Comparative terrestrial planet thermospheres

3. Solar cycle variation of global structure and winds at solstices

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Abstract. The comparison of planetary upper atmospheres using global databases has entered a new era with the advent of recent aerobraking measurements of the Mars thermosphere [e.g., Keating et al., 1998a]. The present maturity of available modeling capabilities also permits us to contrast the Earth and Mars thermosphere structures, winds, and controlling processes using global three-dimensional models [e.g., Bougher et al., 1999b]. This present effort focuses upon the comparison of the combined seasonal solar cycle responses of the thermospheres of Earth and Mars using the National Center for Atmospheric Research (NCAR) Thermospheric General Circulation Model (TGC) utility to address the coupled energetics, dynamics, and neutral-ion composition above ~100 km. Extreme thermospheric conditions are expected at solstices, thereby revealing the changing importance of fundamental physical processes controlling the Earth and Mars thermospheric structures and winds. Seasonal-solar cycle extremes in Mars exobase temperatures are calculated to range from 200 to 380 K, giving rise to maximum horizontal winds of nearly 215 to 400 m/s. Corresponding extremes in Earth exobase temperatures are 700 to 1600 K, with rather small variations in global winds. The orbital eccentricities of Earth and Mars are also shown to drive substantial variations in their thermospheric temperatures. For Mars, dayside exobase temperatures vary by ~60 K (18%) from aphelion to perihelion during solar maximum conditions. Such large temperature variations strongly impact thermospheric densities and global winds. The corresponding Earth dayside temperatures also vary by 60-80 K between solstices. However, the percent temperature variation (5%) over the Earth’s orbit and its overall impact on the thermospheric structure and winds are much smaller. Auroral activity may in fact obscure these orbital variations. Changing dust conditions throughout the Martian year modulate the aerosol heating of its lower atmosphere, yielding considerable variability in the height of the subsolar ionospheric peak about its observed seasonal trend (~115-130 km). Significant further progress in the comparison of Earth and Mars thermospheric features and underlying processes must await expanded Mars global databases expected from Planet-B and Mars Express (2004-2005).

1. Introduction and Motivation for Solstice Studies

The long-term objective of our program in comparative terrestrial planetary atmospheres is to contrast and compare the physical and chemical processes responsible for the structure and dynamics of the thermospheres of Venus, Earth, and Mars. These upper atmospheres are driven by similar forcing agents and subsequently change over time, both naturally and, in the case of the Earth, as a result of human influence. Fundamental planetary parameters for these terrestrial planets are similar yet sufficiently unique to provide independent “laboratories” to examine thermospheric processes controlling energetics, dynamics, and composition. One-dimensional (1-D) global mean heat balance and composition models [see Bougher and Roble, 1991]
paper 1) have been used extensively to compare planetary thermosphere processes. Most recently, three-dimensional (3-D) Thermospheric General Circulation Models (TGCMs) for Venus, Earth, and Mars have been exercised to examine the solar cycle responses of each of these upper atmospheres during equinox conditions [see Bougher et al., 1999b] (paper 2). Both of these previous efforts have sought to simulate the observed responses of these individual planets to solar EUV/UV flux variations. The latter combines thermal and dynamical feedbacks to address the relative importance of common processes that regulate solar cycle behavior.

The acquisition of Mars thermospheric data from recent aerobraking activities [e.g., Keating et al., 1998a,b, 1999a;b; Bougher et al., 1999a] provides the beginnings of a Mars global database required for the detailed comparison of the Earth and Mars upper atmospheres. The Earth database is presently far more complete than that of Mars (see Figure 1). Nevertheless, first order thermospheric trends can be identified for Mars and compared to Earth. These global databases are typically organized as a function of solar cycle and season. The F10.7-cm radio flux index is often used to select solar EUV and UV flux data sets corresponding to various activity levels throughout the solar cycle; that is, the F10.7 cm index is a daily measured proxy for solar EUV and UV fluxes. In addition, Mars seasons are denoted by the “Ls” parameter corresponding to Mars aecocentric longitude. This is an angular measure for the advance of its seasons (Ls = 0, equinox; Ls = 90, northern summer solstice; Ls = 270, southern summer solstice).

In this paper we extend the terrestrial planet TGCM comparisons by examining combined seasonal and solar cycle behavior at solstices for Earth and Mars. Such a comparison is reasonable, since the obliquities of Earth (23.5°) and Mars (25.2°) are quite similar (seasonal effects), and the orbital eccentricities of Earth (0.017) and Mars (0.093) are greater than zero (heliocentric distance variations). These parameters suggest that seasonal-orbital impacts on thermospheric structure and dynamics should be significant for Earth and Mars (see Table 1b from paper 2). The combined seasonal orbital phasing furthermore determines that perihelion presently occurs near southern summer solstice on Earth (early January) and Mars (Ls = 270). It is expected that the Earth and Mars seasonal solar cycle responses give rise to annual extremes in the global temperature and wind structure at solstices. These extremes provide an important validation of the Earth and Mars TGCM codes, revealing the changing importance of fundamental physical processes controlling their thermospheric structures and dynamics. In addition, ongoing Mars aerobraking exercises require an improved understanding of variable thermospheric processes, especially at solstices during aerocesses over the winter polar night [Keating et al., 1999a;b; Bougher et al., 1997a, 1999a].

The physical processes generally responsible for maintaining the Earth and Mars thermospheric structure and dynamics are described in detail elsewhere (see paper 2 and references therein). Processes most important for regulating thermospheric temperatures over the solar cycle include (1) EUV/UV heating, (2) molecular

![Figure 1](https://example.com/figure1.png)

**Figure 1.** Mars Ls versus heliocentric distance. Spacecraft seasonal and solar cycle sampling of the Mars upper atmosphere to date. Solar cycle minimum (min, filled diamonds), moderate (mod, triangles), and maximum (max, squares) observations are indicated.
thermal conduction, (3) CO$_2$-O enhanced 15-µm cooling, (4) NO 5.3-µm cooling (Earth), and (5) global dynamics with the associated adiabatic heating and cooling that follow. We briefly review what has been gleaned recently from TGCMS simulations (paper 2) regarding the relative importance of CO$_2$ cooling and global dynamics in controlling the Earth and Mars thermospheric temperatures.

CO$_2$ 15-µm emission is indeed a major cooling agent in the lower thermospheres of all three terrestrial planets; mesopause temperatures and heights are in large part determined by CO$_2$ cooling [e.g., Bouger et al., 1994, 1999b]. However, its relative importance as a thermostat regulating overall thermospheric temperatures is different for each planet. In general, CO$_2$ 15-µm cooling is highly nonlinearly temperature dependent, resulting in a potential thermostatic control depending on the relative magnitude and vertical placement of the peak cooling rate in these thermospheres [e.g., Bouger and Roble, 1997]. In addition, the relative abundance of atomic oxygen in these lower thermospheres further governs the efficiency of oxygen atom collisions in enhancing 15-µm cooling (see paper 2). For Mars, molecular thermal conduction presently provides the primary cooling offsetting peak EUV heating over 150-200 km; CO$_2$ 15-µm cooling peaks at a slightly lower altitude. This balance determines that Mars thermosphere temperatures are very sensitive to molecular conduction, which operates under vertical temperature gradients. Furthermore, the Mars heliocentric distance (1.38-1.67 AU) controls CO$_2$ photolysis, yielding small atomic oxygen abundances in its lower thermosphere (2-4% at the dayside ionospheric peak at ~60° solar zenith angle). Hence CO$_2$ cooling does not appear to be dominant in the dayside heat budget of the Mars thermosphere, producing rather large exospheric temperature variations (~110 K) over the solar cycle [see Bouger et al., 1999b]. Likewise, Earth CO$_2$ cooling is important below 130 km, well removed from the peak Earth heating at higher altitudes [Roble et al., 1987; Bouger et al., 1994] Instead, molecular thermal conduction generally controls Earth cooling above 150 180 km. During strong auroral conditions, NO 5.3-µm cooling also provides significant IR cooling. This “weak thermostat” yields a large dayside exospheric temperature variation over the solar cycle (~840 K).

Global winds also play a role in modifying Martian thermospheric temperatures, increasingly so as the solar cycle advances [e.g., Bouger, et al. 1990] (see paper 2, Table 2). Exobase horizontal winds nearly double for Mars over the solar cycle, while those for Earth change only slightly. Corresponding Mars vertical (dayside upwelling) winds also double, and are of significant magnitude (~3 m/s). These upwelling winds are predicted to have a significant impact on the Mars dayside thermal budget; that is, adiabatic cooling is calculated to be strong enough to compete with molecular conduction for control of Mars dayside thermosphere temperatures during solar maximum conditions [Bouger et al., 1990, 1999b]. By contrast, the advance of the solar cycle produces larger thermospheric pressure gradients for Earth that are offset by enhanced ion drag, due in part to elevated ionospheric densities. Thus terrestrial thermospheric winds at low latitudes to midlatitudes do not change greatly with the solar cycle and are therefore not important as a thermostat in regulating temperatures.

The net effect of these IR cooling and global wind processes is to determine a unique timescale for thermal equilibration in each of the Earth and Mars thermospheres [see Bouger et al., 1999b]. As global model simulations have shown (paper 2), molecular thermal conduction plus global winds combine to regulate Mars dayside temperatures over the solar cycle. Both of these mechanisms are less efficient than nonlinearly temperature dependent CO$_2$ 15 µm cooling. As a result, the thermal equilibration timescale for the Mars dayside is predicted to be ~4-12 hours (less than 1 day) at the exobase. This implies that solar EUV/UV flux perturbations will result in Mars thermospheric responses about one-half day later; for comparison to Earth, the Earth-Sun-Mars angle must also be taken into account. On Earth the major IR radiators (CO$_2$ and NO) are minor constituents which provide significant cooling only near the bottom of the Earth’s thermosphere. This leaves molecular conduction, a rather inefficient cooling mechanism, to largely balance terrestrial thermospheric heating above ~150-180 km. Thus the Earth thermal equilibration timescale is nearly 2-5 days in the upper thermosphere; solar EUV flux variations require several days to be fully manifested in the Earth’s upper atmosphere following a flare. Do these same thermostatic processes control thermospheric temperatures in a similar manner during solstice conditions as they do during equinox?

The Earth Thermosphere, Ionosphere, Electrodynamics GCM (TIE-GCM) and the Mars Thermosphere GCM (TGCM) codes are exercised in this paper in order to investigate the combined seasonal-solar cycle responses of these thermospheres at both solstices. Standard EUV/UV fluxes are utilized for both models, spanning solar cycle minimum (SMIN; F10.7 = 68), moderate (SMOD; F10.7 = 130), and maximum (SMAX; F10.7 = 200) conditions. Subsequently, seasonal responses are isolated by using constant solar fluxes (F10.7 = 130) for both summer and winter solstices. The goal here is to isolate the seasonal asymmetries of the Earth and Mars thermosphere structures and global winds for various solar flux conditions. Simulations presented in this paper confirm that these seasonal asymmetries are very strong for Mars [Bouger et al., 1990] and rather modest for Earth. Ultimately, these seasonal-solar cycle characterizations provide an improved upper atmosphere climatology for future Mars aerobraking operations [e.g., Keating et al., 1998a; Bouger et al., 1999a].
2. Earth and Mars Three-Dimensional Models and Standard Inputs

The Mars TCGM (MTGCM) and the Earth TIE-GCM codes are described in detail in paper 2. We review here the salient features required to set the context for solstice model simulations to follow.

2.1. MTGCM Hydrodynamic Model

The MTGCM itself is a finite difference primitive equation model that self-consistently solves for neutral temperatures, neutral-ion densities, and three component neutral winds over the globe [Bougher et al., 1988, 1990, 1993, 1997a, 1999a,b]. Prognostic equations for the major neutral species (CO_2, CO, N_2, and O), selected minor neutral species (Ar and O_2), and several photochemical ions (e.g., O^+_3, CO_3, O^+, and NO^+ below 200 km) are included. Zonal, meridional, and vertical velocities, total temperatures, and geopotential heights are also obtained on 33 pressure levels (above 1.32 μbar), corresponding to ~70-300 km (SMAX conditions), with a 5° latitude and longitude resolution. The vertical coordinate is log-pressure, with a vertical spacing of two grid points per scale height. Adjustable parameters which can be varied for individual MTGCM cases include the F10.7 index (solar EUV/UV flux variation), heliocentric distance (orbital variation), solar declination (seasonal variation), and the maximum eddy coefficient (K_e) for eddy diffusion and viscosity. The MTGCM can also be modified to accommodate atmospheric inflation and semidiurnal (2,2) to (2,6) tidal mode amplitudes and phases consistent with dusty conditions present in the Mars lower atmosphere during dust storm events [Bougher et al., 1993, 1997a, 1999a].

The MTGCM has been validated using an assortment of spacecraft observations taken throughout the solar cycle at different seasons [e.g., Barh et al., 1992] (see Figure 1). Careful reconstruction of Mars thermospheric conditions for global modeling exercises requires that seasonal (Ls), orbital, and solar flux inputs (corresponding to available observations) all be properly specified (see Table 1). There is a notable dearth of thermospheric measurements for constraining the MTGCM during SMAX conditions. Most recently, Mars Pathfinder (MPF) descent data [Schofield et al., 1997; Magalhaes et al., 1999] and Mars Global Surveyor (MGS) Accelerometer Experiment data (110-170 km) obtained during its two aerobraking phases (Ls = 184-300; 30-95) [Keating et al., 1998a,b, 1999a,b] have been used to validate the simulated MTGCM vertical, latitudinal, and diurnal structure [Bougher et al., 1999a,b] for SMIN to SMOD conditions (F10.7 = 80-150). Beginning in 2004, the Japanese Planet-B (Nozomi) mission is slated to arrive at Mars and investigate the thermosphere and the solar wind-ionosphere interaction above ~150 km over a Martian year. During this same period, remote measurements from Mars Express will probe upper atmospheric structure (50-140 km). Solar cycle variations of Mars thermospheric temperatures, densities, and airglow intensities have been crudely observed thus far (see Table 1). However, with the advancement of solar cycle 73 (1997-2006), these MGS, Planet-B, and Mars Express missions will make in situ and remote measurements spanning nearly SMIN to SMAX conditions. In this paper we rely upon recent MGS Accelerometer Experiment measurements of Mars thermospheric densities and inferred temperatures to constrain the MTGCM for SMIN and SMOD conditions [e.g., Keating et al., 1998a,b, 1999a,b].

2.2. TIE-GCM Hydrodynamic Model

The National Center for Atmospheric Research TIE-GCM is the latest in a series of three-dimensional gen-

Table 1. Mars Spacecraft Observations of the Upper Atmosphere

<table>
<thead>
<tr>
<th>Mission</th>
<th>Dates(s)</th>
<th>F10.7</th>
<th>Ls</th>
<th>Dsm</th>
<th>SZA</th>
<th>T_{exo}</th>
</tr>
</thead>
<tbody>
<tr>
<td>M4</td>
<td>July 15, 1965</td>
<td>77.0</td>
<td>139.0</td>
<td>1.553</td>
<td>67.0</td>
<td>212.0</td>
</tr>
<tr>
<td>M67</td>
<td>July 31, 1969</td>
<td>197.0</td>
<td>200.0</td>
<td>1.425</td>
<td>0-44.0</td>
<td>315-350.0</td>
</tr>
<tr>
<td>M9N</td>
<td>Aug. 5, 1969</td>
<td>188.0</td>
<td>306.0</td>
<td>1.440</td>
<td>50-60.0</td>
<td>325.0</td>
</tr>
<tr>
<td>M9E</td>
<td>fall, 1971</td>
<td>198.0</td>
<td>30.0</td>
<td>1.630</td>
<td>50-90.0</td>
<td>260.0</td>
</tr>
<tr>
<td>VL1</td>
<td>spring, 1972</td>
<td>100.0</td>
<td>48.0</td>
<td>1.647</td>
<td>44.0</td>
<td>186.0</td>
</tr>
<tr>
<td>VL2</td>
<td>July 20, 1976</td>
<td>69.0</td>
<td>117.0</td>
<td>1.612</td>
<td>44.0</td>
<td>145.0</td>
</tr>
<tr>
<td>M9F</td>
<td>July 3, 1976</td>
<td>76.0</td>
<td>143.0</td>
<td>1.557</td>
<td>135.0</td>
<td>153.0</td>
</tr>
<tr>
<td>MPF</td>
<td>Jan. 1, 1998</td>
<td>93.0</td>
<td>256.0</td>
<td>1.382</td>
<td>73.5</td>
<td>220.0</td>
</tr>
<tr>
<td>MGS2</td>
<td>Oct. 27, 1998</td>
<td>127.0</td>
<td>48.5</td>
<td>1.653</td>
<td>57.0</td>
<td>230.0</td>
</tr>
</tbody>
</table>

F10.7 refers to the 10.7-cm index used to select reference EUV/UV flux data sets; Ls, refers to, the angular measure of the Mars seasons (Ls = 90 is northern summer solstice, Ls = 270 is southern summer solstice, etc.); Dsm refers to the Mars heliocentric distance (AU); SZA refers to solar zenith angle; and T_{exo} refers to exospheric temperature. Spacecraft are indicated as follows: M4 (Mariner 4), M67 (Mariner 6 and 7), M9E (Mariner 9 extended), M9N (Mariner 9 nominal), VL1 (Viking Lander 1), VL2 (Viking Lander 2), MPF (Mars Pathfinder), MGS1 (Mars Global Surveyor phase 1 aerobraking sample), MGS2 (Mars Global Surveyor phase 2 aerobraking sample).
eral circulation models of the Earth’s upper atmosphere and ionosphere that have been developed during the past two decades [e.g., Dickinson et al., 1981, 1984; Roble et al., 1982; Feleson et al., 1986; Roble and Ridley, 1987; Roble et al., 1987, 1988; Richmond et al., 1992]. The model solves for the three-dimensional time-dependent temperature and compositional structure, dynamics, and electrodynamics of the coupled thermosphere and ionosphere between 95 and 800 km altitude (SMAX conditions). The model is a first principles model, and it is driven by a time-dependent specification of solar EUV and UV spectral irradiance and magnetic conjugate auroral particle precipitation and ionospheric convection patterns at the upper boundary of the model. The TIE-GCM computes self-consistently the coupled thermosphere and ionosphere dynamics, the associated dynamo electric field and currents, and the electrodynamic feedback on the neutral and plasma motion and thermodynamics as described by Richmond et al., [1992]. Specifically, this model calculates global distributions of neutral gas temperature, winds, the major constituents N₂, O₂, and O, and several minor neutral constituents (N²D), (N⁴S), NO, He, and Ar). The Eulerian model of the ionosphere solves for global distributions of electron and ion temperatures, O⁺, O²⁺, NO⁺, N³⁺, N⁺, and electron number densities. Mutual couplings between the thermosphere and ionosphere occur at each model time step and at each point of the geographic grid. The TIE-GCM has an effective 5 degree latitude by longitude grid with 29 constant pressure surfaces in the vertical between approximately 95 and 800 km with a vertical resolution of two grid points per scale height and a model time step of 5 min. This model has been used for numerous studies of the thermosphere and ionosphere structure of the Earth’s upper atmosphere and its response to solar and auroral variability. Recently, eccentricity variations have been included in the TIE-GCM code. Further details of the TGCM systematic development strategy leading to the present TIE-GCM code can be found in paper 2 and references therein.

Solar cycle variations of Earth thermospheric temperatures, densities, winds, and airglow intensities have been monitored over the last 40 years. Empirical models have been constructed to capture these solar cycle responses of the terrestrial thermosphere structure and dynamics. The most recent and presently useful empirical models for the Earth’s upper atmosphere are the MSISE-90 (structure) and HWM (winds) codes [Hedin, 1991; Hedin et al., 1991], which are used to constrain the TIE-GCM simulations of this paper.

2.3. Standard Inputs and Forcings

The reference solar X-ray/EUV/UV fluxes (ranging from 2.0 to 200 nm) used by these two-planet thermosphere TGCMs are identical to those specified in paper 2. Standard parameterization methods are employed based upon F10.7-cm and 81-day average F10.7A-cm indices (see section 1). These proxies permit the generation of reference solar fluxes (from published reference spectra [Hinterreiter, 1981]), yielding typical solar F1 minimal and UV fluxes found at F10.7-cm indices of 220, 130, and 60. These indices are generally consistent with typical solar maximum (SMAX), solar moderate (SMOD), and solar minimum (SMIN) conditions, respectively.

These computationally time consuming TGCM models are generally not the proper environment to calculate neutral gas EUV and UV heating efficiencies and the first principles. Therefore use is made of detailed EUV and UV heating efficiencies calculated independently (off line) by others. The MTGCM solstice simulations of this paper follow this strategy and make use of 22% EUV and UV heating efficiencies [see Fox and Dalgarno, 1979], identical to those specified in paper 2. However, for Earth, Roble et al., [1987] and Roble [1995] described the details of EUV and UV energy partitioning and energy balance calculations. Account is taken of nonlocal heating processes. Similar detailed energy balance calculations are performed within the TIE-GCM code for the simulations of this paper.

Atomic oxygen collisions are known to be especially effective in exciting CO₂ (ν=2) vibrational states, resulting in enhanced 15-μm emissions and cooling at thermospheric heights where non-local thermodynamic equilibrium (NLTE) conditions prevail. The importance of this CO₂-O enhancement mechanism upon thermospheric 15-μm cooling rates depends upon (1) atomic O abundances, (2) CO₂ abundances, and (3) the collisional CO₂-O relaxation rate. It is important to consistently include a common CO₂-O rate in calculations of the Earth and Mars thermospheres [see Bouger et al., 1994, 1999b]. For the TIE-GCM and MTGCM simulations of this paper, a “most probable” CO₂-O deactivation rate coefficient of 3 x 10⁻¹² cm³/s (at 300 K) is employed, in accord with recent model studies described in paper 2.

The eddy diffusion coefficient profiles used in both these TGCM codes are each parameterized according to standard aeronomic formulations. A key feature to notice about Mars is that the magnitude of the eddy coefficient is much less (factor of 3-5) than corresponding 1-D continuity diffusion models would suggest [Rodrigo et al., 1990; Krasnopolsky, 1993]. For Mars a maximum eddy coefficient of ~1.0-2.0 x 10⁵ cm²/s (peaking near 125 km) is employed in the MTGCM code, enabling atomic O mixing ratios near the dayside F1 ionospheric peak to match limited observations. Apparently, the global-scale circulation plays a major role in the determination of Mars vertical density profiles, thereby reducing the requirement for vertical eddy diffusion. Corresponding eddy conduction is thus quite limited in its ability to efficiently cool the Mars dayside thermosphere. For the Earth the Roble et al. [1987] 1-D global mean model simulations of observed densities require the maximum eddy coefficient to be ~1.6
x $10^6$ cm$^2$/s (near 100 km), in accord with the minimal role of eddy conduction in the energy budget of the terrestrial mesopause region. A similar eddy coefficient parameterization is employed in the TIE-GCM code.

Tidal parameters are specified at the lower boundaries of the TIE-GCM and MTGCM codes consistent with typical values for summer and winter solstices. These tidal parameters (diurnal and semidiurnal amplitudes and phases) are well documented according to ground-based and spacecraft observations of the Earth’s mesosphere and lower thermosphere (MLT) region [e.g., Forbes, 1982; McLandress et al., 1996]. However, no Mars MLT data presently exist that describe global tidal parameters as a function of season and longitude. This situation will change with the advent of Mars Express MLT structure data from the UV Atmospheric Sounder (SPICAM) instrument (50-140 km) [Bertaux, 1999]. For now, the MTGCM tidal parameters (semidiurnal amplitudes and phases) are taken from the NASA Ames Mars General Circulation Model (MGCM) of the Mars lower atmosphere [cf. Murphy et al., 1995; Bridger and Murphy, 1998], which was exercised for a Mars year for dust-free conditions. The exchange of tidal parameters from an MGCM upper level to the MTGCM lower boundary (at roughly 70 km) is described in detail by Bougher et al., [1993, 1997a]. Dust heating resulting from regional or global storms modifies the Mars tidal parameters significantly, thereby greatly enhancing the tidal impact on the Mars thermosphere during dust storm events [Bougher et al., 1997a, 1999a,b]. This dust influence is most prominent around southern summer solstice (Ls = 200-300), and is discussed further in section 5.2.

Low auroral activity parameters were chosen for implementation in the Earth TIE-GCM for all solar cycle cases conducted. The purpose here is to focus upon the model variations due to solar EUV/UV flux variability, with minimal influence from auroral processes. Such conditions are not uncommon, especially during solar minimum periods. For the TIE-GCM simulations of this paper, the cross-tail potential was held constant at 45 kV, while the integrated global power input was selected to be 16.0 GW.

3. Combined Solar Cycle-Seasonal Variations at Solstices

We take two approaches to describe the Earth and Mars model results for solstice conditions as simulated for various solar fluxes. First, a simple comparison of steady state model fields is presented for combined solar and seasonal-orbital forcing for each planet individually. Specific cases for solar maximum (SMAX) (F10.7=200), solar moderate (SMOD) (F10.7=130), and solar minimum (SMIN) (F10.7=68) conditions at the solstices are contrasted, and their output fields are examined. This provides an overview of the solar cycle plus seasonal responses of the temperatures, neutral densities, and winds for each planet. A comparison of the terrestrial and Martian combined seasonal-solar cycle responses is also provided to further illustrate the thermal and dynamical feedbacks described earlier in paper 2. Second, solar forcing is fixed at SMOD conditions and diurnal, latitudinal, and altitude variations are examined for seasonal extremes (solstices). Earth and Mars simulations are examined for asymmetries between winter and summer solstice responses; these differences reflect the seasonal-orbital variations described in section 1. In this paper, black and white simple line and contour figures are employed to illustrate the key model fields and their variations over the solar cycle and season. In addition, a Web site has been established at the University of Arizona that presents color figures for many of these same TIE-GCM and MTGCM simulated cases and their corresponding fields (see section 6).

3.1. MTGCM Solar Minimum to Maximum

In this paper the MTGCM model is exercised to address solstice conditions at Mars for SMIN to SMAX solar flux inputs (see section 2). Figure 2 illustrates simulated exobase (roughly 190 to 220 km) and homopause (roughly 115 to 130 km) temperatures at the equator for midafternoon conditions (solar local time of ~1500). This corresponds roughly to 45°-60° solar zenith angle (SZA), consistent with most of the available Mars thermospheric measurements (see Table 1). Northern summer solstice (Ls = 90), equinox (Ls = 180), and southern summer solstice (Ls = 270) simulated temperatures are displayed together on SMIN, SMOD, and SMAX curves (solid) for comparison to superimposed spacecraft observations (exobase only). The SMIN to SMAX variation of Mars exobase temperatures is calculated to be 110 K (200-310 K) near aphelion (Ls=90) and 150 K (220 to 370 K) near perihelion (Ls=270). SMAX temperatures range from 310 K (at aphelion) to 370 K (at perihelion), a difference of ~60 K due to the orbital eccentricity of Mars. By contrast, SMIN temperatures range from 200 K (at aphelion) to 220 K (at perihelion), displaying a smaller 20 K variation. It is clear that MTGCM simulated exobase temperatures generally match available observations within ~20 K. Mars Global Surveyor (MGS) topside temperatures plotted (MGS1 and MGS2) both refer to Accelerometer-derived values averaged over several aeropasses at 160 km [Keating et al., 1998b]. It is estimated that these 160-km temperatures are 5 K (SMIN) to 10 K (SMOD) cooler than actual exobase values [Bougher et al., 1999b] (see Table 1); this improves the match of MTGCM exobase and MGS1 and MGS2 estimated exobase temperatures. Notable model-data discrepancies occur for VL2 and M9N measurements, both of which were obtained during periods of significant nonsolar forcing. For the latter a global dust storm was raging during the Mariner 9 nominal mission, significantly impacting thermospheric temperatures [e.g., Stewart et al., 1972, 1992]. VL1 and VL2 temperature profiles were both strongly influenced
Figure 2. Mars exosphere and homopause temperatures at LT = 1500 near the equator. Exospheric temperatures from various spacecraft measurements (solar cycle symbols from Figure 1) are compared to MTGCM simulations (solid lines). Corresponding MTGCM homopause temperatures are indicated by dotted lines, with spacecraft measurements described in the text.

by large amplitude (10°-20° K) and long vertical wavelength (20-40 km) oscillations [Seiff and Kirk, 1977; Magalhaes et al., 1999]. The VL2 profile in particular does not show a distinct topside isothermal thermosphere.

Simulated homopause temperatures are also plotted in Figure 2 (dotted curves). The homopause is defined as the 1.26-nbar pressure level at 45°-60° SZA, which is known to vary by ~15 km (115 to 130 km) from Mars aphelion (LS = 70) to perihelion (LS = 250) [Stewart, 1987; Bougher et al., 1997a]. Available homopause observations are thus far limited to VL1 and VL2 profiles (115 to 150 K) [Seiff and Kirk, 1977], MGS1 aeropass near perihelion (~150 K), and MGS2 aeropasses just prior to aphelion (~140 K) [Keating et al., 1998b]. VL1 and VL2 homopause values are quite low compared to MTGCM predictions. This is largely due to the wave structure evident in VL1 and VL2 profiles that is not addressed by the mean simulations of the MTGCM.

Very limited nightside thermospheric data are available for comparison to MTGCM calculations. Figure 3 shows similar equatorial MTGCM simulated curves for SMAX, SM01, and SM1N conditions, now at LT = 2400. Mars Pathfinder (MPF) passed through the nightside thermosphere at LT = 0300 (see Table 1 for further details). The measured maximum temperature of 153 K [Magalhaes et al., 1999] may be close to an exospheric value, and appears to match MTGCM (SM1N) temperatures quite well. Conversely, phase 2 of MGS aerobraking ended on the nightside near LT = 0100-0200. Topside temperatures derived from Accelerometer data (near its sensitivity limit) at southern mid-latitudes may not be close to exospheric values (G. M. Keating, private communication, 1999). Clearly, the Mars nightside thermosphere is very poorly sampled at the present time. Detailed comparisons with a global circulation model must await more complete nightside thermosphere coverage.

How does the combined variation of solar cycle and season impact Mars atomic oxygen abundances? Figures 4a and 4b compare subsolar latitude variations of the O/CO₂ mixing ratio for the northern summer solstice (near aphelion, Figure 4a) and the southern summer solstice (near perihelion, Figure 4b). Curves are labeled for SM1N, SM0D, and SMAX conditions at LT = 1500 (solid) and LT = 0000 (dashed). The O/CO₂ ratio at the altitude of the F1 ionospheric peak is typically quoted as a measure of the degree of net CO₂ dissociation in the Mars upper atmosphere [e.g., Stewart, 1987; Bougher et al., 1997a]. For aphelion (Figure 4a), the F1 peak is located at ≤120 km, for which the O abundance is 1.6% (SM1N) to 1.9% (SMAX). This variation is rather small, reflecting minimal control by solar-driven photochemistry, and is quite close to the level where a reversal occurs in the solar cycle response. Above ~125 km, maximum O/CO₂ values occur for SM1N conditions, owing to the large effect of dynamics (global advection and diffusion) upon atomic oxygen densities [Stewart et al., 1992; Bougher et al., 1999b]. Winds deplete the midafternoon upper thermosphere of oxygen atoms more effectively during SMAX conditions (strongest winds) than SM1N conditions (weakest winds). As a result, atomic oxygen surpasses CO₂ as the dominant species near 232 km (SMAX) and 190 km (SM1N). For perihelion (Figure 4b), the F1 peak is lo-
cated at 130-135 km; the corresponding O abundance is roughly 2.0% throughout the solar cycle. At this level, both photochemistry (i.e., CO$_2$ photolysis) and global dynamics (advection) combine to mitigate any solar cycle response. In the Mars upper thermosphere, O surpasses CO$_2$ now at $\geq$240 km (SMAX) and 215 km (SMIN).

These MTGCM solstice values must be compared with equinox O abundances at LT = 1500 (2-4% at the F1 peak near 130 km) presented earlier in paper 2. Seasonal variations clearly modify the global dynamics significantly, permitting advection to increase in its importance in determining dayside O abundances at solstices. This implies that the advancing solar cycle and seasons act together to keep the O abundances at the midafternoon F1 peak between 1.6 and $\sim$4%, in general accord with observations [see Bouger et al., 1999b, and references therein]. Further detailed model-data comparisons await in situ Neutral Mass Spectrometer (NMS) composition data from Planet-B.

Figures 5a-5d illustrate the behavior of the MTGCM temperatures and horizontal winds near the exobase for various solstice conditions and solar fluxes. Our purpose here is to compare the combined thermal and wind fields over the course of the Martian year and solar cycle. Extremes in these fields are clearly displayed by perihelion-SMAX (SSLMAX) and aphelion-SMIN (NSLMIN) simulations corresponding to Figures 5a and 5d, respectively. Maximum winds (403 m/s) and exobase temperatures (389 K) occur for the SSLMAX case (Figure 5a); minimum winds (216 m/s) and exobase temperatures (205 K) occur for the NSLMIN case (Figure 5d). VL1 exospheric temperatures were observed to be close to $\sim$190 K, in general accord with the NSLMIN simulated temperature (200 K) near LT = 1600 and LAT = 22.5°N. The MGS1 exobase temperature ($\sim$220 K) [Keating et al., 1998b] is somewhat warmer than the dayside (northern midlatitude) values illustrated in Figure 5b (195-210 K) for perihelion-SSMIN (SSLMIN) conditions. For all model simulations (all four plots), winds diverge (strong upwelling) near LT=1500-1600 at the subsolar latitude, and converge (strong subsidence) along the morning terminator (midlatitudes) and near LT = 2200±2 hours (near equator). Each of these convergence zones results in adiabatic warming. Polar night exobase temperatures are coldest ($\sim$101 K) for the NSLMIN simulation and warmest (124 K) for the SSLMAX case. These polar “vortex” zones are isolated dynamically from the rest of the thermosphere, owing in part to a combination of polar night conditions (darkness) and strong (opposing) zonal winds at the vortex boundaries. In fact, temperatures throughout a vertical column (above the homopause) within each vortex zone are nearly isothermal. These zones can be compared to the nightside “cryosphere” of Venus, which is generally isolated dynamically from its dayside (see review by Bouger et al., 1997b). MGS nightside measurements near the end of phase 2 aerobraking are not sufficient to validate or characterize these vortex zones.

All MTGCM simulations presented thus far in this paper assume a dust-free Mars atmosphere. This is a reasonable assumption for northern spring and sum-
mer conditions when the dust opacities are noted to be rather small (J. Pearl, personal communication, 1999). However, the southern spring and summer Mars atmosphere is notably dusty (tau = 0.3 to 1.0), even outside of regional or global dust storm events [Christensen et al., 1998; Keating et al., 1998a]. Lower atmosphere heating and temperatures are highly dependent upon this dust loading. Correspondingly, upper atmosphere heights of constant pressure surfaces are linked to the integrated inflation/contraction of the atmosphere below, driven by this aerosol heating [Bougher et al., 1997a, 1999a]. The height scale for MTGCM simulations is calibrated according to the height of the 1.26-nbar (reference pressure) level. Values of this height are based upon historical information gleaned from Mariner and Viking upper atmosphere measurements [Stewart,
Figure 5. MTGCM horizontal temperature and wind variations for solstice conditions, consisting of four exobase $T^*(U,V)$ plots. SSL refers to southern summer solstice ($L_s = 90$); NSL refers to northern summer solstice ($L_s = 90$); (a) SSL solar maximum case showing horizontal wind pattern and underlying temperature structure (124 to 389 K); maximum arcw corresponds to 403 m/s; (b) SSL solar minimum case showing horizontal winds and temperature structure (104 to 220 K); maximum winds of 257 m/s; (c) NSL solar maximum case showing horizontal winds and temperature structure (111 to 321 K); maximum winds of 326 m/s; and (d) NSL solar minimum case showing horizontal winds and temperature structure (101 to 205 K); maximum winds of 216 m/s.
1987], as well as recent MGS Accelerometer Experiment data [Keating et al., 1998a].

Figure 6 displays Mars reference pressure level heights as a function of Mars season (Ls), solar zenith angle (SZA) (0°–60°), and SMAX (top curve), SMOD (middle curve) and SMIN (bottom curve) fluxes. The simulated MTGCM curves all show a general seasonal trend (aphelion to perihelion) consistent with the observed rise of the height of this reference pressure level by about ∼15 km [Stewart, 1987; Bouger et al., 1997a]. Most of the thermospheric data points follow this general seasonal trend (independent of dust), albeit with a slight height offset of as much as ∼2-4 km. This discrepancy reflects the background dust contribution of the lower atmosphere to the 1.26-nbar (thermospheric) heights. The implication is that the Mars lower atmosphere is subject to “background” aerosol heating throughout the Martian year; the upper atmosphere responds with a corresponding inflation of the height scale. Furthermore, two data points (VL1 and M9N) illustrate significant modulation of the 1.26-nbar height from “nondusty” conditions. VL1 descent data illustrate a strong wave control of the thermospheric profile [Seiff and Kirk, 1977]. Furthermore, the Viking F1 peak height was found to be near 130 km, ∼10 km higher than expected for this season [Hanson et al., 1977; Stewart, 1987]. M9N data were taken during a global dust storm that raged for nearly 4 months and strongly impacted thermospheric densities and temperatures [Stewart et al., 1972, 1992]. Clearly, the best comparisons of MTGCM simulations and spacecraft data must account for the relative height of the 1.26-nbar level corresponding to specific observations and dust opacities. The nondusty simulations of this paper provide a benchmark for such further studies (see section 5.2). Finally, the changing solar fluxes (SMIN to SMAX) appear to have little impact on the height of the 1.26-nbar level. This confirms that the atmosphere below 1.26-nbar is largely driven by solar near infrared (NIR) heating, having no solar cycle dependence [Bouger et al., 1990].

3.2. TIE-GCM Solar Minimum to Maximum

How important is the Earth’s orbital eccentricity in impacting thermospheric temperatures, winds, and composition? How does this orbital forcing modify thermospheric responses due to solar flux variations over the 11-year cycle? Figures 7a and 7b compare differences in exosphere temperatures for TIE-GCM simulations with and without the changing orbital eccentricity. Here, raw temperature differences are calculated by subtracting the “eccentricity neglected” values from the corresponding “eccentricity included” values. December (Figure 7a) and June (Figure 7b) solstices are plotted, both for SMAX fluxes (F10.7 = 200), in order to isolate the orbital effects. Notice that midafternoon low latitude to midlatitude (±40°) temperatures vary by +36 to +41 K for December (warmer for perihelion),
while corresponding midnight temperatures are warmer by +32 to +37 K when the orbital effects are included. Conversely, for June (near aphelion), midafternoon low-latitude temperatures are 35 to 30 K cooler when the proper heliocentric distance is included; midnight values are 31 to 35 K cooler. Overall, SMAX midafternoon temperatures vary by +80 K over the course of the year, owing solely to the Earth’s orbital eccentricity. SMIN midafternoon temperatures (not shown) are calculated to vary by +40 K to +44 K from June (near aphelion) to December (near perihelion) at low latitudes to midlatitudes. These values are compared with those of Mars (Figure 2), for which corresponding SMAX and SMIN temperature variations (aphelion to perihelion) are calculated to be +60 K and +20 K, respectively. Clearly, these orbital temperature variations at exobase altitudes are of similar magnitude for Mars (20-60 K) and Earth (40-80 K). However, the percentage change in temperatures for Mars (+10% to +18% for SMIN to SMAX conditions) is much larger than for Earth (~5%). This is consistent with very small changes in calculated Earth winds (less than 10 m/s) and oxygen densities (less than 5%) due solely to orbital forcing. In short, Earth thermospheric temperatures, winds, and composition appear to change little (in terms of percentage) owing to orbital variations, while Mars structure and winds vary considerably over the Martian orbit [Bougher et al., 1990].

Figures 8a-8d illustrate the behavior of the TIE-GCM temperatures and winds near the exobase for various solstice conditions and solar fluxes. Again, it is important to compare the combined thermal and wind fields over the course of the Earth year and solar cycle. Extremes in these fields driven solely by solar plus seasonal forcing are difficult to identify unless one confines the comparisons to low latitudes to midlatitudes (±40°) where auroral forcing is negligible. As for the Martian thermosphere, solar cycle plus seasonal extremes in temperatures and winds are simulated for SMAX-perihelion (December) and SMIN-aphelion (June) conditions; exobase simulations for these extremes are illustrated in Figures 8a and 8d, respectively. Low-latitude midafternoon (LT ~ 1500) temperatures range from 1527 K (SMAX-December) to 876 K (SMIN-June), a difference of 651 K. The corresponding horizontal winds change by 82 m/s (including the auroral zone). These Earth temperature variations for solar plus seasonal forcing are enormous compared to Mars (see Figures 5a and 5d). Molecular thermal conduction largely balances solar EUV heating for Earth [Bougher et al., 1999b], giving rise to this large thermal response (see further discussion in section 5.1).

Figures 8a (December) and 8c (June) are compared to extract seasonal differences for constant SMAX conditions. The temperature contours are generally flipped between hemispheres for opposite solstices. However, additional differences are apparent upon closer inspection. For example, December midafternoon (LT = 1500) temperatures at low latitudes (~1500 K) are approximately ≤140 K warmer (9%) than corresponding
Figure 7. TIE-GCM horizontal temperature differences at the exobase for solstice conditions. Raw temperature differences are displayed by subtracting the "no-eccentricity case" temperatures from the corresponding "eccentricity case" values: (a) December solstice (SSL) solar maximum conditions (near perihelion): differences approach +36 to +41 K at LT = 1400 and LAT = +40°; and (b) June solstice (NSL) solar maximum conditions (near aphelion): differences approach -35 to -40 K at LT = 1400 and LAT = ±40°.
Figure 8. TIE-GCM horizontal temperature and wind variations for selected conditions, consisting of four exobase $T^+ (U, \phi)$ plots (including eccentricity) for (a) SSL (December) solar maximum case showing horizontal wind pattern and underlying temperature structure (1028 to 1592 K); maximum arrow corresponds to 449 m/s; (b) SSL (December) solar minimum case showing horizontal winds and temperature structure (679 to 982 K); maximum winds of 374 m/s; (c) NSL (June) solar maximum case showing horizontal winds and temperature structure (1023 to 1585 K); maximum winds of 437 m/s; and (d) NSL (June) solar minimum case showing horizontal winds and temperature structure (653 to 980 K); maximum winds of 367 m/s.
4. Isolated Seasonal Variations at Solstices: SMOD

Solar forcing is now fixed at SMOD conditions for MTGCM and TIE-GCM simulations in order to investigate diurnal, latitudinal, and altitude variations for seasonal extremes (solstices). Earth and Mars simulations are examined for asymmetries between winter and summer solstice responses; these differences reflect the seasonal-orbital variations described in section 1.

4.1. MTGCM Solar Medium

Photochemical equilibrium is a good approximation of the dayside Martian ionosphere below ~180 km. Therefore the altitude of the ionospheric (F1) peak is determined by how deeply into the atmosphere the EUV radiation penetrates, and is consequently dependent upon the neutral density structure and solar zenith angle (SZA) [see Zhang et al., 1990]. The Martian thermospheric temperatures and densities vary considerably with solar cycle and season (see Figures 2, 6). Seasonal effects can be isolated by holding the solar fluxes constant. Figures 9a and 9b show MTGCM electron densities as a function of latitude and height (110-160 km) for LT = 1500. Simulations for southern summer solstice (SSLMED) (Figure 9a), near perihelion, reveal an F1 peak at ~130 km with an electron density of 2.16 x 10^5 cm^-3. Corresponding northern summer solstice (NSLMED) (Figure 9b) simulations, near aphelion, show an F1 peak at ~115 km with an electron density of 1.82 x 10^5 cm^-3. This ~15 km variation of the calculated F1 peak heights reflects the seasonal inflation and contraction of the Mars lower atmosphere (due to IR heating), which regulates the thermospheric altitude scale [Stewart, 1987]. Furthermore, the electron density peaks are simulated to rise with altitude into the summer hemisphere, and generally drop in height into the winter hemisphere. This is consistent with inflation of the summer hemisphere atmosphere due to solar heating, and contraction in the winter hemisphere. Aerosol heating modulates this seasonal variation of F1 peak heights considerably [Stewart, 1987; Keating et al.,

<table>
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<th>F10.7</th>
<th>Month</th>
<th>Ls</th>
<th>Texo(15)</th>
<th>Texo(24)</th>
<th>Thomo(15)</th>
</tr>
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<tbody>
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<td>875.9</td>
<td>745.1</td>
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<tr>
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<td>Dec.</td>
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<td>780.6</td>
<td>162.5</td>
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<tr>
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<td>1192.6</td>
<td>930.7</td>
<td>162.8</td>
</tr>
<tr>
<td></td>
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<td>270</td>
<td>1208.5</td>
<td>988.7</td>
<td>162.4</td>
</tr>
<tr>
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<td>1383.3</td>
<td>1165.4</td>
<td>162.9</td>
</tr>
<tr>
<td></td>
<td>Dec.</td>
<td>270</td>
<td>1526.9</td>
<td>1249.5</td>
<td>162.7</td>
</tr>
</tbody>
</table>

| F10.7 refers to the 10.7-cm index used to select reference EUV/UV flux data sets, Ls refers to the angular measure of the Earth's seasons (Ls = 90 is June solstice, Ls = 270 is December solstice); Texo refers to exospheric temperatures; and Thomo refers to homopause temperatures. Values are given for dayside (LT = 1500) and nightside (LT=2400) conditions at the equator.
Figure 9. MTGCM constant local time slices (latitude versus height) for selected cases with solar moderate conditions at LT = 1500: (a) SSL case showing electron densities (log 10 units) over 110 to 160 km; peak value is $2.158 \times 10^5$ cm$^{-3}$ at $\sim$130 km; (b) NSL case showing electron densities (log 10 units) over 110 to 160 km; peak value is $1.820 \times 10^5$ cm$^{-3}$ at $\sim$115 km; (c) SSL case showing temperatures (106 to 304 K) over 110 to 200 km; polar night topside temperatures drop to $\sim$145 K and polar day values rise to 295 K; and (d) NSL case showing temperatures (111 to 261 K) over 110 to 200 km; polar night topside temperatures drop to $\leq$120 K, and polar day values rise to 255 K.
Therefore these nondusty MTGCM simulations provide a benchmark against which to measure peak heights as a function of season. Regular radio occultation observations of this F1 peak height can be used to provide a monitor of dust activity in the Martian lower atmosphere, making use of nondusty MTGCM plots such as those.

Figures 9c and 9d isolate the seasonal responses of Martian thermospheric temperatures as a function of latitude and height (110-200 km) for LT = 1500. Temperature during southern summer solstice (SMLMED) (Figure 9c) and northern summer solstice (NSLMED) (Figure 9d) reflect the difference in Martian heliocentric distance as well as the solar declination. SMLMED exospheric temperatures peak at 304 K in southern hemisphere midlatitudes, while those for NSLMED rise to 261 K at corresponding northern midlatitudes. The overall seasonal difference of peak exospheric temperatures is nearly 43 K, approximately midway between the SMAX (60 K) and SMIN (20 K) seasonal variations illustrated in Figure 2. Polar nighttime exospheric temperatures drop to 145 K (SMLMED) and 120 K (NSLMED), respectively. Mesopause temperatures, defined as the minima just below the thermospheric rise, are found at 110-120 km (SMLMED) and below 110 km (NSLMED), respectively. This difference reflects the changing seasonal altitude scale described in Figure 6. These midafternoon MTGCM slices confirm that the Mars eccentricity drives a significant seasonal variability of the Martian thermospheric temperatures.

The seasonal variation in mass densities is also quite important for the development of aerobraking strategies for Mars spacecraft. Figures 10a (SMLMED) and 10b (NSLMED) illustrate MTGCM densities as a function of latitude and height (110-200 km) for LT = 1500. The general trend shows that densities decline from the summer to the winter hemisphere as one might expect. In particular, densities at robot aerobraking altitudes (110-130 km) drop from the summer to winter pole by factors of 25-35 (SMLMED) and 11-22 (NSLMED), respectively. This larger variation near perihelion (SMLMED) is consistent with stronger solar heating during this season. Densities spanning near-equatorial latitudes (corresponding to low SZA) are rather flat below 140 km, in accord with MGS observations [Booher and Keating, 1999; Keating et al., 1999a,b]. However, it is noteworthy that large extratropical latitude density gradients are presently simulated for "nondusty" conditions. Proper account of the dusty lower atmosphere, especially during southern spring and summer, will yield a modification of the global wind structure so that polar night densities (and temperatures) are enhanced. The net effect will be smaller pole-to-pole density gradients (see section 5.2). Observed dayside (LT~1500) densities during phase 2 of MGS aerobraking revealed a 60°N to 90°S gradient of a factor of 7.5 at 130 km [Booher and Keating, 1999; Keating et al., 1999a,b]. This is a factor of
Figure 10. MTGCM constant local time slices (latitude versus height) for solstices, showing total mass densities for solar moderate conditions at LT = 1500: (a) SSL case showing densities (log 10 units) over 110 to 200 km; polar night 130 km values drop to ~0.32 kg/km³, and corresponding polar day values rise to 10.0 kg/km³; and (b) NSL case showing densities (log 10 units) over 110 to 200 km; polar night 130 km values drop to ~0.056 kg/km³, and corresponding polar day values rise to 1.26 kg/km³. These latitude variations are extreme since dust opacity effects are not yet included.
3 smaller than calculated by MTGCM simulations for northern summer solstice (SMOD) conditions [Bougher and Keating, 1999]. Clearly, high-latitude winds (thus far poorly understood) play a strong role in the maintenance of Martian thermospheric polar night densities and temperatures.

4.2. TIE-GCM Solar Medium

In general, ambipolar diffusion and advection control the location of the terrestrial F2 ionospheric peak heights, with the primary ion being O⁺. Electric fields can also be important; that is, at low latitudes they produce the Appleton anomaly, ionization peaks at tropical latitudes which develop during the daytime as a result of upward $E \times B$ drifts. These ionization peaks are affected by transsequatorial winds, with the upward wind peak increasing in density (summer hemisphere), and downwind peak decreasing in density (winter hemisphere) relative to no-wind or latitudinally symmetric (equinox) wind conditions. The effect of seasonal changes in transsequatorial winds is revealed in the calculated solstice variations in the tropical electron peaks for constant SMOD conditions (see Figures 11a and 11b). In both plots the highest peaks (with the largest electron densities) occur in the summer hemisphere, that is, 425 km (December at 20° S latitude) and 375-400 km (June at 20° N latitude).

Finally, midafternoon (LT = 1500) thermal cross sections reveal seasonal changes in temperature structure for SMOD conditions. Figures 11c and 11d show that low latitude to midlatitude exospheric temperatures are roughly 80-100 K warmer for December (perihelion) than June (aphelion). This response reflects the solar heating plus thermal conduction control of the dayside thermal structure in conjunction with stronger heating (yielding warmer temperatures) near perihelion. A weaker yet similar response is felt in the equatorial nightside (not shown), for which December temperatures are ~50-60 K warmer due to transport of heat from the dayside (see also Figures 8a-8d). This nightside warming may also be due in part to dynamical effects driven by semidiurnal tides, especially during solar minimum conditions when tidal dissipation is reduced [e.g., Fesen et al., 1986].

5. Seasonal Variations Affecting Thermospheric Processes

5.1. Changing Importance of CO$_2$O Cooling and Dynamics as Thermostats

Paper 2 examined the relative importance of CO$_2$O cooling and dynamics as thermostats for the upper atmospheres of Venus, Earth, and Mars. For both Earth and Mars, it was shown that CO$_2$ cooling is rather weak in its regulation of thermospheric temperatures responding to solar cycle variations in EUV/UV fluxes [Bougher et al., 1999c]. Specifically, Earth CO$_2$ cooling is important only below 130 km, well removed from the peak heating at higher altitudes. This heat is therefore conducted down the vertical temperature gradient toward this IR cooling layer, rendering molecular thermal conduction the dominant cooling mechanism above 150-180 km. This “weak thermostat” yields a large exospheric temperature variation over the solar cycle. The orbital modulation to this solar cycle thermal response (up to 80 K on the dayside) further confirms the strong influence of molecular thermal conduction on the heat budget. The SMAX-December and SMIN-June TIE-GCM temperature extremes are enormous (up to 650 K) (Figures 8a and 8d) owing to the primary balance of variable solar heating and molecular thermal conduction.

On the other hand, Mars solar cycle plus seasonal temperature extremes (Figures 5a and 5d) are much smaller than for Earth (up to 180 K). The smaller Martian EUV/UV heating efficiencies (see paper 2) certainly contribute to this reduced thermal response. However, Mars winds play an increasingly important role in regulating temperatures as the solar cycle advances. MTGCM solstice calculations show that exobase horizontal winds double from SMIN-aphelion to SMAX-perihelion conditions, corresponding vertical (dayside peak upwelling) winds also double. Hence adiabatic cooling (due to upwelling) is predicted to be strong enough to compete with molecular thermal conduction for control of Mars dayside thermosphere temperatures, especially during SMAX-perihelion periods (Figure 5a). Without this dynamical thermostat, Mars exobase temperatures on the dayside would be considerably warmer during SMAX-perihelion conditions. In short, Mars upper atmosphere temperatures are partially determined by a balance of solar cycle plus seasonal heating and molecular thermal conduction. However, global-scale winds provide an important thermostatic control that intensifies with the advance of the solar cycle. Model predictions of the structure of the Martian upper atmosphere must properly account for dynamical forcing mechanisms (EUV heating, planetary and gravity waves, dust storm atmospheric inflation) in order to be realistic (see section 5.2).

5.2. Impact of Dust on the Mars Thermosphere: $L_s = 270^\circ$

Recent Mars Global Surveyor aerobraking data (see section 2.1) confirm a significant coupling of the Mars lower and upper atmospheres that is intensified during “dust events” [Keating et al., 1998a; Bougher et al., 1999a]. These events may include local or regional storms, global encircling storms, or planet-wide dust storms. Dust particles do not reach thermospheric heights. Nevertheless, the remote effects of a dusty Mars lower atmosphere are experienced aloft [see Keating et al., 1998a; Bougher et al., 1999a,b]. The dust-free MTGCM simulations of this paper are meant to
Figure 11. TIE-GCM constant local time slices (latitude versus height) for solstices and solar moderate conditions (SMOD) at LT = 1500 (including eccentricity): (a) SSL (December) case showing electron densities (log 10 units) over 100 to 500 km; peak value is $1.27 \times 10^8$ cm$^{-3}$ at $\sim 425$ km ($\sim 20^\circ$ S latitude); (b) NSL (June) case showing electron densities (log 10 units) over 100 to 500 km; peak value is $7.04 \times 10^6$ cm$^{-3}$ at $\sim 375$-400 km ($\sim 20^\circ$ N latitude); (c) SSL (December) case showing temperatures (150 to 1248 K) over 100 to 500 km; polar night topside temperatures drop to $\sim 875$ K, polar day values rise to 1248 K, and equatorial topside temperatures rise to 1200 K; and (d) NSL (June) case showing temperatures (149 to 1235 K) over 100 to 500 km; polar night topside temperatures drop to $\leq 825$ K, polar day values rise to 1235 K, and equatorial topside temperatures rise to 1100 K.
provide a baseline against which actual Mars thermospheric data can be compared. Dust-free or low-dust conditions are not expected to prevail near perihelion during Mars southern spring-summer. In this section we compare results from a coupled MGCM-MTGCM simulation including dust effects with a nondusty case to illustrate thermospheric impacts that might be expected for Mars near perihelion.

Two southern summer solstice cases (Ls = 270) for SMIN conditions (SSLMIN) are selected for study. This period corresponds to Mars in early February 1998 during MGS phase 1 aerobraking (~P118). One case is run for no-dust conditions; another is run for a visible dust opacity of 0.3. In practice, this means that zonally averaged heights and temperatures are prescribed at the lower boundary of the MTGCM differently for these two cases, in accord with corresponding MGCM values at 1.32 µbars. In addition, semidiurnal tidal amplitudes and phases are prescribed independently for the two cases in conjunction with the changing forcing from the lower atmosphere. With the addition of dust, integrated heating of the lower atmosphere generates an inflation of the entire atmospheric column. As a result, thermospheric densities at a constant height are enhanced above nondusty values. Also, upward propagating semidiurnal tides captured by the MTGCM serve to modify the global thermospheric circulation, thereby impacting the thermospheric density and thermal structure [Hougher et al., 1997a, 1999a,b].

Temperature structure is significantly different for the two MTGCM cases. Figure 12a compares temperature profiles at LT = 1300 and LAT = 47.5°N. This corresponds to the location of the MGS periapsis passage for roughly orbit P118 (February 6, 1998). It is clear that the dusty case generates an exospheric temperature that is ~30 K warmer than the nondusty simulation. The warm exospheric temperature of 215 K compares favorably with the MGS Accelerometer mean value for this season at this location [Keating et al., 1998b; Bougher and Keating, 1999]. Corresponding densities (temperatures) at 130 km are 15-30% (10-13 K) higher (warmer) for the dusty case (47.5°-80° N). This suggests that northern (winter) polar latitudes are warmed and the atmospheric column is inflated dynamically as a result of the changing circulation. In fact, maximum horizontal winds have increased from 257 m/s (see Figure 5b) to 330 m/s. Figure 12b illustrates the thermal balance at LT = 1300 and LAT = 47.5° N for the dusty case. The primary term that has changed is that due to horizontal advection (175 to 300 K/d above 150 km). The implication is that the dusty lower atmosphere drives tidal waves (semidiurnal tides captured here) that modify the thermospheric circulation, yielding warmer winter polar temperatures and enhanced densities at a constant height. The overall impact is to weaken the latitude gradient of densities from the equator toward the winter pole by approximately 40%. Clearly, the best comparison of MTGCM simulations
and observed Mars thermospheric densities and temperatures must take into account these dust effects upon the Mars thermosphere.

6. Summary and Ongoing Work

This current work presents many TGCM simulations for Earth and Mars designed to investigate the seasonal plus solar cycle responses of their thermospheres. Significant asymmetries in densities, temperatures, and winds are clearly visible throughout the Martian year. By contrast, the seasonal asymmetries for the Earth’s thermospheric structure and winds are rather subdued. Also, this paper addresses the role of the dusted Mars lower atmosphere for impacting thermospheric densities, temperatures, and winds for perihelion conditions. Atmospheric inflation/contraction and wave effects significantly modify the Mars thermospheric structure and dynamics during dust events.

This paper is the last in a three-part series examining the terrestrial planet responses to solar cycle forc-
ing for global mean, equinox (Venus, Earth, and Mars) and solstice (Earth and Mars) conditions [Bougher and Roble, 1991; Bougher et al., 1999b; this work]. Many of these Venus, Earth, and Mars TGCM simulations are archived on a Web site at the University of Arizona: http://www.lpl.arizona.edu/~sengel/thermo.html.

As a further validation of the thermospheric processes incorporated in these models, it is important to investigate their time-dependent responses to historical solar storms for which EUV fluxes varied significantly. In addition, a more complete picture of the Mars thermospheric responses to regional and global dust storms is needed to provide an improved upper atmosphere climatology for future Mars aerobraking operations.

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References


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