Fig. 7.7: Zenith atmospheric opacity in the microwave band under typical midlatitude conditions, including the effects of a moderately thick nonprecipitating cloud layer.

Figure 8-7. Absorption spectrum of atmospheric oxygen from 1 to 300 GHz for two pressures (1 bar and 0.1 bar) and at a constant temperature of 273 K (from Chahine et al., 1983).
Sound (verb), to explore or examine, probe.

Basic idea: atmospheric emission is a function of the temperature profile of the atmosphere.

For a multi-look radiometer, the measured brightness temperature is a weighted sum of brightness originating at different depths in the atmosphere.

Note the following:

- There is strong emission of microwave radiation in the lower atmosphere BUT this emission must traverse the entire atmosphere so there is strong attenuation.
- There is weak emission at higher altitudes BUT less attenuation.

Implication: at each frequency there will be a specific layer (or region) in the atmosphere that contributes the most to the total brightness temperature, the layer that has the optimal combination of source strength (gas density) and attenuation above it.

\[
TB^+ (\infty, \mu) \quad \text{brightness temperature observed by satellite,} \quad TB^+(0, \mu) e^{-\frac{\sigma(0,\infty)}{\mu}} + \int_{-\infty}^{\infty} f_k(z^1) T(z^1) e^{-\frac{\sigma(z^1,\infty)}{\mu}} dz^1
\]

\[
\int_{-\infty}^{\infty} f_k(z^1) T(z^1) e^{-\frac{\sigma(z^1,\infty)}{\mu}} dz^1 = \text{atmospheric emission measured by a multi-look radiometer}
\]

\[
W(f_1 \mu) = \text{weighting function} = f_k(\mu, f_1) e^{-\frac{\sigma(f_1, \mu)}{\mu}}
\]

If we know several \(W(f_1 \mu)\), each with different "sweet spots" or layers of the atmosphere to which they are most sensitive, then we can infer \(T(z^1)\).
Ideal Atmospheric Sounding

Unit impulse or "delta function."

\[ s(t) = 0 \quad t \neq 0 \quad \int_{-\infty}^{\infty} s(t) \, dt = 1 \]

Sampling property \( \delta \) unit impulse function.

\[ \text{If } \phi(t) \text{ is some function } s(t) \text{, then } \int_{-\infty}^{\infty} \phi(t) \delta(t) \, dt = \phi(0) \int_{-\infty}^{\infty} \delta(t) \, dt = \phi(0) \]

Consider this ideal weighting function.

\[ w(f_0, z) = s(z - z_0) \]

\[ T_{B_\infty}(f) = \text{atmospheric emission at frequency } f = \int_{0}^{\infty} w(f_0, z) T(z) \, dz = \int_{0}^{\infty} s(z - z_0) T(z) \, dz = T(z_0) \]

\[ T_{B_0}(f) = T(z_0) \]

the brightness temperature of the atmospheric emission at frequency \( f \) corresponds to the atmospheric temperature at \( z = z_0 \)!
Basic principle of sounding illustrated

Simple case: $N_0 = \text{molecules} \times \text{atmosphere}$

Then $k_n(f, z) = k_0(f) e^{-\frac{z}{H}}$

Since $N_0 = p_0 e^{-\frac{z}{H}}$

So, $T_R^+ = \int k_n(f, z) T(z) e^{-\frac{z}{H}} T(z) e^{-\frac{z}{H}} dz = \int k_0(f) e^{-\frac{z}{H}} T(z) e^{-\frac{z}{H}} dz$

$= \int k_0(f) T(z) e^{-\frac{z}{H}} - k_0(f) T(z) e^{-\frac{z}{H}} dz = \int W(f, z) T(z) dz$

where $W(f, z) = k_0(f) e^{-\frac{z}{H}} - k_0(f) T(z) e^{-\frac{z}{H}} = \frac{\text{Im}(f)}{H} e^{-\frac{z}{H}} - \text{Im}(f) T(z) e^{-\frac{z}{H}}$

and $\text{Im}(f) = k_0(f) H$ which determines the layer of the atmosphere that contributes the most to the total brightness temperature.

418 ATMOSPHERIC REMOTE SENSING IN THE MICROWAVE REGION

Figure 9-3. Behavior of the weighting function $W(f, z)$ corresponding to an exponentially decaying atmosphere for different values of the $e_m$ for a down-looking sensor.

Altitude of layer that accounts for the maximum contribution to the brightness temperature $= 2m(f) = H \log(\text{Im}(f))$.

So use different frequencies to sound different altitudes!
Fig. 8.4. Weighting functions for channels 4-14 of the Advanced Microwave Scanning Unit (AMSU).

(a) Magnetic equator, vertical incidence.

(b) O$_2$ near 60 GHz.

(c) O$_2$ near 118 GHz.
Figure 9-3. Behavior of the weighting function $W(v, z)$ corresponding to an exponentially decaying atmosphere for different values of the $\tau_m$ for a down-looking sensor.

Figure 9-10. Unnormalized weighting functions for temperature as a function of height above the surface for observations from the surface looking at the zenith. The curves correspond to the emission by oxygen near the 60 GHz region (from Pearson and Kreiss, 1998).
Fig. 7.7: Zenith atmospheric opacity in the microwave band under typical midlatitude conditions, including the effects of a moderately thick nonprecipitating cloud layer.

Figure 8-7. Absorption spectrum of atmospheric oxygen from 1 to 300 GHz for two pressures (1 bar and 0.1 bar) and at a constant temperature of 273 K (from Chahine et al., 1983).
Figure 9-7. Behavior of a pressure-broadened spectral line as the pressure changes. The upper curves show the change of the absorption as a function of pressure for three different frequencies.

Figure 9-6. Normalized weighting function curves for water-vapor density in the atmosphere at three representative frequencies near and on the 22.235 GHz resonance of water vapor. The curves are derived for brightness temperature measurements from the surface of the Earth. (From Staelin, 1968.)
As we have seen in Sections 7.4.1 and 8.3.1, certain atmospheric constituents — most notably CO₂, water vapor, and oxides, and oxygen — are associated with strong absorption lines and bands that are well mixed throughout the troposphere and stratosphere. That is, they are present at a constant, accurately known mass ratio to all other constituents, such as CO₂ and O₂. The atmospheric opacity over certain ranges of wavelengths is therefore very small.

Figure 8.4 depicts the physical basis for profile retrieval at three levels of idealization, starting with the simplest — and least realistic — case in which the weighting functions are not only represented by single, sharp lines but also overlap. In such a case, the observed intensities will be the summation of contributions from all layers, weighted by the corresponding absorptions. The observed profile and compute the associated intensities for each channel using (8.29). You would then compare the computed intensities with the observed intensities and adjust the profile in such a way as to reduce the discrepancies. This process could be repeated until the differences for all channels fell to within some tolerance, based on the assumed precision of the instrument measurements themselves and of the model calculations of $I_0$.

The above procedure (with certain important caveats; see below) is in fact not too far from what is actually used in routine satellite-based temperature profile retrievals. It is not my purpose here to embark on a rigorous discussion of remote sensing theory, which is best left for a separate course and/or textbook. It is enough for now that you recognize the close connection between the radiative transfer principles discussed earlier in this chapter and an application of immense practical importance to modern meteorology.
tic — on the left: If weighting functions happened to be perfectly sharp — that is, if all emission observed at each wavelength \( \lambda_i \) originated at a single altitude (Fig. 8.4a), then “inverting” the observations would be simple: in this case, the observed brightness temperatures \( T_{B,i} \) would exactly correspond to the physical temperatures at the corresponding altitudes \( h_i \). Your job is then essentially finished without even lifting a calculator. Of course, you wouldn’t know how the temperature was varying between those levels, but you could either interpolate between the known levels and hope for the best or, if your budget was big enough, you could add an arbitrary number of new channels to your sensor to fill in the vertical gaps.

Slightly more realistically, panel (b) depicts the weighting functions as having finite width, so that the observed brightness temperatures correspond to an average of \( B_i(T(z)) \) over a substantial depth of the atmosphere rather than a unique temperature \( T_i \) at altitude \( h_i \). There is now ambiguity in the retrieval, because there is no single atmospheric level that is responsible for all of the emission measured by any given channel. At best, you can estimate the average layer temperature associated with each channel. Nevertheless, the profile retrieval problem itself remains straightforward, because each channel contains completely independent information: there is no overlap between the weighting functions.

Unfortunately, real weighting functions are constrained to obey the laws of physics, as embodied in (8.30). This means that unless you have an unusually sharp change with altitude in the atmospheric absorption coefficient \( \beta_\lambda \), your weighting functions will be quite broad. In the worst case, the mass extinction coefficient \( k_\lambda \) of your chosen constituent will be nearly constant with height, so that the weighting functions will be essentially those predicted for an exponential absorption profile as discussed in Section 7.4.3. The situation is somewhat better for sensor channels positioned on the edge of an absorption line (or between two lines), because pressure broadening (see Chapter 9) then increases \( k_\lambda \) toward the surface, which sharpens the weighting function. Nevertheless, the improvement is not spectacular.

Therefore, given any reasonable number of channels, there is always considerable overlap between adjacent weighting functions, as depicted schematically in Fig. 8.4c. In fact, Fig. 8.5 shows actual weighting functions for the Advanced Microwave Sounding Unit (AMSU), which has 11 channels on the edge of the strong \( \mathrm{O}_2 \) absorption band near 60 GHz (c.f. Fig. 7.7). Although each satellite sounding device has its own set of channels and therefore its own unique set of weighting functions, those for the AMSU are fairly typical for most current-generation temperature sounders in the infrared and microwave bands.
To summarize: It is clear on the one hand that there is information about vertical temperature structure in the radiant intensities observed by a sounding instrument like the AMSU. On the other hand, one shouldn’t underestimate the technical challenge of retrieving temperature profiles of consistently useful quality from satellite observations. In outline form, here are the main issues:

- In general, it takes far more variables to accurately describe an arbitrary temperature profile $T(z)$ than there are channels on a typical satellite sounding unit. This means that you have fewer measurements than unknowns, and the retrieval problem is underdetermined (or ill-posed). The problem is therefore not just that of finding any temperature profile that is consistent with the measurements; the real problem is of choosing the most plausible one out of an infinity of physically admissible candidates.

- Because of the high degree of vertical overlap between adjacent weighting functions, the temperature information contained in each channel is not completely independent from that provided by the other channels. That is to say, if you have $N$ channels, you don’t really have $N$ independent pieces of information about your profile; you have something less than $N$, which makes the problem highlighted in the previous paragraph even worse that it appears at first glance.

- Because any measurement is inherently subject to some degree of random error, or noise, it is important to undertake the retrieval in such a way that these errors don’t have an excessive impact on the final retrieved profile.

- Because of the large vertical width of the individual weighting functions, a satellite’s view of the atmosphere’s temperature structure is necessarily very “blurred” — that is, it is impossible to resolve fine-scale vertical structure in the temperature profile. As a consequence, a great many of the candidate profile solutions that would be physically consistent with the observed radiances $I_{\lambda,j}$ exhibit wild oscillations that are completely inconsistent with any reasonable temperature structure of the atmosphere. Effectively weeding out these bad solutions while retaining the (potentially) good ones requires one to impose requirements on the “smoothness” of the retrieved profile or limits on the allowable magnitude of the departures from the “first guess” profile.

Of course, well-established methods exist for dealing with the above challenges, and satellite temperature profile retrievals are successfully obtained at thousands of locations around the globe every day. These retrievals provide indispensable information about the current state of the atmosphere to numerical weather prediction models. Without the availability of satellite-derived temperature structure data, accurate medium- and long-range forecasts (three days and beyond) would be impossible almost everywhere, and even shorter-range forecasts would be of questionable value over oceans and other data-sparse regions.

8.3.3 Water Vapor Imagery

In the previous subsection(s), we looked at the case that satellite observed emission was associated with a constituent that was well mixed in the atmosphere. Under that assumption, the vertical brightness temperature associated with a constituent such as water vapor in the atmosphere is essentially constant in temperature distribution temperature of the absorber. Under that assumption, the vertical brightness temperature of the absorber is essentially constant in temperature distribution temperature at both time and space. Such is the case for infrared images acquired within the water vapor band centered at 6.3 μm.

The wavelength in this band most commonly utilized for satellite imaging is 6.7 μm, where water vapor absorption is strong enough to block surface emission from reaching the satellite. However, the small amounts of water vapor found in the stratosphere and upper troposphere.