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Daniel Tyler
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simulation**

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San Jose State University, 1994

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Evolution of a Martian CO_2 Atmosphere:
A Numerical Simulation

A Thesis

Presented to

The Faculty of the Department of Meteorology

San Jose State University

by

Daniël Tyler Jr.

August 1994

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ABSTRACT

Evolution of a Martian CO_2 Atmosphere:

A Numerical Simulation

by Daniel Tyler, Jr.

This thesis develops an energy balance model to simulate atmospheric evolution on Mars from the end of heavy bombardment until the present day. The model partitions exchangeable CO_2 consistently between three reservoirs: atmosphere, permanent polar caps, and a planetary regolith. Over time, atmospheric CO_2 escapes to space, and is lost to carbonates via chemical weathering.

Numerous sensitivity experiments are performed, and a wide range of initially available CO_2 is investigated. Evolutionary scenarios are inspected for climates that evolve from a hypothetical warm and wet early Mars to that which is presently observed. The model, however, only reaches the temperatures and reservoir storages assumed for the present climate when initialized with low CO_2 amounts (0.5 to 1 bar). Results indicate it is unlikely that early Mars was warm and wet unless the sun evolved differently, or the Greenhouse Effect of the early atmosphere was stronger than presently believed.

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Chapter 1

Introduction

1.1 The Present Martian Climate System

Thanks to the spacecraft explorations of Mars in the 1960s and 1970s, and the Earth based observations which continue today, some fundamental elements of the present martian climate are rather well determined. The martian year lasts 687 earth days, and the martian day (one “sol”) is 40 min longer than Earth’s (McKay 1983). The mean orbital radius of Mars is 1.52 times that of Earth, so the annually available solar energy, in comparison to Earth, is diminished by a factor of $(\frac{1}{1.52})^2$. The present obliquity of Mars (the angle of its rotational axis with respect to its orbital plane) is 25° , similar to the 23.5° terrestrial obliquity, so Mars has earthlike variations in its latitudinal distribution of solar insolation (seasons). However, since the orbital eccentricity of Mars is 5.5 times the terrestrial equivalent (0.093 compared to 0.017),

the solar flux at perigee is 45% greater than at apogee, and because the southern hemisphere summer solstice presently occurs very near perihelion (McKay 1983), there is a substantial asymmetry in the rates at which solar energy is received between hemispheres (especially polar regions).

In comparison to the northern hemisphere seasons, the southern summer is short and hot, and the southern winter is long and cold. Global dust storms begin in the southern hemisphere only, a consequence of the more intense southern summer insolation (Haberle 1986). These storms are initiated by a sharp surface temperature gradient, which exists at the edge of the receding seasonal CO_2 cap. The temperature gradient causes a strong pressure gradient, which results in winds that can pick up dust.

The martian atmosphere is mainly CO_2 (95%), with an annually and globally averaged surface pressure near 7 mbar (Hess *et al.* 1977). Because the obliquity and rotation of Mars are so similar to terrestrial values, the general circulation is somewhat like that of Earth. However, the absence of oceans on Mars leads to much greater differences between the surface latitudinal temperature gradients in the summer and winter hemispheres. Without oceans, there is no large heat sink to buffer seasonal change in the latitudinal distribution of temperature, and the diurnal insolation becomes the main controlling factor. The gradient is very strong in the winter hemisphere, and to first order nonexistent in the summer hemisphere. A jet stream could then only exist in the winter hemisphere, since the surface temperature

gradient is what causes the pressure gradient aloft and establishes the jet through atmospheric dynamics. Subsequently, the winter jets stream would be very strong. An important consequence is that the transfer of equatorial heat into the polar regions (via baroclinic “storms” that feed on the energy of the jet) would only occur in the winter hemisphere (Haberle 1986).

Although similar to Earth in many ways, there are important differences between the climates of the two planets. The Greenhouse Effect (atmospheric absorption and emission of surface infrared radiation, which causes the mean global surface temperature of a planet to be greater than the effective radiating temperature) is much smaller on Mars than on Earth, 3 K in comparison to 33 K (Haberle 1986). In response to a weak greenhouse and diminished insolation, the present annual mean global surface temperature of Mars is only 215 K (McKay 1983), whereas the terrestrial equivalent is 288 K. Present pressures and temperatures are thus low enough to eliminate the possibility of stable liquid water in the contemporary climate, and Mars is thus cold and dry. Moreover, because of sharp latitudinal temperature gradients during the winter, polar temperatures fall below 150 K and reach the frost point of CO_2 , forcing atmospheric mass to precipitate out or condense upon the surface and form seasonal CO_2 polar ice caps. This distinguishes martian climate from that of Earth (and any other planet in the solar system), since a significant percentage of the major atmospheric constituent must leave and reenter the atmosphere as the polar caps form, grow, recede and disappear seasonally.

The frost point relationship of CO_2 (the correspondence between pressure and temperature for which the gas and solid phases exist in equilibrium) plays an important role in the climate of Mars. The CO_2 ice cap at the south pole seems to never completely disappear throughout the martian year, so it is “permanent,” and has important climatological implications. Kieffer and Zent (1991) write, “The condensation of CO_2 , the primary component of the Martian atmosphere, in the coldest regions of Mars means that the total atmospheric pressure is in balance with the temperature of the coldest regions.” The existence of a permanent cap implies that the annual values for the polar surface temperature and pressure must be on the frost point curve, since the CO_2 ice cap must exist in an environment of equilibrium to be permanent. Consequently, the heat balance at the polar surface is critical to the climate, and the existence of a permanent cap allows the heat balance to set the annual surface pressure through the frost point relationship.

Although it seems that the CO_2 cap never completely vanishes from the south pole (therefore permanent), Jakosky and Barker (1984) argue that it did completely sublime in 1969. Their hypothesis is based on a three fold increase of water vapor in the atmosphere during the southern hemisphere summer on Mars during that year. They argue that the water vapor came from ice that was exposed when the CO_2 covering it disappeared. Whether or not the present permanent cap is underlain by water ice is however unknown (Thomas *et al.* 1991). Nevertheless, most observations indicate that the south pole CO_2 cap survives much more than not, so to first order

it seems permanent. As an integral part of the present climatology, we assume that the south polar CO_2 ice cap is permanent.

1.2 Evidence of Climate Change

Although the present martian climate is cold and dry, the possibility that sometime in the distant past it was warm and wet is suggested from various geological features. These features, which appear to require vast amounts of flowing liquid water at the surface, could not have formed in the present climate. However, such evidence does not constitute proof that the climate was warm and wet, as there are other ways that such features could have formed.

Other surface features indicate more recent climate changes, and seem to imply that the general circulation has experienced some long term cyclical fluctuations. In the next two subsections, the evidence which supports such hypothetical climate changes during the distant and recent past is presented.

a. The Recent Geological Past

The polar regions of Mars are fascinatingly complex. The seasonal CO_2 caps form and disappear on top of residual (permanent) ice caps in both hemispheres. In the southern hemisphere, the residual cap is CO_2 ice (as already mentioned), whereas in the northern hemisphere, the residual cap is water ice. The northern residual cap covers a much greater areal extent than does the southern residual cap. Furthermore,

beneath both residual caps, there is another interesting component in the overall structure of the entire cap. Similarly large regions exist where volatiles have been deposited in alternating layers. These layered deposits extend down to 80° in the north and 70° in the south, and Thomas *et al.* (1991) indicate that both regions have approximately the area allowed a 1200 km diameter. The layering of the deposits is important, and is strong evidence that the prevailing transport of volatiles has changed, indicating the climate has changed.

In both hemispheres, the surfaces of these deposits are covered with troughs, canyons with shallow slopes that were most likely formed by aeolian erosion, and sublimation that was enhanced by solar insolation (Thomas *et al.* 1991). High resolution imagery from the Viking and Mariner 9 missions reveals the stratification of the layered terrain best in the equatorial (sun) facing slopes of the trough walls.

Although the dynamics of trough formation, movement and hemispheric asymmetries pose interesting problems (Thomas *et al.* 1991), it is the vertical layering of these deposits, which is most significant in the study of martian climate. In the high resolution images, the color of the trough walls alternates between reddish brown and white. The deposits are generally accepted to be layers of dust and ice (Kieffer and Zent 1991), which is believed to be evidence that obliquity variations, which modify the latitudinal distribution of annual insolation, have changed the global circulation and the transport of dust and water (Pollack 1979; Toon *et al.* 1980; Fanale *et al.* 1982; Francois *et al.* 1990).

Climate during periods of high obliquity would be quite different from periods at low obliquity. At high obliquity, a permanent CO_2 cap would not be able to survive the elevated annual insolation, whereas polar insolation is sharply reduced at low obliquities, increasing the likelihood of a permanent CO_2 cap. If a low obliquity permanent cap holds large amounts of CO_2 , and disappears during periods of high obliquity, the annual surface pressure could fluctuate greatly (Toon *et al.* 1980, Fanale *et al.* 1982, Francois *et al.* 1990, Kieffer and Zent 1991). The deposition of alternating layers can be hypothetically explained due to the distinctly different climates during the high and low obliquity periods.

When seasonal caps recede during years of high obliquity, the thermal gradient at the boundary between the cold cap and the warm surface would generate severe winds. With increased surface pressure at high obliquity, dust would be easily carried aloft by these winds. As the storm develops into a global dust storm, much of the dust would fall upon the seasonal cap forming at the other pole. As seasonal caps disappear in turn, a good portion of the incorporated dust would remain on the polar surface, and over many years a large deposit (layer) would accumulate upon the polar surface.

During periods of low obliquity, a large permanent CO_2 ice cap would form at one or both poles. The annual surface pressure, buffered by the polar heat balance and the frost point relationship, would be very low. The birth of global dust storms would prove much more difficult, if not impossible. Water vapor would diffuse from the plan-

etary regolith (a porous and heterogeneous conglomerate of rubble/rock/gravel/dirt, the upper crust, which adsorbs and desorbs atmospheric volatiles), and become cold trapped in the permanent CO_2 ice cap. Cold trapping is a consequence of the significantly lower H_2O partial pressure over a cold permanent cap, in relation to the partial pressure over ground ice in the equatorial regolith. Cold trapping establishes a concentration gradient, and mass diffuses down the gradient, becoming incorporated within the permanent cap. Moreover, since the diffusion coefficient of a minor atmospheric component is inversely proportional to pressure (Kieffer and Zent 1991), the poleward migration of water will occur more easily at low obliquity than the reverse flow at high obliquity. Thus, a layer of ice will remain on top of the dust layer, and as obliquity cycles repeat chaotically (Laskar and Robutel 1993), a complicated polar deposit would surely be the result.

The primary period of oscillation in the obliquity of Mars is 1.2×10^5 yr. From the number of layers observed in these deposits ($N \leq 20$ layers), and the absence of cratering upon these surfaces, the formation of the terrain appears to be during the recent geological past, the past few million years.

b. The Ancient Geological Past

The oldest terrains which appear on the surface of Mars are the heavily cratered highlands, which are mostly found in the southern hemisphere. Using standard cratering rates to date these surfaces, the ancient highland terrain dates back to the end of

late heavy bombardment, 3.8 Gyr ago, only 700 Myr after the sun became a main sequence star. Evidence of a different ancient climate is found upon close inspection of this terrain; the surfaces are riddled with valley networks which resemble terrestrial river valley systems. The networks appear to have a fluvial morphology, and Baker *et al.* (1991) believe they are clearly the result of surface drainage. Although the networks could result from a variety of mechanisms, it is generally accepted that flowing water was involved (Pollack 1979; Sagan *et al.* 1973). The valley networks of the ancient highland terrain are thus evidence that Mars may have had an epoch when liquid water was stable at the surface.

The physical mechanism(s) which introduce water to the surface, and the time-frame for the network formation, are difficult to determine. However, there are essentially only two mechanisms which are argued to produce surface water; they are sapping (the release of water from a subsurface source) and precipitation. If the valley networks were formed by the surface drainage of precipitation runoff, then global temperatures $T_M > 273$ K are possibly required to support the hydrological cycle. This is not the case with sapping, as it is only important that the 273 K isotherm be near the surface to allow the release of subsurface water, a condition which could have been maintained in a cold climate by a large internal heat source. During an ancient epoch, this heat source could have been provided by the dissipation of the gravitational potential energy of accretion, and/or the release of radiogenic heat from sources within the core. However, Craddock and Maxwell (1993) claim that certain

fluvial features on the sides and near the tops of large crater rims, at locations in the highest altitudes of the ancient heavily cratered terrain, could not have formed without precipitation (an aquifer in such locations is very unlikely), requiring Mars to have had an active hydrological cycle to form at least some of the valleys.

Kahn (1985), Fanale *et al.* (1991) and Pollack *et al.* (1987) all speak strongly for the probability of precipitation on Mars, where planetary surface temperatures near 273 K are an accepted corequisite (how near is a good question, although with high salt content in bodies of water 250 K seems reasonable). Baker *et al.* (1991) alternatively claim that the overall morphology of the valley networks is that of sapping. Quite likely both processes were active, but in comparison to Earth, where a precipitation morphology clearly dominates over that of sapping, sapping might have been the dominant process on Mars (V. Gulick 1993, personal communication).

That Mars has had ground water released to the surface, resulting in the creation of fluvial features, is quite certain. The formation of the largest fluvial features on Mars, Valles Marineris and the associated outflow channels, for example, happened with the aid of catastrophic releases of ground water in a more recent epoch (roughly 2 Gyr ago), a time when climate was likely very similar to the present, and the 273 K isotherm was deep (Baker *et al.* 1991; Carr 1986). Many of the valley networks were probably formed in a similar, although greatly downscaled process, when elevated internal heat fluxes and shallow permafrost allowed a gradual release (sapping) of ground water. This process probably required a recharging of local aquifers, because

the amount of water required to remove such large amounts of mass (to form the valley networks) is unlikely without aquifer recharging (Goldspiel and Squyres 1991; Craddock and Maxwell 1993). Recharging could mean precipitation, or a process that fills the aquifers with juvenile water from the inner mantle of the planet (Carr 1986). Clifford (1993) suggests that the recharging of a *global* aquifer could be driven by the hydrodynamic head created from the basal melting of large polar H_2O ice caps in response to internal heating. He speculates that this cycle could be maintained without requiring rainfall or the corequisite warm temperatures. It is not clear, however, that enough water could be supplied by these processes to form the valley networks without precipitation aiding the formation of at least some of them; and moreover, some networks appear to have branching patterns which are tied to higher surface elevations that do not seem likely to be formed through sapping (Fanale *et al.* 1991), supporting the runoff morphology. The extent and behavior of local or planet-wide aquifers is highly speculative; we really have nothing but suggestive hypothetical morphologies to support the existence of such systems. Precipitation cannot be ruled out, and the evidence does seem to show that it was at least a contributing factor in valley formation.

However the valley networks were formed, the mechanism(s) appear to have shut down quickly, at or just after the end of late heavy bombardment, 3.8 Gyr ago. This is argued due to the wide range of erosional states displayed by the valley networks (Baker and Partridge 1986); some are perfectly preserved while others are highly

eroded. That these networks appear almost exclusively on the most ancient terrain implies that they are all also this ancient (Fanale *et al.* 1991). Why, if all networks are so very old, do they display such distinctly different degrees of erosion? The logical conclusion is that erosion rates were much greater in the relatively short period during the formation of surviving networks, and that they dropped sharply and remained low during the remainder of martian history. That the number of craters under 30 km on ancient highland terrain is so similar to the younger terrain is also believed to imply high early erosion rates (Arvidson *et al.* 1980), since smaller craters would still exist if erosion rates were always as they are today. When taken together, the valley networks and high erosion rates provide strong evidence to support the warm and wet early Mars hypothesis (Fanale *et al.* 1991), since high erosion rates likely require liquid water for rapid chemical or physical weathering (erosion) of surface features.

A massive CO_2/H_2O atmosphere could provide a significant Greenhouse Effect, and possibly a warm early climate, allowing for a hydrological cycle and precipitation. However, to have had such an atmosphere persist over a geologically significant amount of time (after the emergence of the geologic record at the end of late heavy bombardment), would require a mechanism to replace/recycle atmospheric CO_2 that would be lost in the formation of carbonate rocks, a consequence of aqueous chemical weathering as on Earth. The elimination of a 1 bar atmosphere on Mars, as a result of chemical weathering, has been estimated by Pollack *et al.* (1987) to take several tens of Myr at a surface temperature of 273 K. Mechanisms that could possibly main-

tain the massive atmosphere over longer (geologically significant 10^9 yr) time periods are: continued outgassing of juvenile atmospheric CO_2 , and/or recycling by volcanic, tectonic and bombardment processes of the CO_2 lost to carbonates (Fanale *et al.* 1991).

Plate tectonics and volcanism are the mechanisms by which CO_2 is recycled back into Earth's atmosphere, and might have been active in the recycling of CO_2 on ancient Mars. Large impacts could also provide the energy (heat) to free CO_2 from carbonates, or simply enable volcanism by cracking the martian crust. The maintenance of a massive atmosphere, during a period prior to the decline of such processes, is an important and relevant consideration, especially with respect to the amount of CO_2 currently locked in carbonates. Since so much of the surface of Mars dates back to the end of late heavy bombardment, it is logical to assume that there has been very little (if any) tectonic recycling after that time. If such processes were active in maintaining a massive early atmosphere, many questions would need to be addressed. For a more complete discussion of CO_2 recycling, see Fanale *et al.* (1991) or Pollack *et al.* (1987).

A massive CO_2/H_2O atmosphere, with or without other greenhouse gasses, could keep the surface of Mars warm enough today to maintain the stability of liquid water (Kasting 1991, 1993; Pollack *et al.* 1987). Calculations by Pollack *et al.* (1987) show that between 0.75 and 5 bars of surface pressure are needed to raise the global temperature of Mars to 273 K, with this range most sensitively dependent upon

the solar flux received. However, the solar insolation received during the ancient period under consideration would have been much less due to the “faint young sun.” and unless we wish to throw out the standard model of stellar evolution (Sackmann *et al.* 1990), we must accept that when the sun came on main sequence 4.5 Gyr ago its luminosity was most likely 30% less than the present value, having grown monotonically from that time forward. Luminosity increases as a result of changing stellar composition, and other properties, a consequence of internal nuclear reactions (Graedel *et al.* 1991). As the waste products of nuclear fusion fall to the core, the heat of the nuclear “fire” moves outward towards the surface, increasing the luminosity. Since the energy from the sun is what fundamentally warms the surface of all the planets, a 25% decrease in the solar flux 3.8 Gyr ago would require a much more massive CO_2/H_2O atmosphere (Greenhouse Effect) than is required today to warm the surface of Mars. According to Pollack *et al.* (1987), a surface pressure of 4 to 5 bars would have been required.

There is however a problem with continually calling upon more massive CO_2 atmospheres, and the resulting greenhouse warming, to explain elevated surface temperatures for ancient Mars (low luminosity). According to Kasting (1991), at low luminosities and higher surface pressures, there can be such a steep atmospheric lapse rate that the frost point of CO_2 is reached in the upper atmosphere, and CO_2 (atmosphere) will condense. As this transition pressure is reached and exceeded, the Greenhouse Effect will become increasingly less effective at warming the surface. The

latent heat of condensation will warm the upper atmosphere, allowing more heat to radiate to space. But, since the overall radiation balance must be maintained, the surface and lower atmosphere must radiate less heat to space, and this occurs when the temperatures of the surface and lower atmosphere are colder, which is a weakened Greenhouse Effect for that pressure.

When taken to extreme, by hypothesizing the warming of the surface of Pluto with a CO_2 greenhouse, an end result for the trend becomes clear. Solar flux at Pluto's great distance from the sun is so diminished that the surface would radiate very little infrared into the atmosphere, and atmospheric temperatures would be very cold throughout. The entire atmosphere would have temperatures far below the frost point of CO_2 , condense, and reduce the surface pressure essentially to zero. However, for higher luminosities this becomes less and less of a problem, to the point that the Kasting (1991) results agree with the Pollack *et al.* (1987) results (same model without atmospheric condensation) for present luminosity.

Kasting's results imply that an upper limit for the surface pressure of a CO_2/H_2O atmosphere on ancient (3.8 Gyr ago) Mars is 2 bar. Above this pressure, the atmosphere of his 1-D model has become entirely saturated, and clearly any additional CO_2 would condense out. The Greenhouse Effect at this pressure is severely weakened, but could still be used as a valid method to warm early Mars by calling upon other (alternative) greenhouse gasses.

Gasses such as methane, ammonia and sulphur dioxide are very strong greenhouse

gasses (even more so than CO_2), and would greatly enhance the greenhouse warming if they were present in significant partial pressure. However, the immediate problem with these gasses, especially in an environment with the enhanced solar EUV flux of the young solar system (Zahnle and Walker 1982), is that they all have short photolytic lifetimes. Furthermore, there is no obvious source strong enough to counteract their rapid photolytic destruction.

1.3 Problem Statement and Approach

Although the actual evolution of martian climate will surely remain a mystery until we can physically explore the surface (probably even after this time as well), it is highly probable, based upon the geomorphic evidence presented in the previous section, that ancient Mars was warmer and wetter than it is today. This warm climate would have been maintained by the Greenhouse Effect, where the most likely early atmosphere was a CO_2/H_2O combination. This hypothesis (a warm and wet early Mars due to a CO_2/H_2O greenhouse) is supported by many investigators, and is probably the most likely mechanism capable of creating a favorable environment for the formation of the valley networks (Sagan *et al.* 1973; Pollack 1979). The central question in this work then, is whether or not the climate could evolve from this warm and wet Mars, which would have been left after the processes which maintained the atmosphere became negligible, into the climate we observe today? Furthermore, if Mars did have a massive CO_2/H_2O atmosphere, what was its fate; where is the CO_2 today? Answers

Figure 1.1: A cartoon of the major reservoirs and interactive processes required to model the evolution of martian climate. A $\Rightarrow$ implies a CO_2 sink, and a $\Leftrightarrow$ implies an exchangeable process.

(or at least a greater understanding of deeper questions) to these questions come from a more complete definition of the processes and interactions that actually affect the evolution of the climate on Mars, along with computer experiments constructed to simulate the involved processes throughout the evolution of the climate on Mars.

Figure 1.1 is a graphic description of the major processes and reservoirs which are required in such a model. These processes were active from the end of late heavy bombardment (when atmospheric recycling mechanisms became ineffective 3.8 Gyr ago), throughout the evolution of climate, until the present day. The atmosphere is a conduit for the exchange of CO_2 , and is one of three reservoirs which will exchange CO_2 in the climate system. The other two exchangeable reservoirs are the planetary

regolith and the permanent polar CO_2 caps. The remaining two reservoirs are sinks, and once CO_2 has escaped to space, or has become locked within carbonate rocks, it has become lost to the exchangeable reservoirs, and from the climate system.

We have constructed such a model, and although highly parameterized, it draws upon published estimates of reservoir storage capacities, and the relationships of exchange between these reservoirs for Mars. The model uses heat balance principles to predict the polar and equatorial surface and atmospheric temperatures, includes the change of luminosity with time, the dynamic transfer of equatorial heat into polar regions, and the Greenhouse Effect in its calculation. It is mass conserving, such that at every timestep the sum total of all reservoir abundances is identical to that which the model was initialized with, although the exchangeable total (what is actually available to the climate system) decreases from losses to space and the formation of carbonates. In addition, all exchangeable reservoir abundances are calculated to be in equilibrium with the computed temperature structure at every timestep.

The model maintains an equilibrium balance between the surface pressure and the planetary regolith. The expression (balance/process) which partitions CO_2 between the regolith and the atmosphere is highly temperature and pressure dependent. Adsorption and desorption take place fundamentally at the molecular level, where the CO_2 molecules “stick” to available surface areas within the regolith. The binding force is a result of weak electromagnetic fields that are generated from asymmetries in the electrostatic structures of the bonding molecules. Lower temperatures and

elevated pressures both increase the amount of CO_2 that can be adsorbed within the regolith. The amount of CO_2 presently stored in the regolith is not very well constrained, although for our model we assume a nominal value of 300 mbar.

A balance is also established between the surface pressure and the storage of CO_2 in the permanent polar cap(s). This is accomplished by maintaining the atmospheric pressure at the frost point pressure of the polar surface temperature when permanent caps exist. For the present climate, the storage in the permanent cap (the southern hemisphere residual) is believed to be less than 5 mbar of CO_2 (Fanale *et al.* 1982).

The model is initialized 3.8 Gyr ago with an amount for the total available CO_2 , and is run until the present day, while the temperature structure and reservoirs evolve in equilibrium. The most important constraint on the model is that it reaches the present climate at the end of the integration. More weakly however, we constrain the model to have a warm early climate, and look for a rapid change in the climate shortly after the integration begins (possibly a change in erosion rates). In Chapters 2 and 3, the model is developed in detail, and Chapter 4 details the experiments performed with the model (parameter space exploration).

Chapter 2

The Calculation/Computation of Martian Temperatures

2.1 The Gierasch and Toon Model

Development of the climate model used in this study began by reproducing the model (and results) of Gierasch and Toon (1973: hereafter G&T). Immediately prior to the Viking era of martian exploration, it was hypothesized that the permanent polar caps of Mars could contain large amounts of CO_2 (G&T; Sagan *et al.* 1973). If so, their complete sublimation would result in a significant increase in the mass of the atmosphere, as well as the annually averaged surface pressure. The sublimation of residual caps could be driven by increases in the annual polar insolation, a consequence of higher obliquities, and a topic that is explored further (with the model developed

in this study) in the final section of Chapter 4.

With the understanding that the existence of a permanent cap is closely linked to the annual polar surface heat balance, G&T were curious how sensitive the heat balance would be to the atmospheric transport of equatorial heat into the polar regions (dynamic heat transfer), and devised a simple heat balance model to study the effects. The G&T model solves a simplified annual polar surface heat balance for the annually averaged polar surface temperature, and the temperature of the polar winter atmosphere, respectively T_{ps} and T_{pwa} . When their model computed polar surface temperatures which agreed with the frost point of CO_2 , G&T interpreted the solution to be a climate with a permanent polar cap. This is so because, when the annual surface pressure and polar surface temperature are on the frost point curve, CO_2 ice is stable on the polar surface in an annual average.

The G&T model provides a good starting point for the development of the model used in this study, since an understanding of the annual polar surface heat balance is critical when determining if permanent polar caps exist, as well as the amount of CO_2 that is stored there. In the remainder of this section the G&T model is described in detail, and in the final two sections of this chapter, modifications and the addition of a Greenhouse Effect are discussed.

The G&T polar surface heat balance is given by

$$\sigma T_{ps}^4 = S_{ps} + RW + CW + RS + CS, \quad (2.1)$$

where S_{ps} is the average annual absorbed solar radiation at the polar surface. RW is the heat radiated from the winter atmosphere to the surface, CW is the heat gained by the surface from atmospheric conduction in winter, RS is the heat radiated from the summer atmosphere to the surface, CS is the heat gained by the surface from atmospheric conduction in the summer and $\sigma = 5.6703 \times 10^{-8} \text{ W K}^{-4} \text{ m}^{-2}$, the Stefan-Boltzman constant. The surface insolation S_{ps} is determined by $S_{ps} = S_p(1 - A_p)$, where $S_p = 79 \text{ W m}^{-2}$, the annually averaged top of atmosphere polar insolation at present luminosity, obliquity and eccentricity, and A_p is the albedo of the polar cap region. Since G&T did not know the value of A_p , they performed experiments using various S values.

The radiation terms RW and RS are modeled with

$$RW = \frac{1}{2} \epsilon \sigma T_{pwa}^4, \quad (2.2)$$

$$RS = \frac{1}{2} \epsilon \sigma T_{psa}^4, \quad (2.3)$$

where T_{psa} and T_{pwa} are respectively the polar summer and winter atmospheric temperatures, and $\epsilon = 0.1$ is the constant atmospheric emissivity. The conduction terms CW and CS are modeled with

$$CW = \frac{1}{2} p \beta (T_{pwa} - T_{ps}), \quad (2.4)$$

$$CS = \frac{1}{2}p\beta(T_{psa} - T_{ps}), \quad (2.5)$$

where p is the annually averaged surface pressure, and β is a derived transfer coefficient (constant for all computations), that is given by

$$\beta = \frac{c_p C_H V}{RT_0} \approx 1.39 \times 10^{-4} \text{ m(K sec)}^{-1}, \quad (2.6)$$

where $c_p = 880 \text{ J(K kg)}^{-1}$ is the atmospheric specific heat capacity (of CO_2), $C_H = 0.0009$ is a heat transfer coefficient, $V = 5 \text{ m sec}^{-1}$ is an average wind speed above the boundary layer, $R = 190 \text{ J(K kg)}^{-1}$ is the gas constant, and $T_0 = 150 \text{ K}$ is a constant mean temperature. The factors of $\frac{1}{2}$ indicate the half year contribution of the individual terms. G&T further assume that a temperature of 200 K characterizes both T_{ea} (the annual temperature of the equatorial atmosphere) and T_{psa} .

Since T_{ps} and T_{pwa} remain unknown, another equation is required. The equation is found by assuming that during mid-winter, over a time period sufficient to damp out meteorological fluctuations, the atmospheric heat balance can be treated as the sum of three terms. The polar winter atmospheric heat balance is given by

$$Q_R + Q_C + Q_D = 0, \quad (2.7)$$

where Q_R is the difference between the atmospheric absorption and emission of radiant energy, Q_C is the heat gained by conduction from the surface, and Q_D represents the

energy (heat) transferred from the equatorial into the polar region by baroclinic waves, which were known to be present in the general circulation simulations of Leovy and Mintz (1969), but which had not actually been observed.

Q_R is given according to Gierasch and Goody (1967) as

$$Q_R = -8\epsilon\sigma T_{ps}^3(T_{pwa} - T_{ps}). \quad (2.8)$$

Q_C is by definition the negative of CW , since what is lost by the surface to conduction is gained by the atmosphere. Removing the factor of $\frac{1}{2}$, since all terms act over the same time winter interval, we find that

$$Q_C = -p\beta(T_{pwa} - T_{ps}). \quad (2.9)$$

The final term, Q_D , is found by integrating the transient eddy heat flux through the latitude circle at ϕ , and dividing the result, F , by the area of the polar region, the surface over which the heat is assumed to be uniformly distributed. Q_D is then given by

$$Q_D = \frac{F}{2\pi a^2(1 - \sin \phi)}, \quad (2.10)$$

where $a = 3.4 \times 10^6$ m is the radius of Mars. F is found with

$$F = 2\pi a \cos \phi \int_0^\infty \rho c_p \overline{v'T'} dz, \quad (2.11)$$

where $\overline{v'T'}$ is the zonally and time averaged winter transient eddy heat flux, and ρ is the density of the atmosphere. The integral, when evaluated using the hydrostatic approximation, yields

$$\int_0^\infty \rho c_p \overline{v'T'} dz = \overline{v'T'} \frac{pc_p}{g}. \quad (2.12)$$

To evaluate this expression, G&T used a parameterization developed by Stone (1972) to approximate $\overline{v'T'}$, which is given by

$$\overline{v'T'} = 0.216 \frac{R^2}{f^2} \left(\frac{T_0 \bar{S}}{g} \right)^{\frac{1}{2}} \frac{(T_{ea} - T_{pwa})^2}{L^2}, \quad (2.13)$$

where $f = 10^{-4} \text{ sec}^{-1}$ is the Coriolis parameter, $T_0 = 190 \text{ K}$ is the mean temperature, $\bar{S} = 0.002 \text{ K m}^{-1}$ is the static stability, $g = 3.72 \text{ m sec}^{-2}$ is the acceleration of gravity, L is the equator to pole distance and $\phi = 45^\circ$. The approximation yields reliable values for Earth, the planet for which it can most easily be checked, but the validity is less certain (and harder to verify) when used for Mars. However, according to Stone (1972), the approximation agrees well with Mariner observations and the modeling of Leovy and Mintz (1969), so it seems at least a sufficient approximation for heat transfer in the present climate, but the validity of the parameterization over a wide pressure range is tenuous. The amount of heat transferred into polar regions by the actions of baroclinic waves is thus approximated with

$$Q_D = \alpha p (T_{ea} - T_{pwa})^2, \quad (2.14)$$

where the transfer coefficient α is given by

$$\alpha = 0.216 \frac{c_p}{ag} \left(\frac{\cos \phi}{1 - \sin \phi} \right) \left(\frac{R}{fL} \right)^2 \left(\frac{T_0 \bar{S}}{g} \right)^{\frac{1}{2}} \approx 1.49 \times 10^{-6} \text{ m K}^{-2} \text{ sec}^{-1}. \quad (2.15)$$

G&T ran this model over a wide pressure range, using numerous values for the polar surface insolation S , and found solutions to the polar surface heat balance for T_{ps} and T_{pwa} . They plotted polar surface temperatures against pressure, and looked for solutions where the resulting T_{ps} curve intersects with the CO_2 frost point curve. These points of intersection suggest annual climates with permanent polar caps, since the annual polar surface temperature is at the frost point, indicating CO_2 ice is stable at the polar surface in an annual average. The model has one such solution with $S = 18 \text{ W m}^{-2}$, at $p = 6 \text{ mbar}$ and $T_{ps} = 147 \text{ K}$, basically the present annual values for the surface of the south pole residual cap. This implies, by working backwards from 18 W m^{-2} , that the albedo of the polar cap is $A_p = 0.77$.

Paige and Ingersoll (1985) used a 1-D atmospheric model to examine the spring retreat of the seasonal CO_2 frosts in both hemispheres. They found, that to match the radiation balances and sublimation rates observed during the Viking missions, the albedos of the frosts differ between hemispheres. In the southern hemisphere $A_p = 0.82 \pm 0.06$, and in the northern hemisphere $A_p = 0.64 \pm 0.06$. The polar albedo suggested by G&T is, within uncertainty, an average of the Paige and Ingersoll values. Since the polar cap in the model is generic, and not specific to either hemisphere, the value seems reasonable. The albedo of a polar cap, however, is very sensitive

Figure 2.1: G&T results for T_{ps} and T_{pwa} at $S = 18 \text{ W m}^{-2}$.

to its dust content (dust decreases albedo), so the determination of a long term climatological albedo is difficult.

Figure 2.1 shows the temperatures predicted by the G&T model over a wide pressure range. The S-shape of the T_{ps} curve follows the frost point curve with remarkable agreement over a rather wide range of pressure. For large pressures, both temperatures asymptote to 200 K, the assumed temperature of the equatorial and summer polar atmospheres. This high pressure result is caused by heat transport and conduction between the atmosphere and surface. Both conduction and heat transport are pressure dependent quantities; they have a strong influence at high pressures.

There are three solutions with permanent caps (3 intersections with the frost point curve) when the polar insolation is 18 W m^{-2} , but the middle solution near

6×10^3 Pa is unstable, while the other two are stable. A solution (climate) on the frost point curve is stable when the slope of the equilibrium curve is flatter than the slope of the frost point curve, and unstable when it is steeper. The reason for the instability can be understood by imagining a small pressure perturbation in the climate from the middle intersection. A small increase/decrease in atmospheric pressure will cause the system to equilibrate by raising/dropping T_{ps} . The response to this would be sublimation/condensation of the CO_2 polar cap, and the system would move systematically to the higher/lower solution. A similar thought experiment at either of the remaining solutions results in return to that solution, demonstrating its stability.

Historically this was an interesting result, when it was believed that there might be enough CO_2 stored in the polar caps to allow such transitions. However, since the present belief is that residual cap storage of CO_2 is less than 5 mbar (global surface pressure equivalent), the only other reservoir that could be considered to hold such a large amount of exchangeable CO_2 is the regolith. If the atmosphere and polar caps did exchange large amounts of CO_2 at some time, then the question is: where has it all gone? The notion of multiple climatological equilibrium states is quite thought provoking however, and offers a reasonable mechanism to support hypotheses of drastically different climates on Mars.

2.2 The Modified Heat Balance Model

a. Overview of Modifications

An assumption in the original G&T model was that equatorial and polar summer atmospheric temperatures (T_{ea} , T_{psa}) could be characterized by 200 K, and was assumed true at all pressures. We suspect that this assumption caused unrealistically large heat balance contributions from the summer radiation and conduction terms, adding to the sensitivity of the model, and that the sensitivity detected by G&T was not solely caused by the heat transport. We have relaxed this assumption by including a self-consistent energy balance for the equatorial regions, which is similar to the balance used for the polar regions. This modification to the G&T heat balance (the addition of a similar equatorial balance) is described graphically with Fig. 2.2.

Since this work explores the atmospheric connection in the formation of the observed surface fluvial features, the calculation of a mean annual and global surface temperature, T_M , is required. This temperature is typically used when quantifying the probable existence of stable liquid water on the surface (Kasting 1991; McKay and Davis 1990). To approximate T_M , a knowledge of the surface temperature in equatorial regions, T_{es} , is also required. Thus, our modified heat balance calculation will involve a system of five equations and five unknown temperatures: T_{ps} , T_{pwa} , T_{psa} , T_{ea} and T_{es} . These equations are developed in the same manner as the equations of G&T, and the specifics are presented in the next subsection.

Figure 2.2: A cartoon which depicts the factors involved in the modified heat balance calculation. The temperatures and constants represented with the symbols are explained above in the G&T model description.

We also remove the assumption that the atmospheric emissivity, ϵ , has a constant value. This modification incorporates the Greenhouse Effect into the model, which is important since CO_2 (95% of present the martian atmosphere) is a strong absorber in the infrared. The large pressure range considered in this work also suggests a variable emissivity, because pressure broadening becomes important with dense atmospheres. We experiment with two “greenhouses” (emissivity parameterizations) in the model, one weak and one strong. The weak greenhouse causes model global temperatures (T_M) to agree with the Kasting (1991) calculation, and the strong greenhouse causes the temperatures to agree with the Pollack *et al.* (1987) calculation, where atmospheric condensation was not included in the 1-D radiative-convective calculation.

The basis and development of these “greenhouses” is the subject of the next section.

b. Governing Equations of the Modified Model

With the modifications discussed above, the heat balance calculations explicitly determine, for a specific pressure and emissivity, the five temperatures: T_{ps} , T_{pwa} , T_{psa} , T_{es} and T_{ea} . Each temperature has a specific heat balance equation. The heat balance equations of the polar surface and the polar winter atmosphere remain unchanged from G&T. although the notation is modified slightly for clarity by adding a subscript of *ps* to the solar input term and a subscript of *winter* to the atmospheric heat balance terms. An equation is added for the polar summer atmospheric heat balance, with subscripts of *summer*. The three equations required for the heat balance of the polar regions are

$$\sigma T_{ps}^4 = S_{ps} + RW + CW + RS + CS, \quad (2.16)$$

$$Q_{R_{winter}} + Q_{C_{winter}} + Q_{D_{winter}} = 0, \quad (2.17)$$

$$Q_{R_{summer}} + Q_{C_{summer}} = 0. \quad (2.18)$$

The mathematical construct of the terms in the polar summer atmospheric heat balance is the same as for those in the winter balance, except that T_{psa} replaces T_{pwa} .

The terms are

$$Q_{R_{summer}} = -8\epsilon\sigma T_{ps}^3(T_{psa} - T_{ps}), \quad (2.19)$$

$$Q_{C_{summer}} = -p\beta(T_{psa} - T_{ps}). \quad (2.20)$$

Since we have assumed there is no heat transfer in the summer atmosphere, this equation requires $T_{psa} = T_{ps}$. This is more in agreement with the assumption of radiative equilibrium in the summer hemisphere than the constant 200 K assumption of G&T.

Heat balances are required for the annual equatorial surface and atmosphere, and are defined by the following two equations

$$\sigma T_{es}^4 = S_{es} + RA + CA, \quad (2.21)$$

$$Q_{R_{annual}} + Q_{C_{annual}} - Q_{D_{annual}} = 0. \quad (2.22)$$

We follow G&T in treatment of equatorial temperatures, assuming annual values are sufficient. RA and CA are the annual radiation and conduction terms, which represent the exchange of energy between the equatorial atmosphere and surface, and are given mathematically by

$$RA = \epsilon\sigma T_{ea}^4, \quad (2.23)$$

$$CA = p\beta(T_{ea} - T_{es}). \quad (2.24)$$

The factor of $\frac{1}{2}$ is missing from the expressions since each term now acts throughout the entire year. For the annual equatorial atmospheric heat balance equation, the

individual terms are

$$Q_{R_{annual}} = -8\epsilon\sigma T_{es}^3(T_{ea} - T_{es}), \quad (2.25)$$

$$Q_{C_{annual}} = -p\beta(T_{ea} - T_{es}), \quad (2.26)$$

$$Q_{D_{annual}} = \frac{1}{2}\alpha_e p(T_{ea} - T_{pwa})^2. \quad (2.27)$$

The factor of $\frac{1}{2}$ is included in the dynamic transfer term to account for an averaging of its seasonal fluctuation throughout the year. The assumption of a sine squared variation in the rate at which heat is removed from the equatorial region during the course of a year, results in an annually averaged value that is half the winter maximum, thus the factor of $\frac{1}{2}$. The change of α to α_e is required by differences in area between the polar and equatorial regions (a conservation of energy requirement), and α_e becomes

$$\alpha_e = \frac{1 - \sin \phi}{\sin \phi} \alpha, \quad (2.28)$$

where ϕ , the latitude separating equatorial from polar regions, is again assumed to be 45° . This adjustment forces the integrated flux into the polar region to be equal to the integrated flux from the equatorial region, and energy is conserved.

Insolation upon the equatorial surface is calculated in the same way as for the polar surface insolation. We assume $S_{es} = S_e(1 - A_e)$, where $S_e = 178 \text{ W m}^{-2}$ is the annually averaged top of atmosphere equatorial insolation at current obliquity and eccentricity, and A_e is the equatorial albedo. The model is tuned to the present

climatological values of T_{ps} and T_M by adjusting the albedos and the emissivity ϵ .

The mean annual global surface temperature T_M is found by averaging an upper and a lower bound. These bounds are determined with a simple heat balance for the entire surface of Mars. The annual energy radiated away by the surface is assumed equal to that absorbed from atmospheric emission and solar insolation, and is given by

$$\sigma T_M^4 = S_{es} \frac{\pi}{4} + \epsilon \sigma T_{atm}^4. \quad (2.29)$$

The equation is solved for T_M , where either T_{pwa} or T_{ea} are used as the global temperature of the martian atmosphere, T_{atm} . This yields respective upper and lower bounds, and T_M is the average.

c. Effect of Modifications

Figure 2.3 shows the temperatures of the modified heat balance when tuned to yield the present day climatological temperatures. The model was tuned by finding values for ϵ , A_e and A_p (respectively 0.2, 0.2598 and 0.75) that give $p = 7$ mbar, $T_{ps} = 148$ K and $T_M = 216$ K, values which reflect the present climatology. We note that the characteristic S-shape of the polar surface temperature curve is maintained, but there is now only one solution with a permanent cap ($p = 7$ mbar, $T_{ps} = 148$ K and $T_M = 216$ K), the present climatological values. The upper solutions have disappeared with the removal of the G&T assumption that held the equatorial and polar summer atmospheric temperatures invariant at 200 K. The global surface temper-

Figure 2.3: Modified heat balance model results for T_{ps} , T_{es} , T_{pwa} , T_{ea} and T_M . The top two curves are equatorial, and the bottom two are polar. T_{psa} (not shown) is identical to T_{ps} .

ature is essentially constant throughout pressure, as would be expected without a Greenhouse Effect. The temperature of the polar summer atmosphere (not shown) is identical to the polar surface temperature, indicating radiative equilibrium in the summer hemisphere.

Figure 2.4 shows the magnitudes of polar heat balance terms for both the modified and original G&T model. The dominance of the summer conduction term in the original heat balance is unrealistic, since there is no viable heat source during summer capable of maintaining such an enormous temperature difference between the polar surface and the atmosphere. The modified model has no summer conduction, since the polar summer atmospheric temperature is equal to the annual polar

Figure 2.4: The absolute values for the radiation, conduction and heat transfer terms of the polar heat balance are shown for either the original (G&T), or the modified heat balance.

surface temperature. The logical source to boost polar temperatures is heat transferred from equatorial regions, and in the modified model, the heat transfer is nearly balanced in magnitude by the winter conduction of heat to the polar surface. Figure 2.3 shows that the polar surface temperature is still being affected by heat transport at high pressure (still rising), since equatorial temperatures remain high (maintaining the latitudinal temperature gradient which allows heat transport). For the highest pressures, equatorial regions lose enough heat to heat transport that the temperatures finally begin to fall towards the rising polar temperatures. The end result is a Venus-like planet with spatially uniform surface and atmospheric temperatures.

Figure 2.5 goes a little further in a demonstration of the importance of the summer conduction term. It shows the polar surface temperatures of the original G&T model, the original model without heat transport, and the modified heat balance model. The polar surface temperature of the original model is affected very little when the heat transport coefficient α is set to zero (no heat transport). However, the changes realized are significantly larger due to the incorporation of equatorial heat balances, the result of removing the constant 200 K assumption of G&T. We believe this modified heat balance to be an improvement over the original model.

2.3 Greenhouse Effect

With atmospheric emissivity constant, the radiation terms of the model vary only when the atmosphere warms or cools, they are not directly affected by pressure

Figure 2.5: Polar surface temperatures predicted with the original G&T model, the original model without heat transport, and the modified heat balance model.

changes. The radiation terms vary indirectly due to the effects of the pressure dependent terms, conduction and heat transport. When pressure variations cause these terms to change, the temperature structure adapts to maintain the heat balance through the surface and atmospheric radiation terms.

The Greenhouse Effect is included by allowing the atmospheric emissivity to vary with pressure. The dependence of emissivity on pressure is ad hoc, but is found in such a way that the global mean surface temperatures predicted by the modified heat balance match that of sophisticated 1-D radiative-convective models. The emissivities are chosen to tune the heat balance model to the greenhouses of these results, and are not meant to approximate the actual bulk emissivities of a CO_2/H_2O atmosphere. Emissivity does, however, increase exponentially with pressure, and since the

parameterizations are exponential in construct, there is some physical basis to the fit.

Because of the emissivity parameterizations, the global surface temperatures predicted with the modified heat balance are trustworthy. However, the strength of this model lies in its calculation of the latitudinal temperature gradient (critically important for polar cap formation), and that this temperature gradient is consistent (at least in our model) with the global surface temperatures of the 1-D radiative-convective models. This is the instrument desired; and, in combination with a CO_2 partitioning scheme (developed in Chapter 3), a climate evolution model is born. Moreover, because of low computational overhead, the model allows for a full examination of the large parameter space, and is therefore easily used to provide insight into the dynamic behavior of the martian climate system in evolution.

a. The Baseline Greenhouse Effect

The baseline parameterization tunes the modified heat balance to the planetary surface temperatures found by Kasting (1991). His work demonstrated that, in a CO_2/H_2O atmosphere, greenhouse warming can be severely handicapped by the onset of atmospheric condensation. The effect is stronger at lower luminosities, when upper atmospheric temperatures are colder, the lapse rate is steeper, and CO_2 condensation occurs at lower pressures. Kasting finds that, under the assumption of a faint young sun, the surface of a very ancient Mars (4.5 Gyr ago) could not have been warmer than 214 K.

Greenhouse warming of the surface, by increasing the amount of CO_2 in the atmosphere cannot continue indefinitely, since at some levels the atmosphere will become saturated and begin to condense. This condensation of atmosphere will form clouds and/or precipitation, and leave the latent heat of condensation. The latent heat would raise local temperatures and stabilize the lapse rate, and because of warmer temperatures aloft, the amount of heat that escapes to space from the upper atmosphere would increase. However, the radiative equilibrium of Mars must be maintained, so the temperatures of the lower atmosphere and surface will drop, so these levels may radiate less heat to space.

Assuming the effective radiating temperature of Mars is not affected, the difference between it and the surface temperature (the Greenhouse Effect by definition) will grow smaller due to the condensation of atmospheric CO_2 . This is of course a big assumption, since clouds would likely change the planetary albedo. Kasting has ignored such effects due to the lack of information and the uncertainty about the optical properties of CO_2 ice clouds. As always, the omission of clouds, and their poorly understood effects on the planetary radiation balance and albedo, remain an important climatological uncertainty.

Developing an emissivity parameterization to tune the modified heat balance to the global surface temperatures of Kasting (1991) was a multi-step process. The first step was to repeatedly run the model over the pressure range, each time using a different constant value of ϵ for each of four luminosity values: $\mathcal{L} \in \{0.7, 0.8, 0.9, 1.0\}$.

The four S-curves generated in each of these model runs were super-imposed upon the global surface temperature distributions (same 4 luminosity values) of Kasting (1991) (4 curves on top of 4 curves). At some surface pressure, the global surface temperature of the modified heat balance agrees (intersects) with the Kasting value (curve) for the same luminosity. The value of ϵ used for that specific iteration becomes a value which the emissivity parameterization must yield at that surface pressure (pressure where curves intersect) and luminosity. This process was performed repeatedly (many plots), using many different emissivities, until a pressure dependent emissivity distribution became obvious for each luminosity. Using the data collected from the iterative process described above, a mathematical parameterization for emissivity (in surface pressure) is fit to the data collected for each luminosity.

Experimenting with many numerical formulations for the parameterization of the data, we settled upon the following form

$$\epsilon = a_1 \exp\left(\frac{\log p - a_2}{a_3}\right)^2 + a_4 \exp\left(\frac{\log p - a_5}{a_6}\right)^2 + a_7. \quad (2.30)$$

A nonlinear data-fitting routine was adapted from *Numerical Recipes* (1989) to fit each of the four generated sets of data points. The values of the constants, a_1 through a_7 , required to match the Kasting temperatures at each of the four luminosities appear in Table 2.1.

Figure 2.6 shows how well the painstakingly generated data is fit with the emissivity parameterizations. As pressure increases, there is a growing difference between

a_i :	$\mathcal{L}=0.7$	$\mathcal{L}=0.8$	$\mathcal{L}=0.9$	$\mathcal{L}=1.0$
a_1 :	0.0810198	0.2393943	0.2273591	0.3489408
a_2 :	5.3194737	5.8392320	5.7950463	6.1547098
a_3 :	0.7450621	0.9678558	0.8345106	1.1971037
a_4 :	0.2627158	0.3387805	1.5071115	0.4947092
a_5 :	6.4794660	7.2065725	11.3121357	7.9286799
a_6 :	3.0622840	-3.4868860	4.9170232	3.7436397
a_7 :	0.1233138	0.1217479	0.1195329	0.1239510

Table 2.1: The constants required by the emissivity parameterization, which causes the modified heat balance model to yield the mean global surface temperatures of Kasting (1991) (weak greenhouse). $\mathcal{L}$ is luminosity.

Figure 2.6: Baseline emissivity parameterization. The data to fit is marked as follows: $\mathcal{L}=0.7$ (square), $\mathcal{L}=0.8$ (circle), $\mathcal{L}=0.9$ (triangle) and $\mathcal{L}=1.0$ (plus). The solid lines are the emissivity parameterizations for each respective luminosity.

the emissivities which are required by the model at adjacent luminosities. This shows how much effect the condensation of atmospheric CO_2 has on greenhouse warming at higher surface pressures, as the differences become clearly significant for surface pressures above 10^5 Pa (1 bar). During model computations, when luminosity falls between any two of the adjacent curves, a linear interpolation is used to construct an appropriate value for the emissivity.

b. A Stronger Greenhouse Effect

In an earlier work, Pollack *et al.* (1987) also performed 1-D radiative-convective (greenhouse) calculations to determine the dependence of global surface temperatures on surface pressure and luminosity in a CO_2/H_2O atmosphere. As a matter of fact, the work of Kasting (1991) was based on a modification of the model used by Pollack *et al.* (1987). In the earlier work, Pollack *et al.* did not take atmospheric condensation or its effect on the lapse rate into account. Thus, the Pollack *et al.* mean global surface temperatures are warmer, especially so when compared at low luminosities and large pressures. We believe that forming another emissivity parameterization, one which causes T_M to agree with the Pollack *et al.* results, is a valuable exercise.

The Kasting (1991) greenhouse cannot provide a warm early Mars, and if the warm and wet early Mars hypothesis is to remain viable, an alternative greenhouse is required to boost ancient surface temperatures. A possible explanation is that greenhouse gasses other than CO_2 were of significant partial pressure in the ancient

atmosphere (Brown and Kasting 1993; Kasting 1991), allowing the partial pressure of CO_2 to remain below the frost point pressure. The mean global surface temperatures could have been significantly greater at lower luminosities without the need to invoke a massive CO_2 atmosphere, and suffer the inevitable effects of atmospheric condensation. We assume that the warmer temperatures of the Pollack *et al.* greenhouse (stronger greenhouse) may be treated as results from a calculation with such alternative greenhouse gasses. If this assumption is justified (alternative greenhouse gasses actually existed in the ancient atmosphere), then these temperatures are likely an underestimate of the actual global surface temperature.

Two such candidate gasses are ammonia and methane, respectively NH_3 and CH_4 . There is a problem maintaining a significant partial pressure for either gas, as both have short photochemical lifetimes. However, their existence in early martian climate is believed possible (Kasting 1991; Brown and Kasting 1993). The percentage of either gas in the early atmosphere would depend upon a balance between its photolytic destruction and the magnitude of its source. A very speculative hypothesis, one which is attractive to those of us who believe Mars once supported life, is that ancient Mars was inhabited by methanogenic bacteria. [A right brain perspective—we don't know exactly where life came from, and when honest with ourselves we truthfully can't define it either; life is an enigma, and with the perspective of the Gaia Hypothesis (*The Ages of Gaia* 1988), it's easily hypothesized that bacterial life was involved in the maintenance of a warm early Mars (Kasting (1993), as seems probable with early

Earth.] Such organisms could have provided a methane source large enough to maintain a significant partial pressure, even when in competition with strong photolytic destruction.

Forming an emissivity parameterization for this “strong greenhouse” was easier than for the baseline “weak greenhouse,” since there is no assumption of atmospheric condensation. This makes the pressure dependency relatively constant in luminosity, and the fit can be accurate using only a single expression. Through trial and error, the constants of the following expression are adjusted, so when used in the modified heat balance, the T_M values agree very closely with the McKay and Davis (1990) parameterization (a good fit) of the Pollack *et al.* mean global surface temperatures

$$\epsilon = \mathcal{L}^\alpha [\epsilon_{max}(1 - e^{f(p)}) + \epsilon_{min}e^{f(p)}], \quad (2.31)$$

where $\alpha = 0.08$, $\epsilon_{min} = 0.115$, $\epsilon_{max} = 1.035$, $f(p)$ is a function of pressure and $\mathcal{L}$ is the luminosity. The function $f(p)$ is given by $f(p) = \kappa p^\gamma$ where $\kappa = -0.0079$ and $\gamma = 0.39$.

Figure 2.7 shows the global surface temperatures of the modified heat balance model which result when using the two greenhouses, and compares these results with the temperatures predicted using the 1-D radiative-convective models of Kasting (1991) and Pollack *et al.* (1987). Tending to present luminosity ($\mathcal{L} = 1.0$), we see that the two parameterizations converge to yield essentially the same temperatures. This shows the decreasing influence of atmospheric condensation upon the Greenhouse Ef-

Figure 2.7: Global surface temperatures T_M of the modified G&T heat balance model (solid lines), which result when using the different emissivity parameterizations (weak or strong greenhouse). The T_M values are overlain upon the mean global surface temperatures of Kasting (1991) (markers), and upon the McKay and Davis (1990) parameterization (dashed lines) of the Pollack *et al.* (1987) mean global surface temperatures (bottom).

fect, since the results with and without condensation agree for present luminosity. In this light, the stronger greenhouse could also be interpreted to indicate that the partial pressures of alternative greenhouse gasses became negligible between $\mathcal{L} = 0.8$ and $\mathcal{L} = 0.9$, since if they remained in the atmosphere, the greenhouse warming would be even greater at present luminosity. There is as yet no evidence that this interpretation has any validity, and we make no attempt to identify when alternative greenhouse gasses became trace constituents; however, the scenario is surely attractive.

c. Heat Balance Model With Greenhouses

Figure 2.8 shows the present luminosity polar surface temperatures for the original G&T model, the modified model with constant emissivity, and the modified model with both greenhouses. All curves pass through the present martian polar climatology (intersect frost point curve at present annual global surface pressure), since each model has been tuned to these values for present luminosity. The Greenhouse Effect dramatically increases the polar surface temperatures with pressure, and causes them to cross the frost point curve twice, once at a stable point (present climate), and once at an unstable point (temperature curve steeper than the frost point curve).

The existence of both a stable and an unstable solution proves very important when the modified heat balance is coupled consistently with parameterizations that partition available CO_2 between model reservoirs (the complete climate model of this study). When the chemical weathering parameterization decreases the exchangeable

Figure 2.8: The polar surface temperatures, T_{ps} , of the original G&T model, the modified G&T model with $\epsilon = 0.2$, and the modified model with either the weak Kasting (1991) greenhouse, or the strong Pollack *et al.* (1987) greenhouse.

CO_2 total, the surface pressure decreases to a point where the polar surface temperature reaches the unstable solution, and the atmosphere will “collapse.” When this happens, the climate will fall quickly to the lower stable solution, and reservoir abundances and temperatures will all change quickly. Permanent polar caps will form during this atmospheric collapse, as CO_2 condenses upon the surface, and the annual climate moves to the lower stable solution, a process which requires approximately 100 yr (G&T). Climatological states between the two solutions are not possible, except in transition. In the next chapter, the parameterizations which partition CO_2 between the model reservoirs are developed, and the computational logic involved during the model execution is explained.

Chapter 3

Partitioning and Removal of CO_2 in Climate Evolution

Now that the modified heat balance model has been created, allowing the prediction of equatorial and polar temperatures, two steps remain to complete the development of our climate model. The first is to form mathematical parameterizations, which mimic the way in which reservoirs share and exchange CO_2 during climate evolution. This is the other half of the climate model, and the parameterizations used to model these physical processes are presented in the first section of this chapter. The final step is to connect these parameterizations to the modified heat balance model, and the second section of this chapter explains the computational logic (the combination of the two halves) used to do this consistently during model integrations.

3.1 Parameterization of Physical Processes

Looking back to Fig. 1.1, we recall the five reservoirs of CO_2 required to model climate evolution on Mars: atmospheric pressure, planetary regolith, permanent CO_2 polar caps, carbonates and escape to space. The first three are exchangeable reservoirs, and the sum total of CO_2 in these reservoirs is the amount of CO_2 available to the climate system at that time during its evolution. The remaining two reservoirs are sinks, which remove atmospheric CO_2 from the system, and lower the exchangeable (available) total.

In the first two subsections below, parameterizations are developed which simulate the behavior of the atmosphere/cap and atmosphere/regolith equilibrium relationships. The final subsections present parameterizations used to simulate the loss of atmospheric CO_2 by chemical weathering and that which escapes to space. All parameterizations are derived from published estimates of how these reservoirs interact and respond to climate changes. The amount of CO_2 stored in each reservoir is referred to in terms of its global surface pressure equivalent, the amount that the surface pressure would increase (in bars) if all the CO_2 in that reservoir were to become atmospheric.

a. Atmosphere/Permanent Polar Cap Equilibrium

The partitioning of CO_2 between the atmosphere and the permanent cap depends upon the polar surface temperature, the atmospheric pressure and the frost point

relationship of CO_2 . We need to determine when a permanent cap reservoir exists, and the amount of CO_2 stored there when it does. In the model, permanent (annual) polar caps must be free to appear or disappear, and grow or shrink with the changing climatic conditions. Mathematically, if the surface pressure is greater than the equilibrium vapor pressure over CO_2 ice at the polar surface temperature [$p > p_{fp}(T_{ps})$], the excess atmospheric pressure (mass) will condense upon the polar surface, and form a permanent cap reservoir (P_{cap}), or just add to the amount of CO_2 already stored there. When a permanent cap exists, the surface pressure will be equivalent to the frost point pressure at the polar surface temperature $p_{fp}(T_{ps})$, since for a polar cap to be stable climatically, the annual surface pressure must be equal to the vapor pressure of CO_2 over the cap at its surface temperature. Thus, when permanent polar caps exist, the atmospheric pressure is buffered (set) by the frost point relationship.

The partitioning in the model is performed by first recalculating the cap storage, and then resetting the surface pressure, as follows:

$$P_{cap} = P_{cap} + (p - p_{fp}(T_{ps})), \quad (3.1)$$

$$p = p_{fp}(T_{ps}). \quad (3.2)$$

When polar caps exist during an iteration, and the surface pressure is less than the frost point pressure, they will sublime in order to make up as much of the deficit pressure as possible, and the same two expressions control the exchange of CO_2 in

this case. However, if there is too little CO_2 in the cap to raise the surface pressure to the frost point value, the permanent cap must disappear, and the resulting pressure will be such that $p < p_{fp}(T_{ps})$. The following expressions show the symbolic details:

$$p = p + P_{cap}, \quad (3.3)$$

$$P_{cap} = 0. \quad (3.4)$$

Permanent polar caps are thus linked thermodynamically to the annual surface pressure and the polar surface temperature, as interpreted through the frost point relationship and the polar heat balance of the modified heat balance model. The model does not address seasonal changes in polar cap size; the reservoir only records storage in a permanent (residual) CO_2 cap.

The frost point relationship for CO_2 has been included in the model with a “look up table.” This is done using a FORTRAN subroutine, developed to find either the frost point pressure at some temperature, or the frost point temperature at some pressure. The subroutine performs linear interpolation using 160 data pairs taken from the *CRC Handbook of Chemistry and Physics* (1974).

b. Atmosphere/Planetary-Regolith Equilibrium

After an equilibrium partitioning is determined for the CO_2 available between the atmosphere and cap reservoirs, the equilibrium partitioning between the atmosphere

and the planetary-regolith must be re-determined. The model assumes a globally homogeneous regolith, and that the amount of CO_2 stored in this reservoir can be determined from the annual surface pressure and the globally averaged surface temperature, respectively p and T_M . These values are assumed to represent the average pressure and temperature within the planetary-regolith. Furthermore, we assume that the regolith and atmosphere are in equilibrium, which is justified by the timestep used ($\Delta t = 7.6$ Myr); it is much greater than the time required by the regolith to respond to surface pressure and temperature variations (Toon *et al.* 1980; Kieffer and Zent 1991). Thus, the actual kinetics of temperature and pressure diffusion are not required in this parameterization. The timescales of the regolith response are discussed further in section 4.4 with the study of obliquity variations.

The amount of CO_2 stored in the regolith reservoir (P_{reg}) is determined with the following expression

$$P_{reg} = Cp^\gamma \exp\left(\frac{-T_M}{\Delta T}\right), \quad (3.5)$$

where $\Delta T = 60$ K, and $\gamma = 0.275$ (Toon *et al.* 1980; McKay *et al.* 1991). Physically, ΔT is the temperature increase required to outgas e^{-1} of the CO_2 stored in the regolith at that time. Small values of ΔT indicate that CO_2 is weakly bound (large fluctuation in storage from small temperature changes), 60 K is a large value, thus regolith CO_2 is strongly bound. The constant C (computed at model start up) tunes the parameterization, so that at present values of p and T_M , the above expression yields a value for P_{reg} that is believed presently stored in the regolith of Mars. We

assume that at $p = 7$ mbar and $T_M = 215$ K, the nominal baseline regolith storage is $P_{reg} = 300$ mbar. The amount of exchangeable CO_2 in the regolith is not well known, but according to published estimates our nominal value is near the upper end of such estimates (Francois *et al.* 1991; Kahn 1985; Fanale *et al.* 1982; Toon *et al.* 1980). Estimates are highly dependent upon the assumed depth and composition of the regolith, quantities which are not yet known.

When the atmosphere/regolith equilibrium is recalculated, the CO_2 in these reservoirs is partitioned between them so that the sum of the values after re-distribution is equal to the sum of the values before re-distribution, thus conserving mass. This requires an iterative numerical approach. A FORTRAN subroutine determines the appropriate value for p , and when this value is returned to the main program, P_{reg} is easily found using Eqn. 3.5. The subroutine assumes that the first guess for the equilibrium value of the surface pressure is x_0 , where

$$x_0 = \frac{p + P_{reg}}{2}, \quad (3.6)$$

half the total amount of CO_2 stored in the two reservoirs. Using a bisection method adapted from *Numerical Recipes* (1989), x is converged to the value which makes $F(x_i) = 0$. The function $F(x_i)$ is a sum of the atmospheric and regolith storages of CO_2 (at pressure x_i) minus the total amount available for re-distribution, and is

given by

$$F(x_i) = Cx_i^\gamma \exp\left(\frac{-T_M}{\Delta T}\right) + x_i - (p + P_{reg}). \quad (3.7)$$

A bisection iteration begins with the evaluation of $F(x_i)$. When $|F(x_i)| < 5 \times 10^{-8}$, the subroutine returns $p = x_i$, and the regolith storage is computed, re-determining the equilibrium. If convergence has not yet been established, the subroutine uses x_i as either the new upper or new lower bound (depending on whether $F(x)$ is respectively positive or negative), and finds x_{i+1} by averaging (bisecting) the new bounds. The value is checked for convergence, and if necessary the process is repeated.

c. Carbonate Formation and Chemical Weathering

Terrestrial climatologists are familiar with a CO_2 cycle which removes the gas from the atmosphere in rainfall, thus forming carbonic acid. The weak acidic rain falls upon, and weathers the surface silicate rock due to a chemical reaction between the rock and the acid. Through erosion and runoff, the products of this reaction (cations and anions) are washed into oceans, where the dissociated components of the CO_2 , with help from shell-forming organisms, become incorporated within a carbonate rock layer (sedimentary deposit) on the ocean floor. This rock layer is subducted by plate tectonics into the depths of the Earth's mantle, where high temperatures and pressures cause metamorphism of the rock and the eventual release (via volcanic activity) of CO_2 into the atmosphere, completing the cycle.

In the absence of shell-forming organisms, this cycle would still occur. The con-

centrations of cations and anions increase in the runoff basins (oceans) until the formation of carbonates is initiated when a level of supersaturation is reached. Once supersaturation is reached, the rate of carbonate formation is the same as the influx rate of the weathering products. Fanale *et al.* (1991) note that, when considering a steady state reaction, “the net production rate of carbonate rocks, and hence the net loss rate of atmospheric CO_2 , is entirely determined by the rate of continental weathering.” These rates are well determined for the terrestrial cycle. Here, we use our knowledge of the terrestrial rate, and the factors which affect it, to gain a quantitative approximation for the loss rate of atmospheric CO_2 on Mars due to chemical weathering.

We have adopted the rate determining parameterization of Pollack *et al.* (1987), where the rate at which atmospheric CO_2 is lost due to chemical weathering on Mars is proportional to the terrestrial rate

$$RATE = \frac{dM_{CO_2_{Mars}}}{dt} = (g_1 g_2 g_3 g_4 g_5) \frac{dM_{CO_2_{Earth}}}{dt}, \quad (3.8)$$

where the factors g_1 through g_5 are constants which relate the relative strengths or abundance of chemical weathering prerequisites in the martian environment to the terrestrial environment, and M_{CO_2} is the mass of CO_2 in a planet’s atmosphere.

The factor g_1 determines the dependence upon the relative partial pressure of CO_2

in the martian atmosphere to Earth with

$$g_1 = \left(\frac{p}{p_E}\right)^{0.3}, \quad (3.9)$$

where $p_E = 0.33$ mbar, the terrestrial partial atmospheric pressure of CO_2 .

The factor g_2 expresses the dependence upon the amount of runoff, where runoff is assumed to vary with temperature, since both runoff and temperature increase with insolation. This is the evaporation/precipitation connection, and the formulation for g_2 is

$$g_2 = \exp(c_2 \delta T), \quad (3.10)$$

where the constant $c_2 = 0.038$ and the difference in planetary average temperature is given by $\delta T = T_M - 288$ K. The magnitude of g_2 is ultimately limited by the amount of sunlight available to evaporate water, and is thus held ≤ 1 , since the mean solar flux at Mars is always $(\frac{1}{1.52})^2$ times the terrestrial equivalent. This limitation occurs because the flux density from a point source (the sun) is inversely proportional to the distance (d) from the source squared. The total energy emitted by the source must pass through a sphere of surface area $4\pi d^2$, and since flux density is defined as total energy over area, the inverse proportionality is established.

The factor g_3 expresses the dependence of the weathering rate on temperature, and is similarly determined

$$g_3 = \exp(c_3 \delta T), \quad (3.11)$$

where $c_3 = 0.049$ is a constant, and δT is as above.

The factor g_4 is the ratio of cation content in martian rock to that of terrestrial rock. Following Pollack *et al.* (1987), we use a nominal value of $g_4 = 2$.

The factor g_5 expresses the dependence of the weathering rate on the fraction of the planet covered by open bodies of water. Since this factor would be zero if the planet were entirely covered with water (no surface silicates to weather) or if Mars had no surface water (aqueous reactions impossible), the following form is chosen

$$g_5 = a(1 - a), \quad (3.12)$$

where a is the fraction of Mars covered by open bodies of water. We use $a = 0.05$ (the same value used by Pollack *et al.*), and assume this factor remains constant throughout time.

Since chemical weathering is also limited by the physical weathering rate (the rate at which fresh material is exposed at the surface to be chemically weathered), Pollack *et al.* put an upper limit on the product of the physical factors ($a g_1 g_2 g_3$), and require that

$$a g_1 g_2 g_3 \leq 0.71 g_{\max}, \quad (3.13)$$

where 0.71 is the percentage of Earth's surface covered by oceans, and Pollack *et al.* nominally choose $g_{\max} = 30$. Using 47% as an estimate for Earth's continental surfaces covered by silicate rocks, and estimates of the current chemical weathering

of these rocks, the terrestrial atmospheric CO_2 loss rate is determined to be

$$\frac{dM_{CO_2 Earth}}{dt} = 0.0051 \text{ kg m}^{-2}\text{yr}^{-1}, \quad (3.14)$$

which allows the atmospheric CO_2 loss rate on Mars to be given by

$$RATE = \frac{dM_{CO_2 Mars}}{dt} = 0.0051 \left(\frac{g_4 g_5 \mathcal{G}}{a} \right) \text{ kg m}^{-2}\text{yr}^{-1}, \quad (3.15)$$

where $\mathcal{G}$ is the minimum of $a g_1 g_2 g_3$ and $0.71 g_{\max}$,

After the late heavy bombardment period ended in the solar system (3.8 Gyr ago), we assume that the recycling of CO_2 by bombardment and volcanic processes had ended, and that CO_2 lost to chemical weathering from that time onward was permanently lost. Thus, the carbonate rock reservoir P_{carb} is a CO_2 sink from which this atmospheric volatile cannot return to the interactive/exchangeable climate system.

At the completion of each timestep, we remove CO_2 from the atmosphere according to the assumption that $\frac{dp}{dt} \propto p$. Employing a finite difference substitution for the derivative, and manipulating the expression algebraically leads to

$$p_i = p_{i-1} \exp\left(-RATE \frac{g \Delta t}{p_{i-1}}\right), \quad (3.16)$$

where p_{i-1} and p_i are the respective surface pressures before and after chemical weathering, Δt is the model timestep and $g = 3.73 \text{ m sec}^{-2}$. This is the expression used in

the model to reduce exchangeable CO_2 due to chemical weathering. The storage in the carbonate reservoir increases with the decrease of exchangeable CO_2 according to

$$P_{carb} = P_{carb} + (p_{i-1} - p_i). \quad (3.17)$$

The above parameterization was formulated under assumptions that CO_2 is removed from the atmosphere in an environment with rainfall (an active hydrological cycle), and globally averaged surface temperatures > 273 K, clearly not the present climate of Mars. The validity of this parameterization, which is used in the model throughout the evolution of climate on Mars, is therefore very uncertain. However, periods may have existed when it was quite appropriate.

d. Atmospheric Escape of CO_2

The major heavy constituent of the martian exosphere is atomic oxygen (Luhmann *et al.* 1992). These atoms can be ionized by solar EUV, or through charge exchange with solar wind protons and electrons. The resulting oxygen ions can either be picked up by the solar wind electric field and escape the gravitational pull of Mars, or they can reenter the atmosphere and collide with other molecular species at the exobase. Luhmann *et al.* (1992) propose that reentering O^+ ions can add enough energy to a CO_2 molecule, or its atomic constituents, so the whole or parts gain escape velocity, resulting in the loss of CO_2 to space. This process is called sputtering, and Luhmann *et al.* (1992) believe that the effect has caused an integrated loss of CO_2 throughout

the evolution of martian climate that is within the estimates of the total amount of outgassed CO_2 . The rate of atmospheric CO_2 lost to space from sputtering (E), as calculated by Luhmann *et al.*, can be approximated with the time dependent expression

$$E = E_0 10^{-\tau(t-t_0)}, \quad (3.18)$$

where $E_0 = 0.179 \text{ bar yr}^{-1}$, $\tau = 8.57 \times 10^{-10} \text{ yr}^{-1}$ and $t_0 = 0.7 \times 10^9 \text{ yr}$, the absolute age of the solar system at the end of heavy bombardment, 3.8 Gyr ago. The escape reservoir (P_{escape}) grows each timestep according to the expression

$$P_{\text{escape}} = P_{\text{escape}} + E\Delta t. \quad (3.19)$$

When integrated over the 3.8 Gyr period, a total of 164 mbar is removed from the climate system and lost to space by this process. The total amount of CO_2 lost does not vary with the amount initially assumed available; the loss rate is not dependent upon surface pressure, as is the chemical weathering parameterization. However, since solar EUV flux does change, the rate of loss changes in time. Solar EUV flux was much higher in the ancient solar system (Zahnle and Walker 1992), which caused the O^+ ions to have greater energies, increasing the effectiveness of sputtering.

3.2 Model Time-Stepping and Code Flow

The climate model, which has been presented mathematically in Chapters 2 and 3, is coded into a computer program to facilitate a solution. Thus, the modified heat balance model is linked to the parameterizations of the physical processes within the nested loops of a sizeable (1050 line) FORTRAN program. The outermost loop of the program is in time, where each timestep is 7.6 Myr; thus there are 500 steps in each 3.8 Gyr integration. Within this loop, the evolution of solar luminosity is determined, the surface pressure is reduced by chemical weathering, CO_2 is lost to space by sputtering, and the evolving temperatures and reservoir abundances are stored in output files. In the inner loop, the annual climatological equilibrium between the temperature structure and the exchangeable CO_2 reservoirs is achieved at each timestep. This equilibrium is highly non-linear, since changes in temperatures affect reservoir storages, and conversely the changes in reservoir storages affect temperatures through the Greenhouse Effect and the polar transfer of equatorial heat. Sometimes many iterations through the inner loop (>500) are required to establish this equilibrium balance.

The inner looping begins with an atmospheric emissivity calculation. The surface pressure used in this calculation is affected by the weathering and escape losses of the outer loop, and is further subject to the changes brought about by the reservoir equilibrium calculation in each successive “trip” through the inner loop. Using the computed emissivity, a temperature structure is then determined by solving the

modified heat balance system of equations at the same surface pressure and luminosity. This has been done by incorporating software from *Numerical Recipes* (1989) (Newton-Raphson Method for Nonlinear Systems of Equations), which solves the nonlinear system of five equations, yielding values for the five temperature variables. The model then calls a subroutine which redistributes exchangeable CO_2 consistently with the new temperatures. The equilibrium between the atmosphere, the regolith, and the permanent cap reservoirs is determined by first finding equilibrium between the atmosphere and the permanent cap, and then between the atmosphere and regolith (as described in section 3.1). The inner looping continues until the change in atmospheric pressure is small enough so that the absolute value of the fractional change in surface pressure

$$\left(1 - \frac{p_{\text{after}}}{p_{\text{before}}}\right), \quad (3.20)$$

is less than a specified small tolerance, where p_{before} is the surface pressure before the passage of the inner loop and p_{after} is the surface pressure after the exchangeable reservoir abundances have been recalculated. Nominally we require that this error be $< 10^{-4}$. A flow chart of the logical structure of this computation is shown in Fig. 3.1.

There is, however, one exception to this computational logic. It occurs because of an implied limit on the maximum surface pressure, which results when the atmosphere reaches total CO_2 saturation. According to Kasting (1991) the condensation of atmospheric CO_2 can become quite significant at high surface pressures and

Figure 3.1: A flow chart, which depicts the logical structure of how the climate model performs an evolution.

low luminosities. In Kasting's model (1-D radiative-convective), the global surface temperature reaches a maximum value at some large surface pressure (this pressure increases with luminosity). The global temperature decreases for pressures above this value, and the model quickly fails to converge. We assume that the surface pressure at this maximum temperature is the maximum obtainable (p_{max}) for that luminosity, and that the atmosphere is saturated. We have fit p_{max} with a quadratic function in luminosity, and assume that any additional atmospheric CO_2 (surface pressure) which the model desires must precipitate out (due to saturation), and become part of a permanent polar cap reservoir P_{cap} . We require $p \leq p_{max}$, and call this phenomena

the “saturation constraint.” The value of p_{max} is found using

$$p_{max}(\mathcal{L}) = 230.0\mathcal{L}^2 - 326.0\mathcal{L} + 117.1, \quad (3.21)$$

where $\mathcal{L} = \frac{S(t)}{S_0}$ is the luminosity at that time, the ratio of the solar insolation at that time to the present day value. The constants were found by fitting the p_{max} values (determined from Kasting (1991) Fig. 7) at $\mathcal{L}$ values of 0.7, 0.8 and 0.9.

Chapter 4

Model Behavior and Sensitivity

Experiments

This chapter describes the many experiments that have been performed with our climate model. Because of the number of uncertainties in the present state and evolution of climate on Mars, there are similar uncertainties in the initial conditions and constants used with the model. Thus, we are faced with a large parameter space to explore.

One of the most important model parameters is the amount of CO_2 that was available (exchangeable) at the end of late heavy bombardment, when recycling by impacts and volcanism became negligible. There have been many estimates of this value, P_{tot} , and the range of such estimates is a full two orders of magnitude, from tenths to tens of bars (Fanale *et al.* 1991). The parameter is explored by examining

the initial and final states that are predicted by the model (with a wide range of P_{tot} values), determining whether the predicted ancient climate was warm and wet, and whether it evolved to what is presently observed. A warm and wet Mars is defined by global surface temperatures $T_M > 250$ K, and surface pressures $p > 30$ mb (Kahn 1985). Although the specific details of a model integration are highly unreliable due to the numerous parameterizations, if a specific integration ends with the present climate (the primary constraint upon the model), the details suggest change possible during the evolution of climate on Mars.

The first section below discusses the possible evolutions of the baseline model, as it was developed with nominal parameter values and assumptions. The only parameter varied is P_{tot} . The second section is a study of the many other experiments that have been carried out with the model to determine its sensitivity to the parameterized processes. The third section discusses the phenomenon of atmospheric collapse, and variable obliquity is the topic of the final section.

4.1 Baseline Model Behavior

With all parameterizations and associated coefficients as described in previous chapters, we impose a linearly evolving solar luminosity (faint young sun), and keep the obliquity and eccentricity constant at their respective present values of 25.2° and 0.0934. Thus, the top of atmosphere solar insolation is a function of luminosity only. When using the weaker greenhouse of the Kasting (1991) radiative-convective

modeling, the combination constitutes our baseline conditions. We assume that the sun reached the main sequence of stellar evolution 4.5 Gyr ago with a luminosity of $\mathcal{L} = 0.7$. Luminosity, according to the canonical view of stellar evolution modeling (Sackmann *et al.* 1990), is linear from that time onward, and in the model the value is varied according to

$$\mathcal{L}(t) = 0.7 + \frac{d\mathcal{L}}{dt}t, \quad (4.1)$$

where $\frac{d\mathcal{L}}{dt} = (\frac{0.3}{4.5 \times 10^9}) \text{ yr}^{-1}$, and 4.5 Gyr ago $t = 0$.

The baseline model was initialized and run from 3.8 Gyr ago ($t=700$ Myr, the end of heavy bombardment) with a broad range of values used for the total available CO_2 . Three representative reservoir and temperature histories are shown in Fig. 4.1. These choices for P_{tot} provide good examples of the three types of evolution predicted by the baseline model, since we find that the type of evolution predicted is dependent upon the total abundance of CO_2 from which the integration begins. Figure 4.2, which is used to distinguish between the three evolutionary regimes of the model, shows initial and final reservoir storages versus P_{tot} , the initial total CO_2 abundance of the integration. Each of the following subsections discusses model behavior in one of the P_{tot} regimes.

a. Low Abundance: $P_{tot} < 1.5$ bar

With low initial abundance, due to the existence of permanent polar caps, evolutions begin in a buffered state. There is too little CO_2 available in the system for the

Figure 4.1: Evolution of CO_2 reservoirs (left) and thermal histories (right) for initial inventories of 1 bar(top), 2 bar(middle) and 5 bar(bottom). The reservoir curves are: p (solid), P_{reg} (dash), P_{cap} (dot), P_{carb} (chain dot). The temperature curves are: T_{es} (upper solid), T_{ps} (lower solid), T_M (dash) and the frost point temperature T_{fp} (dot).

Figure 4.2: Initial (3.8 Gyr ago) and final (present) reservoir storage versus available CO_2 of the baseline model.

atmosphere and regolith to equilibrate with a polar surface temperature above the frost point. The thermal contributions from the Greenhouse Effect and the poleward transfer of equatorial heat are negligible at the low surface pressure (0.6 mbar) required in the atmosphere/regolith equilibrium. Thus, the temperatures are mostly radiatively determined, well below the frost point, and a permanent cap must form. Since caps form initially in this regime, the only change in the initial state with an increasing total initial abundance is the size of the permanent cap reservoir. This continues until there is enough CO_2 available so that an atmosphere/regolith equilibrium is established with polar temperatures above the frost point, indicating the beginning of the next regime. The amount of CO_2 stored in the permanent cap reservoir can be > 1 bar throughout the entire integration when P_{tot} is at the high end in this regime.

Since this entire mass of CO_2 could be deposited in a single polar region (due to large orbital eccentricity) we must consider the possibility of basal melting, as the pressure at the base of a deep polar cap could rise above the triple point pressure of CO_2 , 5.2 bar. For example, if a 0.5 bar polar cap were to form northward of 60° , the pressure at the base of this cap would be

$$P_{base} = P_{cap} \frac{A_{Mars}}{A_{cap}} = \frac{4\pi a^2 P_{cap}}{2\pi a^2(1 - \sin 60^\circ)} = \frac{2(0.5)}{1 - \sin 60^\circ} = 7.46 \text{ bar}, \quad (4.2)$$

where A_{Mars} and A_{cap} are respectively the surface areas of Mars and the polar cap. This value is significantly larger than the triple point pressure, and if early internal heat fluxes were able to warm the base of the cap to 216.5 K (the triple point tem-

perature), it would not be stable. This is an important point when considering the structure of such massive permanent polar caps. since the internal heat fluxes of early times were possibly 6.5 times greater than today (Postawko and Fanale 1993). Thus, for reasonable values of the involved quantities, it seems likely that basal melting could have occurred. Refer to Appendix B for a quantitative discussion of how these factors interact to initiate basal melting.

As CO_2 escapes to space and becomes lost in carbonate rock during atmospheric evolution, the polar cap reservoir shrinks to maintain the frost point relationship at the polar surface, thus feeding the growing atmosphere, regolith and carbonate reservoirs. For $P_{tot} < 600$ mbar, permanent caps will have completely sublimated by the present time, leaving the climate system unbuffered. With greater inventories the permanent cap persists throughout time, and the only change in the present climate with larger inventories is the size of this reservoir. When a permanent cap exists in the final state, the evolution ends with the present polar and global annual temperatures, and an agreement with the present climate is dependent upon the size of this cap (we have assumed it to be < 5 mbar on present day Mars). Thus, according to this modeling, a possible evolutionary scenario has $P_{tot} \approx 600$ mbar (an evolution similar to the 1 bar case of Fig. 4.1, except with a systematically smaller cap), with a cold and dry Mars throughout the past 3.8 Gyr.

b. Intermediate Abundance: $1.5 \text{ bar} < P_{tot} < 3.2 \text{ bar}$

With intermediate initial abundances there is enough CO_2 available for the atmosphere and regolith to find equilibrium with the polar surface temperature above the frost point, allowing the model to have an unbuffered initial state throughout this regime. Initial atmospheric pressures are approximately three orders of magnitude larger than those when buffered by polar caps, and the atmosphere is also the dominant consumer of additional CO_2 with increasing P_{tot} . The global temperature is increased by the Greenhouse Effect, and polar temperatures rise sharply from heat transport and feedbacks through conduction and greenhouse warming. The initial abundance of the regolith stays approximately constant due to the opposing forcings, $\frac{\partial P_{reg}}{\partial T_M}$ and $\frac{\partial P_{reg}}{\partial p}$. Quantitatively the opposing effects are seen by examining Fig. 4.3, the regolith CO_2 storage contours generated with the model's regolith uptake parameterization. Clearly an increase in global surface temperature causes a decrease in regolith storage, while an increase in surface pressure increases storage, and as seen in Fig. 2.5, these quantities increase or decrease in unison at a rate determined by the strength of the greenhouse assumed. Thus, in this middle regime, initial surface pressure and global surface temperature increase with total CO_2 according to the initial luminosity in the baseline greenhouse of Fig. 2.5.

The final state in Fig. 4.2 shows a discontinuous increase from low to intermediate initial abundance for the rock reservoir P_{carb} , which grows throughout this regime. We find, by inspecting the temperature history of the 2 bar evolution in Fig. 4.1, that

Figure 4.3: Regolith storage of CO_2 , P_{reg} (bars), a function of T_M (K) and p (Pa), as given with the model parameterization.

the global temperature is much warmer than for the 1 bar case, and remains nearly constant through time as well. This indicates some balance between the forcing of rising luminosity (increased insolation) and the shrinking Greenhouse Effect (decreasing p), which allows a nearly constant global surface temperature until the atmosphere collapses at 3.9 Gyr. In combination with the elevated pressure before collapse, the higher global surface temperatures cause a sharp increase in the chemical weathering rate, and a discontinuous growth in the final carbonate reservoir when the model first initializes without a permanent cap.

The atmospheric collapse that happens at 3.9 Gyr in the 2.0 bar case of Fig. 4.1 is significant. Climatological change is the result of this collapse, since *everything* but the size of the carbonate reservoir changes discontinuously at that point in the integration. This result implies that the long term mean climate is influenced by the frost point relationship. Collapse occurs when the surface pressure drops so far that heat transport becomes ineffective in warming the polar region, and the polar surface temperature falls to the frost point. When this happens, the mean climate enters a buffered state with a permanent cap, and there is rapid redistribution of exchangeable CO_2 , that in turn affects the temperature structure. This is equivalent to coming down the T_{ps} curve of Fig. 2.6, and intersecting the frost point curve for the first time (G&T; Sagan *et al.* 1973). But this solution is unstable, as any small pressure or temperature perturbation below these values will cause the system to fall quickly to the lower (stable) intersection point, which represents the climate

the model reaches after atmospheric collapse. Although this type of instability had been explored for episodic climate variation (G&T: Sagan *et al.* 1973; Toon *et al.* 1980; Francois *et al.* 1990), the relevance to long term epochal climate evolution had not been previously proposed. The atmospheric collapse phenomenon is a sensitive aspect of the model behavior, and will be seen throughout the experiments of section 4.2. The phenomenon is also discussed independently in section 4.3.

We note that the final sizes of the atmosphere and regolith in this regime are the same as for low initial abundance, since the final state is again buffered by polar caps. Moreover, at any time after collapse, the atmosphere and regolith contain the same amount of CO_2 as in the low inventory regime. Since the temperatures also agree, the climates are identical, which indicates there is only one possible *buffered* climate at any specific time (luminosity).

That the final carbonate reservoir grows (making the final cap reservoir shrink) with increasing inventory is tied closely to the time of collapse. More carbonates are formed with a later collapse since there is more time when the climate is at a higher global surface pressure and temperature. The collapse happens later with larger inventories, because it takes more weathering to reduce atmospheric mass to a point where the collapse will occur.

c. High Abundance: $P_{tot} > 3.2$ bar

For high initial abundances, the evolutions begin with an artificially imposed buffered state due to the large scale (global) condensation of atmospheric CO_2 . This is the saturation constraint that we have adopted to incorporate the results of Kasting (1991). The constraint is implemented by assuming that permanent polar caps form from the remaining CO_2 which neither the atmosphere nor the regolith are able to hold when the surface pressure is limited to $p_{max}(\mathcal{L}_i)$. The polar caps which form are thermodynamically unstable, since the polar surface heat balance shows an annual gain of energy when the polar surface temperature is held at the frost point due to the imposed caps. We interpret that the physical cause of this energy imbalance is a vertical flux of latent heat, a term not explicitly included in our modeling. When the saturation constraint is in effect, the excess energy received at the polar surface sublimates CO_2 , carrying latent heat to the atmosphere where it is released through condensation. Condensed CO_2 then falls back to the surface, and repeats a cycle which results in vertical heat transport.

For the initial state of this regime in Fig. 4.2, the permanent cap is the only reservoir that changes with increased initial inventory. This is because the surface pressure is fixed at $p_{max}(\mathcal{L}_i)$, which forces the emissivity, all temperatures, and the regolith storage to have specific values for all abundances in the initial state of this regime. Examining the 5 bar case of Fig. 4.1, the surface pressure follows $p_{max}[\mathcal{L}(t)]$ quadratically in time (luminosity) until enough CO_2 has been lost to carbonates

and space so the imposed cap is depleted, and thermodynamic equilibrium is again established. The surface pressure continues to rise with luminosity until the imposed cap disappears. With the rise in p , there is a simultaneous rise in temperature and a substantial increase in weathering, which forces the cap to sublime at an increasing rate. At the moment when the cap disappears, the maximum global and equatorial surface temperatures obtainable in a 5 bar baseline run are reached (237 K and 243 K respectively), seemingly too cold for rain. There is a large discontinuous jump in the polar surface temperature when the imposed cap disappears, indicating the return to thermodynamic equilibrium, where the surface heat balance now determines the polar surface temperature T_{ps} .

There is very little change in the final storages of the atmosphere and regolith reservoirs throughout this regime, as the final state of Fig. 4.2 shows that a growing percentage of the total initial CO_2 becomes sequestered within carbonates. This happens because larger inventories allow the surface pressure to reach a greater maximum pressure before the imposed cap actually disappears, and the weathering rate responds to this quadratic increase in the maximum pressure by causing the removal of CO_2 at much larger rates early on in the evolutions. As the pressure approaches its maximum value, the sharp increase in the slope of the carbonate reservoir curve (5 bar case of Fig. 4.1) shows that CO_2 is locked into carbonates early. As evolutions approach the present day, we find that the strong effect of early chemical weathering has reduced the surface pressure for all total inventories to roughly the same value.

d. Comparison to Observations

The baseline model best yields what we believe are the present reservoir storages when the amount of initially available CO_2 is approximately 600 mbar. Below this value the model does not predict polar caps at the present time, and for larger initial abundances the present day cap reservoir seems too large, since we believe present storage to be < 5 mbar. Atmospheric pressure is low throughout the entire evolution with this value of P_{tot} , and can only grow slowly with luminosity as increasing insolation enhances the polar surface heat balance.

The existence of polar caps throughout evolution means adherence to the frost point relationship at the pole, subsequently requiring the low pressure. Polar caps exist in constantly diminishing size throughout such evolutions, although there is a change in the dominant mechanism for the loss of atmospheric CO_2 . The escape of CO_2 to space by sputtering causes the loss early in the integration, but becomes a weak process by 1.5 Gyr. By the end of the evolution, the dominant process has become chemical weathering. We emphasize that escape and weathering share in the demise of the polar caps throughout time.

More importantly, a warm and wet early Mars, and a rapid shift in climate that might indicate a change in erosional style, are missing from this evolutionary scenario. Moreover, early Mars cannot be warm and wet with the baseline assumptions in effect. The baseline model is incapable of drawing an evolutionary connection between a hypothetical warm and wet ancient climate and the present martian climate.

4.2 Modification and Sensitivity to Baseline Parameters

In the following subsections, the model's response to changes in the parameters or assumptions of the baseline case are discussed. Experiments are performed in which the weathering rate, the regolith capacity and the poleward transfer of equatorial heat are doubled and halved from the nominal value. Knowledge of sensitivity is gained while inspecting the model results for qualitatively different climate evolutions. An alternative to the generally accepted linear evolution of solar luminosity is examined using the results of Graedel *et al.* (1991), who propose a warmer early sun. Another greenhouse model is examined, one that simulates the addition of alternative greenhouse gasses in the atmosphere, thus causing a stronger Greenhouse Effect. The model behavior with a variable polar albedo is also explored.

a. Weathering Rate: *RATE*

The loss of atmospheric CO_2 to carbonates by chemical weathering plays a significant role in the final reservoir storages for the middle and high initial abundances. Since the method used to simulate this loss is so highly parameterized, experiments are performed to examine how climate evolutions differ when the weathering rate computed in the model is halved or doubled. No weathering can take place until after the model finds initial equilibrium, since the global surface temperature and

the atmospheric pressure are needed to find the weathering rate. Therefore, for all total inventories, there is no change in the initial state of the model. The final states however do change, since different rates of carbonate formation will alter the decline of the exchangeable CO_2 total, affecting the amount of CO_2 available through time, as well as the equilibrium between exchangeable reservoirs. The time at which an atmospheric collapse occurs is very sensitive to the weathering rate, but the sensitivity is indirect, as the collapse is a consequence of the polar surface heat balance, specifically the polar surface temperature. The exchangeable reservoir total, which is directly affected by the weathering rate, affects the polar heat balance through the pressure dependent polar heat transfer and atmospheric emissivity (the Greenhouse Effect) terms.

Evolutionary differences are seen when inspecting the 1, 2 and 5 bar pressure and temperature histories, in Figs. 4.4 and 4.5 of the respective halving and doubling experiments, when compared to the baseline case of Fig. 4.1. There is very little change in the evolution of the 1 bar case. The only difference between these experiments and the baseline is the amount of CO_2 in the polar cap and carbonate reservoirs, since what is lost by one becomes picked up by the other. With increased weathering, the caps sublime more rapidly to compensate for the more rapid loss of atmosphere, while for decreased weathering the caps sublime more slowly, which also only affects the ongoing and final totals of these two reservoirs. Atmospheric pressure, regolith storage and the temperature structure are the same as the baseline in both

Figure 4.4: CO₂ reservoir and temperature histories, as in Fig. 4.1, for a halved weathering rate: $RATE = \frac{RATE}{2}$.

Figure 4.5: CO_2 reservoir and temperature histories, as in Fig. 4.1, for a doubled weathering rate: $RATE = 2(RATE)$.

experiments, since the existence of polar caps in equilibrium requires a specific annual polar surface climate. The polar heat balance can only satisfy this requirement at a surface pressure when the modified heat balance predicts that T_{ps} is at the frost point.

Evolutions become more complex when the model can initialize without a permanent cap. With a halved rate the 2 bar case will not collapse, whereas for a doubled rate, the 2 bar collapse happens 2.1 Gyr earlier than with the baseline rate. The baseline 5 bar case is just about to collapse at the end of integration, as seen by the approach of T_{ps} to the frost point temperature in Fig. 4.1. With a halved weathering rate, the 5 bar case is far from collapse at the end of integration, although for a doubled rate, collapse occurs well before the end of integration at 2.9 Gyr. These changed evolutions, as seen through the lens of atmospheric collapse, are actually caused by interactions between the reservoirs, the polar heat balance and increasing solar luminosity.

When CO_2 is removed at increased rates to carbonates, the pressure and the polar surface temperature will fall faster than with the baseline rate value. This is controlled through the climatological equilibrium required between the atmosphere and regolith reservoirs. These reservoirs share in the loss of exchangeable CO_2 , and the equilibrium determines the actual drop in surface pressure, which controls the loss of warming in the polar heat balance from heat transport and greenhouse forcings. The involved feedbacks are positive (a pressure drop causes a temperature drop which

causes a pressure drop), since the capacity of the regolith to store CO_2 increases with decreasing global surface temperature. The result is a rapid decrease of the polar surface temperature to the frost point, atmospheric collapse, and finally a buffered climate system. When the loss rate is half the nominal value, the pressure decreases more slowly, while increasing solar luminosity opposes the heat transport transport and greenhouse feedbacks, allowing T_{ps} to remain above the decreasing frost point, which eliminates the collapse phenomena in the 2 and 5 bar evolutions, as seen in Fig. 4.4. If the weathering rate is halved, the atmosphere cannot collapse for any initial inventory. But when doubled, the atmosphere always collapses when the initial inventory is large enough to cause an unbuffered initial state. In Fig. 4.6 the final model states predicted by halving and doubling the weathering rate are shown.

A collapse regime, where collapse has taken place sometime during model integration, is determined by examining the final permanent cap storage of Fig. 4.6. Where this curve resembles a decaying exponential, the model atmosphere has collapsed sometime during integration. Of significance is the disappearance of the discontinuity in P_{carb} with the doubling of the weathering rate. This is due to the effect of increased weathering as a mechanism to push the possible collapse times back to immediately after model initialization. A similar effect will be demonstrated later with the stronger greenhouse, which indirectly causes increased weathering by elevating the global surface temperature T_M .

The final size of the carbonate reservoir responds nonlinearly to the value of

Figure 4.6: Reservoir storages at the final model timestep (present time) versus P_{tot} for the weathering rate halving and doubling experiments.

the weathering rate. For evolutions without collapse (low initial abundances), the response is linear as can be seen by noting the final carbonate storage in each of the 1 bar evolutions of Figs. 4.1, 4.4 and 4.5; it increases with the weathering rate. The evolutions of p and T_M are identical since polar caps exist and the frost point relationship determines the climate. The final size of the carbonate reservoir in this low P_{tot} regime can then only be affected by a change in the weathering rate. However, for the 2 bar evolutions, the baseline case has a greater final carbonate reservoir than either of the other experiments. Moreover, the final P_{carb} value with a halved weathering rate is greater than when the rate is doubled. The reason for this is linked to the time at which atmospheric collapse occurs. When collapse happens early, the atmosphere spends little time at higher pressures, and the pressure and temperature dependencies of weathering limit the CO_2 which goes to carbonates. When the collapse takes place later, there is more time at higher pressures, and even a slower rate of carbonate accumulation adds up to a greater final carbonate reservoir. This nonlinear response only appears when P_{tot} is above the low abundance regime, when collapse is possible.

b. Regolith Capacity

The present (and past) capacity of the regolith for holding exchangeable CO_2 is not well constrained. The regolith is, however, a major reservoir for exchangeable CO_2 , and the amount of CO_2 in the regolith will significantly affect the evolution of climate.

To determine how sensitive the qualitative nature of model evolutions is to our choice for contemporary regolith capacity, we perform two experiments with the model, one with half the nominal 300 mbar present storage, and one with twice this value. Changes in the model initial conditions are mostly shifts in the P_{tot} value where the permanent cap will discontinuously begin or cease to exist. These discontinuities in the P_{cap} curve occur at the beginning and the end of the collapse regime, as seen in Fig. 4.2 of the baseline evolutions. Figures 4.7 and 4.8 show initial and final reservoir sizes for the respective halving or doubling experiment. Inspection of these figures shows that with decreased/increased regolith capacity there is more/less CO_2 available to the atmosphere, and the model can initialize without caps at a smaller/greater total inventory. This shows again how polar surface temperatures respond to changes in heat transfer and greenhouse warming with different atmospheric pressures. The P_{tot} value at the end of the collapse regime is simply a result of the relationship

$$p = P_{tot} - P_{reg} = p_{max}(\mathcal{L}_i) = 1.914 \text{ bar}, \quad (4.3)$$

since the saturation constraint imposes a permanent cap if the model attempts to initialize with a larger surface pressure. With increased/decreased regolith capacity, this expression is satisfied at greater/smaller initial inventory.

For the final state with low P_{tot} (entirely buffered evolutions), the regolith capacity can only affect the final size of the polar cap and regolith reservoirs. Increased/decreased regolith capacity means decreased/increased final P_{cap} , and what

Figure 4.7: Same as Fig. 4.2, except with regolith capacity for present climate halved (150 mbar).

Figure 4.8: Same as Fig. 4.2, except with regolith capacity for present climate doubled (600 mbar).

the final cap loses/gains the final regolith gains/loses. The final state of the model is more complicated with middle and high initial abundances.

The possibility of atmospheric collapse during model evolutions is greatly affected by the regolith capacity with middle and high initial abundances. Halving the regolith capacity pushes the beginning of the collapse regime back to total inventories just greater than 1 bar in Fig. 4.7, and also causes collapse to occur for all greater P_{tot} . With the nominal regolith capacity, a collapse could not occur after the onset of the saturation constraint. By doubling the regolith capacity in Fig. 4.8, the collapse regime disappears entirely.

Halving the regolith capacity allows for the partitioning of more CO_2 to the atmosphere initially, which elevates temperatures through atmospheric feedbacks, and sharply increases the amount of CO_2 lost to carbonates. The ability of the regolith to buffer the ongoing exchangeable CO_2 total is diminished, and the combined effect is an earlier collapse, since surface pressure will now drop more rapidly. In the collapse regime, the halved regolith capacity yields smaller present permanent cap size than with the baseline capacity. This is desirable, but a 200 mbar permanent cap is still too much larger than the < 5 mbar we assume that is presently stored there. Conversely, doubling the regolith capacity removes the collapse phenomena from the evolutions entirely. Doubling the capacity aids the ability of the regolith to buffer the exchangeable total, providing a CO_2 source to replace atmospheric CO_2 that is lost to chemical weathering. This maintains atmospheric pressure, and the warming

which results from pressure feedbacks prevents the collapse.

These experiments with the regolith capacity show that the qualitative behavior of model evolutions is unchanged. Initial pressures and temperatures are still limited by the saturation constraint, again eliminating the possibility of a warm and wet early Mars. And for the model to reproduce the presently believed climate and reservoir storages, the only scenario is again a low pressure buffered evolution.

c. Dynamic Transfer of Equatorial Heat to Polar Region

The accuracy of the heat transfer parameterization is suspect. Although it seems to agree with present martian climatology (as mentioned earlier), accuracy throughout the wide atmospheric pressure range of this study is unlikely. To determine how this parameterization for the transient eddy heat flux affects the qualitative behavior of model evolutions, two experiments were performed, a halving and a doubling of the transfer coefficient α . We see from Figs. 4.9 and 4.10, the initial and final reservoir storages of the respective halving and doubling experiments, that the change which appears in the initial states is a shift in the initial total abundance at which the model will initialize without caps.

Halving/doubling the dynamic transfer coefficient α forces the model to require an increased/decreased total inventory before it is capable of initializing without caps. When halving the transfer coefficient, the polar winter atmospheric temperature is less due to the loss of transferred equatorial heat. The surface cools, finding equi-

Figure 4.9: Same as Fig. 4.2 except with dynamic heat transfer halved ($\alpha = \frac{\alpha}{2}$).

Figure 4.10: Same as Fig. 4.2 except with dynamic heat transfer doubled ($\alpha = 2\alpha$).

librium with the resulting diminished winter conduction and radiation terms in the annual polar surface heat balance. The collapse regime is seen to include all values of P_{tot} above that for which polar caps initially disappear, a consequence which results directly from cooler polar surface temperatures. This effect is also the reason that the final atmospheric pressure falls below 5 mbar when a permanent cap exists, as the frost point relationship must be maintained, and lower polar surface temperatures require lower surface pressures.

With the doubled heat transfer coefficient we find a very different final state in Fig. 4.10; there are no polar caps for any initially available amount of CO_2 . Even all the low P_{tot} integrations, which begin with polar caps, lose them when polar regions are warmed to temperatures too great to allow an equilibrium with caps. This result is caused by the combined effects of rising luminosity and the doubled energy from heat transfer. The polar energy balance is clearly sensitive to heat transfer.

d. Warm Early Sun

A possibility which might allow the early climate on Mars to have been warm and wet is that the luminosity of the sun was greater than that determined using conventional stellar evolution models (linear increase of luminosity in time). A fact, which justifies the search for an alternative to the canonical belief about the evolution of luminosity, is that the conventional models do not agree with the observed depletion of solar lithium. Solar lithium is actually 10^2 times less than predicted with the conventional

stellar evolution models (Boothroyd *et al.* 1991).

Lithium is not synthesized within stars, and mechanisms which could produce it on stellar surfaces are negligible. thus connecting the observed abundance of lithium on stellar surfaces (photosphere) to a cosmological distribution at the time of stellar formation (Dravins *et al.* 1993). Furthermore, stellar surface material is mixed over time into the deeper hotter layers where lithium burns. Therefore, the amount of lithium in the photosphere of a star will decrease with time, which allows the lithium abundance in the photosphere to be used as a means for determining stellar age. For a general review see Michaud and Charbonneau (1991).

Graedel *et al.* (1991) propose a different evolution for the solar luminosity. Their results are based on the involved stellar evolution modeling of Boothroyd *et al.* (1991), who assume an early main sequence solar mass loss (driven by rapid initial rotation and perhaps pulsation). In this model, the destruction (and production) of lithium throughout the interior structure of a star in evolution, and a careful account of changes in photosphere lithium abundance due to mass loss and convection, are both included. They find that for stellar masses equal to or slightly greater than $1.1 M_{\odot}$ (1.1 times the present mass of the sun), the “case of the missing solar lithium” is solved. However, since early solar flux is scaled as the 6.5^{th} power of the solar mass, the flux 4.5 Gyr ago could have been 1.3 ($0.7 \times 1.1^{6.5} = 1.3$) times the present value, representing a much warmer early sun.

In the modeling of Boothroyd *et al.* (1991), the presently observed depletion of

solar lithium is not dependent upon the rate at which solar mass is lost (only on the initial solar mass), so the specific history of the luminosity (how long the sun was warmer) is dependent upon an unconstrainable quantity, the mass loss rate. Depending upon the mass loss rate, the solar flux at the time when the sun came on main sequence 4.5 Gyr ago (1.3 times the present value) drops in an exponential fashion, and joins the standard evolution within 0.5 and 2.0 Gyr. We have parameterized these results with a time dependent luminosity which roughly follows a nominal best guess mass loss timescale. Solar luminosity is parameterized as evolving through two linear segments: from 0.9 at 4 Gyr ago, to 0.77 at 3.4 Gyr ago, to 1.0 at the present time.

In Fig. 4.11, which shows the initial and final reservoir storage for model evolutions with the Graedel *et al.* (1991) luminosity, the change from a buffered to an unbuffered initial state has moved to a lower initial inventory. The partitioning in the regime below this discontinuity is again decided in conjunction with the frost point relationship. As a result of added insolation (luminosity), the frost point relationship and polar surface heat balance are satisfied at a tripled initial surface pressure, 1.5 mbar versus 0.5 mbar of the baseline case (Fig. 4.2); however, the qualitative nature of evolutions at such abundances is as before.

The upper discontinuity nearly disappears as a result of elevated initial luminosity ($\mathcal{L}_i = 0.86$), and initial atmospheric pressures can be quite large. This happens because $p_{max} = p_{max}(\mathcal{L}_i)$, and since initial luminosity is enhanced with the Graedel

Figure 4.11: Same as Fig. 4.2 except with Graedel *et al.* (1991) luminosity.

et al. warm early sun ($\frac{0.86^{\text{new}}}{0.75^{\text{old}}}$), p_{max} grows to 6.5 bar. When the model is initialized with a sufficient initial CO_2 inventory (nearly 8 bar), the surface pressure obtains this value. An important consequence is that a 5 bar inventory will initialize with a 4.2 bar atmosphere, and with the warming of the baseline greenhouse, we find an equatorial temperature of 270 K, yielding a warm and wet early Mars. The reservoir and temperature histories for the 1, 2 and 5 bar integrations with the Graedel *et al.* luminosity are shown in Fig. 4.12. The large initial atmospheres in the 2 and 5 bar integrations are rapidly lost in carbonate rock formation, a result of the strong pressure and temperature forcings in the chemical weathering parameterization.

In Fig. 4.11, we find that the final state of the warm early sun integrations is very similar to those of the baseline, although a significant distinction is the ramp in the carbonate CO_2 reservoir, which leads up to a discontinuity that is smaller than with the baseline luminosity. This results from the model being able to initialize without a permanent cap, which causes greater early weathering losses from higher global surface pressures and temperatures. These integrations, however, end in an early collapse, as exchangeable CO_2 is lost to carbonates, and with the help of decreasing insolation, the polar surface temperature drops rapidly to the frost point. The 2 bar case of Fig. 4.12 barely escapes this early collapse, as the luminosity stops decreasing and begins increasing just before it would happen. The polar surface temperature then follows the frost point closely until the atmosphere collapses normally (with L increasing) 800 Myr ago. We note that when initial inventories are large enough for

Figure 4.12: Same as Fig. 4.1 except with Graedel *et al.* (1991) luminosity.

the baseline case to initialize without caps ($P_{tot} = 1.7$ bar) the ramp ends and the smaller discontinuity occurs, indicating that collapse happens normally with greater initial inventories. For initial inventories “on the ramp,” atmospheric collapse with the warm early sun is caused by decreasing solar luminosity (early collapse). Once above this region, the collapse takes place normally at much later times, allowing for cumulatively greater carbonate accumulations before it occurs.

With a warm early sun, the model can produce a warm and wet early climate when initialized with large amounts of CO_2 ($P_{tot} > 5$ bar), but the 300 mbar permanent cap, and the large carbonate reservoirs of the final states seem excessive. The evolution which best achieves the present climate (the most known constraint on the model) is still a buffered low pressure evolution, where the suggested initial inventory is again 0.6 bar.

e. Results of a Stronger Greenhouse Effect

Since the baseline Greenhouse Effect (as implemented with the saturation constraint) will not allow ancient surface pressures to rise above 1.9 bar when using the baseline model of solar luminosity, a warm and wet early Mars is unlikely unless other assumptions are made which would allow such a climate to exist. One possibility is that the strength of the Greenhouse Effect (the increase of T_M with pressure) was greater than we have assumed.

The possible greenhouse warming from a CO_2/H_2O atmosphere alone, after ac-

cepting surface pressure limits determined by the condensation of atmospheric CO_2 , is not sufficient to allow a warm and wet early climate. Kasting (1991) concludes that the presence of additional greenhouse gasses during the period of network formation is a way to make early Mars warm and wet. The effects of such an assumption on climate evolution are explored in this subsection (the emissivity parameterization for this stronger greenhouse was developed in Chapter 2).

We find by inspecting Fig. 4.13, which shows the initial and final reservoir storages with the stronger greenhouse, that for low initial inventories ($P_{tot} < 1$ bar), the results compare identically with those of the baseline (weak) greenhouse. The different greenhouses agree since atmospheric pressure is always low in this regime, which causes greenhouse warming to be negligible so the heat balance is mostly radiatively determined. However, as initially available CO_2 increases, and surface pressure increases in equilibrium with the regolith, the stronger greenhouse causes the transition into the collapse regime to occur at a lower total inventory. Since the model is initialized with most of the available CO_2 in the atmosphere (the remaining amount is put in the regolith), the model has the opportunity to find an equilibrium between p and P_{reg} with T_{ps} greater than the frost point. Comparison of Figs. 4.2 and 4.13 shows that the weaker greenhouse has a greater discontinuity at this transition. The stronger greenhouse can equilibrate without caps with a 210 mbar atmosphere, whereas the baseline greenhouse requires 580 mbar.

For all greater inventories, the model initializes with increasing storage in the

Figure 4.13: Same as Fig. 4.2 except with the stronger greenhouse (parameterization of Pollack *et al.* (1987) modeling).

atmosphere, and an initial regolith storage which obtains a maximum value at a total inventory of 2 bar. A changing importance in the pressure and temperature forcings for the regolith storage is the cause of the maximum. The partial derivatives of the regolith storage will vary in magnitude and influence over the range of p and T_M , and their behavior is determined by the greenhouse parameterization, Eqn. 3.5.

Figure 4.13 shows that large inventory evolutions always end with the present surface pressure, and with 300 mbar stored in both the permanent cap and regolith. For total inventories > 2 bar, all additional CO_2 is lost in the formation of carbonates by the present time. The chemical weathering rate becomes much larger with the enhanced global surface temperatures of the stronger greenhouse and the greater surface pressures possible when not considering the condensation of atmosphere. Atmospheric CO_2 is thus rapidly lost to carbonates when the surface pressure is large; the exchangeable total falls rapidly, polar temperatures drop, and atmospheric collapse always occurs. The existence of the permanent cap buffers the climate, and since pressures are low the solution is mostly radiative, agreeing closely with the baseline greenhouse integrations at higher intermediate abundances.

Figure 4.14 shows the reservoir and temperature histories for the 1, 2 and 5 bar integrations, and reveals that when $P_{tot} > 2$ bar, the model can produce a warm and wet early Mars. However, the final size of the permanent cap is again much larger than what is presently believed stored in this reservoir. The 2 and 5 bar cases of Fig. 4.14 show that the decrease in the size of the regolith during collapse is much

Figure 4.14: Same as Fig. 4.1 except with the stronger greenhouse (parameterization of Pollack *et al.* (1987) modeling).

greater than the corresponding decrease in the surface pressure, thus the regolith is the major factor contributing to the size of the permanent cap after collapse.

To have maintained the massive atmospheres required to have a warm and wet early Mars persist for a geologically significant period of time would have required mechanisms to recycle lost greenhouse gasses. This is indicated by the rate of carbonate formation (atmospheric loss) early in the climate evolutions of the 2 and 5 bar cases of Fig. 4.14. The lifetimes of alternative greenhouse gasses are not known for early Mars, although it is likely that they are very short due to photolytic destruction. Kasting (1993) speculates that microbial life (methanogenic bacteria) is one way to maintain a sufficient partial pressure of CH_4 . This possibility is of course colorfully suggestive in light of the Gaia Hypothesis proposed by James Lovelock in *The Ages of Gaia* (1988), since the existence of bacterial life would have helped maintain a climate in which it could have survived. Although other additional greenhouse gasses could provide the necessary radiative properties, mechanisms to maintain their partial pressures are not obvious. Kasting (1993) laments. "Thus, if I were a betting scientist, I would wager that either early Mars was inhabited, or the martian channels were formed by recycling of subsurface water under a cold climate." The recycling and circulation of water by a subsurface global aquifer is strongly supported by Clifford (1993).

Using the stronger greenhouse, the model still favors an initial inventory of 600 mbar to reproduce the present climate with a permanent cap storage of < 5 mbar.

Whether 5 mbar is really a realistic approximation (and constraint on the model) for the annually averaged value of the permanent cap storage over the past 7.6 Myr (the last model timestep) is certainly a question worth addressing. In the past 18 kyr on Earth (a planet more stable in obliquity than Mars due to its relatively massive moon), the climate experienced great changes as Earth emerged from its most recent ice age. The averaged size of polar ice deposits over millions of years and numerous ice ages is quite likely much different than at the present time. Analogously, the average storage in the permanent cap on Mars throughout the past 7.6 Myr might not agree with the size of the permanent cap at the present time. Moreover, the size of the permanent cap at the present mean martian obliquity ($\Theta = 24.4^\circ$), the logical first choice to represent the size over the past 7.6 Myr, would be larger than at present obliquity ($\Theta = 25.2^\circ$). The point is that to determine a value which represents the average size of the permanent cap over the last 7.6 Myr of climate evolution on Mars is very difficult at best.

The stronger greenhouse is attractive. It can provide the warming which allows for a warm and wet early climate. However, a problem with the assumption is that there is no known evidence to directly support that the early atmospheric composition included the necessary levels of any alternative greenhouse gasses. Brown and Kasting (1993) discuss preliminary work, and they are presently continuing investigations of the Greenhouse Effect for an early Mars with other such greenhouse gasses. Their results and investigations into the lifetimes of the gasses involved will provide further

understanding.

f. Variable Polar Albedo

When a permanent cap is not predicted by the model, it might be a poor assumption to believe that the albedo of the region is the same as when a permanent cap is present. In response to this concern a variable polar albedo has been incorporated into the model such that when a permanent cap exists, the albedo is $A_p = 0.75$ (as previously maintained throughout integrations), and when there is no permanent cap $A_p = 0.5$. This value is twice the bare ground value of 0.25, and is adopted because it approximates the albedo of dirty H_2O ice. The residual polar cap in the northern hemisphere is a dirty ice cap, which exists perennially in the present climate (Thomas *et al.* 1991). It seems reasonable to assume that similar residual cap compositions have prevailed during other epochs as well, as Mars was likely bestowed with a substantial amount of water (Carr 1986).

An interesting consequence of the lower albedo is seen by inspecting Fig. 4.15, which is a plot of polar surface temperatures for either greenhouse model at the same four luminosities, which shows temperatures with both polar albedos. Polar surface temperatures found with the lower albedo are entirely greater than the frost point. This implies that the model cannot transit to a climate with polar caps from a climate without them, as there is no solution to the heat balance with low polar albedo that would allow the initiation of an atmospheric collapse. A question of model start up

Figure 4.15: The effect of polar albedo on the polar surface temperature T_{ps} , shown versus pressure for $\mathcal{L} \in \{0.7, 0.8, 0.9, 1.0\}$ where $A_p = 0.75$ (dash) and $A_p = 0.5$ (dot). The greater temperatures have greater luminosity, and the remaining curve (solid) is the frost point of CO_2 .

is thus raised, since the initial choice for A_p will determine if the model could ever predict caps. Since we partition CO_2 at start up mostly to the atmosphere (providing the model an opportunity to find an unbuffered solution), the logical approach is to set $A_p = 0.5$ (hot start) when initializing the model.

Figure 4.16 shows (when the low polar albedo is assumed at start up) that the only way a cap will form initially is when required by the saturation constraint. When a permanent cap is imposed by this constraint, the albedo increases to the larger value, and the model evolves exactly as the baseline case until the cap has sublimed. Polar temperatures increase significantly when the permanent cap disappears because the albedo drops, and the polar heat balance is again allowed to find thermodynamic equilibrium. From this moment forward, polar surface temperatures are unable to approach the frost point and a permanent cap cannot form. If $P_{tot} = 700$ mbar, the model predicts the present atmospheric pressure in the final state, although no permanent cap exists.

Model evolutions are affected not only by a variable albedo, but by the initial value of this albedo as well. If the model is started assuming the larger value (cold start), then the initial state is that of the baseline case. The final state will differ mainly due to the removal of the collapse regime; a permanent cap can only exist in the final state if it lasts the entire evolution. This result is similar to that of Calderia and Kasting (1992), who argue that Earth might have remained in a permanent glaciated state if it had been cold early on, instead of warm, as is believed by examining climatological

Figure 4.16: Same as Fig. 4.2 except with variable polar albedo in effect, with $A_p = 0.5$ at initialization (hot start).

records. Lorenz (1990), however, points out. using only a slightly more complicated climate model to model terrestrial climate, that to assume climate is intransitive and will lock into a single mode is suspect at best. A chaotic system (such as the climate of a planet) is highly nonlinear, and can slip away from such seemingly intransitive states due to the numerous interactive feedbacks. However, experimenting with a variable polar albedo does show that the model is highly sensitive to this parameter, and on Mars this sensitivity is probably more important than on Earth because the primary constituent of the martian atmosphere will condense at normal temperatures.

4.3 Atmospheric Collapse

Throughout the previous subsections, the sensitivity and range of model evolutions has been explored by varying the nominal baseline assumptions. The phenomenon of atmospheric collapse is sensitive to, and as a result is greatly affected by, modifications made to the processes which are modeled.

Figure 4.17 shows collapse sequences for both emissivity parameterizations (greenhouses) over different ranges of initially available CO_2 . As can be identified by examining Fig. 4.2, the baseline model will collapse only if P_{tot} is between 1.6 and 3.0 bar. For the stronger greenhouse of Fig. 4.13, the collapse regime is unbounded above, although the model cannot initialize without a permanent cap below $P_{tot} = 1.0$ bar, bounding the regime below by this value. There is also a significant difference between the times for which collapse can happen. With the stronger greenhouse, weathering

Figure 4.17: Atmospheric surface pressure histories for numerous P_{tot} values, showing the collapse sequence for the baseline greenhouse(top), and the stronger greenhouse(bottom). Between individual curves $\Delta P_{\text{tot}} = 0.1$ bar, and the range of P_{tot} is 1.6–3.0 bar for the baseline, and 1.0–3.0 bar for the stronger greenhouse.

removes CO_2 fast enough for collapse to take place entirely before 3 Gyr. With the baseline greenhouse, the approach to collapse is much slower (it cannot occur before 3 Gyr), but could happen up to the present day and beyond.

Atmospheric collapse is a phenomenon which is affected by the nonlinear interactions between chemical weathering, changing insolation (luminosity), dynamic heat transport, the Greenhouse Effect and atmospheric escape. Weathering and escape reduce exchangeable CO_2 , thus altering the polar heat balance. Heat transport and greenhouse warming are reduced, and polar temperatures will fall in spite of the fact that solar luminosity is increasing.

During model integrations, atmospheric collapse is initiated when the surface pressure is larger than the vapor pressure over CO_2 ice at the polar surface temperature, $p > p_{fp}(T_{ps})$. To maintain the required climatological equilibrium requires that this difference in pressure is eliminated by the condensation of sufficient atmospheric mass upon the polar surface. This is the beginning of an atmospheric collapse, which is simultaneously the birth of a permanent polar cap.

However, once this process begins at the upper intersection point of one of the dashed curves in Fig. 4.15 with the frost point curve, it cannot end until the climate of the stable lower intersection point on that curve is reached. As discussed in Chapter 2, the upper intersection point is unstable, and since climates with an annual polar surface temperature below the frost point are not realistic, the intermediate heat balance solutions are quickly transited as the climate falls to the lower intersection,

thus completing the collapse of the atmosphere.

The lower pressures at which collapse occurs in the baseline model, and that atmospheric pressures at which collapse occurs decrease with time, are both caused by increasing luminosity in the polar surface heat balance. When luminosity increases, the pressure of the unstable solution (collapse pressure) in Fig. 4.15 drops. The absolute range of possible collapse pressures for each greenhouse is between the pressures at the upper intersections of the dashed $\mathcal{L} = 0.7$ and $\mathcal{L} = 1.0$ curves of Fig. 4.15. The actual range of collapse pressures is smaller, since collapse at low/high luminosity is not possible for the baseline/stronger greenhouse. The actual range of collapse pressure is the distribution on the pressure axis of Fig. 4.17 of the atmospheric collapse.

A collapse of the atmosphere into a buffered state indicates a significant change in the prevailing mean climate. It is possible that such a climatological change ended the era of valley network formation, since large changes in atmospheric pressure and global temperatures would cause a significant drop in erosion rates, likely terminating any significant or intermittent fluvial activity. This is surely speculation, but epochal atmospheric collapse appears to be a real possibility, and might have played an important role in the evolution of the martian atmosphere and climate.

4.4 Obliquity Variation

The existence of layered terrain in both polar regions of Mars is strong supporting evidence for a cyclic behavior in climate. Pollack (1979) suggests that periods of net

deposition have alternated with periods of net erosion, creating the layered structure that is seen today in these regions. The cause for this cyclic behavior is believed to be variation in the annually-averaged latitudinal distribution of solar insolation, a result of obliquity variation cycles (Fanale *et al.* 1982).

The history of martian obliquity has been examined with the mathematical theories of Astrophysics, modeling the effects of torques on the planet from other astronomical bodies (Ward 1974, Ward *et al.* 1979, Bills 1990). Although there is uncertainty in these calculations (since the mean moment of inertia for Mars is poorly known), such efforts have revealed a coarse structure for the recent history of martian obliquity, Θ . The present obliquity (25.2°) is near the present mean (24.4°), and the primary period of oscillation is approximately 10^5 yr, with an amplitude of 13.6° (Bills 1990). The amplitude is modulated as well, with a period of approximately 1 Myr. Ward *et al.* (1979) suspect that the formation of Tharsis (a large volcanic uplift in the equatorial region) may have caused Mars to pass through resonance conditions, allowing the variation of obliquity to be dramatically large during these times. Although tectonic activity has persisted in the Tharsis region throughout a great majority of the geological history, the most intense activity occurred within the Middle Noachian to Late Hesperian period (Tanaka *et al.* 1991, Banerdt *et al.* 1991), which places the greatest effect on the mean moment of inertia in a broad period that begins with the end of heavy bombardment, and continues until approximately 2.5 Gyr ago (Tanaka *et al.* 1991). Active formation of Tharsis would have continually

changed the mean moment of inertia for Mars, which would have likely caused the planet to pass through resonance conditions. that would have affected both the mean obliquity and the maximum amplitude of its variation.

Since there are many nonlinearities in our simulation of the martian climate, it is prudent to inspect the assumption that weathering over a 7.6 Myr timestep is well represented when using the surface pressure p and the global temperature T_M at the mean obliquity during this period. This task is accomplished by explicitly including time dependent obliquity variations in the model. The model is then run through obliquity cycles, and the average weathering rate is computed. This averaged rate is then compared with the rate predicted in the model when obliquity was held at a mean value (how carbonate losses were computed in all previous integrations). Since obliquity variation can only directly affect solar input in our model, we start by determining the annual insolation as a function of obliquity for the polar and equatorial regions.

a. Obliquity Dependent Insolation

The annually averaged insolation at the top of the atmosphere is a function of latitude and obliquity, and is given by

$$\langle \bar{S} \rangle = \frac{S_0}{2\pi^2} (1 - e^2)^{\frac{1}{2}} \int_0^{2\pi} [1 - (\sin \phi \cos \Theta - \cos \phi \sin \Theta \sin \lambda)^2]^{\frac{1}{2}} d\lambda, \quad (4.4)$$

Figure 4.18: Annual top of atmosphere insolation (Wm^{-2}) as a function of latitude and obliquity.

where the integration is over aerocentric longitude λ (a measure of orbital position, where northern spring equinox is $\lambda = 0$), ϕ is latitude, Θ is obliquity and e is eccentricity (Ward 1974). Figure 4.18 shows $\langle \bar{S} \rangle$ for all latitudes, and obliquities between 0 and 60° . Figure 4.19 shows the annual insolation at the pole and equator over the same range of obliquity. The insolation at the pole is heavily dependent upon the obliquity, whereas the dependency is much weaker at the equator, significantly affecting the sensitive polar heat balance while minimally affecting the equatorial balance. By linearly interpolating between the calculated points on these two curves,

Figure 4.19: Equatorial (circles) and polar (squares) top of atmosphere solar flux in Wm^{-2} as a function of obliquity Θ .

variable obliquity is incorporated into the model when the values of S_{ps} and S_{eq} are dependent upon the obliquity Θ . For purposes of simplicity, the obliquity is treated as being sinusoidal in time. The time dependent value of the obliquity is determined by evaluating

$$\Theta(t) = \Theta_0 + \Delta\Theta \sin\left(\frac{2\pi(t - t_0)}{\tau_\Theta}\right), \quad (4.5)$$

where Θ_0 is the mean obliquity from which variations occur, $\Delta\Theta$ is the amplitude of this variation (which is presently 13.6° and also varies with time), t is the time elapsed since the sun reached the main sequence (the model time), t_0 is the starting time of the integration, and $\tau_\Theta = 1.2 \times 10^5$ yr is the primary period of variation (Toon *et al.* 1980). The period of variation for the magnitude of the obliquity variation is

$\tau_{\Delta\Theta}$, which is 1.3×10^6 yr (Kieffer and Zent, 1991).

b. Model Behavior With Obliquity Variation

Deciding which period of the evolutionary history to examine becomes a concern, since to obtain smooth curves over the 120 kyr cycles requires basically the same number of data points (timesteps) as to obtain smooth curves for the previous 3.8 Gyr model integrations. Thus, to resolve the effects of obliquity variation smoothly over an entire evolutionary history would require more than 15×10^6 timesteps. Moreover, the history of obliquity from the end of heavy bombardment until the present is unknown.

In an attempt to maintain brevity this investigation will focus only on an early period in martian history for two reasons. The first is that the model tends to predict the largest rates of carbonate formation during early periods, when surface pressure and global surface temperature were likely greatest, causing the effects of weathering to be most important at those times. The second is that, during the formation of Tharsis, Mars likely had obliquity variations much larger than at present (Ward *et al.* 1979, Bills 1990, Toon *et al.* 1980). In light of these reasons, the weathering rate averaged over an obliquity cycle would most likely differ at early times from the weathering rate at the mean obliquity. We choose to compare these weathering rates 3.5 Gyr ago.

Following Toon *et al.* (1980), a mean obliquity of 30° is accepted. Although the

maximum amplitude of obliquity variation during this time is unknown. a nominal value of $\Delta\Theta = 20^\circ$ is chosen as a compromise between the present 13.6° variation (Ward *et al.* 1979), and the possible large chaotic fluctuations ($0 < \Theta < 60^\circ$) hypothesized by Laskar and Robutel (1993).

Figure 4.20 shows the model behavior throughout one obliquity cycle. The integration begins 3.5 Gyr ago from a 1.0 bar inventory, with a timestep of 240 yr. The model is at mean obliquity at the beginning and the end of the cycle. Therefore, the climate (and weathering rate) at these moments is what would be predicted by the baseline model 3.5 Gyr ago if the exchangeable total were 1.0 bar of CO_2 .

Following climate changes throughout the obliquity cycle, as obliquity increases and enhances polar insolation, the surface pressure rises as governed by the frost point relationship and the polar heat balance. The combined effects of rising insolation and increased heat transport (due to rising pressure) are finally too much, and the remaining polar cap sublimates within one timestep, freeing 120 mbar of CO_2 to be incorporated into the atmosphere and regolith. This can be thought of as the opposite of atmospheric collapse, and due to the rising insolation, the polar surface temperature can no longer intersect the frost point curve at any surface pressure. Equilibrium can then only be maintained without a permanent cap. The surface pressure grows from 100 mbar to 150 mbar, and the regolith storage increases from 780 mbar to 850 mbar, all in 240 yr. There is a high correlation between surface pressure and weathering rate, as can be seen by comparing Figs. 4.20(a) and 4.20(c). The global surface

Figure 4.20: History of obliquity and weathering rate in subplot (a). Temperature and reservoir histories in respective subplots (b) and (c). Subplot (b): T_{es} (upper solid); T_{ps} (lower solid); T_M (dash); T_{fp} (dot). Subplot (c): p (solid); P_{cap} (dash); P_{reg} (dot).

temperature, the other forcing in the chemical weathering parameterization, remains relatively constant throughout the integration as seen in Fig. 4.20(b).

As obliquity starts to decrease, insolation is lost from the polar heat balance and the atmosphere eventually collapses, but the collapse occurs at an obliquity slightly lower than when the caps disappeared. This hysteresis effect is the result of the surface pressure-caused feedbacks due to greenhouse warming and dynamic heat transport. Before the atmospheric collapse the surface pressure is high, so there must be a reduced insolation (lower obliquity) to initiate it. However, once the atmosphere collapses, the climate is again determined by the frost point relationship and the polar heat balance. But since this happens at a lower obliquity, the cap formed is larger, 220 mbar compared to 140 mbar. Surface pressure drops below 2 μ bar at the lowest obliquities, where all exchangeable CO_2 , except the 40 mbar in the regolith, has become part of the permanent cap. As obliquity returns to its mean value, the climate as well returns to the mean state.

That changes in annual insolation are more pronounced in polar regions than equatorial, is made clear by inspecting the T_{ps} and T_{es} curves of Fig. 4.20(b). The polar surface temperature follows the obliquity, whereas the equatorial surface temperature remains relatively unaffected.

The weathering rate averaged over the obliquity cycle is 2.2×10^{-2} mbar Myr⁻¹, whereas the rate at mean obliquity is 1.4×10^{-2} mbar Myr⁻¹. Thus, with a 1.0 bar exchangeable total 3.5 Gyr ago, including obliquity variation will cause a 57% increase

in the average weathering rate over the baseline value.

A possibly more realistic approach assumes that the polar albedo is variable during an obliquity cycle. During a 240 yr period without a permanent cap, the polar region would likely have a significantly lower albedo than for the same period with a permanent cap. Incorporating the variable polar albedo as discussed previously ($P_{cap} = 0$: $A_p = 0.5$; $P_{cap} \neq 0$: $A_p = 0.75$), Fig. 4.21 shows an integration including variable polar albedo, where all other factors are as in Fig. 4.20. The integration is identical until the polar cap disappears. At that time, in response to the lowered polar albedo, there is a sharp rise in the polar surface temperature. The slight rise in the equatorial and global surface temperature is a little greater, since less heat will be transferred into the polar regions in response to a diminished latitudinal temperature gradient. The surface pressure finds a greater equilibrium value as the regolith must release some CO_2 due to the increased T_M , and the combination of increased pressure and global surface temperature increase the weathering rate. These effects, combined with the prolonged time at high pressure before the decreasing polar insolation can finally cause collapse, allow the averaged weathering rate to be 5.2×10^{-2} mbar Myr⁻¹, 3.7 times the weathering rate at mean obliquity.

The effect of variable polar albedo here is significant, although the quantitative accuracy of the result is surely suspect. Of note is that the model can form a permanent cap in a climate where one had not existed previously, an occurrence which did not happen when using a variable polar albedo with the baseline model. This happens

Figure 4.21: Model behavior with obliquity variation and variable polar albedo, as in Fig. 4.20.

because diminished polar insolation at low obliquity actually allows the greater polar surface temperatures at the lower albedo to intersect with the frost point curve. The important point to be made, however, is that enhanced averaged weathering is highly dependent upon the amount of time during the obliquity cycle when the climate is not buffered by polar caps. There does seem to be an important nonlinearity here, since high obliquity affects both surface pressures and global temperatures, whereas low obliquity only decreases surface pressures. High obliquity tends to increase weathering more than low obliquity tends to decrease it.

A theoretical problem, however, exists with both of the previous examples. It is highly unlikely that the regolith is capable of finding equilibrium with such dramatic (discontinuous) changes in climate within a short 240 yr time period. The regolith must, due to its physical nature, take time to equilibrate to these changes in surface temperature and pressure, time for the respective diffusion processes to act. This is exactly the problem discussed in the next subsection, and a means to overcome the inconsistency in the model is developed.

c. Regolith Equilibrium Time τ_{reg}

The physical nature of the martian regolith is not well known. However, from the little knowledge we do have, certain bounds may be established for its thermal and physical properties. Within these bounds it has been determined that the regolith has a thermal response time (to equilibrate with change in annual surface temperature)

between 10^4 yr and 10^6 yr (Kieffer and Zent, 1991), where the range of values is mostly due to the assumed depth of the regolith. When including obliquity variation, we use a timestep much less than the smallest of these values, implying it is no longer valid to treat the regolith as in constant equilibrium with the global surface pressure and temperature of each timestep. The atmosphere and permanent cap can, however, reach equilibrium within 100 yr (Kieffer and Zent 1991), so a rapid or discontinuous change in the storages of these reservoirs is not inconsistent with understood processes. However, a rapid or discontinuous response in regolith storage is inconsistent with the understood physics.

This inconsistency is dealt with in the model by limiting the rate at which CO_2 can be exchanged between the atmosphere and regolith. This approach mimics the overall effect of the diffusive-like (lagging) behavior of surface temperature and pressure forcings on the regolith storage, by setting the time rate of change of the regolith storage proportional to the difference between the equilibrium and current storage

$$\frac{dP_{reg}}{dt} = \frac{(P_{EQreg} - P_{reg})}{\tau_{reg}}, \quad (4.6)$$

where τ_{reg} is defined as the time constant of the combined surface temperature and pressure forcings on regolith storage, and P_{EQreg} is the equilibrium storage of the regolith (the value used for all previous experiments). A reference to temperature and pressure “waves” in the regolith will be used to indicate that there is a finite time which the regolith requires to equilibrate to changes in surface pressure and

temperature. The changes in surface climate are responded to internally by the regolith with processes that are closely approximated with diffusion theories. It is the movement of this internal response through the regolith which is referred to as a pressure or temperature wave.

The above definition for the time rate of change of regolith storage is transformed into a finite difference approximation

$$\frac{P_{reg}^{n+1} - P_{reg}^n}{\Delta t} = \frac{1}{\tau_{reg}}(P_{EQreg}^{n+1} - P_{reg}^n), \quad (4.7)$$

and this expression is algebraically manipulated to solve for P_{reg}^{n+1} , which yields the expression used in the model

$$P_{reg}^{n+1} = \left(1 - \frac{\Delta t}{\tau_{reg}}\right)P_{reg}^n + \frac{\Delta t}{\tau_{reg}}P_{EQreg}^{n+1}. \quad (4.8)$$

Toon *et al.* (1980) discuss pressure diffusion rates, and note that if the regolith is basalt, the pressure wave is essentially instantaneous in comparison to the temperature wave, as it will penetrate 1.5 km in 10^5 yr, while the temperature wave would only reach 300 m. If however, the regolith were essentially clays, the pressure wave could only penetrate 40 m in the same time interval. Toon *et al.* (1980) point out that pressure diffusion probably will not stop the regolith from responding to changes over climatic time scales, but they do not indicate that internal pressure response should be treated as a rapid process with respect to internal temperature change.

Since Toon *et al.* believe that a pressure wave is probably not slower than a temperature wave, and as the approach used here combines both effects, we choose $\tau_{reg} = 5 \times 10^4$ yr as a nominal value. Moreover, Figs. 4.20 and 4.21 indicate that the larger proportion of forcing was due to pressure changes, since global temperature changes were minimal throughout those experiments. As this treatment addresses the finite lag time required by the regolith to equilibrate with climate, we return to our primary concern. How does the regolith storage being out of equilibrium affect the weathering rate when averaged over the obliquity cycle?

Figure 4.22 shows the model behavior through a single obliquity cycle when using the nominal regolith lag time. The integration takes place 3.5 Gyr ago with an exchangeable total of 1.0 bar, and a polar albedo which remains constant at $A_p = 0.75$.

The model now requires a few obliquity cycles before it will settle down to a repeatable cycle. When it does finally settle down, the regolith storage at mean obliquity is 240 mbar, which is 40 mbar less than the value without the finite response time. The freed CO_2 will be incorporated into the permanent cap, since the surface pressure is fixed by the frost point relationship. The regolith is clearly slow to equilibrate with climate change, as discontinuities and steep slopes are not found in the P_{reg} curve of Fig. 4.22(c). Since the regolith can only adsorb a small amount of additional CO_2 before the increasing obliquity enhances insolation to the point where the permanent cap cannot exist, there is a huge reservoir of CO_2 that is freed when the permanent

Figure 4.22: Model behavior with obliquity variation and finite regolith lag time. Curves are as in Fig. 4.20, $\tau_{reg} = 5 \times 10^4$ yr, and the polar albedo is held constant at 0.75.

cap disappears. This CO_2 can only become atmospheric, as the regolith cannot pick up a large amount of CO_2 in a single timestep. Thus, the entire polar cap storage of 560 mbar goes into the atmosphere, and increases the atmospheric pressure from 100 mbar to 660 mbar. The discontinuities in weathering rate, global surface pressure and temperature are therefore quite large. The regolith attempts to equilibrate, but can only do so minimally before decreasing obliquity enhances the equatorial insolation so that the rising global temperature forces the regolith storage to decline, again adding to the atmosphere. This increase in global surface pressure and temperature create feedbacks through heat transport and greenhouse warming, and the unbuffered state persists even longer than with a variable (lowered) polar albedo. During this period of high p and T_M , the weathering rate reaches a peak of $0.4 \text{ mbar Myr}^{-1}$, and the averaged value is $0.13 \text{ mbar Myr}^{-1}$, nearly a ten fold increase over the weathering predicted at mean obliquity with the baseline model. It is noteworthy that when including the finite regolith response time, the global surface temperature was significantly affected.

Although this investigation only compares the weathering rates of 3.5 Gyr ago, it is useful to compare the rates found with these two different approaches over a range of exchangeable totals. Figure 4.23 has been constructed to provide such a comparison. The discontinuity in the mean obliquity weathering rate appears when the exchangeable total is great enough for the climate to be unbuffered. The five points at large exchangeable total indicate that the saturation constraint has been

Figure 4.23: Comparison of the weathering rate at mean obliquity (baseline value) to the weathering rate averaged over an obliquity cycle. The time is 3.5 Gyr ago, and $\tau_{reg} = 5 \times 10^4$ yr for the averaged rate. The polar albedo is 0.75.

reached, thus the surface pressure cannot increase, so neither can the weathering rate.

For the lowest exchangeable totals both rates are very low, although the averaged rate is minimally larger. This is because a period of time during the obliquity cycle is spent in an unbuffered climate, where the global surface pressure and temperature are larger, therefore making the weathering rate greater. Since the time spent in this unbuffered state increases with the exchangeable total, the averaged rate grows accordingly. For the largest exchangeable totals the rates are similar, as both cases predict an entirely unbuffered climate. The averaged rate is slightly less because large values of obliquity reduce the equatorial insolation, which decreases the global surface temperature, and accordingly the weathering rate. The effect is not compensated for

at low obliquity, since the equatorial insolation is nearly invariant at low obliquities.

The middle regime near the discontinuity in the mean obliquity weathering rate is the interesting portion of Fig. 4.23. Weathering averaged over the obliquity cycle can be more than twenty times that of the buffered mean obliquity rate. Although this result should certainly be explored further before making any generalizing statements, it seems to indicate that actual losses to carbonates might be much larger after collapse than we have predicted in previous model integrations. Losses to carbonates do not appear significantly affected during the approach to atmospheric collapse, but after collapse it seems the mean obliquity rate significantly underestimates this loss. This is a mechanism which might significantly deplete the large permanent caps which remain after an atmospheric collapse in the baseline evolutions.

Chapter 5

Summary, Discussion & Suggestions

5.1 Summary

A fully coupled model, which simulates the evolution of climate on Mars from the end of heavy bombardment to present day, has been developed. The model uses heat balance principles to determine the temperature structure, and a mass partitioning scheme to distribute CO_2 between the model reservoirs. The partitioning of CO_2 between reservoirs is consistent with the computed temperatures, the surface pressure and the solar luminosity.

The heat balances compute surface and atmospheric temperatures for polar and equatorial regions. In addition, a global surface temperature is found from these tem-

peratures which matches the results of state-of-the-art greenhouse models. The model was tuned to match these greenhouse models by incorporating an atmospheric emissivity parameterization which depends upon surface pressure and solar luminosity. The climate model includes the transport of equatorial heat into the polar regions, linking the temperature structure of the equator dynamically to that of the pole. The annual surface insolation is a function of solar luminosity, the albedo (either equatorial or polar), the annual top of atmosphere solar flux (at the present obliquity and eccentricity), and the latitude. Although the model has some characteristics of 2-D models (surface and atmospheric temperatures, and a latitudinal temperature gradient), it is clearly not 2-D; however, its simplicity allows for rapid computation of the martian temperature structure, and its separate treatment of equatorial and polar regions allows for the formation of polar caps (global models cannot).

The CO_2 partitioning scheme utilizes published results for regolith uptake and present storage, chemical weathering rates, and atmospheric escape due to the sputtering of CO_2 . When the annual polar surface temperature falls to the frost point of CO_2 , a permanent polar cap is established, and the annual surface pressure becomes buffered by the frost point relationship. During model integrations, a climatological equilibrium is recalculated between the temperature structure and the reservoirs every 7.6 Myr. The model then loses CO_2 to space and carbonates according to the parameterizations, the time is advanced, the luminosity is recalculated, and the cycle is repeated 500 times to complete a 3.8 Gyr simulation of climate evolution on Mars.

As the model is highly parameterized, there is a large parameter space to explore. Experiments were performed to qualitatively determine the effects of doubling and halving the weathering rate, the present regolith storage and the magnitude of equatorial heat transfer. Other experiments explored the effects of a different (stronger) Greenhouse Effect, a warm early sun (alternative solar evolution), a variable polar albedo and the variation of obliquity. These experiments were all performed over a wide range of available CO_2 .

A major objective in this research has been to determine whether the evolutionary scenarios possible with this climate model either support or discredit the hypothesis of a warm and wet early Mars. Evidence that early Mars was warm and wet is geomorphological (explanations for how the geological features of its surface formed), and is therefore a weaker constraint on the model than a requirement that integrations end at the present climate (a fairly well known climatological data point). To be able to quantitatively link the hypothetical warm and wet early Mars to the present climate (via some possible model integration) would provide further evidence of its existence.

5.2 Discussion

The climate evolutions predicted with the model have been examined in light of three decreasingly important criteria: *i*) integrations should end at the present climate; *ii*) the initial climate equilibrium should be warm and wet; *iii*) the climate should experience a “rapid” change during its evolution. With these criteria in mind, the

initial and final climate states that are predicted by the model are of paramount importance, and model evolutions are thus examined using the endpoints of each integration (the climate at the end of heavy bombardment 3.8 Gyr ago, and that of today).

Using the Kasting (1991) greenhouse and the standard solar evolution, a thorough exploration of the parameter space has shown that to reach the present day climate requires the entire evolutionary history of Mars to have been very cold, with permanent CO_2 polar ice caps in existence for the entire 3.8 Gyr. The surface pressure throughout such integrations remains at the vapor pressure of CO_2 over the polar caps (very low). Temperatures too are always very cold, and change only in response to solar luminosity, reaching maximums with the present climate. Sensitivity experiments have shown that the total abundance of CO_2 which yields this “best case” evolution ranges between 0.5 and 1.0 bar. The qualitative size of this initial CO_2 total, and the overall type of evolution which results, remain unchanged. Moreover, the baseline assumptions of the model (weakened greenhouse due to atmospheric condensation in a CO_2/H_2O atmosphere with a faint early sun) are incapable of allowing early Mars to be warm and wet. Therefore, the maximum mean global surface temperature at these early times is 225 K, a direct consequence of the limited Greenhouse Effect at the low early luminosities (Kasting 1991).

The above problem can be solved by incorporating a different luminosity history, as suggested by Graedel *et al.* (1991). When using their results, and assuming

the early sun was brighter than the standard solar evolution predicts, the baseline greenhouse can provide a warm early Mars when the initial inventory is greater than 2 bar. However, there is a problem with such evolutions. The final state of the model will always have a permanent cap of no less than 300 mbar, a value which seems to be two orders of magnitude too large (the present day polar cap storage of CO_2 is believed < 5 mbar).

An alternative solution to warming early Mars, as Brown and Kasting (1993) have suggested, is the presence of other greenhouse gasses in the early atmosphere. When using a stronger greenhouse in the climate model (which is assumed to result from the existence of additional atmospheric constituents), a warm early climate is possible when initial inventories are greater than 2 bar. However, model integrations with the stronger greenhouse also have final states with no less than 300 mbar in the permanent cap. Both warm early sun and stronger greenhouse evolutions will only reach the present climate and polar cap size when the total inventory is 600 mbar.

The major factor which causes the large size of the permanent cap in the final state is the value assumed for the present day storage of exchangeable CO_2 in the regolith. During an atmospheric collapse into a buffered climate (where a permanent polar CO_2 cap exists) the regolith contributes a large amount of its store of CO_2 in the formation of the cap. This amount is generally much larger than what the atmosphere can contribute. Furthermore, after atmospheric collapse the formation of carbonates is quite small (global surface temperature and pressure drop significantly),

which allows the large permanent cap to remain until the present time. Kahn (1985) considers that a 300 mbar regolith capacity (our nominal value) is an upper limit, and our experimentation with the present capacity of the regolith shows that the final size of the permanent cap is smaller when using a smaller contemporary regolith capacity.

Zent and Quinn (1993) have presented preliminary results which indicate that even half this 300 mbar capacity might be too great a value to assume presently stored in the regolith. They argue that since water molecules have a stronger electrostatic polarity than CO_2 molecules (therefore a greater adsorption affinity), they would likely displace a great amount of CO_2 from the available surface area in the regolith. If this is the case, a significant percentage of the available space in the regolith would be taken up by water. Moreover, since the greatest percentage of the regolith capacity is likely to be in high latitudes (colder temperatures allow greater adsorption), and since most of the water on Mars is at high latitudes, the effective size of the regolith (for CO_2 storage) could possibly be an order of magnitude less than what has been assumed in our model (A. Zent 1993, personal communication). The point is simply a reminder that the present and past capacity of the regolith is not well understood.

Another possibility, which might explain how a large permanent cap could be depleted before the present time, was discovered here with experiments that included the variation of obliquity. According to these experiments, the long term weathering rate does not drop discontinuously (as it does without obliquity variation) when the exchangeable total reaches a value so that the equilibrium climate at mean obliquity

(the climate of the baseline model) becomes buffered by a permanent cap. The experiments in fact show a significant increase in the weathering rate (averaged over an obliquity cycle) near the exchangeable total where the equilibrium climate at mean obliquity collapses. Whether or not the long term weathering rate actually increases near the collapse of the mean obliquity equilibrium climate is questionable, but these experiments demonstrate important nonlinearities, and indicate that weathering will continue after collapse, and not drop discontinuously as predicted without obliquity variation. This could allow permanent caps to become depleted by weathering during periods of high obliquity, resulting in a much smaller storage of CO_2 in this reservoir at the present time.

A further complication with the large initial inventories needed for a warm and wet early Mars is the size of the carbonate reservoir in the final state. With increasing initial inventory the rock reservoir consumes a greater percentage of the total CO_2 , to a point where more than 85% of a 5 bar inventory is locked within carbonates by the present day. If a substantial fraction of a large initial inventory resides in a rock reservoir, then massive sedimentary deposits should have formed. Such deposits may, however, be buried below other materials and difficult to locate with the resolutions of the remote sensing instruments that have been used to date. However, if 1 bar of CO_2 were locked in calcite, the result would be a globally distributed 20 m thick deposit. It is hard to not suspect that, if there are several bars of CO_2 locked in such a reservoir, we would have “seen it” by now. Although Pollack *et al.* (1990)

have apparently discovered that 1-3% (by volume) of airborne dust is composed of carbonates, indicating that such a reservoir probably exists, its size remains highly unconstrained.

Although varying the magnitude of equatorial heat transport did not yield new qualitative results, the climate is very sensitive to this parameter. It is clearly an important factor in the polar heat balance. The warming from equatorial heat transfer is also apparent at lower pressures than is the warming from the Greenhouse Effect. The polar winter atmospheric temperature increases (the result of equatorial heat transfer) at lower pressures than does the global surface temperature (greenhouse warming). However, between 80 and 200 mbar, the range over which atmospheric collapse occurred in our model (see Fig. 4.17), both heat transfer and greenhouse warming are significantly influential in determining the energy balances. As model integrations through obliquity cycles have shown, when the exchangeable total is very near the value at which the equilibrium climate of the mean obliquity collapses, the nonlinear feedbacks from heat transport and greenhouse warming are both very important. The weathering rate averaged over the obliquity cycle at these totals is much greater than the corresponding rate at mean obliquity, which due to these feedbacks, keeps the climate unbuffered for a large portion of the obliquity cycle.

With respect to the evolution of climate, the formation of permanent polar caps on Mars is important in two ways. The first is that the initial surface pressure is a discontinuous function of the amount of initially available CO_2 . For abundances

less than the critical value, the climate will initialize in a buffered state where the available CO_2 is shared between the atmosphere, the regolith and a permanent polar cap. If the abundance is greater than the critical value, the climate initializes in an unbuffered state, and available CO_2 is shared between the atmosphere and regolith only. The evolutions which follow from either side of this critical value will differ significantly, and our experiments suggest that the critical value is between 1 and 2 bar, although this value is very sensitive to the polar heat balance. Secondly, there is an instability in the long term mean climate of Mars. An initially unbuffered but shrinking atmosphere can discontinuously decrease in size. The atmospheric collapse occurs when the annual polar surface temperature falls to the frost point of the annual surface pressure. The surface pressure at which this collapse occurs is dependent upon the factors which influence the polar heat balance: solar insolation, the Greenhouse Effect, atmospheric heat transport and the polar albedo.

5.3 Suggestions: Further/Future Work

This work has provided understanding of how the important physical processes and reservoirs (those involved throughout the evolution of climate on Mars) interact and change with time. There is, of course, a great deal that remains unknown about the evolution of climate on Mars, as well as the present storage of CO_2 in exchangeable reservoirs. In this section we suggest directions to take in further modeling or planetary exploration.

The carbonate reservoir is possibly the greatest unknown store of CO_2 on Mars, and a reliable approximation for the magnitude of this reservoir would be very useful since limits on its size would aid in placing limits on the amount of initially available CO_2 . In 1996, a Thermal Emission Spectrometer (TES) will be flying on the orbiter of the Mars Surveyor mission, and if there are carbonates on the surface the TES should be able to detect them. If the TES does not detect carbonates, it becomes very difficult to justify the large carbonate reservoirs (>2 bar) which appear in the final states of model integrations that have a warm and wet early climate. A reliable estimation for the size of this reservoir will most likely require extensive surface exploration, since only the “tip of the iceberg” is detectable from orbit.

The formation of carbonates at low temperatures ($T_M < 273$ K) and in non-aqueous environments needs to be understood in greater detail. Our modeling has assumed that the chemical weathering parameterization is valid throughout the range of temperatures and pressures experienced by the model, but this approach has ignored that carbonate formation in a sub-freezing and non-aqueous environment is a different process than the parameterization assumes. Kahn (1985) proposes that liquid water can exist intermittently in disequilibrium at low annual pressures and temperatures, which provides the liquid water and allows aqueous reactions in annual climates where this water is unstable. Kahn argues that there is a minimum overburden atmospheric pressure P^* , below which intermittent pockets of water cannot exist, since evaporation rates would be much larger than melting rates below this

pressure, making liquid water impossible. Using present martian insolation (melting ability), Kahn puts a very crude upper bound on P^* of 30 mbar. The value of P^* would have been larger earlier, and would decrease with increasing luminosity. The important consequence is that once the total atmospheric pressure has been reduced to P^* , transitory pockets of water cannot form, and atmospheric CO_2 can no longer be removed by this efficient mechanism. It would be valuable to include these ideas in a scheme that parameterizes chemical weathering in climates ranging from warm and wet, through climates with intermittent disequilibrium liquid water, to the eventual inefficient rate in a completely non-aqueous climate.

As atmospheric heat transport is such an important quantity in the polar heat balance, and since it has been included in so highly a parameterized fashion within our model, a more accurate method to estimate this quantity is needed. The colossal scale of martian topography would most likely cause large climatological stationary eddies, and further complicate the poleward transfer of energy due to the existence of large hemispheric asymmetries. The best solution to this problem would be to carry out a suite of experiments with a 3-D general circulation model at a variety of surface pressures. These results could be fit with a more dependable parameterization, and this parameterization could be used in future climate modeling.

The condensation of CO_2 in the early atmosphere is a phenomena that also deserves further study. Since the work of Kasting (1991) was with a 1-D model, the onset of condensation (which would likely occur in polar regions) and the spread of

this condensation throughout the atmosphere cannot be more than qualitatively discussed. To study atmospheric condensation more carefully, it seems that a 2-D model is required. Moreover, the resulting greenhouse with atmospheric condensation would quite likely differ from that of Kasting (1991).

If early Mars was warm and wet for a significant time, the mechanism to allow this would seemingly be the Greenhouse Effect, one strengthened by gasses in addition to CO_2 , such as methane, sulphur-dioxide and ammonia. However, questions such as what are the most likely gasses, and how were these gasses maintained in the early atmosphere (sources and sinks), would have to be addressed before such theories could become generally viable.

The model parameterizes the escape of CO_2 , via the sputtering of reentering O^+ ions, in time only. However, according to Luhmann (1993), there are other mechanisms which could have also caused significant CO_2 losses after the end of heavy bombardment. If all of the possible escape mechanisms could be parameterized and included somehow in the model, a better estimation of the importance of such losses would be the result.

Finally, this study has entirely ignored the effect of CO_2 clouds in the planetary heat balance, a critical climatological concern (the greenhouses used are based upon model calculations which do not include the effects of clouds). The effect of clouds on the climate of Earth is becoming more apparent as further work is done, but radiative and microphysical properties are complicated and difficult to model. For Mars, there

can be no data concerning the behavior of CO_2 clouds in a massive CO_2 atmosphere. and because the radiative properties of such clouds are relatively unknown, their effects upon ancient climates must also remain unknown. The cloud/albedo feedback, and the warming or cooling effect of such clouds, would certainly be important questions to address with a 3-D model with the appropriate cloud physics. If addressed throughout a wide range of surface pressure and luminosity values, a more realistic greenhouse could be determined.

Mars is a very important planet to study. From the perspective of climate and atmospheric evolution there are many analogies (as well as striking differences) with Earth. For two reasons Mars is a unique laboratory as well: it has survived somewhat intact from the early days of the solar system, and it also holds evidence of recent climatological changes. The surface of Mars, being a potpourri of "snapshots" during the evolution of our solar system, *will* provide valuable assistance in our search to further unravel the mysteries of climate evolution on Mars, Earth, and earthlike planets in general. I for one would join a mission to Mars which had a realistic chance of returning (maybe even one that didn't), and I believe many others would as well, since a trip to Mars would surely help to educate me of the "great mystery" from which I was born.

Appendix A

List of Symbols

a	Mean radius of Mars (3.4×10^6 m)
A_e	Albedo of the equatorial region of Mars
A_p	Albedo of the polar region of Mars
c_p	Specific heat capacity of atmosphere ($800 \text{ J (K kg)}^{-1}$)
C	Constant in regolith parameterization to set present day storage
C_H	Heat transfer coefficient (0.0009)
CA	Conduction of heat between surface and atmosphere (annual)
CS	Conduction of heat between surface and atmosphere (summer)
CW	Conduction of heat between surface and atmosphere (winter)
e	Eccentricity
E	Atmospheric loss rate of CO_2 to space

f	Coriolis parameter (10^{-4} sec^{-1})
F	Equatorial to polar dynamic heat flux (W)
g	Acceleration due gravity on Mars (3.72 m sec^{-2})
$g_1 - g_5$	Factors determining strength of weathering parameterization
$\mathcal{L}$	Solar luminosity, a function of time
$\mathcal{L}_i$	Initial Solar luminosity, 3.8 Gyr ago
p	Surface atmospheric CO_2 partial pressure
P_{cap}	CO_2 stored in permanent polar caps
P_{carb}	CO_2 in carbonate rock due chemical weathering
P_{escape}	CO_2 that escaped to space by O^+ sputtering
$p_{fp}(T_{ps})$	CO_2 equilibrium vapor pressure over CO_2 ice at T_{ps}
P_{reg}	CO_2 adsorbed in the planetary regolith
P_{tot}	CO_2 available to model system at start up
Q_R	Polar winter atmosphere radiation: absorption and emission
Q_C	Polar winter conduction
Q_D	Atmospheric heat transport
R	Ideal gas constant ($190 \text{ J (K kg)}^{-1}$)
$RATE$	Chemical weathering rate
RA	Radiation from atmosphere to surface (annual)
RS	Radiation from atmosphere to surface (summer)
RW	Radiation from atmosphere to surface (winter)

$\bar{S}$	Static stability of atmosphere (0.002 K m^{-1})
S_{es}	Annual insolation absorbed by equatorial surface
S_{ps}	Annual insolation absorbed by polar surface
Δt	Model timestep (7.6 Myr)
ΔT	ϵ folding temperature change in regolith parameterization
T_0	Mean temperature used in model when finding β
T_{ea}	Annual equatorial atmospheric temperature
T_{es}	Annual equatorial surface temperature
T_{fp}	CO_2 frost point temperature
T_M	Annual planetary surface temperature
T_{ps}	Annual temperature of polar surface
T_{psa}	Temperature of polar summer atmosphere
T_{pwa}	Temperature of polar winter atmosphere
$\overline{v'T'}$	Magnitude of heat transfer by transient eddies
α	Coefficient for the strength of the equatorial to polar dynamic transfer of heat
α_e	Coefficient for equatorial regions adjusted for area difference from polar region
β	Heat conduction coefficient between atmosphere and surface
Δt	Model timestep: 7.6 Myr
$\Delta\Theta$	Magnitude of obliquity deviation from the mean
ϵ	Atmospheric emissivity
γ	Pressure sensitivity in regolith storage parameterization

λ	Aerocentric longitude
ϕ	Latitude
ρ	Atmospheric density
Θ	Obliquity
Θ_0	Mean value of obliquity
σ	Stephan Boltzman constant
$\tau_{\Delta\Theta}$	Secondary Θ variation period: $1.3(10^6)$ yr
τ_{reg}	Regolith response time for thermal equilibrium
τ_{Θ}	Primary Θ variation period: $1.25(10^5)$ yr
τ_w	Chemical weathering time scale of martian atmosphere

Appendix B

Basal Melting of a CO_2 Ice Cap

The pressure at the base of a permanent polar cap could possibly surpass the triple point pressure of CO_2 (5.2 bar) in some of the evolutionary scenarios of our model. For a single permanent polar cap of uniform thickness, which extends equatorward to latitude ϕ , the pressure due to its weight at the base is

$$P_{\text{base}} = \frac{2P_{\text{cap}}}{1 - \sin \phi}. \quad (\text{B.1})$$

The important consideration for basal melting (assuming the pressure is great enough) is whether or not the temperature at the base of this cap can reach the triple point temperature (216.5 K). When this temperature is reached, the base of the cap will melt. The consequences of this are not fully understood (B. Haberle 1994, personal communication), but it is probable that such a permanent cap (with temperatures and pressures at the base greater than the CO_2 triple point values) would be climato-

logically unstable. The greater ancient fluxes of internal heat (due to the dissipation of the heat of accretion and radiogenic sources) would allow temperatures at the base of a very massive permanent cap to rise above the triple point.

To quantify this further, we determine the temperature gradient within large permanent caps. Assuming that early internal heat flux through the polar cap can be expressed with

$$\mathcal{H} = \kappa \frac{\partial T_{\text{cap}}}{\partial z}, \quad (\text{B.2})$$

where κ is the thermal conductivity of CO_2 ice, and $\frac{\partial T_{\text{cap}}}{\partial z}$ is the vertical temperature gradient within the polar cap, we integrate from the surface of the cap down to the base (through a vertical distance δz), and find the temperature of the base to be

$$T_{\text{base}} = T_{ps} + \frac{\mathcal{H}}{\kappa} \delta z. \quad (\text{B.3})$$

We use the hydrostatic equation to determine δz , such that

$$\delta z = \frac{P_{\text{base}}}{g \rho_{\text{ice}}}, \quad (\text{B.4})$$

where ρ_{ice} is the density of CO_2 ice, and g is again the acceleration of gravity on Mars. The above equations are combined, and the following expression results:

$$T_{\text{base}} = T_{ps} + \frac{\mathcal{H}}{\kappa} \frac{2P_{\text{cap}}}{g \rho_{\text{ice}}(1 - \sin \phi)}. \quad (\text{B.5})$$

Figure B.1: The mass of CO_2 required in a permanent polar cap to initiate basal melting versus different internal heat fluxes (temperature gradients). Mass curves are shown for six different latitudes to which the permanent cap extends equatorward.

Assuming T_{base} is the triple-point temperature of CO_2 , and using an appropriate value for the polar surface temperature, Fig. B.1 is constructed by solving Eqn. B.5 for P_{cap} . This is done by using different values for the internal heat flux $\frac{H}{\kappa}$ (actually the temperature gradient within the cap), and ϕ (the latitudinal extent to which the cap extends equatorward), and then finding P_{cap} .

These calculations indicate that when the temperature gradient in the cap is large (very early times $> 100 \text{ K km}^{-1}$), the maximum storage of a permanent cap is limited to approximately 2 bar, although for the present day ($< 50 \text{ K km}^{-1}$) a permanent cap could be much larger. Since our model predicts early permanent cap storages of

> 1 bar, this analysis does not rule such situations out. but it indicates that the cap would have to cover a significant areal extent ($\phi < 65^\circ$) to remain stable.

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