Lecture 12.

Applications of passive remote sensing using emission:

Remote sensing of SST.

Principles of sounding by emission and applications.

1. Remote sensing of sea surface temperature (SST).

2. Concept of weighting functions: weighting functions for nadir and limb soundings. Concept of an inverse problem.

3. Sounding of the atmospheric temperature. NOAA sounders.

4. Sounding of atmospheric gases.

Required reading:

S: 7.5, 1.2, 7.5.4, 7.3.1, 7.3.2, 7.7

Additional reading:

SST products (from NOAA AVHRR and other satellite sensors):

http://podaac.jpl.nasa.gov/sst/

4 km SST AVHRR Pathfinder Project:

http://www.nodc.noaa.gov/SatelliteData/pathfinder4km/

NOAA Optimum Interpolation 1/4 Degree Daily Sea Surface Temperature Analysis:

http://www.ncdc.noaa.gov/oa/climate/research/sst/oi-daily.php

Reynolds, R. W., et al. 2007: Daily High-Resolution-Blended Analyses for Sea Surface Temperature. J. of Climate.

http://www.ncdc.noaa.gov/oa/climate/research/sst/papers/daily-sst.pdf

Integrated SST Data Products:

http://www.ghrsst-pp.org/index.htm

Sounders for atmospheric chemistry (NASA Aura mission):

http://aura.gsfc.nasa.gov/

Aura data products

http://aura.gsfc.nasa.gov/science/data_release.pdf

Advanced reading:

C. D. Rodgers, Inverse methods for atmospheric sounding: Theory and practice. 2000.

1. Remote sensing of sea-surface temperature (SST).

The concept of SST retrievals from passive infrared remote sensing:

measure IR radiances in the atmospheric window and correct for contribution from "clear" sky by using multiple channels (called **a split-window technique**)

Using Eq.[11.10] , we can write the IR radiance at TOA:

$$I_\lambda^{\uparrow}(0;\mu) = B_\lambda(\tau^*)\exp(-\frac{\tau^*}{\mu}) + \frac{1}{\mu}\int_0^{\tau^*}\exp(-\frac{\tau'}{\mu})\ B_\lambda(\tau')d\tau' \qquad [12.1]$$

Let's re-write this equation using the transmission function $T_\lambda(\tau^*,\mu) = \exp(-\frac{\tau^*_\lambda}{\mu})$

and that

$$I_\lambda^{\uparrow}(0;\mu) = B_\lambda(T_{sur})T_\lambda(\tau^*,\mu) + B_\lambda(T_{atm})[1-T_\lambda(\tau^*,\mu)] \qquad [12.2]$$

where $\boldsymbol{T_{atm}}$ is an "effective" blackbody temperature which gives the atmospheric emission

$$B_\lambda(T_{atm}) = [1-T_\lambda(\tau^*,\mu)]^{-1}\frac{1}{\mu}\int_0^{\tau^*}\exp(-\frac{\tau'}{\mu})\ B_\lambda(\tau')d\tau' \qquad [12.3]$$

We want to eliminate the term with $\boldsymbol{T_{atm}}$ in Eq.[12.3]. Suppose we can measure IR radiances $\boldsymbol{I_1}$ and $\boldsymbol{I_2}$ at two adjacent wavelengths λ_1 and λ_2

$$I_1^{\uparrow} = B_1(T_{sur})T_1(\tau_1^*,\mu) + B_1(T_{atm})[1-T_1(\tau_1^*,\mu)] \qquad [12.4]$$

$$I_2^{\uparrow} = B_2(T_{sur})T_2(\tau_2^*,\mu) + B_2(T_{atm})[1-T_2(\tau_2^*,\mu)] \qquad [12.5]$$

NOTE: Two wavelengths need to be close to neglect the variation in $\boldsymbol{B_\lambda(T_{atm})}$ as a function of λ

Let's apply the Taylor's expansion to $\boldsymbol{B_\lambda(T)}$ at temperature $\boldsymbol{T = T_{atm}}$

$$B_\lambda(T) \approx B_\lambda(T_{atm}) + \frac{\partial B_\lambda(T)}{\partial T}(T - T_{atm}) \qquad [12.6]$$

Applying this expansion for both wavelengths, we have

$$B_1(T) \approx B_1(T_{atm}) + \frac{\partial B_1(T\ )}{\partial T}(T - T_{atm}) \qquad [12.7]$$

$$B_2(T) \approx B_2(T_{atm}) + \frac{\partial B_2(T\ )}{\partial T}(T - T_{atm}) \qquad [12.8]$$

and thus, eliminating ***T- T_{atm}***, we have

$$B_2(T) \approx B_2(T_{atm}) + \frac{\partial B_2(T\ )/\partial T}{\partial B_1(T)/\partial T}[B_1(T) - B_1(T_{atm})] \qquad [12.9]$$

Let's introduce brightness temperatures for these two channels $T_{b,1}$ and $T_{b,2}$

$I_1 = B_1(T_{b,1})$ and $I_2 = B_2(T_{b,2})$

and apply [12.9] to $B_2(T_{b,2})$ and to $B_2(T_{sur})$

$$B_2(T_{b,2}) \approx B_2(T_{atm}) + \frac{\partial B_2(T\ )/\partial T}{\partial B_1(T)/\partial T}[B_1(T_{b,2}) - B_1(T_{atm})] \qquad [12.10]$$

and

$$B_2(T_{sur}) \approx B_2(T_{atm}) + \frac{\partial B_2(T\ )/\partial T}{\partial B_1(T)/\partial T}[B_1(T_{sur}) - B_1(T_{atm})] \qquad [12.11]$$

Let's substitute the above expressions for $B_2(T_{b,2})$ and $B_2(T_{sur})$ in Eq.[12.5]

$$B_2(T_{atm}) + \frac{\partial B_2(T\ )/\partial T}{\partial B_1(T)/\partial T}[B_1(T_{b,2}) - B_1(T_{atm})] = \qquad [12.12]$$

$$= T_2\ \{B_2(T_{atm}) + \frac{\partial B_2(T\ )/\partial T}{\partial B_1(T)/\partial T}[B_1(T_{sur}) - B_1(T_{atm})]\} + B_2(T_{atm})[1 - T_2]$$

where T_1 and T_2 are transmissions in the channels 1 and 2.

Eq.[12.12] becomes

$$B_1(T_{b,2}) = B_1(T_{sur})T_2 + B_1(T_{atm})[1 - T_2] \qquad [12.13]$$

Using Eq.[12.4], we can eliminate $B_1(T_{atm})$

$$B_1(T_{sur}) = I_1 + \gamma[I_1 - B_1(T_{b,2)})] \qquad [12.14]$$

where $\gamma = \dfrac{1 - T_1}{T_1 - T_2}$; (T_1 and T_2 are transmissions in the channels 1 and 2).

Performing linearization of Eq.[12.14]

$$T_{sur} \approx T_{b,1} + \gamma [T_{b,1} - T_{b,2}] \qquad [12.15]$$

<u>Principles of a SST retrieval algorithm</u>:

SST is retrieved based on the linear differences in brightness temperatures at two IR channels. Two channels are used to eliminate the term involving $\boldsymbol{T_{atm}}$ and solve for T_{sur}.

NOTE: Clouds cause a serious problem in SST retrievals => need a reliable algorithm to detect and eliminate the clouds (called a **cloud mask**).

<u>**NOTE*****: One needs to distinguish the bulk sea surface temperature from skin sea surface temperature:***</u>

<u>**Bulk** (1-5 m depth) SST measurements:</u>

(1) Ships

(2) Buoys (since the mid-1970s): buoy SSTs are much less nosy that ship SSTs

⇩

Data from buoys are included in SST retrieval algorithms

<u>**Skin** SST from infrared satellite sensors:</u>

- SR (Scanning Radiometer) and VHRR (Very High Resolution Radiometer) both flown on NOAA polar orbiting satellites: since mid-1970
- AVHRR (Advanced Very High Resolution Radiometer) series: provide the longest data set of SST: since 1978 (4 channels, started on NOAA-6) and since 1988 (5 channels, started on NOAA-11)

Table 12 .1
AVHRR CHANNELS

AVHRR Channel	Wavelength (μm)
1	0.58 - 0.68
2	0.72 - 1.10
3	3.55 - 3.93
4	10.3 - 11.3
5	11.5 - 12.5

AVHRR MCSST (Multi-Channel SST) algorithm:

$$SST = \boldsymbol{a}\, \boldsymbol{T_{b,4}} + \gamma(\boldsymbol{T_{b,4}} - \boldsymbol{T_{b,5}}) + \boldsymbol{c} \qquad [12.16]$$

where ***a*** and ***c*** are constants.

$\gamma = \dfrac{1 - T_4}{T_4 - T_5}$, T_4 and T_5 are transmission function at AVHRR channels 4 and 5

AVHRR NLSST (Non-Linear SST) operational algorithm (Version 4.0):

$$SST = \boldsymbol{a} + \boldsymbol{b}\, \boldsymbol{T_{b,4}} + c(\boldsymbol{T_{b,4}} - \boldsymbol{T_{b,5}})\, SST_{guess} + \boldsymbol{d}(\boldsymbol{T_{b,4}} - \boldsymbol{T_{b,5}})[\sec(\theta_{sat}) - 1] \qquad [12.17]$$

where

SST_{guess} if a first-guess SST;

$\boldsymbol{T_{b,4}}$ and $\boldsymbol{T_{b,5}}$ are brightness temperature measured by AVHRR channels 4 and 5;

a, ***b***, and ***c*** are coefficients that calculated for two different regimes of ($\boldsymbol{T_{b,4}} - \boldsymbol{T_{b,5}}$):

one set for ($\boldsymbol{T_{b,4}} - \boldsymbol{T_{b,5}}$) < or = 0.7 and another set for ($\boldsymbol{T_{b,4}} - \boldsymbol{T_{b,5}}$) > 0.7

The coefficients ***a, b***, and ***c*** are estimated from regression analyses using co-located in situ buoy and satellite measurements (called "matchups").

Alternative approach

(used in the SST retrieval algorithm in ATSR (Along-Track Scanning Radiometer) on ERS; ATSR has 4 channels 1.6, 3.7, 10.8 and 12 μm)

$$SST = a_0 + \sum a_i T_{b,i} \qquad [12.18]$$

Coefficients $\boldsymbol{a_i}$ are calculated from a fit to a radiative transfer model instead of in situ observations as in the AVHRR algorithm.

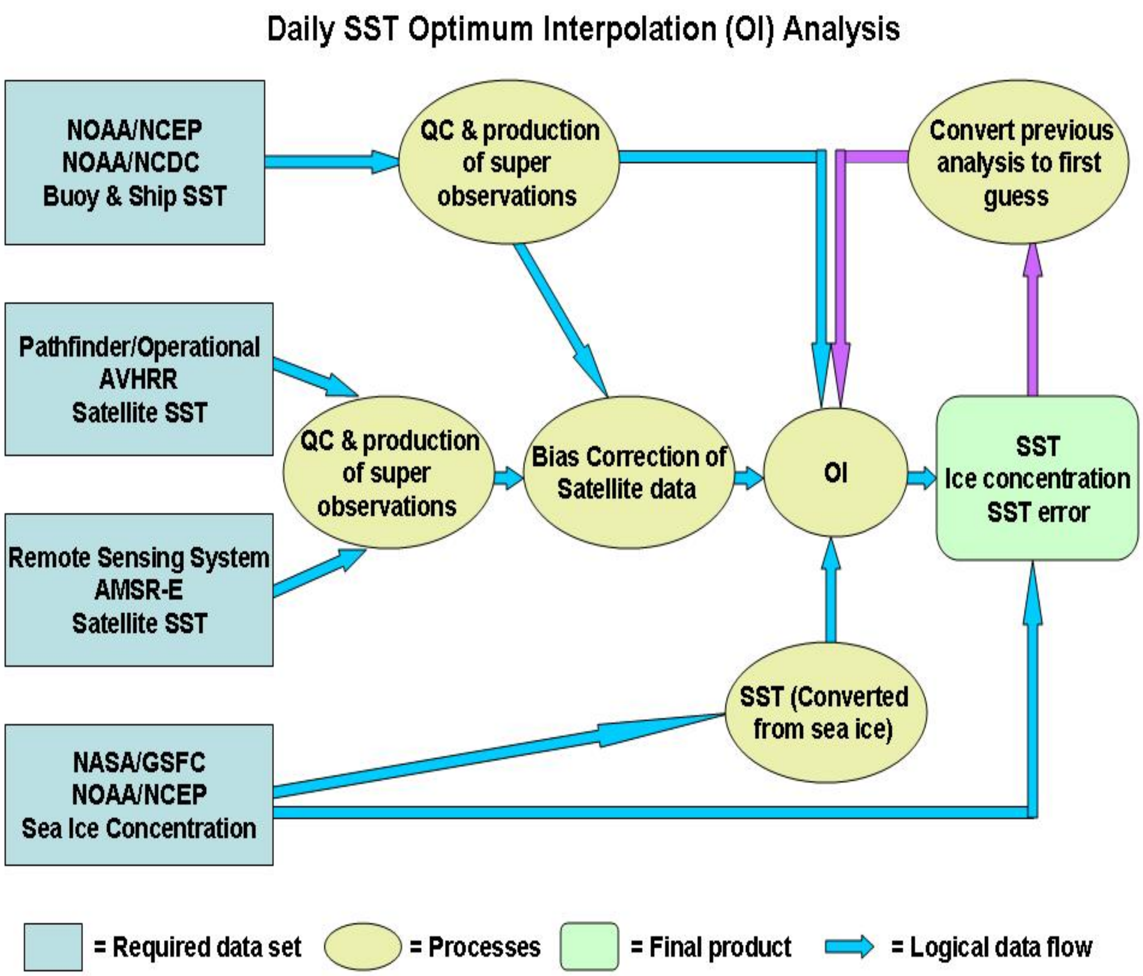
NOTE: both algorithms work for cloud-free pixels => cloud mask is required

Reynolds Optimal Interpolation Algorithm:

Reynolds, R. W., et al. 2007: Daily High-resolution Blended Analyses. J. of Climate.

http://www.ncdc.noaa.gov/oa/climate/research/sst/papers/daily-sst.pdf

1) Merges Advanced Very High Resolution Radiometer (AVHRR) infrared satellite SST data and data from Advanced Microwave Scanning Radiometer (AMSR) on the NASA
2) Data products have a spatial grid resolution of 0.25° and a temporal resolution of 1 day.



✓ **Microwave vs. IR SST retrievals**

Factor affecting radiometry	**Infrared radiometry**	**Microwave radiometry**
Magnitude of emitted radiation from the sea surface	[+] large B(λ, T)	[-] small B(λ, T)
Sensitivity of brightness to SST	[+] large $\frac{1}{B}\frac{\partial B}{\partial T}$	[-] small $\frac{1}{B}\frac{\partial B}{\partial T}$ (B is proportional to T)
Emissivity	[+] ε = about 1	[-] ε = about 0.5
Clouds	[-] Not transparent	[+] Clouds largely transparent (improvement at longer wavelengths)
Sea state (e.g.,, roughness)	[+] Independent	[-] ε varies with sea state
Atmospheric interference	[-] Requires complex correction	[+] Easily corrected with multichannel radiometer
Spatial resolution	[+] A narrow beam can be focused. Diffraction is not a problem in achieving high spatial resolution with a small instrument	[+] Diffraction controls the beam at large wavelengths. Large antenna required for high spatial resolution
Viewing direction on surface	[+] Surface radiance largely independent of viewing direction	[-] ε varies with viewing direction
Absolute calibration	[+] Readily achieved using heated on-board target	[-] Absolute calibration target not readily achieved
Presently achievable sensitivity	0.1 degree K	1.5 degree K
Presently achievable absolute accuracy	0.6 degree K	2 degree K

2. Concept of weighting functions: weighting functions for nadir and limb soundings. Concept of an inverse problem.

Recall Eq.[11.8] that gives the **upward intensity in the plane-parallel atmosphere with emission (neglecting scattering):**

$$I_\nu^{\uparrow}(\tau,\mu) = I_\nu^{\uparrow}(\tau^*,\mu)\exp(-\frac{\tau^*-\tau}{\mu})$$
$$+\frac{1}{\mu}\int_{\tau}^{\tau^*}\exp(-\frac{\tau'-\tau}{\mu})\ B_\nu(\tau')d\tau'$$

and that

$$T_\nu(\tau',\tau,\mu) = \exp(-\frac{\tau'-\tau}{\mu}) \ \text{ and } \ \frac{dT_\nu(\tau',\tau,\mu)}{d\tau'} = -\frac{1}{\mu}\exp(-\frac{\tau'-\tau}{\mu})$$

where μ is the cosine of zenith angle of observation.

Eq.[11.8] can be re-written as

$$I_\nu^{\uparrow}(\tau,\mu) = I_\nu(\tau^*)T_\nu(\tau^*,\tau,\mu) + \int_{\tau^*}^{\tau} B_\nu(\tau')\frac{dT_\nu(\tau',\tau,\mu)}{d\tau'}d\tau' \qquad [12.19]$$

Eq.[12.19] can be expressed in different vertical coordinates such as z, P or ***ln***(P).

Let's denote an arbitrary vertical coordinate by $\tilde{z}$, so Eq.[12.19] becomes

$$I_\nu^{\uparrow}(\tilde{z},\mu) = I_\nu(\tilde{z}^*)T_\nu(\tilde{z}^*,\tilde{z},\mu) + \int_{\tilde{z}^*}^{\tilde{z}} B_\nu(\tilde{z}')W(\tilde{z}',\tilde{z},\mu)d\tilde{z}' \qquad [12.20]$$

where the **weighting function** is defined in the general form as

$$\boxed{W_\nu(\tilde{z}_1,\tilde{z}_2,\mu) = \left|\frac{dT_\nu(\tilde{z}_1,\tilde{z}_2,\mu)}{d\tilde{z}}\right|} \qquad [12.21]$$

<u>Physical meaning of the weighting function</u>:

Radiances emitted from a layer $d\tilde{z}'$ is determined by a blackbody emission, $B_\nu(\tilde{z}')$, of the layer weighted by the factor $W_\nu(\tilde{z}',\tilde{z},\mu)d\tilde{z}'$

Let's re-write the solutions of the radiative transfer equation for upward and downward radiances in the altitude coordinate, z. Recall that the optical depth of a layer due to absorption by a gas in this layer is

$$\tau_\nu = \int_{u_1}^{u_2} k_\nu du$$

Let's express the optical depth in terms of a mass absorption coefficient of the absorbing gas and its density

$$\tau_\nu = \int_{z'}^{z} k_\nu \rho_{gas} \, dz'' \qquad [12.22]$$

Thus transmission between z and z' along the path at μ is

$$T_\nu(z, z', \mu) = \exp\left(-\frac{1}{\mu}\int_{z'}^{z} k_\nu \rho_{gas} \, dz''\right) \qquad [12.23]$$

and

$$\frac{dT_\nu(z, z', \mu)}{dz'} = -\frac{k_\nu \rho_{gas}}{\mu}\exp\left(-\frac{1}{\mu}\int_{z'}^{z} k_\nu \rho_{gas} \, dz''\right) \qquad [12.24]$$

Therefore, for the **upward intensity and downward intensities** at the altitude z, we have

$$I_\nu^{\uparrow}(z, \mu) = I_\nu^{\uparrow}(0, \mu) T_\nu(z, 0, \mu) + \int_0^z B_\nu(T(z')) \left| \frac{dT_\nu(z, z', \mu)}{dz'} \right| dz' \qquad [12.25]$$

$$I_\nu^{\downarrow}(z, -\mu) = \int_z^{\infty} B_\nu(T(z')) \left| \frac{dT_\nu(z, z', \mu)}{dz'} \right| dz' \qquad [12.26]$$

and thus

$$I_\nu^{\uparrow}(z, \mu) = I_\nu^{\uparrow}(0, \mu) \exp\left[-\frac{1}{\mu}\int_0^z k_\nu \rho_{gas} \, dz'\right] + \frac{1}{\mu}\int_0^z \exp\left[-\frac{1}{\mu}\int_{z'}^{z} k_\nu \rho_{gas} \, dz'\right] B_\nu(T(z')) k_\nu \rho_{gas} \, dz' \qquad [12.27]$$

$$I_\nu^{\downarrow}(z, -\mu) = \frac{1}{\mu}\int_z^{\infty} \exp\left[-\frac{1}{\mu}\int_{z'}^{z} k_\nu \rho_{gas} dz'\right] B_\nu(T(z')) k_\nu \rho_{gas} dz' \qquad [12.28]$$

NOTE: Eq.[12.27] is fundamental to remote sensing of the atmosphere. It shows that the upwelling intensity in the IR is a product of the Planck function, spectral transmittance and weighting function. The information on temperature is included in the Planck function, while the density profile of relevant absorbing gases is involved in the weighting function, or vise versa. ⇓

Thus, one can retrieve the profile of temperature if the profiles of relevant absorbing gases are known.

Weighting functions for near-nadir sounding:

From Eq.[12.25], upwelling radiance detected by a satellite sensor at $z = \infty$ is

$$I_\nu^{\uparrow}(\infty,\mu) = I_\nu^{\uparrow}(0,\mu) T_\nu(\infty,0,\mu) + \int_0^{\infty} B_\nu(T(z')) \left| \frac{dT_\nu(\infty,z',\mu)}{dz'} \right| dz' \qquad [12.29]$$

or

$$I_\nu^{\uparrow}(\infty,\mu) = I_\nu^{\uparrow}(0,\mu) T_\nu(\infty,0,\mu) + \int_0^{\infty} B_\nu(T(z)) W_\nu(\infty,z,\mu) dz \qquad [12.30]$$

where

$$W_\nu(\infty,z,\mu) = \left| \frac{dT_\nu(\infty,z,\mu)}{dz} \right|$$

thus from Eq.[12.24] we have

$$W_\nu(\infty,z,\mu) = \left| \frac{T_\nu(\infty,z,\mu)}{dz} \right| = \frac{k_\nu \rho_{gas}}{\mu} \exp\left(-\frac{1}{\mu}\int_z^{\infty} k_\nu \rho_{gas} dz\right) \qquad [12.31]$$

decreases with altitude

increases with altitude

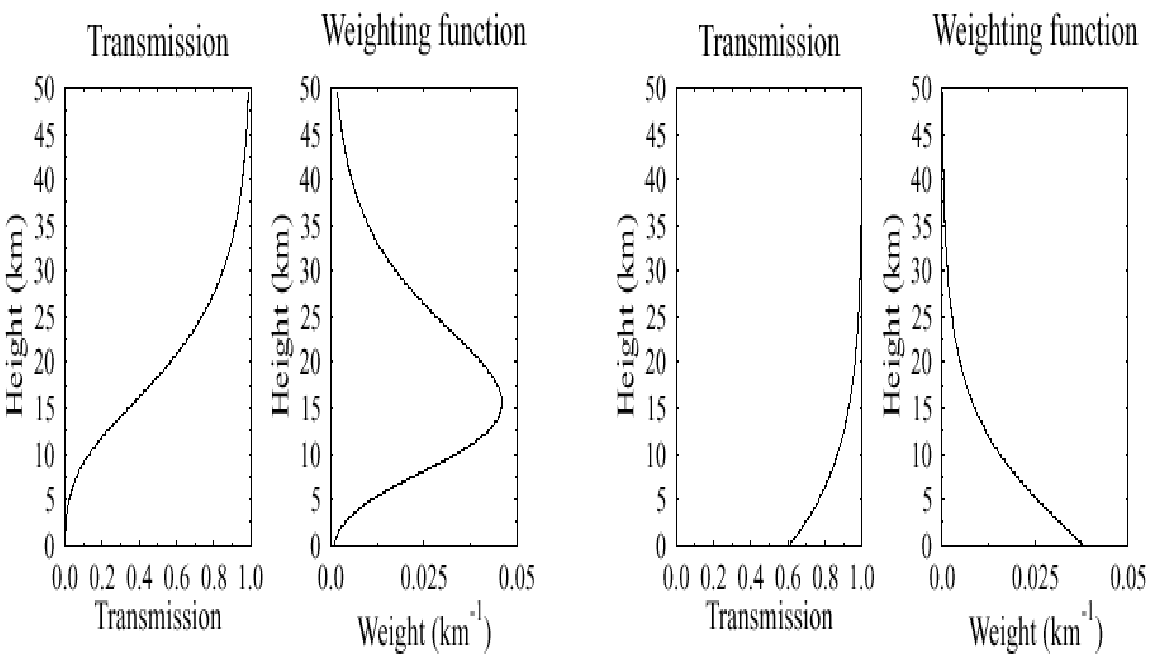


Figure 12.1 Schematic of weighting functions for optically thick and optically thin media.

Weighting functions for limb sounding:

The intensity measured by a satellite for limb viewing geometry is the integral of the emission along a line-of-sign

$$I_\nu(h) = \int_{\infty}^{0} B_\nu(s) \frac{dT_\nu(s,0)}{ds} ds \qquad [12.32]$$

h is the tangent altitude,

T_ν(s, 0) is the transmission function along the path of length s from the outer levels of the atmosphere at $s = \infty$.

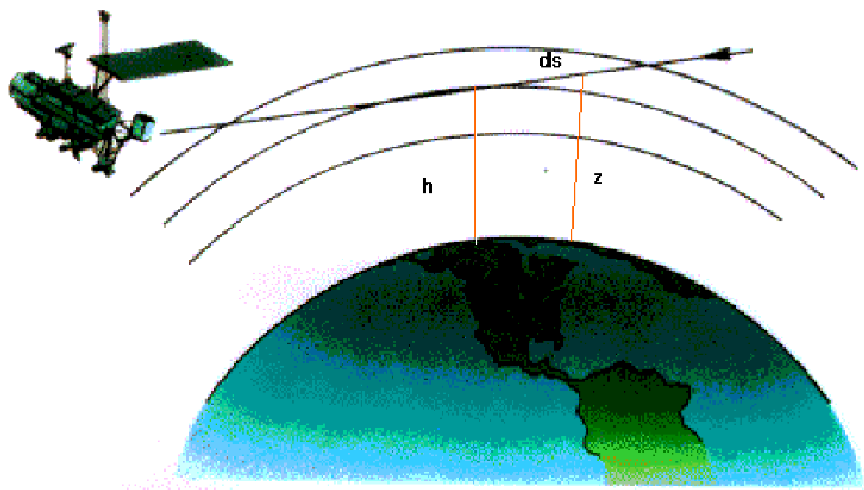


Figure 12.2 Viewing geometry of limb sounding.

We can relate ***s*** and ***z*** as

$$(R + h)^2 + s^2 = (R + z)^2 \qquad [12.33]$$

where R is the radius of the Earth. Using that R>>h, we have

$$s^2 \approx 2R(z - h) \qquad [12.34]$$

Thus Eq.[12.32] becomes

$$I_\nu(h) = \int_{h}^{\infty} B_\nu(z') \frac{dT_\nu(z',\infty)}{ds} \frac{ds}{dz'} dz' \qquad [12.35]$$

and

$$I_\nu(h) = \int_{h}^{\infty} B_\nu(z') W(h, z', \infty) dz' \qquad [12.36]$$

where

$W_\nu\ (h, z, \infty) = 0$ for ***z < h***

$$W_\nu\ (h, z, \infty) = \frac{dT_\nu(z, \infty)}{ds}\sqrt{R/2(z-h)} \quad \text{for } \boldsymbol{z > h} \qquad [12.37]$$

enhancement of the tangent path relative to a vertical path

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Limb sounding has higher sensitivity to emission from trace gases (CO, NO, N_2O, ClO)

Advantages of limb sounding:

- Can measure emission from gases of low concentrations
- Surface emission does not effect limb sounding
- Good vertical distribution since it senses outgoing radiation from just a few kilometers above the tangent height (=> weighting functions have spikes, see S:fig.7.18)

Disadvantages of limb sounding:

- ✓ Can not be used below the troposphere
- ✓ Requires precise information of the viewing geometry

➢ **<u>Concept of an inverse problem.</u>**

Recall Eq.[12. 29] which gives the solution of the radiative transfer equation with emission for the **outgoing intensity** at the top of the atmosphere $z = \infty$

$$I_\nu^{\uparrow}(\infty, \mu) = I_\nu^{\uparrow}(0; \mu,)T_\nu(\infty, 0, \mu) + \int_0^\infty B_\nu(T(z))W_\nu(\infty, z, \mu)dz$$

where

$W_\nu(\infty, z, \mu) = \left|\frac{dT_\nu(\infty, z, \mu)}{dz}\right|$ is the weighting function.

Forward problem is to calculate outgoing radiances for given temperature and gas concentration profiles.

Inverse problem in remote sounding is to determine what temperature and gas concentration profiles could have produced a set of radiances observed at closely spaced wavenumbers.

Let's assume that $T_\nu(\infty, 0, \mu)$ is negligible small. The for nadir sounding we have

$$I^{\uparrow}_{\nu_i}(\infty) = \int_0^\infty B_{\nu_i}(T(z)) W_{\nu_i}(\infty, z) dz \qquad [12.38]$$

where ν_i , ***i*** =1, 2…N, is the wavenumbers (i.e., centers of the finite spectral bands of a spectrometer).

We can ignore the frequency dependence of the Planck function if ν_i are closely spaced. Eq.[12.38] becomes

$$I^{\uparrow}_{\nu_i}(\infty) = \int_0^\infty B(\overline{\nu}, T(z)) W_{\nu_i}(\infty, z) dz \qquad [12.39]$$

Consider that measurements are done in the CO2 absorbing band, so the weighting function are known and $B(\overline{\nu}, T(z))$ is unknown. Thus we want to solve Eq.[12.39] to find $B(\overline{\nu}, T(z))$ and hence $T(z)$.

Eq.[12.38] is a so-called **Fredholm integral equation of the first kind**, which has long been known for many associated difficulties. The general form of the Fredholm integral equation (because the limits of the integral are fixed, not variable) of the first kind (because ***f(x)*** appears only in the integral) is

$$g(y) = \int_a^b f(x) K(y, x) dx \qquad [12.40]$$

where

f(x) is unknown (in our case $B(\overline{\nu}, T(z))$)

K(y, x) is called the kernel (or kernel function), (in our case $W_{\nu_i}(\infty, z)$)

g(y) is known (in our case $I^{\uparrow}_{\nu_i}(\infty)$)

Problems in solving the inverse problem (i.e., problems in solving Eq.[12.39])

- The inverse problem is **ill-posed** (i.e., underconstrained) because there are only a finite number of measurements and the unknown is a continuous function.
- This inverse problem is **ill-conditioned** (i.e., any experimental error in the measurements of radiances can be greatly amplified, so that the solution becomes meaningless)

3. Sounding of the atmospheric temperature.

Let's find the simplest solution of Eq.[12.39], assuming that transmittance (and hence weighting functions) does not depend on temperature. Thus Eq.[12.39] is linear in $B(\overline{\nu}, T(z))$.

A standard approach is to express $B(\overline{\nu}, T(z))$ as a linear function of ***N*** variables, b_j:

$$B(\overline{\nu}, T(z)) = \sum_{j=1}^{N} b_j L_j(z) \qquad [12.41]$$

where $L_j(z)$ is a set of representation functions such as polynomials or sines and cosines.

Eq.[12.39] becomes

$$I_{\nu_i}^{\uparrow} = \sum_{j=1}^{N} b_j \int_0^{\infty} L_j(z) W_{\nu_i}(z) dz = \sum_{j=1}^{N} C_{ij} b_j \qquad [12.42]$$

where the square matrix C whose elements $C_{ij} = \int_0^{\infty} L_j(z) W_{\nu_i}(z) dz$ can be easily calculated. Thus for the N unknown b_j, there are N equations which can be solved. The solution of Eq.[12.42] is called an **exact solution to the linear problem.**

Problems:

Eq.[12.42] is ill-conditioned (i.e., any experimental error in the measurements can be greatly amplified and the solution can be physically meaningless, though it satisfies Eq.[12.42]).

Strategy: instead of **an exact solution**, find the solution that lies within the experimental error of the measurements => it gives us more freedom in a choice of a solution, but a new problem is how to make this choice.

Thus the problem of retrieval can be re-stated as:

Given the measured radiances, the statistical experimental error, the weighting functions, and any other relative information, what solutions for $B(\overline{\nu}, T(z))$ are physically meaningful?

A priori information is often used to provide "other relative information".

A priori information for the retrievals of temperature:

- Radiosonde data
- Forecast atmospheric dynamical models

Example: GOES soundings are generated every hour, using an ETA model forecast as a 'first guess'

NOAA sounders:

NOAA-series polar orbiting satellites:

HIRS/2 (High Resolution Infrared Radiation Sounder 2): 20 channels, resolution 19 km (nadir)

MSU (Microwave sounding Unit): 4 channels, resolution 111 km (nadir)

SSU (Stratospheric Sounding Unit): 3 channels (in the 15-μm CO_2 band), resolution 111 km (nadir)

[HIRS+MSU+SSU] is called ***TOVS*** (TIROS N Operational Vertical Sounder)

NOTE: MSU+ SSU were replaced by AMSU-A and AMSU-B (Advanced Microwave Sounding Units) on NOAA K, L, M series.

Retrieval methods:

- Physical retrievals
- Statistical retrievals
- Hybrid retrievals

Physical retrievals:

use the forward modeling to do the iterative retrieval of the temperature profile

Steps:

1) Selection of a first-guess initial temperature profile.
2) Calculation of the weighting functions.
3) Forward modeling of the radiance in each channel of a sensor.
4) If the calculated radiance agrees with the measured radiance within the noise level of the sensor, the current profile gives the solution.
5) If convergence has not been achieved, the current profile is adjusted, and steps 2 through 5 are repeated until a solution is found.

Advantages:

- Relevant physical processes are taken into account at each step.
- No data is necessary

Disadvantages:

- Computationally intensive
- Requires accurate predictions of the transmittance

Statistical retrievals:

Do not solve the radiative transfer (i.e., no forward modeling). Instead a training data set – a set of radiosonde soundings that are collocated in time and space with satellite soundings – is used to establish a statistical relationship between the measured radiances and temperature profiles. These relationships are then applied to other measurements of the radiances to retrieve the temperature.

Advantages:

- Computationally easy and fast (does not require to solve the radiative transfer equation).

Disadvantages:

- A large training data set (covering different region, seasons, land types, etc.) is required.
- Physical processes are hidden in the statistics.

Hybrid retrievals: (also called inverse matrix methods)

Use the weighing functions (but do not solve the radiative transfer equation) and available data (but do not require a large data set).

One of many linear methods:

Minimum variance method is used for routine soundings from the TOVS (TIROS N Operational Vertical Sounder):

Strategy: to seek the solution which minimizes the mean-square differences between the retrieved profile and true profiles.

Start with a matrix equation

$$l = At + e$$

where **l** is the column vector of radiance deviations from the true profile,

t is the vector of temperature deviations from the true profile;

e is the column vector of measurement errors;

A is the matrix containing the weighting functions, the Planck sensitivity factors dB/dT and it can be calculated with a knowledge of the transmittances.

Let's assume at the moment that we have a set of collocated radiosonde and satellite observations

$$L = AT + E \qquad [12.43]$$

here upper cases means that matrices are N columns for N sounding pairs.

We seek a matrix C that **T = CL.** It can be shown that the matrix C which, in a least-squares sense, minimizes errors in **T = CL** is

$$C = TL^T (LL^T)^{-1} \qquad [12.44]$$

where the superscript T indicates matrix transpose and (-1) indicates the matrix inverse. Substituting Eq. [12.43] in Eq.[12.44], we have

$$C = T(AT + E)^T [(AT+E)(AT+ E)^T)]^{-1} \quad [12.45]$$

Expanding and using that $(AB)^T=B^TA^T$, it becomes

$$C = T(T^TA^T + E^T) [AT T^TA^T +ATE^T + ET^TA^T + EE^T]^{-1} \quad [12.46]$$

Assuming that the measurement errors are uncorrelated with temperature deviations, $\mathbf{TE^T}$ and $\mathbf{ET^T}$ are negligible small and therefore

$$C = S_TA^T[AS_TA^T +S_E]^{-1} \quad [12.47]$$

$\mathbf{S_T}$ the temperature covariance matrix

$\mathbf{S_E}$ the radiance error covariance matrix

Eq.[12.47] is called minimum variance method.

4. Sounding of atmospheric gases.

Strategy: make satellite measurements at frequencies in absorption bands of atmospheric gases.

The principles of the methods for gas retrievals are generally the same as for temperature retrievals, but gas inversion is more difficult:

- The inverse problem is more nonlinear for constituents (because they enter the radiative transfer equation through the mixing ratio profile in the exponent. It is not possible to separate the radiative transfer equation into the product of a simple constituent function and one which is constituent-independent).
- The second main problem is that, in some cases, the radiances are insensitive to changes in the mixing ratio. (For instance, for an isothermal atmosphere at temperature T, any mixing ratio profile will result in the same radiances (i.e. $B_\nu(T)$). In practice, we find this problem in the retrieval of low-level water vapor; because the temperature of the water vapor is close to that of the surface, infrared radiances are relatively insensitive to changes in low-level water vapor).

Examples of sounders and other sensors for remote sensing of gaseous species:

GOME (Global Ozone Monitoring Experiment), ESA-ERS

retrievals of O3, NO2, H2O, BrO, SO2, HCHO, OCLO

http://www-iup.physik.uni-bremen.de/gome/

SCIAMATCHY (Scanning Imaging Absorption Spectrometer for Atmospheric Cartography), ESA-ENVISAT, 2001-present,

retrievals of O2, O3, NO, NO2, N2O, H2O, BrO, SO2, HCHO, H2CO, CO, OCLO, CO2, CH4

http://www.iup.uni-bremen.de/sciamachy/

http://www.temis.nl/products/

MOPITT (Measurements Of Pollution In The Troposphere), NASA Terra, 1999-present

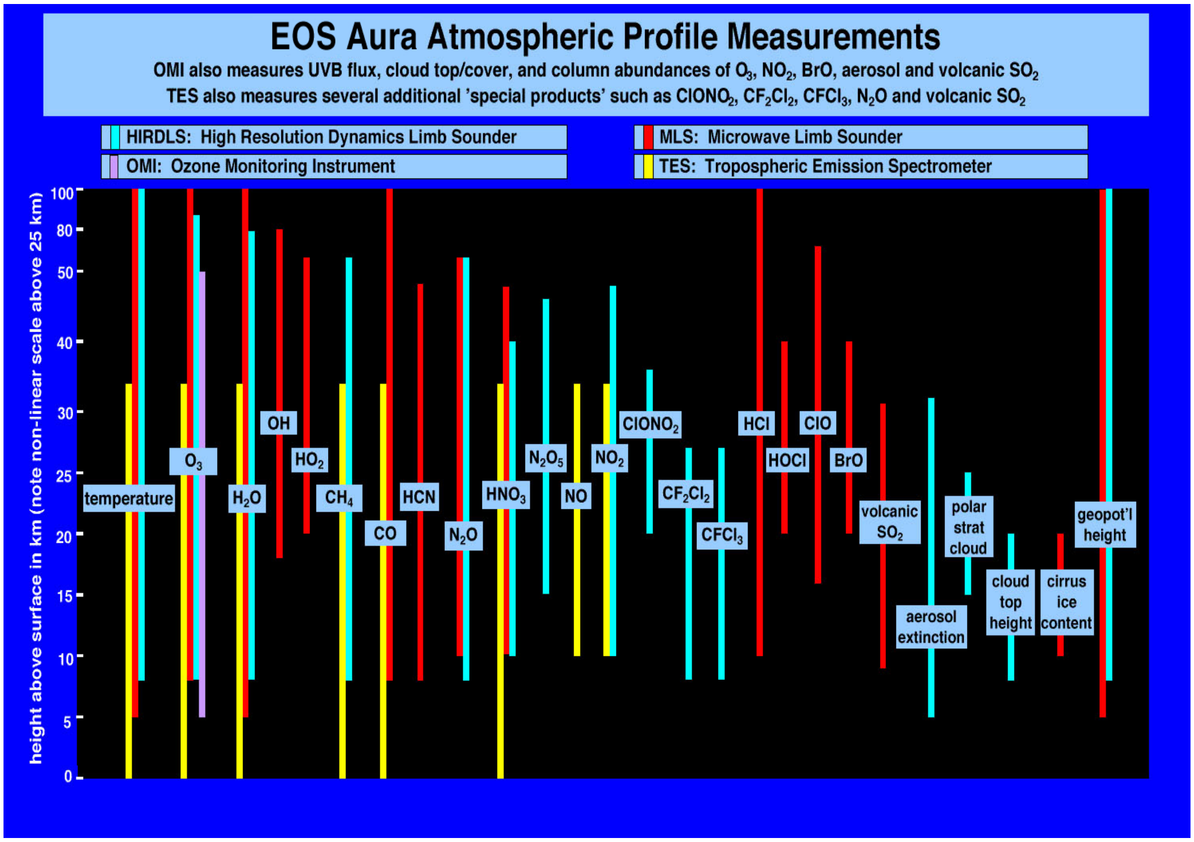
retrievals of CO and CH4 (retrieval algorithm involves weighing function)

http://www.eos.ucar.edu/mopitt/

NASA satellite missions:

EOS Aura: launched July 15, 2004.

http://aura.gsfc.nasa.gov/



HIRDLS (High Resolution Dynamics Limb Sounder) is an infrared limb-scanning radiometer designed to sound the upper troposphere, stratosphere, and mesosphere to determine temperature; the concentrations of O3, H2O, CH4, N2O, NO2, HNO3, N2O5, ClONO2, CFCl2, CFCl3, and aerosols; and the locations of polar stratospheric clouds and cloud tops. HIRDLS performs limb scans in the vertical at multiple azimuth angles, measuring infrared emissions in 21 channels ranging from 6.12 µm to 17.76 µm.
NOTE: HIRDLS has some series technical problems.

MLS (Microwave Limb Sounder) measures lower stratospheric temperature and concentrations of H_2O, O_3, ClO, BrO, HCl, OH, HO_2, HNO_3, HCN, and N_2O.

MLS observes in spectral bands centered near the following frequencies:

118 GHz, primarily for temperature and pressure;

190 GHz, primarily for H_2O, HNO_3, and continuity with UARS MLS measurements;

240 GHz, primarily for O_3 and CO;

640 GHz, primarily for N_2O, HCl, ClO, HOCl, BrO, HO_2, and SO_2; and

2.5 THz, primarily for OH.

OMI (Ozone Monitoring Instrument) continues the TOMS record for total ozone, also will measure NO_2, SO_2, BrO, OClO, and aerosol characteristics.

TES (Tropospheric Emission Spectrometer**)** is a high-resolution infrared-imaging Fourier transform spectrometer with spectral coverage of 3.2 to 15.4 µm at a spectral resolution of 0.025 cm^{-1}. TES provides simultaneous measurements of NO_y, CO, O_3, and H_2O.