Properties of water ice clouds over major Martian volcanoes observed by MOC

Jennifer L. Benson
The University of Toledo

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A Dissertation

entitled

Properties of Water Ice Clouds over Major Martian Volcanoes Observed by MOC

By

Jennifer L. Benson

Submitted as partial fulfillment of the requirements for the

Doctor of Philosophy Degree in Physics

_____________________________
Advisor: Dr. Philip B. James

_____________________________
Graduate School

The University of Toledo

November 2006
An Abstract of

Properties of Water Ice Clouds over Major Martian Volcanoes Observed by MOC

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The Mars Global Surveyor Mars Orbiter Camera obtained global maps of the Martian surface with equatorial resolution of 7.5 km/pixel in two wavelength ranges: blue (400-450 nm) and red (575-625 nm). The maps used were acquired between March 9, 1999 ($L_S=107.7^\circ$) and May 31, 2003 ($L_S=194.5^\circ$), corresponding to approximately two and a quarter Martian years. Using the global maps, cloud area was measured daily for water-ice clouds topographically corresponding to Alba Patera, Olympus, Ascraeus, Pavonis, Arsia, and Elysium Mons. Seasonal trends in cloud activity have been established for each volcano. For volcanoes other than Arsia Mons, interannual variations that could be associated with the 2001 planet-encircling dust storm are minimal. At Arsia Mons, where cloud activity was continuous in the first two years, clouds disappeared totally for $\sim 85^\circ$ of $L_S$ ($L_S=188^\circ-275^\circ$) due to the dust storm. The altitudes of several of these clouds have been measured from the locations of the visual cloud tops and cloud center locations were determined.
We use MOC wide-angle blue (425 nm) images supported by a 3-component (surface, cloud layer, dust layer) radiative transfer model to retrieve and map water-ice cloud properties. TES dust measurements are incorporated into the model after adjusting for wavelength and correcting for topography. Surface reflectance is inferred from MOC images of the same locations acquired during cloud free periods with minimum dust loading.

We chose six aphelion-season MOC image strips, two each for Olympus, Ascraeus, and Elysium Mons, that exhibit well-defined clouds and include TES cloud opacity measurements within the MOC field of view. The resulting maps of 425 nm cloud optical depth reveal large opacity gradients at spatial scales of 10-20 km. Maximum cloud opacity lies between 0.7 and 0.8 for Elysium Mons, 0.8 and 0.9 for Ascraeus Mons, and exceeds 1.0 at Olympus Mons. The largest optical depths lie along the volcano flanks below the summit, and the smallest values are seen near the cloud edges. Cloud water content was estimated by utilizing the opacity measurements, and the total H$_2$O volume of the six clouds studied was found to be on the order of $10^{11}$ cm$^3$. 
Acknowledgements

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Chapter 1

Introduction

Clouds in the Martian atmosphere have been observed and studied for many decades. Nathaniel E. Green identified clouds in the Martian atmosphere in 1877 when he observed white spots near the limb of the planet and concluded that they were morning and evening clouds high in the atmosphere (Kieffer et al. 1992). The Tharsis region is known for the “W” clouds (Slipher 1962), which are prominent from northern mid-spring to midsummer. Smith and Smith (1972) studied the diurnal and seasonal behavior of the clouds associated with Olympus Mons using ground based data collected between 1924 and 1971 and found that cloud activity was generally confined to late spring and early summer seasons in the northern hemisphere, corresponding to $L_S = 60^\circ$-$150^\circ$ (seasons on Mars are referred to in terms of $L_S$, the areocentric solar longitude measured in degrees from the Martian northern vernal equinox; see Table 1.1). They also observed the brightening of the Olympus Mons cloud from early to late afternoon. Based on a comparison of their data with seasonal water abundances as measured by Barker et al. (1970), Smith and Smith concluded that Type I discrete white clouds are composed of water ice. Peale (1973) analyzed flyby observations of “W” clouds by Mariners 6 and 7; he concluded that the brightening was produced by water ice particles that condensed when the atmosphere became saturated by uplift due to topography. Curran et al. (1973) gave the first spectroscopic evidence confirming the presence of water ice clouds on
Mars in the Tharsis region using data collected with the infrared interferometer spectrometer on the Mariner 9 spacecraft. The Mariner 9 spectral shape was consistent with a mean radius of the water ice particles of ~ 2.0 \( \mu \text{m} \). Leovy et al. (1973), using Mariner 9 observations, observed that clouds over Olympus Mons and the Tharsis volcanoes show repeatability, strong topographic control, and evidence for convective structure.

<table>
<thead>
<tr>
<th>Seasons</th>
<th>Northern</th>
<th>Southern</th>
<th>Length (Days)</th>
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<tr>
<td>0° - 90°</td>
<td>Spring</td>
<td>Fall</td>
<td>199</td>
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<tr>
<td>90° - 180°</td>
<td>Summer</td>
<td>Winter</td>
<td>183</td>
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<td>Fall</td>
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Several studies used Viking orbiter data to study the properties of Martian clouds. Extensive fogs containing long, wavelike features were observed in the early morning hours (Hunt et al. 1981, Kahn and Gierasch 1982) and were interpreted as katabatic (downslope) flows induced by the rapid cooling of the low thermal inertia volcanoes. Hunt et al. (1980) examined the diurnal behavior of the Tharsis clouds from dawn until shortly after noon and described how the morning fogs gave way to the large afternoon clouds that characterize the “W”; Pickersgill and Hunt (1981) studied the structure and formation of the occasionally spectacular lee waves and plumes associated with the major volcanoes. Kahn (1984) compiled a catalog of all Viking cloud observations, noting that these clouds occur mainly in northern spring and summer, in agreement with earlier work.
More recent studies have led to the knowledge that water ice clouds play a greater role in the Martian atmosphere and general circulation than previously expected. James et al. (1994) reported Hubble Space Telescope (HST) visible and ultraviolet observations of clouds in the Martian atmosphere. HST observed clouds in the Tharsis region during 1993 and 1995, and global observations in the latter year at $L_S = 68^\circ$ identified an equatorial cloud belt (ECB), extending from $-10^\circ$ to $+30^\circ$ latitude (James et al. 1996a). Orographic clouds such as those over Tharsis, Olympus, and Elysium Mons were observed to be a particularly bright component in this season and latitude region. HST observations in 1997 indicated that this belt was again present between $L_S = 60^\circ$ and $140^\circ$ (Wolff et al. 1999). The cloud belt was subsequently found to have also been present during the Viking years (Tamppari et al. 2000) and is clearly evident in both Mars Global Surveyor (MGS) Thermal Emission Spectrometer (TES) and Mars Orbiter Camera (MOC) data (Wang and Ingersoll 2002, Smith 2004). The growth and decline of the cloud belt was documented extensively from TES measurements, in the form of longitudinally averaged maps extending over several Mars years (Smith 2004). Wang and Ingersoll (2002) determined from MOC images that the ECB began to develop near $L_S \sim 44^\circ$, became longitudinally continuous by $L_S 57^\circ \sim 70^\circ$, and dissipated by $L_S 134^\circ \sim 149^\circ$. This cloud belt therefore seems to be a persistent seasonal feature on Mars near aphelion, which corresponds to northern summer. Clancy et al. (1996) hypothesized that the tropical cloud belt resulted from cooling and low altitude saturation of water vapor trapped in the upward branch of the Hadley cell; this conjecture was subsequently confirmed in a rigorous model by Richardson et al. (2002).
Benson et al. (2003) studied the seasonal behavior of topographic water ice clouds in the Tharsis region using MOC daily global map (DGM) images. They established seasonal trends in afternoon cloud activity for the three Tharsis volcanoes (Ascraeus Mons, Pavonis Mons, and Arsia Mons), as well as Olympus Mons and Alba Patera. They found that Olympus, Ascraeus, and Pavonis Mons exhibited cloud activity during a large fraction of the Martian year (between $L_S = 0^\circ$ and $220^\circ$) with a peak in cloud area near $L_S = 100^\circ$. Alba Patera cloud activity was double peaked, with peaks in cloud area at $L_S = 60^\circ$ and $140^\circ$ and a minimum near $L_S = 100^\circ$. Arsia Mons, the southern-most volcano examined, exhibited nearly continuous cloud activity, confirming the observation of Wang and Ingersoll (2002).

As noted in the previous paragraph, Arsia Mons is unique inasmuch as clouds are observed there in all seasons. A study of MGS TES data by Noe Dobrea and Bell (2005) also found evidence that Arsia is cloudy throughout the Martian year. They noted that these clouds are often located specifically over the southwestern flank of Arsia Mons, a relatively flat feature at an elevation intermediate between the volcano summit and the surrounding plains. Wang and Ingersoll (2002), using MOC images, observed that clouds over Arsia Mons exhibit a unique morphology between mid to late northern summer. These “Aster” clouds consist of a central disk surrounded by radial filaments, which are lower in altitude than the disk and are probably related to the local upslope winds associated with the volcano.

Water ice cloud particle sizes have been found to vary. Noe Dobrea and Bell (2005) found that Arsia Mons clouds exhibit an ice particle size of $r_{\text{eff}}$ about $3.0 \pm 0.5 \, \mu m$, independent of $L_S$. Clancy et al. (2003) define two broad size classifications for
cloud ice aerosols based on MGS TES emission phase function observations. Type 1 ice aerosols have particle sizes of $r_{\text{eff}} = 1$-$2 \, \mu m$ and are most prevalent in the southern hemisphere during southern winter, but they are also present in topographic ice clouds in all seasons. Type 2 ice aerosols with $r_{\text{eff}} = 3$-$4 \, \mu m$ are typical of clouds within the equatorial cloud belt between latitudes of $10^\circ$ S to $30^\circ$ N for the $L_s$ range of $30^\circ$-$140^\circ$. It should be pointed out that these conclusions carry a strong local time bias, since the MGS orbit restricts TES measurements to observations near $\sim14:00$ LST.

Glenar et al. (2003), utilizing near-infrared ground-based observations of clouds over Elysium Mons and in the equatorial cloud belt, found that particle size varied with time of day. Over Elysium Mons, they found that large particles ($> 2 \, \mu m$) dominated in the morning (9:50 LST), particle sizes rapidly decreased in the late morning and early afternoon (10:50 & 12:00 LST), and particle sizes increased in the mid-afternoon (13:20, 13:40, & 14:50 LST), with the largest particles ($> 4 \, \mu m$) becoming dominant as the afternoon progressed. In the equatorial cloud belt, they found large particle sizes associated with morning (9:40 LST) cloud opacity but that particle sizes decreased thereafter.

In addition to particle size, other diurnal characteristics of topographic clouds have been observed. Clouds increase in both size and brightness as the afternoon progresses, reaching maximum extent in the mid- to late-afternoon (Smith and Smith 1972, Benson et al. 2003). Cloud top height rises during the afternoon hours (Benson et al. 2003), which is consistent with increased convective uplift due to surface heating. Furthermore, cloud optical depth has been observed to increase and reach maximum opacity during afternoon hours (Glenar et al. 2003).
A number of studies of orographic cloud extinction optical depths have been carried out using telescopic and spacecraft observations at multiple wavelengths. Christensen and Zurek (1984) studied water ice clouds in the Martian atmosphere with the Infrared Thermal Mapper on the Viking I spacecraft. This paper focused mainly on polar clouds, but they likewise found evidence for clouds on the flanks of Olympus Mons at $L_S \sim 29^\circ$ with a visual optical depth near unity. James et al. (1996a) found the optical depth of the Alba Patera cloud to be 0.46 in 1995 ($L_S = 63.7^\circ$) and that of the Olympus Mons cloud to be 0.48 in 1993 ($L_S = 20.2^\circ$); both values determined from HST F410M images (central wavelength of 410 nm). Akabane et al. (2002), using telescopic data with peak wavelengths between 400-440 nm, estimated the optical depths of Olympus Mons clouds in 1995 and 1997 during northern summer ($L_S = 68^\circ$ to $110^\circ$), finding values of 0.4 – 0.8 at Martian local time near 1400 LST. This study also revealed that afternoon clouds become more active as the season advances from the first half of late spring to early summer. At Arsia Mons Pearl et al. (2001) found values of 825 cm$^{-1}$ absorption optical depth as high as ~0.6 during northern fall ($L_S = 196^\circ$ and $L_S = 210^\circ$), derived from periapsis passes over the volcano by MGS/TES. A subsequent study of Arsia Mons clouds by Noe Dobrea and Bell (2005) showed that clouds with 825 cm$^{-1}$ extinction optical depth between 0.3 and 0.7 persist throughout the Martian year. Benson et al. (2003) measured extinction optical depths from MOC wide-angle blue images for several volcanoes over a full annual cycle. They found optical depth values between 0.11-1.0 for Olympus Mons, 0.10-1.4 for Ascræus Mons, 0.16-1.3 for Pavonis Mons, 0.08-0.82 for Arsia Mons, and 0.16-0.75 for Alba Patera. These measurements showed that the largest
optical depth values occurred during the time when cloud area at each volcano was largest.

There are several meteorological factors that may contribute to the formation and seasonal variability of Martian volcano clouds. Cloud formation over the volcanoes is fostered by the same factors that lead to the formation of the equatorial cloud belt, namely the seasonal maximum in north hemisphere water vapor, combined with cold aphelion temperatures and low saturation altitude. Mesoscale model simulations (Michaels et al. 2006) show that the specific cause of afternoon cloud formation over volcanoes is due to a combination of thermally-induced upslope flow and the rising branch of mountain (gravity) waves which drive water vapor up the flanks, adiabatically cooling and condensing it.

Prior to MGS, observations of Mars have been limited in ways such that continuous, daily monitoring of the planet was not possible. The small angular size of Mars during most of the synodic cycle has limited effective ground-based observations to the period near Mars oppositions. The season on Mars at opposition changes in a roughly 15 year cycle, and it is hard, therefore, to separate seasonal effects from interannual variations. It is also difficult to observe diurnal effects since, due to the similarity of the rotation periods of Earth and Mars, a single ground-based telescope cannot monitor a single region continuously. The Mariner 9 mission lasted less than one year and did not include the seasons when the clouds are most active. Although Viking orbiters imaged Mars for two Martian years, they lacked the synoptic coverage and repetition frequency necessary to give a complete picture of Martian atmospheric phenomena such as the “W” clouds. HST observations are restricted to the period when the elongation of Mars from
the Sun exceeds 50º to prevent exposure of instruments to solar illumination and are also limited by scheduling priorities.

Mars Global Surveyor has conducted synoptic observations of Mars for several Martian years. MOC has two wide-angle (WA) cameras, which provide low-resolution (7.5 km/pixel) maps of the entire Martian surface on a daily basis. TES provides observations in the thermal infrared that complement the visual images. Together, TES and MOC provide the first complete data set with which to study the properties and seasonal variability of clouds throughout the entire Martian year. MGS instruments do not provide much diurnal coverage in the equatorial regions, however, because the orbit is fixed between ~ 1:30 and 2 pm at the nadir. Only the early afternoon part of the cloud cycle, in which previous observations have suggested growth in the large cloud features, can be studied, but this does make it possible to determine the seasonal behavior, augmented by limited local time coverage.

In previous analyses, cloud optical depth estimates were generally quoted as mean values or coarsely binned averages over large spatial scales (Curren 1973, Smith et al. 2001, Pearl et al. 2001, Wolff and Clancy 2003, Glenar et al. 2003), as a consequence of either low spatial resolution, limited time sampling, signal to noise limitations or some combination of all of these. A quantitative study of topographically related cloud properties at spatial scales of a few km has thus been lacking.

This thesis expands on the work of Benson et al. (2003) by continuing the cloud area measurements of the Tharsis volcanoes, Olympus Mons and Alba Patera for an additional Martian year (August 2001 - May 2003) and by also including Elysium Mons measurements from March 1999 through May 2003. This additional coverage enables
the study of interannual variability in cloud activity and the determination of the effect of
the 2001 planet encircling dust storm on cloud activity. Additionally, we produce and
analyze high spatial resolution maps of 425 nm optical depth derived from MOC WA
blue images of aphelion clouds near Olympus Mons, Ascreaus Mons, and Elysium Mons.
In Chapter 2, I will describe the MGS MOC and TES instruments. Chapter 3 details the
method used to obtain the cloud area measurements. In Chapter 4, I will discuss the
cloud area results and the seasonal and interannual variability in cloud activity. Chapter
5 presents measurements of cloud top heights and cloud center locations. Chapter 6
describes our cloud modeling approach for determining optical depth measurements. In
Chapter 7, I present and analyze our optical depth maps and discuss cloud water content
measurements. A summary of the main results of this thesis is given in Chapter 8.
Chapter 2

Instruments and Data Sources

The primary mission of Mars Global Surveyor commenced on March 9, 1999 and ended on January 31, 2001, when the Extended Mission began. MGS has a nearly circular, 2-hour, polar mapping orbit with an altitude of 380-420 km. The orbit is oriented relative to the sun so that local time at the equator is nominally 1400 at the sub-spacecraft point (due to the eccentricity of the orbit of Mars and variation in the precession rate of the orbit of MGS, the local time can vary by over an hour from this nominal value).

Mars Orbiter Camera on MGS includes two wide-angle cameras, one in the blue (400-450 nm) bandpass and the other in the red (575-625 nm) bandpass, each with a 140° field of view (Malin et al. 1992). This coverage is optimal for separating clouds that are composed of condensates from those that contain dust; while dust clouds and ice clouds often appear of similar brightness in red, condensate clouds are very bright in blue while dust clouds are dark at the shorter wavelength (see Cantor et al. 2001). The detectors of both cameras are 3456-pixel line scan CCD arrays with maximum resolution of 230 m/pixel at the nadir and 1.5 km/pixel at the limb. Images are acquired via a “push broom” technique in which the motion of the spacecraft in its orbit creates the spacing between adjacent lines in an image. The WA camera images are bounded by the horizons, corresponding to nominal local times of 1217 to 1543 at the equator. A
variable summing algorithm is used to obtain continuous observations with both WA cameras so that in a 24-hour period a complete global map is obtained at a resolution of approximately 7.5 km/pixel. About 12 image strips per day are recorded and spaced in longitude by about 29 degrees. However, only the central ∼ 1/3 of the MOC WA field of view has been calibrated (flatfielded with absolute responsivity knowledge) accurately enough for quantitative retrieval work. Because of the large field of view, light from a larger area is focused into the various pixels depending on the distance from nadir. The 7.5 km/pixel resolution is set by the resolution at the edge of the calibrated field of view. Pixels closer to nadir are increasingly summed to maintain this resolution. The absolute calibration was done using simultaneous MOC and HST observations. For this work we used the MSSS (Malin Space Science Systems) calibration, which differs somewhat from the USGS calibration. The absolute calibration should be better than ± 15%, but there is the uncertainty due to the discrepancy with USGS which must be methodological. The relative calibration is better except in the outer portions of the blue image, which are very uncertain, hence we use only the central 1/3 of the MOC blue image for quantitative reflectance measurements.

Because MOC images have a fixed nadir of ∼ 1400 LT the complete cycle of diurnal variations cannot be observed. However, because the orbital period of MGS is not commensurate with the rotation period of Mars, the location of a feature fixed with respect to the Martian surface oscillates on a MOC image relative to the central nadir position on consecutive passes over the feature. The local time at the particular feature also oscillates because it is a function of the position on the image strip. As a result, a continuous series of cloud area measurements covers local times spanning roughly 3.5
afternoon hours – the part of the diurnal cycle generally associated with fast changes of the cloud area. For example, a feature near the left edge of the image strip is at an earlier afternoon time (~ 1300) than a feature on the right edge (~ 1600). An example of the geometry of the images is shown in Fig. 2-1.

Figure 2-1: Simulated MOC Wide Angle global map swaths on three consecutive days. Variable crosstrack summing is used to obtain approximately equal ground sampling which compensates for the effects of foreshortening and planetary curvature. Note the variation of the location of a feature on the image strip from one day to the next (e.g., Ascraeus Mons, label “B”). (a) Orbit 827. (b) Orbit 839. (c) Orbit 852.

The Thermal Emission Spectrometer is boresighted with MOC WA, so that each pass produces a narrow track (three pixels wide) near the center of the MOC WA field of view. TES has six detectors in a 3 (across) by 2 pixel array with a single detector resolution of ~ 8.3 mrad (3.3 km at nadir). The spectrum is obtained by scanning a Michelson Interferometer, and the scan time is 2 seconds (1 ick; TES ick’s are TES time intervals = 2 seconds/ick; The ICK “Incremental Counter Keeper” is reset to zero at the nightside equator crossing.). TES products cover the spectral range from 200 to 1600 cm\(^{-1}\) (6-50 µm) (Christensen et al 1992). However, standard processing is applied on an ick-by-ick basis (i.e., average of all 6 detectors). Thus, TES-derived quantities can be superimposed on MOC images with cross-track resolutions of ~9 km and along-track
resolution of 10-20 km. The excess smearing in the second axis is partly the result of spacecraft motion. Calibration of the instrument is accomplished with periodic views of an onboard blackbody and cold space. The use of TES measurements in this thesis is limited to dust aerosol values used for cloud optical depth modeling (Chapter 6) and as a comparison of cloud optical depth profile shape (Chapter 7).
Chapter 3

Cloud Area Measurement Method

To construct a global map, only the center portion of each orbital swath’s coverage is used, which we define as an image strip. Every two hours a new image strip is acquired and added to the previous strips. The daily global maps (DGMs) that we have used for this work are composite cylindrical projections colored with a false green channel and were created by Malin Space Science Systems. Figure 3-1 shows an example of a DGM.

Figure 3-1: A Daily Global Map of Mars on Dec. 13, 2000, used for cloud area measurements. The red box indicates the volcanoes (from north to south): Alba Patera, Olympus Mons, Ascraeus Mons, Pavonis Mons, and Arsia Mons. The blue box indicates Elysium Mons. Clouds can be seen over each volcano. The map is a composite cylindrical projection colored with a false green channel.
This work emphasizes clouds associated with the topographic features of Olympus Mons, Ascraeus Mons, Pavonis Mons, Arsia Mons, Elysium Mons, and Alba Patera, as highlighted in Fig. 3-1 (see also Table 3.1). The data for Elysium Mons discussed here were obtained between March 9, 1999 ($L_s = 107.7^\circ$) and May 31, 2003 ($L_s = 194.5^\circ$), corresponding to approximately two and a quarter Martian years. Data for Alba Patera, Olympus, Ascraeus, Pavonis, and Arsia Mons were obtained between August 1, 2001 ($L_s = 205.8^\circ$) and May 31, 2003 ($L_s = 194.5^\circ$) and provide an additional Martian year of coverage beyond that studied by Benson et al. (2003).

<table>
<thead>
<tr>
<th>Topographic Feature</th>
<th>Central Latitude</th>
<th>Central Longitude</th>
<th>Summit altitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alba Patera</td>
<td>40° N</td>
<td>104° W</td>
<td>~ 6.5 km</td>
</tr>
<tr>
<td>Elysium Mons</td>
<td>25° N</td>
<td>213.5° W</td>
<td>~ 14 km</td>
</tr>
<tr>
<td>Olympus Mons</td>
<td>18° N</td>
<td>133° W</td>
<td>~ 21 km</td>
</tr>
<tr>
<td>Ascraeus Mons</td>
<td>11° N</td>
<td>104° W</td>
<td>~ 18 km</td>
</tr>
<tr>
<td>Pavonis Mons</td>
<td>0°</td>
<td>113° W</td>
<td>~ 14 km</td>
</tr>
<tr>
<td>Arsia Mons</td>
<td>10° S</td>
<td>121° W</td>
<td>~ 17 km</td>
</tr>
</tbody>
</table>

Clouds were identified by visual inspection of the DGMs, and then these features were matched with the calibrated blue strips. The clouds, composed of water ice grains, appear more prominent in blue images due to the relatively high single scattering albedo ($\sim 1.0$) of these particles relative to the low surface albedo ($\sim 0.04$) at these wavelengths.

We chose to use the area of the clouds, obtained directly from the images, as the principal index of cloud activity. Cloud area measurements can be a valuable quantity because, when combined with optical depth measurements, they provide a way of
estimating the cloud water content, and hence the fraction of the total atmospheric water contained in the clouds. To measure the areas of the clouds, image-processing software was adapted from Cantor et al. (2001), which they used to measure the area of Martian dust storms. Our program used a subset of each DGM that corresponded to the particular surface features studied and measured the number of pixels in the blue global map image which had a value greater than a threshold digital number (pixel brightness) value. Because the resolution varies with latitude for a cylindrical projection, each line is multiplied by a correction factor. These lines were then added and multiplied by the resolution of the projection at the equator to give the true area of the cloud. Choosing this particular DN threshold value was somewhat subjective; the best value was determined by careful comparison of the cloud boundaries that were determined by the threshold criterion with cloud boundaries obtained by visual inspection of the blue images. Areas were calculated for several clouds in the calibrated blue images after visual inspection of cloud boundaries. Comparison of those test cases with results using the threshold criterion showed agreement to within 15% (i.e., the area measurements from both methods do not differ more than 15%). The center of the cloud is the center of the ellipse that is the best fit to the cloud contour.
Chapter 4

Cloud Area

4.1. Cloud Area and Interannual Variability

Cloud area was measured daily for afternoon clouds associated with the six topographical features listed in Table 3.1. Examples of the images being analyzed are given in Figs. 4-1, 4-2, and 4-3. Figure 4-1 shows two consecutive days of observations of Olympus Mons during the season of maximum cloud extent. Figure 4-2 is a composite showing Arsia Mons, Alba Patera, Ascraeus Mons, and Pavonis Mons, during the same season. Elysium Mons is seen in Figure 4-3.

Figure 4-1: Cloud over Olympus Mons, observed on two consecutive days and shown in simple cylindrical projection using the same intensity scale. In image A the observed feature is near the left edge of the image strip, and in image B the feature is near the right edge of the strip, at a later local time (see Chapter 2). The cloud brightens from early to mid-afternoon. Latitude and longitude coordinates of the upper left and lower right corners of each image are 37º N, 157º W, and 10º N, 115º W respectively. (A) Jan. 20, 2001 (Ls = 105º), M23-01316, cloud area = 6.5 \times 10^5 \text{ km}^2, the top of the volcano is seen through the cloud. (B) Jan 21, 2001 (Ls = 106º), M23-01376, cloud area = 6.3 \times 10^5 \text{ km}^2.
Figure 4-2: (A) Arsia Mons. This cloud consists of a central disk surrounded by radial filaments, which are lower in altitude than the disk. The radial filaments are between 150 to 300 km long and 25 to 50 km wide. These clouds are probably related to the local upslope winds associated with the volcano. This morphological characteristic of Arsia Mons appears between mid to late northern summer and has been dubbed an “aster cloud” by Wang and Ingersoll (2002) due to its flower-like appearance. Latitude and longitude coordinates of the upper left and lower right corners are 1º S, 130º W, and 10º S, 110º W respectively. May 13, 2001 (L_S = 160º), E04-00756, cloud area = 4.3 × 10^5 km^2. The cloud in the upper right corner is that of Pavonis Mons. (B) Alba Patera. Latitude and longitude coordinates of the upper left and lower right corners are 50º N, 128º W, and 30º N, 99º W respectively. Dec. 28, 2000 (L_S = 95º), M22-02248, cloud area = 2.9 × 10^5 km^2. (C) Ascræus Mons. East of Ascræus Mons is a cloud formation that resembles a wave or a ripple. This morphological trait occurs from late spring through mid-summer. Latitude and longitude coordinates of the upper left and lower right corners are 25º N, 129º W, and 2º N, 88.5º W respectively. Feb. 28, 2001 (L_S = 123.5º), E01-02146, cloud area = 5.8 × 10^5 km^2. (D) Pavonis Mons. Latitude and longitude coordinates of the upper left and lower right corners are 11º N, 127º W, and 8º S, 105º W respectively. Feb. 16, 2001 (L_S = 118º), E01-01081, cloud area = 2.6 × 10^5 km^2. The cloud in the upper right corner is that of Ascræus Mons.
Figure 4-3: Cloud over Elysium Mons. Latitude and longitude coordinates of the upper left and lower right corners of each image are 39º N, 237º W, and 16º N, 205º W respectively. The cloud at Elysium extends westward from the volcano summit. Jan. 17, 2001 (L$_S$ = 104.2º), M23-1141, cloud area = $3.7 \times 10^5$ km$^2$.

Figures 4-4 through 4-8 show cloud area vs. L$_S$ for Olympus Mons, Ascraeus Mons, Pavonis Mons, Arsia Mons, and Alba Patera, respectively. Data from March 1999 (L$_S$ = 110.0º) through July 2001 (L$_S$ = 205.2º) (Benson et al. 2003) are also shown in these figures for comparison. For clarity, to distinguish between the three Martian years discussed in this study, we will use the Clancy et al. (2000) designation of “Mars Year 1” beginning on April 11, 1955. The new Mars Year is defined as beginning at L$_S$ = 0º. The three years discussed here are Mars Year 24 (MY 24) beginning July 14, 1998, MY 25 beginning May 31, 2000, and MY 26 beginning April 18, 2002. Smith (2004) also used this notation for the same time period covered in this study.

Cloud area and optical thickness are heavily influenced by the amount of atmospheric dust which changes sporadically from year-to-year. Planet-encircling dust
storms are the most striking and unpredictable of all weather phenomena on Mars. Within a week or two, a small, localized dust storm can intensify and expand until dust fills the atmosphere over nearly the entire planet (Zurek 1982). After active dust raising stops, atmospheric dust loading can remain at significantly elevated levels for months (Conrath 1975, Martin and Richardson 1993). Furthermore, because the optical depth of the dust is unity or greater in both the visible and infrared during a planet-encircling dust storm, the thermal structure and resulting atmospheric temperature and circulation patterns are significantly altered (Conrath 1975, Pollack et al. 1979, Wilson and Richardson 2000). Planet-encircling dust storms were historically thought only to occur in the season around perihelion (which currently occurs just before southern summer solstice) when solar heating is at a maximum (Martin and Zurek 1993). However, more recent dust storm observations have shown that the dust storm “season” is extended – at least $L_S = 180^\circ$ to $L_S = 310^\circ$, and seems to occur preferentially in early southern spring due to South Polar Cap – Hellas Basin interactions. Nevertheless, planet-encircling dust storms do not occur every year, for reasons not yet understood.

Such a storm evolved on Mars in 2001 (MY 25), beginning in mid-June ($L_S = 185^\circ$) in the Southern hemisphere (Smith et al. 2002). As seen from the MGS TES data, atmospheric temperatures near the Tharsis region began to rise around June 25 and the dust storm had expanded into the region by July 2 ($L_S = 188^\circ$). By July 11 ($L_S = 193.5^\circ$) the dust storm completely encircled the planet (Smith et al. 2002). By about $L_S = 310^\circ$, dust optical depth had returned to normal levels (Smith 2004). However, the effect of the dust storm on surface and atmospheric temperatures continued long after the dust had settled out of the atmosphere. Both daytime surface and atmospheric temperatures were
lower for the entire Martian year after the dust storm than they were in the previous year before the dust storm (Smith 2004).

Richardson et al. (2002) demonstrated that cloud ice content decreases strongly as a function of increasing dust optical depth. The resulting temperature increase lowers the relative humidity and inhibits the condensation of water at lower altitudes, suppressing cloud activity (Pearl et al. 2001). Hunt et al. (1980) noted a period of cloud inactivity in the Tharsis region due to two major dust storms observed by the Viking spacecraft. Therefore, in this section we also consider the effect of the 2001 planet-encircling dust storm in the context of cloud activity and cloud area measurements.

Figure 4-4: Olympus Mons cloud area versus areocentric longitude of the sun ($L_s$). The red line represents data from March 1999 – May 2000 (MY 24), the blue line represents data from June 2000 – April 2002 (MY 25), and the green line represents data from April 2002 – May 2003 (MY 26). Cloud area at Olympus Mons is very similar in all three years of this study. Cloud area peaks near $L_s = 100^\circ$ and goes away by $L_s \sim 190^\circ$. Cloud areas are reliable to within 15%.
Olympus Mons: Cloud area at Olympus Mons in MY 26 is very similar to MY 24 and MY 25 (see Fig. 4-4). Clouds start to appear at $L_S = 0^\circ$, cloud area increases to a maximum near $L_S = 100^\circ$ then gradually declines until clouds disappear at about $L_S = 188^\circ$. This is the same trend found by Benson et al. (2003). Clouds disappeared earlier in MY 26 than in MY 24 ($L_S = 200^\circ$), but later than MY 25 ($L_S = 183^\circ$), the year of the planet encircling dust storm. Clouds are slightly larger in MY 26 between $L_S = 10^\circ$ and $L_S = 50^\circ$ and a little smaller between $L_S = 135^\circ$ and $L_S = 155^\circ$ than the previous years. During the time of maximum cloud area clouds are very bright and prominent, covering most of the summit and stretching out to the west-northwest of the volcano (see Fig. 4-1). Overall there is little interannual variability and no apparent influence of the 2001 planet encircling dust storm on subsequent cloud activity at Olympus Mons.

Ascraeus Mons: There are a few noticeable differences in cloud area trends at Ascraeus Mons in MY 26 compared to the previous years (see Fig. 4-5). Clouds are larger from $L_S = 30^\circ$ - $40^\circ$ but smaller from $L_S = 65^\circ$ - $75^\circ$, $L_S = 90^\circ$ - $110^\circ$, and $L_S = 140^\circ$ - $160^\circ$ in MY 26 than the previous years. Clouds went away at $L_S = 194^\circ$ in MY 26 compared to $L_S = 210^\circ$ in MY 24 and $L_S = 185^\circ$ in MY 25. The overall trend in cloud activity at Ascraeus Mons is similar to that found by Benson et al. (2003); specifically, clouds start to appear at $L_S = 0^\circ$, cloud area increases to a maximum near $L_S = 100^\circ$, and then starts to decline. A local minimum in cloud area is seen near $L_S = 140^\circ$, especially in MY 24 and MY 26. Montmessin et al. (2004) investigated the cloud behavior of Ascraeus Mons using a Mars General Circulation Model (MGCM). Their model results of Ascraeus Mons are quite similar to our observations, reproducing the peak in clouds near $L_S = 100^\circ$ and even the minimum near $L_S = 140^\circ$. When the clouds are at their
largest, they covered the entire volcano, extending north and south and also stretching westward (see Fig. 4-2 C).

Figure 4-5: Ascraeus Mons cloud area versus areocentric longitude of the sun (Lₜ). The red line represents data from March 1999 – May 2000 (MY 24), the blue line represents data from June 2000 – April 2002 (MY 25), and the green line represents data from April 2002 – May 2003 (MY 26). Ascraeus Mons cloud area is smaller near Lₜ = 100° in MY 26 than in the previous year. There is a minimum in cloud area in MY 26 between Lₜ = 140° - 160°. Cloud areas are reliable to within 15%.
Figure 4-6: Pavonis Mons cloud area versus areocentric longitude of the sun ($L_S$). The red line represents data from March 1999 – May 2000 (MY 24), the blue line represents data from June 2000 – April 2002 (MY 25), and the green line represents data from April 2002 – May 2003 (MY 26). Cloud activity at Pavonis Mons is similar in all three years of this study. A minimum in cloud area near $L_S = 140^\circ$ is seen in all three years of the study. The 2001 planet encircling dust storm caused clouds to go away sooner in MY 25 ($L_S = 185^\circ$) than the previous year. Cloud areas are reliable to within 15%.

Pavonis Mons: Cloud area at Pavonis Mons in MY 26 is comparable to MY 24 and MY 25 (see Fig. 4-6). Cloud area peaks near $L_S = 100^\circ$ but decreases to a minimum at $L_S = 140^\circ$ and then increases again. Clouds are smaller from $L_S = 115^\circ$ - $130^\circ$ in MY 26. The minimum in cloud area at $L_S = 140^\circ$ is seen in all three years of observations. The minimum at $L_S = 140^\circ$ also occurs to a lesser extent at Ascraeus Mons as noted above, and it coincides with a peak in cloud area at Alba Patera (see below). Clouds disappeared at $L_S = 185^\circ$ in MY 25 (Benson et al. 2003) due to the increased dust and
atmospheric temperature from the planet-encircling dust storm (Smith et al. 2002) and substantial clouds did not form again until $L_S = 0^\circ$ in MY 26. Some very small clouds were seen between $L_S = 300^\circ$ and $L_S = 360^\circ$ in MY 25, however they were smaller than clouds observed the previous year (Benson et al. 2003); this is most likely an effect of the dust storm. As noted above, the increased dust and atmospheric temperature promote the dissipation of water ice clouds. Therefore when clouds do reappear, it is reasonable to expect lower ice content until atmospheric conditions return to normal, consistent with the observations of smaller clouds. Although clouds had not disappeared at the end of our observing period ($L_S = 194.5^\circ$) in MY 26, they were smaller at this time than clouds observed in MY 24. During the time of maximum cloud area ($L_S = 80^\circ - 110^\circ$), clouds cover the western edge of the volcano and extend westward (see Fig. 4-2 D).

**Arsia Mons:** Observations of Arsia Mons reveal more-or-less continuous cloud activity (Fig. 4-7). Of all of the volcanoes observed in this study, cloud activity at Arsia Mons was the most greatly affected by the dust storm, possibly because of its southerly location. Cloud activity was continuous in MY 24 (Benson et al. 2003), but clouds disappeared completely for about $85^\circ$ of $L_S$ ($L_S = 188^\circ - 275^\circ$) in MY 25. When clouds reappeared, they were smaller in size than the previous year. However, by $L_S = 0^\circ$ in MY 26 clouds were similar in size to the year before. From $L_S = 0^\circ$ through $L_S = 194.5^\circ$ in MY 26 cloud sizes were comparable to the previous years. During the peak activity, the cloud surrounds the volcano and extends slightly southward (see Fig. 4-2 A).
Figure 4-7: Arsia Mons cloud area versus areocentric longitude of the sun (L$_{S}$). The red line represents data from March 1999 – May 2000 (MY 24), the blue line represents data from June 2000 – April 2002 (MY 25), and the green line represents data from April 2002 – May 2003 (MY 26). Effects of the 2001 planet encircling dust storm are seen at Arsia Mons. Clouds disappear completely for $\sim 85^\circ$ of L$_{S}$ ($L_{S} = 188^\circ - 275^\circ$) in MY 25. When clouds reappear, they are smaller in size than the previous Martian year. Cloud areas are reliable to within 15%.

Alba Patera: Cloud area trends at Alba Patera are similar in all three years (see Fig. 4-8). Alba Patera cloud activity displays an interesting double peak feature that was not observed in any of the other regions. The double-peaked feature with peaks at L$_{S} = 60^\circ$ and $140^\circ$, first reported by Benson et al. (2003), is seen again in MY 26. The minimum in the seasonal activity of Alba Patera occurs near the time of maximum cloud areas associated with Olympus and Ascraeus Mons. The peaks in the seasonal activity of Alba Patera occur at the times when the cloud areas associated with Olympus and
Ascraeus Mons are rapidly increasing or decreasing. Clouds had not gone away at the end of our observing period in MY 26, however they were comparable in size at this time to those in MY 24 which went away at $L_S = 211^\circ$ (Benson et al. 2003). Clouds disappeared earlier in MY 25 at $L_S = 189^\circ$ due to the planet-encircling dust storm. Clouds at Alba Patera tend to be elongated in the east-west direction (see Fig. 4-2 B); this could be related to the topography of the volcano, which itself extends farther east and west than north and south.

![Alba Patera Cloud Area](image)

**Figure 4-8:** Alba Patera cloud area versus areocentric longitude of the sun ($L_S$). The red line represents data from March 1999 – May 2000 (MY 24), the blue line represents data from June 2000 – April 2002 (MY 25), and the green line represents data from April 2002 – May 2003 (MY 26). Cloud area at Alba Patera is similar in each year of the study. The double-peaked feature is seen in all three Martian years. Cloud areas are reliable to within 15%.
Figure 4-9: Elysium Mons cloud area versus areocentric longitude of the sun (Ls). The red line represents data from March 1999 – May 2000 (MY 24), the blue line represents data from June 2000 – April 2002 (MY 25), and the green line represents data from April 2002 – May 2003 (MY 26). Cloud area at Elysium Mons is very similar in all three years of the study. Cloud area peaks near Ls = 130°, and then goes away quickly by Ls = 180°. Cloud areas are reliable to within 15%.

Elysium Mons: Cloud area trends at Elysium Mons are very similar in all three years of this study. A plot of cloud area vs. Ls for Elysium Mons is presented in Fig. 4-9. Cloud activity at Elysium Mons is comparable to that at Olympus Mons, although there are some differences. At their peak, cloud areas are similar at both volcanoes. Cloud area gradually increases from Ls = 0° and peaks near Ls = 130° at Elysium compared to Ls = 100° at Olympus. After the peak, cloud area sharply declines and goes away in all years at about Ls = 180°. The clouds at Elysium disappear earlier than those of the other volcanoes studied; therefore the timeline for cloud activity at this location in MY 25
conveys no useful information on the effects of the 2001 planet-encircling dust storm. Clouds at Elysium Mons extend westward from the volcano summit (see Fig. 4-3).

At all other sites besides Elysium Mons, cloud activity ends earlier in MY 25 than the other two years observed, but this can be seen most clearly at Pavonis and Arsia Mons. This earlier cessation of cloud activity correlates with the start of the planet-encircling dust storm. This is also consistent with Wang et al.’s (2005) observations of an earlier cessation of frontal storms in the north polar region in MY 25. Arsia Mons showed the greatest interannual variation due to the dust storm. At Arsia, where cloud activity was continuous in MY 24, clouds disappeared totally for ~ 85° of $L_S$ ($L_S = 188° - 275°$) in MY 25. The Arsia clouds reappeared later in MY 25 but were smaller in size than the previous year. This may be related to the reappearance of flushing dust storms in MY 25, as observed by Wang et al. (2005), after the planet-encircling storm. It is of interest to speculate on events that might forecast major dust storms, but we detected no behavior in the cloud data that could be inferred to be a precursor of the 2001 storm.

The general lack of interannual cloud area variability in MY 26 (after the 2001 dust storm) compared to MY 25 (before the dust storm) is not an unexpected result. Smith (2004) analyzed TES data and also found almost no change in the amplitude and distribution of water ice cloud optical depth in the equatorial cloud belt between MY 25 and MY 26 from $L_S = 105° - 125°$. Some possible explanations for this result suggested by this study are: 1) the high elevation of the volcanoes forces a repeatable pattern from year to year due to the local circulation over them; 2) the change in the water condensation level between MY 25 and MY 26 was minimal at the latitude and pressure level of these clouds; 3) the cloud belt (which encompasses the volcanoes studied here)
contains only a small amount of the total atmospheric water content, so observed interannual variability in water vapor may not lead to changes in water ice cloud optical depth. These factors also provide a valid explanation why there is little interannual variability seen in our results except for during and immediately after the 2001 dust storm.

4.2. Seasonal Behavior

The seasonal trends in cloud activity established by Benson et al. (2003) for the Tharsis volcanoes, Olympus Mons, and Alba Patera are corroborated here with an additional year of coverage. Our data for Olympus Mons and Elysium Mons agree qualitatively with those of Smith and Smith (1972). For Olympus they show a peak in cloud activity near L$_S$ = 100° and no activity from about L$_S$ = 200°-355°. Their Elysium observations indicate a peak in cloud activity at L$_S$ ~ 130°, that clouds begin to abruptly dissipate at L$_S$ ~ 140°, and that clouds have disappeared by L$_S$ = 180°. The average dimension of the Olympus cloud determined by Akabane et al. (1987) for the period L$_S$ = 93°-110° in 1982, about 5 × 10$^5$ km$^2$, is consistent with our values for the same time period. The seasonal dependence at Ascraeus Mons, the most northerly of the Tharsis peaks, is similar to that of Olympus Mons; this suggests that latitude is an important factor in the seasonal trends of cloud activity. Montmessin et al. (2004) investigated the cloud behavior of Ascraeus Mons using a Mars General Circulation Model (MGCM). Their model results for Ascraeus Mons are quite similar to our observations, reproducing the peak in clouds near L$_S$ = 100° and the minimum near L$_S$ = 140°. Cloud area at Olympus and Ascraeus Mons begins decreasing at L$_S$ = 120°. This decrease is likely to
be related to the rise in air temperature due to the season (moving away from aphelion) and increased atmospheric dust (Smith et al. 2001).

The cloud areas of Olympus, Ascraeus, and Pavonis Mons peak at around $L_S = 100^\circ$, corresponding to the largest optical depths in the equatorial cloud belt (Wolff et al. 1999, Smith et al. 2001). At this season, the northern polar ice cap is exposed and injecting water vapor into the northern summer atmosphere (Jakosky and Farmer 1982, Smith et al. 2002). The atmosphere is relatively cold at this time because there is less dust and because there is less solar insolation as Mars is near aphelion. These cold temperatures promote relatively low altitudes of cloud formation (Clancy et al. 1996, Wolff et al. 1999). This season ($L_S \sim 100^\circ$) also corresponds to maximum upward advecting velocities in the solstice Hadley circulation that provide the uplift that promotes condensation (Haberle et al. 1993, Richardson et al. 2002). However, as noted many years ago by Peale (1973), effects due to localized solar heating also enhance the clouds greatly near the volcanoes. The heating and cooling produced by relatively low thermal inertia volcanoes sets up a local circulation that is upslope during the daytime and downslope at night (Hunt et al. 1980, Lindal et al. 1979). These contribute to the formation of the large, bright clouds discussed here during the day and to the fogs and “bore waves” at night. Mesoscale model simulations (Michaels et al. 2006) reveal that the primary cause of afternoon cloud formation over volcanoes is thermally-induced slope flow which drives water vapor up the flanks, adiabatically cooling and condensing it.

There is a brief period of cloud inactivity at Arsia Mons ($L_S = 227^\circ-235^\circ$) in MY 24 which corresponds exactly to the largest regional dust storm reported by Cantor et al.
(2001) and by Smith et al. (2001). The TES observations showed that the dust storm expanded into the region around Arsia Mons at $L_S = 228^\circ$ and was accompanied by a roughly 15 K increase in temperature. By $L_S = 237^\circ$ the dust storm had migrated much further south of Arsia, and the atmospheric temperature in the region had decreased. As noted above, the increased dust and atmospheric temperature promote the dissipation of water ice clouds. Pearl et al. (2001) noted a similar suppression of cloud activity over Arsia Mons in the previous Martian year (1997) caused by the Noachis Terra regional dust storm.

Montmessin et al. (2004) investigated the cloud behavior of Alba Patera using a Mars General Circulation Model, yielding results that corroborate our observations. They found that Alba Patera exhibited two peaks in cloud formation near $L_S = 60^\circ$ and $140^\circ$, although the peaks are less marked than those observed. The authors attribute the peak at $L_S = 140^\circ$ to increased humidity caused by the subliming residual north polar cap and the peak at $L_S = 60^\circ$ to a combination of water evaporation from the receding seasonal north polar cap accompanied by baroclinic eddies. As the seasonal cap recedes further, Alba Patera becomes too far from the cap to be influenced by the eddies, explaining why clouds decrease soon after $L_S = 60^\circ$. Tyler and Barnes (2005), using a mesoscale model, identify short lived, strong transient eddy circulations in a “storm track” corridor at $L_S = 135^\circ$ that extends from the northern slopes of Alba Patera to the North polar residual cap (NPRC). The transient eddies form on the shoulder of Alba Patera, grow, and move poleward before decaying at the edge of the NPRC. They suggest that this “storm track” may be important in the transport of water vapor into
lower latitudes at $L_S \sim 135^\circ$, which may provide water vapor to enhance cloud formation over Alba Patera and lead to the second peak observed at $L_S = 140^\circ$.

Richardson et al. (2002) have modeled water ice clouds in the Martian atmosphere. Their figure 9a displays water ice cloud distribution as a function of latitude and season. At the latitude of Alba Patera ($40^\circ$ N), this map shows a peak in cloud ice near $L_S = 135^\circ$, which corresponds to a peak in cloud area observed with MOC. At the latitude of Olympus and Ascreaus Mons ($18^\circ$ N and $11^\circ$ N), their map displays a peak in cloud ice between $L_S = 90^\circ$-$135^\circ$, which also corresponds to the times of maximum cloud area of these two volcanoes. Figure 15 of Richardson et al. (2002) presents the annual cycle of cloud ice as a function of latitude and longitude at $45^\circ$ intervals of $L_S$. The amount of cloud ice at each interval near the location of the volcanoes we studied generally correlates well with our cloud area measurements. For example, at the location of Arsia Mons, their figure shows at least some cloud ice at all times, which agrees with our continuous cloud activity over this volcano (the issue of water vapor sources for Arsia Mons is discussed further on p. 41). Also, the observed changes in cloud area over Olympus, Ascreaus and Pavonis Mons generally correspond to the changes in cloud ice generated from their model.

While the cloud area at Olympus and Ascreaus Mons begins decreasing at $L_S = 120^\circ$, the cloud area at the southern volcano, Arsia Mons ($10^\circ$ S), starts to increase after $L_S = 140^\circ$ (Fig. 4-7). At this time, the north polar cap supply of water vapor has decreased from its peak (between $L_S = 110^\circ$-$120^\circ$) because temperatures at the cap fall to the point where it no longer provides a significant supply of water vapor (Smith 2002). Thus, the pole to equator gradient in water vapor content disappears as a sizable fraction
of the water recondenses onto the northern polar cap, while water amounts decrease globally. Of the remaining water vapor, the geographical maximum (although a smaller maximum than earlier in the season) shifts to the northern tropics (Smith 2002). Additionally, that work reports evidence of a net transport of water vapor from the northern to the southern hemisphere that is most active from $L_S = 150^\circ$-230$^\circ$. This enhancement of water in the tropics relative to the northern mid-latitudes and the increased water transport to the south are most likely related to the difference in cloud development behavior at Olympus and Ascræus Mons versus Arsia Mons. This may also explain the increase in cloud area after $L_S = 140^\circ$ at Pavonis Mons (Fig. 4-6), which lies on the equator.
Chapter 5

Cloud Top Height and Cloud Center Locations

5.1. Cloud Top Height

Cloud top heights have been determined for Olympus Mons, Ascraeus Mons, Arsia Mons, Pavonis Mons, and Alba Patera for several days throughout their cloudy season and are listed in Table 5.1. The images used for cloud height determination were chosen as a representative cross-section of all the DGM images according to season (more-or-less even sampling in Ls) and local time at the topographic feature. The heights were estimated by visual inspection of the cloud at its highest point of contact with the volcano on the daily global maps. This location was compared to a topographical or contour map for each volcano. The elevation of the highest point of contact determined from the topographical or contour map was estimated to be the height of the cloud. Cloud top height error depends on several factors, namely the resolution of the topographical and contour maps, the accuracy of the visual location of the cloud on the daily global maps, and the ability to determine this location on the topographical or contour map. These considerations lead to height error estimates of ± 0.5 km for Olympus Mons and ± 1 km for Ascraeus Mons, Pavonis Mons, Arsia Mons and Alba Patera. Contour maps for Olympus, Ascraeus, Pavonis, and Arsia Mons, created using Mars Orbiter Laser Altimeter (MOLA) data, were obtained from the website http://www.mars3d.com. The topographic map used in determining the Alba Patera cloud heights is the MOLA high-
resolution global map created by the MOLA science team (Smith et al. 2001). For these maps, the zero elevation is defined as the equipotential surface whose average value at the equator is equal to the mean radius as determined by MOLA. Also included in Table 5.1 is a column which qualitatively describes the location of the highest point of the cloud over the volcano (e.g. Summit, upper north slope, etc.) and a column of the approximate local time of the cloud at this highest point.

Cloud tops generally seemed to fall along the volcano flanks and were lower in height than the summits. Clouds that seemed to cover the entire summit are listed in Table 5.1 with the summit height and a plus (+) symbol indicating that the top of the cloud is above the summit, and therefore we could not identify the exact height using this method. Cloud top heights lie between 19 – 21 km for Olympus Mons, 15 – 18 km for Ascraeus Mons, 12 – 14+ km for Pavonis Mons, 16 – 17.4 km for Arsia Mons, and 5.5 – 6.5+ km for Alba Patera.

From this study of cloud heights, we were able to extract limited diurnal information (though restricted in local time coverage). On the other hand, no significant seasonal trend was observed. Cloud tops tended to lie at a higher altitude when the volcano associated with the cloud was at the right edge of the image strip, and hence at a later afternoon time. On consecutive days, the cloud height could vary 1-2 km depending on local time at the topographic feature. Examples of this trend can be seen in Table I for Olympus Mons at $L_S = 123.46$, 123.94 and Arsia Mons at $L_S = 159.36$, 159.89. This is consistent with increased convective uplift due to surface heating. Leovy et al. (1973), using Mariner 9 observations, detected clouds over Olympus Mons at $L_S = 54^\circ$ that were 20-30 km above the surface, consistent with those observed here.
The measured cloud heights are also generally consistent with the model predictions of Richardson et al. (2002). Their model predicts maximum density of ice at ~ 15 km, with the tops of the cloud deck at 20-25 km. The lower elevation of the Alba clouds is consistent with other unique features: Alba has much lower relief, it is located outside of the ECB (and Hadley uplift) zone, and the seasonal variation of its clouds is unique. The cloud elevation data reinforce the evidence that suggests that the physics of the Alba clouds is different from the clouds in the Tharsis region.

### TABLE 5.1: Cloud Top Heights

<table>
<thead>
<tr>
<th>Orographic Feature</th>
<th>Ls</th>
<th>Cloud Top Height (km)</th>
<th>Location</th>
<th>Local Time (± 0.5 hr.)</th>
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</thead>
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<td></td>
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<td>Location</td>
<td>Local Time (± 0.5 hr.)</td>
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</table>

5.2. Cloud Center Locations

Cloud center locations were determined from the daily cloud observations for each of the six volcanoes in this study. The center of the cloud is the center of the ellipse that is the best fit to the cloud contour. Figure 5-1 shows cloud center locations for the Tharsis volcanoes, Olympus Mons, and Alba Patera. Clouds over Olympus, Ascraeus, and Pavonis extend out to the west-northwest. The largest clouds during late spring and early summer extend the farthest in that direction; small clouds, as seen in early spring and early fall, generally are centered over the volcano summit. Clouds at Alba Patera are stretched out in the east-west direction; this could be related to the topography of the volcano, which itself extends farther east and west than north and south. Michaels et al. (2006) have shown with a mesoscale model for Olympus Mons that once cloud particles
have formed in the lee of the volcano (due to thermally-induced slope flow), the particles
are transported to the west by large-scale horizontal (east-to-west) winds, forming cloud
plumes. Thus the relative locations of the cloud centers convey information on the
direction and magnitude of the winds at the cloud altitudes (17-20 km for all but Alba
Patera).

Figure 5-1: The circles denote cloud center locations for (from north to south) Alba
Patera, Olympus, Ascraeus, Pavonis, and Arsia Mons. Clouds over Olympus, Ascraeus,
and Pavonis Mons extend out to the west-northwest, whereas Arsia clouds extend to the
south over the southwestern flank of the volcano. Clouds at Alba Patera extend over the
volcano in an east-west direction. The background image used is a MOLA topographic
map.
Clouds at Arsia Mons are mostly centered over the volcano, not extending westward like clouds at the other Tharsis volcanoes. However, Arsia clouds do extend to the south over the southwestern flank of the volcano, as reported by Noe Dobrea and Bell (2005). Figures 5-2 A and B show cloud center locations just over Arsia Mons separated into two bins of L_S. Figure 5-2A shows clouds from L_S = 30° - 150° and B shows clouds from L_S = 150° - 30°. Clouds in Fig. 5-2B are mostly centered south of the volcano summit, towards the southwestern flank, and fall within a small latitude range (10° - 12° S). On the other hand, clouds in Fig. 5-2A are mostly centered over the summit with only a few clouds over the southwestern flank, and are more widely scattered in latitude (7° - 14° S). An Arsia cloud simulation at L_S = 100° by Michaels et al. (2006) shows a similar cloud position to the cloud center locations found here for the L_S = 30° - 150° bin. It is not obvious why clouds are preferentially associated with the southwestern flank of Arsia Mons. Noe Dobrea and Bell (2005) speculate that there may be enhanced subsurface water sources in this region associated with buried ice deposits or past/present volcanic activity and these sources continually resupply the vapor which condenses as it is carried upslope by the local winds. They also suggest that the surface in this region could have an enhanced porosity which might lead to enhanced vapor diffusion within the regolith and therefore a larger local source of atmospheric water vapor relative to less porous surrounding regions. Although Richardson et al. (2002) showed some cloud ice present at Arsia at all times, their cloud model required the use of high sedimentation rates (large ice particle sizes) which are unrealistic. Therefore, the physical processes which create the clouds in their model without the necessity of a water vapor source from the regolith are brought into question.
Figure 5-2: Cloud center locations for Arsia Mons from (A) $L_S = 30^\circ - 150^\circ$ and (B) $L_S = 150^\circ - 30^\circ$. From $L_S = 150^\circ - 30^\circ$ clouds are mostly centered south of the volcano summit, towards the southwestern flank, and fall within a small latitude range ($10^\circ - 12^\circ$ S). From $L_S = 30^\circ - 150^\circ$ clouds are mostly centered over the summit with only a few clouds over the southwestern flank, and they are more widely scattered in latitude ($7^\circ - 14^\circ$ S). The background image used is a MOLA topographic map.

Figure 5-3 shows cloud center locations for Elysium Mons and Hecates Tholus (small volcano to the north of Elysium). Clouds at Elysium extend westward from the volcano summit. Clouds form to the east of the Hecates Tholus summit but were only seen between $L_S = 180^\circ$ and $L_S = 360^\circ$ when clouds are absent over Elysium. No clouds were observed over Albor Tholus, the small volcano south of Elysium.
Figure 5-3: Cloud center locations for Elysium Mons and Hecates Tholus (small volcano to the north of Elysium). Clouds at Elysium extend westward from the volcano summit. Clouds form to the east of the Hecates Tholus summit but were only seen between $L_S = 180^\circ$ and $L_S = 360^\circ$. The background image used is a MOLA topographic map.
Chapter 6

Cloud Modeling and Optical Depth Retrieval Methodology

6.1. Choice of Cloud Images

We chose six aphelion-season MOC wide-angle blue (425 nm) image strips, two each for Olympus, Ascraeus, and Elysium Mons, which exhibit well-defined clouds and which include TES cloud optical depth measurements within the MOC field of view (see Table 6.1). These images were selected by first surveying the TES measurement database and applying location (west longitude, latitude) and season (Ls) filters in order to find TES tracks that lie within our aphelion season search window. In three of these cases, the tracks traverse regions of pronounced cloud structure near the volcanoes and can be meaningfully compared with the MOC measurements along the same paths. The selected MOC image strips are accompanied by backplanes which specify observing geometry (\( \mu, \mu_0, \Delta \theta \)) at each pixel as required for radiative transfer (RT) modeling. As is customary, \( \mu_0 \) and \( \mu \) are the incident and reflection angle cosines and \( \Delta \theta \) is the site-plane azimuth angle between incident and reflected beams. MOC images are binned to a resolution of approximately 7.5 km/pixel, as discussed in Chapter 2.

Figure 6-1 shows the selected MOC WA blue images displayed in a latitude – west longitude grid with topographical contours extracted from a 1/8° resolution Mars Orbiter Laser Altimeter (MOLA) elevation map superimposed on the images. The white
boxes outline the cloud region for which optical depth calculations were made, and the red lines in three of the images indicate the TES orbital track. Observations of Olympus Mons are seen in Figs. 6-1a and b, Ascraeus Mons in 6-1c and d, and Elysium Mons in 6-1e and f. The clouds generally extend westward from the summit, although the Elysium cloud in fig. 6-1f also extends north and south. The bright cloud cores form along the volcano slopes and cloud plumes extend westward. The Olympus Mons cloud in fig. 6-1a and Ascraeus Mons cloud in fig. 6-1d are good examples of the diffuse cloud plume extending a great distance from the volcano summit.

TABLE 6.1: Observations Specific to the Optical Depth Measurements

<table>
<thead>
<tr>
<th>Topographic Feature</th>
<th>Date</th>
<th>Mars Year a (MY)</th>
<th>Ls</th>
<th>MOC image ID b</th>
<th>TES OCK c</th>
</tr>
</thead>
<tbody>
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<td>Nov. 8, 2002</td>
<td>MY 26</td>
<td>92.37</td>
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<tr>
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<td>MY 26</td>
<td>89.16</td>
<td>e22-00026</td>
<td>17997</td>
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<td>109.39</td>
<td>e23-00759</td>
<td>18550</td>
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<td>Jan. 17, 2001</td>
<td>MY 25</td>
<td>104.22</td>
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<td>MY 26</td>
<td>123.84</td>
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</table>

a Mars Year (MY) indicates the Clancy et al. (2000) designation of “Mars Year 1” beginning on April 11, 1955. The new Mars Year is defined as beginning at L_s = 0°. The MOC and TES data analyzed here come from MY 25 beginning May 31, 2000, and MY 26 beginning April 18, 2002.

b MOC image ID: MOC images are identified by a 3-character mission subphase descriptor followed by a 5-digit numerical identification. The 3-character mission subphase is denoted by a letter and a 2-digit number. The letters stand for: m – Mapping mission, e – Extended mission, and r – Relay mission. The 2-digit number indicates the month number from the start of the mission subphase. The 5-digit numerical identification indicates the numerical order in which the image was commanded during a subphase.

c TES OCK (Orbit Counter Keeper) is the orbit number of the observation. The beginning of an orbit occurs at the descending equator crossing.
Figure 6-1: Selected MOC WA blue images with topographical contours extracted from a 1/8° resolution Mars Orbiter Laser Altimeter (MOLA) elevation map superimposed on the images. The white boxes indicate the cloud region for which optical depth calculations were made. The red line in figures b, e, and f denotes the TES orbital track. (a) Olympus Mons; Nov. 8, 2002; MOC ID e22-411. (b) Olympus Mons; Feb. 25, 2001; MOC ID e01-1830. (c) Ascreaus Mons; Nov. 1, 2002; MOC ID e22-26. (d) Ascreaus Mons; Dec. 16, 2002; MOC ID e23-759. (e) Elysium Mons; Jan. 17, 2001; MOC ID m23-1141. (f) Elysium Mons; Jan. 16, 2003; MOC ID r01-790.
6.2. Description of the Model and Optical Depth Retrieval Method

Optical depth retrievals were performed for each cloud image using the discrete ordinate, multiple scattering radiative transfer (DISORT) code of Stamnes et al. (1988). DISORT solves the basic transfer of monochromatic radiation in a scattering, absorbing, and emitting plane parallel medium with specified bi-directional reflectivity at the lower boundary.

The basic equation of transfer is given by

\[ \mu \frac{d u_\nu(\tau_\nu, \mu, \phi)}{d \tau} = u_\nu(\tau_\nu, \mu, \phi) - S_\nu(\tau_\nu, \mu, \phi) \]  

(1)

where \( u_\nu(\tau_\nu, \mu, \phi) \) is the specific intensity along the direction \( \phi, \mu \) at optical depth \( \tau \), perpendicular to the surface of the medium, \( \phi \) is the azimuthal angle, and \( \mu \) is the cosine of the polar angle. \( S_\nu \) is the source function, given by

\[ S_\nu = \frac{w_\nu(\tau_\nu)}{4\pi} \int_0^{2\pi} \int_{-1}^{1} P_\nu(\tau_\nu, \mu, \phi; \mu', \phi') u_\nu(\tau_\nu, \mu', \phi') + Q_\nu(\tau_\nu, \mu, \phi) \]  

(2)

where \( w_\nu(\tau_\nu) \) is the single scattering albedo, \( P_\nu(\tau_\nu, \mu, \phi; \mu', \phi') \) is the phase function, and for thermal emission in LTE, \( Q_\nu \) is the source term given by

\[ Q_\nu^{(\text{thermal})}(\tau_\nu) = [1 - w_\nu(\tau_\nu)] B_\nu(T(\tau_\nu)) \]  

(3)

where \( B_\nu(T) \) is the Planck function at frequency \( \nu \) and temperature \( T \). The phase function is expanded in a series of \( 2N \) Legendre polynomials, where \( N \) is the number of directions along which the relative radiances are calculated, so that (1) and (2) are replaced by \( 2N \) independent equations. Discrete ordinate approximation of these leads to a series of \( 2N \) coupled differential equations with non-constant coefficients for which analytic solutions do not exist. The medium is assumed to consist of \( L \) adjacent homogeneous layers, with the single-scattering albedo and phase function constant in each layer, but not necessarily
between the layers. Gaussian quadrature, which makes phase function renormalization unnecessary, is used to transform the coupled differential equations into matrix form. Solutions are obtained by computing the eigenvalues and eigenvectors using a single asymmetric matrix. Further details of this method can be found in Stamnes et al. (1988).

Optical depth retrievals were made by equating the measured cloud reflectance at each MOC WAB pixel with I/F (defined below) values computed for a surface-plus-atmosphere model consisting of surface, cloud, and dust aerosol. Clouds were confined to 10 to 25 km above the surface (see Section 5.1), and dust was uniformly mixed through the atmosphere with dust particle density proportional to pressure (Wolff and Clancy 2003). Reflectance calculations were carried out using DISORT and by adjusting cloud optical depth so that the computed value of I/F agrees with the measured reflectance, expressed in the same units. (Here, I/F is defined as the reflected intensity in the observing direction, with solar irradiance equal to pi in the illumination direction; see Hapke (1993) for a summary of standard reflectance definitions.) This method is similar to that used by Benson et al. (2003) for calculating optical depth at discrete locations, but instead includes 64 streams and 64 Legendre coefficients to represent the angular distributions of atmospheric and surface scattering in the manner of Wolff and Clancy (2003) and Clancy et al. (2003). In addition to the known observing geometry at each pixel, the code requires: radiative transfer (RT) properties (single scattering albedo and phase function shape) for the dust and water ice aerosol layers; the optical depth of the dust aerosol; and the surface reflectance at each pixel position. Though small, the Rayleigh scattering component ($\tau = \lambda^{-4}$) is also included. For the purpose of deriving cloud optical depths, we adopt:
1. The TES solar-band (0.3 – 3 µm) normalized single-scattering phase function shape for “Type 2” aphelion ice cloud particles, derived from TES emission phase function (EPF) measurements by Clancy et al. (2003). However, we also consider in Section 7.2 the effects of applying alternate phase function shapes (Clancy et al. "Type 1" particles and Henyey-Greenstein forms) on the retrieved ice optical depths. We note that the Clancy et al. Type 1 and Type 2 cloud phase function shapes are the only such measurements for Mars ice clouds at near-IR wavelengths. These shapes, retrieved from TES EPF measurements, could not be measured in either the forward or back scattering directions (at phase angles greater than 130° and less than 20°, respectively); hence the inferred shape in the forward scattering direction was extrapolated using Mie models, and the shape in backscattering direction was inferred from a linear extrapolation (see Fig. 6-2).

2. A value of unity for the ice aerosol single scattering albedo, due to the fact that water ice is essentially non-absorbing at visible wavelengths. Lower values for this quantity will also be considered in the context of retrieval uncertainties (Section 7.2) since ice grains are known to be seeded by dust particles, effectively lowering their albedo.

3. An isotropic surface reflectance function (i.e., reflectance proportional to $\mu_0$). We also considered a Hapke reflectance function for a surface of isotropically scattering particles (not equivalent to an isotropically reflecting surface; see Hapke 1993). Numerical simulations showed that ice optical depths inferred using these two forms for surface reflectance were not significantly different over the range of MOC observing geometries.
Figure 6-2: The black traces show the shapes inferred from MGS TES emission phase function measurements (Clancy et al. 2003) of two distinct cloud types: Solid line – Type 1 clouds (1-2 μm grain radii) which are widely distributed in season and latitude; Dashed line – Type 2 clouds (3-4 μm grain radii), typical of the northern sub-tropical aphelion cloud belt. Values below 20º and above 130º, especially the strong forward scattering peak, are extrapolated using Mie models.

6.3. Dust Aerosol Corrections

As discussed in Chapter 4, cloudy regions generally contain non-negligible dust opacity, which must therefore be accounted for in the RT modeling. Dust optical depth at the time and location of each cloud measurement was determined by averaging the 1075 cm⁻¹ dust absorption measurements along each TES orbital track near the cloud sites (see Fig. 6-1) and assuming regionally constant dust opacity (dust absorption optical depth divided by surface pressure). This assumption was found to be reasonably accurate, due to the lack of any significant north-south dust opacity gradient along the orbital track.
MOLA-derived elevations at the location of each MOC pixel were then used to create maps of 1075 cm$^{-1}$ dust absorption optical depth, which were then scaled to 425 nm dust extinction optical depth using a spectral model of dust RT properties derived from TES EPF measurements (Wolff and Clancy 2003), with dust phase-function shape taken from Tomasko et al. (1999) [dust single scattering albedo = 0.67 and asymmetry parameter g = 0.84 at 425 nm]. It should be pointed out that none of the measurement samples exhibited the “ghosting” (anticorrelation with $\tau_{\text{ice}}$) which is a previously observed artifact of the TES optical depth retrieval algorithm.

6.4. Surface Reflectance

Maps of “clear” surface albedo were constructed by searching the full TES archive of aerosol opacity measurements and flagging MGS passes over the cloud sites that exhibited both negligible cloud absorption (values of $\tau_{\text{ice}}$ comparable to or smaller than the TES measurement uncertainty of $\sim 0.05$) and minimum dust loading (dust absorption optical depth $< 0.15$). Of this group, we chose a single image for each volcano that showed maximum overlap with the subject cloud images. The MOC images selected for use as surface albedo maps were corrected for residual dust opacity on a pixel-by-pixel basis using tables of precalculated dust-correction factors (Wolff and Clancy 2003) and restricted to $\mu > 0.5$ and $\mu_0 > 0.5$ in order to avoid large optical paths. The resulting “clear” surface maps were converted to normal albedo assuming an isotropic reflectance function and regridded to the latitude-west longitude points of the cloud images.
Chapter 7

Optical Depth Maps and Cloud Water Content

Figures 7-1a through 7-6a show the retrieved 425 nm extinction (absorption plus scattering) optical depth maps. For convenience we also include with each map (Figs. 7-1b – 7-6b) the corresponding portions of the original cloud images (Fig. 6-1), with the same scale and boundaries. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and are superimposed on the maps, showing the altitudes of each volcano in meters. The altitudes shown correspond to the topography of the volcano but not necessarily to the cloud height.

Olympus Mons shows the largest cloud optical depth of the three volcanoes observed here. The maximum optical depth for the Olympus Mons observation at $L_S = 92.37^\circ$ is ~ 1.0 (Fig. 7-1a), and at $L_S = 122.09^\circ$ it is ~ 1.4 (Fig. 7-2a). In both cases the largest optical depth lies along the slopes of the volcano between ~ 7 and 12 km, to the north of the summit. To the west of the summit there is a region where the optical depth is also above the mean optical depth of each cloud; it is ~ 0.75 for the $L_S = 92.37^\circ$ observation and ~ 1.0 at $L_S = 122.09^\circ$. The cloud optical depth at the volcano summit is not as large; for the observation at $L_S = 92.37^\circ$ it is between ~ 0.40 – 0.50, and at $L_S = 122.09^\circ$ it is ~ 0.50 – 0.60. The edges of the cloud have the lowest optical depth.
Figure 7-1a: Extinction optical depth map of cloud over Olympus Mons at \( L_S = 92.37^\circ \) measured from MOC WA blue image e22-411. The largest optical depth (~ 1.0) lies along the slope to the north of the summit. The cloud edges have the smallest optical depth. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and are superimposed on the map, showing the altitude of the volcano in meters.
Figure 7-1b: MOC WA blue image (e22-411) of Olympus Mons cloud region used to measure optical depths in fig. 7-1a. This is the region within the white box in fig. 6-1a. Bright white cloud is observed to the north of the volcano summit which produces the largest optical depth seen in fig. 7-1a; the rest of the cloud is more diffuse. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and superimposed on the image.
Figure 7-2a: Extinction optical depth map of cloud over Olympus Mons at $L_S = 122.09^\circ$ measured from MOC WA blue image e01-1830. The largest optical depth (~ 1.4) lies along the slope to the north of the summit. The cloud edges have the smallest optical depth. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and are superimposed on the map, showing the altitude of the volcano in meters. The gray bar between 137° and 139° west longitude denotes the TES orbital track.
Figure 7-2b: MOC WA blue image (e01-1830) of Olympus Mons cloud region used to measure optical depths in fig. 7-2a. This is the region within the white box in fig. 6-1b. A bright ‘U’ shaped area of cloud is visible north of the volcano summit, which corresponds to the largest optical depths in fig. 7-2a. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and superimposed on the image. The white line between 137° and 139° west longitude denotes the TES orbital track.

The maximum cloud optical depth for both Ascraeus Mons observations is about 0.80 (Figs. 7-3a and 7-4a). These clouds are formed to the west of the summit and at the summit there is little to no cloud visible. The optical depth near the summit at L_S = 89.16° is very low (~ 0.20). In both cases the largest optical depth (~ 0.70 – 0.80) lies in an arc stretching from north of the summit toward the west along the slopes between ~ 7 and 13 km. The cloud optical depth gradually decreases from this region towards the cloud edges.
Figure 7-3a: Extinction optical depth map of cloud over Ascraeus Mons at $L_S = 89.16^\circ$ measured from MOC WA blue image e22-26. The largest optical depths ($\sim 0.70 – 0.89$) lie in an arc from north of the summit toward the west along the slope of the volcano. The cloud edges have the smallest optical depth. Topographical contours are extracted from the $1/8^\circ$ resolution MOLA elevation data and are superimposed on the map, showing the altitude of the volcano in meters.
Figure 7-3b: MOC WA blue image (e22-26) of Ascræus Mons cloud region used to measure optical depths in fig. 7-3a. This is the region within the white box in fig. 6-1c. A bright white arc of cloud midway down the northwestern slope of the volcano leads to the largest optical depth measured (fig. 7-3a). Topographical contours are extracted from the 1/8° resolution MOLA elevation data and superimposed on the image.
Figure 7-4a: Extinction optical depth map of cloud over Ascraeus Mons at $L_S = 109.39^\circ$ measured from MOC WA blue image e23-759. The largest optical depths ($\sim 0.70 - 0.81$) lie in a narrow arc from north of the summit toward the west along the slope of the volcano. The cloud edges have the smallest optical depth. Topographical contours are extracted from the $1/8^\circ$ resolution MOLA elevation data and are superimposed on the map, showing the altitude of the volcano in meters.
Figure 7-4b: MOC WA blue image (e23-759) of Ascraeus Mons cloud region used to measure optical depths in fig. 7-4a. This is the region within the white box in fig. 6-1d. A bright white arc of cloud midway down the volcano slope from the north toward the west leads to the largest optical depth measured (fig. 7-4a); the rest of the cloud is very diffuse. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and superimposed on the image.

The cloud structures of the two Elysium Mons observations differ more than those of the Olympus or Ascraeus Mons clouds, however the maximum cloud optical depth for both Elysium observations is ~ 0.70. The Elysium cloud observed at $L_S = 104.22^\circ$ (Fig. 7-5a) has two separate regions where optical depth is largest. One region is an arc that extends clockwise from north of the summit around to the east along the slope between ~ 3 and 10 km. The other region lies west of the summit at ~ 1 km. Both regions have optical depths between ~ 0.63 – 0.75. The cloud area connecting the two bright regions has an optical depth of ~ 0.50. At the volcano summit, the optical depth is ~ 0.45. The
region of largest cloud optical depth in the Elysium Mons observation at $L_S = 123.84^\circ$ (Fig. 7-6a) is in an arc extending counterclockwise from north of the summit around to the east along the slope between ~ 2 and 11 km. The optical depth in this region is between ~ 0.60 – 0.70. The optical depth at the summit is ~ 0.55. The edges of both clouds have the smallest optical depth.

![Extinction optical depth map of cloud over Elysium Mons at $L_S = 104.22^\circ$](image)

Figure 7-5a: Extinction optical depth map of cloud over Elysium Mons at $L_S = 104.22^\circ$ measured from MOC WA blue image m23-1141. The largest optical depths (~ 0.63 – 0.75) lie in two separate regions. One is an arc that extends clockwise from north of the summit around to the east along the slope. The other region lies ~ 5° west of the summit. The cloud edges have the smallest optical depth. Topographical contours are extracted from the $1/8^\circ$ resolution MOLA elevation data and are superimposed on the map, showing the altitude of the volcano in meters. The gray bar between $212^\circ$ and $214^\circ$ west longitude denotes the TES orbital track.
Figure 7-5b:  MOC WA blue image (m23-1141) of Elysium Mons cloud region used to measure optical depths in fig. 7-5a.  This is the region within the white box in fig. 6-1e.  The bright cloud west of the summit and the arc of cloud along the volcano slope from the north toward the east correspond to the largest optical depth measured (fig. 7-5a).  Topographical contours are extracted from the 1/8° resolution MOLA elevation data and superimposed on the image.  The white line between 212° and 214° west longitude denotes the TES orbital track.
Figure 7-6a: Extinction optical depth map of cloud over Elysium Mons at $L_S = 123.84^\circ$ measured from MOC WA blue image r01-790. The largest optical depths ($\sim 0.60 – 0.70$) lie in an arc that extends counterclockwise from north of the summit around to the east along the slope. The cloud edges have the smallest optical depth. Topographical contours are extracted from the 1/8° resolution MOLA elevation data and are superimposed on the map, showing the altitude of the volcano in meters. The gray bar between 216° and 220° west longitude denotes the TES orbital track.
Figure 7-6b: MOC WA blue image (r01-790) of Elysium Mons cloud region used to measure optical depths in fig. 7-6a. This is the region within the white box in fig. 6-1f. The bright white cloud on the southeast slope produces the largest optical depth measured (fig. 7-6a). Topographical contours are extracted from the 1/8° resolution MOLA elevation data and superimposed on the image. The white line between 216° and 220° west longitude denotes the TES orbital track.

7.1. Optical Depth Measurement Uncertainties

Uncertainties in retrieved cloud optical depth are dominated by systematic errors rather than by point-to-point fluctuations, as evidenced by the close qualitative agreement in the shapes of the MOC and TES optical depth profiles shown in Figure 7-7. In addition to the uncertainty in the absolute calibration of the MOC camera, errors in our retrievals arise from (in order of decreasing importance): (1) the uncertainty in the ice phase function shape; (2) the range of possible values for single scattering albedo (SSA)
Figure 7-7: Optical depth profiles following the TES tracks shown in Figures 7-2, 7-5, and 7-6. Red triangles: TES absorption optical depth values as listed in the Planetary Data System; Blue circles: MOC Extinction optical depth profiles following the TES tracks, and computed using the Clancy et al. (2003), Type 2 cloud phase function shapes.
of dust-seeded ice grains; (3) surface albedo, and (4) errors in TES-measured dust optical depth as well as the IR-absorption to vis-band extinction ratio. To determine the amount of uncertainty present in our results, we have chosen to carry out a form of model perturbation analysis on those four parameters by changing each one separately and noting how these changes affect the model optical depths, while holding all other parameters constant.

**Phase Function Shape:** The single scattering phase function shape for cloud aerosol particles dominates the optical depth uncertainties since it strongly controls the relation between cloud reflectance and optical depth and since the forward scattering properties for orographic cloud particles are still basically unknown. Figure 7-8 shows the relation between I/F and cloud optical depth for a variety of phase function shapes, including several Henyey-Greenstein (HG) forms. The usefulness of these functions is that they are readily converted into Legendre polynomials as required by the scattering code (DISORT) and provide a way of testing the sensitivity of model reflectance to forward scattering amplitude. Figure 7-8 compares model reflectance curves (tau_ice vs. I/F) calculated using Type 1 and Type 2 ice phase functions as defined by Clancy et al. (2003) with those using HG phase functions with g = 0.64, 0.60, and 0.48. The HG phase function with g = 0.60 gives similar results to the Type 2 phase function, and likewise the HG phase function with g = 0.48 gives similar results to the Type 1 phase function. The HG phase function with g = 0.64 produces larger values of ice optical depth than Type 2. For low ice optical depth, the variation due to phase function is small, but as the ice optical depth increases, so does the uncertainty. Phase function uncertainty relations are defined by the variation in tau_ice resulting from the g = 0.64 – 0.48 HG phase function
measurements. For I/F values less than 0.07, the cloud optical depth varies by ± 0.1 or less. For values of I/F between 0.07 and 0.12, ice opacity variation is between ± 0.1 and ± 0.2. For I/F values larger than 0.12, the variation is between ± 0.2 and ± 0.3.

Figure 7-8: Relation between cloud optical depth and I/F for a variety of phase function shapes. Three Henyey-Greenstein (HG) phase functions with g = 0.64, 0.60, and 0.48 are represented by squares, crosses and diamonds, respectively. The Clancy et al. (2003) Type 1 and Type 2 phase functions are represented with red plus symbols and blue circles, respectively. We used the Type 2 phase function in our optical depth modeling; the HG phase function with g = 0.60 gives similar results to the Type 2 phase function.

Single Scattering Albedo: Figure 7-9 shows the relation between I/F and cloud optical depth for various ice single scattering albedos. Ice optical depth increases as the SSA decreases. For small I/F values, the ice optical depth is not greatly affected by choice of SSA. For values of I/F less than 0.10, the cloud optical depth increases by 0.1
or less; hence the variation of small ice opacities with different SSA is little. For large values of I/F, the cloud optical depth increases by as much as 0.4.

Figure 7-9: Relation between I/F and cloud optical depth for various ice single scattering albedos (SSA). Ice SSA of 1.0, 0.98, and 0.96 are represented by plus symbols, asterisks, and diamonds, respectively. Ice optical depth increases as the ice SSA decreases.

**Surface reflectance:** This quantity also contributes to the uncertainty in ice optical depth. In order to establish the effect on ice optical depth due to this parameter, we multiplied our baseline surface albedo (from the surface albedo maps; Section 6.2) by 0.8 and 1.2. Figure 7-10 shows the results of this exercise for one of our cloud maps that is representative of all (e23-759). In general, the larger the ice opacity, the less the change in surface albedo affects the measured ice optical depth, and vice versa. In other words, the surface reflectance is not as important when clouds are more optically thick because
scattering from the surface is less important. When clouds are thin, the scattering from the surface is of greater consequence, and hence contributes more heavily to the uncertainty of the output ice optical depth. For the largest ice opacity retrievals for each cloud image, the variation due to surface albedo is between $\pm 0.035$ and $\pm 0.05$; for the smallest ice opacity retrievals, the variation is between $\pm 0.08$ and $\pm 0.1$.

![Image of a graph showing the effect on ice optical depth due to a change in surface albedo.](image)

**Figure 7-10:** The effect on ice optical depth due to a change in surface albedo is shown here by multiplying our baseline surface albedo by 0.8 and 1.2. The straight diagonal line represents our baseline ice optical depth results. In general, the larger the ice opacity, the less the change in surface albedo affects the measured ice optical depth, and vice versa. The scatter in ice optical depth due to a change in surface albedo is larger at lower ice opacities, and vice versa.
**Dust Optical Depth**: The amount of dust present at the cloud location also affects the ice optical depth measurements. We created maximum and minimum dust optical depth maps in the manner discussed in Section 6.1 by adding and subtracting a small dust opacity “bias” given by the standard deviation of the measurements from each TES pass. It was found that the variation in dust opacity had a minimal effect on the retrieved ice optical depth (± 0.03) but seemed to change both large and small values of ice optical depth uniformly. Figure 7-11 illustrates these results.

From our perturbation analysis on the four boundary conditions above, we added the model deviations along with the MOC calibration uncertainty in quadrature, which assumes that the uncertainties are independent. The RMS uncertainty for low values of ice optical depth (~ 0.3) is ± 0.18, for medium ice optical depth values (~ 0.7) is ± 0.23, and for high values of ice optical depth (~ 1.0) is ± 0.25.
Figure 7-11: The effect on ice optical depth due to a change in dust optical depth is shown here. The red diagonal line represents our baseline ice optical depth results. The top panel represents our minimum dust optical depth case, and the bottom panel represents our maximum dust case. The variation in dust opacity had a minimal effect on the retrieved ice optical depth.
7.2. Cloud Water Content

For each of the cloud optical depth maps it is possible to estimate the cloud water content. First, H$_2$O column abundances were calculated for each pixel of the six optical depth maps. Next the individual H$_2$O column abundances were summed, and finally the summed total was multiplied by the pixel area $[(7.5 \text{ km})^2]$ to give the total cloud water content. The optical thickness of a cloud, for the thin cloud case, can be approximated by $	au = N C_{\text{ext}}$, where $N$ is the column number density and $C_{\text{ext}}$ is the extinction cross section of the ice particle. The mass of H$_2$O ice in a vertical column per unit area is $m = (4/3)\pi r^3 N \rho = (4/3)\pi r^3 (\tau/C_{\text{ext}}) \rho$, where $r$ is the mean ice particle radius, and $\rho$ is the mass density of ice. We assumed a mean ice particle radius of $r = 3.0 \mu m$ and also considered limiting values of $r = 2.0$ and $4.0 \mu m$ (Clancy et al. 2003), and the density of ice $\rho = 0.917 \text{ gcm}^{-3}$. The values used for the extinction cross section, at a wavelength of 0.4 $\mu m$, is $20.279 \mu m^2$ ($2.0 \mu m$ particle), $44.558 \mu m^2$ ($3.0 \mu m$ particle), and $78.303 \mu m^2$ ($4.0 \mu m$ particle) (Michael J. Wolff, private communication). The H$_2$O column abundance ($m$) is measured in precipitable microns ($\text{pr} \mu m$), where $1 \text{ pr} \mu m = 1 \times 10^{-4} \text{ gcm}^{-2}$. Results of the cloud water content calculations are summarized in Table 7.1. The total H$_2$O volume of the clouds is on the order of $10^{11} \text{ cm}^3$. 
### TABLE 7.1: Cloud Water Content Measurements

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<td></td>
<td>4.0</td>
<td>$1.01 \times 10^4$</td>
<td>$5.70 \times 10^{11}$</td>
<td>$6.22 \times 10^{11}$</td>
</tr>
</tbody>
</table>

### 7.3. Discussion

The modeling results of Michaels et al. (2006) have theoretically supported our observations. They used a 3-D, non-hydrostatic mesoscale Mars atmospheric model with detailed microphysics to model the formation of discrete afternoon clouds over Olympus Mons. They found that the primary cause of afternoon cloud formation over volcanoes is due to a combination of thermally-induced upslope flow and the rising branch of a mountain (gravity) wave which drive water vapor up the volcano flanks, adiabatically cooling and condensing it in the lee of the mountain. Just below the volcano’s peak, cloud particles begin to be transported downstream (to the west) and out of the updraft
core by large-scale horizontal winds. Furthermore, they found that a considerable amount of the cloud mass (the larger particles) quickly falls to lower elevations above the flank of the volcano, but the smaller particles have settling times long enough to create a cloud plume which extends to the west of the summit. The faster settling of larger particles leads to a size gradient within the cloud where the effective radii of the condensate particles decreases with height and lateral distance form the volcano. The cloud plume that Michaels et al. (2006) find in their modeling results is evident in each of our MOC images (Fig. 6-1), but the plumes of Olympus Mons in Fig. 6-1a and Ascraeus Mons in Fig. 6-1d are especially prominent and extend a great distance from the volcano summit. The cloud mass that falls to lower elevations above the volcano flanks is seen in each of our cloud optical depth maps as the regions of largest opacity (cloud cores). Cloud opacity is lower above the volcano summit because many particles do not reach the height of the summit before being carried westward, but those that do are smaller particles that have just recently formed and not yet settled. The cloud edges have the lowest optical depth because these consist of small particles that have been transported downstream.

In cloud optical depth simulations at $L_S = 100^\circ$ for Olympus and Ascraeus Mons, Michaels et al. (2006) found that the largest region of optical depth formed in an arc concentric with the summit caldera on the leeward (west) side of the volcano and that the optical depth above the summit is quite low. As discussed above, we found that the largest cloud optical depth formed in an arced region to the west of Ascraeus and Elysium Mons; however the region of largest optical depth at Olympus was to the north of the summit. Michaels et al. (2006) also saw a secondary region of large optical depth
to the southwest of the Olympus Mons summit which corresponds to the region of “above average” optical depth seen directly west of the summit in Figs. 7-1a and 7-2a.

The bright cloud to the west of the Elysium Mons summit in Fig. 7-5b is unusual in that it is most likely not formed due to upslope flow because it is not over the volcano itself. The elevation of the terrain where the cloud has formed is ~ 1 km as seen in Fig 7-5a. It is possible that this region has an enhanced subsurface source of water vapor associated with buried ice deposits, as is speculated for Arsia Mons by Noe Dobrea and Bell (2005). Feldman et al. (2004) produce a map of water-equivalent hydrogen in the uppermost meter of Martian soil determined from epithermal neutron counting rates using the Neutron Spectrometer aboard Mars Odyssey. This map shows a small enhancement of water-equivalent hydrogen near the same region where the bright cloud is observed, compared to the surrounding terrain. The Hadley circulation is the mechanism that provides uplift of water vapor to create the equatorial cloud belt (Clancy et al. 1996), which includes this region. Thus, the Hadley circulation, which is strongest at this time ($L_s \sim 100^\circ$), may provide the means by which a source of water vapor could be carried high enough into the atmosphere to condense and form this bright cloud.
Chapter 8

Summary and Future Work

8.1. Cloud Area and Cloud Centers

The Mars Global Surveyor Mars Orbiter Camera was used to obtain global maps of the Martian surface with equatorial resolution of 7.5 km/pixel. The maps used were acquired between August 2001 - May 2003 for Olympus Mons, Ascraeus Mons, Pavonis Mons, Arsia Mons, and Alba Patera, and March 1999 - May 2003 for Elysium Mons. Using the global maps, cloud area has been measured daily for water ice clouds topographically associated with the volcanoes. We investigated the interannual variability in cloud area and effect of the 2001 planet-encircling dust storm on cloud activity.

The seasonal trends in cloud activity established by Benson et al. (2003) for Alba Patera, Olympus, Ascraeus, Pavonis, and Arsia Mons are found to closely repeat here with an additional year of coverage. Namely, Olympus, Ascraeus, and Pavonis Mons show cloud activity from about $L_S = 0^\circ - 220^\circ$ with a peak in cloud area near $L_S = 100^\circ$; Arsia Mons shows year-round cloud activity; and Alba Patera shows a double peaked feature in the cloud area with peaks at $L_S = 60^\circ$ and $140^\circ$ and a minimum near $L_S = 100^\circ$.

Elysium Mons cloud activity has a similar trend to that of Olympus Mons, however there are some notable differences. The peak in cloud area of Elysium Mons is near $L_S = 130^\circ$, rather than $L_S = 100^\circ$. Also, cloud area sharply declines after the peak
and goes away in all years at about $L_S = 180^\circ$. Cloud area in all three years of observations was very similar. The clouds at Elysium disappear earlier than at any of the other volcanoes studied; therefore the cessation of cloud activity at this location in MY 25 was not affected by the 2001 planet-encircling dust storm.

At all other sites besides Elysium Mons, cloud activity ends earlier in MY 25 than the other two years observed, but this can be seen most clearly at Pavonis and Arsia Mons. This earlier cessation of cloud activity correlates with the start of the planet-encircling dust storm. At Arsia, where cloud activity was continuous in MY 24, clouds disappeared totally for $\sim 85^\circ$ of $L_S$ ($L_S = 188^\circ - 275^\circ$) in MY 25. The Arsia clouds reappeared later in MY 25 but were smaller in size than the previous year. It is of interest to speculate on events that might forecast major dust storms, but we detected no behavior in the cloud data that could be inferred to be a precursor of the 2001 storm.

Cloud top heights were determined for Alba Patera, Olympus, Ascraeus, Pavonis, and Arsia Mons for several days throughout their cloudy seasons. Cloud heights lie between 19.0-21.0 km for Olympus Mons, 15.0-18.0 km for Ascraeus Mons, 12.0-14.0 km for Pavonis Mons, 16.0-17.4 km for Arsia Mons and 5.5-6.5 km for Alba Patera. Clouds tended to be at a higher altitude later in the afternoon. On consecutive days, the cloud height could vary 1-2 km depending on local time at the topographic feature.

Clouds over Olympus, Ascraeus, and Pavonis Mons extend to the west-northwest. Clouds at Alba Patera extend over the volcano in an east-west direction, and Elysium clouds extend westward from the volcano summit. Arsia clouds extend to the south over the southwestern flank of the volcano, and cloud centers are seen to be distributed into two distinct regions depending on the time of year. From $L_S = 150^\circ - 30^\circ$ clouds are
mostly centered south of the volcano summit, towards the southwestern flank, and fall within a small latitude range (10° - 12° S). From Ls = 30° - 150° clouds are mostly centered over the summit with only a few clouds over the southwestern flank, and they are more widely scattered in latitude (7° - 14° S).

8.2. Optical Depth Mapping

The maps of 425 nm optical depth produced here for Olympus Mons, Ascraeus Mons, and Elysium Mons are the first of their kind; no previous study has examined the optical depth of Martian clouds at a high spatial resolution. The detail of the cloud structure is observed for the first time, and large optical depth gradients at spatial scales of 10-20 km are revealed. Cloud water content calculations were made by utilizing the optical depth measurements, and the total H2O volume of the clouds was found to be on the order of $10^{11} - 10^{12}$ cm$^3$.

Maximum cloud optical depth lies between 0.7 and 0.8 for Elysium Mons, 0.8 and 0.9 for Ascraeus Mons, and exceeds 1.0 at Olympus Mons. The largest optical depths lie along the volcano flanks below the summit, and the smallest optical depths are seen near the cloud edges. At Ascraeus and Elysium Mons the part of the cloud with the largest optical depth formed an arced region at a lower altitude than summit. The Elysium Mons observation at Ls = 104.22° shows a separate region of high optical depth to the west of the summit which may be due to an enhanced subsurface source of water vapor associated with buried ice deposits which is uplifted by the upward branch of the Hadley cell where the particles condense.
The modeling results of Michaels et al. (2006) theoretically support our cloud and optical depth observations. We observed cloud plumes extending to the west of the volcano summits due to cloud particles being transported downstream. Large cloud particles fall to lower elevations above the volcano flanks which give rise to the regions of largest opacity seen in our optical depth maps. Cloud opacity is lower above the volcano summit because many particles condense below the height of the summit and are carried westward. The cloud edges have the lowest optical depth because these consist of the smallest and lightest particles that have been transported downstream.

8.3. Future Work

As a follow-up to the work presented in this Thesis, we intend to continue to investigate the properties of Martian clouds including optical depth, particle size, and water content. A multi-year seasonal study of the optical depth of the volcano clouds is planned in order to investigate how cloud optical depth and particle size varies throughout the entire aphelion season when cloud activity is at its maximum ($L_S \sim 80^\circ - L_S \sim 125^\circ$). It would also be interesting to study the properties of these clouds during the less active season and to extend the study to the volcanoes of Pavonis and Arsia Mons and Alba Patera. Finally, it would also be of interest to examine properties of clouds that are within the equatorial cloud belt but not topographically related to the volcanoes.
Appendix A

Martian Polar Caps

This research was conducted while working towards the Ph.D. degree, and therefore is included here for completeness.

A.1. Introduction

The seasonal Martian polar caps wax and wane in response to the condensation and sublimation of CO$_2$ resulting from seasonal insolation changes on Mars. Carbon dioxide condenses on the surface during the polar winter, and the latent heat released is a major source for the energy radiated to space by the cap. The solid CO$_2$ sublimes in spring, absorbing the latent heat from the insolation and returning CO$_2$ to the atmosphere.

Abundant ground-based data exist on observations of the sublimation phase of the polar caps in the visible portion of the spectrum. Ground-based observations of the polar cycles have been limited to recessional phases because the caps are tilted away from the Sun and Earth during their formation. Fischbacher et al. (1969) compiled mean regression curves based on the photographic material in the Lowell Observatory archive covering the oppositions from 1905 to 1965. Several observers have documented telescopic observations of north polar cap recessions: Dollfus (1973) reported recessions in oppositions between 1946 and 1952; Miyamoto (1963) described the 1962-1963 recession; and C. Capen and V. Capen (1970) reported observations between 1962 and 1968. The behavior of the north cap during the 1970’s, 1980’s, and 1990’s was discussed

Beginning with Mariner 7, spacecraft measurements have documented the seasonal evolution of both caps with much higher spatial resolution. Data from Mariner 7 established that the polar caps are composed of CO$_2$ ice (Herr and Pimental 1969). Mariner 7 also supplied images, which revealed the irregularity of the south polar cap boundary as well as the variations in albedo and frost cover in the cap’s interior (James et al. 1992). Mariner 9 observed late phases of the south polar cap recession after the summer solstice (Sharp et al. 1971) and the spring recession of the north polar cap from Ls = 40º to shortly after summer solstice (Soderblom et al. 1973). Viking orbiter 2 acquired an entire recession of the south cap; however, there were fairly large gaps in the coverage, especially during mid-spring (James et al. 1979). The Viking orbiters acquired more extensive observations of the north cap; however, the orbits of the Viking spacecraft were not favorable for synoptic observations of the north polar cap recession. The observations included oblique views of the seasonal cap during two years by Viking Orbiter 1 (James 1979) and high-resolution images of the residual cap in two consecutive summer seasons and of the seasonal cap during one spring season from Viking Orbiter 2 (James 1982). The Hubble Space Telescope (HST) observed part of one south cap recession in 1992 (James et al. 1996b); however, these images were of limited value.
because Mars was near minimum angular size and because of the WFPC1 aberration. HST also observed several north polar recessions during the 1990’s (Cantor et al. 1998).

James et al. (2000) used observations from the Mars Global Surveyor (MGS) Mars Orbiter Camera (MOC) during the first aerobraking phase to study the 1997 spring regression of the south polar cap. In March 1999, MGS achieved its mapping orbit, and MOC has now observed three complete spring recessions of the south polar cap and two of the north. James et al. (2001) reported on the 1999 regression of the south polar cap during the first mapping year and James and Cantor (2001) documented the 2000 north polar recession. The additional years of MOC mapping allow for a multi-year intercomparison of both the north and south polar recessions measured using the same instrument and technique. This comparison is especially interesting because an extensive planet-encircling dust storm occurred during the second Martian year while there was no such large storm in the first or third years (Smith et al. 2002). The differing dust histories enable the study of the effects of dust on the interannual variability of the polar caps.

The late winter and early spring portions of the north cap recession have been phases for which the largest interannual variability has been reported. The observations of Capen and Capen (1970) in the 1960’s suggest large departures from regression curves in other years. Compared to other ground-based and spacecraft (Mariner 9, Viking, & HST) observations, the regression curves of Capen and Capen are roughly 5° of L_S slower from L_S=10° to L_S=90° and approximately 3° slower after L_S=90° (Cantor et al. 1998). However, their use of limb to limb cap measurements are subject to problems due to limb darkening and morning and evening clouds, which are known to occur. Capen and Capen did note that at times polar hood clouds obscuring the cap affected their measurements.
A plateau in the north polar regression curve during early to mid-spring has been reported by ground-based observers (Capen and Capen 1970, Iwasaki et al. 1979, 1982, 1999) and was also observed by the Viking orbiters (James 1979). During this anomaly, the boundary of the cap remains at a fixed latitude of about 65° for several weeks before the recession resumes. Other datasets, however, do not show evidence of a plateau (James 1982, Cantor et al. 1998, James and Cantor 2001). The shape of the cap in mid-spring has also taken on a polygonal shape in some years (Soderblom et al. 1973, Cantor et al. 1998, James and Cantor 2001) but not in others.

Historically, the recession phase of the south cap has been more repeatable from year to year than that of the north cap. Comparisons between regression curves documented by ground-based observers have shown that they are essentially indistinguishable until Ls ≈230°, at which time some variability begins to be detected (James et al. 1987b). The 1956 south-cap recession, however, was advanced relative to others (James et al. 1987b). James et al. (2000) noted that observations of the 1997 recession during the MGS aerobraking phase are consistent with analyses of the ground-based record. James et al. (2001) observed that the 1999 and 1997 regressions are very similar but that they are faster than those in 1971 and 1977 (James et al. 1979).

Multi-year observations by MGS/TES have documented a large amount of interannual variability in the Martian dust cycle (Smith 2004). Intuitively, this would influence the annual CO₂ condensation and sublimation cycle at the poles. However, polar cap recessions in Martian years with very different dust histories sometimes vary only slightly. During the large, planet-encircling dust storms that have been observed during some Martian years, the atmospheric opacity for the entire planet is elevated for
several months. Atmospheric dust affects the energy input to the polar cap surface in two ways that work in opposite directions: dust absorption in the visible reduces the insolation reaching the surface but the resulting heating of dust grains increases the energy load in the thermal infrared. Dust that is deposited on the surface during sublimation or with CO$_2$ and H$_2$O during condensation can affect the albedo of the ice deposits and influence the rate of sublimation. Hansen (1999) reported that TES spectra indicate dust amounts of the order of 0.1 – 1% are intermixed with CO$_2$ in the polar caps. His modeling showed that 1 wt% dust mixed with CO$_2$ will lower the 0.4 – 1.4 µm albedo of the deposit to less than 0.5, only slightly brighter than the spectrum of pure dust, whereas dust mixing ratios between 0.01% and 0.1% result in albedos between 0.6 – 0.7.

This Section describes observations of the north and south seasonal and residual Martian polar caps obtained by MGS MOC between 1999 and 2003. We look for interannual variability of the north and south polar caps during the years observed by MOC. Due to the 2001 planet-encircling dust storm, we have the opportunity to compare the polar caps between consecutive years with different dust storm histories using the same instrument and technique; therefore it may be possible to determine the effects of dust on the condensation and sublimation of carbon dioxide in the cap. The observations and techniques are described in Section A.2; Section A.3 presents our results and a discussion of them; and major conclusions are summarized in Section A.4.

**A.2. Observations and Techniques**

The primary mission of Mars Global Surveyor commenced on March 9, 1999 and ended on January 31, 2001, when the Extended Mission began. MGS has a nearly
circular, 2-hour, polar mapping orbit with an altitude of 380-420 km. The Mars Orbiter Camera (MOC) on MGS includes two wide-angle cameras, WAB in the blue (400-450 nm) bandpass and WAR in the red (575-625 nm) bandpass, each with a 140º field of view (Malin et al. 1992). The detectors of both cameras are 3456-pixel line scan CCD arrays with maximum resolution of 230 m/pixel at the nadir and 1.5 km/pixel at the limb; images are acquired via a “push broom” technique in which the motion of the spacecraft in its orbit creates the spacing between adjacent lines in an image. The cameras continuously acquire a horizon-to-horizon image using a variable-summing algorithm that produces a constant resolution of approximately 7.5 km/pixel. MGS makes roughly 12.6 orbits per sol, and the mapping orbit is ideal for polar observations. Because of the 87º inclination of the orbit, MGS passes within 3º of the poles on every orbit. The overlap of the horizon-to-horizon strips acquired in the daily global map coverage provides good diurnal coverage near the poles, and mosaics of the polar regions with modest emission angles can be created from the separate strips.

The software used for image processing was developed at Malin Space Science Systems for radiometric and geometric processing. The individual WAR images were calibrated then divided by the cosine of the incidence angle so that the pixel values (DNs) are proportional to the Lambert albedo of the surface. These were projected into polar stereographic projections, and the resulting wedges were mosaikced into full views of the polar caps. There are uncertainties in the absolute calibration due to the unknown phase function, the effects of clouds, as well as by the zero-phase brightening observed in the MOC images, which is fairly extended; comparisons of MOC albedos with those acquired elsewhere suggests that WAR albedos are accurate only to within ~ 15%. The
absolute calibrations differ somewhat from those used in USGS ISIS software, a commonly used image processing package, although barring changes in the MOC cameras, the relative albedos should be comparable.

The results discussed in this Section are based on sets of WAR mosaics for roughly every fourth day. In the south, the mosaics used were acquired between $L_S = 172^\circ$ and $272^\circ$ in 1999, $L_S = 170^\circ$ and $270^\circ$ in 2001, and $L_S = 177^\circ$ and $271^\circ$ in 2003; in the north they were obtained between $L_S = 343^\circ$ and $91^\circ$ in 2000 and $L_S = 339^\circ$ and $91^\circ$ in 2002. This data set has two main advantages over previous observations. First is the temporal continuity of the mosaics; Mariner 9 and Viking polar imaging was sporadic. For example, mosaics of the north polar cap in spring made from Mariner 9 images required mosaicing images from a broad range of $L_S$ except in one case. The only observation in which the cap was portrayed in a single image was overexposed. Likewise, Viking imaging of the south polar cap was especially sporadic in early spring, and there was only one good set of observations between $L_S = 200^\circ$ and $237^\circ$, when the cap undergoes significant changes. The second advantage is the superior viewing geometry for MOC. Because of the multiple orbits per day, MGS images show the entire cap area after the equinox when the pole moves into daylight; before that, there is a hole in the center of the cap where there is no insolation. Conversely, images acquired by the polar orbiting Viking Orbiter 2 during its mission were obtained from near the spacecraft periapsis and therefore covered only a limited portion of the cap edge at one time. Another advantage is the low emission angles in the MOC images, which are beneficial in separating atmospheric and surface effects. In contrast, during the second Mars year of Viking, Viking Orbiter 1 images were obtained from a nearly equatorial orbit and
therefore have large emission angles. In practice, small features in Viking images can be clearly identified in the MOC mosaics.

Interactive Data Language (IDL) was used to locate the coordinates of approximately 100 points spaced roughly equally around the cap’s perimeter on the polar stereographic projection. The cap edge was distinguished from the surrounding terrain by visual inspection of the albedo variation. In this method, a crater or other appendage was included within the cap until it was clearly separated and then it was excluded. At times, dust clouds obscured portions of the cap edge; in this case no points were measured in the obscured region. These individual points were used to determine the average latitude of the north polar cap edge. For the south polar cap, however, the individual points measured were used to determine the radius and the center of the optimum circle fitting the observed cap perimeter as described in James and Lumme (1982). This method was adopted for the south polar cap because of the large displacement of the cap center from the geographic pole. The main error associated with the cap boundary latitude is the accuracy of the visual determination of the cap edge. Atmospheric effects, such as dust storms and clouds, limit the accuracy of locating the cap edge. We estimate the uncertainty of the cap boundary latitude to be ±0.2°.

A.3. Results & Discussion

A.3.1. Seasonal North Cap Recession

We have measured the seasonal recession of the north polar cap in 2000 and 2002 using MOC WAR images from roughly every fourth day between April 29 and December 18, 2000 (L_S = 343° - 91°) and March 10 and November 6, 2002 (L_S = 339° - 91°). The average latitude of the cap edge is plotted as a function of the areocentric
longitude of the Sun ($L_S$) in Fig. A-1. Note that there were no MOC observations between $L_S=10^\circ$ and $L_S=20^\circ$ in 2000 and $L_S=48^\circ$ and $L_S=57^\circ$ in 2002 when Mars was at conjunction.

Figure A-1: Regression of the north polar cap in 2000 ($\Delta$) and 2002 (+). The latitude of the cap edge on a stereographic projection is plotted versus areocentric solar longitude. Recessions are similar, however, there are slight variations between $L_S = 7^\circ$ and $L_S = 50^\circ$.

As noted in the Introduction, a halt in cap regression in early to mid-spring has been reported in some ground-based studies as well as observations in the first Viking year. Our measured regression curves from the two years are similar, with small differences between $L_S = 7^\circ$ and $L_S = 50^\circ$; there is no sign of a halt in cap recession in either year. The discontinuity in the regression curves near $L_S = 85^\circ$ is due to the emergence of
Olympia Planitia near 180° W which separates residual frost outliers from the main part of the cap.

Between Ls=7° and Ls=50°, the average recession appears to be slightly slower in 2002 than 2000 by 1° to 2° of latitude (see Fig. A-1). In order to determine if the difference at this time varied with longitude between the two years, we examined the individual regressions in 30° longitude bins between Ls=8° and Ls=48°. The interannual variation in the recession did vary slightly with longitude. In the longitude bins 90-120° W, 150-180° W, 180-210° W, 240-270° W, and 330-360° W, the regression rates were generally similar in 2000 and 2002; if they varied, it was normally by 1° of latitude or less. In the longitude bins 30-60° W, 210-240° W, and 270-300° W, the 2002 regression rate was slower than 2000 by about 1° to 2° of latitude. In the longitude bin 0-30° W, the regression rates were similar until ~Ls = 27°; after that the 2002 regression rate was slower by about 1° to 2° of latitude. In the longitude bin 60-90° W, the regression rates were comparable until ~Ls = 24°, after which time the 2002 regression rate was about 1° to 2° of latitude slower than 2000. In the longitude bin 120-150° W, the regression rates were similar until ~Ls = 39°, after which the 2002 regression rate was slower by ~1° of latitude. Finally, in the longitude bin 300-330° W, the regression rates were similar until ~Ls = 33°; afterwards the 2002 regression rate was generally about 1° to 2° of latitude slower than 2000. Figure A-2 graphically shows the cap boundaries in 2000 and 2002 between Ls=8° and Ls=48°. To summarize, between Ls=8° and Ls=48°, the regression rate of individual cap segments was either similar between 2000 and 2002 or the 2002 regression rate was ~1° to 2° of latitude slower than 2000, which mimics the behavior of the average cap regression rate in Fig. A-1. The only times when this was not the case, as
seen in Fig. A-2, were at $L_S=8^\circ$ between 70º W and 150º W and at $L_S=48.18^\circ$ between 180º W and 240º W. At these locations, the 2000 recession was slightly slower than 2002.

The recession also varied with longitude within a single year between $L_S=7^\circ$ and $L_S=48^\circ$. In both 2000 and 2002, the regression rate varied by as much as 3º to 4º of latitude depending on the longitude segments compared. This is consistent with the results of James and Cantor (2001) for the 2000 recession of the north polar cap. The longitude segments between 330º to 60º W appear to be the slowest to regress in both years, while the 90º to 120º W segment regresses the quickest (see Fig. A-2). A MOLA map of the north polar region shows that the former area (330º to 60º W) corresponds to the lowest elevations in the circumpolar region (Zuber et al. 1998). This asymmetry is therefore consistent with an enhancement of CO$_2$ condensation in lower regions that have higher pressure. The region around 90º W is slightly above average in elevation between 60º and 65º N latitude.

The 2001 planet-encircling dust storm occurred during the northern fall season on Mars, when CO$_2$ is starting to condense on the surface of the north polar cap. Atmospheric dust could affect the rate of CO$_2$ condensation, and dust that is deposited on the surface during the condensation process could influence the rate of sublimation and hence the regression rate. South polar cap modeling by Bonev et al. (2003) showed that for a CO$_2$ grain size of 1 mm, the normalized CO$_2$ sublimation flux increases 14-fold (from ~ 0.02 to 0.28) as the admixed dust concentration increases from 0.001 to 1 wt%. Their sublimation flux was normalized to the total flux incident to the top of the
Figure A-2: Schematic representation of the recession of the north polar cap between $L_S = 8^\circ$ and $L_S = 48^\circ$ in 2000 (red line) and 2002 (blue line). The edge of the cap is defined by visual inspection of MOC WA red mosaics. (A) $L_S = 8^\circ$, (B) $L_S = 20.90^\circ$, (C) $L_S = 24.17^\circ$, (D) $L_S = 28.79^\circ$, (E) $L_S = 33.82^\circ$, (F) $L_S = 37.0^\circ$ (G) $L_S = 44.63^\circ$, (H) $L_S = 48.18^\circ$. 
atmosphere and was calculated as a difference between the spectrally integrated fluxes absorbed and emitted by the surface. The exact values would differ for the north cap, however, the same general behavior would be expected; that is, the larger the amount of intermixed surface dust, the greater the sublimation flux. As noted above, the average regression and the regression of individual longitude segments does not vary much between 2000 and 2002, but at times the 2002 regression rate was ~1° to 2° of latitude slower than 2000. This seems to contradict what is expected, if in fact during the condensation process more dust than normal was deposited on the surface due to the planet-encircling dust storm. Therefore, it is unclear whether the dust storm had an influence on the cap recession in 2002.

A.3.2. Residual North Cap: Albedo Measurements

We have measured the average Lambert albedo for a 225 × 225 km² region around the geographic north pole in MOC WA red images (see Fig. A-3) for 1999, 2001, and 2003. The Lambert albedo is the ratio of the measured reflected intensity at a certain viewing geometry to the intensity of an ideal, perfectly reflecting, uniformly diffusing surface. The Lambert albedo is plotted as a function of $L_S$ in Fig. A-4. The 2001 data have been corrected for a change in sensitivity of the MOC WA red camera during the fall of 2001 by applying a correction determined using a high albedo region in Isidis (15° N, 275° W). The camera returned to its normal sensitivity near $L_S = 159°$. The general behavior of the albedo in this central region of the cap seems to be similar in 1999 and 2003. The main exceptions are two data points from 1999 near $L_S = 135°$, which have significantly lower albedo values; however, this may be due to a change in camera state in 1999 as described in the following paragraph. There is a gap in the MOC WA red
mapping subsequent to these events in 1999 due to the Geodesy Campaign, so the question of the duration of this suppression is not answered by the red images alone. Following this gap, the 1999 albedos again match the 2003 data quite well. The albedo appears to be somewhat higher in 2001, but this could be due to an overall normalization error remaining after the correction. The four data points occurring after \( L_S = 159^\circ \) do appear to correspond with the 1999 and 2003 data. The decrease after \( L_S > 160^\circ \) is probably due to the Lambert approximation failing at the large incidence angles in late summer.

![MOC WA red image of the residual north polar cap in 2003 (\( L_S = 109.55^\circ \)) showing the region (black box) around the geographic north pole used to measure the average Lambert albedo in MOC WA red and blue camera images.](image.jpg)
Figure A-4: Average Lambert albedo for a 225 × 225 km² region around the geographic north pole as a function of areocentric solar longitude for 1999 (*), 2001 (+), and for 2003 (Δ), measured from MOC WA red camera images.

In order to investigate the duration of the suppression of cap albedo in 1999 near $L_S = 135^\circ$, we have used the blue filter mapping images, which continued through the Geodesy Campaign. We measured the average Lambert albedo around the geographic north pole in MOC WA blue images (see Fig. A-3) from $L_S=132^\circ$ to $L_S=152^\circ$, which includes the time just prior to, during, and following the Geodesy Campaign. Figure A-5 shows these Lambert albedos plotted as a function of $L_S$. Near $L_S=134^\circ$ a substantial drop in albedo occurred, corresponding to the drop measured in the red images. After that the albedo gradually declines until about $L_S=144^\circ$ when it then begins to rise; however, the albedo does not rise as high as the values before $L_S=134^\circ$. The blue
Lambert albedo at two other areas on the cap and several areas off of the bright ice cap exhibit the same trend. The length of the albedo suppression and its geographical extent seems to rule out the possibility that it could be the result of a regional dust event. Measurements of MOC WA blue camera images at a high albedo region in Isidis (15° N, 275° W) suggest the same trend as that seen at the north polar cap, however there is a gap in the data just prior to the observed albedo drop. This suppression could be due to a problem with the gain tables used in processing or from another change in sensitivity of both MOC cameras.

Figure A-5: Average Lambert albedo for a 225 × 225 km² region around the geographic north pole as a function of areocentric solar longitude for 1999, measured from MOC WA blue camera images during the Geodesy Campaign. There is a drop in albedo starting at L_s = 134°.
A.3.3. Seasonal South Cap Recession

We have measured the seasonal recession of the south polar cap in 1999, 2001, and 2003 using MOC WAR images from roughly every fourth day from June 19 to December 30, 1999 ($L_S = 172^\circ - 273^\circ$), June 1 to November 11, 2001 ($L_S = 171^\circ - 270^\circ$), and May 1 to October 2, 2003 ($L_S = 177^\circ - 272^\circ$). The method of measuring the regression is described in Section A.2. The best-fit cap radii are plotted as a function of $L_S$ in Fig. A-6.

![South Polar Regression](image)

Figure A-6: Regression of the south polar cap in 1999 (*), 2001 (+), and 2003 (◊). The dimensionless average cap radius on a stereographic projection is plotted versus areocentric solar longitude. Recessions are nearly identical.
The regression curves from all three years are very similar, and the average cap recession does not appear to be influenced by the planet-encircling dust storm of 2001. Only minor effects that may be due to dust are seen in early spring in 2001, and small differences are seen in 2003 between $L_S=240^\circ$ and $L_S=250^\circ$. The apparent insensitivity of the average cap recession to dust storm activity is consistent with modeling results of Bonev et al. (2003). When averaged over the entire cap, the loss of flux in the visible is balanced by surface heating due to increased dust absorption in the IR, so the net effect of atmospheric dust on the cap thermal balance is small.

The region between $75^\circ$ S and $85^\circ$ S latitude and $150^\circ$ W and $310^\circ$ W longitude, termed the Cryptic region (Kieffer et al. 2000), appears to lose its CO$_2$ frost in early spring. TES observations of the south cap in early spring (Kieffer et al. 2000) reveal that the surface of the Cryptic region is not defrosted ground but is occupied by CO$_2$ deposits that are at the CO$_2$ condensation temperature but have a low albedo, typical of unfrosted ground. This region is interior to the periphery of the cap, which remains bright, until $L_S=235^\circ$; therefore, the Cryptic region does not affect our regression curve, which only includes measurements around the edge of the cap, before this time. When the periphery of the cap reaches the Cryptic region near $L_S=235^\circ$ our visible albedo radius drops quickly, while infrared measurements will gradually decrease. Therefore, there is a period of time when our average cap radius will not agree with TES (or other infrared) measurements.

The Mountains of Mitchel (MM) is an example of a region with high albedo at visible wavelengths that is preferentially sensitive to thermal IR heating from overlying dust. Bonev et al. (2002) showed examples of MOC images, which indicated that the
Mountains of Mitchel had regressed faster in 2001 by about 5º of Ls compared to 1999. They explained this occurrence through a model, which predicts that the CO2 sublimation rate increases with dust optical depth for a surface with high visual albedo. Figure A-7 shows a comparison of the Mountains of Mitchel in 1999, 2001, and 2003 at Ls=262º. This comparison shows that in 1999 and 2003 at Ls=262º the Mountains of Mitchel are nearly identical, while the 2001 image shows a more advanced regression stage, supporting the conclusion that MM sublimation is slower in a year without a major dust storm.

![Image of Mountains of Mitchel at Ls=262º in 1999, 2001, and 2003]

Figure A-7: MOC red filter images of the Mountains of Mitchel at Ls=262º in (A) 1999, (B) 2001, and (C) 2003. This comparison shows that in 1999 and 2003 the Mountains of Mitchel are nearly identical, while the 2001 image shows a more advanced regression stage, supporting the conclusion by Bonev et al. (2002) that the Mountains of Mitchel sublimation is slower in a year without a major dust storm.

**A.3.4. Residual South Cap**

The recession of the south polar cap terminates at about Ls = 300º in its “residual cap” configuration, and only minor changes in the cap then occur during the rest of the summer. Viking IRTM observations of this residual south cap showed that it was
composed at least partially of carbon dioxide ice stabilized against the high insolation of southern summer by its high albedo (Kieffer 1979). There was a substantial difference in the appearance of the residual south polar cap seen in 1972 Mariner 9 images compared to 1977 Viking images (James et al. 1979). In 1972 the cap was much further along in the sublimation process than that in 1977 (see Fig. A-8). James et al. (2001) compared the 1999 residual cap configuration at $L_S=306^\circ$ to those of 1977 and 1972 and found that it was similar to the cap seen by Viking; although in detail the 1999 cap was a little ahead in the sublimation process compared to 1977 at this time. However, by $L_S=340^\circ$ the Viking and MGS residual caps were almost identical in extent and surface structure; in very late summer the MGS cap was slightly more frost covered than that of Viking.

Figure A-8: (A) Mariner 9 and (B) Viking images of the south residual cap at $L_S=306^\circ$. The images have been reproduced from James et al. (2001) (their Figure 5b and c).

In Fig. A-9, we compare MOC images of the residual south cap in 1999, 2001, and 2003, all at $L_S=306^\circ$. The color images are composed of red and blue images obtained with the MOC wide angle red and blue cameras, plus an artificially generated
green image that is the arithmetic mean of the red and blue images. The cap configurations from the three years appear to be nearly identical. There are only a few small differences near the edges of the cap. All three images are similar to the 1977 cap in the Viking image in Fig. A-8B at \( L_S = 306^\circ \) though they are slightly advanced in the sublimation process. That is, the inner structure of the caps are nearly indistinguishable, however, more \( \text{CO}_2 \) appears to have sublimed near the edge of the cap and at the outliers in 1999, 2001, and 2003, than in 1977.

The overall extent and structure of the residual south polar cap (RSPC) does not appear to have been influenced by the 2001 global dust storm. This seems to correlate with the average recession of the seasonal cap that year which was also unaffected. Because the RSPC is very bright, one would have possibly expected more rapid sublimation in 2001 similar to the Mountains of Mitchel. There are at least two possible explanations why the 2002 RSPC behavior is insensitive to prior dust storm activity: the solar incidence angles are quite small during the 2001 storm, thus minimizing the effects on surface energy flux; dust does not appear to penetrate into the core of the cap, perhaps because of net outflow from the seasonal cap at this time.

Within this context, it is uncertain what caused the differences seen in the south residual cap configuration in 1972 versus 1977, 1999, 2001, and 2003. The sublimation of the cap was enhanced in 1972 compared to the other four years observed by Viking and MOC. One idea may be that the residual south polar cap is more sensitive to late spring/summer dust storms. The 1971-1972 dust storm began at \( L_S = 260^\circ \) and lasted until about \( L_S = 330^\circ \), thus occurring while the residual cap was completely exposed. The 2001 dust storm occurred from \( L_S = 193^\circ \) to about \( L_S = 230^\circ \). On the other hand, the occurrence
Figure A-9: Color images of the residual south polar cap at $L_s=306^\circ$ on (A) February 22, 2000, (B) January 9, 2002, and (C) November 28, 2003. These are composed of red and blue images obtained with the MOC wide-angle red and blue cameras, respectively, plus an artificially generated green image that is the arithmetic mean of the red and blue images. The cap configuration from all three years appears to be nearly identical, with only a few small differences near the edges of the caps.
of a late dust storm in 1977 ($L_S=268^\circ$~350$^\circ$) does not support such a direct correlation between late global dust storms and residual cap behavior. Another, more speculative idea attributes observations of enhanced water vapor in 1969 (Jakosky and Barker 1984) to partial or complete sublimation of the CO$_2$ residual polar cap that year. In this case, the 1972 cap would be “recovering” from some event of the previous Martian year.

The small-scale appearance of the residual south polar cap did change between 1999 and 2001. Using MOC narrow angle images, Malin et al. (2001) showed changes in the configuration of pits, ridges and mounds in the residual cap and determined that these escarpments eroded one to three meters in one Martian year, but they did not attribute these changes to dust storm activity. Byrne et al. (2003) did speculate that the initial formation of the pits could be due to a change in dust distribution on a planetary scale.

A.4. Conclusions

The Mars Global Surveyor Mars Orbiter Camera wide-angle cameras were used to obtain images of the north and south seasonal and residual polar caps between 1999 and 2003. Wide-angle red camera images were used in assembling mosaics of the north and south polar recessions and regression rates were measured and compared. Albedo values of the geographic north pole were measured using wide-angle red and blue camera images, and the residual south polar cap configuration was compared between the three years observed by MOC. The main results presented here are:

1. There are small variations in the north polar recession between 2000 and 2002, especially between $L_S=7^\circ$ and $L_S=50^\circ$, although we found no sign of a halt in cap recession in either year. Examination of the regression in 30$^\circ$ longitude bins showed
that between $L_S=8^\circ$ and $L_S=48^\circ$, the regression rate of individual cap segments was either similar between 2000 and 2002 or the 2002 regression rate was $\sim 1^\circ$ to $2^\circ$ of latitude slower than 2000, which mimics the behavior of the average cap regression rate. The only times when this was not the case, were at $L_S=8^\circ$ between 70$^\circ$ W and 150$^\circ$ W and at $L_S=48.18^\circ$ between 180$^\circ$ W and 240$^\circ$ W. At those locations, the 2000 recession was slightly slower than 2002. The longitude segments between 330$^\circ$ to 60$^\circ$ W appear to be the slowest to regress in both years, while the 90$^\circ$ to 120$^\circ$ W segment regresses the quickest. Both of those occurrences appear to be correlated to topography.

2. The albedo of the geographic north pole generally varies between 0.5 and 0.6 as measured from MOC wide-angle red images. The behavior of the albedo in 1999 and 2003 is similar. In 2001 the albedo appears to be somewhat higher, but this could be due to an overall normalization error remaining after a correction was made for a change in sensitivity of the MOC WA red camera.

3. The south polar recession changes very little from year to year, and the 2001 planet-encircling dust storm had little if any effect on the average cap recession that year. This is consistent with modeling results of Bonev et al. (2003). The Mountains of Mitchel in 2003 appear very similar to this feature in 1999 at the same time period ($L_S=262^\circ$), however, they are smaller in 2001, indicating faster regression due to the major dust storm, as first recognized by Bonev et al. (2002).

4. There were only minor variations near the edges of the residual south polar cap between the three years examined.
References


