

Thermal infrared remote sensing of the atmosphere

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Thermal infrared (TIR, 200 to 2000 cm^{-1}) sensors on satellites provide global and diurnal measurements of the Earth's top-of-the-atmosphere (TOA) radiation. These sensors are most sensitive to the mid- and upper tropospheric regions of the atmosphere, where monitoring of greenhouse gases and pollution plumes has become increasingly important. This Brief describes how thermal infrared radiation can be used to determine atmospheric concentrations from space.

The spectral distribution of Earth's thermal radiation is governed by Planck's law, which describes the intensity of radiation, $B(\lambda, T)$, emitted by a black body (a perfect absorber and emitter of radiation) in thermal equilibrium as a function of wavelength (λ) and temperature (T). For a given temperature, the peak intensity of radiation for each Planck function is related to a particular wavenumber (coloured lines in **Figure 1**) (T). In reality, different types of surfaces, such as land, ocean, and snow, do not emit and absorb radiation efficiently: they are not perfect black bodies. This property is described by the emissivity (ϵ), the ratio of the radiation emitted by an object to the radiation that it would emit if it were a perfect emitter/absorber. A typical Earth spectrum, with a particular surface temperature and emissivity, observed from a TIR space sensor would therefore look like the black line in **Figure 1**. This spectrum is not smooth like the Planck curves, but rather exhibits a series of absorption lines and features. These lines contain the spectral fingerprint of many molecular species in the atmosphere. The region from 800 cm^{-1} to 1000 cm^{-1} is often called the "atmospheric window", where the atmosphere appears transparent. This region can be used to gain information on the Earth's surface, clouds and aerosols (2).

Figure 2 shows how different atmospheric species contribute to the radiation intensity observed in a TIR

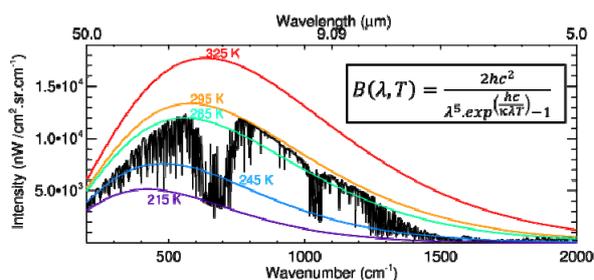


Figure 1. Intensity of the Planck radiation spectrum, $B(\lambda, T)$, for a range of terrestrial temperatures (coloured lines) and a simulation of a typical atmospheric spectrum (black line). In Planck's law (top right), k is Boltzmann's constant, h the Planck's constant, and c the speed of light.

spectrum. Some species such as ozone (O_3) and methane (CH_4) have strong spectral bands. Other species like carbon dioxide (CO_2) and water vapour (H_2O) have a broader series of spectral lines (or "continuum" spectra) present throughout most of the TIR spectral region.

The intensity of TIR radiation and the type of feature observed in a TIR spectrum is dictated by a number of processes. Firstly, absorption (or emission) spectral lines are only produced when a photon is absorbed (or emitted), creating a transition of energy states on a molecular level (3, 4). In the TIR region, such spectral lines correspond to particular molecular transitions that are described as rotational, vibrational or simultaneous rotational-vibrational (5). Secondly, photons can only be absorbed to produce vibrational or rotational transitions if the charge around the molecules is separated so as to produce a permanent electric dipole. Thus, the geometry of the molecules and the abundance of the species in the atmosphere affect the type of spectral feature observed. Now that we recognise what a TIR spectrum looks like, how can we relate interactions between molecules and TIR radiation to the radiation detected by a space-borne sensor?

Radiative transfer

Imagine a beam of radiation travelling through a small section of air. The air is made up of changing

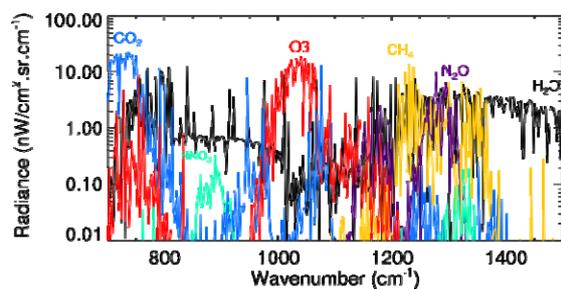


Figure 2. TIR contribution of a range of atmospheric species. The species labels are colour-coded to match the spectral features for each species.

concentrations of different species, with all molecules absorbing and emitting thermal radiation at different rates. As the radiation travels through different layers of the atmosphere, the intensity of radiation will constantly be modified by both absorption and emission processes, as given by

$$\delta I = \delta I_{\text{absorption}} + \delta I_{\text{emission}}$$

where δI is a net change in radiation. This equation is a simplification of the *Schwarzschild's Equation*, which tells us how radiation changes through its travelled path due to all of simultaneous absorption and emission processes. Now imagine that this radiation is detected by a TIR sensor, such as a Fourier Transform Spectrometer (6), looking down at the Earth along a particular line of sight. The net radiation measured by the satellite sensor would be that which is attenuated through each layer (as small increments of absorption and emission) from the surface to the top of the atmosphere *plus* the radiation emitted directly from the surface. In this case, this process can be described by the radiative transfer equation (RTE):

$$I(\text{sensor}) = I(\text{surface}) + I(\text{atmosphere})$$

In remote sensing, I is usually called the radiance (with units $\text{W sr}^{-2} \text{m}^{-2}$). The equation is shown here for a very simplistic scenario. In reality the atmosphere is highly scattering (an effect often assumed to be negligible in the TIR unless clouds and aerosols are present). In these cases, the RTE becomes much more complex to account for additional processes affecting the radiation (I).

Retrieving atmospheric species

In practice, information about an atmospheric species is *retrieved* using a computer algorithm to solve the RTE. The algorithms use an inversion technique, i.e. they mathematically invert a TOA satellite radiance measurement to infer the atmospheric state.

The problem of solving the RTE to determine an

atmospheric concentration is imperfectly defined, because the satellite measurement only provides a snapshot of information about the atmosphere and we must make assumptions about the atmospheric and surface state for that time. There is thus no unique solution, and only the most probable solution can be determined within some uncertainty range (7). Generally, for TIR atmospheric retrievals, an optimal estimation (OE) technique is employed. A typical OE retrieval algorithm requires calibrated spectra from a satellite measurement (covering the spectral range which has the largest sensitivity to the target species); once spectra are extracted, suitable prior data are selected. The prior data are used to simulate (with a radiative transfer model) the satellite measurement and when a good simulation is obtained, the algorithm will solve relevant equations to extract the target species and an uncertainty estimate.

For the retrieval to be as accurate as possible, the prior data (i.e. vertical profiles of temperature and pressure, the target species, any interfering species, surface temperature and emissivity) should match the measurement time and location as closely as possible. The simulation must incorporate up-to-date spectroscopic line and cross section information. A poor representation of prior data, spectral lines and cross sections will result in an inadequate simulation compromising the accuracy of the retrieved quantity. For specific examples of remote sensing of different tropospheric species in the Earth's atmosphere see (8).

References

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