A Dissertation

Entitled

A Method to Obtain Dust and Ice Cloud Optical Depths over the Cold Polar Surfaces of Mars

By

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Submitted as partial fulfillment of the requirements for The Doctor of Philosophy Degree in Physics

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An Abstract of
A Method to Obtain Thermal Spectra of Martian Dust Storms over Cold Polar Surfaces
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The Mars Global Surveyor Thermal Emission Spectrometer (TES) instrument has returned over 200 million thermal infrared spectra of Mars taken between March 1999 and August 2004. This represents one of the most complete records of spatial and temporal changes of the Martian atmosphere ever recorded by an orbiting spacecraft. Previous reports of the standard TES retrieval of aerosol optical depth have been limited to those observations taken over surfaces with temperatures above 210 K, limiting the spatiotemporal coverage of Polar Regions with TES. Here, we present an extension to the standard TES retrieval that models the effects of cold surfaces below 200K. This modification allows aerosol optical depth to be retrieved from TES spectra over a greater spatiotemporal range than was previously possible, specifically in Polar Regions. This new algorithm is applied to the Polar Regions to show the seasonal variability in dust and ice optical depth for the complete temporal range of the TES database (Mars Year 24, $L_s=104^\circ$, 1 March 1999 to Mars Year 27, $L_s=82^\circ$, 31 August 2004).

The University of Toledo

May 2007
To Stephanie, for being there at the darkest times

“When things look bad and it looks like you’re not gonna make it. Then you gotta get mean, I mean plumb mad dog mean. ‘Cause if you lose your head and you give up then you neither live nor win.... That’s just the way it is.” -:Clint Eastwood-: The Outlaw Josey Wales, 1976
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Chapter 1

Introduction

The planet Mars has played a long and detailed role in the rich history of astronomy. It is a planet that has generated immense curiosity and has been a subject of interest for scientists and the public alike for centuries. Mars has sparked the public imagination for as long as humans have charted the positions of objects in the heavens. This curiosity has fueled a long standing human desire to observe, monitor and more recently, robotically explore the surface and atmosphere of our interplanetary neighbor. In fact, Mars may rank as the most intensely studied planet in the history of science next to our own world and it’s Moon, yet there is still much to be learned about its surface properties and atmospheric dynamics.

In order to make new discoveries, new technologies and improved observing methods need to be developed. As the hardware that we use to gather observations evolves, so must our methods of extracting information from the raw data they provide.

Infrared spectra provided by the Mars Global Surveyor (MGS) Thermal Emission Spectrometer (TES) instrument have been used extensively to retrieve dust and ice optical depths from the atmosphere of Mars. Acting on the radiance spectra obtained by TES, optical depth retrievals have been performed by a radiative transfer and fitting algorithm capable of modeling the effects of the Martian atmosphere. While this algorithm has performed well during the course of many studies of Mars, a fundamental
limitation lies in its inability to model the specific effects that extremely cold underlying surfaces will have on atmospheric infrared spectra.

This thesis details the development of a new method to extract accurate dust and ice optical depths from infrared spectra retrieved over the frozen poles of Mars. This method represents the next stage in the evolution of the TES retrieval algorithm and the expansion of the spatiotemporal window over which reliable TES spectra may be obtained.

1.1 History of Mars Observations

Naked eye observations of the position and appearance of Mars are as old, if not older than astronomy itself. Evidence of observations from numerous past cultures can be identified today. The Maya people of South America created elaborate structures dating from 1200BC in recognition of the planet and its significance to their religion and calendar. A more qualitative example of ancient Mars observations lies in the remains of the great Eurimanki Ziggurat of Babylon. This structure was built with seven individual steps; each step dedicated to a different planet and assigned its own characteristic color. A brilliant red brick represents Mars, which occupied the third stage of the structure (Budge, 1925). Observations continued with many cultures conducting observations of Mars and noting its distinct red hue, this coloration most often drew direct association with the spilling of human blood, hence the planets current moniker: Mars, the Roman god of war. The historical quality of Mars observations has slowly evolved with progressing sensory technologies. The dynamic weather patterns of Mars play a long and detailed role in the history of modern astronomy, with many prominent figures in the
history of science contributing to our current state of knowledge. Galileo is recorded to have observed the phases of Mars in 1610, Christiaan Huygens in 1659 produced the first observations of surface features. As time progressed the polar caps were discerned by improving technologies in the possession of scientists such as Giovanni Cassini and Robert Hooke. Mapping of the planet has continued to improve, and with it our recorded history of the surface features and weather of Mars. As a consequence, such features as the varying observational properties of the prominent Martian poles have been under scrutiny for several hundred years.

The first reliable observations of Martian weather were probably conducted by William Herschel, who may be considered one of the most notable astronomers in history given the scope of his contributions to the science. Herschel devoted much of his early career to the study of Mars and was the first to track bright cloud features in many locations on the Martian disc (Cattermole, 1992). Herschel paid particular regard to the changing face of the South Pole and the question of whether Mars possessed an atmosphere. Herschel drew comparisons between the poles of Mars and the Earth (Sheehan, 1996).

This excerpt from Herschel’s own notes shows his thoughts on the nature of the south pole: “If, then, we find that the globe we inhabit has its polar regions frozen and covered with mountains of ice and snow, that only partly melt when alternately exposed to the sun, I may well be permitted to surmise that the same causes may probably have the same effect on the globe of Mars; that the bright polar spots are owing to the vivid reflection of light from frozen regions; and that the reduction of those spots is to be ascribed to their being exposed to the sun”. (Herschel, 1784) Herschel conducted
numerous observations of Mars to determine the existence of an atmosphere and its nature. It was around this time that Beer and Madler were also mapping brightness changes on the surface of the planet, noting prominent changes in shade between observations. The mid-nineteenth century saw a surge in the interest of Martian observation by both amateurs and professionals. It was around this period that astronomers such as Antoniadi, Flammarion, Schroter and Schiaparelli noted the distinctive changes in the color of the Martian disc from dark red to brilliant white (Sheehan,1996). Schiaparelli also reported to have observed a large scale dust storm in 1806 (Martin & Zurek, 1993). The consistent observation of Mars since the late 1800’s has allowed the planet’s weather activity to be followed closely over time. The occurrence of dust storms has now been experienced by a multitude of different observers, both amateur and professional. A complete discussion of all large scale dust phenomenon on record is beyond the scope of this document. However, the interested reader is referred to Martin and Zurek (1993) which gives a very complete account of all major dust activity recorded since the late 1800’s.

The earliest observation of an actual dust storm may be credited to Flaugergues in 1796 (Mckim, 1996), who discussed the rapidly changing features of Mars. Many have surmised that he actually witnessed a dust storm in progress, if correct this would be the earliest report of an actual dust storm on Mars. Beer and Madler may also have observed Martian dust storms in the 1830’s during their meticulous campaign to observe brightness changes on the planet’s disc. It is not confirmed beyond reasonable doubt that they observed dust storms or even realized they may have been observing such phenomena at the time. The credit for the first satisfactory observation of a Martian dust storm or
“yellow clouds” as they were known goes to Giovanni Schiaparelli in 1877. Controversy abounded in the late 19th century over changing Martian features with the now infamous Percival Lowell and many of his contemporaries observing “linear features” or canals as they have come to be known today. This era of Mars exploration is almost dominated by discussion of the changing features of Mars. Thanks in part to Lowell’s exposition of synthetic Martian canals and the literary concoctions of H.G.Wells, public interest in the red planet began to increase. While it did draw attention to some serious science, a rising tide of sensationalists, canal observers and Martian hunters emerges. It must be noted that Schiaparelli’s “canale” (Italian) referred to “channels” and were thought to be of natural origin by most scientists (Lowell and his camp excepted).

Despite the overtly colorful imaginations of the public and the internal squabbling of astronomers, Martian astronomy did see an increase in both public and scientific curiosity over this period. This surge in attention along with improved instrumentation lead to countless observations of both cloud patterns and dust storms, along with improved measurements of the North and South polar cap annual regression cycles. It was not until 1924 that Antoniadi suggested that the “yellow clouds” were most likely caused by dust raised by the action of wind (Sheehan, 1996). Today, we have access to better observing tools and greater background knowledge of the atmospheric mechanisms at work on both Earth and Mars.

Historically, observers had to contend with the age old problems of atmospheric disturbances and the significant variation in the angular size of Mars as it progresses along its slightly eccentric orbit. Today we are fortunate enough to have access to far more sophisticated observing methods, allowing Martian weather to be observed in a
similarly detailed manner to the weather of our own planet. The space age has changed
the way in which Martian planetary science is conducted. It is now possible to place
spacecraft capable of conducting systematic mapping of the planet at a variety of
wavelengths into stable orbits around Mars.

This thesis utilizes infrared spectra obtained by the Mars Global Surveyor (MGS)
thermal emission spectrometer (TES). These data were obtained from 1 March 1999 to 31
August 2004, (Section 1.3 and appendix E contain information on Martian time keeping.)
to detect and carry out detailed thermal observations of Martian dust storms over the cold
polar surfaces of Mars. However, thermal imaging of Mars is not a new or unusual
procedure. Infrared measurements of Mars are routinely conducted from ground based
telescopes and instruments capable of observing Mars at IR wavelengths have flown on
several spacecraft prior to MGS.

1.2 Orbital Infrared surveys

Mariner 9 made orbit in 1971 and became the first spacecraft to conduct
systematic mapping of Mars and obtain good infrared spectra and temperature readings of
Mars with both the infrared Interferometer Spectrometer (IRIS) and the Infrared
Radiometer (IRR). IRIS was the principal IR instrument designed to measure surface and
atmospheric compositions. Instruments such as these had previously flown on the
Mariner 6 and 7 missions, but they had only succeeded in mapping ≈1% of the Martian
surface. In fact scientists prominently involved in the Mariner 9 mission have stated that
the volume of data returned by MGS in a single day is equal to that returned by Mariner 9
over its complete lifetime.
The Viking orbiters which operated from 1976-1980 also observed the planet at Infrared wavelengths using the Infrared Thermal Mapper (IRTM) instrument, this device returned valuable information on both the geology and atmosphere of Mars. The Viking Orbiters had the opportunity to measure a great deal of atmospheric activity with the IRTM and observed a number of dust storms. While the spatial and spectral resolutions of the IRTM were not equal to TES, this data became the best available until Mars was revisited in the 1990’s.

Mars global surveyor (MGS) was launched on November 7, 1996 and achieved Martian orbit in September 1997 (Albee, 1998). The spacecraft has conducted systematic mapping of the planet since March 1999. MGS has conducted long term monitoring of the entire planet on a day-to-day basis in very good spatial and temporal resolution. The thermal emission spectrometer instrument mounted on the nadir deck has captured record numbers of good quality thermal spectra from both the atmosphere and surface of Mars. The TES instrument began mapping operations as the spacecraft entered mapping orbit on 1 March 1999 and obtained a complete IR survey of the Martian globe for each Martian day until the instrument ceased to function effectively on 31 August 2004. In Mars terms this has provided observations from $L_s=90^\circ$ (MY24) to $L_s=90^\circ$ (MY27). The specifications of the spacecraft and its mapping orbit are fully discussed in later sections along with details of the TES instrument. Appendix E and section 1.3 contain further information on both the Mars Year (MY) and Areocentric Longitude ($L_a$) methods of timekeeping.

Currently, the Mars Reconnaissance Orbiter (MRO) is conducting high quality infrared observations of Mars. MRO achieved orbit on March 10, 2006, but has not yet
accumulated the long consistent observational record of thermal infrared observations that have been obtained with MGS.

1.3 Martian Seasonal Cycle and Time Measurement

The orbit of Mars has an eccentricity of around 0.093 as compared to 0.017 for the Earth. This pronounced eccentricity results in significant variation in the distance between Mars and the sun over the course of a single Martian year. The Mars - Sun distance ranges from 1.36AU at perihelion to around 1.64AU at aphelion, causing a direct fluctuation in the level of insolation at the planet. At perihelion Mars is 20% closer to the sun compared to aphelion and so receives 45% more incident radiation at its closest point. (Cattermole, 1992). This relatively high eccentricity value also causes a pronounced inequality in the length of Martian seasons; as dictated by Keplers’ second law, the time any planet spends at perihelion is shorter than its time at aphelion. It is during perihelion that the Martian South pole is preferentially oriented towards the sun.

The combination of orbital eccentricity and axial inclination results in shorter, hotter summers in the Southern hemisphere as compared to the Northern. In fact, spring and summer combined are 77 terrestrial days shorter than the sum of the autumn and winter seasons.
Table 1.1 The Martian Seasons

<table>
<thead>
<tr>
<th>$L_s$(Degrees)</th>
<th>Northern Season</th>
<th>Southern Season</th>
<th>Mars calendar days</th>
<th>Earth equivalent days</th>
<th>Earth calendar days</th>
</tr>
</thead>
<tbody>
<tr>
<td>0º - 90º</td>
<td>Spring</td>
<td>Autumn</td>
<td>194</td>
<td>199</td>
<td>92.9</td>
</tr>
<tr>
<td>90º-180 º</td>
<td>Summer</td>
<td>Winter</td>
<td>178</td>
<td>183</td>
<td>93.6</td>
</tr>
<tr>
<td>180º - 270º</td>
<td>Autumn</td>
<td>Spring</td>
<td>143</td>
<td>147</td>
<td>89.7</td>
</tr>
<tr>
<td>270º - 360º</td>
<td>Winter</td>
<td>Summer</td>
<td>154</td>
<td>158</td>
<td>89.1</td>
</tr>
<tr>
<td></td>
<td>Total Days</td>
<td></td>
<td><strong>669</strong></td>
<td><strong>687</strong></td>
<td><strong>365.3</strong></td>
</tr>
</tbody>
</table>

Table 1.1 shows the relative length of the Martian seasons for their respective $L_s$ values in terrestrial days. When studying seasonal changes on Mars it is important to define a method of timekeeping which relates directly to the conditions on the planet caused by its changing orbital position. As a consequence, the areocentric longitude of the mean sun ($L_s$) becomes the preferred method when discussing the progress of time with respect to Mars. Areocentric longitude or $L_s$ is defined as the angle measured counterclockwise in degrees (Figure 1.1) from the vernal equinox (Martian North Spring equinox.). Perihelion occurs at around $L_s=250^\circ$ during the tail end of Southern spring (Northern Autumn). The increased insolation discussed previously is responsible for increased weather activity during this season, with more extreme pressure and temperature gradients resulting in higher wind speeds. While we discuss the actual seasonal progression in terms of $L_s$, it is convenient to be able to label each Mars Year (MY from here onwards) with a unique number.
Figure 1.1 Relative Motions of the Earth and Mars with respect to the sun. \( L_s \) (Areocentric Longitude of the Sun) is clearly labeled showing the definition of \( L_s \) as the angle from the line of equinoxes to the Mars-sun line. The outer red track shows the Martian orbit marked in \( L_s \), while the inner blue circle represents Earth’s orbit.

Appendix E contains more information on Martian time keeping. This paper utilizes the system of Clancy et.al 2000, in which ‘Year 1’ is arbitrarily selected to begin on April 11 1955. To put this system into the perspective of prior Mars missions, Mariner 9 would have occurred over MY 9-10, Viking from MY 12-15, Phobos from 19-20 and Pathfinder MY23. Using this system the first MGS TES observations would have taken place from: MY 24, \( L_s=104^\circ \) (1 March 1999) to MY 27, \( L_s=82^\circ \) (31 August 2004).
1.4 Martian Topography

1.4.1 The Martian Polar caps

Mars exhibits a pronounced hemispherical dichotomy in topographic elevation (see figure 1.2 and table 1.2). The majority of the southern hemisphere is covered by cratered highland terrain, while the northern hemisphere is composed of low lying terrain and exhibits fewer impact craters, indicating more recent resurfacing. The boundary of the dichotomy is marked by a steep escarpment between the cratered southern highlands and the smoother northern lowlands. This dichotomy contributes to the behavior of the Mars global circulatory system and the transport of volatiles between hemispheres. This thesis deals with the acquisition of infrared atmospheric observations over cold surfaces.
such as the poles. The polar caps and their physical characteristics have a significant impact on the overall circulation patterns of the planet.

Table 1.2 Comparison of North and South Polar features

<table>
<thead>
<tr>
<th>Parameter</th>
<th>North Pole</th>
<th>South Pole</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum Elevation*(m)</td>
<td>-1905.2</td>
<td>4802.5</td>
</tr>
<tr>
<td>Maximum Relief* (m)</td>
<td>2744.8</td>
<td>3052.5</td>
</tr>
<tr>
<td>Mean Elevation* (m)</td>
<td>1015</td>
<td>783.3</td>
</tr>
<tr>
<td>Area</td>
<td>1.12x10^6km^2</td>
<td>1.16x10^6km^2</td>
</tr>
<tr>
<td>Volume</td>
<td>1.505x10^6km^3</td>
<td>1.66 x10^6km^3</td>
</tr>
</tbody>
</table>

* = Measurement relative to mean surface height of Mars

The polar caps are of sufficient albedo and spatial extent that it is possible to obtain accurate observations of their seasonal expansion and regression even with rather modest ground based telescopes. Consequently, a long record of Martian polar observations exists. However, due to the eccentric nature of Mars’ orbit and the length of the Martian year, complete observation coverage of both poles for the entire Martian year is impossible without orbital monitoring.

The terrain at both poles is composed of a combination of layered geological terrain and residual ice deposits. However there are significant differences in the elevation and physical size of each cap. Figure 1.2 shows Mars Orbiter Laser Altimeter (MOLA) maps for the North and South poles and effectively illustrate the contrasting elevations present at each pole.
The polar caps harbor the majority of the volatiles in the Martian system and are subject to current geological activity. Volatiles are channeled towards and away from the poles during the autumn-winter and spring-summer seasons respectively. A network of subsurface chasms and valleys are believed to extend from the poles to latitudes of around 80°N and 80°S respectively.

These areas exhibit extensive dune fields at this latitude, which bear the effects of extensive wind erosion. Examination of the North and South Polar terrain indicates their vital role in relation to the dust, water and carbon dioxide cycles that form the basis of the Martian weather system. The Martian polar caps are heavily influenced by seasonal heating and their expansion and regression is well documented. The South Pole is preferentially oriented towards the sun at perihelion, resulting in longer and more intense summer seasons in the southern hemisphere compared to the North. This differential in seasonal insolation results in a greater variation in size of the south polar cap compared to the North.

The polar caps actually consist of two components, an ephemeral and a permanent residual ice cap. The seasonal cap is formed as CO$_2$ condenses out of the Martian atmosphere during the autumn season. This remains on the ground through winter and dissipates in the spring. During the condensation process, heavy CO$_2$ clouds linger over the cap (See figure 1.3) and serves to retard condensation.

The ice cap retreats towards the pole during the spring season, leaving remnants of polar water ice at lower latitudes which can last for several days. With the disappearance of the residual cap, only an area of water ice deposits interspersed with deep swirling valleys remains. The North residual polar cap is around 1000km across at
Figure 1.3. The changing features of the North Polar cap from October 1996 to March 1997. Note the reduction in the overall extent of the polar caps over this period. (See Appendix D for more detailed comparison.)

its widest point, while the south residual cap can shrink down to approximately 350km at the height of summer (Benson, 2003).

1.4.2 North Polar Cap Physiography

The North pole is approximately 3km higher than surrounding terrain; however, the entire Northern hemisphere is depressed by around 5km in comparison to the Southern hemisphere. (Zuber et al., 1998c) The North polar cap exhibits complex structure with a network of chasms spiraling around the pole. MOLA data indicates that these chasms provide a sink to wind blown dust in the area (Zuber et al., 1998c). A large proportion of the ice cap terrain is very smooth exhibiting regional slopes which exhibit a
gradient of around 0.2° over a distance of several tens of kilometers. Local topography appears to be relatively smooth down to the meter spatial scale. Terrain in the vicinity of the outer cap is relatively similar to the main cap with more frequent small amplitude variations than the central cap, indicating the presence of stagnant ice. Comparisons to gravitational surveys of the area suggest subsurface lifting is taking place or a complex structure exists below the surface (Johnson et al., 2000).

The aforementioned spiraling canyon systems are seen to extend outwards into the Olympia Planitia region, supporting the theory that this area was once contiguous with the polar cap itself. The North residual cap is rich in water; typical summer surface temperatures in these areas are around 205K (-68°C). This temperature is too high for CO₂ frost to form, however it is sufficiently close to the frost point of water for an atmosphere exhibiting such a small proportion of precipitable moisture.

1.4.3 South Polar Cap Physiography

The south polar cap lies at an average elevation of around 783km above the mean surface height; however the Southern ice cap lies around 6km higher than its northern counterpart (Smith & Zuber, 1996; Smith et al., 1998c). This is due in part to the hemispherical dichotomy that exists on Mars. The highest point at the South Pole is located at 87S 10E, well within the residual ice deposits (See table 1.2). The appearance of the local terrain suggests that sub-surface ice extends beyond the edge of the polar cap. The polar cap itself occupies a much smaller spatial extent than the North Pole. The same type of layered terrain seen in the North Pole is also present in the South. In the southern case this layering extends much further from the cap and possesses an asymmetric
distribution compared to the North (Zuber et al., 1998b). The rotational pole of Mars is not coincident with the center of the southern ice cap; residual ice which is present throughout the year is actually 35° to 40° east of the rotational pole. The south polar ice cap undergoes extreme variations in size, from almost disappearing in the summer months to extending as far as 50°S during the winter season. Little water is present in southern residual cap. This cap is composed primarily of CO₂ as winter temperatures in the region are far colder than the Northern case, at around 160K (-13°C).

1.5 The Martian Atmosphere

Martian weather systems are very dynamic, with driving forces that are not dissimilar to those in operation on Earth. Differentials in pressure and temperature induce
atmospheric movement on a variety of scales, resulting in atmospheric phenomenon ranging in size from small local dust devils to local and even globe encircling storms (See figure 1.4).

The Martian atmosphere is extremely tenuous compared to that of Earth; while Earth exhibits a mean surface pressure of around 1 bar, the corresponding value for Mars lies at around 6-7mbar (Owens, 1992). An effect of the low atmospheric pressure is that the atmosphere reacts far more quickly to incident energy than the atmosphere of the Earth. This low pressure combined with the thermal inertia properties of the surface give rise to large variations in surface temperatures over the course of a single diurnal cycle (Golombek et al, 1997). The thin atmosphere of Mars allows larger amounts of incident radiation to reach the surface of the planet than Earth. This means that the atmosphere of Mars responds far more quickly to direct solar heating than on Earth. Consequently, Mars is subject to very large atmospheric tides. The low atmospheric pressure at the surface of Mars also results in diminished surface stresses; consequently elevating dust from the planet’s surface becomes far more difficult. While dependent on the properties of the individual dust particulates, it is generally thought that sustained wind speeds on the order of 18-22ms$^{-1}$ are required to raise significant amounts of dust into the atmosphere (Cattermole, 1992).

Although the exact cause of global scale Martian dust storms remains unknown, common theories involve the convergence of global and local scale circulation patterns inducing high wind speeds, harmonic resonance patterns in the atmosphere or air circulating rapidly between hemispheres via tropical circulation mechanisms. The commonly accepted temporal window for the occurrence of global dust storms or the
“classical” dust storm season as it is known, is usually between \( L_s = 150° - 340° \) (Martin and Zurek, 1993).

Despite our lack of knowledge concerning the formation of large scale dust events, we have reliable measurements of the effects that increased dust activity has on the seasonal evolution of the Martian climate.

### 1.6 Atmospheric Aerosol Activity

Despite the relatively small amount of \( \text{H}_2\text{O} \) suspended in the Martian atmosphere, it is constantly approaching the water vapor saturation point. This results in large water based clouds and surface fogs which persist throughout all seasons. Larger cloud formations are easily observed even from ground-based observatories and can often result in the visual brightening of entire regions. Such is the case with the Hellas Basin, which historically has been known to be mistaken for the polar cap itself. In relation to the poles, the most extreme case of water ice cloud formation is the occurrence of the North polar hood clouds. This cloud feature forms around the North Pole during northern autumn (\( L_s = 180° – 270° \)) and has been noted to extend as far south as 50°N. This behavior is not observed at the southern pole which sees smaller more broken cloud during autumn along with a random temporal distribution.

The South Pole experiences its heaviest cloud cover at the same time as the North Pole, during the southern spring season (\( L_s = 180° - 270° \)). Polar hood clouds in both cases are composed of water ice with trace amounts of \( \text{CO}_2 \). Cloud formations are a common occurrence in the vicinity of major topographic features such as impact craters and
Figure 1.5. Cloud Formation over topographically preferred areas. Note the presence of prominent cloud in the vicinity of the Tharsis Rise volcano chain and the pattern formed around the equator.

volcanoes. In the case of volcanoes, clouds are formed by growth of orographic clouds (See figure 1.5).

This type of cloud is heavily dependent on diurnal variations, developing slowly in the early to late morning and reaching maximum development in the late afternoon. Convective clouds can also form over regions of high elevation around midday. Topography dependent cloud formation is the result of atmospheric instabilities formed by intense heating of the surface. Low lying regions such as canyons exhibit hazes and fogs at dawn and dusk; these are formed by the same mechanisms as their earthly counterparts as ground frost melts due to vaporization by solar radiation.
Although both poles exhibit relatively uniform topography compared to other Martian locales, local surface variations may nevertheless be responsible for changes in local weather phenomenon. Martian clouds may form up to altitudes of 55km above the surface; however MOLA results indicate that the majority of Martian clouds usually reside at altitudes between 2 and 8km (Pettengill and Ford, 2000).

Atmospheric activity such as surface winds may result in the advection of dust particles into the atmosphere. Fine dust particles may reach significant altitudes extending into the upper troposphere. The actual size/mass of dust raised and its maximum altitude is dependent on the strength of the circulation current/storm that rendered the particles airborne. In the colder upper atmosphere these dust particles may serve as nucleation sites for atmospheric vapors (CO$_2$ or trace H$_2$O), allowing ice particulates to form around them, coating the dust (Maattanen and Granada, 2006).

Figure 1.6 shows a comparison of nucleation rate to atmospheric temperature. Nucleation rates climb rapidly with falling temperature. This effect is well observed on Earth and replication of Martian atmospheric conditions in the laboratory show this mechanism to be in effect in the Martian atmosphere (Glandorf et al 2002).

1.7 The Role of Dust

The atmospheric circulation patterns of Mars are heavily influenced by dust particulates. Dust suspended in the atmosphere is an important source of positive
Figure 1.6. Heterogeneous Nucleation rates in the Martian atmosphere for water and Carbon Dioxide (Reproduced from Maatanen and Granada, 2006)

feedback for atmospheric circulation (Haberle et al., 1993b). Dust particles absorb incident solar radiation far more readily than the CO$_2$ in the atmosphere. Dust particulates then reemit this energy in the form of infrared radiation which can be absorbed by CO$_2$. This reemission serves to warm the air surrounding the dust, creating local temperature gradients and encouraging atmospheric circulation.

This enhanced circulation results in stronger gusts and overall higher speed winds at the surface, which serve to enhance dust raising even further, rendering more dust airborne. Through this mechanism dust is redistributed about the surface on local and regional scales, the distance transported being dependent upon circulatory conditions at that time. Redeposited dust also has an effect on local vertical and horizontal temperature and pressure gradients, modifying the local circulation. The transported dust settles out of the atmosphere onto the ground, resulting in changes to the thermal response of the surface as deposited dust will alter the albedo and thermal inertia of the surface.
Figure 1.7 shows a dust storm in progress over the boundary of the North polar cap. Bonev et al (2002) suggest that deposition of dust over polar cap ice during dust storms actually influences the annual regression of the polar cap. In fact polar recession rates are seen to accelerate during years exhibiting heavy dust activity. Large scale regional or global dust storms can raise enormous amounts of dust into the atmosphere.

The effects of this dust while airborne and its subsequent relocation onto the ground can result in truly global changes to atmospheric circulation patterns.

1.8 Dust Storms and Dust Raising Mechanisms

Dust is rendered airborne in a variety of ways; the exact mechanism being dependent upon the size of the individual grains (James, 1985). Surface winds, saltation and dust devils all contribute to dust lifting. Laboratory wind tunnel experiments such as those by Greeley et al (1980) have simulated Martian surface conditions and have
constrained the minimum wind speeds required to raise dust particles from the surface. An average wind speed of 50ms$^{-1}$ is sufficient to render dust particulates 100µm in size airborne. On more differentiated terrain (surfaces with numerous boulders and pebbles) lower wind speeds of only 25-30ms$^{-1}$ may result in the lifting of dust particles (Greeley et al., 1980). This is consistent with meteorological observations obtained by the Viking landers in the 1970’s. It has been suggested that local dust storms with significant grain sizes (~100µm) are generally short lived as the mass of the dust particles causes the particles to succumb more readily to the effects of gravity and settle out much more quickly than lighter dust grains (James, 1985). These heavier particles will also be more likely to remain near the surface level, as wind speeds are insufficient to raise them to significant altitude. Consequently, dust that is elevated to significant altitudes (as in a global dust storm) is likely to be of finer composition. Saltation and dust devil processes are suggested as the main elevators of fine dust grains.

The process of saltation involves the lifting of relatively large particles on a short ballistic trajectory. The dust particle collides with the ground or other geological obstruction, launching secondary dust particles form the surface and possibly even fragmenting itself. Particles then undergo further collisions resulting in fine grained dust being rendered airborne (see figure 1.8).

Only large dust particulates would contain sufficient mass to cause saltation. Greeley et al., 1980 suggests that surface wind speeds on the order of 25-30ms$^{-1}$ would be sufficient for saltation to occur over differentiated terrain. Dust devils are frequent occurrences on the surface of Mars and may be one of the chief dust raising mechanisms,
Figure 1.8 The Process of Saltation. Surface winds incident from the right repeatedly lift particles, impacting them upon the planets’ surface. This serves to raise dust and create finer particulates through fragmentation of the ballistic particle.

as a wide range of particle sizes may be rendered airborne through the action of atmospheric vortices (Greeley et al., 1981).

Suspended dust remains in the atmosphere for timescales of hours to hundreds of days in the case of a global storm, this timescale being dependent on both grain size and wind speed.

Although local dust storms occur frequently throughout the Martian year (Cantor 2001), very large scale storms tend to occur randomly within a specific and repeating temporal window. These large scale storms are seen to cause an enormous increase in global optical depth. Historical records indicate that global dust storms occur preferentially between $L_s=161^\circ$ to $L_s=326^\circ$, which to the Martian perihelion season of late Southern winter through early Southern summer. All global dust storms observed to date have occurred over this period (Smith et al, 2002).
1.9 Atmospheric circulation patterns

The general circulation of Mars has much the same driving forces as Earth. Mesoscale and local winds act with baroclinic waves, planetary waves and Hadley circulation cells to drive bulk atmospheric movements. Martian winds move in varying directions and are induced predominantly through variations in temperature and pressure gradients. Martian winds may be diurnal, semi-diurnal or much longer lived.

1.9.1 Local Circulation Patterns

Atmospheric temperature and pressure gradients may result from differentials in the thermophysical properties of the surface. This differential in surface thermal response (due to albedo and thermal inertia) is especially prominent at the poles, where frost cover is non-uniform. Variations in frost coverage along the boundary of the polar caps coupled with the sublimation of ices as the polar caps recede over the spring/summer period encourages temperature differentials to form. Changes in topographic elevation will also contribute to this effect. Winds are believed to flow downslope from the colder, higher elevation centre of the cap to the lower outlying polar areas, creating winds which are predominantly away from the pole (Haberle, et al., 1993b). These variations in local surface temperature may be responsible for the characteristic slope (see section 2.4 on thermal contrast) observed in the brightness temperature spectra retrieved at the poles. This effect may be due to the presence of “patchy” CO$_2$ and H$_2$O frost coverage, where a range of surface temperatures are present in a single field of view. Local zones of low
Figure 1.9 The General Circulation of the lower Martian atmosphere. a) Double Hadley circulation formed at equinoxes ($L_s=270^\circ-360^\circ$). The polar cap winds interact with low latitude thermal tides. b) Single Hadley cell in operation at the solstices ($L_s=0^\circ-180^\circ$). This cell is more reliant on cross equatorial thermal tides and less on polar cap winds (from Haberle in Kieffer et al. 1992).

thermal emission may also exist due to the condensation of CO$_2$ ice in the atmosphere, this effect may be especially prominent at night and is discussed more fully later on.

1.9.2 Global Circulation patterns

Hadley cells are an important component of the Martian circulation patterns and are responsible for the majority of cross equatorial atmospheric exchanges. The Martian Hadley cell is composed of both rising and descending branches, from this cell remains centered at 30°N and 30°S from $L_s=270^\circ-360^\circ$. The ascending branch of the Hadley cell is responsible for lifting warm southern air to altitudes of around 40km, channeling it North across the equator. The majority of the air cycled by the rising branch flows
through the lower few kilometers of the atmosphere (Haberle, et al., 1993b). Figure 1.9 shows a schematic diagram of the two Hadley circulation cell systems.

The air placed on a Northerly trajectory will cool at high altitudes and descend at around 30°N. Low altitude air flowing South towards the ascending branch of the Hadley circulation cell from the Northern hemisphere is subject to the coriolis forces resulting from the rotation of Mars as it transits the equator. This effect serves to enhance easterly airflow. The circulatory system of Mars shows far more temporal variation than its Earthly counterpart. Mean zonal flow rates can vary by up to an order of magnitude from autumn to winter (Haberle et al 1993b). Upper atmosphere zonal wind speeds in the summer and winter hemispheres are around 60 ms\(^{-1}\) and 120 ms\(^{-1}\) respectively.

This circulation system is weakest at the Spring and Autumnal equinoxes \((L_s=0^\circ\) and 180°), where the Martian system forms two distinct Hadley cells (similar to the Earth system). At this time the predominant winds flow in an easterly direction around the tropical regions and west at mid-latitudes. Atmospheric circulation increases greatly at the solstices and the two individual cells merge into a single circulation cell spanning just under 50% of the planet. While this single cell is in operation, the summer hemisphere is subject to mainly westerly winds and the winter hemisphere easterly wind. Dust introduced to the atmosphere may exaggerate these winds at the surface level and increase atmospheric temperatures (Newman, 2005). This serves to create a statically stable system, where vertical transport of air is inhibited and the system is driven by meridional temperature gradients, resulting in strong zonal winds.

A great deal of work has been done to model the atmospheric circulation of Mars for evidence of repeating patterns (Forget & Pollack, 1996, Newman et al., 2002).
Detailed atmospheric models have existed for the Earth for a number of years (Leovy & Mintz, 1969) and have been used as a basis to model the Martian system.

Fortuitously, the task of modeling the Martian atmosphere is simplified by the absence of the large bodies of water possessed by the Earth.

Due to the absence of water, the Martian surface responds far more quickly to solar heating than the Earth. Topography however, does play a significant role in the formation of Martian global circulation patterns. The thermal response of Mars is very rapid and this reaction to variations in insolation imposes a distinct seasonal dependence on the planet’s circulation patterns. The Martian circulation system is driven by thermal gradients that exist between the respective summer and winter hemispheres. Warm air rises over the preferentially heated summer hemisphere and descends over the cooler winter hemisphere.

This circulation mechanism has severe consequences on polar weather conditions. Around 30% of the Martian air condenses over the winter pole, generating a high latitude low pressure region. This creates extreme pressure gradients which induce strong global circulation in the direction of the winter pole. Surface ice in the vicinity of the respective winter pole then undergoes an expansion as CO\textsubscript{2} condenses over the region. This flow of condensation dominates the wind directions at all latitudes.

The winter hemisphere usually experiences prevailing westerly winds in the mid to high latitude range. Both cyclonic and anti-cyclonic weather systems move across the planet, figure 1.10 shows a cyclonic weather feature in the vicinity of the South Pole. Mid-latitude weather follows patterns not dissimilar from Earth.
Despite the differences between the Earth and Martian weather systems, the immutable laws of physics still dictate the overall behavior of mass air flow. Figure 1.11 shows a comparison of a dust storm on Earth in the vicinity of North Africa and a Martian dust storm over the South Pole. Despite the differences in the mechanism of air movement in each case, the results are remarkably similar in appearance.

The effects of the summer season are radically different in each hemisphere, circulation patterns in each area also change significantly between summer and winter. Northern summers are generally less active than their southern counterpart, with only slight east-west zonal air flow.

Cyclonic systems rarely develop during the northern summer period. In contrast to this, southern summers are prone to frequent dust activity. Southern summer dust storms are often sufficient to interfere with the predominant north-south temperature
Figure 1.11 A comparison of dust storms morphology on Earth over Northern Africa (left hand side) and the south pole of Mars (right hand side)

gradient disrupting global circulation patterns. Storms of this magnitude usually occur in conjunction with the retreat of the south polar cap as discussed earlier.

The southern summer polar cap gives rise to large pressure gradients between the newly exposed surface and the retreating ice (James, 1999). The thin nature of the Martian atmosphere affords little heat capacity. This means that the Martian atmosphere warms and cools much more quickly than Earth’s. The predominant atmospheric species is CO$_2$, which absorbs well in the IR. Consequently the lower levels of the atmosphere see a large range of diurnal variation. When there is little to no suspended atmospheric dust, the Martian air absorbs very little of the direct solar energy. This means the observed temperature profile is mainly affected by heat transfer in the form of both conduction and convection from the planet’s surface. Although this gives rise to large diurnal fluctuations in surface temperature (surface temperature can often vary by as
much as 50°C) these fluctuations rapidly diminish with increasing height. This results in weak diurnal temperature variations and tidal winds as only atmospheric layers near the surface are affected.

High levels of suspended dust significantly change the response of the atmosphere. Dust grains absorb incoming solar radiation so direct heating of the surrounding air takes place. The atmosphere cools more readily at night, resulting in a nearly isothermal temperature profile. This effect reverses the case of a clear atmosphere, narrowing the temperature range at low altitude and increasing the range at greater altitude (see temperature profile in figure 1.12).

It is under these conditions that tidal winds reach their peak strength. Local variations in wind speed and strength may be induced by topographic features and the composition of the local features. The rapid response of the Martian atmosphere to
surface temperature makes weather activity far more dependent on the local surfaces’ albedo, topography and thermal inertia than the Earth. Temperatures are strongly dependent on reflectivity and the local insolation cycle, consequently the ambient temperature at the base of a large Martian volcano is roughly equal to the temperature at its summit. Large horizontal temperature and pressure gradients will give rise to strong winds on the flanks of the volcano. These will cycle between downslope winds at night and upslope winds in the daytime. This type of topography induced weather is present at the poles; however elevation does not vary to the same extent and generally the rise is much gentler than for features in the Tharsis region.

1.10 Instrumentation

The thermal emission spectrometer (TES) consists of a Michelson interferometer spectrometer and obtains thermal infrared spectra (over the range 6 to 50µm, 1655-200cm$^{-1}$), thermal bolometric (0.3-100µm) and solar reflectance (0.3-3.9µm) measurements. Data are obtained from six 8.5mrad instantaneous fields of view, providing a maximum spatial resolution of 3km from the MGS mapping orbit of around 350km. Figure 1.13 shows a schematic of the TES instrument. For full technical specifications of the TES instrument, see appendix B. Despite the comprehensive database of Martian thermal emission measurements available to us, observations are bound by the physical limits of the detectors used to record our observations. In the case of any thermal spectra, a fundamental limitation lies in the thermal contrast of the
Figure 1.13 Schematic of the TES instrument (MGS mission plan. Final version, Rev B (MGS 542-405)

atmosphere. The thermal contrast is defined as the difference in temperature between the surface and the atmosphere as observed through a plane parallel atmosphere.

The minimum contrast range over which TES thermal emission spectra are deemed reliable down is from 20-40K. As surface temperatures fall, noise levels within the data tend to increase, yielding less reliable results. At surface contrasts lower than 20K, noise effects tend to dominate spectra. In most studies, spectra retrieved from areas with a surface temperature of less than 220K are considered too noisy to be of use. For this reason most TES observing projects are limited to areas which exceed 220K, limiting observations to daytime temporal windows and frost or ice free spatial windows.
Chapter 2

Initial Observations

Figure 2.1 TES daytime (local time ≈ 1400) zonally averaged aerosol optical depth and water vapor abundance. Shown is the zonal average of each quantity. Top: Dust optical depth at 1075 cm$^{-1}$ scaled. Middle: Water ice optical depth at 825 cm$^{-1}$. Bottom: Water vapor column abundance in precipitable microns (pr-µm)

This thesis attempts to quantify the effects of cold surfaces on TES atmospheric observations and to develop methods to compensate for them, utilizing detailed spectral models and radiative transfer codes. The goal of this project is to allow reliable observations to be obtained over areas and times previously neglected due to the cold surface limit.
Figure 2.1 shows a survey of interannual variability in zonally averaged dust and ice optical depth along with water column vapor intensities from Smith (2004). These results highlight the limitations of the previous retrieval algorithm, as evidenced by the absence of data in the extreme north and south. As data were unreliable over surfaces below 220K, a distinct seasonal dependence was placed on the spatiotemporal extent of the useable observations. The annual expansion and contraction cycle of the polar caps extends the area over which reliable observations could not be obtained. Consequently, the maximum interference due to poor thermal contrast resulting from a cold underlying surface occurs at $L_s=270^\circ$ in the Northern hemisphere (the North winter solstice) and $L_s=90^\circ$ in the south (during the south winter solstice).

2.1 Reconnaissance Observations

TES data is housed in a database accessible through the use of the “Vanilla” program, this command line program accepts user defined spatial, temporal and quality parameters and retrieves the requested TES derived values from the database.

In this case, calibrated radiance values were recalled along with surface and atmospheric temperature and pertinent spatiotemporal information such as latitude, heliocentric longitude and local time. Here we concentrate on deriving dust and ice aerosol opacities over areas of low surface-atmosphere thermal contrast, so data retrievals were limited to areas with surface temperatures less than 210K. Daytime spectra ($\approx 1400\text{hrs}$) were used throughout.

As we are focusing on the retrieval of optical depths over cold surfaces, it becomes unnecessary (and time consuming) to observe the entire planet in detail.
Figure 2.2 Zonally averaged northern hemisphere surface temperatures and corresponding temperature contrasts between the surface and the atmospheric temperature at the one scale height level for the complete TES record for 2:00pm local time. The temporal scale is in degrees of continuing $L_s$ running from $L_s=90^\circ$, MY24 to $L_s=90^\circ$, MY27.

Therefore initial reconnaissance observations were conducted to determine suitable areas over which to retrieve characteristic cold surface spectra and to develop and test new methods. While we are confident that the coldest locations on Mars are in the vicinity of the poles, it is useful to quantify surface temperatures and surface-atmosphere thermal contrasts for each location on the Martian globe.
Figure 2.3 Zonally averaged southern hemisphere surface temperatures and corresponding temperature contrasts between the surface and the atmospheric temperature at the one scale height level for the complete TES record. The temporal scale is in degrees of continuing L$_s$ running from L$_s$=90°, MY24 to L$_s$=90°, MY27.

Figures 2.2 and 2.3 show surface temperature and corresponding surface-atmosphere thermal contrast values for the Northern (figure 2.2) and Southern (figure 2.3) hemispheres compared to time in continuing Ls. Reliable observations are available over areas of prominent positive or negative contrast. Any region exhibiting a thermal contrast of less than 10K will be of insufficient contrast to allow atmospheric features to be discerned and a flat relatively featureless continuum will be observed.
Figure 2.4 South polar cap regression feature in MY24 observed in terms of temperature contrast. The high values of temperature contrast over the polar region depict the case of a cold surface and warmer atmosphere. Sample spectra were retrieved from this area to quantify the effects of surface temperature.

Regions with Tsurf<210K in both the extreme North and South are the areas previously neglected in the survey of Smith 2004 (recall figure 2.1). The Maps presented in figures 2.2 and 2.3 were utilized to determine the spatial and temporal ranges for initial studies. Key features such as the expansion and recession of the polar caps were targeted due to the presence of contrasting surface and atmospheric temperatures in these regions. Individual spectra were then recalled from selected times/zones and the effects of surface and/or atmospheric temperature were examined.

Figure 2.4 above shows one such target area, the south polar cap expansion and regression feature is seen here in terms of surface-atmosphere thermal contrast from L_s=
Figure 2.5 Characteristic Brightness temperature spectra and corresponding radiances for initial reconnaissance survey of the southern hemisphere at $L_s=356^\circ$.

Spectra from two areas are shown: 85°S (Left hand side) and 70°S (Right hand side). Brightness temperature in Kelvin (K), Radiance in $\text{Wsr}^{-1}\text{m}^{-3}$.

100° to 360° MY24. Spectra were retrieved from a variety of spatial and temporal locations around this polar evolutionary feature. Their morphology and dependency upon time, location and thermal contrast were then studied to form the basis of an effective model of cold surface spectra.
Figure 2.6 TES spectra showing strong dust and ice features. Dust dominated spectra appear as solid lines, Ice dominated spectra appear as dotted lines. The Spectral signatures of gaseous CO$_2$ (667 cm$^{-1}$), Water ice clouds (230 and 825 cm$^{-1}$) and dust aerosols (1075 cm$^{-1}$) are present (Reproduced from Pearl et.al. 2001).

Panel (a) shows the brightness temperature response to the presence of aerosols while Panel (b) showcases the radiance spectra

The 2 spectra shown in figure 2.5 were obtained on the same date L$_s$=356° from two different polar locations, 70°S and 85°S. These spectra were selected due to their proximity to a zone in which surface temperature was relatively low.
2.2 Martian Spectra

The Martian infrared spectrum contains vital information not just on the thermal structure of the atmosphere but also of the atmospheric composition. The presence of atmospheric gasses and suspended aerosols has distinct and significant absorption features within the infrared spectrum of Mars. Figure 2.6 and appendix I contain examples of daytime TES spectra containing strong features typical of the main aerosols present in the Martian atmosphere.

The most prominent feature in the IR spectrum of Mars is the narrow 667 cm\(^{-1}\) (15 \(\mu\)m) \(\text{CO}_2\) absorption band. This feature is the most reliable gauge of atmospheric temperature in the Martian spectrum (Conrath et al., 2000) and is easily identified in figure 2.6(a). Dust and water ice are also prominently displayed. Dust exhibits itself through a broad feature centered at 1075 cm\(^{-1}\) (9 \(\mu\)m), dust absorption is largely positive throughout the spectrum up to 1300 cm\(^{-1}\) where a slight reduction occurs. Water (\(\text{H}_2\text{O}\)) is detectable through a broad feature at around 825 cm\(^{-1}\) and another slightly sharper feature at 229 cm\(^{-1}\), this feature is especially prominent in figure 2.6(b) (see appendix I).

2.3 Polar spectra

Both the North and South poles exhibit zones of significantly reduced emission, resulting in low brightness temperatures for these areas. Polar spectra generally exhibit more anomalies than those observed in other regions of Mars. These anomalies in polar spectra are more common over the colder southern polar cap and are believed to be caused by the atmospheric condensation of \(\text{CO}_2\) (Forget et al., 1995). The presence of precipitating \(\text{CO}_2\) clouds with particle radii on the order of 10 \(\mu\)m along with localized
deposits of CO₂ snow (particle size ≈ millimeter) may account for many of the aberrant features. Forget & Pollack (1996) determined the true surface temperature of the polar caps by compensating for the effects of these low emission events in the Viking IRTM data. Studies such as these along with results from the Viking and Pathfinder regarding surface conditions were used by the TES observing team to increase the accuracy of the TES surface temperature measurements.

It was found that temperatures decrease systematically approaching the pole; the southern cap was estimated to be around 5K colder than its northern counterpart. Average temperatures of 145.5K were recorded in the North and 141.1K in the Southern case. The low emission zones are affected by the presence of dust, as dust aerosol levels increase, zones of low emission decrease. This effect has been observed by both Viking and MGS. It is possible that the strong IR emissivity of airborne dust particles strongly increases the cooling rate of the atmosphere during the polar night. This would cause the condensation of CO₂ ice into the atmosphere as opposed to on the surface, reducing the strength and number of low emission zones. Figure 2.7, reproduced from Christensen et al.1998, shows 5 averaged TES spectra obtained over the south pole during the early phases of the MGS mission. This figure demonstrates the differences inherent to Martian spectra at different locations over the pole. Each spectrum is composed of an average of multiple spectra to filter out noise effects inherent to observations of polar IR spectra. The hot surface shows a very deep 825cm⁻¹ CO₂ absorption line (indicative of the atmosphere). This absorption diminishes proportionally with decreasing temperature, which itself is dependent on proximity to the centre of the polar ice cap. Spectra obtained purely over the polar ice caps demonstrate a CO₂ band that is in emission. This is due to
Figure 2.7 The variation observed in the form of Mars brightness temperature spectra over locations exhibiting different surface temperatures and surface composition. The number attached to each spectrum identifies the number of raw spectra averaged to create each curve. (Christensen et al. 1998).

an inverse temperature contrast where the temperature of the surface is colder than that of the atmosphere.

2.4 Thermal contrast

The limiting factor in the observation of Martian atmospheric aerosol features is the thermal contrast. Thermal contrast is the difference in temperature between the surface and the atmosphere at a particular point. This is demonstrated in figure 2.8 which highlights the differences between a relatively warm brightness temperature spectrum with a flat continuum lying at around 206K and a colder surface temperature exhibiting a
Figure 2.8 Comparison of spectra over surfaces of different temperature. Note the difference in thermal contrast. Panel (a) exhibits a thermal contrast of 53K compared to the smaller 12K thermal contrast in panel (b).

positive gradient with brightness temperatures ranging from 155K – 166K. The Martian surface is responsible for the majority of the detected infrared radiance with a smaller atmospheric IR contribution.

The atmosphere is composed predominantly of CO\textsubscript{2} (see appendix F) and is responsible for the large 825cm\textsuperscript{-1} absorption feature. By comparing the brightness temperature of the atmosphere (≈153K) to the value of the overall continuum, we may estimate the thermal contrast of the system. In this case the 206K continuum is indicative of the surface temperature and the 153K trough of the CO\textsubscript{2} absorption line represents the atmospheric temperature at optical depth unity for this location. Despite the presence of a relatively cold surface, the atmosphere above is even colder with a reasonable thermal contrast of 53K allowing the observation of atmospheric absorption features. This
Figure 2.9 Samples of Brightness temperature spectra over progressively colder regions of the polar cap from 42°S to 88°S latitude, $L_s=140^\circ$. a) $T_{\text{surf}}>T_{\text{atm}}$

b) $T_{\text{surf}}\approx T_{\text{atm}}$
c) $T_{\text{surf}}<T_{\text{atm}}$

pronounced contrast ($T_{\text{atm}}<T_{\text{surf}}$) allows surface and atmospheric details to be resolved with a minimum of difficulty.

Figure 2.8 illustrates the decline in thermal contrast for spectra obtained over cold surfaces. Figure 2.8(a) shows a spectrum with a good thermal contrast of around 53K whereas 2.8(b) being more typical of polar spectra has a much lower thermal contrast of only 12K. In the cold surface spectrum of figure 2.9(b) the continuum has a non-zero slope and the introduction of stronger absorptions at higher wavenumber, compared to the relatively flat and featureless warm spectrum of 2.9(a). (For more examples of the effect of cold surfaces on brightness temperature spectra refer to appendix I).
For the case of a cold 200-210K surface, the temperature of the atmosphere is sufficiently close to the surface temperature ($T_{\text{atm}} \leq T_{\text{surf}}$) that spectral details indicative of the atmosphere will generally be obscured.

Figure 2.9 demonstrates the effects of varying thermal contrast in the vicinity of the polar cap. Figure 2.9(a), shows a spectrum obtained at 42°S, $L_s=140°$ with a thermal contrast of 20K so $T_{\text{surf}} > T_{\text{atm}}$. This allows atmospheric features to be differentiated from the surface. The spectra presented in figure 2.9b were retrieved from closer to the polar cap at 65°S. This spectrum exhibits almost no thermal contrast ($T_{\text{surf}} \approx T_{\text{atm}}$) and the spectral features of the surface and the atmosphere are almost identical. It is impossible to extract useful information on the disposition of the atmosphere for this case. For 2.9(c) the spectrum was obtained directly over the polar cap at 88°S, $L_s=140°$, reversing the situation seen in fig 2.9(a). The surface is now colder than the atmosphere above it ($T_{\text{surf}} < T_{\text{atm}}$), this is a case of inverse thermal contrast. In this case we can distinguish between surface and atmospheric details due to the atmosphere being in emission. The spectral data can be fit in the same manner as for warm spectra; however, a key difference lies in the slope of the brightness temperature spectrum. As the surface becomes colder, a distinct positive gradient manifests itself in the spectra (Compare graphs (a) and (c) in figure 2.9). This “continuum slope” may be caused by the presence of a range of surface temperatures in the field of view of the detector. Likely causes are the presence of differentiated terrain, areas which have thicker surface frost/ice deposits than the adjacent locale. Interference from ice condensing in the lower levels of the atmosphere may also be responsible. While the fitting routine of the retrieval algorithm can easily cope with inverse thermal contrast, this gradient represents a significant source
of error. One major innovation of this thesis compared to previous retrieval algorithms, lies in compensating for this continuum slope.

Thermal spectra obtained from March 1, 1999 ($L_s=103.77^\circ$, MY24) to August 31 2004 ($L_s=82^\circ$, MY27) have been retrieved in order to determine appropriate times and locations for an in depth study, dust optical depth, water ice optical depth and corresponding surface temperature were observed for all available times. These quantities are to be used to determine the conditions both globally and locally along and allow the presence of water ice to be determined for a given set of conditions.
Chapter 3

Data Processing and the Opacity Retrieval Algorithm

The opacity retrieval is composed of a radiative transfer algorithm and a least squares fitting routine. The column integrated opacity is calculated as a function of wavenumber and the total contribution of dust and ice present in the opacity spectrum is obtained by fitting for two predetermined dust and ice spectral shapes.

3.1 Radiative Transfer

If we neglect scattering at this juncture, the method chosen to retrieve temperature values will be as follows: If we consider the contribution from solar radiation in the thermal infrared bands (direct from the sun or by scattered/reflected sunlight) to be negligible, then thermal infrared spectra observed from orbit are a result of photons emitted by the surface of the planet, and the corresponding radiance will follow a standard blackbody distribution. These photons originate from Martian surface and are absorbed by gasses or aerosols in the atmosphere. The CO$_2$ rich atmosphere subsequently emits the absorbed energy at its own characteristic temperature. This temperature is influenced by the amount of direct solar energy absorbed and the radiation emitted from the Martian surface. The photons emitted through vibrational transitions of CO$_2$ will follow a thermal emission spectrum characteristic of the atmospheric temperature of the
molecule or aerosol particle. This means that the thermal emission we observe from orbit is the sum of all of the emitted contributions along a line of sight.

The value retrieved by TES is representative of the radiance of a particular area of the planet. We assume that we are dealing with nadir observations of a plane parallel atmosphere and that scattering due to aerosols is negligible, the radiance along a particular line of sight is equation 1:

\[
I(\nu) = \int_0^{\tau_0} \left( B_{(\text{Surf}, \nu)} e^{-\tau_0 e} + \varepsilon(\nu) B_{(\tau, \nu)} e^{-\tau} \right) d\tau
\]

where:

- \(I=\text{Radiance}\)
- \(B(\tau)=\text{Planck radiance for given temperature at optical depth } \tau\)
- \(\tau=\text{Atmospheric optical depth}\)
- \(\tau_0=\text{Total optical depth along LOS}\)
- \(e^{-\tau}=\text{attenuation (absorption) of the reemitted radiation along observers path}\)
- \(\varepsilon=\text{Surface emissivity} \approx 1\)

This integral is performed over the optical depth (\(\tau\)) and repeated frequency by frequency over the desired range. Both optical depth and the Planck function are functions of radiance. An integration conducted from the spacecraft to the surface of the planet, results in the column integrated opacity (the result of equation 1) for a specific value of \(\tau_0(\nu)\). Radiance is retrieved for each wavenumber and used as an input along with the surface and atmospheric temperatures retrieved by TES. Using equation 1 the expected radiance is calculated for increasing opacity until pair of opacity values is found which bracket the observed radiance. The specific value of opacity for which the observed and calculated radiances agree is then obtained through conventional root
Figure 3.1 Contributions to the observed radiance of Mars from successive atmospheric levels and attenuation mechanisms as observed from Mars Global Surveyor. (Note the change in layer spacing between the surface and the upper atmosphere.) Where, $\varepsilon =$ Surface emissivity $\approx 1$.

finding techniques. Two aerosol spectral features, dust at 1075 cm$^{-1}$ and water ice at 825 cm$^{-1}$ were fit for to obtain dust and ice column densities. Given the spectral shapes of dust, $f_{dust}$ and ice, $f_{ice}$ and their respective scaling factors $A_{dust}$ and $A_{ice}$, the total optical depth is:

$$ \tau_0(\nu) = A_{dust} f_{dust}(\nu) + A_{ice} f_{ice}(\nu) \quad (2) $$

The atmosphere can be hypothetically divided into plane-parallel isothermal layers. Each layer is deemed to be very thin, hence the constant temperature. This method is most reliable when the increment in optical depth is less than unity.
The algorithm developed to undertake this task follows the method originated by Smith et al.2000a, utilizing 90 atmospheric levels to model the Martian atmosphere. The distribution of each atmospheric ‘layer’ is such that the layers are distributed further apart with increasing distance from the surface (i.e. consecutive layers near the surface are much more closely spaced than layers in the outer atmosphere as in figure 3.1). A height \( z_p \) above the surface is defined for each layer of the atmosphere and is measured in units of the scale height (the length scale over which atmospheric pressure drops by a value of \( e \)), where \( z_p \) has a value of zero at the surface of the planet. For example, the vertical height and subsequent pressure of the \( j \)th level of the atmosphere would be found by the expressions:

\[
z_p(j) = -\log(1 - e^{-4(\text{levels} - j + 1/\text{levels})})
\]

(3)

where:

\( n = \text{number of levels} \)

\( j = \text{level number} \)

Note: \( z_p = 0 \) at surface level

and

\[
p(j) = P_{\text{surf}}(e^{-z_p(j)})
\]

(4)

Where: \( p(j) = \text{Pressure at } j \text{th level} \)

\( P_{\text{surf}} = p_0 \) (surface pressure)

As the atmospheric pressure falls exponentially with increasing altitude, \( z_p \) has a rough proportionality to linear height above the surface. If we make the reasonable assumption that the scale height of Mars is around 10km, the linear height above the
surface can be approximated from our atmospheric level numbering system using the following simple relation:

$$z(j) \approx 10 \times z_p(j) \text{km}$$ \hspace{1cm} (5)

Spectra are computed by summing contributions from each atmospheric layer. The term “level” refers to the infinitely thin divisions between each “layer”. 91 levels are deemed to exist in this model separating 90 discrete layers. The contribution from each layer is summed between level $i=1$ and the layer $i+1$. By trapezoid rule we can quantify this relation as:

$$I_v = \sum_{i=1}^{N_{\text{levels}}-1} \left( \frac{B(T(i), \nu) + B(T(i+1), \nu)}{2} \right) e^{-\tau(i)} - e^{-\tau(i+1)} \hspace{1cm} (6)$$

Where: $\nu$=wavenumber (cm$^{-1}$)

$i$=layer number

$\tau$=optical depth

$T$=Temperature (K)

This relationship is summed from the limits $i=1$ to $i=n_{\text{levels}}-1$. The levels are specifically chosen so that contributions are negligible for the region between the surface layer and $i=1$ and for $i=n_{\text{levels}}$ to infinity.

$$B_{(T, \nu, i)} = \frac{(\alpha_i) \nu^3}{e^{\alpha_z \left( \frac{\nu}{T} \right)} - 1} \hspace{1cm} (7)$$
We may calculate the brightness temperature from the corresponding Planck function.

\[
T_b(v, I) = \frac{1.438775v}{\log(1 + 1.1910428 \times 10^{-12} v^3 / I)}
\]  

(8)

The final intrinsic variable to be calculated is the optical depth. If we make the assumption that any aerosols would be well mixed with the atmosphere, the number density of the aerosols should be proportional to the number density of atmospheric CO\textsubscript{2} gas.

By hydrostatic equilibrium we can safely assume that the atmospheric mass above a given level is proportional to the atmospheric pressure at that level, so the atmospheric mass column density at that level will also be proportional to the integrated number density. All of these factors pertain directly to the optical depth. The value of \( \tau \) looking from any given level out to infinity will thus be proportional to the atmospheric pressure. Hence:

\[
\tau_{dust}(i) = \frac{\tau_{dust0} P(i)}{P_{surf}} = \tau_{dust0} e^{-\tau_{p}(i)}
\]  

(9)

The same relation is also true for water ice optical depth,

\[
\tau_{ice}(i) = \tau_{ice0} e^{-\tau_{p}(i)}
\]  

(9a)

where: \( \tau_{dust0} \) = dust optical depth at 1075 cm\textsuperscript{-1},
\( \tau_{\text{dust}} \) = dust optical depth at a given level,

\( \tau_{\text{ice0}} \) = water ice optical depth at 825 cm\(^{-1}\),

\( \tau_{\text{ice}} \) = water ice optical depth at a given level,

\( z_p \) = scale heights above surface.

It should be noted that although the approximation for well mixed aerosols is reliable for dust particles, ice clouds are formed by condensation processes in the atmosphere and are not well mixed.

Complications arise in the case of clouds as a condensation cloud is a discrete “package” of material floating at a given altitude in the atmosphere. Consequently the optical depth of a given cloud depends only on the contributions from the altitude range occupied by the cloud (from the top to the bottom of the cloud); the atmosphere above and below may be much clearer. Appendix H shows MOLA derived altitudes for discrete Martian condensate clouds (Pettengill and Ford 2000).

The dust optical depth measurement is also more complicated than originally assumed as the above relations for dust will only hold true for zenith observations. A corrective term must be used if limb observations or indeed observations from any other incidence angle are used. For our assumption of a plane parallel atmosphere, the line of sight optical depth will be changed by

\[
\tau_{\text{dust}}(i) = \frac{\tau_{\text{dust0}} e^{-z_p(i)}}{\sin \theta} \quad (9b)
\]

where: \( \theta \) = Angle of LOS from zenith.
The factor \( \frac{1}{\sin \theta} \) is considered to be the “airmass” and is a measure of the amount of atmosphere that we are looking through compared to the zenith. A further complication arises from the fact that the optical depth is a function of the wavenumber. Given our standard spectral shape functions (to which all of our observations are fitted) will give an optical depth at each wavenumber relative to the optical depth at a specific reference wavenumber. These points lie at the peak optical depth for each shape, \( 1075\text{cm}^{-1} \) and \( 825\text{cm}^{-1} \) for dust and ice respectively. The standard spectral shapes are provided by the MGS TES team. (The form of these spectral shapes and possible sources of error in their determination are discussed at length in section 3.2, the spectral shapes themselves can be seen in figure 3.1).

Using the total opacity for the atmospheric column, we may perform a fit for individual dust and ice spectral shapes and determine the contributions made by dust and ice to the total optical depth, from this we may derive our values for individual dust and ice optical depths.

Adjusting the algorithm to accommodate for the spectral shapes, our expression will take the form,

\[
\tau_{\text{dust}} (i, v) = \frac{\tau_{\text{dust}} f_{\text{dust}} (v)e^{-z_{\text{dust}}(i)}}{\cos \theta_{\text{emission}}} \tag{10a}
\]

and

\[
\tau_{\text{ice}} (i, v) = \frac{\tau_{\text{ice}} f_{\text{ice}} (v)e^{-z_{\text{ice}}(i)}}{\cos \theta_{\text{emission}}} \tag{10b}
\]
It should be so noted that $\tau_{dust}$ and $\tau_{ice}$ are not a function of wavenumber or height and the dependence on frequency is contained within the spectral shape functions $f_{dust}$ and $f_{ice}$ while $z_p(i)$ is responsible for our height dependence.

The absorption strength of gas is given by an “absorption coefficient”. Each molecule of a given gaseous species of the same type will exhibit the same absorption coefficient. Consequently, the optical depth will be proportional to the product of the absorption coefficient and the number of molecules. More complicated calculation which we may enter into in future projects will require the use of correlated-k coefficients to characterize the distribution of opacity coefficients within the spectral interval of interest. This aids in large calculations by drastically reducing the number of integrations needed. However the main purpose of this project is the development of a new algorithm and this kind of in depth approach is beyond the scope of this study.

In order to solve for aerosol opacity, the radiance, atmospheric and surface temperature are required as inputs to the radiative transfer code. The TES database contains atmospheric temperatures for 38 successive pressure levels in the Martian atmosphere. Atmospheric temperatures were previously obtained by the TES team using an inversion of the 15 micron CO$_2$ radiance band along with corresponding surface temperatures found through the observation of surface radiance at around 1300cm$^{-1}$ (Conrath et.al, 2000). These values archived in the TES database were utilized as inputs to the algorithm.

The assumption is made that aerosols are well mixed throughout the atmosphere. This means that the optical depth at any given altitude will be proportional to the atmospheric pressure at that point, simplifying the solution of equation 1.
3.2 Inputs and Assumptions

![Figure 3.2](image)

Figure 3.2 The spectral dependence of dust optical depth ($f_{\text{dust}}$)(solid line) and water ice optical depth ($f_{\text{ice}}$) (dotted line). The 500-800 cm$^{-1}$ gap is the 15$\mu$m absorption band is excluded from the fitting routine.

Observations of dust storm vertical structure (Jaquin 2000, Smith 1997) support the assumption that dust particulates are well mixed with the CO$_2$ gas of the atmosphere. TES dust opacity data is also known to scale well with surface pressure and emission angle (Smith, 2000b), thus dust opacity may be considered proportional to the amount of gas along the line of sight.

It is assumed that the spectral functions ($f_{\text{dust}}, f_{\text{ice}}$) accurately depict the spectral characteristics of dust and ice particulates and that these features do not change with respect to time or position. The spectral dependencies of both dust and ice opacity are shown in figure 3.2. Several spectral shapes exist for water ice; here we only fit for a
Figure 3.3 Errors due to the use of differing spectral shape for the surface emissivity (solid lines) and water ice (dotted lines). Using the “type 2” (andesitic-basalt) surface emissivity shape (Reproduced from Smith, 2004)

single case. Errors in retrieved optical depth resulting from the use of fixed spectral shapes are quoted to be on the order of 0.01 to 0.02. (Smith, 2005)

Small variations are present in the spectral dependence of both dust and ice for all seasons and at all locations and these effects can be observed in the TES spectra. (Clancy et al., 2003, Wolff and Clancy, 2003) The analysis of Smith 2004 indicates that any effects on retrieved optical depth due to this variation in spectral shape will be relatively small (see figure 3.4). An in depth study of this effect can be seen in Smith 2004, where
250,000 spectra were subjected to changes in the fitted spectral shape and surface emissivity. In these tests the regular water ice spectrum (characteristic of smaller 2µm water ice particles) was replaced with a spectrum depicting particles of a larger radius (∼4µm).

The error induced by this change in a single spectrum was found to be 0.004 for water ice and 0.0003 for the case of dust optical depth. The results of these experiments can be seen in figure 3.3. Consequently, the errors introduced to the final optical depth value by this effect are deemed negligible. The use of differing surface emissivity values also results in slight errors of around 0.005 for dust and 0.003 for ice optical depth. Uncertainties due to the application of the mixing factor to match the brightness temperature gradient will be the same as for the fitting of the spectrum. Figure 4.3 shows the relationship between mixing factor and surface temperature.

The exclusion of scattering from the algorithm is also erroneous. Both dust and ice particulates do cause scattering, however, the single scattering albedo in the TES range for both cases is around 0.1 - 0.4, (Wolff and Clancy, 2003). Due to the extensive amount of extra computations that would be required to fully account for scattering in tens of millions of cases and the limited computational resources at our disposal, this effect was ignored. However, by neglecting scattering it may be considered that we are obtaining the dust opacity resulting purely from absorption, as opposed to the full extinction of the dust aerosol which would include scattering. The effect of scattering on retrieved quantities is minimized through the sole use of nadir observations where scattering effects are very small. The method detailed in Smith 2000b on which this algorithm is based, quotes an underestimation of optical depth by 10-20% when
Figure 3.4 The relationship between absorption and extinction optical depth for dust (solid line) and water ice (dotted line) for the equatorial region, MY24 ($L_s=180^\circ$)

scattering is neglected. Figure 3.4 shows the results of numerical experiments from Smith 2004.

Figure 3.4 show the results of numerical experiments designed to compare values of absorption and extinction optical depth. For both aerosols the ratio of the absorption and extinction optical depths is almost constant. Consequently, the spatial and seasonal trends observed in only absorption optical depth represent the same trends we would obtain if scattering were included. For relatively low optical depths (<0.5) the extinction optical depth is around 1.3 times the absorption optical depth and 1.5 times larger for the case of water. The value of the conversion factor between absorption and extinction optical depth is dependent on the refractive index of the particular aerosol. This value depends on size, shape and composition and furthermore is not a constant function of
wavenumber. For the majority of dust and water ice aerosols, the conversion factors at peak absorption are considered by Wolff and Clancy (2003) to be 1.3 and 1.4 respectively.

Assuming well mixed aerosols, allows atmospheric pressure to be directly related to total dust opacity. This proportionality between opacity and pressure allows the replacement of the optical depth integral with an integration over atmospheric pressure.

The assumption that we are dealing with well mixed aerosols is valid for the case of dust particulates, but may not hold as well for water ice clouds, which form at discrete altitudes where temperatures are favorable for condensation (Pearl et al., 2001). Due to their occupation of discrete altitudes, water ice aerosols are not present at every level of the atmosphere, as assumed in the algorithm. As brightness temperature is determined from the point of greatest optical depth, the derived temperature for a cloud is only an indication of the temperature at the altitude (or range of altitudes) occupied by the cloud. This may cause a bias in the retrieval algorithm, which assumes that the water ice aerosol is well mixed from the surface up to that altitude.

As ice optical depths are derived from a temperature profile which is limited to a narrow altitude range, the calculated temperature profile will be distorted to reflect the temperature at the altitude range occupied by the cloud. This may lead to underestimates of the ice optical depth at high altitudes (higher altitude, lower temperature) and overestimates at low altitudes (lower altitude, higher temperature).

However, elevated ice optical depths can still be accepted as positive detections of water ice clouds and will identify relative increases in ice optical depth within the context of this study.
3.3 Uncertainties

Accounting for all of the effects discussed in the previous section, an individual temperature profile would exhibit an uncertainty of around 2K. However, estimates of atmospheric temperature directly above the surface do exhibit slightly larger uncertainties. The retrieved optical depth has two predominant sources of error, random and systematic errors inherent to the instrument and errors in calibration and errors which may exist in the retrieval algorithm itself due to approximations and assumptions made to aid the retrieval process.

Instrumental noise is stated to be on the order of $3 \times 10^{-8}$ W cm$^{-2}$ sr$^{-1}$ cm$^{-1}$. (Christensen, 1992) however fitting over multiple frequencies reduces noise effects significantly. Smith 2004 suggests that with significant averaging, instrumental errors will produce an uncertainty in the final retrieved optical depth value of ($<0.02$).

Uncertainty in temperature/pressure determination at the surface and in the atmosphere, introduces errors in retrieved optical depth on the order of 0.02. The use of fixed spectral shapes along with the assumption of well mixed aerosols may also constitute a source of error, especially for the case of water ice; these effects are believed to contribute to an error margin of between 0.01 and 0.02 depending on the deviation of the fixed spectral shape to the true case.

The retrieval algorithm itself may contain errors, as detailed in the above discussion. Imprecision in the atmospheric temperature profile and surface pressure contribute errors of around 0.02, as we are dealing with very cold surfaces this value may be up to around 0.05. The use of singular spectral shapes to describe water and ice aerosols respectively introduces errors between 0.01 and 0.02. The assumption that
aerosols are well mixed also constitutes a source of error. However, as stated previously
this will only affect comparisons to other surveys and in the context of this work can be
largely disregarded. Planned future additions to the algorithm should remedy this
shortcoming. There is relatively little compensation for the effects of surface emission
which may have some effect depending on the solar angle. This effect is reduced
however by the use of afternoon nadir pointing spectra. Many of the higher optical depth
values, especially over the poles during the height of winter may be artificially elevated
by around 10% due to the increased albedo of the surface. This again will be addressed in
future versions of the algorithm.

The least squares fitting and minimization procedures employed to fit the spectral
gradient are identical to those used to fit the spectral characteristics. Therefore any error
or uncertainties in the fitting of this feature should be identical to those for the spectral
shapes.

Accounting for all of the above effects leads to an upper value estimate of around
15% for the overall uncertainty of the retrieved optical depth value. As explained above,
the discrete nature of water ice clouds may introduce slightly larger margins of error for
water ice optical depth than dust optical depth. The averaging of data reduces the
uncertainties in the retrieved values, specifically in the case of the 2°x 2° zonally
averaged plots of interannual variability. Systematic errors (such as instrumental
calibration) are inherent to the data and will remain, even in aggressively averaged data.
It should also be noted that the absorption optical depth is presented throughout. Which
may result in underestimates of the quoted optical depths by as much as 30% for dust and
50% for water ice aerosols when compared to the full extinction optical depth.
Chapter 4

TES data and the retrieval algorithm

For the purposes of spatial and temporal mapping, only nadir pointing observations have been utilized. Nadir observations not only represent the bulk of the TES database but also form a systematic map of the planet as the spacecraft progresses in its orbit. During its mapping phase MGS provides two complete planetary maps at 12 hour intervals, one at approximately 1400hrs local time and another at 0200hrs. Retrievals are performed for each 3 x 2pixel TES footprint for a mapping strip 9km in width, giving a resolution between 9 x 10km and 9 x 20km after compensation for image smear (Smith 2000a).

The method presented here is based on the algorithm of Smith 2000a and is designed to retrieve atmospheric dust and ice optical depth. Sufficient modifications have been applied to obtain reliable optical depths over areas and times where surface temperature is below 210K.

4.1 Spectral Shape Fitting

Spectra retrieved over low temperature surfaces exhibit significant positive gradients in their brightness temperature continuum. This gradient is a result of the detection of a range of temperatures in the field of view of the detector, a feature
indicative of frost or ice on the cold underlying surface. Rigorous testing was carried out on numerous spectra utilizing a test algorithm designed to model the continuum slope and determine the appropriate temperature effects to be compensated for. The Planck function was calculated at 148.3 cm\(^{-1}\) using the TES surface temperature retrieval as an input. The algorithm is designed to increase this temperature by 10 K and reevaluate the Planck function, this serves to create an artificial continuum slope in the brightness temperature relation. A statistical weighting or “mixing factor” is then applied to adjust the distribution of points in the modeled Planck function until a suitable fit to the slope of the observed spectrum is attained. Equation 11 shows the expression used to accomplish this.

\[
I = fB_{\nu,(\text{temp}1)} + (1-f)B_{\nu,(\text{temp}2)}
\]  

(11)

Where: \(T_1 > T_2\)

The relationship between mixing factor and brightness temperature can be seen in figure 4.1, this figure shows a variety of Planck distributions with various mixing factors applied along with the corresponding effect on brightness temperature for 15 x 10\(^3\) spectra.
Figure 4.1 The effect of varying mixing ratio from 0.02 to 0.1 on radiance and the corresponding brightness temperature spectrum.

Modeling of multiple Planck functions in this fashion along with their effect on the spectral slope of the brightness temperature continuum showed the average mixing factor for cold surface cases to be from 0.8 to 0.95. This indicates a bias toward lower temperatures. This method was incorporated into the fitting routine of the opacity retrieval algorithm, allowing each spectrum to be analyzed for the effects of the continuum slope and fit for accordingly.
Figure 4.2 Real cold surface spectra subjected to numerical modeling to determine the response of the brightness temperature spectrum to changes in radiance. The radiance on the upper left corresponds to the upper right brightness temperature.

Figure 4.2 shows the results of numerical experiments designed to match the form of real cold brightness temperature spectra. These spectra were subjected to idealized radiance modeling and fitting routines in order to determine a suitable mathematical expression for the form of the continuum slope. The form of the brightness temperature spectrum was determined through manipulation of the radiance spectrum as described previously. Models such as these were utilized to examine the form of the brightness temperature spectrum and its response to changes in radiance. These models provided detailed
Figure 4.3 Mixing factor: Surface Temperature. This figure shows the correlation between surface temperature and the value of the mixing coefficient chosen by the algorithm for $23 \times 10^3$ individual spectra from 120K to 220K.

Observations of the variations in brightness temperature brought about by the application of different mixing coefficients to the radiance spectrum. Uncertainties introduced through the application of the mixing factor to match the brightness temperature gradient vary for individual surface temperatures.

Figure 4.3 shows the relationship between mixing factor and surface temperature. Mixing coefficients for surface temperatures between 160K and 200K between 0.85 and 0.95, arrange of 0.1. Mixing factors varied from 200 to 220K covering a mean range of 0.2 with the majority of points falling into a range from 0.98 to 0.78. Very low temperatures from around 135K to 150K showed the greatest variation at 145K reaching 0.4 and ranging from 0.97 to 0.57.
Figure 4.4 Comparison of TES observed surface temperatures and surface temperatures retrieved through the algorithm. This figure demonstrates the reliability of temperatures generated by the algorithm to those recorded by TES for $15 \times 10^3$ spectra with temperatures ranging from 120K to 180K.

These ranges correspond to the level of uncertainty in the determination of the spectral slope at each temperature. However, temperatures retrieved from the algorithm at a level near the surface match well with TES observations of surface temperature. The close correlation between surface temperatures observed by TES and temperatures retrieved from the algorithm at a level near the surface can be seen in figure 4.4 for $14 \times 10^3$ points over a temperature range of 110K to 180K.
Chapter 5

Results

The algorithm discussed above was used to retrieve dust and ice optical depth values for each spectrum in the TES database with a surface temperature below 210 K. Optical depth for dust is given in terms of a reference frequency of 1075 cm$^{-1}$, while water ice optical depth is given for a reference frequency of 825 cm$^{-1}$. Figure 5.1 shows the zonally averaged seasonal and latitudinal variations in surface temperature over the north and south poles along with corresponding distributions of dust and water ice optical depth. The data shown have been placed in 2° x 2° in latitude and L$_s$ bins. Results are also zonally averaged to account for the exclusion of the longitude axis in this plot. The temperature of the planets surface and atmosphere show a greater dependency on latitude than longitude. This is due to the variation of insolation with latitude and the changing of surface temperatures due to surface frosts/ices. The amount of energy received per unit area will remain the same over constant latitude; consequently it is acceptable to present zonal averages.

Comparison figures 5.1 and 2.1 demonstrates how the findings of this project complement earlier TES surveys, allowing observations to be obtained from the extreme North and South Polar Regions for the autumn-winter periods.
Figure 5.1. Seasonal and Latitudinal variation of Surface Temperature, Ice and Dust aerosols for the period MY24-MY27. The horizontal axis shows continuing and cycling values of Ls. The vertical axis shows latitude. Data are organized into $2^\circ \times 2^\circ$ bins.
Figure 5.2 The evolution of dust optical depth from $\theta=50^\circ$N to $90^\circ$N for MY24-MY27. The differing increments between observations were chosen to span differing temporal ranges of features in each season. 1) MY24-25. 2) MY25-26. 3) MY26-27. Note the relative drop in dust opacity from $\theta=45^\circ$N to $60^\circ$N at $L_s=249^\circ$-292$^\circ$, MY24. $L_s=258^\circ$-305$^\circ$ and MY26 $L_s=234^\circ$-281$^\circ$. 
Figure 5.2 shows the variation in dust optical depth for the Northern hemisphere for equally spaced times throughout the entire seasonal cycle. The inconsistency in the $L_s$ of each observation is due to the slight differences in the length of the seasonal feature (see figure 5.1). The most striking feature in the vicinity of the north pole is the occurrence of the “Annular bands” of low dust activity. Chapter 7 discusses these features in more detail.

**Mars Year 24**

Dust optical depth increased by 0.2 at around $L_s=180^\circ$, this activity initially stretched from $\theta=40^\circ$ to $70^\circ$ latitude until $L_s=210^\circ$ when high dust optical depths became concentrated into two discrete bands from $43^\circ$ to $47^\circ$ and $60^\circ$ to $70^\circ$ latitude. A band of low dust opacity on the order of 0 – 0.1 then remained in a latitudinal zone from $47^\circ$ to $60^\circ$ until a brief resurgence in activity occurred from $L_s=295^\circ$ to $344^\circ$.

At around $L_s=310^\circ$ dust optical depth increased from $\theta=65^\circ$ to $90^\circ$ attaining optical depths in the 0.5 to 1.0 range. Dust opacity rapidly subsided to the 0.1 level at $L_s=14^\circ$ in MY 25.

**Mars Year 25**

Mars year 25 exhibited similar behavior to MY24, however dust optical depth values were higher and began to increase slightly earlier at $L_s=160^\circ$. As in MY24 high dust opacities separated into two latitudinal bands, creating a range of low dust opacity on the order of 0 to 0.1 from $L_s=221^\circ$ to $311^\circ$. Dust activity increased at $L_s=311^\circ$. By $L_s=347^\circ$ dust opacities ranging from 0.5-1.0 dominated the pole. Particularly strong
activity was detected at 60° to 90° latitude from $L_s=347°$ to 10° (MY26). However, dust opacity had returned to the 0-0.1 level by $L_s=37°$ (MY26).

**Mars Year 26**

MY26 demonstrated far higher opacities over the 55° to 90° latitude range than previous years. Dust optical depths on the order of 0.5 to 0.8 were recorded from $L_s=139°$ to 189°, $\theta=68°$ to 90°. Dust opacity increased at $L_s=194°$ exhibiting values of 0.5 -1.0 and stretching from 40° to 90° in latitude with the majority of very high opacities present above $\theta=60°$. As in previous years, a band of low dust optical depth formed between 46 and 59° latitude, ceasing to exist at $L_s=307°$ when dust levels increase. This low opacity band differed slightly from previous years. Although optical depths remained were comparable to MY24 at approximately 0.5, the feature was an average of 6° in latitudinal range and shorter lived with activity falling to the 0-0.1 level by $L_s=294°$. Dust optical depth in the upper latitude band (60°N-90°N) was much higher and longer lived than previous years. With optical depths on the order of 0.5 to 0.1 detected almost consistently from $L_s=193°$ to 29° (MY27). Distinct increases in optical depth took place from $L_s=266°$ to 313°, 60°N to 90°N and $L_s=360°$ to 29° (MY27) at 62°N to 90°N. Dust opacities had fallen to negligible levels by $L_s=35°$ (MY27).
Figure 5.3 The evolution of dust optical depth over MY24-MY27 from θ=50°S to 90°S. The differing increments between observations were chosen to span the differing temporal ranges of features in each season. 1) MY24-25. 2) MY25-26. 3) MY26-27.
5.2 South Polar Observations

Figure 5.3 shows the variation in dust optical depth for the southern pole at equally spaced times through the entire seasonal cycle. Again, the inconsistency in the $L_s$ of each observation is due to the slight differences in the length of the seasonal feature (see figure 5.1).

Mars Year 24-25

The TES dataset begins at $L_s=104^\circ$ for MY24 corresponding to mid-winter in the southern hemisphere. For the period $L_s=104^\circ$ to $231^\circ$ dust opacities remain relatively low with only slight sporadic increases to around 0.6 from $90^\circ$S to $60^\circ$S at $L_s=120^\circ$ to $166^\circ$. A slight increase on the order of 0.2 occurs as surface temperatures begin to climb from $L_s=198^\circ$ to $231^\circ$. Southern dust then remained low for the remainder of MY24.

During the onset of MY25 dust optical depth increased drastically, reaching values in the 0.5 to 1.0 range and extended northwards reaching from $90^\circ$S to $47^\circ$S by $L_s=96^\circ$. Dust opacities were generally high from $90^\circ$S to $55^\circ$S in the southern hemisphere with values between 0.6 and 1.0 from $L_s=138^\circ$ to $216^\circ$. Dust opacities increased at low latitudes from $90^\circ$S to $80^\circ$S near the conclusion of the year from $L_s=344^\circ$ to $360^\circ$.

Mars Year 26-27

Dust optical depth increases from the outset of MY26 reaching values of 0.4 to 1.0 between $\theta=90^\circ$S and $50^\circ$S from $L_s=340^\circ$ to $104^\circ$. Aside from a slight absence of dust activity from $L_s=0^\circ$ to $90^\circ$ opacities remain at around 0.5 extending to around $50^\circ$S latitude, declining from $L_s=173^\circ$ to $190^\circ$. As temperatures fall below 220K at $L_s=274^\circ$,
dust opacities of around 0.2 to 0.3 are recorded up to 80°S latitude. While dust opacities were high, increases were generally weaker and shorter lived than those observed in MY25, \( L_s = 26° \) to 98°. Dust optical depth rapidly increases at the conclusion of the year \( (L_s = 360°) \) reaching values up to 1.0 and expanding northward to a latitude of 65°S. Opacities fall at around \( L_s = 30° \) (MY27) but remain relatively high up to 50°S until the conclusion of the year.

### 5.3 Interannual comparison

**North Pole**

Dust optical depth increased over the north pole for the latter part of each Martian year and early into the subsequent year from \( L_s = 320° \) to 20°, with high optical depths recorded from around 60°N to 90°N latitude over this time. North Polar dust optical depth underwent a further increase each year at around \( L_s = 200° \). High dust optical depths stretched from the upper pole to around 40°N latitude in all three years.

Dust optical depths drop to very low levels from \( \theta = 45° \)N to 60°N at around \( L_s = 220° - 240° \) each year (slightly later in MY25) elevated dust opacities were present above and below this range from 60°N to 90°N and 40°N to 45°N.

Each year a 45°N- 60°N latitudinal zone remains devoid of strong dust signals from around \( L_s = 220° \) to 310°. High optical depths continue to be present above and below this latitude. Despite small interannual changes in overall activity, the distribution and timing of dust optical depth fluctuations is consistently repeated. The North Pole showed lower overall dust activity in MY24 than subsequent years. Both MY25 and 26 exhibited very strong dust responses from 60°N to 90°N latitude between \( L_s = 160° \) to
240°. Although, MY24 exhibited lower levels of dust activity compared to subsequent years.

Dust opacity levels above 60°N latitude remained high from \( L_s = 160° \) to 360° for MY 25 and 26. In the case of MY26 discrete increases in dust optical depth are more prominent in both strength and duration. Two such increases took place between \( L_s = 197° \) to 215° and 266° to 315°. MY 25 and 26 also exhibited higher dust optical depths at lower latitudes than MY24, from \( \theta = 40°N \) to 90°N, \( L_s = 288° \) (MY 25) and \( L_s = 206° \) (MY26). A common feature across all surveyed years was an increase in dust opacity around the onset of northern spring (\( L_s = 360° \)) from 50°N to 90°N latitude. Planet encircling storm 2001a took place from \( L_s = 185° \) to 310° in MY25 beginning around the Hellas region (Smith, 2004). This corresponds to the spring-summer period in the southern hemisphere. This survey is focused on optical depths over cold surfaces which appear in the autumn-winter months; consequently the effects of this storm are not encompassed by the temperature range of this investigation. Dust optical depths for this \( L_s \) range in MY26 were around 10% higher and longer lived than preceding years. As the polar cap is receding at this time, thermal gradients may form between the cold polar ice and comparatively warmer bare ground. This may result in a front that lifts dust across the seasonal polar cap (James, et.al 1999).
Figure 5.4 The evolution of dust optical depth MY24-27 from $\theta=50^\circ$N to $90^\circ$N. The interannual variability of dust optical depth is presented at equivalent times for each Mars Years from MY24-27.
Water ice detections over the North Pole were widespread from 35°N to 80°N latitude, reaching average optical depths of around 0.3-0.6. Water ice optical depth drops to negligible levels in each year between L_s=10° and 70°. Higher water optical depths were also detected at around 37°N each year from L_s=300°-330°. Bands of low opacity from 45°N- 60°N at L_s=220° to 310° are also present for water ice retrievals, however these bands cover slightly greater latitudinal and temporal ranges than the corresponding reductions in dust opacity.

Figure 5.4 shows a comparison of dust optical depth retrievals from 50°N to 90°N at the same value of areocentric longitude from Mars years 24 to 27. This plot demonstrates the repeatability of the Martian seasons in the vicinity of the poles. Each year exhibits the same basic pattern in optical depth fluctuations with some minor differences. As seen in figure 5.1, background optical depth were generally lower in MY26 than MY25, with MY 26 exhibiting large localized and short lived increases in dust optical depth compared to the consistently high dust presence of MY25. The MY26 elevations stretched southwards from the pole to around 50°N and generally lasted for periods of around 20° to 30° of L_s. This effect may account for the relative drop in dust signals seen at L_s=249° (MY26), while dust optical depths were high in MY26. It is likely that the MY26, L_s=249° observation occurred during a temporary lapse in dust activity. The latitudinal band of low opacity from 45°N to 60°N lacks the prominence displayed in figures 5.2 and 5.3, particularly in the case of the 3rd season (MY26-27). However dust optical depths are still shown to be reduced below the 50°N latitude level in most cases. Mars Year 27 does display an unusual feature at L_s=18°. At this point dust activity in the 45°N to 60°N zone is effectively the opposite of that previously described.
as dust activity increases and is localized over this range of latitudes. A similar pattern also forms at this time during MY25, however the MY27 feature is much stronger, forming a complete annular band around the planet compared to the weaker dust response and more broken spatial coverage of the MY25 band.

**South Pole**

Distinct seasonal cycles in dust activity were also observed in the southern hemisphere. In each surveyed year, early summer ($L_s=300°-350°$) dust opacities remained below 0.1, rapidly increasing at around $L_s=350°$ each year. High dust opacities then expand northward, reaching a peak latitude of around $50°S$ by $L_s=90°$ in the subsequent year. Post $L_s=90°$, dust opacity values fell and became more sporadic in space and time, subsiding at around $L_s=180°$. MY 25 showed higher overall dust opacities compared to subsequent years. Dust opacities in the vicinity of the South Pole were significantly higher in MY25 compared to MY26. Data for MY27 was only available until $L_s=82°$ of that year, however this period displays slightly lower dust opacities during early summer compared to prior years. Figures 5.1 and 5.3 show the spatial distribution of dust opacity for regular intervals through the southern winter for each surveyed year.

Prominent repeating water ice detections occur at $25°S$ to $45°S$ latitude during late autumn to early winter of each year ($L_s=70°$ to $140°$). Water ice activity was stronger directly over the South Pole from $90°S$ to $60°S$ latitude, $L_s=350°$ to $171°$ in MY25.

Dust optical depth fluctuations scale well with the seasonal fluctuations of insolation and the corresponding expansion and contraction cycles of the polar cap.
Chapter 6

Comparison of North and South Hemisphere Optical Depths

Dust optical depths were generally higher around the North Pole than the South. Figure 6.1 shows a comparison of daily averaged dust optical depth for the northern and southern hemisphere for surface temperatures < 210K, along with a 10 day adjacent average trend. Care should be taken when comparing figure 6.1 to the results of figure 5.1. The consistently low average optical depth for the southern hemisphere is slightly misleading and is a result of conducting a time average on already averaged data. Optical depths for the southern hemisphere were equal in strength to the northern hemisphere and exceeded northern optical depths in many cases. However, this figure does serve to highlight the sudden surges in northern optical depths compared to the slowly increasing and expanding increases in the southern hemisphere. These spatially and temporally localized optical depth increases are particularly noticeable in the annually recurring dust optical depth surges at around $L_s = 360^\circ$ each year (see figures 5.1 and 6.1).

This localized activity was not observed in the South Polar Region, where elevations in dust optical depth were generally longer lived and occupied a much larger spatial area than in the North. (For example, the high optical depth values from $L_s = 360^\circ$-$450^\circ$ and $720^\circ$-$810^\circ$)
Figure 6.1. Comparison of dust activity in the Northern (Red Line) and Southern (Black Line) hemispheres for the duration of the TES temporal record. Top Panel: Daily average dust optical depth activity. Bottom Panel: 10 Day adjacent average trend.
Dust activity in the vicinity of the North Pole begins at around $L_s=160^\circ$ each year and undergoes a slower increase in strength and spatial distribution relative to the south. The North Pole exhibits background dust optical depths of around 0.5 for the majority of times and latitudes from $L_s=160^\circ$ to $30^\circ$ (subsequent Mars year), with sporadic increases to much higher values particularly at higher latitudes directly over the polar cap (typically $\theta=60^\circ$N-$90^\circ$N). These increases occur at around $L_s=200^\circ$, $290^\circ$ and $360^\circ-10^\circ$ (subsequent Mars year). The elevation in dust optical depth from $L_s=360^\circ-10^\circ$ corresponds to the beginning of the increase in south polar dust optical depth. This may be due to the formation of the double Hadley circulation cell which occurs at the solstices ($L_s=0^\circ$ and $180^\circ$) this may channel material away from the mid-latitudes and toward the poles, as the Hadley circulation system switches from a single to a double cell. This may result in stronger winds at the surface in the vicinity of the poles as well as the lifting of material from mid-latitudes to much higher/lower regions. Dust activity in the north at this time tends to be shorter lived but much more intense (see figures 5.1 and 6.1) compared to the strong, long lived storms of the south pole seen from $L_s=0^\circ$-$180^\circ$ in both MY25 and MY26, which reach their peak strength and latitudinal extent at around $L_s=90^\circ$.

The south polar area exhibits little in the way of detailed activity and appears to have no analogue to the annular patterns in optical depth observed around the North Pole. While dust in aerosols around the North Pole form discrete bands from $L_s=230^\circ$ to $315^\circ$, dust optical depth in the southern hemisphere remains distributed over a wide range of latitudes from $90^\circ$S to $50^\circ$S. This is indicative of the different circulation patterns at work in each hemisphere.
Chapter 7

The Annular Bands

Figure 7.1. Interannual comparison of prominent annular low optical depth bands from Mars Year 24 to 26 from θ=50°N to 90°N.

The most striking difference between dust activity around each polar region is the formation of the “clear annular bands” seen in figures 5.2 and 5.3 from 45°N to 60°N,
L_\text{s}=220^\circ-310^\circ$ each year. Figure 7.1 shows a comparison of these annular bands for each completely surveyed Mars Year.

While these observations are not temporally synchronized, they do demonstrate that the annular banding follows the same morphology in each year. These bands of low dust and water ice optical depth (almost certainly a result of global circulation patterns) are completely absent in the vicinity of the South Pole.

A possible explanation for the formation of these low optical depth bands lies in the formation of a relatively new category of dust storm, a “flushing” dust storm. Frontal dust storms which resemble standard baroclinic fronts are capable of growing into regional scale storms. These storms are known to form along the edges of the North polar cap from L_\text{s}=200^\circ-240^\circ (North Autumn/South Spring.) and L_\text{s}=300^\circ-340^\circ (North winter/South summer) (Cantor et al, 2001; Wang et al, 2003), coincident with the observations of low optical depth. General Circulation Models (GCM’s) have successfully modeled the effect of a flushing storm (Newman et al., 2003). This phenomenon is caused by the synchronization of low level winds in frontal systems and winds associated with the diurnal tide. Dust raised by cap edge winds is fed into the global Hadley cycle and channeled southwards; dust channeled into the Hadley cell alters the thermal dynamics of the system and served to increase circulation strength.

This southward flushing is likely the cause of the drop in optical depth at discrete northern latitudes and is likely the first observation of this effect in the vicinity of the pole.
Chapter 8

Conclusions

• Reliable observations have been obtained of dust and ice aerosol optical depth over the Martian Polar Regions for three complete Martian years. An algorithm now exists that is capable of retrieving reliable optical depth measurements over surfaces with temperatures below 210K. This has been successfully created and tested. These observations contribute to the overall completeness of the TES dataset.

• Mars Year 26 exhibits greater and longer lived increases in dust optical depth during northern winter than other observed years, while MY 25 presented the highest average dust optical depths for southern winter. Repeating patterns in dust and ice optical depth are present in both polar regions. Dust optical depth increased during early spring in the northern hemisphere from Lₘ=320° to 20°, extending down to 60°N latitude. Each year exhibits an annular band of low dust and ice optical depth relative to the surrounding areas, occurring between 45°N and 60°N latitude from mid-autumn (Lₛ=220°) to mid-winter (Lₛ=310°).

• Annual repeating observations of water ice clouds were present at around 35°S latitude during Lₛ=79° to 140° of each year.
The greatest observed dust activity takes place poleward of 40°N latitude from \( L_s = 170° - 30° \) in the north and \( L_s = 340° - 180° \) from \( \theta = 90°S \) to \( 50°S \) in the south. Dust optical depth climbed sharply near the South Pole at the beginning of each year between \( L_s = 340° \) and \( L_s = 90° \).

Consistently repeating annular bands of low optical depth form in the vicinity of the North Pole from \( \theta = 50°N \) to \( 90°N \), \( L_s = 220°-310° \). This behavior is not seen around the South Pole and is likely to be an effect of frontal storms “flushing” aerosols to upper and lower latitudes, effectively clearing an entire zone for an appreciable fraction of the Martian year.

The formation period of the annular bands from \( L_s = 220°-310° \) coincides with MOC observations of frontal dust storms over the polar cap, which take place in two distinct temporal windows, \( L_s = 200°-240° \) (during the northern autumn and southern spring) and \( L_s = 300°-340° \) (northern winter, southern summer) (Cantor et al, 2001, Wang et al, 2003).

These results may represent observational evidence of low level winds and the diurnal tide interacting with the Hadley cell.
Chapter 9

Future Work

This thesis represents only the development and initial testing phase of the algorithm, many more applications exist than those presented here.

This algorithm is capable of retrieving reliable optical depth measurements over surfaces with temperatures below 210K. As discussed previously, around 50% of the spectra contained within the TES database have been neglected due to the cold surface limit. This algorithm will be used to produce reliable optical depth retrievals from this data adding to the completeness of the TES database. This information may be made available through the PDS (Planetary Data System) at some point in the future.

Much of the data presented here (specifically the zonally averaged figure 5.1) was subject to heavy averaging due to constraints on time and computing hardware. Given more lenient time constraints or better resources, these retrievals may be repeated in better spatial and temporal resolution.

With the initial algorithm completed and successfully tested, the method can be modified to further enhance the accuracy of the retrievals, more spectral types may be fit for, including different ice and dust spectral shapes. The use of correlated-K coefficients may better account for statistical weighting of aerosols in the atmosphere and serve to enhance the radiative transfer model of the atmosphere, moving away from the assumption of a well mixed atmospheric aerosol and depicting the atmosphere in more
realistic terms. The algorithm could be further adapted to provide more accurate observations of ice clouds over the poles through the use of different emission angles to better constrain the altitude and discrete nature of the cloud.

Observations may be conducted to observe the synchronization of polar frontal storms with the Hadley circulation cycle.

Further observations of the northern annular bands may prove interesting, specifically a comparison of long term observations of the effect with the results of one of the many advanced GCM’s that are currently in existence.

In view of the recent surge in interest regarding the poles of Mars, specifically with regard to their overall water content and the impending Phoenix Lander (see appendix G) surveys of the polar caps have become more important of late. Advance observations of these areas ahead of such missions may prove vital to their success.
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_Press release November 1996_


Appendix A

Vanilla and data processing procedures.

Vanilla is a program specifically designed to recall data from the PDS Mars Global Surveyor records. Vanilla utilizes a command line interface that reads binary TES TSDR file formats and correlates data between various stored tables of data. This data is then recalled and output in columns of ASCII values. These abilities allow Vanilla to recall records that match the user specified criteria input on the command line.

Preparation to use Vanilla.

Instructions for the use of Vanilla can be found in the file userdoc.txt provided when the program is downloaded. However, it is the opinion of this author that vanilla is an awkward program to use, with wholly inadequate instructions and support provided by the authors, especially for the DOS version. Much of the syntax provided in the instructions is different to that outlined in the text provided.

Vanilla refers to a dataset “.lst” file to identify the data files to be acted on and treats those files as a set of reference tables. Each table consisting of a system of data columns

Vanilla requires a series of catalogue files in order to access the online PDS archive tables. Before using Vanilla, the appropriate .fmt and .var files will need to be downloaded from the desired volumes of the PDS website, these files are located with the
target data or can be found in the **TES TSDR** data archive **SIS**. But it is probably simpler to refer to the files found with the target data. **SIS** archive information is available in the TES-TSDR **Software Interface Specification** archive volume manual which is available from the MGS TES website.

Volumes of catalogue files are sorted by MGS mission orbit number and Earth date on the PDS archive page of the TES website, consequently the user must find the date or orbit number corresponding to their desired Lₘ before accessing the website.

**Recalling data with Vanilla.**

1. Download the Vanilla program and run the appropriate executable for your operating system.

2. The program is executed by calling “**vanilla**” in the command line (“**vanilla.exe** for DOS version.) Simply typing vanilla recalls a list of all data which can be recalled through the program.

3. Information is recalled according to the parameters entered by the user.

TES data is organized into columns in an online database, each column of data has a header which can be searched for using the “**field**” query.

Vanilla requires the user to define the **fields** they wish to search. This is done through the command: `-**fields**` with the query in quotation marks: `-**fields** “**Insert required field here**”` Field queries must be presented as a single string. Although a space is required between each field, the fields query must be in quotes and if multiple fields are entered they must be separated by a space.

A list of fields that can be recalled using Vanilla is available in the **Userdoc.txt** file.
Vanilla also needs selection criteria entered in the form –select “insert required selection here” a space is required between each requested selection criteria.

**Examples**

The following command extracts the opacity value directly below the spacecraft at the nadir point for the Northwest survey area in the Hellas observing project. Note: the .exe is a DOS command and the suffix is not required on UNIX based machines.

TES archives have already been split into the main areas of interest, by differing the value of the array called, one can recall different wavelengths or temperatures at varying depths through the atmosphere. *Nadir_temperature_profile* can be recalled from any number of 38 different arrays at differing planetary radii.

*Nadir_opacity* can be called from 4 different arrays: 1: Dust, 2: Water ice, 3: Hot CO$_2$ and isotope bands and 4: Surface.

To recall all 4 of the above fields for the Northwesterly survey area for a period over Southern spring, the following command must be entered into Vanilla correctly:


The user then defines the pathway to the text file they wish the data to be written. Vanilla returns spectral values as a list of amplitudes, which must then be multiplied by a standard spectral shape in order to convert these values into meaningful information. All values were averaged over each degree of L$_s$ to better express the changes over time and to reduce the overall size of the datasets.

The data often needs to be reformatted for use with previously written programs; These values were then multiplied by the standard spectral shapes using IDL routines to
obtain the various aerosol opacities required. Standard spectral shapes are an array of values provided in the unfortunately named STD folder which is located along with each dataset. After multiplication these datasets can be proved to be extremely long. This often results sometimes in quite long processing times and very large datasets. The following is part of a typical nadir opacity dataset for central Hellas as returned by Vanilla. Note the poor formatting of the returned data.

<table>
<thead>
<tr>
<th>nadir_opacity[1]</th>
<th>latitude</th>
<th>longitude</th>
<th>solar_longitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.3460000639</td>
<td>-40.099999104</td>
<td>279.73999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.3460000639</td>
<td>-40.109999103</td>
<td>279.66999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.3460000639</td>
<td>-40.109999103</td>
<td>279.58999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.3460000639</td>
<td>-40.159999102</td>
<td>279.74999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.243000059</td>
<td>-40.369999098</td>
<td>279.78999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.243000059</td>
<td>-40.369999098</td>
<td>279.70999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.243000059</td>
<td>-40.369999098</td>
<td>279.62999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.1330000538</td>
<td>-40.519999094</td>
<td>279.81999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.1330000538</td>
<td>-40.519999094</td>
<td>279.73999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>1.1330000538</td>
<td>-40.519999094</td>
<td>279.65999375 184.18999588</td>
<td></td>
</tr>
<tr>
<td>0.78300003719</td>
<td>-40.71999909</td>
<td>279.84999374 184.18999588</td>
<td></td>
</tr>
<tr>
<td>0.783000003719</td>
<td>-40.72999909</td>
<td>279.77999375 184.18999588</td>
<td></td>
</tr>
</tbody>
</table>

If searching for surface temperatures for the same area, one would type:

```
Vanilla.exe -fields “spectral_surface_temperature latitude longitude solar_longitude” -select “longitude 295 300 latitude -43 -38 solar_longitude 180 200”
```
Surface temperature data will be returned in the same format as that above with the *nadir_opacity* headed column reading *spectral_surface_temperature*

If the user has specific requirements, vanilla can be told to refer only to the desired arrays. In this example only a portion of the temperature array is required, this query returns only temperatures in central Hellas that were between 235 and 240K.

```
```

Some arrays are of an undefined length, for variable length arrays, the array brackets should be left open, changing our query to:

```
```

The user can also specify the *quality* of the data they wish to recall, the command “0 0” recalls only the best quality data. (i.e. orientation and processing effects that can affect the information are not present in the information recalled)

```
Vanilla.exe .  –fields “spectral_surface_temperature latitude longitude solar_longitude”  –select “longitude 295 300 latitude -43 -38 solar_longitude 180 200 quality: spectral_surface_temperature_rating 0 0”
```

If necessary, results from a specific detector can be requested. The command will be of the form:

```
Vanilla.exe .  –fields “detector calibrated_radiance [ ] latitude longitude solar_longitude”  –select “detector 2 2 target_temperature longitude 295 300 latitude -43 -38 solar_longitude 180 200”
```
Data in this survey were recalled over a user specified spatial and temporal range. Data were then averaged over the 5x5 degree surface area then over 1 and 5 degree timescales in order to best illustrate the change in temperature over time in the given survey areas, using specially composed IDL routines.
Appendix B

MGS/ MOC Specifications

Figures reproduced from NASA/JPL Press kit (Lee et al 1993)

Figure B-1. Mars Global Surveyor Spacecraft in mapping configuration. (Image reproduced from Mars Global Surveyor Arrival Press kit, September, 1997.)
The Mars Global Surveyor (MGS) spacecraft arrived at Mars on September 12, 1997 and began mapping Mars in detail in March 1998, utilizing the Mars Orbital Camera (MOC), Thermal Emission Spectrometer (TES) and Mars Orbital Laser Altimeter (MOLA) devices. These are attached to the nadir deck of the spacecraft, which can be clearly seen in the image above. The components discussed in this appendix are labeled in the MGS image.

The Mars orbital camera consists of two independent cameras: a Narrow angle camera (NAC) and a wide angle camera (WAC) mounted together. Cameras are supported by 2x32 bit processors and a 12MB buffer system for temporary storage purposes. The narrow angle camera is 70cm tall and is an f/10 Ritchey-Critien reflector with a 3.5m focal length, with two 2048 element CCD in the focal plane mounted at 90º to the spacecraft direction of motion. The NAC is capable of 1.4m/pixel resolution.
The wide angle camera consists of two 9.7mm focal length, 140° FOV fish eye lenses on a single focal plane with two 3456 element CCD arrays. Color images are produced by two lenses: one dedicated to the red filter (575 – 625nm) and one to the blue (400 – 500nm). Wide angle images of the surface are obtained with resolutions that can vary by position between 250 m/pixel and 2km/pixel at the limb resolution. Wide angle images are obtained with MOC’s global monitoring mode used for full planet monitoring and provide daily planetwide maps, allowing weather patterns to be spotted and tracked with relative ease.

Figure B-3. Schematic Cutaway Diagram of TES: Thermal Emission Spectrometer.

TES has two nadir pointing telescopes. One 15.24cm diameter Cassegrain feeding into a two port Michelson interferometer spectrometer with a spectral range of 6.25µm to 50µm. The smaller of the two is an off axis 1.5x10^{-2}m parabolic telescope that has two bolometric channels (0.3µm to 3.9µm and 0.3µm to 100µm) and is used to detect the bolometric thermal radiance. Each telescope contains 6 detectors, each with an
8.3 x 8.3 mrad fov. The 6 detectors together form a grid 3 frames wide x 2 frames deep. A pointing mirror capable of rotating 360° allows TES to be adjusted in order obtain views of either horizon in order to effect limb observations.

TES is calibrated utilizing imagery of space obtained with the directional mirror and via an internal reflectance surface. Two redundant neon lamps are also present, capable of generating emission lines at 703.2nm.

The thermal emission spectrometer is a combined infrared spectrometer and radiometer used to measure heat energy radiated from the surface and atmosphere of Mars. TES can be used to determine surface properties, such as mineralogy and thermal properties and can also be used to determine atmospheric properties such as temperature at varying depths, dust opacity and cloud type.

The surface temperature is calculated by performing a simple algorithm on each recalled TES spectrum. The surface kinetic temperature calculated through this method is only useful however when a first order approximation of the temperature is required due to the fact that atmospheric effects and variations in surface temperature are not accounted for. The algorithm scans the entire spectrum and locates the region of highest emissivity where brightness temperature provides the closest approximation to the surface kinetic temperature. The highest emissivity is quite often in the short wavelength parts of the spectrum at around the 8µm level where significant noise exists.

Consequently, TES data gives unreliable readings for low temperatures. Page 17 of the Data processor user guide by P.R.Christensen gives the full details of this algorithm, this has been reproduced on the following page.
Following is an excerpt from section 5 page 17 SURFACE TEMPERATURE DETERMINATION detailing the algorithm used in surface temperature determination. *(Mars Global Surveyor Thermal Emission Spectrometer Data Processing User’s Guide by P.R.Christensen et.al)*

1) Convert the calibrated radiance to brightness temperature at each wavenumber assuming that: a) the emissivity is unity (temp. = TB); and b) the emissivity is 0.97 and dividing the calibrated radiance by this value before determining the brightness temperature (temp. = TB’). Filter the brightness temperatures using a unity-weight filter seven samples wide to reduce noise effects.

2) Find the maximum brightness temperature over the sample ranges from:
   a) 300 to 1350 cm⁻¹, excluding the region from 500 cm⁻¹ to 800 cm⁻¹ where atmospheric CO₂ has strong absorptions. This range was selected to include both the long and short wavelength portions of the spectrum, and to include the wavenumber typically with the highest brightness temperature (~1300 cm⁻¹) as determined by both the Mariner 9 IRIS and the preliminary TES data.
   b) 300 to 500 cm⁻¹ only. Range covers only the long wavelength portion of the spectrum.

3) If TB is ≥ T2 (225 K), set T_surface to TB; If TB’ is ≤ T1 (215 K) set T_surface to TB’.

   Otherwise, provide a smooth transition between these to cases by setting T_surface to weighted average of TB and TB’. Weighting is determined by:

   Weight1 = 1 - ( (T2-TB) / (T2-T1) )
   Weight2 = 1 - ( (TB’-T1) / (T2-T1) )

   If Weight1 or Weight2 < 0, then they are set to 0.

4) Finally: T_surface = ( (TB*Weight1) + (TB’*Weight2) ) / (Weight1 + Weight2)
Appendix C

Mars Polar Topographic Maps

Figure C-1. North Pole Topographic Stereographic Projection (USGS)
Figure C-2. South Pole Topographic Stereographic Projection (USGS)
Appendix D

Figure D-1. Seasonal variation in the extent of the polar caps
Appendix E

Martian Timekeeping: Areocentric Longitude and Mars Year

Figure E-1. Areocentric Longitude

This figure shows the orbital tracks of Both the Earth (Blue circle) and Mars (Red circle). $L_a$ is measured anticlockwise in degrees from $0^\circ$ to $360^\circ$ along the progress of the orbit. The $0^\circ$ degree point is defined by the line of equinoxes.
Areocentric longitude ($L_s$) is the preferred system of time measurement with regard to the planet Mars. $L_s$ is defined as the angle measured counterclockwise in degrees from the vernal equinox (Martian North Spring equinox.). Perihelion occurs at around $L_s=250^\circ$ during the tail end of Southern spring (Northern Autumn)

**Mars Year (MY) System**

The Mars Year system is another system of time measurement in common use with respect to Mars and used extensively in this work. This system uses a numerical designation of Mars Years or MY beginning at MY 1 from the arbitrarily selected date: April 11 1955. Each Mars mission leading up to MGS has occurred in the years summarized in the table below.

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<tr>
<th>Mission</th>
<th>Duration (MY)</th>
<th>Duration (Earth Date)</th>
</tr>
</thead>
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<tr>
<td>Mariner 9</td>
<td>9-10</td>
<td>May 30 1971 - October 27 1972</td>
</tr>
<tr>
<td>Viking</td>
<td>12-15</td>
<td>August 20 1975 - April 11 1980</td>
</tr>
<tr>
<td>Phobos</td>
<td>19-20</td>
<td>July 12 1988 - March 27 1989</td>
</tr>
<tr>
<td>Pathfinder</td>
<td>23</td>
<td>December 4 1996 - September 27 1997</td>
</tr>
<tr>
<td>Mars Global Surveyor</td>
<td>23-29</td>
<td>November 7 1996 - November 2 2006</td>
</tr>
</tbody>
</table>
## Appendix F

### Basic Mars Data

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<tr>
<th>Property</th>
<th>Value</th>
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<td>Identifying Symbol</td>
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</tr>
<tr>
<td>Diameter</td>
<td>6787km</td>
</tr>
<tr>
<td>Mass</td>
<td>$6.4185 \times 10^{23}$ kg (0.107 Earth Masses)</td>
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<tr>
<td>Volume</td>
<td>$162.6 \times 10^{12}$ km$^{-3}$</td>
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<tr>
<td>Mean density</td>
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<td>Center of mass offset</td>
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<tr>
<td>Surface Gravity</td>
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<td>Visual geometric albedo</td>
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<tr>
<td>Orbital Period</td>
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</tr>
<tr>
<td>Minimum solar distance</td>
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</tr>
<tr>
<td>Length of sidereal day</td>
<td>24h 27m 22s</td>
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</tbody>
</table>
Length of mean solar day: 88775.2s

Ellipticity: 0.0059

Eccentricity: 0.0934

Surface temperature range: 148-310K

Mean surface pressure: 6.1mbar

Magnetic field strength: 1/800th Earth magnetic field

Temperature: 148-310K

Mean Pressure: 6.1x10^3bar

Table F-2: Basic Atmospheric Constituents

CO₂: 95.32%  H₂O: 0.03%
N₂: 2.7%  Ne: 2.5ppm
Ar: 1.6%  Kr: 0.3ppm
O₂: 0.13%  Xe: 0.08ppm
CO: 0.07%  O₃: 0.03ppm

Table F-3: Martian Mean Orbital and Rotational Elements

Areocentric Longitude of Perihelion (Lₚₒ): 250.98°

Anomalistic Year (perihelion to perihelion): 686.995days=668.6141sols

Tropical Year (fictitious mean sun): 686.9725days=668.5921sols

Mars Solar Day (Sol): 1.02749125days=24hrs39m35.244s

Orbital Eccentricity: 0.0934°

Orbital Longitude of Perihelion: 336.04°

Mean Solar Distance (semi-major axis): 1.52366AU
Appendix G

Mars Exploration Timeline for the coming decade

Figure G-1 Mars Exploration Timeline
Appendix H

Table H-1: Discrete Martian Polar cloud top heights reproduced from Pettengill and Ford 2000.

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<th>Orbit Number</th>
<th>Lat °N</th>
<th>Lon °E</th>
<th>Intrinsic Period/sec</th>
<th>Height km</th>
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Appendix I

Sample Mars spectra retrieved over Surfaces of temperature < 210K

Figure I-1. Examples of typical daytime brightness temperature spectra as retrieved by TES, exhibiting the characteristic features of CO$_2$ gas, H$_2$O ice and Dust aerosols.
Figure I-2. Spectra obtained over a surface of around 200K showing a strong CO$_2$ absorption band.

Figure I-3. Example of a cold spectrum obtained from the edge of the Spring North polar cap. Note the change in strength of the CO$_2$ emission feature. The atmospheric temperature is approaching that of the surface.
Figure I-4. Extremely cold spectrum retrieved from directly over the South pole. 

Note the inverse temperature contrast as the atmosphere is now significantly warmer than the surface. Also note the pronounced slope present in the continuum.