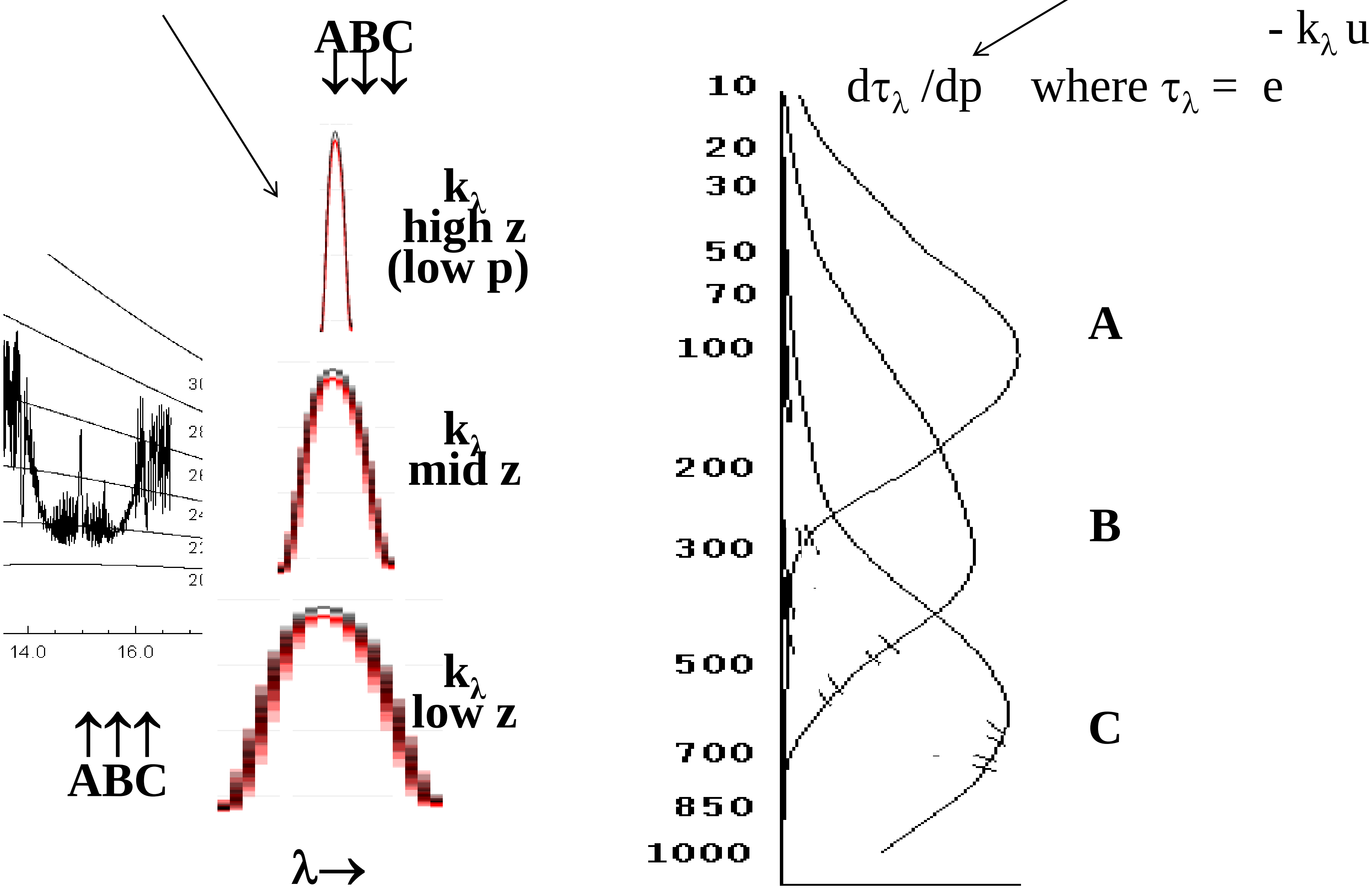
Intro to High Spectral Resolution IR Measurements

Lectures in Madison

27 March 2013

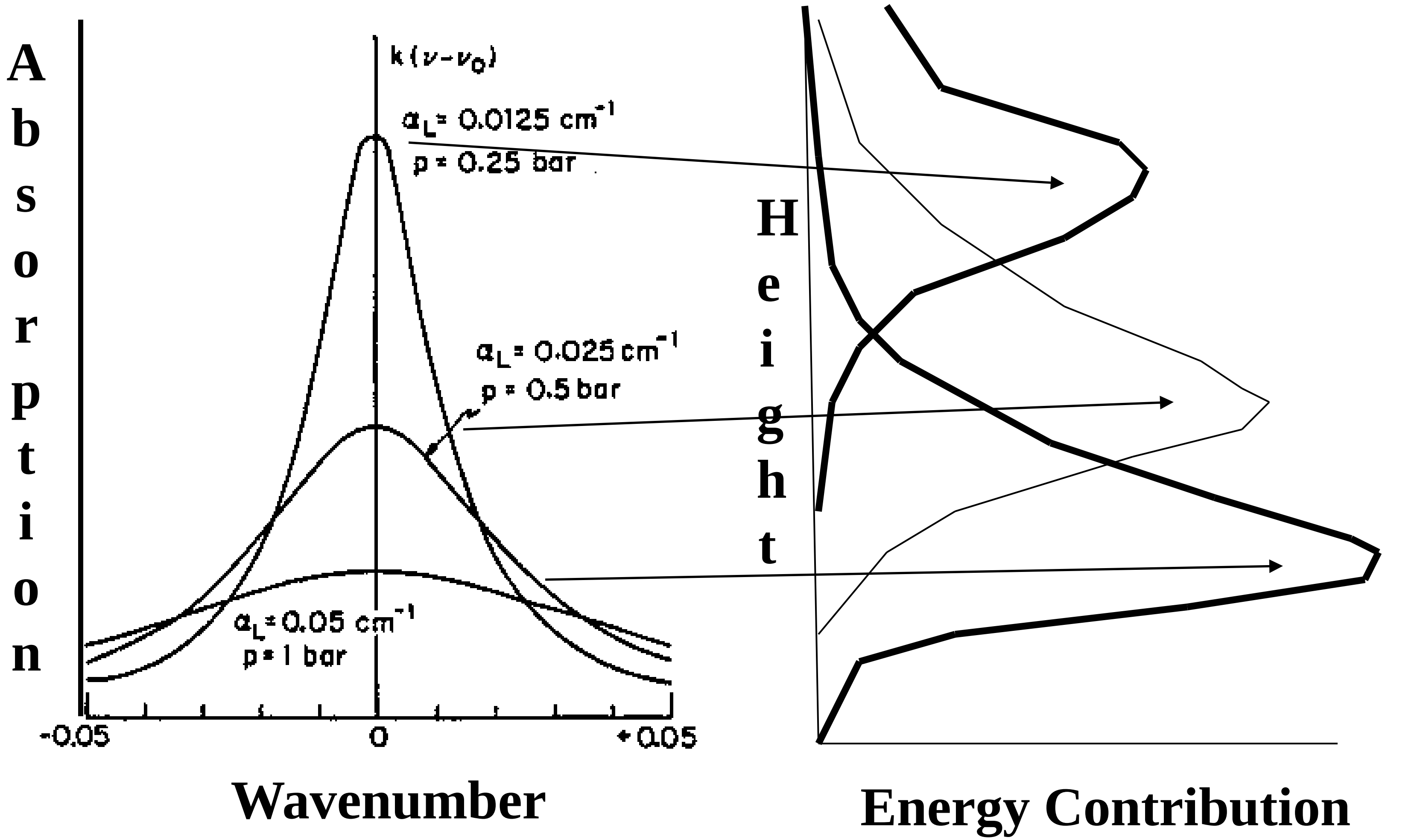
Paul Menzel
UW/CIMSS/AOS

line broadening with pressure helps to explain weighting functions

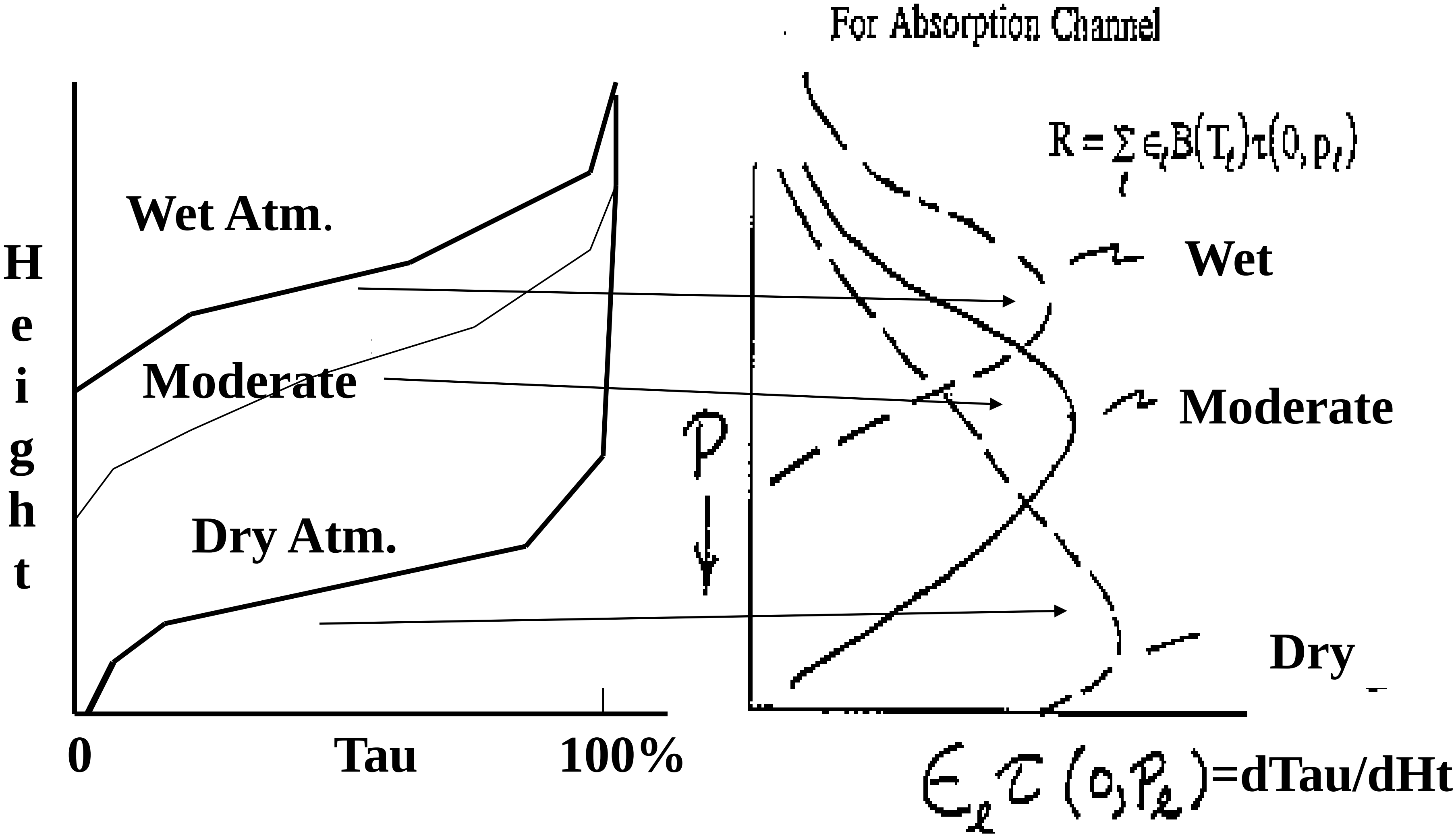


line broadening with pressure helps to explain weighting functions

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

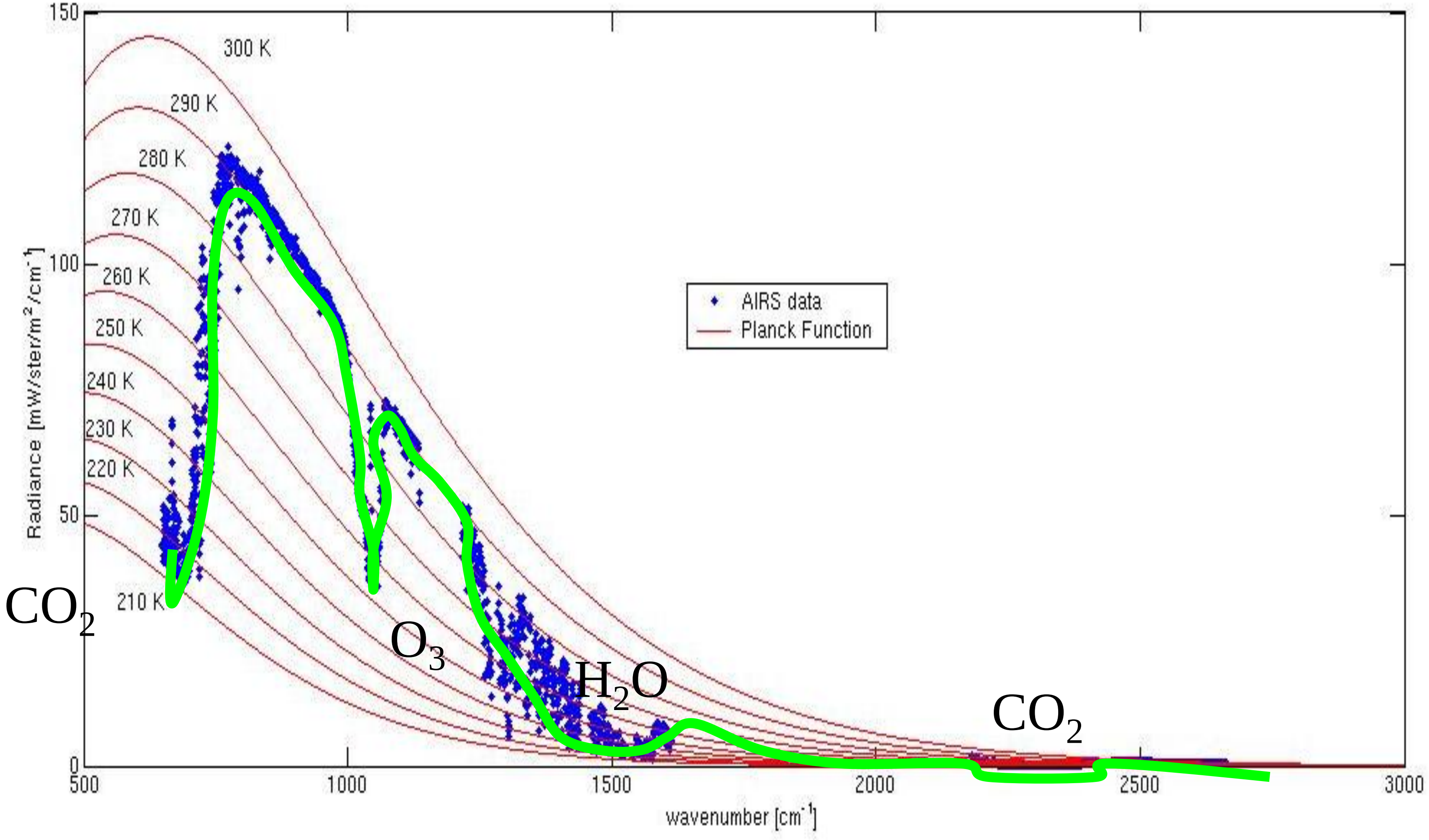


For a given water vapor spectral channel the weighting function depends on the amount of water vapor in the atmospheric column



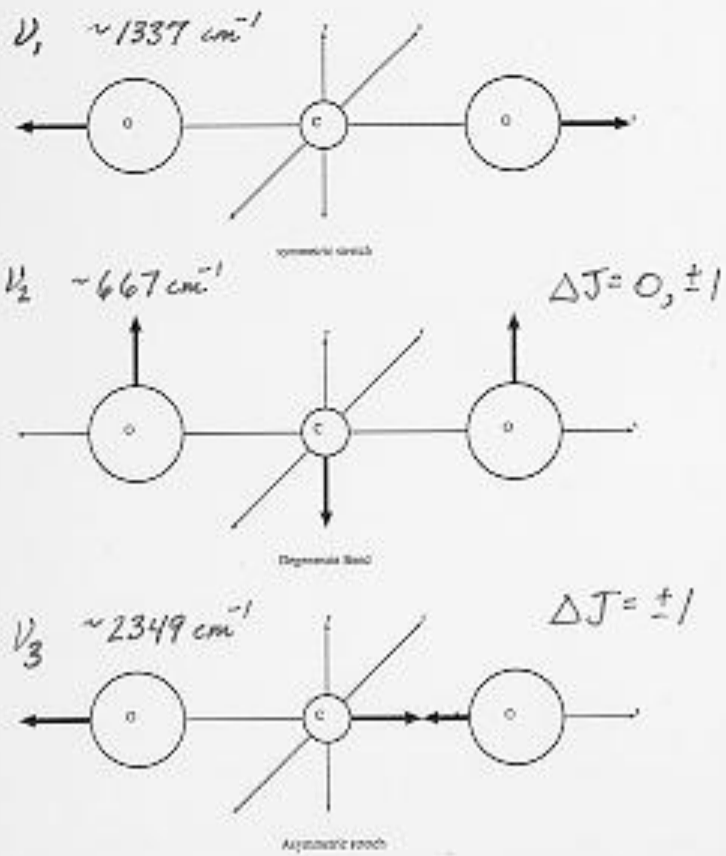
CO2 is about the same everywhere, the weighting function for a given CO2 spectral channel is the same everywhere

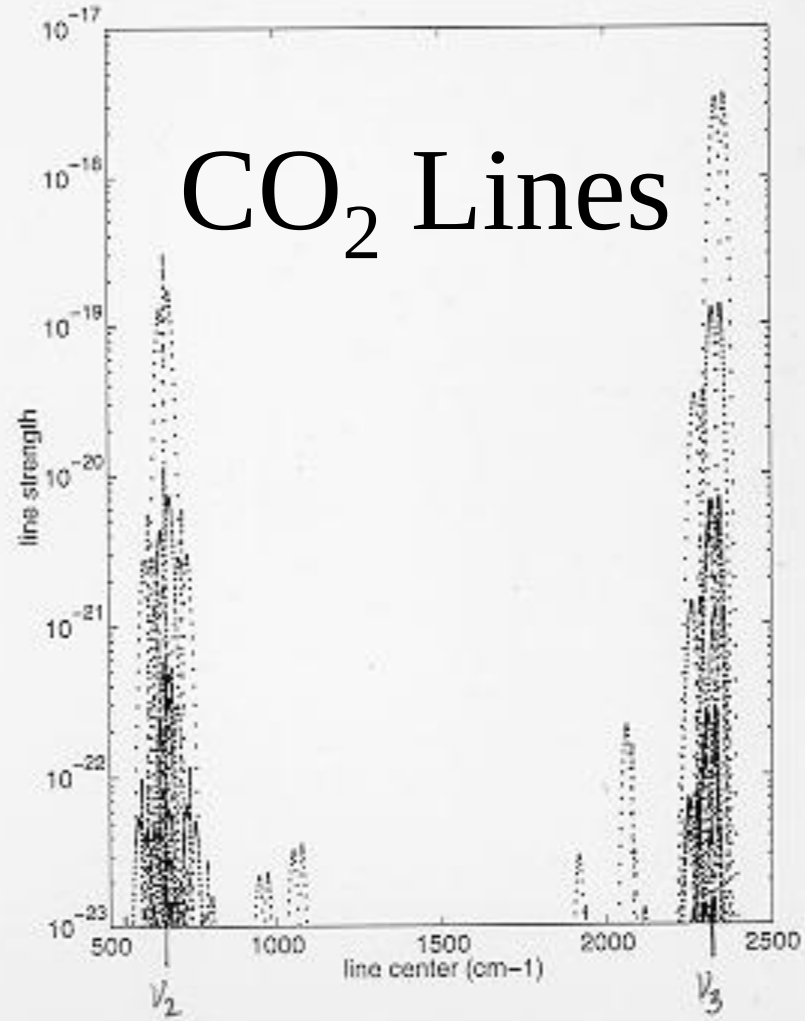
Vibrational Bands



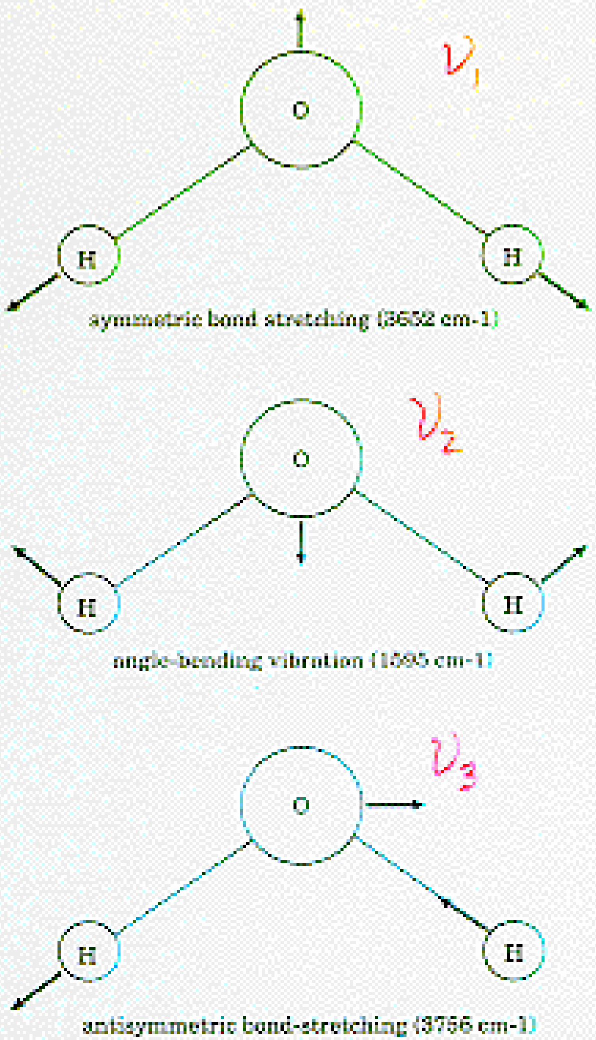
CO_2 Vibration – Rotation Spectra

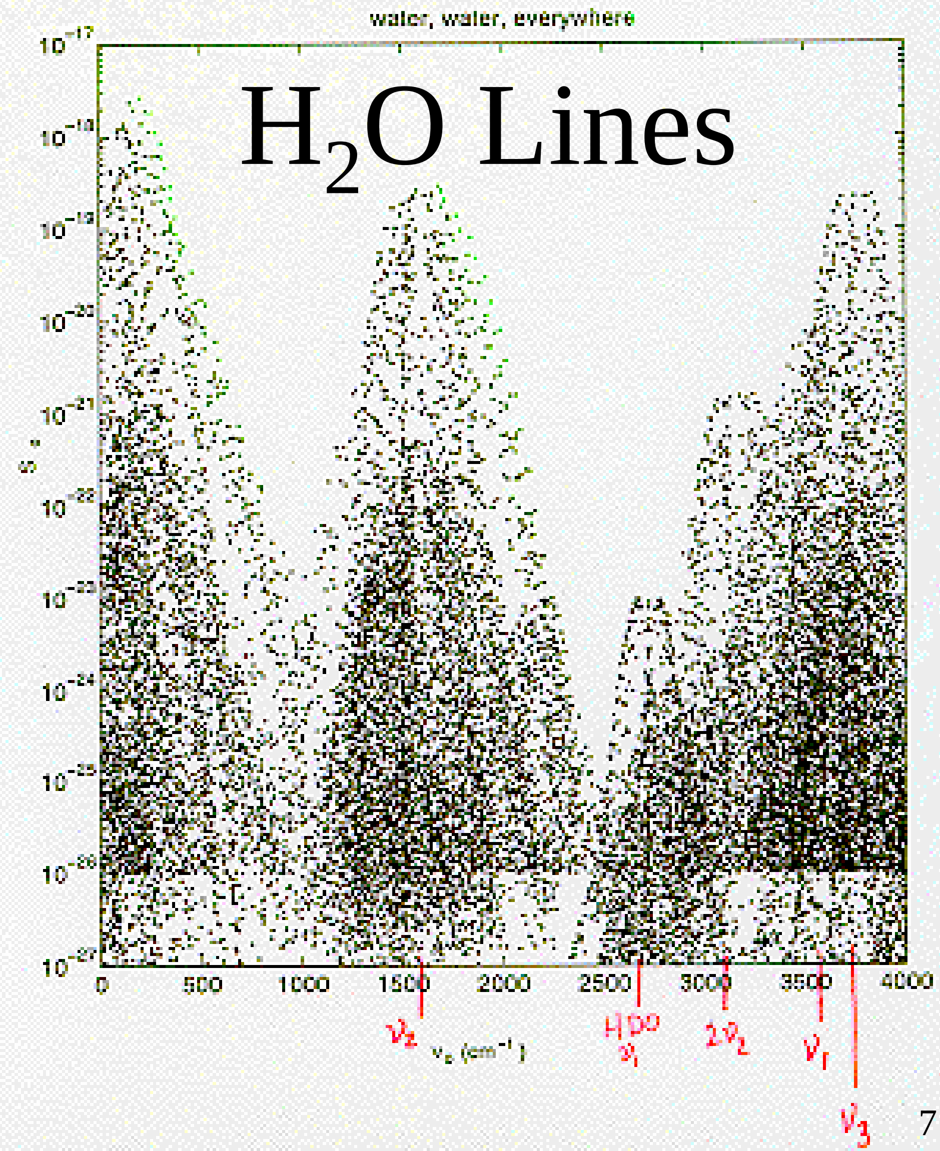
$$E(\upsilon, J) = \underbrace{h\nu(\upsilon + \frac{1}{2}) - xh\nu(\upsilon + \frac{1}{2})^2 + \ldots}_{\text{vibration}} + \underbrace{B_\nu[J(J+1) - \ell^2] - D_\nu[J(J+1) - \ell^2]^2 + \ldots}_{\text{rotation}}$$

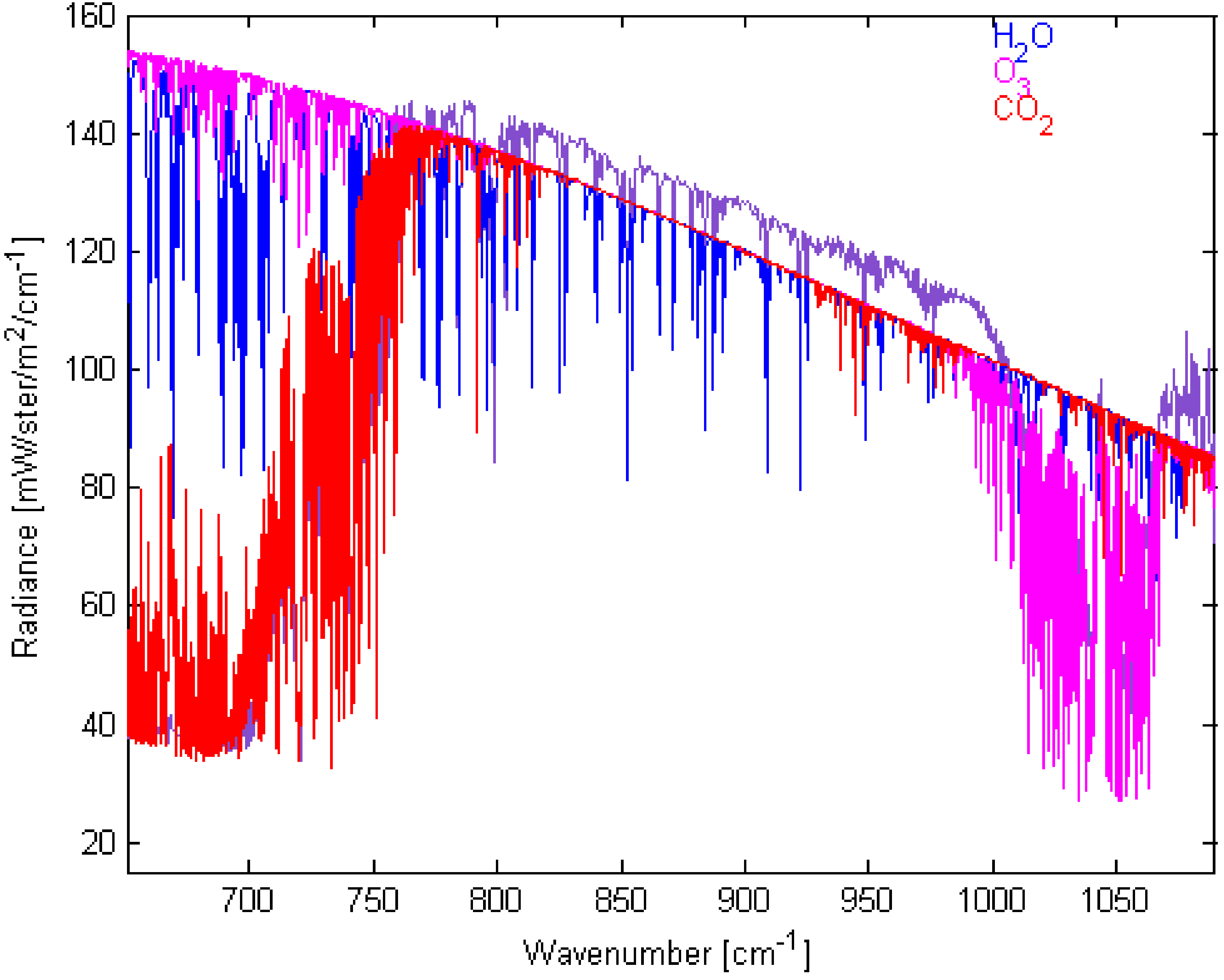




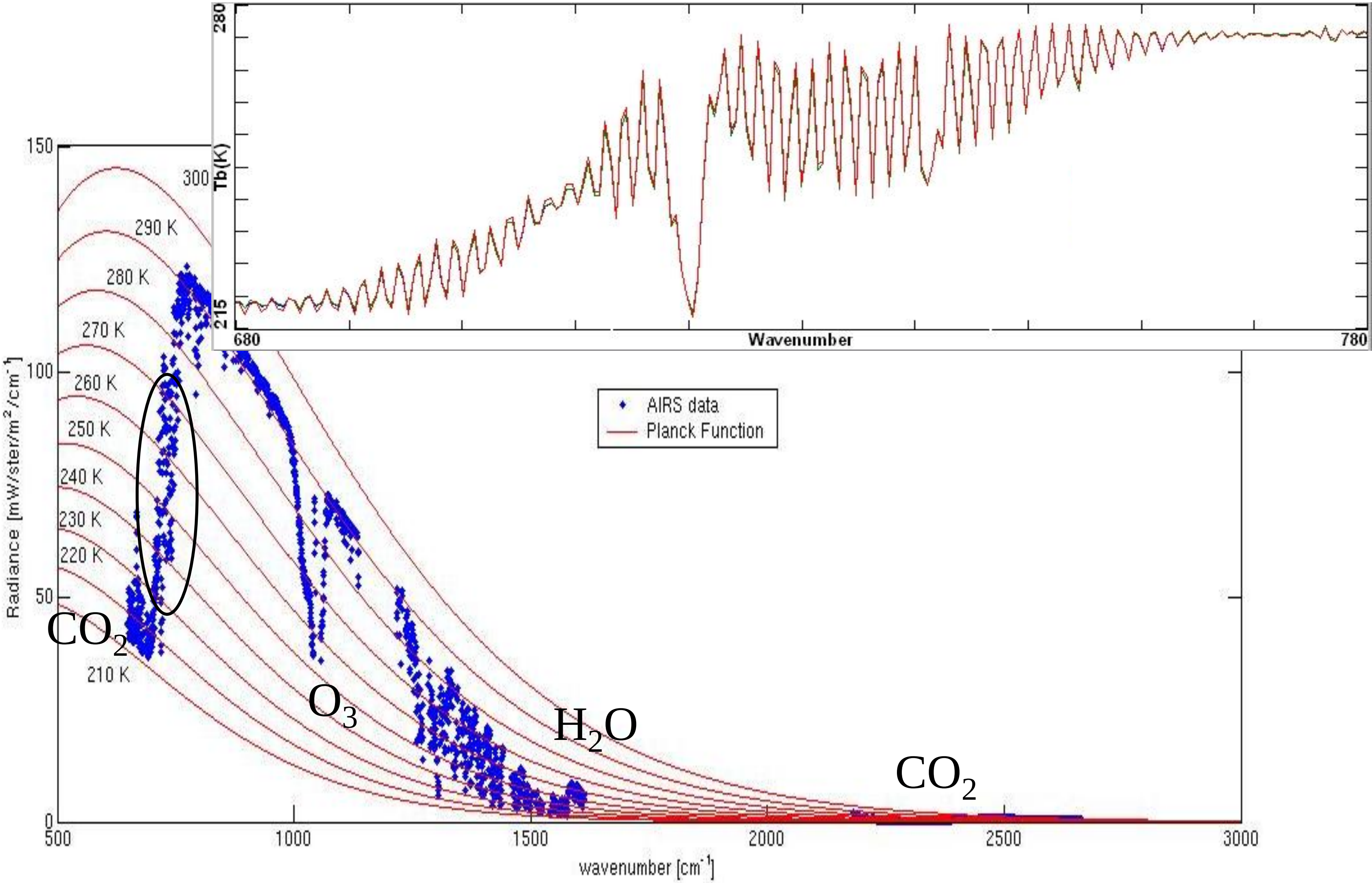
H_2O Vibration – Rotation Spectra



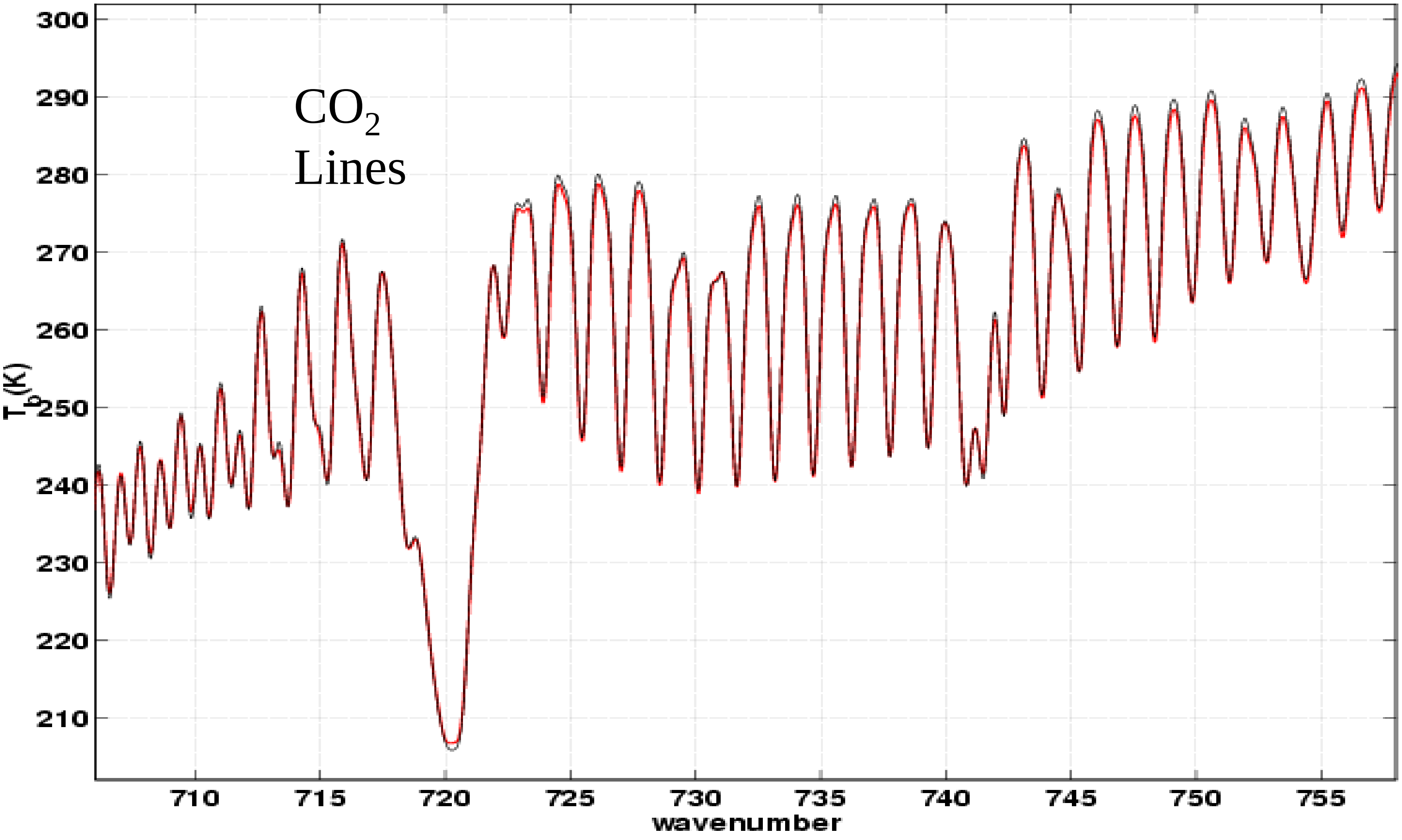



H2O
O3
CO2
Radiance [mW/ster/m2/cm-1]
Wavenumber [cm-1]
160
140
120
100
80
60
40
20
700
750
800
850
900
950
1000
1050

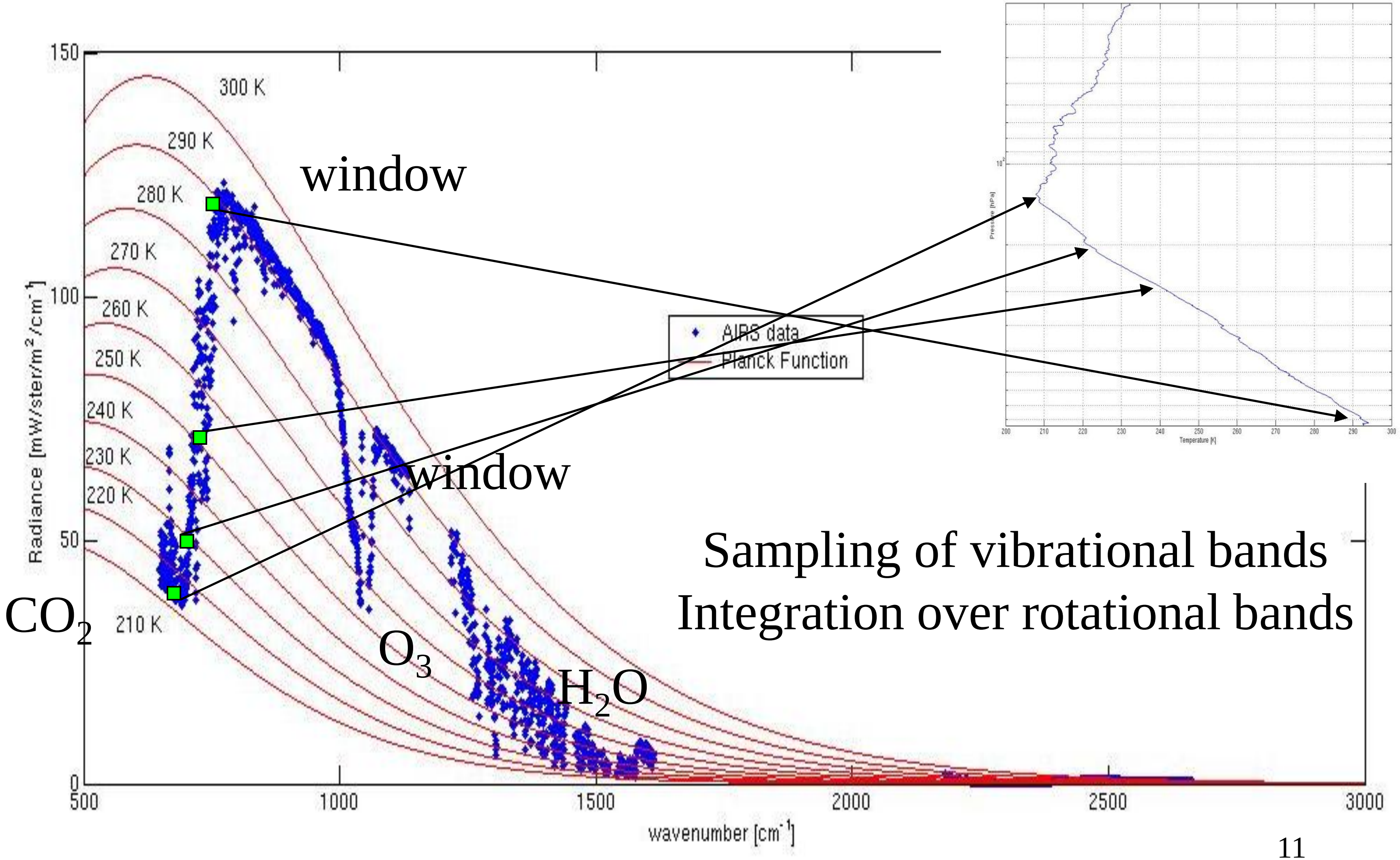
Rotational Lines



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

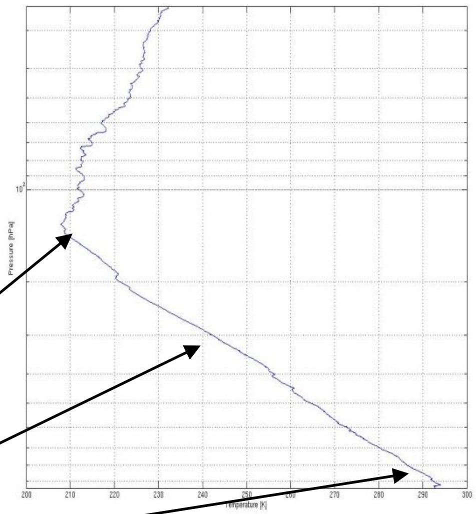


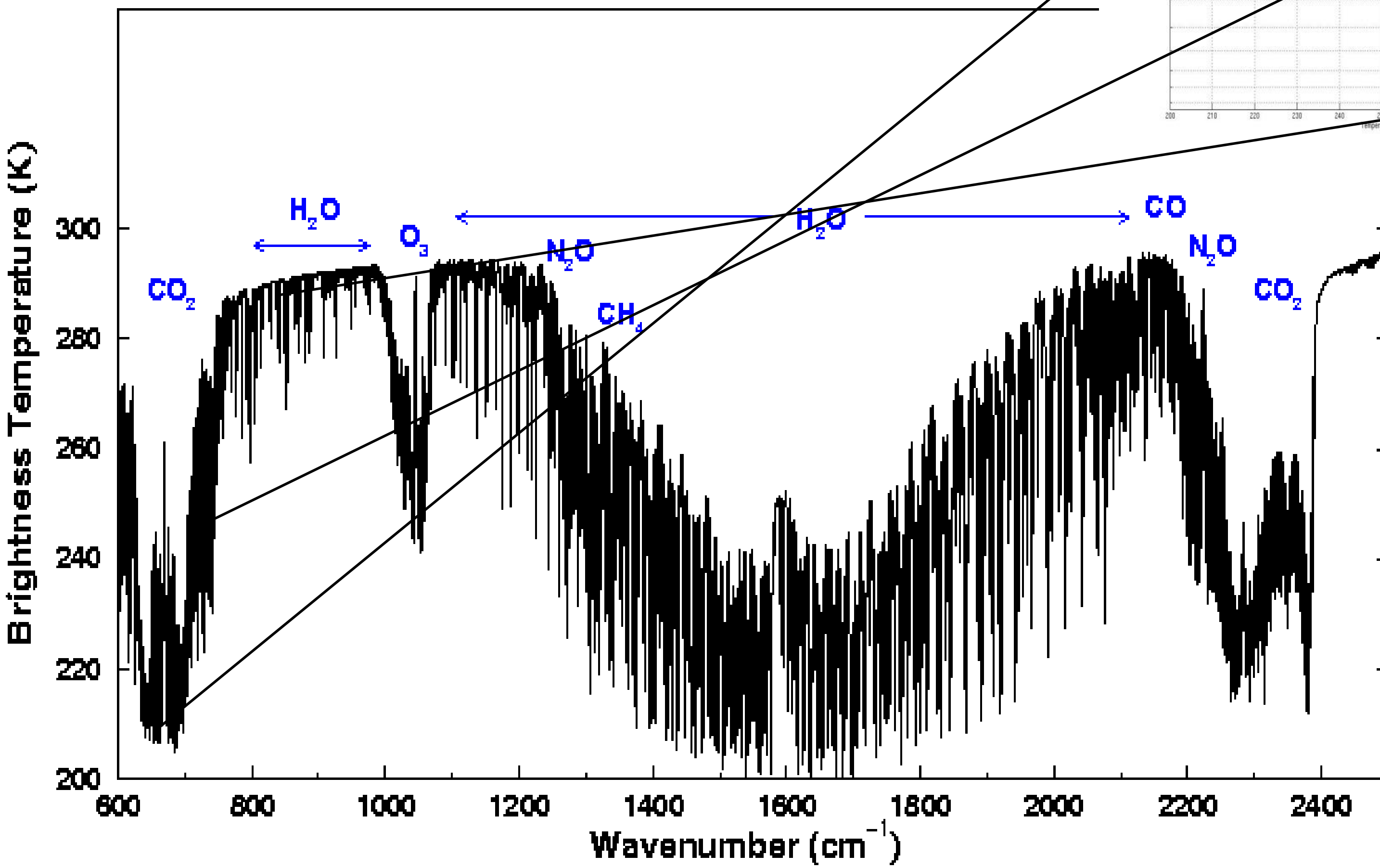
Broad Band



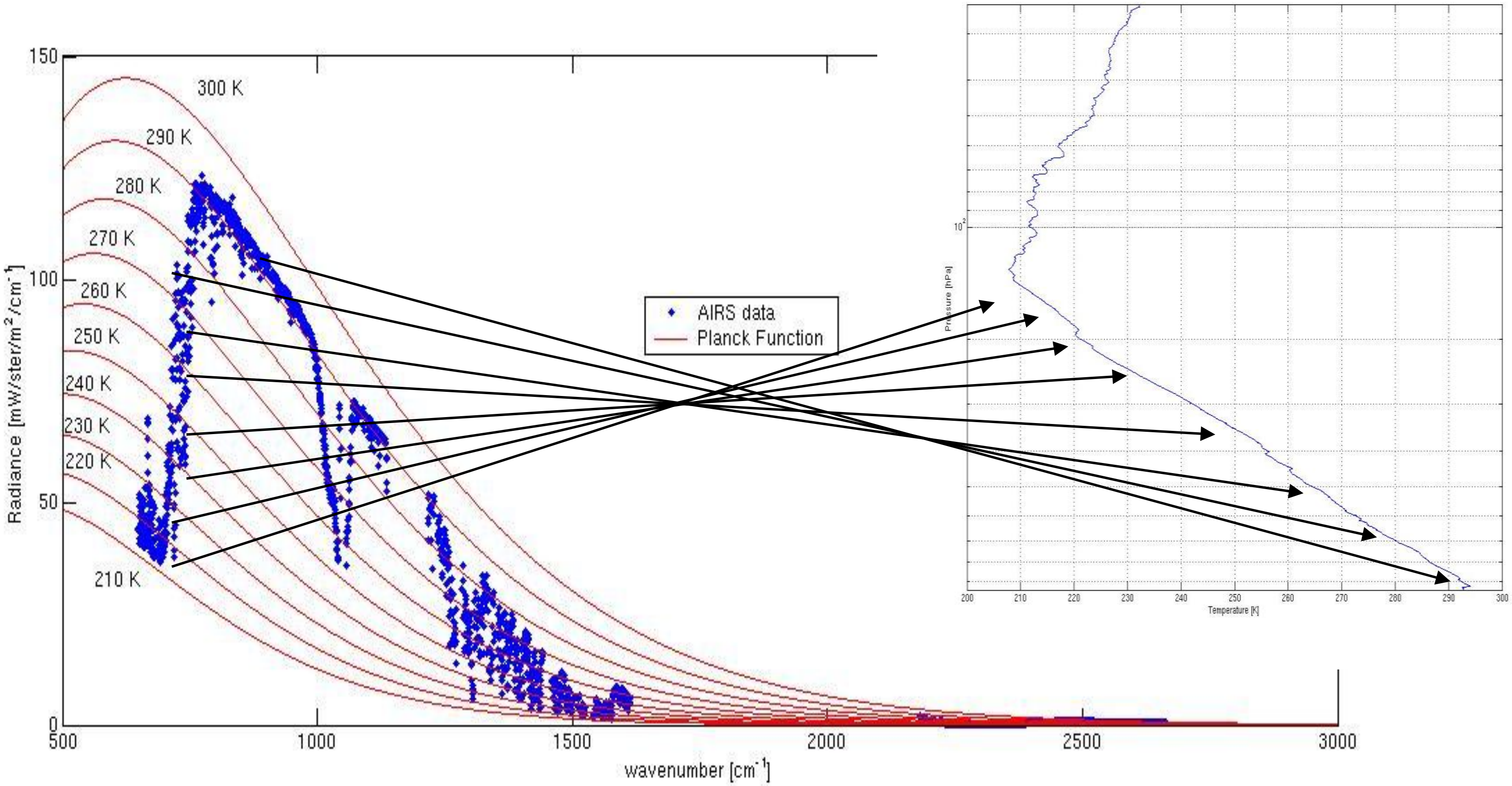
Sampling of vibrational bands
Integration over rotational bands

… in Brightness Temperature





High Spectral Resolution



Sampling over rotational bands

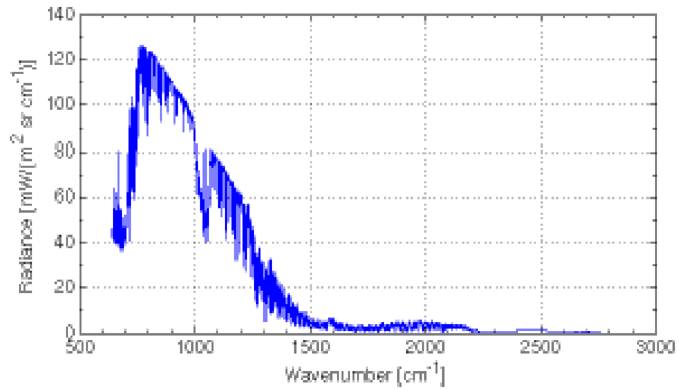
Infrared Radiance and Brightness Temperature Spectrum

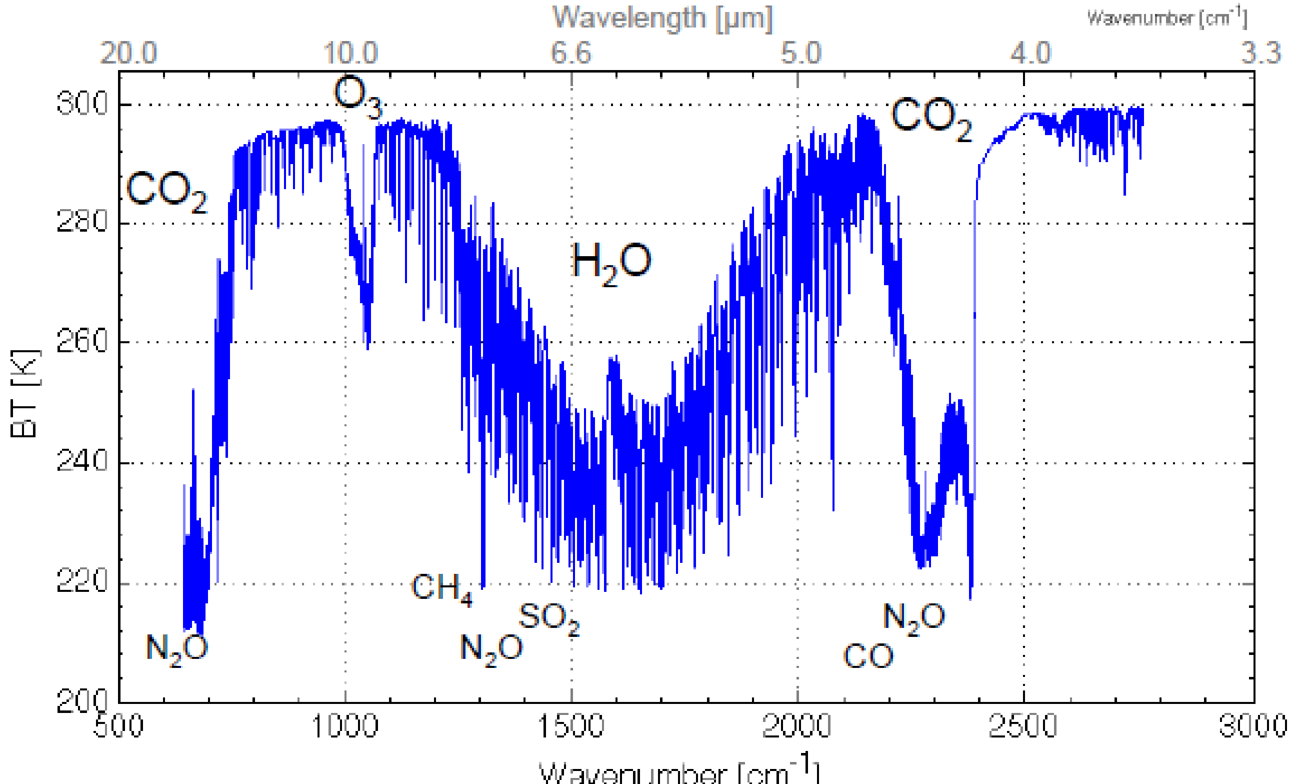
Planck Function

$$B_\nu(T) = \frac{2hc^2\nu^3}{\exp(\frac{hc\nu}{kT}) - 1}$$

Upwelling IR radiation

$$R_\nu = \int_{z_0}^{\infty} B_\nu(T(z)) \frac{d\tau_\nu(z)}{dz} dz$$



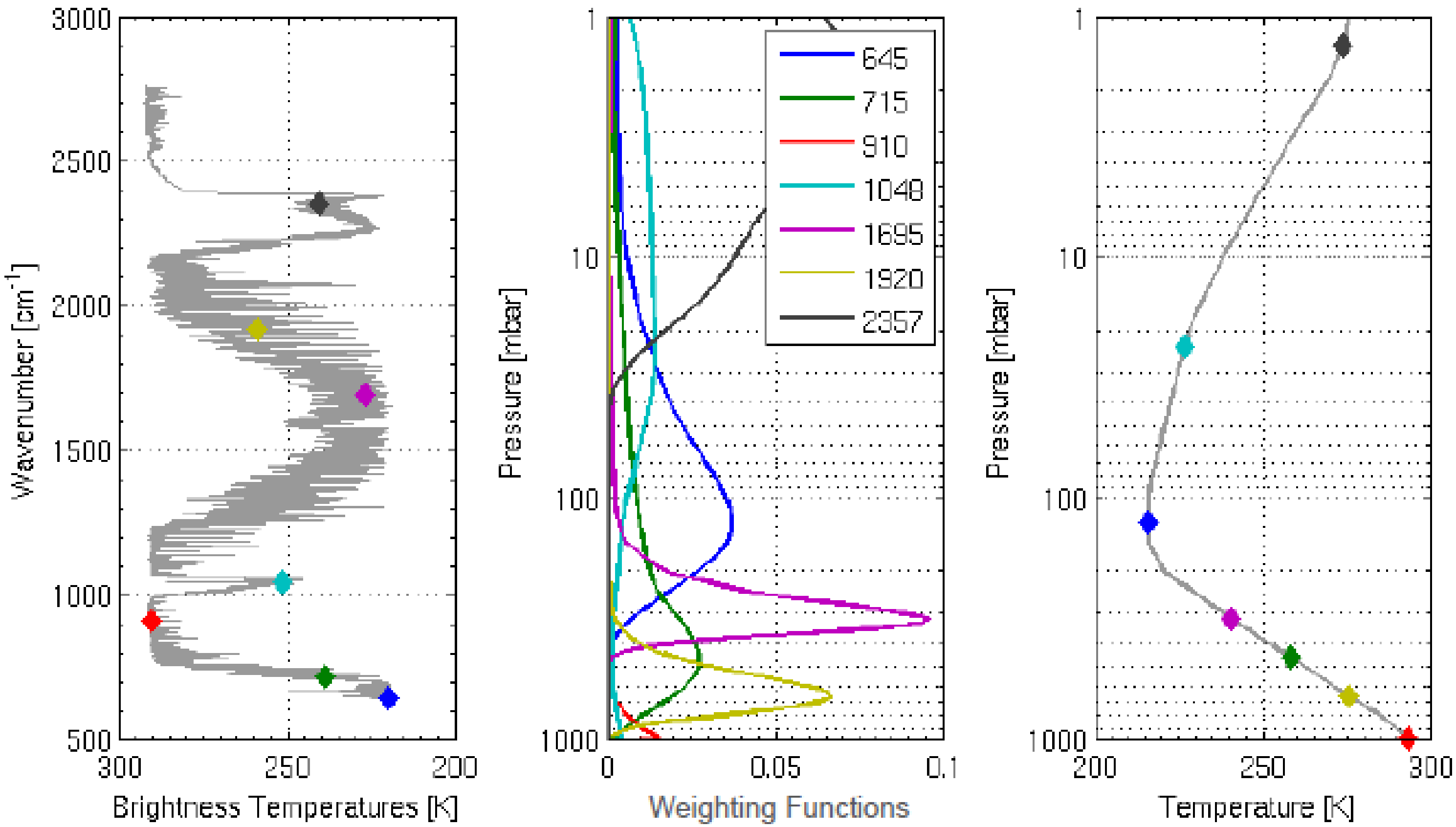


From E. Weisz

Atmospheric Temperature Profile Retrieval

$$R_\nu = \int_{ps}^{0} B_\nu(T(p))W_\nu(p)\,dp$$

$$W_\nu(p) = \partial \tau_\nu(p) \Big/ \partial \ln p$$



High-spectral measurements → Profiles at high-vertical resolution

From E. Weisz

Regression Retrieval Summary

Training Set

↓

Simulated Radiances

↓

Regression Coefficients

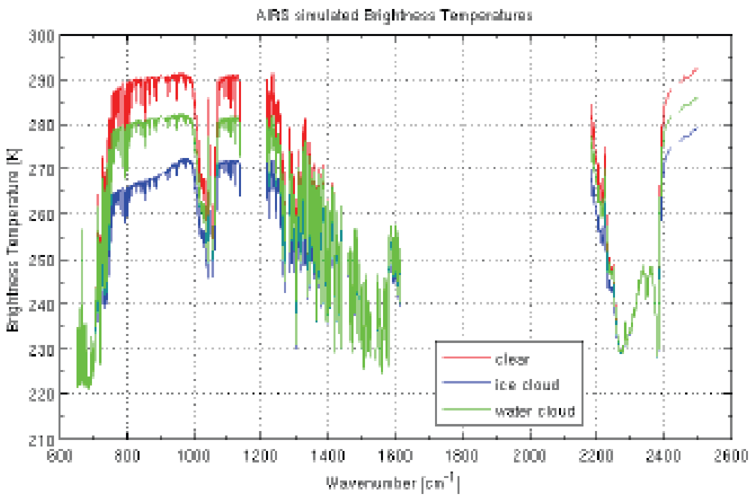
↓

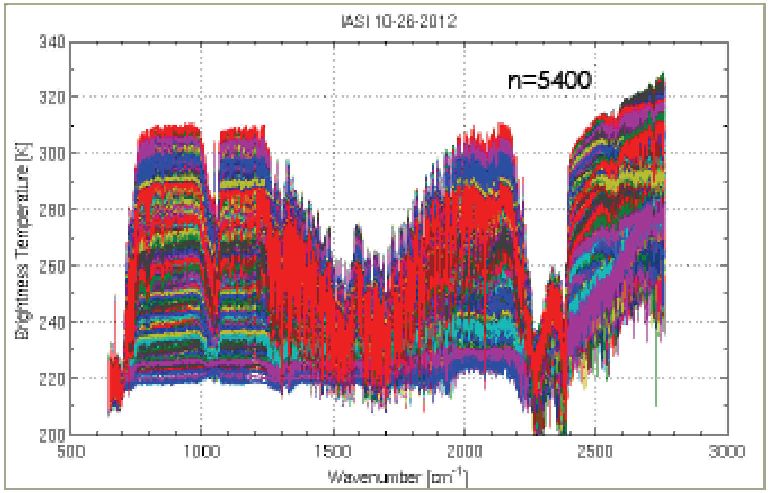
Retrieval

↓

Temperature, Moisture and Ozone profiles,
Surface and cloud parameters ….

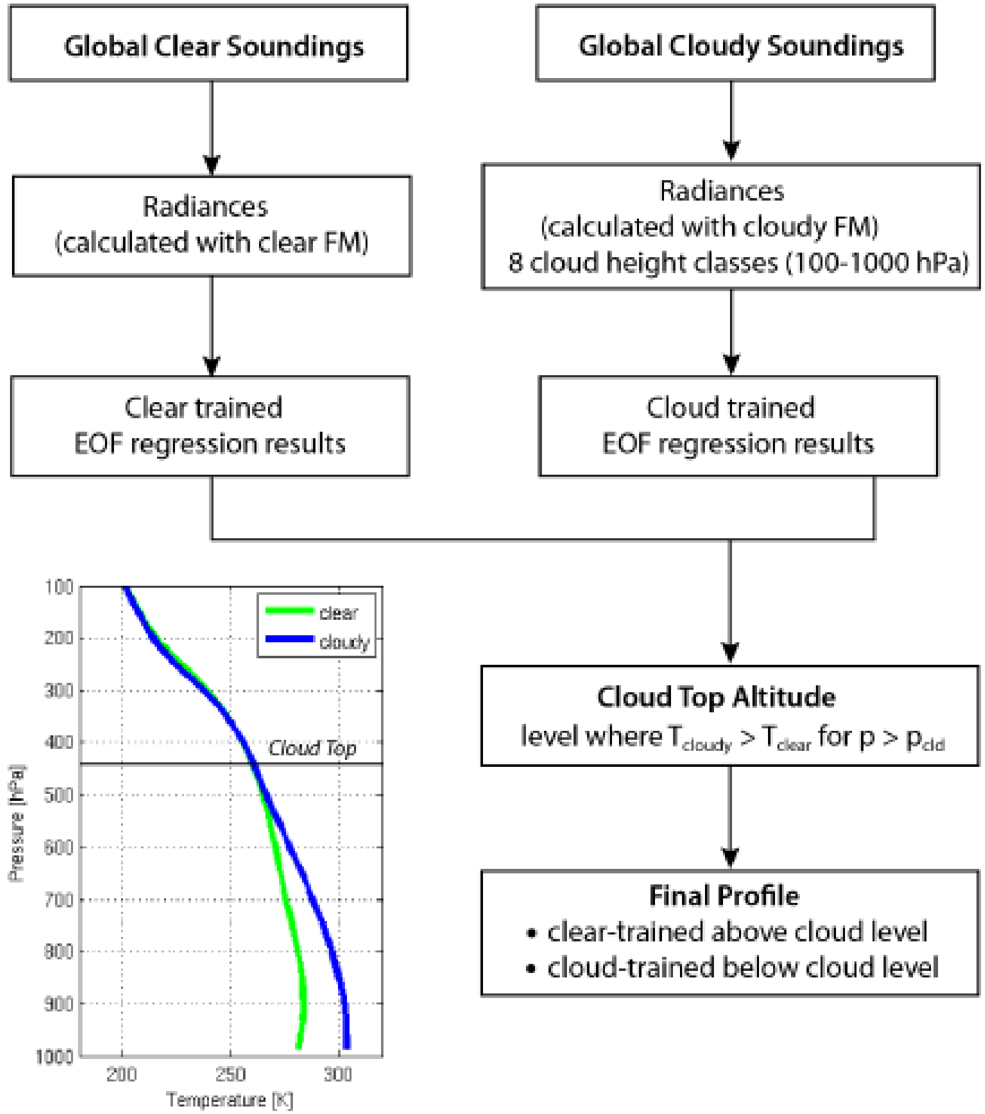
Radiances
AIRS L1B, IASI L1C, CrIS SDR



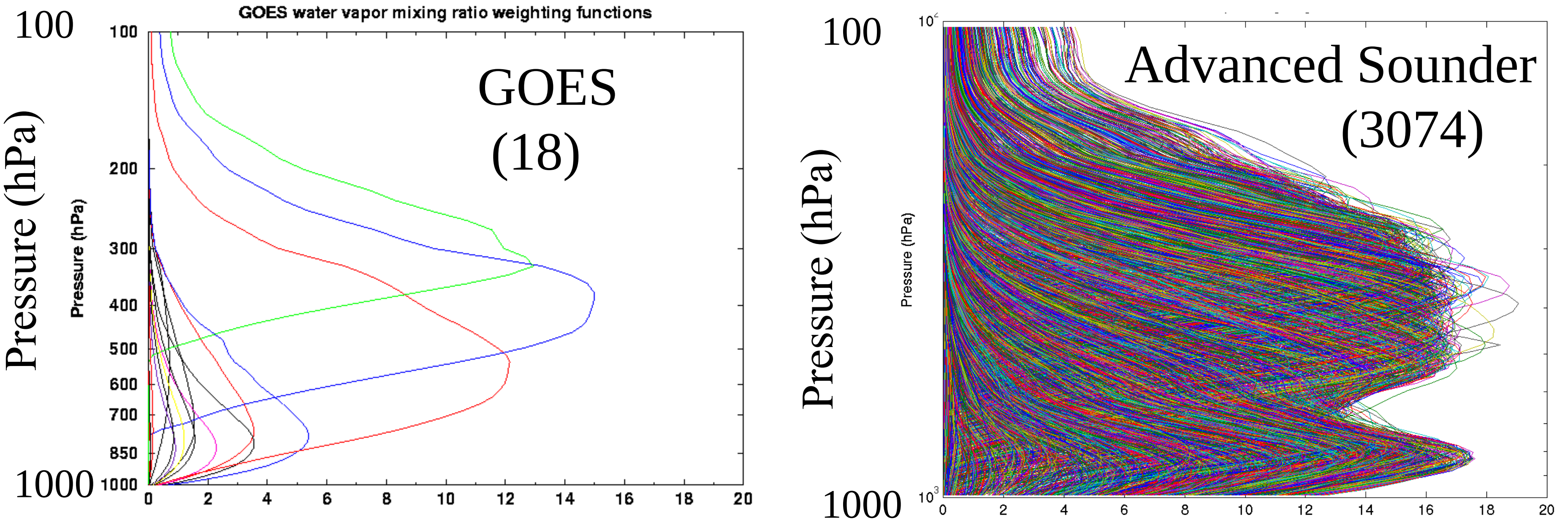


From E. Weisz

Dual-Regression Retrieval

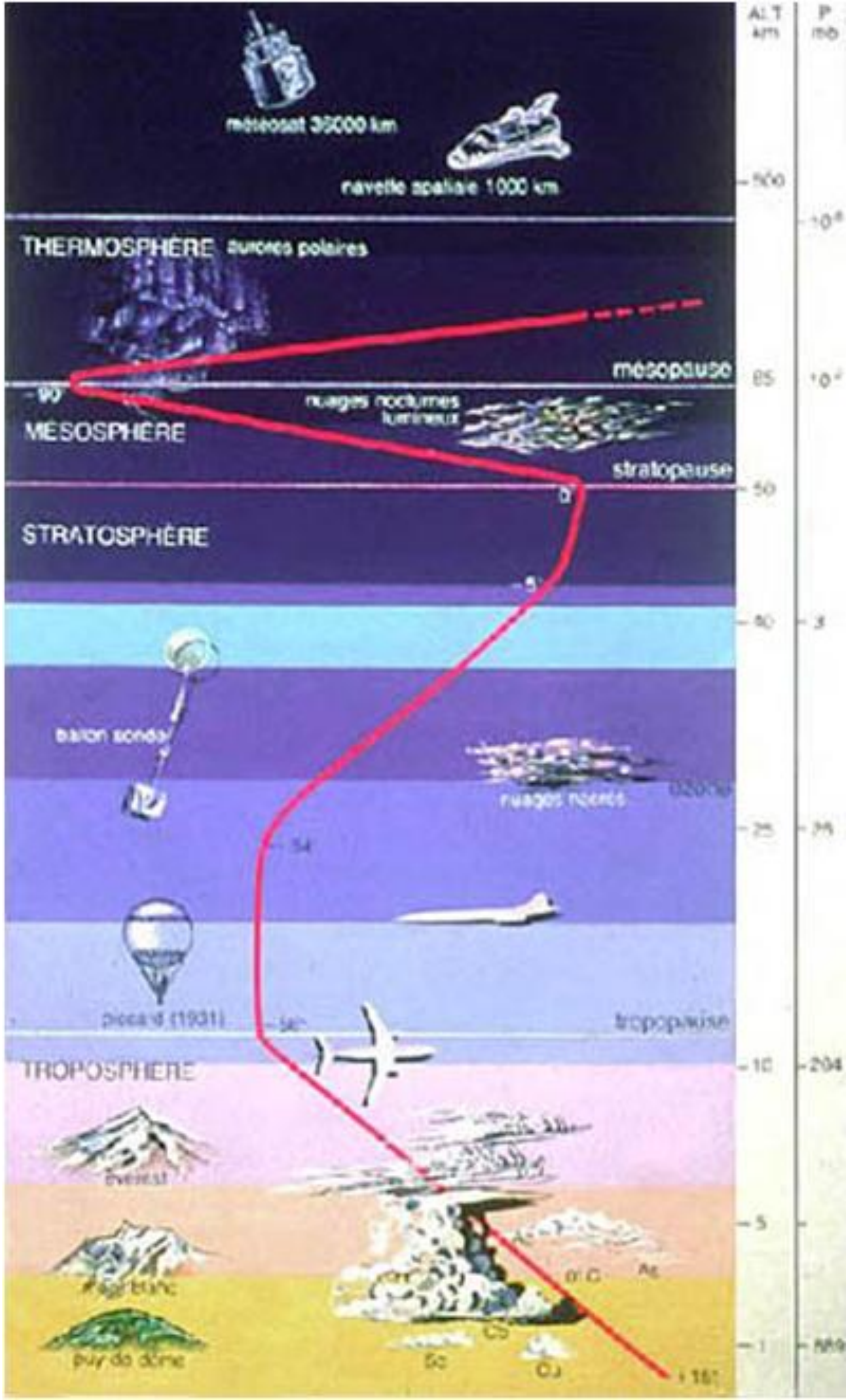


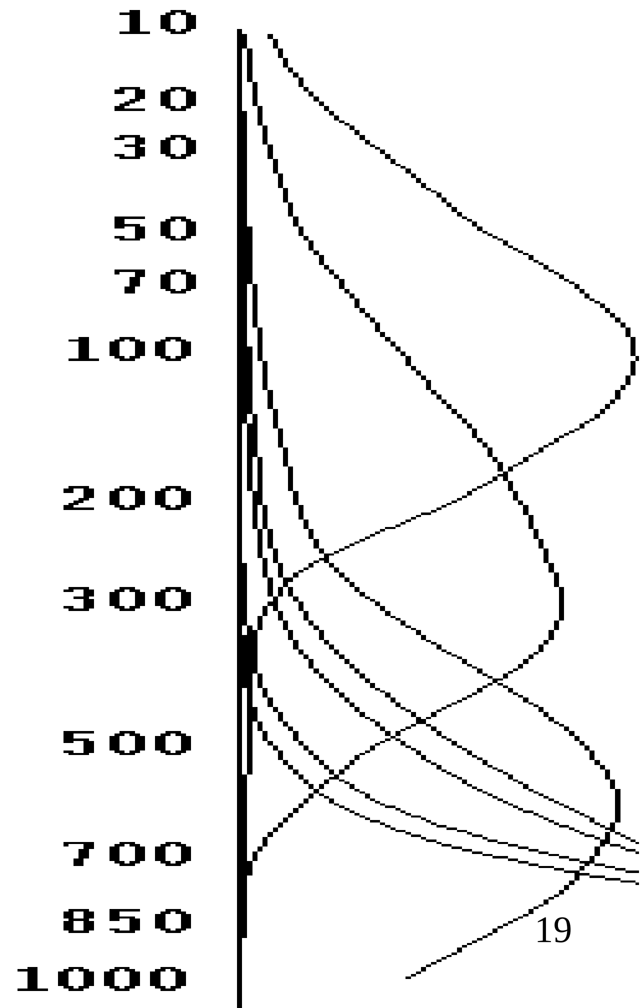
From E. Weisz



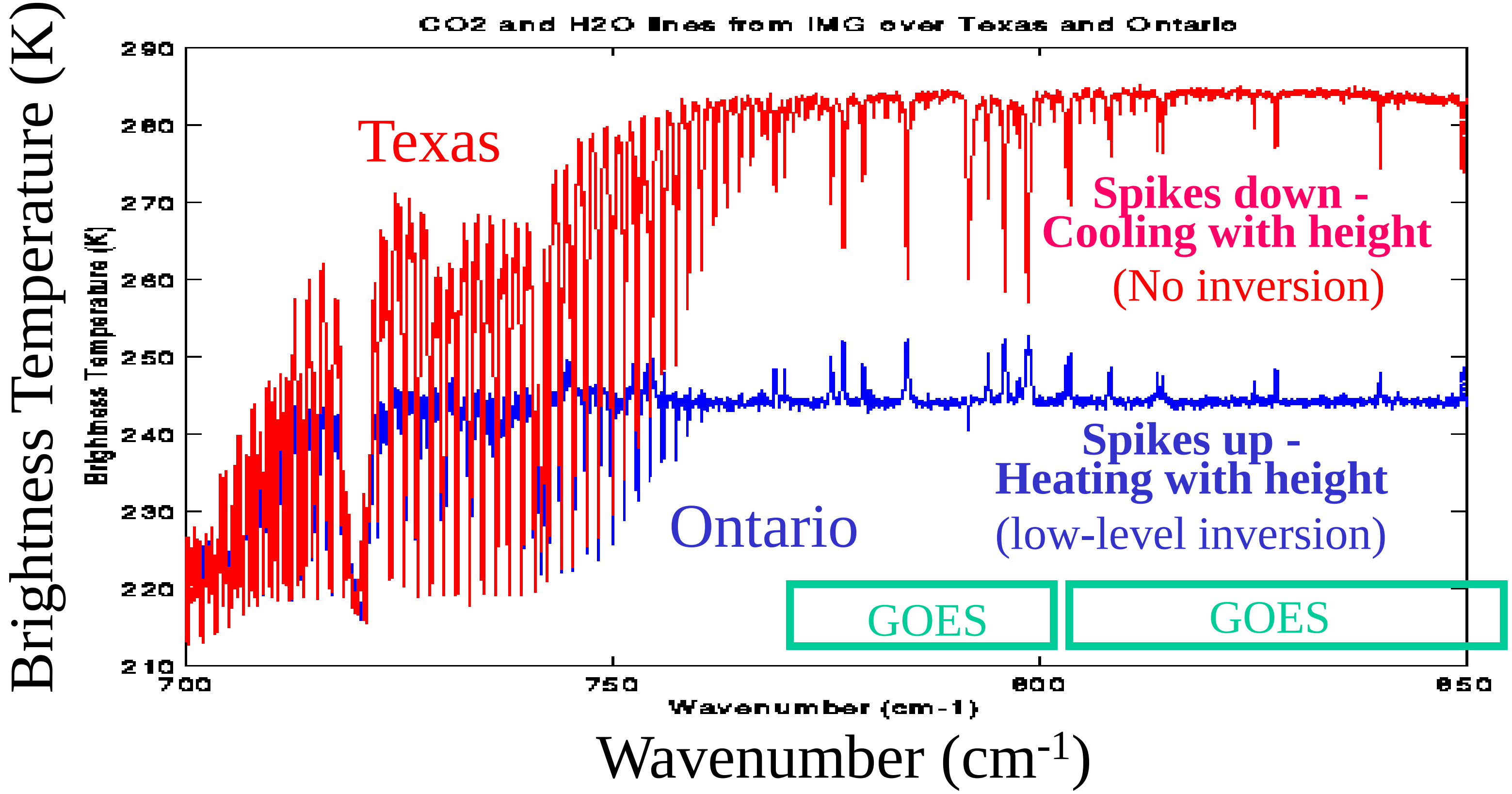
Moisture Weighting Functions

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
météosat 36000 km
navette spatiale 1000 km
500
THERMOSPHÈRE
aurores polaires
mésopause
85
MÉSOSPHÈRE
nuages nocturnes lumineux
stratopause
50
STRATOSPHÈRE
40
ballon sonde
nuages nacrés
25
tropopause
TROPOSPHÈRE
10
Everest
5
puy de dôme
1

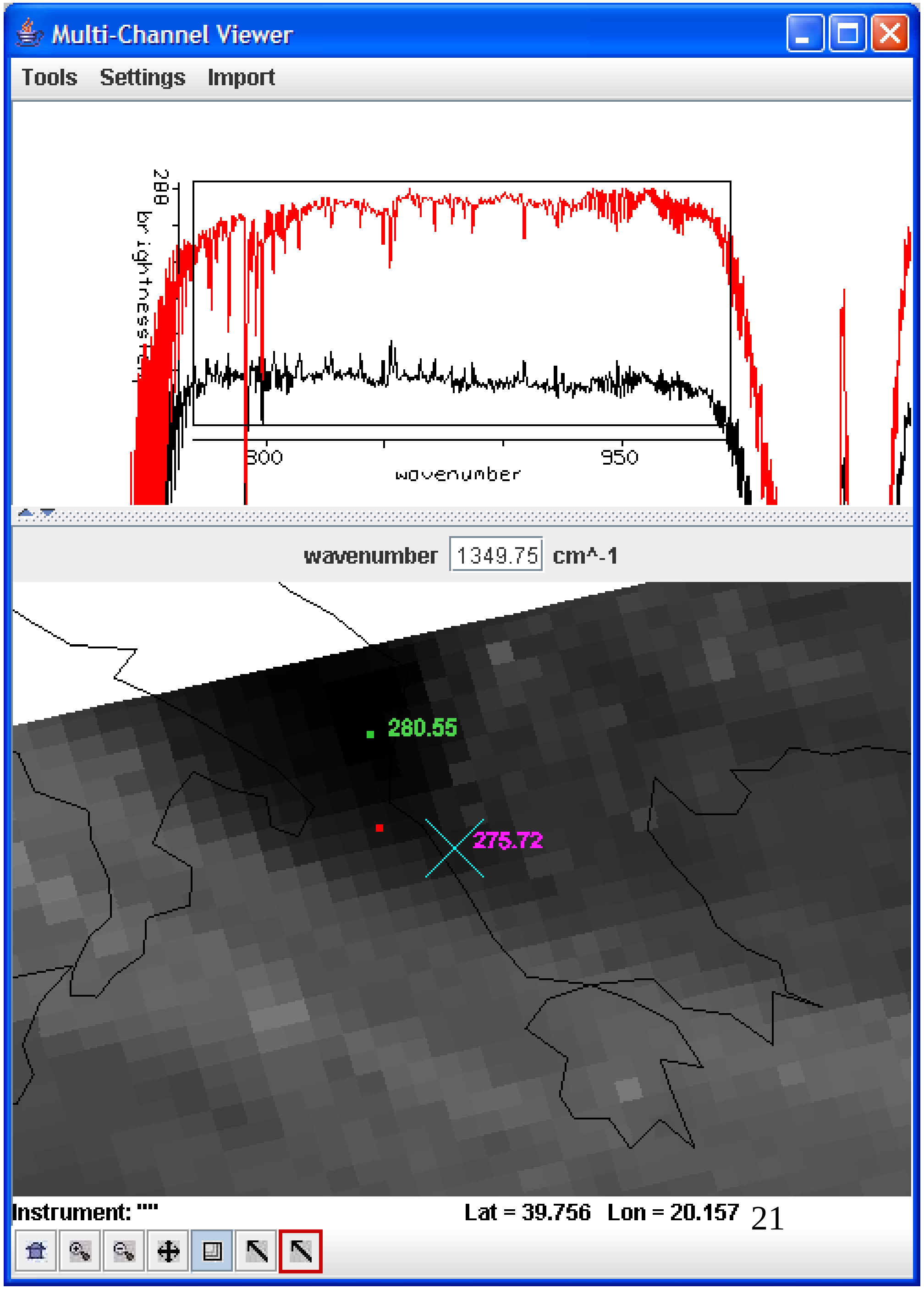

10
20
30
50
70
100
200
300
500
700
850
1000

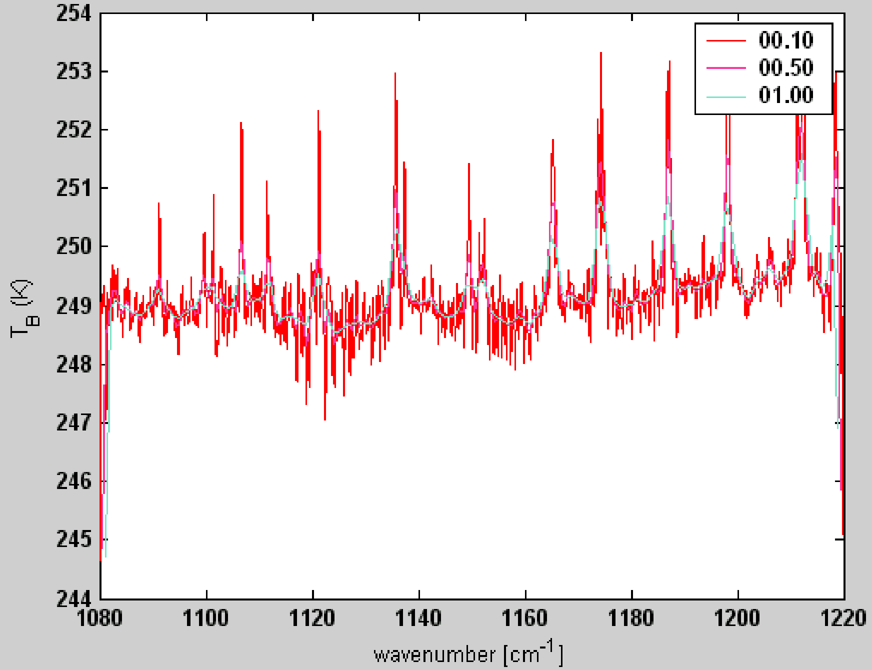
Resolving absorption features in atmospheric windows enables detection of temperature inversions

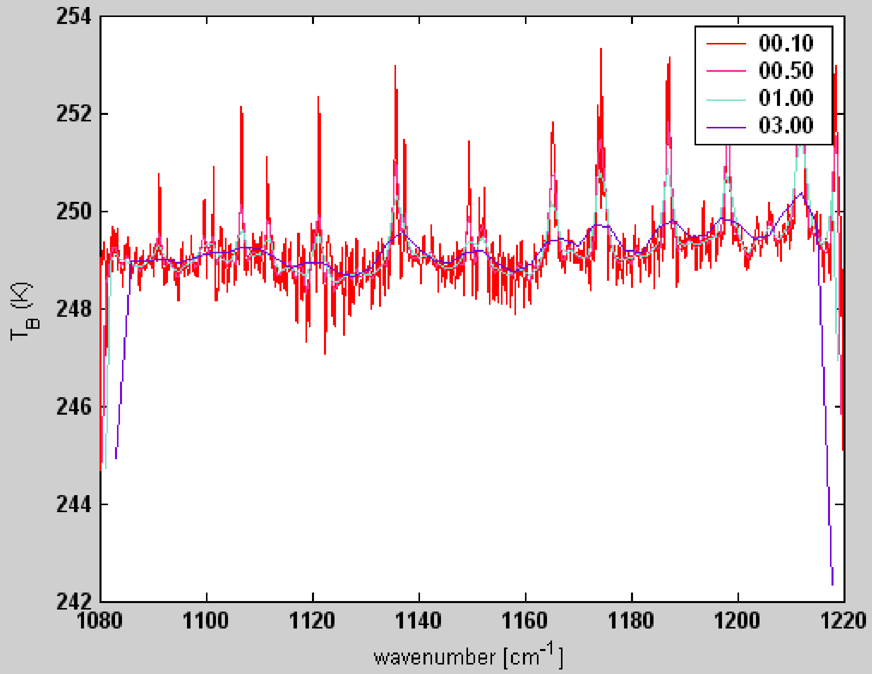


Detection of inversions is critical for severe weather forecasting. Combined with improved low-level moisture depiction, key ingredients for night-time severe storm development can be monitored.

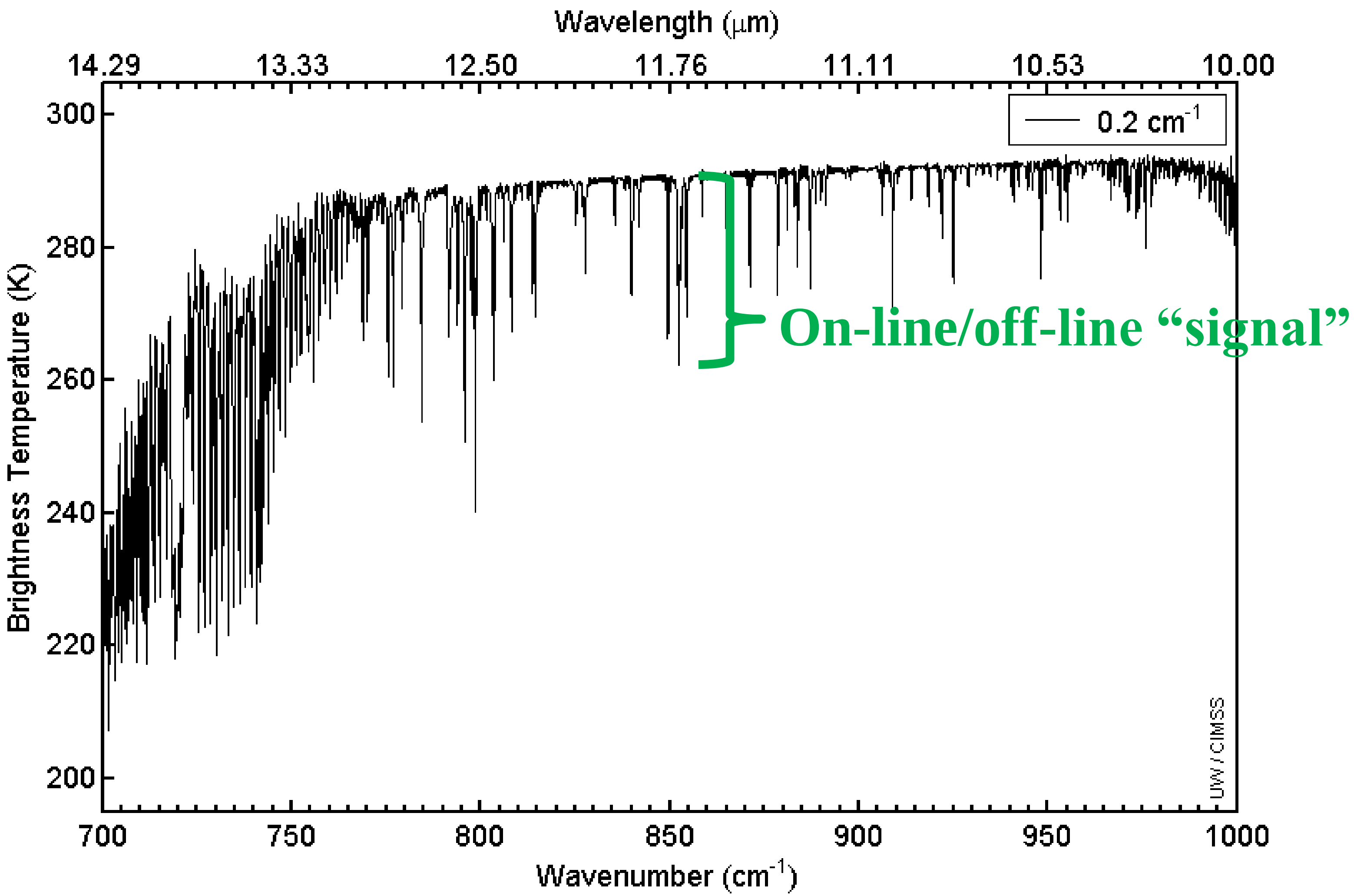
IASI detection of temperature inversion (black spectrum) vs clear ocean (red spectrum)



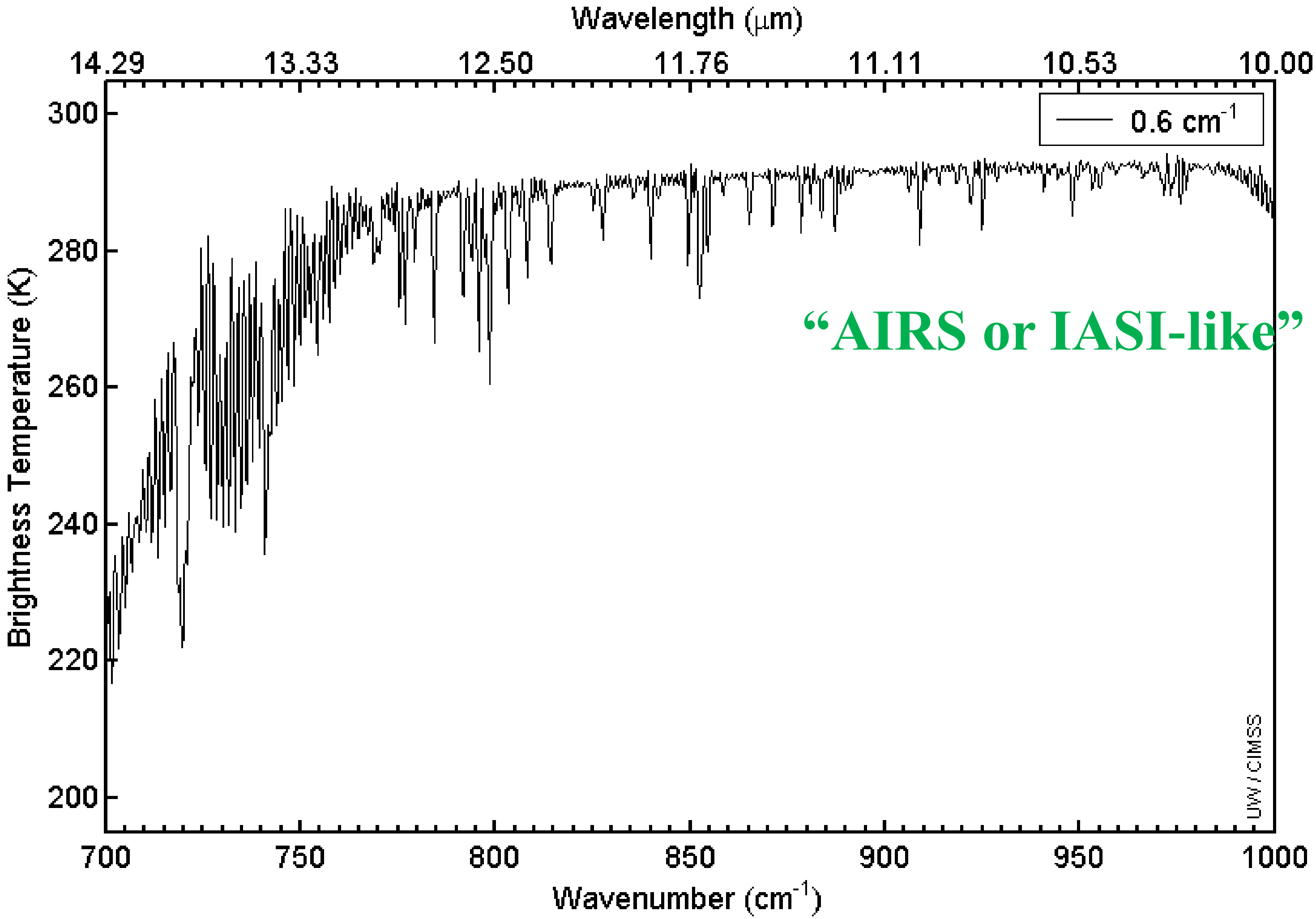




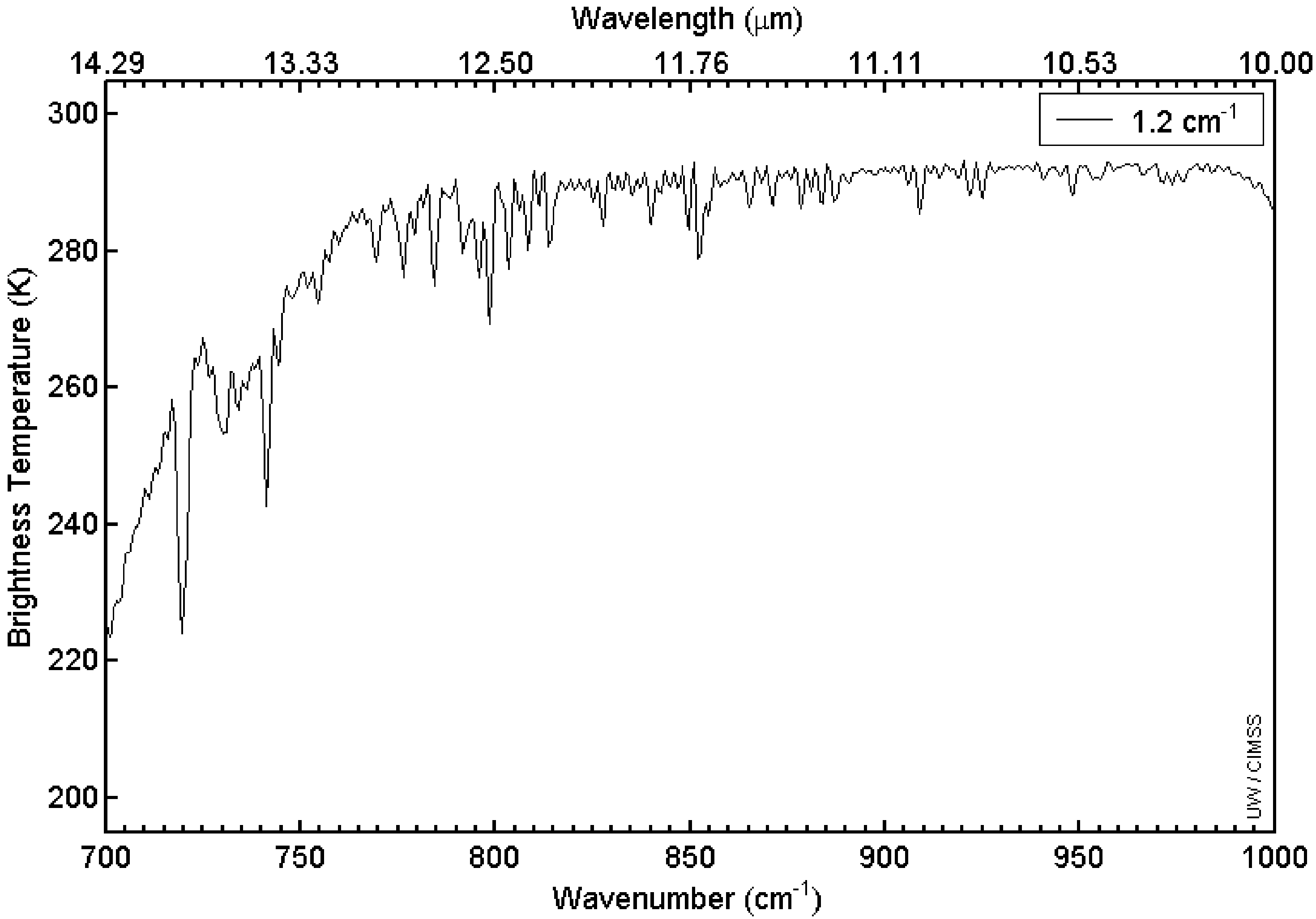
Ability to detect inversions disappears with broadband observations (> 3 cm-1)



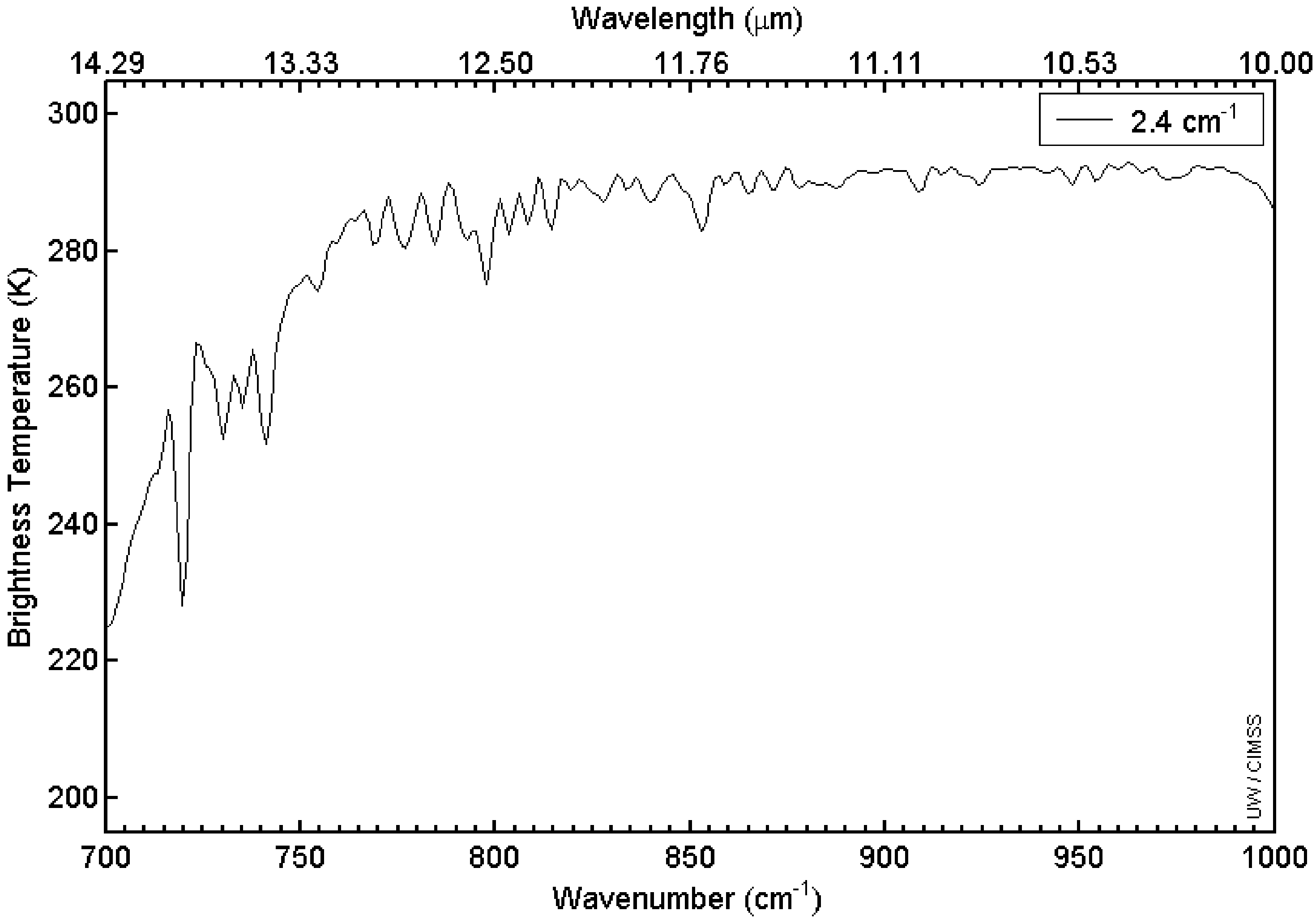
Longwave window region



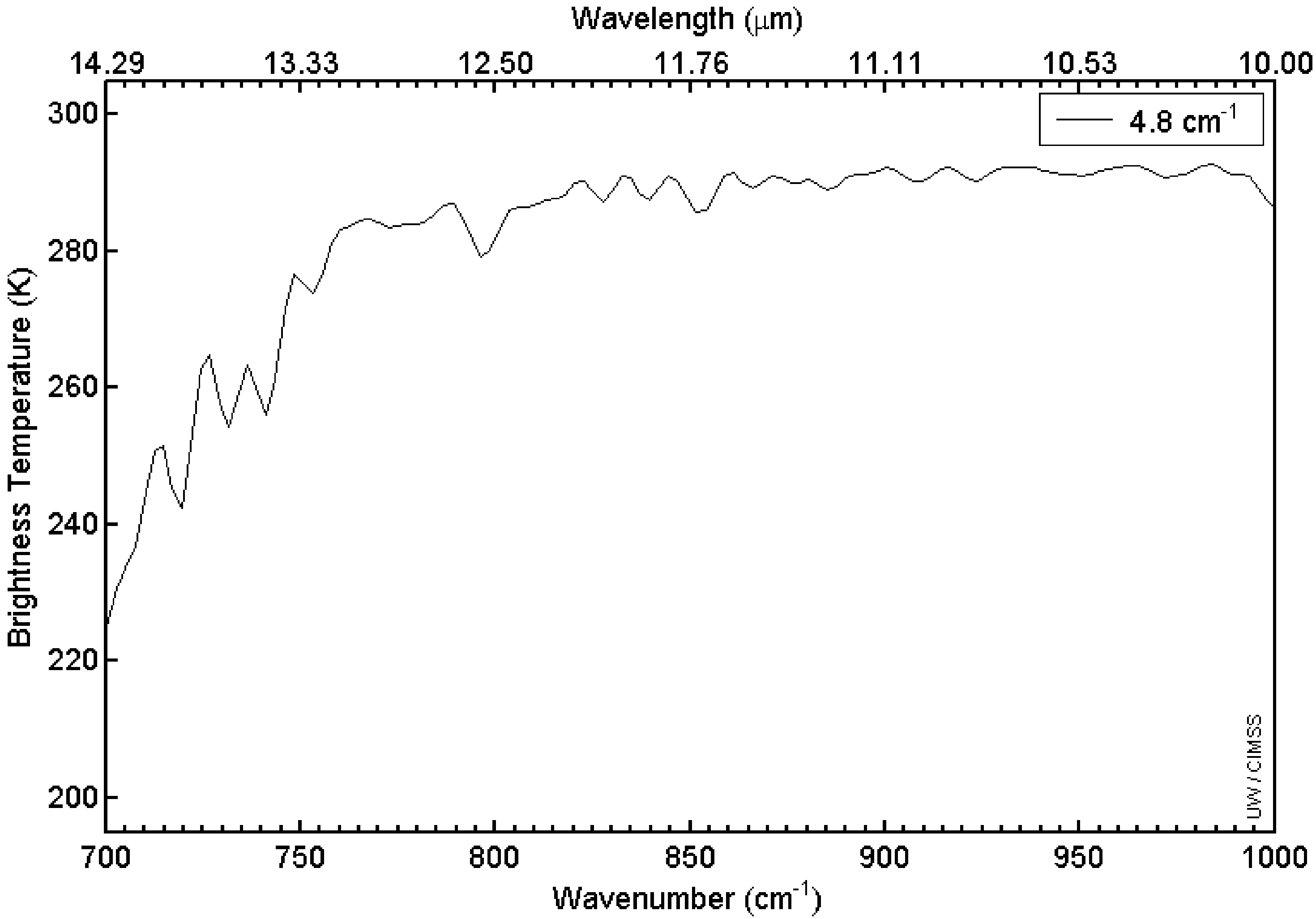
Longwave window region



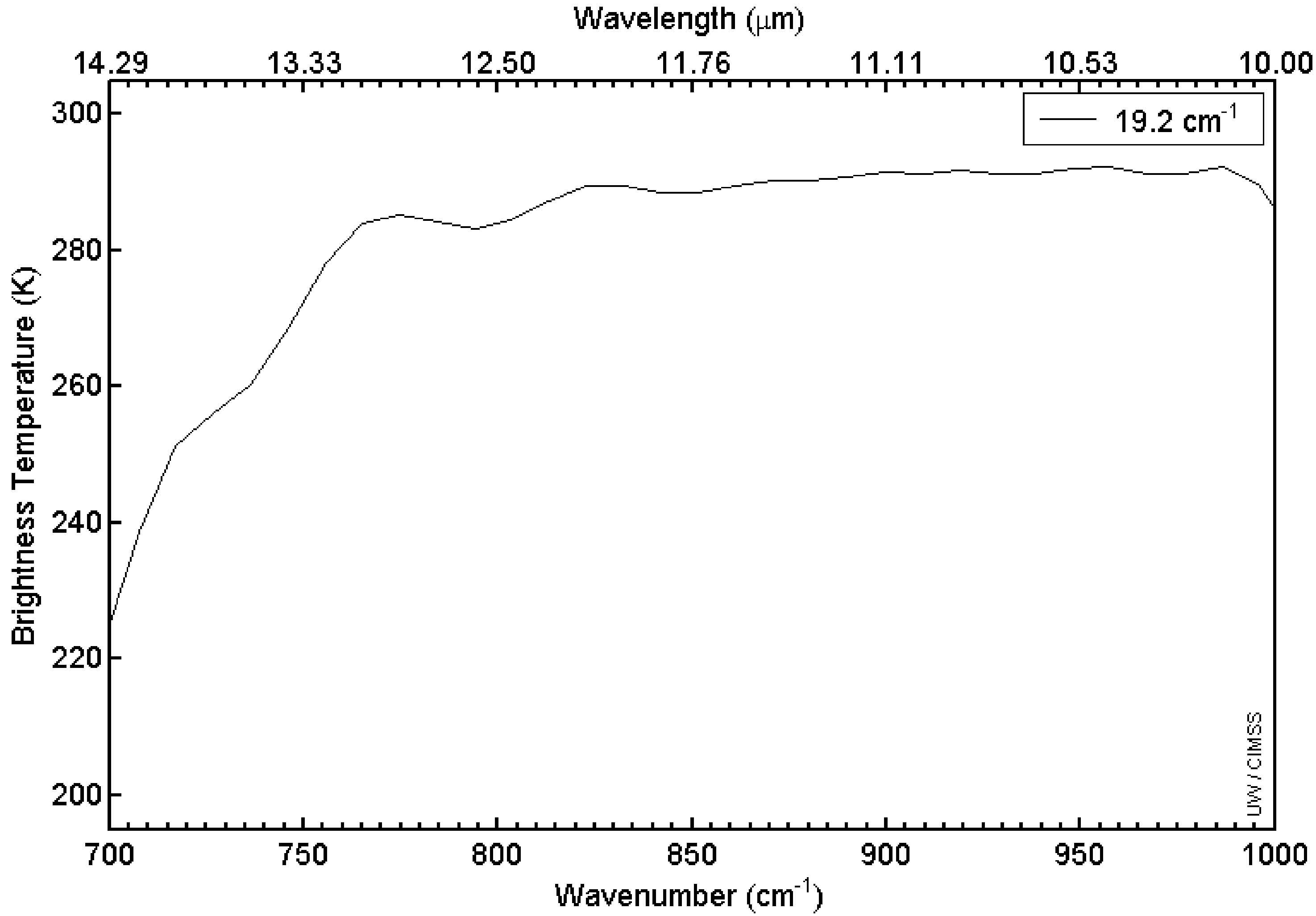
Longwave window region



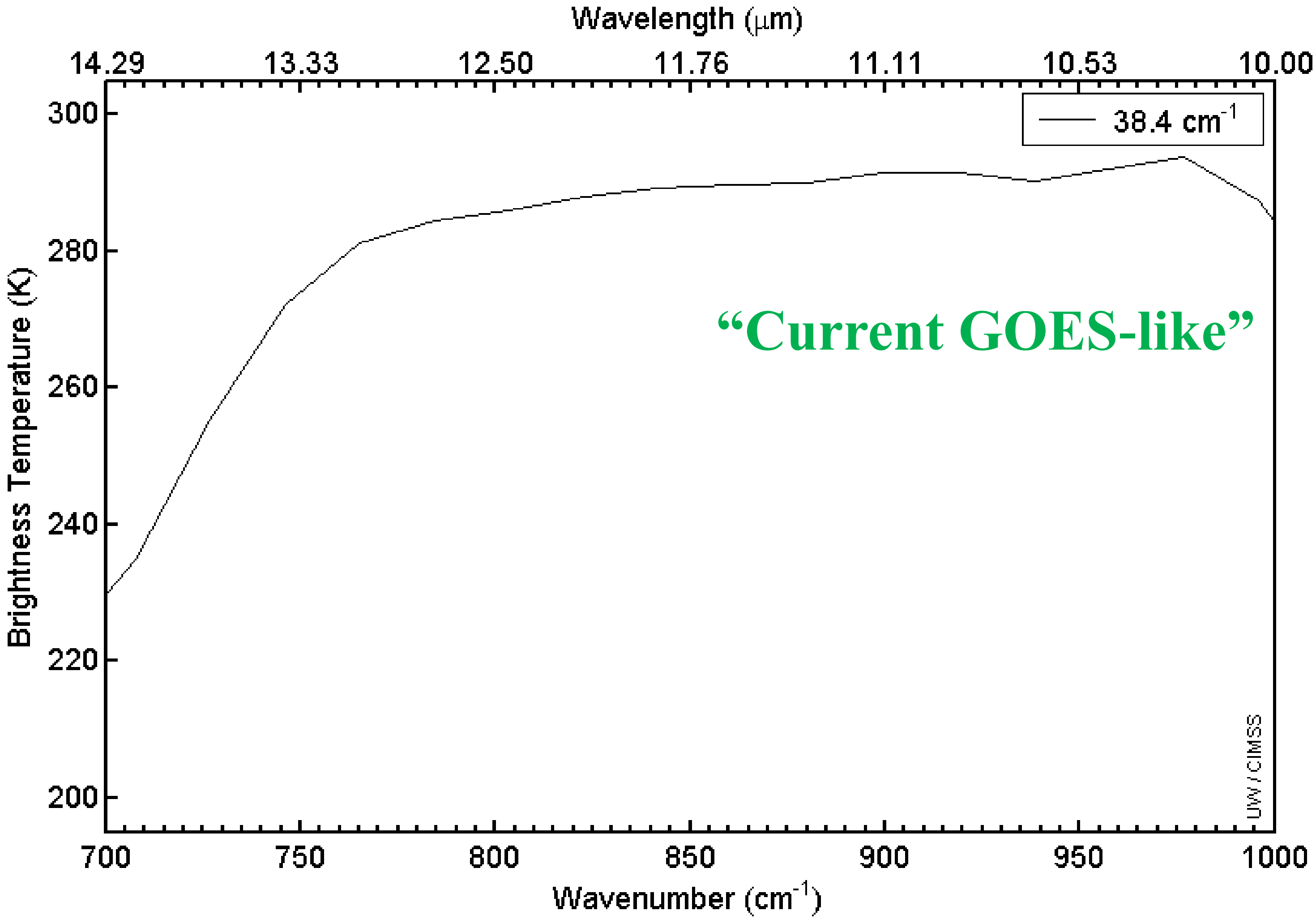
Longwave window region



Longwave window region



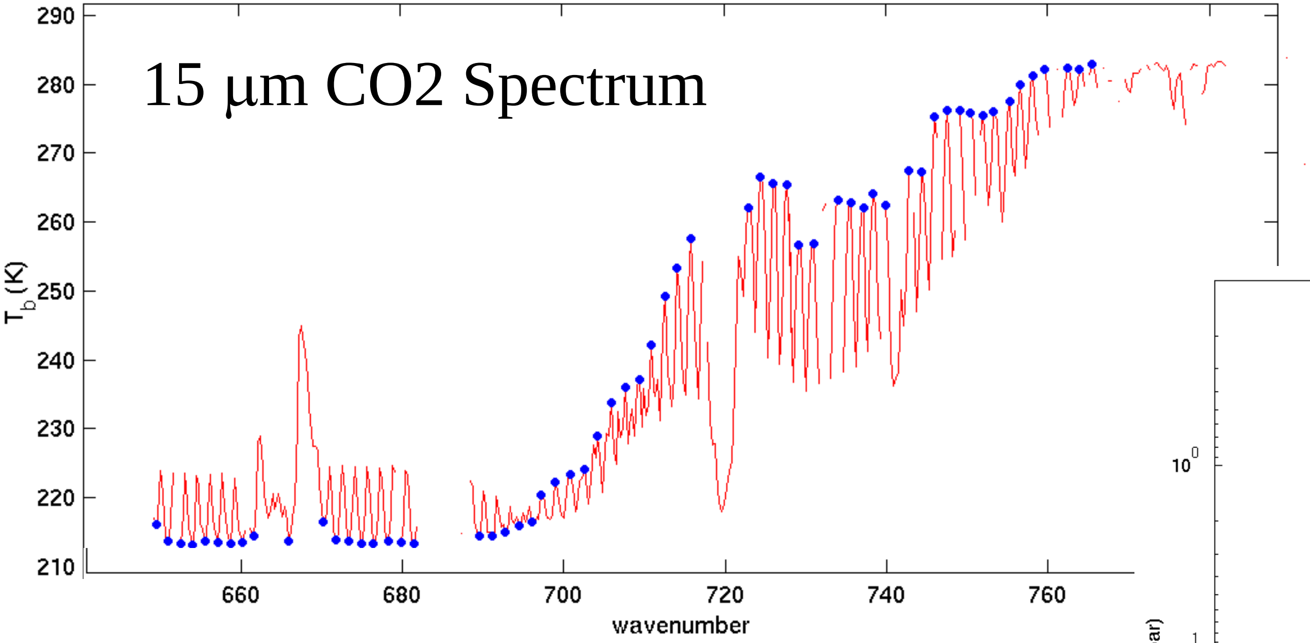
Longwave window region

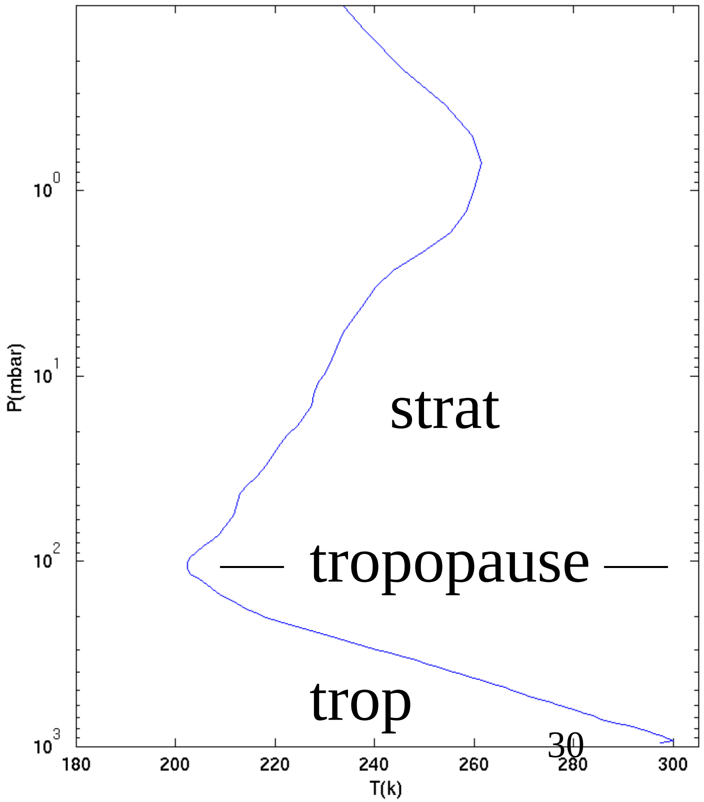


Longwave window region

Twisted Ribbon formed by CO_2 spectrum:

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



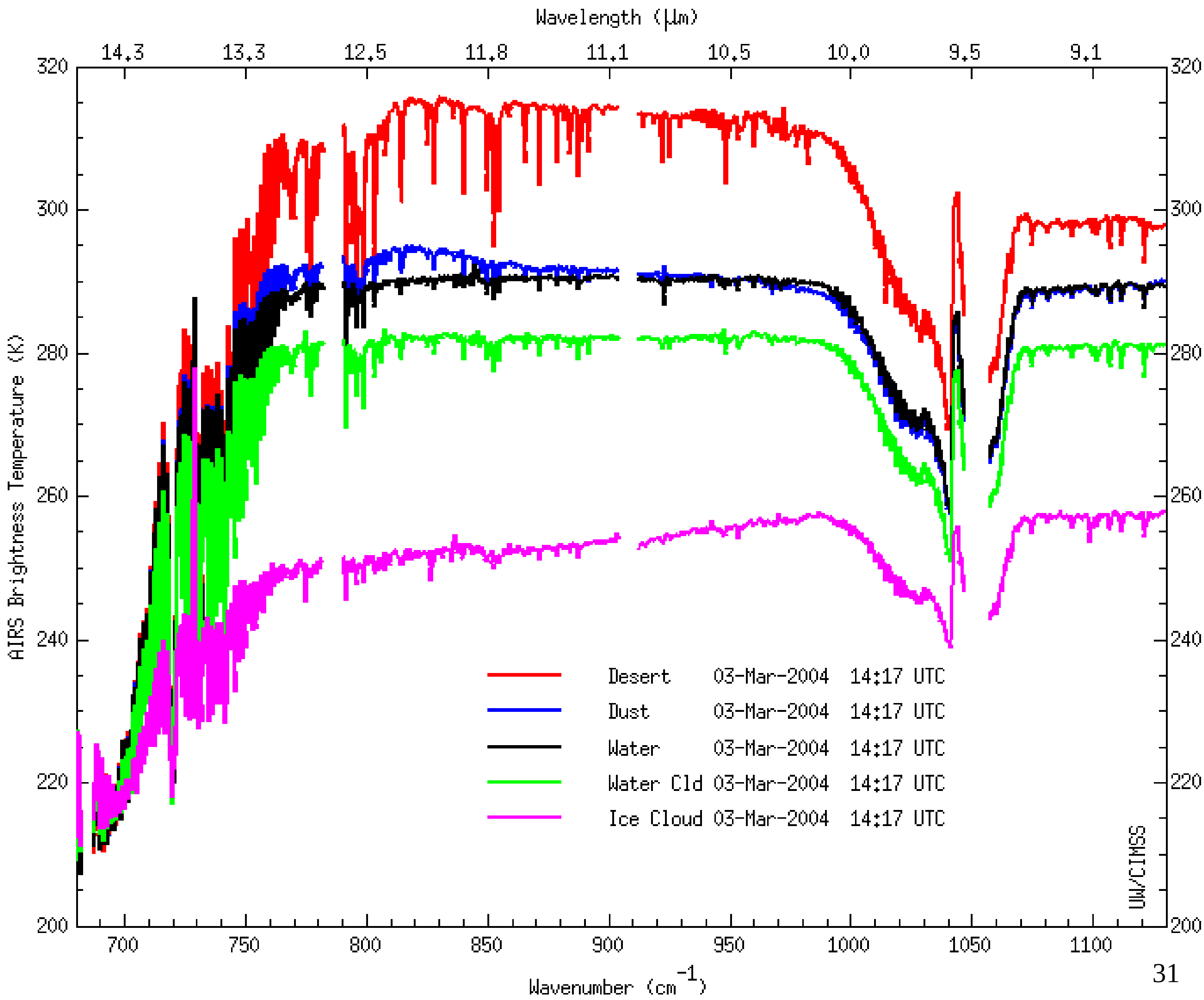


Blue between-line T_b

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

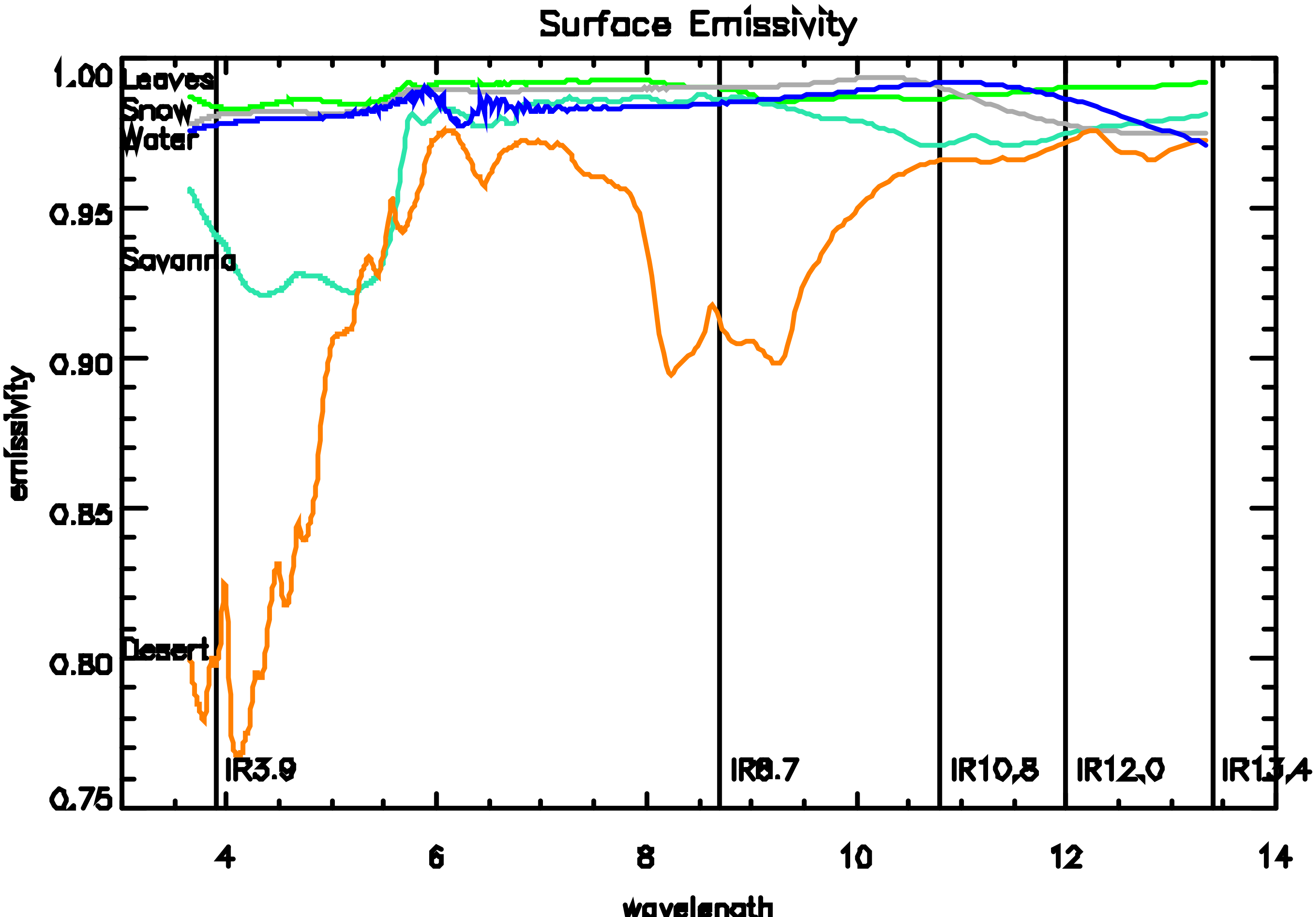
<u>colder</u> for stratospheric channels

Signature not available at low resolution



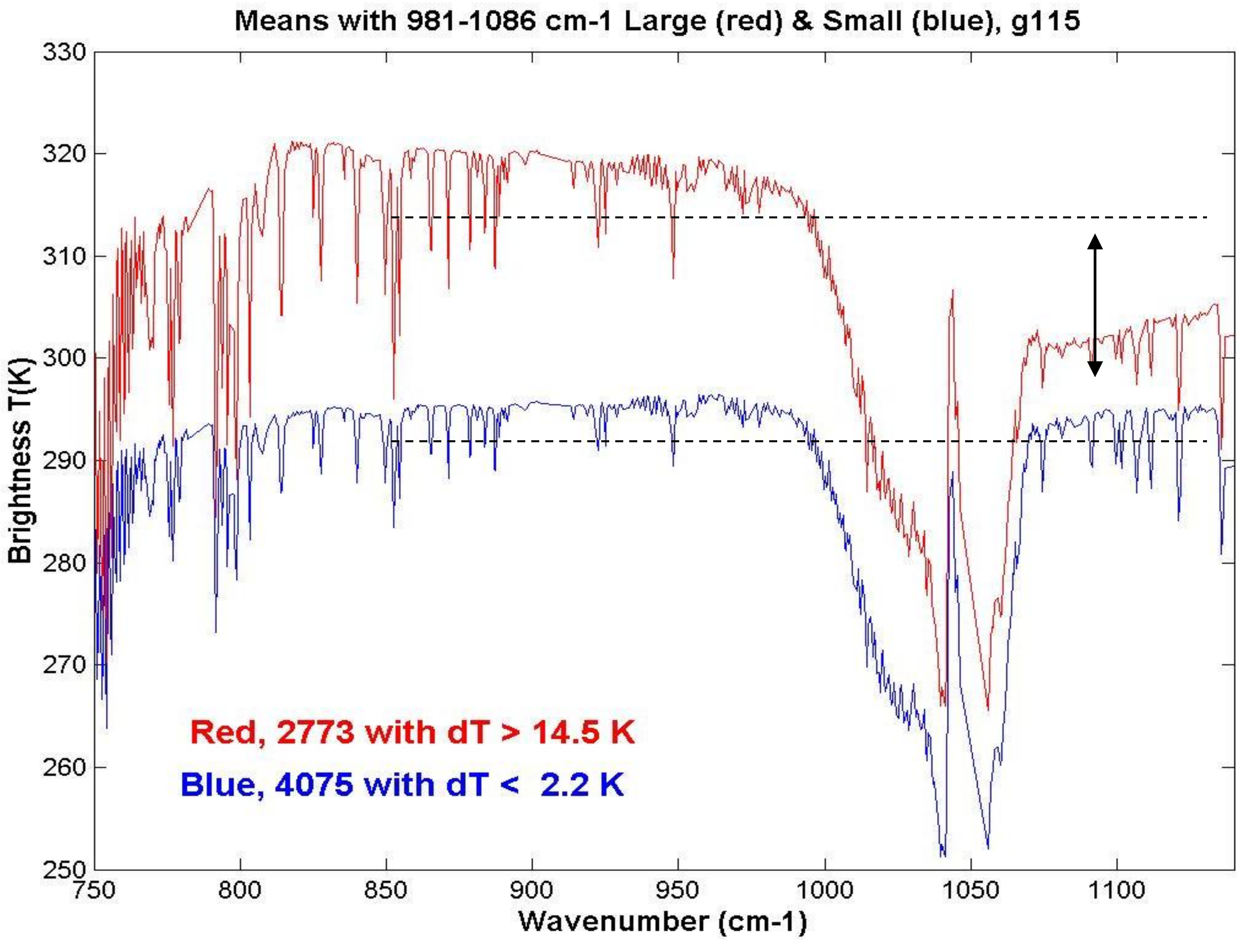
Wavelength (µm)
14.3
13.3
12.5
11.8
11.1
10.5
10.0
9.5
9.1
AIRS Brightness Temperature (K)
320
300
280
260
240
220
200
Desert 03-Mar-2004 14:17 UTC
Dust 03-Mar-2004 14:17 UTC
Water 03-Mar-2004 14:17 UTC
Water Cld 03-Mar-2004 14:17 UTC
Ice Cloud 03-Mar-2004 14:17 UTC
UW/CIMSS
700
750
800
850
900
950
1000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

2500 1000 715 cm-1



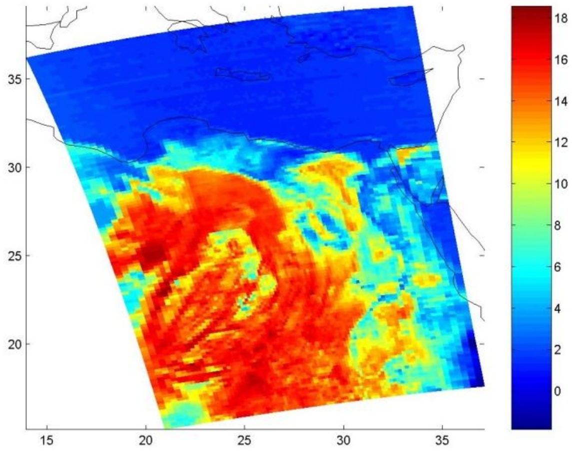
Inferring surface properties with AIRS high spectral resolution data
Barren region detection if T1086 < T981

Barren vs Water/Vegetated

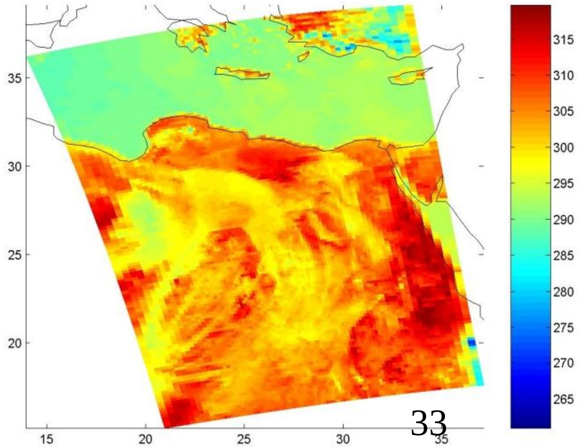


AIRS data from 14 June 2002

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

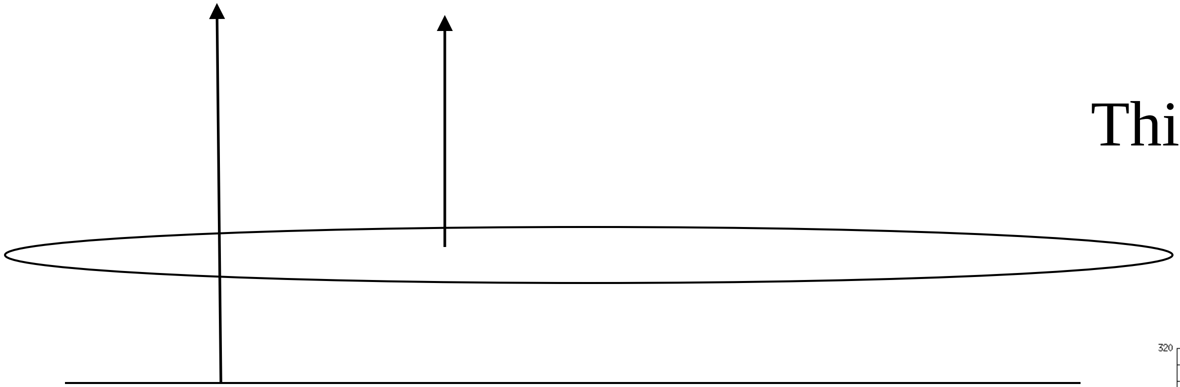


T(1086 cm^{-1})



from Tobin et al.

Thin ice cloud over ocean



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

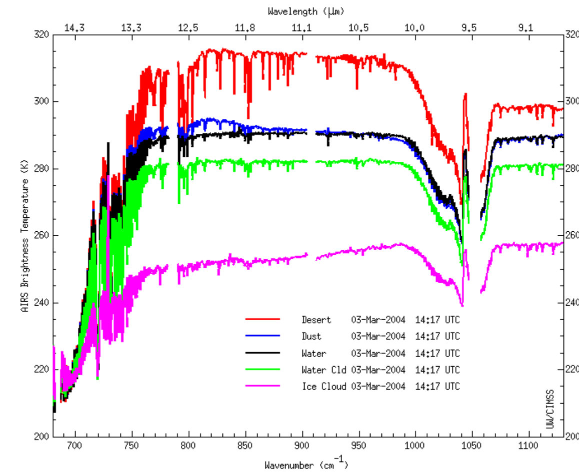
So difference of thin ice cloud over ocean
from clear sky over ocean is given by

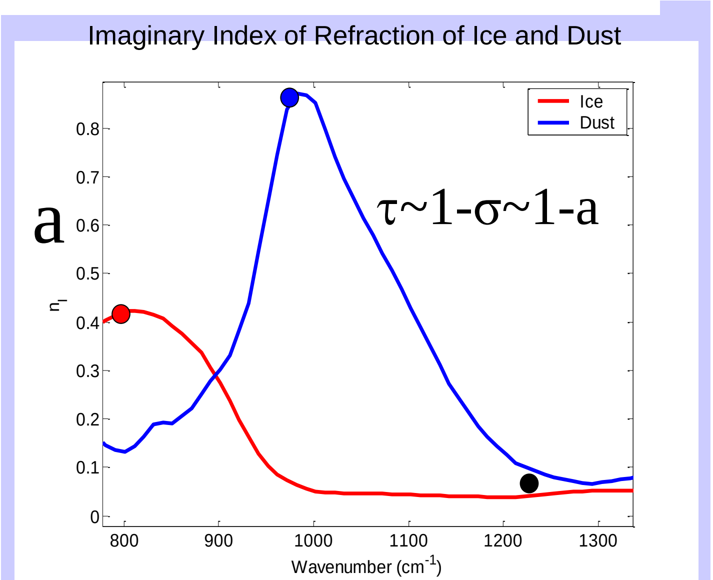
$\Delta R = - \boldsymbol{\varepsilon}_s \boldsymbol{\sigma}_c B_s + \boldsymbol{\sigma}_c B_c$

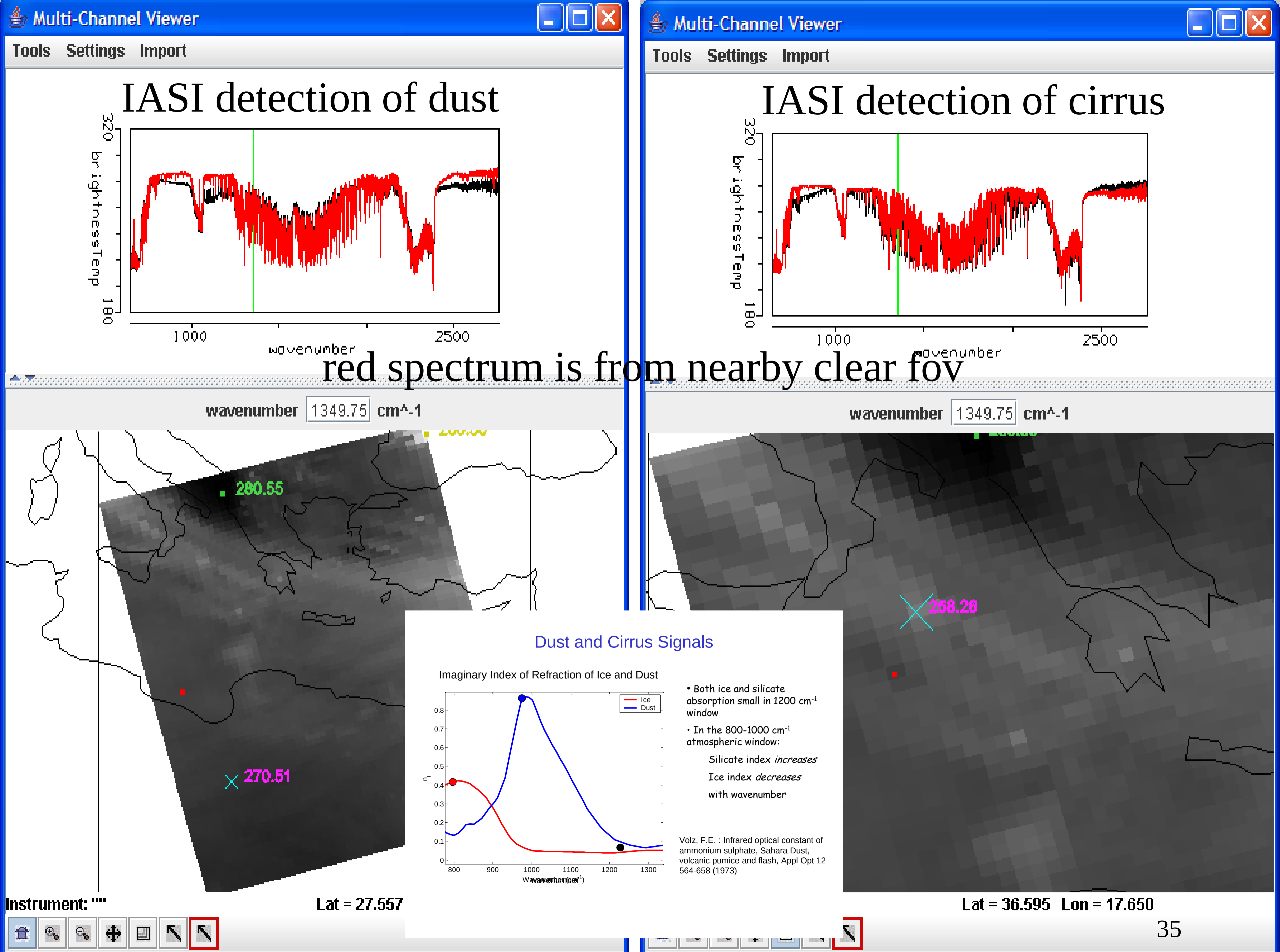
For $B_s > B_c$ and $\boldsymbol{\varepsilon}_s \sim 1$

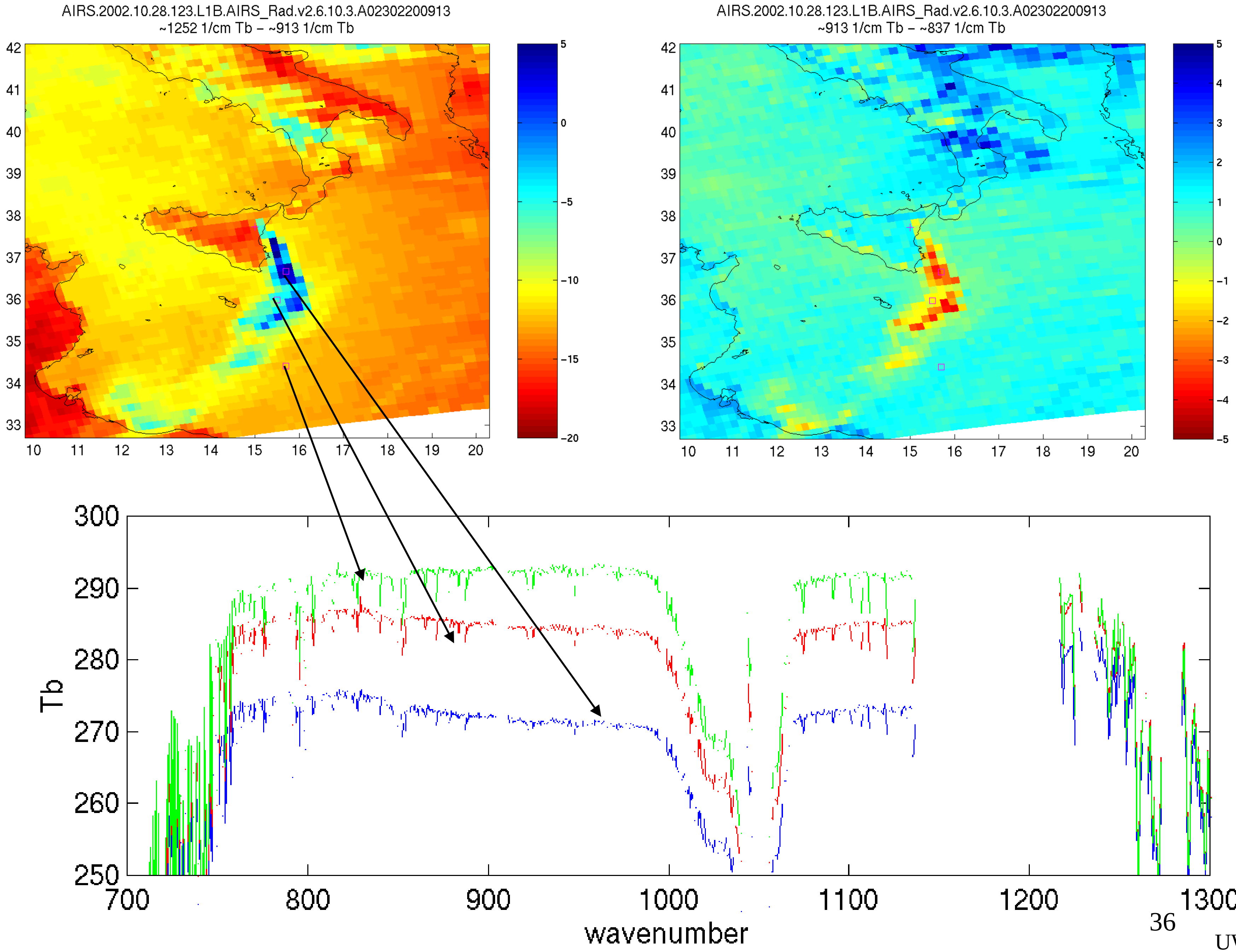
$\Delta R = - \boldsymbol{\sigma}_c B_s + \boldsymbol{\sigma}_c B_c = \boldsymbol{\sigma}_c [B_c - B_s]$

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





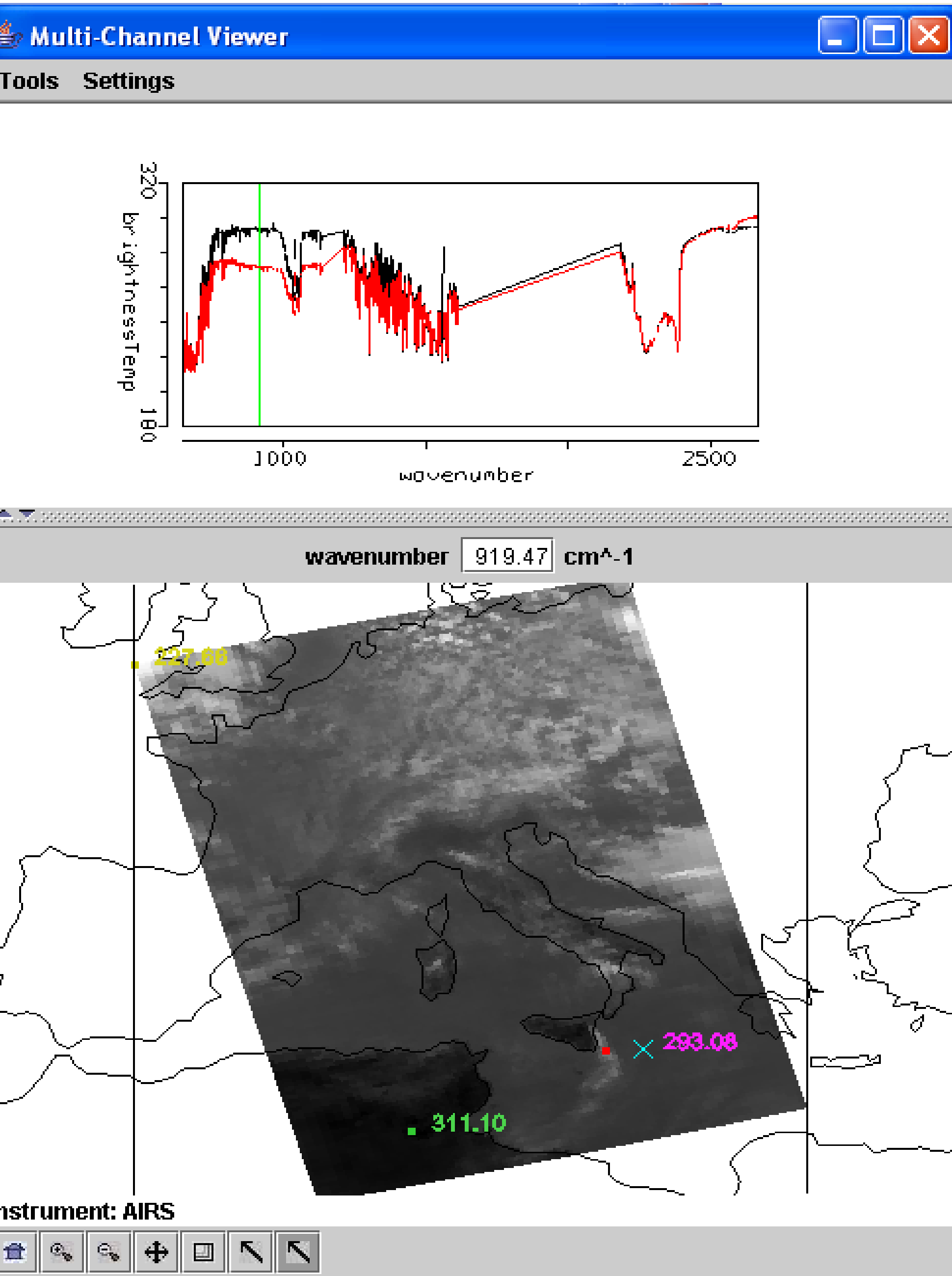

Multi-Channel Viewer
Tools Settings Import
IASI detection of dust
brightnessTemp
320
180
1000
2500
wavenumber
wavenumber 1349.75 cm^-1
280.55
270.51
Instrument: ""
Lat = 27.557
Multi-Channel Viewer
Tools Settings Import
IASI detection of cirrus
brightnessTemp
320
180
1000
2500
wavenumber
wavenumber 1349.75 cm^-1
258.28
Lat = 36.595 Lon = 17.650
red spectrum is from nearby clear fov
Dust and Cirrus Signals
Imaginary Index of Refraction of Ice and Dust
Ice
Dust
• Both ice and silicate absorption small in 1200 cm-1 window
• In the 800-1000 cm-1 atmospheric window:
Silicate index *increases*
Ice index *decreases*
with wavenumber
Volz, F.E. : Infrared optical constant of ammonium sulphate, Sahara Dust, volcanic pumice and flash, Appl Opt 12 564-658 (1973)

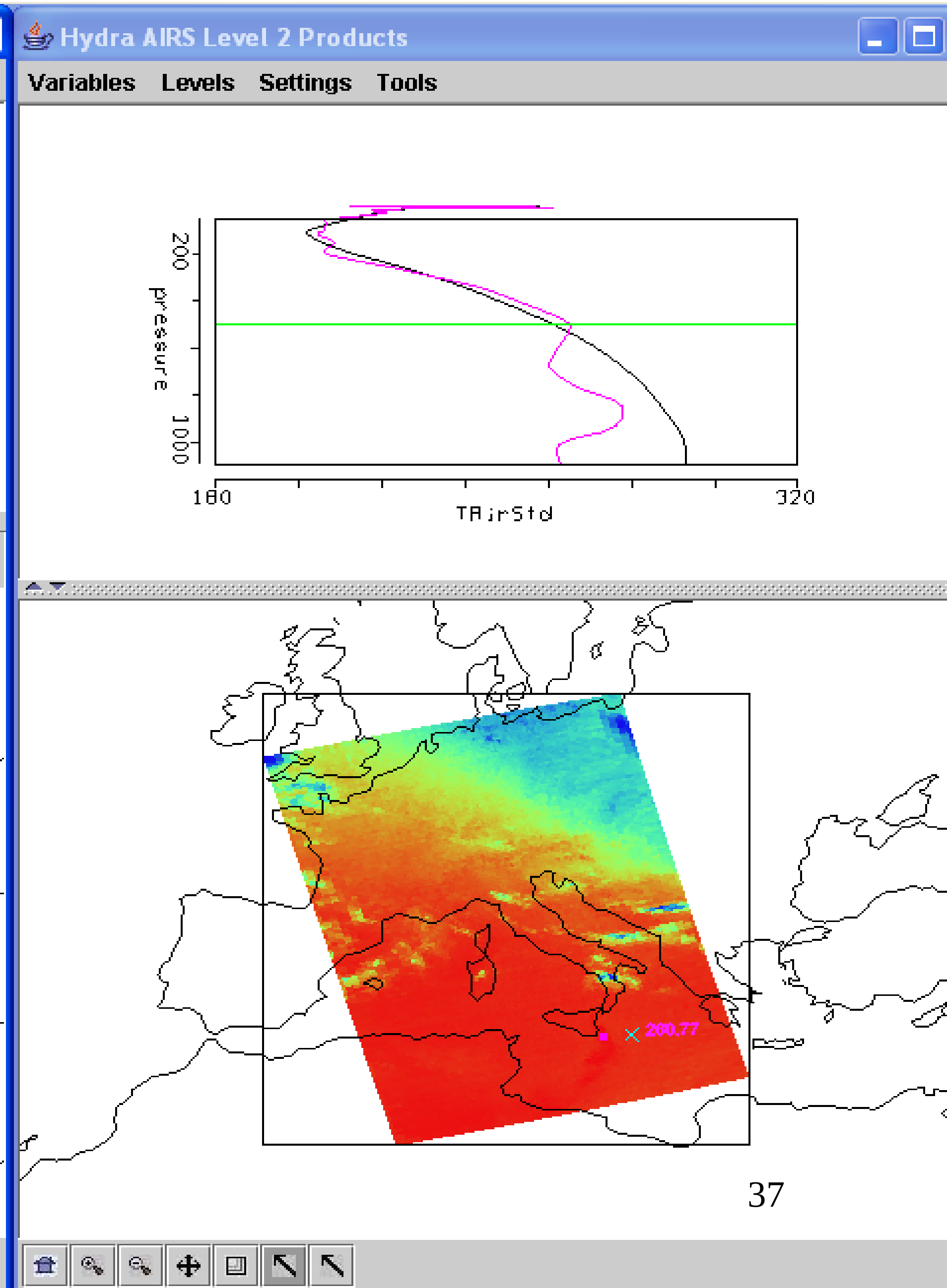


AIRS.2002.10.28.123.L1B.AIRS_Rad.v2.6.10.3.A02302200913
~1252 1/cm Tb – ~913 1/cm Tb
AIRS.2002.10.28.123.L1B.AIRS_Rad.v2.6.10.3.A02302200913
~913 1/cm Tb – ~837 1/cm Tb
Tb
wavenumber

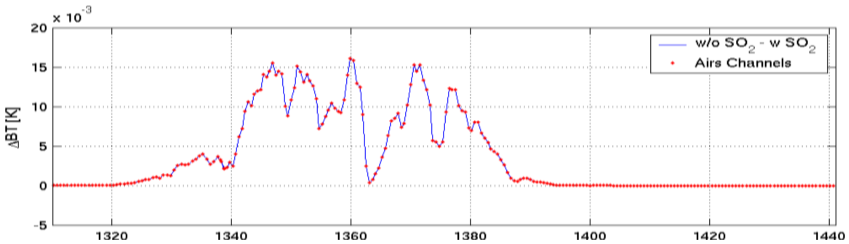
UW
CIMSS

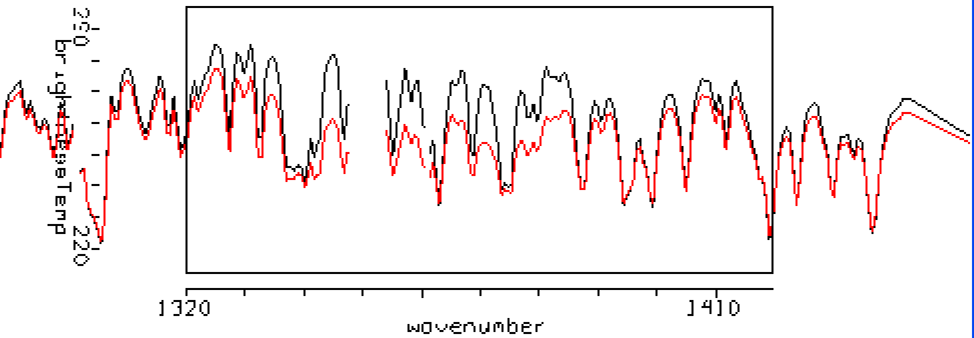
Mt Etna Ash cloud at 500 hPa

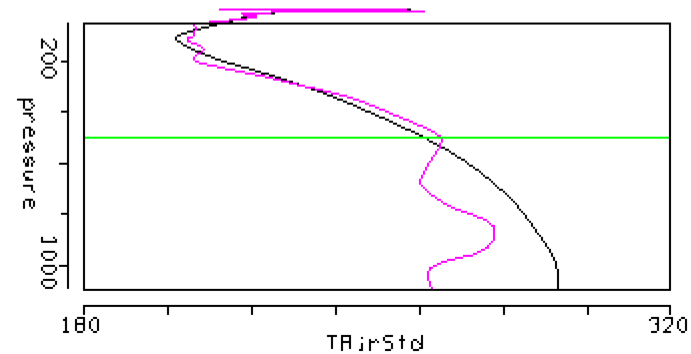


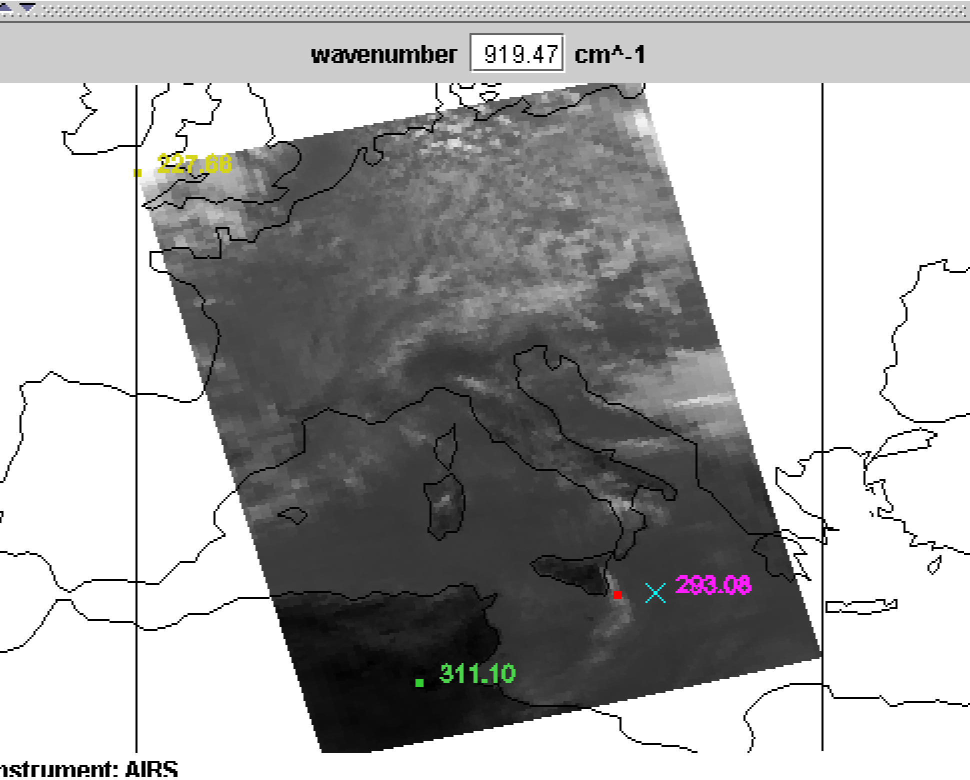


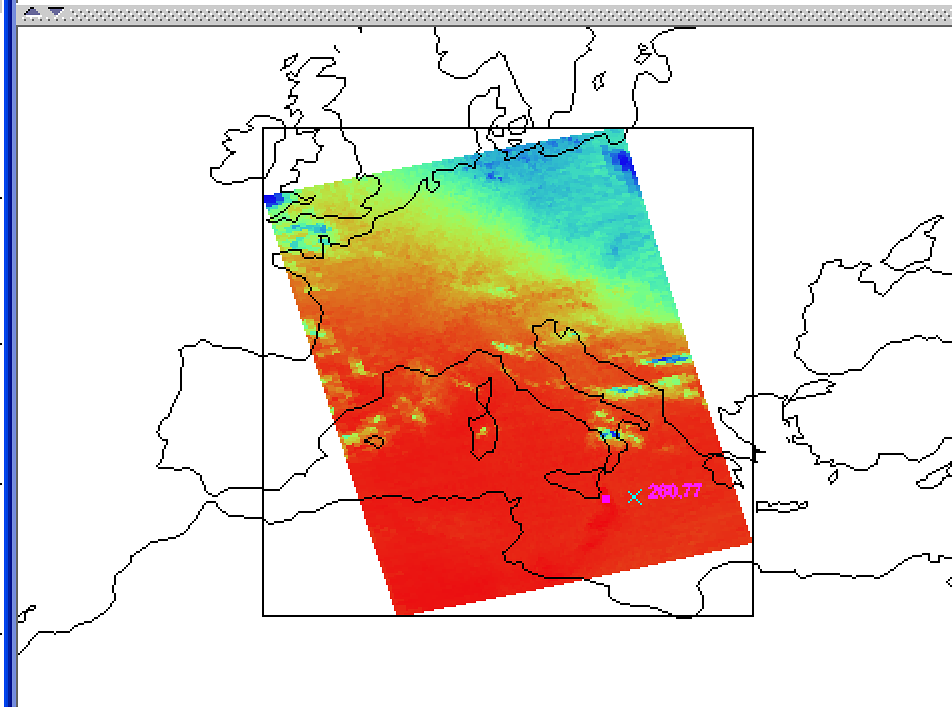
Ash cloud and clear sky spectra

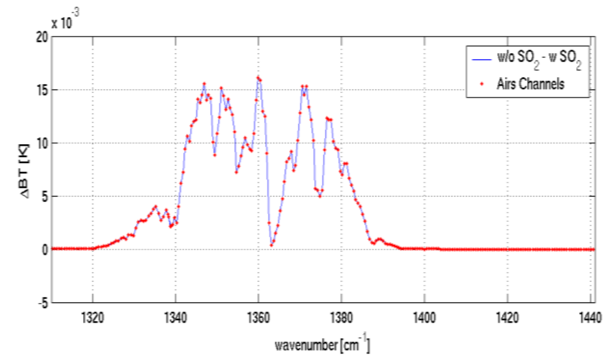




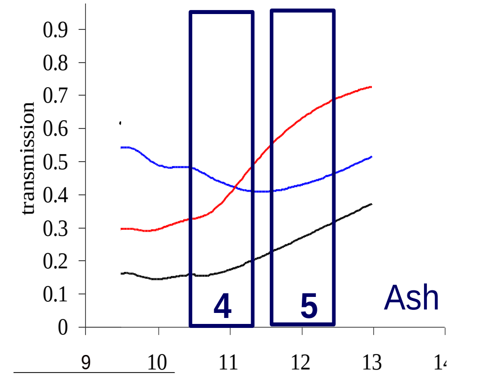


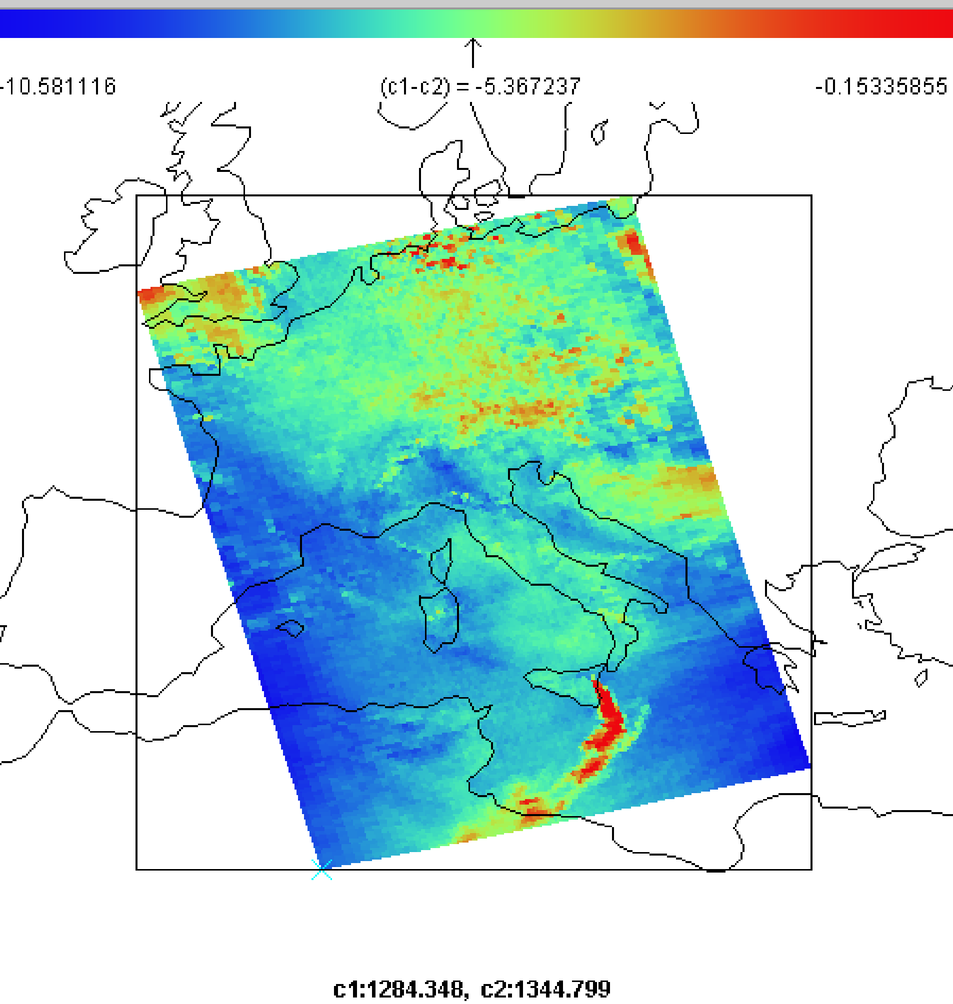


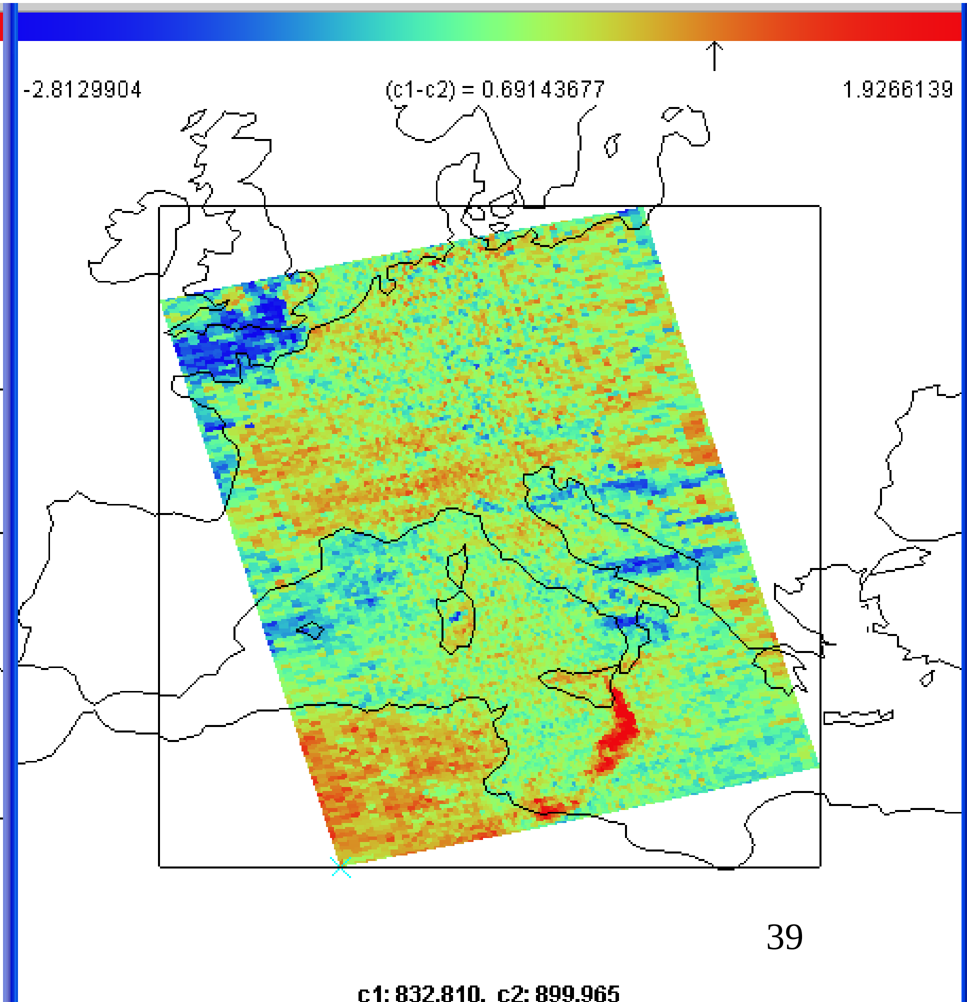




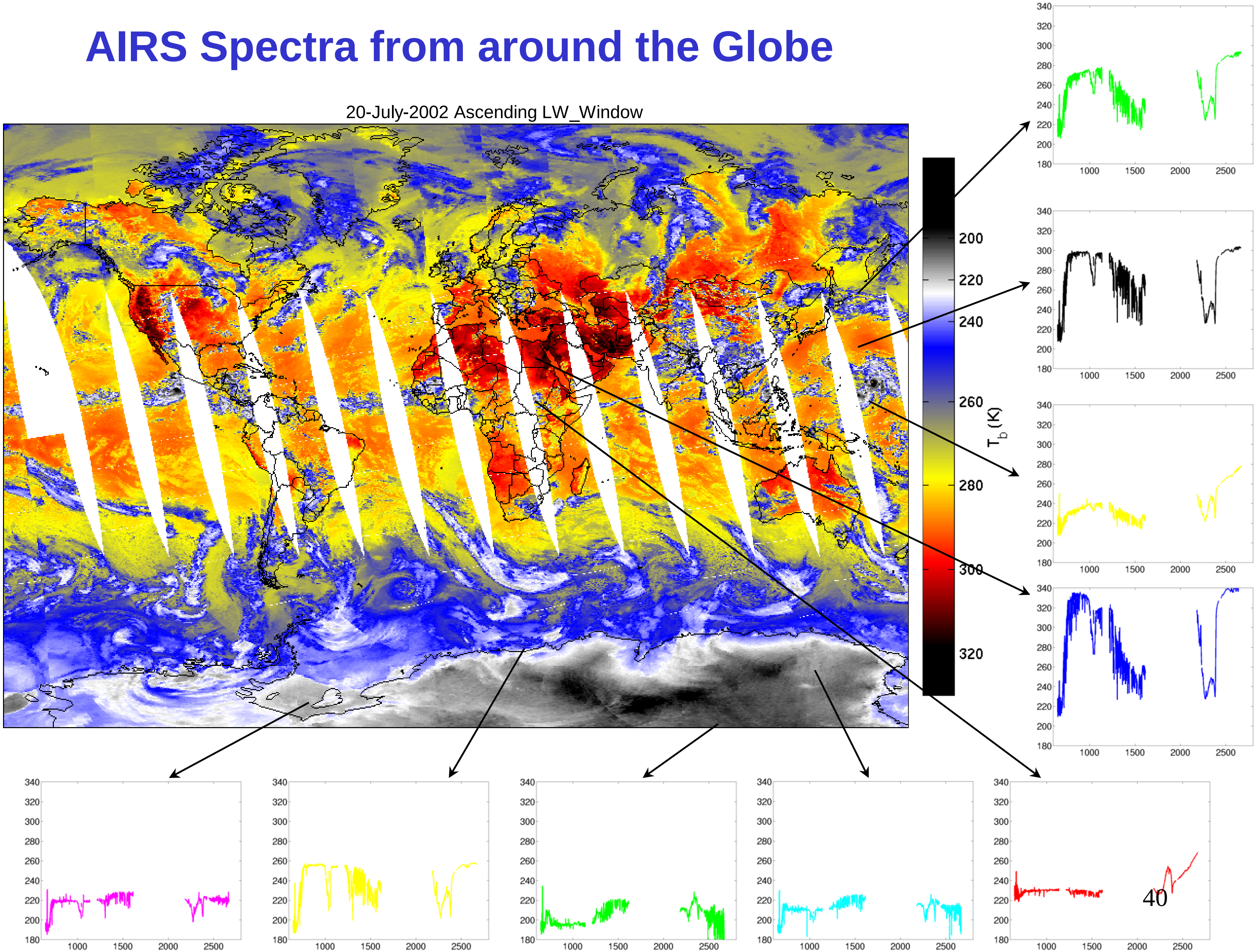
Mt Etna volcanic plume
SO2 (left) from 1284-1345
Ash (right) from 832-900







AIRS Spectra from around the Globe



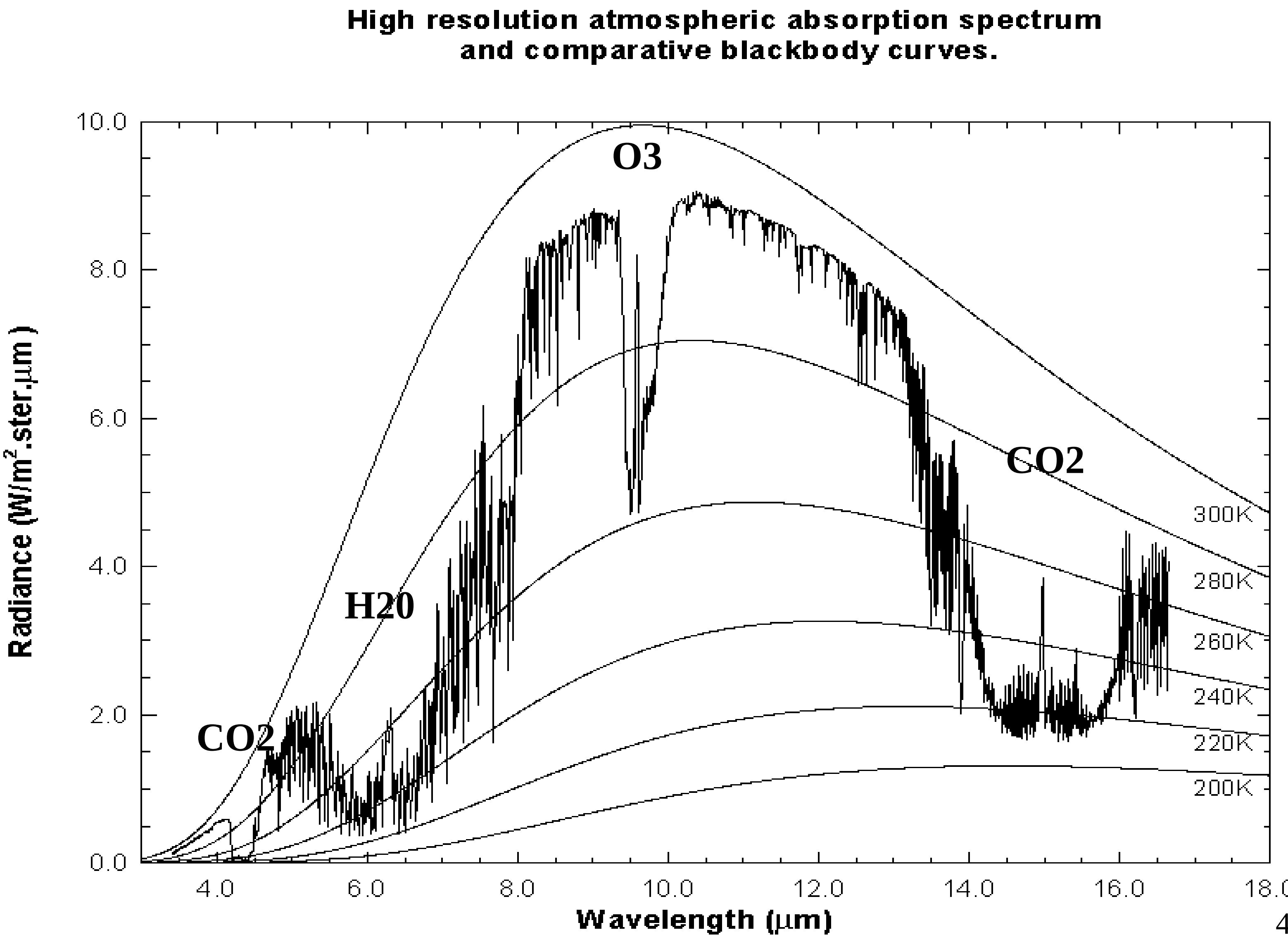
Intro to Microwave and Split Window Moisture

Lectures in Madison

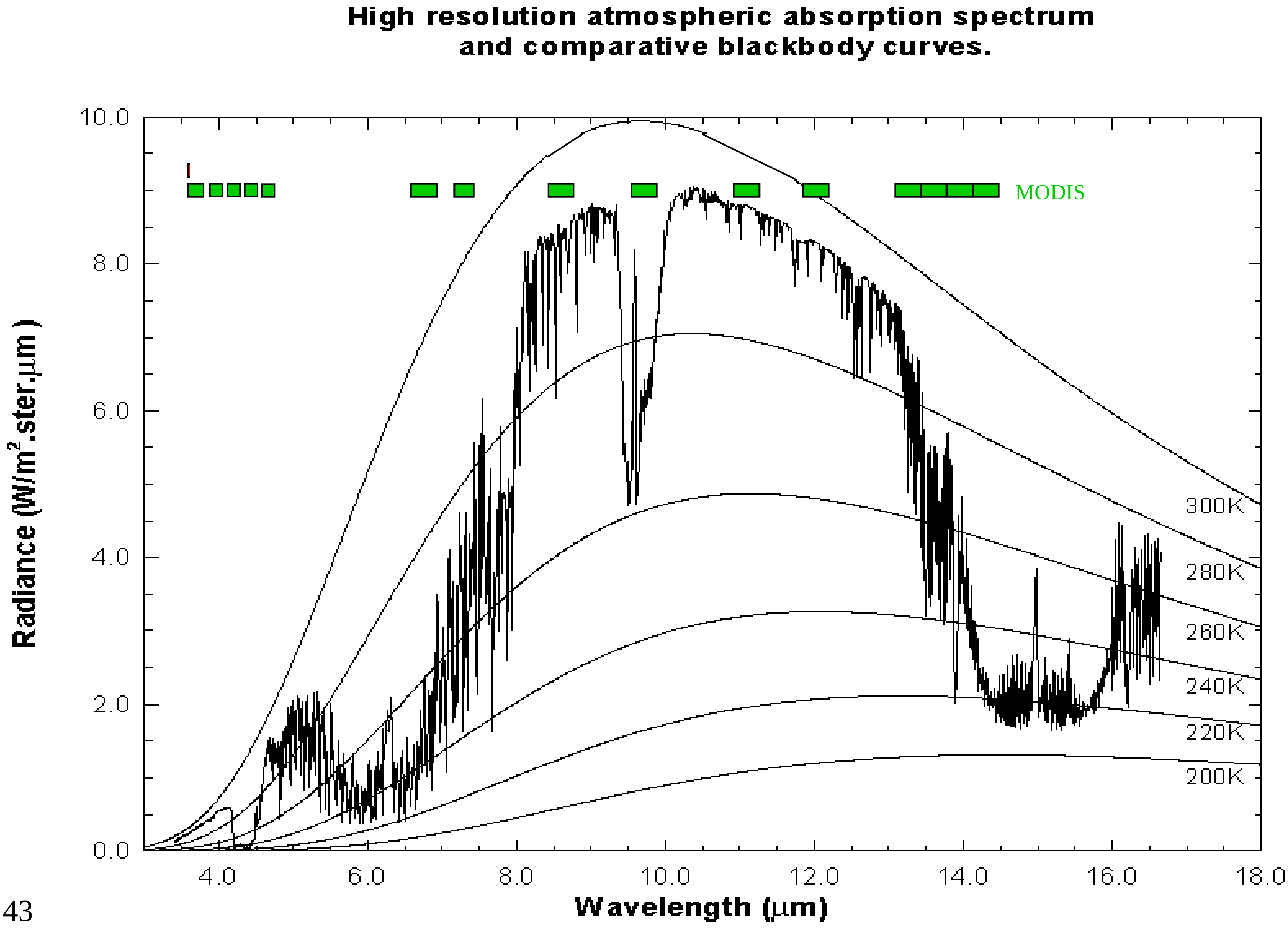
27 March 2013

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

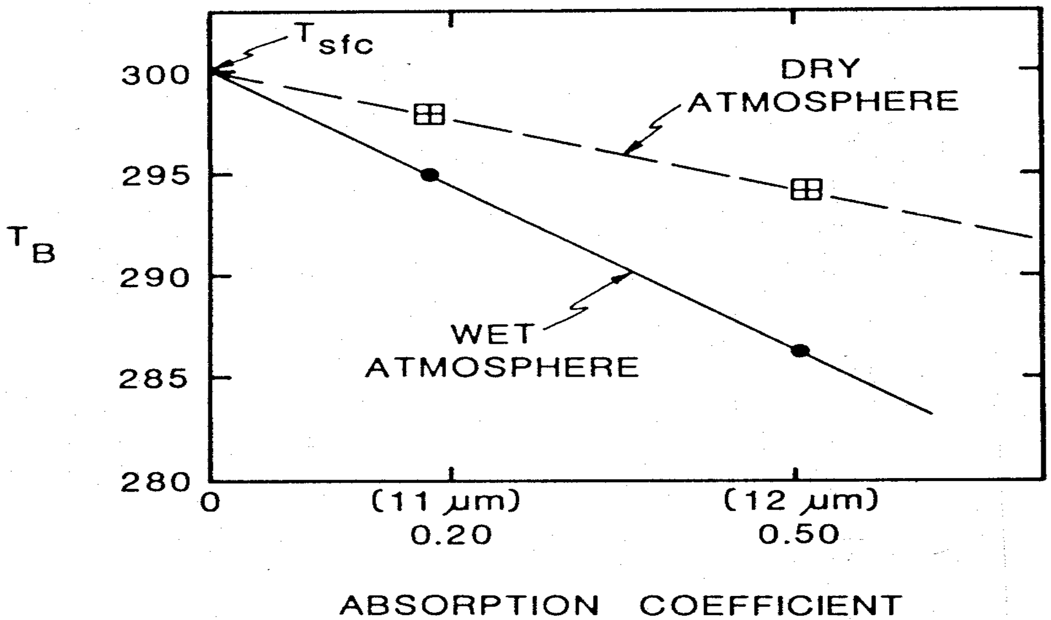
Moisture attenuation in atmospheric windows varies linearly with optical depth.

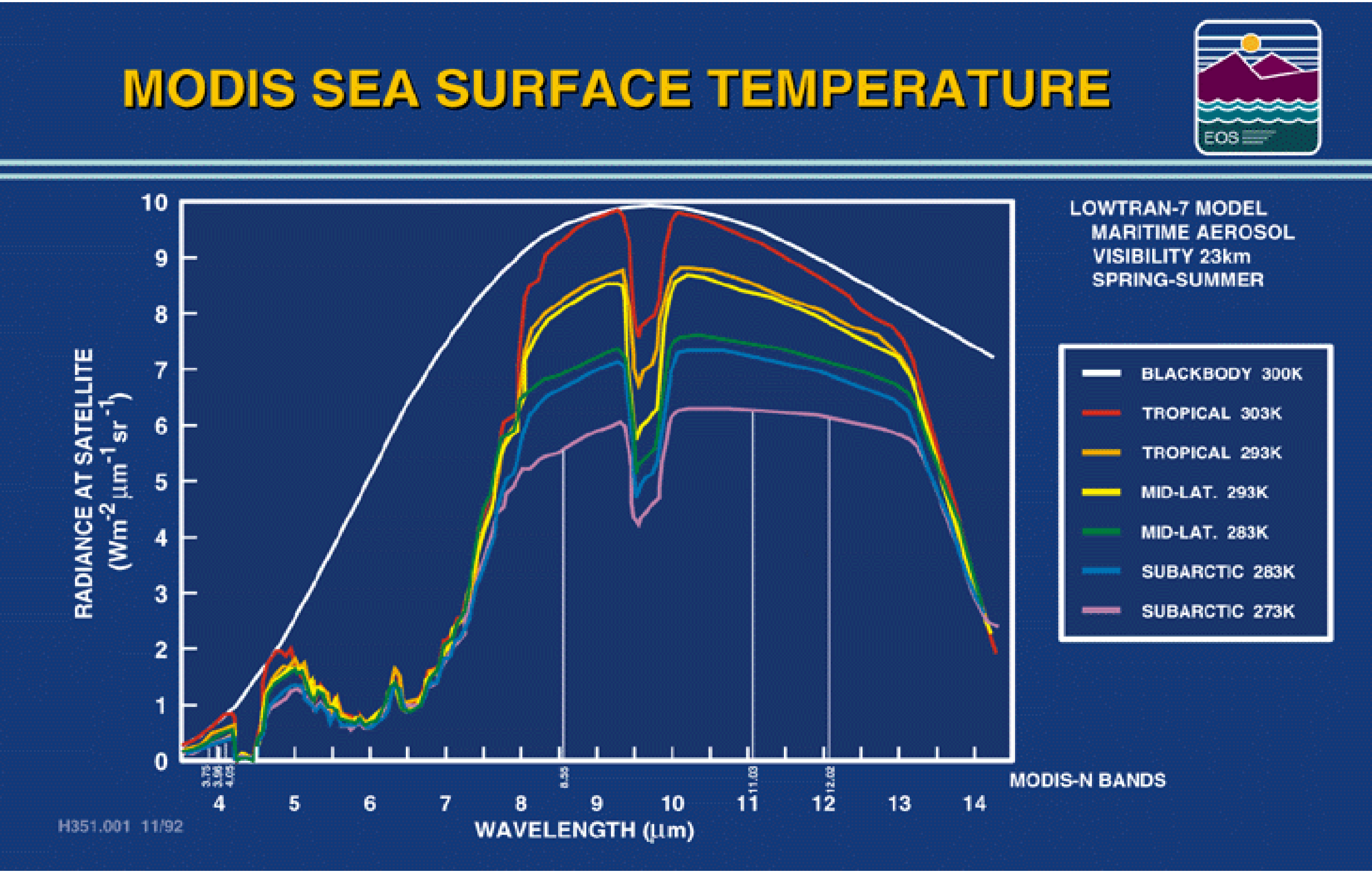
$$\tau_\lambda = e^{-k_\lambda u} = 1 - k_\lambda u$$

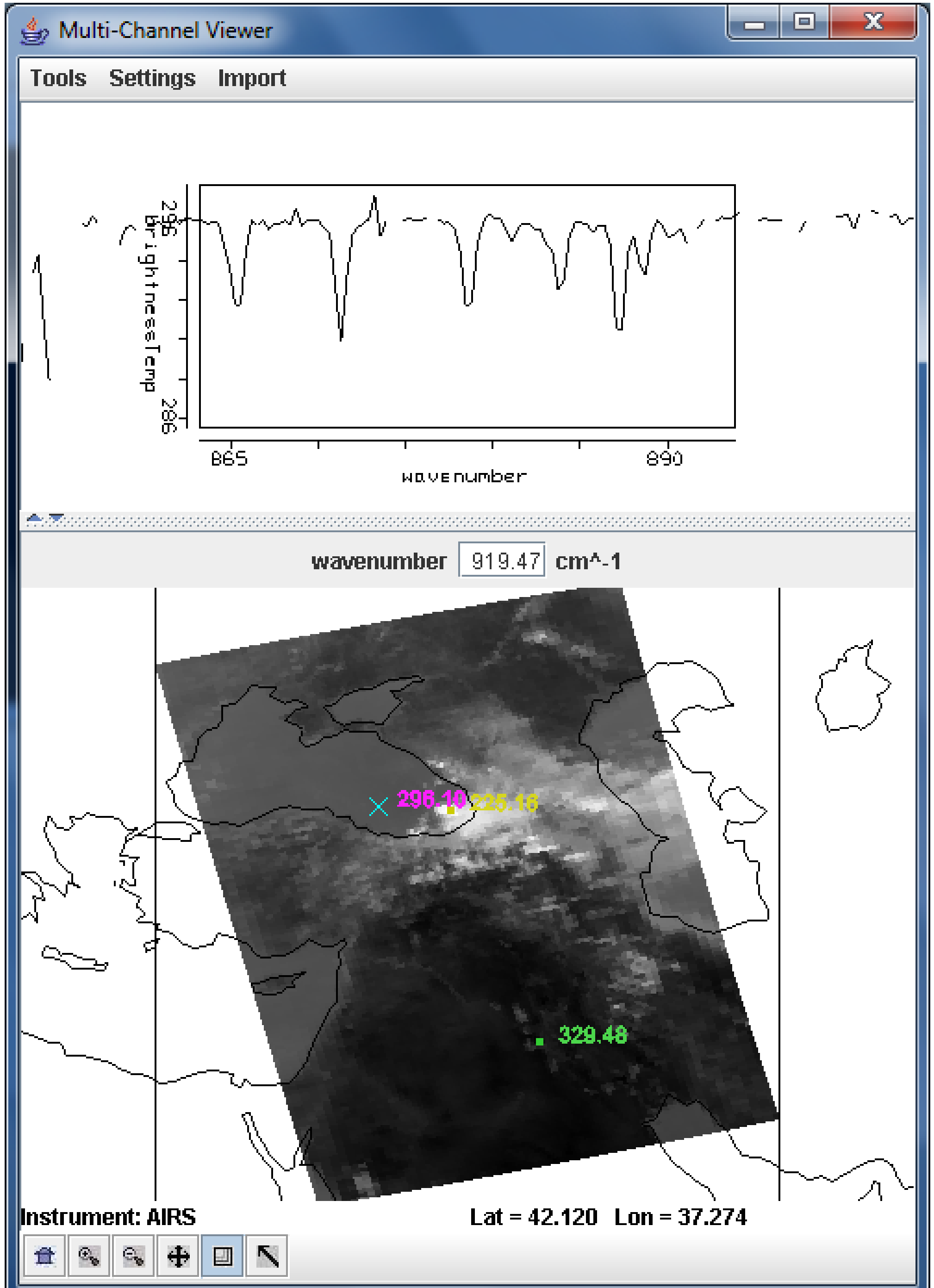
For same atmosphere, deviation of brightness temperature from surface temperature is a linear function of absorbing power. Thus moisture corrected SST can inferred by using split window measurements and extrapolating to zero k_λ

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

Moisture content of atmosphere inferred from slope of linear relation.

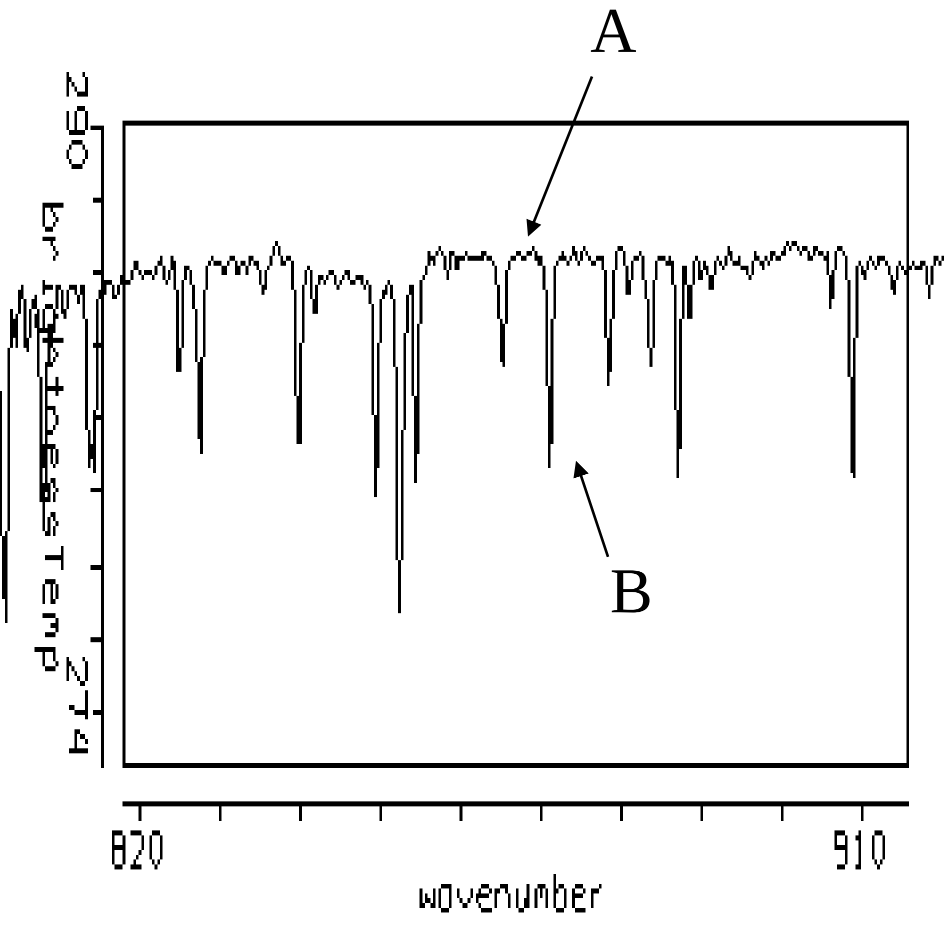



MODIS SEA SURFACE TEMPERATURE
EOS
LOWTRAN-7 MODEL
MARITIME AEROSOL
VISIBILITY 23km
SPRING-SUMMER
BLACKBODY 300K
TROPICAL 303K
TROPICAL 293K
MID-LAT. 293K
MID-LAT. 283K
SUBARCTIC 283K
SUBARCTIC 273K
RADIANCE AT SATELLITE
(Wm-2 µm-1 sr-1)
0
1
2
3
4
5
6
7
8
9
10
MODIS-N BANDS
4
5
6
7
8
9
10
11
12
13
14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

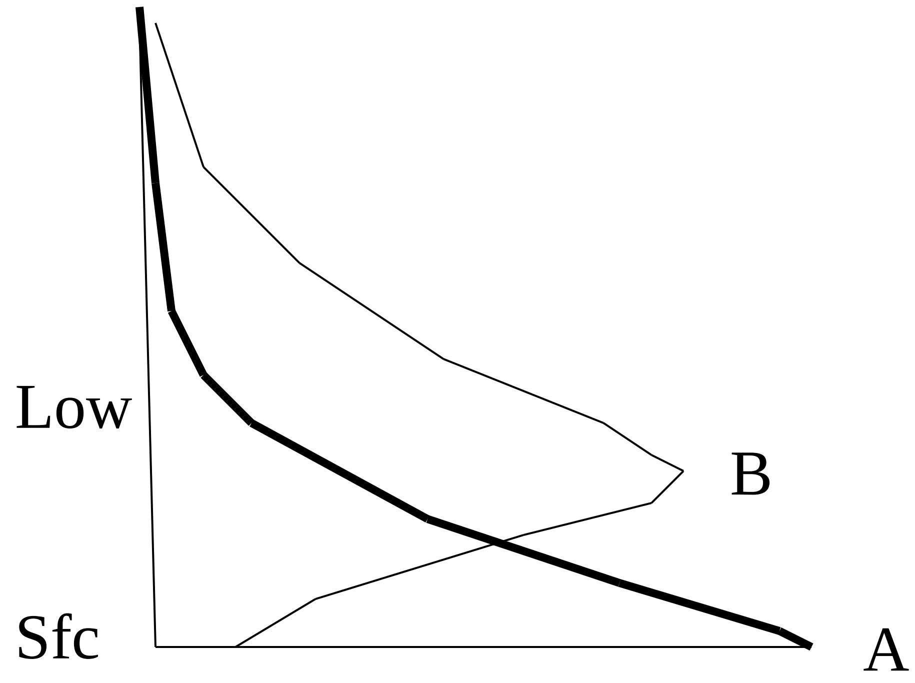


Multi-Channel Viewer
Tools Settings Import
296
BrightnessTemp
286
865
890
wavenumber
wavenumber 919.47 cm^-1
296.10
225.16
329.48
Instrument: AIRS
Lat = 42.120 Lon = 37.274

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

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

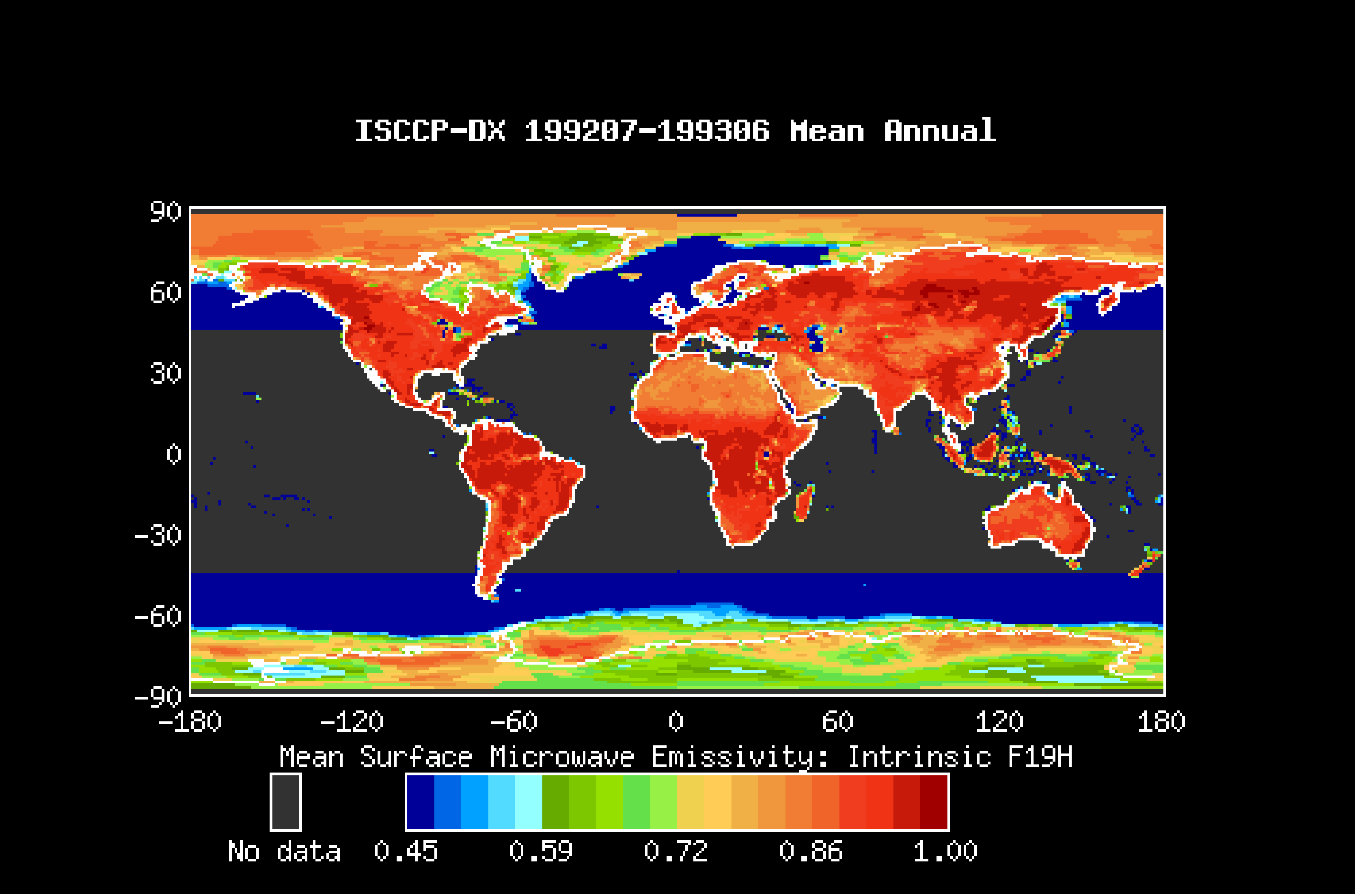
In microwave region $c_2/\lambda T << 1$ so that

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

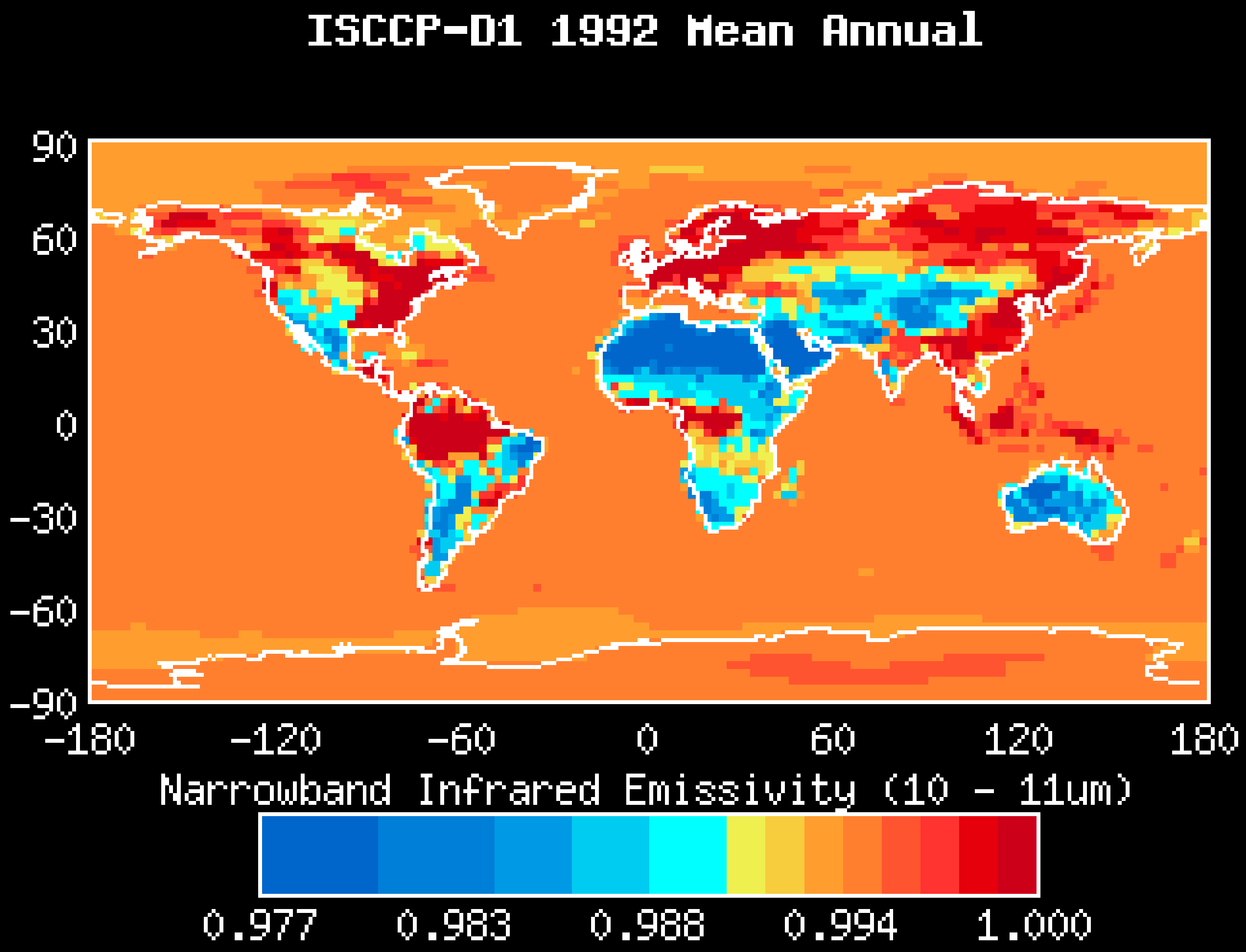
And classical Rayleigh Jeans radiation equation emerges

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

Radiance is linear function of brightness temperature.

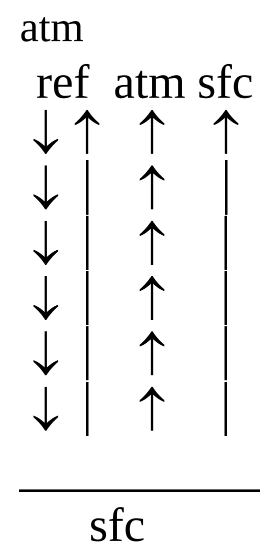

ISCCP-DX 199207-199306 Mean Annual
90
60
30
0
-30
-60
-90
-180
-120
-60
0
60
120
180
Mean Surface Microwave Emissivity: Intrinsic F19H
No data
0.45
0.59
0.72
0.86
1.00

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

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

Transmittance

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

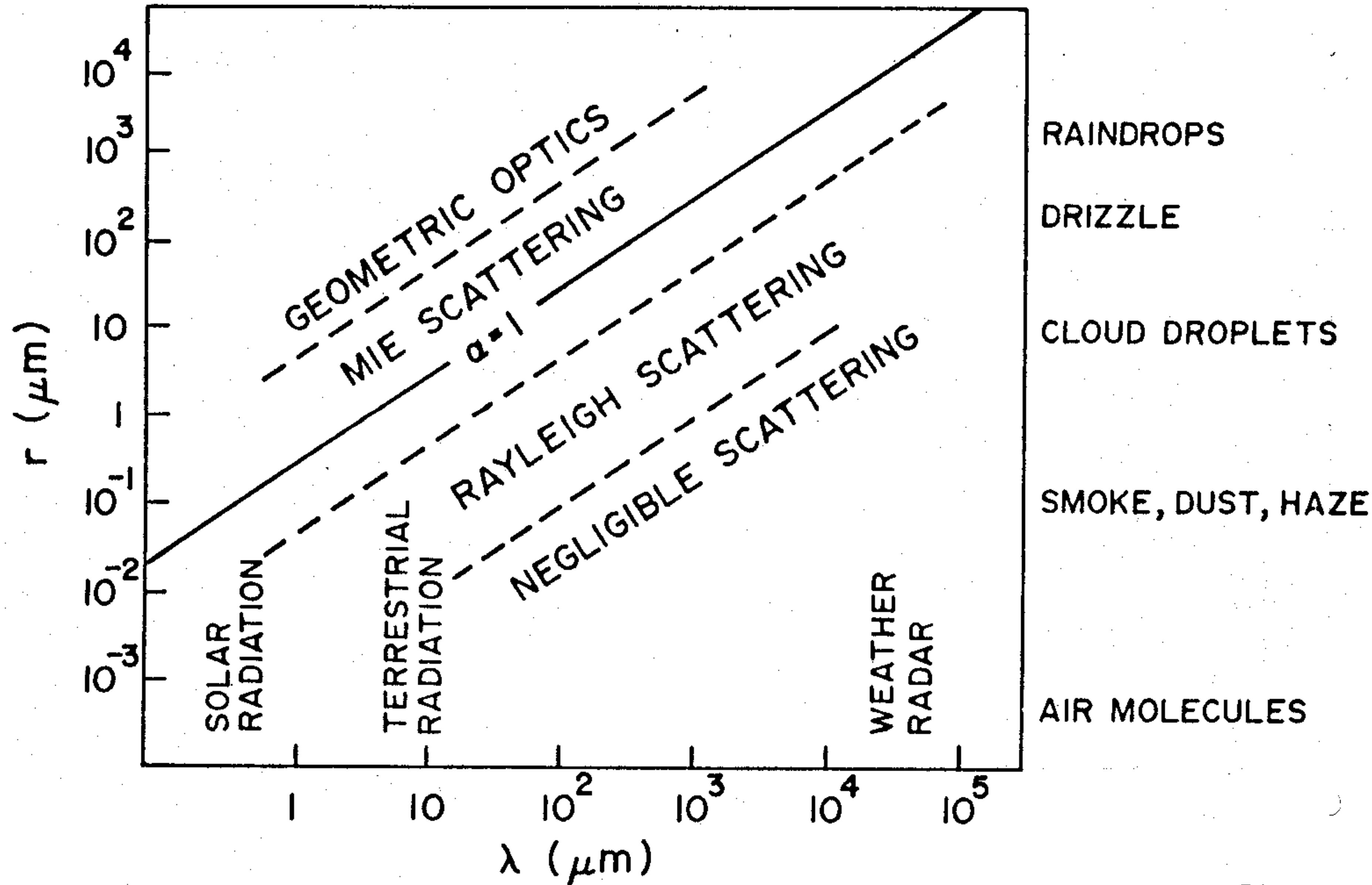
Thus downwelling in terms of upwelling can be written

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

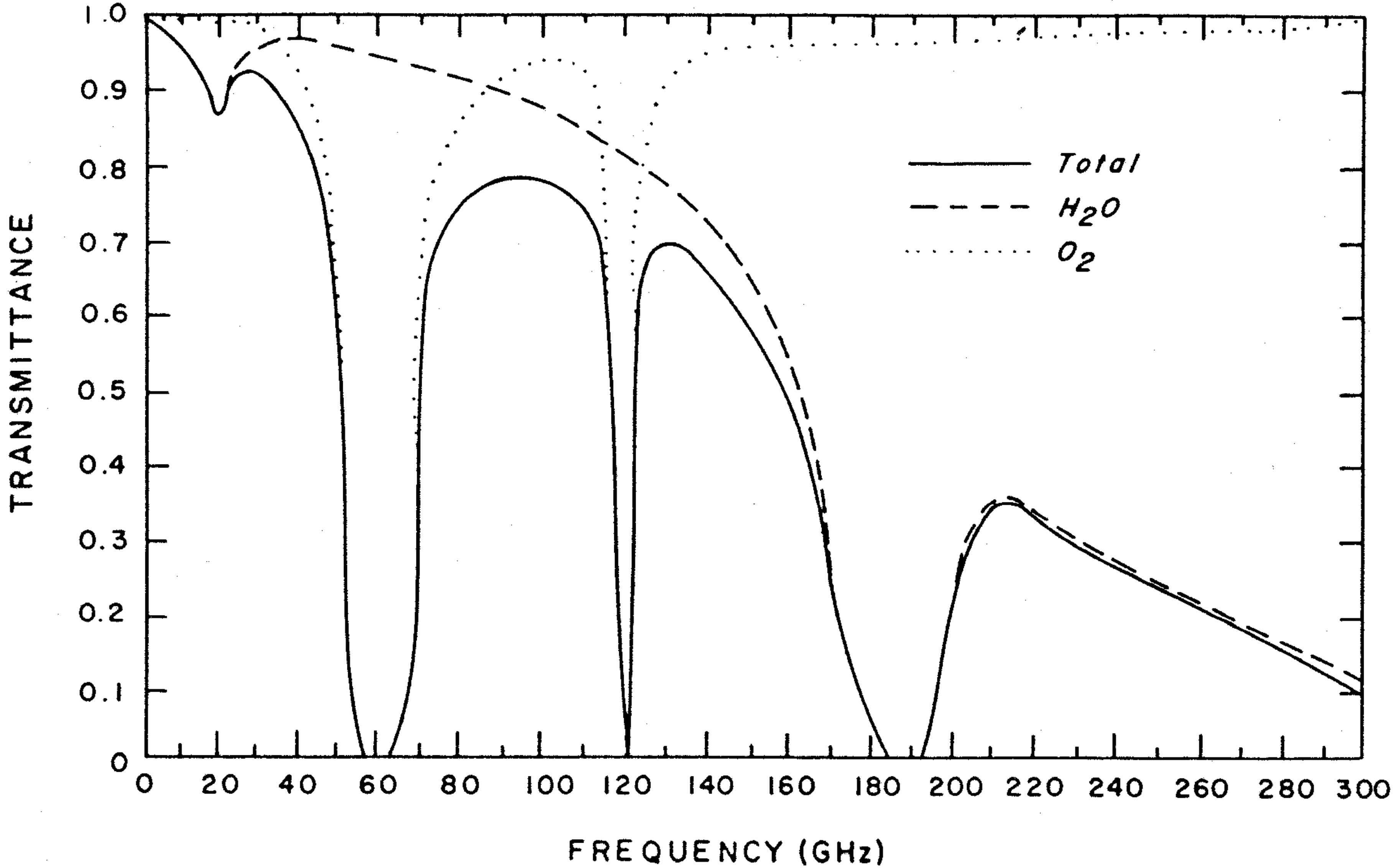
and

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

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



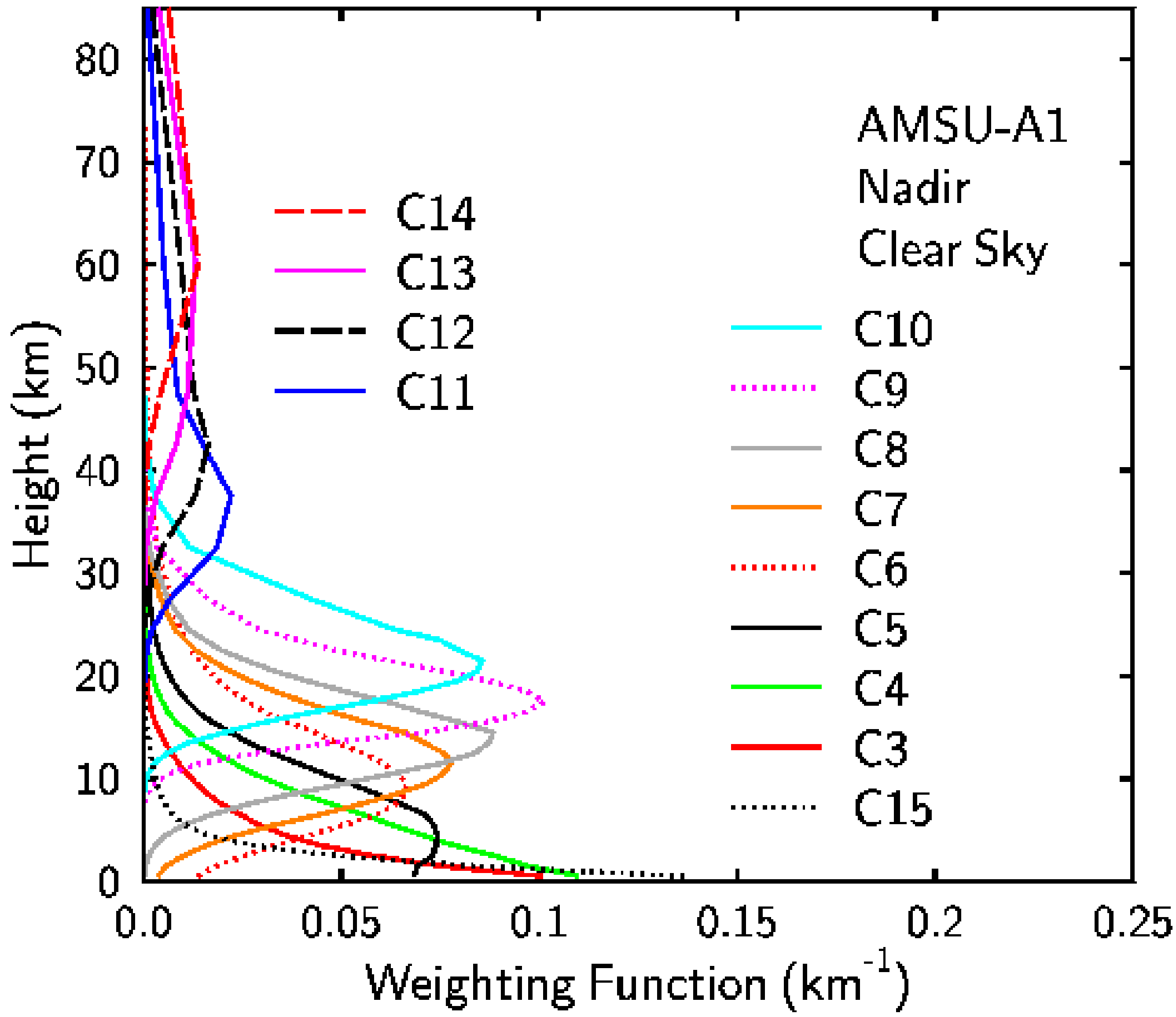
GEOMETRIC OPTICS
MIE SCATTERING
α=1
RAYLEIGH SCATTERING
NEGLIGIBLE SCATTERING
SOLAR RADIATION
TERRESTRIAL RADIATION
WEATHER RADAR
RAINDROPS
DRIZZLE
CLOUD DROPLETS
SMOKE, DUST, HAZE
AIR MOLECULES
r (μm)
λ (μm)
10^4
10^3
10^2
10
1
10^-1
10^-2
10^-3
1
10
10^2
10^3
10^4
10^5



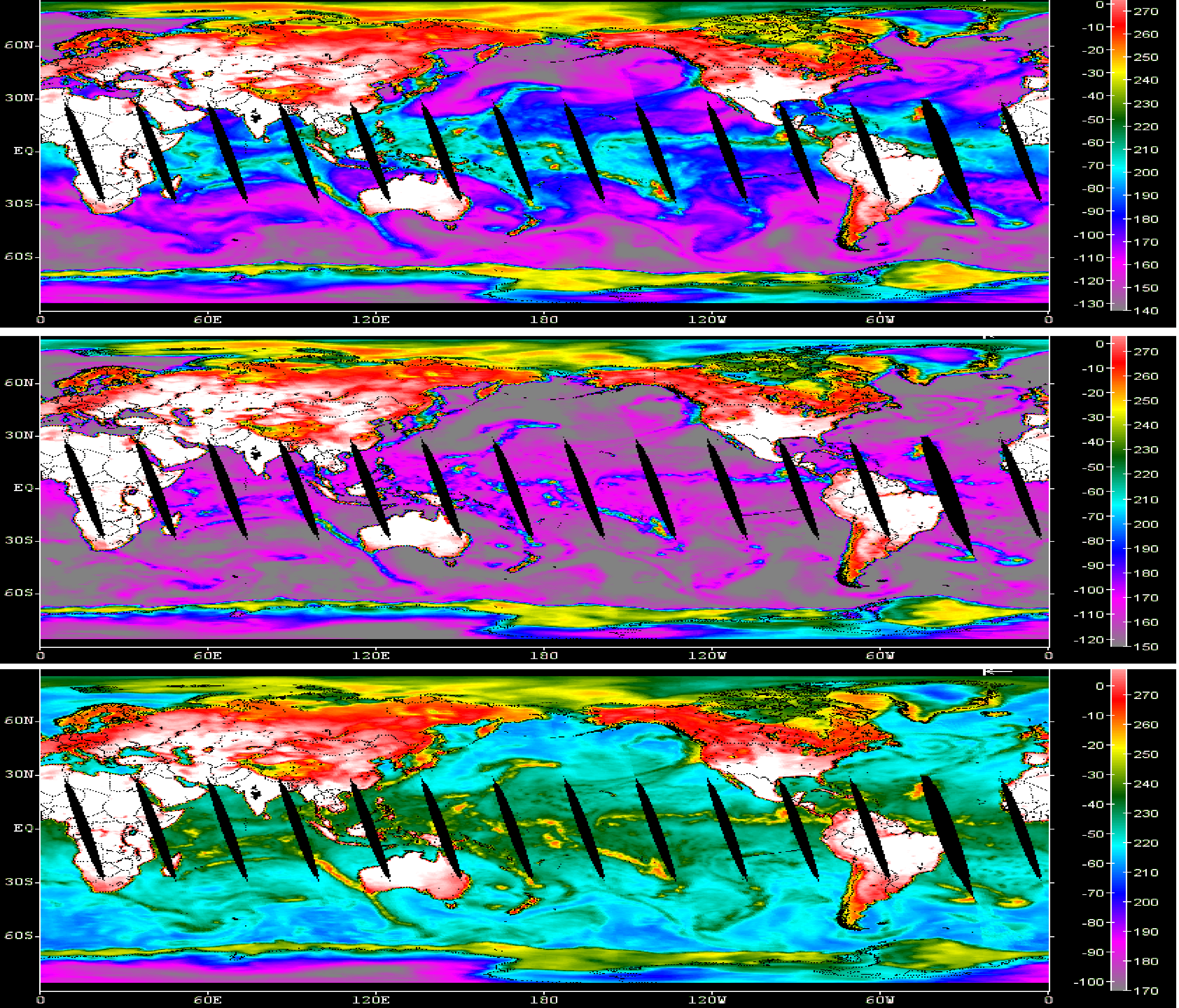
TRANSMITTANCE
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0
20
40
60
80
100
120
140
160
180
200
220
240
260
280
300
FREQUENCY (GHz)
Total
H_2O
O_2

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



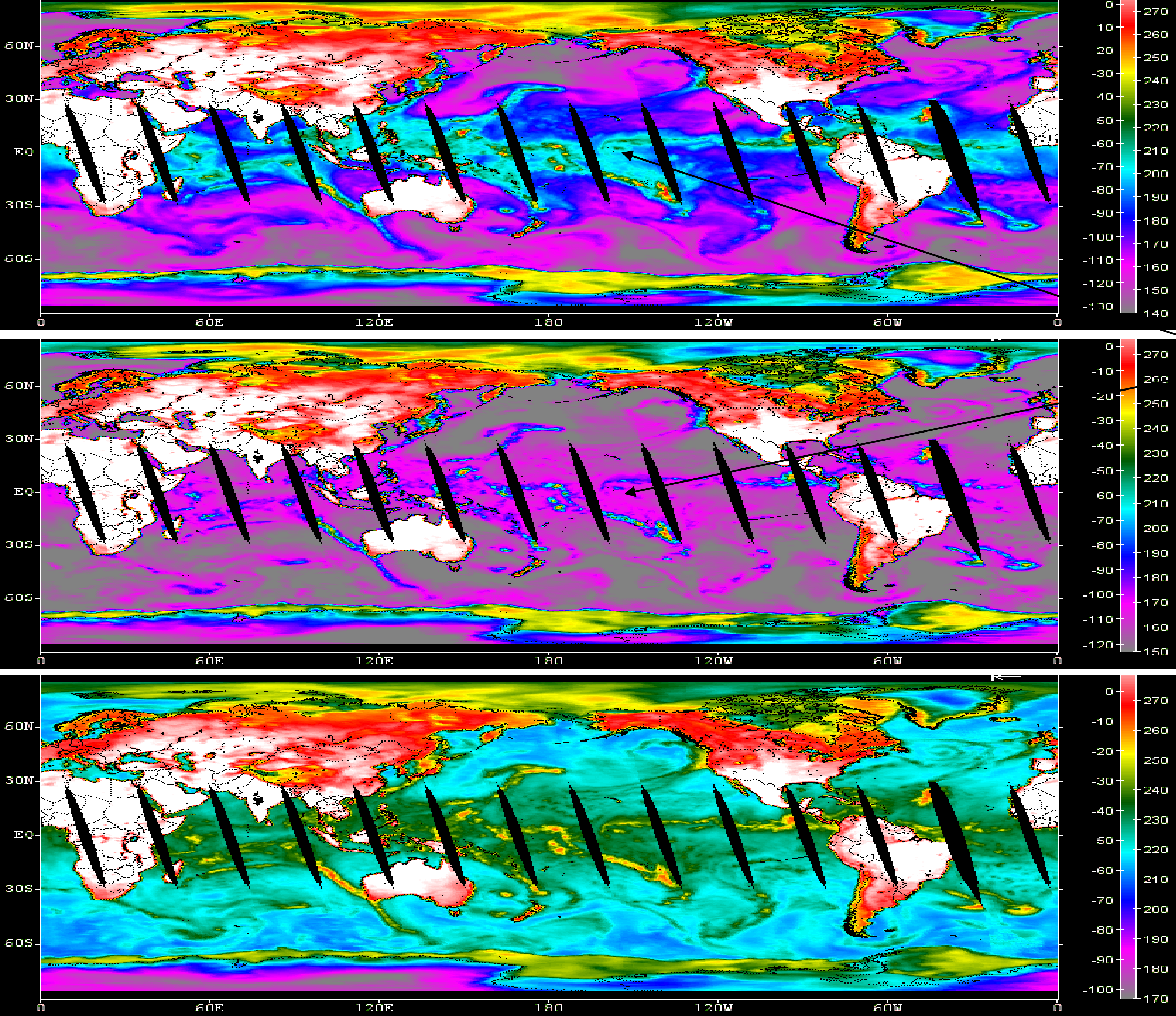
AMSU

23.8

31.4

50.3 GHz

AMSU

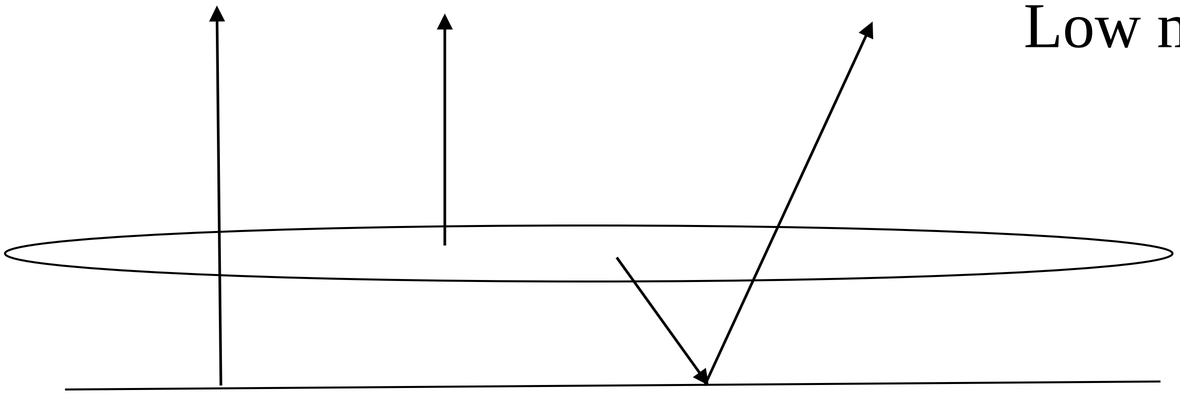


23.8 dirty window

atm Q warms BT

31.4 window

50.3 GHz



Low mist over ocean (MW)

τ_m , ε_m

r_s , ε_s

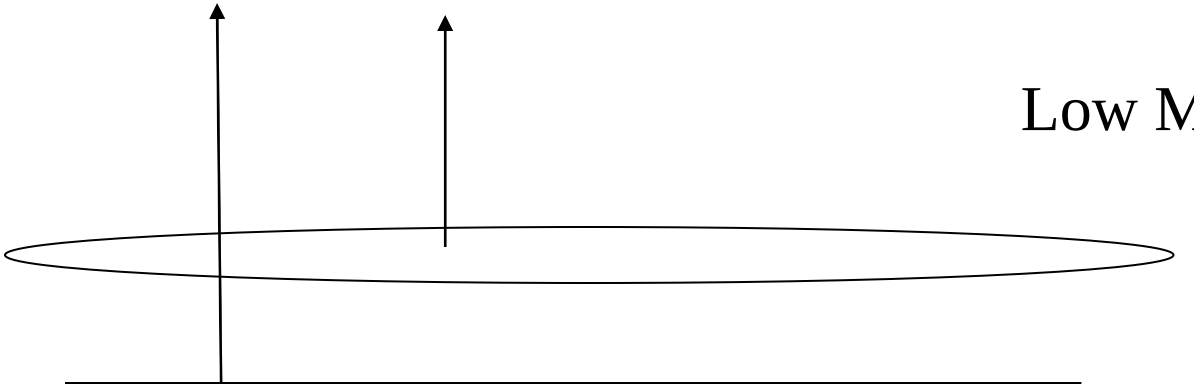
$T_b = \varepsilon_s T_s \tau_m + \varepsilon_m T_m + \varepsilon_m r_s \tau_m T_m$

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

$R = \boldsymbol{\varepsilon}_s B_s (1-\boldsymbol{\sigma}_m) + \boldsymbol{\sigma}_m B_m$ using $e^{-\sigma} = 1 - \sigma$ and $\tau \sim 1-\sigma \sim 1-a$

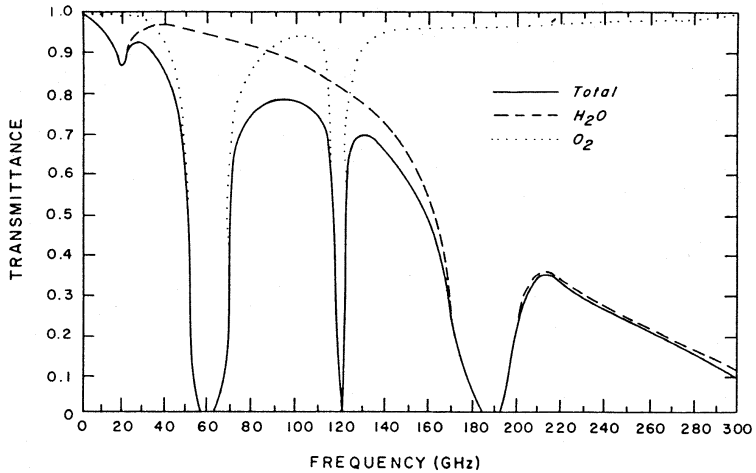
So difference of low mist over ocean
from clear sky over ocean is given by

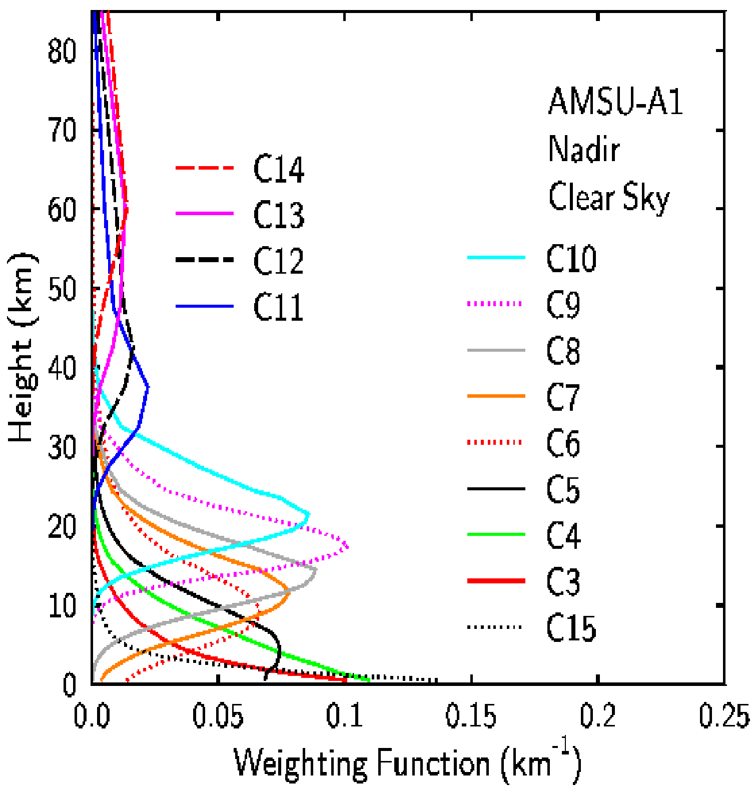
$\Delta R = - \boldsymbol{\varepsilon}_s \boldsymbol{\sigma}_m B_s + \boldsymbol{\sigma}_m B_m$

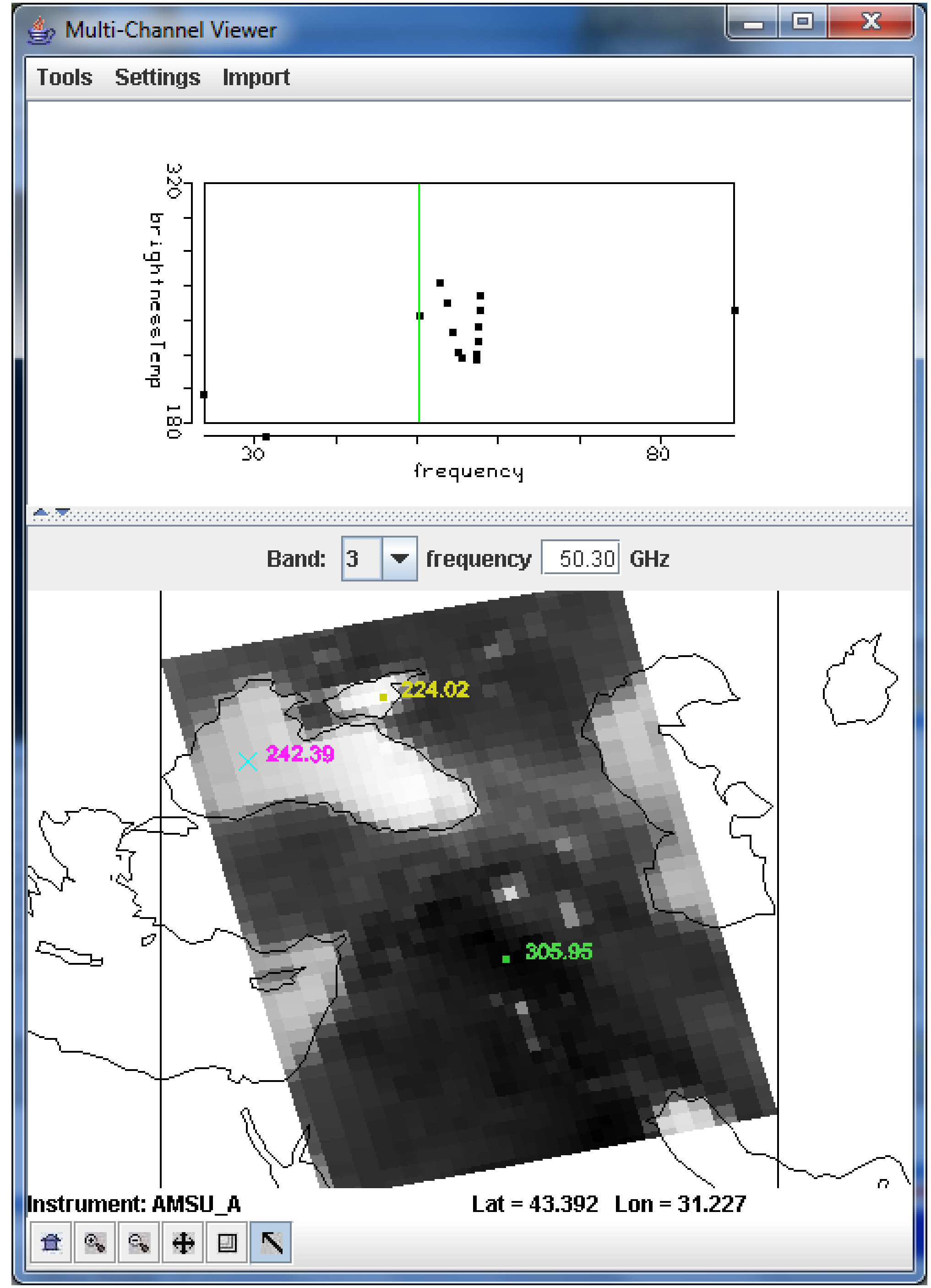
For $\boldsymbol{\varepsilon}_s \sim 1$

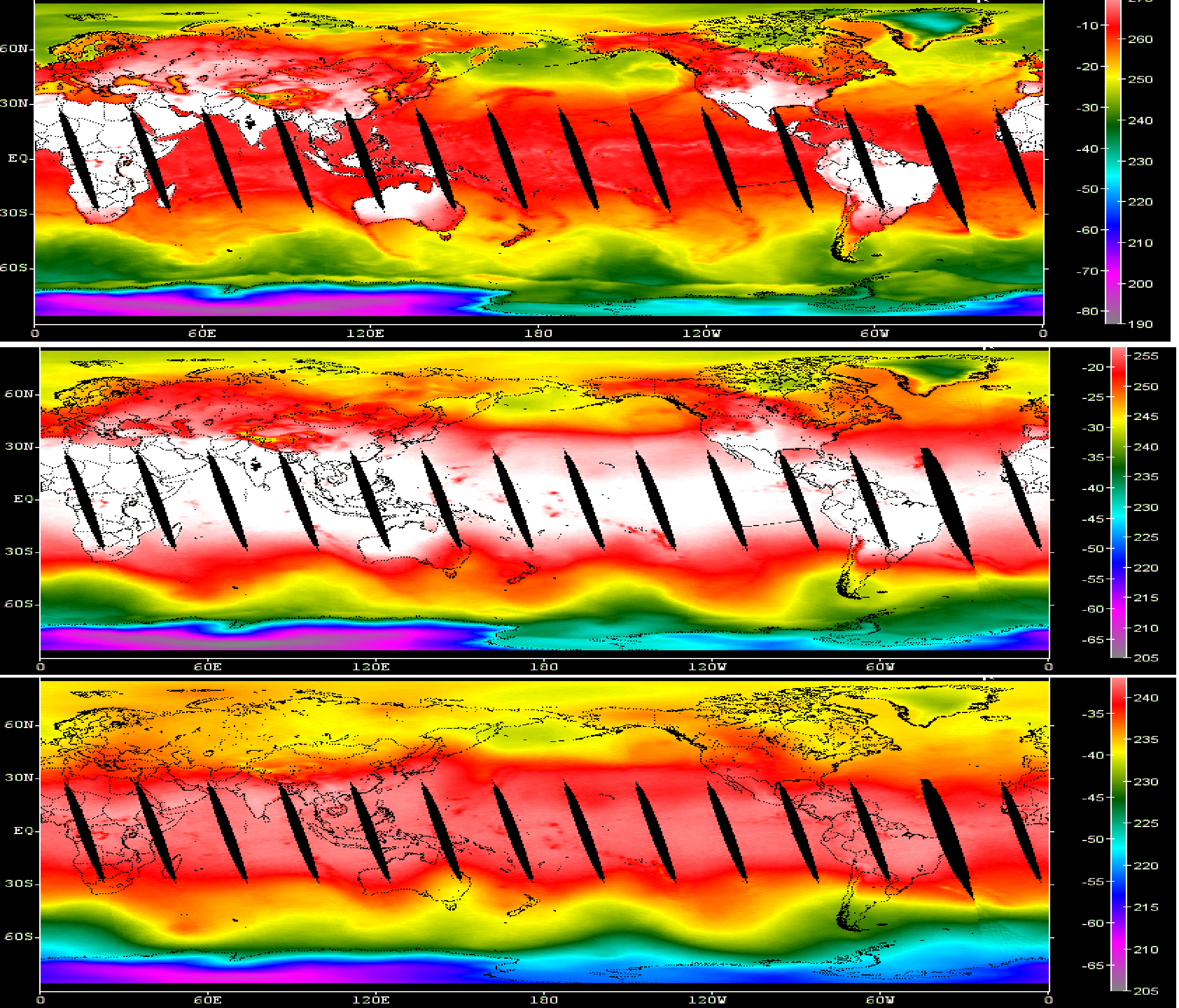
$\Delta R = - \boldsymbol{\sigma}_m B_s + \boldsymbol{\sigma}_m B_m = \boldsymbol{\sigma}_m [B_m - B_s]$

So if $[B_m - B_s] < 0$ then as $\boldsymbol{\sigma}_m$ increases ΔR becomes more negative


TRANSMITTANCE
FREQUENCY (GHz)
Total
H_2O
O_2
1.0
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0 20 40 60 80 100 120 140 160 180 200 220 240 260 280 300


AMSU-A1
Nadir
Clear Sky
C14
C13
C12
C11
C10
C9
C8
C7
C6
C5
C4
C3
C15
Height (km)
Weighting Function (km^{-1})
80
70
60
50
40
30
20
10
0
0.0 0.05 0.1 0.15 0.2 0.25


Multi-Channel Viewer
Tools Settings Import
brightnessTemp
320
180
frequency
30
80
Band: 3 frequency 50.30 GHz
224.02
242.39
305.95
Instrument: AMSU_A
Lat = 43.392 Lon = 31.227

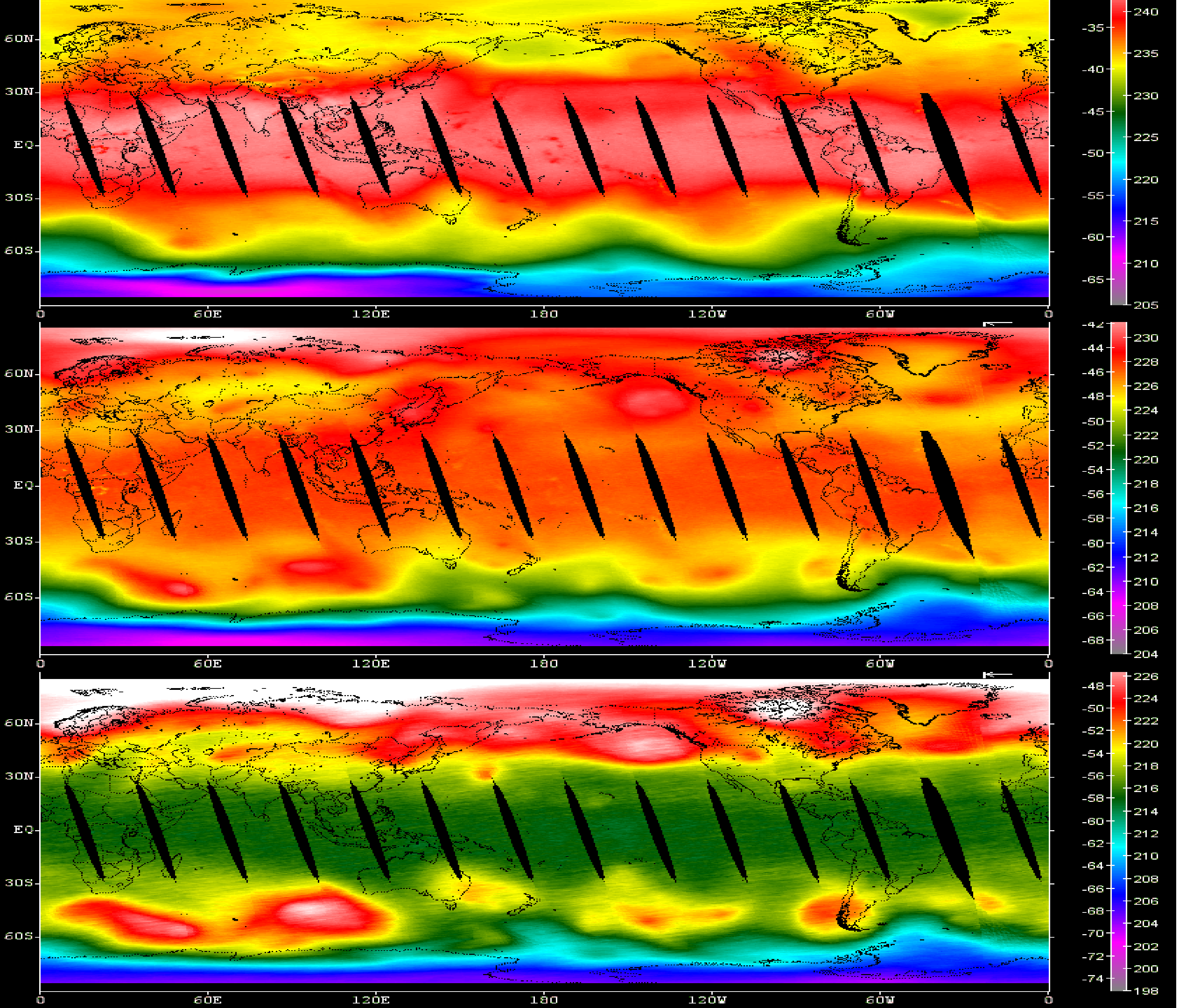


AMSU

52.8

53.6

54.4 GHz



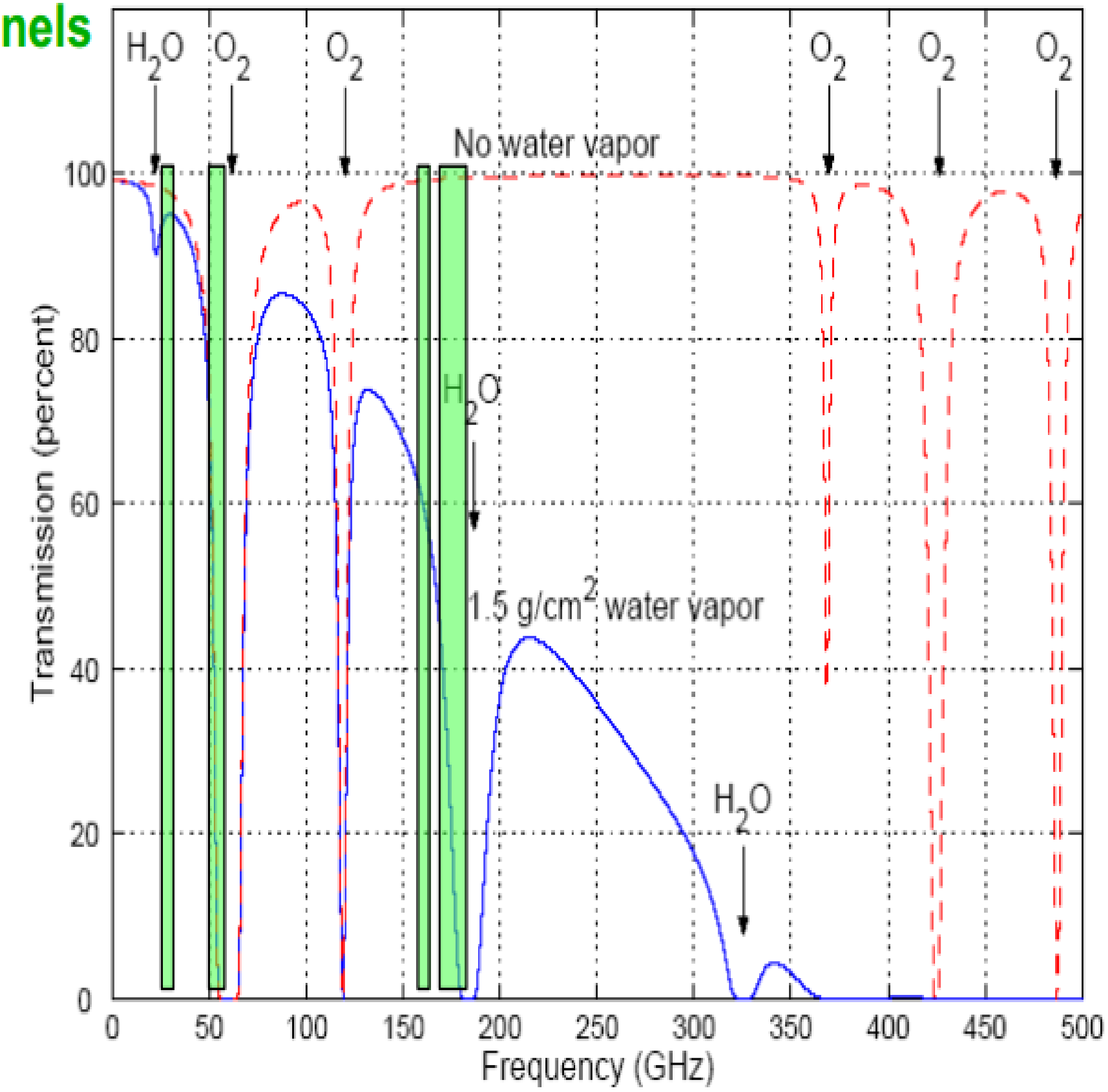
AMSU

54.4

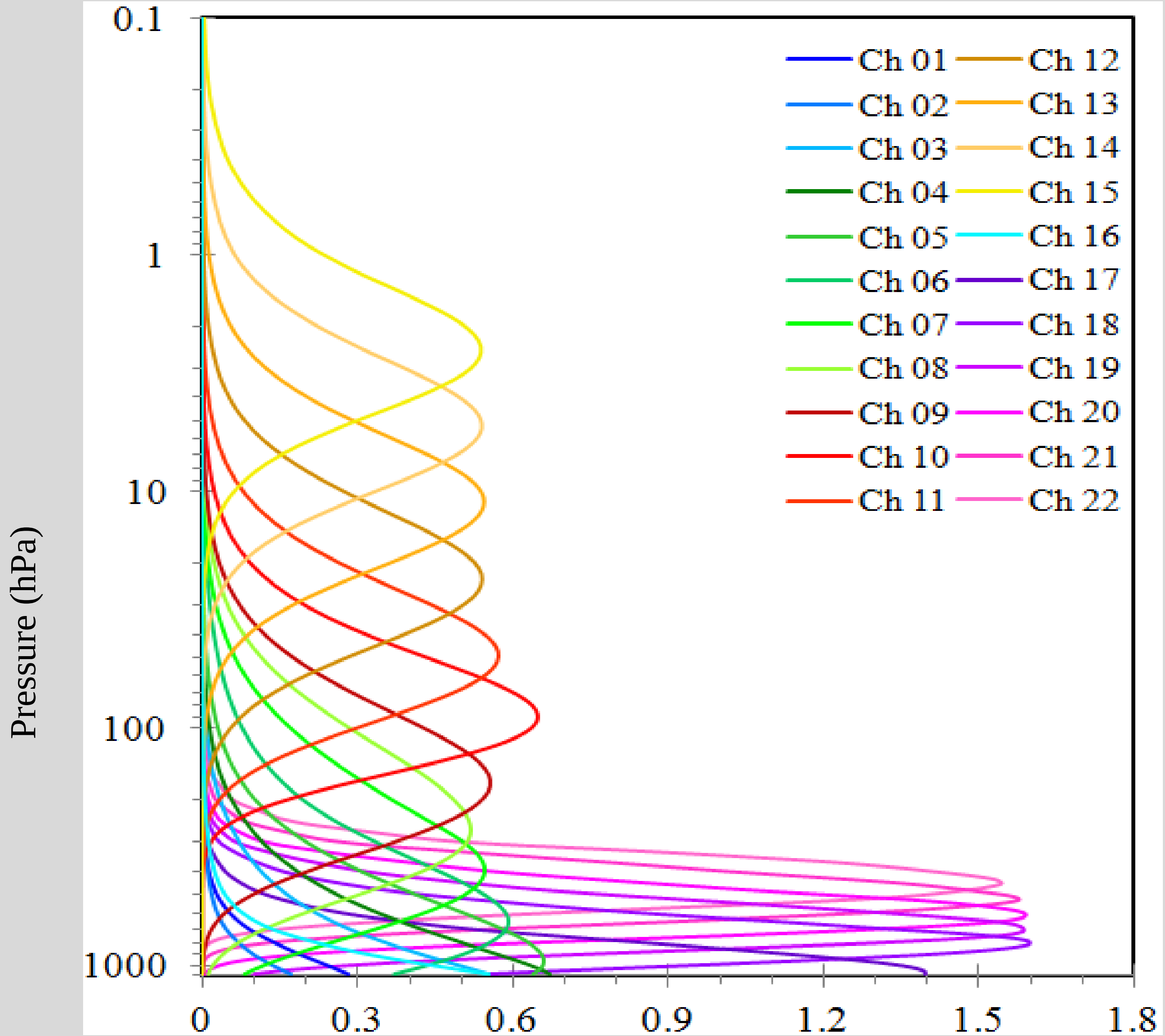
54.9

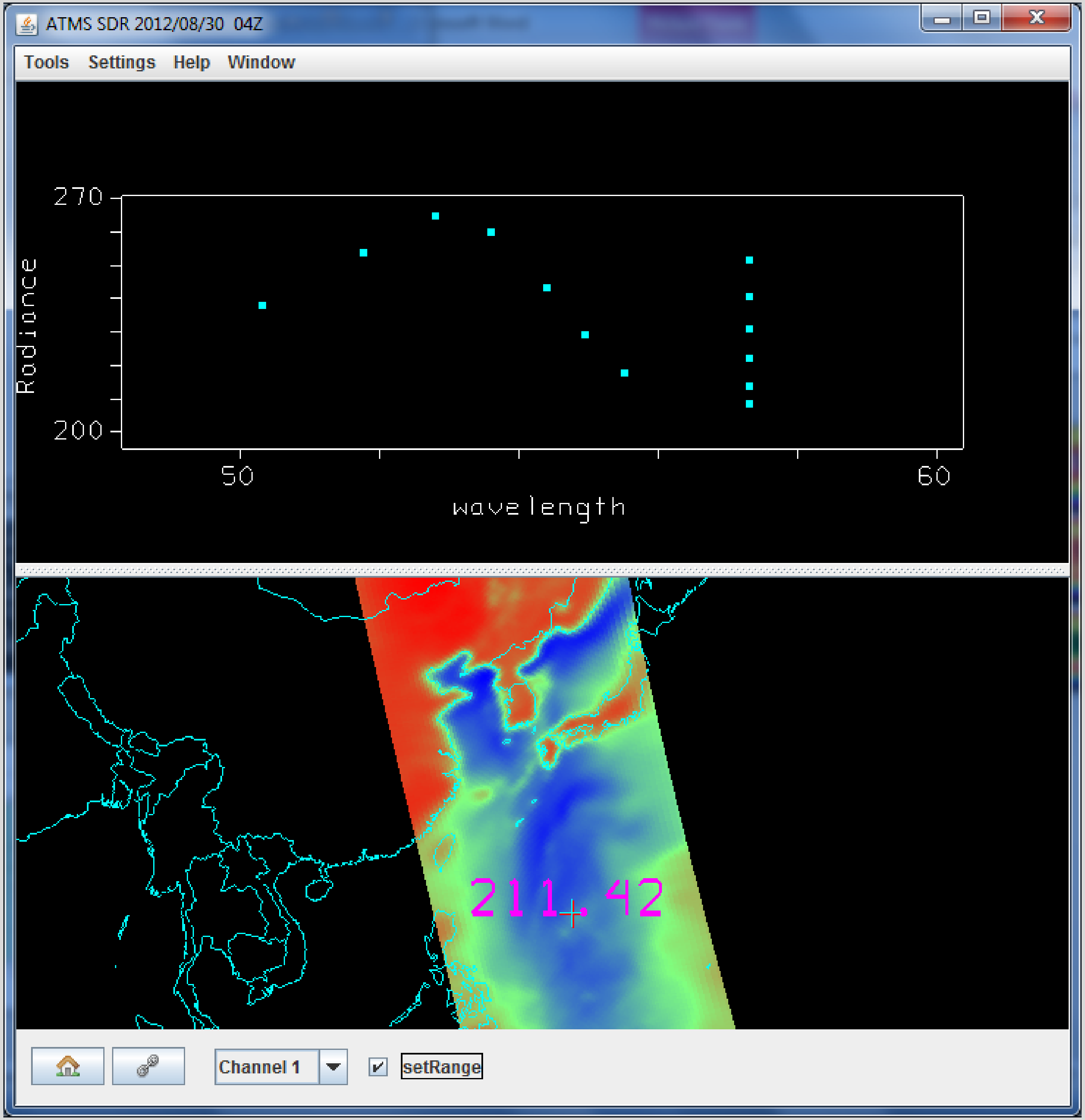
55.5 GHz

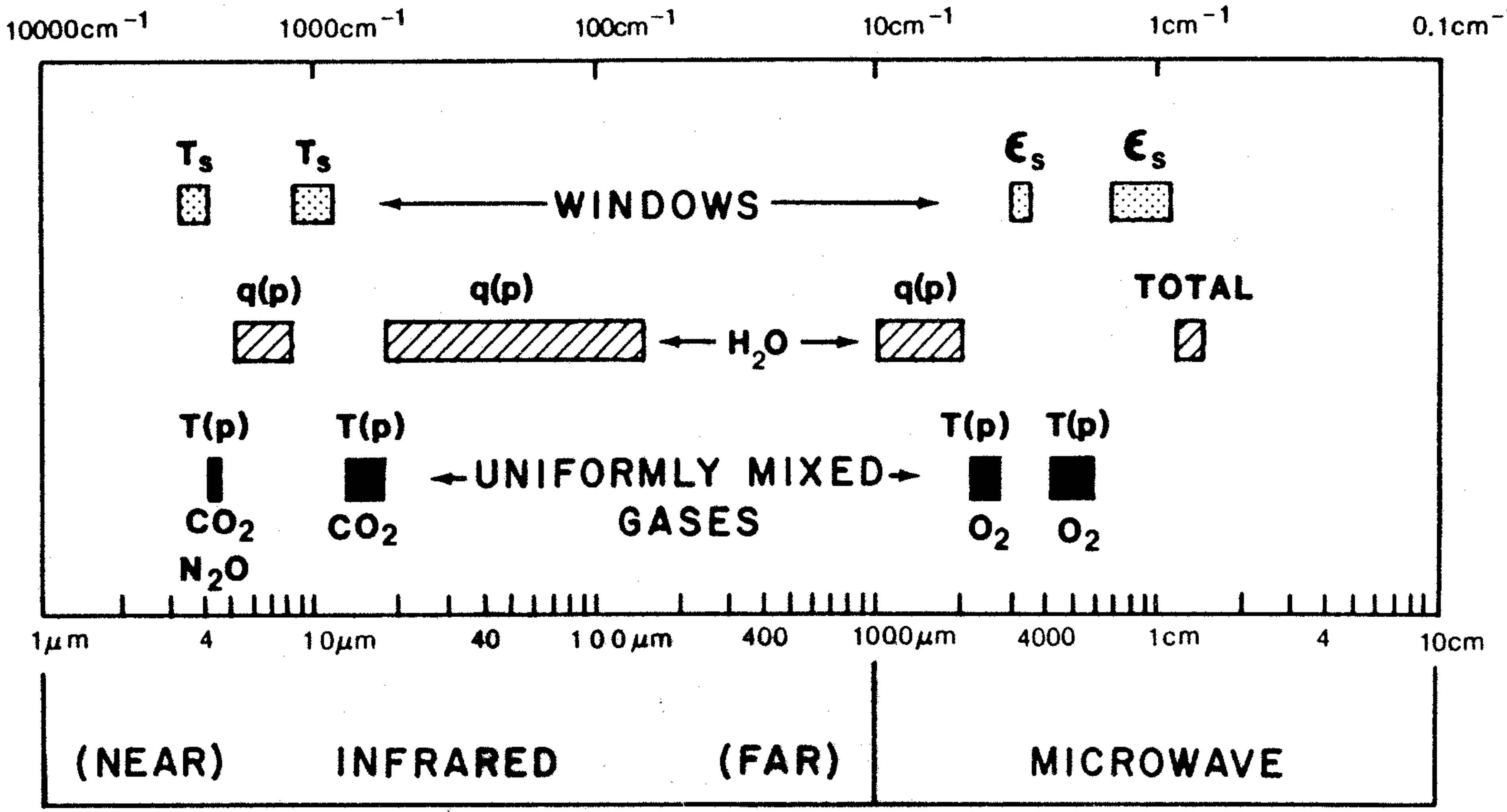
ATMS channels



ATMS Weighting Functions




ATMS SDR 2012/08/30 04Z
Tools Settings Help Window
270
200
Radiance
50
60
wavelength
211.42
Channel 1
setRange



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.