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# A SIMPLE 1DVAR APPROACH TOWARD CLOUDY INFRARED RADIANCE ASSIMILATION

Sylvain Heilliette* and Louis Garand
Meteorological Service of Canada, Dorval, Quebec

## 1. INTRODUCTION

Approximately 75 % of the fields of view (FOV) observed by a typical infrared sounder (FOV size ≈14-17 km) are cloudy (Wylie and Menzel 1999). In order to retrieve information on temperature and water vapor mixing ratio profiles, radiances which are affected by clouds are currently not assimilated by numerical weather prediction (NWP) centers. Some centers only assimilate clear sky observations while others allow the assimilation of channels which are insensitive to low or mid level clouds. To stay on the safe side, criteria for the determination of pixels which are suitable for assimilation tend to be very restrictive. Consequently, there is a substantial lost of information in regions thought to be of highest meteorological interest.

In this work, the assimilation of cloudy infrared radiances is attempted. The proposed method is well suited for the new generation of hyperspectral infrared sounders such as AIRS (Atmospheric Infrared Radiance Sounder) and IASI (Infrared Atmospheric Sounding Interferometer). Here, 1D-variational assimilation experiments using AIRS data are carried out. The approach is simple in the sense that no attempt is made to assimilate cloud information or to retrieve cloud water profiles. Rather, the equivalent cloud top height and emissivity are retrieved and the assimilation proceeds with these two parameters fixed. An important part of the problem is the determination of situations where this simplified treatment of clouds allows reliable atmospheric soundings down to the cloud top and potentially below the cloud.

## 2. CLOUD EMISSIVITY MODEL

The simplest way to model the impact of clouds on infrared radiances is given by the following equation :

$$I_{cld}(\nu)=N\varepsilon(\nu)I_{ovc}(\nu)+(1-N\varepsilon(\nu))I_{clr}(\nu) \quad (1)$$

where $\nu$ is the channel frequency, $I_{cld}$ is the modeled cloudy radiance, $I_{ovc}$ is the overcast radiance corresponding to an opaque cloud, $I_{clr}$ is the clear sky radiance, and $N\varepsilon$ is the effective cloud emissivity. This last parameter is the product of the "geometrical" fractional cloud cover N by the frequency dependent cloud emissivity $\varepsilon$. In the following, all radiative transfer calculations are performed using the fast RTTOV-8 code (see Matricardi et al. 2004 for a detailed description).

Equation (1) is based on the hypothesis that the geometrical depth of the cloud is negligible. This equation, together with the assumption of a frequency independent cloud emissivity, forms the basis of many cloud parameter retrieval methods such as the minimal residual method (see e.g. Eyre and Menzel 1989), or the widely used $CO_2$ slicing (Smith at al. 1978). The latter technique is based on equation (1) using two spectrally closed channels to eliminate $N\varepsilon$ and solve for the cloud top pressure $P_c$. In our implementation for the AIRS instrument (Garand and Beaulne 2004), several estimates are obtained for various pairs of channels and the mean is retained. In the context of cloud parameter retrievals, the assumption of a constant cloud emissivity is valid only if spectrally adjacent channels are used, which is not optimal for such a multi-spectral instrument. The MLEV method, a cloud emissivity and cloud top pressure retrieval method (see Huang et al. 2003) may seem more appropriate because it allows a frequency dependent emissivity. However, this method works best with a nearly continuous spectra between 750 and 950 $cm^{-1}$. This requirement is often not satisfied at NWP centers where only a subset of 281 or 324 AIRS channels is available.

Instead of retrieving the cloud emissivity directly from the cloudy radiance spectra as in the MLEV method, this study takes advantage of a cloud emissivity model. This generalizes the retrieval of cloud parameters to possible mixed phased situations with applicability at all wavelengths and

*Corresponding author address: Sylvain Heilliette
Meteorological Service of Canada,
Dorval, Quebec H9P 1J3, Canada.
Email: sylvain.heilliette@ec.gc.ca

consideration of optical properties.

In the context of the no-scattering approximation often used in the thermal infrared spectral range, the cloud emissivity is given by :

$$N\varepsilon(\nu)=1-\exp[-k_{abs}\,\delta\,\sec\theta] \qquad (2)$$

where $k_{abs}$ is the cloud specific absorption coefficient, $\theta$ stands for the viewing angle, and $\delta$ is the effective cloud depth related to the cloud water content and depth. However, radiative transfer calculations, performed with the widely used DISORT code (Stamnes et al. 1988), have shown that effects of scattering cannot usually be neglected, especially for the shortwave channels. A rigorous treatment of multiple scattering effects is out of reach at the present time in the context of data assimilation in NWP. A simple but efficient way to account for scattering is therefore required. In this study, scattering is accounted for by replacing the absorption coefficient $k_{abs}$ in equation (2) by the modified cloud absorption coefficient $k_{cld}$ following Chou et al. 1999 :

$$k_{cld}=k_{ext}[(1-\omega)+b\,\omega] \qquad (3)$$

where $k_{ext}$ is the extinction (scattering plus absorption) coefficient, $\omega$ the single scattering albedo and b the backscattered fraction. If we follow the common assumption of a Henyey-Greenstein phase function, an analytical expression of b in terms of the asymmetry factor g is obtained (Wiscombe et al. 1976).

The optical properties necessary for $k_{cld}$ calculations (i.e. $k_{ext}$, $\omega$ and g) depend on the size distribution of water droplets and ice particles via a size parameter $r_e$. In this study, we have chosen to use Lindner et al. (2000) parameterization for pure liquid water cloud with $r_e$=11 $\mu$m and Fu et al. (1999) parameterization for pure ice clouds with $r_e$=25 $\mu$m. For mixed phase clouds, an estimation of the liquid water fraction $f_w$ is necessary. In this study, the simple parameterization proposed by Rockel et al. (1991) giving $f_w$ as a function of cloud temperature was used. Given $f_w$, the optical properties of pure ice and water are combined following rules given by Cess (1985).

## 3. CLOUD PARAMETER RETRIEVAL

Experience showed that direct 1DVAR assimilation of cloudy radiances using the cloud emissivity model described in the previous section is very difficult without a good first guess of the cloud parameters. Estimations of these parameters using $CO_2$ slicing are usually relatively accurate. But in some situations, especially for very low clouds, the $CO_2$ slicing method fails to give a meaningful result. In such a situation, an effective cloud height is obtained using a window channel radiance.

Here the aim is to describe another approach to estimate the cloud parameters using all channels together to obtain the best estimate of $P_c$ and $N\varepsilon$. The method is based upon the minimization of the following cost function :

$$J_{cld}(P_c,\delta)=\sum_i\left[\frac{BT^i_{obs}-BT^i_{calc}(P_c,\delta)}{\sigma_i}\right]^2 \qquad (4)$$

where $BT^i_{obs}$ stands for the observed brightness temperature of the ith channel, $BT^i_{calc}(P_c,\delta)$ for the calculated brightness temperature of the ith channel and $\sigma_i$ for the observation error of the ith channel. The procedure of determination of $\sigma_i$ is described in Garand et al. 2005.

This method is similar to the minimum residual method except that it uses a spectrally dependent cloud emissivity model instead of a constant emissivity. It may also be understood as a 1D variational approach without background term since the a priori knowledge on cloud parameters is very low. A similar approach, applied to the GOES longwave channels, is described in Li et al. 2001.

Radiative transfer calculations are performed assuming the background thermodynamic profiles as perfect. Numerical experiments performed have shown that water vapor sensitive channels should be excluded from this first step because the a priori knowledge about water vapor is too poor to be helpful at this stage. Minimization of equation (4) is not as numerically expensive as it may appear. A single call to the RTTOV-8 code allows to compute once and for all $I_{ovc}(\nu)$ and $I_{clr}(\nu)$ for a cloud located at each RTTOV-8 pressure level. A linear interpolation in logarithm of pressure is sufficient to obtain $I_{ovc}(\nu)$ at the required cloud top pressure. The error introduced by this simple interpolation method on the calculated overcast brightness temperature is most of the time lower than 0.1K and never greater than 0.2 K. On average, the minimization process by itself, using the M1QN3 code (Gilbert et LeMarechal 1989), excluding the RTTOV-8 calculation, takes approximately 6 ms per FOV on the Canadian Meteorological Center (CMC) supercomputer IBM POWER4.

## 4. ERROR ANALYSIS, SENSITIVITY STUDY

There are essentially two potential sources of uncertainties in this cloud parameter retrieval method.

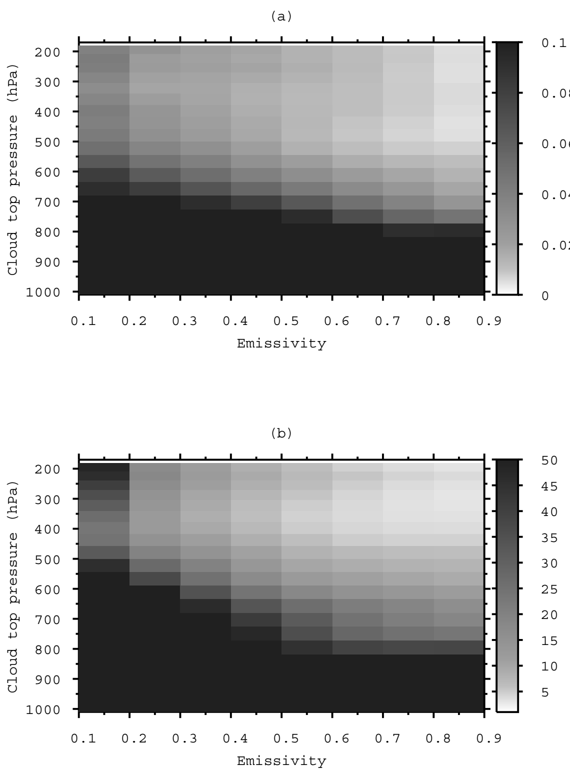


Figure 1 : RMS error for a) 15 μm cloud emissivity b) cloud top pressure resulting from the Monte-Carlo experiments

The first one is linked to the error in the assumed thermodynamic profiles (temperature and water vapor) and surface properties. In order to estimate it, the thermodynamic profiles were perturbed using a multivariate normal random generator of mean 0 and covariance matrix B. In this study, B was chosen as the background error covariance matrix used in the operational model of the CMC described in Gauthier et al. (1998). Simulated brightness temperatures calculated with the true profiles with cloud top pressures ranging from 175 hPa to 1000 hPa and 15 μm cloud emissivity ranging from 0.1 to 0.9, were given as input to the estimation procedure and perturbed profiles were used to make the retrieval.

Each cloud configuration was perturbed 400 times. On figure 1, the variation of the RMS error resulting from these Monte-Carlo experiments versus the two parameters $\varepsilon$(15 μm) and $P_c$ are shown.

According to these graphs, the retrieval of reliable cloud top pressures seems to be possible in the range from the lower tropopause to 850 hPa. In the tropopause, as the temperature profile is almost isothermal, the sensitivity to the cloud top pressure is very low; two solutions with two significantly different cloud top pressure and very close cloud water content may exist in this situation. When the cloud is close to the surface, the contrast between the cloud top temperature and the skin temperature is so weak that it is not possible to estimate the cloud top pressure accurately.

Similarly to the cloud top pressure, the $\delta$ parameter is almost impossible to retrieve accurately near the surface. In the upper troposphere, it is still possible to retrieve an accurate $\delta$ even if the corresponding $P_c$ is poorly estimated. The accuracy of the cloud parameter retrieval is greater for an apaque cloud than for a semi-transparent cloud.

The second source of error is associated with the noise in observed brightness temperatures. The effect of noise on the retrieval error is expected to be low because of the high number of channels used and the appropriate weighting of the cost function. In a similar fashion, in order to estimate this error, brightness temperatures with cloud top pressures ranging from 50 hPa to 1000 hPa and effective cloud water contents ranging from 0.01 to 0.09 were simulated. A random noise generated from a normal distribution of mean zero and variance $\sigma_i^2$ was added to each brightness temperature. These simulated "noisy" brightness temperatures were used as input of our cloud parameter estimation procedure. The resulting error is lower than the error coming from the background profiles error, but its variation with the two cloud parameters is qualitatively similar.

## 5. VALIDATION WITH REAL AIRS DATA

The cloud parameter retrieval algorithm described in section 3 was applied to observed AIRS brightness temperatures using CMC 6 hour forecasts to provide temperature and water vapor profiles. 105 AIRS channels were selected among the 281 distributed channel set in order to get information on temperature and water vapor profiles using various criteria : Jacobian function shape, low sensitivity to ozone, good background statistics. On figure 2 these channels are presented.

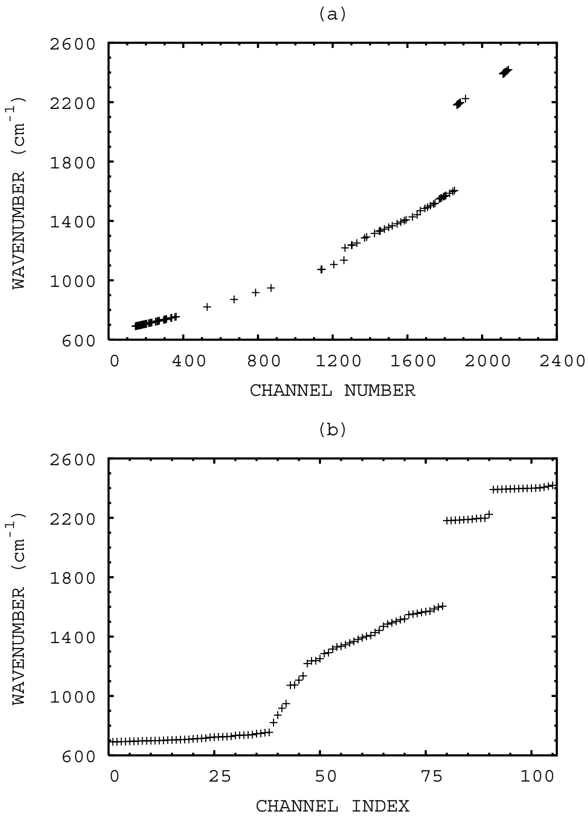


Figure 2 : (a) Wavenumber (cm$^{-1}$) versus AIRS channel number (1-2378) and (b) versus the channel index (1-105) defining the subset used in this study

Among these 105 channels, only 68 channels are used in this first stage because water vapor sensitive channels are discarted.

The comparison presented here is limited to night situations for which the shortwave channels are not affected by solar radiation. This choice allows the validation of the cloud emissivity model for the largest possible spectral.

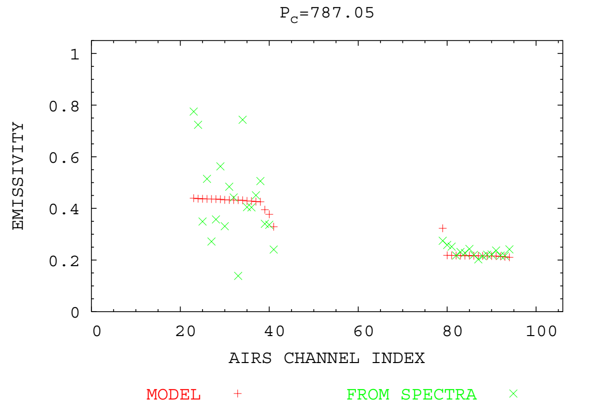


Figure 3A: Comparison between retrieved emissivity based on the proposed minimization (Equ. 4) and emissivity calculated directly from Equ. 5 for situation (A).

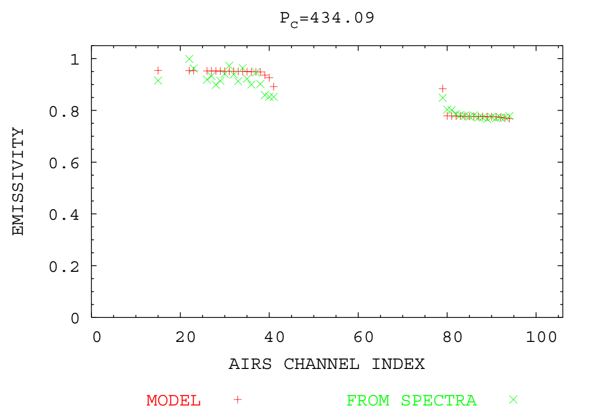


Figure 3B: same as figure 3A for situation (B)

In order to validate this cloud emissivity model, it is interesting to compare model emissivity with emissivity directly computed from the spectra using the following equation derived from equation (1) :

$$N\varepsilon(\nu) = \frac{I_{clr}(\nu) - I_{obs}(\nu)}{I_{clr}(\nu) - I_{ovc}(\nu)} \qquad (5)$$

This equation must be used with great care because the denominator may vanish. The emissivity of high clouds is therefore obtained with a greater accuracy than the emissivity of low clouds. It is also important to keep in mind that this emissivity is dependent on the retrieved cloud top pressure.

On figures 3A and 3B, two examples of comparisons between model emissivity and emissivity calculated from spectra using equation (5) after determination of cloud parameters are plotted for two typical situations. In these two situations, there is a relatively good agreement between the two emissivities. The dispersion of the emissivity around the model prediction for channels 20 to 40 is most of the time greater than the same dispersion for channels 80 to 96, because this first set of channels is less sensitive to cloud than the second one.

## 6. 1D VARIATIONAL EXPERIMENTS

Once the cloud parameters are, as far as possible, accurately determined using the above described method, 1D-var experiments may be performed using all 105 selected AIRS channels leading to retrieved temperature and humidity profiles. In this second and last step, the following cost function is minimized :

$$J(x) = \frac{1}{2}\left\{(x - x_b)^t B^{-1}(x - x_b) + (H(x) - y)^t O^{-1}(H(x) - y)\right\} \qquad (6)$$

In equation (6), x is a vector containing temperature and water vapor profiles discretized on 35 levels (the 28 eta levels of CMC's GEM model plus 7 additional levels corresponding to the 7 highest RTTOV-8 pressure levels), surface temperature and pressure and cloud parameters ( $\delta$ and $P_c$). y is the vector of observations, here the AIRS brightness temperatures of the 105 AIRS channels selected. H(y) is the radiative transfer model.

B is the CMC background error covariance matrix used in the operational model augmented to account for error statics of $\delta$ and $P_c$. In these experiments, it was assumed that each of these two parameters are uncorrelated with the other parameters. The assumed errors are of 20 hPa for $P_c$ and 0.005 for $\delta$.

For each situation considered, the cloud parameters were in a first step determined using the procedure described in section 3 using as first guess an estimation resulting from the $CO_2$ slicing method. In this second step, the atmospheric profiles are allowed to vary during the minimization process. Two experiments were performed. In the first one, $\delta$=0 was forced (i.e. no clouds) and only clear channels insensitive to clouds were used. In the second experiment, all 105 selected channels were used. In figures 4 and 5, some examples of temperature and water vapor increments resulting from these experiments for the same two cases of figures 3A and 3B are presented. As expected, the clear sky increments tend to zero below the cloud level. The cloudy increments may be significantly different from zero below the cloud level if the cloud is semi-transparent. The two increments are similar in the upper part of the atmosphere.

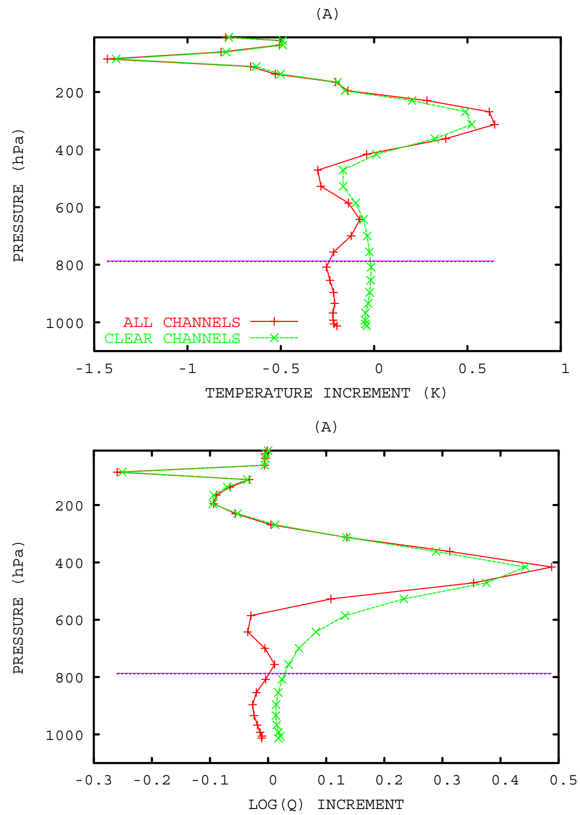


Figure 4 : Comparison between 1Dvar temperature ( top panel) and water vapor (bottom panel) increments obtained using only clear channels and those obtained using all channels and the cloud emissivity model for situation (A).

It was also tried to perform the same experiments with fixed cloud parameters, by assigning to them a very low error. The temperature and water vapor increments obtained during these experiments were not very different except in some rare cases.

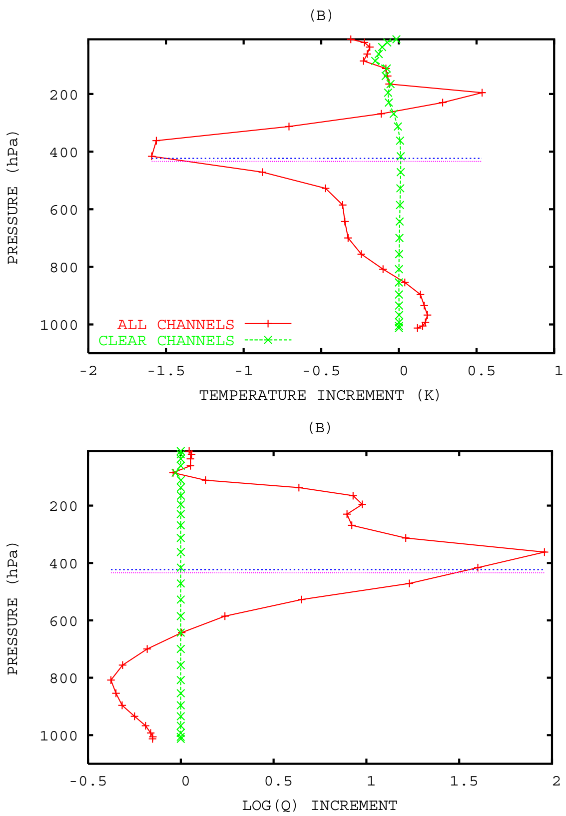


Figure 5 : Same as figure 4 for situation (B)

## *7.* CONCLUSION

A simple two parameter infrared cloud emissivity model and an estimation procedure for the retrieval of these two parameters were presented. By providing a realistic description of the frequency dependent cloud emissivity, this model allows the simultaneous use of many channels corresponding to different wavelengths. A realistic and robust estimation of the cloud parameters is a prerequisite for the assimilation of cloudy radiances. Simple 1Dvar experiments with this model were performed using real AIRS data. The temperature and water vapor increments obtained during these experiments appear reasonable. A possible way to validate the retrieved temperature profiles would be to compare these to temperature profiles retrieved from 1D-var assimilation of collocated microwave radiances of the AMSU-A instrument on board the AQUA satellite which are insensitive to non-precipating clouds. Once validated, the next step towards the assimilation of cloudy AIRS radiances using this approach is to perform 3D var experiments. These experiments will show the impact of cloudy radiance assimilation on 3D analysis fields.

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