Ground and Satellite-Based Remote Sensing of Mineral Dust Using AERI Spectra and MODIS Thermal Infrared Window Brightness Temperatures

A dissertation submitted in partial satisfaction of the requirements for the degree Doctor of Philosophy in Atmospheric Sciences

by

Richard Allen Hansell Jr.

2008
This dissertation of Richard Allen Hansell Jr. is approved.

Jochen Stutz

Suzanne Paulson

Charlie Zender

Annmarie Eldering

Kuo Nan Liou, Committee Chair

University of California, Los Angeles
2008
# Table of Contents

<table>
<thead>
<tr>
<th>Chapter</th>
<th>Page No.</th>
</tr>
</thead>
<tbody>
<tr>
<td>List of figures.</td>
<td>vi</td>
</tr>
<tr>
<td>List of tables.</td>
<td>ix</td>
</tr>
<tr>
<td>VITA</td>
<td>x</td>
</tr>
<tr>
<td>Abstract</td>
<td>xii</td>
</tr>
<tr>
<td>1. Introduction</td>
<td>1</td>
</tr>
<tr>
<td>2. Background and Motivation</td>
<td>9</td>
</tr>
<tr>
<td>3. UAE$^2$ Field Campaign Overview</td>
<td>33</td>
</tr>
<tr>
<td>4. Instrumentation</td>
<td>39</td>
</tr>
<tr>
<td>5. The Thermal IR Radiative Transfer Equation</td>
<td>52</td>
</tr>
<tr>
<td>And Radiative Transfer Models</td>
<td></td>
</tr>
<tr>
<td>6. Dust Microphysics and Single Scattering Properties in The Thermal IR Window</td>
<td>64</td>
</tr>
<tr>
<td>7. Satellite Dust Detection Using Thermal IR Window</td>
<td></td>
</tr>
<tr>
<td>Radiances: Methodology, Case Studies, and Validation</td>
<td>79</td>
</tr>
<tr>
<td>8. Ground-based AERI Dust Detection and Retrieval: Sensitivity and</td>
<td></td>
</tr>
<tr>
<td>Modeling Studies, Methodology, Case Studies, and Validation</td>
<td>94</td>
</tr>
<tr>
<td>9. Surface Radiative Forcing of Dust Aerosol - Direct Effect</td>
<td>145</td>
</tr>
<tr>
<td>10. Conclusions</td>
<td>208</td>
</tr>
<tr>
<td>11. Bibliography</td>
<td>213</td>
</tr>
</tbody>
</table>
List of Figures

Figure 1.1 IPCC 2001 Report ----------------------------------------------- 8
Figure 2.1 Three Modes of Dust Particle Transport ----------------------------- 27
Figure 2.2 Global Distributions of TOMS Dust Sources ----------------------- 28
Figure 2.3 Micrograph Image of Collected Dust Particles ---------------------- 30
Figure 2.4 X-ray Peak Spectra for Individual Dust Particles ------------------ 31
Figure 2.5 Retrieved volume size distributions from AERONET -------------- 32
Figure 3.1 a. Map of UAE b. Satellite Image of Al-Ain, UAE ------------------- 38
Figure 4.1 a. MAARCO b. SMART --------------------------------------------- 48
Figure 4.2 SMART AERI ----------------------------------------------------- 49
Figure 4.3 Approximate Dwell Times during AERI Measurement Cycle---------- 50
Figure 5.1 Surface Dust Cloud Model ---------------------------------------- 62
Figure 6.1 Imaginary Component of Dust Refractive Indices ----------------- 75
Figure 6.2 Dust Aspect Ratio Cumulative Probability Plot-------------------- 76
Figure 6.3 Re-binned APS Size Distributions Measured during the UAE2,-------- 77
Figure 6.4 Dust single scattering properties in the thermal IR window------- 78
Figure 7.1 Simulations of the $BTD$ and $D^*$-parameter versus Optical Depth($\tau$) ---- 92
Figure 7.2 Comparison of Daytime/Nighttime Integrated Dust Detection with (1) MODIS visible image plots, (2) MPL Profiles and (3) NASA Deep Blue Scheme for Various Locations ----------------------------------------------- 93

Figure 8.1 Sensitivity of AERI ‘clear-sky’ sub-band spectra to changes in PWV - 133
Figure 8.2 Sensitivity of AERI BT spectra to changing RH and Temperature (T) profiles
for three days during UAE2

Figure 8.3 (a) BTD 10-11 vs. dust optical depth for dust (Q) or cirrus only conditions. (b) BTD 10-11 for dust and cloud (c) BTD clear-sky offset uncertainty vs. PWV (d) Dust and cloud (mixed scenario) and BTD offsets.

Figure 8.4 Particle size and shape $BT$ sensitivity across (a) upper sub bands (USB: 13-17) and (b) lower sub-bands (LSB: 1-11).

Figure 8.5 BT sensitivity to dust optical depth at 962 cm$^{-1}$.

Figure 8.6 BT sensitivity to dust cloud height and thickness.

Figure 8.7 AERI BT sensitivity to mineral composition.

Figure 8.8 Estimated radiance error due to MCT detector non-linearity and AERIPLUS retrieved temperature and WVMR profiles.

Figure 8.9 a. AERI detected cloud/dust vs. MPL for 9/30/04 b. AERI retrieved IR optical depths for 9/22/2004 with/without non-linearity corrections applied. c. AOT scatter plot of AERI versus AERONET retrieved optical depths at 0.5µm.

Figure 8.10 UAE2 AERI dust/cloud separation results versus the MPLNET MPL at SMART for two days without the non-linearity correction.

Figure 8.11 AERI daytime and nighttime dust and cloud separation during NAMMA.

Figure 8.12 AERI optical depth retrievals vs. AERONET sun-photometer after cloud screening.

Figure 9.1 (a) The arrangement of SW and LW radiometers at UCLA surface site. (b) Cimel 318A sun photometer at UCLA surface radiation site. (c) Example of downward SW flux profile on a clear day at UCLA surface site. (d) Example of downward shortwave flux profile on a partially cloudy day at UCLA surface site.

Figure 9.2 (a) PSP/CG1 nighttime output correlation (b) Correction-PSP nighttime signals.

Figure 9.3 (a) Scatter ratio time series for PRIDE. (b) Comparison of total solar fluxes with and without application of the cloud screening scheme for PRIDE.
Figure 9.4 TOMS satellite image of African Continent

Figure 9.5 Retrieved Cimel/MFRSR Aerosol Optical Depths

Figure 9.6 Angstrom shaping factor

Figure 9.7 Model vs. Observed Surface Fluxes and SARF

Figure 9.8 Estimated Aerosol Forcing

Figure 9.9 Estimated Forcing Efficiency

Figure 9.10 (a) Model vs. observed LW surface fluxes during UAE2 (b) Absolute differences between model and observed.

Figure 9.11 (a) AERIPLUS vs. MAARCO ambient temperature profile for September 22, 2004 during UAE2. (b) Profile of residuals for (a).

Figure 9.12 Surface forcing sensitivity to mineral composition versus dust optical depth for dust particles with effective size of 0.2 µm.

Figure 9.13 Surface forcing sensitivity to particle shape/size as a function of dust optical depth (a) effective particle size is equal to 0.2 µm (b) effective particle size is equal to 1 µm.

Figure 9.14 Surface meteorology measured at SMART during UAE2 (a) Pressure (b) ambient temperature and (c) relative humidity.

Figure 9.15 Modeled vs. observed LW surface fluxes/forcing as a function of time for 9/22/2004 during UAE2 at SMART site. (a) Model versus observed PIR fluxes. (b) Absolute error for both dust models. (c) Instantaneous LW surface forcing.

Figure 9.16 Linear Variation of Average IR Forcing with AERI Retrieved Optical Depth and Residuals of Linear Fit with Observations

Figure 9.17 Modeled vs. observed LW surface fluxes/forcing as a function of time for 9/23/2004 during UAE2 at SMART site. (a) Model versus observed PIR fluxes. (b) Absolute error. (c) Instantaneous LW surface forcing.

Figure 9.18 Dust LW Surface Forcing Efficiency
List of Tables

Table 2.1 Common Dust Component Densities ------------------------------ 29
Table 4.1 Sun-photometer Channels -------------------------------------- 51
Table 5.1 Fu-Liou RTM - spectral Bands ---------------------------------- 63
Table 5.2 Fu-Liou RTM Surface Specifications ------------------------------ 63
Table 6.1 FDTD Size Parameters ------------------------------------------ 74
Table 8.1 AERI Sub-bands ---------------------------------------------- 129
Table 8.2 Summary of Dust and Cloud Detection Sensitivity --------------- 130
Table 8.3 Summary of Dust Size/Shape Sensitivity for $\tau_{\text{dust}}=1$ -------- 131
Table 8.4 AERI Dust Optical Depth Retrieval Summary --------------------- 132
Table 9.1 Primary Instruments and Measurements for PRIDE, SAFARI and ACE-ASIA ----------------------------------------------- 187
Table 9.2 MFRSR Solar Spectral Constants ------------------------------- 191
Table 9.3 $\tau_a$ Statistics -------------------------------------------- 194
VITA

1988  
BSEE  
Loyola Marymount University  
Los Angeles, California

1988 – 1998  
Electronics Engineer  
Naval Surface Warfare Center  
Port Hueneme Division  
Port Hueneme, California

1999 – 2000  
Research Assistant  
San Fernando Observatory  
California State University, Northridge  
Northridge, California

2000  
MS Physics  
San Fernando Observatory  
California State University, Northridge  
Northridge, California

1999 – 2001  
Research Assistant  
Department of Atmospheric Sciences  
University of California, Los Angeles

2001-2003  
Teaching Assistant  
Department of Atmospheric Sciences  
University of California, Los Angeles

2004  
MS Atmospheric Sciences 2004  
Department of Atmospheric Sciences  
University of California, Los Angeles

PUBLICATIONS AND PRESENTATIONS


Paper presented at 11th Conference on Atmospheric Radiation, AMS, Utah, June 3-7


ABSTRACT OF THE DISSERTATION

Ground and Satellite-Based Remote Sensing of Mineral Dust Using AERI Spectra and MODIS Thermal Infrared Window Brightness Temperatures

By

Richard Allen Hansell Jr.

Doctor of Philosophy in Atmospheric Sciences

University of California, Los Angeles, 2008

Professor Kuo Nan Liou, Chair

The radiative effects of dust aerosol on our climate system have yet to be fully realized and remain to be a topic of contemporary research. To investigate these effects, detection/retrieval methods for dust events over major dust outbreak and transport areas have been developed using satellite and ground-based approaches. To this end both the shortwave (SW) and longwave (LW) surface radiative forcing of dust aerosol was investigated.

The ground-based remote sensing approach uses the Atmospheric Emitted Radiance Interferometer (AERI) brightness temperature spectra to detect mineral dust events and to retrieve their properties. Taking advantage of the high spectral resolution of
the AERI instrument, absorptive differences in prescribed thermal IR window sub-band channels were exploited to differentiate dust from cirrus clouds. AERI data collected during the United Arab Emirates Unified Aerosol Experiment (UAE\textsuperscript{3}) at Al Ain UAE, (8/13-9/30, 2004) was employed for dust retrieval. Assuming a specified dust composition model a priori and using the light scattering programs of T-matrix and the finite difference time domain (FDTD) methods for oblate spheroids and hexagonal plates, respectively, dust optical depths have been retrieved and compared to those inferred from a collocated and coincident AERONET sun-photometer dataset. The retrieved optical depths were then used to determine the dust LW surface forcing during the UAE\textsuperscript{3}. Likewise, dust SW surface forcing is investigated employing a differential technique from previous field studies.

The satellite-based approach uses MODIS thermal infrared brightness temperature window data for the simultaneous detection/separation of mineral dust and cirrus clouds. Based on the spectral variability of dust emissivity at the 3.75, 8.6, 11 and 12 µm wavelengths, the $D^*$-parameter, $BTD$-slope and $BTD_{3-11}$ tests are combined to identify dust and cirrus. MODIS data for the three dust-laden scenes have been analyzed to demonstrate the effectiveness of this detection/separation method. Detected daytime dust and cloud coverage for the Persian Gulf case compare reasonably well to those from the “Deep Blue” algorithm developed at NASA-GSFC. The nighttime dust and cloud detection for the cases surrounding Cape Verde and Niger, West Africa has been validated by comparing to coincident and collocated ground-based micro-pulse lidar measurements.
BLANK PAGE
CHAPTER 1

1. INTRODUCTION

The effects of both anthropogenic and naturally occurring mineral dust on our climate system have yet to be fully realized and remain to be a topic of contemporary research. The current assessment of the radiative effects of mineral dust in the working group summary of the IPCC 2001 report shows large uncertainties in both the magnitude and sign of the direct radiative forcing which are primarily due to uncertainties inherent in the key dust properties, and in the competing effects of both long wave (LW) and short wave (SW) radiative interactions. Based on the works of previous investigators [Hanson 1998, Sokolik and Toon 1996, Tegen and Fung 1995 and others] a tentative global annual mean range of -0.6 to +0.4 Wm\(^{-2}\) was reported with large uncertainties stemming mostly from model assumptions, lack of observational data, and inherent difficulties in the measurement techniques which can often mean the difference between a positive and negative forcing.

Historically aerosol science has focused more on the SW effects of dust aerosol mainly because of the larger availability of dust data in the visible wavelengths. More recently however, there has been a surge of interest in the LW effects which although are much smaller than the SW can still exert a sizeable impact on the net radiative forcing. To accurately assess the radiative impact of dust aerosol, a comprehensive understanding of its complex radiative properties is absolutely critical, particularly in the thermal IR
where common dust minerals ranging from silicates to clays exhibit a wide range of unique spectral features [Sokolik et al. 1999].

Exploiting the LW properties of dust for measuring and modeling high resolution dust spectra has been the focus of recent research efforts including, Desouza-Machado et al. [2006], Hong et al. [2006], Pierangelo et al. [2005] and Pierangelo et al. [2004]. Although significant progress is being made, accurate dust parameterizations for remote sensing and climate applications will strongly depend on the availability of quality global dust data from major source regions around the world.

In response to the growing needs for global dust data, there has been a major increase in field studies designed to measure key dust properties in regions largely affected by dust aerosol. Unique to the work presented in this dissertation was one such field campaign: the United Arab Emirates Unified Aerosol Experiment (UAE2) conducted from August to September, 2004 where over two dozen research organizations convened in the United Arab Emirates (UAE) to participate in field experiments designed to improve our knowledge of dust aerosol. In particular the field study employed a diverse network of ground-based and aircraft mounted instruments including passive/active radiation sensors and aerosol in-situ equipment for measuring dust particle size, shape and composition, all critical parameters for elucidating the radiative effects of dust aerosol. Some of the key instruments used for this study included the atmospheric emitted radiance interferometer (AERI), aerodynamical particle sizer (APS) and LW precision infrared radiometer (PIR). Details of these and other instruments are discussed in chapter 4.
In addition to surface and aircraft mounted sensors, the UAE\textsuperscript{2} also provided an excellent forum for testing/validating satellite dust detection and retrieval algorithms over water and desert surfaces. These included the multi-spectral dust enhancement algorithm [Miller, 2003] which combines dust coloration properties with the MODIS negative 11-12µm brightness temperature difference (BTD 11-12) for the daytime and nighttime detection of dust over dark and bright surfaces and the NASA-GSFC Deep-Blue algorithm [Hsu et al., 2004] which uses the blue wavelengths (λ< 0.5µm) for retrieving aerosol properties over bright reflecting source regions. Other current remote sensing techniques include a method for retrieving dust optical depth and altitude using AIRS measurements [Pierangelo et al., 2004] and MISR retrievals over dark water [Kalashnikova et al., 2005] using optical dust models for complex and non-spherical dust mixtures. Several new dust detection algorithms have also been developed. Roskovensky and Liou [2005] define a dust parameter for differentiating dust from cirrus by combining a dust reflectance ratio in the visible and near IR channels, with a split window difference of the 11µm and 12µm channels. More recently, Hansell et al. [2007] developed an integrated method for the simultaneous detection/separation of mineral dust and clouds for both daytime and nighttime conditions using MODIS thermal infrared window (3.75, 8.6, 11 and 12µm) brightness temperature (BT) data. Details of the integrated dust detection method along with examples of its application to several MODIS day and night dust scenes are presented in chapter 7.

This study uses data from the two ground-based facilities (super sites) deployed during the UAE\textsuperscript{2}: The Naval Research Laboratory’s Mobile Atmospheric Aerosol and
Radiation Characterization Observatory (MAARCO) located along the coast approximately 60 km northeast of Abu-Dhabi at (24°N; 54°E) and the NASA Goddard Space Flight Center’s Surface-sensing Measurements for Atmospheric Radiative Transfer (SMART) located in the interior desert at Al-Ain airport near (24°N; 55°E) just west of Oman. Both sites were strategically located to include the UAE coastal and desert regions to provide comprehensive instrumental coverage for measuring both aerosol properties and surface radiation. Details of the super sites including primary science objectives and pertinent background information related to the region’s meteorology, geology and unique oceanographic features are given in chapter 3.

This dissertation focuses on the following major topics. First a detailed and comprehensive discussion of the dust optical and microphysical properties is given. Dust composition is prescribed using the refractive index datasets for minerals commonly observed around the UAE region including quartz, kaolinite and kaolinite internally mixed with hematite and calcium carbonate. For comparison with the pure minerals, the refractive indices of the Volz [Volz, 1973] bulk dust mixture are also evaluated. Dust microphysical models are constructed using particle size and shape in-situ data from the UAE² and prior field campaigns. Dust single scattering properties for oblate spheroids and hexagonal plates, two particle geometries routinely interpreted in electron microscopy and spheres were computed using the T-matrix, FDTD and Mie light scattering programs.

Next, a new satellite-based technique is presented for the simultaneous detection/separation of mineral dust and clouds for daytime and nighttime conditions
using MODIS thermal infrared window Brightness temperature (BT) data. Several dust and cirrus cases over regions near heavy dust sources are given to demonstrate the effectiveness of the technique. In each case, a collocated and coincident micro-pulse lidar is used for validating the technique’s dust/cloud classification results. This technique offers a variety of promising applications some which include: (1) scene classification for dust radiative forcing studies and (2) assessment of the nighttime dust hazard for improving transportation safety and mitigating dust’s adverse health effects.

A newly developed ground-based dust detection and IR optical depth retrieval method using daytime AERI BT data from the UAE2 field campaign is then presented. The methodology is based on detailed sensitivity and modeling studies of the AERI BT spectra to key dust and atmospheric parameters which include: (1) total column water vapor (2) vertical distribution of column water vapor, (3) vertical distribution of temperature, (4) dust cloud thickness, (5) dust cloud altitude, (6) mineral dust composition, (7) dust optical depth, (8) dust particle size and (9) dust particle shape. The physical basis for the approach relies on the complex spectral variability of the IR optical properties for common mineral dust components. Dust detection follows the physical principles of dust and cloud particle absorption across the thermal IR window while the retrieval scheme employs a $\chi^2$ statistical optimization approach in the AERI ‘clean’ sub-bands for determining the dust IR optical depths.

The combined detection/retrieval method is applied to several UAE2 cases which have prevalent dust and cloud coverage to demonstrate the feasibility of the approach. Collocated and coincident MPLNET micro-pulse lidar and AERONET sun-
photometer measurements provide the observed dust/cloud profiles and visible dust optical depths for validating the technique’s detection and retrieval results respectively.

The method’s application for nighttime remote sensing using recently acquired AERI BT spectral data from the NAMMA field campaign (August-September, 2006) is also discussed. Combining both satellite and ground-based approaches will help provide for complementary dust detection in the atmospheric column which could benefit the tracking of major dust events at both regional and global scales.

Lastly, the dust LW surface radiative forcing during the UAE2 field campaign is determined. The daily retrieved AERI IR optical depths are used in a 1-D column radiative transfer model for computing the LW surface fluxes for both dust and dust-free atmospheres. Model surface fluxes are validated using measured downwelling LW fluxes of a precision infrared radiometer (PIR) located at the GSFC NASA SMART site. The LW surface forcing efficiency of dust aerosol using observations in a differential technique [Hansell et al. 2003] is then calculated for a 10-day period during the UAE2 and is compared to the model instantaneous daily forcing calculations.

This dissertation is organized as follows. Chapter 2 gives the background and motivation for the study. Chapter 3 provides a brief overview of the UAE2 field campaign with emphasis on science objectives, logistics and critical information related to the region’s meteorology, geology and oceanographic features. Chapter 4 gives an overview of the mobile ground-sites and the primary instruments used in the study. Chapter 5 gives a brief overview of the fundamentals of thermal IR radiative transfer and describes the structure of the radiative transfer models employed. Chapter 6 examines the detailed
dust microphysics and single scattering properties in the thermal IR window. Chapter 7 addresses satellite dust detection using MODIS thermal IR window radiances including methodology, case studies and validation. Chapter 8 presents the ground-based AERI dust detection and retrieval technique, which includes detailed sensitivity studies, methodology, selected case studies and validation. Chapter 9 addresses the SW and LW surface radiative forcing of dust aerosol and lastly, a final discussion is given in Chapter 10 which outlines the key elements of the study along with its implications for future research employing the current approach.
Figure 1.1 IPCC 2001 report showing the global and annual mean radiative forcing (direct and indirect) for aerosols, gases and clouds from 1750 through 2001. Level of scientific understanding is an indicator for the amount of certainty in a given forcing.
CHAPTER 2

2. BACKGROUND AND MOTIVATION

Before examining the detailed physics of the dust models employed in this study, it is important to investigate several key features of dust aerosol. First is the notion of how dust gets transferred to the atmosphere in the first place. A short summary underlining the chief physical mechanisms responsible for the horizontal and vertical transport of dust aerosol is given. Once airborne, the dust particles depending on their effective size may either fall out due to gravitational settling, or remain suspended in the atmosphere long enough to interact with the local radiative fields. Significant to this study are those particles whose lifetimes are long enough (i.e. having small settling velocities) to optically couple with the LW fields. In the context of these dust-radiative interactions, a brief overview of the experimental and theoretical development of the dust optical and microphysical properties is given which outlines key issues and identifies areas requiring additional research. Lastly, the radiative forcing of dust aerosol is briefly examined

2A. Horizontal and vertical transport of dust particles

This section identifies the basic physical processes responsible for the lifting of sand and dust particles into the atmosphere and their horizontal transport. For a more
detailed treatment of these mechanical processes, please refer to the website http://dust.ess.uci.edu/facts/aer/aer.html titled “Natural Aerosols in the Climate System” by Zender [2004] and also Pye [1987].

In the context of the current discussion, let it be sufficed to say that depending on the wind speed, soil characteristics and the size of the dust particles, the sediment can be moved via 3 transport modes: creep, saltation and suspension. Each process governs the way particles of different sizes respond to an applied wind field. Creep typically describes the movement of larger sized particles characterized by a rolling or sliding motion along the ground’s surface. Saltation is a collisional process of smaller particles, whereby the particle’s kinetic energy is continuously transferred to other nearby particles at the surface, in the direction of the applied wind field. The collisions displace some of the particles vertically, but the overall bulk motion is characterized by horizontal advection due to the stress exerted by the wind. This mode of particle transport is best described as a chain reaction forced by the turbulent winds. So long as the surface wind stress, or wind-friction velocity, exceeds a specified threshold for dust mobilization (typically \( v \geq 5 \text{ m/s} \)), the saltation process will govern the horizontal transport of the dust particles. Lastly, suspension, when coupled with turbulent eddies and updrafts causes the smaller sized particles to remain lifted, as the larger particles fall out due to gravitational settling. Figure 2.1 adopted from Pye [1987], shows the three modes of dust particle transport with the particle size ranges being characteristic of moderate sized dust storms.

Soils from around the world are highly variable in terms of their composition,
moisture content, and vegetative cover. In arid and semi-arid regions, like the UAE, moisture and vegetative cover are typically sparse to none, due to the lack of precipitation, leading to soils characterized by dry and loosely packed grains and mineral aggregates. Optically active particles in the visible and IR regions of the spectrum, mostly clay-sized ($D_p < 2.5\mu m$) and silt-sized ($2.5\mu m < D_p < 60\mu m$) particles are typically bound to these soil aggregates but are released on impact when other free particles, mostly sand-size ($D_p < 60\mu m$) particles, collide with the aggregate on the downwind side of their saltation trajectories. The process is referred to as sandblasting and is the primary mechanism by which smaller mineral particles are injected into the atmosphere. Figure 2.2 shows the global distribution of major dust source regions using frequency of occurrence distributions of the aerosol absorbing index (AAI) derived from the Total Ozone Mapping Spectrometer (TOMS) sensor aboard the Nimbus 7 satellite [Prospero et al. 2002]. One can easily identify the primary arid/semi-arid regions having the necessary soil conditions for contributing to the global dust burden.

Once airborne the suspension time of dust can be determined by direct application of Stoke’s law which permits evaluating the particle’s terminal settling velocity while it gravitationally settles in still air. For irregular aerosol particles, Hinds [1982] defines the terminal settling velocity as:

$$V_{ts} = \frac{\rho_p d_e^2 g}{18 \eta X}$$

(2.1)

where $\rho_p$ is the particle density, $d_e$ is the particle diameter, $g$ is the acceleration due to
gravity (980 cm s\(^{-1}\)), \(\eta\) is the viscosity of air (182 \(\mu\)P at STP) and \(\chi\) is the dynamic shape factor of the particle. Ranges of particle densities (\(\rho\)) for common minerals found in dust aerosol are given in table 2.1[from Reid et al. 2003].

Since dust particles exhibit a wide range of particle shapes which are typically non-spherical and highly irregular [Koren et al. 2001], a dynamic shape factor (\(\chi\)) is employed to account for changes in the particle’s drag coefficient. Davies [1979] for example found that \(\chi\) can range from 1.36-1.82 for quartz which is characterized by a tetrahedral shaped crystal lattice structure [Huggins, 1922]. By virtue of Eq. 2.1, the settling velocities of irregular dust particles will be smaller than those of their equivalent volume spheres hence allowing more time for the particles to interact with the radiative fields. Eq. 2.1 is valid for particle diameters greater than 1\(\mu\)m (most coarse-mode dust particles), however must be modified for sub-micron particles by multiplying through by the Cunningham slip correction factor (\(C_c\)). This accounts for the nonzero relative velocity of air at the particle’s surface which becomes important since the size of the particle approaches the mean free path of the gas.

Neglecting the vertical velocity component of the wind, one can estimate the settling velocity and hence residence time of dust particles in the atmosphere. Applying Eq. 2.1 for example to a giant quartz sphere with an effective radius 500 \(\mu\)m at a height of 1 km, yields a drop time of less than 2 minutes before striking the surface. If however the effective radius of the same quartz sphere is reduced to 5 \(\mu\)m, the particle takes well over a day to reach the surface given the same altitude and will therefore have more time to be optically active. So long as the vertical velocity component of the wind exceeds the
particle’s settling velocity, the particle will remain in suspension.

The underlying dynamics, which control the surface winds and stresses are quite variable and are highly dependent on the local topography, the surface characteristics and on thermal circulations. To properly address the effects of surface stresses on dust production in model simulations, accurate parameterizations must be made to account for these dependent variables. Liu et al. [2001], Alfaro et al. [2001] and Zender et al. [2003] among others demonstrate these effects. Overall, many modeling studies including those being conducted at the Naval Research Laboratory (NRL) in Monterey, California [Westphal et al. 1988] are being explored to improve the predictive skill of forecasting dust storms. Understanding the physical mechanisms of dust production is crucial for such studies and is ultimately tied in to the radiative and climatic effects, significant to this study.

2B. Optical and microphysical properties of dust particles

The early works of Volz [1973], Fisher [1976], Patterson et al. [1977 and 1981], Levin et al. [1980], and d’Almeida [1987], among others, have been instrumental in providing the dust research community with climatological datasets of the refractive indices and particle size distributions of dust aerosols that are so widely referenced in today’s scientific literature and used in radiative transfer and global climate models. Since then there has been on-going research, both in the laboratory and field [Volten et al. 2001, Reid J.S. et al. 2003 and Reid E.A. et al. 2003, for example] to improve our
knowledge of dust properties, including particle shape which only until recently with further advances in numerical light scattering techniques, has been modeled as a sphere. It is also important to point out the significant progress in mineral spectroscopy [Salisbury et al. 1991, Rousch et al. 1991, and Popova et al. 1972, to name a few] where datasets of mineral reflectance spectra commonly found in dust such as quartz and many of the common clays (i.e. kaolinite, montmorillinite and illite) have been made available to the scientific community. Collectively, these works have vastly improved our knowledge of dust properties, and have prompted more studies to be conducted. Many questions however still remain open, particularly those dealing with data interpretation [Reid et al. 2003], model assumptions of dust radiative properties [Sokolik et al. 1999] and technical difficulties associated with the differences found using various measurement techniques [Reid et al. 2003]. Also is the limited availability of data in the IR which is clearly needed to fully understand the LW impacts of the direct radiative forcing [Vogelmann et al. 2003]. These uncertainties make it unclear on how to best parameterize the dust properties for use in radiative transfer and global climate models and is certainly an area in need of further research. Examples of some of the experimental methods and theoretical development used for determining the dust optical and microphysical properties are given below. First considered are the chemistry and composition of dust particles.
2. B. 1. Chemistry and composition of dust particles

Elemental analysis of dust particles often shows extreme variation in their mineralogical composition. This is due both to soil differences in the global source regions and to the chemical transformation processes, which occur during dust transport. The chemical constituents of a dust particle are determined utilizing individual particle analysis (IPA). This is opposite to the bulk sample analysis technique which uses X-ray diffraction to find averaged compositional information about the sample (i.e. percentages of clays to non-clays for example). This is useful for identifying source regions but does not yield any information about the individual elements. IPA commonly uses the scanning electron microscope (SEM) equipped with an energy dispersive analysis (EDA) system for identifying the particle’s chief chemical species. Figure 2.3 shows two example SEM images of dust samples from backscattered electrons collected during (a) PRIDE, 2000 and (b) UAE² 2004. Note the varying complexity of sampled particle shapes, with particular attention to the angular and flat sided nature of dust.

The SEM consists of a filament from which electrons are emitted and are directed to the target material (i.e. filter sample) via an accelerating voltage ($\approx 30$ kV). The stronger the voltage, the higher is the penetration depth of the electrons into the sample. Sample imaging is accomplished by way of three types of interactions of the electron beam with the target material. First are secondary generated electrons that are produced by weakly attached surface electrons knocked loose by the primary beam. This interaction is more dependent on the surface area of the particle and hence is useful for
characterizing the particle’s topographic features. Second are the backscattered electrons, which are high-energy electrons from the primary beam that backscatter from the atomic nucleus of the target into a detector for further processing. The energy of these particles, or the intensity of the detected signal, depends on the mean atomic number (Z) of the target’s nucleus and hence is useful for identifying the different chemical species of the dust sample. Third the primary electron beam knocks loosely bound electrons from their respective orbits, leaving behind vacancies. Higher energy electrons from the outside orbits then fill these vacancies, giving up their energy in the form of X-rays which are characteristic of the material’s energy level transitions. This provides an efficient means for identifying the material’s chemical composition. An EDA system employing an X-ray spectrometer can provide accurate chemical spectra of individual particles in a dust sample. Figure 2.4, taken from Falkovich et al. [2001], shows example SEM-EDS chemical spectra of individual dust particles collected in a dust storm over Israel in 1998. Shown are pure quartz (a) and dolomite aggregated with quartz (b).

Large uncertainties are inherent in single particle analysis which include (1) system errors introduced by analyzing a three-dimensional particle on a two-dimensional surface, (2) matrix effects arising from interference of the incident electron beam with the emitted x-rays from the target medium (requires ZAF corrections), (3) imaging of the polycarbonate substrate due to small diameter particles (≤ 2µm) and/or loose aggregate particles (4) biases due to adjacent particles being counted as a single particle and (5) the electron beam favoring one side of a particle over the other.

The SEM/EDA technique has been widely used to help identify many chemical
species of dust samples throughout the world. For example, Falkovich et al. [2001] found that the dust particles collected over Israel during the 1998 dust storm typically occurred in aggregated form with varying mineralogical composition, consisting mostly of Ca, Mg, Al, and Si. Analysis also indicated that sulfur and iron adhered to the particle’s surface, suggesting that this must have occurred from processes interior to the source region. Dust traveling inside marine air layers and interacting with low-level water clouds also tends to increase the levels of NaCl and H₂SO₄ in the particle and hence can be used as tracers for atmospheric transport. Gao et al. [2001] characterized the composition of dust originating in China during April – May 1999 using three sampling stations located at different environmental sites in China. In Beijing for example, the dominant aerosol mixture contained sodium and iron, along with traces of silicate and sulfate species. As pointed out by Gao et al. [2001], the Beijing sample was most likely influenced by urban aerosol, such as coal combustion, which would have led to the elevated levels of observed iron in the dust. Reid E. A. et al [2003] characterized African dust transported to Puerto Rico by individual particle and size segregated bulk analysis where over 70% of the dust particle mass was attributed to aluminosilicate clay minerals such as illite, kaolinite and montmorillonite.

The chemical interactions of the mineral dust with the environment can have significant radiative consequences. For example, gaseous SO₂ emissions from industrial processes can mix with dust converting the SO₂ to sulfates in the course mode particles. Sulfates, in turn, have a high single scattering albedo (ω) in the visible wavelengths, which causes the incoming solar energy to be reflected back to space, thus leading to a
The optical properties of dust can also be modulated by the presence of acidic sulfate, which can alter the hygroscopicity, i.e. its affinity to water, of the aerosol thus changing its absorbing and scattering properties.

Heterogeneous reactions between dust particles and NO₂ and HNO₃ have also been investigated [Underwood et al. 2001] through combined laboratory and modeling studies. The observed and modeled surface reactions show significant effects in the production of tropospheric ozone, for example. The surfaces of dust particles provide important pathways for mass transport of key pollutants and is a major health concern requiring further investigation. Next considered are the optical properties and refractive indices of dust particles.

2. B. 2. Optical properties and refractive indices of dust particles

The techniques commonly employed to measure the refractive indices of dust aerosol (absorption + scattering) rely on collecting transported (i.e. remote) or localized bulk dust samples, which often possess uncertain or unknown mineralogical composition for limited geographical locations [Sokolik et al. 1999]. For example, Volz [1973] and Fouquart et al. [1987] performed laboratory measurements of the absorption and refractive index of collected Saharan dust samples, at Barbados and Niger respectively, using a transmission technique in which dust was suspended in a potassium bromide (KBr) pellet. Others have relied on thin-film transmission techniques and reflectance methods coupled with Kramers-Kronig theory. Factors, such as transport and mobilization processes, and even particle size distribution, which can significantly alter
the chemical composition, further complicate the ability to accurately quantify the dust optical properties. Overall, the available data is rather poor in quality and does not adequately represent all the major dust source regions. Furthermore, the atmospheric conditions under which the samples were collected typically represent single points in space and time. Hence, accurately modeling the spatial and temporal variability of dust particles during transport cannot be ascertained on the basis of these optical constants alone. Moreover, the equipment uncertainties and the technical difficulties associated with the measurement process make data interpretation extremely difficult [Reid et al. 2003].

Recent theoretical studies [Sokolik et al. 1999 and Kalashnikova et al. 2004] have investigated an alternative approach for modeling the dust radiative properties which use the measured refractive indices of individual minerals rather than those from a bulk mixture [e.g. Volz, 1973]. The thermal IR in particular displays a wide range of spectral features for common dust minerals [Sokolik et al. 1999] yielding a very complex spectrum of single scattering properties. Sokolik et al. [1999] point out that because of the nonlinear dependence of dust optical properties on the refractive index, it is unclear whether or not refractive indices measured as an average for a mixture are appropriate to use in modeling studies. Rather than assuming an averaged set of optical properties from a mixture with variable mineralogical composition, Sokolik et al. [1999] derived the optical properties of bulk samples using the individual optical properties of the component species. The dust models used in this study follow this approach although for comparison the refractive indices from the Volz Saharan dust model are also evaluated.
The mineral datasets employed in this study, involve the works of many previous investigators. Next considered are particle size measurements of dust.

2. B.3 Dust particle size

Previous studies have shown [D’Almeida et al. 1991 and Patterson et al. 1977 among others] that the sizes of collected samples of mineral dust tend to follow a lognormal distribution, which is characterized by an effective radius $r_g$, and effective variance $\sigma_g$. This distribution is often expressed in terms of several size modes which takes the form:

$$ N(r) = \sum_j N_j ((2\pi)^{1/2} \ln(\sigma_j) r)^{-1} \exp[-\ln(r / r_{0j})^2 / 2 \ln(\sigma_j)^2] $$

where the summation is given over the $j$ size modes, $N( r )$ is the particle number concentration for a given mineral, and $N_j$ is the particle number concentration of the $j^{th}$ mode. Each size mode is characterized by the effective radius $r_g$, and variance, $\sigma_g$.

The size fractionation of dust particles collected during dust events often exhibit bi-modal or even tri-modal characteristics. For example in a large dust storm characterized by strong surface winds, particularly near the source region, course-mode particles with sizes ranging from 10$\mu$m to as high as 100$\mu$m is common, however sub-micron sized particles (fine mode) may also be found. The distribution of particle size is highly dependent on the local meteorology and the resulting mechanical processes that
lift the dust particles into the atmosphere (section 2A). As one moves away from the source region, the dust plume will disperse, becoming optically thinner with increased distance. The larger particles will fall out due to gravity, leaving behind a dust cloud composed of smaller and much lighter particles. It is suspected in the case of the UAE that the suspended dust is primarily due to the mechanical stirring of localized desert regions; hence course mode particles should dominate. Remnants of transported dust, i.e. mostly fine and accumulation mode particles, may have originated from neighboring desert regions across the Arabian Gulf or as far away as the Saharan Desert. Figure 2.5 illustrates the bimodality of retrieved volume size distributions from the UAE using NASA GSFC AERONET [Holben et al. 1998]. The distributions are normalized to the total volume integrated over the particle size range.

Determining particle size experimentally can often yield different sets of size parameters even when operating under similar conditions. These measurement differences can be attributed to differences in the methodologies employed and also due to the uncertainties inherent in each technique [Reid et al. 2003]. A short description of some of the sizing methodologies follows.

Particle sizing instruments can generally be classified according to its relation to some physical parameter of the particle. This includes (1) Geometric sizing, which employs electron and light microscopy techniques, and is based on the particle’s physical size, however can be quite subjective when it comes to sizing irregularly shaped particles, such as dust (2) Aerodynamic sizing, which uses cascade impactors and aerodynamic particle sizers (APS) and relates the particle’s mass to its aerodynamic drag. Particles
with unusual shapes, like dust, can cause large biases and hence distort the measured size distributions. These biases depend on such factors as flow rate which control the cutoff size ranges and also on plate impaction physics which can lead to particle ‘bounce’ resulting in interstage losses [Hinds, 1982]. According to Reid et al. [2007], the APS is the most viable option to date for studying the dynamics of dust particle size distributions and is one of the reasons for employing the APS in this study. (3) Optical particle counters (OPC) such as the Forward Scattering Spectrometer Probe (FSSP) or the Passive Aerosol Spectrometer Probe (PCASP), sizes particles by measuring the amount of coherent light scattered by the particle as it traverses the beam onto a detector (4) Optical inversion techniques from remotely sensed data such as that from AERONET, which uses spectral extinction and angular scattering measurements over the atmospheric column to determine a best fit to the modeled size distributions. Reid et al. [2003] show comparisons of the mass and volume median diameters obtained from previous studies using these methods, and report similarities between the aerodynamic and optical inversion methods, but note differences with the optical counters, yielding median diameters approximately 2-3 times higher than the former methods. Reid et al. [2003] caution the interpretation and application of the measured size data, as this will have significant consequences on radiation budget studies. Lastly we consider measuring dust particle morphology.

2. B.4 Dust particle shape

Contrary to the non-spherical and irregular nature of dust particles [Okada et al.,
researchers usually assume dust is composed of spherical particles having homogeneous compositions which permit the use of a Lorenz-Mie scattering code for calculating the dust single scattering properties. Further advances in IPA employing SEM have made it possible to investigate the sensitivity of dust optical properties to non-spherical dust particles. Currently most shape analysis techniques assume an elliptical dust model where shape parameters such as the particle’s aspect ratio (ratio of maximum projection to width (a/b)) and circularity (C), defined by the particle’s deviation from a perfect circle or sphere (C =1), are uniquely determined from the 2-D SEM image (refer to fig. 2.3). Since 3-D particle shapes are determined from a 2-D image, information regarding the particle’s height or thickness is lost and hence does not adequately describe the true shape of the particle. In addition, analysis of the SEM image is subject to errors due to particle size, where smaller particles require greater magnification of the image in order to count the pixels; hence the shapes of smaller particles may be biased low. Irregularly shaped dust particles typically have circularities in the range of 1–2. Recent studies show that as dust particle size increases, as in the case of heavy dust outbreaks, so too does its circularity, indicating dust is rarely spherical [Koren et al 2001]. Typical aspect ratios for dust particles are usually in the range of 1.4-2.0, as observed from SEM images collected during studies in China [Okada et al. 2001] and PRIDE [Reid J. et al. 2003] from which particle shape distributions are based in this study.

2. B.5 Dust single scattering properties

After the dust optical and microphysical parameters have been prescribed, the
corresponding single scattering properties can be determined. The single scattering properties which include the extinction coefficient ($\beta_{\text{ext}}$), single scattering albedo ($\bar{\omega}$), and the asymmetry parameter ($g$) uniquely define the dust model(s) to be used in the radiative transfer simulations from which radiances/fluxes are calculated for remote sensing and climate applications respectively. Details of the methods and light scattering codes used to calculate the single scattering properties of the various dust models employed in this study are discussed in chapters (5-6).

Given the many uncertainties inherent in the key dust parameters, as was previously illustrated, one can imagine the corresponding uncertainties in the formulated dust models and the effect this will have on how remote sensing and climate data is ultimately interpreted. Uncertainties in mineralogy for example, can lead to differences in the sign of the dust radiative forcing which can have significant climatic implications. Sokolik et al. [1999] showed that the forcing can either be negative or positive leading to a cooling or warming atmosphere depending on how the minerals are mixed. Lastly dust surface radiative forcing is examined.

2. B.6 Dust LW surface radiative forcing

Historically, aerosol science has focused more on the SW effects of dust aerosol mainly due to the larger availability of data. In the solar wavelengths, dust blocks the incoming light, scattering it back to space. In response to the dust load, the SW flux is reduced at the surface and is enhanced at the top of the atmosphere (TOA) which leads to
a negative and positive SW forcing respectively. Here aerosol forcing is defined as the change in radiative flux ($\Delta F$) between a dust-laden and dust-free atmosphere. *Hansell et al.* [2003] assessed the surface SW forcing of dust aerosol from several field campaigns using only surface observations in a differential technique. The reported SW surface forcing, for example during PRIDE 2000, for values of air-mass of 1.5 and 2.0 were approximately -12.0 and -20Wm$^{-2}$ respectively given a mean range of aerosol optical depth (AOT) between 0.21-0.24. If the dust load corresponded to one unit of AOT (i.e. $\tau=1$), referred to as the forcing efficiency, the resulting decrease in SW flux would have amounted to -121.8 and -92.1Wm$^{-2}$ for the same air mass values which can significantly cool the surface.

More recently, however, the impacts of aerosols in the LW regions of the spectrum have garnered a significant amount of attention within the scientific community. Based on observations, previous investigators *[Vogelmann et al. 2003* for example] have shown that the LW contributions are similar in magnitude to the surface warming due to green-house gas emissions. Various modeling studies indicate that the surface forcing in mild dust conditions can vary from approximately 3-7 Wm$^{-2}$ and can exceed 15 Wm$^{-2}$ for heavy dust episodes. For normal background dust conditions, the positive LW forcing, albeit small compared with the SW, can still have an effect at the surface. Since most dust particle sizes around the source region (i.e. coarse and giant sized modes) are comparable to that of the incident wavelengths, scattering and absorption in the thermal IR will be non-negligible *[Dufresne et al 2002*] and will ultimately affect the local radiative fields in a positive or negative way depending on how
the dust particle’s component minerals are mixed, their size distribution and the shapes of
the particles. *A priori* knowledge of these key dust properties is absolutely critical for
properly assessing the radiative effects of dust aerosol and thus reducing the uncertainties
of the direct forcing.
Figure 2.1 Three modes of dust particle transport (creep, saltation and suspension) due to wind flow (left to right) and turbulent eddies/updrafts responsible for horizontal and vertical advection respectively. From Pye [1987].
Figure 2.2 global distributions of TOMS dust sources. The map is a composite of selected monthly means of AAI frequency of occurrence statistics from 1980-1992. Plotted is the total number of days in which the AAI exceeded a prescribed threshold of 0.7 or 1.0 for regions outside and inside the dust belt (i.e. from Northern West Africa through the Middle East to China) respectively. Plot adapted from Prospero et al. [2002].
Table 2.1 Common dust component densities

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Density (ρ) g cm(^{-3})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Amorphous silicon oxide</td>
<td>2.1-2.3</td>
</tr>
<tr>
<td>Illite/muscovite</td>
<td>2.7-3.1</td>
</tr>
<tr>
<td>Montmorillonite</td>
<td>2.2-2.7</td>
</tr>
<tr>
<td>Quartz</td>
<td>2.65</td>
</tr>
</tbody>
</table>

From Reid et al. [2003]
Figure 2.3 (a) Taken from Reid E.A. et al. [2003] shows a micrograph image of collected dust particles from the Saharan Air Layer during PRIDE. (b) Same but from the UAE² courtesy of J.S. Reid.
Figure 2.4 X-ray peak spectra using a SEM-EDS system for individual dust particles corresponding to a. pure quartz and b. dolomite aggregated with quartz collected in a dust storm over Israel in 1998. Data adapted from Falkovich et al. [2001]
Figure 2.5 Retrieved volume size distributions from AERONET almucantar measurements on September 1, 2004 during UAE$^2$ illustrating two volume size modes one for fine sub-micron particles ~0.2µm and another for coarse-mode particles ~5-6µm. Courtesy of AERONET Project.
3. UAE\(^2\) FIELD CAMPAIGN OVERVIEW

3. A. Background

The UAE\(^2\) field campaign was conducted during the summer of 2004, from August 06 through October 06 in the United Arab Emirates (UAE). A map of the area and surrounding regions is given in Figure 3.1(a) with the white/dark gray squares denoting the positions of the two major ground sites: NRL MAARO and GSFC NASA SMART respectively. Figure 3.1(b) shows an ASTER satellite image of Al-Ain, (N 24 14’ 56”; E 55 36’ 43”), UAE on October 5, 2004. The arrow points to the airport where SMART was deployed. The campaign was organized by several activities, including the Naval Research Laboratory (NRL), the University of Witwatersrand in Johannesburg, South Africa, GSFC NASA, Jet Propulsion Laboratory, and the Department of Water Resources, Office of the President (DWRS) in the UAE. Other research groups included the TNO Physics and Electronics Laboratory from The Hague, Netherlands, Warsaw University, Warsaw, Poland, Universite de Shebrooke, Sherbrooke, Quebec, Canada, NCAR in Boulder, Colorado, and about 4 U.S. Universities. The following is referenced from the project’s science plan [Reid et al. 2004].

The project was an outgrowth from previous cloud-seeding experiments [Salazar et al. 2001], which began in 2001 in collaboration between the DWRS, NCAR and the South African research group. Logistically, the dust study was a natural transition, since
the supporting aircraft and required resources were already on-site. As the program
developed, the mission objectives and the number of participants expanded. Ultimately
the field study included the two ground-based mobile observatories (MAARCO and
SMART), a distributed network of 15 AERONET sun-photometers, both polarized and
standard models, two research aircraft (South African 690A Aerocommander and NCAR
OUP), multiple satellite coverage from MODIS, SeaWiFs, MISR, and AVHRR and other
scientific support equipment including the University of Hawaii’s airborne scanning lidar
and TNO’s 7-wavelength transmissometer.

The UAE² field campaign provided the unique opportunity to conduct a series of
experiments from various platforms located along the coastal regions and the inland
deserts with the local support of the DWRS. Details of the test platforms and the primary
instruments employed during the study are discussed in Chapter 4. During the 2-month
period, over two dozen research organizations convened in the United Arab Emirates
(UAE) to participate in field experiments designed to include the following objectives:
(1) to provide ground truth for satellite and model products in the region’s highly
heterogeneous environment; (2) to evaluate the properties of dust particles from the
numerous sources that converge in the UAE region; (3) to determine the impact of
aerosol particles on the local radiation budget; and (4) to determine the effect of the
radiative perturbations on regional mesoscale flow patterns. Of particular interest to this
work were objectives (2) and (3). The following is intended to serve as a brief overview
to some of the UAE’s key atmospheric, oceanographic and geographic features, which
explains in part the motivation behind the field study.
Anecdotal reports and supporting satellite data show that the Arabian Gulf and its surrounding areas is a major location for the emission and deposition of many types of aerosol. Primary aerosols include dust from the local deserts and remote arid regions, advected smoke from the Indian Subcontinent, and local pollution from the petroleum industry. Dust, interestingly enough, is most often transported into the UAE from sources to the Northeast in Iraq and Saudi Arabia as well as from dry lakebeds and marshes in Iran and Afghanistan. It could also be brought in from regions as far away as Africa. The combined optical depths from each aerosol result in a reduced visibility of less than 10km, which typically persists throughout the spring and summer months, coinciding with the Summer Monsoon and the onset of the Southwesterly Shamals. Probably the earliest study of this region’s aerosol properties was conducted by Ackerman and Cox [1982] during the Summer Monsoon Experiment (SMONEX) where observations showed a highly variable dust loading for the region, which decreased with height, extending to a maximum altitude of about 6km. The atmospheric heating rates in the solar spectrum were found to have doubled, with a much smaller LW forcing.

3. B. Regional meteorology and oceanographic overview

The aerosol load is further enhanced with the region’s complex meteorology. Seasonally, the Arabian Gulf undergoes a two-monsoon cycle: the winter Northeasterly Monsoon and the summertime Southwesterly Monsoon which is characterized by the intense Shamal winds. Northeasterly winds develop off the coast of the Indian Subcontinent due to relatively constant sea surface temperatures (SST) and a cooling of
the Indian and Middle Eastern landmasses. The wintertime Monsoon brings most of the precipitation for the lower elevation areas in the UAE, although annual rainfall amounts vary considerably. The last several years in the UAE have been very dry, making dust storms more frequent. During the spring months, a reversal in the wind patterns switches from a Northeasterly flow to a Southwesterly flow, making the transition into the Summer Monsoon cycle. Low upper level moisture combined with upper level subsidence leads to very dry conditions in the UAE. At times, isolated convective cells can pass through the area with an increase in precipitation. This was observed on several occasions during the field campaign, coupled with increased dust activity which can trigger large dust fronts or haboobs [Miller et al. 2007]. Variable sea surface temperatures (SST), the region’s sharp topographical features, and the strong land-sea gradients and mesoscale circulations contribute to the area’s complex environment.

During the first half of the campaign in August, the temperatures were quite elevated, reaching at times as high as 46°C with humidity levels ranging from 20% in the inland deserts to more than 70% along the coast. During the second half of the campaign in September, the temperatures fell about 10°C with humidity levels remaining fairly constant. The Gulf region was typically cloud free with cloud cover on the order of about 20-30% of which less than 10% was cirrus.

The Arabian Gulf exhibits a very complex topography with many banks, shoals and small islands. Sea surface temperatures are very high which can reach over 35°C during the summer months. The warm waters and positive dust fertilization contribute to the region’s high biological activity and increased populations of phytoplankton.
Chlorophyll concentrations are also considerably high, with elevated concentrations of microorganisms.

3. C. Regional geology:

About 80% of the UAE is desert with elevations less than 200m. The remaining 20% is composed of mountains, with elevations of 2000m in the northern region near the eastern border of Oman. The coastal regions and the UAE’s drainage basins or sabkha (shallow salt water table), contain minerals such as gypsum, anhydrite, calcite and dolomite, which can precipitate out, trapping wind-blown sand and silt. Evaporites, including salts and nitrate minerals typically form on the surface where they can be eroded away. The non-Sabkha regions mostly contain aeolian materials such as gravel, sand, silt and clays. On occasion local dust storms, including dust devils can entrain dust in the Sabkha’s and in the desert plains, however as previously noted, most of the dust is transported from surrounding regions. This makes modeling dust exceedingly difficult, due in part to the variable composition of the parent soils, not to mention the particles’ complex shapes and size distributions.
Figure 3.1 a. Map of UAE and surrounding regions. White and dark gray squares denote positions of the NRL MAARCO and GSFC NASA SMART sites respectively. b. ASTER satellite image of Al-Ain, UAE. The arrow points to the airport where SMART was deployed. The darkened tan region at the bottom right is the Hajar Mountain range where an AERONET sun-photometer was stationed.
CHAPTER 4

4. INSTRUMENTATION

The 3 primary measurement platforms used during the UAE2 field campaign were the: (1) The University of Witwatersrand South African Aerocommander 690A research aircraft, (2) The NRL Mobile Atmospheric Aerosol and Radiation Characterization Observatory (MAARCO) and (3) The GSFC NASA Surface Sensing Measurements for Atmospheric Radiation Transfer (SMART). Note that since the UAE2, GSFC NASA has added a second mobile research facility called COMMIT (Chemical Optical Microphysical Measurements In-Situ Troposphere) dedicated to in-situ measurements of trace gases and aerosols. Please refer to http://smart-commit.gsfc.nasa.gov/ and http://www.nrlmry.navy.mil/aerosol/Docs/maarco.html for more information regarding each of the mobile facilities. Figure 4.1 shows (a) MAARCO at NRL in Monterey, California and (b) SMART deployed during the NAMMA.

During UAE2, the MAARCO observatory was located along the coast (N 24, 41’ 59”; E 54 39’ 32”) near the city of Abu-Dhabi, while SMART was deployed in the desert plains about 160km inland at Al-Ain Airport (N 24 14’ 56”; E55 36’ 43”). These “super sites” were further augmented by an extensive AERONET deployment of 15 sun-photometers strategically located throughout the region. Since the data collected at the MAARCO and SMART sites are primary to this dissertation, this section only addresses the relevant instruments used at these sites. At SMART the following instruments were
employed for this study: (1) Atmospheric Emitted Radiance Interferometer (AERI-Bomen MR100), (2) Eppley precision infrared radiometer (model PIR), (3) AERONET sun-photometer (standard model CE-318A), and (4) MPLNET micro-pulse lidar. At MAARCO the following instruments were employed: (1) the aerodynamical particle sizer (APS 3321) which was primary to this study, (2) daily radiosonde launches and (3) AERONET sun-photometer (standard model CE-318A). Additional radiosonde data from Abu-Dhabi Airport was also used throughout the study. An overview of the primary instruments follows.

4A. SMART

The SMART container is a self-sufficient ground-based mobile laboratory whose sole purpose is to conduct active and passive remote sensing of atmospheric radiation. SMART passive instrumentation includes, a suite of rooftop mounted broadband radiometers in the visible and IR, including a 7-channel multiple filter rotating shadow-band radiometer, a solar tracker equipped with a pyroheliometer and shaded pyranometers for direct normal and diffuse irradiance measurements and an analytical spectral devices (ASD) spectroradiometer. In addition, the SMART system includes: (1) an ultraviolet radiometer for ozone determination, (2) a Fourier transform infrared spectroradiometer (FTIR) or AERI for measuring atmospheric emissions, (3) a scanning microwave radiometer (SMIR) for measuring column precipitable water (PWV), (4) standard meteorological sensors for monitoring both surface (mounted 1m AGL) and skin temperature (thermistor was made to contact surface repetitiously at the SMIR
frequency), (5) a total sky imager (TSI) for recording daytime cloud coverage, and (6) two AERONET sun-photometers which included both the standard and polarized instruments. Lastly, SMART is outfitted with a MPLNET micro-pulse lidar. Below are detailed descriptions of the primary instruments used in this study.

4A1. AERI

The AERI system was initially developed to support the ARM Program [Revercomb et al. 1993] and was an outgrowth of the High-Resolution Interferometer Sounder (HIRS) Program at the University of Wisconsin [Smith et al. 1993]. The Bomem model MR-100 AERI measures spectrally resolved downwelling radiation from 3.3-19µm. The AERI hardware layout is depicted in figure 4.2. It contains a standard 2-port Michelson-type interferometer, 2 blackbody external calibration sources, one set to drift at ambient temperature (~25°C) and the other heated to 65°C and a mirror optics assembly programmed to alternate between the zenith scene view and each blackbody reference once about every 15 minutes. It is designed to yield radiometric accuracies within 1% of the ambient radiance [Knuteson et al., 2004]. Approximate scene and blackbody dwell times are shown in fig. 4.3. The successive peaks are the blackbody scans, which last for approximately 5 minutes each. Bracketed by each reference scan are the scene views which have a temporal frequency of about every 30 seconds. Successive mirror scans are averaged together via a co-adding process (to ensure an adequate signal to noise ratio) to produce an interferogram.

AERI is a 2-channel instrument which uses dual photoconductive detectors
arranged in a sandwiched configuration. The indium antimonide (InSb) detector (channel 2) is used for sensing radiation between 3.3-5.5 µm and the HgCdTe (MCT) detector (channel 1) is used for longer wavelength sensitivity between 5.5 and 19µm. A stirling cooler is used to reduce the thermal noise from the instrument. Turner et al. [2004] discuss the non-linearity effects associated with the mercury cadmium telluride (MCT) detector which can become significant (~ 0.5% of ambient radiance) under clear-sky conditions and also under optically thin scattering media. Further discussions about the detector non-linearity and its estimated effects on AERI dust detection and retrieval is given in chapter 8.

Data acquisition was controlled by an IBM-PC compatible computer running the Bomem control software. The control software apodizes the interferogram and performs a fourier transform to produce an uncalibrated spectrum with a spectral resolution of 1 cm⁻¹. To convert the raw data into scene radiances, the Bomem radiometric correction functions were employed which compute the real/imaginary gain and offset coefficients for both forward and reverse mirror directions during the measurement. The complex gain and offset coefficients are calculated from the hot/ambient blackbody reference spectra and so they are a direct representation of the instrument response function (which includes phase information). AERI calibration was performed just before and after each scene view to account for possible drift in the instrument's temperature during the measurement sequence. All data calibration and post processing was performed at the Atmospheric and Oceanic Sciences Department, UCLA.

During the UAE², AERI logged nearly 130 hours of daytime spectral data over
21 days interspersed within the August 13 – September 30, 2004 timeframe. Periodic
downtimes included times of heavy airborne dust during which the instrument’s fore
optics was covered to prevent damage and also for equipment maintenance and data
storage handling. Refer to http://www.abb.com/ for further details on the Bomen AERI
instrument.

4A2. PIR

The Eppley PIR, a hemispherical sensing instrument, is designed to measure
either the downwelling or upwelling broadband LW radiation in the 3.5-50 µm spectral
range. Further instrument details can be found at http://www.eppleylab.com/. In the case
of UAE², the PIR was configured to measure the downwelling radiation due to cloud,
aerosol, and gas emissions. The PIR employs a thermopile detector for converting the
incoming LW radiation into an analog voltage response and two thermistors mounted
near the detector and dome for monitoring and correcting for temperature gradients inside
the instrument. Radiometric data is corrected using procedures described by Ji et al.
[2000] to account for the dome effect which can lead to flux uncertainties of 10 Wm⁻² or
more. All flux data was recorded 24-hours each day during the entire operating period
and was logged by a Campbell Scientific CR-7 data logger. The UAE² PIR dataset is
used to compare the computed model fluxes for the radiative forcing calculations
discussed in Chapter 9.
4A3. Sun-photometer

The AERONET program, an organized global network of ground-based remote sensing sun-photometers (Cimel Electronique 318A spectral radiometer) used for aerosol research was established by GSFC NASA and LOA-PHOTONS (CNRS) in 1993. Collaborators from around the world, including universities, government institutions, individual scientists and partners, participate in this expanding network (over 200 sun-photometers currently deployed on all continents) to maintain a continuous climatology of aerosol optical, microphysical and radiative properties to help better characterize aerosols in radiative transfer and global climate models and for validating satellite retrieval algorithms. During UAE², a dense network of about 15 sun-photometers was deployed throughout the UAE region which included the standard, polarized and extended models. Each instrument has an approximate 1.2° full angle field of view and two detectors for measuring direct sun, aureole (3° from sun) and all-sky radiances. Ion-assisted deposition interference filters are mounted on a filter wheel that is rotated by a stepping motor to measure the radiance over a narrow spectral bandwidth of approximately 10 nm. The corresponding channels for each sun-photometer model are given in table 4.1. The SMART site operated both the extended and polarized sun-photometers. Crucial to this work was the level 1.5 cloud-screened and level 2.0 cloud-screened/quality assured datasets from the extended model.

The measurement protocol consists of two basic measurements, either direct sun or all-sky. The direct sun measurements are made over each channel (table 4.1) in a
series of triplet observations which take approximately 2 minutes to complete. This measurement sequence is performed during morning and afternoon Langley calibrations and during every 15 minute interval. Triplet observations are collected to facilitate cloud-screening since the time variation of clouds is typically larger than that of aerosol. From the direct measurements, aerosol optical depth can be retrieved via the Beer-Lambert law.

Sky measurements are performed at 440,670,870 and 1020nm over two basic sky observation sequences, almucantar and principal plane. This is done so as to acquire sufficient aureole and sky radiance observations through a large range of scattering angles to retrieve particle size distribution and phase function.

Data transmission from the remote sites is automated via a satellite link to GSFC NASA using a Stevens/Vitel Data Collection Platform (DCP) or a Sutron SATLINK2 data logger/transmitter with a frequency of every 30-60 minutes. For further information, please refer to http://aeronet.gsfc.nasa.gov/ and Holben et al. [1998].

4A4. Micro-pulse lidar

SMART operates a NASA Micro-pulse Lidar Network instrument (MPL) [Welton et al. 2001], which is a key instrument for profiling the vertical structure of aerosol and clouds. MPLNET is comprised of ground-based lidar systems that are collocated with the AERONET sun-photometer and are strategically located at key sites around the world. The MPLNET provides critical information for various validation programs and provides useful information related to aerosol transport studies.
The lidar is completely autonomous, transmitting low power laser pulses from a Nd:YLF laser of 10\(\mu\)Joules at a frequency of 2.5Khz. The vertical and time resolutions are 75m and 60s respectively, providing both cloud and aerosol profiles in normalized relative backscatter intensity. The MPL operates at a wavelength of 0.523\(\mu\)m.

During normal operations, the MPL was shutdown around solar noon to prevent damage to the instrument’s optics. This study makes use of the level 1.0 normalized relative backscatter data described in Campbell et al. [2002] to facilitate validating the AERI dust detection scheme. For further information, please refer to http://mplnet.gsfc.nasa.gov/.

4B. MAARCO

The MAARCO container is a self-sufficient mobile laboratory for conducting aerosol chemistry and radiation measurements both in the visible and IR wavelengths. Key instruments include an array of broadband radiometers mounted to the MAARCO rooftop and a solar tracker for measuring diffuse and direct irradiance components. Daytime cloud coverage is also continuously observed with a total sky imager (TSI). An aerodynamic particle sizer (APS 3321), a 3-wavelength nephelometer and modified ground-based aerosol optical counters are integrated into MAARCO’s inventory. Aerosol chemistry is performed through a variety of filters and gas monitors, including a total suspended particulates filter (TSP), a tapered element oscillating mass balance.
(TEOM) for measuring particle mass, a MOUDI sampler for collecting size fractionated particles for gravimetric and chemical analysis, and SO$_2$ and O$_3$ monitors. MAARCO also maintained 1 standard sun-photometer for continual aerosol optical depth coverage. Lastly MAARCO has a radiosonde system and weather station in place for continual monitoring of surface-based atmospheric parameters and vertical profiling. Descriptions of the primary instruments used in this study are given below.

4B1. APS

The APS is a time of flight spectrometer for measuring both particle aerodynamic diameter and light scattering intensity. The particle sizes are segregated into 52 size bins ranging from 0.5-20 µm. Size segregation is performed based on the particle’s mass to drag characteristics in the accelerated air-flow of the instrument. Data collection during UAE$^2$ occurred continuously from August 11 – September 30, 2004, with a sampling frequency of every one hour, recording a total of 1200 size spectra over the entire study. J.S. Reid et al. [2007] discusses the technical details of the instrument and data collection protocol. Its dataset has been used by other investigators during UAE$^2$ including Ross et al. [2007] to infer fine-mode particle size distributions of localized air pollution sources and E.A. Reid et al [2007] for separating particle size distributions into common mineralogy groups. For further information, please refer to http://www.tsi.com and J.S. Reid et al. [2007].
Figure 4.1 a. MAARCO with remote sensing instruments located on top of container. b. SMART with remote sensing instruments on top of container and on the surface. Not shown is the equipment interior to the containers.
Figure 4.2 SMART AERI. Input port (arrow pointing to top aperture) directs the atmospheric emissions into the instrument’s interferometer. The blackbodies (black cylinders) shown in foreground are the calibration sources for converting the raw scene data into radiance spectra.
Figure 4.3 Approximate dwell times (vertical axis) during AERI measurement cycle. Peaks represent the blackbody reference scans used during calibration. In between each blackbody scan are the AERI scene views. Data corresponds to September 22\textsuperscript{nd} 2004.
<table>
<thead>
<tr>
<th>Model</th>
<th>Channels (nm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Standard</td>
<td>1020, 936, 870, 675, 500, 440, 380, 340</td>
</tr>
<tr>
<td>Extended</td>
<td>1640</td>
</tr>
<tr>
<td>Polarized</td>
<td>1020, 870P1*, 675, 440, 870P2*, 870, 936, 870P3*</td>
</tr>
</tbody>
</table>

*P1-3: polarized filters
CHAPTER 5

5. THE THERMAL IR RADIATIVE TRANSFER EQUATION AND RADIATIVE TRANSFER MODELS

In the context of a ground-based zenith-viewing AERI system, the following is an overview of the basic theory governing the generalized thermal IR radiation transfer for a finite plane parallel atmosphere in thermodynamic equilibrium. Here the formal solution for the downward surface intensities and fluxes are derived for application to dust remote sensing and radiative forcing respectively. An overview of the radiative transfer models (RTM) employed in this study is then presented along with a description of the model input parameters.

5. A. THE THERMAL IR RADIATIVE TRANSFER EQUATION

Over climatological time scales (periods of a year or more), the temperature of the earth-atmosphere system is relatively constant due to a global balance of the SW and LW radiative fluxes. Perturbations to the energy balance by cataclysmic events such as volcanic eruptions or massive dust storms, can force the system out of thermal equilibrium leading to a net heating or cooling of the atmosphere. Moreover, climate feedbacks in the system can further amplify or dampen these perturbations leading to
even more drastic climate responses. A comprehensive understanding of both SW and LW effects and their competing interactions in the earth-atmosphere system are therefore critical.

Significant to this study are the surface radiative effects of dust aerosol in the LW regions of the electromagnetic spectrum, in particular the thermal IR window (8-12 µm) where the atmosphere is relatively transparent and hence radiometrically cold in terms of its $BT$. This facilitates the detection and retrieval of dust aerosol since dust strongly absorbs across the window region and will therefore appear to be much warmer against the cold background of space. For the current discussion, dust is considered a non-scattering medium in the LW domain since absorption by dust minerals dominates much of the LW regions of the spectrum due to the inherently strong molecular vibrational processes [Salisbury et al. 1991] of the component minerals. From the surface, the downward transfer of IR energy is therefore expressed in terms of simultaneous emission and absorption processes which occur in the dust layer(s). Although scattered LW energy is neglected in the current discussion, its effects have been shown to be non-negligible [Dufresne et al. 2002] and are therefore considered in the radiative transfer models used in this study. To evaluate the downward radiative transfer of IR energy through a dust layer(s), a simple conceptual model is given below.

Consider a dust cloud near the surface as shown in figure 5.1, which is divided into ‘n’ homogeneous layers with respect to the dust single scattering albedo ($\omega$), asymmetry parameter (g) and optical depth ($\tau$). For this model, the dust properties are prescribed to be uniform across each layer.
On the basis of conservation of energy, the transfer of monochromatic IR radiation through a non-scattering dust layer(s) is given by:

\[ A_\nu + T_\nu = 1 \]

where \( T_\nu \) and \( A_\nu \) are the transmitted and absorbed components of the energy respectively. Note, the effects of surface scattering and multiple scattering inside the dust layer(s) would be included by adding a reflectivity term \( R_\nu \) to Eq. 5.1. Surface reflections (\( I_{\text{ref}} \)) and thermal emissions (\( I_{\text{clear}}/I_{\text{dust}} \)) inside the dust layers are illustrated in fig. 5.1.

Through its interactions with dust, the IR energy which is represented by its intensity or radiance (I) will be both attenuated and strengthened simultaneously through absorption and emissions processes of the dust medium respectively. Written in terms of dust optical depth \( \tau_n \) to avoid the height dependence of the absorption coefficient and gas density, the basic thermal IR radiative transfer equation takes the form:

\[
\mu \frac{dI(\tau, \mu)}{d\tau} = I(\tau, \mu) - B(\tau)
\]

(5.2)

where the terms on the right-hand side describe the reduction and increase in radiant intensity respectively. The variable \( \mu \) is the cosine of the zenith angle (\( \theta \)) which for the downward intensity is defined from \( \pi/2 \leq \theta \leq \pi \) (fig 5.1) and B is the source function given by the Planck intensity following Kirchoff’s law under the assumption of local
thermodynamic equilibrium. Note $\mu$ can be replaced by $-\mu$ for the downward case. Subscripts denoting the wavenumber ($\nu$) dependency of intensity have been omitted for clarity but is implied in Eq. (5.2) and in the following equations.

Because AERI is a zenith-viewing instrument of the down-welling thermal radiation, the formal solution for the downward intensity ($I_{\downarrow}$) is given by:

$$I_{\downarrow}(\tau, -\mu) = \int_0^\tau B(\tau') \frac{\exp\left(-\frac{(\tau - \tau')}{\mu}\right)}{\mu} d\tau' \quad 5.3$$

subject to the boundary conditions

$$I(\tau^*, \mu) = B(T(\tau^*)) \quad 5.4$$

$$I_v(0, -\mu) = B_v(TOA) \approx 0 \quad 5.5$$

which correspond to the isotropic emissions from earth’s surface (Eq. 5.4), an approximate blackbody in the IR, and at the top of the atmosphere (TOA) which for this case is approximately zero.

The exponential term in the integrand of Eq. 5.3 can be re-written to define a monochromatic transmittance or transmission function $T$, which is given by:

$$T(\tau / \mu) = \exp[-(\tau / \mu)] \quad 5.6$$

where $T$ includes the attenuating effects of all primary atmospheric gases, mostly $H_2O$ and $CO_2$ and trace gas species, including $CH_4$, $NO_x$ etc in the thermal IR. Upon taking
the derivative of $T$ with respect to $\tau$, the solution to $I_{\downarrow}(\tau, \mu)$ can then be expressed as:

$$I_{\downarrow}(\tau, -\mu) = \int_{0}^{\tau} B(\tau') \frac{d}{d\tau'} T[(\tau - \tau') / \mu] d\tau'$$ \hspace{1cm} 5.7

where the downward intensity $I_{\downarrow}(\tau, -\mu)$ becomes a sum of products of the Planck intensity $B$ and its corresponding weighting function $dT/d\tau$ for each layer of atmosphere at a given wavenumber $\nu$. The weighting functions as a function of height $z$ go from a highly peaked function near the surface where the strength of emissions is greatest to a broad function around 3km. Most of the AERI detected energy will therefore be in the lower layers near the surface [Feltz et al. 2003]. Following prescription of the dust layer properties, Eq. 5.7 can then be applied to determine the surface intensity.

To compute LW surface radiative forcing, the downward LW flux densities (herein referred to as fluxes) are required which is defined as the angular sum of the downward directional intensities given by Eq. (5.7). The angular sum yields the hemispherical flux over a solid angle of $2\pi$ steradians. Following the plane-parallel assumption, the downward surface flux $F_{\downarrow}$ for a single wavenumber can be written as:

$$F_{\downarrow}(\tau) = 2\pi \int_{0}^{\tau} I_{\downarrow}(\tau, -\mu) \mu d\mu$$ \hspace{1cm} 5.8

To facilitate the derivation of the flux equation, a physical parameter referred to as the diffuse transmittance, a measure of the exponential attenuation, is defined as:
\[ T(\tau) = 2 \int_0^\infty T(\tau/\mu) \mu d\mu \]  

which when combined with Eq. (5.7) yields the basic downward flux equation:

\[ F \downarrow (\tau) = \int_0^\infty \pi B(\tau') \frac{d}{dt} T(\tau - \tau') d\tau' \]

for a given level at wavenumber, \( \nu \). For the surface, the downward flux is made up of contributions from all overlying atmospheric layers. When accounting for the total flux across the LW domain, a spectral integration over all wavenumbers of all monochromatic fluxes yields the fundamental equation for broadband applications:

\[ F \downarrow (z) = \int_0^\infty F \downarrow (z) d\nu \]

where \( F \downarrow \) is written in terms of the height coordinate \( z \) since the optical depth \( (\tau) \) is a function of waveumber. Equations 5.7 and 5.11 form the basis for generalized IR radiation transfer for remote sensing and climate applications respectively.

5. B. RADIATIVE TRANSFER MODELS (RTM)

This study employs two radiative transfer models for investigating dust aerosol in the thermal IR. First the downward spectral radiances of AERI are calculated using a highly resolved line-by-line radiative transfer model (LBLRTM) which includes multiple
scattering and the second computes the corresponding broadband surface fluxes. A description of each model along with its associated input parameters follows.

5. B.1 CHARTS RTM

The one-dimensional radiative transfer code, the Code for High Resolution Accelerated Radiative Transfer with Scattering [CHARTS, Moncet et al., 1997], is used for simulating the downwelling AERI radiances. CHARTS employs the adding-doubling method for aerosol and cloud scattering, coupled with a line-by-line radiative transfer model [LBLRTM, Clough et al. 1992] with a line resolution ($\Delta \nu$) of 0.00015 cm$^{-1}$ for calculating gaseous absorption/transmission using the HITRAN 2000 line parameter database [Rothman et al. 1992]. CHARTS also employs the CKD2.4 water vapor continuum model [Clough et al. 1989]. The dust parameters inputted into CHARTS include the dust scattering ($\beta_s$) and absorption ($\beta_a$) coefficients, the dust single scattering albedo ($\varpi$) and the asymmetry parameter (g) all defined over the thermal IR window with a spectral resolution of 1 cm$^{-1}$. CHARTS calculates the AERI spectral radiances which are then later converted into their equivalent $BT$ spectra. Model observations are made from level 1 at the surface and the viewing zenith angle ($\theta$) is defined to be at 0 degrees with respect to the local zenith. To adequately represent the regional atmospheric state, the temperature and relative humidity profiles employed in the model atmospheres are taken from MAARCO and Abu-Dhabi airport radiosondes which are defined over 43 vertical layers with a grid spacing of 1 km. The surface boundary is characterized by an averaged spectral emissivity over the window region corresponding to a dark brown
quartz surface. Assuming the lowest dust layer is at the surface (i.e. near the source), the radiative transfer in a dust-laden atmosphere can be partitioned into three components as illustrated in Fig. 5.1: (1) Transmitted clear sky emissions mainly from water vapor ($I_{\text{clear}}$); (2) Emissions by the dust layer(s) ($I_{\text{dust}}$); and (3) Reflections of the surface emissions by dust layer(s) ($I_{\text{ref}}$). Re-writing Eq. 5.7 in terms of the pressure coordinate, the total downward IR radiance reaching the surface is given by:

\[
I_{\downarrow} = \int_{p_{\text{surf}}}^{p_{\text{top}}} B(T(p')) \frac{d\Im_{\text{dust}}(p-p')}{dp'} dp' + \Im_{\text{clear}} \int_{p_{0}}^{p_{\text{top}}} B(T(p')) \frac{d\Im_{\text{dust}}(p-p')}{dp'} dp' + \\
\Im_{\text{dust}} \int_{p_{\text{surf}}}^{p_{\text{top}}} B(T(p')) \frac{d\Im_{\text{dust}}(p-p')}{dp'} dp' + \\
r(\Im_{\text{dust}}) B(T_{\text{surf}}) \Im_{\text{dust}} + \int_{p_{\text{surf}}}^{p_{\text{top}}} B(T(p')) \frac{d\Im_{\text{dust}}(p-p')}{dp'} dp'
\]  

5.12

where $I_{\downarrow}$ is the downward surface radiance, $B[T]$ is the Planck radiance and $\Im_{\text{clear}}$ and $\Im_{\text{dust}}$ are the transmitted radiances for both clear and dust layers respectively. The variables $p_{\text{surf}}$, $p_{\text{top}}$ and $p_{0}$ represent the surface pressure, pressure at the top of the dust layer(s) and pressure at the TOA respectively. Lastly, $r$ is the reflectance from both the surface and the dust layer(s) which we assume to be negligible compared with the emission terms since the dust is at the surface. As before, the wavenumber ($\nu$) dependency of all the variables is implied in Eq. (5.12). The model dust layers range from the surface up to a maximum of 5 km, consistent with most field observations and the dust properties are prescribed to be uniform and homogeneous inside each layer.
5. B.2 FU-LIOU RTM

To model broadband LW surface fluxes in dust and dust-free conditions, the Langely modified Fu-Liou radiative transfer code (FL0403, April 15, 2003 - Rose et al. 2002; Fu and Liou et al. 1992) is employed. The model is run using a two/four-stream scheme with correlated k-distributions to account for absorption across 18 spectral bands (6 SW and 12 LW bands). The spectral bands are given in table 5.1.

This version of code allows for multiple/overlapped cloud inputs which are disabled for this study, and also includes various changes made to the boundaries in several of the spectral bands. The code also employs a parameterized version of the LW water vapor continuum model (CKD2.4) to account for the strong water vapor absorption. LW trace gas absorption in the model includes contributions from CO2, O3, N2O, CH4, and the CFCs. Modifications to the code allow for time series of retrieved AERI IR optical depths (scaled to 0.55µm), AERIPLUS atmospheric profiles and solar zenith angle (SZA) to be input for determining the daytime temporal variability of the LW surface forcing. Details of the AERI retrieved optical depths and AERIPLUS [Feltz et al. 2003] profiles are discussed in chapter 8 while the SZA is determined from local AERI observation times. Spectral surface emissivity ($\varepsilon_\nu$) across the 12-LW bands is specified using the JPL Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) spectral library, assuming a dark-brown quartz desert surface. The spectral land-surface albedo ($\alpha_\nu$) across the 6-SW bands is prescribed using the analytical spectral devices (ASD) spectroradiometer reflectance measurements observed during
UAE\textsuperscript{2} (\textit{Ji et al.}) where the albedo for each band represents an approximate time average during the measurement period. Table 5.2 lists the spectral emissivity and albedo input values used in the RTM.

A total of 15 albedo values are inputted into the code, where the first 10 correspond to the sub-bands of the first SW band. Observed surface-skin temperatures at SMART derived from a technique developed by \textit{Ji et al.} were also used as input where temperature measurements were logged using a thermistor attached to a scanning microwave radiometer (SMIR).

Dust aerosol inputs to the code allow for several options including the Tegen \& Lacis (1996) climatological dust model (referred to as TL96) based on \textit{Volz} [1973], and the OPAC database \cite{Hess et al. 1998}. For this study the TL96 dust model was employed to calculate the surface fluxes during typical dust conditions, however various sensitivity studies to dust composition, size and shape were performed using the dust microphysical models developed for this study (refer to chapter 6), including those for pure quartz \cite{Shettle and Fenn, 1979}, kaolinite, illite, montmorillonite, and an internal mixture of quartz mixed with 10\% hematite. The dust parameters represented a wide range of properties that were hardwired directly into the RTM code. The single scattering properties were computed as an average over the Fu-Liou LW spectral bands given in table 5.1. Lastly, a solar zenith angle of 0.9 was used for all forcing calculations.
Figure 5.1 A surface dust cloud is divided into n homogeneous layers with respect to the dust single scattering properties. Note, the first layer is defined at the surface increasing in the upwards direction. Arrows represent downward intensities at each layer detected by AERI (triangle) at the surface. The pressure scale is given along the RHS.
Table 5.1 Fu-Liou RTM - spectral bands

<table>
<thead>
<tr>
<th>Spectral region</th>
<th>Fu-Liou bands</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>SW (µm)</strong> *</td>
<td>0.2 - 0.7, 0.7 - 1.3, 1.3 - 1.9, 1.9 - 2.5, 2.5 - 3.5, 3.5 - 4.0</td>
</tr>
</tbody>
</table>

* first SW band is further subdivided into 10 sub-bands

Table 5.2 Fu-Liou RTM surface specifications

<table>
<thead>
<tr>
<th>Surface characterization</th>
<th>Fu-Liou bands</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Albedo (α)</strong></td>
<td>0.2, 0.2, 0.2, 0.2, 0.2, 0.2, 0.2, 0.2, 0.2, 0.30, 0.40, 0.35, 0.20, 0.2</td>
</tr>
<tr>
<td><strong>Emissivity(ε)</strong></td>
<td>0.91, 0.97, 0.98, 0.98, 0.84, 0.90, 0.96, 0.96, 0.96, 0.96, 0.96</td>
</tr>
</tbody>
</table>
Conventional approaches for calculating dust’s single scattering properties in the thermal IR window depend on one’s *a priori* knowledge and/or assumptions about dust composition, shape and size distributions. The following describes the modeling of each key dust parameter, and its implementation in the current study along with the theoretical basis and rationale.

**6-A. Dust composition**

Individual dust particles are usually mixtures of different types of minerals. These include silicates, such as quartz and clays (e.g. kaolinite, montmorillonite and illite), and non-silicates, which include carbonates (e.g. calcite and dolomite), iron oxides (e.g. hematite and magnetite) and sulfates (e.g. gypsum). Significant work in modeling dust composition has been performed by Sokolik *et al.* [1999] and Kalashnikova and Sokolik [2004].

To model dust composition for application to AERI dust detection/retrieval during the UAE², the refractive indices of typical minerals observed around the UAE region are employed, consisting of pure silicates (mainly quartz) and clays (kaolinite and illite).
Quartz is selected to be the major component of dust, since dust is mostly composed of quartz in terms of mass [Sokolik et al., 1999]. Kaolinite is included, because dust samples collected from Saudi Arabia, which is just south of UAE, were found to contain kaolinite at approximately 55% by weight [Sokolik et al., 1999]. Illite is also evaluated on the basis of work performed by Aba-Husayn et al [1977]. Lastly to consider particles with a heterogeneous composition, dust mixtures consisting of hematite (10%) and calcium carbonate (50%) internally mixed with kaolinite and quartz are examined. The calcium carbonate is used to account for the large distributions of distinct limestone sources found throughout the UAE region [E.A. Reid et al., 2007] and hematite for the possibility of traces of iron oxide in the soil. The Bruggemann effective medium approximation is used to calculate the refractive indices of internally mixed dust minerals [Sokolik et al., 1999]. Back trajectories from a kinetic trajectory code, provided by NASA-GSFC and made available by the AERONET program [Holben et al. 1998], show evidence for Saharan dust in the UAE region. For this reason, the Volz [1973] refractive indices are used to represent transported Saharan desert dust.

Figure 6.1 is a log-plot of the imaginary components (absorptive term) of the complex refractive index for common dust minerals in the thermal IR window (800-1200 cm\(^{-1}\)). The refractive indices from bulk dust samples including Volz [1973], Afghanistan dust [Sokolik et al. 1993] and D’Almeida [1991] is shown for comparison. For convenience, the curves are color-coded according to mineral class i.e. the clays are in red, quartz is in green, and sulfates and carbonates are in gray and light blue respectively. Those from the bulk dust samples are displayed in black. Note the large spectral
variability between mineral classes, whereas for the bulk dust samples, the spectra are smoothed out due to internal mixing of the component minerals. Unlike liquid water and ice (green and blue marked curves respectively), minerals exhibit strong absorption features as a result of their dominant molecular vibrational motions. The differences in the refractive indices between minerals and liquid water and ice in the thermal IR window form the basis for this paper’s dust/cloud detection technique. The AERI “clean” sub-bands employed in this study are denoted by the light gray vertical bars. The sub-bands are further discussed in chapter 8.

6-B. Dust size and shape distributions

6-B-1 Particle size

Particle size is a crucial parameter over which the single scattering properties for individual sizes and shapes are integrated to yield bulk dust optical properties. In this study, dust particle sizes are characterized using synthesized log-normal size distributions (Eq. 2.2) with size parameters obtained from the UAE^2 – MAARCO aerodynamical particle sizer (APS) dataset. Figure 6.2 shows re-binned APS size distributions according to surface area mean radius based on particle count measured during the UAE^2. The MAARCO site operated a TSI APS 3321 spectrometer during the UAE^2 from August 11 – September 30, 2004 (refer to chapter 4). J.S. Reid et al. [2007] discuss the technical details and data collection protocol for the instrument.

For unique APS size groups that were found to be characteristic of specific dust
source regions, the volume median diameters (VMD) and their geometric standard deviations ($\sigma_{gv}$) were identified [J.S. Reid et al. 2007]. Each of these groups was associated with a specific source location. The average VMD and $\sigma_{gv}$ were $3.8 \pm 0.4 \ \mu m$ and $2.0 \pm 0.1, \ \mu m$ respectively. To account for the possibilities of the presence of fine and more coarse-mode desert dust at SMART, the range of VMD’s were extended to include smaller ($r_{eff} =0.75 \mu m$) and larger ($r_{eff} = 5 \mu m$) sized particles each with a $\sigma_{gv}=1.9-2.0$. The effective size range employed in this study was $r_{eff} = 0.75-5.0 \mu m$. Since the AERI retrieved IR optical depths will be dominated more by the presence of coarse-mode dust compared with the fine-mode influence in the visible wavelengths, the extended range of size parameters is significant for determining, if any, the sensitivity of the AERI $BT$ spectra to particle size. Details of this are given in chapter 8. The size distributions were constructed using the averaged integrated coarse-mode (0.8-10$\mu m$) dust particle concentrations ($N=116cm^{-3}$) corresponding to the large dust episode that occurred on September 12, 2004.

6-B-2 Particle shape

Dust particles are rarely spherical as evidenced from recent analysis [Koren et al. 2001] which shows that as dust particle size increases, as in the case of heavy dust outbreaks, so too does its circularity, a measure of its deviation from a perfect sphere. This study uses both spherical and non-spherical dust particles for examining the sensitivity to key dust and atmospheric parameters.

To investigate the effect of sharp-edged dust particles [Koren et al., 2001] a single
randomly oriented particle geometry was chosen: a hexagonal flat plate (i.e. compact hexagonal column) to represent for example clay particles, with aspect ratio L/2A =1 where L is the axial length parameter and A is the half width. Table 6.1 gives the L and A values used in the calculation based on the volume-equivalent spherical radii: 0.1, 1.0 and 2.5µm.

This study also assumes dust particles are randomly oriented oblate spheroids, following the work of J.S. Reid et al. [2003] and Okada et al. [2001] who found typical aspect ratios for dust particles to be in the range of 1.4-1.9, based on samples collected during PRIDE and in China respectively. Aspect ratio distributions from PRIDE [E.A. Reid et al., 2003], which preliminarily seem to match the findings during the UAE² field campaign, are used in this study to account for the range of particle shapes when calculating the dust single scattering properties. Figure 6.3 [from E.A. Reid et al., 2003] shows the cumulative probability plot of the particle aspect ratios as a function of the particle average diameter derived from aircraft particle samples. The plot indicates the aspect ratios are rather large and broad-based with median values averaging 1.9±0.9 with the largest particles (>10µm) representing approximately 3% of the sampled population.

6C Dust single scattering properties

The classical approach for computing dust single scattering properties is to assume that dust particles are spheres, and the Lorenz-Mie scattering code is applied. The non-spherical and irregular nature of dust particles [Okada et al., 2001], however, demands that a more detailed light scattering code, which is able to resolve the non-
spherical particle geometry, be employed for determining the single scattering properties. Several such codes are available.

The T-matrix method [Waterman, 1971; Mishchenko et al., 1994] an analytical approach to light scattering, is useful for simulating spheroids, circular cylinders and chebyshev particles. For this study, dust particles are treated as oblate spheroids based on the works of J.S. Reid et al. [2003] and Okada et al. [2001]. Dust particles with extremely large aspect ratios and imaginary refractive indices or large sizes may cause the T-matrix method to produce unrealistic results (e.g. single scattering albedos $\varpi > 1$) or fail to converge [Mishchenko et al., 1998]. Several experiments were conducted to test the code’s sensitivity to these extreme parameters. It was found that using common silicate minerals with $L/2A < 2.3$ yields physically realistic single-scattering properties. When the particle asphericity was further increased, the single scattering albedo was found to exceed 1, most likely due to numerical instability caused by extreme values of the particle parameters. For this reason, the single-scattering properties were integrated over the range of aspect ratios of 1.2-2.2, corresponding to particles that are nearly spherical to those that are elongated (oblate) spheroids.

The finite difference time domain method [FDTD – Yee, 1966; Yang and Liou, 1995], is also employed which is a numerical approach for simulating light scattering by particles of more complicated shapes. For this study, dust particles are represented as compact hexagonal columns (i.e. flat plate-like structures –E. A. Reid et al., 2003), similar to the common clay, kaolinite [Kalashnikova and Sokolik, 2004].

For the particle sizes 0.5-20 µm based on APS measurements, and the
wavelengths in the thermal IR window (8-12 µm), the size parameter ($x = 2\pi a/\lambda$, where ‘a’ is the particle radius and $\lambda$ is the incident wavelength), ranges from ~0.15-15. The numerical processing time of FDTD can be on the order of weeks for large size parameters of the target particle, because of the large memory requirements. For example, on a Unix-based Sun system (Solaris 5.8), up to 500 run-time hours were required to calculate the visible scattering properties at $\lambda=0.55$ µm of hexagonally shaped dust columns with size parameter, $x=6$. Since the peak of most APS size distributions is located at less than 10µm, the size parameters in the thermal IR window are so small, that only FDTD is needed. However, for size distributions observed near the source regions of large dust outbreaks, which contain more large sized particles, the application of FDTD alone can be computationally inhibitive. One must resort to a modified geometric optics program [GOM - Yang and Liou, 1996b]. The coupling of GOM and FDTD in the context of a unified theory for light scattering is described in detail in Liou [2002].

Both the T-matrix and FDTD methods were used to calculate the single scattering properties for individual particle sizes and shapes over the thermal IR window based on the observed dust parameters. The single scattering properties are the extinction and scattering coefficients ($\beta_e$ and $\beta_s$), the single scattering albedo ($\varpi$), and the asymmetry parameter ($g$). $\beta_e$, $\beta_s$, $\varpi$, and $g$ are integrated over the UAE$^2$ - APS size distributions and the aspect ratio shape distributions from PRIDE [E. A. Reid et al., 2003] to derive the bulk (mean) optical dust properties using the following expressions:
\[
\langle \beta_e \rangle = \sum_{i=1}^{2} \sum_{j=1}^{2} \sum_{k=1}^{N} W_k(a_0) \sigma_c(a_0) j n(a_0) \Delta a_0 \quad (6.1)
\]

\[
\langle \beta_s \rangle = \sum_{i=1}^{2} \sum_{j=1}^{2} \sum_{k=1}^{N} W_k(a_0) \sigma_s(a_0) j n(a_0) \Delta a_0 \quad (6.2)
\]

\[
\langle \varpi \rangle = \frac{\langle \beta_s \rangle}{\langle \beta_e \rangle} 
\]

\[
\langle g \rangle = \frac{\sum_{i=1}^{2} \sum_{j=1}^{2} \sum_{k=1}^{N} W_k(a_0) g(a_0) j \sigma_s(a_0) j n(a_0) \Delta a_0}{\langle \beta_s \rangle} \quad (6.4)
\]

where \( \langle \beta_e \rangle, \langle \beta_s \rangle, \langle \varpi \rangle, \) and \( \langle g \rangle \) are the bulk mean extinction and scattering coefficients, single-scattering albedo, and mean asymmetry parameter respectively. Particle extinction and scattering cross sections are denoted as \( \sigma_c \) and \( \sigma_s \) respectively. In Eqs. (6.1), (6.2), and (6.4), the first summation from \( \lambda_1 \) to \( \lambda_2 \) is over the wavelength spectral domain, while the second summation from \( a_1 \) to \( a_2 \) is over the APS size intervals \( [n(a_0)] \), where \( n \) is particle count per \( \text{cm}^3 \) evaluated at the center of each APS size bin \( a_0 \) (\( \mu \text{m} \)). The third summation from 1 to \( N \) is over the shape distribution for \( N \) possible dust shapes with \( W(a_0) \) being the aspect ratio weighting factor. It is noted that the size integration of optical properties for the oblate spheroid was performed over size bins ranging from 0.1-5\( \mu \text{m} \) in steps of 0.2\( \mu \text{m} \). Due to the large memory requirements and computational time required by FDTD, the size integration for the compact hexagon was performed over three size bins from 0.1-2.5\( \mu \text{m} \) (Table 6.1).

Figures 6.4(a) and 6.4(b) show the calculated single scattering albedos (\( \varpi \)) and asymmetry parameters (\( g \)) for pure and internally mixed mineral compositions employing the T-matrix and FDTD methods over the window region. Oblate spheroids and compact
hexagons are labeled (OS) and (CH) respectively. Four (4) mineral compositions are shown including the Volz model (V), 100% quartz (Q), 10% hematite/kaolinite mixture (K/H) and a 50% calcium carbonate/kaolinite mixture (K/C). Pure kaolinite is not shown since this does not vary much compared with the mixtures. Figures 6.4(c) and 6.4(d) show the corresponding size variation for the Volz $\varpi$ and $g$ respectively. As was also reported by Highwood et al. [2003], the single scattering properties show little change with size, compared to mineral composition, although some slight differences are noted. For example, Fig 6.4(c) shows that as particle size increases, so too does $\varpi$ ($\leq 4\%$), possibly due to the increase in particle surface area. Radiative transfer simulations show that increased scattering reduces the magnitude of the emissivity effects of the resulting $BT$ spectrum, when observed from a ground-based AERI system.

The spectral average and the root mean square (RMS) variance of $\varpi$ and $g$ are $0.4629\pm0.0231$ and $0.1563\pm0.0162$, respectively. In Figs. 6.4(a) and 6.4(b), note the complex spectral variability in the optical properties for each mineral. For comparison, the solid black curves represent the FDTD optical parameters for the Volz compact hexagon with effective size distribution $r_{\text{eff}} = 2\mu$m.

Consistent with Highwood et al. [2003], the single scattering properties for the cases analyzed exhibit greater spectral variability with refractive index compared to particle size or shape. This illustrates the significance of having $a$ priori knowledge of particle composition for accurately retrieving dust properties.

In Fig. 6.4(a), the compact hexagons in the 8-9$\mu$m spectral region scatter more (~factor of 2) than oblate spheroids. In the intermediate part of the window region, both
shapes have similar single scattering albedos, whereas in the 11-12 μm region, the hexagonal particles exhibit greater absorption. These differences may be attributed to the reduced size integration for the compact hexagon, where contributions from larger sized particles are not considered. Differences may also be due to the integration over particle shape for the oblate spheroid, whereas, we only consider a single shape using FDTD. In Fig. 6.4(b), the compact hexagons are shown to have a significantly greater amount of forward scatter than do oblate spheroids. It is noted that the asymmetry parameter for FDTD is scaled down 3X for comparison. Generally the asymmetry parameter for each composition is found to decrease with wavenumber.
Table 6.1 FDTD Size Parameters

<table>
<thead>
<tr>
<th>Volume-Equivalent Spherical Radius (µm)</th>
<th>L(µm)</th>
<th>A(µm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1</td>
<td>1.48E-01</td>
<td>2.96E-01</td>
</tr>
<tr>
<td>1.0</td>
<td>1.48E+00</td>
<td>2.95E+00</td>
</tr>
<tr>
<td>2.5</td>
<td>3.69E+00</td>
<td>7.39E+00</td>
</tr>
</tbody>
</table>
Figure 6.1. Imaginary component of refractive indices for common dust minerals versus liquid water and ice in the thermal IR window. Afghanistan and D’Almeida dust models and calcium carbonate are scaled 5X and 100X respectively for comparison.
Figure 6.2. Dust aspect ratio cumulative probability plot from PRIDE 2000 (E. A. Reid et al., 2003) used in constructing the study’s dust model shape distribution. Each curve corresponds to the probability of finding a particle’s aspect ratio within the specified size range.
Figure 6.3 Re-binned APS size distributions measured during the UAE$^2$. N1, N3 and N6 are the size distributions with effective radii, 0.8, 1.0 and 2.0µm respectively.
Figure 6.4 Dust single scattering properties for compact hexagons (CH), and oblate spheroids (OS) with variable compositions in the thermal IR window. (a) Single scattering albedo for 4 compositions: V=Volz, Q=quartz, K/C=50% kaolinite/calcium carbonate, and K/H=10% kaolinite/hematite. (b) Asymmetry parameter for the same dust parameters. (c) Volz size variation of single scattering albedo. (d) Volz size variation of asymmetry parameter.
CHAPTER 7

7.0 SATELLITE DUST DETECTION USING MODIS THERMAL IR WINDOW RADIANCES: METHODOLOGY, CASE STUDIES, AND VALIDATION

Abstract
An integrated method for the simultaneous detection/separation of mineral dust and clouds for both daytime and nighttime conditions using MODIS thermal infrared window brightness temperature data has been developed. Based on the spectral variability of dust emissivity at 3.75, 8.6, 11 and 12 µm wavelengths, three heritage approaches are combined to identify dust and cirrus. MODIS data for three dust-laden scenes have been analyzed to demonstrate the effectiveness of this detection/separation method. The detected daytime dust and cloud coverage for the Persian Gulf case compares reasonably well to those from the “Deep Blue” algorithm developed at NASA-GSFC. Validation of the nighttime dust and cloud detection method has been carried out by using the cases surrounding Cape Verde and Niger, West Africa on the basis of the coincident and collocated ground-based micro-pulse lidar measurements.

7.1 Introduction

Global detection of tropospheric mineral dust is often hampered by the radiative effects of clouds, in particular thin cirrus [Roskovensky and Liou 2005]. The effects of not properly detecting dust outbreaks may result in 1. Large biases in satellite retrieved cloud and surface parameters [Reid et al. 2003], 2. Impeding military [Miller 2003] and
commercial operations due to severe reduction of near-surface visibility and 3. Causing adverse health effects [Prospero 1999].

Recently, a number of visible and IR channel techniques have been developed for dust detection. Hutchison and Jackson [2003] showed that the reflection by sand increases steadily with wavelength in the 0.4 to 1.0 μm spectral region and implemented a 0.41µm threshold test to identify clouds over desert surfaces in the future NPOESS/VIIRS cloud mask. Miller [2003] implemented a multi-spectral dust enhancement algorithm that combines dust coloration properties with the MODIS negative 11-12µm brightness temperature difference (BTD11-12) for the daytime and nighttime detection of dust particles over dark and bright surfaces. Roskovensky and Liou [2005] followed the conceptual approach of Hutchison and Jackson and used a dust detection parameter in conjunction with the cirrus detection parameter [Roskovensky and Liou 2003] to separately identify thin cirrus clouds from airborne dust.

This chapter focuses on the detection of dust over major dust outbreak and transport areas using MODIS infrared brightness temperatures (BT). An integrated approach has been developed to simultaneously detect and separate airborne mineral dust from clouds along with distinguishing cloud phase to facilitate dust aerosol remote sensing, with emphasis on nighttime dust events. This approach is based on the spectral variability of dust emissivity at 3.75, 8.6, 11 and 12 μm wavelengths. MODIS data for three dust-laden scenes, including a daytime dust case over the Persian Gulf and two nighttime dust events over the Cape Verde Islands and Niger, West Africa, have been analyzed to demonstrate the effectiveness of this approach. The organization of this chapter is as follows. Section
7.2 discusses the dust detection methodology. Section 7.3 presents the MODIS data and details of each case study. Finally, a summary along with potential applications is given in section 7.4.

7.2. Dust detection methodology

The integrated dust detection technique consists of three independent approaches described below. The first is a threshold detection technique referred to as the “$D$ (dust) – parameter method” following the concept developed by Roskovensky and Liou [2005]. The $D$-parameter is defined as

$$D = \exp \{[rr \cdot a + \text{BTD}_{11-12} - b]\}, \quad (7.1)$$

where $rr$ represents the 0.54 µm/0.86 µm reflectance ratio, $a$ is a scaling factor, and $b$ is the $\text{BTD}_{11-12}$ offset. Roskovensky and Liou showed that the $D$-parameter is highly sensitive to cirrus cloud optical depth, since it is an exponential function of $rr$ and $\text{BTD}_{11-12}$. Because the $D$-parameter test requires solar channel data, it can only be applied to local daytime conditions. To detect nighttime dust, a $D^*$-parameter test has been devised in terms of $\text{BTD}_{11-12}$ and $\text{BTD}_{8-11}$ (the $\text{BTD}$ between 8.6 µm and 11µm) as follows:

$$D^* = \exp \left[\frac{\text{BTD}_{11-12}-C}{\text{BTD}_{8-11}-E}\right], \quad (7.2)$$
where parameters $C$ and $E$ are the thermal offsets for $BTD_{11-12}$ and $BTD_{8-11}$, respectively. These offsets are adjustable based on analyses of dust-laden MODIS scenes and simulations to enhance the dust signal against a cloud-filled background. In comparison to the $D$-parameter test, the $BTD_{8-11}$ term replaces the reflective ratio $rr$, providing an additional constraint on dust detection. This is significant since many silicate minerals that have strong restrahlen bands often absorb more at 8.6$\mu$m than at 11$\mu$m, leading to a negative $BTD_{8-11}$. In contrast, cirrus cloud (ice) particles absorb more at 11$\mu$m than at 8.6$\mu$m, producing a positive $BTD_{8-11}$, except when there is a large amount of water vapor, which could make $BTD_{8-11}$ negative due to the enhanced water vapor absorption at 8.6$\mu$m.

The detection test is so designed that $D^* > 1$ indicates dust and $D^* \leq 1$ indicates cloud. This method is herein referred to as dust filter 1 (DF1). DF1 does not discriminate cloud phase, however. Theoretical calculations similar to those conducted by Roskovensky and Liou have been performed to evaluate the $BTD$ terms in Eq. (7.2) for the four IR channels as a function of dust and cloud optical depths. The $BTD$s for both cirrus/water clouds and various compositions of dust including quartz, quartz mixed with hematite, and several clays have subsequently been analyzed.

Figures 7.1(a), 1(b) and 1(d) show the calculated $BTD_{8-11}$, $BTD_{11-12}$ and $D^*$-parameter as functions of optical depth for cirrus and dust conditions over land. The dust was assumed to be pure quartz with a spherical shape, and its optical properties were generated from a Lorenz-Mie scattering code based on the spectral refractive indices given by Shettle and Fenn [1979]. Cirrus clouds were modeled following the approach
developed in Roskovensky and Liou [2005]. A typical Persian Gulf summer atmospheric profile was used and the land surface reflectivity was modeled by using the JPL Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) spectral library, assuming a dark brown quartz surface. For the computation of $D^*$, the offsets $C$ and $E$ in Eq. (7.2) were set as -0.5 and 15, respectively, based on prior analyses of MODIS dust-laden scenes and simulations. Theoretical simulations suggest that offset $C$ can vary between -0.5 and 0.35, as $BTD_{11-12}$ is sensitive to dust composition, surface type, and column water vapor. Further work is required to develop a dynamical adjustment to these offsets. However, for the purposes of this study, the chosen constant values are sufficient to demonstrate the effectiveness of the present approach.

For dust, $D^*$ increases exponentially with optical depth ($\tau_d$) due to a large negative $BTD_{11-12}$ [Fig. 7.1(b)]. Because of this strong sensitivity, the $D^*$ test using the prescribed model parameters is shown to be effective in separating dust from cirrus and thus appears to be an excellent initial approach for detecting dust.

Sensitivity tests were performed to evaluate the $D^*$-parameter’s response to changes in dust layer height, particle asphericity, low clouds and other aerosol. For dust layer height, theoretical simulations suggest that increasing the altitude decreases the $D^*$-parameter’s sensitivity, because an increase in thermal emission from a vertically extended dust layer tends to increase $BTD_{11-12}$. For particle asphericity, oblate spheroids were chosen with aspect ratios of 2 and 3, consistent with previous research results [Reid et al. 2003]. The $D^*$-parameter was found to be slightly larger than that over the spherical value on the order of 0.5-1% due to an increase in $BTD_{8-11}$ by ~1K. For
spheroids, particle asphericity effects are minimal. For low clouds, theoretical simulations over a quartz surface indicate that day/nighttime coarse-mode dust will not be detected when it is located beneath a cloud layer of moderate optical depth ($\tau_{\text{cld}} \sim 4$). Elevated dust above clouds, however, particularly at night, can be detected provided the dust has a minimum optical depth ($\tau_d \geq 0.07$). Lastly, dust was replaced by other aerosol: coarse-mode sea-salt (RH=80%), ammonium sulfate and black carbon in cloud-free conditions. For the cases tested, the $D^*$-parameter fell below the threshold ($D^* < \varepsilon$) where $\varepsilon=1$ using the prescribed thermal offsets in Eq. (7.2). Adjustment of these offsets could potentially be used to separate dust from these other aerosol in cloud-free conditions.

The second approach is the $BTD_{8-11}$ vs. $BTD_{11-12}$ slope method, which follows the principle of the cloud phase algorithm developed by Strabala et al. [1994]. The algorithm was modified and used the slope of the best-fit straight line for the scatter plot of $BTD_{8-11}$ vs. $BTD_{11-12}$ to identify dust and cloud. Specifically, if a grid domain contains less than a prescribed number of clear pixels, using the MODIS cloud mask (MOD-35, Level 2), then both $BTD_{8-11}$ and $BTD_{11-12}$ are calculated, a linear regression is performed, and the slope of the best-fit line [i.e. $\Delta BTD_{8-11}/\Delta BTD_{11-12}$ where $\Delta$ denotes a differential change of the parameter] is determined. If the slope is negative, the domain pixels are identified as dust. Scatter data plots over the dust-filled regions from selected MODIS scenes and radiative transfer simulations show that the slope is negative for dust. Similar to cirrus clouds, although of opposite sign, the change in $BTD_{8-11}$ is much larger than the corresponding change in $BTD_{11-12}$, which follows from the
differences in the spectral absorption coefficients for common dust minerals. If the slope is positive, however, the pixels are flagged as cloud. For the identified cloudy pixels, if the slope is between 0 and 1, the domain pixels are flagged as water cloud based on thresholds established from a sample of observed cases. Otherwise, the domain pixels are labeled as cirrus cloud (for slope greater than 1). This test is referred to as dust filter 2 (DF2). It takes the output of DF1 which initially identifies the dust, and then applies the slope test to confirm the presence of dust. If both $D^*$-parameter and $BTD$ slope tests identify a pixel as dust, it is classified as dust, otherwise it is classified as cloud and its thermodynamic phase is subsequently determined.

The third approach is the $BTD_{3-11}$ ($BTD$ between 3.75 µm and 11 µm) method. Based on limited observations and model simulations, data points of $BTD_{3-11}$ vs. the 11 µm $BT$ fall into: separate clusters for clear-sky, dust, and water and cirrus clouds. For the same 11 µm $BT$, $BTD_{3-11}$ for cirrus clouds generally exhibit the largest values due to a significant difference in water vapor and cloud spectral optical properties for the two bands. The $BTD_{3-11}$ for dust and water cloud is smaller than that for cirrus due to less variation in their spectral optical properties. Thus, dust and cirrus cloud can be separated based on the differences in $BTD_{3-11}$. This method is quite sensitive to the assumed dust composition and surface parameters as well as column water vapor. Figure 7.1(c) shows that $BTD_{3-11}$ for dust is a function of the dust optical depth ($\tau_d$). This dependency can be further used to qualitatively separate optically thin ($\tau_d < 1$) and thick ($\tau_d > 1$) dust regions, which are identified by DF1 and DF2 tests, while detection thresholds can be established on the basis of simulations and observational data.
7.3. Data and case studies

Three MODIS dust scenes were selected to demonstrate the effectiveness of the integrated dust detection approach presented above. The MOD-35 cloud mask was used in the current approach to identify clear pixels based on the 99% confidence flag. This forms the basis for the detection analysis since the algorithm is applied where the MOD-35 flags clouds. A 1km land/water mask was also employed to define land and ocean surfaces. A data reduction method subdivides the MODIS granule into 5 x 5 grids covering an area of approximately 25 km². If MOD-35 identifies at least 75% clear sky pixels in the grid space it is labeled as clear, otherwise the algorithm proceeds to the BTD tests. Using the central frequency of each channel, the measured radiances are converted to equivalent $BT$.

The first case is a daytime MODIS/Aqua overpass across the Persian Gulf on September 12, 2004 at 1005UTC. The MODIS visible image shown in Fig. 7.2(a) is provided for reference. The MODIS scene [Fig. 7.2(b)] captures a large dust outbreak area (white circle), transported over the northern Persian Gulf and along the Strait of Hormuz. The MODIS cloud mask product indicates that clear sky is predominant over most of the land areas. Consequently, there is almost no detected dust over land, although uncertainties in the cloud mask may lend itself towards detecting optically thin dust over land surfaces, denoted by the light gray color shown in Fig. 7.2(b). Dust detection over seas, on the other hand, is quite apparent. Figure 7.2(b) shows that the present integrated approach detects dust (red/yellow) over the northern Persian Gulf, the Strait of Hormuz, and the Gulf of Oman. Light to moderate dust loadings ($\tau_d < 1$) shown in red were found
to occupy about 42% of the area over water. Likewise, about 2% of the area (yellow) was found to contain high dust loads ($\tau_d > 1$). Clusters of water clouds (green) are seen over the Strait of Hormuz, the Gulf of Oman and over portions of the Oman mountain range. Qualitative comparisons of the spatial coverage of this method’s detected clouds with those in the visible image and the MODIS cloud/phase mask (MOD-06, Level 2) shows reasonable agreement. Note the clouds in Fig. 7.2(a) have more structure compared to dust which helps further distinguish between the two fields particularly around the Strait of Hormuz. This method’s detected dust complements the MOD-06 water clouds identified over the Persian Gulf since this region was heavily obscured by dust. This is evident in the Deep Blue results [Fig. 7.2(c)] which show substantial dust coverage. Scattered cirrus clouds (blue) were found over parts of the dust plume in the Persian Gulf. Although the visible MODIS image shows evidence for clouds in this region, some may actually be due to dust where the $D^*$-parameter falls below its threshold due to a positive increase in $BTD_{11-12}$. This can be attributed to changes in mineral composition, surface type and water vapor amounts. For example, the clay, illite, can cause the $D^*$-parameter to fall below its threshold making it necessary to adjust thermal offset $C$.

Figure 7.2(c) shows the contours of the retrieved dust optical depth at $\lambda = 0.50\, \mu m$ for the same date presented by Hsu et al. [2004] using a daytime solar algorithm, referred to as Deep Blue. For details of the Deep Blue algorithm, please refer to Hsu et al. [2004]. Comparing Fig. 7.2(b) to Fig. 7.2(c) shows the integrated approach produces similar dust coverage, particularly for the regions over the Persian Gulf. In Fig. 7.2(c), for example, the results from Deep Blue capture the dust outbreak at the northern tip of
the Persian Gulf (white circle) and a significant dust loading found along the Strait of Hormuz. The white patches indicate areas of clouds where no retrieval was performed. The integrated approach also detects clouds near to those of Deep Blue, particularly along southern Iran’s coastal regions. The retrieved optical depths ($\lambda = 0.50 \mu m$) at the AERONET Sir-Bu-Nuair site (25°N, 54°E; within the white square), collocated and coincident with the Aqua overpass, were found to be about 1.018. For the same area, the Deep Blue retrieved $\tau_d$ was about 1.20, while the integrated approach identifies the dust region as optically thin with an estimated $\tau_d$ close to 1 in close agreement with those from Deep Blue and AERONET. The integrated approach can also be useful in the sun-glint area by employing only DF1 and DF2 since $BTD_{8-11}$ and $BTD_{11-12}$ are not subject to variations in solar reflection. This capability in the current approach could complement MODIS daytime dust detection algorithms.

The second case chosen is associated with a nighttime dust scene over the Cape Verde Islands during the recent NASA African Monsoon Multidisciplinary Analysis (NAMMA) field campaign on September 7, 2006 at 0245UTC. The MODIS IR image shown in Fig. 7.2(d) is provided for reference. The white box at the bottom of Fig. 7.2(e) shows the location of the islands at 16.732°N, 22.935°W. Evident in this scene is a prominent dust plume with small to moderate optical depths (red) that follows the African coastline. The division of dust along the coastline further illustrates the difficulties with over land detection. As in the previous case, MOD-35 indicates clear sky over most of the land areas, so very little dust is detected over land leading to the coastal divisions. Much of the area including Sal Island is labeled ‘uncertain’ (gray), to
account for uncertainties in the MOD-35. There also appears to be a continental outflow of dust (red) around 18°N, 15°W, near Mauritania and the Western Sahara, both of which are major dust source regions. Further inland, streaks of optically thicker dust (yellow) are found. Along the northwest and southeast corners of the MODIS granule are bands of mostly water clouds mixed with some cirrus. Validating cloud phase away from surface sites proves difficult due to the lack of data; however we suspect on the basis of the current approach that clouds occupied these areas with slopes indicative of both water and ice. It is clear more cases are needed to have a complete validation, however this approach demonstrates a clear potential for dust and cloud separation during nighttime conditions.

The Surface-sensing Measurements for Atmospheric Radiative Transfer and Chemical Optical and Microphysical Measurements for in situ Troposphere (SMART-COMMIT) developed at Goddard Space Flight Center/NASA was deployed at the Sal Island during NAMMA. SMART’s micro-pulse lidar (MPL), a key instrument for profiling the vertical structure of aerosol and clouds, operated continuously, with the exception of local noon. Figure 7.2(f) shows the MPL normalized relative backscatter (NRB) signal for September 7, 2006. At the Aqua overpass time (0245UTC) marked by the solid white line, a strong dust aerosol signature up to 4km is evident. This is likely associated with the Saharan air layer transport, similar to the lidar observations presented by Sassen et al. [2003]. Almost two hours before the Aqua overpass, low level clouds around 1km were detected. The current detection scheme for the dust layer over the Sal Island is consistent with the MPL observations. In addition, coincident measurements
from the SMART’s AERI instrument observed higher brightness temperatures of about 5K at the 10µm wavelength, where dust aerosol absorbs more IR radiation than other wavelengths in the window region.

In addition to the preceding two cases, we have selected another nighttime dust scene over Niamey, Niger in West Africa, observed by the ARM Mobile Facility (AMF) on March 8, 2006 at 0100UTC. The MODIS IR image shown in Fig. 7.2(g) is provided for reference. The AMF experienced a significant dust episode, beginning on March 7, 2006 and lasted several days. The white box in Fig. 7.2(h) marks the location of AMF at 13.477°N, 2.175°E. Satellite images indicate that the dust storm originated in the central Sahara (Niamey News, ARM website). The detection results in Fig. 7.2(h) show a prominent dust band with light to moderate optical depths (red) traversing the north central parts of the continent, particularly through Niger and Chad. A significant amount of cirrus clouds (blue) was also found over the detected dust, with an extensive mix of both water (green) and cirrus clouds to the south. An MPL has been continuously operating at the AMF since November 2005. Figure 7.2(i) shows the instrumentally corrected NRB signal for March 8, 2006. At the MODIS/Aqua overpass time (0100UTC), marked by the solid white line, a significant amount of dust aerosol can be seen up to 4km, including a high-level cirrus cloud at 13km. Again, the present scheme has the capability to detect dust directly over the AMF site and to identify overlying cirrus clouds consistent with the MPL observations.
7.4. Summary

An integrated method for the detection and separation of mineral dust from cirrus clouds using the MODIS thermal IR window band data has been presented. The $D^*$-parameter, $BTD$ slope, and $BTD3-11$ methods were combined to detect dust and cirrus for both nighttime and daytime scenes. This integrated approach was applied to three MODIS dust cases to demonstrate its feasibility and reliability. The first is a daytime case over the Persian Gulf. By comparing the results determined from the integrated method and the NASA Deep Blue solar retrieval algorithm, the reliability of the present approach in detecting dust events during daytime was demonstrated. The two other cases are nighttime scenes over the Cape Verde Islands and Niamey, Niger. By comparing the detection results to the MPL observations at each site, it was illustrated that the present scheme can be used to detect dust and to separate dust from cirrus during nighttime conditions with some confidence. Of course, more cases are needed to have a complete validation of the present approach. Nevertheless, it offers a variety of promising applications, including: (1) scene classification for the dust radiative forcing studies, (2) dust aerosol correction for improved sea surface temperature retrievals and dust data assimilation in the coupled ocean-atmosphere model, (3) application of the detection results to the current operational daytime aerosol retrieval algorithm, and (4) assessment of the nighttime dust hazard to improve transportation safety and mitigate dust’s adverse health effects.
Figure 7.1. Simulations of the BTD and $D^*$-parameter versus optical depth ($\tau$) for dust (pure quartz) and cirrus over land surface whose soil composition is prescribed to be quartz. Solar and satellite zenith angles used were 30° and 0°, respectively. (a) BTD8-11, (b) BTD11-12, (c) BTD3-11, and (d) $D^*$-parameter
Figure 7.2. (a) MODIS visible image of the Persian Gulf on September 12, 2004 at 1005UTC (b) Daytime integrated dust detection over the Persian Gulf (c) Retrieved visible optical depths using the Deep Blue scheme over the Persian Gulf (d) MODIS IR image showing the Cape Verde Islands on September 7, 2006 at 0245UTC (e) Nighttime integrated dust detection over the Cape Verde Islands (f) SMART MPL normalized backscatter at Sal on the Cape Verde Islands during NAMMA (g) MODIS IR image showing Niamey, Niger on March 8, 2006 at 0100UTC (h) Same as (e), except over Niamey, Niger (i) AMF MPL normalized backscatter at Niamey
CHAPTER 8

8.0 GROUND-BASED AERI DUST DETECTION AND RETRIEVAL: SENSITIVITY AND MODELING STUDIES, METHODOLOGY, CASE STUDIES, AND VALIDATION

Abstract

Numerical simulations and sensitivity studies have been performed to assess the potential for using $BT$ spectra from a ground-based AERI during the UAE$^2$ for detecting/retrieving dust aerosol. A methodology for separating dust from clouds and retrieving the dust IR optical depths was developed by exploiting the differences between the spectral absorptive power for dust and cloud in prescribed thermal IR window sub-bands. The dust microphysical models were constructed using in-situ data from the UAE$^2$ and prior field studies. Dust composition using refractive index datasets for minerals commonly observed around the UAE region including quartz, kaolinite, illite, calcium carbonate and hematite. The Volz Saharan dust refractive indices were also examined. Dust single scattering properties for oblate spheroids and hexagonal plates, two particle geometries routinely interpreted in electron microscopy and spheres were computed using the T-matrix, FDTD and Mie light scattering programs. Sensitivity of the modeled AERI spectra to key dust and atmospheric parameters were investigated using the CHARTS radiative transfer program.

Four daytime UAE$^2$ cases were evaluated to demonstrate the strength of the
detection/retrieval technique and its sensitivity to the estimated MCT detector non-linearity was assessed. Preliminary results were compared to collocated AERONET sun-photometer/ MPLNET micro-pulse lidar measurements.

8.1. Introduction

The effect of mineral dust on earth’s climate system remains highly uncertain (IPCC 2001) due in part to a lack of data and a comprehensive understanding of its complex radiative properties, particularly in the thermal IR where common dust minerals exhibit a wide range of spectral features [Sokolik et al. 1999]. Exploiting the long wave properties of dust for measuring and modeling high resolution dust spectra has been the focus of recent research efforts including, DeSouza-Machado et al. [2006], Hong et al. [2006], Pierangelo et al. [2005] and Pierangelo et al. [2004]. Although significant progress is being made, accurate dust parameterizations for remote sensing and climate applications will strongly depend on the availability of quality global dust data from major source regions around the world.

In response to the need for dust data, there has been a large increase in field studies designed to measure key dust properties in areas affected by dust aerosol. One such study was the UAE² which hosted a diverse array of instrumentation at the surface and onboard aircraft and satellite platforms. Among the ground-based instruments deployed was the SMART AERI which has much potential for dust remote sensing applications. The AERI’s high spectral resolution allows for exploiting the absorptive differences of dust minerals across prescribed thermal IR “clean” window sub-bands
enabling one to (1) differentiate dust from cirrus and liquid water clouds and (2) retrieve dust IR optical depths. AERI has been used by previous investigators for providing near-continuous profiling of temperature, moisture and atmospheric stability [Felz et al., 2003], for cloud phase determination and retrieval of cloud properties [Turner et al., 2003], for measuring cirrus cloud visible-to-infrared spectral optical depth ratios [Deslover et al., 1999], for evaluating surface aerosol IR forcing [Vogelmann et al., 2003], and for measuring IR emissions from the Saharan air layer [Nalli et al. 2006] using the M-AERI system, a highly accurate sea-going IR spectroradiometer.

This chapter investigates the potential for detecting/retrieving dust aerosol using AERI BT spectra with applications to the UAE$^2$ field experiment. To this end, detailed sensitivity studies of the modeled spectra to key dust and atmospheric parameters are examined which include mineral composition, dust particle size and shape, dust optical depth, dust layer thickness and altitude, and the vertical distributions of water vapor and temperature.

A methodology is given for detecting/separating dust from cloud and retrieving the dust IR optical depths. The physical basis for the approach relies on the complex spectral variability of the IR optical properties for common mineral dust components. Dust detection follows the physical principles of dust and cloud particle absorption across the thermal IR window while the retrieval scheme employs a $\chi^2$ statistical optimization approach in the AERI ‘clean’ sub-bands for determining the dust IR optical depths.

To illustrate the detection/retrieval methodology, AERI data from several daytime UAE$^2$ cases were evaluated to: (1) test the method’s ability to successfully
separate dust from cirrus cloud under mostly cloudy skies and (2) retrieve dust optical depths during typical dust conditions. Estimated errors on the retrieval due to the MCT detector non-linearity were also evaluated. The results were then compared to collocated MPLNET micro-pulse lidar/AERONET sun-photometer measurements respectively.

Although the UAE$^2$ AERI measurements are daytime only, the technique can also be applied to nighttime measurements. To illustrate the potential for nighttime applications, the dust and cloud detection scheme is run using AERI data from the recent NASA African Monsoon Multidisciplinary Analyses (NAMMA) field campaign (2006). The main objectives of NAMMA were to: (1). characterize the evolution and structures of African Easterly Waves, (2). examine the composition and structure of the Saharan dust layer, and (3). study the effects of aerosols on cloud precipitation and their impact on cyclone development.

Potential research areas that may benefit from this work include (1) daytime and nighttime dust hazard mitigation, (2) assessment of the diurnal effects of regional dust radiative forcing, and (3) validating satellite-based dust aerosol remote sensing products.

This chapter is organized as follows. Sensitivity studies to key dust and atmospheric parameters and estimated non-linearity errors are given in section 2. Section 3 discusses the model atmosphere and clear-sky spectra. Section 4 discusses the AERI detection and retrieval methodology. Results of the case studies including the non-linearity error assessment and validation are presented in section 5. Lastly, a summary is given in section 6.
8.2. Sensitivity studies: Pristine and dust-laden atmospheres

Detailed simulations were performed to evaluate the sensitivity of the AERI surface spectra to a number of critical dust and atmospheric parameters over the thermal IR window. The AERI spectral response is first evaluated to changes in precipitable water vapor (PWV) as a potential constraint in the methodology. Following this, the effect of the vertical distribution of water vapor and temperature on AERI clear sky spectra is investigated for cloud and dust-free atmospheres. Next the spectral response to various dust-cloud scene scenarios for developing an AERI detection algorithm is examined. Following this the sensitivity of AERI spectra to key dust parameters including mineral composition, particle size and shape, dust cloud thickness and altitude and dust optical depth are evaluated. Lastly, an estimated radiance error spectrum is constructed corresponding to the non-linear response of the MCT detector which is used for testing the sensitivity of the dust detection/retrieval methodology. Results are presented in section 5.

8.2-A Sensitivity of BT to water vapor and temperature

The thermal IR window is a relatively ‘clean’ spectral region, useful for retrieving aerosol and cloud properties [Turner et al., 2003; Deslover et al., 1999; Pierangelo et al. 2004/2005 and DeSouza-Machado et al. 2006]. With the selection of narrow ‘sub-bands’ [Deslover et al. 1999] inside the window region (see table 8.1 for sub-band locations) the effects of line absorption by water vapor and other trace gases are further minimized, however absorption due to the water vapor continuum remains to be a source
for error in IR remote sensing applications. Moy et al. [2005] conducted an AERI noise analysis which included instrumentation and water vapor uncertainties and showed the water vapor error to clearly dominate the AERI signal in the 10µm region. Accounting for the dynamic effects of water vapor is therefore critical for IR remote sensing applications.

8.2-A-1 Sensitivity of BT to PWV

To test the sensitivity of the clear-sky spectra to changes in total column water vapor amount, radiative transfer simulations using CHARTS were performed using scaled water vapor profiles from averaged MAARCO radiosonde data during September 2004 for a range of PWV consistent with AERONET measurements (~1.5-4 g cm⁻²). Profiles were averaged to represent the atmospheric state during this relatively active dust period. Figure 8.1 shows the sensitivity of the AERI BT spectra at each sub-band (denoted by the markers) to changes in total column water vapor inside the window region. The curves roughly corresponding to the minimum, maximum and average PWV profiles during the UAE² are labeled 1.57, 3.96 and 2.43 g cm⁻² respectively. The plot shows the sensitivity is strongly dependent on wavenumber, where large changes in BT are more evident in the shorter wavenumber regions (i.e. 800-900 cm⁻¹) due to the larger water vapor continuum absorption coefficients [Grant 1990].

Observed changes in retrieved PWV derived from the 0.94µm channel of the AERONET sun-photometer [Bruegge et al., 1994] can vary as much as 15-30% in as little as several hours. Such changes over short timescales could under or overestimate
the effects of dust if not properly accounted for in the detection/retrieval scheme. For example, a 17% increase in PWV from 2.8 to 3.3 g cm\(^{-2}\), will increase the average clear-sky BT spectra by \(~3.9\%\) or 9.3K, around the same order of magnitude as dust [refer to Highwood et al. 2003]. Furthermore, the ±10% uncertainty in the PWV retrieval [Bruegge et al., 1994] for the range of PWV considered in this study, translates to an uncertainty in the clear-sky BT spectra of about ± 4-6K, which according to radiative transfer calculations, corresponds to an IR optical depth (\(\tau\)) of \(~0.075\) at 962 cm\(^{-1}\) or \(~0.15\) at 0.55µm using a Volz extinction coefficient ratio for 1µm sized particles.

To minimize bias, a look up table (LUT) of PWV and BT spectra can be constructed to remove the continuum effects of water vapor (i.e. a clear sky BTD offset) from the spectral measurements. This is further discussed in sections 4 and 5 in connection with AERI dust detection. Lastly, there is also the uncertainty in the vertical distribution of water vapor having the same PWV. Sensitivity of the BT spectra to this and the uncertainty in the vertical temperature distribution are discussed next.

8.2-A-2 Sensitivity of BT to water vapor and temperature vertical distribution

UAE\(^2\) radiosonde data from Abu-Dhabi Airport were analyzed to examine the sensitivity of AERI BT spectra to variable water vapor distributions having the same PWV (2.3±0.01 g cm\(^{-2}\)). Figure 8.2(a) shows the relative humidity (RH) profiles for three days: 8/12, 8/15, and 9/19 where water vapor peaks near the surface. The BT spectra for each case were computed and were differenced with an arbitrarily chosen reference spectrum (\(\Delta BT\)), in this case 8/12 due to its smaller BT’s, to plot the relative
differences. Figure 8.2(b) shows the sensitivity of two days (8/15 and 9/19) with respect to the reference spectrum. Given the selected water vapor distributions, the three days in terms of their mean differences over the window region, were within ~0.005 to 3.37K of each other. For a fourth case analyzed, not shown, the mean spectral difference was about 8K due to a peak shift in water vapor from the surface to a height of nearly 5km, likely attributed to a passing dust storm.

Using the same sounding data, a similar analysis was conducted by evaluating the $BT$ sensitivity to changes in the vertical temperature profiles while holding water vapor constant. Figures 8.2(c) and 8.2(d) show the temperature profiles and relative $BT$ differences between the same three days respectively. The variability of temperature inside the boundary layer is shown in the inset to Fig. 8.2(c), where a maximum difference of ~8°C is evident. For these cases, the relative $BT$ differences were within ~1.18-3.0K of each other. Similarly for the fourth case analyzed, a maximum difference of ~6K was observed likely due to the same dust storm.

8.2-B Sensitivity of $BT$ for cloud/dust discrimination

A sensitivity study of the impact of various dust/cloud scenarios on AERI $BT$ spectra is presented. Sub-bands that are most sensitive to clouds or dust are selected for developing a dust detection algorithm using AERI spectral radiances. A description of the cloud and dust models follows with the specified parameters adequately representing the observed dust/cloud conditions during the UAE$^2$. 
8.2-B-1 Cloud and dust model descriptions

The FDTD IR database [Yang et al., 2005] was used to model the cirrus cloud optical properties where ice crystals were assumed to be hexagonal columns with an effective size of 24µm based on size distribution data collected by the Cloud, Aerosol, and Precipitation Spectrometer (CAPS) aboard the WB-57 aircraft during CRYSTAL-FACE. Using UAE² MPL data to identify cirrus cloud positions, the model cirrus cloud-top heights were located at 7 and 9km with a geometrical thickness of 2 and 1km respectively. Cirrus cloud optical depths were varied from 0.05-3 however the main focus is on the $BT$ effects of thin cirrus ($\tau \leq 1$) having spectra very similar to dust.

Liquid water clouds were assumed to be composed of spherical particles with a monodispersed droplet radius of 32µm. Based on UAE² MPL data, the water cloud-top heights were positioned at 3 and 5 km, with a geometrical thickness of 2 km. Water cloud optical depths were varied from 1-10.

The model dust layers consisted of four mineralogical compositions: Volz (V), quartz (Q), kaolinite (K) and kaolinite carbonate (K/C) mixture. Dust particle shape and size distributions were characterized by the parameters defined in chapter 6. The dust was uniformly distributed along the lowest 2-3 km in all scenarios with dust optical depths ranging from 0.05-1.0.
8.2-B-2 Model simulations

Theoretical calculations were performed for a number of possible dust/cloud scenarios: (1) dust only, (2) cirrus cloud only, (3) dust with cirrus cloud, and (4) dust with liquid water cloud.

The model results for the dust only [Fig. 8.3(a)] and dust with optically thin ($\tau=0.5$) cirrus [Fig. 8.3(b)] cases show a common trend in the $BT$ slope from $\lambda=10\mu m$ (sub-band 11) and $\lambda=11.16\mu m$ (sub-band 7), herein referred to as a $BTD_{10-11}$. For simplicity, the dust model results are given in terms of quartz. Since most minerals with the exception of pure calcium carbonate tend to absorb more at wavelengths near $10\mu m$ than at $11.16\mu m$ (Fig. 6.1), their $BTD_{10-11}$ are negative. This behavior is similar to that of the slope of MODIS $BTD_{8-11}$ vs. $BTD_{11-12}$ which is used by Hansell et al., [2007] for distinguishing dust and cloud.

The results for scenarios 1 (dust only) and 2 (cirrus only) are presented in Fig 8.3(a) where the $BTD$ 10-11 is plotted as a function of optical depth. The error bars on the dust curve (black) depict $BTD$ 10-11 variability due to the uncertainty in PWV (1.8-3 g cm$^{-2}$). As PWV increases, so too does the $BTD$ 10-11, which may cause clouds to be overestimated. Likewise, the $BTD$ slope for water vapor when PWV $\leq 1.84$ g cm$^{-2}$ is similar to mineral dust which may cause dust to be overestimated. For optical depths greater than 0.12, dust has a negative $BTD$ 10-11, whereas when dust is optically thin ($\tau \leq 0.12$), the effects of water vapor are dominate and the $BTD$ 10-11 becomes positive. The cirrus only case on the other hand displays a positive $BTD$ 10-11 since ice water absorbs more strongly with longer wavelengths.
The results for the dust and cloud mixed cases are presented in Fig. 8.3(b). The two dust/cirrus cases (black curves) describe the effect on $BTD_{10-11}$ when the cloud optical depth (COD) increases from 0.5 to 3. When cloud and dust are optically thin, the effects of water vapor dominate and the $BTD_{10-11}$ is positive. As dust optical depth increases under a thin cirrus cloud, the $BTD_{10-11}$ becomes increasingly negative (lower black). When cirrus cloud optical depth increases (upper black) or as the height of the cirrus cloud is reduced (e.g. from 9 to 7 km), the $BTD_{10-11}$ becomes more positive which makes it more difficult to detect dust even when dust optical depths are high. Fig. 8.3(b) for example shows that under an optically thin cirrus cloud, dust would require a minimum optical depth threshold ($\tau = 0.12$) to be detected. If the COD were to increase, the threshold for detecting dust would be well over 0.6.

Fig. 8.3(b) also shows an example of dust and water cloud (gray curve) with a COD of 10. The $BTD_{10-11}$ is positive for all dust optical depths, i.e. the dust will not be detected. This same result occurred when the layers of water cloud and dust were reversed (e.g. in a transported dust plume). Not shown is the case for a thin water cloud ($\tau = 1$), where $BTD_{10-11}$ is negative for dust optical depths greater than 0.05. Lastly, the sensitivity to cloud height was evaluated by elevating the water cloud to 5-6 km. No significant change in the $BTD_{10-11}$ was observed; hence the main variable controlling the $BTD_{10-11}$ behavior of water cloud with dust is the COD. A summary of the results presented is given in table 8.2.
8.2-C Sensitivity of $BT$ to dust parameters

8.2-C-1 Particle size and shape

To assess the sensitivity of the AERI $BT$ spectra to dust particle size and shape, spheres (S), oblate spheroids (OS) and compact hexagons (CH) are evaluated each having a lognormal size distribution with an effective radius that is increased from 0.5-2.0 µm in steps of 0.5µm. Dust composition is prescribed using the Volz [1973] refractive indices. For each size step, the dust optical depths are adjusted to 0.05, 0.4 and 1.0 to assess optically thin and thick dust layers. The results shown in Figs. 8.4(a)-8.4(b) are expressed as a function of both effective size and shape and are plotted in terms of the $BTD$ slope across the upper ($BTD$ 17-13) and lower ($BTD$ 11-1) sub-bands respectively. For presentation purposes, the spheres and compact hexagons are plotted for a dust optical depth of 0.4.

Comparing Fig. 8.4(a) with 8.4(b) shows that particle shape sensitivity appears to be greater across the upper sub-bands (USB), while that for particle size is greater across the lower sub-bands (LSB). For example, the $BTD$ between the sphere and compact hexagon in fig. 8.4(a) varies from 4-8K over the given particle size range whereas in fig. 8.4(b), the same $BTD$ varies from only 0.3-2K. Likewise the $BTD$ between particle sizes, although much greater for sub-micron particles, particularly the CH, varies from less than 0.5K across the USB, to between 0.5-1.5K across the LSB.

Not shown is the $BTD$ sensitivity to dust optical depth. It was found that when optical depth increases, the particle size sensitivity also increased particularly for
effective radii less than 1µm. This was most evident for compact hexagons. For example; the $BTD$ for a compact hexagon between 0.5 and 1µm with unit optical depth ($\tau=1$) is ~2.8K and 5.8K for the USB and LSB regions respectively. Beyond 1µm however, the BTD quickly falls off to ~0.38K and 0.80K as the size approaches 2µm. When the optical depth is reduced to 0.05, the corresponding $BTD$’s decrease by a factor of 1.2 and 5X for the USB and LSB respectively. It appears first off that that the potential for separating particle size in the thermal IR window diminishes as the effective radius increases (i.e. particles become coarser) and the optical depth decreases, however there appears to be shape dependence where angular particles like the compact hexagon display greater sensitivity compared with spheres and oblate spheroids.

An inspection of the $BTD$ between particle shapes reveals similar differences, which like size, is greater for particles with effective radii less than 1µm. This is evident in the USB region for the submicron particles shown in Fig. 8.4(a). Similar to particle size, the shape sensitivity also decreases as optical depth becomes smaller.

Overall, it appears that both size and shape sensitivity of the $BT$ spectra are strongly dependent on dust optical depth and particle size. Relative to each other, both size and shape parameters are significant inside the thermal IR window, although for the geometries and sizes considered, the absolute values of the mean $BTD$’s (i.e. the average $BT$ over each dust parameter) appear to be greater for particle shape. Refer to table 8.3 for a summary of the average size and shape $BTD$’s.
8.2-C-2 Dust optical depth

The theoretical basis for the optical depth retrieval method lays in the sub-band slope dependence of dust optical depth, particularly from 1100-1200 cm\(^{-1}\) which covers sub-bands 13-17. This spectral region is emphasized due to the higher \(BT\) sensitivity to PWV from 800-1000 cm\(^{-1}\) (Fig. 8.1) and also the strong ozone absorption at 1086 cm\(^{-1}\).

Two cases are presented in Fig 8.5 to illustrate the slope sensitivity from sub-bands 13 to 17 employing two dust microphysical models: (1) Volz hexagonal plates and (2) kaolinite/carbonate mixture of oblate spheroids. The curves are dust \(\Delta BT\) spectra where \(\Delta BT\) is defined as:

\[
\Delta BT = BT_{dust} - BT_{clear}
\]  

\(BT_{dust}\) and \(BT_{clear}\) represent the modeled or observed dust and clear sky spectra respectively across the AERI sub-bands. To a first approximation, \(\Delta BT\) over the window region can be used as a rough qualitative indicator of the dust IR forcing, since this is where most dust absorption occurs. The markers from left to right denote the sub-band locations 1-17. Dust optical depths are varied from 0.05-1.0 in increasing order from bottom to top. For both cases, AERI observed spectra (broken gray curves) are shown for comparison. Note the reasonable fit between sub-bands 13-17; however larger differences occur particularly in the 800-900 cm\(^{-1}\) region which can be attributed to water vapor and uncertainty in dust composition. Similar results were also found using other dust models.

Both figures show a positive increase in the sub-band slope as optical depth
becomes larger due to enhanced particle absorption at these wavelengths. Also note the sharp spectral folding [Fig. 8.5(a)] and the much smoother spectral roll-off [Fig. 8.5(b)] near 1100 cm$^{-1}$ for each dust model respectively. This spectral behavior follows the imaginary terms of the refractive indices (Fig. 6.1), which shows that most minerals exhibit peak absorption around 1000-1100 cm$^{-1}$ with smaller secondary peaks scattered throughout the window region. As the absorption increases, so too does the spectral peak which causes the slope to further increase.

8.2-C-3 Dust altitude

To assess the sensitivity of the AERI $BT$ spectra to dust altitude, a homogeneous dust layer near the surface with a geometrical thickness of 1km is used and its attitude is progressively increased in 1 km increments until a layer height of 5km is reached. This is analogous to dust being lifted and transported from its source. The dust model consists of spherical particles with a lognormal size distribution ($R_{\text{eff}} = 2.0\mu\text{m}; \sigma_{\text{gv}} = 2.0$) and mineral composition defined by the Volz refractive indices. Results corresponding to IR optical depths of 0.05 and 0.4 for layer heights of 1, 3 and 5 km are presented in figure 8.6(a)-8.6(c) respectively. As the dust layer’s altitude increases, the magnitude of the resulting $BT$ spectrum decreases and is a function of the dust optical depth. For example, fig. 8.6(a) shows that optically thin dust ($\tau=0.05$) with an altitude change of 5 km, results in a relatively small $BT$ decrease of $\sim$2K. This means that AERI is not very sensitive to the altitude changes of light dust. Increasing the optical depth to 0.4 [figs. 8.6(b)] however for the same altitude change, results in much larger $BT$ decrease of 10K because
of enhanced dust absorption. This has significant implications for detecting/retrieving transported dust plumes such as in the case of dust in the Saharan Air Layer.

8.2-C-4 Dust thickness

The effects of dust layer thickness on AERI surface spectra are examined by successively incrementing the number of dust layers used during each simulation. Beginning with the first dust layer near the surface, additional layers are added until the geometrical thickness of the dust cloud reaches 5 km. This range in dust thickness is consistent with observations during the UAE² where dust top heights were often capped by a strong inversion around 5 km. The dust properties are uniform and homogeneous across each layer and are based on the Volz refractive indices. Dust particle shape and size are characterized by spheres with a lognormal size distribution of $R_{\text{eff}}=2\mu m$ and $\sigma_{gv}=2$. A vertical distribution of dust optical depths from the UAE² level 1.5a MPLNET [Welton et al. 2001] dataset on September 22, 2004 was used [fig. 8.6(c)]. The black and gray curves denote the MPL retrieved optical depth profiles at 0.55 and 10µm respectively where the latter was obtained by scaling the visible optical depths using the Volz extinction coefficient ratio between 0.55 and 10µm. The circles on the gray curve represent the scaled IR optical depths corresponding to the centers of each dust layer. The results are given in figure 8.6(d). Note that as more layers are added, the $BT$ differences decrease. The largest difference occurs between the first and second layers (~8-10K) where the optical depths given by fig. 8.6(c) are much larger. Beyond a thickness of 2km, the $BT$ sensitivity decreases due to the reduced optical depths of the
layers where differences are on the order of ~2-5K.

**8.2-C-5 Mineral composition**

The sensitivity of the AERI BT spectra to mineral composition is investigated for the thermal IR window. For comparison, the spectral region 2500-2700 cm\(^{-1}\) (3.7-4µm) was also examined. For simplicity it is assumed that dust particles are spherical with effective size \(R_{\text{eff}} = 2.0\mu\text{m}\) and optical depth (\(\tau\)) of 0.20. The particles’ refractive indices are varied to coincide with pure quartz, an internal mixture of quartz and 10% hematite, illite, and calcium carbonate. Figure 8.7(a) shows the BT spectrum (\(\Delta BT\)) for each composition for the thermal IR window after removing the clear-sky contributions. Figure 8.7(b) shows the BT spectrum for the same minerals for the spectral region 2500-2700 cm\(^{-1}\). Because sensitivity in the thermal IR window is exploited for retrieving optical depth, the focus is on the latter spectral region which shows a clear separation of the BT spectra around 2500-2540 cm\(^{-1}\) (~10K) for potentially identifying dust mineral components. This spectral region could be exploited and used in a complementary scheme following the dust optical depth retrieval. Note there is little sensitivity between quartz and the quartz/hematite internal mixture.

**8.2-C-5 AERI MCT detector non-linearity**

The estimated radiance errors due to the MCT detector non-linearity are based on the prior works of Turner et al. [2004], and Knuteson et al. [2004]. Turner et al. [2004] studied the impact of the AERI MCT detector non-linearity on observed radiance spectra,
using the corrections discussed in Knuteson et al. [2004]. The study found that the detector error is largest in clear-sky conditions for radiometrically cold scenes where the non-linear response in the detector’s electronics causes the observed radiances to be underestimated. The effect is most evident in the window region, where the radiance residuals are in the approximate range of ~0.3–0.75 RU where 1 RU = mW m⁻² sr⁻¹ cm⁻¹.

The radiance residuals in Turner et al. [2004] and Knuteson et al. [2004] were used to construct an estimated radiance correction spectrum constructed over the thermal IR window with random noise added to account for the underestimated signal. Figure 8.8(a) shows the estimated non-linearity corrections. Although the corrections are not exact (only during AERI calibration can these be uniquely determined) these will provide a sense as to the potential effect they might have on the detection/retrieval methodology. Details of the results are given in section (6).

It is noted that while the dissertation was being prepared, the SMART AERI underwent calibration at the Southern Great Plains (SGP) central ARM site in Lamont, Oklahoma. The non-linearity will be characterized through a cross comparison with the SGP AERI and applying the methods discussed in Knuteson et al. [2004].

8.3 Model atmosphere and clear-sky spectra

To characterize the dynamical state of the column atmosphere, the retrieved temperature and moisture profiles from the UW-SSEC AERIPLUS physical retrieval algorithm [Feltz et al. 2003] for the PBL are combined with regional sounding data up to
a height of 18 km. The AERIPLUS retrievals using only AERI data are limited below 2.5-3 km since most of the IR signal detected at the surface comes from the lower atmospheric emissions [Feltz et al. 2003]. Because the AERI instrument measures the near-surface emissions very precisely, the AERIPLUS retrieved parameters well represent the atmospheric state of the PBL.

Figures 8.8(b)-8.8(c) show the AERIPLUS retrieved time-height temperature and water vapor mixing ratio (WVMR) profiles for September 22, 2004 during UAE$^2$ which includes the estimated correction for the detector non-linearity [fig. 8.8(a)]. The same profiles were also run without the correction and the residuals were computed for each parameter. The temperature bias was on the order of 0.1°C while the WVMR bias was generally less than 0.5 g kg$^{-1}$ except for at the surface which was found to be $\sim$2 g kg$^{-1}$.

8.4 Dust Detection and Retrieval Methodologies

8.4-A Detection methodology

Following the discussion in section 2, the AERI cloud/dust detection scheme employs the $BTD_{10-11}$ for separating cloud and dust using a dynamic threshold ($BTD_{10-11}=0$) where positive/negative $BTD_{10-11}$ is labeled as cloud/dust respectively. To compensate for water vapor in the algorithm and the potential for misclassifying dust, a $BTD_{10-11}$ clear-sky offset is introduced which is a function of PWV.

The $BTD$ offset is determined by comparing coincident measurements of AERONET derived PWV with AERI to a LUT which is calculated using the data from
fig. (8.1) and then interpolated using a PWV spacing of 0.1 g cm\(^{-2}\). The offset is then subtracted from the AERI BTD10-11 so as to remove the effects of water vapor. For the cases examined, the PWV uncertainty (total column and vertical variability) in water vapor can produce clear sky BTD errors in the range of \(~0.5-6\)K respectively.

The effects of column water vapor uncertainty are shown in Figure 8.3(c) which shows the BTD10-11 clear-sky offset as a function of PWV. The horizontal error bars denote the \(\pm 10\%\) PWV uncertainty in the AERONET measurements, while the vertical error bars represent the corresponding BTD uncertainty. For atmospheres with PWV \(\leq 1.84\) g cm\(^{-2}\), the offsets are negative meaning the BTD slope for water vapor and dust is similar (refer to fig. 8.1). For these cases, the offset is added. For values of PWV \(> 1.84\) g cm\(^{-2}\), the BTD slope of water vapor and dust is of opposite sign and the offset is subtracted.

Fig. 8.3(d) shows the impact of PWV uncertainty on the AERI detection scheme for a mixed case of dust and cirrus assuming a PWV of 2.4 \(\pm 0.1\)g cm\(^{-2}\). The top solid curve represents the BTD10-11 for dust and cirrus before the offsets are applied. Note that for each dust optical depth, the BTD10-11 lies above the zero-point threshold; hence dust will not be detected. For the specified range of PWV, adding the offsets from Fig. 8.3(c) yields a range of corrected BTD10-11 shown by the broken curves. The offset effectively reduces the magnitude of BTD10-11, however depending on the error in the PWV measurement; the optical depth threshold for dust detection can vary. For example, if the retrieved PWV is on the high end (+10\%), dust may be overestimated where the BTD10-11 for optical depths greater than 0.15 will be labeled as dust. Conversely, if
PWV is on the low end (−10%), dust may be underestimated.

To further compensate for uncertainties in PWV and detecting cloud and dust near the $BTD_{10-11}$ threshold, the algorithm accounts for high frequency temporal instabilities in the $BT$ spectra, an approach similar to the AERONET cloud screening procedure [Smirnov et al., 2000]. This is accomplished by evaluating the change in the $BTD$ slope during each AERI measurement where rapid changes are indicative of cloud transits.

Lastly in the event that cloud is misclassified as dust, the magnitude of the $BT$ sub-band spectrum averaged over the window domain is checked to see if it exceeds a prescribed threshold which is taken to be 270K based on radiative transfer simulations. The algorithm labels the event as cloud when $BT < 270K$.

8.4-B Optical depth retrieval methodology

The retrieval methodology relies on a statistical optimization approach whereby a search is conducted for the maximum likelihood that $\Delta BT$ (observed) takes on the functional form of $\Delta BT$ (calculated) evaluated over the prescribed sub-bands. The retrieval algorithm is given by minimizing a residual sum, $\chi^2$, defined as follows:

$$\chi^2 = \sum_{i=a}^{b} [\ln(\Delta BT_{\text{calc}}^i(\tau, a_e, T, \nu)) - \ln(\Delta BT_{\text{aeri}}^i(T, \nu))]^2$$

(8.2)

where the summation is performed over sub-bands a to b and $\Delta BT$ is a function of optical depth ($\tau$), effective radius ($a_e$), temperature ($T$), and wavenumber ($\nu$). The retrieval is
repeated for each cloud-free AERI measurement following the dust/cloud detection scheme to yield a best fit IR optical depth at 962 cm\(^{-1}\). Other significant works utilizing this and similar approaches using thermal IR observations include Rathke et al. [2000], and Pierangelo et al. [2004].

The retrieval methodology employs a LUT of modeled dust \(\Delta BT\) for the dust microphysical models presented in chapter 6. To this end, a LUT is constructed for each critical dust parameter including 5 mineral compositions: quartz, kaolinite, kaolinite mixed with hematite, kaolinite mixed with carbonate and the Volz dust model and 3 particle shapes: compact hexagon, oblate spheroid and sphere. This allows for comparing the performance of each dust model in terms of a best fit with the observed \(BT\) data. In the current scheme, particle size is not retrieved due to the relatively small variability in the window region (chapter 6). The IR optical depths of the modeled dust spectra were interpolated over a range extending from \(\tau=0.05-1.0\) in steps of 0.10.

The atmospheric profiles used in the retrieval scheme rely on two model clear sky approaches. One which uses the AERIPLUS derived state parameters (section 8.4) and the other which employs a fixed profile representing the averaged atmospheric state. Since the model clear sky in the former approach is continuously updated for each AERI observation, PWV is not considered a free-parameter in the retrieval, and is therefore not included in the LUT. In the second approach however, PWV is considered a free-parameter and PWV is added to the LUT. The PWV resolution in the LUT is somewhat coarse (~13K), however is shown to demonstrate the technique.

The retrieval scheme also accounts for local temporal changes in the ambient
surface temperatures by subtracting a \( BT \) thermal offset from each sub-band where a table of thermal offsets is constructed by differencing the clear-sky spectrum for a range of local times ranging from 0600-1700 UTC with a reference ‘cold’ spectrum at 0600 UTC.

8.5 UAE\(^2\) case studies and validation

The detection and retrieval results are divided into two sections. The first coincides with the results obtained using the AERIPLUS clear-sky spectra with and without the estimated detector non-linearity corrections. The second section gives the results using clear-sky spectra with a LUT which includes an expanded PWV.

8.5.1 AERIPLUS clear-sky spectra with non-linearity correction

Two UAE\(^2\) cases were selected to illustrate the detection/retrieval methodology. September 22 and September 30, 2004 were chosen based on the prevalent dust and cirrus cloud conditions for the two days respectively. The strength of the dust detection approach is first demonstrated. Figure 8.9(a) shows the SMART MPLNET micro-pulse lidar (MPL) normalized relative backscatter (NRB) signal for September 30, 2004. Note the persistent cirrus cloud coverage from 0600-1600UTC. The AERI detection results are denoted by the red and white ‘x’ labels at the bottom of the MPL plot where the colors indicate AERI data with (red) and without (white) the estimated non-linearity corrections applied. The top and bottom rows of ‘x’ labels represent cloud or no cloud
conditions respectively, where no cloud denotes typical ‘background’ dust conditions. Consistent with the MPL, the AERI detection algorithm detected mostly cirrus from 0930-1415 UTC with occasional breaks in clouds noted at 0950UTC, 1300-1330UTC and 1400UTC. Several points around 1200UTC were also identified as dust in the presence of thick cirrus.

Next the dust retrieval technique is tested using cloud-screened AERI radiances from September 22 during the 4-hour period 0600-1000UTC. This day was most likely associated with a dust episode called a haboob (refer to Miller et al., [2007]) which can be triggered from the outflow of intense thunderstorm cells which often develop over the Al Hajar mountain range bordering UAE and Oman. The AERI fore optics were covered at the onset of the dust episode, a precautionary measure, to preclude the possibility of damage to the instrument. Though the peak dust loads were missed, AERI still captured the ‘normal background’ conditions leading up to the dust event.

The retrieval is performed employing the channel 1 AERI dataset with and without the estimated MCT non-linearity corrections. Figure 8.9(b) shows the AERI retrieved optical depths at 962 cm\(^{-1}\) (10.4\(\mu\)m). The black curve represents the data without corrections while the light and dark gray curves correspond to the corrected data. The AERI optical depths are scaled to 0.55\(\mu\)m for comparison with AERONET (red) using the Volz visible-to-IR extinction coefficient ratio \(\beta_{0.55\mu m} / \beta_{10.4\mu m}\) which varies from 2-3 for particle sizes in the range of 0.75-1.5\(\mu\)m. The error bars depict the variability in the retrieved optical depths due to the uncertainty in particle sizing. The scatter plot in Fig 8.9(c) shows the linear trend between the scaled AERI and the
AERONET optical depths with a correlation coefficient of 0.78, indicating the retrieval is able to capture the variability in the dust load with some success. If the MCT detector non-linearity is not accounted for, the estimated errors in the retrieved optical depths range from ~6-12% over the 4-hour period.

8.5.2 Clear-sky spectra without the non-linearity correction

To assess the performance of the detection/retrieval methodology using clear-sky spectra without the non-linearity correction and a LUT including the extended PWV, several cases were selected: September 21-23 and September 30.

Validation of the dust/cloud detection results is accomplished using collocated and coincident MPL backscatter images at SMART. Qualitative comparisons of the detection results are made using image plots from the MPLNET website (http://mplnet.gsfc.nasa.gov/). To further demonstrate the effectiveness of the detection approach and to illustrate the potential for nighttime applications, the MPL dataset from NAMMA is used. Lastly, validation of the optical depth retrieval method is accomplished using the collocated and coincident UAE\(^2\) AERONET sun-photometer dataset at SMART.

8.5.2. a. Dust/cloud detection results and validation

Qualitative comparisons were made using the UAE\(^2\) MPL NRB image plots for
the dust/cloud days considered in this study with the exception of September 23, due to the unavailability of the image plot and September 30 which was addressed in section 8.5.1. The AERI detection results are presented in Fig. 8.10 which defines an index of 1 for cloud and 0 for no cloud, where the no cloud flag represents typical ‘background’ dust conditions. On September 21, the AERI detection algorithm detected clouds from about 1100-1300UTC [Fig. 8.10(a)] which are shown in the MPL plot (white circle – Fig. 8.10b) during which time low clouds were identified around 4-5 km. The remaining AERI observations were dominated by the ‘background’ dust aerosol, again evident in the MPL plot shown by the thick dust layers reaching ~5 km. It is noted that several AERI observations flagged as dust are likely to be cloud (~1200-1300UTC) which may be due to the effects of reduced water vapor and/or optically thin cloud overlying thicker dust layers (section 2). The MPL plot shows a very thin layer of low cloud at 5 km from about 1230-1500UTC which may be causing the misclassification of dust. September 22, was classified as all dust just before the dust front/convective storm reached SMART around 1020UTC [Fig. 8.10(c)]. The MPL plot shows the dust only conditions (white circle – Fig. 8.10d) leading up to the storm.

To demonstrate nighttime detection, the MPLNET data from SMART during the recent NAMMA field campaign was used. SMART was deployed at Sal, on the Cape Verde Islands just off the coast of West Africa in September 2006 during which notable dust cycles and clouds were observed over 24 hour periods. Fig. 8.11(a) presents the MPL NRB profile from September 7, which shows significant dust loading in the early morning (0000-0700UTC) and evening hours (1900-2400UTC) with overlying persistent
low clouds during the day. From about 1200-1400UTC, the MPL was shaded to prevent
damage to the instrument at solar noon. Also, the white asterisks denote the cloud
positions. The AERI detected clouds and dust employing the BTD 10-11 technique are
shown in Fig. 8.11(b). It is noted that the BTD 10-11 clear-sky offset was not applied
since the AERONET sun-photometer at Cape Verde was not operational during
NAMMA. The ‘x’ and diamond marks represent the AERI observations and the
superimposed MPL cloud locations, which are shown for comparison. The negative BTD
10-11 is flagged as dust while the positive BTD 10-11 is flagged as cloud. The AERI
detected clouds and dust compare favorably with those detected by the MPL, with AERI
detecting approximately 83% of the clouds observed by the MPL. After applying the BT
threshold criteria in the detection scheme to the AERI data, the comparison with the MPL
increases to about 85%. Where AERI and the MPL do not agree is most likely due to not
accounting for the effects of water vapor. These results indicate that this approach can be
used to separate dust from cloud during daytime and nighttime conditions with some
confidence. Some improvement in the performance of the detection algorithm is
expected after application of the BTD 10-11 clear-sky offset. Although more cases are
needed to have a complete validation of the current approach, it offers promising
potential for daytime and nighttime detection applications. A more robust methodology
for handling water vapor and mixed dust/cloud scenarios is under investigation.
8.5.2. b. Dust optical depth retrieval results and validation

Following the dust and cloud separation, the algorithm retrieves the dust IR optical depths at 962 cm$^{-1}$ (10.4$\mu$m) using only cloud-free observations. Validation is accomplished using AERONET retrieved visible optical depths at 0.55$\mu$m. To compare optical depths, visible-to-IR extinction coefficient ratios ($\beta_{0.55\mu m}/\beta_{10.4\mu m}$) for three dust models were calculated over a particle size range of 0.1-2.0 $\mu$m.

Both AERI and AERONET retrieved optical depths for each of the four cases are presented in Fig. 8.12. The AERONET data is shown in red with error bars denoting the uncertainty in the optical depth retrieval (0.02) due to calibration uncertainties in the field instruments (see AERONET web-site – [Holben et al., 1998]). The AERI retrieved optical depths shown in dark gray/black are plotted as the average optical depth over the particle size range retrieved by AERONET (V2, Level 2). The corresponding error bars depict the variability in optical depth due to particle sizing. For each case the effects of the retrieval on several mineral dust compositions and particle shape were examined. The effects of the retrieval on PWV using the cirrus case were also examined to demonstrate the sensitivity to water vapor.

The first case on 9/21/04 shown in Fig. 8.12(a) gives the AERI retrieved optical depths, with error bars, using the Volz model for two particle shapes: oblate spheroid (OS) and compact hexagon (CH). Differences between the two dust models were found to be between 17-20% with optical depths from the compact hexagon to be higher, although for this case, the oblate spheroids seem to yield a better fit with AERONET.
This is probably due to accounting for larger dust particles in the size distribution and also to using the in-situ based shape distributions. This could also be due to 1. Dust shapes were not hexagonal, and/or 2. Dust shape was described by a more complicated distribution of particle geometries.

For both instruments the optical depths reveal higher dust loadings in the early morning hours and then fall-off substantially around 0830UTC with AERI optical depths decreasing by about 40%, almost 2X that over AERONET. This decrease is thought to be due to the large thermal sensitivity that AERI has to the ambient environment, which includes (1) changes in the total dust load, and (2) changes in the total water vapor which for this period was found to have decreased by ~10%. Overall the algorithm’s cloud detection routine performed quite well, being able to account for most of the clouds observed by the MPL between 1130-1300UTC. One exception was noted around 1130UTC, where the AERI optical depths were found to have increased by over a factor of 2. This was due to the detection algorithm flagging cloud as dust. If the misclassified spectra is removed, the results from both microphysical models correlate well with AERONET with an averaged correlation coefficient, r=0.920. The relative errors of AERI versus AERONET are given in table 8.4. Generally, the errors are within 20-30% which is fairly reasonable given the uncertainties in each dust microphysical model and in the water vapor amounts.

The results to the second dust case, 9/22/04 are given in Fig. 8.12(b). Three dust models are compared to AERONET: pure quartz, kaolinite internally mixed with calcium carbonate and Volz, all consisting of oblate spheroids. The sensitivity of the retrieved
optical depths to changes in the respective dust composition is assessed. The dust front approached the SMART site about 1000UTC reaching peak intensity around 1020UTC. Near 1000UTC, the AERI fore optics were covered and data logging was terminated.

The retrieved results correlate well with AERONET, on average around 0.92 with optical depth differences between minerals reaching between 28-30%. Both instruments track higher optical depths earlier in the morning and then reach a minimum between 0700-0800 UTC. Optical depths begin to gradually increase just before the onset of the dust episode. The relative errors of each dust model with AERONET are given in table 8.4. Here the errors are generally within 10% for the Volz and quartz microphysical models and increasing to about 20% for kaolinite and kaolinite mixed with carbonate. The best comparison is again achieved using the Volz (OS) model. Since the Volz dataset represents a mixture of common dust minerals (mostly the clays: kaolinite and illite along with traces of quartz) based on in-situ measurements, this comparison comes as no surprise. Perhaps this mixture is an adequate representation of the mineral composition around the SMART site.

The results to the third dust case, 9/23/04 are given in Fig. 8.12(c). The AERI results correlate well with AERONET from about 0530-0930 UTC (~0.75), however, from 0930-1030UTC, the AERI optical depths are shown to decrease while those from AERONET remain essentially unchanged (total correlation is about 0.31). This is probably due to the increased sensitivity of the AERI instrument to the local dust loading. From 1030-1130UTC, AERI and AERONET agree quite well, both showing an increase in dust optical depth which is consistent with the observations by on-site personal of an
approaching dust episode. The relative errors of each dust model with AERONET are given in table 8.4. Here the errors are reasonable. For the cases shown (Volz (OS) and quartz (OS)), the relative errors are nearly within 20%. For pure and internally mixed kaolinite, and Volz (CH), the errors increase about 3X, suggesting as with the previous cases, that 1. Compact hexagons may not be accurate descriptions of regional dust particle shapes; 2. The local dust probably consists of a wide distribution of irregular shapes and; 3. There are likely other minerals that are not accounted for in the model. The best comparisons are achieved using the Volz and quartz (OS) dust models.

The final day, 9/30/04, is a cirrus case. Shown in fig. 8.9(a), the MPL at SMART observed the cloud base heights near 7km with variable cloud thickness ranging from 1-2km. The cloud coverage was fairly persistent from about 0800-1600UTC with AERI operating from around 0930-1415UTC. Dust optical depths measured by AERONET (τ = 0.18) were a minimum for the entire study period. Fig. 8.12(d) shows the AERI retrieved results employing the quartz (OS) dust model for three values of PWV: 1.10, 1.35 and 1.57 g cm\(^{-2}\). The sensitivity of the downwelling BT to a change in PWV over this range is ~10K. The dust/cloud detection routine flagged 90% of the AERI observations as being due to cirrus cloud. Fig. 8.12(d) shows a morning and afternoon retrieval at ~0945 and 1315 UTC respectively. The afternoon retrieval was a case where cirrus was classified as dust. This is also evident in the MPL profile, fig. 8.9(a). This is likely due to optically thinner cirrus at lower altitudes overlying a dust layer with minimal optical thickness which yields a negative BTD 10-11 (section 2). Nevertheless, the morning retrieval compares quite favorably with AERONET, particularly with the
quartz (OS) dust model when PWV was set to 1.35 g cm\(^{-2}\). Relative errors (table 8.4) are within 10%. The PWV was later perturbed to 1.10 and 1.57 g cm\(^{-2}\) which caused the retrieved optical depths to increase/decrease by over 100% and 43% respectively.

Overall, given the uncertainties in dust model parameters and atmospheric water vapor, the comparisons with AERONET are favorable, generally within 30%. Similarly, comparisons of the retrieved optical depths from the NASA-GSFC Deep Blue algorithm with those from two desert AERONET sites (Nigeria and Saudi Arabia) also indicated good agreement, within 30% [Hsu et al., 2004], lending confidence to the current retrieval approach.

8.6 Summary

Detailed sensitivity studies to key dust and atmospheric parameters in the thermal IR window were conducted to examine the potential for developing methodologies to detect and retrieve dust aerosol using ground-based AERI spectra during the UAE\(^2\). To this end, dust microphysical models characterizing mineral composition, particle shape, and size were constructed by combining \textit{in-situ} data from the UAE\(^2\) with that from prior field studies. Using FDTD and T-matrix light scattering codes, the dust single scattering properties were calculated and were found to be most sensitive to the refractive index of the component minerals, consistent with previous studies. Comprehensive studies of the uncertainties in water vapor measurements and its vertical variability were critically analyzed for its impact on the AERI applications. As a result, a dynamic \textit{BTD} clear-sky offset was formulated for dust detection and AERIPLUS retrieved temperature and
relative humidity profiles were employed to characterize the dynamical state of the lower troposphere for dust retrieval. The vertical variability in temperature and the large surface temperature gradients observed during UAE2 also necessitated the need for a thermal offset in the retrieval methodology.

Detailed sensitivity studies of critical dust parameters revealed significant affects on AERI BT spectra. Unique to AERI detection was the BTD10-11 dynamic threshold which exploits the differences in the absorptive properties of minerals and liquid/ice water for separating dust from cloud. Particle size and shape effects on the AERI spectra using compact hexagons, oblate spheroids and spheres for sizes in the range of 0.5-2µm, exhibited unique spectral dependence over the sub-bands and were found to increase with dust optical depth. Independent of shape, the potential strength for particle size separation in the window region appears more likely for particles less than 1µm with decreasing sensitivity towards 2µm. For the cases analyzed, compact hexagons showed the greatest sensitivity which emphasizes the significance of particle asphericity in dust remote sensing applications. AERI sensitivity to dust optical depth displays strong sub-band slope dependence particularly around 1100-1200 cm⁻¹ for most minerals including quartz and the kaolinitecarbonate dust mixture. AERI sensitivity to selected silicates, clays and carbonates is evident around 2500-2540 cm⁻¹ where mineral separation is on the order of ~10K between each mineral class. AERI sensitivity to dust layer altitude and thickness was investigated and is dependent on the vertical structure of dust optical depth. The spectra in elevated dust layers such as those found in transported dust plumes was found to be reduced in magnitude between 2-10K for the range of optical depths 0.05-0.4.
The largest spectral change to dust thickness occurs in the first two layers (~8-10K) and is reduced in the remaining layers with relative differences amounting to ~2-5K.

Based on the sensitivity results, a combined detection/retrieval methodology was formulated and tested using several daytime dust and cirrus cases from the UAE\textsuperscript{2}. Two clear-sky approaches were evaluated including use of (1) AERONET clear-sky profiles and (2) profiles with an extended PWV in the LUT. Consideration to the MCT detector non-linearity was also addressed to examine the methodology’s sensitivity to such errors.

By comparing the AERIPLUS detection/retrieval results with collocated AERONET sun-photometer/ MPLNET micro-pulse lidar measurements, it was shown that the present scheme can be used to separate dust from cloud and retrieve dust IR optical depths during daytime conditions with some confidence. Adding the estimated MCT detector non-linearity corrections to the AERI data did not affect the detection results, although did increase the magnitude of the retrieved optical depths by ~6-12%. The potential for nighttime applications was also demonstrated where AERI detected at least 85% of clouds observed by the MPL during NAMMA.

Employing a LUT with extended PWV, the potential for using the AERI $BT$ spectra to retrieve dust IR optical depths was investigated. Comparisons of the AERI retrieved optical depths with those from AERONET indicate reasonable agreement generally within 30%. Both the Volz and quartz microphysical models containing oblate spheroid shape distributions yielded the best results for all cases examined. It is suspected that because the size distribution of the compact hexagon did not contain particles larger than 2.5µm (coarse-mode particles) and was only composed of a single
shape, caused the retrieved optical depths to be biased high compared with the oblate spheroid. The relative difference between shapes was on the order of 17-20%. The differences in retrieved optical depths between the selected mineral compositions were on the order of 28-30%, which reflects the varied nature of the dust minerals in their complex refractive indices and single scattering properties. The effects of water vapor if not properly accounted for can dramatically impact the retrieval by as much as 100%.

The current approach offers a variety of promising applications for AERI including: (1) improved tracking and monitoring of regional dust episodes to complement satellite-based detection methods, (2) providing the dust optical parameters needed for estimating regional long wave surface radiative forcing, (3) providing a means for identifying mineral composition, and (4) the ability to provide for the detection and retrieval of nighttime dust episodes.
Table 8.1: AERI sub-bands

<table>
<thead>
<tr>
<th>Sub-band index</th>
<th>$\nu$(cm$^{-1}$)</th>
<th>$\lambda$(µm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>809.0-812.9</td>
<td>12.30</td>
</tr>
<tr>
<td>2</td>
<td>815.3-824.5</td>
<td>12.10</td>
</tr>
<tr>
<td>3</td>
<td>828.3-834.6</td>
<td>12.00</td>
</tr>
<tr>
<td>4</td>
<td>842.8-848.1</td>
<td>11.80</td>
</tr>
<tr>
<td>5</td>
<td>860.1-864.0</td>
<td>11.60</td>
</tr>
<tr>
<td>6</td>
<td>872.2-877.5</td>
<td>11.43</td>
</tr>
<tr>
<td>7</td>
<td>891.9-895.8</td>
<td>11.16</td>
</tr>
<tr>
<td>8</td>
<td>898.3-905.5</td>
<td>11.08</td>
</tr>
<tr>
<td>9</td>
<td>929.6-939.7</td>
<td>10.60</td>
</tr>
<tr>
<td>10</td>
<td>959.9-964.3</td>
<td>10.39</td>
</tr>
<tr>
<td>11</td>
<td>985.1-998.1</td>
<td>10.00</td>
</tr>
<tr>
<td>12</td>
<td>1076.7-1084.9</td>
<td>9.25</td>
</tr>
<tr>
<td>13</td>
<td>1095.0-1098.2</td>
<td>9.12</td>
</tr>
<tr>
<td>14</td>
<td>1113.5-1116.1</td>
<td>8.97</td>
</tr>
<tr>
<td>15</td>
<td>1124.4-1132.6</td>
<td>8.86</td>
</tr>
<tr>
<td>16</td>
<td>1142.2-1148.0</td>
<td>8.73</td>
</tr>
<tr>
<td>17</td>
<td>1155.3-1163.5</td>
<td>8.63</td>
</tr>
</tbody>
</table>
Table 8.2: Summary of dust and cloud detection sensitivity

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Cloud Phase</th>
<th>COD</th>
<th>PWV (g cm$^{-2}$)</th>
<th>Height (Km)</th>
<th>BTD 10-11 (K) (averaged)*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Q</td>
<td>NC</td>
<td>---</td>
<td>2.4</td>
<td>1-3</td>
<td>-5.4751±2.2219</td>
</tr>
<tr>
<td>Q</td>
<td>NC</td>
<td>---</td>
<td>1.8</td>
<td>1-3</td>
<td>-11.3730±2.0232</td>
</tr>
<tr>
<td>Q</td>
<td>NC</td>
<td>---</td>
<td>3.0</td>
<td>1-3</td>
<td>-1.8393±1.3787</td>
</tr>
<tr>
<td>V</td>
<td>NC</td>
<td>---</td>
<td>2.4</td>
<td>1-3</td>
<td>-6.0017±2.2654</td>
</tr>
<tr>
<td>K</td>
<td>NC</td>
<td>---</td>
<td>2.4</td>
<td>1-3</td>
<td>-3.8720±2.0337</td>
</tr>
<tr>
<td>K/C</td>
<td>NC</td>
<td>---</td>
<td>2.4</td>
<td>1-3</td>
<td>-4.8389±1.6848</td>
</tr>
<tr>
<td>ND</td>
<td>IC</td>
<td>---</td>
<td>2.4</td>
<td>---</td>
<td>6.4306±1.7167</td>
</tr>
<tr>
<td>ND</td>
<td>WC</td>
<td>---</td>
<td>2.4</td>
<td>---</td>
<td>1.4078±0.7115</td>
</tr>
<tr>
<td>V</td>
<td>IC</td>
<td>0.05</td>
<td>2.4</td>
<td>1-3</td>
<td>-2.1280±3.3132</td>
</tr>
<tr>
<td>V</td>
<td>IC</td>
<td>0.05</td>
<td>2.4</td>
<td>1-3</td>
<td>-1.9182±3.3180</td>
</tr>
<tr>
<td>V</td>
<td>IC</td>
<td>3.0</td>
<td>1.8</td>
<td>1-3</td>
<td>2.4109±2.4702</td>
</tr>
<tr>
<td>V</td>
<td>IC</td>
<td>3.0</td>
<td>1.8</td>
<td>1-3</td>
<td>0.7009±3.1364</td>
</tr>
<tr>
<td>V</td>
<td>IC</td>
<td>3.0</td>
<td>3.0</td>
<td>1-3</td>
<td>4.4201±2.3733</td>
</tr>
<tr>
<td>V</td>
<td>IC</td>
<td>3.0</td>
<td>3.0</td>
<td>1-3</td>
<td>4.0073±3.0309</td>
</tr>
<tr>
<td>V</td>
<td>WC</td>
<td>10</td>
<td>1.8</td>
<td>1-2</td>
<td>0.3824±0.2331</td>
</tr>
<tr>
<td>V</td>
<td>WC</td>
<td>10</td>
<td>2.4</td>
<td>1-2</td>
<td>0.6098±0.2393</td>
</tr>
<tr>
<td>V</td>
<td>WC</td>
<td>10</td>
<td>3.0</td>
<td>1-2</td>
<td>0.8364±0.2448</td>
</tr>
<tr>
<td>V</td>
<td>WC</td>
<td>10</td>
<td>2.4</td>
<td>1-2</td>
<td>0.6098±0.2393</td>
</tr>
<tr>
<td>V</td>
<td>WC</td>
<td>1</td>
<td>2.4</td>
<td>1-2</td>
<td>-1.4475±1.2152</td>
</tr>
</tbody>
</table>

Legend: Mineral nomenclature – Q=quartz, V=Volz, K=Kaolinite, KC=Kaolinite/Carbonate ND = no dust, NC=no cloud, IC=ice cloud, WC=water cloud, COD=cloud optical depth

* BTD 10-11 is calculated as an average quantity plus 1 standard deviation (1σ) over the respective particle sizes and optical depths (refer to text for discussion).
Table 8.3: Summary of dust size/shape sensitivity for $\tau_{\text{dust}}=1$

<table>
<thead>
<tr>
<th>Particle size ($BTD$)</th>
<th>Particle shape ($BTD$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CH</td>
<td>0.5µm</td>
</tr>
<tr>
<td></td>
<td>-1.36±1.2K</td>
</tr>
<tr>
<td>OS</td>
<td>1.0µm</td>
</tr>
<tr>
<td></td>
<td>-0.304±0.25K</td>
</tr>
<tr>
<td>S</td>
<td>1.5µm</td>
</tr>
<tr>
<td></td>
<td>-0.46±0.58K</td>
</tr>
<tr>
<td></td>
<td>2.0µm</td>
</tr>
<tr>
<td></td>
<td>-4.38±1.9K</td>
</tr>
</tbody>
</table>

Legend: CH – compact hexagon, OS – oblate spheroid, S-sphere
### Table 8.4: AERI dust optical depth retrieval summary

<table>
<thead>
<tr>
<th>Case</th>
<th>Dust Model</th>
<th>Correlation coefficient (r)</th>
<th>Relative error (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>9/21</td>
<td>V (OS)</td>
<td>0.4703 (0.9264)**</td>
<td>0.0615±0.2133</td>
</tr>
<tr>
<td>9/21</td>
<td>V(CH)</td>
<td>0.5935 (0.9302)**</td>
<td>0.1270±0.1670</td>
</tr>
<tr>
<td>9/22</td>
<td>Q (OS)</td>
<td>0.9188</td>
<td>0.0857±0.0202</td>
</tr>
<tr>
<td>9/22</td>
<td>V (OS)</td>
<td>0.9219</td>
<td>0.0683±0.0209</td>
</tr>
<tr>
<td>9/22</td>
<td>K (OS)</td>
<td>0.9255</td>
<td>0.1612±0.0168</td>
</tr>
<tr>
<td>9/22</td>
<td>K/C (OS)</td>
<td>0.9243</td>
<td>0.1741±0.0155</td>
</tr>
<tr>
<td>9/23</td>
<td>V (OS)</td>
<td>0.3161 (0.7668)**</td>
<td>0.0296±0.1436</td>
</tr>
<tr>
<td>9/23</td>
<td>Q(OS)</td>
<td>0.7468</td>
<td>0.0594±0.1097</td>
</tr>
<tr>
<td>9/23</td>
<td>V (CH)</td>
<td>0.7582</td>
<td>0.2994±0.1451</td>
</tr>
<tr>
<td>9/23</td>
<td>K (OS)</td>
<td>0.7572</td>
<td>0.3268±0.1540</td>
</tr>
<tr>
<td>9/23</td>
<td>K/C (OS)</td>
<td>0.3209 (0.7658)**</td>
<td>0.1969±0.1159</td>
</tr>
<tr>
<td>9/30</td>
<td>V (OS)</td>
<td>-----</td>
<td>0.0703</td>
</tr>
<tr>
<td>9/30</td>
<td>Q (OS)</td>
<td>-----</td>
<td>0.0124</td>
</tr>
<tr>
<td>9/30</td>
<td>Q(OS)</td>
<td>-----</td>
<td>&gt; 100%</td>
</tr>
<tr>
<td>9/30</td>
<td>Q (OS)</td>
<td>-----</td>
<td>0.4300</td>
</tr>
</tbody>
</table>

Legend: Q = quartz, V = Volz, K = kaolinite, K/C = kaolinite/carbonate
OS = oblate spheroid, CH = compact hexagon
** = correlation coefficient before/after points removed
Figure 8.1 Sensitivity of AERI ‘clear-sky’ sub-band spectra to changes in PWV (g cm$^{-2}$). Relative spectral differences in the 800-1000 cm$^{-1}$ region are larger due to greater water vapor absorption.
Figure 8.2 Sensitivity of AERI $BT$ spectra to changing RH and Temperature (T) profiles for three days during UAE$^2$ (a) RH profiles with the same PWV (~2.3 g cm$^{-2}$) (b) $BT$ sensitivity to RH. (c) T profiles for same days with inset showing profiles inside PBL. (d) $BT$ sensitivity to T.
Figure 8.3 (a) \( \text{BTD} \) 10-11 vs. dust optical depth for dust (Q) or cirrus only conditions. Error bars depict \( \text{BTD} \) variability due to uncertainty in PWV (1.8-3 g cm\(^{-2}\)). (b) \( \text{BTD} \) 10-11 for dust and cloud (mixed scenario). Shown are cirrus (black) and water (gray) clouds as functions of dust optical depth. (c) \( \text{BTD} \) clear-sky offset uncertainty vs. PWV. (d) Dust and cloud (mixed scenario) and \( \text{BTD} \) offsets. Dotted black lines denote zero point threshold.
Figure 8.4 Particle size and shape $BT$ sensitivity across (a) upper sub bands (USB: 13-17) and (b) lower sub-bands (LSB: 1-11). Shown are spheres (S), and compact hexagons (CH) with an effective size range of 0.5 to 2µm. Dust optical depth is fixed at 0.4.
Figure 8.5 BT sensitivity to dust optical depth at 962 cm⁻¹. Markers denote locations of AERI sub-bands 1-17 from left to right. (a) Volz (CH) model spectra for four optical depths with best fit AERI spectrum (broken line). Lines connecting sub bands 13-17 show optical depth slope dependence. (b) Same as (a) except for K/C (OS) dust model.
Figure 8.6 (a) $BT$ sensitivity to dust cloud height for altitudes $z = 1, 3$ and $5\text{km}$ at optical depth $\tau = 0.05$ (b) same as (a) but for $\tau = 0.4$ (c) MPL optical depth profile used to model optical depths for each layer. (d) $BT$ sensitivity to dust cloud thickness. $L1$ denotes a 1-layer thick dust cloud; $L1$-$L3$ denotes a 3-layer thick dust cloud etc.
Figure 8.7 AERI BT sensitivity ($\Delta BT = BT_{dust} - BT_{clear}$) to mineral composition over two spectral regions. Dust particles are spherical with a $R_{eff}$ of 2$\mu$m. Dust optical depth of 0.2 was used. (a) Thermal IR window: 800-1200 cm$^{-1}$ and (b) from 2500-2700 cm$^{-1}$. Note BT separation of minerals around 2500-2550 cm$^{-1}$. 
Figure 8.8   a. Estimated radiance error (residuals) due to MCT detector non-linearity with vertical axis given in radiance units [1 RU = mW m⁻² sr⁻¹ (cm⁻¹)].  b. AERIPLUS retrieved temperature profile from 9/22/04 during UAE².  c. Same as (b) but for WVMR.
Figure 8.9 a. AERI detected cloud/dust vs. MPL for 9/30/04. b. AERI retrieved IR optical depths for 9/22/2004 with/without non-linearity corrections applied. c. AOT scatter plot of AERI versus AERONET retrieved optical depths at 0.5μm. See text for details.
Figure 8.10 UAE\textsuperscript{2} AERI dust/cloud separation results versus the MPLNET MPL at SMART for two days without the non-linearity correction. The white circles on the MPL plots show the clouds detected by AERI.  (a) AERI results for 9/21/2004 (b) MPL for same day.  (c) AERI results for 9/22/2004.  The gray box denotes the times corresponding to the dust episode. (d) MPL for same day.
Figure 8.11. AERI daytime/nighttime dust and cloud separation. Both panels roughly coincide in time for comparison. (a) MPL NRB profile on 9/7/2006 during NAMMA showing heavy dust and low clouds. White asterisks denote positions of clouds. Break between ~12-14UTC corresponds to when the MPL was covered during solar noon. (b) AERI BTD 10-11 for the same day showing dust (below zero) and cloud (above zero). Diamond markers represent clouds from the MPL dataset in (a) overlaid for comparison.
Figure 8.12 AERI optical depth retrievals vs. AERONET sun-photometer (red) after cloud screening. AERI IR optical depths are scaled to those at a visible wavelength $\lambda=0.55\mu$m using the D’Almeida and Afghanistan dust models. Error bars on AERI and AERONET retrievals denote uncertainties in particle sizing and instrument calibration respectively. (a) Data from 9/21/2004, showing shape sensitivity to oblate spheroids (OS) and compact hexagons (CH). (b) Data from 9/22/2004, showing sensitivity to mineral composition. (c) Same as (b) but from 9/23/2004. (d) Data from 9/30/2004 showing sensitivity to PWV.
CHAPTER 9

9.0 SURFACE RADIATIVE FORCING OF DUST AEROSOL – DIRECT EFFECT

The following chapter is a detailed discussion of the surface radiative forcing of dust aerosol both in the SW and LW regions of the electromagnetic spectrum. The results of the SW approach are based on observations alone which employ a differential technique [Hansell et al. 2003] while in the LW approach, observations and model simulations are both used.

9.1 Shortwave forcing study

Abstract

An approach is presented to estimate the surface aerosol radiative forcing using collocated cloud-screened narrow-band spectral and thermal-offset corrected radiometric observations during the Puerto Rico Dust Experiment (PRIDE) 2000, South African Fire Atmosphere Research Initiative (SAFARI) 2000, and Aerosol Characterization Experiment Asia (ACE-ASIA) 2001. It is shown that the aerosol optical depths from the Multiple Filter Rotating Shadow-Band (MFRSR) data match closely with those from the Cimel Sun-Photometer data for two SAFARI-2000 dates. The observed aerosol radiative forcings were interpreted based on results from the Fu-Liou radiative transfer model, and in some cases, cross-checked with satellite derived forcing parameters. Values of the aerosol radiative
forcing and forcing efficiency, which quantifies the sensitivity of the surface fluxes to the aerosol optical depth, were generated based on a differential technique for all three campaigns, and their scientific significance is discussed.

9.1.1 Introduction

The quantification of the surface aerosol radiative forcing (SARF) and the minimization of its uncertainty remain to be the two primary goals in today’s studies of radiative effects of aerosols. A major global effort given towards these areas is the aerosol robotic network (AERONET) program [Holben et al. 1998]. Under the initiative and guidance of the National Aeronautics and Space Administration’s Goddard Space Flight Center (NASA GSFC), the AERONET uses the 8-channel sun photometer manufactured by Cimel Electronique at measurement sites all over the world to derive aerosol optical properties to facilitate the estimation of SARF based on observations. The inversion routines of Nakajima et al. [1983] and from AERONET [Dubovik et al. 2000] are used to retrieve aerosol optical depths, single scattering albedos, scattering phase functions, asymmetry parameters and indices of refraction.

There have been numerous efforts to estimate SARF. Most recently, Schafer et al. [2002] show that by using the sun-photometer retrieved aerosol optical depths, $\tau_a$, along with observed surface flux data from field campaigns in Brazil and South Central Africa, aerosol radiative forcing efficiencies due to smoke particulates are determined to be $-145 \text{ W/m}^2/\tau_a$ and $-210 \text{ W/m}^2/\tau_a$, respectively, for the range of instantaneous solar
zenith angles ($\theta_o$) between 25° to 35°. Maximum reductions in surface flux on the order of 337 W/m$^2$/\(\tau_a\), for $\theta_o = 31$° were observed for the heaviest smoke conditions ($\tau_a = 3.0$) in Brazil. The difference in the aerosol attenuation of flux between both sites is attributed to the variable composition of the fire source materials and the type of combustion. Christopher et al. [2002] used retrieved aerosol optical depth from half-hourly Geostationary Operational Environmental Satellite (GOES-8) imager data as well as collocated surface fluxes measured by NASA’s SMART system during the Puerto Rico Dust Experiment (PRIDE) to estimate the top-of-the-atmosphere (TOA) aerosol radiative forcing and SARF. Based on a mean $\tau_a$ value of 0.26, the daytime average SARF and associated standard deviation over a period from 28 June to 26 July 2000 were found to be $-18.13$ and $15.81$ W/m$^2$, respectively. Data from the first field phase of the Indian Ocean Experiment (INDOEX) have also been studied for estimating the SARF over the Indian Ocean [Conant et al. 2000 and Meywerk et al. 1999]. It was shown that for a soot-laden marine region south of India, a 0.1 change in $\tau_a$ leads to a $-4.0 \pm 0.08$ W/m$^2$ change in the 0.4-0.7 $\mu$m surface flux. Differences in the reported forcings are due to the corresponding changes in $\tau_a$ as identified in Schafer et al. [2002], while the monthly mean forcing is obtained by multiplying the efficiencies by observed $\tau_a$.

The quantification of SARF is focused on using radiometric and sun photometer data from PRIDE-2000 [Christopher et al. 2003], South African Fire Atmosphere Research Initiative (SAFARI) 2000 [Eck et al. 2003], and Aerosol Characterization Experiment Asia (ACE-ASIA) 2001 [Redemann et al. 2003] and radiometric measurements at the University of California at Los Angeles (UCLA) surface site. The
focus of PRIDE is the study of the radiative, microphysical and transport properties of
dust from the Saharan desert at the Roosevelt Roads Naval facility in Puerto Rico.
Similarly, the goals of SAFARI and ACE-ASIA are to study aerosol properties at and
around Skukuza, South Africa and Dun Huang, China, respectively. The aerosols over
these areas were generated from local biomass burnings, biological and industrial
sources, blowing sand over desert, and urbanized sources.

The scientific objectives of the works reported are as follows. First, spectral
measurements from the Multiple Filter Rotating Shadow-Band Radiometer (MFRSR),
and the Cimel sun-photometer and surface fluxes from broadband radiometers are
calibrated and processed. The time series of $\tau_a$ and surface radiative fluxes are generated
for selected dates during each field campaign. The retrieved $\tau_a$ from the MFRSR and sun-
photometer are then inter-compared. Similar comparisons of $\tau_a$ from these two
instruments were reported by Augustine et al. [2003] and Jin et al [2002]. Because the
MFRSR’s sampling frequency is about 60 times greater than that of the Cimel sun-
photometer, we chose to use the MFRSR data to determine the SARF to produce
statistically meaningful results.

Second, because clouds were prevalent during PRIDE 2000, it is necessary to
develop an effective cloud screening scheme in order to remove effects of cloud bias in
the retrieved aerosol properties and estimates of SARF. The magnitude of thermal offsets
in the UCLA pyranometer data is also evaluated, and an empirical correction method to
enhance the accuracy of the measured downward solar flux is developed. Finally, using
the irradiance data from collocated broadband radiometers along with the time series of
retrieved $\tau_a$, quantitative estimates of the SARF are made for each campaign. Two parameters are derived. First is the aerosol radiative forcing, which is a measure of the radiative effects of the aerosol. The second is the aerosol forcing efficiency, which is defined as the derivative of surface flux with respect to the aerosol optical depth. In this study, we directly determine SARF parameters from measurements and the results can be used as a tool for validating satellite retrieved aerosol radiative forcing. Knowledge of the regionally estimated SARF parameters will help the study of global climate change and improve the aerosol parameterizations in radiative transfer models.

The arrangement of this section is as follows. Part 2 discusses the primary instrumentation used in the study, including a brief introduction of the UCLA surface radiometric site. Part 3 describes the cloud removal procedure, the cross calibration scheme using the Cimel sun photometer and the $\tau_a$ retrieval scheme subject to the correction of Rayleigh scattering and ozone absorption. In part 4, estimates of surface aerosol radiative forcing and forcing efficiency are presented. Also included in this section are turbidity coefficients and shaping factors determined from angstrom’s empirical relation for SAFARI and model comparisons of the predicted SARF with the observed SARF using the Fu-Liou radiative transfer code [Fu and Liou, 1992]. Finally a summary is given in part 5.
9.1.2. Instrumentation

The primary instrumentation used in this study includes the MFRSR and the Total Solar Pyranometer (TSP-700) from Yankee Environmental Systems (YES), the Cimel Electronique Sun Photometer (CE-318), and Epply’s Precision Spectral Pyranometer (PSP). Both the MFRSR and the CE-318 contain narrowband visible and near infrared interference filters each 10 nm full width half maximum (FWHM). The MFRSR channels cover the following wavelengths: 0.41, 0.49, 0.61, 0.67, 0.86 and 0.94 µm. In addition, it contains one unfiltered broadband silicon pyranometer whose spectral response is from 0.30 – 1.10 µm. The MFRSR employs an automated rotating shadow-band to make simultaneous measurements of the total, diffuse and direct-normal components of the solar spectral irradiance. MFRSR calibration, performed at the YES laboratory, involves evaluating the angular, spectral and absolute responses for each channel. Instrument output is subject to uncertainty due to variations in the ambient temperature, which has been shown by a recent Atmospheric Radiation Measurement Program (ARM) study (May 2000 - ARM Web Site (http://www.arm.gov/docs/instruments/static/mfrsr.html) to be in the range of 1-2% per 1°K change in sensor head temperature. Another uncertainty is the positioning error in the shadow-band, which could lead to a positive bias in diffuse irradiance, and consequently the total component is underestimated. Shadow-band errors occurred at the start of PRIDE, so this data was rejected in the follow on analysis.

The Cimel sun photometer (refer to chapter 4) contains the following channels: 0.34, 0.38, 0.44, 0.50, 0.67, 0.87, 0.93 and 1.02 µm. It is designed to perform automated
direct solar and sky-scanning spectral radiometric measurements every 15 minutes throughout the day. Sky-scanning channels are calibrated using NASA GSFC’s 2-meter integrating sphere while the direct solar scans are compared to a Mauna Loa Observatory Langley calibrated reference instrument. The measured uncertainty in the retrieved optical depth ($\tau$) is estimated to be ~0.01-0.02 [Holben et al. 1998]. Together, both instruments provide excellent data for the reconstruction of the incident solar irradiance and vital information related to the determination of optical depths of water vapor, ozone and aerosols. Other instruments used include the normal incidence pyrheliometer (NIP) and both shaded and unshaded pyranometers that were mounted on the Kipp and Zonen’s (K & Z) two-axis positioning gear drive solar tracker, and a fixed platform. Table 9.1 shows a summary of the primary instruments used in this study.

For the past 8 years, the Atmospheric Sciences Department at UCLA has been operating an array of ground-based radiometers, including a K & Z CM21 Pyranometer, a K & Z CG1 Pyrgeometer, and an Eppley Precision Spectral Pyranometer (PSP). Daily solar and IR surface downward flux measurements by these instruments are recorded every minute and archived. In addition, a Cimel sun photometer, CE-318, has been added to the suite of instrumentation and is a part of the NASA Goddard Space Flight Center’s AERONET Program. Radiometric data collected at UCLA are employed to develop a thermal offset correction scheme and will be utilized in a long-term study to establish the climatology of aerosol radiative forcing in the Los Angeles (LA) basin area. Figures 9.1a and 9.1b depict the arrangement of broadband radiometers and the sun photometer, respectively. In the background of Fig. 9.1a are the SW radiometers mounted
on the K&Z solar tracker. The cloud-screened data measured by these instruments will provide the needed surface flux to compute the SARF. Figures 9.1c and 9.1d show examples of the time series of surface downward solar and IR surface fluxes for a clear day and a partially cloudy day, respectively. Note that the direct component of the surface irradiance has not been angle corrected. The large fluctuations during the morning hours reflect the effects of clouds. The clear-sky broadband data coupled with the sun-photometer $\tau_\alpha$ retrievals will provide for an effective assessment of the SARF.

9.1.3. Data Processing

A. Determination of Thermal Offset

Recently, much attention has been given to the quantification of the thermal offset of pyranometers [Dutton et al. 2001 and Haefelin et al. 2001]. Driven by the differential heating effects of the instrument’s thermopile detector, an output voltage is generated which can then be related to the radiative input via calibration. In addition to the shortwave response of the detector, there is also a longwave response due to intrinsic temperature gradients established within the device. These differences in temperature, particularly between the dome and the body of the instrument are caused by a number of factors. For example, the small thermal conductivity of the dome causes a temperature gradient between the top and base of the dome. The instrument is also subject to the transient conditions of the ambient air and surface temperature as well as to the wind and precipitation.
Several approaches have been developed to characterize this effect with diffuse flux measurements. Previous studies [Charlock et al. 1996; Kato et al. 1997 and Halthore et al. 1997] have shown that the observed diffuse flux density is consistently smaller than that predicted by models. One of the methods developed by Dutton et al. [2001] is based on the relation between the pyranometer’s nighttime offset and the net infrared thermopile signal of a collocated pyrgeometer. This method was adopted for deriving a quantitative estimate of the pyranometer’s thermal offset at the UCLA site, assuming that the pyranometer’s thermal responses are approximately the same for both daytime and nighttime operations.

Both the shaded/ventilated pyranometer (K&Z CM21) and the unshaded/unventilated pyranometer (Eppley PSP) were used alongside the K&Z CG1 ventilated pyrgeometer. Two sets of nighttime data were analyzed covering the periods of 24 May – 15 June 2001 and 02 August – 11 September 2001. During both periods it was found that the dark offset of both pyranometers was less than $5 \text{ W/m}^2$. Linear regressions were used to determine a simple function relating the PSP thermal offset to the CG1 thermopile output. Figure 9.2a shows illustrative results for the Eppley PSP during the May-June period. A slope of 0.039 with intercept of –0.39 was found with the correlation coefficient of 0.9308. These regression coefficients are comparable to the slope found by Haeffelin et al. [2001]. The small slope indicates that the PSP offset is insensitive to the net IR signal. Previous studies [Alberta et al. 1999] suggested that the regression be rotated around the centroid of the data to force it through the origin, so that a net IR flux of zero $\text{W/m}^2$ corresponds to a zero thermal offset.
In this case, because the intercept is considered to be small enough, the use of this empirically derived relationship with non-zero intercept is justified. The simple relationship takes the form:

$$\delta^{\text{PSP}} = 0.039 \times \text{CG1} - 0.39$$  \hspace{1cm} (9.1)

from which the $\delta^{\text{PSP}}$, thermal offset, can be determined based on the CG1, the thermopile output. By subtracting this offset from the irradiance measurements, one obtains a corrected signal, which is greater than the uncorrected PSP irradiances. Figure 9.2b depicts the uncorrected (range between 0 and $-5 \text{ W/m}^2$) and corrected nighttime signals. The corrected signals are generally closer to the zero offset line than the uncorrected signals. Further thermal offset studies involving the direct measurement of the daytime offset by capping the dome of the pyranometer are being investigated.

B. Cloud Screening Procedure

Various statistical cloud-screening techniques have been developed to reduce uncertainties in retrieved aerosol optical depths from a cloudy data set [Conant et al. 2000]. The following cloud screening technique is adopted in this study. For the same solar zenith angle, the direct component of the surface downward solar flux is generally larger for a clear atmosphere than that for a cloudy atmosphere. On the other hand, the diffuse component of the surface downward solar flux is normally smaller for a clear sky.
than for a cloudy sky due to additional scattering by cloud particles. Thus the surface
diffuse flux measured by the shaded pyranometer is used in conjunction with the surface
direct flux measured by the NIP to develop a cloud screening scheme based on their
relative magnitudes.

A parameter, the scatter ratio is defined as the ratio of the observed diffuse flux to
the observed direct flux, normalized by air mass and total irradiance. This parameter is a
measure of the scattering effect of clouds. A threshold of the scatter ratio is determined
by a simple approach. The time series of daily scatter ratios typically shows large scatter
in the early morning and late afternoon hours with varying lower values in between. Data
points associated with solar zenith angles greater than 80° were first discarded. The
screening procedure then takes successive averages of the scatter ratios. During each
iteration, data points greater than the average plus one standard deviation are discarded.
The iteration stops when the average values converge. The final average value is taken as
the threshold. An estimated threshold of 0.0003 was found for PRIDE, while for both
SAFARI and ACE-ASIA, a threshold of 0.0005 was prescribed. This simple approach is
similar to the clear-sky shortwave flux method used by Chou et al. [1997].

Figure 9.3a depicts the scatter ratios for July 8 and 9 during PRIDE. The threshold
is obviously above the bulk of lower values of the scatter ratio. Points above the
threshold, denoted by a dashed line, are considered to be cloud-contaminated, although
some of these points actually may be cloud-free. For the case of heavy Saharan desert
dust event during PRIDE, the threshold value is much higher. Figure 9.3b shows the time
series of surface downward flux measured by the TSP-700 based on filtered data, which
are compared to the initial unfiltered data set for July 8. One can see that those large
drops in irradiance due to cloud contamination are removed after applying the screening
algorithm. The data at low elevation angles is also removed. This technique provides a
simple means for removing cloud-contaminated data points. The computed thresholds are
specified to ensure that there are enough remaining data points to produce statistically
meaningful results. The YES Total Sky Imager (TSI-440) was also used as an
independent check for assessing the cloud coverage. Further developments are required to
make the method more robust.

C. Calibration of MFRSR Irradiance

For clear sky conditions, one can perform a Langley analysis to determine the
exo-atmospheric solar spectral constant $I_{o\lambda}$. A linear fit of the logarithm of the direct
irradiance at wavelength $\lambda$, $I_{\lambda}$, versus the air mass, $m$, yields a straight line with slope
$\frac{d(\log I_{\lambda})}{dm}$. This slope is proportional to the total optical depth of the atmospheric
medium. One can further extrapolate the straight line to zero air mass to obtain the
intercept, which is the exo-atmospheric solar spectral irradiance constant $I_{o\lambda}$ (solar
constant). The mean earth-sun distance must also be accounted for in computing the
solar constant $I_{o\lambda}$. It is assumed that the total optical depth of the atmosphere does not
exhibit significant temporal variations.

For cloudy-sky conditions, alternative methods in determining the solar constant
are sought. Alexandrov et al. [2001] reviewed several retrieval algorithms for processing
MFRSR data for partially cloudy conditions using direct and diffuse measurements. The
current study estimates the solar constant $I_{0,\lambda}$ via two approaches. The first approach cross calibrates the MFRSR with the Cimel data. Using the sun-photometer retrieved aerosol optical depth along with the direct measurements of the MFRSR, an estimate of $I_{0,\lambda}$ is obtained based on the Beer-Lambert law:

$$\frac{I_{\lambda}}{I_{0,\lambda}} = \exp (-\tau_t m) \quad (9.2)$$

where $I_{\lambda}$ and $I_{0,\lambda}$ have been defined above, $\tau_t$ is the total optical depth due to the extinction by atmospheric scattering and absorption and $m$ is the air mass.

The estimated accuracy by which the calibration information is transferred from the Cimel to the MFRSR is primarily constrained by the uncertainty in the Cimel’s $\tau_a$ retrievals as well as the reported errors in the measured spectral irradiances of the MFRSR. Initially, based on the raw data, the values obtained for $I_{0,\lambda}$ vary more than expected. Values of $I_{0,\lambda}$ were then calculated using only the filtered data and the variation of $I_{0,\lambda}$ between campaigns is reduced. As has been done with the unscreened data, a limit was imposed on the $I_{0,\lambda}$ to constrain the variation to no larger than a few percent. A temporal average of $I_{0,\lambda}$ was then obtained from the resulting data sets for each wavelength. The second approach relies on using the values of $I_{0,\lambda}$ derived via the Langley method from the clearer days found during ACE-ASIA.

Table 9.2 shows the resulting values of $I_{0,\lambda}$ for each MFRSR channel using both methods where the constant values are based on screened data only. The values of $I_{0,\lambda}$ are
comparable to the spectral solar constants found by Thekaekara[1974], Neckel and Labs [1981] and Anderson et al. [1995]. Cloud screening significantly improves PRIDE constants by almost 10-15%. The solar constants from SAFARI and ACE-ASIA data remain the same after screening, probably because there were not so many clouds detected as during PRIDE. The ACE-ASIA 0.41 µm constant requires investigation, as it remains to be quite small. Differences in \( I_{0\lambda} \) could be due to the small differences in the location in the band center of the MFRSR and Cimel channels (e.g. 0.41 µm versus 0.44 µm) and to the varying aerosol conditions of each field campaign.

D. Determination of Aerosol Optical Depth

Once the values of \( I_{0\lambda} \) are determined, the total optical depths can be found via the Beer-Lambert Law using the direct beam measurements of the MFRSR and the corresponding air masses. It is assumed that errors in the retrieved MFRSR \( \tau_a \) are approximately linearly proportional to the uncertainties in the calibration of the MFRSR irradiance determined in the previous section. The aerosol optical depths \([ \tau_a(\lambda) ]\) can then be calculated for each channel by subtracting the molecular Rayleigh scattering and ozone absorption optical depths from the total optical depth [Liou, 2002]:

\[
\tau_a(\lambda) = 1/m(\mu_o) [\ln(I_{0\lambda}) - \ln(I_{\lambda})] - \tau_R(\lambda) - \tau_O(\lambda) \\
(9.3)
\]

where \( \mu_o \) is the solar zenith angle. The data of MFRSR 0.94 µm channel was discarded in this study so as to minimize effects of water vapor absorption.
The parameterization of Rayleigh scattering optical depth is expressed as follows:

\[
\tau_R (\lambda) = 0.008569 \lambda^{-4} (1 + 0.0113 \lambda^{-2} + 0.00013 \lambda^{-4}) \frac{p}{p_o} \tag{9.4}
\]

where \( \lambda \) is the wavelength (\( \mu \)m) and \( p \) and \( p_o \) are the pressure at the observation site and standard atmospheric pressure (1013.25 mb), respectively. A corrective term to account for ozone absorption is also applied. If the column ozone amount (in Dobson units- DU) is known \textit{a priori}, its extinction optical depth can be obtained by

\[
\tau_O (\lambda) = a_o (\lambda) C_o \rho \tag{9.5}
\]

where \( a_o (\lambda) \) is the ozone absorption coefficient for wavelength \( \lambda \) (cm\(^2\)/g), \( C_o \) is the column ozone expressed in Dobson units (1DU = 10\(^{-3}\) atm-cm) and \( \rho \) is the ozone number density, given by Loschmidt’s number \((2.69 \times 10^{19} \text{ cm}^{-3})\) which is the approximate number density of particles at standard atmospheric conditions. The column ozone over the region of interest at a given time is either extracted from the NASA Total Ozone Mapping Spectrometer (TOMS) database as was done for this study or calculated by the methods of \textit{Van Heuklon} [1979]. Alternatively, a climatological value of \( C_o \rho \) can also be used to determine the extinction optical depth. The absorption coefficients of other trace gases (i.e \( \text{CO}_2, \text{O}_2, \text{and NO}_2 \)) in the visible spectrum are negligibly small and are therefore not considered in this study. Water vapor absorption in the visible spectrum is also very small and thus neglected.
9.1.4. Observational Results

In this section, illustrative observation and analysis results on the retrieved aerosol optical depths, turbidity coefficients, aerosol size distribution parameters and surface aerosol radiative forcing efficiency are shown.

A. Aerosol Optical Depth

Figures 9.4a and 9.4b show the Earth Probe TOMS satellite images of the aerosol index distribution over the African continent for September 6 and 7, 2000 during SAFARI. The aerosol index is defined as the difference between the observations and model calculations from an aerosol-free atmosphere with the same surface reflectivity and measurement conditions. It is approximately linearly proportional to the aerosol optical depth if the index of refraction, particle size distribution and the aerosol layer height are known from the other measurements, with the residence height being a key assumption. Information on tropospheric aerosols, which absorb ultra-violet radiation, is obtained from the 0.33 and 0.36 μm radiances. The circled regions denote the geographical location of Skukuza, South Africa, where the ground-based radiometers were deployed during SAFARI. As can be seen, the aerosol loading on the 6th was much larger than on the 7th. These two images illustrate the notable change in aerosol index from the 6th to the 7th probably due to the large variations in the concentration of smoke particles from the local biomass burnings.

Figure 9.5a shows the ground-based cloud-screened retrieved 0.5 μm aerosol optical depth for the two dates during SAFARI. It is noted that both the MFRSR and the
Cimel observed higher optical depths on the 6th than on the 7th. The retrieved MFRSR aerosol optical depths correlate well with those of the Cimel sun-photometer with a correlation coefficient of about 0.93. One would think that such a high degree of correlation is expected, because the instruments were cross-calibrated. However, there are slight differences in the datasets that were used to derive aerosol optical depths. This could be attributed to intrinsic differences in the instruments, particularly in the central frequency of corresponding channels. Figures 9.5b and 9.5c show daily time series of retrieved aerosol optical depth for ACE-ASIA and PRIDE after cloud removal, and correlation coefficients for these corresponding time series are 0.98 and 0.80, respectively. The mean and root-mean-square (RMS) differences of the retrieved \( \tau_a \) for both instruments during each field campaign are given in Table 9.3. The smallest mean difference was noted for SAFARI. Therefore this was chosen as the primary data set in subsequent analyses.

B. Aerosol Turbidity Coefficients/Shaping Factor

To infer aerosol size distributions from the sun-photometer data, we use the Angstrom empirical turbidity relation:

\[
\tau = \beta \lambda^{-\alpha}
\]

(9.6)

where \( \beta \) and \( \alpha \) are the turbidity coefficient and the shaping factor, respectively [Liou, 2002]. The turbidity coefficient \( \beta \) is a proportionality constant relating the optical depth
and the wavelength. The shaping factor $\alpha$ provides a measure of how rapidly aerosol optical depth ($\tau_a$) changes with wavelength (i.e. it is a measure of the “steepness” of the $\log(\tau_a)$ vs. $\log(\lambda)$ curve. $\alpha$ also is related to the size of particles. Larger particles generally correspond to smaller $\alpha$ while smaller particles generally correspond to larger $\alpha$.

Plotting the logarithm of $\tau_a$ against the logarithm of the sun-photometer channel wavelength, $\lambda$ and performing a linear regression fit of the data points in the plot would lead to a straight line with its slope and intercept corresponding to $\beta$ and $\alpha$, respectively. However, in this study, an alternative approach is adopted to evaluate the shaping factor $\alpha$. The shaping factor can be approximated by taking the logarithm of the ratio of Angstrom’s power law expression at two different wavelengths. If $\tau_a$ is measured at two wavelengths, then the following expression is obtained:

$$Z = \frac{\tau_a(\lambda_1)}{\tau_a(\lambda_2)} = \left[\frac{\lambda_1}{\lambda_2}\right]^{-(-\alpha+2)} = y^{-(-\alpha+2)} \quad (9.7)$$

Thus the shaping factor can be inferred from:

$$\alpha^* = 2 - \log_{10} Z \quad (9.8)$$

This was done using the wavelength pair of 0.67 and 0.86 $\mu$m for the three field campaigns as a way to understand the relative difference in the size parameters of aerosol particles for different field campaigns.

Shaping factors for both PRIDE and ACE-ASIA were found to be much smaller than those for SAFARI, indicating that the size distribution of wind blown dust particles
over PRIDE and ACE-ASIA sites are more polydispersed than that for the smoke particles over SAFARI sites. According to Eck et al. [2003], typical values of the shaping factor are larger than 2.0 for fresh smoke particles and close to zero for Sahelian/Saharan dust particles. Figure 9.6 shows the comparison of MFRSR and Cimel retrieved shaping factors from SAFARI-2000. Note the peak shaping factors around August 17, which are very likely due to heavier loading of smoke particles than on other dates. Both data sets yield a mean coefficient of about 1.5. Cimel data sets for PRIDE and SAFARI were evaluated for comparison. The dust studies seem to yield shaping factors in the range of about 0.2, while particles produced from biomass burnings yielded the shaping factor around 1.5 and higher.

C. Surface Radiative Fluxes

To interpret and cross check the reliability of the observed surface radiative fluxes, the one-dimensional broadband radiative transfer program developed by Fu and Liou (Fu-Liou Code - Fu and Liou [1992]) and later modified by Charlock and his associates was employed. This code was chosen because of its capability to accurately simulate aerosol radiative effects through a delta-four-stream method. This model was used to compute the total, direct-normal and diffuse surface radiative fluxes for the two chosen dates, September 6 and 7, 2000, during SAFARI. Daily averages of retrieved $\tau_a$ (0.5µm) along with the solar zenith angle was used as input to the radiative transfer code to simulate the diurnal variations of surface fluxes at Skukuza, South Africa, because both dates exhibited nearly constant values of aerosol optical depth. In order to obtain an
approximate match of spectral bands between the Yankee TSP-700 pyranometer (bandwidth 0.3 µm –3.0 µm) and the Fu-Liou Code, fluxes for 5 of the 6 short-wave bands (0.2-0.7, 0.7-1.3,1.3-1.9,1.9-2.5,2.5-3.5 µm) in the Fu-Liou code are computed. Because the spectral interval of the Fu-Liou Code is larger than that of the broadband radiometer (0.3-3.0 µm), the model shortwave fluxes are expected to be slightly larger than the observed surface fluxes.

Both chosen days were relatively cloud free, although a few cloudy patches were observed using the TSI-440 and noted. It’s also difficult to assess whether or not sub-visual cirrus clouds were present. However, in this study we prescribe the cloud optical depth to be nearly zero. A default equatorial sounding profile was used, with an assumed average surface albedo of 0.15 for the spectral range considered. Aerosol input parameters are prescribed as follows. Earlier it was shown that the aerosol loading on the 6th was heavier than on the 7th. To simulate the aerosol composition for the 6th, an external mixture of 80% black carbon (soot) and 20% sulfate particles was assumed. According to Eck et al [2001], 15-20% of the aerosol produced for flaming combustion is black carbon while less than 3% of soot is generated for smoldering combustion processes. Because the biomass burning fires in the savanna ecosystems of South Africa are of flaming nature, the black carbon was chosen to be 20% of the total mixture of aerosols. The aerosol types and single scattering properties are taken from the Optical Properties of Aerosols and Clouds (OPAC – Hess et al. [1998], D’Almeida [1991], Tegen et al. [1996]). The model then partitions the average optical depth according to the assumed percentages of aerosols used in the mixture.
The TOMS image for the 7th, on the other hand, shows a much smaller aerosol loading near Skukuza. For the purpose of running the model, a simple continental background aerosol was used for this date using the aerosol properties from D’Almedia [1991]. Figures 9.7a-b show the comparison of computed surface fluxes with the observed fluxes for both dates. Figure 9.7c displays the estimated solar noontime SARF between the two days where the difference between the calculated and observed forcings is approximately 11 W/m². The differences between computed and observed fluxes shown in Fig. 9.7 can be explained as follows. The model computes surface fluxes under clear sky, which are compared with those observed during the cloudy periods on the 6th (morning) and on the 7th (afternoon). A review of the percentage cloud cover from the TSI-440 was performed which found the percentage of opaque clouds in the morning of the 6th was larger than that in the afternoon of the 7th. Another possible reason for the observed discrepancy is the model’s use of default equatorial temperature and pressure data. Incorporating sounding profiles obtained during campaign into the model would most likely reduce the discrepancy. Having the aerosol in-situ data for each field campaign would also help to resolve these differences. This simple model comparison shows a reasonable agreement between predicted and observed surface fluxes in clear-sky conditions with differences less than 3-5% for a nearly constant aerosol loading based on aerosol composition representative of biomass burning processes. Sensitivity studies will need to be conducted to assess the impact of these variable conditions on the calculated surface fluxes.
D. Estimation of Surface Aerosol Radiative Forcing and Forcing Efficiency

Aerosol radiative forcing is defined as the difference between the observed and aerosol-free radiative fluxes. In this study, the differential method is adopted to estimate the aerosol radiative forcing. Using measured surface fluxes and aerosol optical depths, the forcing efficiency, which is the derivative of surface flux with respect to the aerosol optical depth was first evaluated. By fixing an air mass interval, the forcing efficiency over this interval can be determined by a least-square fitting of total flux measurements versus the corresponding $\tau_a$. The radiative forcing is obtained by multiplying the forcing efficiency with the mean retrieved aerosol optical depth. The advantage of this method is that the aerosol radiative forcing is directly determined from observations, so that it is independent of assumptions made on the aerosol single scattering properties.

3-D surface plots were generated for data obtained during each campaign. Referring to Figs. 9.8a-c, the data are expressed in terms of 3 variables, $\tau_a$, air mass ($m$) and total surface flux. The total surface flux was measured by either the TSP or PSP. The resulting surface plots allow the visual determination of the radiative forcing efficiency for different intervals of air mass. For a fixed interval of air mass, the steeper the surface is, the larger the aerosol forcing efficiency. Figures 9.9a-c depict the air-mass dependent radiative forcing efficiencies for PRIDE, SAFARI and ACE-ASIA respectively. It is noted that the total number of data points for PRIDE is less because the PRIDE data collecting period was shorter than the other campaigns by about two weeks. Linear regressions were applied to data points for each air mass interval, and the slope $[dF/d\tau_a]$ for each fitted straight line corresponds to the forcing efficiency. The slopes
depicted also represent maximum correlation between the total irradiance and the aerosol optical depth. Figure 9.9a shows that there was considerable scatter in the data, probably because the MFRSR’s shadow-band does not completely cover the sensor during diffuse radiometric measurements and sea salt built up on the dome of the radiometer.

Compared to Q. Ji’s (GSFC NASA – University of College Park, Maryland) initial solar noon radiative forcing efficiency (-95 W/m²/τa) based on the Cimel data from PRIDE, the MFRSR derived efficiency is higher by about 25-50 W/m²/τa as shown in Figures 9.8a and 9.9a. The radiative forcing efficiency first decreases with increasing air mass up to θo = 60°, then increases again. For air masses larger than 3, the estimated radiative forcing results are subject to large uncertainties due to less data points in those intervals. The mean τa at 0.5 μm observed during PRIDE was found to be 0.21 for the MFRSR and 0.24 for the Cimel. The estimated radiative forcings for air mass of 1.5 and 2.0 were found to be –12.82 W/m² and –19.68 W/m², respectively. Christopher et al. [2003] used aerosol optical depth retrievals from half hourly GOES-8 imager data along with the measured downward solar flux using the NASA SMART platform to derive a quantitative estimate of the radiative forcing due to dust at both the surface and at the TOA. Surface radiative forcings at approximate air masses of 1.5 and 2.0 were found to be –19.66 ± 16.71 W/m² and –20.36 ± 16.92 W/m², respectively. These ranges of satellite-derived radiative forcings compare well with the results from the present study.

For SAFARI, as shown in Figures 9.8b and 9.9b, after discarding several outliers for large airmass and large τa, the radiative forcing efficiencies are much higher than those for PRIDE. This could be attributed to the large amount of absorbing aerosols
emitted from the local biomass burnings at Skukuza. The surface plot found in Figure 9.8c is from ACE-ASIA. For most of the time, the sun remained low in the sky. The smallest air mass observed was around 1.1 on 09 May corresponding to $\theta_o=70^\circ$. The observed radiative forcing efficiencies are comparable but slightly higher than PRIDE and could be due to the effect of urbanized aerosols detected during ACE-ASIA. This might be consistent, perhaps with the dust aerosols observed for both field campaigns. Further work is required to reduce the uncertainties in the PRIDE data and to make physically more meaningful comparisons to other studies done using both ground and satellite based platforms.

9.15. Summary

Surface aerosol radiative forcing (SARF) parameters have been determined using spectral and radiometric data obtained during PRIDE-2000, SAFARI-2000 and ACE-ASIA-2001. A brief introduction of the UCLA radiometric surface site along with a description of the measured data and current research plans was presented. To remove cloud contamination, a cloud-screening filter was employed to reject data points with scatter ratios larger than an empirically determined threshold. A thermal offset correction procedure was also used to account for the differential heating effects of the pyranometer.

Through a cross calibration scheme and the Beer-Lambert Law, the aerosol optical depths were retrieved from the MFRSR datasets and compared to those obtained from the Cimel sun photometer. A correlation coefficient of about 93% was obtained.
Errors in the calibration transfer are constrained by the uncertainty in the Cimel’s $\tau_a$, retrievals as well as the reported errors in the measured spectral irradiances of the MFRSR. The Angstrom shaping factor was also determined with focus on the SAFARI campaign. A mean value of 1.5 was found by employing a simple wavelength pairing technique using Beer’s Law. Coefficients found from PRIDE and ACE-ASIA were smaller, indicating larger size particles, which could be due to desert dust aerosols. To interpret and cross check the measured radiative fluxes, the Fu-Liou radiative transfer code was used to calculate the total downwelling surface fluxes and hence the forcing corresponding to the conditions found during SAFARI. Surface radiative fluxes were computed using a nearly constant observed aerosol optical depth and solar zenith angles corresponding to Skukuza, South Africa over a two-day period. The predicted surface fluxes are within 3–5% of the measured fluxes. The estimated noontime forcing efficiencies for these two days was found to be around $-139.92 \text{ W/m}^2/\tau_a$ compared to the observed value of $-151.16 \text{ W/m}^2/\tau_a$ with a difference of $11.2 \text{ W/m}^2/\tau_a$.

The retrieved aerosol optical depths were then used along with the total surface flux measurements to derive the forcing efficiencies for different airmass intervals based on linear least squares fitting. The larger efficiencies found for SAFARI ($\sim -176.0 \text{ W/m}^2/\tau_a$) and ACE-ASIA ($\sim -286.0 \text{ W/m}^2/\tau_a$) are most likely due to the highly absorbing aerosols generated from local biomass burnings, biological and industrial sources, and urbanized sources. For air mass intervals of 1.5 and 2.0, and using an averaged $\tau_a$ of 0.21, the calculated SARM for PRIDE was found to be $-12.82 \text{ W/m}^2$ and $-19.68 \text{ W/m}^2$ respectively.
Further investigation is required on SARF studies particularly in the areas of cloud removal schemes, and comparisons of SARF with computations. Measured aerosol optical properties, cloud information and meteorological sounding data will be used to accurately assess the model’s performance with the observed quantities. It is anticipated that the current study will provide for and encourage alternative strategies in the estimation of SARF over various regions with variable aerosol composition.

9.2 Long wave forcing study

Abstract

The surface radiative effects of dust aerosol in the thermal IR are highly uncertain due to the varied nature of its optical and microphysical properties. In this study we calculate the LW surface forcing of dust aerosol during the UAE$^2$ field campaign using AERI retrieved dust IR optical depths (10µm) in a 1-D radiative transfer model. The modeled LW surface fluxes are compared to those from a precision infrared radiometer and on average are found to be within ± 5 Wm$^{-2}$. Employing a dust climatological model, the averaged instantaneous LW surface forcing for a 7-day period over which notable dust activity was observed ($\tau \approx .45$) was found to be $\sim 4.3 \pm 0.91$ Wm$^{-2}$ similar in magnitude to that reported in previous forcing studies. The forcing sensitivity to key dust properties was examined using the dust models developed in chapter 6. Differences in LW forcing due to uncertainties in particle shape/size can vary from less than 1 Wm$^{-2}$ to $\sim 35$ Wm$^{-2}$.
for the sizes/shapes considered and strongly depends on dust optical depth. Similarly differences in LW forcing due to dust composition range from less than 1 Wm\(^{-2}\) to \(~20\) Wm\(^{-2}\). Model calculations suggest that the LW forcing for heavy dust conditions where visible optical depths exceed unity varies from about 8-20 Wm\(^{-2}\). Although smaller than the SW forcing, the LW contribution can be significant and should be considered in global climate modeling studies.

### 9.2.1 Introduction

Previous studies show that the direct radiative effects of dust aerosol are difficult to quantify due to the inherently large uncertainties in the dust particle size, shape and composition [Sokolik et al. 1999]. The lack of measurements both in the field and laboratory along with the difficulties in interpreting the data we do have has significantly hampered any progress in constraining the dust radiative forcing parameters and hence any attempt on reducing the large uncertainties given in the IPCC (2001) report (Fig. 1.1).

Due to the larger availability of dust data in the visible wavelengths (e.g. AERONET), most research efforts have placed greater emphasis on the cooling shortwave (SW) effects of mineral dust [Haywood et al. 2001] with reported values typically in the range of -20 to -60 Wm\(^{-2}\) at the surface depending on the surface albedo and solar zenith angle and up to -20 Wm\(^{-2}\) at the top of the atmosphere (TOA - De Tomasi et al. 2005).

More recently the impacts of aerosols in the long wave (LW) regions of the
spectrum have garnered a significant amount of attention within the scientific community. Vogelmann et al [2003] have recently demonstrated the importance of the surface LW contributions of aerosols, suggesting the LW effects should be included in current climate model simulations. For example, Vogelmann et al. [2003] show that the daytime surface LW forcing observed during the ACE-ASIA (2001) field study, where dust was a major component of the total aerosol burden, can range anywhere from several Wm$^{-2}$ up to 10 Wm$^{-2}$, certainly comparable in magnitude and sign to the surface warming due to green-house gas emissions. Sokilik et al. [1997] further demonstrate that model simulations of surface forcing in mild dust conditions varies from 3-7 Wm$^{-2}$ and can exceed 15 Wm$^{-2}$ for heavy dust episodes. The positive LW surface forcing, albeit usually much smaller compared with the SW, can still exert a sizeable impact to the radiative budget. Since many dust particle sizes near the source region (i.e. coarse and giant sized modes) are comparable to that of the incident wavelengths, scattering and absorption in the thermal IR will be non-negligible [Dufresne et al 2002] and will contribute to a positive or negative total forcing (solar + IR) depending on the how the mineral components are mixed (aggregated or individual), and on the sizes and shapes of the particles. A priori knowledge of these key dust properties is therefore critical for properly assessing the radiative effects of mineral dust and thus reducing the uncertainties of the direct forcing.

In this study, the daytime LW surface radiative forcing of dust during the UAE$^2$ is investigated. This study employs the dust microphysical models developed in chapter 6. The NASA Langley modified Fu-Liou radiative transfer code (FL0403, April 15, 2003 -
Rose et al. 2002; Fu and Liou et al. 1992) is employed to calculate the instantaneous surface forcing parameters for a 10-day period at SMART (13-23 September, 2004) over which notable dust activity was observed. The daytime measurements at SMART and MAARCO were used to constrain the model’s surface boundary conditions and atmospheric state. Model broadband surface fluxes were first compared to those of a collocated Eppley precision infrared radiometer (PIR) to validate the model calculations. A time series of LW surface forcing was then determined by subtracting the flux profiles of the model clean (pristine)-sky from those of the corresponding dust-laden atmosphere using the AERI retrieved optical depths from the dust models. Lastly the averaged instantaneous LW forcing over the 10-day period was computed and compared to the UAE2 SW forcing reported by Markowitz et al. [2006]. The LW forcing efficiency based on the differential technique [Hansell et al. 2003] using only the ground-based observations at SMART is determined and compared with the model derived LW forcing presented in this chapter. Lastly a sensitivity study of the LW surface forcing to key dust parameters was performed.

The organization of this section is as follows. Part 2 discusses the methodology for calculating the LW surface radiative forcing of mineral dust. Part 3 presents the sensitivity of LW forcing to key dust parameters, and PWV. Part 4 provides the data and details of each case study and finally, a summary along with potential applications are given in Part 5.
9.2.2. Dust radiative forcing methodology

In this study the LW surface forcing of mineral dust is defined to be the difference in the downwelling radiative flux at the surface between a dust-laden atmosphere and a clean (pristine) atmospheric reference free of dust. This is given by the following expression:

$$\Delta F \downarrow = F_{\text{dust}} \downarrow - F_{\text{clean}} \downarrow$$  \hspace{1cm} (9.2.1)

where $\Delta F$ represents the calculated forcing in Wm$^{-2}$. Since SMART was located in the desert interior where airborne dust was prevalent nearly every day, observed clean-sky references were not available and therefore had to be calculated using a model atmosphere with dust optical depths set to zero. Following the methods presented in [Hansell et al. 2003], the LW surface forcing efficiency ($F_{\text{eff}}$) of dust is defined to be the derivative of LW surface flux ($F$) with respect to the IR optical depth ($\tau$) at 10µm. This is given by the following expression:

$$F_{\text{eff}} = \frac{\partial F \downarrow}{\partial \tau}$$  \hspace{1cm} (9.2.2)

where $F_{\text{eff}}$ has the units of Wm$^{-2}/\tau$ (per unit optical depth). The IR optical depth at 10µm was chosen on the basis that most mineral components of dust strongly absorb at 10µm [Pierangelo et al. 2005].

Water vapor is a large source of error when calculating aerosol radiative effects in
the thermal IR window due to the strong absorption properties of the water vapor continuum (refer to chapters 6 and 8). Large changes in PWV can significantly alter the downwelling radiances and hence flux if not properly taken into account. The LW forcing is therefore sensitive to the PWV, however for model considerations, the PWV for both dust and clean-sky atmospheres are the same and will effectively cancel out from Eq. 9.21. If however one uses observations via the differential technique [Hansell et al. 2003] to calculate the LW forcing, a moderately large PWV with high temporal variability can potentially bias the results. Several precautions can be met to minimize such impact including 1. Restricting the observed data to those times where the PWV falls below a prescribed threshold or 2. Only employing the observed data where the variability in the PWV is minimal or effectively constant. Both conditions require that the data be properly screened to ensure the LW forcing calculations are accurate.

The main source of detected LW flux at the surface is from dust emissions which radiate thermal energy due to the dust particles’ increased temperature resulting from the strong mineral absorption processes prevalent in the IR. It has also been shown in previous work [Dufresne et al. 2002], that the LW scattering effects of dust can exert a significant influence on the surface forcing where the backscattering of photons from surface emissions is enhanced due to the presence of dust particles in the range of ~0.5-10µm (fine + coarse-mode). Since this size range includes many of the particles found near source regions and in transported dust plumes, the LW scattering effect should be taken into account when modeling dust conditions. For example, Dufresne et al. [2002] show that neglecting the LW scattering produces an absolute error of ~5 Wm⁻² at the
surface or about a 15% reduction in the received surface flux; hence the LW forcing can be underestimated by up to 5 Wm$^{-2}$, a significant amount compared to the total (scattered + absorbed/emitted) LW forcing. In the current study, all possible sources of LW flux in the model simulations are considered due to the scattering and absorption/emission processes inside the dust layers.

The Langely modified Fu-Liou radiative transfer code (FL0403, April 15, 2003 - Rose et al. 2002; Fu and Liou et al. 1992) is used to calculate the LW surface fluxes for both dust and pristine conditions (refer to chapter 5 for model inputs). To calculate the radiative forcing, the TL96 dust climatology model (chapter 5) is used to represent the observed dust conditions versus a specific mineral dust model (e.g. pure quartz). The sensitivity studies for dust composition, size and shape however are performed using the dust models from chapter 6 which represent a range of specific dust parameters that are directly hardwired into the code. A best-fit with the observed surface fluxes (Fig. 9.10) was achieved using the TL96 dust model (within ± 5 Wm$^{-2}$), for accumulation-mode dust particles and therefore was selected for this study. The best fit is within the uncertainty of the PIR measurements (~2-3% or about ±10 Wm$^{-2}$).

The sensitivity studies of LW surface forcing used the dust models described in chapter 6 which included those for pure quartz, kaolinite, illite, montmorillinite, and an internal mixture of quartz mixed with 10% hematite. For each dust model, the single scattering properties were computed as an average over the Fu-Liou spectral bands (see table 5.1). Size and shape parameters were prescribed assuming spherically shaped particles in a lognormal size distribution with an $r_{\text{eff}}$ of 0.2, 0.75 and 1.5µm and $\sigma_g=1.55$-
1.9. The shapes were varied assuming the particles were oblate/prolate spheroids with aspect ratios of 1.4-2.3 and 0.4-0.7 respectively.

Capturing the changing atmospheric state of the lower boundary layer (< 4km) in the model is critical for accurately calculating the LW surface fluxes where ambient temperature ($T_a$) and relative humidity (RH) in the lower levels can drastically impact the received surface radiation. The corresponding weighting functions fall off significantly with height and do not contribute as much to the surface radiation. To accurately characterize the dynamical state of the atmosphere during the daytime AERI measurements, the AERIPLUS [Feltz et al. 2003] ambient temperature ($T_a$) and water vapor mixing ratio (WVMR) profiles derived from channels 1/2 of the AERI radiance data were used. The AERIPLUS retrieved profiles were compared to daily radiosondes launched at MAARCO to lend confidence in the method. Refer to fig. 9.11 for a comparison of AERIPLUS $T_a$ with MAARCO on September 22, 2004. The remaining levels in the atmospheric profiles (> 4km) were specified using the daily soundings from the MAARCO radiosonde (~4-20km) and the climatological mid-latitude summer profile (> 20km). The model atmosphere is defined over a total of 35 layers with a vertical resolution of about 1km up to ~30mb and 5km to the TOA.

Dust optical depths ($\tau_d$) in the model calculations were based on the AERI retrieved $\tau_d$ scaled from 10µm to 0.55µm using an IR to visible extinction coefficient ratio, $\beta_{\text{ext}}(10\mu m)/\beta_{\text{ext}}(0.55\mu m)$, discussed in chapter 8. The vertical distribution of $\tau_d$ was arbitrarily defined such that the bulk of dust occupied the lowest layers (~1-5km) consistent with observations. Beyond 5km, the dust loading is nearly zero. The heaviest
loading is closest to the surface and falls off with height.

Previous studies have suggested that radiative flux calculations are not sensitive to particle asphericity [Mishchenko 1993; Mishchenko and Travis, 1994] i.e. the spherical/nonspherical differences cancel out in the angular integration and that a spherical particle approximation using the Mie scattering code is sufficient for calculating the radiative forcing. More recent work, Kahnert et al. [2006] however shows the errors of assuming spherical particles can be comparable to the uncertainties in the refractive index of the component minerals and that particle asphericity can be important in calculating the radiative forcing.

9.2.3. Sensitivity of LW surface forcing to dust parameters

Studies were conducted to investigate the LW forcing sensitivity to the composition, size and shape parameters of dust particles. To assess the effects of mineral composition, six dust models were employed: quartz (Q), quartz internally mixed with 10% hematite (Q/H), illite (I), montmorillinite (M), kaolinite (K), and lastly the bulk Saharan (S) dust model. The composition is varied while keeping both particle size and shape constant. The dust particles are assumed to be spherical with the size parameters: \( R_{\text{eff}}=0.20 \mu\text{m}, \sigma_g=1.55 \) (accumulation-mode) and number density (N) of 100 particles cm\(^{-3}\) which roughly corresponds to the APS 3321 measurements at MAARCO during a large dust episode on September 12, 2004. The effective radius was chosen based on a goodness of fit with the PIR observations (± 5 Wm\(^{-2}\)). For each composition, the LW
forcing was calculated for the dust optical depths ($\tau$): 0.05, 1.0, 2.0, 4.0, 6.0, 8.0 and 1.0. A built-in desert-atmospheric profile was used, with a constant surface albedo of 0.2 over the 6-SW bands and an averaged surface emissivity of 0.98 over the 12-LW bands.

Figure 9.12 shows the sensitivity of the LW forcing to composition as a function of optical depth. Each curve is marked according to mineral type. As expected the forcing increases as optical depth increases however the functional relationship is not so linear but rather quadratic. This becomes more evident as particle size increases (not shown). Interestingly, the curves corresponding to quartz and the quartz/hematite mixture (asterisk and up-arrow respectively) exhibit the largest forcing compared with the clays by a factor of $\sim$1-3 for all optical depths. For example, in light background dust conditions (e.g., $\sim$\tau=0.1), a dust particle composed of pure quartz will have a LW forcing of $\sim$ 4 Wm$^{-2}$ compared with $\sim$ 1-2 Wm$^{-2}$ for the clays. In heavy dust conditions ($\sim$\tau=1), the quartz particle’s LW forcing increases to 28 Wm$^{-2}$ while that for the clays increases to $\sim$10-17 Wm$^{-2}$. This is due to the dominant absorption features found in quartz. Adding 10% hematite to quartz reduces the forcing by about 1 Wm$^{-2}$ since hematite exhibits weaker absorption in the LW, although is much stronger in the visible. For the particle size chosen, Saharan dust (right pointing arrow) also produces a rather large forcing due to the presence of quartz in the model. However, as particle size increases, the forcing of Saharan dust significantly decreases. It is thought that as dust particle size increases the mineral distribution is such that the clay species tend to dominate over quartz thus reducing the total absorption and thus the forcing. Illite and montmorillinite exhibit nearly the same amount of forcing with relative differences varying from $\sim$0.1-2 Wm$^{-2}$
over the range of optical depths 0.05-1.0 respectively. The exception among the clays is kaolinite whose forcing differs significantly from that of illite and montmorillinite by \( \sim 0.5-8 \) Wm\(^{-2}\) over the same range of optical depths. As particle size increases (\( R_{\text{eff}} = 1.5\mu\text{m} \): coarse-mode particles) the relative differences of kaolinite with the other clays increases even further to \( \sim 2.5-20 \) Wm\(^{-2}\).

To assess the effects of particle shape, the T-matrix light scattering code is used for computing the single scattering properties of both oblate and prolate spheroids (OS/PS) with variable aspect ratios AR = 1.4/2.3 and 0.4/0.7 respectively. The resulting LW forcing is then compared with that from a spherical model. The aspect ratios were chosen such that the sphere is compressed/stretched by the same amount along both axis directions (i.e. for the OS, ‘a’ is 1.4/2.3 times b and for the PS, ‘b’ is 1.4/2.3 times a).

The climatological Saharan dust model is used to constrain composition with the size parameters \( R_{\text{eff}}=0.2\mu\text{m}, \sigma_g=1.55 \) and \( 1.0\mu\text{m}, \sigma_g=1.9 \) for the same particle density of \( N=100 \) and as before the LW forcing is calculated over the same range of optical depths. Figure 9.13 shows the sensitivity of the LW forcing to shape as a function of optical depth. In the case of smaller particles at low optical depths (\( \tau \leq 0.1 \)), the relative differences are not so apparent, however as \( \tau \) increases, the forcing for non-spherical particles further deviates from that of a sphere where for heavier dust episodes (\( 0.5 \leq \tau \leq 1 \)), the absolute differences are on the order of \( \sim 1-3 \) Wm\(^{-2}\). It is noted that for OS (1.4), the forcing over all optical depths is smaller than that of a sphere and only becomes larger when the aspect ratio is further increased. For larger-sized particles, the relative differences become more evident particularly as \( \tau \) increases. Interestingly, the forcing for
the OS (1.4) and PS (0.7) shapes falls below that of a sphere (1-10 Wm\(^{-2}\) and 1-12 Wm\(^{-2}\) respectively) in the order of increasing \(\tau\), however particles with more extreme aspect ratios, have forcings that are higher than the sphere in the range of \(\sim (1-12 \text{ Wm}^{-2})\).

The resulting sensitivity study suggests that particle shape for the smaller submicron particles (accumulation-mode) may not be so significant under most dust conditions (light to normal background levels) however when \(\tau\) reaches values near unity the difference can be as high as \(\sim 3 \text{ Wm}^{-2}\) for particles with more extreme aspect ratios. As particle size increases (coarse-mode), the shape parameter becomes significant under most dust conditions and should be considered when computing surface fluxes. Even for very light dust background levels (\(\tau \leq 0.1\)), the non-spherical/spherical differences can be as high as 2-3 Wm\(^{-2}\) and can reach values of \(\sim 15-20 \text{ Wm}^{-2}\) for heavy dust conditions (\(\tau = 1\)) an amount that is certainly comparable to the uncertainties found in mineral composition. Based on the results of varying particle shape, we find similar conclusions reported by Kahnert et al. [2006] which suggest the errors associated with using a spherical approximation for dust particles can be quite significant comparable to that found in the particle’s refractive indices. Since this is particularly evident in larger-sized particles, it is recommended that extreme caution be used when calculating the surface fluxes of dust particles near the source region as this will have a direct effect on the LW forcing.

9.2.4. Forcing data and case studies

This study evaluates the LW surface forcing at the SMART site from 13-23
September, 2004. First examined is the temporal evolution of the instantaneous forcing for two dust cases from September 22 and 23 where relatively high dust loads were observed associated with the passage of convective thunderstorms that had developed over the nearby Oman Mountain range. These dust fronts (Haboobs-Miller et al. 2007) were likely triggered by the storm’s strong downdrafts. Lastly the averaged LW forcing was calculated during a 7-day period. For comparison, the LW forcing efficiency/forcing over the same period was determined employing the differential technique discussed in Hansell et al. [2003].

The two case studies were chosen based on observations from the on-site personnel who reported considerable dust activity. On both days, between 1000-1200UTC, the dust increased significantly over the background levels when the dust front reached the SMART site just minutes before a convective thunderstorm passed through the area. Unfortunately, the AERI was turned off and covered just before the dust event reached its peak intensity to prevent damage to the instrument’s fore-optics; nonetheless, AERI collected sufficient data to ascertain the radiative effects of the dust for both days. Approximately 4-6 hours of AERI data were recorded for the two events. Figure 9.14 shows the time series of meteorological measurements at SMART for both days with the light-gray boxed regions corresponding to the passage of the dust events. The surface measurements correspond to those typically observed during the passage of a thunderstorm’s gust front [Wakimoto, 1982], which caused the pressure and relative humidity to increase (0.5-2.0 mb and 5-15% respectively), the temperature to decrease (8-10°C), followed by subsequent precipitation. Recorded wind gusts measured at the
local meteorological station at Al-Ain Airport reached ~55 knots.

Figure 9.15 shows the time series of the instantaneous LW surface forcing for 9/22/04. The modeled LW surface fluxes (red) shown in Fig. 9.15(a) were computed using the AERI retrieved optical depths corresponding to the Volz oblate spheroid (VOS) and compact hexagon (VCH) dust models – refer to chapter 8. For comparison, the observed PIR fluxes (black) at SMART for the 24-hour period are also shown. The black vertical line around 1000UTC denotes the start of the dust event which caused a large spike in the measured LW surface flux of almost 40-50W m⁻² due to the increased emissions and reflectance of the dust particles. Overall, the model reproduces the measurements exceptionally well with an absolute error (model – observed) within ±5 W m⁻² [Fig. 9.15 (b)]. The computed LW instantaneous surface forcing is shown in Fig. 9.15(c) for both dust models where on average, the forcing is in the range of +3-6 W m⁻². This range of forcing corresponds to that predicted by Sokolik et al. [1997], i.e. from 2-7 W m⁻² for a low-dust loading scenario. In this case, both the average AERONET and AERI (IR-visible scaled) retrieved optical depths (τ = 0.55µm) over the 4-hour period is ~0.50 [Fig. 8.12(b)] which is taken to be consistent with the study’s ‘normal background’ dust conditions. It is suspected that had AERI been collecting data during the peak dust loads, assuming optical depths near unity, the projected forcing would have been ~8.0 W m⁻² based on the linear relationship between the LW forcing and optical depth given by fig. 9.16. For heavier dust loads (τ≥ 1), it is expected that the LW forcing could have exceeded 10.0W m⁻². The error bars depict the uncertainty in dust particle sizing after applying the IR-visible optical depth scaling factors discussed in chapter 8.
Similar to the magnitude of LW forcing from 9/22/04 was that on 9/23/04 which is shown in fig. 9.17. The dotted black vertical line around 1120 UTC denotes the start of the dust event. The modeled LW surface fluxes (red) shown in Fig. 9.17(a) were computed using the AERI retrieved optical depths corresponding to the Volz oblate spheroid (VOS) dust model – refer to chapter 8. For comparison, the observed PIR fluxes (black) at SMART for the 24-hour period are also shown. The model reproduces the measurements reasonably well with an absolute error (model – observed) within ±10 W m\(^{-2}\) [Fig. 9.17 (b)], still within the uncertainty of the PIR measurements. The instantaneous surface forcing is given in Fig. 9.17(c).

Employing a dust climatological model, the averaged instantaneous LW surface forcing for a 7-day period over which notable dust activity was observed (\(\tau \approx .45\)) was found to be \(~4.3 \pm 0.91\) Wm\(^{-2}\).

Lastly, the LW surface forcing efficiency was determined for a 7-day period during September 2004 using the method presented in Hansell et al. [2003]. Fig. 9.18 shows a scatter plot of the measured PIR flux versus AERI retrieved IR optical depth (chapter 8) for three levels of air mass ranging from 1-1.25, 1.29-1.39 and 1.4-2.0. Several sources of error may be present due to (1) PWV effects not accounted for in the measured PIR fluxes and (2) not enough data points for respective air mass bins, particularly with increasing air mass. By fixing the air mass interval, the forcing efficiency (forcing per unit optical depth) for this interval can be determined by a least-square fitting of total flux measurements versus the corresponding \(\tau_a\). The resulting slopes represent the LW forcing efficiencies. For an average range of LW optical depths
(~0.1—0.2), the equivalent forcing can be determined. For air masses 1-1.25, the forcing ranges from 2.65-5.3 Wm\(^{-2}\). For air masses 1.29-1.39 the forcing ranges from 1.96-3.92 Wm\(^{-2}\) and lastly for air masses 1.4-2.0, the forcing ranges from 1.47-2.94 Wm\(^{-2}\). These forcing values representative of the 7-day period are all very close to the instantaneous values shown in Figs. 9.15 and 9.17 lending confidence in the methods. The derived LW forcing efficiencies are ~0.28-0.5 times those derived for the SW [Markowicz et al. 2006] during the UAE\(^2\).

9.2.5 Summary

The AERI retrieved optical depths (chapter 8) were used in a broadband radiative transfer model for calculating the LW surface radiative forcing of dust aerosol. Employing the Volz dust model, the averaged instantaneous LW forcing for a seven-day period during the UAE\(^2\) was calculated to be ~4.3 ± 0.91 Wm\(^{-2}\) which was found to be within the range predicted by Sokilik et al. [1997] for surface forcing in relatively mild dust conditions, which for the UAE\(^2\) could be due to the normal ‘background’ dust conditions. Model calculations further suggest that the LW forcing for heavy dust conditions where visible optical depths exceed unity varies from about 8-20 Wm\(^{-2}\) again similar to those estimated by Sokilik et al. [1997]. Accounting for the estimated MCT non-linearity corrections, the resulting forcing differs at most by ~0.5 Wm\(^{-2}\). The LW forcing efficiency was computed over the same 7-day period employing the differential technique and the results indicate that the range of LW efficiencies are about 25-50%.
those calculated in the SW suggesting the LW forcing although smaller than the SW, can still be significant and should be considered in global climate modeling studies.
Table 9.1 Primary Instruments and Measurements for PRIDE, SAFARI and ACE-ASIA

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Parameters measured/derived</th>
<th>Spectral characteristics</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eppley PSP</td>
<td>Downward SW Irradiance</td>
<td>Broadband (0.2-4.0 µm)</td>
<td>+/- 5% (+/- 10 W/m²)*</td>
</tr>
<tr>
<td>YES MFRSR</td>
<td>Narrowband Direct, Diffuse and global SW Irradiance/τ_a and column water vapor and ozone</td>
<td>Center wavelengths (0.41, 0.49, 0.61, 0.67, 0.86 and 0.94 µm)</td>
<td>1% (Langley Calibration) With ~1% for τ_a retrieval</td>
</tr>
<tr>
<td>AERONET Cimel Sun-Photometer</td>
<td>Narrowband Direct, Diffuse radiance/τ_a, SSA, asymmetry parameter and volume size distribution</td>
<td>Center wavelengths (0.34, 0.38, 0.44, 0.50, 0.67, 0.87, 0.93 and 1.02 µm)</td>
<td>Sky radiance measurements (1%-3%) ** Direct radiance measurements (≤ 1%) ***</td>
</tr>
</tbody>
</table>

*When calibrated within 1 year to the World Radiometric Reference (WRR), **Reported accuracy of 2-meter integrating sphere¹, ***Reported accuracy for visible wavelengths, λ > 0.44 µm used in study
Fig. 9.1 (a) The arrangement of SW and LW radiometers at UCLA surface site. (b) Cimel 318A sun photometer at UCLA surface radiation site. (c) Example of downward SW flux profile on a clear day (January 20, 2000) at UCLA surface site. (d) Example of downward shortwave flux profile on a partially cloudy day (January 22, 2000) at UCLA surface site.
Fig. 9.2(a) PSP/CG1 nighttime output correlation. (b) Correction-Unventilated PSP nighttime signals.
Fig. 9.3 (a) Scatter ratio time series for July 8/9, 2000, PRIDE. The cutoff (dashed line) is at 0.0003. (b) Comparison of total solar fluxes with and without application of the cloud screening scheme for July 8, 2000, PRIDE.
Table 9.2 MFRSR Solar Spectral Constants

<table>
<thead>
<tr>
<th>λ  (µm)</th>
<th>0.41</th>
<th>0.49</th>
<th>0.61</th>
<th>0.67</th>
<th>0.86</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>PRIDE</td>
<td>1.7688</td>
<td>2.2821</td>
<td>1.7715</td>
<td>1.6601</td>
<td>1.0162</td>
<td>Level 2 Cimel</td>
</tr>
<tr>
<td>SAFARI</td>
<td>1.6611</td>
<td>2.2088</td>
<td>1.6251</td>
<td>1.5740</td>
<td>0.9680</td>
<td>Level 1.5 Cimel</td>
</tr>
<tr>
<td>ACE-ASIA</td>
<td>-----</td>
<td>1.9478</td>
<td>1.8111</td>
<td>1.7323</td>
<td>1.1689</td>
<td>Level 2 Cimel</td>
</tr>
<tr>
<td>ACE-ASIA</td>
<td>1.7674</td>
<td>2.1852</td>
<td>1.85220</td>
<td>1.60919</td>
<td>1.01072</td>
<td>Langley</td>
</tr>
<tr>
<td>AVG</td>
<td>1.7324</td>
<td>2.1559</td>
<td>1.7649</td>
<td>1.6439</td>
<td>1.0409</td>
<td></td>
</tr>
<tr>
<td>STD</td>
<td>0.0168</td>
<td>0.1448</td>
<td>0.0989</td>
<td>0.0687</td>
<td>0.0879</td>
<td></td>
</tr>
<tr>
<td>Thekaekara</td>
<td>1.7694</td>
<td>1.9492</td>
<td>1.6185</td>
<td>1.4415</td>
<td>0.9575</td>
<td>*1973 (Interpolated)</td>
</tr>
<tr>
<td>Neckel and Labs</td>
<td>1.7169</td>
<td>1.9064</td>
<td>1.7140</td>
<td>1.5092</td>
<td>0.9832</td>
<td>*1981 (Interpolated)</td>
</tr>
<tr>
<td>MODTRAN</td>
<td>1.6000</td>
<td>1.8800</td>
<td>1.6800</td>
<td>1.5400</td>
<td>0.9540</td>
<td>Version 4.0</td>
</tr>
</tbody>
</table>

Note: Coefficients have units of W/m²/nm
Fig. 9.4 (a) TOMS satellite image of African Continent on September 6, 2000 showing aerosol loading due to biomass burnings in savanna ecosystems. (b) Same as (a) except for September 7. The circles denote surface observation site at Skukuza, South Africa.
Fig. 9.5 (a) Retrieved 0.5 µm aerosol optical depths (Cimel/MFRSR) on September 6/7, 2000 – SAFARI. (b) Same as (a) except for May 2, 2001, ACE-ASIA. (c) Same as (a) except for July 19, 2000 – PRIDE.
<table>
<thead>
<tr>
<th>Campaign</th>
<th>$R^2$</th>
<th>AVG difference</th>
<th>RMS difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>PRIDE</td>
<td>0.80</td>
<td>0.016</td>
<td>0.004</td>
</tr>
<tr>
<td>SAFARI</td>
<td>0.93</td>
<td>0.001</td>
<td>0.006</td>
</tr>
<tr>
<td>ACE-ASIA</td>
<td>0.98</td>
<td>0.016</td>
<td>0.004</td>
</tr>
</tbody>
</table>
Fig. 9.6 Angstrom shaping factor using 0.87/0.67 μm channel pair of MFRSR and Cimel (SAFARI 2000).
Fig. 9.7 (a) Observed surface fluxes vs. Fu-Liou model results on September 6, 2000 (SAFARI). (b) Observed surface fluxes vs. Fu-Liou model results on September 7, 2000 (SAFARI). (c) Observed SARF vs. Fu-Liou model results at solar noon on September 6 and 7, 2000 (SAFARI).
Fig. 9.8 (a) Estimated aerosol forcing with MFRSR retrieved 0.5 μm aerosol optical depths (PRIDE 2000). Cloud screening applied with scattering ratio threshold of 0.0003. (b) Estimated aerosol forcing at 0.5 μm (SAFARI, 2000). Cloud screening applied with scattering ratio threshold of 0.0005. (c) Estimated aerosol forcing at 0.5 μm (ACE-ASIA 2001). Cloud screening applied with scattering ratio threshold of 0.0005.
Fig. 9.9 (a) Estimated forcing efficiency at 0.5 µm with slices of air mass 1.2, 1.5, 2.0, 3.0, 5.0 and 5.3. (b) Estimated forcing efficiency at 0.5 µm with slices of air mass 1.2, 1.5, 2.0, 3.0 and 5.0. (c) Estimated forcing efficiency at 0.5 µm with slices of air mass 1.1, 1.5, 2.0, 3.0 and 5.0.
Figure 9.10 (a) Model vs. observed LW surface fluxes for 9/22/04 during UAE$^2$ plotted as a function of observation number (b) Absolute differences between model and observed. Observed fluxes measured by SMART PIR.
Figure 9.11 (a) AERIPLUS vs. MAARCO ambient temperature profile for September 22, 2004 during UAE2. (b) Profile of residuals for (a).
Figure 9.12 Surface forcing sensitivity to mineral composition versus dust optical depth for dust particles with effective size of 0.2µm.
Figure 9.13 Surface forcing sensitivity to particle shape/size as a function of dust optical depth (a) effective particle size is equal to 0.2\(\mu\)m (b) effective particle size is equal to 1\(\mu\)m. See text for shape definitions.
Figure 9.14 Surface meteorology measured at SMART during UAE\textsuperscript{2} on September 22/23, 2004 (a) Pressure (b) ambient temperature and (c) relative humidity. Boxes denote times of dust periods with corresponding changes in meteorological variables.
Figure 9.15 Modeled vs. observed LW surface fluxes/forcing as a function of time for 9/22/2004 during UAE² at SMART site. Model fluxes correspond to the AERI retrieved IR optical depths using the Volz dust model for oblate spheroids (VOS) and compact hexagons (VCH).  (a) Model versus observed PIR fluxes. (b) Absolute error for both dust models. (c) Instantaneous LW surface forcing ($F_τ - F_{clr}$). Error bars denote the variability in both absolute (model) error and forcing due to uncertainties in particle sizing.
Figure 9.16  (a) In the VOS case, the average IR forcing (F) varies linearly with AERI retrieved optical depth (τ): F = 7.87τ - 0.026 W m^-2, so for a dust storm with unit optical depth, the IR forcing is nearly 8.0 W m^-2. In cases of severe dust loading (i.e. at or near the source region: τ ≥ 1), F can exceed 10 W m^-2. When τ = 0 (clear-sky), F should be ~ zero.  

(b) Residuals of linear fit with observations.
Figure 9.17 Modeled vs. observed LW surface fluxes/forcing as a function of time for 9/23/2004 during UAE² at SMART site. Model fluxes correspond to the AERI retrieved IR optical depths using the Volz dust model. (a) Model versus observed PIR fluxes. (b) Absolute error. (c) Instantaneous LW surface forcing ($F_\tau - F_{clr}$).
Figure 9.18   Dust LW surface forcing efficiency over 7-day period from 13-23 September 2004. Colors represent different air-masses and their corresponding slopes using the method presented in Hansell et al. [2003]. Units are in Wm$^{-2}$ per unit optical depth.
CHAPTER 10

10. CONCLUSIONS

To investigate the radiative effects of dust aerosol, detection/retrieval methods for dust events over major dust outbreak and transport areas have been developed using ground and satellite-based remote sensing approaches in the thermal IR window. To this end the surface radiative forcing of dust aerosol was investigated for both the SW and LW spectral regions.

Dust microphysical models were constructed using *in-situ* data from both the UAE$^2$ and prior field studies. Particle size and shape distribution parameters were specified from a ground-based APS-3321 aerodynamical particle sizer and electron micrographs of airborne collected filter data during the UAE$^2$ and PRIDE field campaigns respectively. Dust composition models were prescribed using refractive index datasets for minerals commonly observed around the UAE region including pure quartz, pure kaolinite and kaolinite internally mixed with hematite and calcium carbonate using the Bruggemann effective medium approximation. The Volz bulk dust mixture, a Saharan dust model, was also evaluated for comparison with the individual minerals. The differences in the refractive indices between minerals and liquid water and ice in the thermal IR widow form the basis for both ground and satellite based dust/cloud detection techniques. The single scattering properties for oblate spheroids and hexagonal plates, which are two particle geometries routinely interpreted in electron microscopy, were
computed using the T-matrix and FDTD light scattering programs.

Using the CHARTS RTM and dust microphysical models, detailed sensitivity studies to key dust and atmospheric parameters in the thermal IR window were conducted to examine the potential for developing the methodologies for detecting and retrieving dust aerosol using ground-based AERI spectra during the UAE$^2$. To facilitate the method’s sensitivity to dust aerosol, prescribed AERI sub-bands were employed to minimize contamination due to the continuum absorption of water vapor. Estimated errors on the retrieval due to the AERI MCT detector non-linearity were also evaluated.

A methodology was given for detecting/separating dust from cloud and retrieving the dust IR optical depths. The physical basis for the approach relied on the complex spectral variability of the IR optical properties for common mineral dust components. Dust detection followed the physical principles of dust and cloud particle absorption across the thermal IR window while the retrieval scheme employed a $\chi^2$ statistical optimization approach in the AERI ‘clean’ sub-bands for determining the dust IR optical depths. Unique to AERI detection was the $BTD_{10-11}$ dynamic threshold which exploits the differences in the absorptive properties of minerals and liquid/ice water for separating dust from cloud. A $BTD_{10-11}$ clear sky offset was also employed to account for the changing water vapor. Unique to AERI retrieval was the sub-band slope dependence of dust optical depth, particularly from 1100-1200 cm$^{-1}$ which was generally found to be independent of dust composition. To illustrate the feasibility of the detection/retrieval methodology, AERI data from several daytime UAE$^2$ cases were examined to: (1) test the method’s ability to successfully separate dust from cirrus cloud
under mostly cloudy skies and (2) retrieve dust optical depths during typical dust conditions. Estimated errors on the retrieval due to the MCT detector non-linearity were also evaluated. The results were then compared to collocated MPLNET micro-pulse lidar/AERONET sun-photometer measurements respectively. For daytime dust/cloud detection, AERI was able to capture the MPL dust/cloud profiles with some confidence. The same was true for nighttime detection where AERI accounted for 85% of the MPL observations during a NAMMA case study. Adding the estimated MCT detector non-linearity corrections to the AERI data did not affect the detection results.

The potential for using the AERI $BT$ spectra to retrieve dust IR optical depths was investigated employing a LUT with extended PWV and using AERIPLUS clear sky profiles. Comparisons of the AERI retrieved optical depths with those from AERONET indicate reasonable agreement generally within 30%. Both the Volz and quartz microphysical models containing oblate spheroid shape distributions generally yielded the best results for all cases examined. Adding the estimated MCT detector non-linearity corrections to the AERI data increased the magnitude of the retrieved optical depths by ~6-12%.

A similar study was conducted to investigate the potential for using MODIS thermal IR window band data to detect and separate mineral dust from cirrus clouds. An integrated method was presented which combined the $D^*$-parameter, $BTD$ slope, and $BTD_{3-11}$ methods for detecting dust and cirrus cloud for both nighttime and daytime scenes. By comparing the detection results to the MPL observations at each site, it was illustrated that the detection scheme can be used to detect dust and to separate dust from
cirrus during nighttime conditions with some confidence.

Lastly, the surface radiative forcing of dust aerosol was investigated for both the SW and LW spectral regions. An independent study of the dust SW effects was conducted employing a differential technique using ground-based PSP fluxes and MFRSR optical depths from three prior field studies: PRIDE, ACE-ASIA and SAFARI. The dust forcing efficiency and forcing was determined for several air mass intervals during each study. Similarly, the AERI retrieved IR optical depths from the UAE\textsuperscript{2} were used to determine the dust LW surface radiative forcing. The same differential technique was also applied to the LW dataset to calculate the forcing efficiency. Comparisons of the dust LW forcing with independent model calculations from previous research show reasonable agreement.

Both ground and satellite based remote sensing approaches show great promise for detecting/separating dust from cloud, particularly cirrus, during daytime and nighttime conditions. Potential applications which can benefit from this work include (1) scene classification for the dust radiative forcing studies, (2) dust aerosol correction for improved sea surface temperature retrievals and dust data assimilation in the coupled ocean-atmosphere model, (3) application of the detection results to the current operational daytime aerosol retrieval algorithm, and (4) assessment of the nighttime dust hazard to improve transportation safety and mitigate dust’s adverse health effects. The retrieval of dust IR optical depths using the AERI $BT$ data will be particularly useful for providing the dust optical parameters needed for estimating regional LW surface radiative forcing, which is currently a poorly understood quantity that needs to be better
constrained. Other potential applications for using AERI data include (1) identifying mineral composition which was found to be sensitive in the 2600-2700 cm\(^{-1}\) spectral region (2) improved tracking and monitoring of regional dust episodes to complement satellite-based detection methods and (3) the ability to provide for the detection and retrieval of nighttime dust episodes.
Bibliography


215


Reid, E.A., Reid, J.S., M. M. Meier, M.R Dunlap, S.S. Cliff, A. Broumas, K. Perry, and H. Maring (2003a), Characterization of African dust transported to Puerto Rico by individual


Shettle, E. P., and R. W. Fenn (1979), Models for the aerosols of the lower atmosphere and the


