DETECTION OF CLOUDS AND AEROSOLS USING INFRARED HYPERSPECTRAL OBSERVATIONS

Steven A. Ackerman¹, Hung-Lung Huang, Paolo Antonelli, Robert Holz, Henry Revercomb, David Tobin, Kevin Baggett, Jim Davies

Cooperative Institute for Meteorological Satellite Studies (CIMSS) University of Wisconsin, Madison, Wisconsin

1. INTRODUCTION

High spectral resolution infrared measurements provide increased capabilities over imaging radiometers for detecting clouds and aerosols and for characterizing their optical properties. For over 20-years the University of Wisconsin Cooperative Institute for Meteorological Satellite Studies (UW-CIMSS) has been developing algorithms for detecting clouds and aerosols using high spectral resolution infrared measurements (Revercomb *et al.*, 2004). The goal has been to apply these algorithms to current and future hyperspectral sensors on polar orbiting and geostationary satellite platforms with a view to improving our understanding of the Earth's hydrological cycle and energy budget.

This paper reviews the advantages and disadvantages of using high-spectral resolution infrared observations to detect cloud and aerosol and, where feasible, retrieve their microphysical properties. Presented is an overview of these algorithms and a demonstration of their capabilities as applied to aircraft (HIS/S-HIS/NAST-I) and satellite (AIRS) instruments.

Cloud retrievals can generally be broken down into four processes: cloud detection, cloud altitude (or emitting temperature), cloud emissivity, and cloud microphysical properties (Figure 1). While all of these cloud properties can be retrieved simultaneously, for simplicity this section discusses each step separately.

2. CLOUD DETECTION

Various approaches to cloud detection have been proposed (e.g. Revercomb et al., 2004; Ackerman *et al.*, 1990; Huang *et al.*, 2003; DeSlover *et al.*, 1999). Some cloud detection approaches rely directly on the radiance measurements, comparing observed radiances with expected radiances from a clear sky scene. Others compare the observed radiance measurements to clear sky simulations.

Cloud detection is a function of the contrast between cloud and clear scenes. Contrast variations can be in space, time or spectral domain. The high spectral resolution capability improves spectral contrast. Cloud detection using passive infrared radiometric observations is difficult in scenes with low temperature

¹ University of WI-Madison 1225 West Dayton St Madison WI 53706 contrast, such as low clouds in polar winter conditions. Figure 2 demonstrates the advantage of high spectral resolution measurements for detecting stratus clouds over ice. The HIS instrument made these observations during the FIRE-ACE campaign. Narrow spectral regions were averaged for a few wavenumbers at 10.4 μm (958-962 cm⁻¹), 12.7μm (786-790 cm⁻¹), 8.7μm (1150-1160 cm⁻¹). Observations near these three wavelengths have been used to attempt to separate ice clouds from liquid water clouds. Figure 2 is a time series of the corresponding brightness temperatures of a scene of stratus clouds over ice. Because of the high spectral resolution, we can average over smaller spectral regions and apply the technique to distinguish water clouds from the ice below. Also shown is the brightness temperature for 3.7µm (2602-2700 cm⁻¹) that includes solar reflection and provides a means of validating scenes with cloud (warm BT) from clear ice scenes.



Figure 1. Flow diagram for retrieving cloud properties

An approach to cloud detection over cold surfaces is to look for inversions using on-line/off-line spectral measurements (Tanamachi, 2001). This is a good approach for detecting clouds in polar winter conditions, as the surface is often colder than the atmosphere. This on-line/off-line approach to cloud detection over the Polar Regions has been applied to IMG observations. The method is graphically depicted in Figure 3. The two brightness temperature spectra are from two views taken over the Barrow AK region, from two different IMG (Interferometric Monitor for Greenhouse Gases) sequences, showing the inversion present in clear scenes (the on-line and off-line regions are boxed.) The spectral measurements detect the presence of the surface temperature inversion as seen in the larger brightness temperature on the absorption line than off the line. The on-line feature shows a colder brightness temperature in the cloudy spectra.



Figure 2. Detection of stratus over ice using observations from HIS during FIRE-ACE. Stratus clouds occur in regions where the $BT_{12.7}$ is more similar to the $BT_{10.4}$ than the $BT_{8.7}$.



Figure 3. Two brightness temperature spectra from two sky views taken over the Barrow, AK region from two different IMG sequences, showing the inversion present over clear scenes for an absorption line and off-line.

Based on this experience with HIS, SHIS and IMG, we have developed a simplified cloud mask and applied it to AIRS observations. The algorithm is a combination of threshold tests based on infrared brightness temperatures. To validate the approach we have compared the AIRS cloud amount to the cloud amount derived from collocated MODIS observations.

To conduct the comparison each AIRS FOV were represented as circles with a diameter ranging from 16km at nadir to 25-km at a 50 degree scan angle. The count of 1-km MODIS pixels within each AIRS FOV was determined and classified using the MODIS cloud mask algorithm (cloudy, uncertain, probably clear and confident clear). We then created AIRS clear flags using thresholds on fractions of clear MODIS pixels as shown in Figure 4.



Figure 4. Collocation of AIRS (circles) and MODIS cloud mask results.

The results are compiled in Figure 5. When the AIRS cloud mask said it was clear, approximately 72 percent of the time 90-100 percent of the MODIS pixels within those AIRS FOVs were determined to be confident clear (green) by the MODIS cloud mask, approximately 12 percent of the time 90-100 percent of the MODIS pixels within those AIRS FOVs were determined to be probably clear (cyan) by the MODIS cloud mask, approximately 0 percent of the time 90-100 percent of the MODIS pixels within those AIRS FOVs were determined to be uncertain (red) by the MODIS cloud mask, and approximately 3 percent of the time 90-100 percent of the MODIS pixels within those AIRS FOVs were determined to be uncertain (red) by the MODIS cloud mask, and approximately 3 percent of the time 90-100 percent of the MODIS pixels within those AIRS FOVs were determined to be cloudy (white) by the MODIS cloud mask.



Figure 5. A comparison of MODIS and AIRS cloud mask results.

3. CLOUD ALTITUDE

There are a few approaches to deriving cloud altitude using passive infrared measurements. The CO₂ slicing algorithm has been well explained and studied (Frey et al., 1999). Cloud altitude determination using the CO₂ slicing algorithm is applied routinely to MODIS observations. CO₂ slicing using high spectral resolution observations can produce a better estimate due to the higher spectral resolution, though imagers tend to produce a product at higher spatial resolution. А different algorithm using the CO2 absorption band between 650-800 wavenumbers (Holz et al., 2004) has recently been developed. A selected clear sky scene is sorted relative to channel brightness temperature and a sorted index is created. The result is a smoothly increasing function of brightness temperature starting with the coldest most opaque channels to the warmest transparent channels in the CO₂ band. In essence, the sorting has ordered the channels by the atmospheric level at which the channels weighting function is peaked. When the sorted index is applied to a cloud filled field of view the cloudy sorted spectrum will deviate from the clear sky sorted spectrum at channels whose weighting functions peak at or below the cloud attitude. The brightness temperature of this inflection point is related to the cloud emitting temperature. Using an atmospheric temperature profile the cloud top altitude can be inferred.

Figure 6 is a comparison of the different infrared passive retrieval approaches with measurements from a lidar. The capabilities of each approach are a function of the cloud optical thickness. During the CRYSTAL-FACE field experiment the NAST-I (Blackwell et al., 2001) and Cloud Physics Lidar (CPL) (McGill et al., 2002; Spinhirne et al., 1996) flew on different platforms. The NAST-I flew on the Proteus while the CPL flew on the ER2. To evaluate the performance of the cloud top altitude retrieval algorithms applied to the NAST-I, a comparison with lidar sensed cloud top altitudes is necessary. Because the instruments flew on different platforms during CRYSTAL-FACE a direct comparison was not possible. Instead, data period histograms of the frequency of occurrence of cloud top altitude retrieved using the HIS algorithms (CO₂ slicing, CO₂ sorting, and MLEV) was compared to the frequency of occurrence of the cloud top altitude derived from CPL during THORPEX between Feb 22 01:07:00 - 01:45:00 UTC. The CPL data was collocated with the SHIS so each data point in the CPL histogram is the average of all the CPL shots that fell inside the SHIS field of view. On average, this is about 8 shots per SHIS field of view.

The results are presented in Figure 7. The CPL has a narrow distribution of sensed cloud heights, with ice clouds ranging from 10.5 to 13 km. The CO_2 slicing approach has the widest distribution and tends to underestimate the cloud heights when compared to the CPL results. This is consistent with previous studies. The MLEV compares well with the clouds tops in the range of 11 and 12 km, though there is a frequent retrievals at 14 km, which is not observed by the CPL. When the CPL distribution is compared to the CO_2 sorting it is apparent that the CO_2 sorting shows a tendency to overestimate the cloud height sensed by the CPL.



Figure 6. A comparison of cloud top retrievals from three infrared approaches with lidar. High-spectral resolution measurements were made by the S-HIS during THORPEX. Lidar data was kindly provided by Matt McGIII of GSFC.



Figure 7. Comparison of three passive and one active remote sensing of cloud top.

4. CLOUD EMISSIVITY

Once the cloud altitude is known, first the temperature then the cloud emissivity is determined. Cloud altitude and effective emissivity can also be estimated using the Minimum Local Emissivity Variance (MLEV) algorithm, which takes advantage of high-spectral resolution observations in the 8-12 micron region. Neglecting scattering, the emissivity is defined as:

$$N\boldsymbol{e}_{n} = \frac{\boldsymbol{I}_{n} - \boldsymbol{I}_{n}^{clr}}{\boldsymbol{I}_{n}^{blc} - \boldsymbol{I}_{n}^{clr}}, \qquad (1)$$

where N is the cloud fraction, I is the radiance for the cloudy scene, superscript *clr* labels a clear radiance,

$$I_{\mathbf{n}}^{blc} = B_{\mathbf{n}}(T_{cd}) t(p_{cd}) - \int_{0}^{p_{cd}} B_{\mathbf{n}}(p) dt(p), \quad (2)$$

is the radiance that would result from a black cloud with effective cloud top temperature T_{cd} at pressure p_{cd} for the same atmosphere. In the definition of the black cloud radiance, *B* is the Planck function and *t* is the transmission from pressure *p* to space. The clear radiance can either be obtained from a clear observation or calculated from a nearby temperature and water vapor retrieval or from a radiosonde measurement. The black cloud radiance is calculated using the same atmospheric state.

Errors in cloud emissivity are introduced from uncertainty in the water vapor and temperature vertical profiles, the surface emissivity, vertical cloud and aerosol distribution and emissivity, and spatial heterogeneity over the studied area. The error on effective emissivity is defined by the error in the radiance (either measured or calculated), which depends on the reference values and measurement error. Mathematically:

$$\frac{\Delta \mathbf{e}}{\mathbf{e}} = \frac{\Delta \mathbf{I}}{(\mathbf{I}^{\text{thar}} - \mathbf{I})} + \frac{\mathbf{I} - \mathbf{I}^{\text{thad}}}{(\mathbf{I}^{\text{thar}} - \mathbf{I})} \Delta \mathbf{I}^{\text{thar}} + \frac{\Delta \mathbf{I}^{\text{thad}}}{(\mathbf{I}^{\text{thar}} - \mathbf{I}^{\text{thad}})} \quad (3)$$

As an example, a +/- 0.5 K error has been estimated in the equivalent brightness temperature, either due to a noise or calibration error (including noise and bias in modeling the radiances). Calculations of emissivity uncertainty plotted against cloud emissivity and wavenumber at 225K shows the potential error in magnitude as well as by percent.

The uncertainty in the emissivity calculation is a function of many variables, including emissivity, wavelength. cloud effective temperature. and measurement uncertainties of clear radiance, cloudy radiance, and cloud temperature. For these figures, measurement uncertainties are parameterized to characteristic values so we can consider the error as a function of wavenumber (wavelength) and cloud emissivity. The errors are a strong function of emissivity, becoming very high (several times the measured values) for low emissivity cases. In fact, the errors become even greater for warmer cloud layers, with errors nearly doubling for cloud temperatures of 255K. As a result, this method can only be used reliably for higher emissivity cases, say above 0.1 where errors are generally less than 30%. It is this effect that leads to the spurious sub-zero values often retrieved.



Figure 8. The magnitude of the derived emissivity uncertainty as a function of wavenumber and cloud emissivity, assuming a cloud temperature of 225K.

The impact of selecting cloud effective temperature on the retrieved emissivity is demonstrated in Figure 9, which is a scatter plot of calculated emissivity versus cloud thickness. The blue dots are the retrieved cloud emissivity using the cloud mean temperature as determined by the cloud lidar and a near by radiosonde. The red error bars represent the error analysis as discussed above. Generally, smaller errors are associated with high emissivity cases and the relatively large errors associated with smaller emissivities and thicker clouds. The green x's represent the retrieved emissivity assuming the cloud top temperature and demonstrate the importance of accurately determining cloud effective temperature in the retrieval.



Figure 9. Scatter plot of calculated emissivity vs. cloud thickness using cloud mean temperature is represented by the blue dots. The red error bars represent the error analysis as discussed above. The generally smaller errors associated with high emissivity cases and the relatively large errors associated with smaller emissivities and thicker clouds are evident.

5. CLOUD MICROPHYSICAL PROPERTIES

Methods to derive cirrus particle size have also been developed using a combination of measurements in the 8-12µm window region. Ackerman et al. (1990) used HIS observations to characterize particle sizes as large, small, and intermediate. Smith et al. (1993 and 1998) and Chung et al. (2000) retrieved particle size by matching observed radiance with theoretical calculations. Observations at high-spectral resolution in the infrared are particularly valuable as the bulk transmittance of water and ice displays a strong spectral absorption variation in the 800 to 1200 cm^{-1} (8 to $12 \mu \text{m}$) region. This is also a spectral region over which the atmosphere is fairly transparent. This window region has been used to detect cloud water phase (Ackerman et al., 1990; Baum et al., 2000; Strabala et al., 1994). The sensitivity of the infrared spectral observations to cloud water path has been demonstrated and algorithms have been developed. Retrievals of cloud water path require knowledge of the clear-sky radiance obtained from the observations or computations using representative temperature and gas concentration profiles. Retrieval of the cloud microphysical properties using high-spectral resolution IR measurements is approached through minimization of the differences between observed and calculated radiances in 'superwindows' (Smith et al., 1998; Sunggi et al., 2000; Ackerman et al, 1990; Ackerman et al., 1998; DeSlover et al., 1999; Smith and Frey, 1990). The spectral variation across this window region due to the presence of small ice crystals results in large brightness temperature variations. This is demonstrated in Figure 10, made by the HIS during the SUCCESS experiment.



Figure 10. HIS observations of clear and cirrus clouds during the SUCCESS program demonstrating the impact of small crystals on the window brightness temperature.

6. AEROSOL DETECTION

Detection is often expressed in terms of spectral variations of the brightness temperature. The brightness temperature is determined by converting the observed radiance to a blackbody temperature. Detection of clouds and aerosols is common using brightness temperature differences between approximately 8.5, 11, and 12 μ m (BT₈, BT₁₁, and BT₁₂, respectively).

Aerosol detection techniques using infrared wavelengths have a distinct advantage over visible techniques by providing coverage over bright surfaces and during the night. The detection of volcanic and soil-derived aerosols using infrared at the approximate wavelengths 8.5, 11 and 12 μ m is particularly useful. The spectral variation of the index of refraction plays an important role in the success of these methods. Measurements of the index of refraction of a material are very difficult to make (Bohren and Huffman, 1983).

The scattering and absorption properties of an aerosol also depend on the particle size distribution. The real and imaginary parts of the index of refraction of carbon are independent of wavelength. Thus, spectral difference techniques may not be very successful in detecting aerosols with a large carbon component, unless the aerosol is optically thick or the surface below the aerosol layer very hot. Measurements in the region between 8 and 12 μ m are useful for detecting aerosols containing large components of quartz, hematite, and sulfuric acid.

Volcanic debris can be a serious hazard to air traffic. This type of debris has a signal similar to that of sand and dust with negative values in the brightness temperature difference BT₁₁–BT₁₂. Algorithms to detect the presence of volcanic ash plumes have been developed and can be applied as real-time algorithms. Split window IR techniques have primarily been applied to volcanic aerosols, particularly those from sulfur-rich eruptions (e.g. Barton et al. 1992). Several aerosol remote sensing techniques have been developed using observations from the Advanced Very High Resolution Radiometer (AVHRR). The volcanic ash plumes often generate negative brightness temperature differences between BT₁₁ and BT₁₂. Prata (1989) has demonstrated the detection of volcanic aerosols using two infrared channels. Figure 11 is a MODIS RGB image of the eruption of Mt Etna on 28 October 2002. An example application using high-spectral resolution IR is shown in Figures 12 and 13. The ash plume is clearly visible in the brightness temperature differences of BT₈₆₂ - BT₉₄₂ and $BT_{1392} - BT_{1345}$. The first brightness temperature difference is associated with a soil-derived aerosol.



Figure 11. MODIS image of the Mt. Etna eruption on 28 Oct. 2002. The left panel is an RGB image and the right image is a brightness temperature difference image.

Detection of smoke using infrared measurements alone has not been successful because of the small spectral variation in the absorption of the smoke. Highspectral resolution infrared observations allow the advantage of looking for fires or trace gases produced by the smoke. Figure 14 is an example of the fire spectral feature observed during the second ER-2 overpass. The spectral feature at 2395 cm⁻¹ coincides with the presence of a fire in the S-HIS field-of-view (FOV). Figure 15 locates the reference FOV on the MODIS Airborne Simulator (MAS) image. The target fire, with its attendant smoke plume, can be clearly observed near the center of the image. A larger, older burn scar lies to the north of the image (see inset). The S-HIS observation path passes directly down the center of the MAS image. MAS images such as this were crucial to identifying the regions of interest present in S-HIS SAFARI data. Absorption by atmospheric CO2 in the 2300 cm⁻¹ to 2380 cm⁻¹ spectral region (i.e. 4.2 μm to 4.35 µm) is 100 % over long path lengths. transparency becomes large at Atmospheric approximately 2400 cm⁻¹. The fire signature occurs in this transition region between strong and weak absorption. This spectral region has been used to examine hot exhaust plumes, and the observed feature near 2400 cm⁻¹ is referred to as the 'blue spike' (Heland and Schäfer, 1997). Looking for this spike is an indication of an active fire and hence smoke.

Biomass burning is a major source of atmospheric trace gases and aerosols. Carbon dioxide emission with an absorption band from 4.2 μ m to 4.5 μ m and carbon monoxide with absorption from 4.45 μ m to 4.95 μ m are both gaseous by-products of fires.



Figure 12. AIR brightness temperature difference $(BT_{862}-BT_{942})$ image of the Mt Etna eruption..



Figure 13 AIRS brightness temperature difference $(BT_{1392} - BT_{1345})$ image of the Mt. Etna eruption.

In this section, we present a treatment of the physical and mathematical aspects of the method used to detect changes in the column amount of carbon monoxide. The primary assumptions are stated and justified for this effective optical depth data product. The current method makes use of high spectral resolution radiance observed by the S-HIS spectrometer to derive an optical depth using weak absorption lines. It is less sophisticated than other approaches, but is very useful for a survey result of localized events.



Figure 14. HIS observations of a fire. The FOV is located on the MAS observations in figure 15.



Figure 15. A MAS image of a control burn during the SAFARI. The white circle indicates the FOV of the HIS measurement shown in Figure 14.

A ratio of on-line to off-line emission for selected lines of the gas of interest provides a measure of the gas amount. The spectral channels selected for this purpose are listed in Table 1. The form of the equation used is derived from a single-layer atmosphere approximation;

$$N_n^{Obs} = t_n \cdot N_n^{Surf} + N_n^{Arm} + t_n \cdot N_n^{Refl}$$

$$t_n = \exp[-t_n] \text{ and } t = \text{optical depth}, \qquad (4)$$

where N^{Obs} is the S-HIS observed upwelling spectral radiance at an altitude of 20 km, t_v is the atmospheric transmission for wavenumber v, N^{Surf} is the emission from the surface, N^{Atm} is the atmospheric emission, N^{Refl} is the contribution from surface reflection. *B* is defined to be the Planck emission function at a temperature T_{Atm} that approximates the mean atmospheric temperature.

In this study T_{Atm} is held to a fixed value of 260K. Neglecting reflected radiances and assuming that the surface emitted radiance is equal for both on-line and off-line spectral channels and using the formal definition of transmission to define an effective optical depth, τ , the resulting expression becomes

$$\Delta \boldsymbol{t} = \boldsymbol{t}_{on} - \boldsymbol{t}_{off} = -\ln[\frac{N_{on}^{Obs} - B_{on}(T_{Atm})}{N_{off}^{Obs} - B_{off}(T_{Atm})}]$$
(5)

This equation allows us to plot time series of changes in the effective optical depth of trace gases using only the radiance observations. This result is susceptible to any dramatic change in the path distance between the emitter and the detector. Changes in the aircraft altitude and surface topography are accounted for but not the presence of thick cloud in the FOV.

Gas	On-line	Off-line
	Wavenumber	Wavenumber
	2150.80	2151.77
	2154.667	2153.698
	2158.05	2159.02
	2158.52	2164.80
CO	2165.29	2166.75
	2165.77	2168.193
	2169.157	2170.126
	2172.54	2173.506
	2173.02	

Table 1. Spectral wavenumbers used to estimate CO column amounts.

As a primary combustion product, an increase in CO column density and thus effective optical depth would be observed when the ER-2 flew over the fire. The retrieved effective optical depths are consistent with this expectation. Figure 16 shows a significant peak in the CO effective optical depth derived from the S-HIS data in each of the three ER-2 overpasses coincident in time with the controlled fire. The peak increase of the CO optical depth in the first overpass is about 40% over that of the background. The peak increase in the second and third overpasses is approximately 100% over the background level.

Methane absorption was also studied in order to provide an independent measure of the uniformity of the atmospheric column during the overpasses. The methane region 1230-1240 cm⁻¹ is used here as a measure of the homogeneity of the atmosphere since it is a gas that is not a significant by-product of the combustion process. The retrieved methane effective optical depth was not observed to vary significantly over any of the three overpasses of the fire site. This uniformity of methane optical depth can also be seen as an indicator that the mean atmospheric temperature assumed in our algorithm is sufficiently constant over the area of interest. Figure 14 shows the results of the optical depth algorithm for both carbon monoxide and methane. For comparison to the fire detection method, Figure 14 also includes a time series of the mean brightness temperatures for the region 2390-2395 cm⁻¹.

7. SUMMARY

The University of Wisconsin Space Science and Engineering Center (SSEC) and the Cooperative Institute for Meteorological Satellite Studies (CIMSS) have collaborated for over 20 years to develop new observations of the atmosphere using high-spectral resolution measurements in the infrared region. Instrument developed at UW include the HIS, SHIS and Observations from these instruments have AERI. improved our ability to observe and thus better understand the atmosphere. The scientific goals of these programs have been to improve the characterization of the atmosphere and its components (including clouds and aerosols) and the surface and develop new methods of analysis of these measurements. This paper has summarized some of the results of applying these measurements to the study clouds and aerosols.



Figure 16 Effective optical depths derived from S-HIS radiance observations for three ER-2 overpasses of a controlled fire on 7 September 2000. Derived carbon monoxide effective optical depths correlate with the observed brightness temperature in the 2392-2397.5 cm⁻¹ spectral range, while methane amounts are insensitive to the presence of the fire.

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