

PhD Thesis

The Water Cycle on Mars

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# Preface

This thesis is submitted in partial fulfilment of the requirement for the Danish PhD defence. The work was performed at the Center for Planetary Science and the former Department of Geophysics (now: Ice and Climate Group) at University of Copenhagen, Denmark. The project was supervised by Professor Dorte Dahl-Jensen and Associate Professor Christine Schøtt Hvidberg, whose help and advice during the project is greatly appreciated.

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# Introduction

From past, present and upcoming missions to Mars, a wealth of data has been obtained and will continue to be collected for many years to come. This enables us to increase our understanding of the red planet and also learn more about our own. Understanding the water cycle on Mars will teach us about climate dynamics in general, under conditions different from those on Earth. One way of preparing for the future missions is to develop methods for data analysis which can be ready for the new data. For example, building models for the atmospheric transport or the ice caps of Mars enables us to probe our basic understanding of the processes controlling the climate and water cycle on Mars. This is achieved by comparing model results to measured data.

In this thesis the seasonal cycle of water vapour in the Martian atmosphere is studied through data analysis and model studies. The current state of the north polar ice cap is assessed with a model based on the seasonal mass balance of the ice cap.

The thesis begins with a background chapter about water and climate on Mars and a description of missions to Mars (Chapter 2). Next follows a chapter with a description of the data sets used in the models described in the later chapters (Chapter 3).

In Chapter 4, the current state of the north polar ice cap will be investigated. The main focus will be on two questions: Is it in steady state, growing or disappearing? And how does the climate influence the ice cap? A new model

has been made, where the influence of the geometrical shape of the ice cap on the mass balance has been incorporated.

Chapter 5 contains an analysis of water and wind data on Mars. A statistical numerical method of dividing the water fluxes in the atmosphere into different modes of transport has been applied. There is currently no wind data covering the surface of Mars globally, so calculated wind velocities from a database consisting of results from a general circulation model have been used. The outcome of the calculations allows us to analyse the mechanisms in the circulation, which actually transport the water on Mars, and shows the differences between the atmospheric branch of the water cycle on Mars and Earth.

In Chapter 6, two different models of the atmospheric transport are described and tested. They are then coupled to a model for exchange of water with surface sinks and sources. Using the results from the models, the mobility of water in the atmosphere and the degree of interaction with the sources and sinks are assessed.



# Background

Even though there is little water on the surface of Mars today, the landscape shows abundant evidence of erosional features created by liquid water, indicating a warmer and wetter climate in the past than at present [e.g. Carr, 1996]. At present there are minute amounts of water in the atmosphere, and liquid water is unstable at the surface. Ice has only been detected with certainty at the poles, however, it is most likely also present in the ground at high latitudes. Despite the sparse amount of water compared to past times, there is an active water cycle on Mars even today. Spacecraft have observed the formation of ice clouds and water frost on the surface and have measured the atmospheric water vapour content, which shows seasonal and spatial changes in the atmospheric water distribution through the Martian year.

## 2.1 Mars

Mars and Earth are the two planets in the Solar System which are the most alike in climatical aspects. They are both terrestrial i.e. they consist of rock and metal, and the gravity on the surface of Mars is  $0.38 \cdot g$  compared Earth ( $g = 9.82 \frac{\text{m}}{\text{s}^2}$ ). Mars rotates at approximately the same rate as Earth (a Martian day called a *sol* is 24.66 Earth hours), and its orbit around the Sun takes 669 sols, which is a little less than two Earth years. The obliquity of the two planets (the inclination of the rotational axis to the ecliptica)

is very similar;  $25^\circ$  for Mars compared to  $23.5^\circ$  for Earth. However, the eccentricity of Mars' orbit is greater than that of Earth, which results in a greater difference between the maximum solar insolation on the northern and southern hemisphere. At present Mars is closest to the Sun at the end of southern spring, so southern summers are shorter but warmer than those in the north. [Carr, 1996]. See Appendix A for more facts on the two planets.

Atmospheric motion on the two planets is governed by the same laws of physics, yet there are still some differences in the circulation on Mars, which can be explained by the differences between Earth and Mars as described below.

**Atmospheric composition:** Mars' atmosphere consists predominantly of  $\text{CO}_2$  with some noble and trace gasses. See Appendix A for a summary of the atmospheric composition.

**Temperature:** Mars is colder than Earth due to two primary reasons. 1) The larger distance to the Sun causes the surface of Mars to receive less solar radiation and 2) the less dense atmosphere causes a smaller greenhouse warming. During winter time the atmosphere will become so cold that the atmospheric  $\text{CO}_2$  will condense on the surface in the polar region, which induces a mass driven flow of air towards the winter hemisphere.

**Topography:** Mars has a greater variation in its topography. The highest point, the highest volcano in the solar system, is Olympus Mons at 26 km above the areoid, and the huge impact basin Hellas forms the lowest point at 8.2 km below the areoid. Mars has a distinct dichotomy, the average altitude of the southern highlands being several km higher than the smooth northern hemisphere, which only has few craters. See Figure 2.1 for a topographic map of Mars.

**Airborne Dust:** Mars has recurring huge dust storms, which sometimes cover the whole planet, the largest ones occur during the southern summer. The airborne dust affects the radiative transfer in the atmosphere significantly, which leads to a huge potential for feedback processes between the dust distribution and the general circulation.

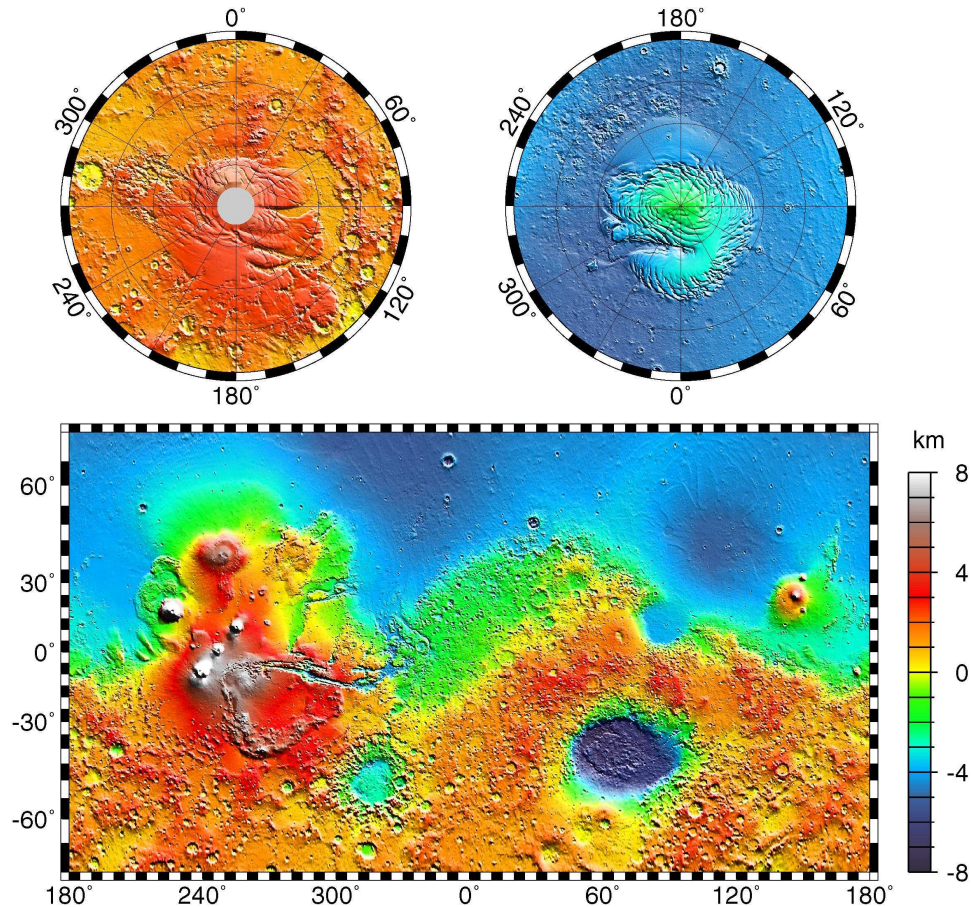
**No liquid water:** Mars has no liquid water on the surface today due to two causes. 1) The low atmospheric pressure (which is usually just below the triple point of water, which is given by  $T = 273.16$  K and  $p = 611.73$  Pa. 2) The low surface temperature, which is usually below the freezing point of water, although the temperature does sometimes rise above it. The lack of an ocean simplifies the climate dynamics and possible mechanisms of lateral heat transport on Mars compared to Earth.

## 2.2 Weather and Climate

In the northern and southern hemispheres CO<sub>2</sub> begins to condense out of the atmosphere at high latitudes in autumn as the temperatures drop below the freezing point of CO<sub>2</sub>. The CO<sub>2</sub> forms a snowcover on top of the perennial ice cap and the surrounding terrain. Throughout the period of deposition, the frost line migrates from the polar regions towards the equator and reaches mid-latitudes in late winter [James et al., 1992]. During the northern spring the CO<sub>2</sub> seasonal cap covering the north polar ice cap recedes, and the perennial H<sub>2</sub>O cap emerges gradually until the last CO<sub>2</sub> is gone [Cantor and James, 2001]. In the southern hemisphere the CO<sub>2</sub> does not disappear completely but continues to cover the south polar ice cap. However, recently H<sub>2</sub>O has been observed in gaps in the CO<sub>2</sub> cover, which indicates interannual variation in the coverage of the CO<sub>2</sub> snow. The timing of the recession of the CO<sub>2</sub> cap varies, which is a sign of interannual variability in the annual cycles of the volatiles in the atmosphere i.e. CO<sub>2</sub> and H<sub>2</sub>O. [Cantor et al., 1998, James and Cantor, 2001, Cantor and James, 2001, Calvin and Titus, 2004].

Global dust storms have been observed to occur repeatedly, but not every year [Kahn et al., 1992]. The global dust storms seemingly result from growth and coalescence of several local dust storms. They rapidly grow due to a positive dynamic feedback involving the intensification of winds due to heating of dust suspended in the atmosphere, and raising of more dust by the intensified winds [Zurek et al., 1992]. All of the observed global dust storms and the largest local ones have occurred during southern spring and summer, and originate in the south subtropical regions of Mars [Zurek et al., 1992].

There is reason to believe that major climate changes have happened on Mars, as so many landforms have been shaped by liquid water, indicating a warmer and wetter climate in the past than at present. A probable cause of the climate changes could be the variations in incoming solar radiation caused by variations in Mars' orbit around the Sun. Mars' orbit around the Sun changes in time due to gravitational interaction with the other planets in the Solar System. The periodical changes in the eccentricity of the orbit, the obliquity and the precession of the rotational axis are called Milankovitch cycles. As on Earth, the Milankovitch cycles cause variations in the climate, as the position and orientation of the planet relative to the sun determines the amount, seasonality, and geographical distribution of the incoming solar radiation at the surface.

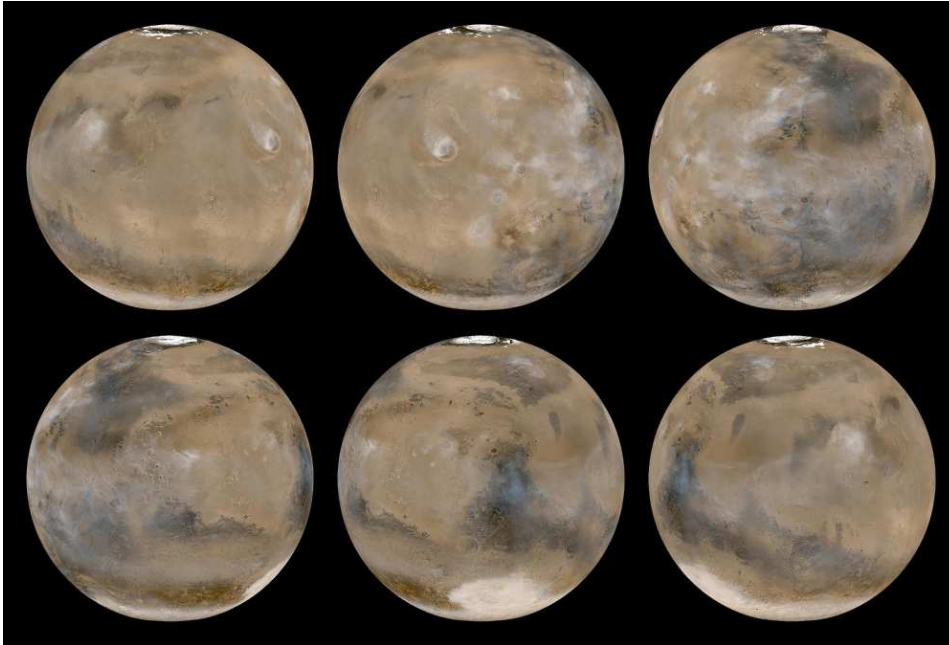


**Figure 2.1:** Topographic maps of Mars based on data from the Mars Orbiter Laser Altimeter (MOLA). The colour bar indicates the topographic height in kilometres above the areoid (the equipotential surface which is equivalent to the “mean sea level” on Earth). The latitude and longitude refer to an areocentric coordinate convention with east longitude positive. Image credit: MOLA Science Team.

Lower panel: Mercator projection map covering the latitudes between  $70^{\circ}\text{S}$  (indicated by minus sign) and  $70^{\circ}\text{N}$ . The Martian dichotomy is clearly seen with the distinct division between the low-lying smooth young plains of the northern hemisphere and the cratered highlands in the South. Prominent features seen are the deep canyon system Valles Marineris ( $10\text{--}20^{\circ}\text{S}$ ,  $265\text{--}325^{\circ}\text{E}$ ); the impact basin Hellas ( $45^{\circ}\text{S}$ ,  $70^{\circ}\text{E}$ ); and the volcanic highlands Tharsis (equator,  $220\text{--}300^{\circ}\text{E}$ ) with the largest volcano in the Solar System Olympus Mons to the northwest ( $18^{\circ}\text{N}$ ,  $225^{\circ}\text{E}$ ).

Upper right: Polar stereographic map centered around the north pole. The light blue and green area is the north polar ice cap, which lies several km higher than the surrounding smooth terrain. The north polar cap is cut through by a deep gorge named Chasma Boreale. Around the edges of the polar cap a spiral pattern is seen which consists of flat plateaus and steeper scarps. The center of the ice cap is very smooth and almost devoid of craters indicating a very young surface.

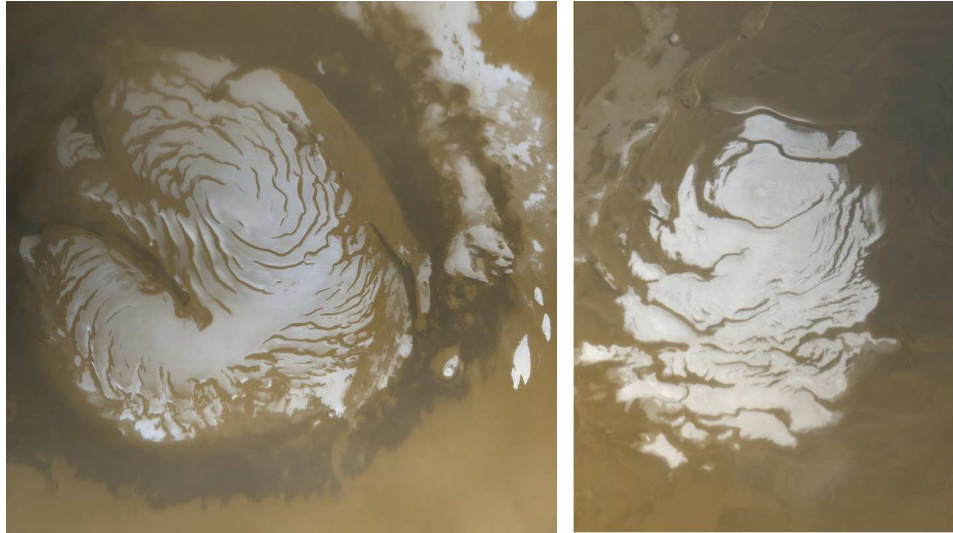
Upper left: Polar stereographic map centered around the south pole. The south polar ice cap and the polar layered deposits are seen as the red area surrounded by orange. Note that the spiral structures close to the pole are similar to the features on the north polar cap. The grey area indicates insufficient data.



**Figure 2.2:** The Mars Orbiter Camera (MOC) has since 1999 daily recorded images of the weather patterns of the planet. The figure shows six different views composed of 24 images of Mars acquired on February 14, 2003, when it was in the middle of the summer on the northern hemisphere. Water ice clouds can be seen around the equator and the lower latitudes, and orographically induced water ice clouds can be seen hanging over the highest points in the topography, which are the large volcanoes in the Tharsis and Elysium regions (upper left and centre and lower right). The north polar ice cap is fully exposed in the northern hemisphere, whereas the southern polar regions are covered by a seasonal layer of CO<sub>2</sub> frost. The frost extends itself furthest to the north in the Hellas Basin, where the lower topography and therefore higher atmospheric pressure in the basin allows the CO<sub>2</sub> to freeze at slightly higher temperatures than elsewhere in the southern hemisphere. Image credit: NASA/JPL/Malin Space Science Systems.

## 2.3 Water, Ice, and Permafrost

The atmospheric water vapour varies geographically and in time. An example of water ice clouds can be seen in Figure 2.2. Both the Viking Mission's Mars Atmospheric Water Detector (MAWD) and Mars Global Surveyor's Thermal Emission Spectrometer (TES) observed the Martian atmospheric water abundance through the Martian seasonal cycle. The MAWD data exists for a period of one Mars year and TES for three years. A repeating dominant feature in the atmospheric water distribution is a large peak in the concentration at high northern latitudes in the northern summer, when the permanent H<sub>2</sub>O ice cap is exposed to the atmosphere. This also shows

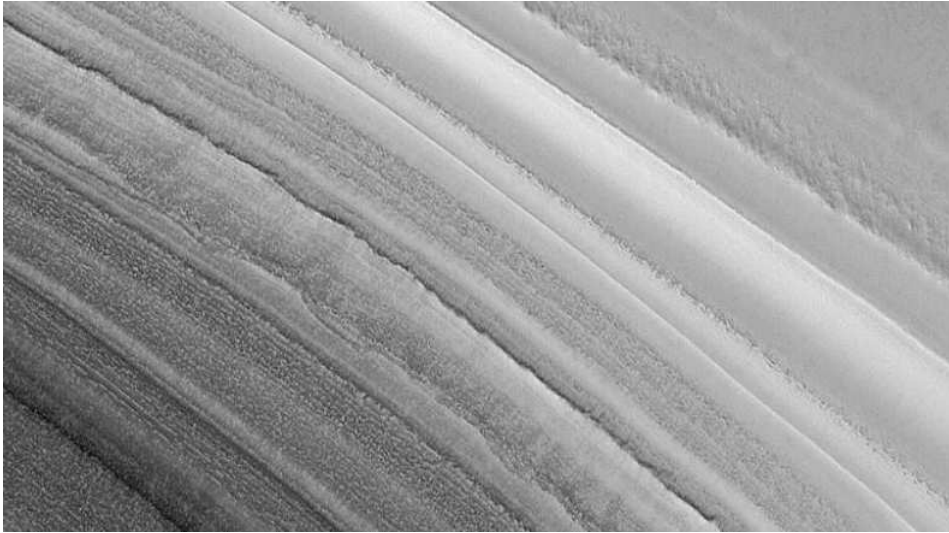


**Figure 2.3:** Left: The north polar ice cap in the northern summer time. The white areas are residual water ice that remains through the summer season. The nearly circular band of dark material surrounding the cap consists mainly of sand dunes formed and shaped by wind. The north polar cap is approximately 1100 kilometers across. Right: The south polar ice cap in the southern summer time. The image shows the south polar cap at its minimum extent. Even though it is summer, observations made by the Viking orbiters have shown that the south polar cap remains cold enough for the  $\text{CO}_2$  frost to remain. Recent observations however have shown that gaps in the  $\text{CO}_2$  frost occur in some years allowing the  $\text{H}_2\text{O}$  ice below to be detected [Bibring et al., 2004]. Image credit: NASA/JPL/Malin Space Science Systems.

that the northern ice cap is an active water reservoir in the present Martian water cycle. A somewhat smaller peak in the concentration is seen in the southern summer, which can be explained by two possible reasons. 1) The signal in the water vapour may be obscured in the measurements by the heavy planet-encircling dust storms. 2) The southern ice cap may be completely or partly covered by a layer of  $\text{CO}_2$  frost, which inhibits evaporation. However, both earthbased and satellite measurements have shown that the ice cap may be episodically exposed, and a peak of high water concentration can be present in the atmosphere at the south pole.

Studies of the MAWD and TES data indicate that there is a net transport of water from the north to the south during northern summer, which is not balanced by a similar northward transport at other seasons [Jakosky and Farmer, 1982, Smith, 2002].

The north polar ice cap of Mars (Figure 2.3) is one of the largest reservoirs of water at present. It contains layers of  $\text{H}_2\text{O}$  and  $\text{CO}_2$  ice and dust, as seen in Figure 2.4 which reflect the climate changes of the past. The elevation of



**Figure 2.4:** Layers seen in a slope of the north polar ice cap at  $81.5^{\circ}\text{N}$  in late northern spring. The image is taken by the Mars Orbiter Camera (MOC) and covers an area of approximately 3 km in width. The layers consist of  $\text{H}_2\text{O}$  and  $\text{CO}_2$  ice and dust. The brightest areas in the image are covered by frost. Image credit: NASA/JPL/Malin Space Science Systems.

its highest point relative to its surroundings is  $2950 \pm 200$  m and it covers an area of  $1.04 \cdot 10^6 \text{ km}^2$  [Zuber et al., 1998]. The large areas at the top of the ice cap are smooth and fresh-looking, indicating an active resurfacing process such as a presently occurring net deposition. The MOLA instrument on board Mars Global Surveyor measured the topography of the northern ice cap, and it was calculated by Nye et al. [2000] that it could not consist of  $\text{CO}_2$  ice, as this would have made it flatter, due to the differences in the rheology of  $\text{H}_2\text{O}$  and  $\text{CO}_2$  ice.

## 2.4 Missions

### Past Missions

Modern exploration of Mars began with the Mariner 4 mission in 1965 and was followed by Mariner 6 and 7, which all indicated a cold, dry, and dead planet. This picture changed with the arrival of Mariner 9 in November 1971 as the first artificial satellite of Mars. The mission revealed the geological diversity of the planet and revived the interest for its biological potential [Carr, 1996]. Between 1971 and 1973, the Soviet Union sent the missions Mars 2–7, where only Mars 5 lasted long enough to send back high quality

data and pictures.

The next major step was the Viking mission, which consisted of two spacecraft, each comprising an orbiter and a lander which all arrived in 1976. The primary goal of the mission was to look for signs of life. The two landers conducted three biology experiments designed to detect possible signs of biological activity, but found no evidence for this. They returned data on the organic and inorganic chemistry of the soil, the composition of the atmosphere, and the local meteorology. The orbiters systematically photographed the surface, mapped its thermal properties, and measured the water content of the atmosphere in space and time. The Viking mission was planned to last 90 days, but the landers continued to transmit data back to Earth until 1980 and 1982, respectively, and the orbiters until 1978 and 1980 [NASA's Mars Exploration Program, 2005].

After the Viking Missions a series of missions was launched, but only a small amount of data came out of the attempts. In 1988 the Phobos 1 and 2 missions were launched with the aim of studying Mars' satellite Phobos, but both were lost during the mission. However, some data of thermal and spectral properties of the Martian surface and on the erosion of the upper Martian atmosphere by the solar wind resulted from the mission [Carr, 1996].

In 1997 the Mars Pathfinder landed on Mars, carrying the small Sojourner Rover. During the mission scientific studies were made of the geology and geomorphology, mineralogy and geochemistry, magnetic properties, and surface material properties [JPL Mars Pathfinder, 2005]. The atmosphere and local meteorology was also studied through measurements of the varying temperature, pressure and near-surface wind. In order to get more precise data for studies of the rotational and orbital dynamics, the rotation of Mars was measured via the radio communication signals between Pathfinder and Earth.

In 1999 two spacecraft, Mars Climate Orbiter and Mars Polar Lander, were lost on arrival. The orbiter was supposed to study the weather and atmosphere and work as a communication relay for the lander. The lander should have studied surface meteorology as well as examined samples collected from the surface for water content and chemistry analyses [NASA's Mars Exploration Program, 2005].

## Current Missions

Currently there are four satellites orbiting Mars and two rovers driving on the surface of the planet, which is an unprecedented high number of missions occurring simultaneously.

Mars Global Surveyor has been orbiting Mars since 1997 and has delivered several groundbreaking new findings from its low-altitude near polar orbit. The mission has studied the entire Martian atmosphere, surface and interior. The Mars Orbital Camera (MOC) has taken high-resolution images of gullies and debris flows, suggesting that occasional sources of liquid water were once present at or near the surface of the planet. The Mars Orbiter Laser Altimeter (MOLA) has provided highly accurate topography data. Particularly the elevation measurements of the North Polar Ice Cap have led to new information regarding the structure, volume and extent, as well as the composition of the ice cap.

The 2001 Mars Odyssey spacecraft has provided measurements of thermal and radiative properties of the surface materials, enabling the creation of maps of minerals and chemical elements.

In January 2004 two robotic rovers landed on opposite sides of the planet. since then, Spirit and Opportunity have travelled several km across the Martian landscape studying the geology and making atmospheric observations. The rovers have made detailed microscopic images of rocks and soil and measured their elemental and mineralogical composition by use of four different spectrometers. On a robotic arm they carry a special rock abrasion tool allowing studies of the interiors of the rocks found on the surface. Bright patches of soil have been found to consist of evaporative minerals indicating deposition in an aqueous environment. Another indication that Mars had surface water in the past is the finding of layered rocks which may be sediments from running water in the past.

Mars Express arrived at Mars in 2003 and has since then been conducting investigations about the geology, atmosphere, surface environment, the history of water and potential for life on Mars. The mission is in particular dedicated to the search for sub-surface water from orbit.

The most recent spacecraft in orbit around Mars is the Mars Reconnaissance Orbiter, which arrived in March 2006. It carries a camera with the highest resolution so far, which will provide detailed images of the surface with the purpose of learning more about the planet as well as providing context for the rovers and determine whether future possible landing sites are smooth and safe. Studies will also be made of the transport of water and dust in the atmosphere, which will provide more insight into the climate dynamics on Mars.





# 3

## Model Input Data

Direct observations of the environment on Mars are very sparse, both in temporal and spatial coverage, so modelled quantities from a Mars general circulation model have been used in this thesis as a supplement to observations. Data from the Viking Mission, Mars Global Surveyor and Mars Odyssey Missions have been used. This chapter describes the data used in this thesis.

### 3.1 Mars Climate Database

At the Laboratoire de Météorologie Dynamique du CNRS in Paris and the University of Oxford a collaboration in developing and validating two General Circulation Models (GCM) has resulted in a compilation of a set of time-averaged GCM results for various representative scenarios of the Martian atmosphere into the Mars Climate Database (MCD) [Forget et al., 1999, Lewis et al., 1999]. The climatological statistical output in the MCD as well as a possibility of applying statistical-dynamical models of variability on a large scale (e.g. due to weather systems) and a small scale (e.g. due to vertically propagating gravity waves) allow for quick and easily accessible extraction of realistic atmospheric states. These can be used for studies of the Martian climate and meteorology from diurnal to annual timescales as well as modelling entry and landing of spacecraft on the surface of Mars.

The datasets in the MCD are stored so that they capture the diurnal and

seasonal changes in the weather pattern and cover the whole planet in a sufficiently high spatial resolution to be useful for both climate studies and modelling of spacecraft entry into the Martian atmosphere. The structure of the MCD datasets are described below.

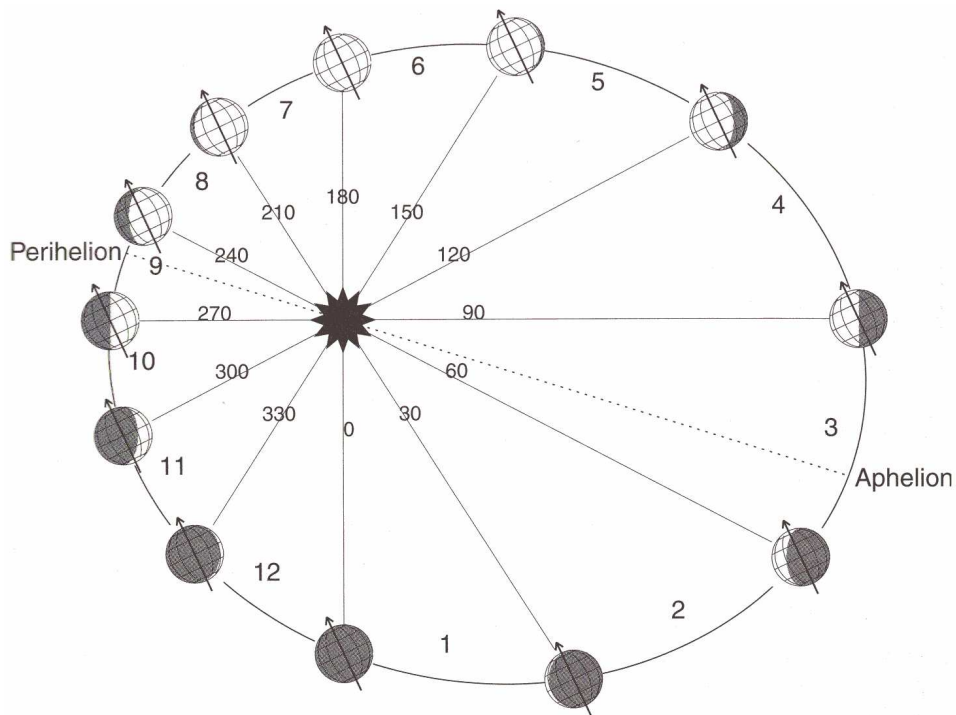
**Seasonal Resolution:** In the MCD the year is divided into 12 *seasons* (equivalent to the division of the Earth year into months), which are equally spaced in  $^{\circ}L_s$  or areocentric longitude of the perihelion illustrated in Figure 3.1. A whole year consists of  $360^{\circ}L_s$ , and each season corresponds to  $30^{\circ}L_s$ . The counting begins at the vernal equinox, so  $0^{\circ}$  corresponds to the northern spring,  $90^{\circ}$  to northern summer,  $180^{\circ}$  to northern autumn, and  $270^{\circ}$  to northern winter. Due to Mars' eccentricity the length of the seasons vary through the year. In Table 3.1 the correspondence between season number,  $^{\circ}L_s$ , sols, and a description of the seasons is listed. The data sets in the MCD are stored as season mean values for each season in order to represent the seasonal cycle of the climate within a manageable amount of data.

**Diurnal Resolution:** Each sol is divided into 24 “hours”, which are simply units of time of length  $\frac{1}{24}$  of a sol. For each of the different seasons the data fields have been stored every second hour, i.e. 12 times per sol, in order to ensure representation of the diurnal cycle.

**Horizontal Resolution:** The data sets are divided into 72 longitude points and 36 latitude points, which gives a spatial resolution of  $5^{\circ}$  in both latitude and longitude.

**Vertical Resolution:** The data sets are given for 32 vertical levels. The vertical spacing is stretched in order to enhance the resolution near the surface and thereby better represent the atmospheric boundary layer. At the upper layers the vertical resolution is about half a scale height, which for Mars is about 10 km. The vertical levels in the MCD are given by a terrain following  $\sigma$  coordinate system, where  $\sigma$  is pressure divided by the surface pressure; thus  $\sigma = 1$  at the surface and  $\sigma = 0$  at infinite height.

**Dust Scenarios:** The datasets are stored for 5 different dust scenarios with varying amounts of dust suspended in the atmosphere. The Viking dust scenario has been used in this study to ensure that the extracted wind speeds match the atmospheric state under which the MAWD water vapour data were collected during the Viking mission.



**Figure 3.1:** Illustration of the division of the Martian year into 12 seasons equally spaced in  $^{\circ}L_s$  for seasonal data stored in the MCD. Image credit: [Read and Lewis, 2004].

**Table 3.1:** Seasonal temporal structure for MCD

| Season number | $^{\circ}L_s$ | Sol interval | Season Length in sols | Description     |
|---------------|---------------|--------------|-----------------------|-----------------|
| 1             | 0–30          | 1–61         | 61                    | NH spring       |
| 2             | 30–60         | 62–127       | 66                    | Late NH spring  |
| 3             | 60–90         | 128–193      | 66                    | Early NH summer |
| 4             | 90–120        | 194–258      | 65                    | NH summer       |
| 5             | 120–150       | 259–318      | 60                    | Late NH summer  |
| 6             | 150–180       | 319–372      | 54                    | Early NH autumn |
| 7             | 180–210       | 373–422      | 50                    | NH autumn       |
| 8             | 210–240       | 423–468      | 46                    | Late NH autumn  |
| 9             | 240–270       | 469–515      | 47                    | Early NH winter |
| 10            | 270–300       | 516–562      | 47                    | NH winter       |
| 11            | 300–330       | 563–613      | 51                    | Late NH winter  |
| 12            | 330–360       | 614–669      | 56                    | Early NH spring |

## 3.2 The Meridional Wind Field

Measurements of wind on Mars are limited to surface winds at lander sites and vertical profiles from the entry of spacecraft. The winds can also be inferred from cloud movement as well as from profiles of vertical temperatures calculated on the basis of thermal emission spectra or earthbased microwave measurements. Wind directions can be inferred from the orientation of a variety of aeolian surface features, which however are limited in their usefulness due to a lack of understanding of how and when they have been formed. All of the above-mentioned ways of understanding the winds on Mars are limited by the lack of systematic global coverage. Until future ideas and ambitions of a global meteorological network are realised, GCM models are good substitutes at providing the global winds from calculations validated by the few datasets available.

Most of the year the Martian meridional circulation is dominated by an equatorial Hadley cell, with an ascending branch in the summer hemisphere and a descending branch in the winter hemisphere at about  $30\text{--}60^\circ$  latitude [Leovy, 2001]. The Hadley circulation is a form of thermal convection driven by the difference in solar heating between the equatorial and polar regions. Generally this circulation consists of zonally symmetric overturning, in which heated equatorial air rises and flows poleward where it cools, sinks and flows equatorward again. The Coriolis force deflects the poleward moving air at the upper levels to the east and the equatorward moving air near the surface to the west.

On Earth, Hadley cells exist on both hemispheres between the equator and  $30^\circ$  (the most prominent and permanent cells) and between  $60^\circ$  and the poles. Between  $30^\circ$  N/S and  $60^\circ$  N/S a circulation cell called a Ferrel cell forms at the mid-latitudes to balance the transport by the Hadley and polar cells. The Ferrel cell has air motion opposite to planetary rotation. At the surface, this forms the southwesterly prevailing westerlies. On Mars, the Hadley circulation during solstice seasons consists of a strong cross-equatorial surface flow with rising motion in the summer hemisphere, return flow at high altitude, and descent in mid-latitudes of the winter hemisphere.

The meridional wind field has been extracted from MCD and diurnally and zonally averaged to give the desired data set, which is illustrated in Figure 3.2.

The Hadley circulation on Mars is strongest during the northern hemisphere winter solstice due to the eccentricity of the Martian orbit, which causes large seasonal differences in the insolation. A quasi-global Hadley circulation extends from  $60^\circ$  S almost to the north pole, and strong meridional winds above

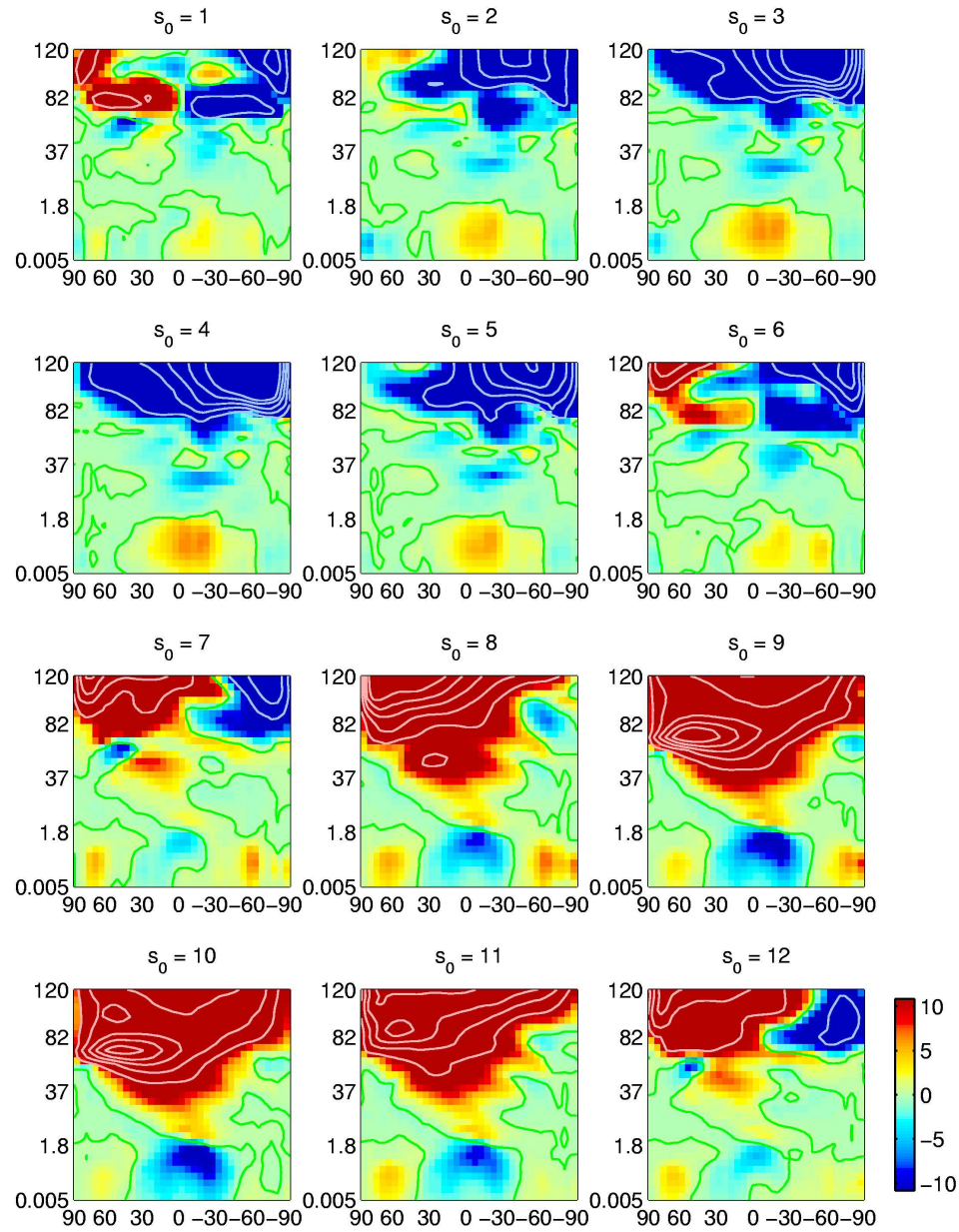
50 km altitude induce a convergence of mass above the polar region during northern winter, see seasons 9 and 10 in Figure 3.2. The convergence causes a descent of air in the polar region which leads to an adiabatic warming of air at lower altitude [Forget et al., 1999]. In the northern summer (seasons 3 and 4) the Hadley circulation is less intense due to a smaller thermal forcing caused by a smaller dust load in the atmosphere. At the equinoxes during northern spring and fall, the general circulation is dominated by a single Hadley cell in each hemisphere driven by the equator-to-pole heating gradient.

### 3.3 The Atmospheric and Surface Pressure

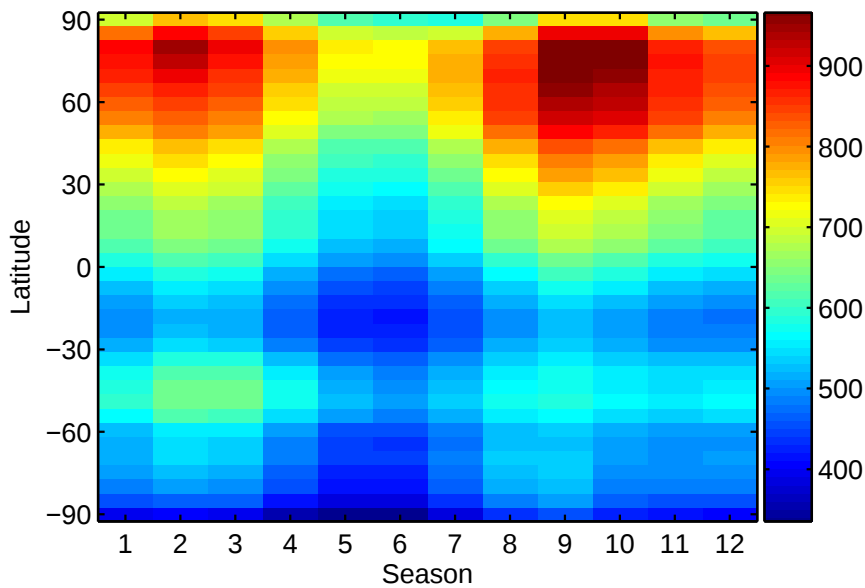
The Viking landers observed the seasonal, diurnal, and even hourly variation of the surface pressure during several Mars years. A seasonal cycle was observed and ascribed largely to the condensation and sublimation of  $\text{CO}_2$  with addition of a dynamical component. The seasonal cycle is modulated by high-frequency oscillations due to weather systems passing the landers [Zurek et al., 1992]. The Viking pressure measurements together with Mars Pathfinder's 83 sols of data have been used for tuning GCM models of the Martian atmosphere.

In Figure 3.3 the zonally averaged surface air pressure data from the MCD is shown. The influence of the topography can be seen as the pressures are higher on the northern hemisphere than on the southern hemisphere, where the average altitude is approximately 6 km higher than that of the northern hemisphere in average. Note also the increased pressure at 30-60°S, which is caused by the topographic depression of the impact basin Hellas. The average global surface pressure varies on Mars depending on how much of the atmospheric  $\text{CO}_2$  condenses out as snow or frost in the seasonal frost layer in the polar regions.

Figure 3.4 illustrates the surface pressure as a function of latitude and time of year relative to the annual mean pressure at the same latitude. The general level of each of the curves shows roughly how much  $\text{CO}_2$  is present in the atmosphere; it is seen that the highest levels occur during the northern late spring and early summer ( $s_0 = 2, 3$ ) and southern late spring and early summer ( $s_0 = 9, 10$ ). The pressure reaches the lowest levels during northern late summer and early autumn ( $s_0 = 5, 6$ ), when a large part of the atmosphere condenses out onto the surface at southern high latitudes. During the northern late spring and throughout the summer ( $s_0 = 2-5$ ), the surface pressures are relatively larger in the south midlatitudes compared to the north causing atmospheric motion towards the northern midlatitudes, which



**Figure 3.2:** Meridional wind data for the Viking dust scenario from MCD. The horizontal axis represents the latitude [°] and the vertical axis the altitude above the Martian surface [km]; note that the vertical scale is non-linear. Each plot corresponds to a season which lasts  $30^\circ L_s$ , the season number  $s_0$  is shown above each plot. The colour bar shows wind velocity [m/s], where positive values represent northward motion. The green contour line represents 0 m/s, the light red contour lines represent 20, 40, 60, 80, 100 m/s (northward motion), and the light blue contour lines represent -20, -40, -60, -80, -100 m/s (southward motion).



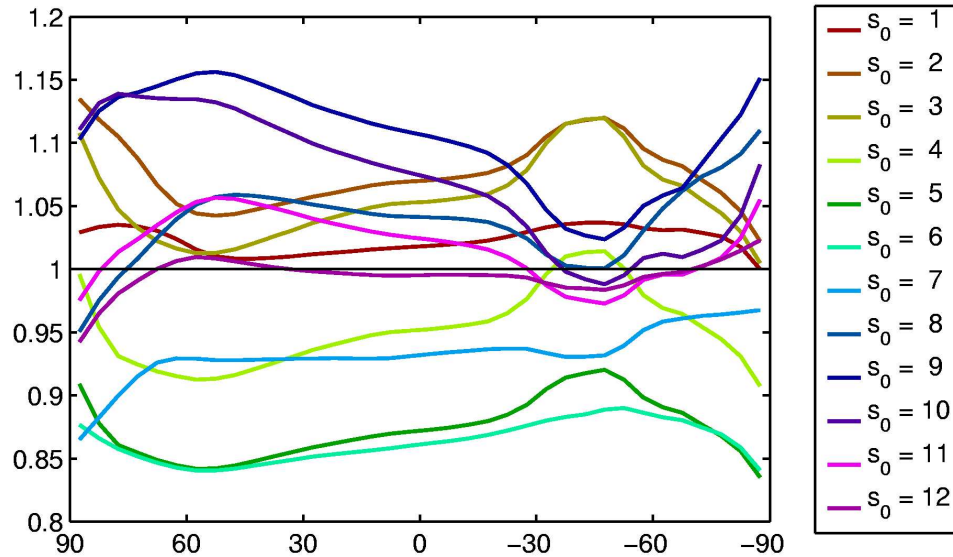
**Figure 3.3:** Zonally averaged surface pressure data for the Viking dust scenario from MCD. The colour scale covers the interval 326–933 Pa, which corresponds to the minimum and maximum values of the zonally averaged surface pressure.

is in agreement with the Hadley circulation. The opposite happens during southern late spring and summer ( $s_0 = 8\text{--}11$ ), where the general pressure is high at the beginning of this period compared to the annual mean, but the pressures are higher in the northern hemisphere than in the southern hemisphere.

The zonally averaged atmospheric pressure as a function of altitude above the surface and season is shown in Figure 3.5. The difference in topography causes higher pressures at low altitudes in the northern hemisphere than in the southern one. The atmospheric air pressure has been extracted from the MCD and averaged and stored for later use in the same manner as the meridional wind field.

### 3.4 The Atmospheric and Surface Temperature

The temperature on both the surface of a planet and the atmosphere is ultimately determined by the heating from the sun and the cooling to space. Heating takes place as a result of absorption of visible sunlight in the atmosphere and in the ground. Cooling is caused by direct infrared cooling of the

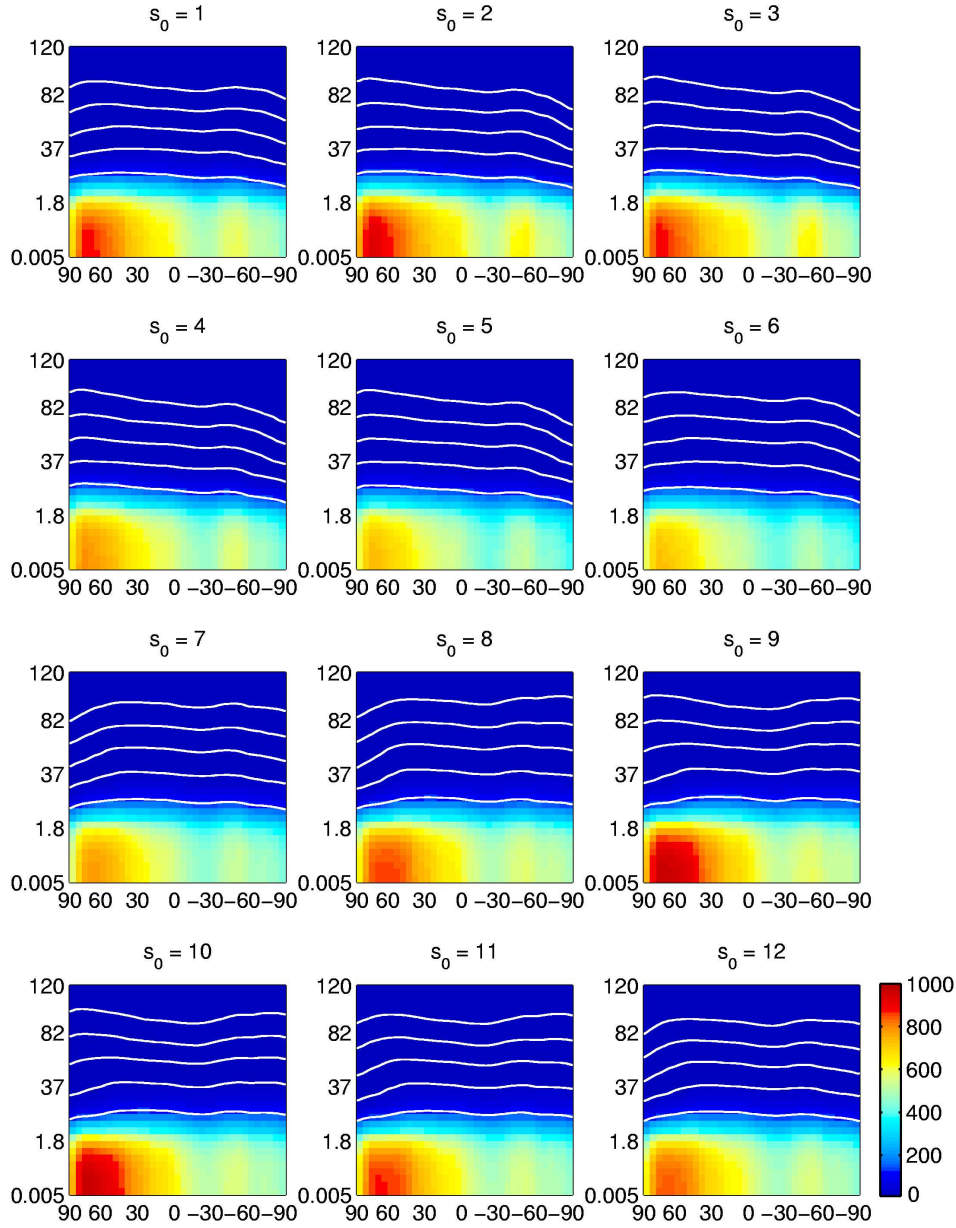


**Figure 3.4:** Illustration of the seasonal variation in the zonally averaged surface pressure relative to the annual average as a function of latitude (horizontal axis) and time of the year (see legend for season numbers).

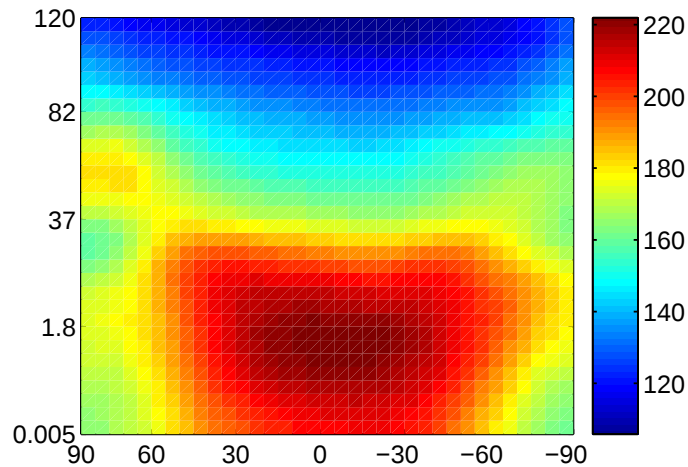
surface and indirect infrared cooling from the atmosphere — on Mars mainly in the  $15\ \mu\text{m}$  band in  $\text{CO}_2$ . To a first order the cooling is relatively uniform with latitude but the heating is a strong function of latitude, leading to a net heating in the equatorial region and a net cooling in the polar regions which can be seen in the annual average atmospheric temperatures in Figure 3.6. Due to atmospheric exchange of heat and energy across latitudes, the temperature difference between the high and low latitudes observed on Mars are much smaller than if the atmosphere was static and transport did not occur.

In the GCMs, which are the basis for the MCD the radiative transfer takes into account the effects of the presence of  $\text{CO}_2$  gas and mineral dust suspended in the air at both solar and infrared wavelengths [Forget et al., 1999]. The radiative effects of water vapour and water ice particles were neglected because of the low vapour column abundance and the limited occurrence of water ice clouds in space and time. The effect of  $\text{CO}_2$  condensation and sublimation is included in the calculation of the energy balance.  $\text{CO}_2$  condenses when the local temperature drops below the condensation temperature and releases the corresponding latent heat, and can conversely sublimate when it is heated. The sedimenting  $\text{CO}_2$  ice particles can thus sublimate when they reach warmer atmospheric layers as they descend to the ground.

In Figure 3.7 the diurnally averaged atmospheric temperature is shown as



**Figure 3.5:** Atmospheric pressure data for the Viking dust scenario from MCD. The horizontal axis represents the latitude [degrees] and the vertical axis the altitude above the Martian surface [km]; note that the vertical scale is non-linear. The colour bar shows the atmospheric pressure [Pa]. The number above each plot indicates the season number, each of which lasts  $30^\circ L_s$ . The white contour lines represent pressures of 0.01, 0.1, 1, 10, and 100 Pa (pressures decreasing with altitude).



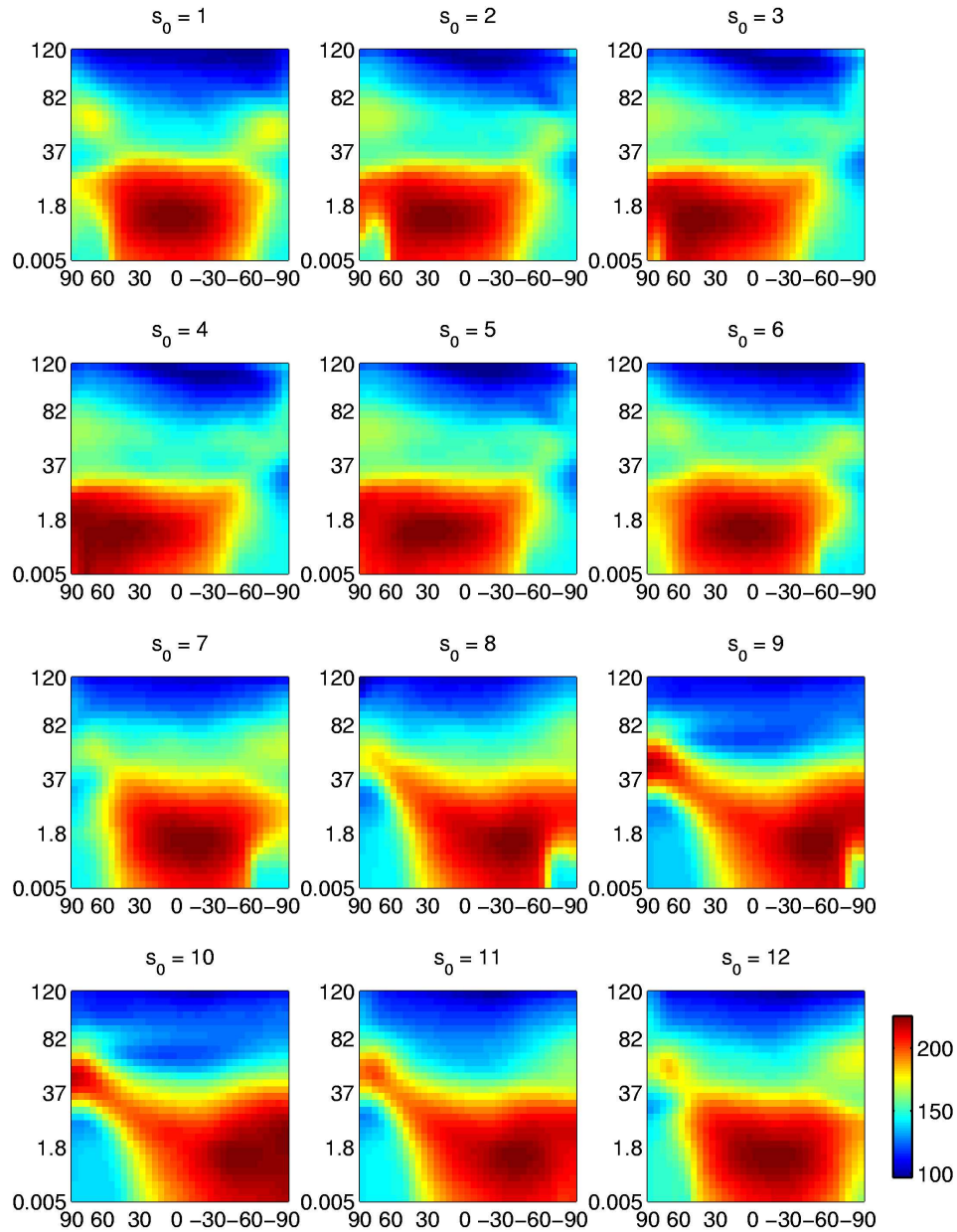
**Figure 3.6:** Zonally averaged annual mean atmospheric temperature data from the MCD, the colour bar indicates the temperature in [K]. The horizontal axis shows the latitude [°] and the vertical axis the altitude [km] on a non-linear scale.

a function of latitude, altitude, and time of the year. The heating at north polar high altitudes due to convergence of air during the northern winter is clearly seen (see also section 3.2). It can also be seen that the seasonal coverage of CO<sub>2</sub> frost on the surface causes cold temperatures in the lower atmosphere above the polar regions until the frost has disappeared by early northern summer ( $s_0 = 2, 3$ ) and early southern summer ( $s_0 = 8, 9$ ).

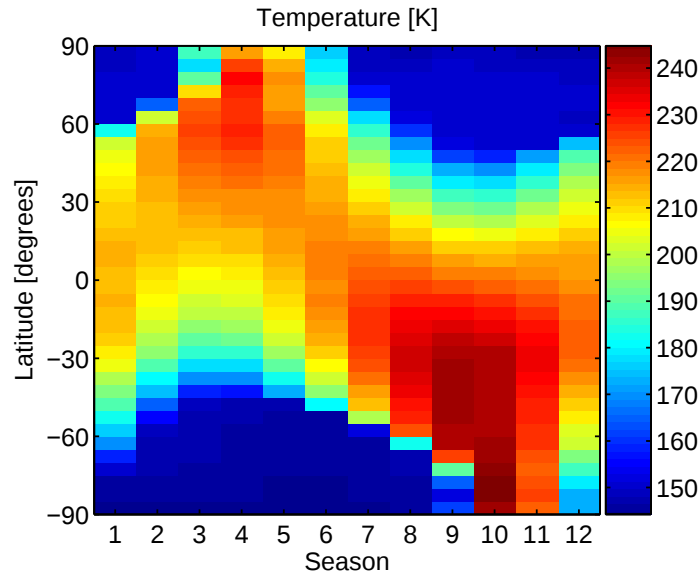
The calculated surface temperature is governed by the balance of the incoming fluxes of energy and heat (solar insolation, thermal radiation from the atmosphere and the surface itself, and turbulent fluxes) and the thermal conduction in the soil [Forget et al., 1999]. In the GCMs the thermal conduction in the soil was parameterised by a 11-layer soil model [Hourdin et al., 1993] with a vertically homogeneous soil and a spatially varying thermal inertia and albedo based on Viking IRTM observations.

The zonally averaged mean diurnal surface temperature  $T_s(\phi, L_s)$  is shown in Figure 3.8. The effect of the eccentricity of Mars' orbit is seen in the difference in the peak summer temperature on the two hemispheres. A rapid increase in the polar surface temperature can be seen after the disappearance of the seasonal CO<sub>2</sub> frost at the beginning of the summer on both hemispheres.

The atmospheric and surface temperatures have been extracted from the MCD, averaged and stored for later use.



**Figure 3.7:** Zonally averaged diurnal mean atmospheric temperature data from the MCD, the colour bar indicates the temperature in [K]. The horizontal axis shows the latitude [ $^{\circ}$ ] and the vertical axis the altitude [km] on a non-linear scale.



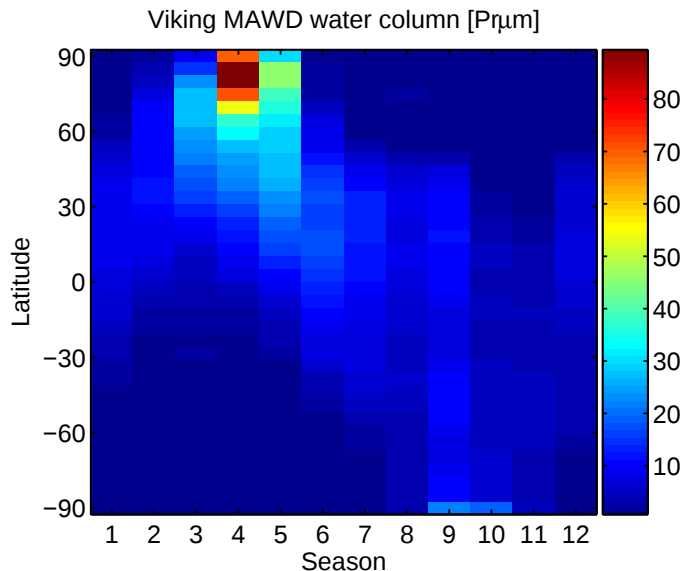
**Figure 3.8:** Zonally averaged diurnal mean surface temperature data from MCD, the colour bar indicates the temperature in [K]. The horizontal axis show the season number and the vertical axis the latitude [°].

### 3.5 The Water Vapour Column Abundance

The Mars Atmospheric Water Detector (MAWD) experiment on board the Viking Orbiters consisted of near-infrared reflectance spectrometers and mapped the total water column abundance in the Martian atmosphere for a period of a little more than a full Martian year. The vapour abundance along the observational path was determined from the relative absorption and translated into a column abundance using the known observational geometry and assuming a universal temperature of 200 K and an effective pressure of half the surface pressure [Jakosky and Farmer, 1982].

The measurements can be biased from scattering of dust or other atmospheric aerosols, as the observed sunlight will not have sampled the entire atmospheric column equally. The uncertainty in the data arising from the assumptions of effective pressure and temperature will be up to 30%, but is probably  $>15\%$ . The atmospheric dust will probably reduce the observed column by  $>10\%$  except during global dust storms and dusty time periods [Jakosky and Haberle, 1992]. During the recording, there were occasional fall-outs or missing data. Furthermore, the data does not cover the polar night on the two hemispheres as there is no sunlight present.

The MAWD data used in this study has been extracted from a dataset by



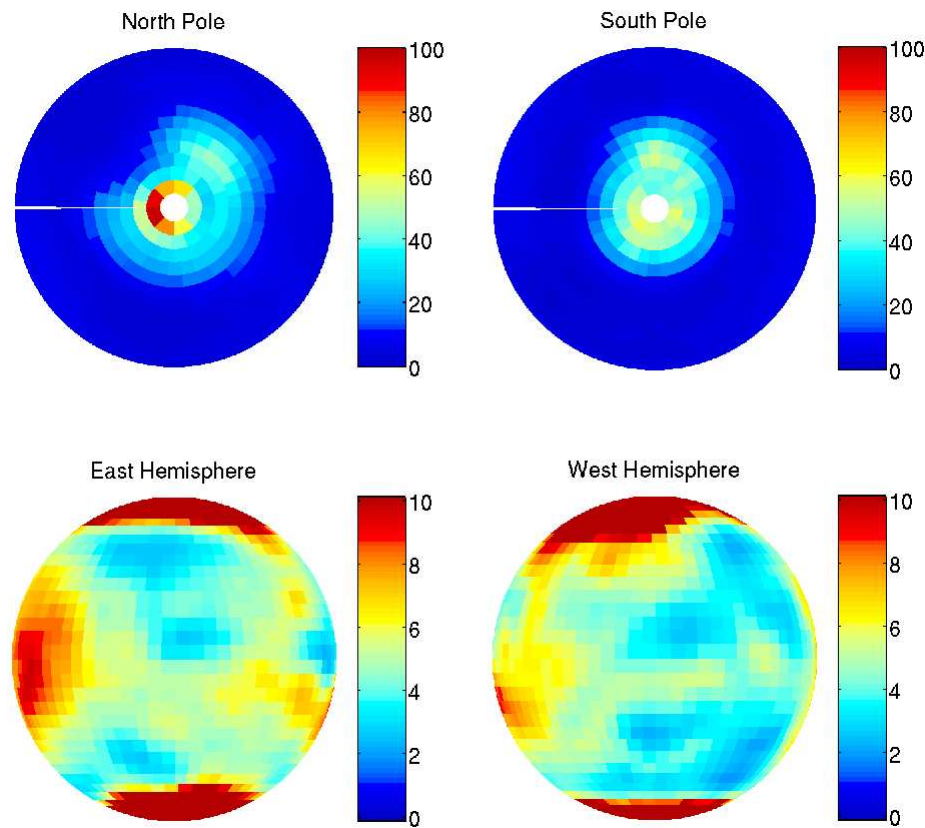
**Figure 3.9:** Zonally averaged total water column abundance through the Martian year. The colour bar indicates the amount of water present in  $[\text{Pr}\mu\text{m}]$ .

Bass and Paige [2000] based on the original Viking MAWD dataset. The total water column abundance was binned and averaged over a  $5^\circ$  resolution latitude-longitude mesh identical to the one used for the MCD, in order to have data of identical format for later calculations. In Figure 3.9, the zonal mean values of vapour column abundance are shown as a function of season and latitude. The unit of the data is  $\text{Pr}\mu\text{m}$  (precipitable microns), which is a measure for the height of the collapsed total water column in water equivalent microns. The geographical distribution of vapour throughout the Martian year can be seen in Figure 5.3.

### 3.6 The Ground Ice in the Regolith

The Mars Odyssey Gamma Ray Spectrometer suite of instruments has detected the water content of the near subsurface on Mars [Feldman et al., 2002, Mitrofanov et al., 2002, Boynton et al., 2002, Mitrofanov et al., 2003]. The original data set consisted of binned values in a grid that is resolved in  $5^\circ$  in latitude and has a varying resolution in longitude in terms of degrees but quite similar in terms of length. If the binning had been equally spaced in longitude, the grid would be too narrow to allow for a statistically valid number of data points in the binning of data [Feldman et al., 2004]. The data is a combined set of summertime measurements at both northern and

southern high latitudes as well as the average of the whole data set at mid-to equatorial latitudes. This is needed because during the winter the CO<sub>2</sub> frost cover obscures the measurements, wherefore the map in Figure 3.10 is not representative for a single time shot of Mars, but represents what the neutron detector would measure in a frost-free situation.



**Figure 3.10:** Ground Ice percentages from the MGS Neutron Spectrometer. The colour bars show the values of ground ice estimated from water-equivalent-hydrogen in mass-%. Notice the different scales for the polar- (upper panel) and low-latitude plots (lower panel). For the latter two, the scale is saturated for values greater than 10%. The globes each show a part of the Martian surface. UL: The north pole region down to equator. The 0° meridian is pointing from the pole to the left. UR: The south pole region down to equator. The 0° meridian is pointing from the pole to the left. LL: The eastern hemisphere. The 0° meridian is located along the left edge of the globe. LR: The western hemisphere. The 0° meridian is located along the right edge of the globe.



# 4

## Mass Balance of the North Polar Ice Cap

In order to better understand the current climate on Mars, it is important to investigate the state of the ice caps. Relevant questions to ask include: Are the ice caps gaining or losing mass? On which time-scales does the interaction with the atmospheric water occur? Do the ice caps flow? How can we extract the past climate history from the layers in the ice caps? The understanding of the interaction between today's water cycle and the ice caps may be the key to understanding the past buildup of the ice caps and thereby the past climate. For this thesis a new model has been developed in order to investigate the current state of the north polar ice cap.

On the northern ice cap of Mars, the pattern of areas gaining or losing mass is much more complicated than on a terrestrial ice sheet. It is heavily influenced by the scarp distribution. Accumulation occurs on the white areas of the ice cap, and ablation occurs from the scarps. The net mass balance of a point on a glacier is controlled by many different parameters. For high altitudes there is likely less deposition than at lower altitudes, however, there is also significantly less sublimation, which results in a positive net mass balance for higher altitudes. Near the pole there is less incoming solar radiation than at more southern latitudes, so the sublimation increases with distance from the pole. The local slope of the ice sheet also influences the incoming solar radiation, so a steep equatorward-facing slope has higher sublimation than a less steep one.

## 4.1 Current State of the North Polar Ice Cap

The northern ice cap has a maximum elevation of 3 km, and the highest point is within a few kilometres of the rotation pole [Zuber et al., 1998]. The cap has an extent of around 1000 km in diameter and a volume of  $1.2 \cdot 10^6$  to  $1.7 \cdot 10^6$  km<sup>3</sup>, which corresponds to approximately half the volume of the Greenland ice cap. The whole northern hemisphere lies 5 km below the areoid, which is the Martian equivalent of “sea level”, and locally the Martian surface outside the ice cap gently slopes downward toward the pole at all longitudes in the northern hemisphere [Zuber et al., 1998].

The main visual features of the northern ice cap of Mars are the dark spirals around a central white smooth looking area and a large valley or reentrant called Chasma Boreale. The spiral pattern consists of troughs, with steep southfacing slopes called scarps, which spiral outward in a counterclockwise direction from the pole. The troughs are up to 1 km deep according to Hvidberg [2003] and 200-800 m deep according to Fisher [1993]. They are 5-15 km wide, hundreds of kilometres long, and occur with a spacing of about 50 km between neighbouring troughs [Hvidberg, 2003]. The surface appears white along the horizontal and north-facing surfaces, while the steep south-facing scarps are dark due to numerous horizontal layers exposed with varying dust content. These layers are thought to reflect climatic variations in dust storm intensity and ice accumulation. The inclination of the steep slopes are  $10^\circ - 15^\circ$  according to Hvidberg [2003],  $5^\circ - 8^\circ$  according to Fisher [1993], and  $15^\circ - 20^\circ$  according to Zuber et al. [1998].

While it has been known for decades that the seasonal caps covering high latitudes during winters are composed of CO<sub>2</sub> frost, the composition of the permanent polar cap remained uncertain for much longer. Studies have shown that an ice cap consisting of CO<sub>2</sub> ice would have lower elevations and a flatter profile than an ice cap consisting of H<sub>2</sub>O ice [Nye et al., 2000], so it is now widely accepted that the major constituent is water ice. In addition, the cap may contain unknown amounts of dust and possible layers of solid CO<sub>2</sub>. Thermal mapping of the north polar cap with the Viking Orbiter Infrared Thermal Mapper (IRTM) showed that the summer temperature rises to 295 K, well above the 148 K sublimation temperature of CO<sub>2</sub> [Kieffer et al., 1976]. Thus the seasonal CO<sub>2</sub> cover is presently being completely removed during northern summer. In previous climate epochs, however, some CO<sub>2</sub> may have survived the summer, and layers of CO<sub>2</sub> may still remain within the cap.

Winter snow accumulation on the northern cap is thought to occur when H<sub>2</sub>O ice crystals with dust particles at their centres pick up sufficient CO<sub>2</sub> condensate to become heavy enough to fall. In the summer, the CO<sub>2</sub> sub-

limates away, leaving a mixture of water ice and dust [Pollack et al., 1979]. Summer ablation is thought to occur by preferential sublimation from the steep sunward facing scarps. Some of the sublimated water is thought to recondense on the gentler ( $1^\circ - 2^\circ$ ) north facing colder high-albedo slopes. [Fisher, 1993]. As the ablation face recedes it becomes dirty with exposed dust [Howard, 1978, Howard et al., 1982].

Topographic profiles measured by Mars Orbiter Laser Altimeter (MOLA) across the ice cap show a sharp rise by about 1000 m at the edge of the ice cap. MOLA, which operates optimally in a nadir-oriented configuration, is an instrument on the Mars Global Surveyor (MGS). However, the orbit of the MGS is inclined in relation to the rotation axis, which results in an  $\sim 450$  km diameter gap in coverage, centred on the pole. In order to obtain measurements of the topography at latitudes above  $86.3^\circ\text{N}$ , the MGS spacecraft was tilted  $\sim 50^\circ$  for 10 tracks [Zuber et al., 1998].

The layers seen in the polar ice caps have been mapped and investigated in several studies [Byrne and Murray, 2002, Milkovich and Head, 2004]. The same layers have been identified in different scarps indicating that the layers have a wide extent and may even be present throughout in the ice caps; however, it is not clear whether this is the case for all layers. Fisher [1993] suggested that the scarps may consist of alternating young and old layers. This would cause the chronology to be broken by unconformities, however such features have not been identified in the ice caps yet, except for the boundary between the distinct basal unit identified by Fishbaugh and Head [2005] and the more ice-rich layers of the bottom part of the ice cap.

Hvidberg [2003] made a 2D flow model study of the north polar ice cap. In this study it was investigated how fast the scarp pattern would close due to ice flow. In order for the scarps to maintain their present morphology it was found that a sublimation rate of 5 mm ice per year was necessary at the dark scarps.

A recent study by Ng and Zuber [2005] proposed that the scarp pattern is caused by an instability in the ice/dust-albedo feedback. The dust is constantly blown onto the surface of the ice sheet, but will be covered by fresh ice at flat surfaces and uncovered at steeper surfaces, revealing dust layers imbedded in the ice sheet.

The Thermal Emission Spectrometer (TES) on board Mars Global Surveyor has measured the total column abundance of water vapour in the Martian atmosphere during 3 Mars years. The data shows the highest peak in the water vapour content at the north pole during mid summer, coinciding with the highest temperatures. At lower latitudes than at the pole, the highest concentration occurs at a later time of the year. This delay could either

be due to a transport of vapour from high to low latitudes or due to later appearance of water from the subsurface at the low latitudes. In Smith [2002] it was reported that the polar ice caps are likely to be the sources for the atmospheric vapour peak seen in the summer hemispheres.

## 4.2 A Model Study of the North Polar Ice Cap

A new model has been constructed for the north polar ice cap on Mars. The mass balance is parameterised so that it reflects variation in insolation and exchange of vapour with the atmosphere. In order to investigate the present state of the ice cap, the flow velocities for the ice cap are calculated under the assumption of steady state.

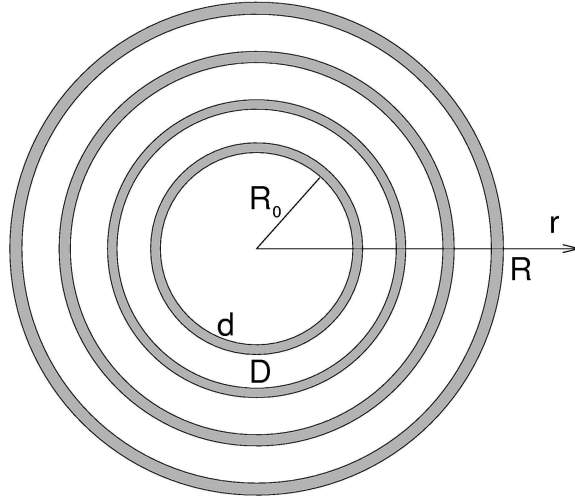
### 4.2.1 Mass Balance

For the local net mass balance a simple model is proposed, which includes the effects of the altitude, the latitude, and the slope of the surface. The mass balance  $b$  is given by

$$b = \alpha \cdot r + \beta \cdot s(r) + \gamma \cdot h(r) \quad (4.1)$$

where  $r$  is the radial distance from the centre of the ice cap,  $s$  is the local surface slope of the ice cap,  $h$  is the height of the ice cap, and  $\alpha$ ,  $\beta$ , and  $\gamma$  are tunable parameters. The mass balance  $b$  is positive when there is net accumulation (deposition) and negative for net ablation (evaporation).

The distance from the centre of the ice cap  $r$  is directly linked to the latitudinal variation of the insolation. The longer distance from the pole, the more insolation, which gives a negative contribution to the net mass balance due to enhanced evaporation, hence  $\alpha$  has a negative value. The slope  $s$  is defined to be negative when the altitude decreases with the radial distance from the centre of the ice cap, i.e. when the slopes are southfacing. A steeper equatorward slope will enhance the local insolation and therefore give an overall negative contribution to the net mass balance, whereas a poleward slope will protect the ice from sublimating away, hence  $\beta$  has a positive value. The local height of the ice cap  $h$  determines the annual average temperature. The temperature decreases with altitude and therefore the height of the ice cap  $h$  will ensure less evaporation at high altitudes. The height will thus have a positive effect on the net mass balance, and  $\gamma$  has a positive sign.



**Figure 4.1:** The geometry of the new mass balance model. The geometrical parameters shown in the figure are: The radial distance from the centre of the ice cap  $r$ , the radius of the ice cap  $R$ , the radius of the central white area  $R_0$ , the width of scarps (grey areas)  $d$ , and the distance between scarps  $D$ .

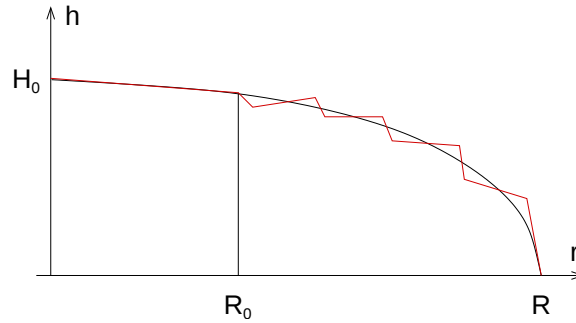
#### 4.2.2 Ice Cap Geometry

The geometry of the north polar ice cap is simplified in the model to a circular form seen from above, see Figure 4.1. In the central inner white area, net deposition is believed to occur. This area corresponds to 20% of the area of the whole ice cap [Fisher, 2000]. Hence, for an ice cap with radius  $R = 500$  km, the radius of the central inner area corresponds to a value of  $R_0 = 220$  km. On the remaining parts of the ice cap there is either net sublimation from the scarps (indicated by the grey areas in Figure 4.1), or deposition on the flat white areas in between.

The ice cap is assumed to have a profile of an ideal plastic ice sheet, which is the usual assumption for terrestrial ice sheets [Paterson, 1994]. The height  $h$  of a cylindrical symmetric ideal plastic ice sheet as a function of the radius  $r$  is given by

$$h^2 = \frac{2\tau_0}{\rho g} (R - r) = \frac{H_0^2}{R} (R - r) \quad (4.2)$$

where  $\tau_0$  is the yield stress at the base of the ice sheet,  $\rho$  is the density of the ice,  $g$  is the gravitational acceleration,  $R$  is the maximum extent of the ice sheet, and  $H_0$  is the central maximum height of the ice sheet.



**Figure 4.2:** The slopes used in the model (red lines) overall follow the ideal plastic profile (black line) but are steeper for the scarps and less steep for the flat white areas. The vertical axis is exaggerated.

The slope of the ideal plastic ice cap is

$$\frac{dh}{dr} = \frac{d}{dr} \left( \sqrt{\frac{H_0^2(R-r)}{R}} \right) = -\frac{1}{2} \sqrt{\frac{H_0^2}{R(R-r)}} \quad (4.3)$$

The slope at a given point of the ice cap in the model depends on the point being on a scarp or not. The local slope at a given point in the model is given by

$$s = \frac{dh}{dr} + s_i \quad (4.4)$$

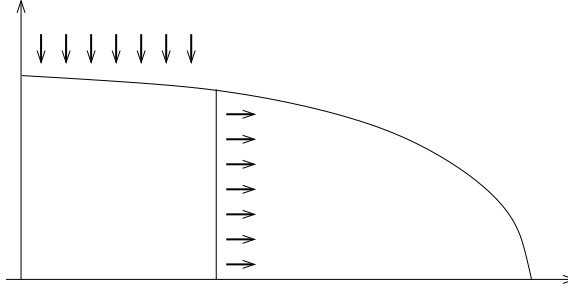
where  $s_i = s_s$  for scarps and  $s_i = s_n$  for non-scarp areas outside  $R_0$ . For the central region, where no scarps are present, the slope of the ice cap follows the one of the ideal plastic ice cap, as seen in the schematic in Figure 4.2.

The ice sheet is assumed to be in steady state everywhere. This means that what is deposited on the surface of the ice cap north of a certain latitude, will flow out through the cylindrical cross-sectional area at the same latitude, as illustrated in Figure 4.3. For an ice sheet in mass balance, the total net accumulation of an area north of a given  $r$ -value must thus be equal to the amount of ice moved by the horizontal velocity  $u$  through the surface area of the cylinder with height  $h(r)$  and circumference  $2\pi r$ . This gives the following expression of the mean horizontal velocity.

$$u(r) = \frac{\int_0^r 2\pi r \cdot b(r) dr}{h(r) \cdot 2\pi r} \quad (4.5)$$

### 4.2.3 Model Experiments

Several experiments were made in order to investigate the sensitivity of the model to the different geometrical parameters illustrated in Figure 4.1



**Figure 4.3:** Cross section of the centre to the edge of the ice cap. Steady state is obtained when the mean horizontal flow for a given radius balances the total net mass balance of the ice cap poleward of the point.

**Table 4.1:** Mean deposition and sublimation rates from various calculations and mass balance models

| Deposition rate          | Sublimation rate         | Reference                 |
|--------------------------|--------------------------|---------------------------|
| $2.6 \cdot 10^{-4}$ m/yr |                          | [Pathare and Paige, 2005] |
|                          | $0.3 \cdot 10^{-3}$ m/yr | [Jakosky et al., 1993]    |
| $0.5 \cdot 10^{-3}$ m/yr | $5 \cdot 10^{-3}$ m/yr   | [Hvidberg, 2003]          |
| $0.1 \cdot 10^{-3}$ m/yr | $0.3 \cdot 10^{-3}$ m/yr | [Fisher, 2000]            |

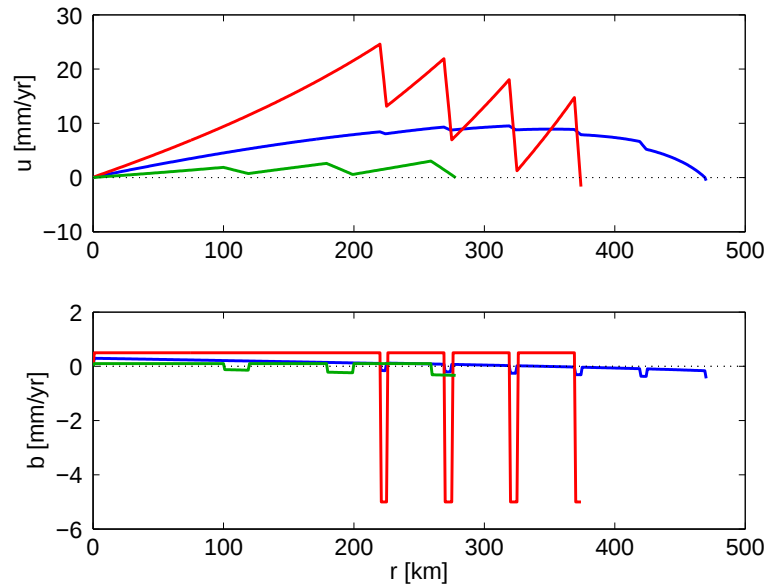
and Figure 4.2. Likewise, the influence of the climatological parameters in Equation 4.1 on the results was investigated.

The present day mass balance has been estimated by several authors. Two different models for mass balance and two calculations of the present rates of deposition and evaporation have been used in this study. These mean deposition and sublimation rates for the models are shown in Table 4.1.

The deposition and evaporation rates in Table 4.1 were used to study the influence of the climatological parameters on the ice cap in the model. For this study the evaporation rate from Pathare and Paige [2004] and the deposition rate from Jakosky [1993] have been chosen for estimation of the parameters  $\alpha$ ,  $\beta$ , and  $\gamma$ . If the height of the ice cap is in the interval  $h = [0, H_0]$  km and the effects of radius and slope for now are disregarded,  $\gamma$  can be estimated to be in the order of

$$\gamma \sim \frac{\text{dep. rate}}{H_0} = \frac{0.3 \cdot 10^{-3} \frac{\text{m}}{\text{yr}}}{3 \text{ km}} = 1 \cdot 10^{-7} \text{ yr}^{-1} \quad (4.6)$$

$\alpha$  and  $\beta$  are estimated in a similar way, with the maximum sublimation rate instead of deposition. If the radius of the ice cap is in the interval  $r = [0, R_0]$



**Figure 4.4:** Top panel: Mean horizontal velocity for the deposition and sublimation rates listed above. Lower panel: Net mass balance for the same calculations. The radius of the steady state ice cap is:  $R = 470$  km for the new mass balance model (blue line),  $R = 374$  km [Hvidberg, 2003] (red line), and  $R = 278$  km [Fisher, 2000] (green line).

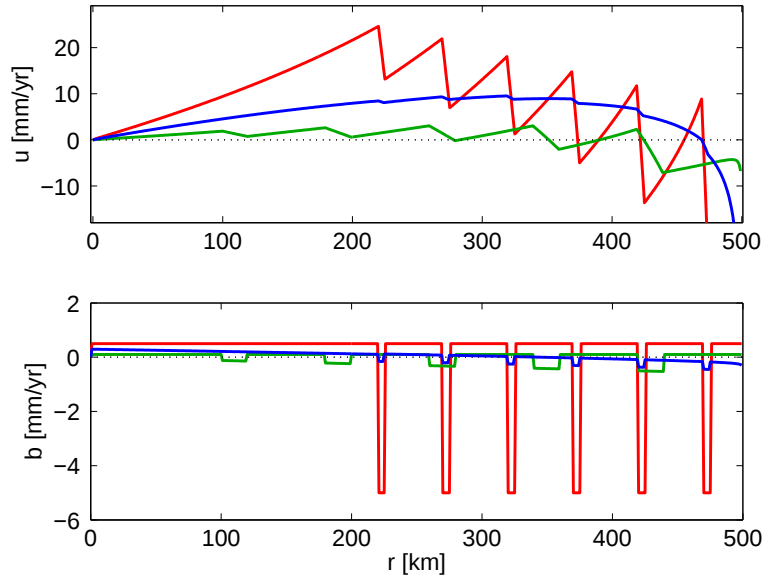
km,  $\alpha$  is estimated to be

$$\alpha \sim \frac{\text{sub. rate}}{R_0} = \frac{-2.6 \cdot 10^{-4} \frac{\text{m}}{\text{yr}}}{500 \text{ km}} = -5 \cdot 10^{-9} \text{ yr}^{-1} \quad (4.7)$$

### 4.3 Results

The mean horizontal flow velocity  $u$  is calculated for the three different mass balance patterns using Equation 4.5. The radius  $R$  of the ice cap is defined to be the value of  $r$ , where  $u$  becomes zero or negative. From Figure 4.4, it can be seen that the new model for the mass balance produces an ice cap in steady state of similar size to the present ice cap. For these values the ice cap seems to be close to steady state under present conditions.

The new model used with the sublimation rate from Pathare and Paige [2004] and deposition rate from Jakosky et al. [1993] shows that the ice cap may be close to steady state at the moment. An estimation of the values of  $\alpha$ ,  $\beta$ , and  $\gamma$  was made such that the model would give a value of the distance from the centre of the ice cap where  $u$  is 0, is close to the actual extent of



**Figure 4.5:** Top panel: Mean horizontal velocity for the deposition and sublimation rates listed above. The calculations are continued to the edge of the present ice cap at  $r = 500$  km. Lower panel: Net mass balance for the same calculations. Colour coding: Same colour coding as figure 4.4.

the north polar ice cap. These parameters were then used with the model and the results compared to the mass balance patterns used by Fisher [2000] and Hvidberg [2003].

The mass balance models by Hvidberg [2003] and Fisher [2000] assume there is a net zero mass balance and estimate the deposition or sublimation rate from the other. Their mass balances are thus tuned to yield an ice cap near steady state. For the two mass balances by Hvidberg and Fisher, there are several intervals where the calculated ice velocity drops to zero or negative values. These intervals can be interpreted as disconnected zones from the central part of the active ice cap. The zones arise if the calculations are forced to cover an ice cap of the present size at  $R = 500$  km in radius.

A sensitivity study of the influence of the geometrical parameters was made within the intervals listed in Table 4.2. The geometrical parameters tested are: The area of the inner white area (inside  $r = R_0$ ), the width of scarps  $d$  and the distance between them  $D$ , the slope  $s$  of scarps and the areas between them. The model showed minor sensitivity against changes in extent of the ice cap  $R$  and the radius of the central smooth area  $R_0$ . However, the model was not as robust when testing it by varying the inclination of the surface  $s$  and the width of the scarps  $d$ . Based on Mars observations and

**Table 4.2:** The intervals in which the geometrical parameters have been varied and the chosen standard values.

|       | Description                     | Interval                 | Standard value |
|-------|---------------------------------|--------------------------|----------------|
| $D$   | Distance between scarps         | 10–100 km                | 50 km          |
| $d$   | Width of scarps                 | 1–15 km                  | 5 km           |
| $R_0$ | Radius of central area          | 11–500 km                | 220 km         |
| $s_s$ | Additional slope on scarp       | $-5^\circ$ – $-20^\circ$ | $-15^\circ$    |
| $s_n$ | Additional slope between scarps | $0^\circ$ – $2^\circ$    | $1^\circ$      |
| $H_0$ | Central height of ice cap       | 2–4 km                   | 3 km           |
| $R$   | Radius of ice cap               | 300–600 km               | 500 km         |

the sensitivity studies a standard value for each parameter was chosen and used for the remainder of this study.

When altering the climatological parameters the effects described in the following are expected. For a decrease in  $\alpha$ , an enhanced latitude effect is expected, which means that the sublimation is more effective, leading to a decreased net mass balance. Increasing  $\gamma$  enhances the altitude effect, which gives higher accumulation in the central parts of the ice cap. The combination of the changes of these two parameters makes the slope of the net mass balance steeper. The steeper slope of the mass balance causes the velocity field to be more effective with higher ice flow velocities in order to keep up with the enhanced accumulation in the central parts and the enhanced sublimation near the edge of the ice cap.

## 4.4 Conclusion

A new model was made for the north polar ice cap on Mars. The model incorporates the influence of the geometrical shape of the ice cap on the mass balance.

The model was tested for the individual importance of the geometrical parameters, and the far most important ones are those controlling the total scarp area,  $d$  and  $D$ . This implies that factors related to the ablation such as the surface inclination (and most likely the albedo, which has not been included in this study) probably are the most important factors in controlling the ice cap. The distinct pattern of spirals on the Martian ice caps is unique compared to glaciers and ice caps on Earth and is likely a governing control of the present state of the ice cap.

It cannot be definitely concluded whether the northern ice cap on Mars is

presently in steady state or in imbalance, but the new calculations using the deposition and sublimation rate from Pathare and Jakosky suggest that the ice cap is near mass balance, however flowing slowly, at the present stage. When the deposition and sublimation rates from Hvidberg and Fisher are used, the calculations suggest that the ice cap at present is stagnant or receding.

The fact that the ice cap presently is stagnant or receding shows that the mass balance must have been different in earlier times in order to form the ice cap initially. We have seen that by changing the tunable parameters  $\alpha$ ,  $\beta$ , and  $\gamma$  it is possible to increase the flow velocities. A more vigorous ice flow during build up would have given the ice cap its overall shape, which it still has contained even though it is now cut through by scarps.

An enhanced accumulation at high altitude in the central parts of the ice cap and an enhanced ablation near the edges at lower elevations and latitudes would result in the net mass balance  $b$  being a steeper function of the radius  $r$ . This would then lead to higher flow velocities in order for the ice cap to achieve steady state. The enhanced ablation at lower latitudes could be caused by a stronger insolation, likely caused by a change in Mars' orbital parameters. The enhanced accumulation at high altitudes near the north pole could have been caused by a higher atmospheric content of water vapour, which might be a consequence of a stronger sublimation at lower latitudes.



5

# Atmospheric Water Transport — Mechanisms

In this chapter the relative importance of different mechanisms for meridional transport of water vapour is identified. In order to examine the water fluxes in the atmosphere during the Martian year, wind data from the MCD and MAWD measurements of the water vapour column are analysed with an empirical statistical method by Peixoto and Oort [1992].

The wind system in the Martian atmosphere consists of several components: The Hadley circulation; the  $\text{CO}_2$  condensation flow; the planetary waves, due to atmospheric instabilities in the zonal mean temperature and wind fields; and stationary planetary waves, caused by eastward zonal winds and the large scale topography [Leovy, 2001, Tokano, 2005].

As the wind field and atmospheric water content vary in time and geographical distribution, the resulting transportation of vapour will change. In addition to the wind, the transport of water depends on the timing and location for exchange of vapour between the atmosphere and the surface reservoirs.

## 5.1 The Empirical Method by Peixoto and Oort

In order to examine mechanisms of the seasonal meridional transport of water in the atmosphere, an empirical statistical method by Peixoto and Oort [1992] has been used. The meridional water vapour flux is decomposed into three parts: MM — mean meridional, TE — transient eddies, and SE — standing eddies.

When subdividing the water vapour transport for Earth, most of the water vapour transport at low latitudes takes place with the mean meridional motion (Figure 5.1). The transient and the stationary eddies each contribute with smaller amounts, however located at higher latitudes. The transient eddies and the mean meridional circulation are close to being hemispherically symmetric, whereas the stationary eddies transport far more water in the northern hemisphere than in the southern.

### 5.1.1 Definitions of Averages

The time average of a quantity is given by

$$\overline{A} \equiv \frac{1}{\tau} \int_0^{\tau} A dt \quad (5.1)$$

and the deviation from the average is denoted as  $A'$ . The instantaneous value of  $A$  is

$$A = \overline{A} + A' \quad (5.2)$$

where it can be noted that  $\overline{A'} = 0$ .

The zonal average of a quantity is given by

$$[A] \equiv \frac{1}{2\pi} \int_0^{2\pi} A d\lambda \quad (5.3)$$

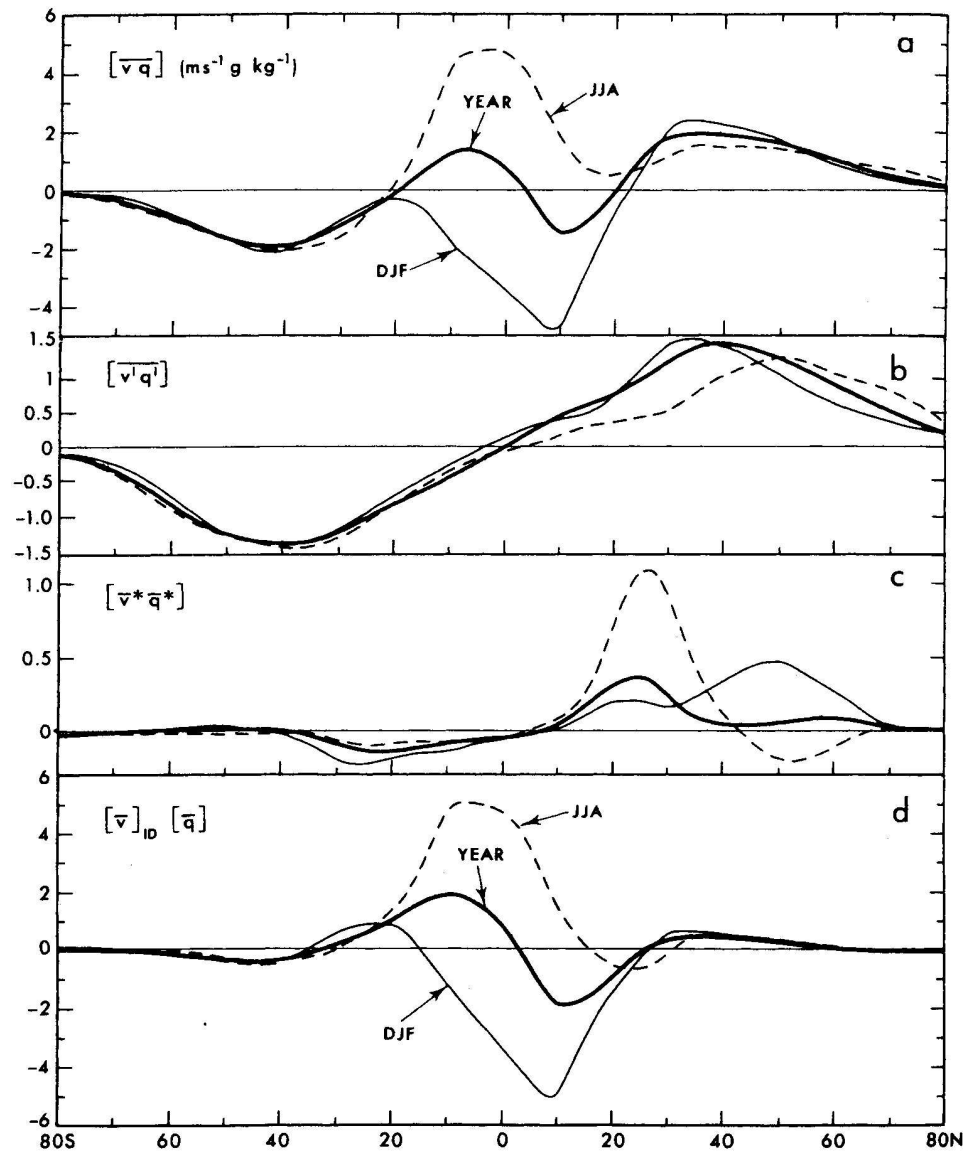
where  $\lambda$  is longitude and the deviation from the average is denoted as  $A^*$ . The instantaneous value of  $A$  is

$$A = [A] + A^* \quad (5.4)$$

where it can be noted that  $[A^*] = 0$ .

The two types of averaging are permutable so

$$[\overline{A}] = \overline{[A]} \quad (5.5)$$



**Figure 5.1:** The zonally averaged water cycle on Earth decomposed into various fluxes and times during the year. In all the subplots the thick solid line shows the annual result, the dashed line the northern summer, and the thin solid line the northern winter. a) The total meridional flux of vapour. b) The flux resulting from the transient eddies. c) The flux resulting from the standing eddies. d) The flux resulting from the mean meridional motion. From [Peixoto and Oort, 1992].

### 5.1.2 Partitioning of Fluxes

The time average of the product of two quantities is

$$\overline{AB} = \overline{AB} + \overline{A'B'} \quad (5.6)$$

The zonal average of the product of two quantities is

$$[AB] = [A][B] + [A^*B^*] \quad (5.7)$$

Taking the time average of the zonal averaged product gives

$$[\overline{AB}] = [\overline{A}][\overline{B}] + [\overline{A'}][\overline{B'}] + [\overline{A^*B^*}] \quad (5.8)$$

where  $[\overline{A}][\overline{B}]$  is the mean meridional circulation MM,  $[\overline{A'}][\overline{B'}]$  is the transient mean meridional circulation, and  $[\overline{A^*B^*}]$  is the spatial eddy circulation.

Taking the zonal average of the time averaged product gives

$$[\overline{AB}] = [\overline{A}][\overline{B}] + [\overline{A^*B^*}] + [\overline{A'B'}] \quad (5.9)$$

where  $[\overline{A}][\overline{B}]$  is the mean meridional circulation MM,  $[\overline{A^*B^*}]$  is the stationary eddy circulation SE, and  $[\overline{A'B'}]$  is the transient eddy circulation TE.

### 5.1.3 Derivation of the Decomposed Transport

The product of the two quantities  $A$  and  $B$  can be expanded with their zonal averages and corresponding deviations yielding

$$AB = ([A] + A^*)([B] + B^*) = [A][B] + [A]B^* + A^*[B] + A^*B^* \quad (5.10)$$

Taking the zonal average of the product eliminates the products with only one zonal deviation and gives us the result in Equation 5.7, which is then expanded with the time averages and deviations giving

$$\begin{aligned} [AB] &= [A][B] + [A^*B^*] \\ &= ([\overline{A}] + [A'])([\overline{B}] + [B']) + [A^*B^*] \\ &= [\overline{A}][\overline{B}] + [\overline{A}][B'] + [A'][\overline{B}] + [A']B' + [A^*B^*] \end{aligned} \quad (5.11)$$

Taking the time average of this eliminates the terms with only one time deviation and leaves us with the following decomposition of the fluxes, which is the same as written in Equation 5.8. This leads to

$$[\overline{AB}] = [\overline{A}][\overline{B}] + [\overline{A'}][\overline{B'}] + [\overline{A^*B^*}] \quad (5.12)$$

In order to derive Equation 5.9 we follow a similar approach as in the derivation of Equation 5.8. This time we begin with the time average expansion and introduce the zonal average in the step after this, which gives

$$AB = (\bar{A} + A') (\bar{B} + B') = \bar{A}\bar{B} + \bar{A}B' + A'\bar{B} + A'B' \quad (5.13)$$

The terms with only one time deviation are eliminated when we take the time average of the product, which is the result seen in Equation 5.6, and further expand it with the zonal averages, which gives

$$\begin{aligned} \overline{AB} &= \overline{AB} + \overline{A'B'} \\ &= ([\bar{A}] + \bar{A}^*) ([\bar{B}] + \bar{B}^*) + \overline{A'B'} \\ &= [\bar{A}] [\bar{B}] + [\bar{A}] \bar{B}^* + \bar{A}^* [\bar{B}] + \bar{A}^* \bar{B}^* + \overline{A'B'} \end{aligned} \quad (5.14)$$

Taking the zonal average of this eliminates the terms including only one zonal average, and leaves us with the following decomposition of the fluxes

$$[\overline{AB}] = [\bar{A}] [\bar{B}] + [\bar{A}^* \bar{B}^*] + [\overline{A'B'}] \quad (5.15)$$

This expression is the same as written in Equation 5.9 and is the one used for the flux calculations for Mars in the following section.

## 5.2 Water Fluxes on Mars

In order to decompose the meridional water flux on Mars, data sets of the atmospheric water content as well as the meridional wind field are needed as functions of location and time. The two data sets used for the flux calculations are the MAWD data set (Viking — Mars Atmospheric Water Detector), which is described in section 3.5, and the meridional wind field from the Mars Climate Database (MCD) based on the LMD/AOPP Mars GCM [Forget et al., 1999, Lewis et al., 1999] described in section 3.2.

The meridional wind is given as  $v(\phi, \lambda, z, L_s)$  [ $\frac{m}{s}$ ], where  $\phi$  is latitude,  $\lambda$  is longitude,  $z$  is altitude, and  $L_s$  is the time of year. The wind data has a resolution of  $5^\circ$  in latitude and longitude as well as 32 vertical layers of varying altitude. The values are given for 12 different times during Martian year in order to represent a full annual cycle as described in section 3.1, and have been diurnally averaged before being used in the calculations. In Figure 5.2, the vertically averaged values of  $v$  are shown as function of latitude, longitude, and time of year.

The water vapour is given as the total column abundance  $w(\phi, \lambda, L_s)$  [ $\text{Pr}\mu\text{m}$ ], where the unit  $\text{Pr}\mu\text{m}$  or *precipitable microns* corresponds to the height of

**Table 5.1:** Altitude of some vertical layers from the MCD.

| Number of layer, $N_z$           | 10  | 11   | 12    |
|----------------------------------|-----|------|-------|
| Altitude at centre of layer [km] | 7   | 9.6  | 14.2  |
| Thickness of layer [km]          | 3.1 | 3.6  | 4.95  |
| Altitude top of layer [km]       | 8.3 | 11.9 | 16.85 |

the total water column in water equivalent. One  $\text{Pr}\mu\text{m}$  corresponds to a volume of  $10^{-6}\text{m} \cdot 1\text{m}^2 = 10^{-6}\text{m}^3$  water, which has a mass of  $10^{-3}$  kg. So 1  $\text{Pr}\mu\text{m}$  corresponds to  $10^{-3} \frac{\text{kg}}{\text{m}^2}$ . In Figure 5.3, the total vertical column of water vapour  $w$  is shown as function of latitude, longitude, and time of year.

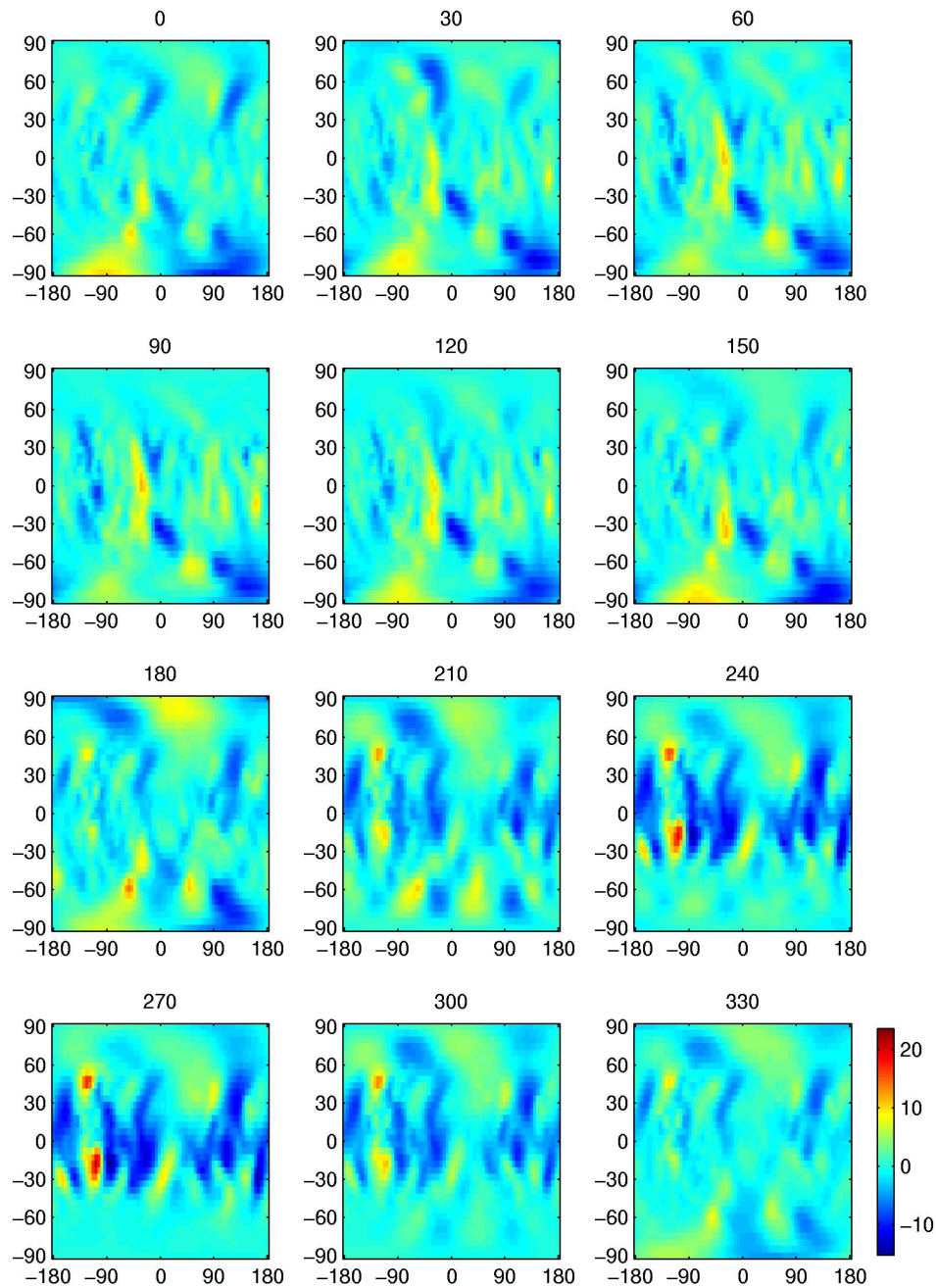
The water is assumed to move as a passive tracer with the wind. At a certain location the wind may be going at different strengths and possibly different directions at two different levels of altitude, so in order to calculate the transport of vapour with the wind, the vertical distribution of the water has to be known.

### 5.2.1 Vertical Distribution of Vapour

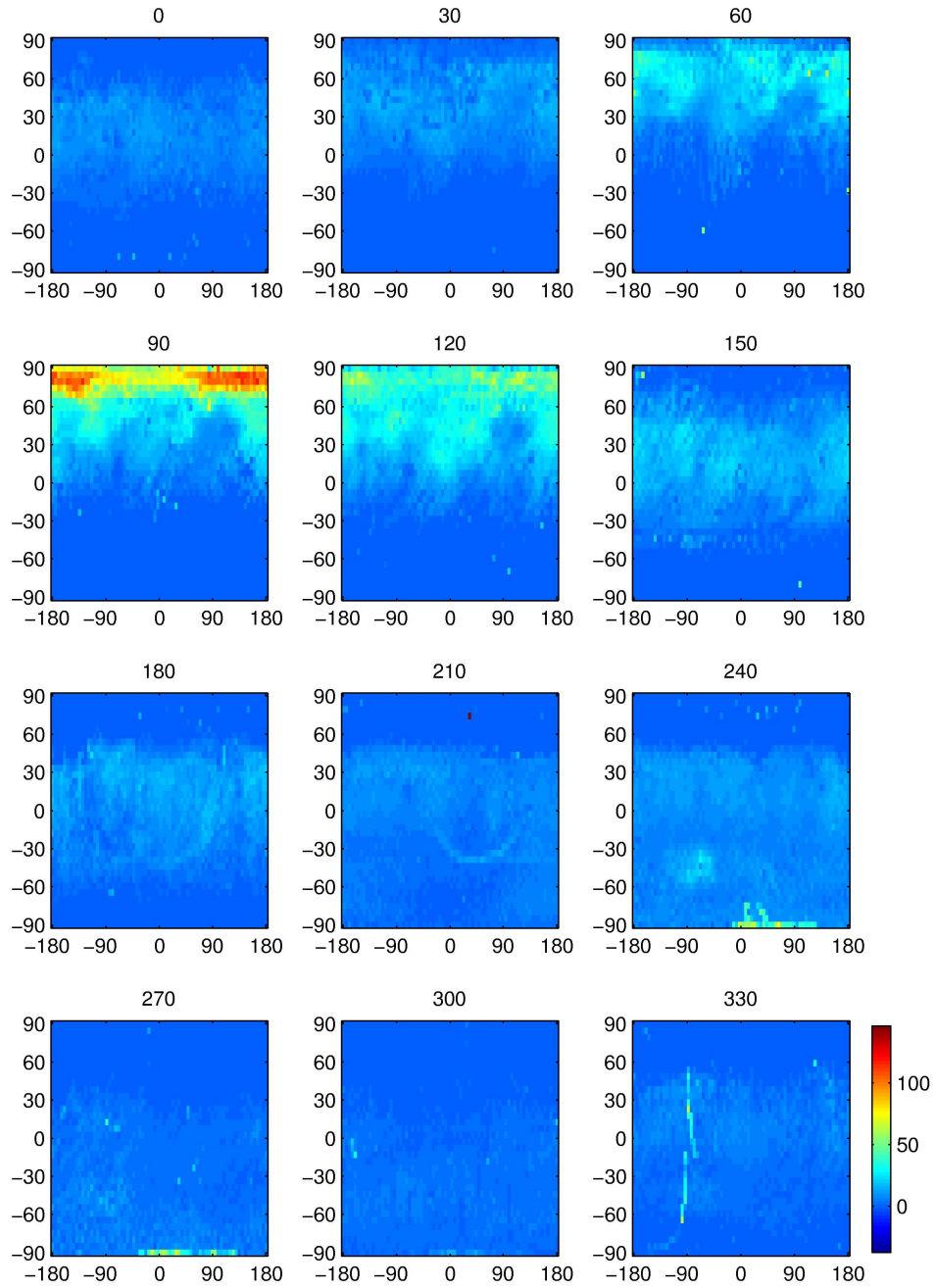
Earthbased studies have shown that the water vapour is located in the lower 25–30 km of the atmosphere on Mars [Davies, 1979, Rodin et al., 1997, Gurwell et al., 2000, Encrenaz et al., 2001]. Davies [1979] states on the basis of Viking MAWD data that the vapour in the Martian atmosphere is distributed fairly uniformly in the lowest 10 km, and that it is unlikely to be uniformly mixed above 15 km altitude.

In the MCD database, the vertical layer of  $N_z = 10$  is centred at an altitude of 7 km and has a thickness of 3.1 km. The altitude of the top of this layer is therefore at 8.3 km, which corresponds to the column height of the lowest 10 layers. The same calculation has been made for layers 11 and 12 and can be seen in Table 5.1. These altitudes are chosen as the maximum altitude for vertical mixing of water vapour for many of the calculations in this study, since they correspond to the most likely values of 10–15 km altitude mentioned earlier. The total amount of water vapour  $w$  for a given location and time of year is assumed to be distributed over the vertical layers with a constant mixing ratio.

The distance between the atmospheric layers in the vertical direction from the MCD is neither uniform nor linearly depending on the altitude. The layers are closer near the surface, where greater details are needed for the



**Figure 5.2:** The vertically averaged wind data from the MCD shown for each of the 12 seasons. The number above plot is the start of the season given in  $^{\circ}L_s$ . The vertical average was made over the 10 lowermost layers in the vertical direction. The horizontal scale denotes the longitude, and the vertical the latitude. The colorbar shows the wind velocity in m/s, where positive values are northward.



**Figure 5.3:** The total water vapour column from MAWD shown for each of the 12 seasons. The number above plot is the start of the season given in  $^{\circ}L_s$ . The horizontal scale denotes the longitude, and the vertical the latitude. The colorbar shows the atmospheric water content in  $\text{Pr}\mu\text{m}$ .

calculations, and grow with altitude until about 10 km altitude, where the layer spacing becomes approximately uniform. The altitude of the “centre of layer” in Table 5.1 thus refers to a weighted altitude level between the top and bottom of the given layer.

In this study, the assumption is made that the water vapour for each latitude band is distributed in the air column such that all the water is contained in the atmospheric column from the surface and up to a certain level  $z_{TOW}$  (top of the water column). The mixing ratio of water with respect to Martian air is constant within this region.

The ideal gas law states

$$pV = NRT \quad (5.16)$$

where  $p$  is pressure [Pa],  $V$  is the volume of the gas [ $\text{m}^3$ ],  $N$  is the number of molecules in the air measured in mol,  $R$  is the universal gas constant, and  $T$  is the temperature [K]. The mass density  $\rho$  of air [ $\frac{\text{kg}}{\text{m}^3}$ ] can be found from the ideal gas law as

$$\rho = \frac{Mp}{RT} \quad (5.17)$$

where  $M$  is the molecular weight of the air [ $\frac{\text{kg}}{\text{mol}}$ ].

Dalton’s law states that the partial pressure  $p_Y$  of a gas  $Y$  in a mixture of gasses with pressure  $p$  is proportional to the mixing ratio  $X_Y$  [ $\frac{\text{mol}}{\text{mol}}$ ] of the gas

$$p_Y = X_Y p \quad (5.18)$$

The water vapour concentration by mass  $C$  in the Martian air is identical to the density of the  $\text{H}_2\text{O}$  gas  $\rho_{\text{H}_2\text{O}}$  that can be calculated from Dalton’s law, which gives the expression

$$C = \frac{M_{\text{H}_2\text{O}} p_{\text{H}_2\text{O}}}{RT} = \frac{p X_{\text{H}_2\text{O}}}{R_{\text{H}_2\text{O}} T} \quad (5.19)$$

where  $R_{\text{H}_2\text{O}} = R/M_{\text{H}_2\text{O}}$  is the gas constant for water vapour.

The total column mass of water  $m_{\text{H}_2\text{O}}$  [ $\frac{\text{kg}}{\text{m}^2}$ ] can be found as

$$m_{\text{H}_2\text{O}} = \frac{1}{A_s} \int_{z=0}^{z=z_{TOW}} C dV \quad (5.20)$$

where  $A_s$  is the surface area of the bottom of the column.

The expression for the total column mass of water (Equation 5.20) can be combined with the density of air (Equation 5.17) and approximated with a sum over the  $N_z$  vertical layers from the surface and up to  $z_{TOW}$ , which is the altitude limit of the vapour. From this combination, a discrete expression

for the total column mass of water can be found under the assumption of the mixing ratio being constant with altitude

$$\begin{aligned}
 m_{H_2O} &= \frac{1}{A_s} \sum_{k=1}^{N_z} C_k V_k \\
 &= \frac{1}{A_s} \sum_{k=1}^{N_z} \frac{p_k X_{H_2O}}{R_{H_2O} T_k} \Delta z(i_z) \\
 &= X_{H_2O} R_{H_2O}^{-1} \frac{1}{A_s} \sum_{k=1}^{N_z} \frac{p_k}{T_k} V_k
 \end{aligned} \tag{5.21}$$

The mixing ratio can then be calculated from the total column mass of water  $m_{H_2O}$  in combination with the atmospheric temperature and pressure profiles from the MCD.

$$X_{H_2O} = m_{H_2O} R_{H_2O} \left( \frac{1}{A_s} \sum_{k=1}^{N_z} \frac{p_k}{T_k} V_k \right)^{-1} \tag{5.22}$$

For a given latitude and a given time of the year, the mixing ratio can then be calculated, based on the columnar mass of water vapour at the given point in time and space. By use of Equation 6.45, the vertical distribution of the vapour can then be calculated.

Inserting the data fields described above in Equation 5.9 the total meridional transport of water vapour  $[\overline{wv}]$ , the mean meridional circulation MM  $[\overline{w}][\overline{v}]$ , the stationary eddy circulation SE  $[\overline{w^*v^*}]$ , and the transient eddy circulation TE  $[\overline{w'v'}]$  are calculated.

The fluxes can be divided into seasons depending on the interval used for the time averaging, whereby it can be seen which processes are the most important ones during the different seasons.

### 5.3 Results

Experiments were made with water distributed over a different number of layers in the vertical direction in order to see the effect of the vertical distribution of the water vapour. In the three experiments the water was distributed evenly over the lower 10, 11, and 12 layers respectively, which corresponds to the altitudes shown in Table 5.1.

Not many differences were seen between the experiments, and the results thus seem robust. As other studies have shown that the water vapour in

the Martian atmosphere is likely contained within the lower 10 km of atmosphere, the experiment with the lowermost 11 vertical layers containing the water is chosen as the best representation of the results and described below.

The partitioning of the water transport into the different modes of transport is illustrated in Figure 5.4. In the top panel of Figure 5.4 it is seen that there is a net annual transport of water vapour from northern mid latitudes to low southern latitudes. The transport occurs during the northern fall and winter and is opposed by a transport of smaller magnitude going in the opposite direction during northern spring and summer. There is no observed net flux of water from the polar regions, indicating that the ice caps are not losing mass at the present climate regime. In the three lower panels the net transport due to different mechanisms is shown.

The second panel of the figure illustrates the transport due to transient eddies e.g. storms, and it is seen that there is a convergence of water at northern mid latitudes receiving water from northern low latitudes as well as the north polar region. The transient eddies are not transporting much water on the southern hemisphere except during the northern winter and fall, where water is transported from the south pole to lower latitudes. This takes place at the same time as the huge dust storms occur during the southern summer. The Viking MAWD data set used in these calculations was recorded during such a global dust storm, so the results are likely to illustrate net transport which only occurs during such events. It is possible that there could be less transport during a year without a global dust storm. Also the water vapour measurements were obscured by the suspended dust in the atmosphere, which only allowed for a fraction of the atmospheric column to be sampled for vapour data.

The stationary eddies in the third panel seem not to transport much water in the meridional direction except a south-going transport in the south polar region, counteracting the transport away from the south pole by the transient eddies. Again, this may be the case only for years with a global dust storm or a result of a possible data bias during the dust storm season.

The mean meridional transportation in the fourth panel shows a small net annual transport from the northern mid-latitudes to the northern high latitudes, which counteracts the south-going transport from the transient eddies. It is seen that the mean meridional motion is responsible for the majority of the net annual transport across the equator seen in the first panel.

When comparing the results for Earth in Figure 5.1 and Mars in Figure 5.4 it is expected that there are major differences. In order to be able to

compare the results in the two figures, the units must be converted. The unit for the water vapour fluxes in Figure 5.1 is in  $\text{m s}^{-1} \text{ g kg}^{-1}$ , which can be converted to  $\text{m s}^{-1} \text{ Pr}\mu\text{m}$  in the following way:  $1 \text{ g kg}^{-1}$  corresponds to  $10^{-3} \text{ kg water per kg air}$ . An estimate of the mass of the air column  $\text{pr m}^2$  on Earth is the weight of the atmosphere  $5.3 \cdot 10^{18} \text{ kg}$  divided by the surface area  $510.1 \cdot 10^{12} \text{ m}^2$  which gives  $1.039 \cdot 10^4 \text{ kg air per m}^2$ . As  $1 \text{ Pr}\mu\text{m}$  corresponds to  $10^{-3} \text{ kg water m}^{-2}$ , we end up with  $1 \text{ m s}^{-1} \text{ g kg}^{-1}$  equals  $1.039 \cdot 10^4 \text{ m s}^{-1} \text{ Pr}\mu\text{m}$ . This means that one unit on the vertical axis in the figure with the water fluxes on Earth is about 10000 times bigger than in the Mars plot.

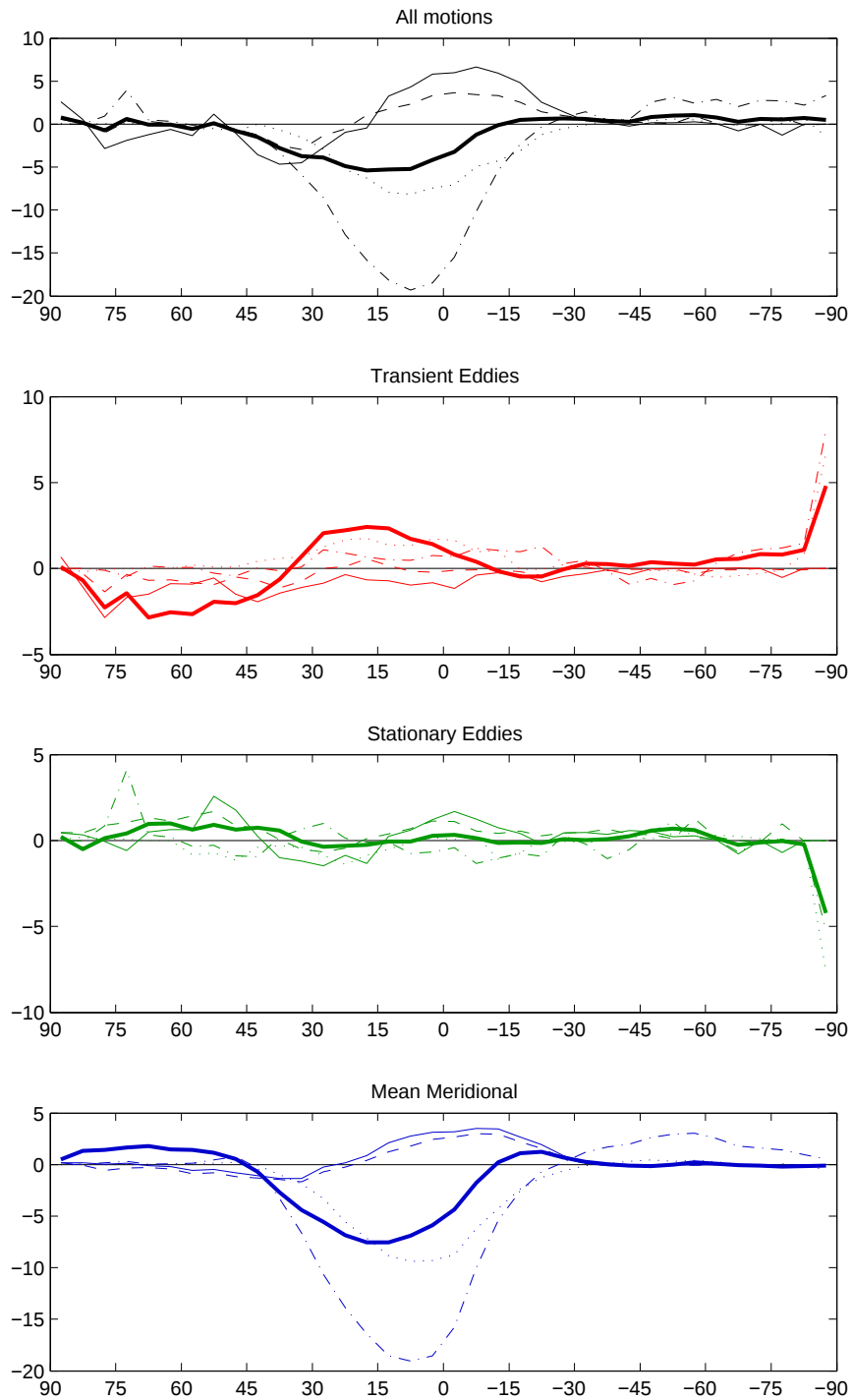
It is seen that the movement of water on Mars only takes place by atmospheric transport in the present climate. On Earth, it is possible to maintain a steady water cycle, by balancing a net transport of water in the atmosphere across latitudes with counteracting transport by ocean currents. Another major difference is that the water fluxes on Earth are 1000-10000 times greater than on Mars. A third observation is that on Earth, the net annual results for almost all of the fluxes is that there is no transfer of water from one hemisphere to the other, since the value of the flux changes sign at the equator. This is in contrast to the results for Mars, where there is a net transport across equator.

## 5.4 Conclusion

An empirical statistical method developed by Peixoto and Oort [1992] was applied to MAWD atmospheric water data and wind velocities from the MCD.

Analysis of transport of water in the Martian atmosphere by partitioning the atmospheric motions into different modes of fluxes has shown that the water cycle in general is local. Most of the transport occurs near the equator, where the water is transported southward during the northern winter and fall, which to some degree is counteracted by a similar north-going flow during spring and summer. The fluxes during the different seasons of the year are not equally sized, and with the data available for this study there is a net southward transport at low latitudes across the equator.

There does not seem to be significant transport of water from the ice caps on a net annual basis. As a net atmospheric transport from one hemisphere to the other will result in depletion of one reservoir and build-up of another it seems that some re-localisation of near-equatorial permafrost is occurring on Mars today.



**Figure 5.4:** The water cycle on Mars decomposed into various fluxes and times during the year. The vertical Positive numbers indicate a northward flux. Thick line: Annual average. Dashed: Spring. Thin: Summer. Dash-dotted: Autumn. Dotted: Winter.



# 6

## Atmospheric Water Transport — Models

Two types of simple models for transport of water vapour have been examined in order to understand the present water cycle on Mars better. The atmospheric transport of water is investigated with a one-dimensional diffusion model and a 2D advection model and compared to the observed seasonal and spatial variations in Mars' atmospheric water content. After initial testing, the two models are coupled to a surface interaction model in order to study the exchange of water with the subsurface reservoir of ground ice and the polar ice caps.

In the first two sections of this chapter, the two transport models made for this thesis will be described. The chapter will then continue with the description of the surface interaction model, and finally with the coupling of the transport models to the surface exchange model.

### 6.1 One-Dimensional Diffusion Model

The Hadley circulation is dominant in the dynamics of the Martian atmosphere and under certain assumptions a simple representation of the water cycle can be made with a one-dimensional eddy diffusion model.

The Hadley circulation will produce a down-gradient transport of water

vapour if the water is uniformly mixed vertically. If the water is uniformly mixed, the upper branch of the cell will contain about as much water as the lower one [Jakosky, 1983], and the presence of a latitudinal gradient in the column abundance of water vapour will result in a net transport down the gradient. If, on the contrary, all the water is located in one of the branches of the Hadley cell, transport will go in the direction of the motion in that branch, independent of the water gradient.

Jakosky [1983b] argues that use of a simple mixing time is not an unreasonable approximation to the water transport. When mixing between the two branches of the Hadley cell occurs due to eddy mixing, the result is a transport of water which is similar to simple diffusion. An atmospheric time constant  $\tau_a$  was estimated, which is the  $e$ -folding time during which the vapour in two adjacent  $10^\circ$  wide latitude bands will mix together and the column abundances of the two bands will approach the same intermediate value (as determined using the appropriate spherical geometry). The resulting estimation of the global mixing times is roughly 1–7 sols.

Generally, an estimation of the diffusion coefficient for eddy diffusion  $D$  can be based on the length scale  $L$  over which the mixing takes place, and the time scale  $\tau$  for this mixing or from the speed of the mixing  $u$ .

$$D \approx \frac{L^2}{\tau} \approx uL \quad (6.1)$$

The length scale used for the mixing is the length of  $10^\circ$  latitude on Mars corresponding to about 591 km, which was also used by Jakosky [1983b]. A typical wind velocity in the lower Martian atmosphere is 1–5 m/s. Inserting the numerical values of the mixing time and length scales into Equation (6.1) gives upper and lower values of the diffusion coefficient

$$D_{up} \approx \frac{(10^\circ \cdot \frac{\pi}{180^\circ} R_M)^2}{1 \text{ sol} \cdot 88775 \text{ s/sol}} \doteq 3.94 \cdot 10^6 \frac{\text{m}^2}{\text{s}}$$

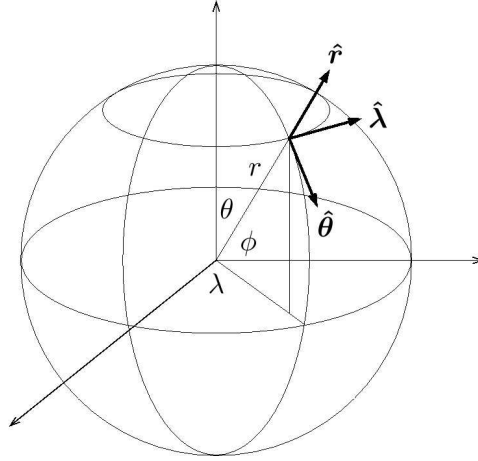
$$D_{low} \approx \frac{(10^\circ \cdot \frac{\pi}{180^\circ} R_M)^2}{7 \text{ sol} \cdot 88775 \text{ s/sol}} \doteq 5.63 \cdot 10^5 \frac{\text{m}^2}{\text{s}}$$

and using the length and velocity scales give

$$D_{up} \approx 10^\circ \cdot \frac{\pi}{180^\circ} R_M \cdot 5 \text{ m/s} \doteq 2.98 \cdot 10^6 \frac{\text{m}^2}{\text{s}}$$

$$D_{low} \approx 10^\circ \cdot \frac{\pi}{180^\circ} R_M \cdot 1 \text{ m/s} \doteq 5.92 \cdot 10^5 \frac{\text{m}^2}{\text{s}}$$

Thus values of  $D$  ranging from  $5 \cdot 10^5$  to  $4 \cdot 10^6 \frac{\text{m}^2}{\text{s}}$  seem to be reasonable to use for the calculations.



**Figure 6.1:** The spherical coordinate system.  $\phi$  is the latitude,  $\theta = 90^\circ - \phi$  is the co-latitude,  $\lambda$  is the longitude, and  $r$  is the radius.

### 6.1.1 The Diffusion Equation and Discretization

The diffusion equation is generally written as

$$\frac{\partial C}{\partial t} = D \nabla^2 C \quad (6.2)$$

where  $C$  is the water vapour density in the atmosphere [ $\frac{\text{kg}}{\text{m}^3}$ ],  $t$  is time, and  $D$  is the diffusion constant. In spherical coordinates (see Figure 6.1 for notation) Equation 6.2 is written as

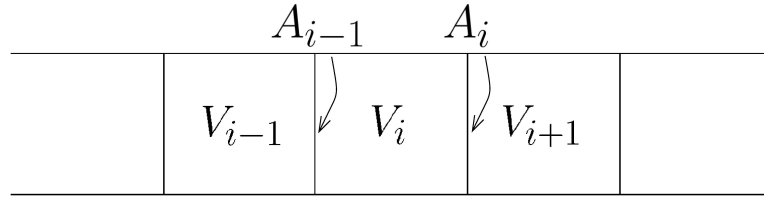
$$\frac{\partial C}{\partial t} = \frac{1}{r} \frac{\partial}{\partial r} \left( r^2 \frac{\partial C}{\partial r} \right) + \frac{1}{r^2 \sin^2 \theta} \frac{\partial^2 C}{\partial \lambda^2} + \frac{1}{r^2 \sin \theta} \frac{\partial}{\partial \theta} \left( D \sin \theta \frac{\partial C}{\partial \theta} \right) \quad (6.3)$$

If it is assumed that any vertical diffusion process is described by the one-dimensional diffusion, the  $r$  dependency can be neglected. Similarly, when averaging in the zonal direction, the  $\lambda$  dependency disappears, and Equation 6.3 reduces to

$$\frac{\partial C}{\partial t} = \frac{1}{r^2 \sin \theta} \frac{\partial}{\partial \theta} \left( D \sin \theta \frac{\partial C}{\partial \theta} \right) \quad (6.4)$$

Instead of using Equation 6.4 directly, we can take one step backwards and derive an expression for the diffusion based on the fluxes. Fick's law states that the flux  $f$  [ $\frac{\text{kg}}{\text{s m}^2}$ ] passing through a surface of unit area is given by

$$f = D \frac{\partial C}{\partial x} \quad (6.5)$$



**Figure 6.2:** Index notation for the box volumes and areas between boxes. The numbering begins with  $V_1$  and  $A_0$  at the north pole and ends with  $V_N$  and  $A_N$  at the south pole.

where  $D$  is the diffusion constant, and  $\frac{\partial C}{\partial x}$  is the change of concentration with distance.

The total flux  $F$  [ $\frac{\text{kg}}{\text{s}}$ ] passing through the area  $A$  [ $\text{m}^2$ ] is given by  $F = fA$ , and can be discretized as

$$F_i = A_i D \left( \frac{C_{i+1} - C_i}{x_{i+1} - x_i} \right) \quad (6.6)$$

where  $F_i$  and  $A_i$  are located at the latitude  $\phi_i$  between boxes  $i$  and  $i+1$ .  $C_i$  is the concentration in box  $i$  between latitudes  $\phi_{i-1}$  and  $\phi_i$ . The distance from the North Pole at the centre of box  $i$  is given by  $x_i = \frac{\pi}{180^\circ} \frac{1}{2} (\phi_{i-1} + \phi_i) R_M$ , where  $R_M$  is the radius of Mars. The index numbering for the variables gridded on the box centres (e.g. the volume  $V_i$ , the concentration of water vapour  $C_i$ , or the distance from origin  $x_i$ ), and on the box boundaries (e.g. the surface area between boxes  $A_i$ , the latitude of the box boundary  $\phi_i$ , or the total flux  $F_i$ ) is illustrated in Figure 6.2.

The area  $A_i$  between the two boxes  $i$  and  $i+1$  is given by

$$A_i = \pi(2R_M + h) \cos \phi_i h \quad (6.7)$$

and the box volume of the  $i$ th box is

$$V_i = \frac{2}{3} \pi [(R_M + h)^3 - R_M^3] (\sin \phi_{i-1} - \sin \phi_i) \quad (6.8)$$

where  $R_M$  is the radius of Mars, and  $h$  is the height of the atmosphere. See section B.1 in the appendix for derivation of the expressions in Equations 6.8 and 6.7.

The net change in concentration in box  $i$  is given by the amount of water transported in and out of the box i.e. the difference between the fluxes at its boundaries

$$\Delta C_i = \frac{F_i - F_{i-1}}{V_i} \Delta t \quad (6.9)$$

The concentration in the  $i$ th box at the time step  $j + 1$  can thus be written in finite difference form

$$C_i^{j+1} = C_i^j + \frac{F_i - F_{i-1}}{V_i} \Delta t \quad (6.10)$$

or in Crank-Nicholson form, where the average of the concentration change in the two time steps is used

$$C_i^{j+1} = C_i^j + \frac{(F_i - F_{i-1})^j + (F_i - F_{i-1})^{j+1}}{2V_i} \Delta t \quad (6.11)$$

The expressions for the fluxes from Equation 6.6 can be combined with the expression in Equation 6.11, which gives the following expression, when the time steps  $j$  and  $j + 1$  are separated

$$\begin{aligned} & C_i^{j+1} + \frac{D\Delta t}{2V_i} \left[ A_{i-1} \frac{C_i^{j+1} - C_{i-1}^{j+1}}{x_i - x_{i-1}} - A_i \frac{C_{i+1}^{j+1} - C_i^{j+1}}{x_{i+1} - x_i} \right] \\ = & C_i^j + \frac{D\Delta t}{2V_i} \left[ -A_{i-1} \frac{C_i^j - C_{i-1}^j}{x_i - x_{i-1}} + A_i \frac{C_{i+1}^j - C_i^j}{x_{i+1} - x_i} \right] \end{aligned} \quad (6.12)$$

The latitudinal spacing of the boxes is the same throughout the work presented here, so  $x_i - x_{i-1} = x_{i+1} - x_i = \Delta x$ . With  $Q_i = \frac{D\Delta t}{2V_i\Delta x}$  Equation 6.12 can be written on a shorter form

$$\begin{aligned} & C_i^{j+1} + Q_i A_{i-1} (C_i^{j+1} - C_{i-1}^{j+1}) - Q_i A_i (C_{i+1}^{j+1} - C_i^{j+1}) \\ = & C_i^j - Q_i A_{i-1} (C_i^j - C_{i-1}^j) + Q_i A_i (C_{i+1}^j - C_i^j) \end{aligned} \quad (6.13)$$

Rearranging this equation gives

$$\begin{aligned} & (-Q_i A_{i-1}) C_{i-1}^{j+1} + (1 + Q_i (A_{i-1} + A_i)) C_i^{j+1} + (-Q_i A_i) C_{i+1}^{j+1} \\ = & (Q_i A_{i-1}) C_{i-1}^j + (1 - Q_i (A_{i-1} + A_i)) C_i^j + (Q_i A_i) C_{i+1}^j \end{aligned} \quad (6.14)$$

The system of equations for all the boxes can be written as two tridiagonal  $N \times N$  matrices and column vectors of dimension  $N$ , which is written in Equation 6.15.

$$\begin{aligned}
& \begin{bmatrix} 1 + Q_1 A_1 & -Q_1 A_1 & 0 & \dots & 0 \\ -Q_2 A_1 & 1 + Q_2(A_2 + A_1) & -Q_2 A_2 & 0 & \vdots \\ \vdots & & & & \vdots \\ \vdots & \ddots & \ddots & \ddots & \vdots \\ \vdots & & & & \vdots \\ \vdots & 0 & -Q_{N-1} A_{N-2} & 1 + Q_{N-1}(A_{N-1} + A_N) & -Q_N A_N \\ 0 & \dots & 0 & -Q_N A_{N-1} & 1 + Q_N A_N \end{bmatrix} \begin{bmatrix} C_1^{j+1} \\ C_2^{j+1} \\ \vdots \\ C_{N-1}^{j+1} \\ C_N^{j+1} \end{bmatrix} \\
& = \begin{bmatrix} 1 - Q_1 A_1 & Q_1 A_1 & 0 & \dots & 0 \\ Q_2 A_1 & 1 - Q_2(A_2 + A_1) & Q_2 A_2 & 0 & \vdots \\ \vdots & & & & \vdots \\ \vdots & \ddots & \ddots & \ddots & \vdots \\ \vdots & & & & \vdots \\ \vdots & 0 & Q_{N-1} A_{N-2} & 1 - Q_{N-1}(A_{N-1} + A_N) & Q_N A_N \\ 0 & \dots & 0 & Q_N A_{N-1} & 1 - Q_N A_N \end{bmatrix} \begin{bmatrix} C_1^j \\ C_2^j \\ \vdots \\ C_{N-1}^j \\ C_N^j \end{bmatrix} \\
& \hspace{25em} (6.15)
\end{aligned}$$

Or in a shorter form

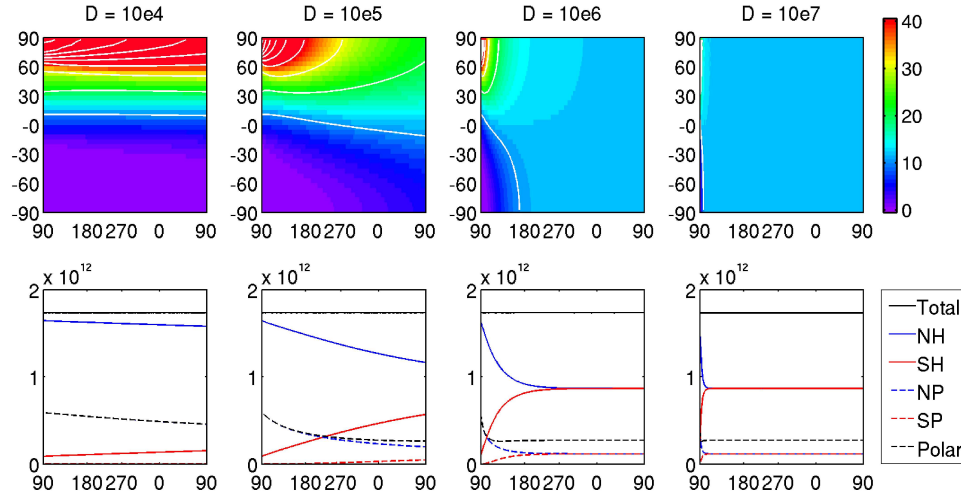
$$\underline{\underline{\mathcal{G}}} \mathbf{C}_{j+1} = \underline{\underline{\mathcal{H}}} \mathbf{C}_j \quad (6.16)$$

where  $\underline{\underline{\mathcal{G}}}$  and  $\underline{\underline{\mathcal{H}}}$  are the coefficient matrices for the time steps  $j+1$  and  $j$  respectively, and  $\mathbf{C}_{j+1}$  and  $\mathbf{C}_j$  are the concentration vectors for the same time steps. Equation 6.16 can be rearranged and used to find the concentrations for the next time step from

$$\mathbf{C}_{j+1} = \left( \underline{\underline{\mathcal{G}}} \right)^{-1} \underline{\underline{\mathcal{H}}} \mathbf{C}_j = \underline{\underline{\mathcal{K}}} \mathbf{C}_j \quad (6.17)$$

### 6.1.2 Results

It is expected that the mixing of the atmosphere occur more rapidly for increasing values of  $D$ . Figure 6.3 illustrates an example of the transport of atmospheric water vapour for different values of the diffusion constant  $D$



**Figure 6.3:** Test of the effectivity of the diffusive transport in the Martian atmosphere for four different values of the eddy diffusion constant  $D$  (columns from left to right:  $10^4$ ,  $10^5$ ,  $10^6$ ,  $10^7 \frac{\text{m}^2}{\text{s}}$ ). Initial vapour distribution is from the MAWD data set for  $L_s = 90^\circ$  ( $s_0 = 4$ ). Upper panel: The water vapour column abundance in  $\text{Pr}\mu\text{m}$  as function of latitude  $[\circ]$  (vertical axis) and time of year  $[\circ L_s]$  (horizontal axis). The contour lines represent values of 20, 40, 60, 80, and 100  $\text{Pr}\mu\text{m}$ . Lower panel: Vapour contents of different regions [kg] as a function of time of year  $[\circ L_s]$ . The total amount of vapour in the atmosphere (“Total”, black solid line), the total amount of vapour in the northern hemisphere (“NH”, blue solid line), the total amount of vapour in the southern hemisphere (“SH”, red solid line), the total amount of vapour in the atmosphere of the polar regions (“Polar”, black dashed line), the total amount of vapour in the northern polar atmosphere (“NP”, blue dashed line), the total amount of vapour in the southern polar atmosphere (“SP”, red dashed line).

for the one-dimensional diffusion model. In all cases, the total amount of vapour in each of the hemispheres goes towards the same intermediate value with time, however the time for a steady state to be reached increases with decreasing value of the diffusion constant. The constant uniform amount of water vapour in the atmosphere reached at steady state corresponds to a global uniform amount of water vapour column of about 12  $\text{Pr}\mu\text{m}$ . This in turn corresponds to the initial amount of vapour evenly spread out over the globe. It is clear that the eddy diffusion cannot explain the present water cycle on Mars without interaction with sources and sinks at the surface, as diffusion alone only serves to even out the water distribution with time.

In Figure 6.4 results from the diffusion calculations for different values of the diffusion constant are shown. The plots illustrate which latitudinal regions will lose vapour, and which will gain vapour. The calculations were made for different values of the eddy diffusion constant and an initial water distribution as given by MAWD data. Each tick mark on the horizontal axis

represents the onset of one of the twelve seasons of  $30^\circ L_s$ , the width of the tick marks varies as the seasons have different lengths due to the eccentricity of Mars' orbit. At the beginning of each season the distribution of vapour is set to the MAWD values at the given season, which gives rise to the vertical white stripes in the plots. Red values show loss of vapour at the given latitude and time, and blue values show a net gain, the colour scale is given in kg.

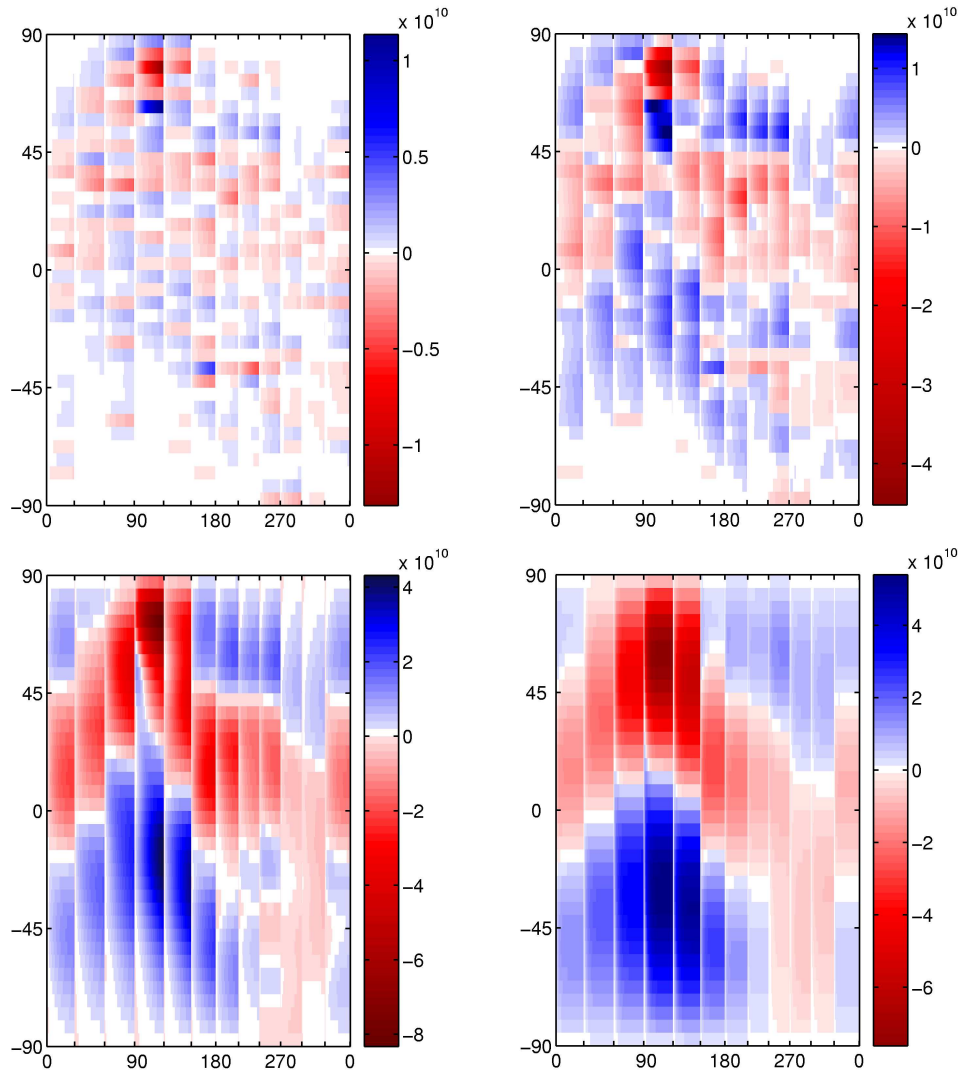
For the values of  $D$  above  $10^5 \frac{\text{m}^2}{\text{s}}$  it is seen that water is lost from the high northern latitudes during northern summer (around  $L_s = 90^\circ - 120^\circ$ ) and gained south of this region. During southern summer (around  $L_s = 270^\circ - 300^\circ$ ) the southern polar areas lose water as well as southern midlatitudes and northern low latitudes, and a rise in the water content is seen in the northern midlatitudes. It is seen that for  $D = 10^4 \frac{\text{m}^2}{\text{s}}$  the diffusion process is not able to transport water globally, as the vapour is stranded in smaller regions on the timescale of the seasonal variation. For an increasing value of  $D$  it is seen that the diffusion of vapour occurs more rapidly and over increasingly larger regions.

## 6.2 Advection Model

The water transport in the Martian atmosphere is examined with a model calculating the advection of water vapour explicitly. It is assumed that the water moves with the wind as a passive tracer, meaning that the water cycle itself does not influence the dynamics of the atmosphere as is the case on Earth.

The advection model is two-dimensional in its spatial coordinates, meaning that the calculations take place for a zonally averaged latitude-altitude plane  $(\phi, z)$ . A one-dimensional model as in the case of the diffusion model in the previous section would not have been able to fulfil the continuity requirement, as air would have accumulated in some boxes at the expense of others. The initial water vapour distribution is given in total column abundance as a function of latitude from the MAWD data set, and needs to be converted into a vertical profile of vapour. A simple assumption is that the water is distributed in the air within a given column so that the mixing ratio of water is constant with altitude. If there is sufficient vertical mixing, the simplification made should be a valid assumption.

As input the advection model uses the zonally averaged data fields of the meridional winds  $v(\phi, z, t)$ , an initial distribution of water with latitude  $Pr_0(\phi)$ , the atmospheric temperature  $T(\phi, z, t)$ , the atmospheric pressures  $p(\phi, z, t)$ , and the surface pressure  $p_s(\phi, t)$ .



**Figure 6.4:** Changes in atmospheric water content resulting from diffusional transport as a function of time (horizontal axis) and latitude (vertical axis). The colour bar shows the change in water content in kilogrammes [kg]. The tick marks on the horizontal axis, which is linear in sols, show the beginning of each season; the numbers correspond to time of the year in  $^{\circ}L_s$ . For each season the initial amount of water vapour content is set to be the measured values from MAWD. The four results shown here are for different values of the diffusion constant: Upper left:  $D = 10^4 \frac{\text{m}^2}{\text{s}}$ . Upper right:  $D = 10^5 \frac{\text{m}^2}{\text{s}}$ . Lower left:  $D = 10^6 \frac{\text{m}^2}{\text{s}}$ . Lower right:  $D = 10^7 \frac{\text{m}^2}{\text{s}}$ .

In the following sections, first the the latitudinal exchange and next the vertical exchange are described. After the model is thoroughly tested without sources or sinks, the model is coupled to the surface interaction model, which exchanges water between the atmosphere and the ice caps and permafrost in the regolith.

### 6.2.1 The Continuity Equation

The continuity equation for the concentration of water vapour  $C$  in the Martian atmosphere is given by

$$\frac{\partial C}{\partial t} + \nabla \cdot (C\mathbf{v}) = 0 \quad (6.18)$$

where  $\mathbf{v}$  is the wind velocity. The equation can be expanded using the vector identity

$$\nabla \cdot (C\mathbf{v}) \equiv C\nabla \cdot \mathbf{v} + \mathbf{v} \cdot \nabla C \quad (6.19)$$

The divergence of a vector field  $\mathbf{F} = (F_r, F_\theta, F_\lambda)$  in spherical coordinates is

$$\nabla \cdot \mathbf{F} = \frac{1}{r^2} \frac{\partial}{\partial r} (r^2 F_r) + \frac{1}{r \sin \theta} \frac{\partial}{\partial \theta} (\sin \theta F_\theta) + \frac{1}{r \sin \theta} \frac{\partial F_\lambda}{\partial \lambda} \quad (6.20)$$

and the gradient of a scalar  $f$  is

$$\nabla f = \hat{\mathbf{r}} \frac{\partial f}{\partial r} + \frac{1}{r} \hat{\boldsymbol{\theta}} \frac{\partial f}{\partial \theta} + \frac{1}{r \sin \theta} \hat{\boldsymbol{\lambda}} \frac{\partial f}{\partial \lambda} \quad (6.21)$$

If we only look at a thin spherical shell with negligible changes in the  $r$ -direction, it can be assumed that there is no  $r$  dependency. Similarly, when a zonal average is made, the  $\lambda$  dependency disappears and the equations reduce to

$$\nabla \cdot \mathbf{F} = \frac{1}{r \sin \theta} \frac{\partial}{\partial \theta} (\sin \theta F_\theta) \quad (6.22)$$

and

$$\nabla f = \left( 0, \frac{1}{r} \frac{\partial f}{\partial \theta}, 0 \right) \quad (6.23)$$

Combining Equations 6.18 and 6.19 for the concentration of water vapour in the Martian atmosphere  $C$  [ $\frac{\text{kg}}{\text{m}^3}$ ] and the meridional wind velocity  $\mathbf{v}$  [ $\frac{\text{m}}{\text{s}}$ ] with the zonal symmetric thin shell Equations 6.22 and 6.23, we get the following expression for the rate of change of concentration of water vapour

$$\frac{\partial C}{\partial t} = - \left( C \left[ \frac{1}{r \sin \theta} \frac{\partial}{\partial \theta} (\sin \theta v_\theta) \right] + v_\theta \frac{1}{r} \frac{\partial C}{\partial \theta} \right) \quad (6.24)$$

The differential equation for the temporal change in concentration in Equation (6.24) can be discretized for the purpose of solving the equation numerically. The first order temporal derivative of the quantity  $f$  in the  $i$ th box at the time step  $j$  is approximated to the finite difference

$$\left[ \frac{\partial f}{\partial t} \right]_i^j = \frac{f_i^{j+1} - f_i^j}{\Delta t} \quad (6.25)$$

The first order spatial derivative of the quantity  $f$  centered in the  $i$ th box at the time step  $j$  is approximated to

$$\left[ \frac{\partial f}{\partial \theta} \right]_i^j = \frac{f_{i+1}^j - f_{i-1}^j}{2\Delta\theta} \quad (6.26)$$

where  $\Delta\theta$  is the spacing between two box centres in radians. This leads to the following discretized equation for Equation (6.24)

$$\begin{aligned} \frac{C_i^{j+1} - C_i^j}{\Delta t} = & - \left[ C_i^j \frac{1}{r_i \sin \theta_i} \left( \frac{\sin \theta_{i+1} v_{i+1} - \sin \theta_{i-1} v_{i-1}}{2\Delta\theta} \right) \right. \\ & \left. + v_i \frac{1}{r_i} \left( \frac{C_{i+1}^j - C_{i-1}^j}{2\Delta\theta} \right) \right] \end{aligned} \quad (6.27)$$

which can be arranged in such a way that the solution for the new time step  $j+1$  can be explicitly calculated from the values of the previous time step.

$$C_i^{j+1} = C_i^j + \Delta C_i^j \quad (6.28)$$

where

$$\Delta C_i^j = - \frac{\Delta t}{r_i 2\Delta\theta} \left[ \frac{C_i^j}{\sin \theta_i} (\sin \theta_{i+1} v_{i+1} - \sin \theta_{i-1} v_{i-1}) + v_i (C_{i+1}^j - C_{i-1}^j) \right] \quad (6.29)$$

For the boxes at the poles (box number 1 and  $N$ ) the spatial derivative can not be centered around the box, wherefore a right and a left difference is used instead. This leads to Equation. (6.30, 6.31) instead of Equation (6.29)

$$\Delta C_1^j = - \frac{\Delta t}{r_1 \Delta\theta} \left[ \frac{C_1^j}{\sin \theta_1} (\sin \theta_2 v_2 - \sin \theta_1 v_1) + v_1 (C_2^j - C_1^j) \right] \quad (6.30)$$

$$\Delta C_N^j = - \frac{\Delta t}{r_N \Delta\theta} \left[ \frac{C_N^j}{\sin \theta_N} (\sin \theta_N v_N - \sin \theta_{N-1} v_{N-1}) + v_N (C_N^j - C_{N-1}^j) \right] \quad (6.31)$$

The FTSC-scheme used here (forward in time, centered in space) is not numerically stable but if the Courant-Friedrichs-Lewy (CFL) criterion is

fulfilled, there is hope to get at better solution. This criterion states that in order for the scheme to be stable, the wind speed, the length of the time step, and distance between points in space must be related to each other in the following way

$$\frac{|v|\Delta t}{\frac{\pi}{180^\circ}\Delta\phi R_M} \leq 1 \quad (6.32)$$

As the wind speeds and spatial resolution are given by the input data and its resolution, the only thing that can be adjusted to fulfil the criterion is the length of the time step, which must be short enough to meet the criterion.

### 6.2.2 Alternative Discrete Representation of the Advection

The advection process may also be parameterized based on a calculation of the fluxes from box to box, where the variation of box size with latitude is taken into account.

The flux  $f$  [ $\frac{\text{kg}}{\text{m}^2\text{s}}$ ] from one box to another is given by

$$f = (vC)|_{\text{box boundary}} = \frac{1}{A} \frac{dM}{dt} \approx \frac{1}{A} \frac{\Delta M}{\Delta t} \quad (6.33)$$

where  $A$  is the area of the boundary between the two boxes, and  $\Delta M$  is the mass exchanged between them in the time interval  $\Delta t$ . The  $i$ th boundary is located between the boxes  $i$  and  $i + 1$ . Rearranging Equation (6.33) and inserting parameter values for transport across the  $i$ th boundary gives the following changes in concentration for the  $i$ th and  $(i + 1)$ th boxes

$$\Delta C_i = \frac{A_i}{V_i} (vC)|_{i\text{th box boundary}} \Delta t \quad (6.34)$$

$$\Delta C_{i+1} = -\frac{A_i}{V_{i+1}} (vC)|_{i\text{th box boundary}} \Delta t \quad (6.35)$$

where the sign conventions are that northward fluxes and velocities are positive and that the box number increases from north to south.

The total change in concentration in the  $i$ th box is given by the amount of water exchanged with the  $(i - 1)$ th box and the  $(i + 1)$ th box. The wind velocity at the  $i$ th boundary is taken as the linear average of the values from the two neighbouring boxes, i.e.  $v|_{i\text{th box boundary}} = \frac{1}{2}(v_i + v_{i+1})$ . The concentration  $C|_{i\text{th box boundary}}$  transported across the boundary is equal to the concentration in the box the water is transported from, which is determined by the sign of  $v|_{i\text{th box boundary}}$ . E.g. if the velocity is positive, the transport goes from south to north and  $C|_{i\text{th box boundary}} = C_{i+1}$ , if it is negative the transport goes from north to south and  $C|_{i\text{th box boundary}} = C_i$ .

Hence, the total change for the  $i$ th box for the  $j$ th time step is given by

$$\begin{aligned}\Delta C_i^j = & -\frac{A_{i-1}}{V_i} \frac{(v_{i-1} + v_i)}{2} C|_{(i-1)\text{th box boundary}} \Delta t \\ & + \frac{A_i}{V_i} \frac{(v_i + v_{i+1})}{2} C|_{i\text{th box boundary}} \Delta t\end{aligned}\quad (6.36)$$

The box boundary areas  $A_{ik}$  and box volumes  $V_{ik}$  depend on both latitude  $\phi$  (index  $i$ ) and altitude  $z$  (index  $k$ ) and are given by (see section B.2 in the appendix for derivation of the expressions)

$$A_{ik} = \pi [(R_M + z_{k+1})^2 - (R_M + z_k)^2] \cos \phi_i \quad (6.37)$$

$$V_{ik} = \frac{2}{3} \pi [(R_M + z_{k+1})^3 - (R_M + z_k)^3] (\sin \phi_i - \sin \phi_{i+1}) \quad (6.38)$$

The change in concentration for the box  $i$  at timestep  $j$  can be found by inserting the expression for  $A_i$  and  $V_i$  from Equation. (6.37) and (6.38) into (6.36) and reducing the expression to

$$\begin{aligned}\Delta C_{ik}^j = & \frac{3 (R_M + z_{k+1})^2 - (R_M + z_k)^2}{4 (R_M + z_{k+1})^3 - (R_M + z_k)^3} (\sin \phi_i - \sin \phi_{i+1})^{-1} \\ & \cdot \left[ -(v_{i-1} + v_i) C|_{(i-1)\text{th box boundary}} \cos \phi_{i-1} \right. \\ & \left. + (v_i + v_{i+1}) C|_{i\text{th box boundary}} \cos \phi_i \right] \Delta t\end{aligned}\quad (6.39)$$

For each of the polar boxes there is only exchange with one box, which gives the concentration

$$\begin{aligned}\Delta C_{1k}^j = & \frac{3 (R_M + z_{k+1})^2 - (R_M + z_k)^2}{4 (R_M + z_{k+1})^3 - (R_M + z_k)^3} (1 - \sin \phi_2)^{-1} \\ & \cdot \left[ (v_1 + v_2) C|_{1\text{st box boundary}} \cos \phi_2 \right] \Delta t\end{aligned}\quad (6.40)$$

$$\begin{aligned}\Delta C_{Nk}^j = & \frac{3 (R_M + z_{k+1})^2 - (R_M + z_k)^2}{4 (R_M + z_{k+1})^3 - (R_M + z_k)^3} (\sin \phi_{N-1} + 1)^{-1} \\ & \cdot \left[ -(v_{N-1} + v_N) C|_{(N-1)\text{th box boundary}} \cos \phi_{N-1} \right] \Delta t\end{aligned}\quad (6.41)$$

This can be compared with the corresponding expression for the spherical continuity equation in Equation (6.29). Notice that one equation is in co-latitude centered in the boxes and the other in latitude centered at the

box boundaries. When the terms in the two equations are examined more closely, it is seen that the difference between the two expressions is basically the method of averaging applied.

### 6.2.3 Vertical Exchange

It is assumed that the water vapour for each latitude band is distributed in the air column in such a way that all the water is contained in the atmospheric column from the surface and up to a certain level  $z_{TOW}$  (top of the water column), and the mixing ratio of water with respect to Martian air is constant within this region.

The ideal gas law states

$$pV = NRT \quad (6.42)$$

where  $p$  is pressure [Pa],  $V$  is the volume of the gas [ $\text{m}^3$ ],  $N$  is the number of molecules in the air measured in mol,  $R$  is the universal gas constant, and  $T$  is the temperature [K]. The mass density  $\rho$  of air [ $\frac{\text{kg}}{\text{m}^3}$ ] can be found from the ideal gas law as

$$\rho = \frac{Mp}{RT} \quad (6.43)$$

where  $M$  is the molecular weight of the air [ $\frac{\text{kg}}{\text{mol}}$ ].

Dalton's law states that the partial pressure  $p_Y$  of a gas  $Y$  in a mixture of gasses with pressure  $p$  is proportional to the mixing ratio  $X_Y$  [ $\frac{\text{mol}}{\text{mol}}$ ] of the gas

$$p_Y = X_Y p \quad (6.44)$$

The water vapour concentration by mass  $C$  in the Martian air is identical to the density of the  $\text{H}_2\text{O}$  gas  $\rho_{\text{H}_2\text{O}}$  that can be calculated from Dalton's law, which gives the expression

$$C = \frac{M_{\text{H}_2\text{O}} p_{\text{H}_2\text{O}}}{RT} = \frac{p X_{\text{H}_2\text{O}}}{R_{\text{H}_2\text{O}} T} \quad (6.45)$$

where  $R_{\text{H}_2\text{O}} = R/M_{\text{H}_2\text{O}}$  is the gas constant for water vapour.

The total column mass of water  $m_{\text{H}_2\text{O}}$  [ $\frac{\text{kg}}{\text{m}^2}$ ] can be found as

$$m_{\text{H}_2\text{O}} = \frac{1}{A_s} \int_{z=0}^{z=z_{TOW}} C dV \quad (6.46)$$

where  $A_s$  is the surface area of the bottom of the column.

Equation (6.46) can be combined with Equation (6.43) and approximated with a sum over the  $N_z$  vertical layers from the surface and up to  $z_{TOW}$ ,

which is the altitude limit of the vapour, to give a discrete expression under the assumption of the mixing ratio being constant with altitude

$$\begin{aligned}
 m_{H_2O} &= \frac{1}{A_s} \sum_{k=1}^{N_z} C_k V_k \\
 &= \frac{1}{A_s} \sum_{k=1}^{N_z} \frac{p_k X_{H_2O}}{R_{H_2O} T_k} \Delta z(i_z) \\
 &= X_{H_2O} R_{H_2O}^{-1} \frac{1}{A_s} \sum_{k=1}^{N_z} \frac{p_k}{T_k} V_k
 \end{aligned} \tag{6.47}$$

The mixing ratio can then be calculated from the total column mass of water  $m_{H_2O}$  in combination with the atmospheric temperature and pressure profiles from the MCD.

$$X_{H_2O} = m_{H_2O} R_{H_2O} \left( \frac{1}{A_s} \sum_{k=1}^{N_z} \frac{p_k}{T_k} V_k \right)^{-1} \tag{6.48}$$

For each time step the columnar mass is calculated after the horizontal exchange of water across the latitude at each layer. From the columnar mass, the mixing ratio is calculated, and finally the vertical distribution of the vapour can then be calculated from Equation 6.45.

The atmospheric pressure data is thus used together with the atmospheric temperature to distribute the water vapour vertically within an atmospheric column for each time step during the calculations. After calculation of the advection of vapour, the total column of vapour is found by adding the contribution for each of the 32 vertical levels and then redistributing this sum back into the layers in so that the mixing ratio of water is constant through the column.

## 6.2.4 Results

The two different numerical schemes described in sections 6.2.1 and 6.2.2, respectively, produce very similar results, and term by term analysis of Equations 6.29 and 6.39 showed that the difference between the two schemes is the method of averaging made for the discretization. On this basis, it is inconclusive which scheme is the better one to use. However, for calculations for the full Martian year, the scheme based on the alternative discretization (see section 6.2.2) shows to be computationally faster than the scheme based on the spherical continuity equation (see section 6.2.1). Without the

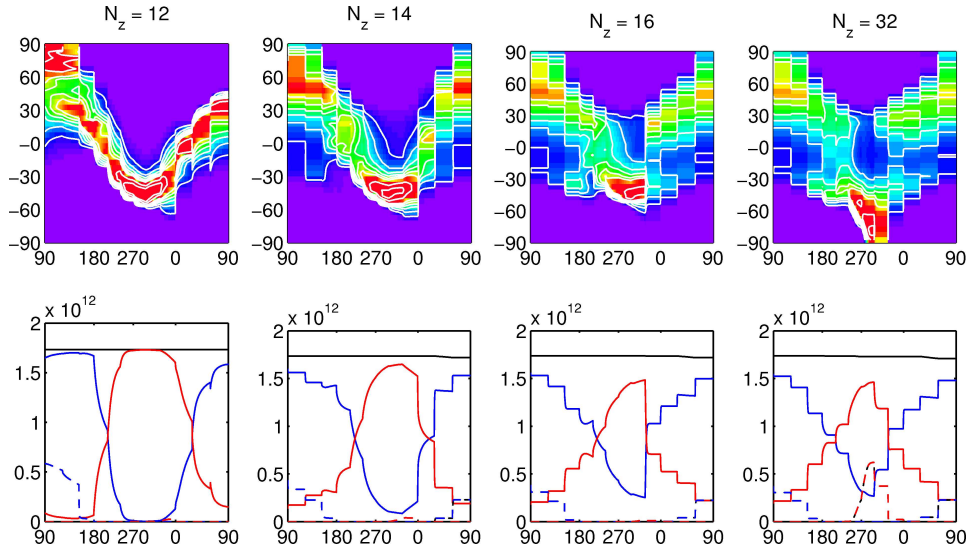
normalization mechanism, the alternative model is closer to mass conserving than the spherical continuity equation. This could be due to the lack of a sufficient method of detecting and correcting numerical problems, where more water is taken from one box to another than is actually available in the spherical continuity scheme. This sort of control mechanism is more easily incorporated in the alternative model, where each transport from box to box is accounted for.

In order to examine the influence of the vertical mixing in the Martian troposphere, experiments were made with four different values for the water column height. The entire atmospheric water inventory was placed in the lower 12, 14, 16, and 32 vertical layers in the model, where the last one corresponds to the entire atmospheric column. These numbers of boxes correspond to a water column height of 11.9, 22.25, 34 and 123 km, respectively. In Figure 6.5 it is seen that only when the water vapour is mixed over the entire vertical column will the water be transported across all latitudes. In the experiment with water only distributed in the lower atmosphere, the water is only present in a narrow range of latitudes during the different seasons, whereas more atmospheric mixing of vapour occurs in the other experiments.

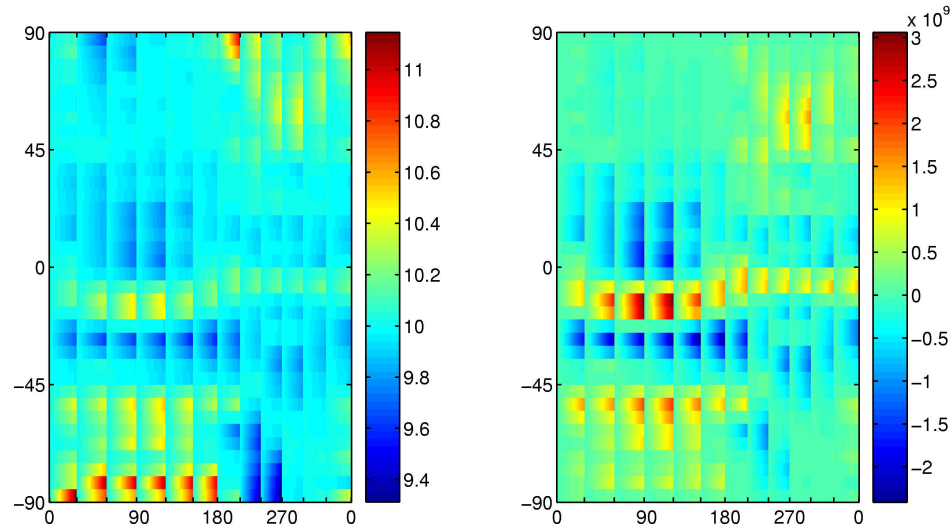
The ability of the advection process to transport water in the Martian atmosphere was investigated using some different approaches. The Figures 6.6 and 6.7 both illustrate the change in water content at a certain latitude arising from meridional transport for a given season (season here corresponds to Mars “month” where the Martian year is divided into 12 intervals of  $30^\circ L_s$ ). The model used both MAWD data (Figure 6.7) and a uniform global distribution of  $10 \text{ Pr}\mu\text{m}$  (Figure 6.6) as input for the calculations.

Using a uniform global distribution of  $10 \text{ Pr}\mu\text{m}$  water as initial condition, the ability of the meridional winds to transport water across the latitudes can be examined. In Figure 6.6 it is seen that water is lost from the summer pole and gained at the winter pole. This is in agreement with the expected pattern, in which the condensation of  $\text{CO}_2$  on the winter hemisphere drives a flow in the atmosphere towards the winter pole in order to compensate for the  $\text{CO}_2$  that has condensed out from the atmosphere to the surface. It is also seen that the water amount increases in a band approximately at the equator and decreases in the nearby regions.

When using the MAWD measured values of atmospheric water content it is illustrated which transport of atmospheric water actually takes place on Mars in today’s climate. In Figure 6.7 it is seen that water indeed is removed from high northern latitudes in the northern summer time and vice versa for the southern summer. The strong increase in water vapour in the winter hemisphere is not so pronounced for the case of the MAWD input data as



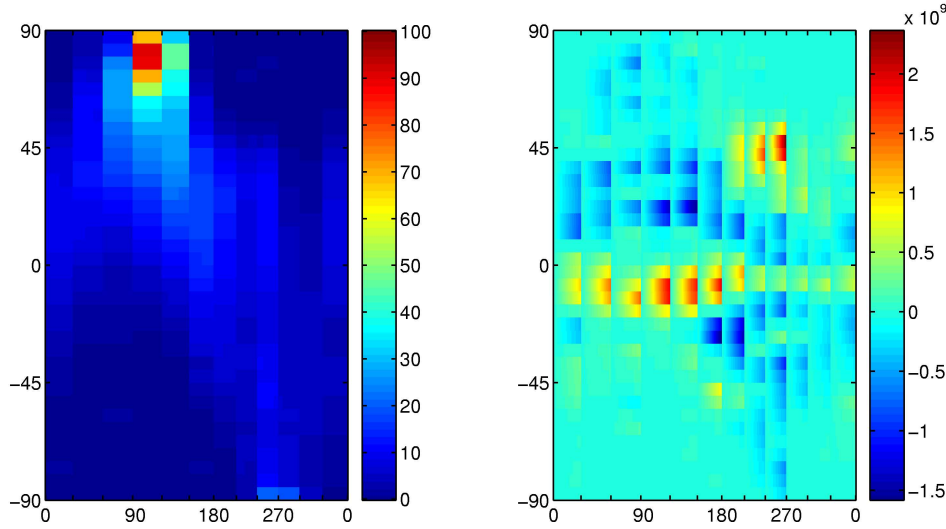
**Figure 6.5:** Comparison of transport with different maximum heights of the water column.  $N_z$  refers to number of layers in the vertical direction, see the text for corresponding heights of vertical atmospheric mixing. Upper panel: The water vapour column abundance in  $\text{Pr}\mu\text{m}$  as function of latitude  $[\circ]$  (vertical axis) and time of year  $[\circ L_s]$  (horizontal axis). The contour lines represent values of 20, 40, 60, 80, and 100  $\text{Pr}\mu\text{m}$ . Lower panel: Vapour contents of different regions [kg] as a function of time of year  $[\circ L_s]$ . The total amount of vapour in the atmosphere ("Total", black solid line), the total amount of vapour in the northern hemisphere ("NH", blue solid line), the total amount of vapour in the southern hemisphere ("SH", red solid line), the total amount of vapour in the atmosphere of the polar regions ("Polar", black dashed line), the total amount of vapour in the northern polar atmosphere ("NP", blue dashed line), the total amount of vapour in the southern polar atmosphere ("SP", red dashed line).



**Figure 6.6:** Trends in meridional transport as a function of time of the year [ $^{\circ}L_s$ ] and latitude [ $^{\circ}$ ]. At the beginning of each season, the initial water vapour content is set to 10  $\text{Pr}\mu\text{m}$ . The deviation from this value with time indicates the change in water content arising from in meridional transport for the given season. Left: The total water column in the atmosphere as function of time and latitude; values shown are in  $\text{Pr}\mu\text{m}$ . Right: The deviation from the initial condition of 10  $\text{Pr}\mu\text{m}$ ; values shown are in kg.

for the uniform 10  $\text{Pr}\mu\text{m}$  data. This is because the 10  $\text{Pr}\mu\text{m}$  data here overestimates the atmospheric water content, which gives rise to a potential larger transport than is seen in reality. The differences between the two figures illustrate that only if there is vapour present at the time and location of high wind velocities will the water be transported from one location to another.

When running the advection model without surface interaction it is seen that the water will be trapped at the poles and in a narrow band near the equator. This will be the case in zones of convergence, where the winds are directed towards such a zone e.g. the equatorial regions at certain times of the year. If the transport is more effective at times of convergence than at divergence or transport across the zones, the water will eventually be accumulated in these locations. When the MAWD input data is used for the initial condition, the final location of the water is seen to be influenced by initial distribution, i.e. when a large amount of water is initially located in the northern hemisphere, a larger amount will also end up at the north pole compared to the south pole.



**Figure 6.7:** Trends in meridional transport as a function of time of the year [ $^{\circ}L_s$ ] and latitude [ $^{\circ}$ ]. At the beginning of each season, the initial water vapour content is set to be the measured values from MAWD. Left: The total water column in the atmosphere as function of time and latitude, values shown are in  $\text{Pr}\mu\text{m}$ . Right: The deviation from the initial value for each season in the unit of kg.

## 6.3 Exchange with Surface Reservoirs

As seen in the previous sections (6.1 and 6.2), an exchange with surface reservoirs of water must be included in order to model the present water cycle on Mars more realistically. A simple approach has been taken, where sublimation and deposition are based on the balance at the phase transition between vapour and ice.

The total sublimation rate from an exposed surface of clean ice is following the sublimation model by Ivanov and Muhleman [2000]. Here  $dM/dt$  is the rate of change of water vapour in the atmosphere and is given by

$$\frac{dM}{dt} = D_{ex} (\epsilon_{\text{ice}}(T) - p_{\text{H}_2\text{O}}(t)) \sqrt{\frac{m_{\text{H}_2\text{O}}}{2\pi kT}} \quad (6.49)$$

where  $\epsilon_{\text{ice}}$  is the saturation vapour pressure over ice [Pa],  $p_{\text{H}_2\text{O}}$  is the partial pressure of vapour in the atmosphere [Pa],  $m_{\text{H}_2\text{O}}$  is the molecular mass of vapour [kg],  $T$  is temperature [K],  $k$  is the Boltzman constant, and

$$D_{ex} = A_{\text{drag}} u \sqrt{\frac{2\pi m_{\text{H}_2\text{O}}}{kT}} \quad (6.50)$$

Here  $A_{\text{drag}}$  is a dimensionless drag coefficient, and  $u$  is the near surface wind

speed. When combined the above expressions lead to the equation

$$\frac{dM}{dt} = (\epsilon_{\text{ice}}(T) - p_{\text{H}_2\text{O}}(t)) \frac{m_{\text{H}_2\text{O}}}{kT} A_{\text{drag}} u \quad (6.51)$$

The saturation vapour pressure over ice  $\epsilon_{\text{ice}}$  only depends on the temperature  $T$  and is calculated from the formulation by Hardy [1998] for the ITS-90 temperature scale.

$$\epsilon_{\text{ice}} = \sum_{i=0}^4 k_i T^{i-1} + k_5 \ln T \quad (6.52)$$

where

$$\begin{aligned} k_0 &= -5.8666426 \cdot 10^3 \\ k_1 &= 2.232870244 \cdot 10^1 \\ k_2 &= 1.39387003 \cdot 10^{-2} \\ k_3 &= -3.4262402 \cdot 10^{-5} \\ k_4 &= 2.7040955 \cdot 10^{-8} \\ k_5 &= 6.7063522 \cdot 10^{-1} \end{aligned}$$

Sublimation can be inhibited if CO<sub>2</sub> frost is present at the surface. In case CO<sub>2</sub> frost is present, there will be no sublimation of water away from the surface into the atmosphere, but deposition can take place as the CO<sub>2</sub> frost functions as a cold trap for the water condensate.

For the surface interaction model, the zonally averaged values for the ground ice have been used. The original data set covered the region between 87.5°N and 87.5°S, so for this study, a value of 100% water ice has been ascribed to the North Pole and equivalently a value of 50% has been used for the South Pole. These values can be justified from examination of images which show that the residual north polar ice cap covers the whole region north of 85°N, and the south polar ice cap covers approximately half the corresponding region on the southern hemisphere.

### 6.3.1 Experiments

In order to examine to which degree different sections of the surface reservoir are interacting with the atmospheric water content, a set of experiments have been constructed giving understanding of the influence of the different scenarios.

In the experiments a certain percentage of the surface at a given latitude is clean H<sub>2</sub>O ice, and the amount corresponds to the percentage of ice in the

subsurface. The remaining part of the surface is assumed to be rocky and does not contribute to or influence the atmospheric exchange.

Six different experiments of exchange with an ice surface (Figure 6.8) have been examined and compared to the reference value where no surface exchange is calculated through the model.

**GRS:** The surface area in each box has a fraction which is clean ice corresponding to the values for the regolith measured by the GRS instruments, see section 3.6. This is believed to be the best representation for the ice exchange.

**Poles:** Only the poles (boxes number 1 and  $N$ ) exchange water with the atmosphere. For the north polar box the surface consists of 100% ice and the south polar box of 50% ice. The remainder surface of the south polar box is believed to be dust-covered and does not interact with the atmosphere.

**Poles and GRS 60°+:** Only latitudes higher than 60° on both hemispheres exchange with the atmosphere in the same manner as described above.

**Poles and 50% GRS:** The poles and 50% of the GRS surface of the rest of the latitudes interact with the atmosphere.

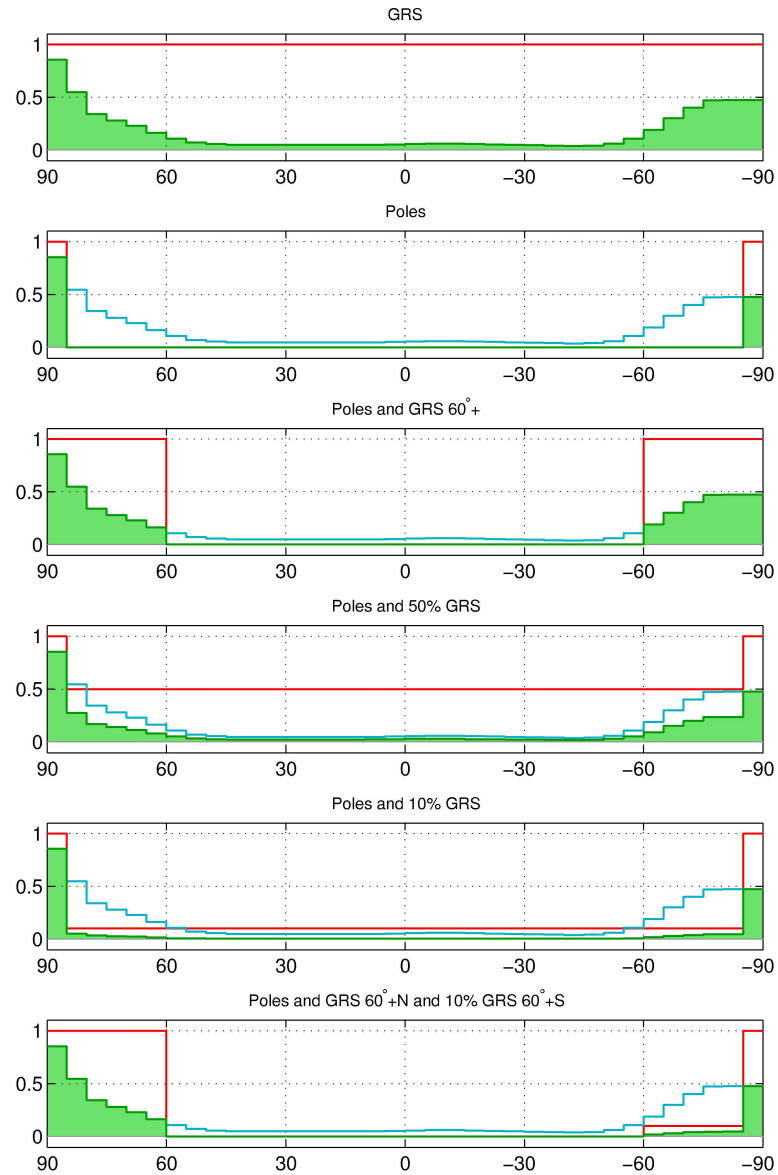
**Poles and 10% GRS:** Same as above except only 10% of the GRS surface is used.

**Poles and GRS 60°+N and 10% GRS 60°+S:** The latitudes higher than 60°N, the South Pole, and 10% of the rest of the GRS surface south of 60°S exchange vapour with the atmosphere.

### 6.3.2 Results for Coupling with the Diffusion Model

In the lower panel of Figure 6.9 results are shown for the coupling of the surface exchange model with the one-dimensional diffusion model. The model reproduces most of the features which can be seen in the data of the latitudinal water distribution during the Martian year.

The model reproduces the timing and magnitude of the peaks in water column abundance during the northern and southern summer, when the ice caps become free of CO<sub>2</sub> ice and release H<sub>2</sub>O into the atmosphere as the temperature rises. The southward propagation of vapour from the North Pole during northern summer and fall is also seen in the model results. The water peak at 150-210  $L_s$  during the northern fall at around 30°N seen in



**Figure 6.8:** An overview of the 6 different experiments for the surface exchange model. The figures show the fraction (between 0 and 1) of the surface that is clean ice as a function of latitude [°]. The blue line shows the ice percentage of the near-surface regolith from GRS for reference, see section 3.6. The red line shows the experiment specific factor which is multiplied with the GRS level. The green line, which is the product of the blue and red lines and partially covers them in the plot, shows the percentage of the surface which is considered to be clean ice in each experiment.

both the TES and MAWD data sets (Figure 6.9, upper and middle panel respectively) is not reproduced to the same degree in the model results. The model results generally show the water to be distributed more smoothly both geographically and in time than the measured data. During the southern fall and winter, there is almost no water present in the southern latitudes according to both TES and MAWD data. In the model results however, there is still a remainder of about  $10 \text{ Pr}\mu\text{m}$  left.

When coupling the surface exchange model with the one-dimensional diffusion model, the modelled results are in much better agreement with the measured data compared to running the diffusion model alone. This illustrates that there is an active exchange between the atmospheric water and the surface reservoirs.

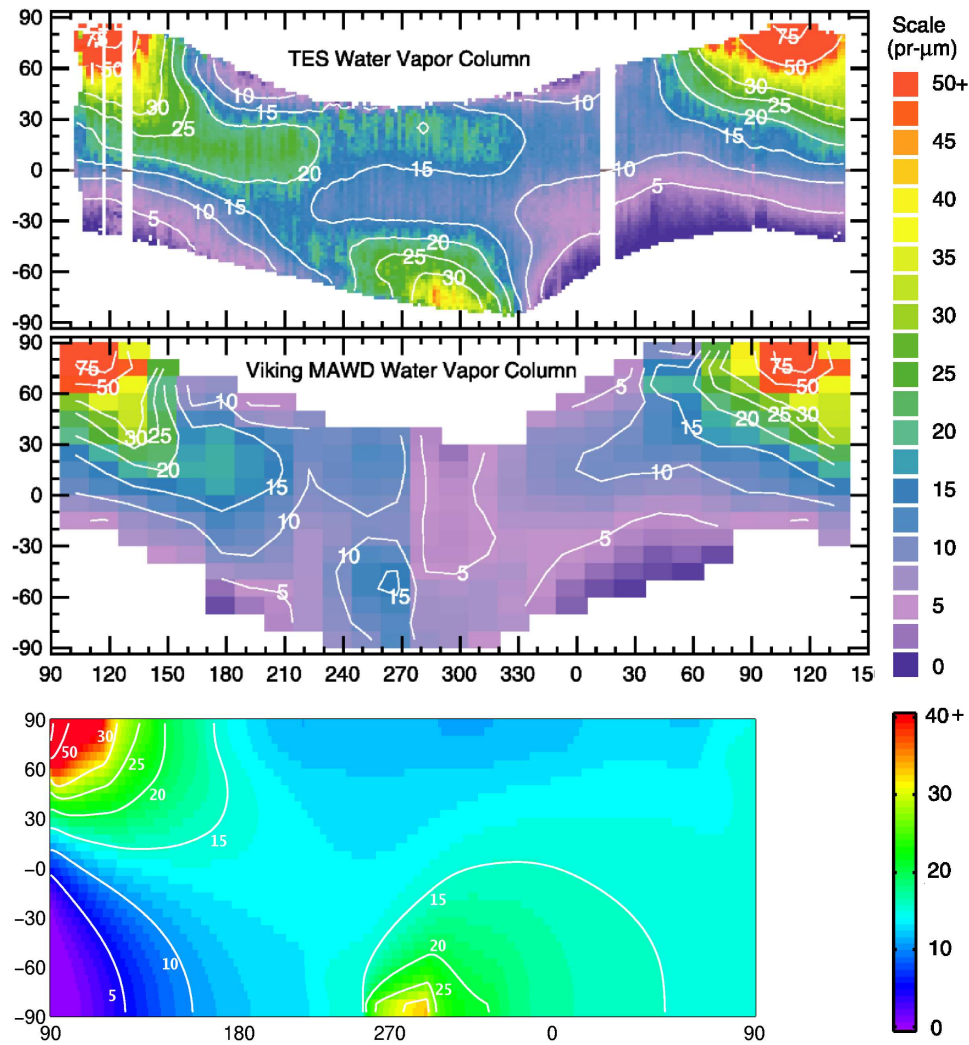
The best fitting results were obtained with the experiment where a diffusion constant of  $D = 5 \cdot 10^5$  and the surface exchange experiment *Poles and GRS  $60^\circ + N$  and  $10\%$  GRS  $60^\circ + S$*  were used.

Figure 6.10 illustrates the total amount of water vapour in the Martian atmosphere during the year. When comparing the model results to the TES data, it is seen that the general level of the amount of water agrees well. The variations on a short timescale in the TES data are not caught in the model, however, the overall trends are still represented. The most marked difference is the large amount of deposition which occurs during the southern fall, as seen in the TES data. This feature is not found to the same degree in the model results.

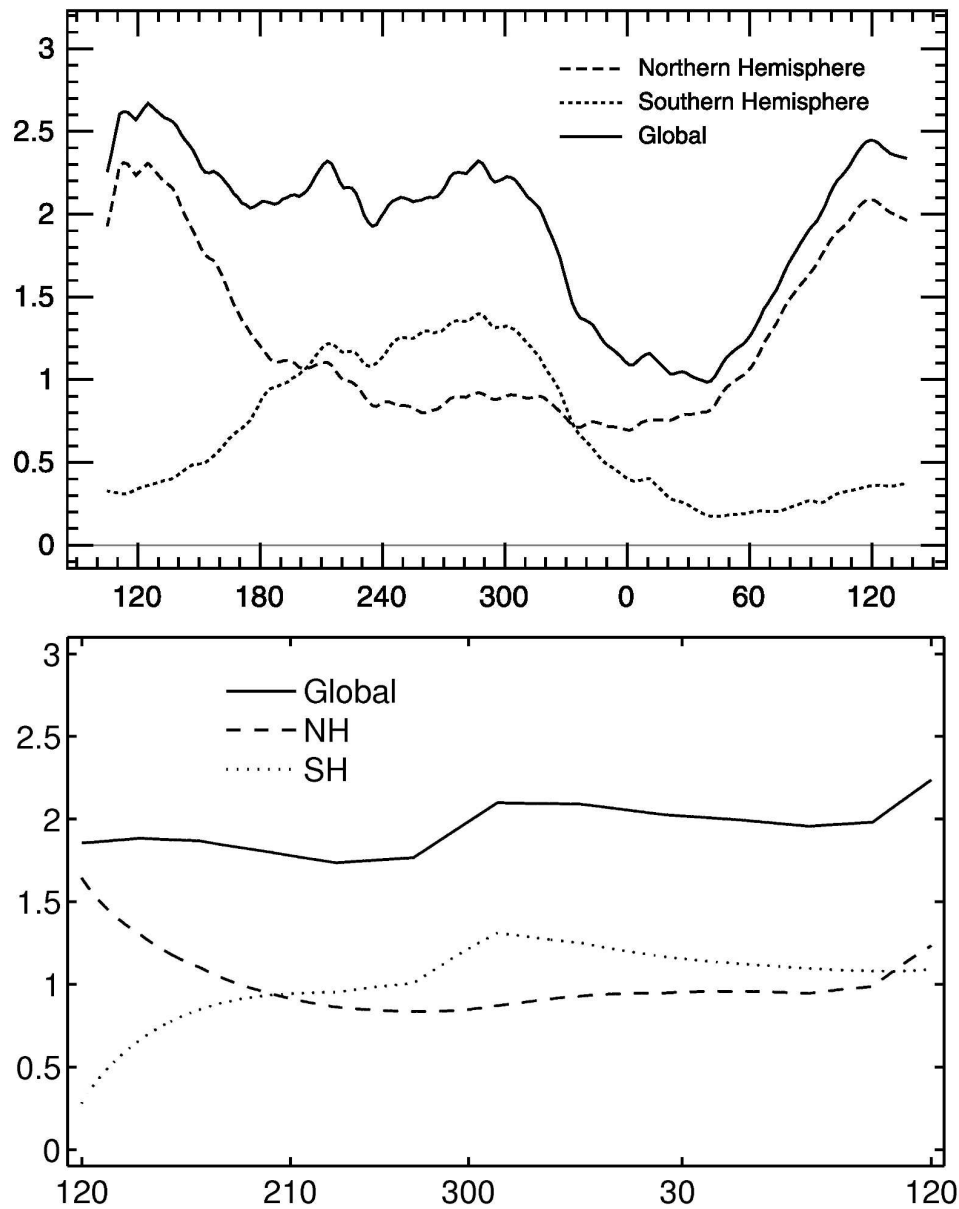
### 6.3.3 Results for Coupling with the Advection Model

The MAWD measurements of the latitudinal distribution of water for the 12 seasons are used as initial values for different model runs. The different initial values correspond to the MAWD measurements for the 12 seasons at which each calculation is started. The calculations were run for 20 Martian years, where the surface exchange was coupled to the advection model.

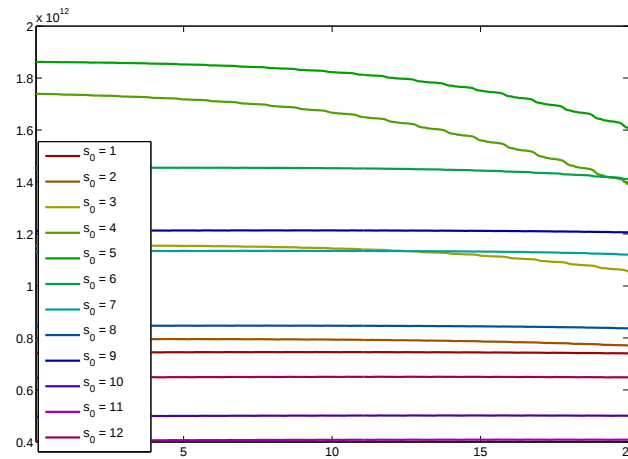
In Figure 6.11 the development of the total mass of water in the atmosphere is shown when only the north polar is allowed to exchange water with the atmosphere. A clear drop in mass is seen for the seasons  $s_0 = 3, 4$ , and  $5$ , and is also seen more subtly for seasons  $s_0 = 2$  and  $6$ . The drop is caused by an increase in the freezing out of vapour as it is transported to the North Pole and deposited there during the winter season. This is illustrated by Figure 6.12, showing the atmospheric exchange of water with the surface for the same calculation. In fact it is the case for all the different initial seasons that the total water content is decreasing in the 20 years period,



**Figure 6.9:** Comparison of the model results with the measured data of the water vapour distribution in the Martian atmosphere. The horizontal axis shows that time of the year [ $^{\circ}L_s$ ] and the vertical the latitude [ $^{\circ}$ ]. The colour bars show the vertical column of water in  $\text{Pr}\mu\text{m}$ . Upper panel: TES water vapour data obtained in 1999-2001 covering a little more than one Martian year. Figure from Smith [2002]. Middle panel: MAWD water vapour data obtained during 1976-1979. Figure from Smith [2002]. Lower panel: Results from the diffusion model coupled with the surface exchange model. The initial conditions correspond to the MAWD vapour data at northern summer, and a diffusion constant of  $D = 5 \cdot 10^5 \frac{\text{m}^2}{\text{s}}$  was used. The surface exchange model was used in the mode "Poles and GRS 60°+N and 10% GRS 60°+S" described in section 6.3.1 and illustrated at the bottom of Figure 6.8.



**Figure 6.10:** Comparison of the model results with the measured data of the total water vapour content of the Martian atmosphere. The horizontal axis shows that time of the year [ $^{\circ}L_s$ ] and the vertical the total mass of water vapour in the atmosphere [ $10^{12}$  kg]. Upper panel: Total amount of water vapour in the Martian atmosphere (solid line), in the northern hemisphere (dashed line) and the southern hemisphere (dotted line). Data from TES [Smith, 2002]. Lower panel: Results for the same plot as above for the same model experiment as illustrated in Figure 6.9.

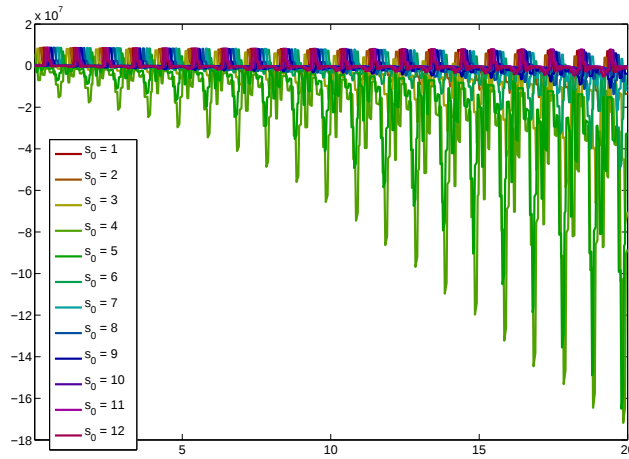


**Figure 6.11:** Total water vapour inventory in the atmosphere for 20 years' programme runs of the advection model with exchange of water with the North Pole box only. The horizontal axis shows the time [year] and the vertical the total mass of water vapour in the atmosphere in kilogrammes [kg]. The legend indicate the season number  $s_0$ .

although most of the cases only do this after an initial increase, see Figure 6.13. This implies that the north polar cap is not by itself able to supply the atmosphere with the water needed for the water cycle to be in balance.

Running a similar experiment with the coupled model, where only the South Pole is exchanging water with the atmosphere, an increase is seen for most of the initial conditions, see Figures 6.14, 6.15 and 6.16. However, the rate of the increase is diminishing, which indicates that as for the north polar exchange scenario there will also be a decrease in the total mass of vapour in the Martian atmosphere over time. The reason for the increase in the atmospheric vapour is the large sublimation occurring in the summer months, which is much larger than the corresponding winter freeze-out as can be seen in Figure 6.15, which shows the amount of water exchanged with the south polar box. In this figure it is also seen that an gradually larger amount of vapour freezes out during the winter, as more of the atmospheric vapour is transported to the southern hemisphere by the wind as described in section 6.2.4 about the advection model.

A simple representation for exchange with the water in the regolith can be made by letting the surface ice cover vary with latitude in the same manner as the ice content of the near-subsurface regolith from GRS measurements does. An experiment was run with the exchange model, which is based on the GRS data shown in Figure 6.8 and described in section 6.3.1. It is seen that these calculations overestimate the amount of water exchanged with



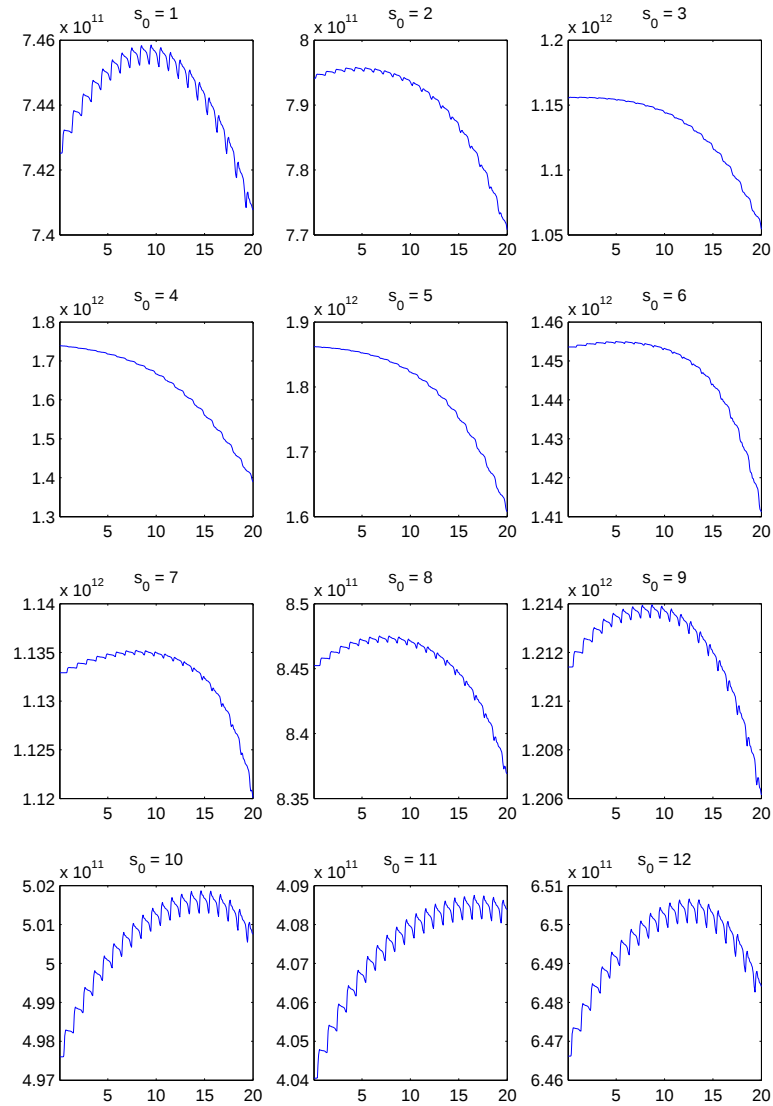
**Figure 6.12:** Surface exchange of water vapour between the the atmosphere and the North Pole only for 20 years' programme runs of the advection model. The horizontal axis shows the time [year] and the vertical the total mass of exchanged water vapour in the atmosphere in kilogrammes [kg]. The legend indicate the season number  $s_0$ . Positive numbers on the vertical axis indicate evaporation from the surface to the atmosphere and negative deposition of ice to the surface.

the surface and that the total amount of vapour in the atmosphere increases (Figure 6.17) mainly due to a high sublimation (Figure 6.18).

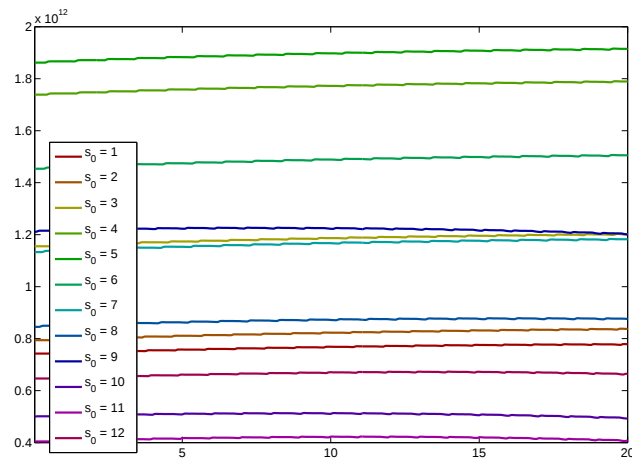
## 6.4 Conclusion

Two new models for transport of water vapour in the Martian atmosphere were presented. The first model was based on the eddy diffusion process related to the Hadley circulation. In the second model water vapour was transported as a passive traced by the wind. A model for exchange of water with the surface reservoirs was coupled to the two transport models and the results analysed.

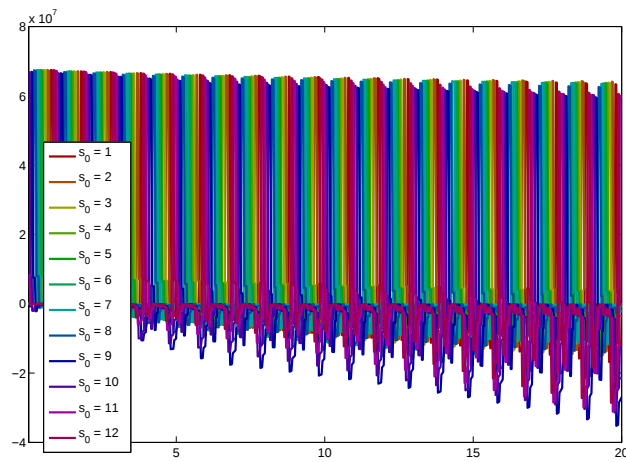
The diffusion model described in section 6.1 showed that depending on the value of the diffusion constant the water in the Martian atmosphere could be distributed globally in the Martian atmosphere. The model is very sensitive to the assumptions on the vertical distribution of the vapour. If the model should be made more realistic, the diffusion constant  $D$  should vary in space and time as is the case for the Hadley circulation, which is represented by the diffusion process in the model.



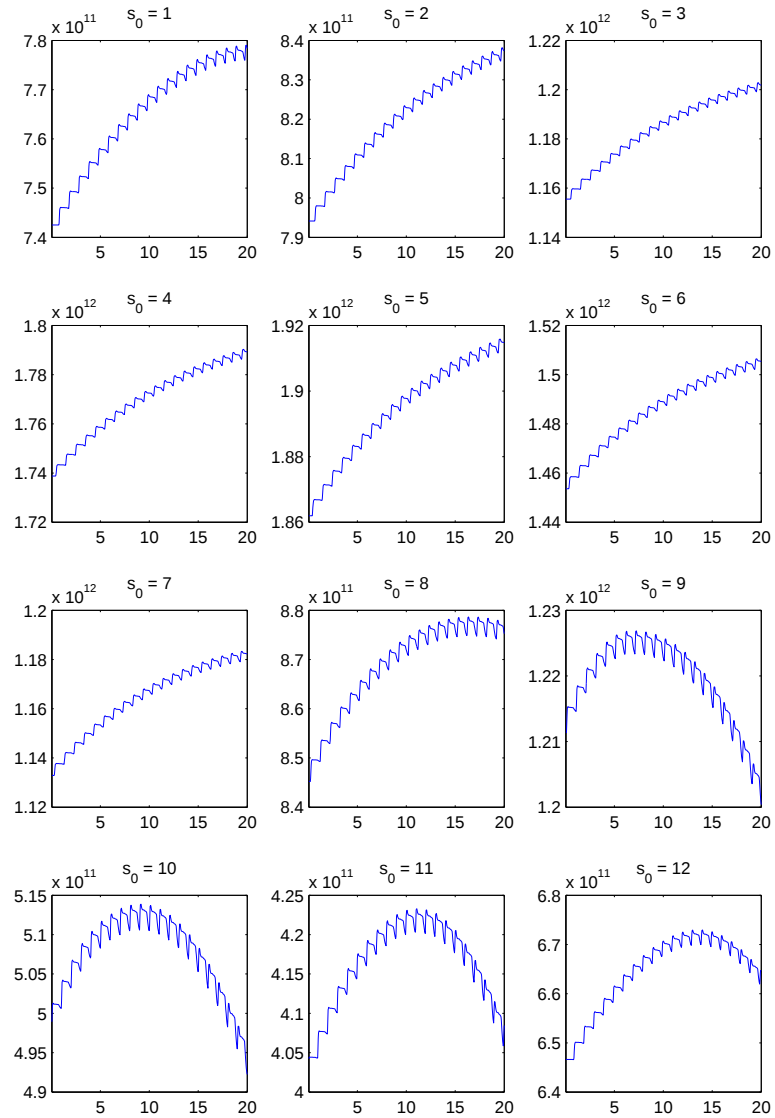
**Figure 6.13:** Total water vapour inventory in the atmosphere for 20 years' programme runs of the advection model with exchange of water with the North Pole box only. The horizontal axis shows the time [year] and the vertical the total mass of water vapour in the atmosphere in kilogrammes [kg]. Same results as in Figure 6.11, only each model run is plotted separately in order to show details of the development over time.



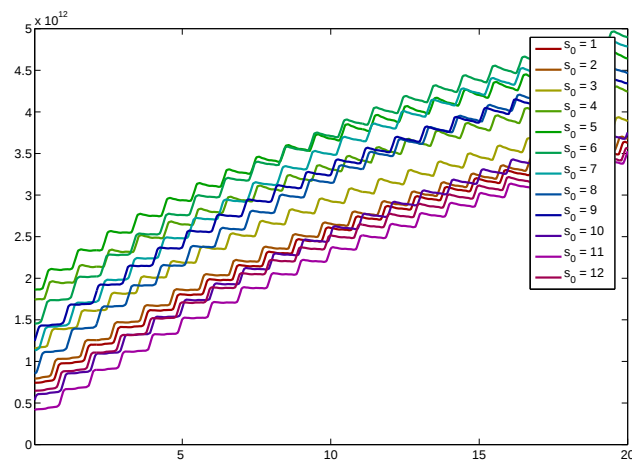
**Figure 6.14:** Total water vapour inventory in the atmosphere for 20 years programme runs of the advection model with exchange of water with the South Pole box only. The horizontal axis shows the time [year] and the vertical the total mass of water vapour in the atmosphere in kilogrammes [kg]. The legend indicate the season number  $s_0$ .



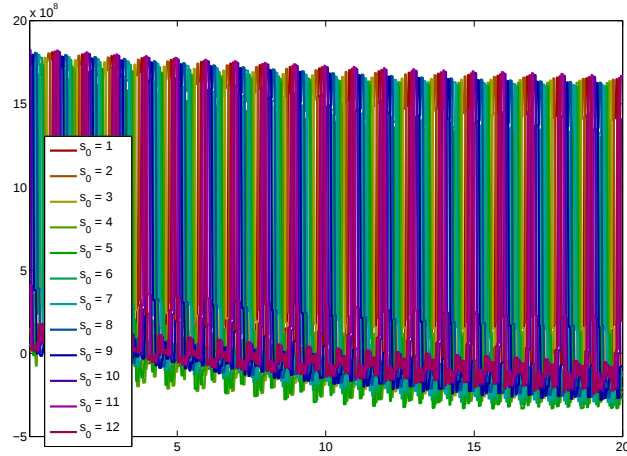
**Figure 6.15:** Surface exchange of water vapour between the the atmosphere and the South Pole only for 20 years' programme runs of the advection model. The horizontal axis shows the time [year] and the vertical the total mass of exchanged water vapour in the atmosphere in kilogrammes [kg]. The legend indicate the season number  $s_0$ . Positive numbers indicate evaporation from the surface to the atmosphere, and negative deposition of ice to the surface.



**Figure 6.16:** Total water vapour inventory in the atmosphere for 20 years' programme runs of the advection model with exchange of water with the South Pole box only. The horizontal axis shows the time [year] and the vertical the total mass of water vapour in the atmosphere in kilogrammes [kg]. Same results as in Figure 6.14, only each model run is plotted separately in order to show details of the development over time.



**Figure 6.17:** Total water vapour inventory in the atmosphere for 20 years' programme runs of the advection model with exchange of water according to the GRS regolith water data from Feldman et al. [2004]. The horizontal axis shows the time [year] and the vertical the total mass of water vapour in the atmosphere in kilogrammes [kg]. The legend indicate the season number  $s_0$ .



**Figure 6.18:** Surface exchange of water vapour between the the atmosphere and the GRS-experiment in the exchange model for 20 years' programme runs of the advection model. The horizontal axis shows the time [year] and the vertical the total mass of exchanged water vapour in the atmosphere in kilogrammes [kg]. The legend indicate the season number  $s_0$ .

The advection model described in section 6.2 showed to be able to transport water vapour across the equator and back again during the annual cycle. The water was primarily collected in a narrow band of latitudes which shifts north and south of equator. Again the vertical exchange is seen to be important, as the results vary considerably with different values for the height of vertical mixing.

The diffusion as well as the advection model are not singlehandedly able to describe the water cycle as seen in Martian data, but need coupling to sources and sinks.

In section 6.3.2 the diffusion model coupled with the surface exchange model showed that it could reproduce most of the characteristics of the measured distribution of water in latitude and time. Some features could not, however, which illustrates the deficits in the simplistic model.

The water is more smoothly distributed in space and time in the model results than in the data sets, which is illustrated by the less pronounced water peak in the northern fall at medium latitudes and the failure to reproduce the removal of water in the southern hemisphere in the southern latitudes during southern fall and winter. These differences between the data and the model results can be explained by two reasons. The first is an insufficient deposition of water from the atmosphere to the surface in the surface model due to the simple mechanism of exchange, which is essentially sublimation of water onto an ice surface. In reality there could be other mechanisms involved as well, including freezing of water into cracks in the surface, which would involve a larger surface area than used in the exchange model; or adsorption of water onto the surface of grains in the Martian soil. The second reason is the nature of the diffusion processes compared to advection. In a diffusion process the difference in concentrations are equalized in time until a steady state is established. The conclusion must therefore be that diffusion alone cannot describe the mechanism for transport of water in the Martian atmosphere, even combined with the possibility for exchange of vapour with surface sinks and sources.

Best results were obtained with the exchange model *Poles and GRS 60° + N and 10% GRS 60° + S*, which resembles the interaction of a water exchange by the ice caps on both poles as well as most of the northern hemisphere and some of the southern hemisphere.

The delay in peak concentration in the atmospheric water, seen in both TES and MAWD data compared to the model results, could be caused by a deeper position of the ice in the regolith in reality than in the surface exchange model. The ground ice likely lies deeper at lower latitudes than in the polar regions due to the warmer annual average temperatures. This

would cause a delay in release of vapour to the atmosphere, as it takes longer time for the temperature wave to penetrate to the deeper lying ground ice.

In section 6.3.3 the advection model was coupled with the exchange model. It was shown that coupling of each of the ice caps alone with the atmospheric transport of water was an unstable situation in time. Thus, the ice caps are insufficient as sources for maintaining the present amount of atmospheric water vapour on their own. When coupling the transport with the exchange model experiment corresponding to the GRS data, the amount of water in the atmosphere would rise relatively quickly to a higher level than measured by the TES and MAWD instruments. This indicates that the water is more bound to the surface reservoirs than the example in the exchange model, where a clean ice surface exchanges water directly with the atmosphere by sublimation and deposition.





# 7

## Conclusion

In Chapter 4 a model was made in order to calculate under which circumstances the north polar ice cap would be in mass balance. The conclusion of the study was that currently the north polar ice cap seems to be stagnant and not flowing. However, there are geological signs that the ice cap has been flowing in the past, so it was investigated under which conditions it would begin to flow. It was found that a more vigorous circulation of water vapour with increased evaporation at lower altitudes and latitudes and enhanced deposition near the pole would result in higher flow velocities in the ice cap.

The water flux in the atmosphere was analysed in Chapter 5 with a statistical numerical method, which enables the seasonal water flux to be separated into different meteorological modes of transport. The study was based on water vapour data from Viking MAWD and atmospheric wind velocities from a climate database based on GCM results. It was found that water moves from the northern to the southern hemisphere at low latitudes, but that no significant net transfer takes place at higher latitudes in the current climate scenario. It was concluded that the atmospheric water at higher latitudes than  $30^\circ$  is currently bound in regions. Furthermore, the water flux study showed that the primary mechanism for the transhemispherical transport of water was the meridional flux, which corresponds to the mean Hadley circulation.

In Chapter 6 two new models for atmospheric transport and mixing of water vapour across latitudes were presented. The models were not capable of

explaining all spatial and temporal details in the measured and mapped characteristics of the vapour distribution in the Martian atmosphere. So it was concluded that the models should be coupled to a surface exchange model representing the interaction of the atmosphere with surface sources and sinks for water vapour.

In Section 6.3.2, an atmospheric diffusion model, representing atmospheric mixing caused by the Hadley circulation on Mars, was coupled with a model where the ice caps and most of the regolith in the northern hemisphere would exchange water with the atmosphere. In general, the results agreed well with the measured data. The magnitude and the general pattern of the spatial and temporal water distribution were reproduced. However, some features were not reproducible, and it was concluded that the model representing an advection process, where vapour is transported like a tracer with the wind, would be better suited to describe the annual water cycle.

The surface exchange model coupled with the atmospheric advection of water vapour in Section 6.3.3 showed that the ice caps alone are not sufficient for retaining the present level of water in the atmosphere. In an experiment based on the GRS measurements of ground ice in the regolith, it was shown that the model using the GRS values as percentages of a clean ice surface would overestimate the evaporation of water to the atmosphere. This indicates that the water in reality is bound more strongly to the reservoirs than in this experiment with the model.

## 7.1 Outlook

It would be beneficial to examine further to what degree the surface and subsurface water reservoirs on Mars are interacting with the atmosphere. A question that would be interesting to answer is: Is all the water on Mars involved in active processes today, or are there hidden protected reserves of water in the subsurface? As time passes, the lighter isotopes of Oxygen and Deuterium in the atmospheric water will be depleted by loss of molecules to space.

Continuing the work from Chapter 5, the lifetime of a water molecule in the Martian atmosphere can be calculated. If combined with fractionation processes for the phase transitions between the different reservoirs, the isotopic depletion of Oxygen and Deuterium in the water reservoirs may be estimated and compared to measurements in the future. Therefore the isotopic composition of the surface and subsurface water reservoirs may enable us to reach an understanding of the water cycle on Mars in the past.

With the Phoenix mission a mass spectrometer will be carried to the high northern latitudes on Mars. Measurements of the isotopic composition will be made of the atmosphere and of the ice samples that will be taken from the permafrost.

With the new data coming from Phoenix and MRO, there will be a wealth of new information. It is a very exciting time for Mars research.



# A

## **Key Planetary and Atmospheric Parameters**

**Table A.1:** Planetary and atmospheric data for Mars and Earth

|                                                                  | Earth                     | Mars                 |
|------------------------------------------------------------------|---------------------------|----------------------|
| Mean orbital radius [ $10^{11}\text{m}$ ]                        | 1.50                      | 2.28                 |
| Distance from sun [AU]                                           | 0.98–1.02                 | 1.38–1.67            |
| Orbital eccentricity                                             | 0.017                     | 0.093                |
| $L_s$ of perihelion                                              | $281^\circ$               | $251^\circ$          |
| Planetary obliquity                                              | $23.93^\circ$             | $25.19^\circ$        |
| Rotation rate, $\Omega$ [ $10^{-5} \text{ s}^{-1}$ ]             | 7.294                     | 7.888                |
| Solar day, sol [s]                                               | 86400                     | 88775                |
| Year length (sol)                                                | 365.24                    | 668.6                |
| Year length (Earth days)                                         | 365.24                    | 686.98               |
| Equatorial radius [ $10^6\text{m}$ ]                             | 6.378                     | 3.396                |
| Surface gravity, $g$ [ $\text{m s}^{-2}$ ]                       | 9.81                      | 3.72                 |
| Surface pressure [Pa]                                            | 101300                    | 600 (variable)       |
| Atmospheric Constituents [molar ratio]                           | $\text{N}_2$ (77%)        | $\text{CO}_2$ (95%)  |
|                                                                  | $\text{O}_2$ (21%)        | $\text{N}_2$ (2.7%)  |
|                                                                  | $\text{H}_2\text{O}$ (1%) | Ar (1.6%)            |
|                                                                  | Ar (0.9%)                 | $\text{O}_2$ (0.13%) |
| Gas constant, $R$ [ $\text{m}^2 \text{ s}^{-2} \text{ K}^{-1}$ ] | 287                       | 192                  |
| Mean solar constant [ $\text{W m}^{-2}$ ]                        | 1367                      | 589                  |
| Bond albedo                                                      | 0.306                     | 0.25                 |
| Equilibrium temperature, $T_e$ [K]                               | 256                       | 210                  |
| Scale height, $H = RT_e/g$ [km]                                  | 7.5                       | 10.8                 |
| Surface temperature [K]                                          | 230–315                   | 140–300              |
| Dry adiabatic lapse rate [ $\text{K km}^{-1}$ ]                  | 9.8                       | 4.5                  |

# B

## Calculation of Areas and Volumes

Derivation of the box boundary areas and box volumes in the models in Chapter 6.

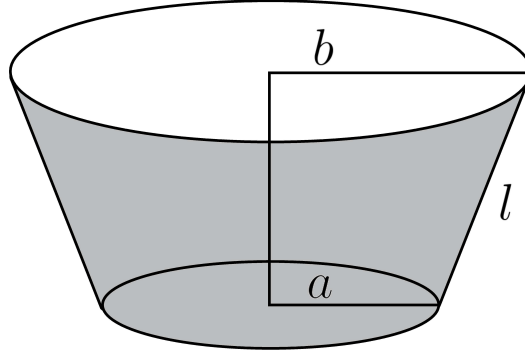
### B.1 Geometrical Variables in the Diffusion Model

The diffusion model described in section 6.1 is one-dimensional along the latitude, and the discretized version of the equations are given for  $N$  latitude points.

The surface area of the boundary of two boxes is given by the slant area of a conical frustum. A conical frustum is created by slicing the top of a cone with the cut made parallel to the base of the cone [e.g. Weisstein, 1999a], see Fig. B.1. Let  $l$  be the slant height and  $a$  and  $b$  the bottom and top radii, then the surface area  $A$  not including the top and bottom circles is

$$A = \pi(a + b)l \tag{B.1}$$

For the geometrical shape of the discretized diffusion equation in section 6.1,



**Figure B.1:** Conical frustum. The surface area  $A$  not including the top and bottom circles is shaded.

the bottom  $a_i$  and top  $b_i$  radii of the conical frustum at latitude  $\phi_i$  are

$$\begin{aligned} a_i &= \cos \phi_i R_M \\ b_i &= \cos \phi_i (R_M + h) \end{aligned}$$

where  $R_M$  is the radius of Mars, and  $h$  is the height of the atmosphere. By insertion into Eq. B.1 the area  $A_i$  between the two boxes  $i$  and  $i+1$  is found as

$$A_i = \pi(2R_M + h) \cos \phi_i h \quad (\text{B.2})$$

$$= \pi [(R_M + h)^2 - R_M^2] \cos \phi_i \quad (\text{B.3})$$

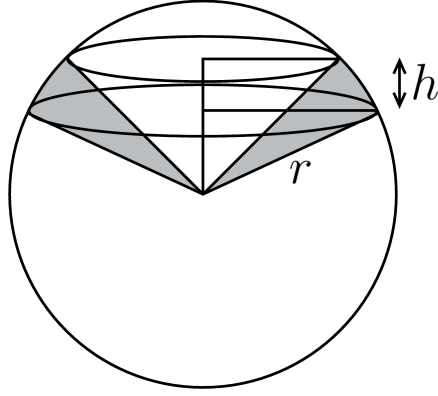
The volume of a box is calculated from the difference between two spherical sectors. A spherical sector is a solid of revolution enclosed by two radii from the center of a sphere [e.g. Weisstein, 1999b]. The volume  $V$  of a spherical sector illustrated in Fig. B.2 is given by

$$V = \frac{2}{3} \pi r^2 h \quad (\text{B.4})$$

where  $r$  is the radius of the sphere and  $h$  is the vertical distance between where the upper and lower radii intersect the sphere.

The volume of the geometrical shape of the boxes for calculation of one-dimensional diffusion in the Martian atmosphere can be calculated by subtracting a spherical sector with a radius corresponding to the surface of the planet from a spherical sector with a radius corresponding to the top of the atmosphere. The distance along the rotational axis between latitudes  $\phi_i$  and  $\phi_{i+1}$  at the surface  $h_{s,i}$  and of the top of the atmosphere  $h_{\text{TOA},i}$  is given by

$$\begin{aligned} h_{s,i} &= R_M (\sin \phi_i - \sin \phi_{i+1}) \\ h_{\text{TOA},i} &= (R_M + h) (\sin \phi_i - \sin \phi_{i+1}) \end{aligned}$$



**Figure B.2:** Spherical sector. The shaded wedges show a cross section in the plane of the paper of the spherical sector.

When the above expressions are inserted into Eq. B.4 the volume of the  $i$ th box is found as

$$V_i = \frac{2}{3}\pi [(R_M + h)^3 - R_M^3] (\sin \phi_i - \sin \phi_{i+1}) \quad (\text{B.5})$$

## B.2 Geometrical Variables in the Advection Model

The advection model described in section 6.2 is two-dimensional in the latitude-altitude plane, and the discretized version of the equations are given for  $N$  latitude points and  $N_z$  altitude points.

The expressions for the surface area between two neighbouring boxes and volume of a box can be generalized from the cooresponding expressions for the one-dimensional case in Eqs. (B.3) and (B.5).

The bottom  $a_{ik}$  and top  $b_{ik}$  radii of the conical frustrum at latitude  $\phi_i$  and altitude  $z_k$  are

$$\begin{aligned} a_{ik} &= \cos \phi_i (R_M + z_k) \\ b_{ik} &= \cos \phi_i (R_M + z_{k+1}) \end{aligned}$$

where  $R_M$  is the radius of Mars and  $z_k$  is the altitude between the  $k$ th and  $k + 1$ th box in the vertical direction of the atmopshere. By insertion into Eq. B.1 the area  $A_{ik}$  between the two boxes  $i$  and  $i + 1$  at the  $k$ th altitude level is found as

$$A_{ik} = \pi(2R_M + z_k + z_{k+1}) \cos \phi_i (z_{k+1} - z_k) \quad (\text{B.6})$$

$$= \pi [(R_M + z_{k+1})^2 - (R_M + z_k)^2] \cos \phi_i \quad (\text{B.7})$$

The distance along the rotational axis between two latitudes  $\phi_i$  and  $\phi_{i+1}$  at the bottom  $h_{k,i}$  and top  $h_{k+1,i}$  of the  $k$ th altitude level is given by

$$\begin{aligned} h_{k,i} &= (R_M + z_k)(\sin \phi_i - \sin \phi_{i+1}) \\ h_{k+1,i} &= (R_M + z_{k+1})(\sin \phi_i - \sin \phi_{i+1}) \end{aligned}$$

When the above expressions are inserted into Eq. B.4 the volume of the  $i$ th box at the  $k$ th altitude level found as

$$V_{ik} = \frac{2}{3}\pi [(R_M + z_{k+1})^3 - (R_M + z_k)^3] (\sin \phi_i - \sin \phi_{i+1}) \quad (\text{B.8})$$

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