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Atmospheric motion vectors height assignment by IRW and water vapour (H_2O) intercept methods

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Abstract

The atmospheric motion vectors (AMV's) derived from geostationary satellites are valuable tool in weather forecasting especially in data sparse region. This paper presents the results of an inter-comparison of AMVs assigned heights derived from Meteosat -7 & Kalpana -1, geostationary satellite data for both lower and upper levels by Infrared Window (IRW) and Water Vapour (H_2O or IR/WV) intercept methods. The Kalpana -1 satellite data (different sensor and resolution than Meteosat -7) is being processed by similar algorithm as Cooperative Institute of Meteorological Science (CIMSS), USA. In this short study of inter-comparison, the utility of the IR/WV intercept method in assigning the height of derived wind vectors especially at upper level winds is shown graphically. It is observed that actual wind speed direction from radiosonde data at upper levels (300-150 hPa) is higher up to the order of 8-12 m/sec and 5-8 degree for Kalpana – 1 data after applying the semi-transparency correction.

Keywords: Meteosat-7, Kalpana-1, Atmospheric motion vectors, IRW, H_2O intercept, Meteorological data.

Introduction

Kalpana-1 Satellite (Meteorological satellite) was renamed in 2004 as Kalpana after Kalpana Chawla. It has onboard three channels {Visible (0.55µm -0.75µm), Infrared (10.5µm-12.5µm) and WV (5.7µm-7.1µm)} Very High Resolution Radiometer (VHRR) payload at 2x2, 8x8 and 8x8 km spatial resolution at nadir respectively. European Meteorological Satellite (EUMETSAT) operates Meteosat -7 satellite in three channels {Visible (0.45µm - 1.00µm), Infrared (10.5µm-12.5µm) and WV (5.7µm-7.1µm)} with resolution at nadir 2.5, 5.0 and 5.0 km respectively. Kalpana -1 satellite is three axes stabilized, whereas Meteosat -7 is spin stabilized. The position in space of Meteosat-7 satellite is 0° / 57.5° E / 35,800 km and Kalpana -1 satellite 0° / 74° E / 35,800 km. The height assigned to the AMV wind field is the level where the atmospheric temperature reaches brightness of the tracked structures. The brightness of the temperature is computed from the radiance through the Planck function and radiometer filter function. Coakley and Bretherton (1982) scheme is an in-built software program to classify the low, medium and high clouds**.** Normally, opaque or thick cloud heights are calculated from IRW method which is the representative of equivalent black body temperatures derived from the AMV target area and temperature forecast file from the Numerical Weather Prediction (NWP) model. But in high level, sub-pixel or semi-transparent clouds the background contribution of the radiance overestimate the cloud top temperature and assigned the cloud height too low.

There are two modes of radiance measurements, one is theoretically calculated by radiative transfer equation and other is measured by the satellite in IR and H_2O channels. The forward calculations of the radiance measurements with radiative transfer equation is done by the streamer software programme up to the top-of-the-atmosphere. The fundamental assumption of the method is that a linear relationship exists between measurements in two spectral bands observing a single cloud layer. The intersection of the measured and calculated radiances should occur at 0 and 100% cloud cover (*i.e*. clear sky and opaque conditions). The cloud top pressure for semi-transparent or sub-pixel cloud is determined from the intersection of the linear fit to the observations and the curve through the calculated opaque radiances (Schmetz *et al.,* 1993). The brightness temperature of the radiation received by the instrument in the water vapour channel depends upon both temperature and humidity vertical profiles between the sensor and ground and on the spectral response of the instrument in that channel, while it is least modified in the window channel (Szejwach, 1982). The radiation emanating from the upper level clouds (Cirrus clouds) is the sum of the radiation reaching the cloud base, transmitted after absorption and scattering by ice crystals and water vapour present inside the cloud, and the radiation emitted by the absorbing particles at their thermodynamic temperature. The brightness temperature observed in IR and WV channels was found always warmer than the surrounding (cloud thermodynamic) temperature. This indicates the semi-transparency in the cirrus cloud; otherwise in full transparency we would observe the same brightness temperature as the temperature at cloud base level. In fact, the brightness temperature in absorption channel is smaller than the window channel due to the larger value of the cloud emissivity in the absorption channel (H_2O band). Jianmin and Zhang (1996) taken the brightness temperatures of IR and WV channels at different cloud conditions and observed the linear behavior, which later utilized in correcting the semi-transparency in thin or cirrus type clouds. Water vapour winds (derived from water vapour absorption band) appear to be useful to complete and improve wind fields from the Infra red channel for high levels (Szantai & Desbois, 1991). The

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use of the water vapour channel seems to provide complementary information especially at high levels (Laurent, 1990). Tracking features in the water vapour absorption band (Meteosat, WV =5.7 -7.1 µm) would enhance the wind data in the middle and high troposphere (Stewart *et al*., 1985). The separation of middle/high clouds from clear sky /low clouds is done by masking the clouds which is based on the image threshold in (IR and WV bands). The screening of cumulonimbus (Cb) clouds is done by WV -IR temperature difference (Tokuno, 1996).

Data and methodology

The satellite data Kalpana -1 and Meteosat -7 used in the study has been taken from India Meteorological Department, Lodi road, New Delhi and CIMSS, USA respectively. Two approaches have been used for the assignment of the height of cloud tracers. One is mostly suitable for thick and opaque clouds known as Infrared window (IRW) technique, and second approach is H_2O intercept (WV/IR intercept) method, suitable for sub-pixel cloud and semi-transparent cirrus type of clouds. The second approach is suitable for upper levels cloud height assignment as most of the contribution in this H_2O absorption channel comes above 600 hPa. In IRW method the brightness temperature is inferred directly from radiances.

Infrared Window (IRW) method

The IR intercept method is suitable for opaque clouds. The low & medium level tracers are move with the cloud base and their height can be estimated in better way by mean of 25 % of coldest pixel. Thus, for optically thick and large areal extent clouds, Temperature is evaluated directly from the satellite observed radiance. The above concept is applied in IR -window method of AMVs height assignment. In which, for thick opaque clouds and valid tracers the tracer height is assigned by mean of 25 % of coldest pixels temperature (Merrill *et al*., 1991), which is inferred directly from the satellite radiance in IR window channel.

$$T\left(p\right) \cong B_{\nu}^{-1}\left(R_{\nu}^{obs}\right)$$

Where, R_{ν}^{obs} is the satellite radiances in IR window channel.

H_2O Intercept method

The water vapor intercept technique (Szejwach, 1982), based on the fact that radiances influenced by upper tropospheric moisture (H_2O at 6.5 µm) and radiances not sensitive to water vapor (IRW at 10.7 µm) exhibit a linear relationship as a function of cloud amount is used to extrapolate the correct height. If there is an absorbing elements presents in the atmosphere then brightness temperature shows considerable differences. The water vapour contribution of semi-transparent clouds is lying between 8 and 10 km whereas the dotted lines represents the situation with the opaque clouds between 8-9 km, 6-7 km and 4.5 -5.5 respectively (Fig. 19, Holmlund, 1993). But for high -level tracers, broken clouds , thin cirrus or sub-pixel clouds (Not fully in the range of sensor field of view) the contribution from the lower layers of the atmosphere also present, which will assign the tracer height at lower level. This semi transparency is corrected by intercept method of using WV and IR images. In this approach bi-spectral clustering for coldest and warmest clusters, is done by measuring the radiances in IR & WV bands (Nieman *et al*., 1993; Menzel *et al*., 1993; Schmetz *et al*., 1993).

The radiances in IR ($10.7\mu m$) and WV ($6.5\mu m$) vary linearly with cloud amount. These data are used in conjunction with forward calculations of radiance for both spectral channels for opaque clouds at different levels in a given atmosphere specified by a numerical weather prediction of temperature and humidity. The intersection of measured and calculated radiances will occur at clear sky radiances and cloud radiances. The cloud top temperature is extracted from the cloud radiance intersection (Schmetz *et al*., 1993). Presently, in most of the meteorological centers heights are assigned by any of three techniques (IRW, H_2O intercept and CO_2 slicing) when the appropriate spectral radiance measurements are available (Nieman *et al*., 1993). Hence, radiance ratio of WV/IR or H_2O intercept method is the linear regression using all the measured radiances and the calculated radiances of clear and opaque pixels.

$$R_{H_2O}^{cl} = R_{H_2O}^{cs} + \left\{ \frac{R_{IR}^{cs} - R_{IR}^{cl}}{R_{IR}^{cs} - R_{IR}^{m}} \right\} \left(R_{H_2O}^{m} - R_{H_2O}^{cs} \right)$$

Or pixel by pixel level, the pressure level is assigned if the following equation is satisfied.

$$\frac{R_{H_2O}^{cl} - R_{H_2O}^{cs}}{R_{IR}^{cl} - R_{IR}^{cs}} = \frac{R_{H_2O}^{m} - R_{H_2O}^{cs}}{R_{IR}^{m} - R_{IR}^{cs}}$$

Where, ($R_{IR}^{cs}, R_{H_2O}^{cs}$), ($R_{IR}^{m}, R_{H_2O}^{m}$) and ($R_{IR}^{cl}, R_{H_2O}^{cl}$) are clear sky, measured radiances and opaque radiances pairs respectively.

Radiances in the atmosphere for Kalpana -1 satellite data are calculated theoretically by streamer radiative transfer code (Key & Schweiger, 1988). The code uses N-stream approximation to the radiative transfer equations (Stamnes *et al*., 1988) and allows for flexible choice of bands.

Results and discussions

The retrieval of atmospheric motion vectors (AMV's) needs proper height assignment for suitable representation of the winds at that pressure level. The AMV's height assignment of thick or opaque clouds is normally determined by IR window method. Whereas, the height of thin layer cirrus or mainly high level clouds are better represented by H_2O intercepts method. To explain the concept of H_2O intercept method we have randomly selected the area between 30^0N-40^0N, 70^0E-80^0E for

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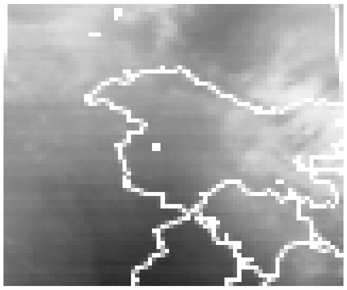

Fig 1. Kalpana 1-Infrared (IR) on 30-01-11 (Area: 30^0N-40^0N, 70^0E-80^0E,0900 UTC)

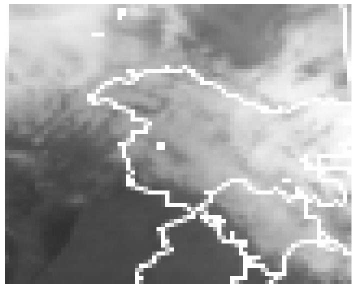

Fig 2. Kalpana -1 Water Vapour (WV) on 30-01-11 (Area: 30^0N-40^0N, 70^0E-80^0E,0900 UTC)

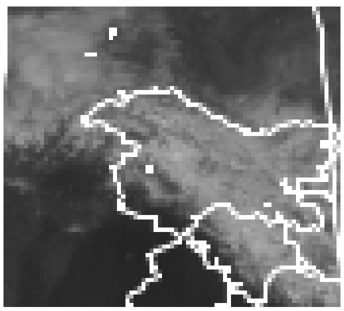

Fig 3: Kalpana -1 Visible (VIS) on 30-01-11 (Area: 30^0N-40^0N, 70^0E-80^0E,0900 UTC)

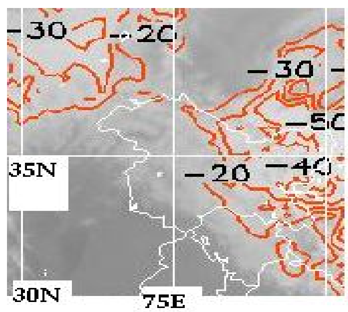

Fig 4: Cloud top temp (CTT) of Kalpana -1 on 30-01-11 (Area: 30^0N-40^0N, 70^0E-0^0E,0900UTC)

consideration of height assignment of upper levels, in which few sets of clear and cloudy pixels are available. Fig. 1-2 shows the IR and water vapour image at 0900 UTC of 30 -01-2011 of Kalpana -1 satellite, showing clear and cloudy pixels area. The radiances of both IR and H_2O images is plotted on X and Y axis respectively (Fig. 9) which show the linear behavior. Fig.3 is the visible band image of the same area and gives more clarity regarding the clear and cloudy (warm and cold) areas separately. Fig.4 shows the cloud top temperature over the above said area which shows the approximate height by seeing the cloud top temperatures. The cloud top temperature is shown in the Fig. 4 is of the order of -30 ^{0}K to -40 ^{0} K which shows deep convection up to upper troposphere of around 9-10 km in vertical. Fig. 5 and 6 show the CMV and WVW winds over the above selected area. Fig. 7 & 8 shows the Meteosat -7 IR + Visible and Upper level winds and their pressure levels over the selected area. Fig. 9 shows the linear behavior of IR and H_2O radiances of warmest and coldest pixels along with the theoretical black body curve as suggested by Nieman *et al*. (1993). In Fig. 9 the intersection point of theoretical curve and linear regression line of warmest and coldest curve represents the semi transparency corrected radiance or brightness temperature. The pressure level is assigned based on the value of brightness temperature at the intersection point and model forecast field.

In this way mostly at upper levels where thin sheets cirrus like clouds are present H_2O intercept method is more appropriate. One other factor of the differences in the cloud heights may be attributed due to different sized Field of Views (FOVs) and the different spectral response functions of Kalpana -1 and Meteosat -7**.** Hence, if the clouds are thick or clear areas in upper troposphere then they are the representative of mean layer flow and defined by the mean layer temperature. In those cases, the AMVs heights are not representative of single level winds but it comes through the mean layer of clouds. This type of average motion is not suitable in the present NWP models.

To highlight more the concept and an inter-comparison between IR window method and H2O intercept method, collocated points of both upper level and lower level AMV's of 0300 UTC (03 December, 2009) and 0900 UTC (09 December, 2009) have been selected randomly. The collocation diagram of both 0300 UTC and 0900 UTC CMV and WVW are shown in the Fig. 14 & 15. The winds (CMV 0300 UTC & WVW 0900 UTC) generated at IMD is shown in the Fig. 10 & 12. The IR and Visible winds at 0300 UTC of 03 December, 2009

Fig. 5. Kalpana -1 (K-1) CMV on 30-01-11 (Area: 30^0N-40^0N, 70^0E-80^0E,0900 UTC)

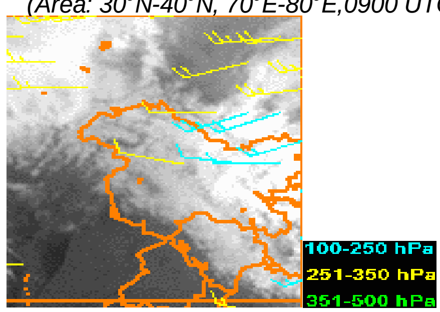

Fig. 6. Kalpana -1 (K-1) WVW on 30-01-11 (Area: 30^0N-40^0N, 70^0E-80^0E, 0900 UTC.)

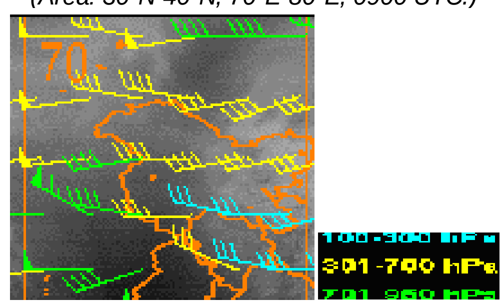

Fig. 7. CIMSS, USA Visible & IR on 30-01-11 Area: 30^0N-40^0N, 70^0E-80^0E ,0900 UTC)

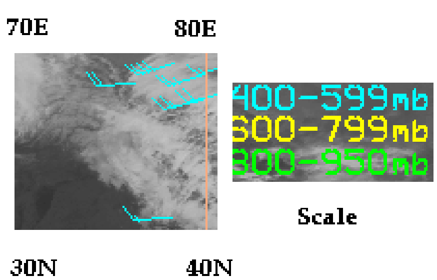

Fig. 8. CIMSS, USA- upper level winds on 30-01-11 (Area: 30^0N-40^0N, 70^0E-80^0E,0900 UTC)

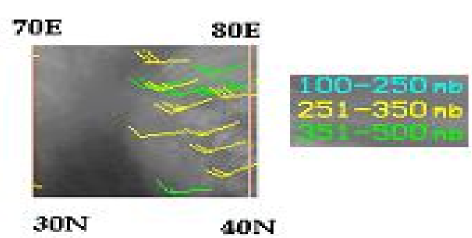

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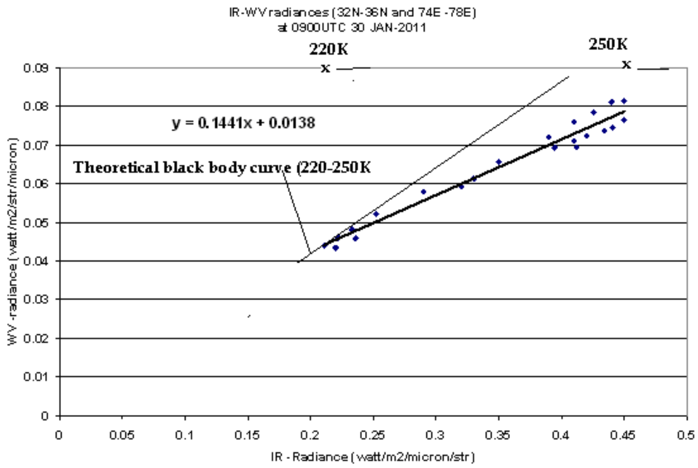
Fig. 9. Linear variation of Water vapour and Infrared radiances of Kalpana -1 data



and Upper level winds of 0300 UTC of 09 December, 2009 generated by the CIMSS, Wisconsin, USA are shown in Fig. 11 and 13 respectively. There collocation shows reasonably good agreement with Meteosat -7 and Kalpana -1 satellite derived winds by the similar software. Mean cloud top pressure of Meteosat -7 and INSAT Kalpana -1 satellite at 0900 UTC on 03 December, 2009 for cloud tracers between 150 -400 hPa with H_2O intercept method is 263 and 276 hPa respectively. The mean values for IRW method for upper level winds (0900 UTC) and 0300 UTC are 246, 240 and 718, 699 for INSAT and Meteosat -7 respectively. The scatter plot for both IRW and H_2O intercept are shown in the Fig. 14-16. Fig. 17 and 18 shows the IR and WV radiances at different cloud conditions (Jianmin, 1996). The Fig. 17 & 18 shows that the not only cirrus type of clouds but cumulonimbus (Cb) type and other middle level clouds H_20 contribution is significant and hence in this case H_2O intercept method give better height assignment as compared to IR window method alone. Fig 19 shows that clear sky contribution which are suitable clear sky are vary from 2 km to 14 km and maximum (90 %) is around 8 km. The semi-transparent type of clouds maximum contribution comes around 10 km. The contribution of

Fig. 10. CMV generated at Satellite division (0300UTC), Lodi Road, New Delhi

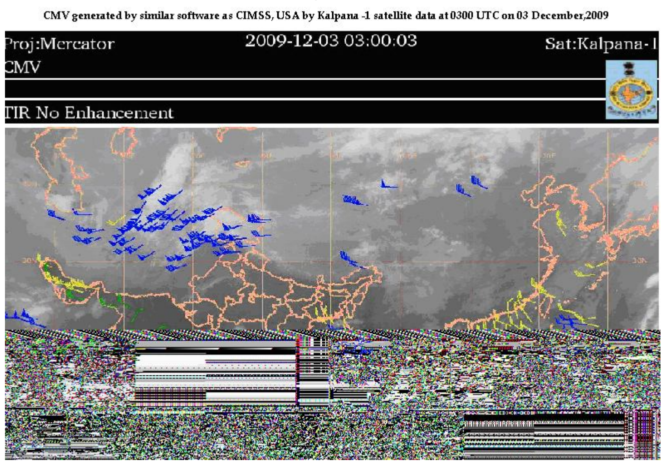


Fig. 11. IR + Visible winds generated at CIMSS (0300UTC)Madison Wisconsin (USA)

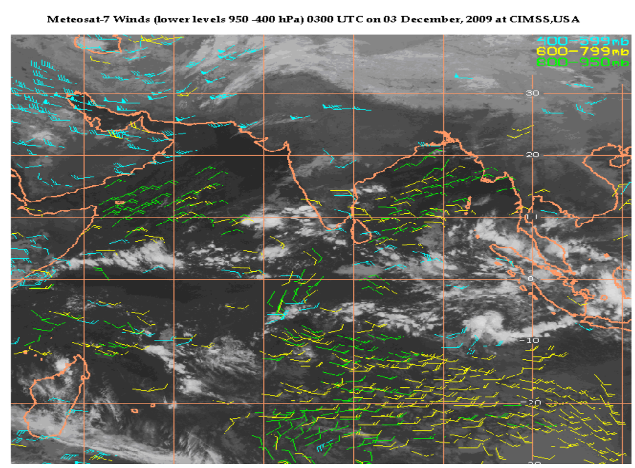


Fig. 12. WVW generated (0900UTC) at Satellite division, Lodi Road, New Delhi

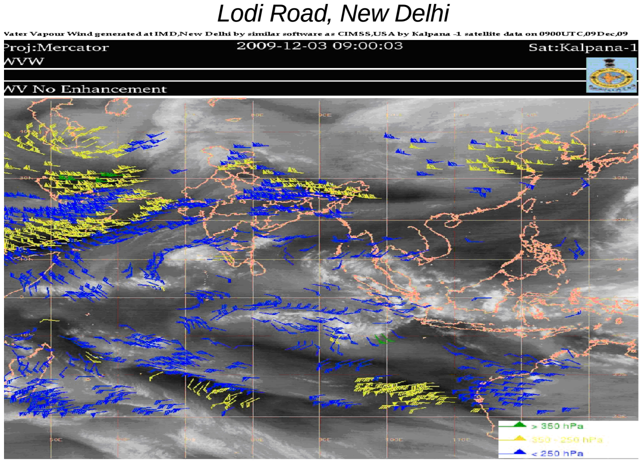


Fig. 13. Upper level winds generated at CIMSS (0900 UTC), Madison Wisconsin (USA)

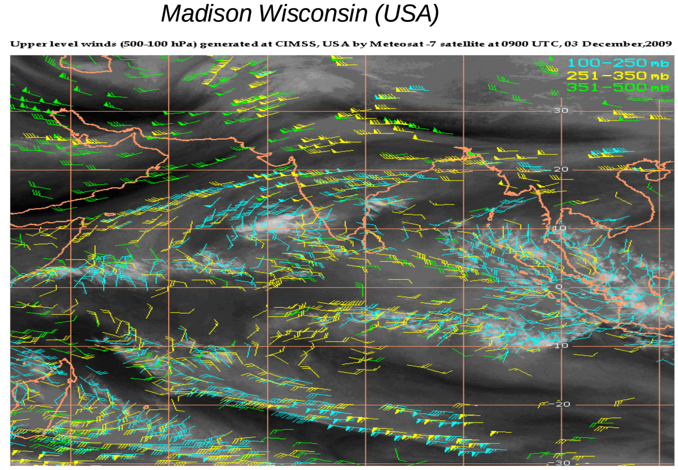


Fig. 14. Collocated CMV (Meteosat-7 and Kalpana-1) 0300 UTC of 09 December, 2009

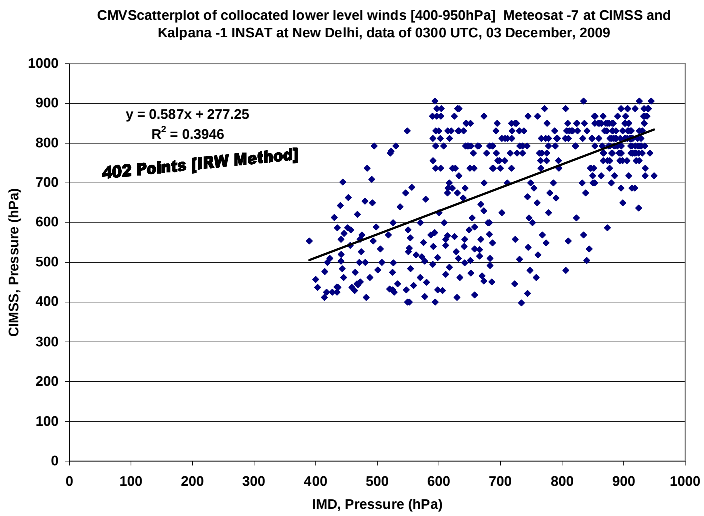


Fig 15. Collocated WVW (Meteosat-7 and Kalpana-1) 0900 UTC of 09 December, 2009 (IRW method)

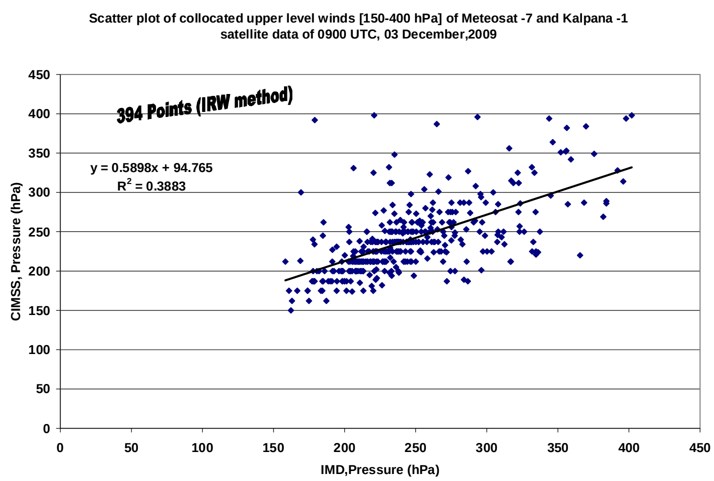
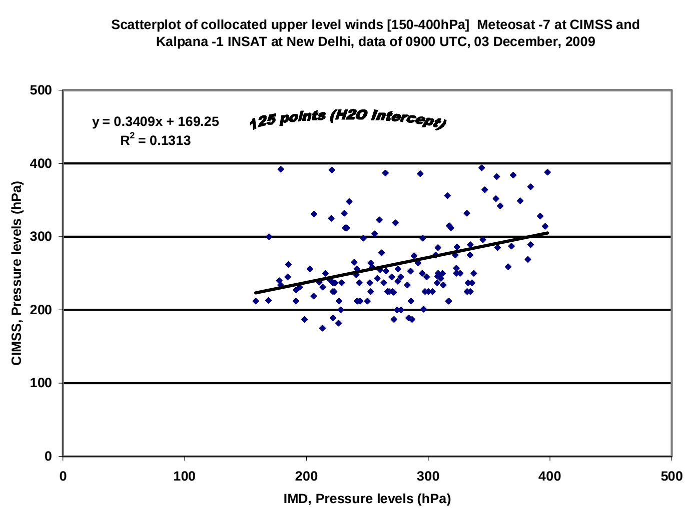


Fig. 16. Collocated WVW (Meteosat-7 and Kalpana-1) 0900 UTC of 09 December, 2009



several opaque clouds which are suitable for IR window method vary from 4-8 km and maximum (70 -80 %) comes around 4 and 7 km respectively. The collocated AMV and RS/RW speed differences for upper levels (< 400 hPa) are shown in the Fig. 20 and 21 and their corresponding direction bias are shown in the Fig. 22. The discrepancy of speed increases with AMV wind speeds (Fig. 21). The directional difference between AMV and RS is shown against RS direction increases gradually after 275^0 (Fig. 22). This indicates that the directional bias is more significant in SEly and NEly direction. Most of the upper level erroneous AMVs are filtered after semi-transparency correction. Inter-comparison shows that the RS wind speed and direction is approximately higher up to 12 m/sec and 8.0 degree around 150-200 hPa. The relevant brief details of sensor response function and H_2O intercept method fundamentals are given in appendices as Part (A) and (B) respectively.

Concluding remarks

In the present study, we have selected two random samples of images, one small area between latitude 30^0N-40^0N and longitude 70^0E-80^0E, which facilitates the understanding of the IRW and H_2O intercept methods of AMV's height assignment. Because, new H_2O intercept algorithm needs regression of warmest and coldest pixels for assigning the height of the tracers over the area. So one random sample of above domain is selected which fulfill the basic requirement of the algorithm. The second domain covers the entire Indian region (latitude 40^0N-30^0S and longitude 40^0E-100^0E) to elaborate further the inter-comparison between two approaches (IRW and H_2O intercept) of height assignments of wind components from Meteosat -7 and Kalpana -1 geostationary satellites, in which the retrieval of AMV's is done by similar software. The sensor characteristics and resolutions and geo-referencing of both the satellites are different. Below are the points observed during the study:

1. The H_2O intercept method is an essential technique for semi-transparency correction in upper level winds especially generated from thin cirrus clouds as they are best suitable single level tracers for NWP assimilation.
2. Radiance ratioing method is viable alternative tool at upper level AMVs because the speed and direction bias in upper levels increases with wind speed and is of the order of 8-12 m/sec and 5-8 degree for Kalpaqna -1 data around 300-150 hPa.
3. The mean values of cloud top pressures of Meteosat -7 and Kalpana -1 data by H_2O intercept method for cloud tracers between 150 -400 hPa is 263 and 276 hPa respectively. Similarly, the mean values of cloud top pressures by IRW method in upper (150-400 hPa) and lower levels (400-950 hPa) are 240, 246 and 699, 718 hPa respectively. The differences in the cloud heights may be attributed due to different sized Field of Views (FOVs) and the different spectral response functions.

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Fig. 17. Origin of IR and WV channel radiance at different cloud conditions (Jianmin, 1996)

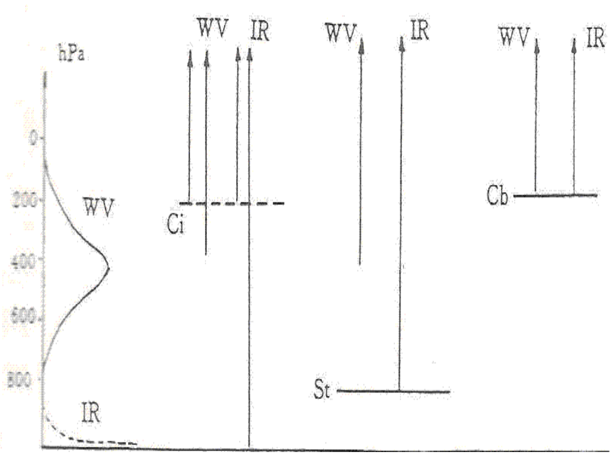
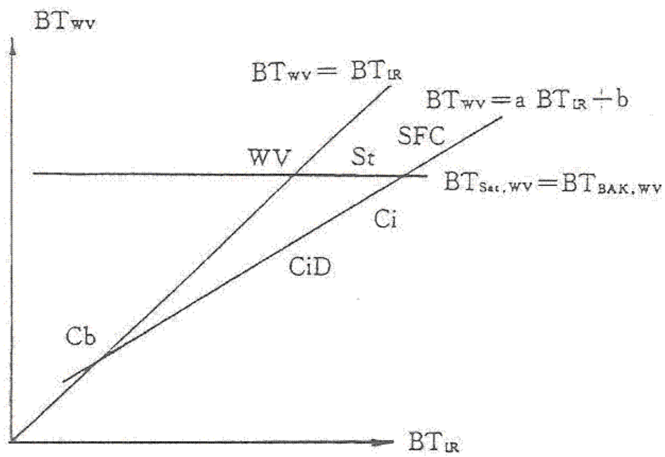


Fig. 18. Scatter diagram of IR and WV brightness temperature at different cloud conditions (Jianmin, 1996)



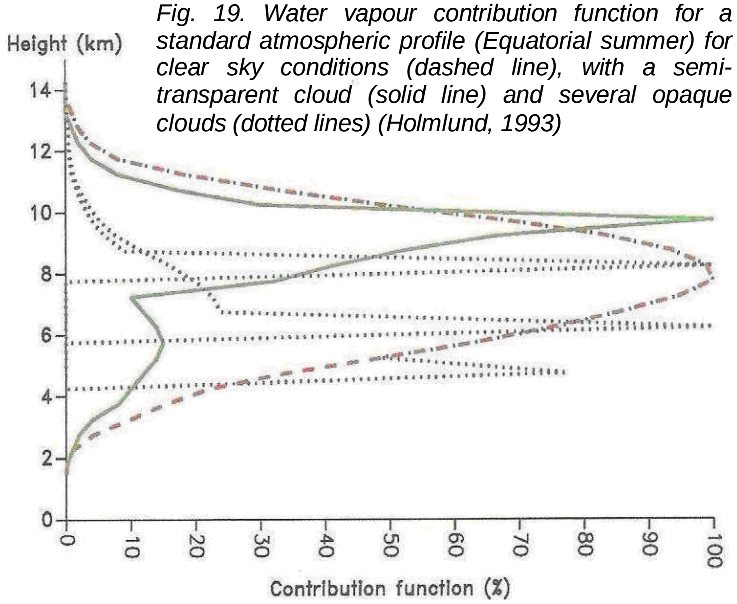


Fig. 19. Water vapour contribution function for a standard atmospheric profile (Equatorial summer) for clear sky conditions (dashed line), with a semi-transparent cloud (solid line) and several opaque clouds (dotted lines) (Holmlund, 1993)

Acknowledgements

The authors are grateful to Dr. Steve Wanzong, Dr. D.Santek of Space Science and Engg. Centre, Wisconsin University, USA and DG IMD for their kind help and support in providing the data.

Appendices:

(A) Spectral response of the detector and brightness temperature

The energy reached to the satellite sensor at the wavelength band (IR, WV or VIS) depends on the Spectral Response Function (SRF) of the detector. Radiance emitted by the black body ($\varepsilon = 1$) at temperature T, measured by a satellite sensor is given below in Eq-1

$$B(T) = \int_{\lambda_1}^{\lambda_2} S(\lambda)\varepsilon(\lambda)B(\lambda,T)\,d\lambda \quad \text{(General form)} \quad (1)$$

Where, ε is the emissivity of the object, between 0 & 1

$$B(T) = \int_{\lambda_1}^{\lambda_2} S(\lambda)B(\lambda,T)\,d\lambda \qquad \text{(1a)}$$

Where, $S(\lambda)$ is the Spectral Response Function (SRF) of the sensor & $B(\lambda,T)$ is the spectral radiance of the black body (Planck function, in $W/m^2 sr \mu m$)

$$B(\lambda,T) = \frac{c_0 \lambda^5}{\left(e^{c_1/\lambda T} - 1\right)} \qquad \text{1(b)}$$

Using the second order Taylor approximation around a central wavelength λ_0 ,Eq-1 leads to:

$$B(T) \approx \int_{\lambda_1}^{\lambda_2} S(\lambda)\left[B(\lambda_0,T) + (\lambda-\lambda_0)\frac{\partial B(\lambda,T)}{\partial \lambda}\right] d\lambda$$

or

$$B(T) \approx B(\lambda_0,T)\int_{\lambda_1}^{\lambda_2} S(\lambda)d\lambda + \int_{\lambda_1}^{\lambda_2} S(\lambda)(\lambda-\lambda_0)\frac{\partial B(\lambda,T)}{\partial \lambda} d\lambda \qquad \text{1(c)}$$

The central wavelength is estimated as the centroid of SRF

$$\lambda_0 \approx \frac{\int_{\lambda}^{\lambda_2} \lambda S(\lambda)\,d\lambda}{\int_{\lambda}^{\lambda_2} S(\lambda)\,d\lambda} \quad (2)$$

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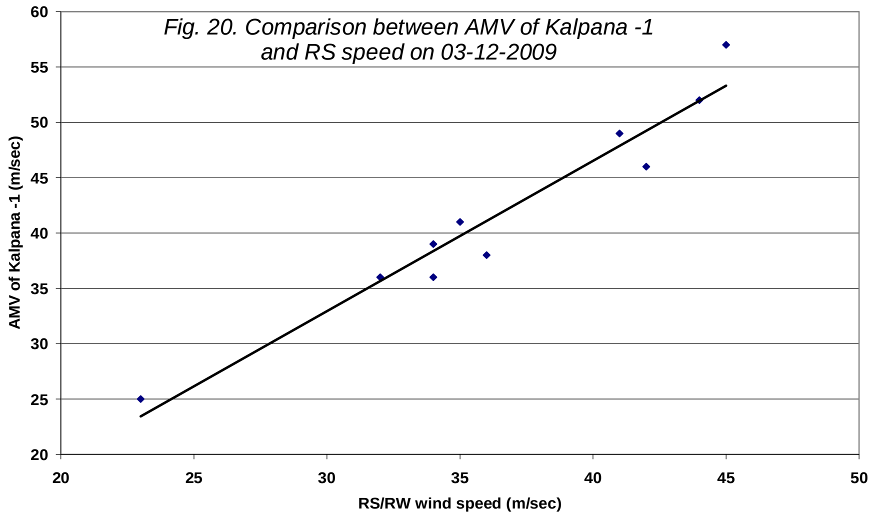

Fig. 20. Comparison between AMV of Kalpana -1 and RS speed on 03-12-2009

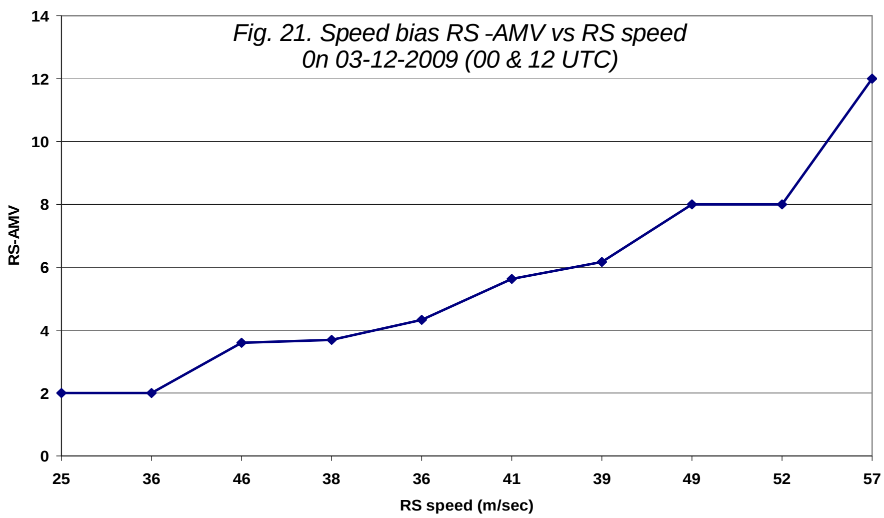

Fig. 21. Speed bias RS -AMV vs RS speed 0n 03-12-2009 (00 & 12 UTC)

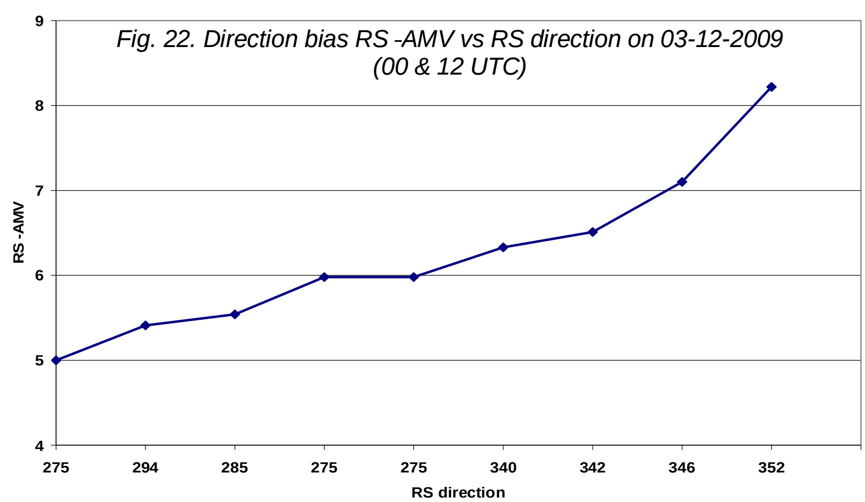

Fig. 22. Direction bias RS -AMV vs RS direction on 03-12-2009 (00 & 12 UTC)

Assuming a SRF sampled on N equidistant wavelengths $\{\lambda_i, S_i\}_{i=1,\ldots N}$. Applying the numerical integration by mid-point rule on Eq-2, we get: or

$$\lambda_0 \approx \frac{\Delta\lambda \sum_{i=1}^{N-1} \lambda_i \left(S_i + S_{i+1}\right)/2}{\Delta\lambda \sum_{i=1}^{N-1} \left(S_i + S_{i+1}\right)/2} = \frac{\sum_{i=1}^{N-1} \lambda_i \left(S_i + S_{i+1}\right)}{\sum_{i=1}^{N-1} \left(S_i + S_{i+1}\right)}$$

Let, the total area under the SRF is A:

$$A = \int_{\lambda}^{\lambda_2} S(\lambda)\, d\lambda \approx \Delta\lambda \sum_{i=1}^{N-1} \lambda_i \left(S_i + S_{i+1}\right)/2$$

Then, Eq- 1(b) reduces to:

$$B(T) \approx AXB(\lambda_0, T) + O(\lambda) \qquad (3)$$

Because of the different response of the Planck function Because at different wavelengths, the radiant temperatures measured in two channels may be expressed m terms of contributions from two temperature fields, each occupying a portion of the pixel, where the portions are not necessarily contiguous.

Most of the thick clouds generally assumed blackbodies $(\varepsilon \approx 1)$ and not introduced much source of error in this approximation. But other elements present between the target (cloud or surface) and the satellite may absorb or diffuse and thus reduce the amount of radiation received by the detector. That is why, Thermal Infra Red (TIR) window channel is taken for the brightness temperature estimation; in this case, attenuation of radiation is almost negligible.

Normally, in the atmosphere at height 'z' the temperature change with height is given below:

$$T = T_0 - \gamma z \qquad (4)$$

Where, T_0 is the temperature at sea level (altitude =0) in ^{0}C or ^{0}K. γ is the lapse rate (6.5 ^0K /Km) and positive in case of inversion and z is the altitude in Km.

For a standard atmosphere (P_0 =1013.25 hPa, T_0 =288.15 ^0K) pressure is a function of temperature:

$$\frac{P}{P_0} = \left(\frac{T}{T_0}\right)^{5.25} \qquad (5)$$

Using Eq (4) above we get:

$$\frac{P}{P_0} = \left(1 + \frac{\gamma z}{T_0}\right)^{5.25} \quad (6)$$

The height or pressure level of the AMVs is assigned with model forecast (Eq-6) by minimum of 25 % coldest pixel of the TIR image, in case of thick and opaque cloud.

We know that, gravity can be represented in terms of the gradient of a potential function, (Φ) called geopotential (defined as work required raising a unit mass to height z from mean sea level:

$$\Phi = \int_0^z g dz \quad (7)$$

or

$$d\Phi = gdz \quad (7a)$$

Hydrostatic equation, $\frac{dP}{dz} = -g\rho$ (7b)

(Where, ρ is the density of dry air in the atmosphere)

From the gas equation,

$p\alpha = R_d T_v$ ($\rho\alpha = 1$)

$p = \rho R_d T_v$

Putting the value of p in Eq- (7b)

$$\frac{dp}{dz} = -\frac{gp}{R_d T_v}$$

or

$$\frac{dpg}{d\Phi} = -\frac{gp}{R_d T_v}$$

$$\frac{T_v R_d dp}{p} = -d\Phi$$

$$\int_{\Phi_1}^{\Phi_2} d\Phi = -\int_{p_1}^{p_2} R_d T_v \frac{dp}{p} \quad (8)$$

$\Phi_2 - \Phi_1 =$ LHS, in which $\Phi_1 = 0$ and $\Phi_2 = \Phi$ =gz,

Let, z is the height of the cloud, then z = z_{cloud}

In that case, Eq-8 becomes:

$$z_{cloud} = -\frac{R_d}{g} \int_{\ln(p_{surface})}^{\ln(p_{cloud})} T_v(p)\, d(\ln(p)) \quad (9)$$

Where, R_d =287 $JK^{-1} kg^{-1}$ gas constant for dry air

g = 9.81 ms^{-2} magnitude of gravity

T_v is the virtual temperature of air at pressure P and can de deduced from temperature and relative humidity profile.

Layer mean temperature is defined as:

$$\langle T \rangle = \frac{\int_{\ln(p_1)}^{\ln(p_2)} T_v(p)\, d(\ln(p))}{\int_{\ln(p_1)}^{\ln(p_2)} d(\ln(p))} \quad (10)$$

(B) Radiative transfer calculation with model guess field (Szejwach, 1982; Nieman *et al*., 1993)

The operationally used method by EUMETSAT and NOAA agencies (Nieman *et al*., 1993; Schmetz *et al*., 1993) known as H_2O intercept method. This method is overestimate the pressure, except in dry areas where the pressure is underestimated (Menzel *et al*., 1993). Observing a single cloud layer the radiance in one spectral band (IR = 11.2 µm) vary linearly with the radiances in another spectral band (H_2O) as a function of cloud amount in the Field of View (FOV). This method requires clear and cloudy sky radiance measurement in both the spectral bands (area roughly 100 km on a side centered on a target point). The linear relationship between both channel radiances is average radiances of cluster of clearest (warmest) and cluster of cloudiest (coldest), define the semi-transparency line. The clear and cloudy cluster selection based on the warmest 10 % or the surface analysis as clear and coldest 25 % as cloud. Theoretically, the radiances are calculated by radiative transfer equation from the guess profile. When the calculated H_2O radiances are not the same as measured, then the calculated H_2O radiances is adjusted with measure clear sky H_2O radiances. This adjustment is not required on warmer side (calculated radiances are more than measured) since low result may be the result of cloud contamination (Menzal *et al*., 1993; Nieman *et al*., 1993).

In a non scattering atmosphere for a given cloud element in a FOV the radiance observed, is given by:

$$R(\nu) = (1 - n\varepsilon)\left\{B(\nu, T(p_s))\tau(\nu, p_s) + \int_{p_s}^{p_c}\left[B(\nu, T(p))\frac{d\tau(\nu, p)}{dp}\right]dp\right\}$$
$$+ n\varepsilon\{B(\nu, T(p_c))\tau(\nu, p)\} + \int_{p_c}^{0}\left[B(\nu, T(p))\frac{d\tau(\nu, p)}{dp}\right]dp \quad (16)$$

Where, n is the fraction of the field coverage. Eq-16 is gross simplification form (dealing with single cloud layer and free from multi-deck situation), because many types and layers of clouds can exists in FOV.

In the above equation the four terms are the radiation emitted from the surface, the contribution of the atmosphere below the cloud, the cloud contribution and the contribution from the atmosphere above the cloud. Assume only the clear sky and opaque conditions then equation (16) can be simplified as Eq-12.

From Eq-12 in terms of wave number is given by:

$$R(\nu)=\left(1-n\varepsilon(\nu)\right)R_{cs}(\nu)+n\varepsilon(\nu)R_{cl}(\nu,p) \quad (12a)$$

or

$$R(\nu)-R_{cs}(\nu)=n\varepsilon(\nu)\left[R_{cl}(\nu,p)-R_{cs}(\nu)\right] \quad (12b)$$

For Clear sky Field of view ($n\varepsilon(\nu)$=0) from Eq-16:

$$R^{cs}(\nu)=B(\nu,T(p_s))\tau(\nu,p_s)+\int_{p_s}^{p_0}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp \quad (16a)$$

Taking the difference of the Eqs {(16 -16a)}, we get:

$$R(\nu)-R^{cs}(\nu)=$$

$$(1-n\varepsilon)\left\{B(\nu,T(p_s))\tau(\nu,p_s)+\int_{p_s}^{p_c}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp\right\}$$

$$+n\varepsilon\left\{B(\nu,T(p_c))\tau(\nu,p)\right\}+\int_{p_c}^{0}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp-B(\nu,T(p_s))\tau(\nu,p_s)-\int_{p_s}^{p_0}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp$$

or

$$R(\nu)-R^{cs}(\nu)=$$

$$-n\varepsilon B(\nu,T(p_s))\tau(\nu,p_s)-n\varepsilon\int_{p_s}^{p_c}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp+n\varepsilon\left\{B(\nu,T(p))\tau(\nu,0)\right\} \quad 16b$$

Where,

$$\int_{p_c}^{0}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp+\int_{0}^{p_s}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp$$

$$=-\int_{p_s}^{p_c}\left[B(\nu,T(p))\frac{d\tau(\nu,p)}{dp}\right]dp$$

Integrating by parts (Eq-16 b) by taking $B(\nu,T(p))$ as first function and $\frac{d\tau(\nu,p)}{dp}$ as second, we get:

$$-n\varepsilon B(\nu,T(p_s))\tau(\nu,p_s)-n\varepsilon\left[B(\nu,T(p))\tau(\nu,p)-\int_{p_s}^{p_c}\left\{\tau(\nu,p)\frac{dB(\nu,T(p))}{dp}\right\}dp\right]+n\varepsilon\left\{B(\nu,T(p_c))\tau(\nu,0)\right\}$$

$$R(\nu)-R^{cs}(\nu)=n\varepsilon\int_{p_s}^{p_c}\left\{\tau(\nu,p)\frac{dB(\nu,T(p))}{dp}\right\}dp \quad (16c)$$

In the above simplification, we have used the rule: $\exp(\alpha+\beta)=\exp\alpha+\exp\beta$ for the transmittance (as the transmittance is exponentially related as optical function, $\varepsilon(p_1p_2)=\exp\left[-\frac{1}{g}\int_{p_1}^{p_2}k_\nu(p)q(p)dp\right]$.

Where, $\tau(\nu,p_s)+\tau(\nu,p)=\tau(\nu,0)$

If the difference $R(\nu)-R^{cs}(\nu)$ is within the range of instrumental noise, roughly 0.5 mW/m^2/ster/cm^{-1} then the ratioing method is not suitable for height assignment. From Eq-16 (c):

$$\frac{R_{H_2O}(\nu_1)-R^{cs}_{H_2O}(\nu_1)}{R_{IR}(\nu_2)-R^{cs}_{IR}(\nu_2)}=\frac{\varepsilon_1\int_{p_s}^{p_c}\left\{\tau(\nu_1,p)\frac{dB(\nu_1,T(p))}{dp}\right\}dp}{\varepsilon_2\int_{p_s}^{p_c}\left\{\tau(\nu_2,p)\frac{dB(\nu_2,T(p))}{dp}\right\}dp} \quad (17)$$

Eq-17 the frequencies $\nu_1 \& \nu_2$ are 6.7 micron and 11.5 micron respectively. Assuming clear sky as cloud amount zero (warm) and opaque cloud as cloud amount one (cold), an intersection line is drawn which intersect the calculated radiances (Using radiative transfer Eq-16 c above) curve. The intersection point is the point of determination of cloud top temperature or cloud height. In the H_2O intercept method, the radiances are primarily emanating from the upper troposphere so height determination below 600 hPa is screened out. In the above case, the warmest and coldest field of view radiances within observational area (around 100 km) are the average radiances and the difference between calculate and measured on colder side is adjusted not the warmer side as colder side is the main region of height uncertainties. This provides a valuable tool of dealing the semi transparency type of situation in the determination of cloud heights.

In more general form, let ($R^{cs}_{IR}, R^{cs}_{H_2O}$), ($R^{m}_{IR}, R^{m}_{H_2O}$) and ($R^{cl}_{IR}, R^{cl}_{H_2O}$) are clear sky, measured radiances and opaque radiances pairs, then Eq -17 can be written in the following way:

$$R^{cl}_{H_2O}=R^{cs}_{H_2O}+\left\{\frac{R^{cs}_{IR}-R^{cl}_{IR}}{R^{cs}_{IR}-R^{m}_{IR}}\right\}\left(R^{m}_{H_2O}-R^{cs}_{H_2O}\right) \quad (18)$$

or

Pixel by Pixel level,

$$\frac{R^{cl}_{H_2O}-R^{cs}_{H_2O}}{R^{cl}_{IR}-R^{cs}_{IR}}=\frac{R^{m}_{H_2O}-R^{cs}_{H_2O}}{R^{m}_{IR}-R^{cs}_{IR}} \quad (19)$$

(c) IR window method:

The general from of IR radiation in a non-scattering atmosphere which is in local thermodynamic equilibrium is given by:

$$R(\nu) = B\left(\nu,T\left(p_s\right)\right)\tau\left(\nu,p_s\right)+\int_{p_s}^{p_c}\left[B\left(\nu,T\left(p\right)\right)\frac{d\tau\left(\nu,p\right)}{dp}\right]dp \quad \text{(a)}$$

In the IR window channel the transmittance in the atmospheric layers is nearly constant ($\frac{d\tau\left(\nu,p\right)}{dp}\approx 0$).

$$R(\nu)=B\left(\nu,T\left(p_s\right)\right)\tau\left(\nu,p_s\right) \quad \text{(b)}$$

Let, the cloud level pressure is p_c at cloud level in the FOV of the detector.

$$R_\nu^{obs}=\left(1-n\varepsilon\right)R_\nu^{cs}+n\varepsilon B\left(\nu,T\left(p\right)\right) \quad \text{(c)}$$

The transmittance of the IR window channel, say $\tau\left(n,p\right)$ then,

If clouds have large areal extent (n=1)

$$R_\nu^{obs}=\left(1-\varepsilon\right)R_\nu^{cs}+\varepsilon B\left(\nu,T\left(p\right)\right) \quad \text{(d)}$$

ε = 1- $\tau\left(n,p\right)$

$\tau\left(n,p\right)$, is the transmittance between cloud layers.

$$R_\nu^{obs}=\tau\left(n,p\right)R_\nu^{cs}+\left(1-\tau\left(n,p\right)\right)B\left(\nu,T\left(p\right)\right) \quad \text{(e)}$$

Let, α be the ground albedo, then $\tau\left(n,p_s\right) = 1-\alpha$

(Using Eq-b, above for R_ν^{cs})

We get,

$$R_\nu^{obs}=\tau(n,p)\left(1-\alpha\right)B\left(\nu,T\left(p_s\right)\right)+\left(1-\tau\left(n,p\right)\right)B\left(\nu,T\left(p\right)\right) \quad \text{(f)}$$

For, optically thick clouds, $\tau(n,p)\approx 0$, ground effect is negligible.

$$R_\nu^{obs}=B\left(\nu,T\left(p\right)\right) \quad \text{(g)}$$

Take the inversion of equation (f), we get,

$$T\left(p\right)\cong B_\nu^{-1}\left(R_\nu^{obs}\right) \quad \text{(h)}$$

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