Correction methods for aerosol and thin cirrus effects on remote sensing

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ABSTRACT

Aerosol and cirrus clouds affect the reflected solar radiation detected by satellite sensors. Both aerosol and cirrus increase the apparent reflectance over dark surfaces such as vegetation in the visible channels by scattering sunlight to the sensor. Aerosol reduces the apparent reflectance over bright surfaces (e.g. vegetation and soils in the near IR) by absorbing reflected solar radiation on its way down to the surface and on its way up to the sensor. Correction, removal or avoidance of the atmospheric effects due to aerosol and thin cirrus clouds (we cannot correct for the effect of thick cirrus) can take several pathways:

- Direct correction for aerosol and thin cirrus: The effect of aerosol or cirrus on radiation is determined first and then used for direct correction of the measured radiance to derive the surface spectral reflectance.
- Indirect correction for aerosol: No direct correction is applied but the remote sensing functions designed to find the surface properties (e.g. the NDVI) are redesigned in order to minimize their dependence on the aerosol presence.
- Circumvention of the aerosol effect: In case of heavy aerosol, switching the remote sensing investigation to longer wavelengths, where the aerosol is transparent and has only a small effect, while the details of the surface cover are still evident.

We shall review these techniques and bring some new insight on their applications and limitations, mainly in light of the launch in the summer of 1998 of the MODIS instrument on the NASA Earth Observing System with an array of spectral channels that will allow these correction processes.

Introduction

With the launch of the Earth Observing System (EOS) AM in summer 1998, and parallel efforts in Japan (ADEOS) and Europe (ENVISAT) we are entering a new era in studying the earth as an interrelated system rather an assembly of independent parameters. The Earth Observing system will have an unprecedented spectral and spatial remote sensing capability with scientific instruments that were designed, characterized, and calibrated for the purpose. For example MODIS on EOS-AM will have 5 spectral bands designed for aerosol and land remote sensing in the solar spectrum with resolution of 500m (470 nm, 550 nm, 1.2 μm, 1.6 μm, 2.1 μm) and 2 additional bands with resolution of 250m (650 nm and 860 nm). MODIS will have a special channel at 1.375 μm for highly sensitive remote sensing of cirrus clouds that together with shorter wavelength channels can be used to derive the cirrus radiative effects and to correct for their presence.

One of the main reason for this large investment in remote sensing is the recognition that climate change research based on the effects of greenhouse gases alone is not sufficient to understand climate change and we need to bring in elements with high spatial and temporal variability like aerosol, changes in land use, clouds and water vapor and ocean productivity. The only way to measure these variable effects is from satellites that can provide daily high resolution data. In this respect the introduction of the new satellite systems starts a new stage in the 160 year old climate change research, from the first recognition by Fourier in 1824 that greenhouse gases in the atmosphere can warm the surface, to the first calculations of the effect of doubling of CO_2 and the water vapor feedback by Arrhenius in 1896, based on the IR spectrum of CO_2 and water vapor derived from Langley's precise measurements of the moon IR emission (see review by Ramanathan and Vogelmann, 1997). Presently it is recognized that the greenhouse gases by themselves cannot solve the climate problem and highly varying processes have to be studied. Satellites are uniquely suitable for studying spatially highly variable processes, but they require high accuracy including accurate derivation of the surface properties.

The atmosphere affects remote sensing of the earth surface from space by scattering and absorption by aerosol particles (e.g. micron size dust particles or submicron pollution particles suspended in the air) and molecules in the atmosphere. The effect of aerosol and molecular scattering is stronger in the shorter solar wavelengths where the particle size is similar to the radiation wavelength. Absorbing gases (e.g. water vapor, ozone and oxygen) absorb in specific spectral bands (e.g. Tanré et al., 1992). Aerosol scattering is broader, it increases the apparent surface reflectance over dark surfaces while aerosol absorption reduces the apparent brightness of bright surfaces (Fraser and Kaufman, 1985; Kaufman, 1989). While the concentrations of atmospheric gases do not vary significantly, the concentration of aerosol and water vapor can vary by more than an order of magnitude.

Correction for the effect of molecular scattering, ozone and oxygen absorption is important, since even though their concentration does not change substantially from climatologic values, their effect on radiation detected by the satellite and on the apparent surface reflectance will vary as a function of the view and illumination direction (e.g. Tanré et al., 1992). The effect of water vapor absorption is very significant for the NOAA-AVHRR near IR channel, that is used together with the visible channel to derive the normalized vegetation index (NDVI) used to monitor the dynamics of global vegetation (Tucker et al., 1985; Holben, 1986; Tanré et al., 1992). The newly designed space instruments, e.g. MODIS, have narrower channels that are almost not sensitive to water vapor absorption. Review of the aerosol properties on remote sensing is given by Kaufman (1989), review of the effect of aerosol and gases on remote sensing from AVHRR by Tanré et al. (1992). The detection of cirrus clouds by MODIS was introduce by B.-C. Gao (e.g. Gao et al., 1996).

Direct correction for the aerosol effect

Direct correction of the aerosol effect is based on determination of the aerosol opacity (expressed by the optical thickness and the resulting path radiance) from specific pixels in the image and applying it to correct the same image (Kaufman and Sendra, 1988; Holben et al., 1992; Vermote et al., 1997). Once the optical thickness is determined, the atmospheric correction can proceed using inversion of the basic radiative transfer equation (e.g. Vermote et al., 1997). To determine the aerosol optical thickness, the surface reflectance, ρ, has to be small and known with good accuracy (e.g. within uncertainty of $\Delta\rho=\pm0.005$). Vegetated surfaces are dark in the red and blue parts of the spectrum (Kaufman et al., 1997a) and their reflectance can be estimated using their reflectance in the mid-IR (2.15 or 3.75 μm - Kaufman and Remer, 1994; Kaufman et al., 1997b). This method of using the mid-IR reflectance for correction in the visible channels, is possible for two reasons:

- Aerosol is mostly transparent in the long wavelengths, because the optically active particles are small in case of biomass burning or urban/industrial emissions. For dust the particles are large but their refractive index is significantly smaller in the mid IR reducing their backscattering (see Fig. 1) and atmospheric effect at 2.1 μm.

- There is a good correlation between the surface reflectance in the visible channels and the mid-IR channels (Kaufman et al., 1997a). The correlation is caused by shadows introduced by vegetation in all the wavelengths. Also while vegetation introduces chlorophyll absorption in the red and blue channels, it introduces liquid water absorption in the mid-IR. Apparently the vegetation chlorophyll and liquid water content are correlated to some extend. In general it was found in several ecosystems that the surface reflectance in the red (0.66 μm) is half of that at 2.1 μm for dark surfaces with reflectance at 2.1 less then 0.15 (Kaufman et al., 1997a).

To derive the aerosol optical thickness from the satellite measured radiance and to use it in atmospheric correction, it is important to apply the same aerosol model in both cases and to have a model that represent the total column of ambient aerosol. For this purpose the aerosol model has to be based on inversion of sky and sun radiance measured from the ground (Kaufman and Holben, 1996; Kaufman and Tanre, 1996). Such model was developed recently for smoke and urban/industrial aerosol (Remer et al., 1996). The model is dynamic in the sense that the composition of the aerosol modes varies with the aerosol loading represented by the aerosol optical thickness. Table 1 summarizes these models. The models are given graphically in Fig. 2.

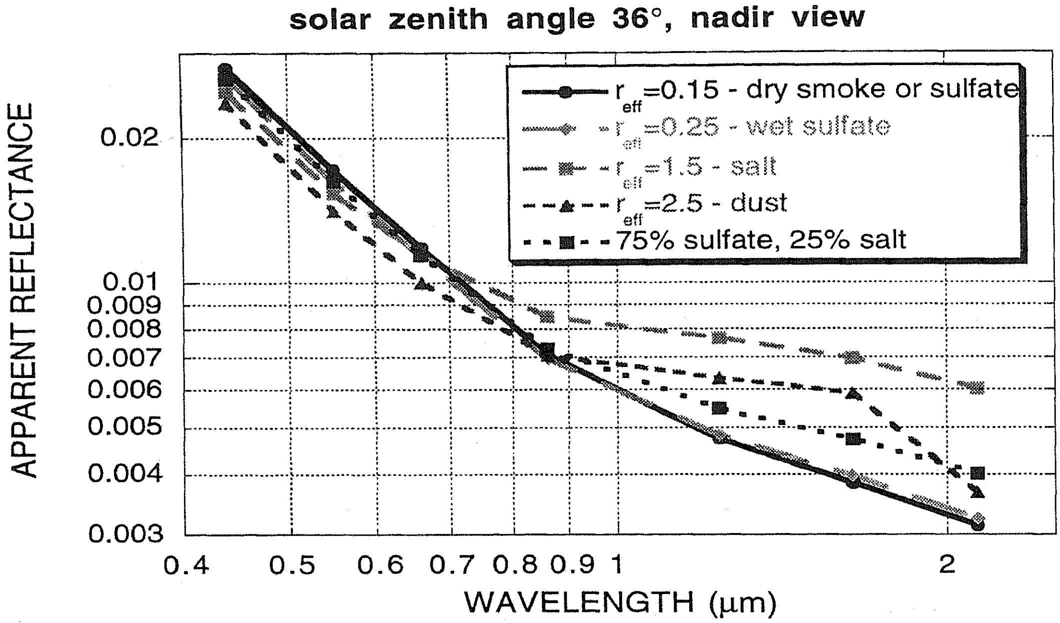

Figure 1: Typical spectral radiance (normalized to apparent reflectance by multiplying the radiance by π/F_O - F_O is the solar flux) for aerosol optical thickness of 0.2 at 0.86 µm over a black surface, for several aerosol types: small accumulation mode particles that correspond to smoke or dry urban/industrial aerosol ("sulfate"); wet larger "sulfate" particles, salt, dust, and a mixture of sulfate and salt (Tanré et al., 1997). Note the reduction with wavelength of the apparent reflectance for all aerosol types. Most aerosol types have a very small effect on the apparent reflectance at 2.1 µm due to small particle size (smoke and sulfates) or small refractive index (dust).

Indirect corrections

In the direct correction for the aerosol effect, the measured radiances is directly corrected for the aerosol effect by deriving the surface reflectance from the measured radiance for the measured aerosol loading and optical properties. In the indirect corrections, a remote sensing function, designed for remote sensing of the surface properties (such as the NDVI used for remote sensing of vegetation - Tucker et al., 1985) is redefined in a new form that minimizes the dependence on the aerosol loading. The aerosol loading and optical properties are no longer needed. One such function is the Atmospheric Resistant Vegetation Index (ARVI) that replaces the red channel used in the NDVI, with a combination of the red and blue channels (Kaufman an Tanré, 1992; 1996). This redefinition of the NDVI as ARVI retains a similar information about the surface properties while being significantly less sensitive to the aerosol effect. We can demonstrate the ability of ARVI to reject the aerosol effects with the following simple calculations. The NDVI is defined by the normalized difference between the apparent reflectance of the surface at 0.86 μm, ρ_{86}, and that at 0.66 μm, ρ_{66}:

$$\mathrm{NDVI} = \frac{\rho_{86} - \rho_{66}}{\rho_{86} + \rho_{66}}.$$

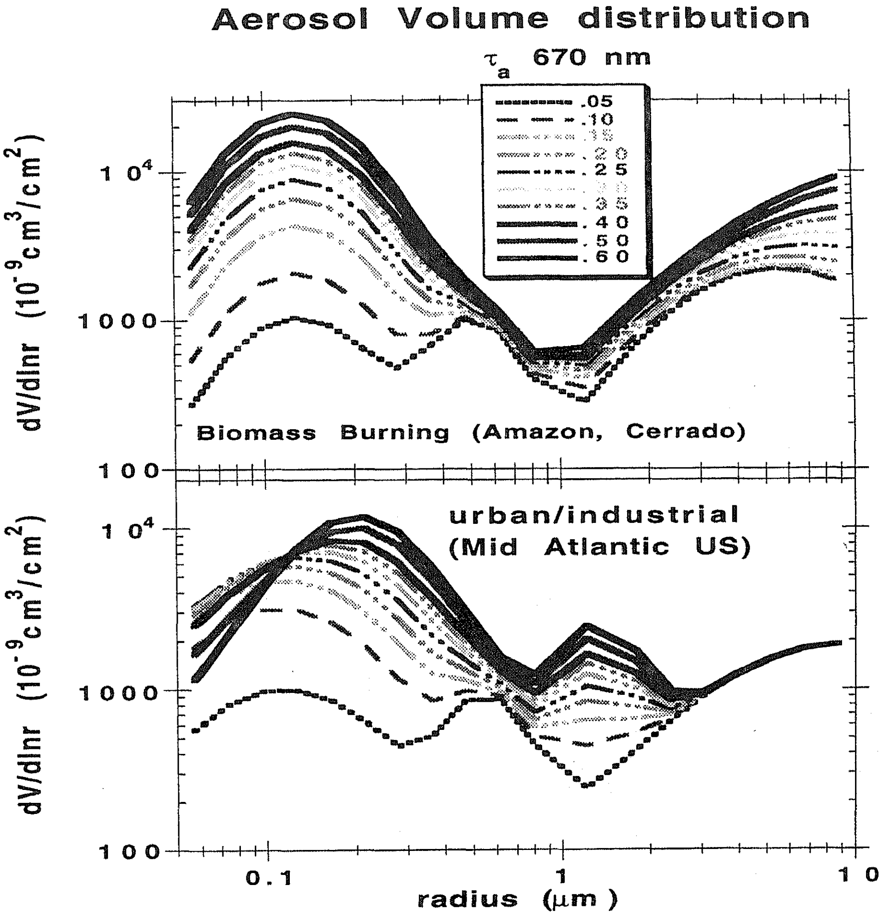

Fig. 2: Aerosol volume distribution for the aerosol models derived from AERONET 1993 deployments, in the Amazon - smoke aerosol and in the Mid Atlantic region of the US - industrial/urban aerosol (Remer et al., 1996). The modes are for: r<0.3 μm - accumulation mode (organic smoke or sulfates); for 0.3 μm<r<0.8 μm stratospheric aerosol; for 0.8 μm<r<2.5 μm maritime salt particle mode; for 2.5 μm<r the coarse particle mode.

The apparent reflectance is the radiance measured in the top of the atmosphere, L, normalized to reflectance units ($\pi L/F_0\mu_0$) using the solar spectral flux F_0 and the cosine of the solar zenith angle μ_0. In this definition the aerosol effect can double the apparent reflectance of vegetation at 0.66 μm for moderate aerosol loading. Therefore a new function is defined, ARVI, based on the same principle of remote sensing of vegetation as NDVI, but replacing the problematic apparent reflectance at

0.66 μm with a corrected reflectance:

$\rho_{66} \rightarrow \rho_{RB} = 2\rho_{66} - \rho_{47}$, and therefore: $\mathrm{ARVI} = \dfrac{\rho_{86} - \rho_{RB}}{\rho_{86} + \rho_{RB}}$,

Note that the apparent reflectances of ARVI are defined after corrections for molecular scattering (Kaufman and Tanré, 1992).

Table 1: Summary of the aerosol dynamic models for biomass burning and urban/industrial aerosol. r_g - the number distribution mode radius, r_v - the volume distribution mode radius, σ - the std. of the logarithm, of the radius, V_o - the volume of the mode and ω_o - the single scattering albedo

continental aerosol (RH=70%)	**r_g(μm)**	**r_v(μm)**	**σ**	**V_o(10^6 μm^3/cm^2)**	**ω_o**
accumulation					
coarse					
biomass burning	**r_g(μm)**	**r_v(μm)**	**σ**	**V_o(10^6 μm^3/cm^2)**	**ω_o**
accumulation	0.061	0.13	0.5	$-2.4+45\tau$	
stratospheric	0.38	0.51	0.31	0.984	
coarse	$1.0-1.3\tau$	$6.0-11.3\tau+61\tau^2$	$0.69+0.81\tau$	$2.4-6.3\tau+37\tau^2$	
industrial/urban aerosol	**r_g(μm)**	**r_v(μm)**	**σ**	**V_o(10^6 μm^3/cm^2)**	**ω_o**
accumulation 1	0.036	0.106	0.60	$-2.0+70\tau-196\tau^2+150\tau^3$	
accumulation 2	0.114	0.21	0.45	$0.34-7.6\tau+80\tau^2-63\tau^3$	
stratospheric	0.43	0.55	0.29	0.73	
salt	0.99	1.30	0.30	$-0.16+4.12\tau$	
coarse	0.67	9.50	0.94	1.92	

Why should introduction of the blue channel, heavily contaminated by aerosol reduce ARVI dependence on the aerosol effect? The main reason ARVI is useful is that the aerosol decreases with wavelength while vegetation reflectance is unchanged and soil reflectance increases with wavelengths. Consider a representative case of aerosol with the aerosol effect in the blue twice of that in the red. The effect of aerosol at 0.86 μm is considered to be negligible due to net zero balance between the aerosol scattering and absorption effects above the bright surface. We assume the vegetation reflectance to be the same in the red and blue channels. Therefore aerosol causes the following transformations:

ρ_{66} ==> $\rho'_{66} = \rho_v + 0.02$, ρ_{47} ==> $\rho'_{47} = \rho_v + 0.04$, ρ_{86} ==> $\rho'_{86} = \rho_{86}$

and therefore:

ρ_{RB} ==> $\rho'_{RB} = \rho_v$ is unchanged and

ARVI ==> ARVI' = ARVI is also unchanged

The transformation would change NDVI:

$NDVI = (\rho_{86} - \rho_v)/(\rho_{86} + \rho_v)$ ==> $NDVI' = (\rho_{86} - \rho_v + 0.02)/(\rho_{86} + \rho_{66} + 0.02)$

While NDVI changed, ARVI is not influenced in this case by the aerosol presence.

Avoiding the aerosol effect- spectrally

Alternative technique to eliminate the aerosol effect instead of the atmospheric direct or indirect corrections is avoidance of the atmospheric aerosol effect by using wavelengths longer than 1 micron, where the aerosol effect is minimal. In the range of 1.2-1.6 μm the land surface is bright, and the aerosol opacity is smaller then in the visible wavelengths. Therefore the net aerosol effect on the apparent surface reflectance is very small (Fraser and Kaufman, 1985; Kaufman et al., 1997a). At 2.1 μm the aerosol effect is very small even for dark surfaces (see Fig. 1). The use of spectral bands beyond 1 μm for remote sensing of vegetation was not demonstrated in general but its capability is shown in Fig. 3. Here a heavy smoke covering the region around Cuiaba, Brazil prohibits the surface observations in an otherwise very clear image. A color image in the longer wavelengths depicts clearly the river, vegetation, bare soils and burn scars and is almost completely independent of the presence of the smoke.

Correction for cirrus clouds using the 1.37 μm channel

Remote sensing in the solar channels traditionally avoided the water vapor absorption bands looking for atmospheric windows to observe the land, ocean, aerosol and clouds. But absorption bands also provide valuable information, in a similar way to the applications of absorption bands in the IR. The 0.89 -0.96 μm water absorption region is used on POLDER and MODIS instruments on ADEOS and EOS respectively for remote sensing of total precipitable water vapor. This technique uses the reflected sunlight that travels through the atmosphere to the surface and back to space through the water vapor to measure the total precipitable water vapor concentration in the vertical column (e.g. Kaufman and Gao, 1992). The 1.375 μm strong water vapor absorption band, discovered by Bo-Cai Gao and adopted for MODIS (Gao et al., 1996) will be used mainly to derive the reflective properties of cirrus clouds and to correct for their optical effects on remote sensing of land, ocean, aerosol and low altitude clouds (Gao et al., 1998). Figure 4 demonstrates this remote sensing application. While the atmospheric window at 0.65 μm allows the observation of the land, ocean and the cirrus clouds, the opacity of the lower troposphere at 1.375 μm due to the strong water vapor absorption, does not allow light to penetrate to the surface and we observe only scattered light from higher altitudes by the cirrus clouds. If the cirrus clouds would not have been present the image would be black. This information can be used to subtract the cirrus effect on the path radiance as shown in the bottom of Fig. 4. But the cirrus cloud effect on the path radiance at wavelengths shorter then 1.375 μm is stronger then that at 1.375 μm due to the effect of the residual water absorption in this channel. Over ocean the effect of the unknown amount of water vapor on the reflection of sunlight by the cirrus clouds can be derived from scatter plots of the cirrus reflectance at the 1.375 μm and the nearby 1.2 μm channel. Since the reflectance of the cirrus crystals does not change substantially between these two wavelengths due to their large size, and spectrally stable refractive index, the ratio between the cirrus effect at 1.375 μm and 1.2 μm is the water vapor transmitance in the viewing and illumination directions of the particular image. Over the land a similar method can be used only if there are some dark surface pixels, e.g. vegetation at 0.66 μm. Fig. 4, following Gao et al. (1998), demonstrates this correction.

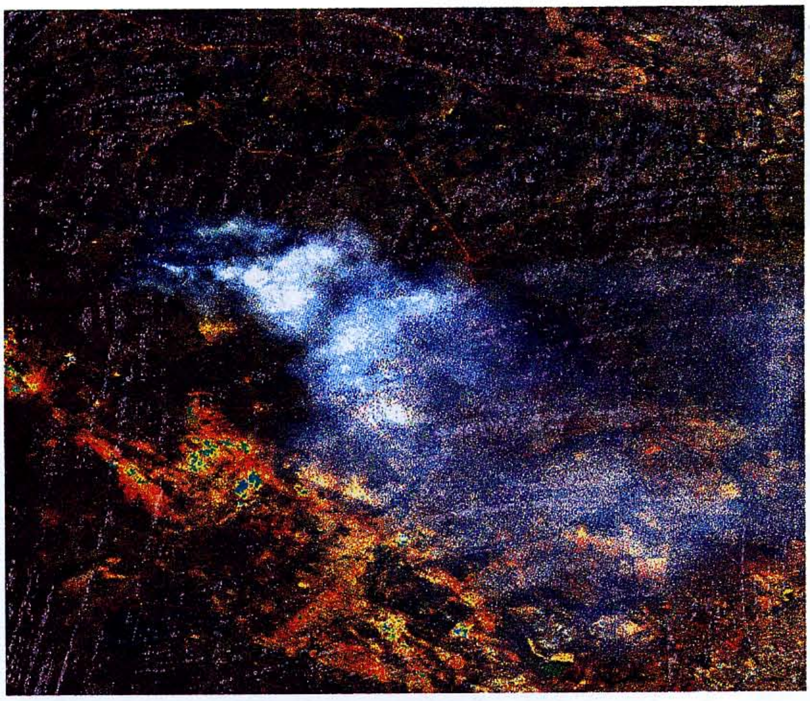

Fig. 3: Large fire near Cuiaba on Aug. 25, 1995 taken from the ER-2 AVIRIS instrument during the SCAR-B experiment. The image is 10x20 km and 20 m resolution. Top image - heavy smoke emitted from the fire and flowing over Cuiaba. The image resembles human vision and is composed of 0.47 μm (blue), 0.55 μm (green) and 0.66 μm (red). Bottom image - is for 2.1 μm (blue), 1.2 μm (green) and 1.65 μm (red). The smoke is almost transparent in these longer wavelengths and the fire is clearly seen with its 3 main temperatures zones (blue - glowing, purple - smoldering - emitting the heavy smoke, and white - the fire front. Note that it is much easier to observe the surface features in the long wavelengths that penetrates the smoke. The solar zenith angle is 36°. The AVIRIS data were provided by Robert Green from JPL.

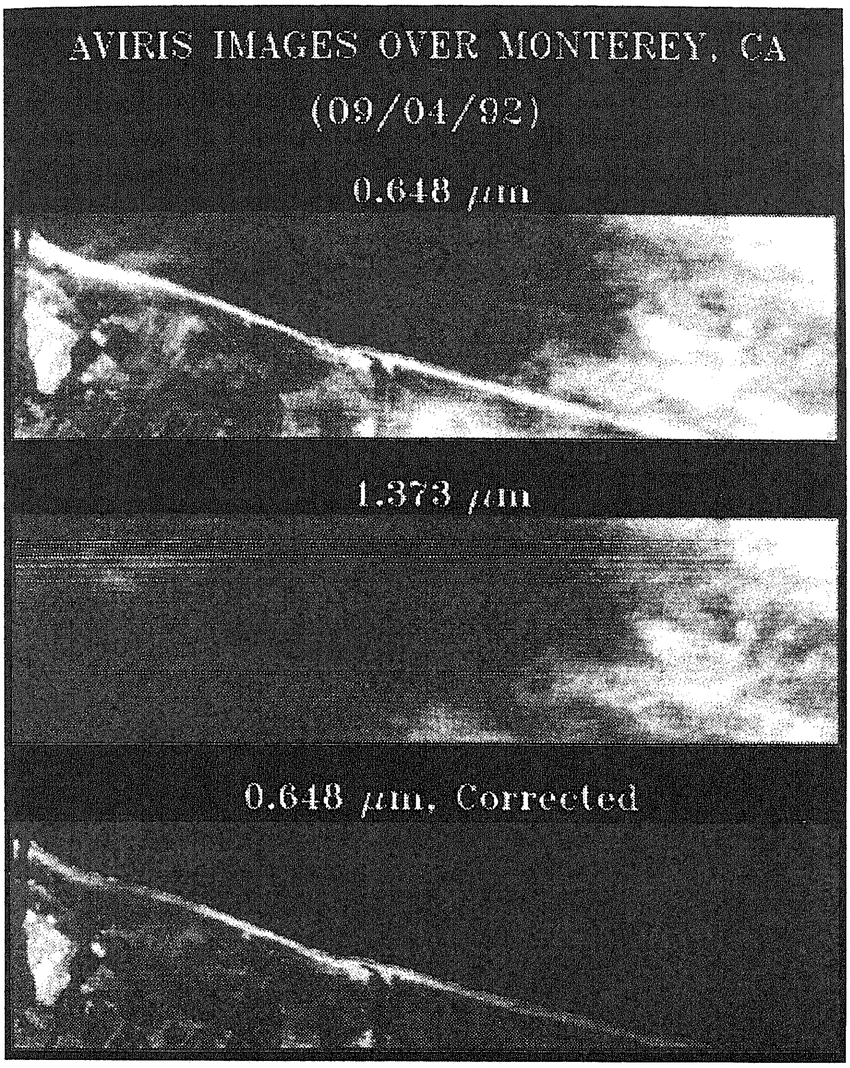

Fig 4: Application of the MODIS new 1.375 μm cirrus channel, using AVIRIS data. Top: image of an ocean-land region at 0.65 μm, indicating the presence of thin clouds over the land and ocean. Middle image - image from the 1.375 μm cirrus channel, observing high clouds only, no land-ocean signal. The difference between the land and ocean reflectance is eliminated by the strong water vapor absorption in the lower troposphere in this channel. Bottom image - is the same as the top image but corrected for the cirrus effect (After Gao et al., 1998).

Conclusions

AEROSOL: Although direct corrections for the aerosol effect over the land are possible, they do not constitute the only, or the best possibility to reduce or get rid of the aerosol effect on remote sensing of the surface. Indirect corrections by redefining functions that represent the surface properties, but are less dependent on the aerosol presence is one possibility. Switching to remote sensing in longer wavelengths (1.2 μm to 2.1 μm) where aerosol has a small effect may be another possibility. Note that the longer wavelengths are not sensitive directly to the chlorophyll concentration but rather to the liquid water in the vegetation associated with it. Thus application of the longer wavelengths for remote sensing of vegetation requires additional studies. The image presented here and others from Brazil show excellent ability to distinguish between different vegetation densities, and lack of sensitivity to the presence of even very heavy and nonhomogneous smoke. The direct correction for the aerosol effect is also helped by the low optical effects of the aerosol in long wavelengths, e.g. 2.1 μm. For smoke or urban/industrial aerosol the high transparency at 2.1 μm is due to the small particle size of the accumulation mode relative to the wavelength. The accumulation mode is the dominant mode affecting the solar radiation. For dust the transparency is due to the low real refractive index of the dust at 2.1 μm. Application for dust of this method was not demonstrated yet. The method for direct correction is completed by finding a correlation between the surface reflectance at the visible channels and the 2.1 μm channel, for part of the pixels in the image (e.g. vegetation). In this way a complete circle is achieved in which the surface properties of some pixels are observed in the mid IR, and used in the visible to derive the aerosol amount and to perform the correction of the entire image. Applications of this technique were performed so far only in limited scope and full demonstration and evaluation awaits the launch of the MODIS instrument on the Earth Observing System.

CIRRUS CLOUDS: The newly found and implemented cirrus cloud channel in the middle of the strong water absorption band at 1.375 μm was shown to be very sensitive to the presence of cirrus clouds. We anticipate to find from MODIS that a larger fraction of the earth is covered by thin cirrus clouds that were not detected with previous satellite systems. To use this channel for atmospheric corrections two steps are needed. For correction for the increase in the path radiance (the radiance measured by satellite for zero surface reflectance) the signal in 1.375 μm needs to be corrected for the residual effect of water vapor. This is done with the help of correlation with the 1.2 μm over water and 0.65 μm over vegetation. But for correction of the cirrus clouds for their effect on attenuation of the solar radiation reflected from the surface, a complete evaluation of the optical properties of the cirrus clouds in needed. Due to the high altitude of the clouds, the presence of the cloud in one pixel affects the observation of the surface in another pixel. A method to correct for these effects still awaits the ingenuity of young scientists.

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