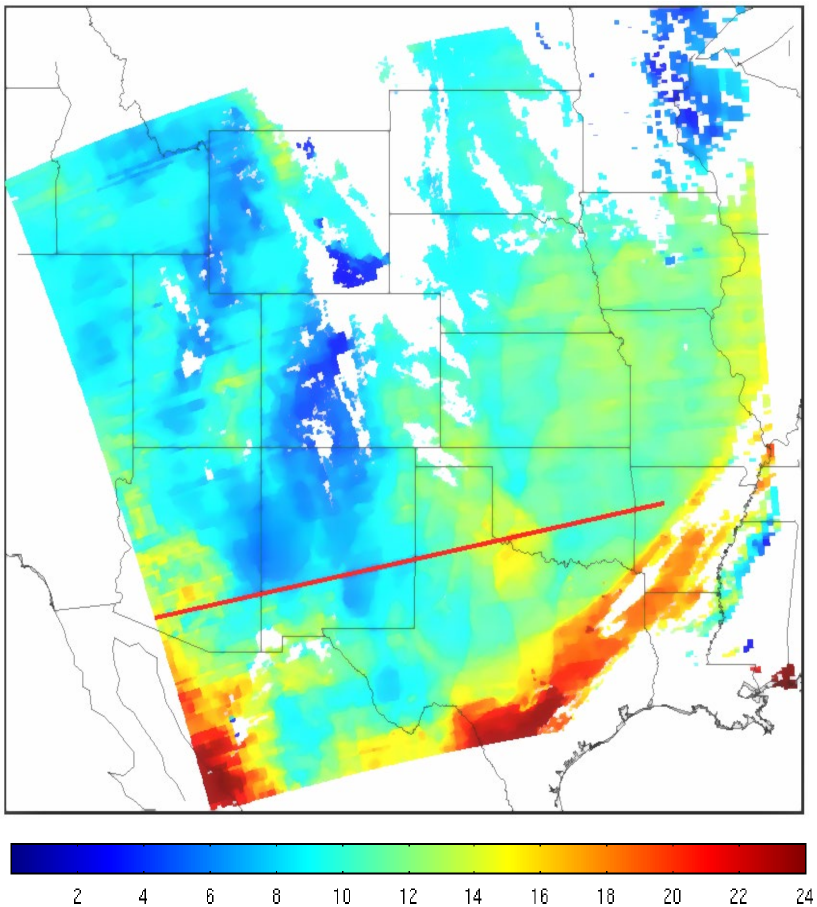
where

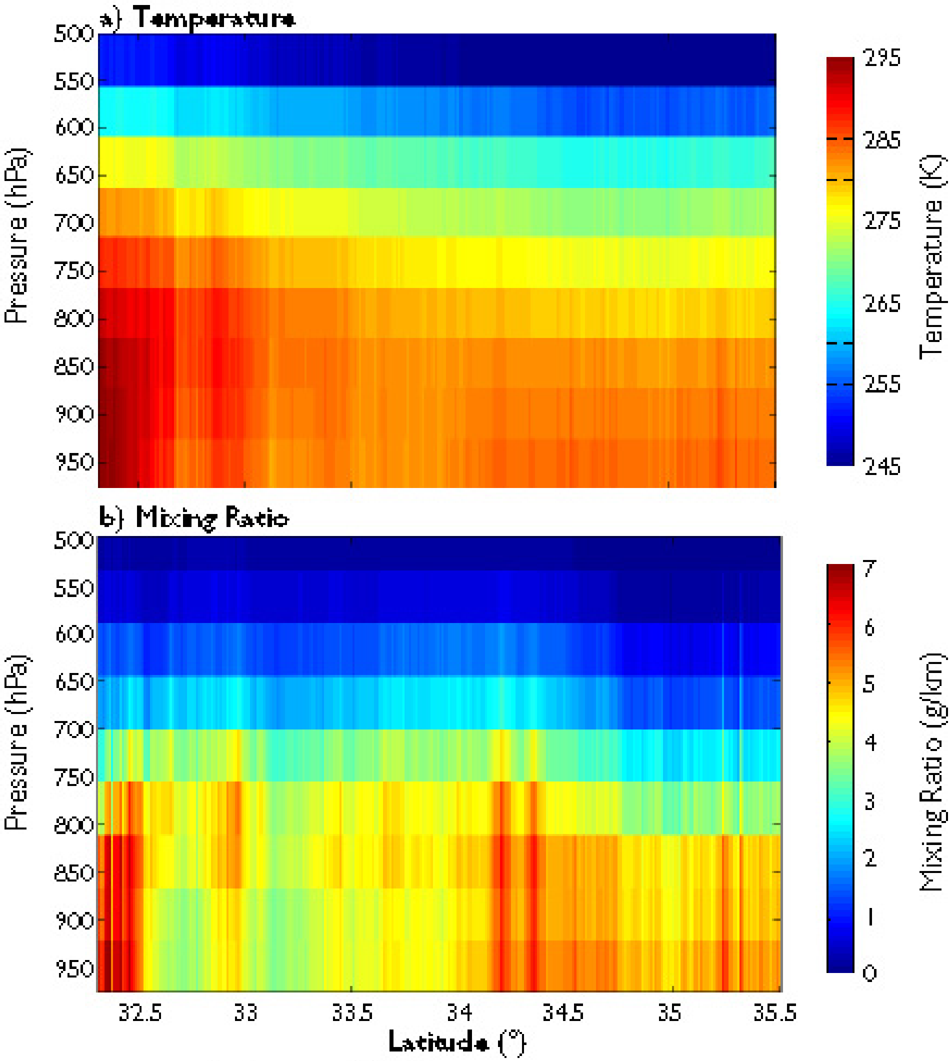
$$F_\lambda(p) = \{ 1 + (1 - \varepsilon_\lambda) [\tau_\lambda(p_s) / \tau_\lambda(p)]^2 \}$$

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.

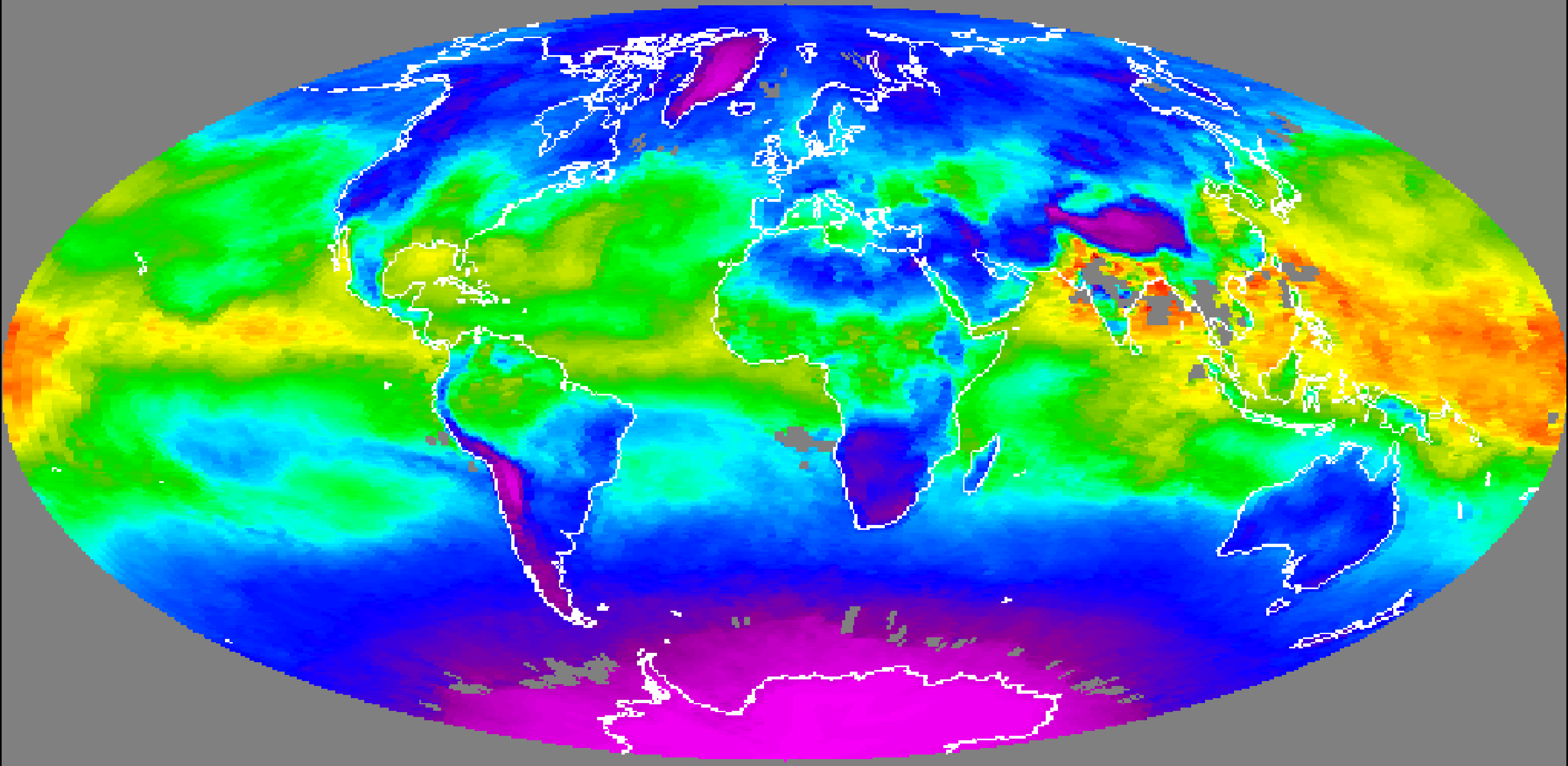
The atmospheric contribution is the weighted sum of the Planck radiance contribution from each layer, where the weighting function is [$d\tau_\lambda(p) / dp$]. This weighting function is an indication of where in the atmosphere the majority of the radiation for a given spectral band comes from.

MODIS TPW





Clear sky layers of temperature and moisture on 2 June 2001



2 AUG 2002 **Global TPW from Seemann**

RTE in Cloudy Conditions

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

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

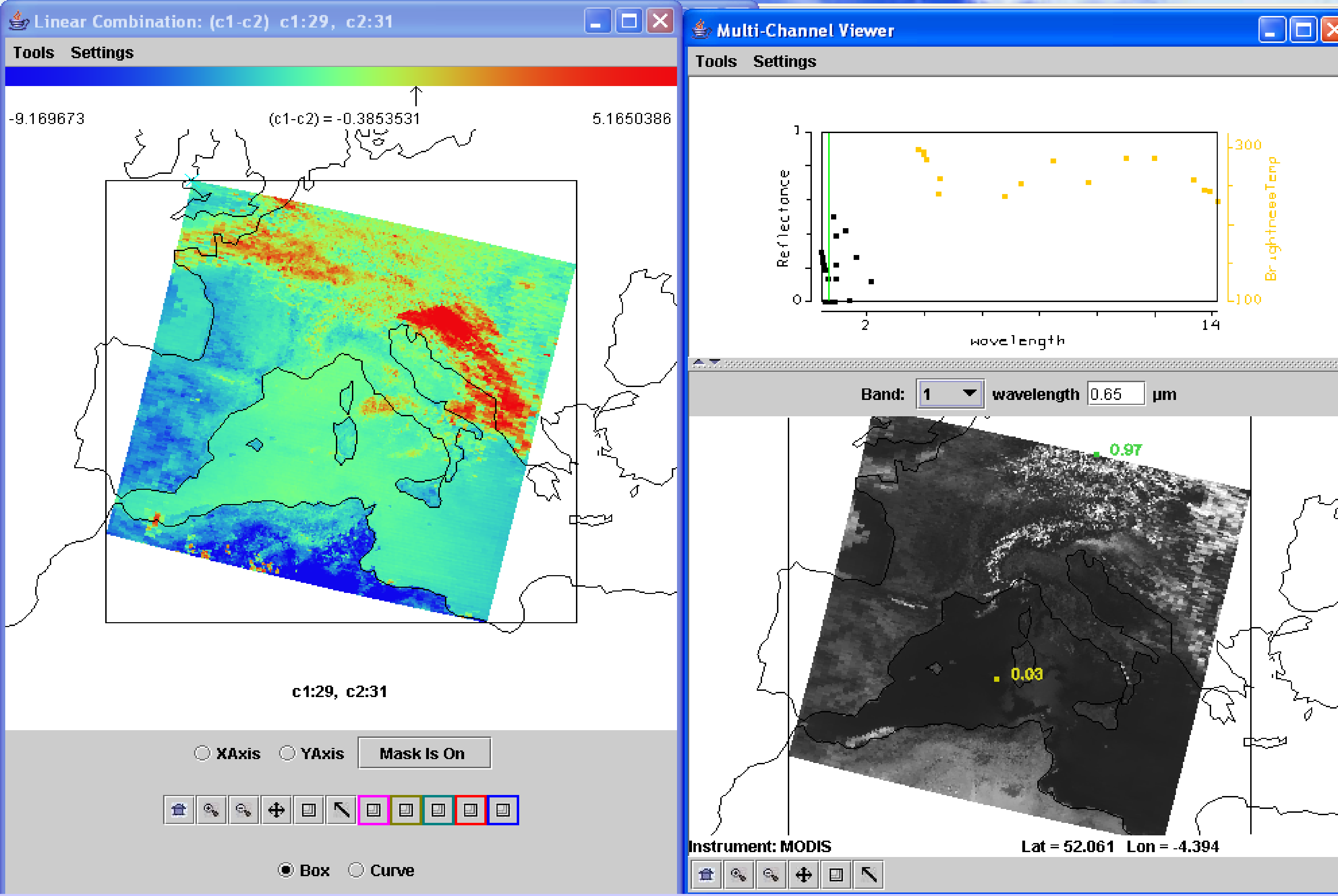
$$I^{cd}_\lambda = (1-\varepsilon_\lambda)\, B_\lambda(T_s)\, \tau_\lambda(p_s) + (1-\varepsilon_\lambda) \int_{p_s}^{p_c} B_\lambda(T(p))\, d\tau_\lambda$$

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

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.



Ice clouds are revealed with BT8.6-BT11>0 & water clouds and fog show in r0.65

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)

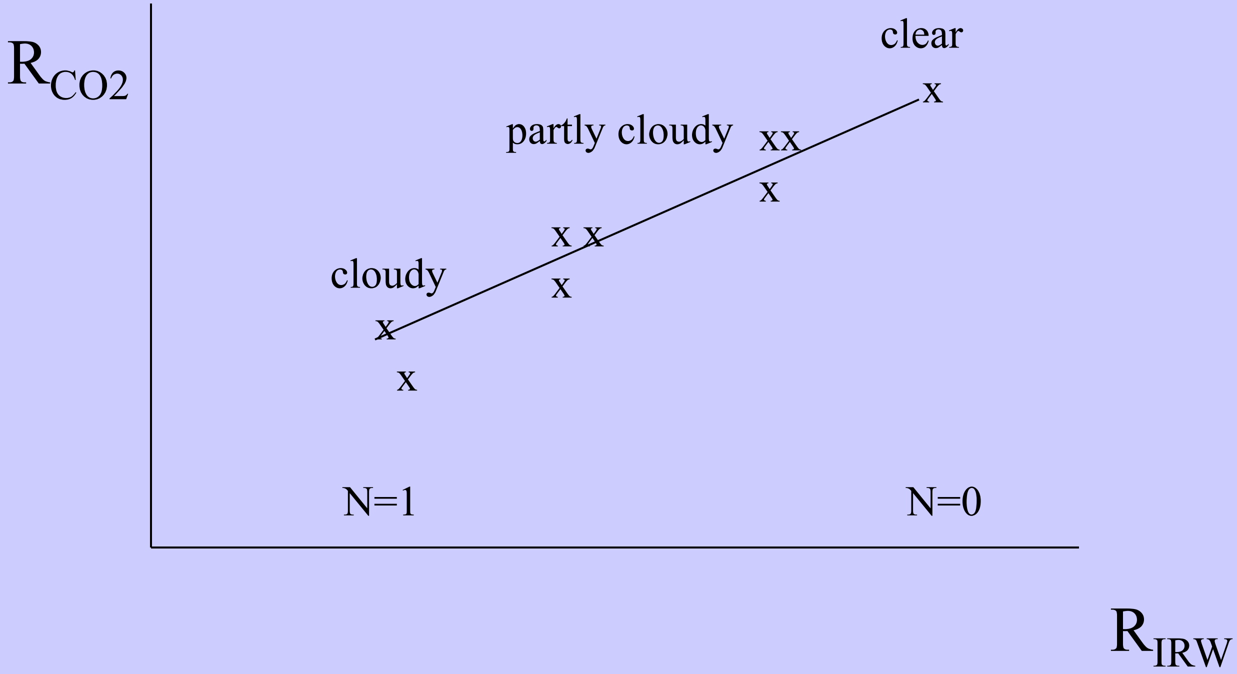
$$(I_\lambda - I_\lambda^{clr}) = \eta\varepsilon_\lambda \int_{p_s}^{p_c} \tau_\lambda \, dB_\lambda \, .$$

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.

Cloud Clearing

For a single layer of clouds, radiances in one spectral band vary linearly with those of another as cloud amount varies from one field of view (fov) to another



Clear radiances can be inferred by extrapolating to cloud free conditions.

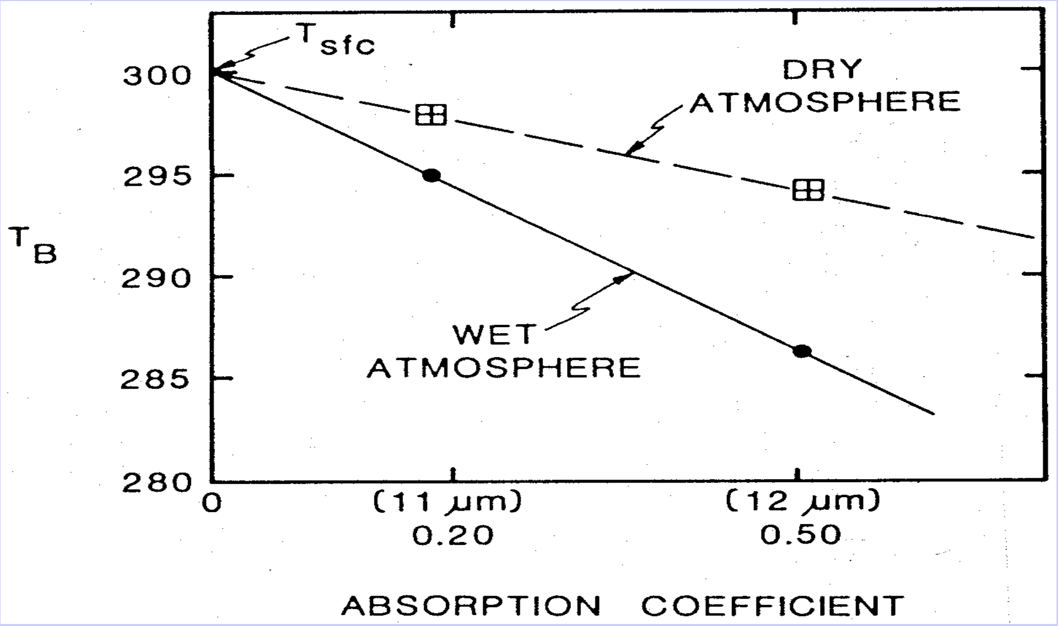
Moisture

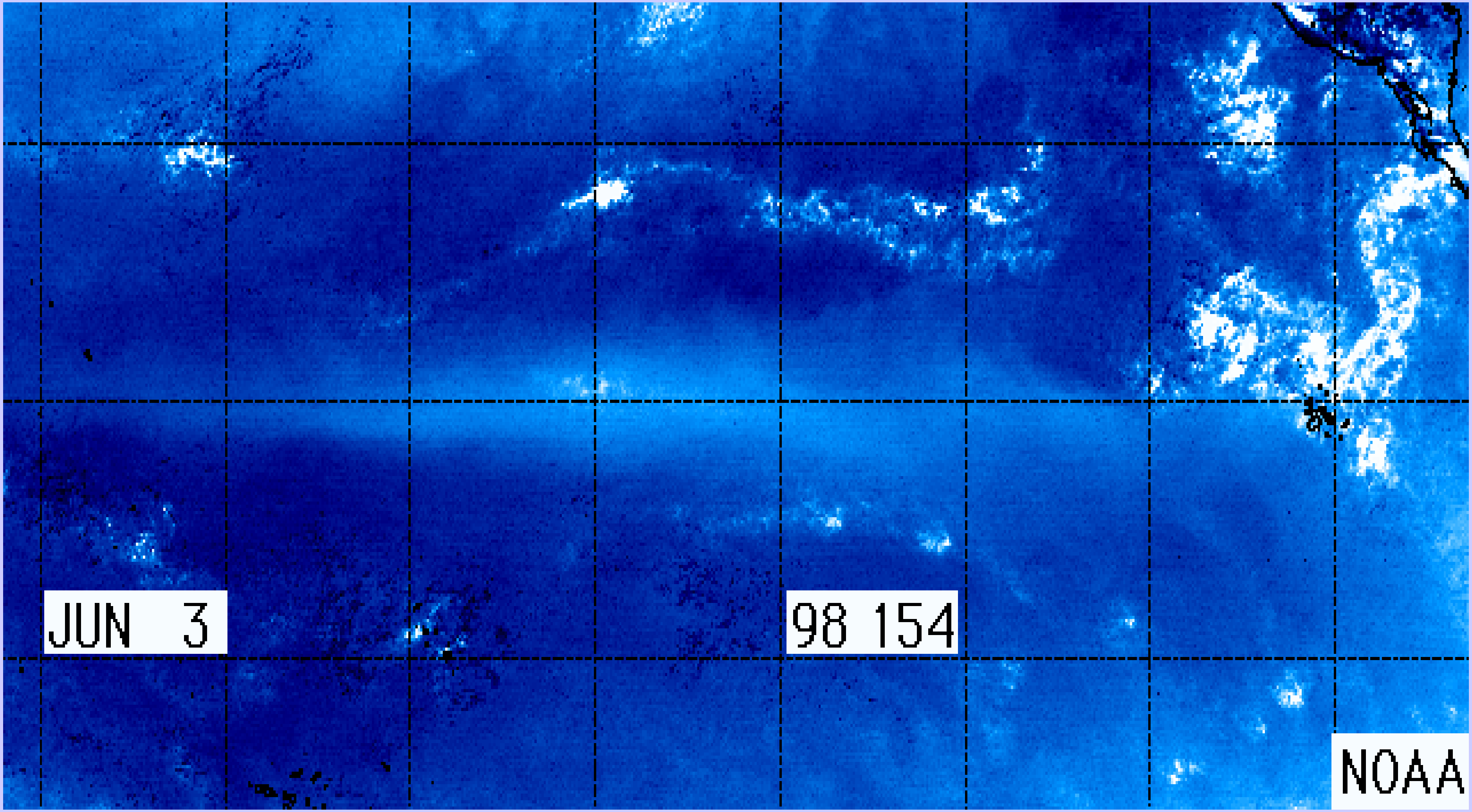
Moisture attenuation in atmospheric windows varies linearly with optical depth.

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

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

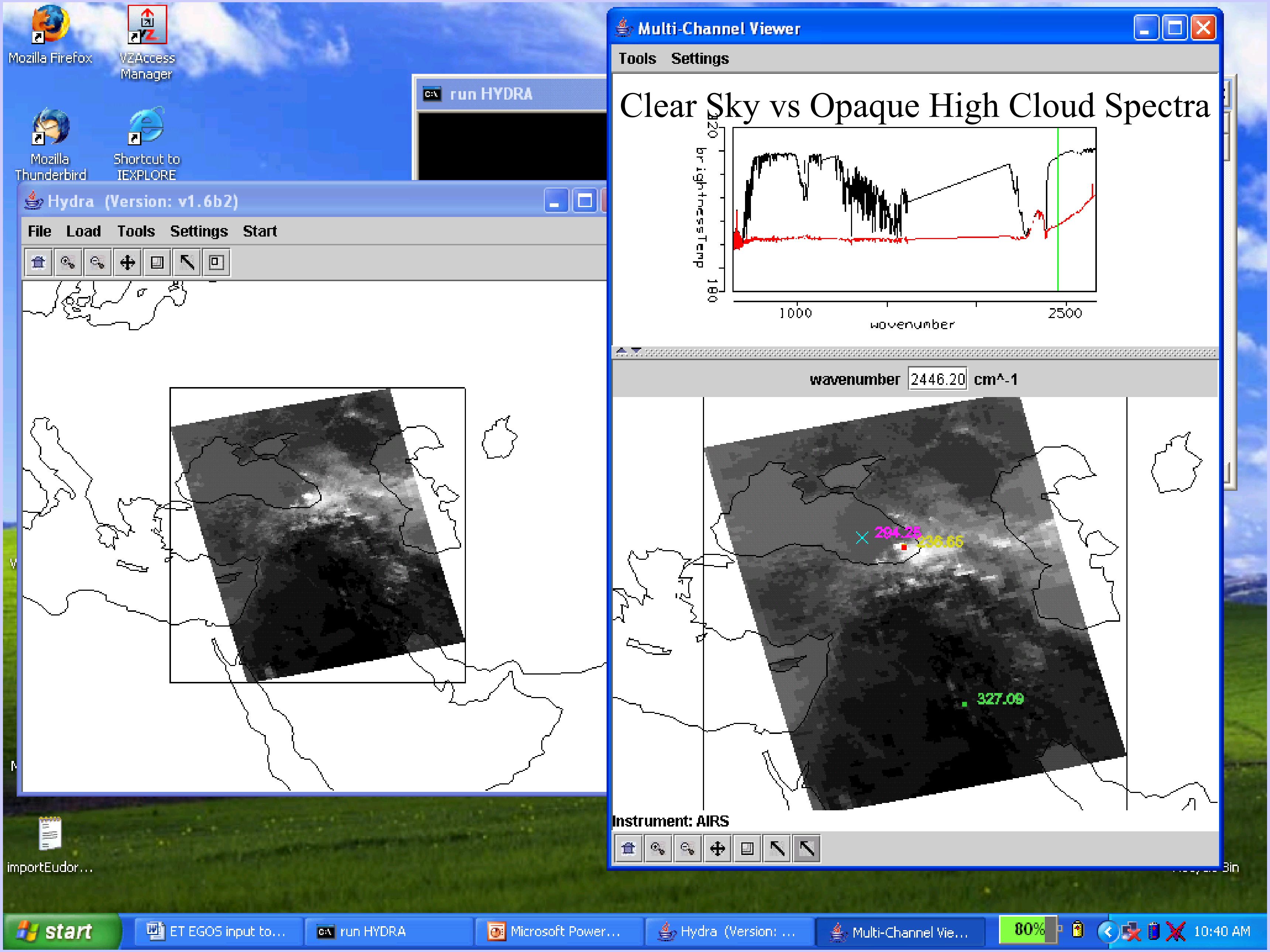
Moisture content of atmosphere inferred from slope of linear relation.

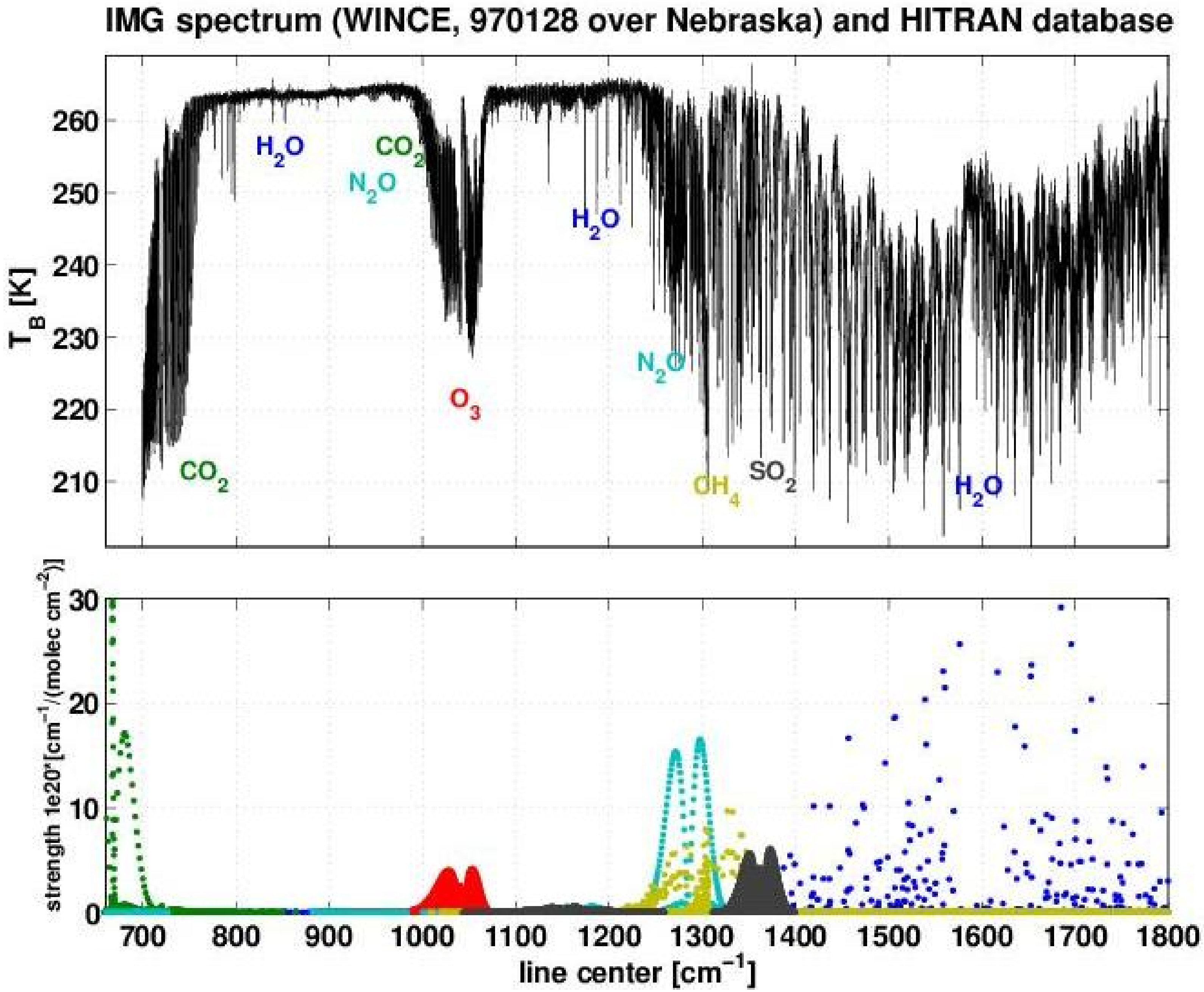




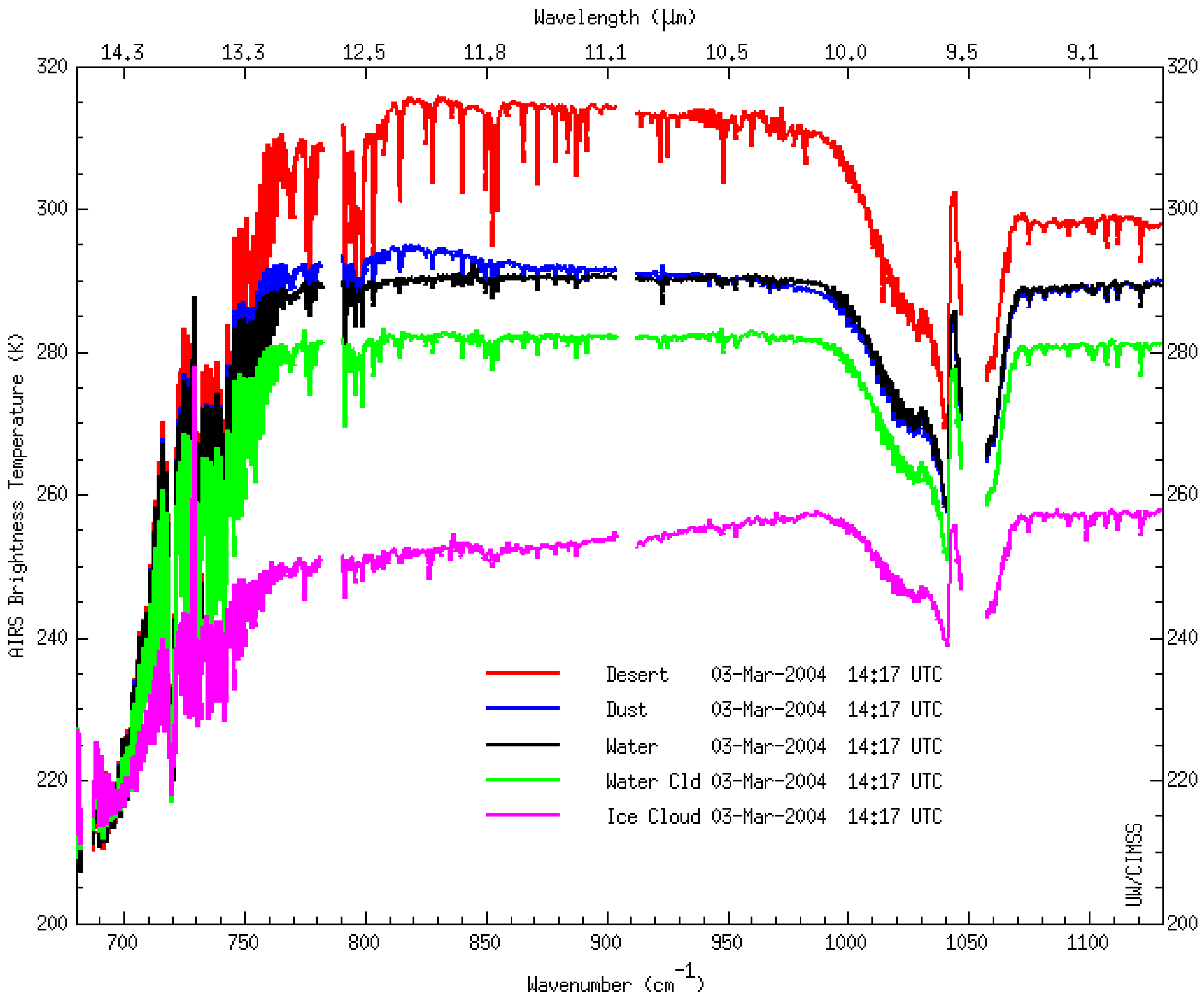
SST Waves from Legeckis

AIRS data from 28 Aug 2005





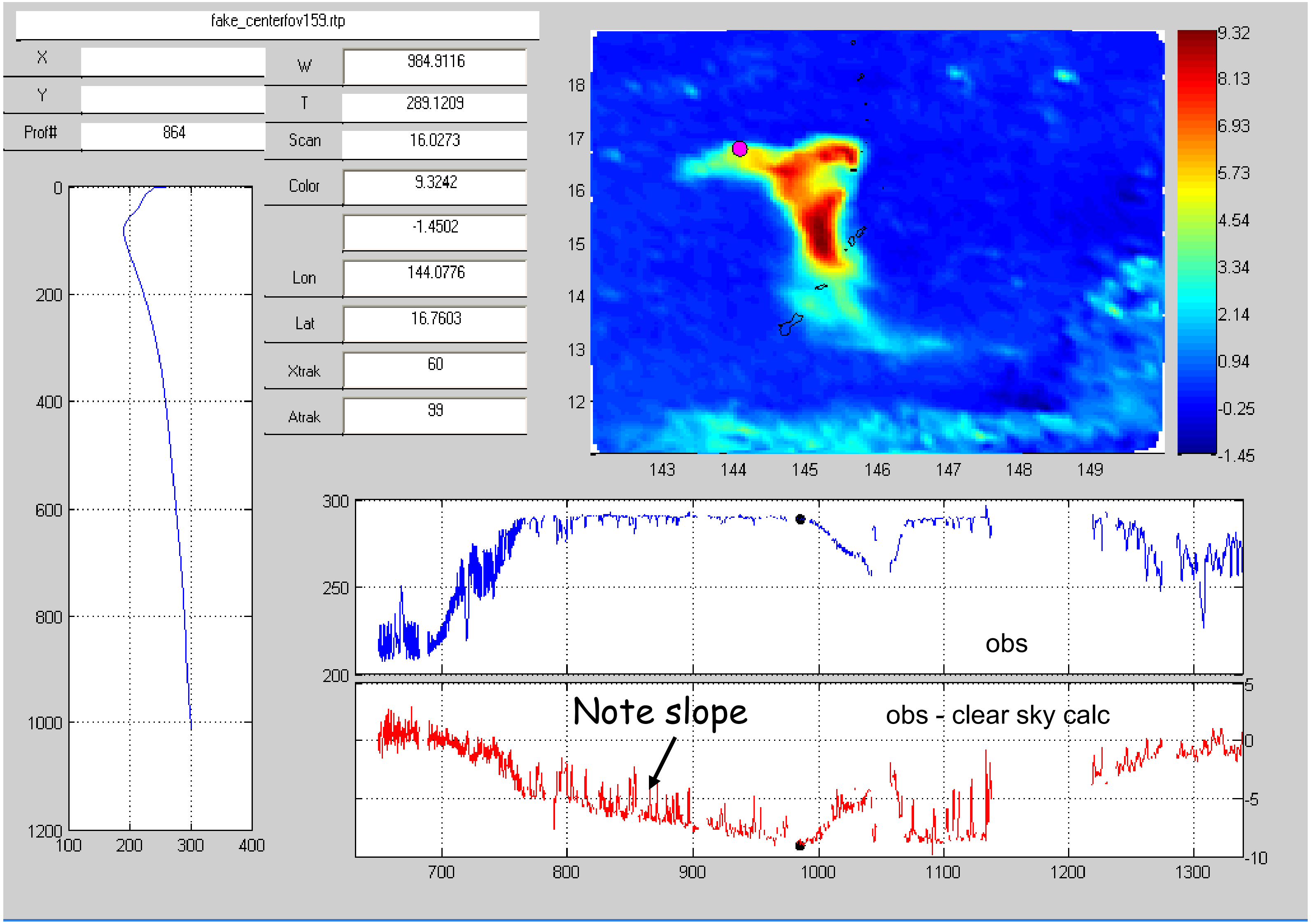
IMG spectrum (WINCE, 970128 over Nebraska) and HITRAN database
T_B [K]
260
250
240
230
220
210
H_2O
CO_2
N_2O
H_2O
N_2O
O_3
CO_2
CH_4
SO_2
H_2O
strength 1e20*[cm^{-1}/(molec cm^{-2})]
30
20
10
0
700
800
900
1000
1100
1200
1300
1400
1500
1600
1700
1800
line center [cm^{-1}]



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)
700
750
800
850
900
950
1000
1050
1100
Wavenumber (cm-1)
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

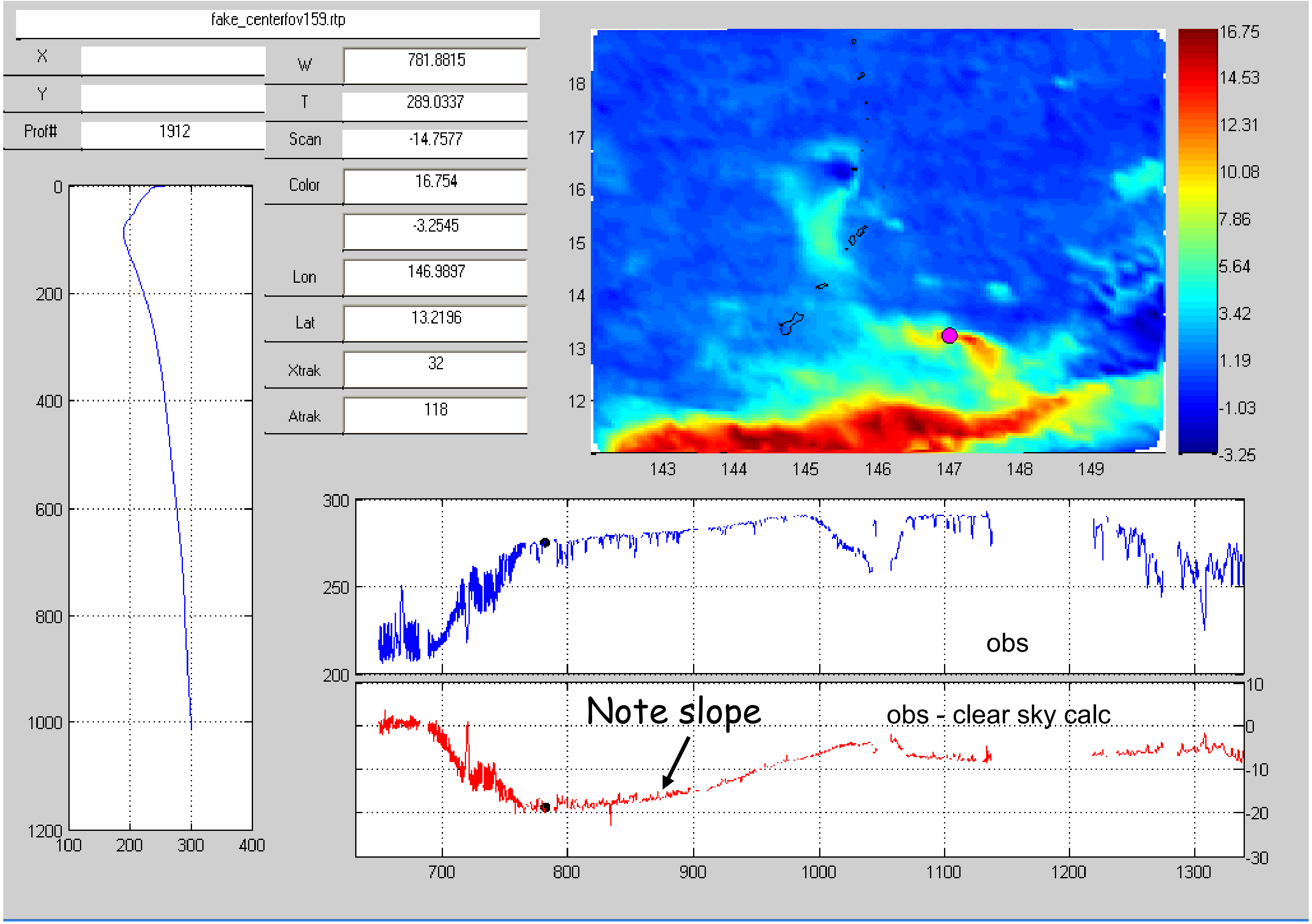
Silicate (ash cloud) signal at Anatahan, Mariana Is

Image is ECMWF bias difference of 1227 cm^{-1} – 984 cm^{-1} (double difference)

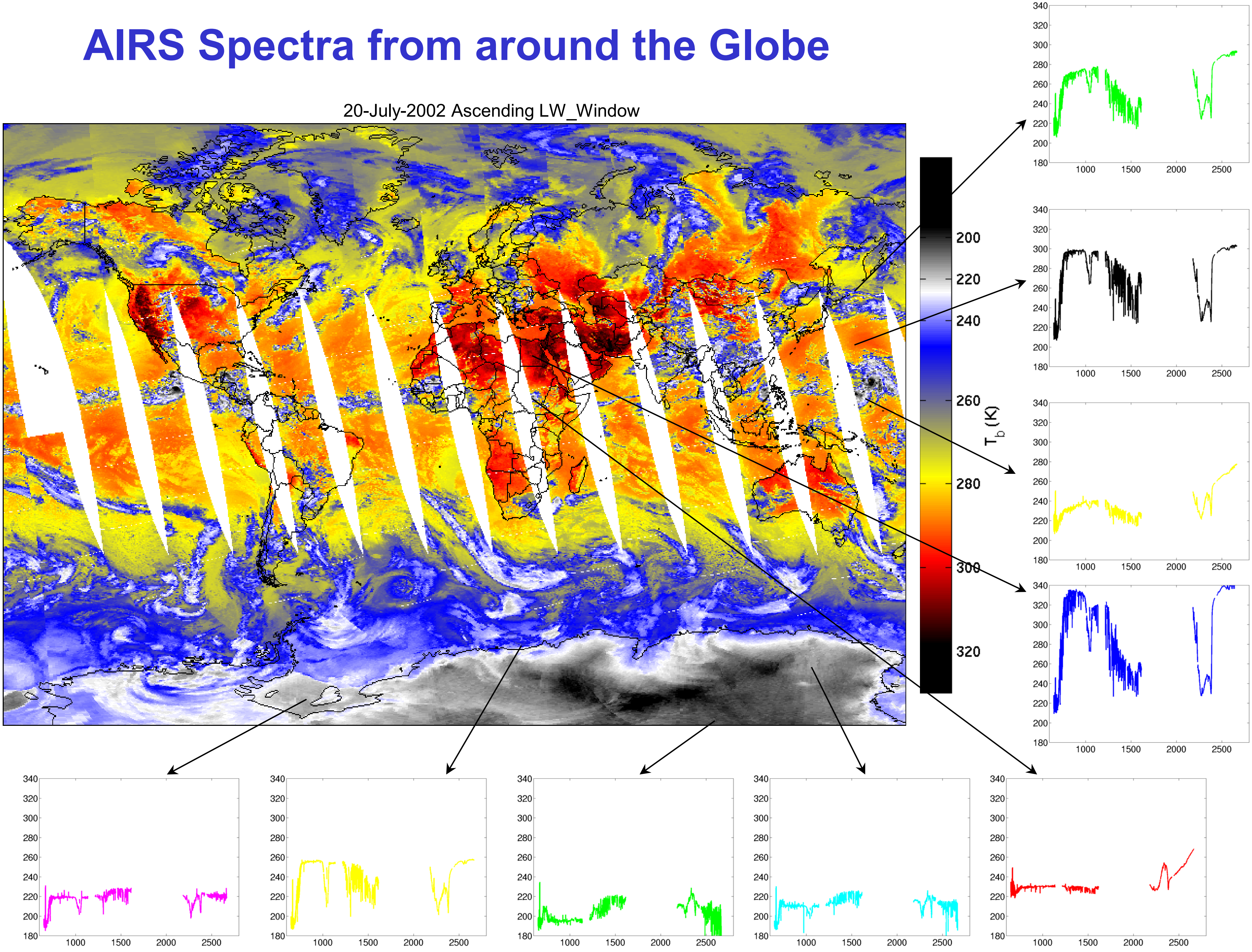


Cirrus signal at Anatahan

Image is ECMWF T_b bias difference of 1227 cm^{-1} – 781 cm^{-1} (double difference)



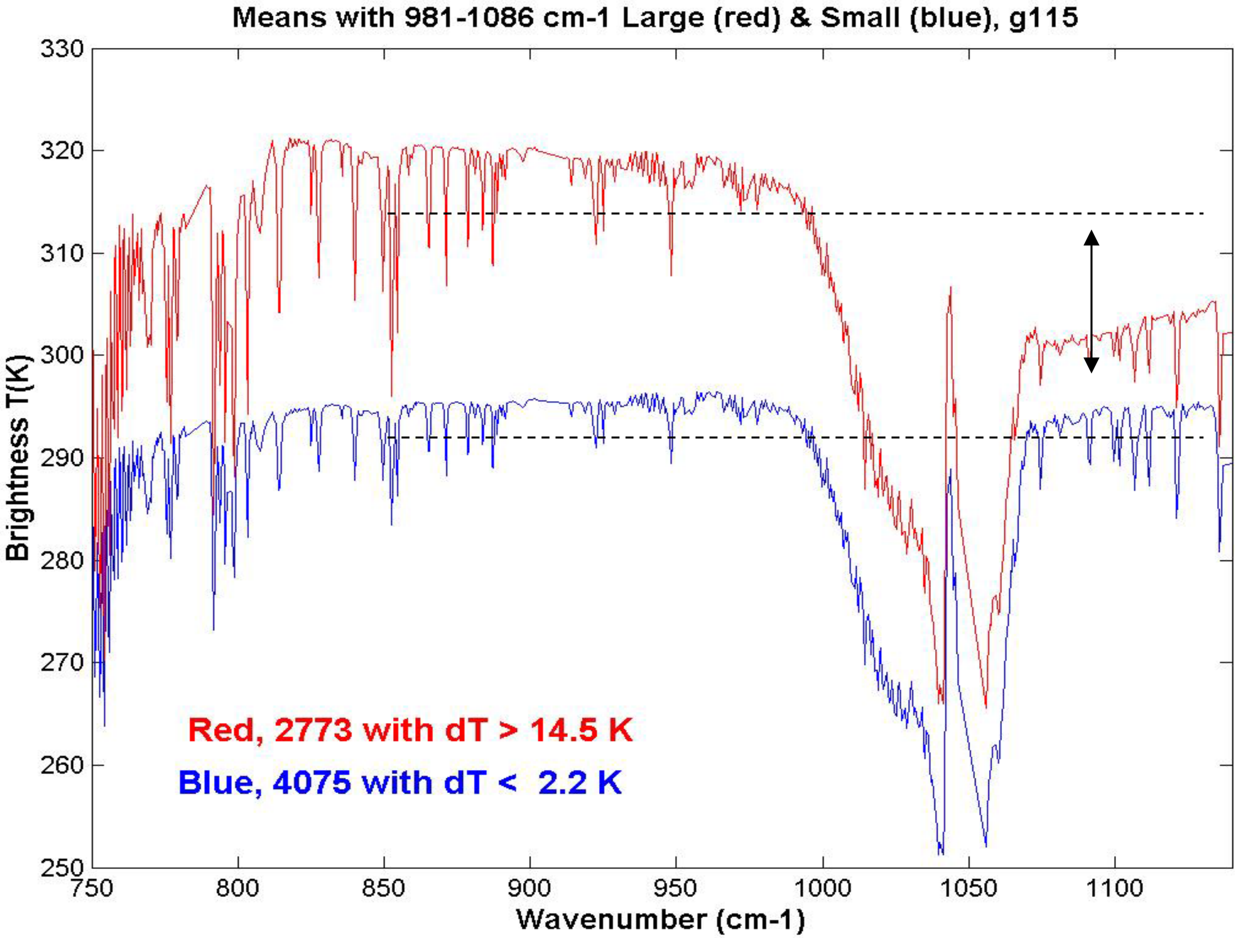
AIRS Spectra from around the Globe



Inferring surface properties with AIRS high spectral resolution data

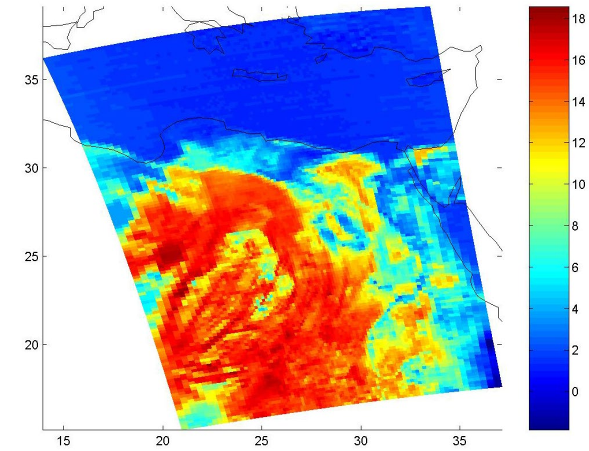
Barren region detection if T1086 < T981

Barren vs Water/Vegetated

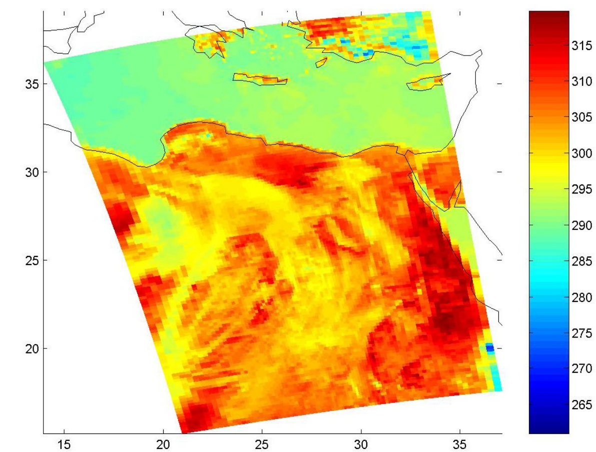


AIRS data from 14 June 2002

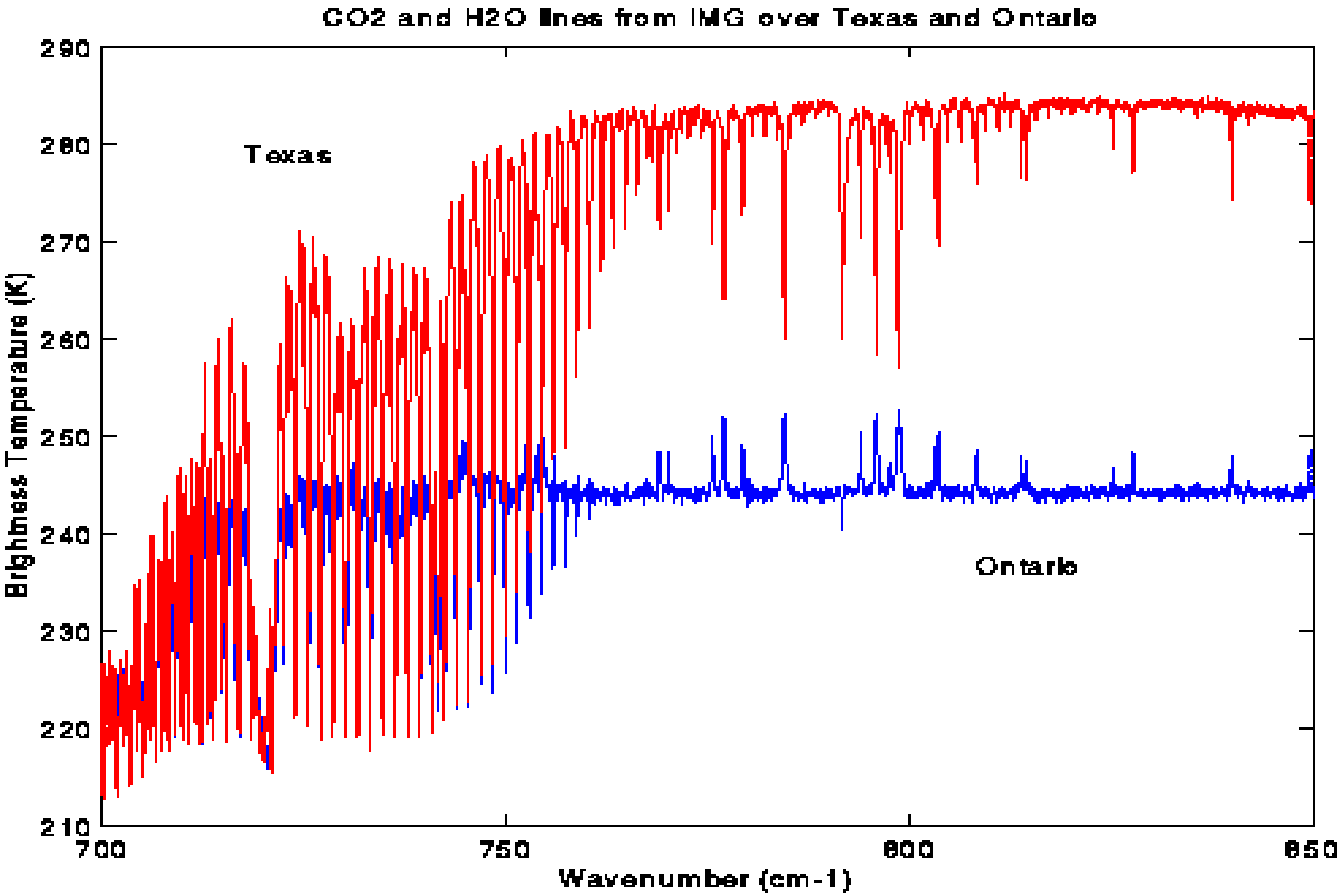
$T(981\ cm^{-1})-T(1086\ cm^{-1})$



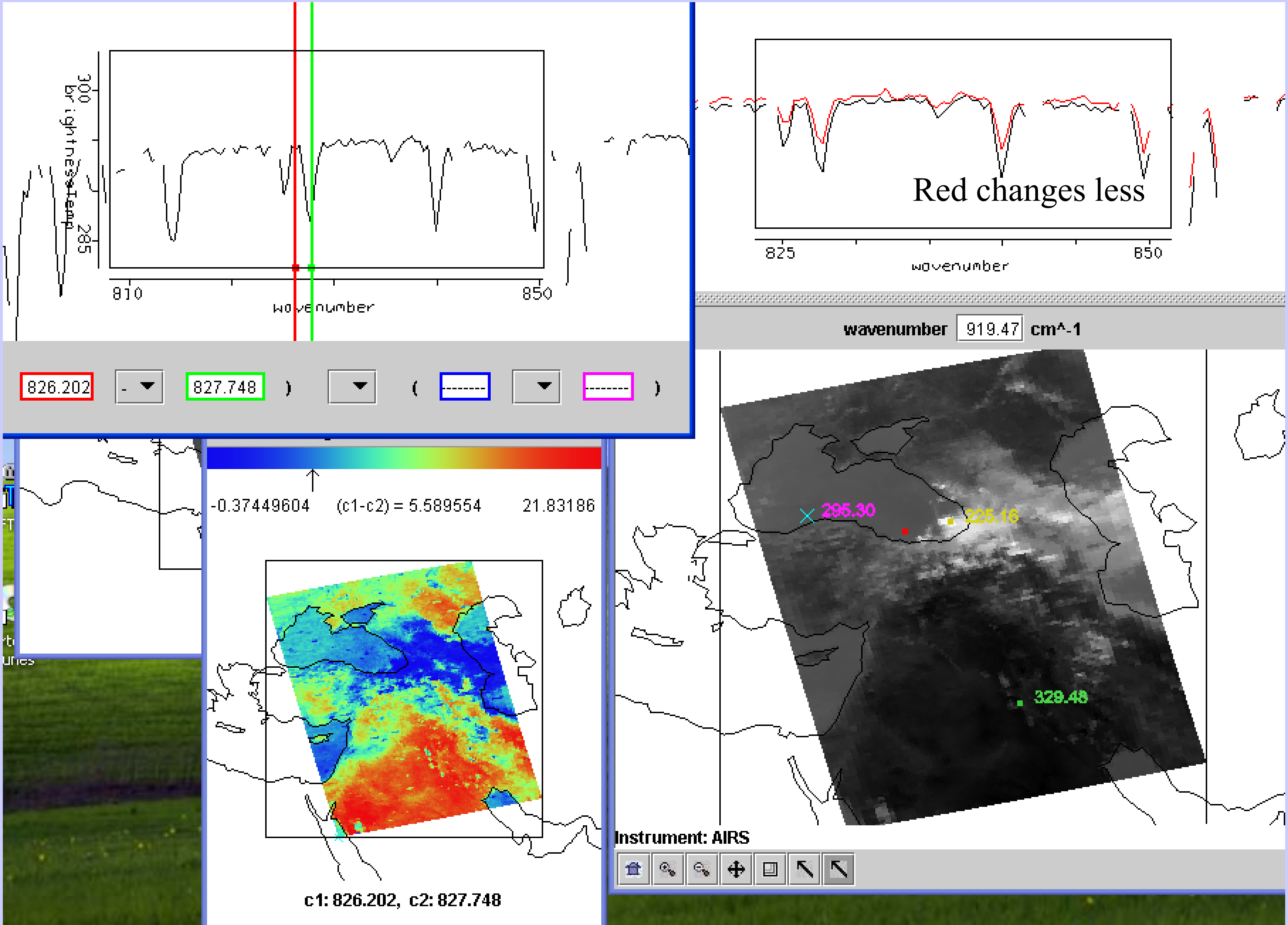
$T(1086\ cm^{-1})$



Sensitivity of High Spectral Resolution to Boundary Layer Inversions and Surface/atmospheric Temperature differences
(from IMG Data, October, December 1996)

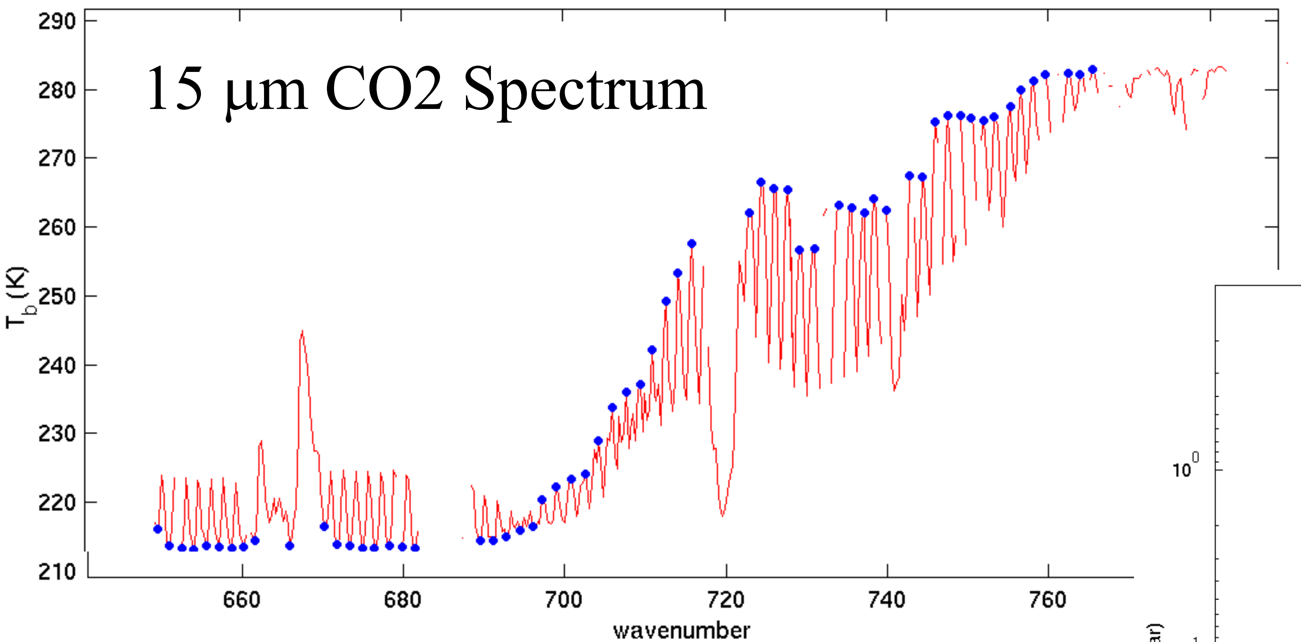


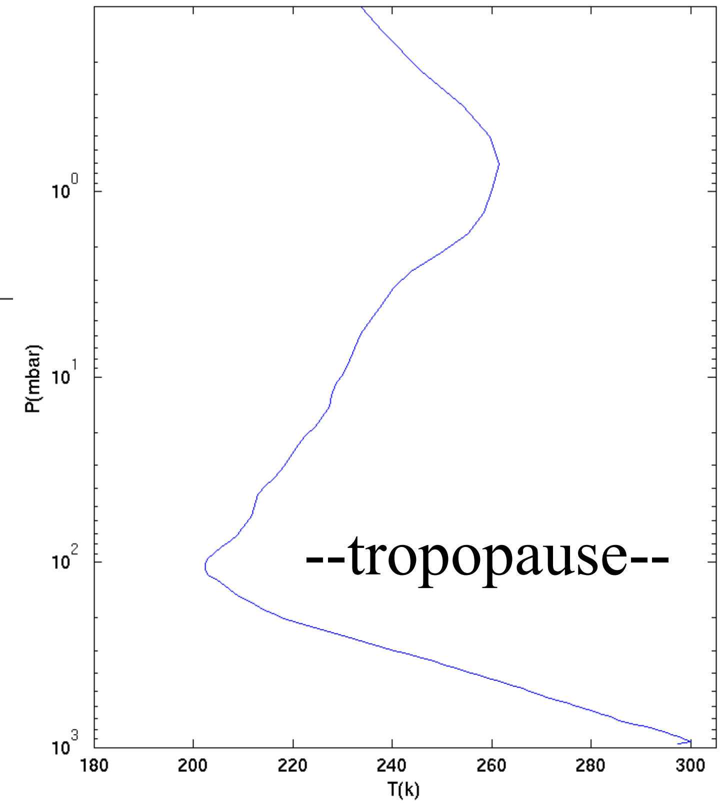
Offline-Online in LW IRW showing low level moisture



Twisted Ribbon formed by CO_2 spectrum:

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



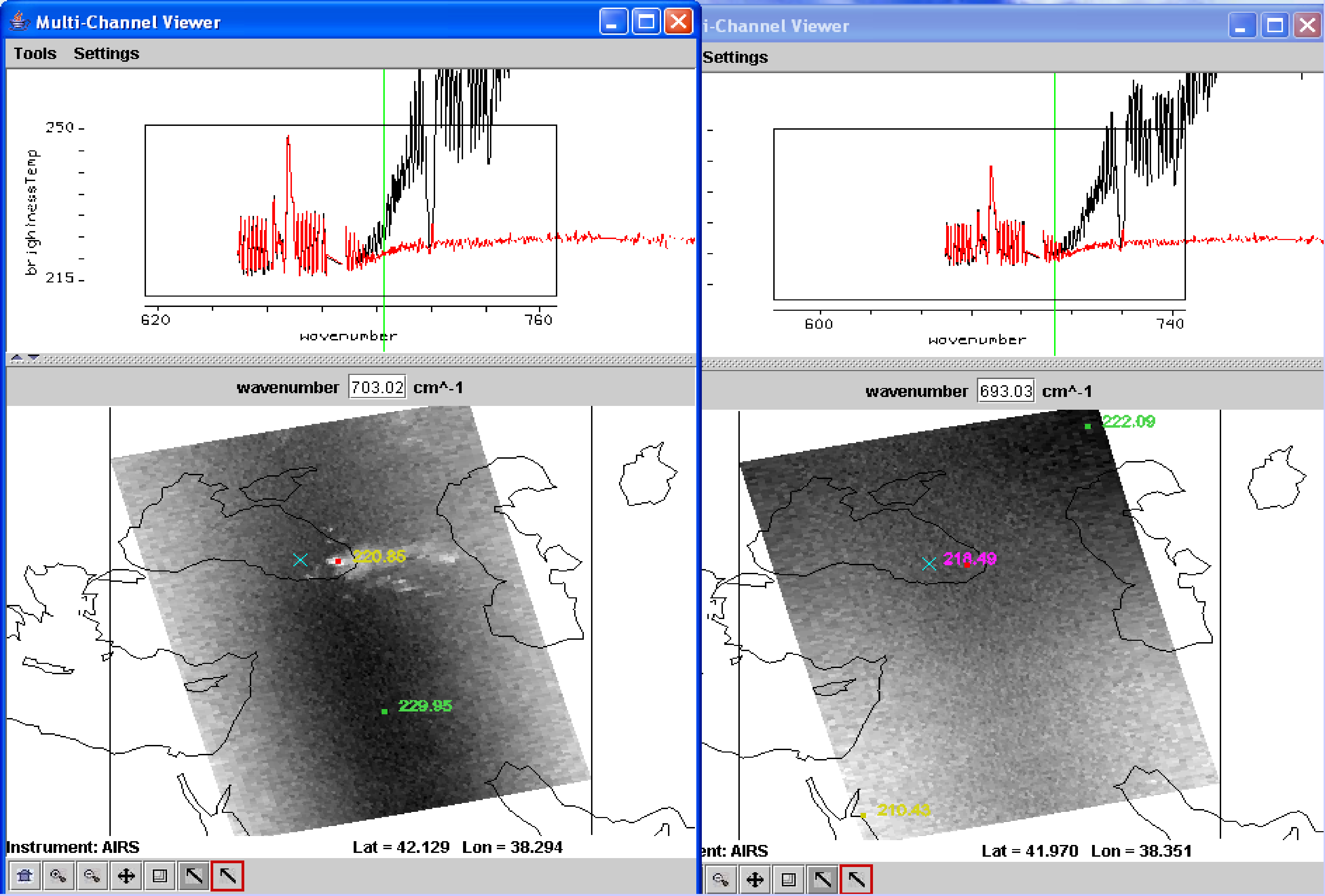


Blue between-line T_b

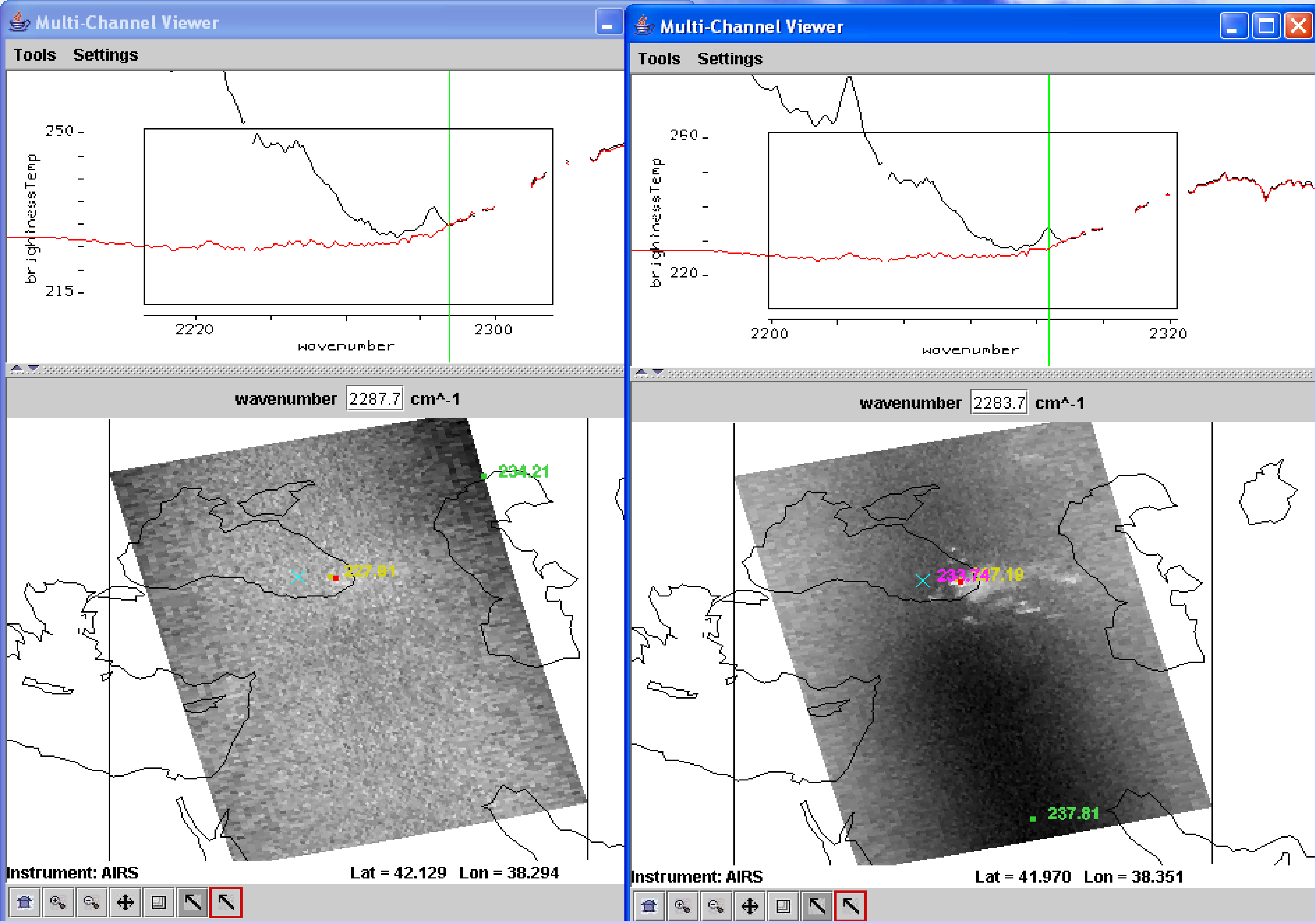
<u>warmer</u> for tropospheric channels,

<u>colder</u> for stratospheric channels

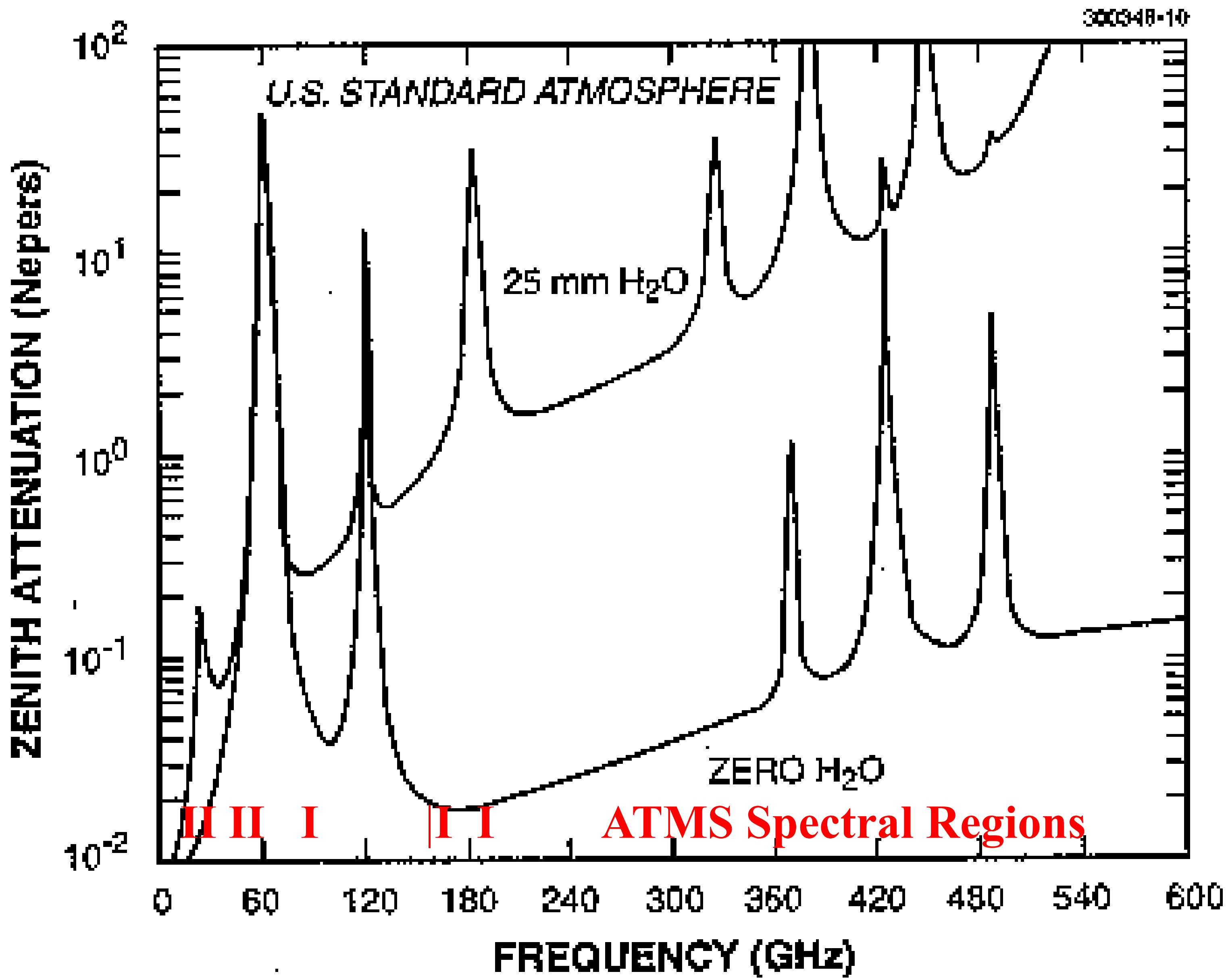
Signature not available at low resolution



Cld and clr spectra in CO_2 absorption separate when weighting functions sink to cloud level



Cld and clr spectra in CO2 absorption separate when weighting functions sink to cloud level



300348-10
U.S. STANDARD ATMOSPHERE
25 mm H_2O
ZERO H_2O
II II I
|I I
ATMS Spectral Regions
ZENITH ATTENUATION (Nepers)
10^2
10^1
10^0
10^{-1}
10^{-2}
0
60
120
180
240
300
360
420
480
540
600
FREQUENCY (GHz)

Radiation is governed by Planck's Law

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

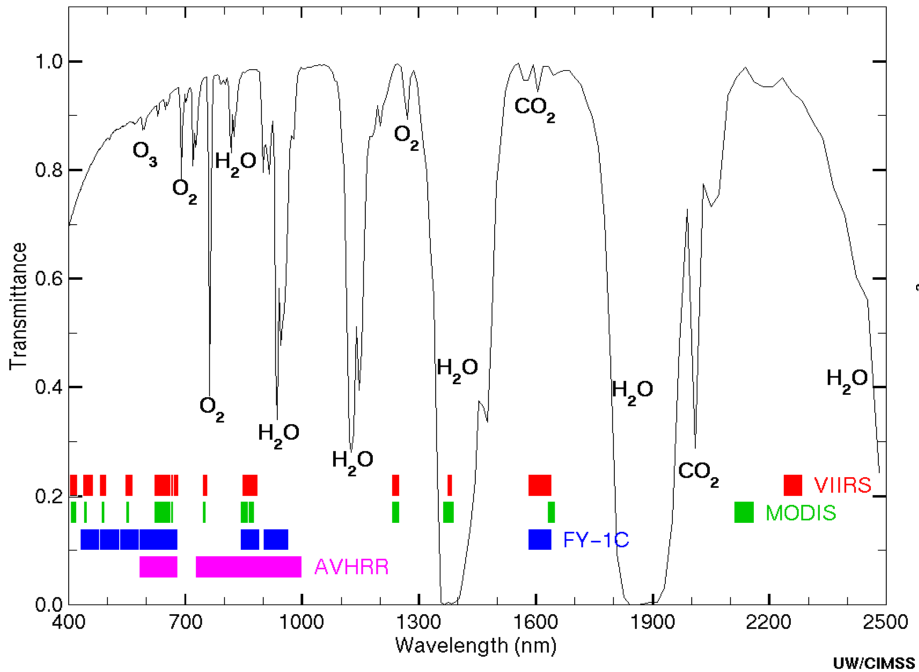
In microwave region $c_2/\lambda T \ll 1$ so that

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

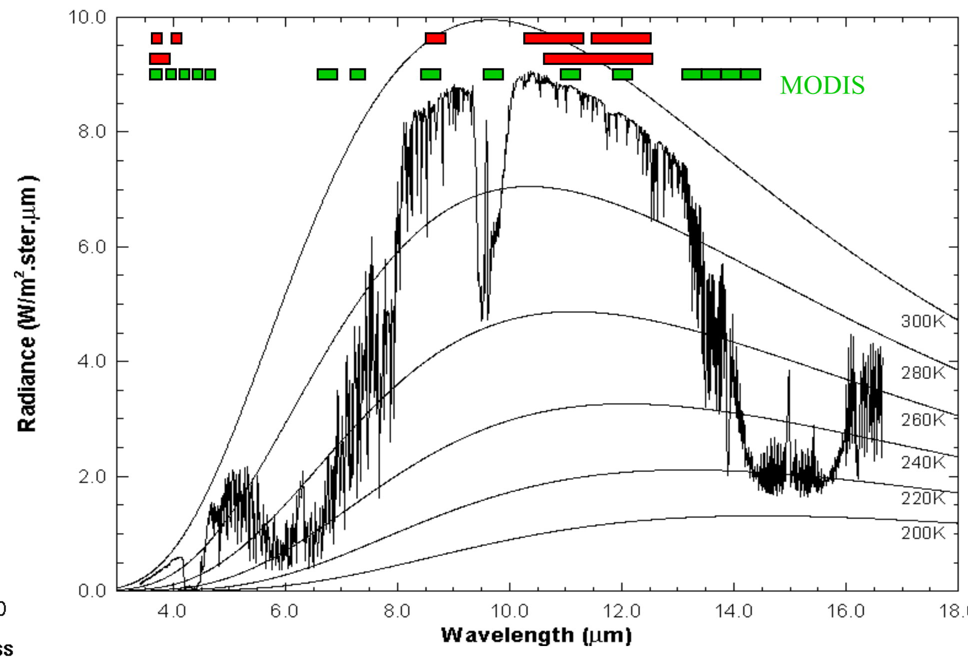
And classical Rayleigh Jeans radiation equation emerges

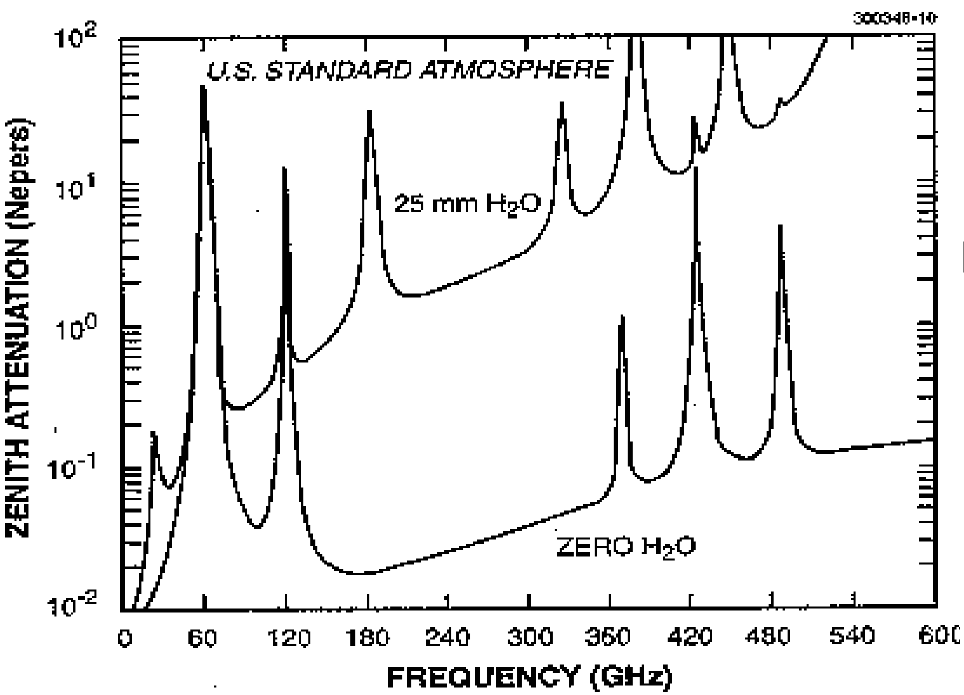
$$B_\lambda(T) \approx [c_1 / c_2] [T / \lambda^4]$$

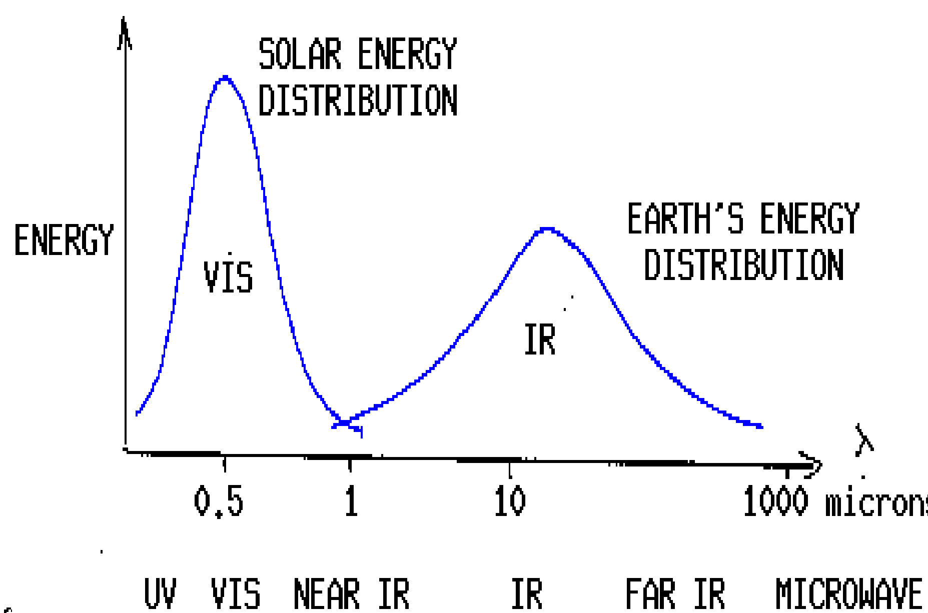
Radiance is linear function of brightness temperature.



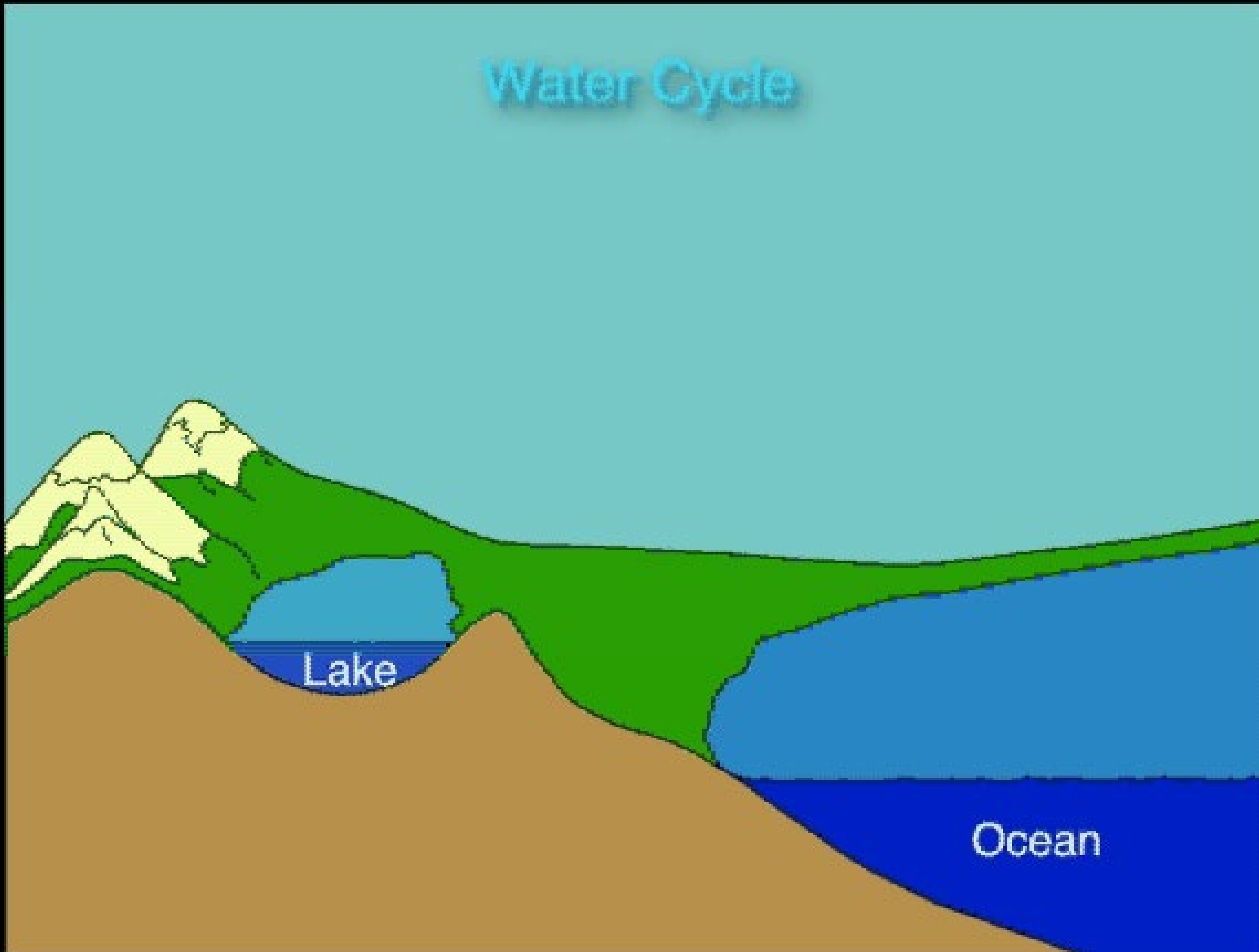
High resolution atmospheric absorption spectrum
and comparative blackbody curves.





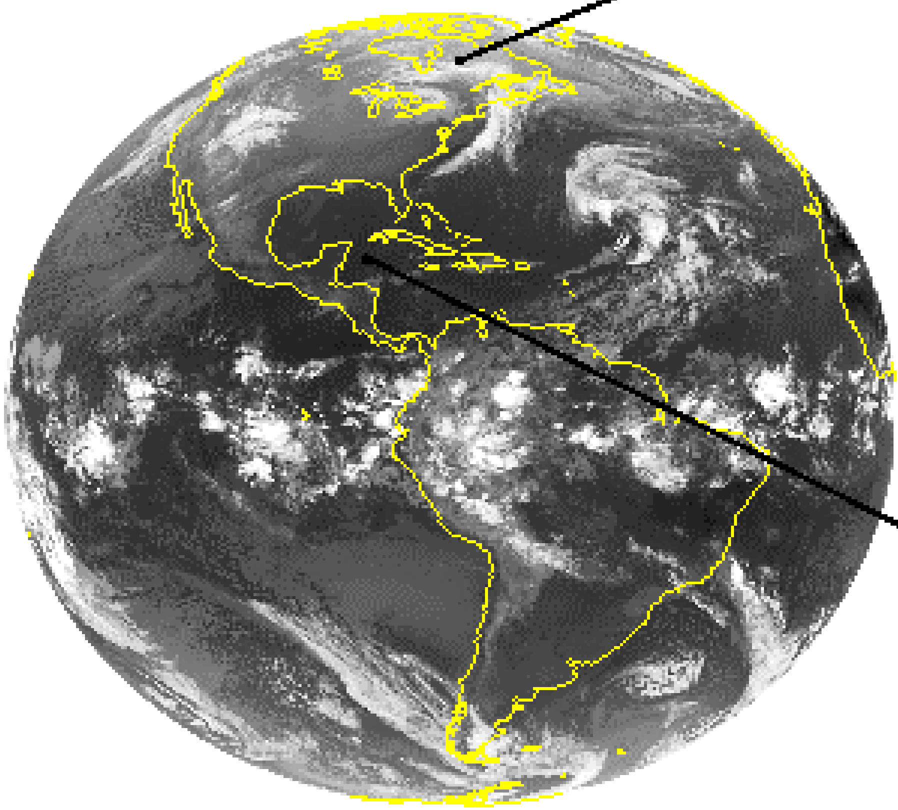


Energy Cycle From Levizzani

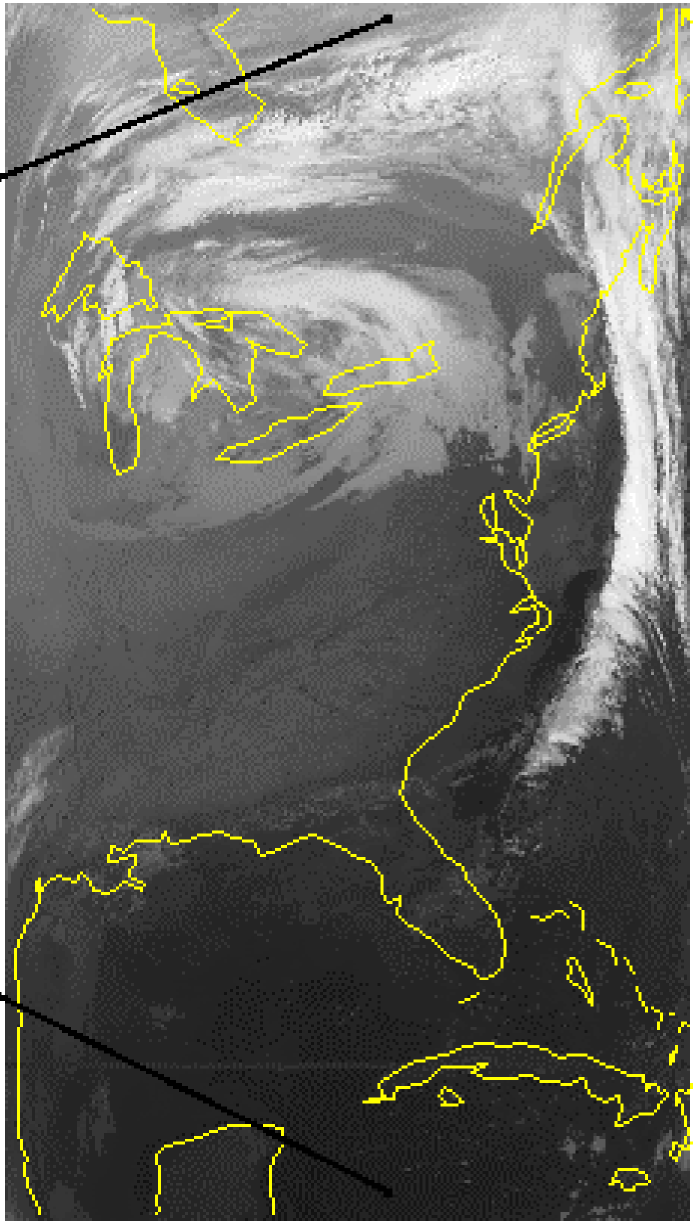


Water Cycle From Levizzani

GEO vs LEO



GOES-8 IMAGER 12UTC 02APR98



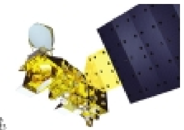
NOAA-12 AVHRR 12UTC 02APR98

Comparison of geostationary (geo) and low earth orbiting (leo) satellite capabilities

Geo	Leo
observes process itself (motion and targets of opportunity)	observes effects of process
repeat coverage in minutes ($\Delta t \leq 30$ minutes)	repeat coverage twice daily ($\Delta t = 12$ hours)
full earth disk only	global coverage
best viewing of tropics	best viewing of poles
same viewing angle	varying viewing angle
differing solar illumination	same solar illumination
visible, IR imager (1, 4 km resolution)	visible, IR imager (1, 1 km resolution)
one visible band	multispectral in visible (veggie index)
IR only sounder (8 km resolution)	IR and microwave sounder (17, 50 km resolution)
filter radiometer	filter radiometer, interferometer, and grating spectrometer
diffraction more than leo	diffraction less than geo

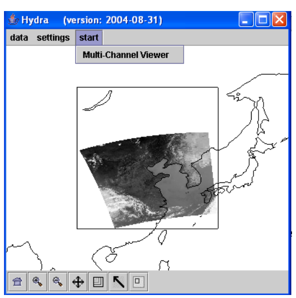
HYperspectral viewer for Development of Research Applications - HYDRA

MSG,
GOES



MODIS,
AIRS

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



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With guidance from
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Liam Gumley
Paul Menzel

C I M S S

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

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

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

For MODIS amd AIRS data orders
http://daac.gsfc.nasa.gov/