CHAPTER 2

BACKGROUND

# 2.1 Introduction

Stratospheric aerosol plays an important role in the global radiative forcing balance by scattering solar irradiation and causing an overall cooling effect that depends on the particle size distribution and the concentration (*Kiehl and Briegleb*, 1993; *Stocker et al.*, 2013). These climate effects are an important and recent focus of research due to the potential contribution of stratospheric aerosol to the so-called global warming hiatus (*Solomon et al.*, 2011; *Haywood et al.*, 2014; *Fyfe et al.*, 2013) and efforts to quantify the variability and trends in the global stratospheric aerosol load are underway with various ground-based and satellite data sets (*Rieger et al.*, 2015; *Ridley et al.*, 2014).

Since its discovery with stratospheric balloon observations (*Junge et al.*, 1961), stratospheric aerosol has been measured with various techniques, although due to the variability of physical composition and particle size, the observations are always limited to some degree. In-situ balloon observations continue to be used and have provided highly valuable data sets, including most notably the long time series of optical particle counter measurements from Laramie, WY (*Deshler et al.*, 2003; 2008; *Kovilakam et al.*, 2015). Aircraft-borne nephelometers (*Beuttell and Brewer*, 1949; *Charlson et al.*, 1969) acquire detailed in-situ measurements, providing, for example, plume composition (*Murphy et al.*, 2014), but are spatially limited to the aircraft track. Ground based lidars have been used to do detailed studies of the extent of volcanic aerosol plumes (*Chazette et al.*, 1995; *Sawamura et al.*, 2012) and provide valuable insight into long term local variability and trends in the aerosol layer. For example, lidar observations were used by *Hofmann et al.* (2009) to first report the observed increase in stratospheric aerosol over approximately the last decade. However, the global distribution, which can only really be obtained with satellite observations, provides invaluable insight into aerosol processes and variability. A good example of this is the use of satellite observations by *Vernier et al.* (2011b) to determine that the increased stratospheric aerosol load reported by *Hofmann et al.* (2009) was in fact due to a series of relatively minor, mostly tropical, volcanic eruptions.

Several recent studies have highlighted the requirement for continued global stratospheric aerosol observations and especially the need to resolve, both vertically and horizontally, aerosol in the lowermost stratosphere and the upper troposphere. This is the case for tracking the evolution of aerosol from volcanic eruptions, which can have a substantial effect on the aerosol optical depth in the lowermost stratosphere (*Ridley et al.*, 2014; *Andersson et al.*, 2015). Furthering the understanding of the transport of aerosol near and across the tropopause would also benefit from higher spatial and temporal resolution observations. This is evident in the case of volcanic plumes, such as that from Nabro in 2011, the transport and origin of which has been studied extensively and somewhat controversially (*Bourassa et al.*, 2012c; 2013; *Vernier et al.*, 2013; *Fromm et al.*, 2013; 2014; *Fairlie et al.*, 2014; *Clarisse et al.*, 2014). However, this is also the case for the formation of background-level aerosol, particularly in the region of the Asian and North American monsoons, which have been identified as a source of substantial, seasonal and highly structured aerosol formation from precursor tropospheric source gases (*Vernier et al.*, 2011a; *Neely et al.*, 2014; *Thomason and Vernier*, 2013). .

Continued stratospheric aerosol observations from space are drastically needed though few, if any, planned missions with such capability are underway. In this work, we present the design and test of a prototype instrument for potential future satellite-based stratospheric aerosol observation. The Aerosol Limb Imager (ALI) concept is a relatively small, low-cost, low-power, passive instrument, suitable for microsatellite deployment with the capability to provide high spatial resolution measurements, both vertically and horizontally, of the visible/NIR aerosol extinction coefficient. The basic idea is to leverage the clear advantages of the limb scatter technique as a passive, and therefore low mass and low power, means to obtain daily global coverage, with a two dimensional hyperspectral imager for filling cross-track observation.

The ALI instrument concept is built around the use of an Acousto-Optic Tunable Filter (AOTF), which is a novel filtering technology that provides the ability to rapidly select the central wavelength of an image with no moving parts. These filters, which have recently been developed as large aperture imaging quality devices, operate very efficiently in the red and near infrared spectral range, which is a well matched spectral range for limb scatter sensitivity to aerosol and cloud (*Rieger et al.*, 2014). Additionally, the spectral bandpass of the AOTF, which is typically between 3-6 nm at these wavelengths, is very suitable for the broadband scattering characteristics of the aerosol limb signal. The two dimensional imaging nature of the design provides the capability to achieve at least sub-kilometer resolution at the tangent point, which is on the order of the scale size of the upper troposphere and lower stratosphere (UTLS) aerosol features mentioned above.

It should be noted that the basic instrument design concept of ALI is very similar to that of the Atmospheric Limb Tracker for the Investigation of the Upcoming Stratosphere (ALTIUS) (*Dekemper et al.*, 2012), which is a Belgian instrument concept from the Belgian Institute for Space Aeronomy (BIRA). ALTIUS is designed to measure limb scattered sunlight; however, it also has solar, stellar, and planetary occultation modes and is scientifically focused on trace gas measurements, particularly for ozone, whereas ALI is optimized for aerosol observation.

# 2.2 Stratospheric Aerosol

The atmosphere of earth is a complex and complicated system and is effected by human activities. In the late 18th century, it was known that atmospheric temperature decreased with altitude and a theory had been raised that at specific certain altitudes the temperature must eventually go to absolute zero. This lead to a series of balloon campaigns to discover this mysterious altitude in the atmosphere which were noisy and unreliable due to the technology available. However, in the late 19th century the technology used in these sounding balloon had improved to a point where the atmospheric temperature could be accurately measured and it was found that at approximately 12 km an inversion point was discovered where the temperature started to increase and thus the tropopause, which separates troposphere and the stratosphere, was discovered (*Hoinka*, 1997). The stratosphere is the region of the atmosphere above the temperature inversion of the troposphere, where atmospheric temperature increases and the lower bound of the stratosphere is in-between 10 and 16 km from the high latitude to the tropics (*Andrews*, 1987). The stratosphere, which extends up to approximately 50 km, is thermodynamically stable and fairly dry (*Boucher*, 2015). The characteristic stability of the region limits vertical transport of the stratosphere, leading to long lifetimes, spanning from months to years, for non-volatile species (*Volk et al,* 1997; *Brasseur and Solomon*, 2005)

The stratosphere undergoes exchange of air with the troposphere though a series of dynamical processes including tropical convection, the polar vortex, and tropopause folding (*Holton et al.,* 1995) and circulation within the stratosphere is dominated by the slow Brewer-Dobson circulation (*Plumb and Eluszkiewicz*, 1999). Some chemicals can cross the tropopause thermal barrier into the stratosphere which allows for chemicals reactions to occur. One such reaction forms stratospheric sulfate aerosol, discovered by *Junge et al* (1961) though stratospheric balloon sondes, which are droplets of hydrated sulfuric acid (H­2­SO­4) formed from the oxidation of sulfate compounds, primary OCS and SO2 (*Brock et al.*, 1995). This stable layer of aerosol exists in the stratosphere from the altitude of the tropopause to approximately 30 km.

## 2.2.1 Aerosol Sources and Microphysics

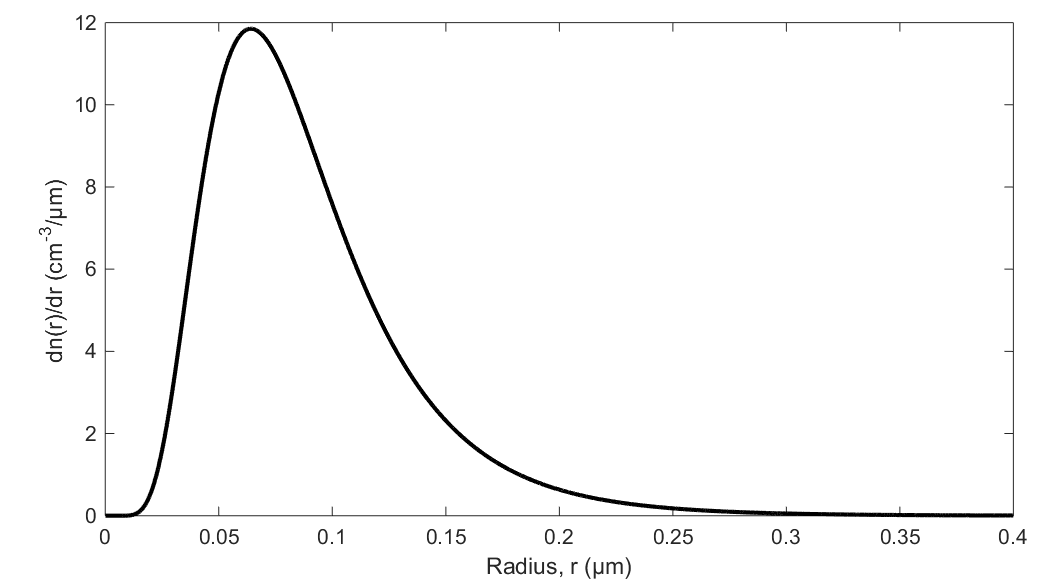
The sources for sulfate aerosol are primarily produced in the troposphere and formed though both natural and anthropogenic processes. One primary source of this sulfur is OCS which originates from marine processes, biomass burnings, and industry (*Kettle et al.*, 2002; *Notholt et al.*, 2003). OCS has a long lifetime in the troposphere and low solubility allowing for a significant portion to reach the stratosphere and some of it oxidizes and hydrates to form sulfate aerosol (H2SO4) and adds to the background aerosol layer (*Crutzen*, 1976).

Another source of sulfur is SO2, one source originates in the troposphere though industry from the burning of fossil fuels. Sulfur dioxide (SO2), which has a short lifetime in the troposphere and its concentration varies regionally (*Thomason and Peter*, 2006), also hydrates into sulfate aerosol. A second source of SO2 is from volcanic eruptions which are highly variable in location and time and can insert a large amount of sulfur directly into the atmosphere. The sulfur amounts that can be injected by volcanos can be so large they in fact dominate the stratospheric aerosol layer perturbing from the background levels. Examples of this perturbation of the aerosol layer were noted during the volcanic eruptions of El Chichon in 1982 (12-20 Tg of sulfur) (*McCormick and Swissler*, 1983; *Hofmann and Rosen*, 1983) or Mount Pinatubo in 1991 (20-30 Tg of sulfur) (*McCormick and Veiga*, 1992). However, after the Mount Pinatubo eruption a volcanically quiescent period occurred where aerosol layers returned to background. Following this period, a series of smaller volcanic eruptions have increased the background aerosol layer in the amount of 4-7% per year from 2000 to 2009 (*Vernier et al.*, 2011).

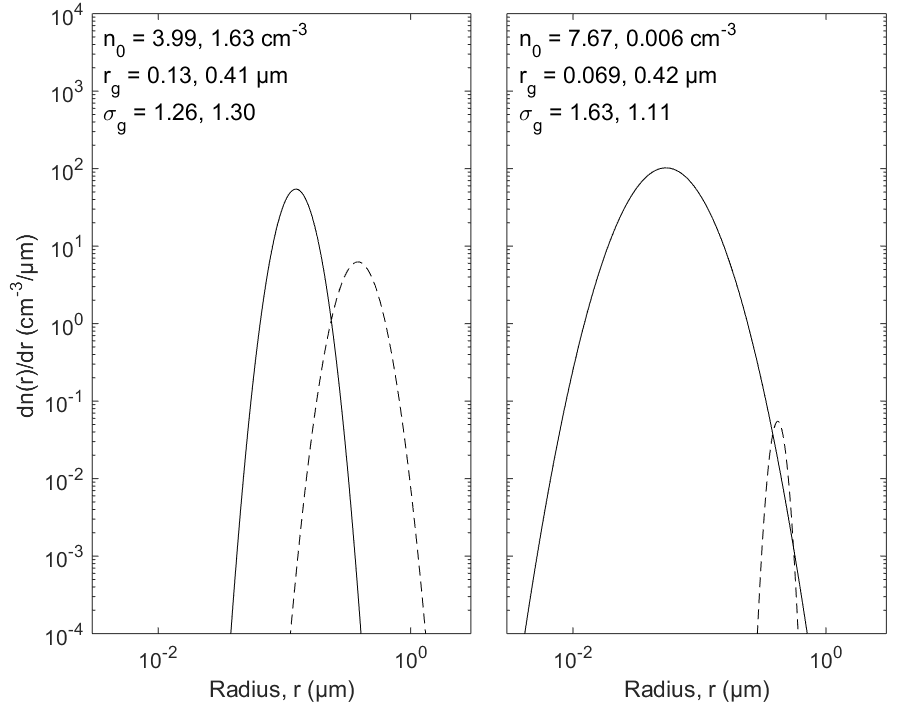
These sulfur sources undergo a series of reactions to be converted into H2SO­4 and have been found to form droplets of aerosol that consists of approximately 25% H2O and 75% H2SO4 (*Rossen*, 1971; Wang *et al.*, 1989). These aerosol particles form droplets of various sizes on the order of 0.05 to 1.0 µm depending on the processes of nucleation, evaporation and condensation that is underwent (*Junge et al.* 1961; *Brock et al.*, 1995; *Bingen et al.*, 2004). A log-normal distribution generally describes the particle size of aerosol in the form of

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| --- | --- |
|  | (2.1) |

where is the aerosol number density, is the mode radius, and is the mode width (*Jäger and Hofmann,* 1991; *Hamill et al.* 1997). The particles are distributed normally over the logarithm of the radius. To fully describe aerosol for a single log-normal distribution the number density, mode radius and mode width is required and for a non-volcanic background aerosol layer an approximate mode radius and mode width of 0.08 µm and 1.6 respectively have been determined (*Thomason*, 1991) shown in Figure 2-1.



**Figure 2-1**: Log-normal distribution for non-volcanic background aerosol layer.



**Figure 2-2**: Bimodal particle size distributions fits from optical particle size. a) Distributions from a volcanic period after the Mount Pinatubo eruption recorded in 1993. b) Distributions from a background aerosol period eruption recorded in 1999. Both of the aerosol distribution measurement are from 20 km altitude with the solid line being the fine mode and the dashed line is the coarse mode. Figure is recreated from figure 5 of *Deshler et al.* (2003).

Optical particle counters have been used on board sondes from Laramie, Wyoming over the past 40 years to measure particle sizes in bins between 0.15 to 2.0 µm. The composition of the particle size distributions have been primarily unimodal, known as a fine mode, over non-volcanic periods and can be used as an acceptable distribution approximation during background periods. But during volcanic episodes, a bimodal log-normal distribution of aerosol particles is more representative of the stratospheric aerosols with the addition of a coarse mode (*Deshler et al.*, 2003; 2008; *Kovilakam et al.*, 2015). The coarse mode has larger particles than the fine mode and complicates determining full aerosol microphysical parameters since the number of required parameters has increased to six: a number density for both the fine and coarse mode, two mode radii, and two mode widths. Figure 5 from *Deshler et al.* (2003), recreated in Figure 2-2, demonstrated two bimodal particle size distributions from balloon optical particle counters. The first distribution is from a volcanic period in 1993 after the Mount Pinatubo eruption and another from a background period in 1999.

## 2.2.2 Climate Effects

Aerosol have several effects on the climate of the planet with a large amount of uncertainty to the overall effect (*Solomon et al.*, 2007). Through a direct effect, aerosol particles scatter incoming visible solar radiation away from earth increasing the albedo causing a cooling effect on the surface of the planet (*Lacis et al.*, 1992). A secondary effect from aerosols, that is dependent on aerosol particle size distributions, is a greenhouse effect that is caused by infrared radiation from the planet being scattered by aerosols (*Kiehl and Briegleb*, 1993). Aerosol also introduces an indirect effect to a radiative balance known as the cloud albedo effect. This is caused by condensation on existing aerosol particles forming cloud condensation nuclei which leads to an increase of the albedo for short wavelengths which also contributes to cooling the planet's surface. These types of cloud forming particles tend to increase the overall lifetime of the cloud increasing the cloud coverage of the planet thus increasing the albedo (*Charlson et al.*, 1992). Overall, the aerosol particles’ cooling effect dominates the warming effects and cools the surface of the planet (*Solomon et al.*, 2011).

During background aerosol periods there is little effect of the cooling from stratospheric aerosols but this can greatly change during periods of volcanic activity where the layer can be greatly perturbed. During the eruption of Mount Pinatubo in 1991 the sulfate aerosol load was increased by a large quantities causing a cooling of the lower atmosphere by 0.5oC (*McCormick et al.* 1995; *Soden et al.*, 2002) and 0.1 to 0.3oC on the surface (*Thompson et al.*, 2009; *Canty et al.*, 2013). And the surface temperatures did not return to pre-Pinatubo level until approximately 3 years after the eruption as the atmosphere filtered out the additional aerosol (*Hansen at al.*, 1996). More recently, a series of small to moderate volcanic eruptions have increased the background stratospheric aerosol layer (*Vernier et al.*, 2011b). The additional volcanic aerosol load may have been linked to a larger cooling effect, known as the global warming hiatus, when compared to background levels (*Solomon et al.*, 2011; *Haywood et al.*, 2014; *Fyfe et al.*, 2013). In order to fully quantify this effect, long term time series with microphysics information are requires to fully understand the aerosol forcing effect on climate modeling and change.

# 2.3 Aerosol Measurements

Two fundamental methods are used to measure aerosol concentrations within the atmosphere. The first of these methods are ground based and in-situ measurements which give good detail and information about a specific localized area. However, these measurements are limited in scope as they do not have global coverage that is inherent in satellite instrumentation. Both ground-based instruments and satellites have important roles in monitoring the planet’s aerosol content, however each of these methods have inherent advantages and disadvantages. An overview will be given on some of the common methods to determine aerosol concentration and why using different methods helps to increase the overall accuracy and precision of data sets.

## 2.3.1 In-Situ Measurements

In-situ measurement have occurred on balloon based platforms and aircrafts. Balloon instruments that use particle counters during the assent directly count the aerosol particles and can determine the particle size distributions. The optical particle counter is an active instrument that uses an incandescent light source internal to the device to optically count aerosol particles. This instrument has been launched from Laramie, Wyoming since 1971 to determine aerosol extinction and particle size (*Deshler et al.*, 2003; 2008; *Kovilakam et al.*, 2015). Another type of balloon instrument uses a passive light source, including the sun, moon, or stars, to determine aerosol extinctions. Instruments that use this type of technology are the Absorption par les Minoritaires Ozone et NOx (AMON) from 1992 to 2003 and Spectroscopie d’Absorption Lunaire pour l’Observation des Minoritaires Ozone et NOx - Nacelle 2 (SALOMON-N2) from 2007 onwards which use starlight and moon light respectively (*Berthet et al.*, 2002). Furthermore, aircrafts have been used to carry nephelometers to acquire in-situ measurements (*Beuttell and Brewer*, 1949; *Charlson et al.*, 1969).

In-situ measurements of aerosol extinction give direct measurement of scattered light from the altitude that the balloon is currently situated and allows for direct measurements of aerosol extinction and cross-section unlike remote sensing applications from satellites. However, these types of instruments do not achieve global coverage and only give aerosol extinction from a very localized region, like the Laramie, Wyoming optical particle counters, or have very few flights, for example AMON which had a total of six stratospheric balloon flights, three mid-latitude northern and three high-latitude northern flights. In order to achieve full global coverage satellite remote sensing instruments were created to fill the spatial gap.

## 2.3.2 Occultation

Satellite instrumentation capable of remote sensing stratospheric aerosol has been in use since the 1970’s, beginning with limb sounding solar occultation measurements, and it operational geometry can be seen in Figure 2-3. Solar occultation measurements have provided a reliable, accurate and essentially continuous long term record of vertically resolved aerosol extinction coefficient measurements, mostly from the series of Stratospheric Aerosol and Gas Experiment (SAGE) instruments (*Russell and McCormick*, 1989; *Thomason and Taha*, 2003). These SAGE measurements, which have a vertical resolution of approximately 1 km, have generally compared well with ground-based and in-situ measurements, although there are challenges associated with comparing the retrieved extinction profiles to other microphysical parameters (*Russell and McCormick*, 1989; *Kovilakam et al.*, 2015). However, solar occultation is generally a robust and stable technique as it directly measures atmospheric optical depth, along with the exo-atmospheric solar spectrum with each scan, allowing for straight forward retrieval of aerosol extinction coefficient (*Damadeo et al*, 2013). The major drawback to occultation satellites is the number of measurements it can record in a single day since the instrument needs to be viewing a sunrise or sunset event which limits the geometry to only 16-48 measurements per day depending on the orbit. Although the currently operational MAESTRO and ACE-Imager instruments on SciSat (*McElroy et al.*, 2007; *Gilbert et al.*, 2007) have had some success producing stratospheric aerosol extinction products (*Vanhellemont et al.*, 2008; *Sioris et al.*, 2010), the era of solar occultation measurements essentially came to an end with SAGE III in 2006. However, a manifestation of SAGE III is planned for deployment on the International Space Station in 2016 (*Cisewski et al.*, 2014).



**Figure 2-3**: An occultation instrument monitoring the atmosphere by scanning the atmosphere by looking directly at the sun.

## 2.3.3 Lidar

Test

## 2.3.4 Limb Scatter

Test

# 2.4 Radiative Transfer

Test

## 2.4.1 Scalar Radiative Transfer

Test

## 2.4.2 Vector Radiative Transfer

Test

## 2.4.3 Rayleigh Scattering

Test

## 2.4.4 Mie Scattering

Test

# 2.5 ALI Prototype and Stratospheric Balloon Flight

Test.