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Study of Atmospheric Aerosols And Their Measurements

V.L.Kiranmai,[1] P. Vishnu Prasanth,[2] T. Kiran Kumar[3]

JNTU-Ananthapur, Ananthapur, Andhra Pradesh, India

[2]Sree Vidyanikethan Engineering College, Rangampet,

Andhra Pradesh, India

[3]Sri Venkateswara University, Tirupathi.

ABSTRACT:

Atmospheric aerosols are considered to be one of the most complex and important parameters affecting the climate change. According to IPCC, Scientific understanding of Atmospheric Aerosol forcing on climate change is very low. Estimating the impact of Atmospheric Aerosols on climate is very complex as they differ in chemical composition and optical properties viz. Scattering and absorbing. Precipitation phenomena is influenced by aerosols hence measurement of atmospheric aerosols over Indian sub-continent is necessary as the Indian economy is highly dependent on Monsoon rainfall. AOD (Aerosol Optical Depth) from MODIS (Moderate Resolution Imaging Spectroradiometer) onboard Terra and Aerosol Index (AI) from OMI (Ozone Monitoring Instrument) onboard Aqua is used to measure the seasonal variation of Aerosols. The Study of Aerosols measurement is highlighted to investigate their behaviour during the seasons. The Topographic effects on Aerosol loading is observed and it is less over elevated regions. Aerosol index from OMI strengthensthe statement, shows that AOD is high during monsoon. The transport of dust desert is also observed from the MODIS true color images may be one of the source mechanisms.

Keywords: MODIS, SSA, AOD, Pyrheliometer, IGB

- **Atmospheric Aerosols and their importance in the Climate System**

Aerosols are a combination of different particles suspended in air with various sizes, shapes, composition. They have different chemical, physical, and radiative properties which vary with location, season and meteorological conditions, their sources and production processes. Based on their size, aerosols are broadly classified into three categories: 1) Nucleation mode (radius with 0.001-0.1 μm), 2) Accumulation or Fine mode (radius with 0.1-1.0 μm), and 3) Coarse mode (radius>1.0 μm). Aerosols in the size range 0.01 μm to 100 μm are important in the study of aerosol-radiation interactions. Aerosols between about 0.1 μm and 2.5 μm in radius are of main interest to climate precipitation, visibility, and human health studies. Most aerosols of interest are found in the troposphere and concentrated toward the Earth's surface. Aerosols can be of both natural and anthropogenic (man made) origins Aerosols of different sizes are shown in detail in the Figure 1.

Aerosols are an important constituent part of the atmosphere not only with respect to climate change, but also to air quality including transboundary transport of air pollution, human health and mortality and visibility (*Davidson et al 2005*).The importance of role of atmospheric aerosols in the Earth's radiation budget and hydrological cycle (*Ramanathan et al 2000*) has enhanced scientific interest in study of aerosol physical, chemical and optical properties. Aerosols provide negative radiative forcing, or cooling through their direct and indirect effects. Anthropogenic aerosols could thus compensate the positive radiative forcing by increasing concentrations of greenhouse gases to a significant, but largely unknown extent. Despite the large numbers of studies on aerosol climate interaction, the uncertainty range of aerosol indirect effects has actually increased from the second to the third assessment of the Intergovernmental Panel on Climate Change (IPCC), and the level of scientific understanding was still classified as 'very low'. The impact of Aerosols on the climate is termed Direct and Indirect and Semi Direct Effect

- **Important parameters for measurement of Atmospheric Aerosols**

The optical activity of aerosols in the atmospheric column can be summarized by the aerosol optical depth, which is a wavelength dependent measure of the total extinction of sunlight due to scattering and absorption by aerosols. Additional information about the light scattering properties of the aerosols is required for the determination of the single scattering albedo (SSA) and asymmetry parameters which are essential in the calculation of aerosol radiative forcing. Still AOD is the single most comprehensive variable to assess the aerosol load of the atmosphere and represents the least common denominator by which ground based remote sensing; satellite retrievals and global modelling of aerosol properties are compared.

Aerosol optical depth(AOD) is a measure of the amount of incident light either scattered or absorbed by airborne particles. Formally, aerosol optical depth is a dimensionless quantity, the integral of the product of particle number concentration and particle extinction cross-section (which accounts scattering and absorption for individual particle), along a path length through the atmosphere, usually measured vertically. In addition to AOD, particle size, composition, and structure, which are mediated both by source type and subsequent atmospheric processing, determine how particles interact with radiant energy and influence the heat balance of the planet. Size and composition also determine the ability of particles to serve as nuclei upon which cloud droplets form. This provides an indirect means for aerosol to interact with radiant energy by modifying cloud properties. Among the main aerosol properties required to evaluate their effect on radiation is the **single scattering albedo(SSA),** which describes the fraction of light interacting with the particle that is scattered, compared to the total that is scattered and absorbed. Values range from 0 for totally absorbing (dark) particles to 1 for purely scattering ones; in nature, SSA is rarely lower than about 0.75. Another quantity, the **asymmetry parameter(*g*)**, reports the first moment of the cosine of the scattered radiation angular distribution. The parameter *g* ranges from -1 for entirely back-scattering particles, to 0 for isotropic (uniform) scattering, to +1 for entirely forward-scattering. The **surface albedo**(*A*), a measure of reflectivity at the ground, which, like SSA, ranges from 0 for purely absorbing to 1 for purely reflecting. In practice, *A* can be near 0 for dark surfaces, and can reach values above 0.9 for visible light over snow. AOD, SSA, *g*, and *A* are all dimensionless quantities, and are in general wavelength dependent. About 10% of global atmospheric aerosol mass is generated by human activity, but it is concentrated in the immediate vicinity, and downwind of sources. These anthropogenic aerosols include primary (directly emitted) particles and secondary particles that are formed in the atmosphere. Anthropogenic aerosols originate from urban and industrial emissions, domestic fire and other combustion products, smoke from agricultural burning, and soil dust created by overgrazing, deforestation, draining of inland water bodies, some farming practices, and generally, land management activities that destabilize the dust to wind erosion. The amount of aerosols in the atmosphere has greatly increased in some parts of the world during the industrial period, and the nature of this particulate matter has substantially changed as a consequence of the evolving nature of emissions from industrial, commercial, agricultural, and residential activities, mainly combustion-related. One of the greatest challenges in studying aerosol impacts on climate is the immense diversity, not only in particle size, composition, and origin, but also in spatial and temporal distribution .For most aerosols, whose primary source is emissions near the surface, concentrations are greatest in the atmospheric boundary layer, decreasing with altitude in the free troposphere. However, smoke from wildfires and volcanic effluent can be injected above the boundary layer; after injection, any type of aerosol can be lofted to higher elevations; this can extend their atmospheric lifetimes, increasing their impact spatially and climatically.

3. Evolution of Aerosol Measurements

Scientific interest in the transparency of the atmosphere started at least 250 years ago with the work of Pierre Bouguer (1698 – 1758), a French mathematician and hydrographer, who in 1729 published his photometric observations of the Sun and the Moonlight and proposed an exponential dependency of the transmission of stellar light through the Earth's atmosphere. The Swiss physicist Horace Benedict de Saussure (1740 – 1799) developed and used a so called **Diaphanometer** to determine the transparency of air as well as a **Cyanometer** to judge the blueness of the sky, which he associated with the cleanliness of the air. With the development of caloric radiometers and the introduction of the compensation principle in the second half of the 19th century, it became possible to replace these visual photometric observations by more accurate quantitative measurements and, by adding spectral filters, to obtain more specific information about the turbidity or transparency of the atmosphere and its extinction components. Claude Pouillet (1791 – 1868) made the first absolute measurements of solar radiation (1838) with his **Pyrheliometer**. Through application of Bouguer's empirical law, he was able to determine the intensity of solar radiation above the atmosphere, the so called solar constant, at 1.79 ly min-1 (1249W m-2) or within 10% of today's accepted value of 1.959 ly min-1 (1367W m-2). The Irish physicist John Tyndall (1820-1893) had examined the optical properties of gases, including absorption spectra of water vapour, ozone and carbon dioxide. He noticed the importance of these atmospheric components for terrestrial climate by saying that "without water vapour, the Earth's surface would be held fast in the iron grip of frost". Lord Rayleigh (John William Strutt, 1842 – 1919) developed his theory for scattering of light by small particles between 1871 and 1899 whereby he could explain the blue colour of the sky as the scattering of sunlight by the molecules of air itself (Rayleigh, 1899). Gustav Mie (1869 – 1957) expanded the electromagnetic scattering theory (Mie, 1908) to particles larger than optical wavelengths and laid the foundation for aerosol optical depth retrieval from radiance measurements.

Since the beginning of the 20th century, routine measurements of the solar radiation were made at a number of astronomical and meteorological observatories in Europe (e.g. Potsdam, Davos, and Uppsala) and USA (e.g. Mt. Whitney) in order to establish climatology of solar radiation or to determine the solar constant. Through the work of Langley and Fowle at the Smithsonian Institution, the solar constant and its spectral distribution became known with sufficient accuracy to permit the determination of atmospheric turbidity from terrestrial measurements of solar radiation. Although the limitation of the Beer-Lambert law to quasi-monochromatic radiation was well known, it was still applied to derive 'effective' or 'complex' extinction coefficients from panchromatic measurements with Pyrheliometer. These simple turbidity coefficients suffered from strong virtual variations depending on solar zenith distance and atmospheric humidity. In order to reduce these diurnal variations, not considered to be part of turbidity, Linke and Boda (1922) proposed their **'Trübungsfaktor'**, which relates the total extinction to the

spectrally integrated Rayleigh extinction. This turbidity factor can be interpreted as the number of pure and dry atmospheres required to match the extinction of the actual atmosphere at a given zenith angle. In a later (Linke, 1942) revision of his Trübungsfaktor, Linke related the total extinction to that of dust-free atmosphere with fixed water vapour content of 1atm-cm. This '**Neuer Trübungsfaktor'** would be representative for dust turbidity, but was still hampered by virtual variations.

More consistent results were obtained from spectrally filtered measurements that allowed the separation of the infrared region dominated by water vapour absorption from total radiation. The infrared part of the solar radiation above 0.625μm was measured through a standard colour glass filter, and subtracted from total measurements to obtain shortwave (Kurzstrahlung) radiation. In analogy to the original Linke factor, a **'Kurztrübungsfaktor'** was then derived that was unaffected by water vapour absorption, but could not eliminate the virtual diurnal variation due to the broadband averaging of the dust extinction.

A different approach to quantify atmospheric turbidity was taken by Anders Ångström (Ångström 1929, 1930) by assuming that the extinction of dust can be modelled similar to the molecular scattering by a power law in wavelength of the form $\beta\lambda$-α. By analyzing the extensive spectrometric data set of the Smithsonian Institution, he could confirm this assumption and found an average value of 1.3 for the exponent α, which he first called size coefficient and associated it with a typical size of the dust particles of about 1μm. He assumed the coefficient β to give a close measure of the total quantity of dust present in the atmosphere and derived its relation to the column number density N experimentally as $\beta = 1.7 \cdot 10\text{-}9\ N$. From the relative stability of α he concluded that the 1μm class of particles dominates the optical extinction of all other size classes and therefore renamed α as wave-length exponent. In fact, the wavelength exponent is closely related to the aerosol number size distribution $n(r)$ which Junge (1952) had found experimentally to be inversely proportional to the particle volume r3. From Mie theory it follows (Junge, 1952) that a power law aerosol size distribution $n(r) = C/r\nu$ will lead to a power law for the extinction with an exponent $\alpha = \nu - 2$. Assuming a constant wavelength exponent α, the turbidity coefficient β could be determined from broadband filter measurements through pre-calculated lookup tables or graphical interpolation, similar to today's satellite retrieval methods.

Schüepp (1949) introduced another model for dust turbidity, closely resembling the Ångström power law, but assuming a wavelength dependency of the wavelength exponent and expressing his coefficient B in decadic form normalized at a wavelength of 0.5μm. For an α exponent of 1.3, Ångström's and **Schüepp's turbidity coefficients** are related by the relation $\boldsymbol{B = 1.07\ \beta}$. Schüepp used a set of Schott filters with sharp cut-off wavelengths and devised a number of graphical procedures (nomograms) to derive his coefficient *B* and the wavelength exponent α. It appears, however, that his laborious method was very sensitive to observational errors and required utmost care in calibration of the radiometers and in cleaning the filters (Robinson, 1966). Valko (1961) later developed a simplified method to determine the Schuepp coefficients as well as the precipitable water content from actinometric measurements.

All of these early efforts based on broadband measurements failed to provide a consistent climatology of atmospheric dust turbidity. The lengthy measurements with filtered actinometers themselves were cumbersome, delicate due to the high precision required in obtaining spectral irradiances from differences of broadband measurements and tedious to evaluate for the variety of corrections to be applied. Consequently, they were pursued at few stations or for short periods only. It was not until Volz (1959) had developed a **handheld sun photometer** that turbidity measurements became more popular. The Volz instrument used a photoelectric selenium cell that is insensitive to infrared radiation, thus eliminating the need for water-vapour corrections and permitted rapid measurements through several filters. With the introduction of metallic interference filters, quasi monochromatic measurements became possible and the evaluation was greatly simplified.

Integrating nephelometer (Beuttell and Brewer, 1949) measure the total amount of light scattered by an aerosol. The integration covers scattering angles from near forward to near backward. To determine the contribution of gases and electronic noise to the scattering signal, the instrument's light scattering response to the filtered air is measured periodically. The contribution of particles to scattering is then determined by difference. When equipped with a photon-counting detector (Charlson et al., 1974), the integrating nephelometer can measure particle light scattering coefficients of less than 0.1 Mm~1, a value equal to about 1% of the light scattering coefficients of particle-free air at normal atmospheric pressure. The design and applications of the integrating nephelometer were reviewed by Heintzenberg and Charlson (1996). Because of its potential for high accuracy, portability, and moderate cost, the nephelometer has been used widely for measurements of light scattering coefficients. Information on wavelength-dependent light scattering provided by multiwave length nephelometers provides useful information on aerosol size distributions (Thielke et al., 1972; van de Hulst, 1981).

Absorption coefficients are most commonly inferred from measurements on particles collected on filters. However, the viability of conducting in-situ measurements using photo acoustic spectrometry has also been demonstrated. These techniques are discussed in this section. Filter techniques are the most common methods for measuring particle absorption coefficients. Because light transmittance through filters is affected by scattering and absorption, the effects of scattering, including multiple scattering, must be accounted for. If light interacts with more than one particle as it passes through the filter, apparent absorption coefficient will exceed the correct value. Also, filter techniques are problematic because the optical properties of deposited particles may be different from those of airborne particles, especially if the particles undergo chemical reactions on the filter.Lin et al. (1973) developed the integrating plate technique for measuring absorption coefficients of particle deposits on filters. This technique was modified somewhat by Clarke (1982) to improve measurement accuracy. Clarke reports that with his modifications this technique can measure absorption coefficients as low as 0.005 Mm^{-1} for a 10-h sampling period. Aerosol absorption coefficients are frequently inferred from measurements of `elemental carbon concentrations. Obtaining an accurate value for absorption

coefficients with this approach requires (1) an accurate measurement of the elemental carbon concentration, (2) knowledge that elemental carbon is the only significantly absorbing particulate species, and (3) knowledge of the "elemental carbon mass absorption efficiency. Foot and Kilsby (1989) compared particle absorption coefficients measured with a filter technique with those measured with a photo acoustic technique. At present Aethalometer is used for measuring the absorption coefficient of Atmospheric aerosols using filter technique.

Ground based AOD observations have high temporal, typically hourly, resolution, but are restricted to a limited number of sites. Although they have been made since at least 50 years at various locations and for different periods, an observational proof of changing atmospheric aerosol burden, on a global scale had not been established by mid 90's (Andreae, 1996). Even worse, a system of sufficient coverage and accuracy did not exist that would allow scientists thirty years from then to assess whether aerosol burdens have changed since the end of the 20th century. Such a system to accurately monitor the global distribution of aerosols requires the combination of continuous observations from satellites, networks of ground-based instruments, and dedicated field experiments (Kaufman et al., 2002).Advances in satellite monitoring capabilities have resulted in the generation of many valuable scientific datasets from local to global scales, which are useful to researchers, policy makers, and the general public. Satellite instruments give us the advantage of making more accurate measurements on a nearly daily basis across a broader geographic area and across a longer time frame.

The first visual observations of atmospheric aerosol effects were made from the manned spacecrafts. Cosmonaut Yuri Gagarin observed clouds and their shadows, as well as optical phenomena due to the presence of aerosols, during the first manned space flight on the spacecraft Vostok on April 12, 1961. These first observations were visual in nature but in subsequent space flights, photography was used by cosmonaut G. S. Titov (Vostok-2, August 6, 1961), cosmonaut V. V. Tereshkova (Vostok-6, June 16, 1963), K. P. Feoktistov (Voskhod, October 12, 1964), A. A. Leonov (Voskhod-2, March 18, 1965), and others. They took photos of the horizon in order to estimate the vertical distribution of aerosols. A. G. Nikolaev and V. I. Sevastyanov (Soyuz-9, June 1, 1970) used **hand-held spectrophotometers** to measure the spectrometry of the twilight and daylight horizons, as well as that of clouds and snow. This instrument was also used in several follow-up missions. Stratospheric aerosol measurements using a hand-held sun photometer were made on the Apollo-Soyuz in 1975 (Pepin and McCormick 1976). Further information on the first instrumental observations of the planet from manned aircrafts is given by Lazarev et al. (1987).

One of the first retrievals of aerosol optical depth from space borne measurements of the spectral intensity of the reflected solar light was performed using observations from the **Multi Spectral Scanner (MSS)** onboard the **Earth Resources Technology Satellite (ERTS-1)** [Griggs, 1975; Mekler et al., 1977] and the first operational aerosol products were generated using data from the radiometer on board the **TIROS-N** satellite launched on 19 October 1978. The **Advanced Very High Resolution Radiometer (AVHRR)** [Stowe et al., 2002] onboard TIROS-N was originally intended for weather observations but its capability was expanded to the detection of aerosols. The **Nimbus-7** was launched on 25 October 1978, carrying the Stratospheric Aerosol Measurement instrument (SAM) (McCormick et al. 1979) and the **Total Ozone Mapping Spectrometer (TOMS)**. *While the TOMS were not originally designed foraerosol monitoring, it has since provided the longest measurement record of global aerosols from space (Herman et al. 1997; Torres et al. 2002).* These launches thus marked the beginning of an era of satellite-based remote sensing of aerosols that has lasted over three decades to date.

Since the turn of the century AOD has been retrieved not only over oceans but also over land from measurements of several dedicated aerosol instruments (e.g. POLDER, MODIS, MISR) on a fleet of Earth observing satellites. **MODIS can provide global coverage at moderate temporal, typically monthly, resolution with estimated errors (Kaufmann et al., 2002) of ±0.03 ±0.05AOD over the oceans and 0.05±0.20AOD over land (Remer et al 2005)** Actual uncertainty of satellite retrievals is probably larger, as Myhre et al., (2005) found AOD over ocean from 9 algorithms to differ by at least a factor of 3, and identified cloud screening as the most probable main reason for this large disagreement. Details of different Satellites used for monitoring aerosols are given in the table.1.

Satellite AOD observations with near global coverage are therefore supplemented by ground based networks, notably AERONET (Holben et al., 1998) or EARLINET (Matthias et al., 2004). These networks can provide more accurate AOD observations in the order of 0.01 optical depths for satellite ground truthing. They are also contributing to the emerging aerosol climatology (Holben et al., 2001; Michalsky et al, 2001) at sites with multi-year observations. A WMO workshop (WMO, 2005a) has recently identified about 90 global surface stations with 4 or more years of continuous observations. Roughly half of these stations are well organized in the AERONET project while the remaining ones are operated by various national meteorological services. AOD data from different satellites and different networks throughout the globe are available at the World Data Centre for Aerosols (WDCA) (http://wdca.jrc.it).

S.No	Platform	Instrument	Number of Bands	Period	References
1	Landsat(ERTS-1)	MSS	4(0.5-11)	1972-1978	Griggs(1975)
2	SMS-1,2	VISSR	5(0.65–12.5)	1974-1981	
3	GOES-1~12	VISSR	5(0.65–12.5)	1975-present	Knapp et al (2002)
4	Apollo-Soyuz	SAM	0.83	1975-1975	McCormic et al(1979)
5	GMS1~5	VISSR	4(0.45–12.5)	1977-2005	-
6	TIROS-N	AVHRR	4(0.58–11.5)	1978-1980	-

7	Nimbus-7	SAM-2 CZCS TOMS	6(0.443–11.5) 6(0.312–0.380)	1978-1993	McCormic et al(1979)
8	AEM-B	SAGE	4(0.385,0.45,0.6,1)	1979-1981	ChC Chu and McCormic et al(1979)
9	NOAA 6~16	AVHRR	5(0.58–12)	1979-Present	Stowe et al (1997)
10	ERBS	SAGE2	4(0.386–1.02)	1984-2005	Miscchenko et al.(1999)
12	SPOT-3	POAM-2	9(0.353–1.060)	1991-1996	Ignatov and stowe (2000)
13	ERS-1	ATSR GOME	4(1.6, 3.7, 11, 12) 4(0.24–0.79)	1991-1999	Randall et al (1996)
14	UARS	HALOE	8(2.45–10.01)	1992-2005	Hervig et al(1998)
15	SSD	LITE	3(0.355, 0.532, 1.064)	1994-1994	Gu et al (1997)
16	ERS-2	ATSR-2 GOME	7(0.55–12) 0.24–0.79	1995-Present	Veekfind et al(1999)
17	Earth Probe	TOMS	6(0.309–0.360)	1996-Present	Torres et al.
18	ADEOS	POLDER ILAS OCTS	9(0.443–0.910) 2(0.75–0.78, 6.21–11.77) 7(0.412–0.865)	1996-1997	Herman et al.
19	Orbview-2	SEAWiFS	8(0.412–0.865)	1997-Present	Gordon and Wang(19940
20	SPOT-4	POAM-3	9(0.354–1.018)	1998-Present	Randall et al (2001)
21	TERRA	MODIS MISR	36 (0.4–14.4) 4 (0.45~0.87)	1998- Present	Remer et al (2005) Kahn et al (2005)
22	METEOR-3M	SAGE-3	9(0.385–1.545)	2001-2005	Thomson et al (2007)
23	PROBA	CHRIS	62(0.4–1.05)	2001- Present	Bransley et al (2004)
24	Odin	OSIRIS	0.274–0.810	2001- Present	Bourassa et al (2007)
25	AQUA	MODIS	36 (0.4–14.4)	2002- Present	Savtchenko et al (2004)
26	ENVISAT	AATSR MERIS SCIAMACHY	7(0.55~12.0) 15(0.4–1.05 0.24–2.4	2002- Present	Grey et al (2006) Vidot et al (2008) Graf and Stammes (2005)
27	ADEOS-2	POLDER-2 ILAS-2 GLI	9(0.443–0.910) 4(0.75–12.85) 36(0.38–12)	2002-2003	Zasetsky and Sloan(2005) Murakami et al (2006)
28	MSG-1	SEVIRI	12(0.6–13.4)	2002- Present	Popp et al (2007)
29	ICEsat	GLAS	2(0.532, 1.064)	2003-2003	Palm et al (2002)
30	AURA	OMI HIRDLS	3(0.27–0.5) 21(6–18)	2004- Present	Torres et al (2007) Froidevaux & douglass (2001)
31	PARASOL	POLDER-3	8(0.44–0.91)	2004- Present	Lier et al(2008)
32	CALIPSO	CALIOP	2(0.532, 1.064)	2006- Present	Winker et al (2003)

Table.1

- **Satellite Measurements of Atmospheric Aerosols**

Atmospheric aerosols are exceptionally well suited for satellite remote sensing, because the sun provides a stable light source and the light scattered by aerosols is easily detectable by spectral radiation sensors with high spatial resolution The science of atmospheric aerosol optics is based on well understood fundamental physical principles: Mie theory of light scattering and the theory of radiative energy transfer through scattering and absorbing media. The very same physical laws that govern satellite aerosol detection are applied to atmospheric visibility research. Satellite aerosol detection operate on the general principle of passive remote sensing: (1) The radiation source is the sun; (2) the target is the Earth's surface; (3) the detector is an array of colour sensitive sensors; (3) in between the reflecting surface and the light sensor is the radiatively active atmosphere consisting of aerosols, air, and clouds. The aerosol optical properties and effects are computed directly from the measured values using physically-based procedures. Satellite aerosol remote sensing on the other hand intends to solve the inverse problem of deriving the properties of atmospheric aerosols from the measured optical effects at the top of the atmosphere. The inherent problem is this kind of inversion is that the retrieval of the unknown aerosol properties requires knowledge of those same properties which, of course, is not available. This retrieval paradox is generally resolved by constructing a discrete set of 'aerosol models' and choosing the aerosol model that best fits the satellite optical data. Thus, art of satellite aerosol remote sensing consists of combining the insights from the fields of *atmospheric aerosols* and of *remote sensing.* Satellite aerosol monitoring can significantly enhance the understanding of atmospheric aerosols, including emissions, transport and spatio-temporal pattern of atmospheric aerosols.

The scattering characteristics of pure air are known and their radiative effects can be calculated. In the absence of clouds, aerosols interact with the radiation. The solar radiation reaches the surface either as a direct beam or indirectly as diffuse light scattered by the air and aerosol particles. A fraction is backscattered toward the space. This backscattered solar radiation is the primary signal that is used for the satellite detection of atmospheric aerosols. Most of the incoming radiation is absorbed at the surface but a fraction, R_0, is reflected, the upwelling radiation is again attenuated by the atmosphere. The upwelling radiation is also enhanced by aerosol scattering into the sensors line of sight. The satellite sensor receives the radiation reflected from the surface and also the amount of scattered into the line of sight by aerosols (shown in figure.1.). The immediate challenge of satellite aerosol detection consists of separating and isolating the faint aerosol signal from cloud scattering and surface reflectance. Below we indicate some of the tasks and methods for this signal decomposition.

The radiative signal received by a satellite is the combined contribution of surface reflectance and aerosol reflectance. The separation of the aerosol signal from the surface reflectance can be achieved by the solution of the appropriate radiation transfer equation as derived in (Husar and White, 1976).

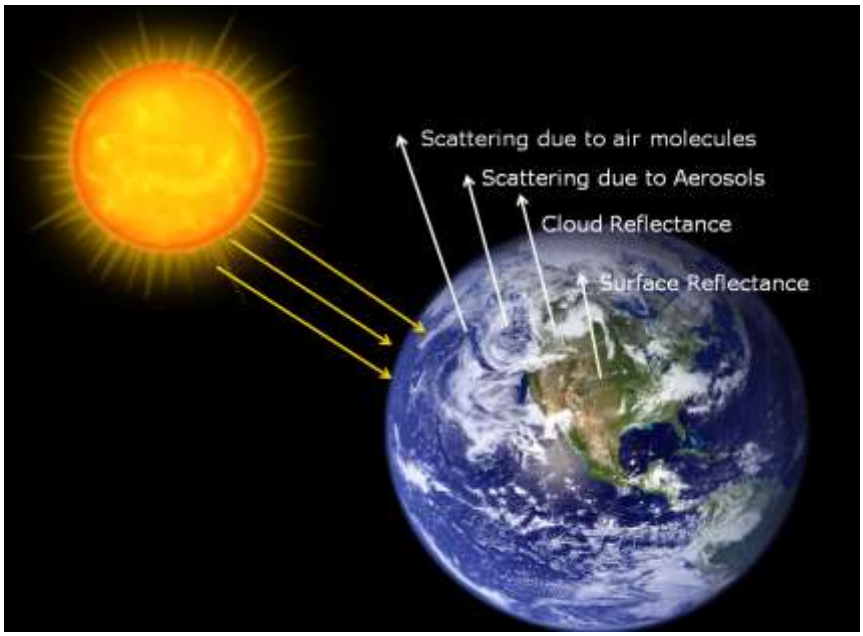

Figure 1. Different types of radiations detected by an Aerosol Satellite sensor

The aerosol contribution to the upwelling radiation consists of two competing effects as shown schematically. On one hand, light scattering particles tend to add reflectance to the brightness of surfaces. This contribution depends both on the aerosol optical thickness, τ, as well as the aerosol scattering phase function, P. The other role of aerosols is to act as a filter which exponentially diminishes the upwelling radiation from a reflecting surface and from aerosol scattering itself. The next interesting case is when the aerosol phase function has the same value as the surface reflectance function, $P = R_0$ in the direction of sensor's line of sight. In this case adding aerosol will not change the apparent reflectance. Since, the aerosol induced extinction of surface radiance is balanced by the aerosol path radiance and aerosol detection is not possible. At $P \sim R_0$, e.g. soil and vegetation at 800 nm, the reflectance is unchanged by typical haze aerosols. The critical measure, whether aerosol will increase or decrease the apparent reflectance is the ratio P/R_0.

The third case occurs when the surface reflectance R_0, is high and the addition of aerosol optical thickness actually diminishes the initial high reflectance of the surface. As the aerosol layer above the bright surface becomes optically thick, $\tau > 4$, the apparent reflectance asymptotically approaches the value of aerosol reflectance. For example, an aerosol layer on top of bright clouds will tend to decrease the brightness of cloud. In this case of an optically thick aerosol layer, the filtering and scattering effects of aerosols are balanced, i.e. the aerosol system reaches radiative equilibrium. The implication is that the retrieval of τ for optically thick aerosol layers ($\tau > 4$) is not possible. Aerosol retrieval is also limited when $P \sim R_0$, which occurs over the bright surfaces of deserts and rocky terrains.

The optical thickness, τ of the aerosol column is derived from the excess reflectance, i.e. the difference between the apparent reflectance of the top of the atmosphere. Through the radiative transfer theory it can be shown that the excess reflectance and the aerosol optical thickness are related by the following simple expression, $\tau = -\ln(R-P)/(R_0-P)$. The value of P is obtained from fitting the observed and retrieved surface reflectance spectra. For example in light haze at 412 nm it is found that $P = 0.38$. Accurate and automatic retrieval of the relevant aerosol P is the most difficult part of the retrieval process.

At blue wavelength (412 nm) the apparent reflectance, R, reaching the satellite sensor is dominated by the haze since the surface reflectance is low over the ocean and vegetation. The aerosol patterns are clearly discernable, while all the surface features are obscured by haze. Hence, the blue wavelength is well suited for aerosol detection over land (Hsu et al., 2006) but surface characterization is difficult. Over the ocean the apparent reflectance in the blue is also dominated by the added aerosol reflectance.

At green wavelength (555 nm), the apparent reflectance is still dominated by haze. At red wavelength (677 nm) , the apparent reflectance is contributed significantly by both surface reflectance and aerosol reflectance. However, the magnitude of the surface reflectance differs greatly for vegetated and solid surfaces, such as urban centres. Hence, at the red wavelength aerosol detection over land is difficult because it is very sensitive to the type of underlying surfaces. Over the ocean the aerosol reflectance is high compared to the water reflectance, and therefore quantitative aerosol detection is possible. The near IR wavelength (865 nm), is uniquely suitable for aerosol detection over the ocean since the ocean reflectance is below 1%. Hence, virtually all the apparent reflectance is due to aerosol scattering. Over land, the reflectance of both vegetated and solid surfaces is high and comparable to the aerosol reflectance P, i.e., $P/R_o \sim 1$ and aerosol detection is not possible.

- **Challenges of Satellite Aerosol Remote Sensing**

As discussed above there are some hindrances for Aerosol satellite remote sensing. As the attenuation of solar radiation due to Clouds is large when compared to that of Aerosols, they can be measured only in the absence of clouds. Gas molecules in the atmosphere also attenuate solar radiation by scattering and hence their effect should be taken in to account while measuring Aerosols. Surfaces of different nature reflect the solar radiation and this reflection may dominate the aerosol effect and thus it is very difficult to measure aerosols over highly reflective regions like deserts and land covered with snow.

5.1.Cloud Effects:

Clouds obscure the detection of both underlying surfaces and atmospheric aerosols. The areas delineated by a 'cloud mask' are unavailable for aerosol detection. Since large portions of the Tropics and also at Northern latitudes are cloudy throughout the year, cloudiness is a major limitation of satellite remote sensing of both aerosols and surfaces. The cloud edges are not easily discernible from aerosols, particularly during humid, hazy conditions. Thin cirrus clouds are also difficult to identify. Consequently, improper cloud masks are a source of error in satellite-derived aerosol fields. Cloud shadows on the Earth's surface also constitute a complication for aerosol retrieval.

5.2.Rayleigh scattering and Molecular Absorption:

Air or Rayleigh scattering of blue light always contributes to the radiation detected by the satellite sensor. The magnitude of air scattering can be corrected by rigorous calculations based on the known optical properties of air molecules, the elevation of the surfaces and by application of well established radiative transfer code, such as offered by (Vermote et al., 2002). Absorption by stratospheric ozone and tropospheric water vapour also attenuate selected bands in the solar spectrum passing through the atmosphere. The stratospheric ozone layer has an interfering absorption band in the visible range (520-740 nm). Whenever possible, the satellite sensor bands are located in transmission windows, i.e. at wavelengths away from major molecular absorption bands.

5.3. Surface Reflectance

From the perspective of aerosol detection, the highly textured and colourful land reflection from the Earth's surface constitutes a significant background signal that needs to be subtracted or minimized. The reflectance, i.e. the fraction of the incoming solar radiation reflected by a surface, depends on many parameters most notably on the nature of the surface itself and the geometric and spectral distribution of incoming and reflected radiation.

Ocean surfaces reflect the incoming radiation, mostly by mirror-like, specular reflection.The areas of 'sun glint' can be identified from the sun-surface-sensor geometry and can be masked out just like clouds. Ocean waves tend to broaden the angular spread of the specular reflection and may also add spurious reflectance from foaming white caps. Over land, the angular dependence of the reflected radiation depends both on the angle of the incoming radiation as well as the angle of reflected radiation and is defined by the Bidirectional Diffuse Reflectance Function (BDRF) (Roujean et al., 1992). Solid land surfaces are mostly diffuse Lambertian reflectors, with minor specular component 2(a). Vegetation is nearly diffuse Lambertian reflector but exhibit an extra reflection 'hot spot' back toward the sun 2.(b).

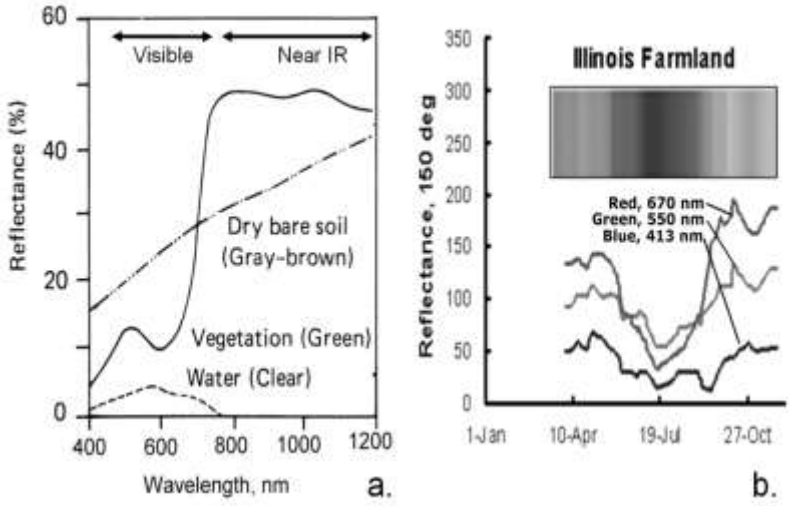

Figure 2. Spectral reflectance of characteristic surfaces.

The BDRF is also strongly wavelength dependent which gives colour of objects (Figure 2). The spectral reflectance of water is low in the visible range and vanishes in the near IR, making it practically black. Vegetation reflectance has a peak in the green (500 nm) and rises sharply to over 40% in the near IR. Typical soil or rock reflectance gradually increases from about 10-15% in the blue to over 40% in the near IR. The spectral reflectance of many surfaces varies with season.

Unlike in situ sensors that can detect the aerosol characteristics in a small volume of air, satellite aerosol sensors detect integral measures of the aerosol over a large volume of air. The ability of satellite sensors to provide global-scale routine monitoring at kilometres scale resolution comes at a cost of limited specificity. The retrieved AOT is an integral over the entire atmospheric column. It is known that aerosols tend to reside in distinct layers of the atmosphere. This integrated signal from all aerosol layers makes it virtually impossible to retrieve the vertical profile of aerosols. However, with space-borne pulsing lidar instruments such as CALIPSO (Winker et al., 2003), even the vertical profile of atmospheric aerosols can be mapped under the narrow path of the spacecraft.

There is a wide array of sensors that have been deployed for the characterization of atmospheric aerosols. Some sensors provide specific aerosol characteristics. For example, the TOMS sensor detects the aerosol absorption which is derived from the extinction of Rayleigh scattering by absorbing aerosols. Since the 1990's, the absorbing Aerosol Index (AI) derived from TOMS (Herman et al., 1997) has been shown to be an extremely helpful indicator of large scale biomass smoke and dust events. However, it is insensitive to non-absorbing aerosols or and cannot detect dust and smoke near the ground.

More sophisticated satellite aerosol sensors allow the retrieval of several aerosol properties including spectral optical thickness and single scatter albedo from which one can infer the aerosol type. These sensors use backscattered solar radiation in multiple narrow wavelength channels (MODIS (King et al., 1992), SeaWiFS, (Gordon and Wang, 1994, McClain et al., 1985), polarization of backscattered radiation (POLDER) (Deuzé et al., 2001), combined angular and spectral scattering (MISR) (Diner et al., 1998). A common goal of the aerosol retrievals is to extract the key properties of the aerosol system, usually the vertical aerosol optical thickness as a function of wavelength. The AOD is a measure of the total aerosol burden and the spectral dependence is a clue on the size-composition of the aerosol column. All the sensors require a mapping of the measured apparent reflectance in a given angle to the intrinsic aerosol property of AOT (λ).This mapping depends strongly on the aerosol properties, pure air scattering and absorption, cloud screening and the spectral reflectance of surfaces. Each of these parameters also depend on the sun-surface-sensor geometry through a highly non-linear radiative transfer processes. The contribution of pure air and atmospheric gases is removed by standard procedures. The cloud masks are also applied based on the spectral and/or texture characteristics of the sensor. The remaining signal still contains both the surface and aerosol reflectance, both contributions are only ill determined and can only be guessed. The task of the inversion routines is to make intelligent guesses at these two parameters. A common technique is to prepare a catalogue of 'models' for aerosol and for the surfaces and to pick the appropriate model based on the best fit approach. The aerosol models and the surface models are typically tailored to the characteristics of the particular sensor. For example, the two-wavelength AVHRR sensor can only accommodate a simple one or two parameter models for aerosols and surfaces, while the multi-spectral MODIS sensor can retrieve multiple aerosol parameters. The most sophisticated MISR aerosol sensor detects both the spectral and angular characteristics of the backscattered radiation and it uses dozens of aerosol models from which a best fit match between the observed and modelled selected. A detailed list of different satellite sensors used for Aerosol monitoring is given in Table.1.

The performance of satellite aerosol sensors is evaluated against observations provided by the AERONET sun photometer network. The federated monitoring network has strategically located sites that characterize typical aerosol regimes such as dusty desert locations biomass smoke areas, marine background, as well as urban industrial sites with complex aerosol mixtures. Over the past decade the co-evolution of the satellite sensors, retrieval algorithms and the continued retrieval verification has yielded a continuous improvement in the performance of satellite aerosol measurement systems.

- **Results and conclusions:**

The Indian subcontinent is one region requiring improvements in characterization of aerosol properties because of an insufficient number of in-situ observations and high aerosol loading. More than one billion people (one sixth of the world's population) living in the Indian sub- continent are exposed to enormous pollution. The countries economy depends mainly on agriculture, which in turn depends on monsoon rainfall. High aerosol loading over the region can affect precipitation [Ramanathan et al., 2001], and hence Aerosol research over the country has gained importance. Investigation of aerosols in India started in 1960s, when Mani et al. [1969] studied Ångström turbidity from solar radiance measurements. During the Indian Middle Atmosphere Program (IMAP), a project was initiated to monitor aerosol characterization over the Indian region by setting up Multi wavelength Radiometers (MWR) at few selected sites. These measurements became operational in 1980s, and have been continued as a part of the Geosphere Biosphere Program (GBP) of Indian Space Research Organisation (ISRO). The ground-based measurements of aerosols in India had been started in 1980s in Trivandrum (south-western coastal city of India) and in few cities in south India (Sikka, 2002) using multi-wavelength solar radiometer (Krishna Moorthy et al., 1999) and in the same year, aerosol vertical distribution measurements by ground based lidar was initiated at Pune (Devara et al., 2002). These events marked the beginning of systematic ground based measurements of aerosol properties in India The Indian Ocean Experiment(INDOEX) (Ramanathan et al., 2001) first addressed the issue of climatic effects of aerosols in this region in a comprehensive manner which initiated ISRO to enhance, coordinated efforts for in situ multi instrument and multiplatform measurements. The result is, the ISRO GBP program has been extended to many other regions with a wide range of Aerosol measuring instruments (Moorthy et al., 2005; Tripathi et al., 2006; Nair et al., 2008). ISRO has conducted several focused campaigns in recent years over India and its surrounding oceans to characterize the chemical, microphysical and optical properties of aerosols. Measurements Aerosol optical and microphysical properties are available in the literature only over a few places in the Indian subcontinent namely Trivandrum [Moorthy et al., 2007a], Pune [Devara et al., 2008], Kanpur [Dey and Tripathi, 2007, 2008], Nainital [Dumka et al., 2008], Ahmedabad [Ganguly et al., 2006a], Hyderabad, Anantpur [Badarinath et al., 2009], Bangalore [Satheesh et al., 2006c], Anantpur [R R Reddy et al] ,Visakhapatnam [Madhavan et al., 2008] Dibrugarh (Gogoi et al ,2011,Kalita et al,2010), Gadanki (Harish and Jayaraman, 2010). A Long term in situ observations at ISROGBP sites along with AErosol RObotic NETwork (AERONET) measurements at Kanpur have provided deep insight into the seasonal variability of aerosol optical properties [Devara et al., 2002; Singh et al., 2004; Sagar et al., 2004; Ganguly et al., 2006a; Adhikary et al., 2007; Moorthy et al., 2007a; Dey and Tripathi, 2008]. For example, long range transport of desert dust during the premonsoon season (March–May) has been found to affect regional aerosol optical properties [Dey et al., 2004; Prasad and Singh, 2007], and mixing of dust with anthropogenic particles [Chandra et al., 2004; Dey et al., 2008] has made the characterization of the aerosols more difficult. Besides these continuous measurements, aerosol microphysical and optical properties were measured in India during four major field campaigns in the last ten years.Indian Space Research Organization-Geosphere Biosphere Program (ISRO-GBP) road campaign experiment was conducted during Feb 2004 (Jayaraman et al. 2006)and a land campaign was carried out in south India during February–March 2004 [Moorthy et al., 2005] followed by a campaign focused in the IGB during the winter season (December–February) of 2004–2005 [Tripathi et al., 2006; Nair et al., 2007]. The major scientific objective of these campaigns was to quantify aerosol direct radiative effects at several locations by using simultaneous in situ measurements of aerosol microphysical, chemical and optical properties. The Integrated Campaign on Aerosol and Radiation Budget (ICARB) was carried out in the pre monsoon season of the year 2006 to measure the physicochemical and radiative properties of aerosols and trace gases and their vertical distributions [Satheesh et al., 2009]. In the pre monsoon and monsoon seasons of the years 2008 and 2009, aircraft and ground based measurements were carried out over the IGB and central India as part of Continental Tropical Convergence Zone campaign to quantify the aerosol indirect effect [Department of Science and Technology, 2008]. All of these studies have greatly improved our under- standing of aerosol microphysical and optical properties and their variability at several locations across the Indian sub- continent and its surrounding waters; however, a complete spatial analysis remains hindered because of the spatially limited nature of these data sets. This is where satellite data become very useful and can complement the in situ measurements. Satellite retrievals of aerosol properties over land have only been available in recent years and a few studies have been done using these data over the Indian subcontinent. Di Girolamo et al. [2004] were the first to study the spatial distribution of AOD over India using Multiangle Imaging Spectroradiometer (MISR) in the winter season during 2001– 2004, where they were able to explain the enormous pollution observed over the IGB based on meteorology, topography and potential aerosol emission sources. All subsequent studies using Moderate Resolution Imaging Spectroradiometer (MODIS) data have confirmed this observation [Jethva et al., 2005; Prasad et al., 2006; Ramachandran and Cherian, 2008] with additional information on the seasonal variability of AOD and fine mode fraction, to some extent. Suresh kumar et al (2010) have made observation over tirupati during May 2009. Hyvärinen et al, 2011a, 2011b studied effect of the summer monsoon on aerosols at stations in Northern India. Spatial distributions and seasonal cycles of aerosols in India are studied by Henriksson et al, 2011. They have reported that the Indian and Chinese aerosol characteristics and seasonal cycles differed from each other. India had higher concentrations of black and organic carbon and China had higher concentrations of sulfate. India experiences a strong natural cycle with a maximum in the spring-summer (May–July) and a strong anthropogenic cycle having a maximum in the winter (November–January).**Their simulations reproduced MODIS-derived AOD seasonal cycles reasonably well for India and China**. In India, monthly average AOD is largest when fine anthropogenic particle pollution is at its lowest.

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