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Impact of Airborne Dust on Sea Surface Temperature Retrievals

Alec Setnor Bogdanoff

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IMPACT OF AIRBORNE DUST ON SEA SURFACE TEMPERATURE RETRIEVALS

By

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A Thesis submitted to the
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I dedicate this manuscript in memory of my grandmother Bernice Setnor.
Her smile brightened everyone’s day.
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ABSTRACT

Sea Surface Temperatures (SSTs) are an important measure of our current weather and climate, as well as an essential variable in both short and long term weather forecasting. Infrared SST retrievals are reliant on passive sensors, and retrieval techniques are influenced by changes in atmospheric composition, including aerosols. Many empirically derived retrieval algorithms are based on matching Top of Atmosphere (TOA) Brightness Temperatures (BTs) from the Advanced Very High Resolution Radiometer (AVHRR) to buoy measurements during clear-sky conditions. Data is cloud-cleared to remove cloud-contaminated data. However, small, but influential, Aerosol Optical Depths (AODs) data may not be flagged as contaminated and the algorithms incorrectly calculate a cold SST due to the radiometer sensing the cooler, elevated aerosol layer temperature.

Many studies on aerosol effects on SSTs focus on aerosols due to volcanic eruptions. However, truly operational tropospheric aerosol corrections for daytime and nighttime retrievals have yet to be implemented. This work constitutes a first step to creating an accurate aerosol correction by exploring the sensitivity of aerosols on SSTs.

The Santa Barbara DISORT Radiative Transfer model is used to quantify the effects of aerosol contamination on retrieved TOA BTs. The calculated radiances are spectrally averaged over each channel, converted to BTs, and used to calculate an SST using the Naval Oceanographic Office AVHRR algorithms. A radiative transfer model is used to evaluate the SST retrieval error due to varying AOD, height of an aerosol layer, and the satellite zenith angle (or viewing angle).

This analysis shows that errors greater than the stated retrieval uncertainty of 0.5 K are observed for AODs greater than 0.25. Two sites with state-of-the-art aerosol measurements are analyzed for AOD variability. The first site, at Anmyon in east Asia, is found to have 14% of the days during the springtime with an AOD greater than 0.25. Based on the AERONET data from a second site in Cape Verde, 65% of the days during the boreal summer are found to have AOD greater than 0.25. Unfortunately, this seasonal peak in dust activity coincides with the active tropical cyclogenesis season for the region, making accurate SSTs even more vital for prediction purposes.
CHAPTER 1

INTRODUCTION

With more than 70% of the globe covered by water, understanding the interaction between the atmosphere and ocean is essential to understanding our climate. The entire depth of the atmosphere contains as much heat capacity per unit area as the upper 2.5 meters of the ocean; therefore, the upper ocean is an important thermal storage mechanism for the climate (Gill 1982). Currently, remote sensing techniques are the only accurate methodology to obtain a high-resolution global snapshot of environmental conditions near the surface of the Earth. Given that satellites are unable to directly observe the structure of the ocean below the surface, Sea Surface Temperature (SST) is one of the best metrics for evaluating the current condition of the global oceans. Satellites provide a better method for global analysis of SST variability on shorter time and space scales than can be resolved by current buoy and ship networks.

While air-sea interaction plays an important role in the global climate, the time scales of the variations in the two fluids differ greatly. SSTs are affected both by the slow variability of the ocean and the relatively faster variability of the atmosphere. SST impacts the current weather and climate, as well as being an essential factor in both short and long term weather forecasting. Specifically, SSTs impact the surface radiation budget, latent heat flux, and sensible heat flux in the surface energy balance (Curry et al. 2004). On long time scales the atmosphere is sensitive to changes in the SST on the order of 1 K or less. For example, the Atlantic basin can become favorable for tropical development with slight (0.5 K) increases in SST (Evan et al. 2008 conclusions from Kossin and Vimont 2007). Depending on the desired use of an SST measurement, different accuracies are required. Numerical Weather Prediction models and mesoscale oceanographers desire accuracies of 0.1 – 0.3 K (Gentemann et al. 2009). Climate research users desire accuracies less than 0.1 K for time series greater than 10 years in length (Vazquez and Heinz 2007). Therefore, it is vital to obtain the best possible estimate of the SST from satellites.
The most commonly used satellite instrumentation for SST datasets is the Advanced Very High Resolution Radiometer (AVHRR). The AVHRR sensor is among the instruments on the National Oceanographic and Atmospheric Administration (NOAA) polar-orbiting environmental satellites, and more recently the European Organisation for the Exploitation of Meteorological Satellites (EUMETSAT) MetOp satellites (Leslie 2005). An advantage of this platform is the continuous availability of data from 1981 until present (Kilpatrick et al. 2001). Another advantage of using infrared sensors for SST retrievals is the ability to have higher spatial resolution (up to 1.1 km on AHVRR) than is currently possible from microwave sensors (roughly 25 km). However, SST retrieval techniques using infrared sensors, such as the AVHRR, are influenced by changes in atmospheric composition, including aerosols (Walton et al. 1998). Aerosol biases in SST retrievals, especially for large aerosol events (e.g. volcanic eruptions, airborne dust from deserts), impede an infrared sensor’s ability to accurately measure surface variables due to enhanced attenuation of the surface signal and emission of the aerosol layer. Additional complicating factors include the variability in attenuation that occurs depending on the type of aerosols and their height within the atmosphere (such as stratospheric or tropospheric). Volcanic aerosols have been extensively studied (discussed in Section 2.2) since the eruption of El Chichón in 1985 and Mount Pinatubo in 1991. The error characteristics associated with dust aerosols, such as those found off the Atlantic African coast, have recently become of greater interest to the scientific community.

SST variability off the West African coast is of importance to the regional, and potentially the global climate (Foltz and McPhaden 2008; Evan et al. 2009). In addition, this area is an important region for tropical cyclone development known as the Main Development Region (MDR). Anomalously high SSTs within the MDR can make conditions more favorable for tropical cyclone development and are linked to intensification, making it paramount to have accurate measurement of SSTs within this region (Evans 1993). Airborne tropospheric dust aerosols, associated with the Saharan Air Layer (SAL) complicate retrievals in this region. Dust is a solid particle approximately 10 microns in diameter created by mechanical disintegration (Hinds 1999). Dust associated with the SAL is lifted into the atmosphere from the Saharan Desert. The boreal summer is the active dust season for the tropical Atlantic Ocean, corresponding to tropical cyclogenesis season, further discussed in Chapter 5. In the SAL, the dust is located in a well-mixed layer above the boundary layer at an altitude generally below 6
km (Karyampudi et al. 1999). Thus, SST retrievals in this region are likely to be strongly affected by the presence of aerosols. Rao (1992) found depressions in individual retrievals ranging up to 4°C due to aerosols, but stated the need for a more comprehensive analysis of troposphere dust phenomenon such as the Saharan Air Layer, as done in this study. In addition to the well-known dust outbreaks off the western African coast, outbreaks in East Asia carry substantial dust off the Pacific coast creating a similar situation of large clouds of dust that impede the accuracy of infrared sensors. East Asia will be another area of focus. However, more emphasis will be placed on African dust and tropical Atlantic Ocean.

The difficulty of using satellite retrievals for determining a surface property should not be underestimated. Schollaert et al. (2003) note, “separat[ing] the atmospheric signal from the much smaller water-leaving radiance signal is a difficult problem...” With satellites measuring surface values from more than 800 km above the surface of Earth (Robel 2009), the extreme challenges in calculating an SST are examined. These difficulties lead to current accuracies lower than the desired accuracies of the geoscience community, and thus investigation into potential methods for improving retrievals is necessary. Although aerosols are only one source of potential errors, it is a challenging problem that requires attention. While researchers are exploring the radiative effects of dust aerosols on the tropical Atlantic (e.g. Foltz and McPhaden 2008; Evan et al. 2009), the inaccuracy of SST retrievals due to aerosols is merely noted as a possible source of error or ignored entirely. By introducing methodology to correct for the biases introduced by aerosols, the accuracy of SSTs remotely sensed by AVHRR can be greatly improved.

The primary goal of this research is to theoretically understand and quantify the impact of airborne dust on infrared satellite retrievals used to derive an SST. This study focuses on the tropical North Atlantic and West Pacific dust events. Due to the limitation of being able to accurately remotely sense aerosols during the daytime only, this study will focus on daytime SST retrievals. Since MetOp-A serves as the primary daytime sensor (Robel 2009), the MetOp-A AVHRR sensor is used as the basis for our study. Furthermore, a better understanding of the impact of dust on SST retrievals will assist in developing historical corrections to the almost 30 years of AVHRR SST products. It is the desire of the Naval Oceanographic Office (NAVOCEANO) and generally of all producers of SST products to use the greatest number of retrievals as possible over the globe from AVHRR to produce the best possible SST product. Currently, NAVOCEANO flags potentially aerosol contaminated retrievals and does not use
them in their SST analysis product. It would be beneficial to correct aerosol related biases, rather than just discarding retrievals that may be aerosol contaminated. The end results of correcting for these biases will be greater coverage and more accurate retrievals.

A secondary goal of this study is to investigate the feasibility of an accurate correction method for climate and operation. In order to properly create a correction to SST retrievals, it is necessary to first fully quantify the impact of airborne dust on retrievals. A radiative transfer model approach is used to examine the error due to a dust layer. The amount of aerosol and height of the aerosol layer is varied in modified SAL and eastern Asia frontal soundings. The radiance output from the model is converted to an SST, and this SST is compared to the known SST inputted into the radiative transfer model. The difference between the computed SST and the known SST is considered the error due to dust. The rest of the thesis has the following structure: further background on SST retrievals and previous corrections are found in Chapter 2. The methodology of the sensitivity study is outlined in Chapter 3. The results of the sensitivity study are presented in Chapter 4. The seasonal cycle of dust in the SAL and East Asia is explored in Chapter 5. Chapter 6 provides in situ and satellite matchups to examine the presence of errors induced by aerosol contamination. Lastly, conclusions and future work are presented in Chapter 7.
CHAPTER 2
BACKGROUND

2.1 Sea Surface Temperature Retrieval Theory

The first generation of AVHRR operational SST algorithms are a linear multi-channel empirically derived parameterization described by McClain et al. (1983), based on the atmosphere-integrated form of the Schwarzchild equation. The Schwarzchild equation is an important solution to the radiative transfer equation, which characterizes the atmosphere by a mean temperature, $T$, and an optical depth, $\tau$ (Martin 2004). An optical depth can be computed for any atmospheric constituent, including the aerosol component of the atmosphere.

The amount of aerosol present in the atmosphere is measured using a non-dimensional Aerosol Optical Depth (AOD). AOD is calculated by monochromatically integrating the aerosol extinction coefficient, $b \, (m^{-1})$, over the depth of the atmosphere. Mathematically, AOD is

$$\tau_a(\lambda) = \int_{z_0}^{z_{TOA}} b(z',\lambda)dz'$$

(2.1)

where $\tau_a$ is the aerosol optical depth integrated from the surface to the TOA and a function of wavelength only (Seinfeld and Pandis 2006). As dust impacts a broad range of the visible and thermal spectrum, it can be detected during the daytime by changes in albedo via visible channels. However, at night, the lack of availability of visible channels requires different methodologies (Merchant et al. 2006). This study focuses on daytime SST retrievals, although the methodology (Chapter 3) can be used to examine the impact of aerosols during day and night.

Noting the optical depth as a function of the zenith angle due to increasing path length with increasing zenith angle, the Schwarzchild equation is written as

$$L(\lambda_i,z_{TOA}) = L(\lambda_i,z_0)e^{-\tau(\lambda_i,\rho)} + f_p(T,\lambda_i)(1-e^{-\tau(\lambda_i,\rho)})$$

(2.2)
where \( L(\lambda_i, z) \) is the radiance (Wm\(^{-2}\)sr\(^{-1}\)) at a specific wavelength or wavelength band, \( \lambda_i \), height, \( z \). \( \theta \) is the zenith angle, \( z_{TOA} \) is the height of the top of the atmosphere, and \( z_0 \) is the ocean surface. The left hand side of the equation is the radiance at the top of the atmosphere, and is measured by the satellite. The first term on the right hand side of Equation 2.2 is the surface radiance that has been attenuated by the atmosphere. The second term on the right is the atmospheric emission (Martin 2004). \( f_p(T, \lambda_i) \) is the Planck function at temperature, \( T \), and at a specific wavelength or wavelength band:

\[
f_p(T, \lambda_i) = \frac{2hc^2}{\lambda_i^5 \left( e^{hc/\lambda_iKT} - 1 \right)}.
\]  

(2.3)

The transmittance of the atmosphere \( e^{-\tau(\lambda_i, \theta)} \), can be approximated by a first-order Taylor series as \( 1 - \tau(\lambda_i, \theta) \). Approximating the radiances with the Planck function, which is appropriate for the infrared portion of the spectrum being explored, Equation 2.2 can be rewritten following McClain et al. (1985):

\[
f_p(T_i, \lambda_i) = f_p(T_{sfc}, \lambda_i)[1 - \tau(\lambda_i, \theta)] + f_p(\bar{T}, \lambda_i)\tau(\lambda_i, \theta).
\]  

(2.4)

\( T_i \) is the equivalent blackbody temperature sensed by the satellite (Petty 2006). \( T_{sfc} \) denotes the temperature of the ocean surface and \( \bar{T} \) is the mean effective blackbody temperature of the atmosphere.

Further simplifying the Equation 2.4 by dropping the \( \lambda_i \) and \( \theta \) notation and rearranging the equation to group the terms with optical depth together, the linearized approximate solution to the equation of radiative transfer is:

\[
f_p(T_i) = f_p(T_{sfc}) + \left[ f_p(\bar{T}) - f_p(T_{sfc}) \right] \tau.
\]  

(2.5)

Optical depth can be parsed into an aerosol portion and the remainder of the atmospheric constituents: \( \tau = \tau_{atm} + \tau_{aero} \). Specifically, \( \tau_{atm} \) is the optical depth of the atmosphere without the aerosol component, \( \tau_{aero} \) is explicitly the AOD. Expanding Equation 2.5 and substituting the mean temperature of the atmosphere term with the mean temperature of the aerosol layer for the AOD term, Equation 2.5 may be rewritten in a two-layer approximation as

\[
f_p(T_i) = f_p(T_{sfc}) + \left[ f_p(\bar{T}) - f_p(T_{sfc}) \right] \tau_{atm} + \left[ f_p(\bar{T}_{aero}) - f_p(T_{sfc}) \right] \tau_{aero},
\]  

(2.6)
where $T_{aero}$ is the average temperature of the aerosol layer. Since the radiative temperature of an aerosol layer may greatly vary from the mean atmospheric temperature, for the aerosol term (third term on the right hand side), a mean aerosol temperature is most appropriate. The second term on the right hand side may be neglected for high AOD cases due to the overwhelming aerosol contribution although for completeness it is included here.

Equation 2.6 provides a good approximation for thermal infrared radiative transfer, and provides a physical explanation for the impact of dust or any aerosol on the satellite retrieval of an infrared signal from the surface of Earth. When both optical depth terms are zero, which assumes the atmosphere does not attenuate or emit radiation, the temperature measured by the satellites will be the blackbody temperature at the surface. Equation 2.6 physically demonstrates the cooling impact of dust aerosols in the atmosphere. The third term of the right hand side will be negative given the temperature of the surface is greater than the mean temperature of the aerosol layer. The greater the temperature difference between the mean atmosphere and surface, the greater the impact on the BT calculated at the TOA. Furthermore, the greater the optical depth of the atmosphere, the greater impact the temperature difference will have on the TOA BT calculation.

### 2.2 Sea Surface Temperature Definition

The term sea surface temperature is quite ambiguous and can define many temperatures in the upper ocean (Donlon 2001). The upper ocean temperature variability throughout the profile is on the order of several degrees and varies diurnally as well (Figure 2.1). The change in SST ($\Delta$SST) noted in the figure corresponds to the difference from temperature at the depth where diurnal variation is no longer present. The temperature at which diurnal variation is no longer present is strictly defined as the Foundation SST (GHRSSST Science Team 2010). Kawai and Wada (2007) note that in extreme cases this daytime $\Delta$SST can be more than 5 K. Donlon et al. (2001) was among the first to note the essential concept of providing specific depths at which the SST is measured, a point further discussed by Clayson and Weitlich (2005), Donlon et al. (2002), Kawai and Wada (2007), and Gentemann et al. (2009).
Many current SST climatologies focus on a bulk or foundation temperature, commonly derived from bucket ship or buoy measurement (e.g. Reynolds Optimally Interpolated SST V2; Reynolds et al. 2007). The term bulk SST generally describes any temperature in the region where turbulent processes dominate, which may be on the order of several meters. Table 2.1 provides further definitions from the Group for High Resolution Sea Surface Temperatures (GHRSST), an international science group focused on providing useful and accurate SST products. It is evident that the temperature measured by infrared satellites is not the same as the temperature measured by in situ instruments, such as buoys. This study uses different SSTs depending on the instrumentation or model used. The SSTs used throughout this study are generally temperatures very close (~ 10 µm to 1 mm depth) to the surface and therefore, diurnal variation should not be a substantial source of error. However, potential diurnal bias in the SST algorithms is discussed in Section 3.5. The $T_{sfc}$ first defined in Equation 2.4 is the Interface SST, which is theoretical, but generally very similar to the Skin SST measured by infrared radiometers.

Figure 2.1 – A nighttime (left) and daytime (right) example of the temperature profile of the upper ocean, demonstrating the cool near-surface skin layer and the potential diurnal variability under low wind speed conditions. Adapted from Donlon et al. (2002) and Gentemann et al. (2009).
Table 2.1 – The five SSTs defined by the Group for High Resolution Sea Surface Temperature (GHRSST). Adapted from Donlon et al. (2001), Kawai and Wada (2007), and GHRSST Science Team (2010).

<table>
<thead>
<tr>
<th>Temperature Type</th>
<th>Depth Represented</th>
<th>Instrumentation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Interface SST</td>
<td>At the infinitely small air-sea interface layer. It is a theoretical temperature</td>
<td>There are no instruments to measure this temperature</td>
</tr>
<tr>
<td>Skin SST</td>
<td>Within layer where conductive and diffusive processes dominate</td>
<td>Infrared radiometers operating at wavelengths 3.7 to 12 µm, measuring approximately 10-20 µm below the air-sea interface</td>
</tr>
<tr>
<td>Subskin SST</td>
<td>Where molecular and viscous processes begin to dominate</td>
<td>Microwave radiometers operating in the 6 to 11 GHz frequency range, measuring 1 – 2.5 mm below the air-sea interface</td>
</tr>
<tr>
<td>Subsurface SST (or SST depth)</td>
<td>Any point below the subskin SST where turbulent processes dominate</td>
<td>Buoys, Ships, etc. Rather than a “bulk” temperature, which has an ambiguous depth, this temperature should be measured at a specific depth (i.e. 5 m)</td>
</tr>
<tr>
<td>Foundation SST</td>
<td>Point at which a diurnal cycle is no longer present</td>
<td>Buoys, Ships, etc.</td>
</tr>
</tbody>
</table>

2.3 Sea Surface Temperature Retrievals

To provide a simple relationship between the TOA BTs and SST, a further simplification of Equation 2.6 (neglecting aerosols, \( \tau_{aero} = 0 \)) yields the “split window” equation, which breaks the solution of radiative transfer into terms corresponding to the channels of the satellite sensors and is the basis of many infrared SST algorithms. From the “split window” equation, the Multi-Channel SST (MCSST) algorithm is empirically derived of the form,

\[
SST = A_1T_4 + A_2(T_4 - T_5) + A_3,
\]

where \( T_4 \) and \( T_5 \) are the brightness temperatures of channels 4 and 5 of the AVHRR retrieval, respectively, and \( A_1, A_2, \) and \( A_3 \) are empirically derived constants from buoy matchups (McClain et al. 1985).
In general, the SST retrieval algorithms previously described neglect reflection of downwelling atmospheric radiance, ignore aerosols, and assume that any variations in transmittance in the lower troposphere are due only to water vapor (Martin 2004). Neglecting the aerosol term in Equation 2.6, the assumption is made that the non-aerosol portion of the optical depth only changes due to water vapor. Furthermore, there is no term in the equation to account for downwelling atmospheric radiance, only upward. More sophisticated nonlinear algorithms still have the same assumptions as made in Equation 2.7, but take into account the zenith angle (or viewing angle) of the satellite and therefore, the changing optical depth of the atmosphere based on path length. Below is the form of the nonlinear MCSST AVHRR algorithm by the Naval Oceanographic Office for daytime:

\[
SST = A_1 T_4 + A_2 (T_4 - T_5) + A_3 (T_4 - T_5)(\sec \theta - 1) - A_4, \tag{2.7}
\]

where \( T_4 \) and \( T_5 \) are the BTs of channels 4 and 5 respectively, and \( A_1, A_2, A_3, \) and \( A_4 \) are empirically derived constants (May et al. 1998). The algorithms are solely a function of the viewing angle (or zenith angle) of the satellite, and the BT of channels 3B, 4, and 5 of the AVHRR (channel 3B is generally neglected during the daytime because it is near infrared).

Table 2.2 displays the six channels of AVHRR, although 3A and 3B cannot be used simultaneously (Robel 2009). Channels 1 and 2 are not used for sea surface temperature retrievals because they are visible channels, and Equation 2.1 is not a good approximation in the visible region of the electromagnetic spectrum. The empirically derived constants are generally based on matching Top of Atmosphere (TOA) BTs from the AVHRR sensors to buoy
measurements during clear-sky conditions. As such, the term $SST$ is used in place of $T_s$ (since the temperature may not be directly at the surface) due to the different depths at which infrared sensors and buoys measure temperature, described in Table 2.1 (Martin 2004).

Dust is not the only source of error. SST retrievals from infrared sensors are susceptible to many error sources. These sources include deficient cloud removal techniques, incorrect atmospheric transmission representations due to faulty physics or incorrect profile information or both, incorrect handling of the issues associated with the varying temperature profile, and sun glint (Figure 2.2). Each of these effects must be considered in the retrievals in order to minimize error.

AVHRR sensors are sensitive to cloud cover, so cloud-contaminated data are discarded (May et al. 1992). Many retrieval methods assume that the cloud-clearing algorithm used also removes all pixels containing enough aerosols to affect the retrieval. However, small, but influential AODs (generally measured at 0.55 $\mu$m) may go unnoticed during cloud-clearing. If

Figure 2.2 – The atmospheric and oceanic processes and constituents involved in infrared sensor retrievals. Sources of potential error are given as well. Image from Martin (2004).
these pixels are not removed, the SST retrieval will be colder due to the radiometer sensing the elevated, cooler aerosol layer temperature (Nalli and Reynolds 2006; Reynolds et al. 2007). As described in Section 2.1, the aerosol layer has a two-part impact to inhibit an accurate SST retrieval. The layer emits an infrared signal that contaminates the surface signal, and causes enhanced attenuation of the surface signal.

Although this study focuses on dust aerosols, another substantial source of error for satellite-derived SSTs is the potential difference between skin SST measured by infrared satellites and the temperature at which a buoy or ship measures, for those retrieval techniques that regress the satellite-measured SST to a buoy or ship measurement. These errors can exceed 5 K during strong diurnal warming events (Kawai and Wada 2007). Buoys are a common in situ measurement used not only for matching TOA BTs for empirically derived constants for SST algorithms, but also in quality control and analysis (Nalli and Reynolds 2007). A careful analysis of errors must include errors due to differing depths of measurements and accompanying variability in the upper ocean temperature profiles due to diurnal or synoptic variations. Moreover, during the daytime, the near infrared channel of AVHRR is not used in the SST algorithms due to potential sun glint errors (May et al. 1998). The use of sun zenith angle filters for differentiating daytime and nighttime retrievals, and the neglecting of the near infrared channel during daytime limit the possibility of error due to sun glint.

### 2.4 Previous Aerosol Corrections to SSTs and Retrievals

Aerosol corrections to SST retrieval algorithms are, in general, not a new methodology. However, the main focus of aerosol corrections has been volcanic aerosols as opposed to dust aerosols, partly because the colder stratospheric aerosols create a more noticeable impact on the retrieval of SSTs rather than the relatively warmer dust layers. Another difference between the stratospheric and tropospheric aerosol is the timing of the input events and residence time of the aerosols. Nalli and Reynolds (2006) note that although volcanic eruptions are sporadic, the residence time of the aerosols injected into the stratosphere can range from months to over a year. In contrast, dust aerosols in the troposphere have much shorter residence times. As an example, Mount Pinatubo erupted in 1991 creating noticeable negative biases in SST anomalies for many months (Figure 2.3). The difference in day and night biases in the figure is due to the
ability to more easily flag the volcanic aerosol for cloud clearing techniques during the daytime using the two AVHRR visible channels (Reynolds 1993).

The magnitude of the error, the near global coverage of the aerosols, and the long length of time of these errors caused researchers to focus on corrections associated with stratospheric aerosols. Griggs (1985) and Walton (1985) were among the first to pioneer an aerosol correction to SST retrievals after the eruption of El Chichón in April 1982. The Griggs correction to the visible AVHRR channels was solely a function of the AOD (obtained from satellite observations) and the satellite viewing angle, which worked well as an initial gross estimate of the aerosol effect. Walton (1985) provided an alternate method by attempting to create a retrieval algorithm less sensitive to stratospheric aerosol and more sensitive to moisture impacts and, unfortunately, noise. The algorithm correction for volcanic aerosols of Walton is reexamined after the eruption of Mount Pinatubo in 1991 and noted in Reynolds (1993). Due to the

Figure 2.3 – SST anomaly (from climatology) time series for daytime (solid), nighttime (dashed) and in situ (solid line with circles) for 55 weeks over 20°N to 20°S. Image from Reynolds (1993).
familiarity with the impact of El Chichón on infrared sensors, the corrections to SST errors due to Mount Pinatubo were timelier. Correcting large-scale biases by updating algorithms through comparison with in situ data before performing optimum interpolation allowed for the retrievals to be useful and more accurate. Studies into the effect of volcanic aerosols on AVHRR retrievals still reveal significant negative bias, as much as 2.0 K, in areas impacted by the eruption for as long as two years (e.g. Vargas et al. 2009; Walton 1998), even with extensive work performed to eliminate volcanic aerosol bias.

May et al. (1992) introduced a correction for dust effects on SSTs; this study builds on that correction by providing a more comprehensive assessment of the impacts of dust. Currently, the Naval Oceanographic Office flags and does not use SSTs considered aerosol contaminated by quality control measures (Bruce McKenzie, personal communication, 2010). Other modern corrections include statistical methods such Merchant et al. (1999), Diaz et al. (2001), Nalli and Stowe (2002), Vázquez-Cuervo et al. (2004), Nalli and Reynolds (2006). However, to the author’s knowledge, a robust operational correction for daytime and nighttime retrievals has yet to be implemented due to the lack of a real-time AOD dataset and thorough understanding of the impact of tropospheric aerosols on SST retrievals from infrared sensors. The Navy maintains an operational aerosol transport model, the Navy Aerosol Analysis and Prediction System (NAAPS) that will eventually be explored as the operational source of AOD information. Merchant and Le Borgne (2004) use a radiative transfer model approach to create a physical retrieval of SSTs. The use of these physical retrievals provides greater perception of the retrieval process and provides a methodology to explore retrieval errors, making it the ideal approach for use in this study.
CHAPTER 3

METHODOLOGY

The methodology outlined in this chapter is a radiative transfer model (RTM) approach for exploring the impact of aerosols on infrared sensors. Although this study specifically focuses on infrared sensors, this approach can be used to investigate any portion of the electromagnetic spectrum limited only by the wavelength range of the selected RTM. Therefore, this physically based approach can be used for any sensor type. It does not limit the aerosol properties nor limit the selection of methodologies for evaluating AOD.

To examine the impact of airborne dust on infrared satellite retrievals, the procedure outlined in the flowchart of Figure 3.1 is used. The atmospheric profile of temperature, moisture, ozone, and dust are input into the radiative transfer model along with the aerosol optical properties. The model is run over the windows of Channels 4 and 5 of the AVHRR sensor, which is used for daytime SST retrievals. For the analysis, the Santa Barbara DISORT Radiative (SBDART) Transfer model is utilized. The model outputs a top of atmosphere radiance at each wavelength interval for a particular channel, which is converted to a spectrally averaged brightness temperature using the spectral response function of each channel. A further description of SBDART is provided in Section 3.1. The NAVOCEANO MetOp-A Daytime SST algorithm provides the mechanism for the SST calculation, which is then compared to the inputted or known SST.

The use of a radiative transfer model allows for the examination of SST changes due to varying AOD, the vertical distribution of an aerosol layer, the vertical location of an aerosol layer, and the satellite zenith angle (or viewing angle). Six vertical distribution types have been created based on observed vertical distributions of dust. Five different vertical placements of dust are tested (two for SAL and three for Eastern Asia), creating 28 different cases. Each case is run for 28 different AODs over the window of both Channels 4 and 5, providing 1,568 different model runs.
Figure 3.1 – Flowchart overview of the methodology used for this sensitivity study.
3.1 Radiative Transfer Model

The Santa Barbara DISORT Atmospheric Radiative Transfer model, further described by Ricchiazzi et al. (1998), is a FORTRAN program written to explore radiative transfer problems in radiation budget calculations and remote sensing. SBDART was selected due to the flexibility of changing aerosol properties, including the optical properties, as well as the output flexibility. The user can provide a sounding with self-defined levels, or use one of the six climatological soundings provided in the model. For this study, the soundings described in Section 3.2 are used. DIScrete Ordinate Radiative Transfer (DISORT) is used to numerically integrate the equation of radiative transfer for the atmosphere. It is a numerically stable algorithm to solve an inhomogeneous plane-parallel atmosphere. It is important to note that the treatment of aerosols is not perfect or most physically correct, but the error due to uncertainty in the optical information of dust will be greater error due to the treatment of absorption, scattering, and extinction within the model, with model errors typically on the order of milliwatts (Ricchiazzi et al. 1998).

SBDART allows for the inclusion of user defined aerosol optical properties and atmospheric profiles, as previously noted. The solar zenith and solar azimuth angles are set to 0° for all runs for consistency. To explore changes in path length (and therefore effective aerosol optical depth) due to the satellite viewing or zenith angle, all zenith angles between 0° and 55°, incrementing by 5°, are modeled. The satellite zenith angle is the nadir angle from the satellite facing Earth’s surface defining the location in a swath of an instrument. The radiance at each specified atmospheric layer inputted can be outputted at specified wavelengths. Therefore, the highest level (~approximately 10 mb for the soundings) used in this analysis is considered to be the TOA radiance. The TOA radiance is calculated by the radiative transfer model over the 1.0 μm spectral range of Channels 4 and 5 of AVHRR every 0.02 μm corresponding to the spectral response functions for MetOp-A provided by Robel (2009). The MetOp-A satellite AVHRR is used because it currently serves as the AM (morning time) Primary satellite.

3.2 Input Variables

Input Soundings

Two modified soundings are generated from measured atmospheric profiles during the dust season from both the Saharan Air Layer region (Sal Island, Cape Verde - GVAC on
06/02/2009 at 1200 UTC; 16.73°N, 22.95°W) and a profile near the coast of Eastern Asia (Qingdao, China - ZSQD on 03/15/2009 at 1200 UTC; 36.06°N, 120.33°E). Modified soundings from over land are used to eliminate the use of modeled soundings for this study. Soundings over the open ocean are generally nonexistent. Therefore, actual soundings are used to replicate the actual atmosphere as close as possible. The temperature profiles in the boundary layer of the soundings are adjusted to reduce discontinuities between the SST in that region and the surface temperatures over land, since the soundings are taken over land. Figures 3.2 and 3.3 are Skew-T diagrams of the modified soundings for the Atlantic Ocean case and the Eastern Asian case, respectively. Adjusting the sounding did not have any substantial impact on the results of the study, and henceforth the study uses the adjusted soundings.

The Saharan Air Layer sounding has a superadiabatic surface layer, with the rest of boundary layer being well-mixed and nearly adiabatic. The quintessential signal of the SAL is a well-mixed nearly adiabatic layer in which the dust is generally present, with the strong inversion above the boundary layer. The first case study explores the impact of dust located in the Saharan Air Layer. Using the same sounding, a second case study explores dust located in the boundary layer, another common vertical location of dust found in the tropical Atlantic region. The East Asian sounding displays a slight surface inversion below a well-mixed adiabatic layer. This sounding was taken shortly after the passage of a cold front that had relatively strong wind that carried the dust to the east. The vertical placement of dust varies greatly off the coast of East Asia; therefore, dust is placed in three differing heights in the soundings to explore the impact of the vertical placement of dust on SST retrieval error.

**Temperature and Humidity.** Temperature and humidity are specified by the two adjusted soundings. The temperature, in Kelvin, can be input on any pressure level (but must include the height of the pressure level), and humidity is input as the water vapor density in grams per cubic meter. The water vapor density was calculated by multiplying the mixing ratio in grams per kilogram by the density calculated from the ideal gas law using virtual temperature.

**Ozone.** The ozone profile is required for the SBDART. Climatological atmospheric profiles of ozone are provided by SBDART. The profiles provided were linearly interpolated to the heights of the input soundings. A tropical ozone sounding is used for the Saharan Air Layer, and a mid-latitude winter sounding for the Eastern Asian sounding. Varying the selected ozone profile however, did not make a noticeable difference in the calculations at the wavelengths used.
Figure 3.2 – The Saharan Air Layer sounding of temperature (red) and dew point temperature (green).
Figure 3.3 – The Eastern Asia sounding of temperature (red) and dew point temperature (green).
for SST retrievals. Radiatively, varying the ozone profile does not make a large difference, but the most appropriate ozone profile is used for completeness.

**Sea Surface Temperature.** Over the ocean, the surface temperature is at the theoretical interface that cannot be measured and therefore, the skin or subskin sea surface temperature is the best measure of the surface temperature. The SST has been selected by examining the NAVOCEANO’s SST analysis from the day of the profile, using the closest land-free pixel. The SSTs are 23.55°C and 13.95°C for the SAL and Eastern Asia, respectively.

**Aerosol Loading.** Both the vertical location and vertical distribution of dust is varied in each of the soundings to test the impact of different dust profiles. As stated previously, for the Saharan Air Layer sounding, the impact of dust in the SAL and the boundary layer is investigated. For Eastern Asia, the impact of dust is tested for three case studies with dust placed in a low, elevated, and high layer. Six vertical distribution profiles were created to examine the importance of correctly placing the dust loading within a layer (Figure 3.4). The first three types are defined by the magnitude of the gradient between no aerosol and maximum loading. Type 1 has an “infinite gradient,” which is physically a gradient depth of 0.01 km for the model. Types 2 and 3 have gradients of 0.5 km and 1 km, respectively. Since the boundary layer placement of dust in the Saharan Air Layer and the low level placement of dust in the East Asia sounding have a vertical thickness of less than or equal to 2 km, Type 3 is not possible. Type 4 has a constant gradient of AOD from no aerosol to a point of maximum loading in the middle of the layer. A Type 5 profile has maximum aerosols loading at the bottom and a constant gradient to zero AOD at the top of the layer. Flipping Type 5 vertically, Type 6 has maximum loading at the top with a constant gradient to zero AOD at the bottom of the layer.

**Aerosol Optical Depth.** The aerosol optical depth for each profile, for each vertical placement of aerosol, and for each aerosol vertical profile is varied, providing 28 different cases as shown in Table 3.1. The AOD for SBDART is inputted for 0.55 µm, and the model interpolates to the appropriate wavelength. The AOD at 0.55 µm is a surrogate for the AOD in the infrared region. Even though the AOD in the infrared region will be smaller than at 0.55 µm due to the wavelength dependence (see Figure 3.6b), AOD is most commonly given at 0.55 µm, and will be used throughout this study for consistency. There were 28 different AODs tested for every case: 0, 0.1, 0.2, 0.3, 0.4, 0.5, 0.6, 0.7, 0.8, 0.9, 1.0, 1.1, 1.2, 1.3, 1.4, 1.5, 1.6, 1.7, 1.8, 1.9, 2.0, 2.1, 2.2, 2.3, 2.4, 2.5, 5.0, 10.0.
Figure 3.4 – The six developed vertical distributions of aerosols.
Aerosol Optical Properties

For this study, we use the Forecast of Aerosol Radiative Optical Properties (FAROP), based on the Optical Properties of Aerosols and Clouds (OPAC) dataset of dust aerosol properties further described by Hess et al. (1998). The dataset provides the single scatter albedo, asymmetry parameter, and mass extinction coefficient necessary to run the aerosol portion of the radiative transfer model with user defined inputs. The Henyey-Greenstein (H-G) phase function is used to handle multiple scattering. Although the H-G phase function is purely analytic, it is a reasonable fit to the actual phase function (Thomas and Stamnes 2002). Furthermore, changing the phase function only marginally changes the results, even though Ricchiazzi et al. (1998)

![Graph of Single Scatter Albedo & Asymmetry Factor Spectral Dependence](image_url)

Figure 3.5 – Single scatter albedo and asymmetry parameters from FAROP for dust plotted against the interpolated single scatter albedo and asymmetry parameter from SBDART.
Figure 3.6 – The spectral (a) and normalized spectral (b) dependence of AOD on wavelength.
note the H-G phase function as less reliable for radiance calculations compared to radiative flux calculations. The single scatter albedo and asymmetry parameter as a function of wavelength are shown in Figure 3.5. The FAROP lines are the data directly inputted into the model, and the SBDART lines show how the model interpolates the optical properties. In the infrared region, the range of this sensitivity study, the differences are small.

The aerosol optical depth is directly provided to the model as a user input. The spectral dependence of AOD is explicitly calculated by the model from the optical properties of the aerosols as provided by the model user. For SBDART, the vertical aerosol optical depth is nominally noted at 0.55 µm, previously noted as the wavelength at which to denote AOD. Therefore, the AOD is provided for 0.55 µm and extrapolated to all necessary wavelengths for calculations (Figure 3.6a). Although changing the magnitude of AOD appears to change the spectral dependence on wavelength, all AODs approximately fit the normalized spectral dependence shown by Figure 3.6b. The curve is normalized to the AOD at 0.55 µm. Due to the ease of changing the AOD in SBDART and maintaining consistent optical properties of aerosols, a wide range of AODs is examined without creating bias due to user changes in the aerosol properties.

### 3.3 Spectral Weighting and Conversion to Brightness Temperature

Since the model is run at wavelengths corresponding to the wavelengths of the spectral response functions for each channel, spectral weighting proves to be fairly straightforward. Quite simply, the spectrally averaged radiance \( W \, m^{-2} \mu m^{-1} sr^{-1} \), \( I_{\lambda} \), from lower wavelength, \( \lambda_1 \), to upper wavelength, \( \lambda_2 \), is

\[
I_{\lambda_1, \lambda_2} = \frac{\sum_{\lambda=\lambda_1}^{\lambda_2} I_\lambda \delta_\lambda}{\sum_{\lambda=\lambda_1}^{\lambda_2} \delta_\lambda},
\]

where \( I_\lambda \) is the monochromatic radiance for a specific wavelength within the spectral window and \( \delta_\lambda \) is the corresponding normalized weight for that wavelength from the spectral response function from NOAA (Figure 3.7). A spectrally weighted radiance for each channel of the AVHRR is calculated using Equation 3.1. Therefore, one radiance value is calculated for each channel of the AVHRR.

Using the central wavelength of each channel the radiance values are converted to a brightness temperature using the method outlined in Robel (2009) using the Planck function,
where $h$ is the Planck constant, $K$ is the Boltzmann constant, $c$ is the speed of light, $\lambda_c$ is the central wavelength of the channel, and $T_b$ is the spectrally averaged brightness temperature for the channel. The Planck function calculates a blackbody temperature for the given spectrally averaged radiance and central wavelength. Brightness temperatures are commonly given as a measure of radiance from a satellite due to the more easily understood nature of a temperature value. To maintain consistency in the calculations and minimize the systematic error in using a satellite-specific SST retrieval algorithm, the term “effective” blackbody temperature is used to describe the BT calculated by using the inverse Planck function. This temperature is converted to

![MetOp–A AVHRR/3 Relative Spectral Response](image)

Figure 3.7 – Spectral response function ratio for MetOp-A AVHRR/3 for Channels 4 and 5 as provided by the NOAA KLM Users Guide (Robel 2009).
an AVHRR blackbody temperature via a linear correction to the “effective” blackbody temperature. This linear correction is important to include in this study to reduce the chances of a systematic error. The AVHRR BT is the temperature used in the SST algorithm.

3.4 Sea Surface Temperature Retrieval

The SST algorithm used throughout this study is the Naval Oceanographic Office’s Nonlinear SST (NLSST) algorithm with the Multi-Channel SST (MCSST) as the “first guess” SST. The inputs necessary for the daytime SST algorithm are the Channels 4 and 5 spectrally weighted brightness temperatures and satellite zenith angle. The daytime algorithms for the AVHRR onboard MetOp-A used throughout the study are:

\[ \text{MCSST} = 1.0241T_4 + 2.2458(T_4 - T_5) + 0.9717(T_4 - T_5)(\sec \theta - 1) - 280.0106 \]

and

\[ \text{NLSST} = 0.9690T_4 + 0.0772 \text{MCSST}(T_4 - T_5) + 1.0318(T_4 - T_5)(\sec \theta - 1) - 263.3489 \]

All of the brightness temperatures are entered in Kelvin. The satellite viewing angle is converted to radians before being inputted into the algorithm, if necessary. The temperature output by the algorithms is in degrees Celsius.

3.5 Sea Surface Temperature Comparison

As discussed in Chapter 2, the term sea surface temperature is quite ambiguous. Since a known SST is provided to the radiative transfer model, a simple comparison between the known SST and the SST calculated using the algorithms in Section 3.4 can be made. The known SST is most appropriately noted as a skin temperature for comparison sake. The coefficients in Equations 3.3 and 3.4 are derived empirically from regressing infrared satellite BTs to in situ observations, creating ambiguity as to the exact depth of the derived SST. The derived NLSST is generally noted as a “bulk” SST, even though an infrared sensor views a depth of about 10 µm. The “bulk” SST is noted at a depth deeper in the ocean than the known SST, which is close to the skin SST. Therefore, diurnal bias may be present in the algorithm. A warmer known SST compared to the derived SST could be due to this diurnal bias in the empirical parameterizations. However, when the model is run at 0.0 AOD, the derived SST is generally warmer than the
known SST (described in Chapter 4) noting a source of potential bias in the algorithms. This bias may be due to the background troposphere aerosol present (an AOD of approximately 0.15) in the matchups used to calculate the empirically derived coefficients. Other sources of error are extensively discussed throughout Chapter 4.

3.6 Case Studies

Table 3.1 lists the 28 cases studied. Each case is run over 12 satellite zenith angles and 28 AODs. As previously noted, if the depth of the aerosol layer was smaller than 2 km, then the Type 3 vertical distribution of aerosols is not included in the analysis, since the gradient portions are greater than the entire depth of the layer. Although 28 simulations are performed, composites of the runs for a specified profile type and location are used to average potential error due to the loading of the dust in an individual layer. The composites are a mean SST error for all 5 or 6 cases of each case study. This provides 5 case studies of dust contaminated atmospheric profiles: Saharan Air Layer sounding with dust in the Saharan Air Layer, Saharan Air Layer sounding with dust in the boundary layer, Eastern Asia sounding with a low level dust layer, Eastern Asia sounding with an elevated dust layer, and Eastern Asia sounding with a high level dust layer. The differing case studies provide new insight into the structure of SST error due to dust by varying the vertical location, vertical distribution, and amount of dust present in the atmosphere.
Table 3.1 – List of all cases, the vertical profile type (either Saharan Air Layer or Eastern Asia), the vertical location and height of the dust layer, and the distribution type (shown in Figure 3.4).

<table>
<thead>
<tr>
<th>Case #</th>
<th>Profile Type</th>
<th>Location of Dust</th>
<th>Height</th>
<th>Distribution Type*</th>
<th>Abbreviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Saharan Air Layer</td>
<td>Saharan Air Layer</td>
<td>1 – 4 km</td>
<td>1</td>
<td>SALSAL1</td>
</tr>
<tr>
<td>2</td>
<td>Saharan Air Layer</td>
<td>Saharan Air Layer</td>
<td>1 – 4 km</td>
<td>2</td>
<td>SALSAL2</td>
</tr>
<tr>
<td>3</td>
<td>Saharan Air Layer</td>
<td>Saharan Air Layer</td>
<td>1 – 4 km</td>
<td>3</td>
<td>SALSAL3</td>
</tr>
<tr>
<td>4</td>
<td>Saharan Air Layer</td>
<td>Saharan Air Layer</td>
<td>1 – 4 km</td>
<td>4</td>
<td>SALSAL4</td>
</tr>
<tr>
<td>5</td>
<td>Saharan Air Layer</td>
<td>Saharan Air Layer</td>
<td>1 – 4 km</td>
<td>5</td>
<td>SALSAL5</td>
</tr>
<tr>
<td>6</td>
<td>Saharan Air Layer</td>
<td>Saharan Air Layer</td>
<td>1 – 4 km</td>
<td>6</td>
<td>SALSAL6</td>
</tr>
<tr>
<td>7</td>
<td>Saharan Air Layer</td>
<td>Boundary Layer</td>
<td>0 – 1.5 km</td>
<td>1</td>
<td>SALBL1</td>
</tr>
<tr>
<td>8</td>
<td>Saharan Air Layer</td>
<td>Boundary Layer</td>
<td>0 – 1.5 km</td>
<td>2</td>
<td>SALBL2</td>
</tr>
<tr>
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<td>Boundary Layer</td>
<td>0 – 1.5 km</td>
<td>4</td>
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</tr>
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<td>10</td>
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<td>Boundary Layer</td>
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<td>5</td>
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<td>11</td>
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<td>Boundary Layer</td>
<td>0 – 1.5 km</td>
<td>6</td>
<td>SALBL6</td>
</tr>
<tr>
<td>12</td>
<td>Eastern Asia</td>
<td>Low</td>
<td>1 – 3 km</td>
<td>1</td>
<td>EASIALOW1</td>
</tr>
<tr>
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<td>1 – 3 km</td>
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<td>EASIALOW2</td>
</tr>
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<td>1 – 3 km</td>
<td>4</td>
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<td>15</td>
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<td>Low</td>
<td>1 – 3 km</td>
<td>5</td>
<td>EASIALOW5</td>
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<tr>
<td>16</td>
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<td>1 – 3 km</td>
<td>6</td>
<td>EASIALOW6</td>
</tr>
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<td>EASIAELEV1</td>
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<td>Elevated</td>
<td>2.5 – 5 km</td>
<td>5</td>
<td>EASIAELEV5</td>
</tr>
<tr>
<td>22</td>
<td>Eastern Asia</td>
<td>Elevated</td>
<td>2.5 – 5 km</td>
<td>6</td>
<td>EASIAELEV6</td>
</tr>
<tr>
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<td>High</td>
<td>3.5 – 7 km</td>
<td>1</td>
<td>EASIAHIGH1</td>
</tr>
<tr>
<td>24</td>
<td>Eastern Asia</td>
<td>High</td>
<td>3.5 – 7 km</td>
<td>2</td>
<td>EASIAHIGH2</td>
</tr>
<tr>
<td>25</td>
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<td>High</td>
<td>3.5 – 7 km</td>
<td>3</td>
<td>EASIAHIGH3</td>
</tr>
<tr>
<td>26</td>
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<td>High</td>
<td>3.5 – 7 km</td>
<td>4</td>
<td>EASIAHIGH4</td>
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<tr>
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</tr>
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<td>28</td>
<td>Eastern Asia</td>
<td>High</td>
<td>3.5 – 7 km</td>
<td>6</td>
<td>EASIAHIGH6</td>
</tr>
</tbody>
</table>

*Distribution types from Figure 3.4*
CHAPTER 4

RESULTS

Airborne dust aerosol influences the surface infrared radiation signal, limiting the accuracy of remotely sensed surface values, such as sea surface temperature. Based on this study, averaged over all zenith angles and cases, an AOD of 0.25 corresponds to a 0.5 K error in SST due to aerosol, a threshold discussed throughout the results. Even at relatively small AODs, airborne dust contaminated retrievals would have increased error above the desired accuracy of 0.5 K for daytime. The desire is to make the potential retrieval error due to dust less than 0.1 K by creating an accurate correction. As previously noted, a 0.15 AOD is generally noted as the background tropospheric AOD level, and AODs greater than 0.15 would indicate an increased level of aerosols. Thus, 0.15 AOD is a second threshold value that will be discussed throughout the chapter. The results discussed in this chapter focus on the five case studies, two for the eastern tropical Atlantic Ocean and three for the East Asian coast. Furthermore, the error structure of SSTs relative to AOD, satellite viewing angle, height of the dust layer, and vertical distribution of the dust layer are explored in detail.

4.1 Eastern Tropical Atlantic Ocean

The magnitude and structure, with respect to AOD and satellite zenith angle, of the SST error depends upon the vertical location of the dust, whether located the SAL or in the boundary layer. Of all the case studies, dust present in the boundary layer for the SAL profile has the smallest magnitude of errors. Furthermore, the strong capping inversion shown in Figure 3.2 traps the dust relatively close to the surface, and creates a unique SST error structure for this case study. The shape of the error curves for the SAL boundary layer case study (Figure 4.1) is different from the other case studies due to the temperature profile within the boundary layer, the inversion layer located at the top of the boundary layer, and the similarity of the average
Figure 4.1 – The contour of SST error (K) for cases of the Saharan Air Layer profile with a boundary layer dust layer. The titles correspond to the case in Table 3.1.
temperature of the aerosol layer and the SST. The temperature within the inversion layer at the top of the boundary layer in the SAL sounding is warmer than sea surface temperature input. In general, the temperatures above the surface are cooler than the surface temperature causing a cool bias in the SSTs when dust is present. Since the inversion layer is fairly shallow the dust still causes a cooler retrieved SST, but the cooling is modulated by the location of the dust. At larger zenith angles, the difference between the aerosol layer temperature and the SST is saturated out at smaller AODs than seen in the other case studies.

The different vertical distribution types have an impact on the error distribution of each case. Contours of the error based on satellite viewing angle and AOD provide insight into the structure of SST error due to aerosols. Type 2 and 4 (SALBL2 and SALBL4) have almost identical error curves, while the infinite gradient of Type 1 (SALBL1) creates a reduced error overall due to the increase in the average temperature of the dust layer. The infinite gradient type weights the temperature of the surface and inversion layer more than the larger gradients Type 2 and 4. Type 5 has the least error overall due to the dust loading at the surface, which has a temperature very close to that of the surface. When the dust is concentrated at the top of the boundary layer (Type 6) the resulting curve is similar to those of SALBL2 and SALBL4, where the dust is more evenly distributed. This is due to the shallowness of the layer and the maximum loading of the cases located in close proximity; therefore, the characteristics of the dust layers are similar.

The creation of a composite from the 5 vertical profiles of dust reveals a general structure of error for dust located in a boundary layer in the SAL (Figure 4.2). On average, increasing the AOD and increasing the satellite viewing angle causes an increase in the SST error. For an AOD of 0.5, for example, the error ranges from -0.5 K to -1.5 K. For an AOD of 1.0, the error ranges from -1 K to -3 K. Examining SST error at constant AOD, 0.5 for example, a 1 K difference between across zenith angles demonstrates a strong zenith angle dependence. This is due to the relationship between satellite viewing angle and the path length through the atmosphere a surface signal must be transmitted. At some values of AOD and zenith angle the signal received by the satellite is completely dominated by the aerosol layer. The shading of Figure 4.2 displays the spread of all the cases included to make the composite. In general, greater SST error is associated with greater the spread of SST error. The greater AODs and larger zenith angles cause more pronounced differences between the different vertical distribution types.
Figure 4.2 – Composite SST error contours (K) from all five boundary layer dust cases for the Saharan Air Layer sounding. The shading is the spread of the SST error (K) of all cases included to make the composite.
Figure 4.3 – The contour of SST error (K) for cases of the Saharan Air Layer profile with a Saharan Air Layer dust layer. The titles correspond to the case in Table 3.1.
Dust located within the SAL is one of the most prevalent locations of dust over the ocean and one of the most extensively studied locations of dust aerosols. The vertical distribution of Types 1 - 4 (SALSAL1, SALSAL2, SALSAL3, and SALSAL4) for the SAL dust case study induce almost identical error distributions (Figure 4.3). In all cases the error increases with increasing AOD and satellite viewing angle, as the viewing angle increases the path length through the atmosphere and therefore the effective AOD of the dust layer. When the maximum loading is located at the bottom of the dust layer (SALSAL5), the overall error is reduced due to

Figure 4.4 – Composite SST error contours (K) from all six Saharan Air Layer dust cases for the Saharan Air Layer sounding. The shading is the spread of the SST error (K) of all cases included to make the composite.
the warmer temperature that is sensed at the bottom of the layer where the temperature is closer to the surface temperature. In contrast, the overall error is increased when the maximum dust loading is located at the top of the dust layer (SALSAL6). The temperature of the cooler elevated dust is sensed.

Over all, the differences between dust located in the SAL and dust located in the boundary layer are due to the atmospheric temperature of the layer in which the dust is located (typically a function of the height of the dust layer). The inversion located at the top of the boundary layer contains warmer-than-surface temperatures, therefore allowing the temperature of the warmer-than-surface aerosols to be sensed. The inversion layer is also responsible for the larger spread of error in the SAL dust case study. When the maximum dust loading is located at the bottom and top of the dust layer (SALSAL5 and SALSAL6), the minimum and maximum magnitudes of SST error occur, respectively. Disregarding the SALSAL5 and SALSAL6 cases, the spread of error is much smaller, with a maximum spread across all AODs and zenith angles of 1.0 K. The inversion layer also causes the error structure with respect to AOD and viewing angle to be different between the two case studies due to the temperature differences. It cannot be understated that a simple nonlinear correction that is merely a function of AOD and viewing angle does not provide the best possible correction in all plausible regimes of airborne dust.

4.2 East Asian Coast

Due to the much colder than surface temperature of the dust layers of the Eastern Asia sounding, the error structure of the case studies in the Eastern Asia sounding provides a contrast to those in the SAL region. In these case studies, the temperature profile has a fairly constant decrease with height. Thus, the relative differences between the three case studies are solely a function of the vertical distribution of dust, or type of profile (Figures 4.5 – 4.7). The low-level, elevated, and high dust layers of the East Asia sounding contain overall errors greater than the SAL sounding due to the higher altitude and lower temperature of the dust layer of all three case studies. Due to the depth of the layer, a Type 3 vertical profile is not tested for the low-level dust case of the Asian sounding. Across Types 1-3 (Type 3 is not tested for the low-level dust case) similar SST error magnitudes are found for all three case studies. Type 5 has reduced error, and Type 6 has enhanced error, due to the smaller difference temperature and greater difference from the surface temperature, respectively. This provides an important result of the sensitivity study: it
Figure 4.5 – The contour of SST error (K) for cases of the East Asian profile with a low-level dust layer. The titles correspond to the case in Table 3.1.
Figure 4.6 – The contour of SST error (K) for cases of the East Asian profile with an elevated dust layer. The titles correspond to the case in Table 3.1.
Contour of SST Error – East Asia/High

Figure 4.7 – The contour of SST error (K) for cases of the East Asian profile with a high-level dust layer. The titles correspond to the case in Table 3.1.
is necessary to know the altitude of maximum loading accurately, but not the gradient of the aerosol loading.

For all cases, the zero error line occurs at an AOD of roughly 0.15, which is the background tropospheric AOD level present in the matchup data used for the empirical SST algorithms. Although the background aerosol is the combination of all aerosol types present in the atmosphere, for this study dust is the only type of aerosol present. Recall that all AODs are noted at 0.55 \( \mu \)m. This further demonstrates the inherent AOD present in the empirical SST algorithms, which match up the TOA BTs from AVHRR to buoys and other in situ observations. The increasing error with viewing angle is due to the change in effective optical depth because of the increased path length of the atmospheric emission of the dust layer signal. Most notably of the all three case studies is the relatively small AOD of 0.17 for which the error due to aerosols for the high dust layer at a 0° viewing angle is greater than 0.5 K. The higher dust layers can have a substantial impact at small AODs due to the large temperature difference between the aerosol layer and the surface. In all cases, the change in error due to changing the vertical profile is a function of AOD and viewing angle. The greater the AOD or zenith angle, the greater potential difference between the SST error of different cases. This is further shown by the spread of the error shaded in the composites of each case study (Figure 4.8 – 4.10). At generally observed AODs during retrievals (non-cloud flagged, less than 0.5), the maximum difference in error among all vertical profile types is 0.6 K. Excluding Types 5 and 6, the maximum difference in error is 0.3 K. With the error due to aerosols greater than the 0.3 K potential error due to an incorrect vertical profile of dust, a correction will still bring improved accuracy to an SST retrieval.

Compositing the three case studies of the East Asia profile further exhibits the consistency in the error structure (Figures 4.8 – 4.10). The changes in error from the 0° zenith angle at a constant AOD with respect to satellite zenith angle are a function of the secant of the zenith angle due to the increasing path length. In comparison, at a constant zenith angle, the error with respect to AOD from a 0.0 AOD is linear. Equation 2.6 shows the linear impact of AOD on the TOA radiance measured by the satellite sensor. It is clear that the difference in the magnitudes of the error between the three case studies is partially due to zenith angle and AOD, and mainly due to the temperatures differences between the aerosol layers and the surface temperature.
Figure 4.8 – Composite SST error (K) from all five low-level dust cases for the East Asian sounding. The shading is the spread of the SST error (K) of all cases included to make the composite.
Figure 4.9 – Composite SST error (K) from all six elevated dust cases for the East Asian sounding. The shading is the spread of the SST error (K) of all cases included to make the composite.
The SST error due to dust is a linear function of AOD and a function of the secant of the satellite viewing angle. This provides a useful starting point for an empirical correction, if that is the desired approach. Upon examining the zenith angle dependence of the NAVO SST Algorithm for MetOp-A at very small AODs, it becomes apparent that an inherent AOD is found in the retrieval algorithm, as noted in Nalli and Reynolds (2006). In fact, due to the inherent

4.3 Zenith Angle and Aerosol Optical Depth

The SST error due to dust is a linear function of AOD and a function of the secant of the satellite viewing angle. This provides a useful starting point for an empirical correction, if that is the desired approach. Upon examining the zenith angle dependence of the NAVO SST Algorithm for MetOp-A at very small AODs, it becomes apparent that an inherent AOD is found in the retrieval algorithm, as noted in Nalli and Reynolds (2006). In fact, due to the inherent
AOD in current empirical SST retrieval algorithms at AODs below approximately 0.15, an overestimation of the SST is possible. Therefore, with the continued use of current empirical algorithms, it is necessary to only correct AODs above the background levels. The structure of aerosol error in SST with respect to zenith angle and AOD is well studied and well-known (e.g. Rao 1992). The extent to which these results are in line with previous studies is further exemplified by the use of an effective AOD to examine SST error (Figure 4.11). Each of the best-fit lines for the case studies is a good fit for a linear function. When neglecting the Type 5 and 6 vertical distribution types, the best-fit line has much smaller residuals with the errors in the line fitting less than the dust correction for all cases above an AOD of 0.15, the background tropospheric AOD.

![Figure 4.11 – Scatter plot of all model runs grouped by the profile and vertical placement of dust with best-fit lines for each case study.](image-url)
4.4 Vertical Placement and Distribution of Dust

Dust impedes an infrared sensor’s ability to accurately retrieve surface values by masking the surface with the characteristics of the dust layer. Therefore, an SST retrieved in a dust-contaminated atmosphere will adjust to the characteristics of the dust layer. The most important results of this sensitivity study are the relationship between vertical placement and distribution of dust with SST error. A simple linear correction that is solely a function of effective AOD for SST error due to dust does not work for all possible cases. It is evident through the five different best-fit lines for the five different case studies show in Figure 4.11 that each requires a separate empirical parameterization. This is due to the varying temperature differences between each aerosol layer and the surface, shown by Equation 2.6. As such, any useful corrections of SST error due to dust aerosols must not only be a function of effective AOD, but also the temperature difference between the aerosol layer and the surface.
CHAPTER 5

SEASONALITY OF DUST

Airborne dust located above the eastern tropical Atlantic Ocean and off the eastern Asian coast exhibit a strong season cycle. For this study, it is essential to understand the regularity of aerosol optical depths that could potentially be corrected using this new methodology. Without a correction, the removal of dust-contaminated retrievals can lead to seasonal and regional biases. To evaluate the variability of AOD over the regions of interest, state-of-the-art ground-based aerosol sensors are used.

5.1 Data

The Aerosol Robotic Network (AERONET) is a global network of ground-based remote sensing instruments. AERONET instruments are sun-sky scanning radiometers historically known as sun photometers. The photometers sense solar radiation over the spectral range of 340-1020 nanometers using multiple spectral bands, which are used to diagnose the aerosol optical depth of an atmospheric column (Holben et al. 1998). The ground-based network not only provides better temporal sensing of AODs than historical satellite based measurements, but also provides a ground truth for the development of satellite based retrievals (Holben et al. 1998). The autonomy of automatic sensors allows for continuous data coverage over extended time periods, which is especially important for remote locations. Furthermore, the AERONET database provides data with uniform cloud screening, calibration, quality control mechanisms, and distribution to make sure the data is consistently robust across all sensors in the network (Smirnov et al. 2000).

We use AERONET Level 2 data, which is strictly quality-controlled, cloud-screened, and cross-calibrated (Smirnov et al. 2000), in order to provide the smallest possible uncertainties. Level 2 data is the highest quality AERONET data. Typically AERONET Level 2 AOD data have uncertainties of +/- 0.02 (O’Neill et al. 2003). To best represent the eastern tropical Atlantic
Ocean and eastern Asian coast, the Cape Verde and Anmyon sites are examined, respectively (Figure 5.1). These sites are used due to the relatively long length of time data is available. The Cape Verde AERONET station (16°N, 22°W) is examined from 10/21/1994 until 1/31/2008, a period of roughly 14 years. The Anmyon AERONET site (36°N, 126°E) is examined over the period of 10/17/1999 until 11/25/2007, roughly spanning 8 years. Figure 5.2 displays the number of days with observations for each day number for both stations over the time periods studied. The apparent cycle in the number of days with observations is partly due to downtime of the instrumentation and the presence of greater cloud cover during certain times of the year. However, it is important to note that the data is not continuous over the time periods, as demonstrated by Figure 5.3. There are yearlong time periods with no data due to the unavailability of the instrumentation for maintenance, other technical issues, or scientific mission changes. Moreover, the number of observations on a given day also varies due to the atmospheric conditions and length of day (which determines the number of observations possible). As stated previously, the AERONET photometers cannot properly sense in cloudy conditions resulting in fewer observations for cloudy days. Therefore, the seasonal cycle of cloud cover impacts the number of observations available in Level 2 data. Due to the availability of
Figure 5.2 – Number of observation per each day of the year over the time period of the AERONET data for the Cape Verde (a) and Anmyon (b) stations.
more days of data for the Cape Verde station, the conclusions for that station are considered more robust. Using seasonal distributions, rather than monthly distributions reduces biases due to inconsistent data coverage throughout the year. This allows for more robust results, as there are more observations over a season than from just one month.

5.2 Methodology

Using the available wavelengths, the quadratic interpolation method of O’Neill et al. (2003) is used to parse the data into fine (approximately less than 1 µm diameter particles) and coarse (approximately greater than 1 µm diameter particles) AODs at 0.55µm. AOD is canonically reported at 0.55µm, which is in the center of the visible spectrum. AOD varies with respect to wavelength based on the optical properties of the aerosols. In general, dust aerosols are irregular particles ranging from below 1 micron to over 100 microns (Hinds 1999). The coarse
AOD is assumed to be dust, especially off the east coast of Africa. Fine mode aerosol optical depth is neglected due to the negligible impact on the infrared portion of the electromagnetic spectrum measured by the AVHRR sensor. The size of the particles is much smaller than the wavelength of the infrared radiation measured to calculate an SST. Throughout processing, the temporal distribution of data and amount of data available for each season is checked to make sure that results are not due to sampling differences among the seasons. It is also important to note that since AERONET Level 2 data is cloud-cleared, it can be assumed that AVHRR would view similar distributions of dust since AVHRR SST retrievals are also cloud-cleared, although the cloud clearing methodology is different. In order to explore the seasonality of dust while reducing the error due to sampling, a percentage of days with AOD above a threshold value for a particular season is explored. This methodology is a type of normalization by the number of observations made each season and each day. These percentages are the average chance over a season to have a day with an AOD above a threshold value. In other words, the percentages noted in Tables 5.1 and 5.2 can be taken as the chance that a given day within a season or time period will measure an AOD above the threshold value.

5.3 Eastern Tropical Atlantic Ocean

Continental African deserts are the largest source of dust in the world (Huang et al. 2010). The dust from these regions becomes airborne and travels westward across the Atlantic Ocean. Dust over the Atlantic is generally found in the Saharan Air Layer between 1 and 5 km from the surface, but can also be found in the marine boundary layer (Karyampudi et al. 1999). Here we will focus on the boreal summer, when the largest concentrations of Saharan dust are found west of northern Africa (Swap et al. 1996; Huang et al. 2010). Again, only the coarse aerosol results are shown for the Cape Verde AERONET station. The average percentage of days with a coarse AOD above designated thresholds is shown in Table 5.1. A seasonal cycle in coarse aerosols, or dust, is clearly evident. An AOD of 0.25 corresponds to an error of roughly 0.5 K in SST (see Chapter 4), which is greater than the absolute accuracy of SSTs quoted by Donlon et al. (2007). Therefore, the percentage of days with a coarse AOD greater than 0.25 can be paralleled with the percentage of days that may have a retrieval with an error due to aerosols greater than the current estimate of absolute accuracy of SST retrievals from AVHRR. On an annual basis the average percent of days with an observed AOD greater than 0.25 is 31.3%, but the percentage is much
higher during the months of June, July, and August when 64.7% of the days have an observed AOD greater than 0.25. On average, 64.7% of the days will have a retrieval error due to aerosols greater than the absolute accuracy of the retrieval if the aerosol-removal scheme does not flag these pixels. Many applications require an accuracy of at least 0.5 K, especially during the Atlantic hurricane season, and thus SST retrievals with a potential error of greater than 0.5 K should be corrected if at all possible.

The annual distributions of AOD for the Cape Verde AERONET station are shown in Figure 5.4. Many of the larger observed AODs are coarse aerosols, which can be assumed to be dust. Annually, 31.3% of the days, on average, will have an observed coarse AOD greater than 0.25. Not only are the means visibly different (see Figure 5.5 and last row of Table 5.1), but the means are also statistically different. Comparing the means of December-January-February (DJF) to JJA coarse AODs using an unpaired two-tail t-test, the difference in the means is statistically significant at the 99% confidence interval. This reveals a significant season cycle, with the mean of the JJA distribution greater than the DJF. During JJA (Figure 5.6JJA), a majority of the observations are of AODs greater than 0.25. The JJA distribution is also more similar to a normal distribution than any other season. Not only are the upper range values

<table>
<thead>
<tr>
<th>AOD Threshold</th>
<th>Annual</th>
<th>DJF</th>
<th>MAM</th>
<th>JJA</th>
<th>SON</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.15</td>
<td>51.8</td>
<td>29.3</td>
<td>44.2</td>
<td>84.1</td>
<td>49.2</td>
</tr>
<tr>
<td>0.25</td>
<td>31.3</td>
<td>15.1</td>
<td>20.1</td>
<td>64.7</td>
<td>25.3</td>
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<td>0.35</td>
<td>17.1</td>
<td>8.5</td>
<td>11.1</td>
<td>37.2</td>
<td>11.6</td>
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<tr>
<td>0.5</td>
<td>5.6</td>
<td>3.4</td>
<td>4.8</td>
<td>11.1</td>
<td>3.0</td>
</tr>
<tr>
<td>0.6</td>
<td>2.9</td>
<td>1.9</td>
<td>3.6</td>
<td>4.5</td>
<td>1.4</td>
</tr>
<tr>
<td>Average AOD</td>
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<td>0.176</td>
<td>0.228</td>
<td>0.372</td>
<td>0.231</td>
</tr>
<tr>
<td>Variance</td>
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<td>0.028</td>
<td>0.047</td>
<td>0.030</td>
<td>0.033</td>
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</table>

Table 5.1 – Average percentage of days with an observed coarse aerosol optical depth above a threshold value over the 14-year period for the Cape Verde AERONET station. DJF, MAM, JJA, and SON are the first letter of each month taken into account for the season. The 0.25 AOD threshold is noted by grey shading. The last two rows are the mean coarse AOD and variance for the time period.
Figure 5.4 – Annual histograms and empirical Cumulative Distribution Functions (CDF) for Cape Verde AERONET station for total, coarse, and fine aerosols.
Figure 5.5 – Seasonal histograms and empirical Cumulative Distribution Functions (CDF) for coarse aerosols at Cape Verde AERONET station. Vertical blue dotted line denotes 0.25 AOD.
generally higher, but also the lowest values are very infrequent. This means that dust is almost always present during the peak season. This seasonal variability is similar to results from previous studies (Mbourou et al. 1997; Husar et al. 1997). Both the aerosol seasonal cycle and the spatial distribution of the differences in seasons are due not only the annual cycle of the shifting Inter-Tropical Convergence Zone, but also to the seasonal rainfall differences across Africa (Mbourou 1997).

5.4 East Asian Coast

The Gobi Desert is the major source of dust for East Asia (Kurosaki and Mikami 2003). The Aeolian process’ seasonality is related to the strong cyclone activity during the springtime in that region. Airborne dust generated in the Gobi Desert is advected eastward to the Pacific Ocean. The presence of dust off the eastern coast of Asia is less frequent than in the eastern tropical Atlantic, as can be seen by comparison of Table 5.2 with Table 5.1. It is important to note that the dashes indicate the lack of sufficient measurements exceeding the threshold during that time period to have any conclusive numbers. The greatest probability of observing a coarse AOD above any threshold values is during March-April-May (MAM). Similarly, during MAM, on average 13.6% of the days will have a retrieval error due to aerosols greater than the desired

Table 5.2 – Average percentage of days with an observed coarse aerosol optical depth above a threshold value over the 8-year period for the Anmyon AERONET station. The last two rows are the mean coarse AOD and variance for the time period.

<table>
<thead>
<tr>
<th>AOD Threshold</th>
<th>Annual</th>
<th>DJF</th>
<th>MAM</th>
<th>JJA</th>
<th>SON</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.15</td>
<td>8.6</td>
<td>4.7</td>
<td>27.3</td>
<td>2.6</td>
<td>-</td>
</tr>
<tr>
<td>0.25</td>
<td>4.1</td>
<td>1.7</td>
<td>13.6</td>
<td>1.1</td>
<td>-</td>
</tr>
<tr>
<td>0.35</td>
<td>2.2</td>
<td>2.7</td>
<td>7.4</td>
<td>1.1</td>
<td>-</td>
</tr>
<tr>
<td>0.5</td>
<td>0.4</td>
<td>0.2</td>
<td>1.5</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>0.6</td>
<td>0.3</td>
<td>0.3</td>
<td>1.1</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>0.7</td>
<td>0.2</td>
<td>0.9</td>
<td>0.9</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Average AOD</td>
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<td>0.073</td>
<td>0.173</td>
<td>0.046</td>
<td>0.039</td>
</tr>
<tr>
<td>Variance</td>
<td>0.016</td>
<td>0.007</td>
<td>0.034</td>
<td>0.005</td>
<td>0.001</td>
</tr>
</tbody>
</table>
accuracy of SST retrievals. The boreal spring has the highest concentrations of dust, as the baroclinic zone moves and makes conditions more favorable for frontal outbreaks that transport dust. The observation of an AOD greater than 0.25 is 12 times more likely during MAM than JJA, and 8 times more likely during MAM than DJF. Therefore, it is more likely that regional and seasonal biases in SST are present during the boreal spring due to the impact of airborne dust on SST retrievals.

Figure 5.6 displays the total annual distribution of aerosol observations from the AERONET site Anmyon. There is a greater presence of fine aerosols in Asia, compared to the Cape Verde AERONET station, primarily due to biomass burning and anthropogenic sources from Asia (Chang and Song 2010; Oh et al. 2004). The seasonal cycle of dust, shown by Figure 5.7, clearly peaks in MAM with the observations of larger AODs occurring almost entirely during the boreal spring. As noted, this is due to the synoptic climatology of Asia carrying dust from the Gobi Desert, and demonstrated by the reduced negative skewness. MAM has a significantly different mean from the other seasons at the 99% confidence level, as the mean is driven by the higher frequency of events in the tail.

5.5 Comparison of Stations

Eastern Asian dust exhibits a seasonal cycle different than that of African dust. Only 13.6% of the days for the Anmyon station during MAM will, on average, experience an observation of coarse aerosols greater than 0.25, compared to 64.7% during JJA for the Cape Verde station. This also means there may be a greater bias due to aerosols in the tropical Atlantic, as more of the retrievals may be contaminated by dust. The season of maximum dust frequency also differs, as synoptic variability drives the seasonality of dust for the two locations. The means of the season of maximum frequency are statistically significantly different from the other seasons for each station at the 99% confidence interval. The synoptic variability not only changes the frequency of dust events, but as well as the amount of dust transported in those regions. Since there is a far greater chance of having a dust contaminated SST retrieval from the tropical Atlantic than from off eastern coast of Asia, much of the focus of this study is on African dust. Although the final goal is an operational correction to be used to correct any retrieval that may be dust contaminated, it is important to first explore actual retrieval biases, to compare the results of this study with currently observed errors.
Figure 5.6 – Annual histograms and empirical Cumulative Distribution Functions (CDF) for Anmyon AERONET station for total, coarse, and fine aerosols.
Figure 5.7 – Seasonal histograms and empirical Cumulative Distribution Functions (CDF) for coarse aerosols at Anmyon AERONET station. Vertical blue dotted line denotes 0.25 AOD.
CHAPTER 6

AEROSOL ERROR IN CURRENT SST RETRIEVALS

The signal of aerosols in AVHRR SST retrievals is investigated to complement modeling the impact of airborne dust on sea surface temperature retrievals. The first issue to be explored is the quality of the flagging mechanisms used by the Naval Oceanographic Office. The second issue is the error between AVHRR SST retrievals and ground-truth SST observations, with a focus on the extent to which aerosols may be causing some or all of this error. Since microwave based SSTs have been used in previous studies as a remotely sensed SST close to the depth of an infrared retrieved SST, but not impacted by dust aerosols (Nalli and Reynolds 2006), the Advanced Microwave Scanning Radiometer for the Earth Observing System (AMSR-E) is used as an “ground truth” SST.

6.1 Datasets and Matchup Methodology

Three datasets are used to observe the aerosol signal in AVHRR SST retrievals: the AVHRR/In Situ matchup dataset, an AMSR-E-derived SST, and a MODIS Level 3 Aerosol Optical Depth. The data is examined for July 2010. This time period is chosen due to the significant variability and amount of AODs observed during this month (see Chapter 5). These datasets and the methodology are described in detail, followed by the results of the comparisons.

Relaxed AVHRR/In Situ Matchup Dataset

The Naval Oceanographic Office (NAVOCEANO) and Naval Research Laboratory, Monterey, CA (NRLMRY) provided the matchup dataset of SST retrievals from AVHRR sensors including MetOp-A, NOAA-18, and NOAA-19 to in situ measurements, including ship and buoys. NAVOCEANO has a series of tests performed during their SST processing to remove any retrieval that may be incorrect. An AOD is not used in their flagging methodology. The relaxed matchups are created to include data that is normally flagged as potentially aerosol
contaminated by two tests during the strict quality control used by NAVOCEANO in their AVHRR retrievals, as well as the data flagged as normal (Bruce McKenzie, personal communication, 2010). Besides the time, location, SST of the retrieval, and the in situ observation, this dataset includes the type of retrieval from NAVOCEANO’s quality flagging: possible aerosol contamination, normal day, less-stringent daytime and normal night. The less-stringent daytime retrievals are only done when normal retrievals are not possible due to quality control measures, to attempt at least one retrieval in a target area (May et al. 1998). The satellite zenith angle, solar zenith and azimuth angle, BTs for the near and infrared channels, reflectance values of the visible channels, and cloud flagging category (clear, probably clear, possibly cloudy) are also provided. Lastly, the difference in distance and time of the matchups are provided, as well. Therefore, this matchup dataset allows for the exploration of retrievals that would normally be excluded from an SST analysis product. Since the infrared SST is calculated using a regression algorithm tuned to in situ observations, the AVHRR data is not truly a skin SST, but rather a “bulk” SST with an ambiguous depth. The in situ SSTs are either from buoys or ships, taken at a depth from about 0.5 to 3 meters. Therefore, in situ SST measurements are not directly compared to the AVHRR SST retrievals, but the microwave SSTs are used to reduce possible bias from the differing depths of the temperature measurements.

**AMSR-E SST**

The Advanced Microwave Scanning Radiometer for the Earth Observing System SST is a microwave based SST retrieval defined as a subskin SST from the Aqua satellite. The final version 5 Daily SST product from Remote Sensing Systems is used in this study. The $\frac{1}{4}^\circ \times \frac{1}{4}^\circ$ grid of the SST product matches up with the AOD dataset with 16 SST boxes per AOD box. The valid time of each $\frac{1}{4}^\circ \times \frac{1}{4}^\circ$ SST is provided. Gentemann et al. (2008) notes a mean discrepancy between the AMSR-E SST and an accurate skin SST measurement from a field campaign of 0.12 +/− 0.39 K. The discrepancy is much less than the uncertainty in the AVHRR retrievals. Meissner and Wentz (2010) note concerns with the 18.7 GHz channels of the AMSR-E sensor during 2010. However, SST is relativelty insensitive to the 18.7 GHz channel and therefore any potential error due to this issue would be small. The AMSR-E data are sponsored by NASA Earth Science MEaSUREs DISCOVER Project and the AMSR-E Science Team. Data are available at www.remss.com.
MODIS Level 3 Aerosol Optical Depth

The Moderate Resolution Imaging Spectroradiometer (MODIS) Level 3 Aerosol Optical Depth product was provided by NRLMRY. The product is six hourly and averaged to a 1° x 1° latitude-longitude grid. The MODIS AOD is quality assured through quality control procedures and empirical corrections designed to reduce uncertainties in the operational MODIS Level 2 AOD relating to cloud contamination, microphysics, and lower boundary condition characteristics (Zhang and Reid 2006; Yingxi et al. 2010) are used to create the Level 3 product. Data is only available for the boxes in which there are satellite retrievals, and no interpolation is done for the Level 3 product. The Level 3 AOD is used in the Navy’s aerosol forecast system, the Navy Aerosol Analysis and Prediction System (NAAPS).

Matchup Methodology

The AVHRR retrieval is considered the ground point for all matchups. This means that all other products and data are matched up in time and location compared to the AVHRR retrieval location. The Relaxed AVHRR/In Situ Matchup dataset is created by NAVOCEANO, while the rest of the matchup are done specifically for this study. The AMSR-E SST is matched only if the AVHRR retrieval falls within a ¼° x ¼° in which there is an AMSR-E retrieval within the two hours preceding or following. The AOD dataset is matched to the AVHRR retrieval by finding the 1° x 1° box in which the retrieval falls, spatial and temporally. The AVHRR retrieval is matched in time with the AOD dataset by selecting the six-hour window in which the AVHRR retrieval occurs, and using that grid. A matchup occurs if data exists for all three products: AVHRR SST retrieval, AMSR-E SST, and Level 3 AOD. All matchups occur during the daytime due to the MODIS AOD daytime only retrieval capabilities. All three remotely sensed values (AVHRR SST, AMSR-E SST, and MODIS AOD) are from sensors with different footprint sizes and spatial resolutions. The matchup methodology is designed to reduce potential error from these varying resolutions by only using high-resolution product values that fall within the footprint of lower-resolution product values. Although there is error associated with this methodology, the results of these matchups are noted to be qualitative. The location of the matchups of all three products over the globe for July 2010 is shown in Figure 6.1; a magnified view of the tropical north Atlantic region with its dust-filled Saharan Air Layer is shown in Figure 6.2.
Figure 6.1 – Geographic location of matchups for MetOp-A, NOAA-18, and NOAA-19 for July 2010. Color bar represents the AOD observed for that retrieval.

Figure 6.2 – Geographic location of matchups for MetOp-A, NOAA-18, and NOAA-19 for July 2010 in the tropical north Atlantic only. Color bar represents the AOD observed for that retrieval.
6.2 Aerosol Flagging Efficiency

The Naval Oceanographic Office performs strict quality control procedures on the AVHRR SST retrievals to ensure data is not cloud or aerosols contaminated. For the relaxed AVHRR/In Situ matchup, NAVOCEANO provides retrievals that would be flagged as potentially aerosol contaminated and not used (Bruce McKenzie, personal communication, 2010), as well as all retrievals flagged as normal. Since the aerosol flagging is statistical rather than based on observed AOD values, it is important to examine the effectiveness of the flagging by using the matched AOD of the each retrieval and whether the retrievals are categorized (normal or potentially aerosol contaminated) correctly.

Globally, out of 36,071 retrievals, 373 are flagged as possibly contaminated. Of the 35,698 retrievals not flagged as possibly aerosol contaminated only 1,624, or 4.6%, are matched with an observed AOD greater than 0.15. An AOD greater than 0.15 is a good threshold value since 0.15 is generally noted as the background tropospheric AOD level (Nalli and Reynolds 2006), and AODs greater than 0.15 would indicate an increased level of aerosols. However, of the 25,628 retrievals noted as normal daytime (not less-stringent daytime or possibly contaminated), only 316 were found to have a corresponding AOD greater than 0.15. For the tropical north Atlantic, 438 of the 1,560 of the retrievals not flagged as possibly contamination have a matched AOD greater than 0.15. However, looking at only the 661 normal day SST matchups, only 12 have a matched AOD greater than 0.15. Of the 373 retrievals flagged as possibly aerosol contaminated 261, or 70%, are matched with an observed AOD less than 0.15. The daytime retrieval quality control techniques appear to properly flag aerosol contaminated SST retrievals, with a proportion of correct of 94.8%. The false alarm rate of the aerosol flagging is 70.0%. This methodology will allow for a more robust aerosol flagging and correction mechanism, rather than just throwing away potentially contaminated data. The less-stringent SST matchups need to be further evaluated, as 80% of the retrievals not flagged as possibly aerosol contaminated matched with an AOD greater than 0.15 are denoted as a less-stringent matchup.

Overall, the NAVOCEANO quality flagging properly denotes uncontaminated retrievals. Focusing on contaminated retrievals, of the possibly contaminated retrievals across the globe 70% have an observed AOD of less than 0.15. Globally, most of the retrievals denoted as possible aerosol contamination are not matched with elevated AODs. Delving further into the tropical north Atlantic, none of the retrievals flagged as possibly contaminated have an AOD less
than 0.15. The quality flagging tends to do well for dust aerosols. This provides further motivation to specifically focus on a dust correction for now, and using the Saharan Air Layer region in the tropical north Atlantic as the focus for this study. Once an operational aerosol correction is in place, aerosol-flagging mechanisms can be relaxed and more retrievals can be corrected and used.

6.3 Observed Error

The SST error, defined as the difference between the AVHRR SST and the AMSR-E SST, versus Aerosol Optical Depth from the MODIS Level 3 product is plotted in Figures 6.3 and 6.4. The global distribution of SST error for possibly contaminated retrievals shows a clear negative trend for increasing AOD (see Figure 6.3). This figure only includes retrievals that are flagged as possibly aerosol contaminated. Although the flagging efficiency is low for correctly denoting retrievals with elevated aerosol levels, the larger AODs show a negative bias consistent with contamination. This is expected as greater aerosol amounts emit greater amounts of infrared radiation and mask the surface signal. Trends due to satellite viewing angle are hard to distinguish because there are other sources of uncertainty for the SST error estimate, such as the height of the aerosol layer (see Chapter 4). However, since only one month of data was provided, July 2010, caution should be used for generalizing the results over all locations and time periods. More matchups over a longer period of time would increase the robustness of the analysis. However, the results are suggestive of the point that aerosols are contaminating the retrievals and have a substantial impact.

Shifting the focal point to the tropical north Atlantic and exploring the matchups in that region, the negative trend of SST error is again visible with increasing effective AOD, denoted as the AOD multiplied by the secant of the satellite zenith angle (Figure 6.4). All quality types of SST are used for investigation in this region, as compared to only possibly contaminated retrievals for the global analysis to provide an adequate number of analysis points. Many of the larger observed AODs occur with retrievals located in the dust prone SAL region; however, this figure focuses on retrievals with an AOD less than 0.5 (not effective AOD). At larger AODs, the temperature of the aerosol layer becomes of great importance and can cause differences greater than 1 K (see Section 4.4). Among retrievals with an AOD less than 0.15, there is spread in SST
error. This may be due to sources of uncertainty discussed in Chapter 2 and variable such as the temperature or height of the aerosol layer.

A regression line fitted to the data has a slope of -2.11 ± 0.11 and a y-intercept of 0.15 ± 0.2. The SST error and effective AOD have a statistically significant correlation at the 99% confidence interval. The slope of the line corresponds to a depression of roughly 0.5 ± 0.03 K per 0.25 AOD increase, which is similar to the results found in this study. Furthermore, the positive y-intercept value physically means there is a positive bias in SST retrievals at small AODs, which is due to the empirical derivation of the algorithms in an atmosphere with a background tropospheric AOD of roughly 0.15. It is important to note that at small AODs a positive bias is possible. In the SAL for example, annually 48.2% of the days do not have an observed AOD greater than 0.15, with a much higher percentage in the boreal winter. Although this study focuses on a correction for large AOD values, a correction for smaller AOD values would be just as feasible. A simple linear fit of SST error as a function of the effective AOD would provide a correction and improve the accuracy of an SST retrieval, but for the best possible retrieval, a more complex correction is necessary. The matchups display the expected trend of greater error with increasing AOD, and provide further motivation for this study.
Figure 6.3 – Sea surface temperature error as a function of AOD and satellite zenith angle (color bar) for the global matchup of possible aerosol contamination data.
Figure 6.4 – Sea surface temperature error as a function of effective AOD and satellite zenith angle (color bar). These matchups fall within the region noted by Figure 6.2 as the tropical north Atlantic and include all quality types. The linear best fit line is the solid black line.
CHAPTER 7

CONCLUSIONS

7.1 Summary

Current SST algorithms are very sensitive to aerosol contamination. Errors larger than the current stated accuracy of 0.5 K for SSTs are possible at AODs that occur frequently, especially during the summertime in the tropical Atlantic and springtime off the east coast of Asia when large-scale dust outbreaks occur. The errors due to aerosols are not negligible, and thus should not be ignored. While this study focuses on dust of the coast of east Africa and the east coast of Asia, the results are applicable to other regions and the methodology can be used to assess different types and distributions of aerosols. Furthermore, this study compares well with the results of Rao (1992), and provides a more modern and in depth investigation of dust aerosols. As a first step towards the ultimate goal of using modeled or observed AODs to create a bias correction or enhanced retrieval technique, it is necessary to understand the sensitivity of SST retrievals to aerosol structure and amount. Moreover, the methodology of this sensitivity study further provides a basis for creating an operational aerosol correction to SST retrievals.

With a strong seasonality of dust noted above the eastern tropical Atlantic Ocean and off the eastern Asian coast, as well as the differing heights at which aerosols are located in these regions, the locations provide a good range of possible inputs. Recall that an AOD of 0.25 roughly corresponds to an SST error of 0.5 K. In the tropical Atlantic area of interest, an AOD greater than 0.25 is observed to occur 64.7% of the days during an average June-July-August. Therefore, during June-July-August there can be an SST retrieval error due to dust of greater than 0.5 K 64.7% of the days. Based on the seasonal distribution of AOD from the Cape Verde AERONET site, assuming a 30° zenith angle, and using the results from the SAL dust layer in the SAL sounding, the potential seasonal bias in SSTs for JJA is as great as -0.27 K. In comparison, the maximum dustiness occurs during March-April-May for the Asian coast area of
interest, with only 13.6% of the days, on average, observing an AOD greater than 0.25. This corresponds to a potential 0.5 K error in SST retrievals due to dust 13.6% of the days in March-April-May. Based on the seasonal distribution of AOD from the Anmyon AERONET site, assuming a 30° zenith angle, and using the results from the elevated dust layer off the east coast of Asia, the potential seasonal bias in SSTs for MAM is as great as -0.11 K.

These cooling patterns evident in the idealized model simulations were also evident using the infrared and microwave SST differences (Chapter 6). Qualitatively, with increasing effective AOD the difference between the infrared SST and microwave SST increases. The microwave SST is noted as the “ground truth” and therefore the difference between the two is noted as error. For example, for all matched up retrievals at an AOD of 0.5 +/- 0.05, observed differences in the SST are greater than -1 K in some cases, and not a single matched up retrieval has a positive SST error. Thus the trends and magnitudes evident in the model case studies appear to be realistic.

The four main variables in the sensitivity studies are the AOD, satellite viewing angle, height of the aerosol layer, and vertical distribution of the aerosol layer. The relationship between AOD, satellite viewing angle, and SST error are widely known (e.g. Rao 1992) and demonstrated by Equation 2.6. SST error increases linearly with AOD and as a function of the secant of the satellite viewing angle. The two less studied parameters are the height of the aerosol layer and the vertical distribution of the dust layer. As long as the correct location of maximum loading is used, the actual gradient of the aerosol loading does not have a large impact on the results. However, the most important parameter as determined from this study in correcting for the effects of aerosols is the height of the aerosol layer, and therefore the temperature of the height at which the aerosol layer is found. It is evident that the height of the dust layer has a substantial impact on the error in SST retrieval. To the author’s knowledge, no current aerosol correction to SST retrievals use a temperature related to the height of the aerosol layer as a necessary input for error correction. Any aerosol correction, whether physical or empirical, to SST retrievals should be a function of the atmospheric temperature of the aerosol layer.

Current empirical corrections are generally a function of the effective AOD. Figure 4.11 shows that a simple empirical correction cannot entirely account for the temperature difference between the surface and aerosol layer. Since a dust layer impedes an infrared sensor’s ability to measure the surface by masking the signal with the aerosol layer characteristic, the satellite will
retrieve a signal from the dust layer. The larger the difference between the temperature of the aerosol layer and the surface, the greater the error in the SST retrievals due to dust. As such, this study is not only a basis for a physical correction using a radiative transfer model, but also the basis for a next generation empirical correction that would use a temperature measure of the aerosol layer. Moreover, the methodology used in this study is sensor independent and easily modified for most sensors.

Regardless of correction methodology, more information than is currently available from satellites is necessary. A modeled atmosphere and aerosol distribution must be used. Potential sources of atmospheric profiles are the Navy’s Operational Global Atmospheric Prediction System (NOGAPS), a global weather forecasting model (Bayler and Lewit 1992), for users associated with the Naval Research Lab in Monterey, CA. The vertical profile and placement of the dust aerosol could be taken from the Navy Aerosol Analysis and Prediction System (NAAPS), which provides dust, smoke, and sulfur emissions 6 hourly on a 1° x 1° grid, for 24 vertical levels. The model is a modified version of the Christensen (1997) model, and is the world’s first truly operational and global aerosol model. Before these model simulations could be used, further work quantifying their errors and the impacts on the retrievals would need to be performed, but the possibility for improved SST retrievals using these model simulations is real.

This study quantifies the error due to airborne dust and provides a usable methodology to create an accurate physical correction. Rather than disregarding retrievals that may be aerosols contaminated, this study shows it is possible to quantify the error allowing for the correction of these retrievals. It is the desire of the SST community to provide the best possible global coverage of SSTs with the best possible ( < 0.5 K error) accuracy. The results of this study show it is necessary to deal with the possible errors due to airborne dust, and that the errors are correctable.

### 7.2 Future Work

The methodology used in this study performs simulations of the radiative transfer model for 402 different wavelengths (201 for each AVHRR channel) for each SST retrieval. Investigation into the ability to use solely the central wavelength of the AVHRR channels may reveal substantially increased computational efficiency with only minimally reduced accuracy. This would allow for this methodology to be used as an operational aerosol correction to SST
retrievals. This study also focuses on only two specific case studies of airborne dust. The inclusion of more profiles will provide further confirmation of the results of this study. The solar zenith and azimuth angles are held constant for all case studies in this work; the impact of varying the solar angles on error due to dust in SST retrievals should be further explored. Moreover, this study focuses on daytime SST retrievals, and the same methodology can be used to explore nighttime retrieval error due to airborne dust.

This study provides a strong foundation for an operational correction to SST error due to airborne dust aerosols. However, prior to creating an operational correction cases studies of actual retrieval error compared to modeled retrieval error are necessary. Figure 7.1 provides a possible schematic of an operational physical aerosol correction using NAAPS and NOGAPS. Radiative transfer models with full aerosol scattering require extensive computing power, especially if run globally and daily. Therefore, empirical corrections to the SST error due to aerosols will also be explored. This study has demonstrated that previous empirical corrections for aerosol should not only be a function of AOD and satellite viewing angle, but should also take into account the difference between the temperature aerosol layer (or temperature at maximum loading) and the SST.
Figure 7.1 – Potential future flowchart of an operational aerosol correction using a radiative transfer model approach and “ground truth” measurements not impacted by airborne dust.
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BIOGRAPHICAL SKETCH

Alec Setnor Bogdanoff

Born and raised in sunny South Florida, Alec’s love for weather started just as quickly as afternoon thunderstorms during the summertime. His family’s dining room table became a fort and the Weather Channel was his form of afternoon cartoons. At a young age he became fascinated with weather and it is this fascination that brought him to Florida State University. In spring of 2008, Alec graduated from FSU with a Bachelor of Science in Meteorology with Honors, Summa Cum Laude. Under the advisement of Dr. Jon Ahlquist, Alec completed his Honors Thesis, receiving a Bess Ward Honors Thesis Grant to assist in his undergraduate research. As an undergraduate Alec served as a Resident Assistant, Orientation Leader, First Year Experience Instructor, Freshmen Interest Group Leader, and was an active member of the North Florida Chapter of the American Meteorological Society.

In the fall of 2008, Alec began his Master of Science in Meteorology under the direction of Dr. Carol Anne Clayson. Alec’s research interests include radiative transfer, satellite remote sensing, turbulence, and air-sea interaction. He plans to pursue his Doctorate.