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Solar Tides in the Middle and Upper Atmosphere of Mars

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Key Points:

- Multi-year latitude vs. Ls climatology of Mars tidal spectrum is derived from MRO/MCS measurements at 76 km.
- MCD tidal spectrum compares well with MCS climatology, and with longitude structures in MGS aerobraking densities near 112 km.
- MCD latitude-height tidal structures are strongly influenced by mean winds and dissipation, and provide new insights into density variability at 100-170 km altitude.

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Abstract

Solar tides are responsible for much of the spatial-temporal variability of Mars' upper atmosphere (100 - \sim 200 km). However, the tidal spectrum, its latitude versus Ls variability, and its vertical evolution remain uncertain. In this paper, Mars Climate Sounder (MCS) temperature measurements at 76 km above Mars' areoid are used to construct a multi-year latitude versus Ls climatology of the tidal spectrum. The most important spectral components include the solar-synchronous ("migrating") components DW1, SW2, and the solar-asynchronous ("non-migrating") tides DE3, DE2, DE1, SE1, S0, and SW1. The Mars Climate Database (MCD), which provides predictions from the Laboratoire de Météorologie Dynamique (LMD) Global Climate Model, captures particularly well the amplitudes and key structural features of the solar-asynchronous tides at 76 km that furthermore underly the large longitudinal structures in density that are observed between 100-200 km. Height-latitude and latitude-Ls structures of MCD density perturbations are therefore examined between 76-172 km and interpreted in terms of mean wind and dissipation effects. In particular, due to the smaller radius and more intense zonal-mean zonal winds at Mars compared to Earth, Doppler-shift effects are significantly exaggerated compared to Earth. Evidence is also provided for non-negligible contributions to density variability from stationary planetary waves which arise from tide-tide non-linear interactions. It is moreover shown that MCD captures the salient amplitude and phase characteristics of the $\sim \pm 30$ -60% longitudinal density perturbations measured by the Mars Global Surveyor (MGS) accelerometer. This, and the excellent MCD-MCS agreement at 76 km, lends credibility to the ability of MCD to provide new insights into thermosphere density variability at Mars due to vertical coupling by solar tides.

1 Introduction

The importance of solar-driven tides to the dynamics of Mars' atmosphere is now common knowledge. Due to their vertical propagation characteristics and growth with height, they serve the important role of transmitting the variability associated with their lower-atmosphere sources and other meteorological processes to much higher altitudes (e.g., Angelats i Coll et al., 2004; Forbes et al., 2002; Moulden and Forbes, 2008a; England et al., 2016, 2019), even to the edge of space (i.e., the exobase, roughly near 200 km). Additionally, their dissipation above about 80 km altitude deposits net momentum, thus contributing as a driver for the mean circulation in this altitude region (e.g., Moulden

and Forbes, 2008b), and in turn redistributing minor chemical constituents including water (Shaposhnikov et al., 2019).

The sources for tidal heating include infrared radiative absorption and transfer by CO₂ near the surface, radiation absorption by dust and water ice clouds, and near-infrared radiation absorption by CO₂ above 70 km. These processes are strongly influenced by topography, thermal properties of the surface, and by the dynamical processes responsible for redistributing the absorbing species. Consequently, both the solar heating and the atmospheric responses are dependent on latitude, longitude, local solar time (LST) and day of year. Throughout this paper, LST refers to local true solar time.

Planetary rotation furthermore admits periodic dependencies on time and longitude, so that solar tides are expressed mathematically as

$$\sum_s \sum_n A_{n,s}(z, \theta) [\cos(n\Omega t + s\lambda - \phi_{n,s}(z, \theta))] \quad (1)$$

where t = universal time (UT), $\Omega = 2\pi\text{sol}^{-1}$, z = altitude, θ = latitude, integer s is the zonal wavenumber, integer n defines the frequency or period of the oscillation, $A_{n,s}$ is the amplitude, and $\phi_{n,s}$ is the phase (i.e., longitude λ of maximum at 0000 UT, or time of maximum at $\lambda = 0$). The outer sum captures the longitude dependence of any tidal frequency (either source or response), while the inner sum captures the diurnal variation of the source (or response) at any given longitude. Setting the quantity in parentheses equal to zero and taking the derivative defines the zonal phase speed of any tidal component: $C_{ph} = \frac{-n\Omega}{s}$. For the assumed mathematical expression (1), this implies $s < 0$ for eastward-propagating waves ($C_{ph} > 0$) and $s > 0$ for westward-propagating ($C_{ph} < 0$) waves. Tides with $s = 0$ do not propagate zonally and are called zonally-symmetric; this means that the atmosphere oscillates at a given tidal frequency in unison at all longitudes. Note also that if $s = n$, then a tide with any frequency migrates westward with the apparent phase speed of the Sun to a ground-based observer; these are solar-synchronous and often referred to as “migrating” tides. Tides with $s \neq n$ are solar asynchronous and often referred to as “non-migrating” tides, and it is these tides that capture the longitude dependence of the tidal response (or source).

The notation DWs or DEs is used to denote a westward or eastward-propagating diurnal tide, respectively, with zonal wavenumber = s . For semidiurnal oscillations, S replaces D. The zonally-symmetric oscillations are denoted D0, S0. In the present pa-

per we will confine ourselves to the diurnal ($n = 1$) and semidiurnal ($n = 2$) solar tides, which generally exceed higher-order tides in magnitude and importance.

Much of what we know, observationally, about tides in the middle atmosphere (~ 50 - 100 km) of Mars has been learned from temperature measurements by the Mars Climate Sounder (MCS) instrument (McCleese et al., 2007) on Mars Reconnaissance Orbiter (MRO), which has been observing the atmosphere and surface of Mars since September 2006. (However, see also observations of Mars solar tides between 70-110 km by the SPICAM instrument on Mars Express reported in Withers et al., 2011). MCS temperature measurements extend from near the surface to about 80 km altitude between $\pm 87^\circ$ latitude. In its original in-track observing mode, one of the difficulties attached to the MCS measurements, insofar as tides are concerned, is the limited LST coverage offered by the $\sim 0300/1500$ LST solar-synchronous orbit of MRO. Expressing (1) in terms of local time (t_{LST}),

$$\sum_s \sum_n A_{n,s}(z, \theta) [\cos(n\Omega t_{LST} + (s-n)\lambda - \phi_{n,s}(z, \theta))] \quad (2)$$

Expression (2) demonstrates that the signatures of tides in solar-synchronous ($t_{LST} \approx \text{constant}$) satellite data are identifiable in terms their longitude structure, specifically their “space-based wavenumber” $k_s = |s-n|$. However, there is ambiguity inherent in these signatures, since multiple combinations of s and n yield the same k_s . The first tidal analyses of MCS temperatures using the in-track observing mode (Lee et al., 2009; Guzewich et al., 2012) exploited the fact that the ascending and descending parts of the orbits are 12 hours apart (equatorward of $\pm 75^\circ$ latitude). Since tides with n odd are in anti-phase at local times 12 hours apart, the difference between ascending and descending observations is expected to contain only odd tides (diurnal, terdiurnal, etc.) and the sum is expected to contain only even ones (semidiurnal, quatradiurnal, etc., as well as the $n = 0$ zonal mean and stationary planetary waves (SPW)). Assuming that higher-order tides are negligible relative to diurnal and semidiurnal tides, spectral analysis can then be applied to the resulting sum and difference fields to reveal the amplitudes of the diurnal or semidiurnal tides contributing to each solar-synchronous zonal wave number, k_s . For a given n and $k_s = |s-n|$, there are generally two mathematical solutions, and a more physically reasonable can often be identified. These authors presented evidence for DE3, DE2, DE1, D0, DW1, and SE1 tides, as well as SPW with $s = 1$ and $s = 2$ (hereafter SPW1, SPW2, respectively).

110 Starting September, 2010, a special cross-track observing mode was also initiated
111 which expanded the local time coverage (Kleinböhl et al., 2013). Kleinböhl et al. (2013)
112 and Wu et al. (2015) exploited this capability, and were able to derive both diurnal and
113 semidiurnal tides by directly fitting the MCS temperature measurements with respect
114 to LST and longitude. Kleinböhl et al. (2013) provided the first delineation of the SW2
115 semidiurnal tide in Mars' middle atmosphere, and through comparisons with a Mars GCM,
116 concluded that SW2 is a dominant feature throughout the Martian year, and that ra-
117 diatively active water ice clouds (Wilson et al., 2007; 2008; see also Wilson, 2014) pro-
118 vide the most plausible explanation for the observed amplitudes. Wu et al. (2015) pro-
119 vided a new and expanded view of middle atmosphere tides, focusing on vertical tidal
120 structures in the middle atmosphere at discrete latitudes and solar longitudes (Ls). Their
121 study also included DW2, DW3 and SW1 in addition the aforementioned tidal compo-
122 nents. These latter tides were hypothesized to originate from nonlinear interactions be-
123 tween solar-synchronous ($s = n$) tides and SPWs.

124 At higher altitudes (100-200 km), information on tides has been inferred indirectly
125 from longitude structures revealed in density measurements made by accelerometers on
126 the Mars Global Surveyor (MGS), Mars Odyssey (MO), and Mars Reconnaissance Or-
127 biter (MRO) satellites (e.g., Wang et al., 2006; Wilson, 2002; Withers, 2006; Withers et
128 al., 2003), or more recently, in neutral mass spectrometer (e.g., England et al., 2016; Liu
129 et al., 2017) and ultraviolet imager (e.g., England et al., 2016; Lo et al., 2015; Gröller
130 et al., 2018) measurements on the MAVEN (Mars Atmosphere and Volatile Evolution)
131 and Mars Express (Withers et al., 2011) missions. These studies have been based on ei-
132 ther ascending or descending orbital data alone, and thus did not involve any constraints
133 insofar as LST is concerned, in contrast to the aforementioned MCS tidal studies. In a
134 few isolated cases where LST changed sufficiently fast over restricted ranges of latitude
135 and Ls, more definitive tidal inferences have been made (e.g., England et al., 2019; Forbes
136 et al., 2004; Forbes and Zhang, 2018).

137 Modeling frameworks for interpretation of the aforementioned data in terms of tidal
138 coupling between the middle (~ 50 -100 km) and upper (~ 100 -200 km) atmosphere of Mars
139 have been relatively few and incomplete. Forbes et al. (2002) used a linear model to ex-
140 tend several solar-asynchronous tidal fields at 70 km from the Ames General Circula-
141 tion Model to 200 km altitude for $L_s = 30^\circ$ and $L_s = 270^\circ$. Angelats i Coll et al. (2004)
142 and Moudeden and Forbes (2008a) used GCMs to study the full spectrum of solar-asynchronous

143 tides up to 120 km and 160 km, respectively, for $L_s \approx 75^\circ$, and both obtained some rea-
144 sonable agreement with longitude structures of density from MGS accelerometer mea-
145 surements. While Mars GCMs now exist that extend from the surface to Mars exosphere
146 at ~ 200 km (e.g., Bougher et al., 2011, 2015; Gonzalez-Galindo et al., 2009, 2013, 2015),
147 they have not yet provided a more complete characterization and interpretation of ver-
148 tical tidal coupling between Mars' middle and upper atmospheres.

149 By employing MCS temperature observations at 76 km altitude coupled with a gen-
150 eral circulation model of Mars' atmosphere extending to ca. 200 km, the objective of the
151 present study is to significantly advance our understanding of how the tidal spectrum
152 serves to couple the middle and upper atmospheres of Mars. Towards this end, we pro-
153 vide a climatological representation of the MCS tidal spectrum at 76 km above Mars'
154 areoid from nearly pole to pole and as a continuous function of L_s . The veracity of the
155 Mars Climate Database (MCD) derived from the LMD Global Climate Model (Forget
156 et al., 1999; Millour et al., 2018; Gonzalez-Galindo et al., 2015, and references therein)
157 is demonstrated against this tidal climatology, and the MCD is then applied to clarify
158 how this tidal spectrum evolves with height. A specific goal is to shed light on the tidal
159 origins of longitude variability in atmospheric total mass densities as observed by satellite-
160 borne accelerometers during aerobraking operations (e.g., MGS, MRO, MO), as well as
161 measurements from the MAVEN mission at higher altitudes (~ 135 -200 km).

162 The following section describes the MCS data that are utilized, and how they are
163 processed to extract diurnal and semidiurnal tides. Section 3 provides results, which in-
164 clude an MCD-MCS comparison of tidal spectra as a function of latitude and L_s at 76
165 km altitude, examination and assessment of MCD vertical tidal structures, and MCD
166 predictions of total mass density variations at 108 and 172 km. A summary of results
167 and conclusions comprises Section 4.

168 **2 MCS Data and Processing, and MCD Model Description**

169 **2.1 Data employed**

170 The data used in this study consist of temperature profiles (version 5; Kleinböhl
171 et al., 2009, 2011, 2017) from the MCS instrument on board MRO, and were obtained
172 from the Planetary Atmospheres Node of the Planetary Data System (PDS) ([http://pds-
173 atmospheres.nmsu.edu/](http://pds-atmospheres.nmsu.edu/)). MCS is a limb- and on-planet scanning infrared radiometer

174 that launched aboard MRO and became fully operational at the end of September 2006
175 (McCleese et al., 2007). Temperature retrievals use the CO₂ absorption feature at 15 mi-
176 cron and typically provide profile information from the near-surface up to ~80 km al-
177 titude with a vertical resolution of ~5 km. MROs orbit is nearly polar and solar-synchronous
178 with an inclination of 92.66° and orbital period of 112 minutes. At any given time the
179 spacecrafts local solar time (LST) is near 1500 LST or 0300 LST during the ascending
180 or descending parts of the orbit, respectively, except poleward of 75° latitude where the
181 spacecraft shifts from 1500 LST or 0300 LST and vice versa in the opposite polar hemi-
182 sphere. Note that the actual local true solar time can deviate from the nominal local mean
183 solar time by up to 0.7 Mars hours over the course of the Mars year. In addition, the
184 local mean solar time varied in ranges between 0.2 and 0.5 Mars hours over the dura-
185 tion of the MRO mission.

186 As noted previously, prior to September 2010 the nominal MCS data-taking mode
187 was in-track, which means that at latitudes equatorward of about ±75° measurements
188 are only available at 1500 and 0300 LST. However, the cross-track observing mode that
189 was initiated in September 2010 (Kleinböhl et al., 2013) provides measurements at ad-
190 dditional LSTs, and data collection is still ongoing. At low latitudes cross-track measure-
191 ments are offset by ~ ±1.5-2.0 hours in LST from in-track. This local time difference
192 increases to over 3 hours towards high latitudes, such that analyses that rely on this lo-
193 cal time difference, e.g. the extraction of semidiurnal tide parameters, are most robust
194 at high latitudes. As shown by Kleinböhl et al. (2013), the LST spacing afforded by this
195 observational mode is adequate to extract the semidiurnal tide, which these authors did
196 for the solar asynchronous component. Wu et al. (2015) additionally extracted diurnal
197 and semidiurnal solar-asynchronous tides from MCS temperature measurements made
198 in this operational mode. Since 12-14 longitudes are sampled in a given day, diurnal and
199 semidiurnal tides with zonal wavenumbers as large as 6 (both eastward and westward)
200 are in principle capable of being extracted from these data.

201 For the present study, tidal specifications are sought at 76 km since the expecta-
202 tion is that all tidal components capable of growing exponentially to 100 km and beyond
203 would have their maximum amplitudes in the MCS data set at this altitude. (Although
204 MCS data often extend beyond 80 km, utilizing data at this altitude would introduce
205 additional gaps.) All available MCS temperature data at 76 km between Sept 2010 - Jul
206 2019 are utilized except for the global dust storm period of Jun - Aug 2018, since tides

207 are anomalous during global dust storms, and not part of normal climatology. Other dusty
208 periods are included since they are arguably part of climatology.

209 2.2 Tidal fitting

210 In our experimentation with these data, we found that aliasing between diurnal and
211 semidiurnal components could occur. The robustness of tidal fits is likely affected by vari-
212 ability associated with uncertainties in the temperature measurements at 76 km. The
213 following steps were therefore taken to ameliorate these issues. To improve statistics, all
214 available MCS data between Sept 2010 - Jul 2019 are binned and averaged with respect
215 to latitude, longitude, LST and Ls. Multi-year averaging is a reasonable approach since
216 prior studies have noted the repeatability of salient structural features in MCS tides from
217 year to year (Moudden and Forbes, 2015; Wu et al., 2015). Similar multi-year averag-
218 ing to delineate tides has been successfully applied at Earth. Also, we found that fitting
219 across zonal wavenumbers $(s-n) = -6$ to $+6$ to diurnal and semidiurnal components
220 separately, instead of together, eliminated aliasing between the two while yielding the
221 nearly the same diurnal-mean temperatures.

222 Typical portrayals of data distributions and fitting are provided in Figure 1. The
223 distribution of data points in latitude-LST space is shown in Figure 1(a), which is typ-
224 ical of other Ls periods. The distribution of points poleward of $\pm 60^\circ$ latitude is rather
225 good. However, local time coverage equatorward of $\pm 40^\circ$ is potentially of concern for ex-
226 tracting semidiurnal tidal components, particularly SW2 which can alias with the zonal-
227 and diurnal-mean temperature; at a fixed local time, both SW2 and the zonal mean tem-
228 perature are longitude-independent (cf. Equation (2)). In addition, potential systematic
229 uncertainties between in-track and cross-track observations can introduce errors in the
230 amplitude of SW2 as well as the zonal mean temperature when tidal modes are fit over
231 a large local time gap. The local mean solar time of MCS observations varied in ranges
232 between 0.2 and 0.5 Mars hours over the duration of the MRO mission; also the actual
233 local true solar time can deviate from the nominal local mean solar time by up to 0.7
234 Mars hours over the course of the Mars year (see Figure S1 in the Supporting Informa-
235 tion File). Hence using multi-year averages of the data reduces random errors, assists
236 with local time coverage at low latitudes, and also allows to have some data bins being
237 not exclusively populated by in-track or cross-track measurements, which somewhat mit-
238 igates possible systematic uncertainties between these observation types. To deal with

239 the disparity between the numbers of data points in some bins versus others, we weighted
240 the bin-averaged data points according to the cube root of the number of data points.

241 In practice this results in little overall difference with results in which all bin-averaged
242 data points were weighted equally. This insensitivity serves as a secondary validation of
243 the robustness of the tidal determinations.

244 As shown in Section 4, there is often a remarkable correspondence between MCD
245 and MCS seasonal-latitudinal structures and amplitudes that serves as a kind of mutual
246 validation of the MCS tidal analysis on one hand and the MCD model results on the other.

247 As demonstrated there, for a given space-based zonal wavenumber, the corresponding
248 diurnal and semidiurnal tides have markedly different latitude structures, which is fur-
249 ther evidence that aliasing between them does not exist to a measurable degree. How-
250 ever, for the reasons noted above, the extraction of the solar synchronous semidiurnal
251 tide (SW2) is still problematic as it is potentially aliased with the zonal mean. There-
252 fore, the seasonal-latitudinal variability of SW2 derived from MCS and depicted in this
253 paper must be viewed with this caution in mind.

254 Figures 1(b)-1(f) show some typical examples of the fits to MCS temperatures (resid-
255 uals from zonal mean) with respect to LST at a few latitudes, Mars longitudes and Ls.
256 Each panel illustrates the fitted data points with their attendant standard deviations,
257 the semidiurnal fits to these data points subject to the above weighting, the diurnal fits,
258 and the combined diurnal and semidiurnal fits. The fits enable latitude versus Ls depic-
259 tions of amplitudes and phases of each tidal component to be constructed, which are per-
260 formed within windows of length 30° Ls moved forward in increments of 15° Ls. The largest
261 amplitude tides consist of DW1, SW2, DE3, DE2, DE1, SE1, S0, and SW1. In Section
262 4, the latitude versus Ls depiction of each of these tides at 76 km is compared with that
263 from the MCD, and the MCD height-latitude structure of each tide and its vertical cou-
264 pling characteristics are addressed. First, however, a brief description of the MCD is pro-
265 vided in the next subsection.

266 **2.3 MCD model description**

267 The Mars Climate Database (MCD) provides meteorological fields that are derived
268 from a Mars general circulation model (MGCM) developed at the Laboratoire de Météorologie
269 Dynamique (<http://www-mars.lmd.jussieu.fr>). The LMD-MGCM is rooted in the early

270 work of Forget et al. (1999), but since then has adopted more sophisticated and real-
271 istic treatments of dust and water cycles, photochemistry, radiative transfer, and solar
272 radiative fluxes. Particularly relevant to the current work, the modern version extends
273 into the thermosphere and ionosphere Gonzalo-Galindo et al. (2009, 2013, 2015). In ad-
274 dition to simulations pertaining to individual Martian Years (MY 24-33), the MCD cur-
275 rently offers users eight baseline database scenarios related to dust and solar EUV con-
276 ditions. In this study, we use the climatology scenario from MCD Version 5.3 (Millour
277 et al., 2018), in which dust distributions are based on those observed in Mars years when
278 no global dust storms occurred, and average solar EUV conditions are assumed. The MCD
279 outputs are provided in increments of 30° Ls, which are interpolated to 15° to be con-
280 sistent with the fitting window of the MCS observations.

281 2.4 Brief review of tidal theory

282 Following the terrestrial example set by Truskowski et al. (2014), in this paper we
283 will find it instructive to employ some aspects of classical tidal theory to interpret the
284 height-latitude structures and propagation characteristics of tidal components in Mars'
285 atmosphere. Classical tidal theory refers to solutions of the momentum, thermal energy,
286 continuity and state equations of the atmosphere that are linearized relative to a horizontally-
287 stratified basic state without dissipation (e.g., Chapman and Lindzen, 1970). The early
288 development of tidal theory with specific application to Mars is attributable to Zurek
289 (1976). Assuming dependent variables such as horizontal and vertical winds, tempera-
290 ture, density and pressure or geopotential to be periodic in t and λ as in (1) eliminates
291 derivatives with respect to t and λ . This permits consolidation of the problem into a sin-
292 gle second-order partial differential equation for the perturbation geopotential, and other
293 variables to be expressed in terms of the geopotential. This equation is separable in z
294 and θ , leading to an eigenfunction-eigenvalue problem where the θ dependence is embod-
295 ied in Laplace's tidal equation and the height dependence is embodied in a "vertical struc-
296 ture equation" which also contains specification of the thermal source.

297 For a given tidal period and zonal wavenumber, the latitude structure of the source
298 (or its response) is captured by an expansion of orthogonal Hough functions ("tidal modes")
299 which are the eigenfunction solutions to Laplace's tidal equation. Each Hough function
300 has attached to it an eigenvalue called an "equivalent depth", which appears in the ver-
301 tical structure equation for that Hough function. The vertical structure equation, with

302 the same form as a 1-dimensional time-independent Shrodinger equation, can be solved
 303 analytically in the case of an isothermal background state, or numerically in the case of
 304 a more general vertical distribution of temperature. In this context an approximate ver-
 305 tical wavelength can be derived as follows:

$$\lambda_z \approx \frac{2\pi H_0}{\sqrt{\frac{H_0}{h_n}(\kappa + \frac{dH}{dz}) - \frac{1}{4}}} \quad (3)$$

306 where $\kappa = \frac{R}{c_p} = .223$ at Mars, R is the gas constant and c_p is the specific heat at con-
 307 stant pressure, $H_0 \approx 8.5$ km is the mean scale height and $\frac{dH}{dz} \approx -0.046$ below 100 km
 308 at Mars. These values are consistent with those computed for $L_s = 180^\circ$ zonal-mean val-
 309 ues from MCD v5.3, and those reflected in Viking 1, Viking 2, Curiosity and Opportu-
 310 nity Lander profiles. Given the approximate nature of classical tidal theory in the con-
 311 text of interpreting a GCM, nothing more sophisticated than this approach is consid-
 312 ered useful. Thus, a single approximate vertical wavelength can be assigned to each tidal
 313 mode at Mars, applicable below 100 km. Note also that for $h_n > 6.0$, the quantity un-
 314 der the square root is negative, yielding an imaginary λ_z , thus implying a non-propagating
 315 (evanescent) solution. This occurs for very large vertical wavelengths. Classical tidal the-
 316 ory also admits negative equivalent depths for diurnal tides (Chapman and Lindzen, 1970;
 317 Lindzen, 1966; Zurek, 1976) which also correspond to evanescent solutions, and which
 318 capture diurnal responses at mainly middle-high latitudes. See further discussion on this
 319 point in subsection 3.1.

320 The first two to four Hough functions for each of the tidal components considered
 321 in this study can be found in Figure S2 of the Supporting Information file, and some are
 322 illustrated explicitly within several of the figures to follow. The vertical wavelengths cor-
 323 responding to these tidal modes according to (3) are tabulated in Table 1, which are only
 324 intended to provide rough estimates. In addition to $h_n > 6.0$, Hough modes with $h_n <$
 325 0.44 ($\lambda_z < 30$ km) are indicated by symbols in Table 1. The former do not propagate
 326 vertically while it was determined by viewing MCD tidal structures that the latter do
 327 not effectively propagate above about 100 km due to their susceptibility to dissipation.
 328 The remaining tidal modes in Table 1 are those that most effectively couple the lower
 329 and upper atmospheres of Mars, although the absolute effectiveness of any given mode
 330 also depends on the magnitude of its forcing. In the following, we will use classical tidal
 331 theory as a starting point for interpretation of the MCD height-latitude structures, and

332 then considering dissipation and Doppler-shifting effects by the mean wind field as needed
 333 to provide qualitative but more realistic interpretations.

334 In addition to neglecting mean wind effects and dissipation, classical tidal theory
 335 also neglects the nonlinear terms in the momentum and thermal energy equations. As
 336 shown for Earth (Truskowski et al., 2014) and Mars (Moudden and Forbes, 2008a), tide-
 337 tide and tide-SPW interactions can yield tides with different periods and/or zonal wavenum-
 338 bers than the original interacting waves. This follows from Teitelbaum and Vial (1991)
 339 who more generally show that the nonlinear interaction between two primary waves with
 340 [frequency, zonal wavenumber] = $[\sigma_1, s_1]$ and $[\sigma_2, s_2]$ gives rise to two secondary waves
 341 (SW) with the “sum” and “difference” frequencies and zonal wavenumbers of the pri-
 342 mary waves: $SW^+ = [\sigma_1 + \sigma_2, s_1 + s_2]$ and $SW^- = [\sigma_1 - \sigma_2, s_1 - s_2]$. The two interact-
 343 ing waves can be any combination of tides, traveling planetary waves, or SPWs (for which
 344 the frequency is zero). For example, consider the interaction between SPW1 ($[0.0d^{-1},$
 345 $+1]$) and SW2 ($[2.0d^{-1}, +2]$):

$$SPW1 \times SW2 \rightarrow SW3 + SW1 \quad (4)$$

346 where SW3 = $[2.0d^{-1}, +3]$ and SW1 = $[2.0d^{-1}, +1]$; or the interaction between DW1
 347 ($[1.0d^{-1}, +1]$) and DE2 ($[1.0d^{-1}, -2]$):

$$DW1 \times DE2 \rightarrow SE1 + SPW3 \quad (5)$$

348 where SE1 = $[2.0d^{-1}, -1]$ and SPW3 = $[0.0d^{-1}, +3]$

349 Although Teitelbaum and Vial (1991) suggest that self-interactions are not likely to be
 350 efficient, there are indications in the literature that this is not necessarily true for prop-
 351 agating waves if the background wind conditions are favorable (e.g., Pogoreltsev, 2001).
 352 This appears to be the case for the 6-hour $s = +4$ solar-synchronous tide generated by
 353 SW2 self-interaction ($SW2 \times SW2$), which achieves meridional wind amplitudes up to
 354 27 ms^{-1} at 130 km at 20°S in the terrestrial calculations of Angelats i Coll and Forbes
 355 (2002). As shown in the simulations of Huang et al. (2007), a SW2 secondary-wave tem-
 356 perature amplitude of 18K at 106 km over the terrestrial equator can result from the DW1
 357 self interaction ($DW1 \times DW1$).

3 Results

In this section, comparisons are performed between temperature amplitudes of the DW1, SW2, DE3, DE2, DE1, SE1, S0, and SW1 tidal components at 76 km from the MCD and those derived from MCS observations at 76 km altitude by the methodology described in Section 2. These tidal components all achieve amplitudes greater than 2.0 K in both MCS and MCD at 76 km. In addition, given the reasonable agreement between MCD and MCS tides at 76 km in many respects, we also explore the vertical propagation of these tides well into the thermosphere, and interpret their height-latitude structures in terms of dissipative processes and background propagation conditions. Due to their relevance to the interpretation of longitude structures measured by, e.g., MGS, MO, MRO and MAVEN, the solar-asynchronous tidal components are furthermore considered in terms of the total mass density perturbations that they produce.

The largest tidal components in both the MCD and MCS observations are the solar-synchronous components, DW1 and SW2. These are considered in the following subsection 3.1. As noted in Section 1 in connection with (2), solar-synchronous tides are longitude-independent when viewed from a quasi-solar-synchronous satellite perspective. The diurnal and semidiurnal solar-asynchronous tides, which do determine the longitude structures measured from such satellites as MGS, MO, MRO and MAVEN, are considered in subsections 3.2 and 3.3, respectively. The MCD demonstrates that SPW can sometimes produce measurable longitude variations above 100 km, and these are discussed briefly in subsection 3.4. Subsection 3.5 deals more specifically with tidal density perturbations due to tides at 108 km and 172 km, heights relevant to interpretation of MGS, MO, MRO and MAVEN respectively. A comparison between MCD longitude structures and those measured by MGS is also performed and assessed in subsection 3.5.

3.1 Solar-synchronous (“migrating”) diurnal and semidiurnal tides

Figures 2(a)-2(d) illustrate comparisons between MCS and MCD latitude versus L_s amplitude structures at 76 km for the diurnal and semidiurnal solar-synchronous tides, DW1 and SW2. The variability in these structures represent some combination of variability due to sources in the lower atmosphere, and propagation conditions between the sources and 76 km altitude. For DW1 (Figures 2(a) and 2(c)), MCS and MCD share the common feature of maxima in the equatorial-region and at higher latitudes in both hemi-

389 spheres. The equatorial DW1 maxima occur around the equinoxes, $L_s = 330\text{--}030^\circ$ and
 390 $L_s = 150\text{--}210^\circ$, in both the MCS and MCD analyses. The non-equatorial maxima oc-
 391 cur around $\pm 75^\circ$ latitude. For MCS the non-equatorial maxima in one hemisphere are
 392 not always accompanied by the counterpart in the opposite hemisphere (i.e., $L_s = 210\text{--}$
 393 300° in the S. Hemisphere). The presence(absence) of DW1 in the MCS data(MCD) be-
 394 tween about $L_s = 30\text{--}140^\circ$ corresponds exactly to the radiatively-active "aphelion cloud
 395 belt", which contributes substantial heating (Wilson, 2014; Wilson and Guzewich, 2014)
 396 and appears to be underestimated in MCD V5.3 (Navarro et al., 2014). In addition, when
 397 MCS non-equatorial maxima do exist, they are sometimes not spread towards the poles
 398 to the degree that they are in the MCD (i.e., $L_s = 0\text{--}240^\circ$ in the N. Hemisphere and L_s
 399 $= 300\text{--}360^\circ$ in both hemispheres). Finally, it is noted that MCD amplitudes generally
 400 achieve $\sim 50\%$ higher amplitudes than those obtained from MCS.

401 For SW2, the MCS (Figure 2(b)) and MCD (Figure 2(d)) amplitudes also tend to
 402 maximize at middle latitudes, with more of a tendency to minimize at equatorial lati-
 403 tudes than DW1. MCD reveals modest-amplitude ($\sim 4\text{--}5$ K) equatorial maxima around
 404 $L_s = 190^\circ$ and $L_s = 350^\circ$, with an MCS equatorial maximum (~ 7 K) occurring around
 405 $L_s = 320^\circ$. MCS SW2 maximum amplitudes generally exceed those in MCD by $\sim 50\%$.
 406 However, as emphasized in subsection 2.2, the MCS SW2 amplitudes and seasonal-latitudinal
 407 structures must be viewed with some caution, since some aliasing with the zonal mean
 408 can potentially still be included within them.

409 Figure 2(e) illustrates the latitude versus L_s structure of the zonal- and diurnal-
 410 mean zonal wind (hereafter zmU or \bar{U}) at 76 km. At this point it should be noted that
 411 the \bar{U} distribution in Mars atmosphere is not well-known, and other models that include
 412 more extensive treatment of parameterized gravity waves (e.g., Gilli et al., 2020; Medvedev
 413 et al., 2011) exhibit significant differences from MCD v5.3 while also sharing many salient
 414 features. The MCD \bar{U} distribution at this altitude and depicted in Figure 2(e) is mainly
 415 characterized by westward winds equatorward of $\sim \pm 40^\circ$ latitude, and eastward winds
 416 at higher latitudes. This provides an initial hint that \bar{U} may be playing a role in deter-
 417 mining some aspects of the DW1 and SW2 structures common to MCS and MCD, since
 418 westward-propagating waves preferentially propagate into eastward wind regimes where
 419 they are Doppler-shifted to higher frequencies and hence are less susceptible to dissipa-
 420 tion. Of course, the effects of \bar{U} are to impose a seasonal-latitudinal modulation on top
 421 of the variability due to the sources of DW1 and SW2, so that establishing cause-effect

relationships is not completely straightforward. Figure 2(f) shows that the middle- to high-latitude maxima in the MCD SW2 amplitude distribution also persists to 172 km altitude, but we will find below that this is not true for DW1. In Figures 3 and 4 and discussion below, it is our aim to clarify the role of \bar{U} on DW1 and SW2 by examining their height-latitude structures along with that of \bar{U} and our knowledge of relevant dissipative processes.

Figures 3(a) and 3(b) illustrate the height-latitude structure of DW1 temperature amplitudes and phases, respectively, from MCD for $L_s = 150^\circ$ - 180° . The overall structure in the middle atmosphere resembles the DW1 structure based on MCS data (Lee et al. 2009). The broad high-amplitude signature above ~ 130 km with no phase progression with height is the response to in-situ EUV solar radiation absorption and heating. Similarly the broad temperature response below 10-20 km is the result of near-surface heating. The 3 maxima below 100 km in Figure 3(a) are signatures of the tide that is excited near the surface and propagates to higher altitudes, as indicated by their downward phase progressions with height in Figure 3(b). The equatorial-region maximum propagates upward with a vertical wavelength of order 50 km between 20 and 70 km, and which shortens abruptly around 105 km. Meanwhile the phases associated with the mid- to high-latitude maxima reflect downward phase progression with height (indicating upward propagation) between 20 and 70 km, but nearly evanescent phase behavior above about 70 km. Note also that the mid- to high-latitude maxima are out of phase with the maximum between $\pm 30^\circ$ latitude.

Based on our knowledge of classical tidal theory, which neglects the effects of mean winds and dissipation, the near-surface heating will excite a combination of evanescent and propagating orthogonal modes or “Hough functions” that are solutions to Laplace’s tidal equation. The evanescent modes predominate at middle to high latitudes and remain trapped near the surface, whereas the propagating modes are confined to low latitudes (Chapman and Lindzen, 1970; Zurek, 1976). The first symmetric propagating mode propagates upward as the primary contributor to the response, since higher-order propagating modes have relatively short vertical wavelengths with high susceptibility to dissipation. From Table 1 the vertical wavelength of the first symmetric propagating mode is about 32 km according to classical tidal theory, and has the latitude structure given in Figure 3(c). The horizontal structures and vertical wavelengths quoted in connection with Figures 3(a) and 3(b) differ significantly from these expectations, and from obser-

455 vations and modeling of its terrestrial counterpart (see, e.g., Figure A1 in Forbes et al.,
 456 2018). Therefore, some effort is expended below to explain the nature of DW1 at Mars.

457 In connection with the terrestrial atmosphere, Forbes and Vincent (1989) devel-
 458 oped simple analytic extensions to classical tidal theory that elucidated the effects of mean
 459 winds and dissipation. They found that in regions of eastward(westward) \bar{U} , Doppler-
 460 shift of the diurnal tide to higher(lower) frequencies resulted in increased(decreased) ver-
 461 tical wavelengths and reduced(enhanced) susceptibility to dissipation. In this context,
 462 consider the latitude-height distribution of \bar{U} in Figure 3(d), which is characterized by
 463 a westward jet of order 120 ms^{-1} above 90 km, and eastward jets of order $75\text{-}100 \text{ ms}^{-1}$
 464 extending from roughly 30 km to 100 km at middle to high latitudes in each hemisphere.
 465 At least qualitatively, the Doppler-shifting effects predicted by Forbes and Vincent (1989)
 466 suggest why the vertical wavelengths in Figure 3(b) change the way they do, and why
 467 in Figure 3(a) vertical propagation of the equatorial maximum is apparently impeded
 468 by the westward jet, and the non-equatorial maxima follow the eastward jets.

469 Forbes and Vincent (1989) also showed that the non-equatorial maxima in Figure
 470 3(c) shifted poleward(equatorward) in the presence of eastward(westward) \bar{U} , and am-
 471 plified(diminished) in amplitude. However, they did not discuss this effect further, and
 472 in their example used rather modest Doppler shifts in comparison to the conditions rep-
 473 resented in Figure 3(d). The confinement of vertically-propagating DW1 components to
 474 low latitudes (c.f. Figure 3(c)) and the dominance of vertically-trapped solutions at mid-
 475 dle to high latitudes is related to the ratio of the frequency of DW1 to the planetary ro-
 476 tation rate, by analogy with the behavior of internal gravity waves in a rotating planar
 477 fluid (Lindzen, 1971; Chapman and Lindzen, 1970). Vertical propagation of gravity waves
 478 in a rotating planar system requires the gravity wave frequency σ to be less than the Brunt-
 479 Väisälä frequency and greater than twice the rotation rate (Ω), or $\sigma > 2\Omega$. On a ro-
 480 tating sphere the vertical component of the rotation vector, $\Omega \sin\theta$ (where $\theta = \text{latitude}$),
 481 replaces Ω , so that the requirement for vertical propagation becomes $\sigma > 2\Omega \sin\theta$. For
 482 DW1 ($\sigma = \Omega$), this requirement is only met equatorward of 30° latitude, which is con-
 483 sistent with low-latitude confinement of DW1 in Figure 3(c).

484 Now consider replacing σ by a Doppler-shifted frequency, $\sigma_D = \Omega + s\bar{U}/a \cos\theta$
 485 where the zonal wavenumber $s = 1$ for DW1. As σ_D increases, then the latitude at which
 486 the transition from propagating to evanescent solution should increase with latitude. Put

487 another way, one can ask what the Doppler-shifted frequency (or \bar{U}) must be at a given
 488 latitude for the solution to remain propagating:

$$\begin{aligned}\sigma_D &> 2\Omega \sin \theta \\ \Omega + \bar{U}/a \cos \theta &> 2\Omega \sin \theta \\ \bar{U} &> 241 \cos \theta (2 \sin \theta - 1) \text{ms}^{-1}\end{aligned}\tag{6}$$

489 where $241 \cos \theta \text{ ms}^{-1}$ is the zonal phase speed of DW1 and a is the planetary radius. Note
 490 that $\cos \theta = 0$ at $\theta = 90^\circ$; $(2 \sin \theta - 1) = 0$ at $\theta = 30^\circ$; and $(2 \sin \theta - 1) < 0$ equator-
 491 ward of 30° latitude. Therefore, for DW1 propagating solutions to exist at (for exam-
 492 ple) 45° , 60° and 75° latitude, then \bar{U} must exceed 70, 88, and 59 ms^{-1} , respectively.
 493 Comparing these values with those in Figure 3(d), these conditions are met in the S. Hemi-
 494 sphere, and nearly met in the N. Hemisphere, lending credibility to the argument that
 495 the extension of DW1 maxima to high latitudes as in Figure 3(a) can be explained as
 496 a Doppler-shift effect. At Earth, the zonal phase speed of DW1 is $463 \cos \theta \text{ ms}^{-1}$, mainly
 497 due to differences in the radii of the planets (6370 km for Earth; 3397 km for Mars). This,
 498 combined with the much larger zonal wind speeds at Mars, accounts for the differences
 499 in the latitudinal structures of DW1 at the two planets. This contrast between mean wind
 500 effects at Mars and Earth was also recognized by Takahashi et al. (2006), whose work
 501 we now discuss in the context of DW1.

502 Mars GCM results presented by Takahashi et al. (2006) also exhibited an increase
 503 in DW1 vertical wavelength to about 50 km compared to classical tidal theory, and an
 504 extended DW1 temperature response peaking near $\pm 60^\circ$ latitude between 60 and 80 km.
 505 The corresponding \bar{U} distribution was symmetric about the equator, with eastward winds
 506 similar to those in the N. Hemisphere in Figure 3(d), and maximum westward winds of
 507 order 20 ms^{-1} below 80 km at the equator. Using a linear tidal model, they performed
 508 numerical experiments that showed that the eastward winds at low latitudes were insuf-
 509 ficient to produce the observed increase in vertical wavelength through Doppler shift-
 510 ing, and that instead that it was the term in the zonal momentum equation that included
 511 the meridional shear of \bar{U} (i.e., the zonal-mean vorticity, $\bar{\zeta}$) that was responsible for this
 512 effect. The zonal-mean vorticity only appears with the Coriolis parameter f in the zonal
 513 momentum equation (Andrews et al., 1987), $(f + \bar{\zeta})$, and so does not really translate
 514 to a change in the planetary rotation rate.

515 The importance of $\bar{\zeta}$ to DW1 was first noted by McLandress (2002) who demon-
516 strated its contribution to the low-latitude seasonal variability of DW1 in Earth's mid-
517 dle atmosphere. Neither Takahashi et al. (2006) or McLandress (2002) addresses extreme
518 movement of the DW1 wind or temperature maxima towards the poles. This aspect of
519 the problem is complicated in the Takahashi et al. results, which includes large enough
520 dissipation (parameterized by Rayleigh friction) to couple into trapped modes (see their
521 Figure 4) above about 70 km, which significantly spreads the temperature structure to-
522 wards the poles (see their Figure 3(a)). Therefore the role of $\bar{\zeta}$ in producing these high-
523 latitude maxima is obscured by this effect. The LMD GCM does not include any pre-
524 scribed eddy dissipation in the momentum equations, and therefore the spreading to higher
525 latitudes in the MCD output must be a mean wind effect. Although the GCM includes
526 radiative damping in the thermal energy equation, this does not render the tidal equa-
527 tions inseparable, and thus does not induce any mode coupling.

528 Consider now the discussion at the beginning of this Section concerning the simi-
529 larities and differences between the latitude-Ls structures of DW1 in MCS data (Fig-
530 ure 2(a)) and in the MCD (Figure 2(c)). In particular, the non-equatorial maxima as-
531 sociated with the MCS DW1 latitude structures often do not extend as far poleward as
532 those in the MCD. The above analysis in terms of Doppler shifting effects now seems to
533 suggest that these disparities may be connected with differences in the intensity and/or
534 height-latitude structure of the zonal-mean eastward jets between the MCD and clima-
535 tological conditions at Mars. It is also possible that effects associated with vorticity of
536 the background state could also be playing a role, but these are not easily separated from
537 Doppler shifting effects within a GCM framework, and would require linear-model nu-
538 merical experiments similar to those performed by Takahashi et al. (2006) to disentangle
539 them.

540 The presence of a westward jet above 90 km is likely not the only impediment to
541 vertical penetration of DW1 above 100 km, and in fact this westward jet may exist in
542 part to the the deposition of westward momentum by the dissipating DW1. CO₂ cool-
543 ing can serve to damp DW1 temperature perturbations between 100-140 km. This is also
544 the height region where the importance of molecular dissipation, which varies accord-
545 ing to $1/\rho_0\sigma\lambda_z^2$ where λ_z is the vertical wavelength and ρ_0 is the background density, be-
546 gins to significantly damp atmospheric tides. It has also been shown that nonlinear self-
547 interaction of the terrestrial propagating DW1 tide can diminish its equatorial-region tem-

548 perature amplitudes and enhance amplitudes of its non-equatorial maxima compared to
 549 a linear solution (Huang et al., 2007), a feature that is consistent with the relative mag-
 550 nitudes seen in Figures 3(a) and 3(c). There is also evidence that the diurnal propagat-
 551 ing tide could undergo convective adjustment if the temperature amplitudes become suf-
 552 ficiently large (Lindzen, 1968; Wu et al., 1968; Zurek, 1976). From Figure 3(a) it is clear
 553 that the vertically propagating DW1 undergoes abrupt dissipation in the 80-120 km height
 554 region. However, sorting out the relative importances of these effects is not a straight-
 555 forward task within the confines of a GCM, and would benefit from numerical experi-
 556 ments using a simpler model.

557 Figures 4(a) and 4(c) illustrate examples of MCD SW2 height-latitude structures
 558 for $L_s = 60\text{-}90^\circ$ and $L_s = 300\text{-}330^\circ$, respectively. The structure in the polar middle at-
 559 mosphere resembles the SW2 structure derived previously from MCS data (Kleinböhl
 560 et al., 2013). Note that SW2 ($\sigma = 2\Omega$) satisfies the condition $\sigma > 2\Omega \sin\theta$ all the way
 561 to the poles, and thus in the absence of mean winds all SW2 modes are vertically-propagating.
 562 Contrary to DW1 where a single tidal mode dominates, for SW2 the first four Hough
 563 modes span vertical wavelengths of 47-358 km based on classical tidal theory where it
 564 is assumed that $\bar{U} = 0$. These Hough modes consist of the first two symmetric and first
 565 two antisymmetric modes, and have sufficiently long vertical wavelengths to enable prop-
 566 agation into the thermosphere. (The horizontal structures of these Hough modes are de-
 567 picted in the Supporting Information File.) In the presence of the wintertime eastward
 568 jets depicted in Figures 4(b) and 4(d), the horizontal structures of these modes will spread
 569 to higher latitudes, analogous to that illustrated for DW1. Vertical wavelengths will also
 570 lengthen, perhaps admitting even higher-order modes to those capable of propagating
 571 into the thermosphere, although use of the term “modes” may be questioned since they
 572 are likely no longer orthogonal. What all of this means is that the superposition of these
 573 Doppler-shifted latitudinal structures will be able to capture much more complex hor-
 574 izontal and vertical structures than was the case for DW1.

575 Figure 4(a) is an example of such a complex structure. Note that the winter-hemisphere
 576 eastward jet approaches $+200 \text{ ms}^{-1}$ and extends well into the thermosphere. It is flanked
 577 by the polar boundary and a westward jet extending from about 30°S to 60°N , with wind
 578 speeds approaching -150 ms^{-1} . This configuration essentially forms a propagation chan-
 579 nel for SW2, which extends well into the thermosphere where molecular dissipation lim-
 580 its its growth with height around 150 km. At $L_s = 300\text{-}330^\circ$ (Figures 4(c) and 4(d)) the

581 winter hemisphere eastward jet does not exhibit extension into the thermosphere, and
 582 a propagation path does not exist to the extent that it does during $L_s = 60-90^\circ$. The
 583 modest SW2 amplitudes above about 100 km and between about $\pm 30^\circ$ latitude, with
 584 larger values prevailing around perihelion, likely originate from heating due to in-situ ab-
 585 sorption of EUV solar radiation, by analogy with the thermosphere maximum seen in
 586 Figure 3(a).

587 3.2 Solar-asynchronous (“non-migrating”) diurnal tides

588 The focus of this subsection is on the solar-asynchronous diurnal tides with MCS
 589 temperature amplitudes exceeding 2K at 76 km, and these consist of DE3, DE2 and DE1,
 590 as illustrated in Figure 5. Moudden and Forbes (2008a) performed numerical experiments
 591 with a Mars GCM (however, only for $L_s = 75^\circ$) to better understand the sources of solar-
 592 asynchronous tides. They concluded that the most important source of DE2(DE1) was
 593 the interaction of $m = 3(m = 2)$ topography with the solar-synchronous diurnal com-
 594 ponent (DW1) of near-surface solar heating, analogous to the interaction between SPW3(SPW2)
 595 and DW1: $SPW3(SPW2) \times DW1 \rightarrow DE2(DE1) + DW4(DW3)$. The $m = 2$ topography-
 596 DW1 heating interaction that produces DE1 was in fact first anticipated by Zurek (1976)
 597 and its associated temperature response was detected in Mariner 9 IRIS data by Con-
 598 rath (1976). From Table 2, DE2 and DE1 possess vertical wavelengths in the range cor-
 599 responding to efficient vertical coupling, whereas DW4 and DW3 consist of short ver-
 600 tical wavelength modes and do not contribute to vertical coupling. Moudden and Forbes
 601 (2008a) also conclude the following to be a secondary source of DE2: $SPW1 \times DE1 \rightarrow$
 602 $D0 + DE2$, and that zonal variations in surface thermal inertia and albedo served as
 603 secondary contributors to DE1. These authors did not mention DE3, but it seems log-
 604 ical to conclude that it arises from the same DW1 solar heating-topography interaction
 605 as DE2 and DE1, except this time with $m = 4$.

606 The general characteristics of DE3, DE2 and DE1 were comprehensively delineated
 607 using MGS Thermal Emission Spectrometer (TES) temperature measurements near 25
 608 km by Wilson (2000). It is immediately apparent from Figure 5 that DE3, DE2, and DE1
 609 are confined to latitudes equatorward of $\pm 40^\circ$ latitude, have comparable amplitudes with
 610 those of MCD, and that the MCS structures tend to be shifted poleward in the N. Hemi-
 611 sphere in comparison to MCD. These tides are sometimes referred to as diurnal Kelvin
 612 waves (specifically DK3, DK2, DK1; Wilson, 2000), although strictly speaking only the

613 first symmetric Hough mode of a given zonal wavenumber is a Kelvin wave; latitudinal
614 asymmetries and other deviations from Kelvin wave shapes (such as the shifting in lat-
615 itude of an otherwise Gaussian-like latitude structure), are captured by higher-order Hough
616 modes. In fact, we have determined that all the displayed structures in Figure 5 are well
617 captured by the first 4 Hough modes (see Supporting Information File), the first 3 of which
618 are illustrated in the Figure 6. However, the reconstructed structures (not shown) based
619 on these Hough functions are slightly more confined (by about 10° in each hemisphere),
620 compared to those displayed in Figure 5. This is due to the Doppler-shifting effect of mean
621 winds, which are now discussed in connection with with DE3, DE2 and DE1 in the con-
622 text of their vertical propagation characteristics as displayed in Figure 7.

623 The MCD height-latitude structures of DE3, DE2, and DE1 are shown in Figure
624 7(a), (b) and (c), respectively, for $L_s = 60-90^\circ$, along with the corresponding \bar{U} in Fig-
625 ure 7(d). In Figure 7 as well as upcoming Figures 9-11, the plotted variable is percent
626 total mass density perturbation relative to the zonal- and diurnal-mean. Total mass den-
627 sity is the quantity plotted due to the relevance of these tides to longitude structures in
628 density that are evident in MGS, MO, and MRO accelerometer measurements and MAVEN
629 mass spectrometer measurements. The displayed latitude structures are increasingly broad
630 for DE3, DE2 and DE1, consistent with expectations based on classical tidal theory as
631 illustrated in Figure 6. In terms of mean wind effects, propagation of these eastward-
632 propagating tidal components favors the propagation channel produced by the prevail-
633 ing westward wind regime that is shifted slightly into the N. Hemisphere.

634 In the following, we refer to the symmetric and antisymmetric components of DE3,
635 DE2, and DE1 about the equator, and also the phases of these tidal components. These
636 plots are included in Figures S3, S4 and S5, respectively, of the Supporting Information
637 file.

638 The vertical wavelengths for DE3 in Table 1, and diagnosis of the symmetric and
639 antisymmetric structures of DE3 about the equator (not shown, refer to Figure S2), re-
640 veal a few insights relevant to the interpretation of Figure 7(a). The vertical wavelength
641 of the symmetric part is about 80 km below 100 km, consistent with the value of 64 km
642 listed in Table 1 and some lengthening due to Doppler-shifting effects of the prevailing
643 westward winds. The vertical wavelength lengthens much further beginning at about 120
644 km where dissipation begins to dominate, and the symmetric part of DE3 reaches its max-

645 inum value of 12% at the equator. The antisymmetric part of DE3 also peaks near 120
646 km but at amplitudes of 6% near $\pm 30^\circ$ latitude, as compared with $\pm 20^\circ$ from classical
647 tidal theory (Figure 6), which is consistent with the latitude broadening mentioned above.
648 Its vertical wavelength can be best discerned from examination of Figure 7(a), noting
649 that the structure is shifted to the S. Hemisphere around 70 km, but a maximum shift
650 towards the N. Hemisphere occurs at about 120 km. This yields an effective vertical wave-
651 length of ~ 100 km, which includes the lengthening effects of both Doppler shifting and
652 dissipation. Thus, both the first symmetric and antisymmetric modes of DE3 play im-
653 portant roles in determining its latitude-height structure.

654 A similar situation exists with regard to DE2 (Figure 7(b)), but with distinct dif-
655 ferences. The symmetric part (Figure S4) also peaks near 120 km, but with a much larger
656 equatorial maximum of 26% and vertical wavelength below 100 km of about 107 km. This
657 value for the vertical wavelength suggests that the value of 122 km in Table 1 may be
658 an overestimate, but as we noted in connection with (3), Table 1 is only intended to serve
659 as a rough guide. The antisymmetric part of DE2 (Figure S4) has maximum amplitudes
660 that increase monotonically with height from 1% at 50 km to about 9% near 170 km.
661 The latitudinal maxima shift from about $\pm 20^\circ$ latitude near 50 km, to ± 30 - 50° near 120
662 km, to ± 20 - 40° latitude at 170 km. We interpret this degree of spreading to be consis-
663 tent with Doppler-shift modification of the first antisymmetric mode for DE2 plotted in
664 Figure 6(b). As with DE3, the influences of the antisymmetric part of DE2 are evident
665 in the total structure plotted in Figure 7(b). The maximum hemispheric shifts occur to-
666 wards the S. Hemisphere at 80 km, to the N. Hemisphere at 120 km, and back towards
667 the S. Hemisphere 170 km. These yield vertical wavelengths of order 80 km between 80-
668 120 km and 100 km between 120-170 km.

669 DE1 is particularly interesting since the first symmetric component is a resonant
670 oscillation (or “normal mode (NM)”) in Mars’ atmosphere (Zurek, 1976, 1988). Indeed,
671 the symmetric part of Figure 7(c) (refer to Figure S5) shows little evidence for phase pro-
672 gression with height, as would be expected for a NM. It achieves a maximum value of
673 21% near 120 km. The antisymmetric part of Figure 7(c) (refer to Figure S5) shows ev-
674 idence of the first antisymmetric Hough mode with maxima near $\pm 30^\circ$ latitude, and this
675 accounts for most of the asymmetry seen at altitudes above 150 km and between 50 and
676 80 km. However, there is a broad antisymmetric component that extends nearly pole to
677 pole that can only be accounted for by evanescent tidal components which arise through

678 “mode coupling” induced by dissipation of DE1 between 100 and 150 km in the pres-
 679 ence of strongly asymmetric mean winds. The presence of these evanescent components
 680 accounts for the shift of the maximum at 120 km into the N. Hemisphere, but the max-
 681 imum above 100 km near the S. Pole arises from the nonlinear interaction

$$SPW1 \times D0 \rightarrow DW1 + DE1 \quad (7)$$

682 according to the numerical experiments conducted by Moudden and Forbes (2008a). The
 683 MCD D0 shows a temperature(density) maximum of 5-10K(3-6%) in this region for L_s
 684 = 60-90° (not shown). The SPW1 counterpart is illustrated in subsection 3.4 (Figure
 685 10(a)). Forbes et al. (2018) interpreted winter hemisphere $s = 1 \sim 10-15\%$ polar-region
 686 density perturbations in both MO and MRO accelerometer measurements in terms of
 687 some combination of D0 and SPW1 .

688 3.3 Solar-asynchronous (“non-migrating”) semidiurnal tides

689 The MCS and MCD latitude- L_s structures at 76 km for the semidiurnal solar-asynchronous
 690 tides that exceed 2K in amplitude are shown in Figure 8, and consist of SE1, S0 and SW1.
 691 Examples of height-latitude structures are also shown in Figure 9 for $L_s = 240-270^\circ$. For
 692 the convenience of interpreting structures in both Figures 8 and 9, a subset of Hough
 693 modes for SE1 and S0, and the height-latitude distribution of zonal-mean zonal wind are
 694 also provided in Figure 9.

695 The existence of SE1 in Mars atmosphere has been recognized in several previous
 696 analyses of MCS temperatures (Lee et al., 2009; Guzewich et al., 2012; Moudden and
 697 Forbes, 2015; Wu et al., 2015), and Wu et al. (2015) were the first to report the pres-
 698 ence of SW1 which they interpreted in terms of $SPW1 \times SW2$ nonlinear interaction (i.e.,
 699 expression (4)). S0 was suggested as a possible contributor to observed longitude wave-
 700 2 density variability measured by MRO above 80° latitude around $L_s = 75-80^\circ$ by Forbes
 701 et al. (2018); however, it was indistinguishable from DE1 due to the nature of MRO LST-
 702 longitude sampling. From their numerical experiments Moudden and Forbes (2008) con-
 703 cluded that S0 and SE1 were primarily generated by modulation of the SW2 component
 704 of solar radiative heating by zonal wavenumber $m = 2$ and $m = 3$ components of to-
 705 pography, respectively. They also found SW1 in their simulation, but could not distin-
 706 guish whether this was generated by SW2 solar radiative interaction with $m = 1$ to-
 707 pography, or nonlinear interaction between SW2 and SPW1.

708 MCS and MCD SE1 temperature amplitudes in Figures 8(a) and 8(b) are distin-
709 guished by their 3-peaked structures in latitude (for MCS, only between $L_s = 180\text{-}360^\circ$),
710 and their similarities in amplitude. From Table 1, the first symmetric and antisymmet-
711 ric modes of SE1 are evanescent, and the second symmetric and antisymmetric modes
712 have vertical wavelengths of 134 km and 74 km, respectively. Therefore, compared to
713 the diurnal tides in the previous subsection, a more structured response is expected. The
714 MCS structures between $L_s = 30\text{-}150^\circ$ require multiple Hough functions (refer to Fig-
715 ure S2) to replicate, including the 4-peaked second antisymmetric mode and first and
716 second symmetric modes. Interestingly, MCD reflects the strong presence of the second
717 antisymmetric mode during this same period, but combines with the second symmet-
718 ric mode (not shown) to produce a much different structure than MCD. Between $L_s =$
719 $210\text{-}330^\circ$ the second symmetric mode is important for both MCD and MCS, giving rise
720 to the 3-peaked structures.

721 For S0, it is important to note that non-zero temperature amplitudes at the poles
722 are possible, and the fundamental mode is antisymmetric, in contrast to the complete
723 set of non-zonally symmetric tides. The MCS and MCD amplitude distributions of S0
724 are remarkably similar, and with comparable amplitudes. Between about $L_s = 180\text{-}360^\circ$,
725 MCS and MCD are both dominated by a 4-peaked structure with a null at the equator,
726 whereas between $L_s = 0\text{-}180^\circ$ both model and data consist of 3 peaks but with most of
727 the activity in the N. Hemisphere. As with SE1, the first two Hough modes possess evanes-
728 cent vertical structures (see Table 1), and the second antisymmetric and symmetric modes
729 have vertical wavelengths from classical tidal theory of 99 km and 63 km, respectively,
730 thus leading to the expectation that SE1 will consist of a higher-order structured response.
731 Indeed, both MCS and MCD consist in large part of the second antisymmetric mode,
732 with maxima near $\pm 20\text{-}30^\circ$ and near-polar latitudes, with a null at the equator.

733 SW1 is somewhat similar to SW2 in that it consists of at least four modes with ver-
734 tical wavelengths that are capable of efficiently coupling the middle atmosphere with the
735 thermosphere. The means that they can combine in a various ways to produce a vari-
736 ety of complex structures, including an isolated response in one hemisphere. There are
737 several similarities between Figures 8(e) and 8(f) including the elongated structure be-
738 tween $L_s = 0\text{-}240^\circ$ in the S. Hemisphere, complementary maxima in the N. Hemisphere
739 between $L_s = 90\text{-}210^\circ$, the maximum around $L_s = 300^\circ$ and 20°N , and the elongated
740 structure poleward of this maximum between $L_s = 240\text{-}330^\circ$, albeit at 60°N for MCD

741 and about 75° for MCS. The maximum amplitudes in Figure 8(f) are about 40% lower
 742 for MCD than MCS, so it takes a little effort to see these similarities on plots with the
 743 same color scales.

744 3.4 Stationary planetary waves

745 SPWs are generally not thought to be capable of propagating into the thermosphere,
 746 primarily based on the lack of reported observations of such penetration at Earth, as well
 747 as terrestrial modeling results (Pogoreltsev et al., 2007); perhaps also because early in-
 748 terpretations of SPWs in aerobraking densities (Keating et al., 1998) were shown to be
 749 unlikely in favor of a solar-asynchronous tide interpretation (Forbes and Hagan, 2000;
 750 Joshi et al., 2000; Wilson, 2002). However, Forbes et al. (2018) recently discussed pos-
 751 sible interpretation of winter polar-region $s = 1$ features in MO and MRO densities,
 752 and suggested a connection with either SPW1 or a displaced polar vortex, as well as D0.
 753 SPW can also be generated by nonlinear wave-wave interactions, as exemplified in (5)
 754 and demonstrated in numerical simulations by Angelats i Coll et al. (2004) and Moud-
 755 den and Forbes (2008a). A few examples of SPWs in MCD perturbation densities from
 756 the zonal mean are shown in this section.

757 Height-latitude depictions of MCD SPW1, SPW2, and SPW3 density perturbations
 758 are shown in Figures 10(a), (b), and (c). SPW1 is particularly large, approaching 30%
 759 near 100 km. Moulden and Forbes (2008a) also obtained a similar SPW1 structure in
 760 their $L_s = 75^\circ$ simulation, but with maximum amplitudes of about 13% near 80 km. Up-
 761 ward propagation of SPWs in the prograde jet is well known from previous studies of
 762 Mars' lower and middle atmosphere (Hollingsworth and Barnes, 1996; Guzewich et al.,
 763 2012). SPW2 in Figure 10(b), with amplitudes of order 5-10% near 170 km, possesses
 764 a similar structure to DE1 (see Figure 7(c)) above 100 km, and it likely arises from

$$765 \quad DW1 \times DE1 \rightarrow S0 + SPW2 \quad (8)$$

765 SPW3 in Figure 10(c), with maximum amplitudes of order 8%, appears to have its ori-
 766 gins in the nonlinear interaction

$$767 \quad DW1 \times DE2 \rightarrow SE1 + SPW3 \quad (9)$$

767 This is consistent with the coincidence of the large amplitudes of DE2 in Figure 7(b) with
 768 those of the in-situ forced DW1, which is prevalent in the thermosphere.

769 Figures 10(d), (e) and (f) illustrate the latitude versus Ls structures of density per-
 770 turbations associated with MCD SPW1, SPW2, and SPW3, respectively at 172 km al-
 771 titude. This height is chosen due to its relevance to neutral mass spectrometer measure-
 772 ments from MAVEN. Figures 10(g), (h) and (i) show the same depictions for 108 km.
 773 This altitude is chose since it is representative of altitudes where accelerometer-measured
 774 densities originated from MGS, MO and MRO. Amplitudes decrease with height for SPW1
 775 ($\sim 10\text{-}15\%$ vs. $\sim 2\text{-}5\%$), but increase with height for SPW2 and SPW3 ($\sim 2\text{-}4\%$ vs. $4\text{-}8\%$).
 776 In addition, there are significant differences between the latitude-Ls structures between
 777 108 km and 172 km. Both of these features are consistent with the hypothesis that in-
 778 situ generation by nonlinear wave-wave interactions is occurring in the thermosphere.
 779 It also implies that, at least insofar as SPW are concerned, longitude structures evident
 780 in aerobraking densities are not good indicators of what to expect in MAVEN mass spec-
 781 trometer measurements.

782 3.5 Additional insights into thermosphere density variability

783 Figure 11 presents latitude-Ls depictions of MCD DE3, DE2, DE1, SE1, S0 and
 784 D0 density perturbations at 108 and 172 km. DE1 and DE2 are the dominant tides, each
 785 with 15-20% density perturbations at 108 and 172 km. At 108 km, amplitudes associ-
 786 ated with DE3, SE1, S0 and D0 are all of order 4-7%, while at 172 km these tides gen-
 787 erally contribute on the order of 5-12% each except for DE3 which is in the range 4-7%.
 788 Note that SW1 in our list of important solar-asynchronous tides is replaced here with
 789 D0, since D0 has about twice the amplitudes of SW1 at these altitudes. Examination
 790 of its height-latitude structures suggest that it does not originate to any significant de-
 791 gree in Mars' lower or middle atmosphere, but rather through nonlinear interactions with
 792 large-amplitude tides in the thermosphere:

$$793 \quad SPW1 \times DW1 \rightarrow DW2 + D0 \quad (10)$$

$$794 \quad SPW1 \times DE1 \rightarrow D0 + DE2 \quad (11)$$

$$795 \quad SPW2 \times DE2 \rightarrow D0 + DE4 \quad (12)$$

796 As shown below, the aggregation of these various tidal components results in quite sub-
 797 stantial ($\sim \pm 40\%$) total density perturbations in the thermosphere that are in line with,
 e.g., those derived from accelerometer measurements.

798 In addition to the MCD-MCS comparisons in Figures 2, 5, and 8, the veracity of
799 the MCD tidal structures is further checked by comparing longitude structures in the
800 model against those revealed in MGS accelerometer measurements between 105-115 km
801 in Figure 12. Consistent with the MCD model results in Figure 11, the MGS densities
802 are cast in the form of percent residuals from the zonal mean within a window of time,
803 and within each window the residuals are fit to sinusoids with respect to longitude with
804 zonal wavenumbers $s = 1, 2, 3, 4$. That window is then moved through the data one sol
805 at a time. It is desired that the window length be as short as possible so that time vari-
806 ations can to some degree be separated from latitude variations, yet still yield robust and
807 relatively clean results for the individual wavenumber components. After trial and er-
808 ror, it was found that requiring the maximum spacing between any two adjacent longi-
809 tudes in sequence within the window be less than 35° in longitude was a good choice.
810 It is also noted that in Figures 12-14, the percent density perturbations are now rela-
811 tive to the zonal-mean densities at the height, latitude and local time of the correspond-
812 ing MGS densities.

813 Figure 12(a) illustrates the latitude-longitude structures obtained from the MGS
814 densities according to the procedure described above, with L_s as a secondary y-axis. Lo-
815 cations of the individual periapsis data points are also included in Figure 12(a). Using
816 the MCD-supplied software, the MCD model is interpolated and sampled at the height,
817 latitude, longitude, LST and L_s of every MGS measurement location indicated in Fig-
818 ure 12(a), and analyzed identically to the MGS data to obtain Figure 12(b). Also shown
819 in Figure 12(b) are the number of points in each fitting window, and the length of each
820 fitting window, that result from the longitude sampling criterion described above. Fig-
821 ure 12(c) illustrates the latitude-longitude structure of density obtained from full lati-
822 tude and longitude sampling of the MCD model at the heights, LSTs, and L_s values of
823 the MGS measurements.

824 The main MGS feature (Figure 12(a)) that is well captured in both amplitude and
825 phase by the sampled MCD (Figure 12(b)), is the wave-2 structure that extends from
826 50°N to 40°S in both figures. However, the maximum amplitudes near 60° and 240° lon-
827 gitude sometimes exceed 60% in both figures, whereas in Figure 12(c) the maximum am-
828 plitudes are in the same location except they are of order 40%. This suggests that the
829 excess amplitudes in Figures 12(a) and 12(b) are connected with the incomplete nature
830 of MGS sampling. The tides that are capable of contributing to wave-2 in Figure 12(c),

831 i.e., those with a space-based wavenumber $k_s = |s - n| = 2$, include DE1 and S0. In
832 the $\pm 30^\circ$ equatorial region, DE1 is the dominant contributor with S0 making a secondary
833 contribution, as suggested by 108 km results in Figure 11. Based on Figure 7(c) which
834 is applicable between $L_s = 60\text{--}90^\circ$, DE1 amplitudes are of order 12-20% between 110-
835 115 km, the location of MGS periapsis during $L_s = 45\text{--}85^\circ$. These values account for only
836 about half the wave-2 maxima/minima between $\pm 30^\circ$ in Figure 12(c), and suggests that
837 the remainder might be accounted for in terms of the constructive/destructive interfer-
838 ence effects of additional harmonics of the total longitudinal structure. Indeed, visual
839 evidence of wave-3 exists in Figure 12(c) over much of the domain, and also wave-1 pole-
840 ward of 40°S . This interpretation is now pursued in connection with Figures 13 and 14.

841 Figures 13 and 14 present the same three categories of plots as in Figure 12, ex-
842 cept for longitudinal wave-1 and wave-2 in Figure 13, and wave-3 and wave-4 in Figure
843 14. Focusing on wave-2 for the moment, the MCD model indicates low-latitude maxima
844 of order 15-22%, consistent with the above interpretation that DE1 and S0 are the main
845 contributors. Examination of DE1 and S0 in Figure 11 also indicate that these tidal com-
846 ponents are the main contributors to the maxima poleward of 60°S , consistent with the
847 interpretation of polar-region MRO data by Forbes et al. (2018). Examination of MCD
848 wave-3 and wave-4 amplitudes in Figures 14(c) and 14(f), respectively, indicate ampli-
849 tudes of order 10-15%, with maxima near 60° and 240° for wave-4 and 90° and 210° for
850 wave-3. These are fairly close to the wave-2 maxima in Figure 13(f), confirming that the
851 aggregate effect of combining wave-3 and wave-4 to wave-2 accounts for the difference
852 in low-latitude amplitude maxima between Figure 12(c) and Figure 13(f).

853 Referring to Figures 10(a), 10(g), and 11(i), the primary contributor to wave-1 in
854 Figures 13(b) and 13(c) is SPW1, with D0 making non-negligible contributions. This is
855 also consistent with interpretations made by Forbes et al. (2018) in the context of polar-
856 region MRO aerobraking density measurements. However, there is about a 150° longi-
857 tude shift compared to the maxima/minima in the S. Hemisphere of the MGS measure-
858 ments in Figure 13(a). This accounts for some of the differences in the structures pole-
859 ward of 40°S between Figures 12(b) and 12(a). In addition, the MCD generally under-
860 estimates wave-1 amplitudes in the S. Hemisphere compared to MGS, which points to
861 a deficiency of the model.

862 The wave-3 MCD amplitudes in Figure 14(b) agree well in phase with those from
 863 MGS in Figure 14(a) between 30°S-55°N, but underestimate amplitudes (roughly, 12-
 864 15% vs. 15-25%). Based on Figures 7(b), 11(b), and 11(e), the main contributor to $k_s =$
 865 $|s-n| = 3$ at these latitudes is DE2 with non-negligible contributions from SE1. Pole-
 866 ward of 60°S SE1 likely assumes the primary role, and its signatures in Figure 14(b) also
 867 appear with similar amplitude and phase in the MGS density perturbations in Figure
 868 14(a). The wave-3 density perturbations from MGS between 40°S-60°S are not replicated
 869 by MCD. The MCD wave-4 densities in Figure 14(e) between 55°S-55°N similarly un-
 870 derestimate those from MGS in Figure 14(d), but the phases are again in quite good agree-
 871 ment except for the 25° longitude phase difference near 50°S. MCD DE3 amplitudes are
 872 of order 5-10% in this height regime at $L_s = 60-90^\circ$ and between 20°S-50°N, but SE2
 873 amplitudes (not shown) are of order 3-5%, extend between 40°S-60°N, and are thus not
 874 negligible in comparison. We conclude that most of the structures seen in Figures 14(d)
 875 and 14(e) between $\pm 55^\circ$ latitude are attributable to the combinations of DE3 and SE2.

876 4 Summary and Conclusions

877 In the first part of this paper, Mars Climate Sounder (MCS) temperature measure-
 878 ments at 76 km were used to construct a multi-year latitude versus L_s climatology of the
 879 tidal spectrum. The most important spectral components included the solar-synchronous
 880 (“migrating”) components DW1, SW2, and the solar-asynchronous (“non-migrating”)
 881 tides DE3, DE2, DE1, SE1, S0, and SW1. A similar analysis was performed on model
 882 output from the Mars Climate Database, and the same tidal components arose as the
 883 most significant. The largest tidal components were found to be DW1 and SW2. The
 884 MCS amplitudes of DW1 and SW2 were comparable ($\sim 4-9\text{K}$), whereas MCD amplitudes
 885 were somewhat larger ($\sim 6-14\text{K}$) for DW1 and smaller for SW2 ($\sim 3-7\text{K}$). Assessment of
 886 the veracity of MCD with regard to SW2 at latitudes equatorward of $\pm 40^\circ$ latitude is
 887 impeded by the particularly sparse local time sampling of MCS at these latitudes, which
 888 introduces an unknown level of uncertainty due to potential aliasing with the zonal- and
 889 diurnal-mean component of temperature. Discrepancies between the measured and mod-
 890 eled estimates of DW1 potentially arise from an underestimation of water-ice cloud ra-
 891 diative forcing in the MCD, and possibly the zonal-mean zonal winds (\bar{U}) in MCD.

892 The height-latitude structures (0-170 km, pole-to-pole) of DW1 and SW2 were ex-
 893 amined within the context of the MCD. Emphasis was placed on a few typical examples

894 that revealed the influences of \bar{U} and dissipation on these structures. DW1 and SW2 re-
 895 sponded to eastward \bar{U} 's at middle to high latitudes through Doppler-shifting to higher
 896 frequencies, latitudinal broadening, lengthening of their vertical wavelengths, and reduced
 897 dissipation. The vertical wavelength of DW1 at low-middle latitudes is of order 50 km,
 898 as compared with 32 km from classical tidal theory. As shown by Takahashi et al. (2006),
 899 this can be traced to the meridional shear in \bar{U} , as embodied in the zonal-mean vortic-
 900 ity $\bar{\zeta}$ term in the zonal momentum equation (see also McLandress (2002) for other ef-
 901 fects on DW1 due to $\bar{\zeta}$ at Earth). Furthermore, the zonal-mean westward jet, which is
 902 a prominent feature at low latitudes above 80–100 km in Mars atmosphere, represented
 903 an impediment to vertical propagation, so that the solar-synchronous tides tend to prop-
 904 agate within propagation channels bounded by this westward jet and the poles. Doppler-
 905 shifting and $\bar{\zeta}$ effects on tides at Mars are much more extreme than in Earth's atmosphere,
 906 due to Mars' smaller radius and the more intense \bar{U} 's that exist in its atmosphere.

907 Significantly better MCD-MCS agreement in terms of seasonal-latitudinal struc-
 908 tures and amplitudes were found for the solar-asynchronous tidal components. This led
 909 us to examine height-latitude structures of their *density perturbations* with respect to
 910 the zonal mean, in light of their relevance to interpretation of accelerometer-measured
 911 densities from MGS, MO and MRO, as well as mass spectrometer measurements on MAVEN.
 912 The largest solar-asynchronous tides, DE3, DE2 and DE1, are all eastward-propagating
 913 and confined to low latitudes. The \bar{U} distribution therefore favored vertical propagation,
 914 so that quite large density perturbations ($\sim 15\text{-}25\%$) were attainable in the 100-170 km
 915 height region. These large amplitudes also led to the secondary generation of tides through
 916 tide-tide and tide-SPW1 nonlinear interactions (for instance, D0 at $\sim 5\text{-}10\%$), and to the
 917 in-situ production of SPW in the thermosphere. Amplitudes of order 5-10% were found
 918 for SPW2 and SPW3 at 170 km. SE1 and S0 amplitudes were found to range between
 919 10 and 20% at 170 km altitude, whereas SW1 was much smaller in general ($<5\%$).

920 The net effect of these solar-asynchronous tides and SPWs is that in aggregation
 921 their constructive/destructive interference can lead to significant longitudinal variabil-
 922 ity in addition to local time variability. To further test the veracity of the MCD in this
 923 regard, a comparison between MCD density perturbations and those derived from Phase-
 924 II aerobraking of MGS was performed. The MGS latitude (55°N to 40°S) versus longi-
 925 tude structure sampled between $L_s = 40^\circ$ and $L_s = 80^\circ$ is dominated by a wave-2 struc-
 926 ture with amplitudes as large as $\pm 65\%$. The MCD comparison was performed with MCD

927 sampled according to the same heights, latitudes, LSTs, longitudes and Ls values as MGS,
928 and analyzed identically. In addition, a comparison was made with the same latitude-
929 longitude structure obtained from full sampling of the MCD. This allowed isolation of
930 the influences of sampling from those of the actual underlying tides.

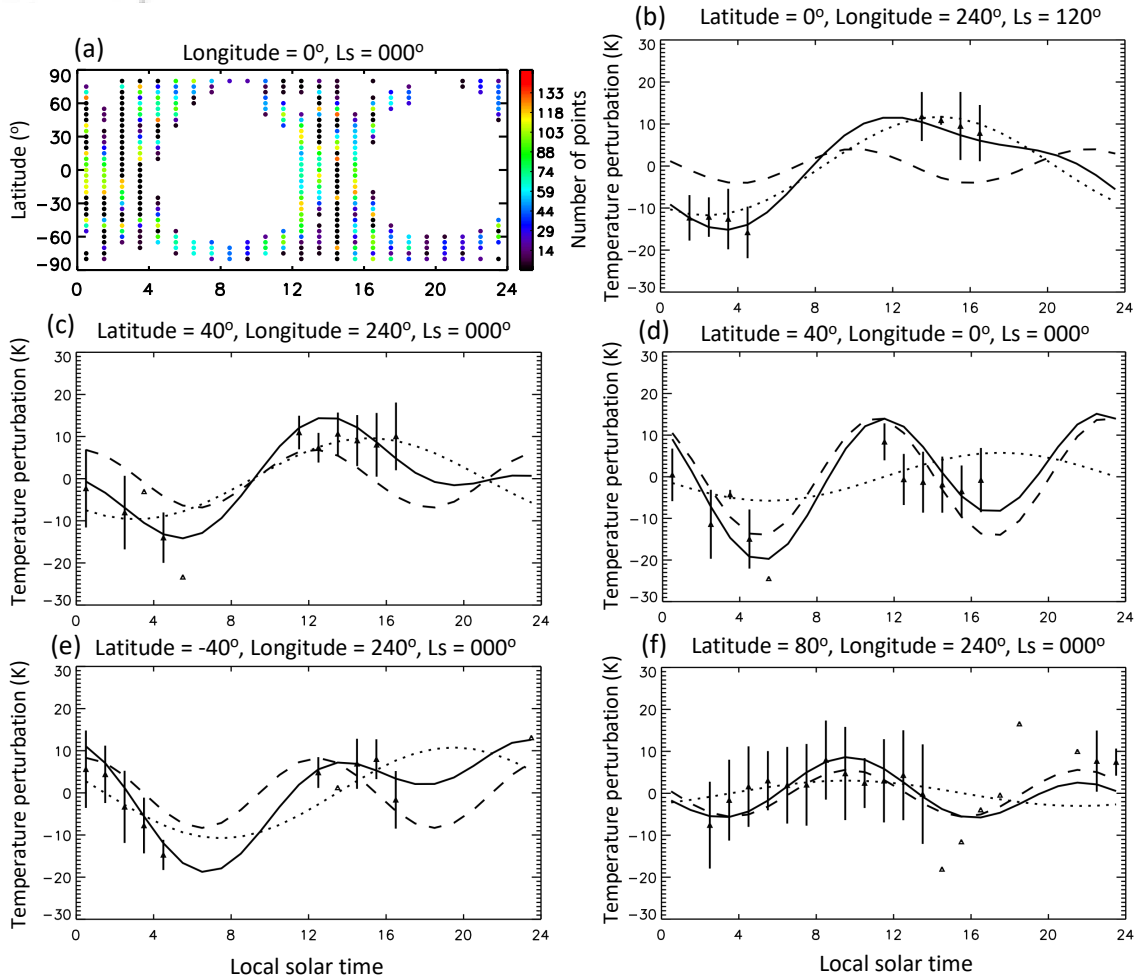
931 First of all, the MGS-sampled MCD captured the dominant wave-2 structure in
932 MGS very well in phase, and achieved almost as large maximum amplitudes. However,
933 these were significantly larger than the $\sim \pm 45\%$ amplitudes from the fully-sampled MCD,
934 indicating that some of the former originated from high-order modes and/or day-to-day
935 variability not captured by the incomplete nature of the MGS sampling. Furthermore,
936 the $\pm 45\%$ amplitudes from the fully-sampled MCD were much larger than those con-
937 tributed by DE1 itself ($\sim \pm 20\%$), which at first sight and in many prior publications
938 is the assumed interpretation of the origin of wave-2. The deficit was in fact found in
939 wave-3 and wave-4 (due mainly to DE2 and DE3, respectively), which combined syner-
940 gistically with DE1 to nearly replicate the larger-amplitude "wave-2" structure in the
941 MGS-sampled MCD and in the MGS observations themselves.

942 Poleward of 40°S the MGS data and the MCD both revealed wave-2 density per-
943 turbations of order $\sim 10\%$, similar in phase. It was determined that this feature origi-
944 nated from a combination of DE1 and S0. MGS and MCD also indicated 15-20% wave-
945 1 density perturbations in the same latitude regime, but with about a 150° longitudi-
946 nal shift in phase. The primary contributor to wave-1 was concluded to be SPW1, with
947 D0 making non-negligible contributions. These same conclusions regarding origins of wave-
948 1 and wave-2 were made by Forbes et al. (2018) in the context of interpreting MRO aer-
949 obraking data at high S. Hemisphere latitudes around $L_s \approx 80^\circ$.

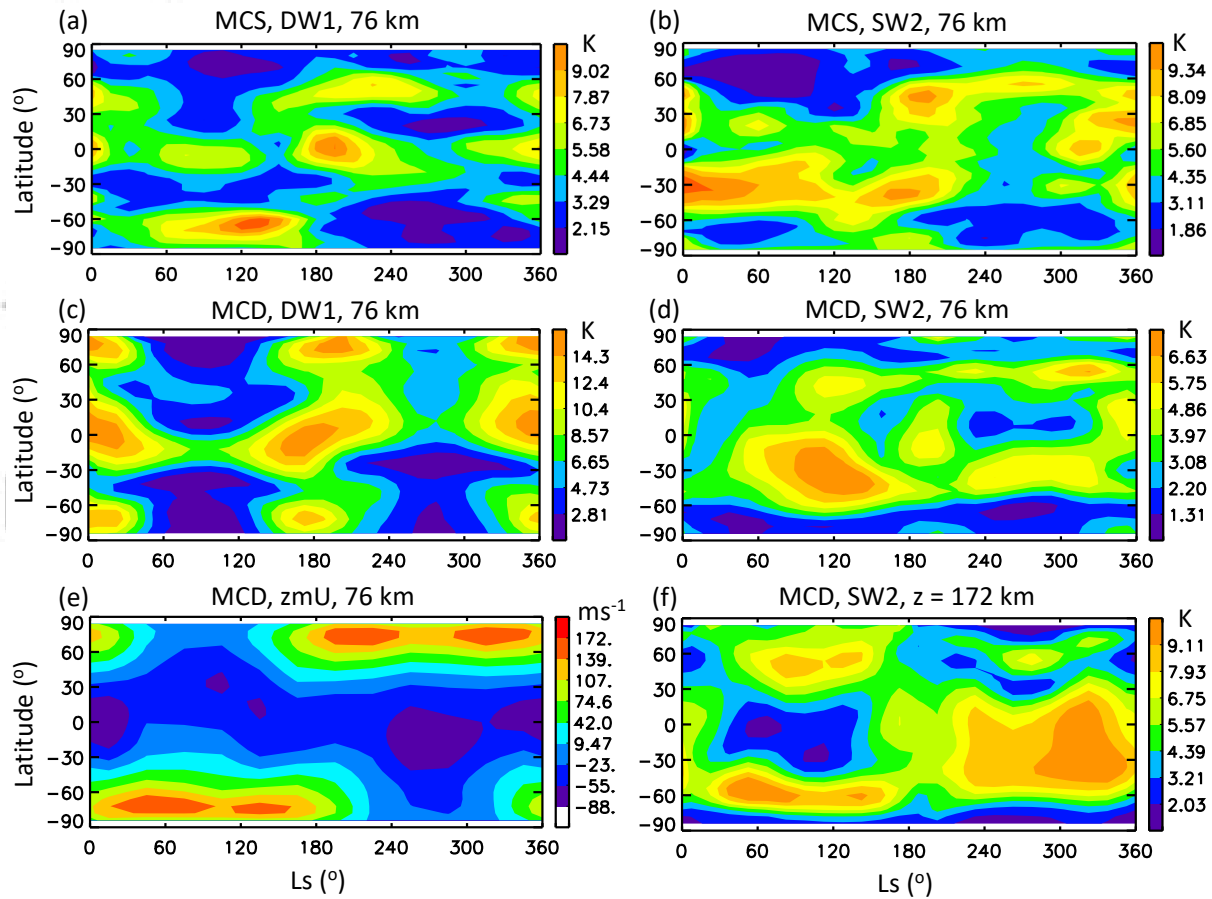
950 With regard to the solar-asynchronous tides that give rise to significant longitu-
951 dinal variability in the thermosphere of Mars, it is concluded that MCD captures the salient
952 amplitude and phase characteristics of the $\sim \pm 30\text{-}60\%$ longitudinal density perturba-
953 tions measured by the MGS accelerometer near 105-115 km, and the seasonal-latitudinal
954 variability of the tidal spectrum derived from MCS temperature measurements at 76 km.
955 This makes credible the interpretations contained herein regarding the height-latitude
956 structures of various tidal components between 76 and 170 km. Additional assessments
957 of the ability of MCD to replicate longitude structures from MO, MRO, and MAVEN
958 are planned for future work.

959 Finally, despite the impressive agreement between the MCD and MRO/MCS and
960 MGS accelerometer data demonstrated herein, noteworthy differences and knowledge gaps
961 remain. The \bar{U} distribution in the middle and upper atmosphere of Mars is not well con-
962 strained by existing observations, and moreover is likely to be affected by gravity waves
963 (e.g., Medvedev et al., 2011) whose properties in Mars' atmosphere are also not well con-
964 strained. Some recent progress on incorporating non-orographic gravity wave effects into
965 the LMD GCM used to produce the MCD have recently been made (Gilli et al., 2020),
966 and will be incorporated into future MCD versions. As acknowledged above, tidal forc-
967 ing by radiatively active water-ice clouds warrants closer examination in the LMD GCM,
968 and because this source is modulated by topography, future modifications will likely in-
969 fluence both solar-synchronous and solar-asynchronous tides. Similarly, the GCM is also
970 known to underestimate the vertical extension of the dust and in particular the forma-
971 tion of detached layers in the atmosphere (see e.g. Wang et al., 2018), and efforts are
972 ongoing to better represent the dust cycle (alongside the water cycle) in the GCM. As-
973 the absorption of heat by dust is one of the primary sources of thermal tides, this should
974 also affect all modeled tides and in particular the partitioning between diurnal and semid-
975 iurnal tides which is controlled by the depth of the heat absorption.

976 Recent successes at Earth in terms of high-resolution whole-atmosphere models that
977 enable resolution of gravity waves and their effects (e.g., Liu et al., 2014), and that as-
978 similate observational data (e.g., Pedatella et al., 2014), may point the way for future
979 Mars model developments. Indeed the present paper suggests that assimilation of MCS-
980 like middle-atmosphere measurements into a whole-atmosphere Mars model could rep-
981 resent a leap forward in our ability to predict thermosphere longitude structures due to
982 solar tides. While significant advances have been made insofar as data assimilation of
983 low- and middle-atmosphere data goes (Navarro et al., 2017; Graybush et al., 2019), it
984 remains to be determined how thermosphere structures are affected. Concerning the need
985 for future observations, major advances would likely result from improved local time sam-
986 pling of Mars atmosphere. This could be accomplished by more than one satellite pre-
987 cessing in local time, or 2-4 satellites in near-polar orbits sampling 4-8 local times simul-
988 taneously (see, e.g., Kleinböhl, et al., 2018). Small satellites and new technologies now
989 make these scenarios feasible.

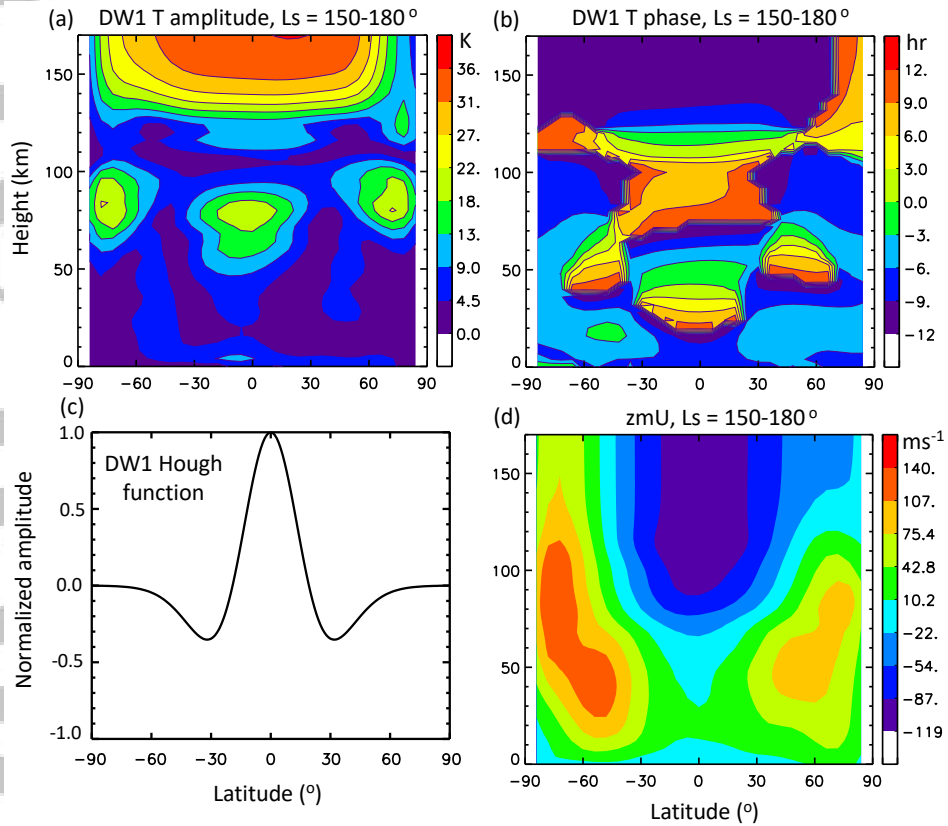


990 **Figure 1.** Typical examples of (a) latitude-local time data distributions, and (b)-(f) tidal
 991 fits. In (b)-(f) dotted lines are diurnal fits, dashed lines are semidiurnal fits, and solid lines are
 992 combined diurnal and semidiurnal fits. The data points in the fits are weighted according to the
 993 cube root of the number of points in the corresponding local time bin at a given latitude and
 994 longitude. The data points without error bars are single data points.

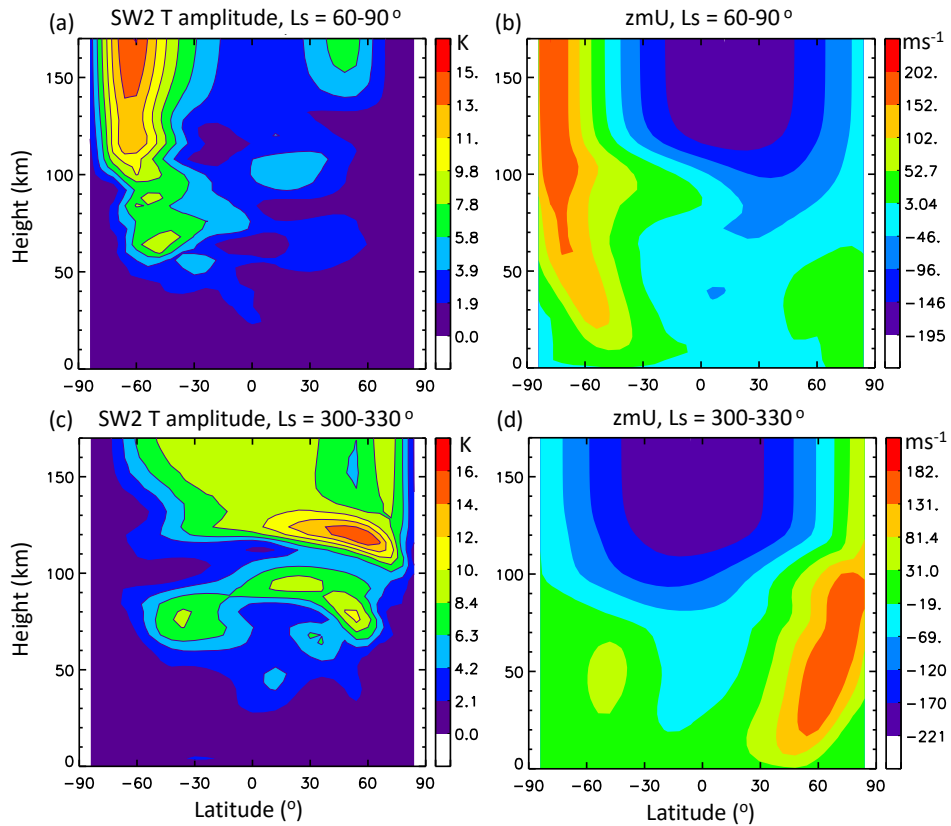


995 **Figure 2.** (a) and (b): Latitude-Ls variability of MCS DW1 and SW2 temperature amplitudes
 996 at 76 km above Mars' areoid. (c) and (d): Same as (a) and (b) except for MCD. (e) Latitude-
 997 Ls variation of zonal-mean zonal winds from MCD at 76 km. (f) Latitude-Ls variation of SW2
 998 temperature amplitudes at 172 km for $L_s = 150-180^\circ$.

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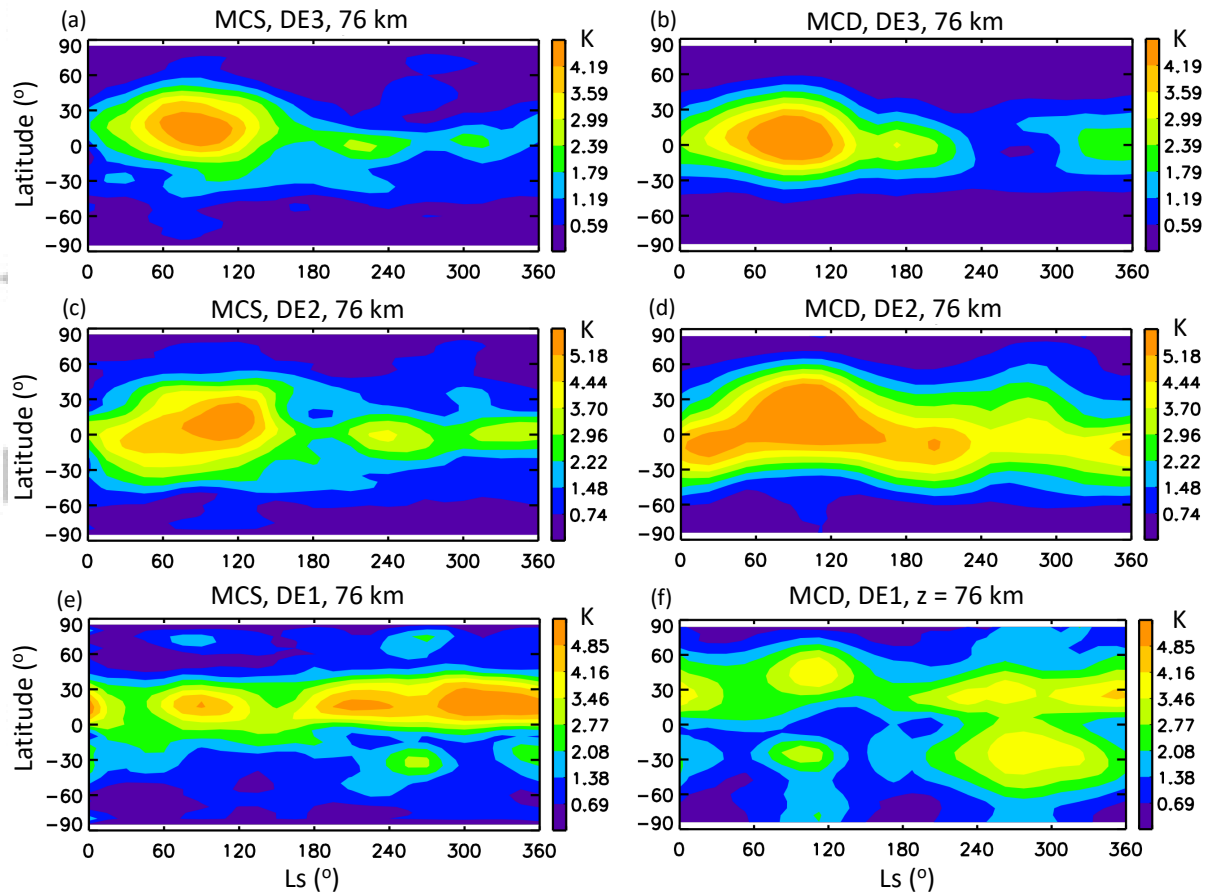


999 **Figure 3.** (a) and (b): Height-latitude variations of MCD DW1 temperature amplitude and
 1000 phase (c) Latitude dependence of DW1 Hough function from classical tidal theory. (d) Height-
 1001 latitude variation of MCD zonal mean zonal wind for $L_s = 150-180^\circ$.



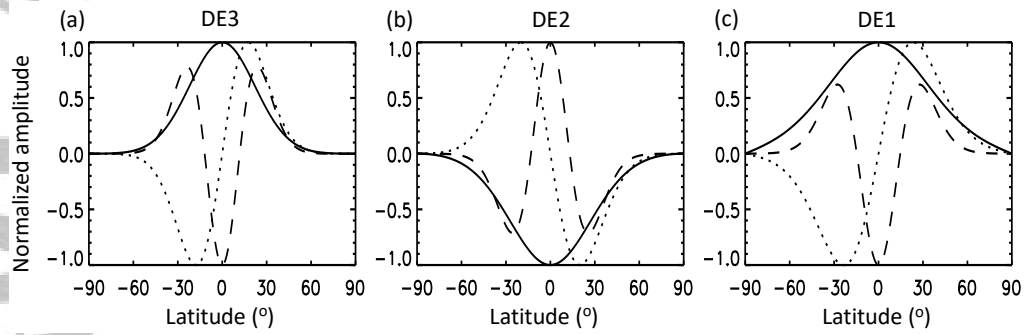
1002 **Figure 4.** (a) and (b): Height-latitude variation of MCD SW2 temperature amplitude and
 1003 MCD zonal-mean zonal wind, for Ls = 60-90°. (c) and (d): Same as (a) and (b) except for Ls =
 1004 300-330°.

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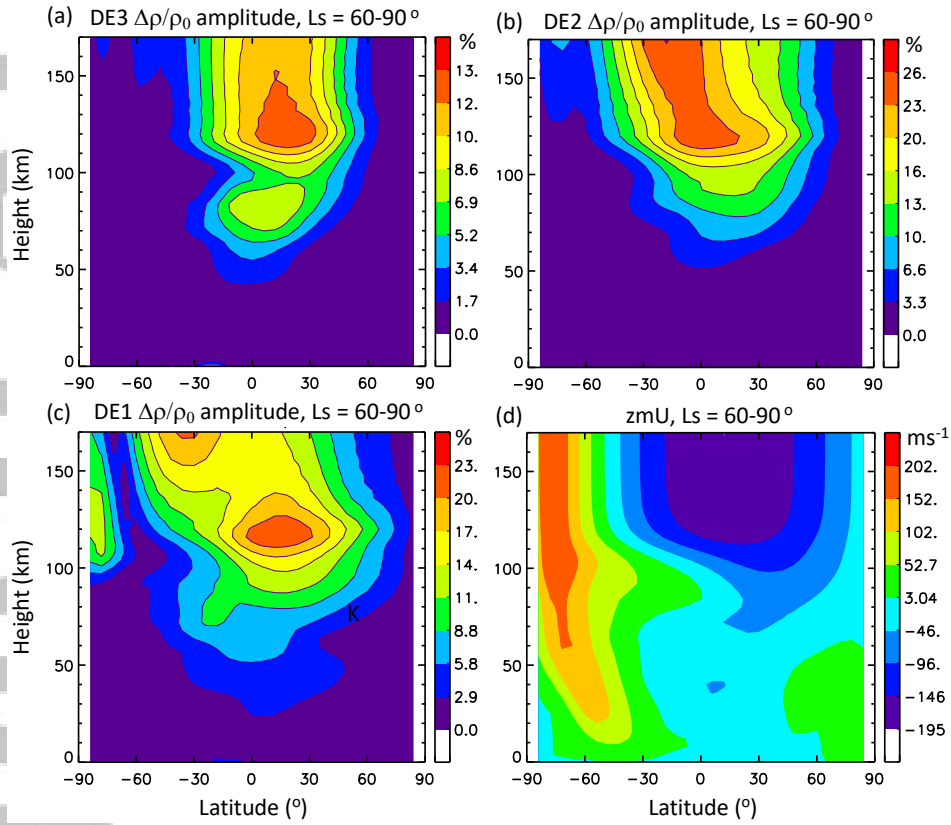


1005 **Figure 5.** (a) and (b): Latitude-Ls variations of MCS and MCD DE3 temperature ampli-
 1006 tudes at 76 km above Mars' areoid. (c) and (d): Same as (a) and (b) except for DE2. (e) and (f):
 1007 Same as (a) and (b) except for DE1.

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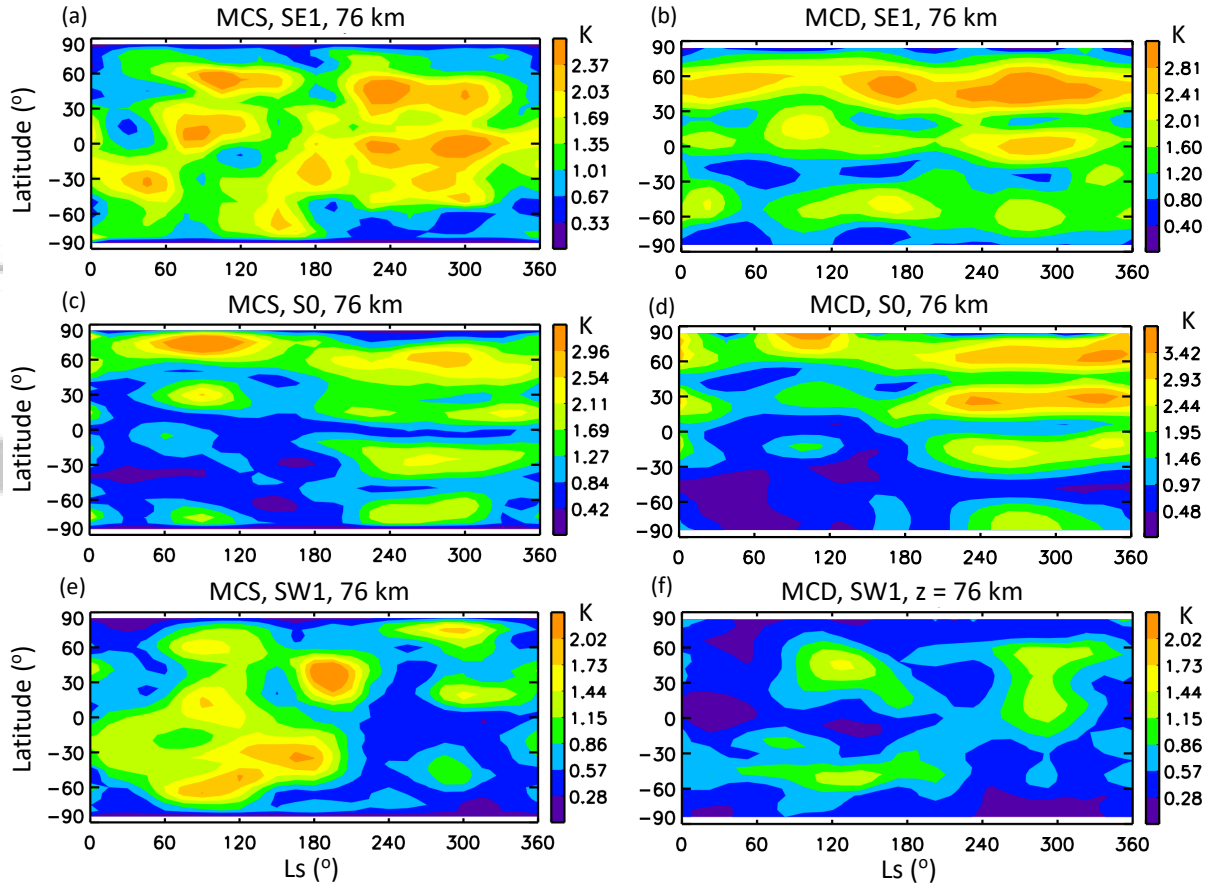


1008 **Figure 6.** First symmetric (solid lines), first antisymmetric (dotted lines), and second sym-
 1009 metric (dashed lines) Hough functions for (a) DE3, (b) DE2, and (c) DE1.



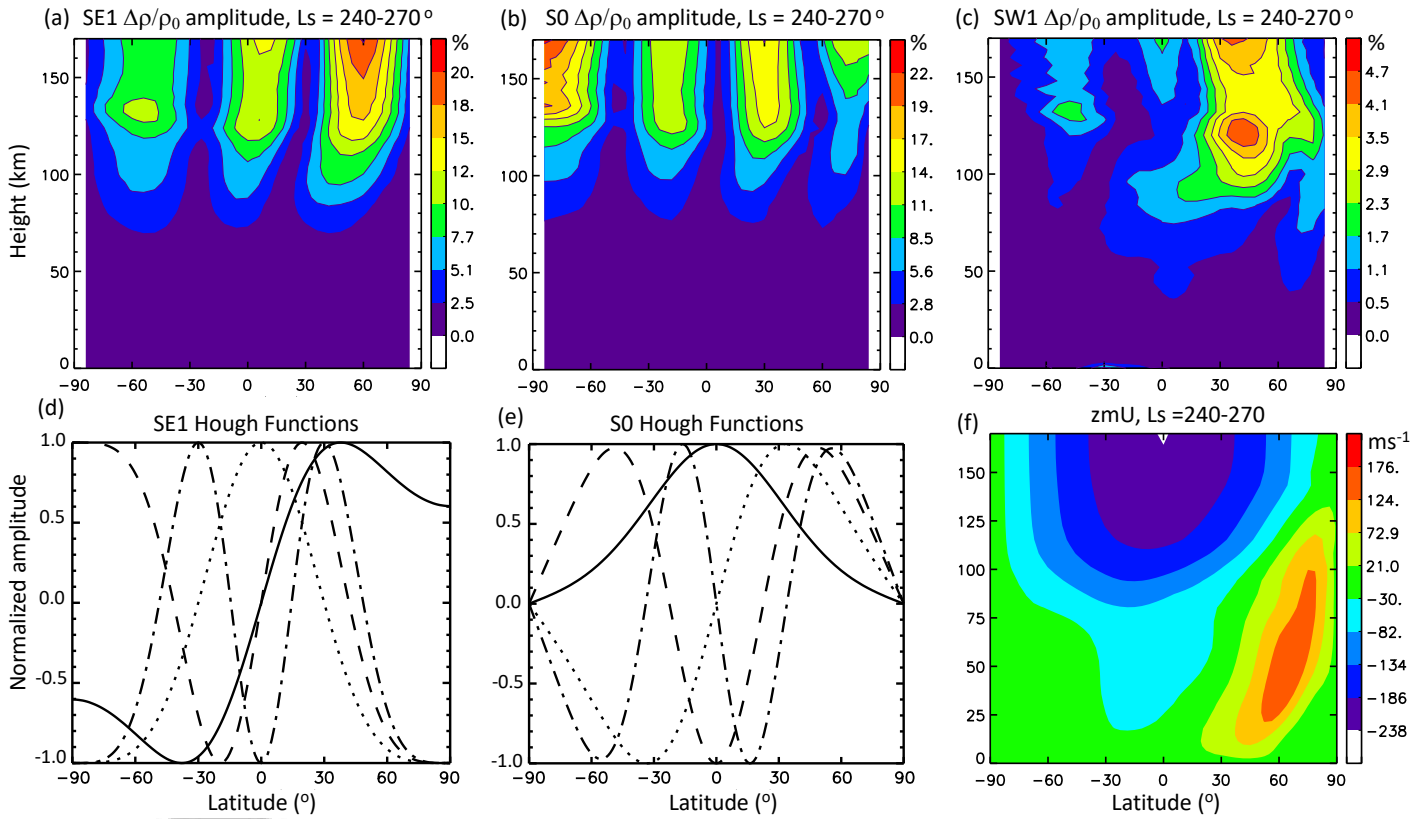
1010 **Figure 7.** Height-latitude structures of MCD % relative density perturbations for $L_s = 60-90^\circ$
 1011 for (a) DE3, (b) DE2, and (c) DE1. Zonal-mean zonal winds are shown in panel (d).

icle

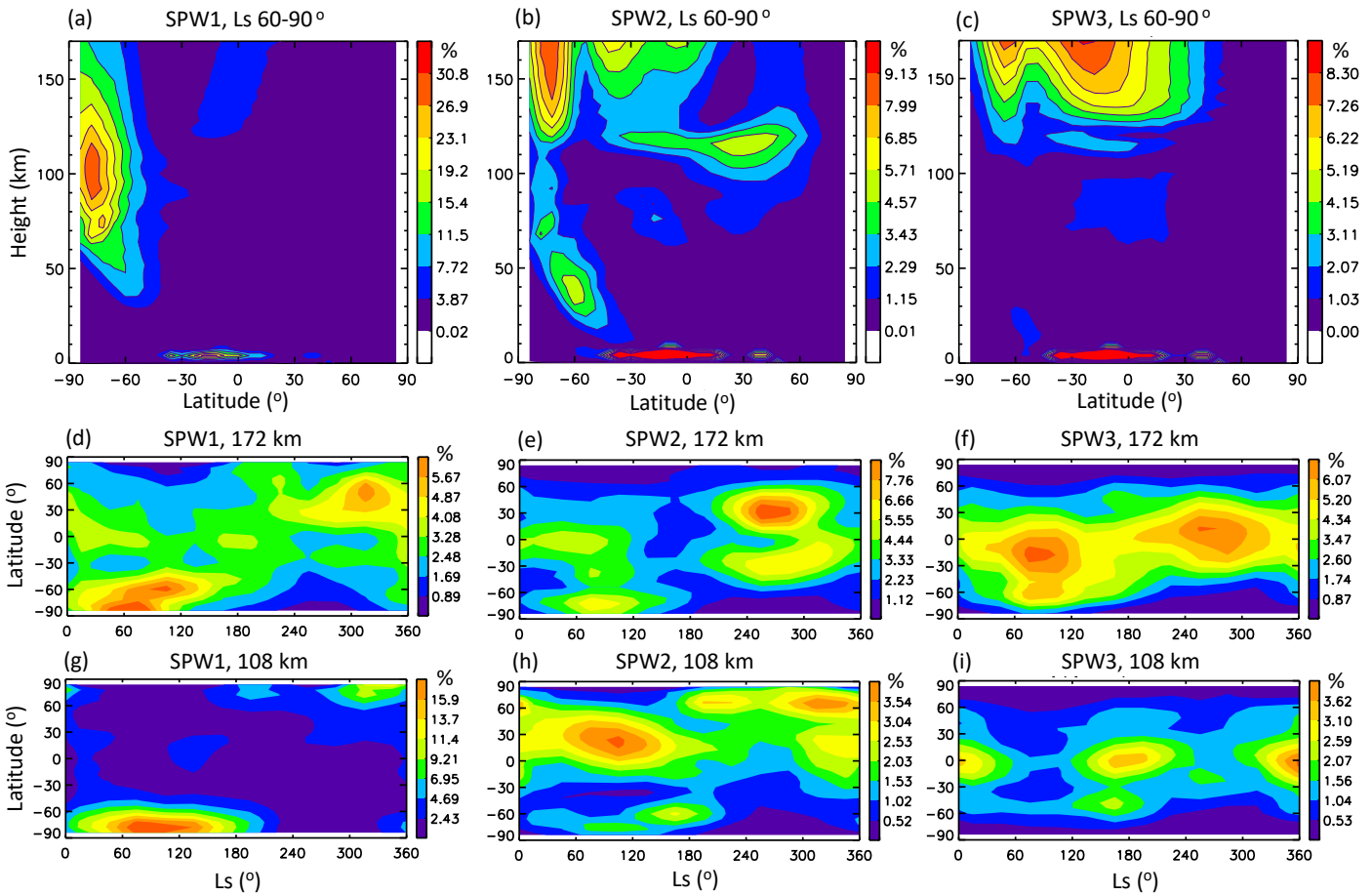


1012 **Figure 8.** (a) and (b): Latitude-Ls variations of MCS and MCD SE1 temperature amplitudes
 1013 at 76 km above Mars' areoid. (c) and (d): Same as (a) and (b) except for S0. (e) and (f): Same
 1014 as (a) and (b) except for SW1.

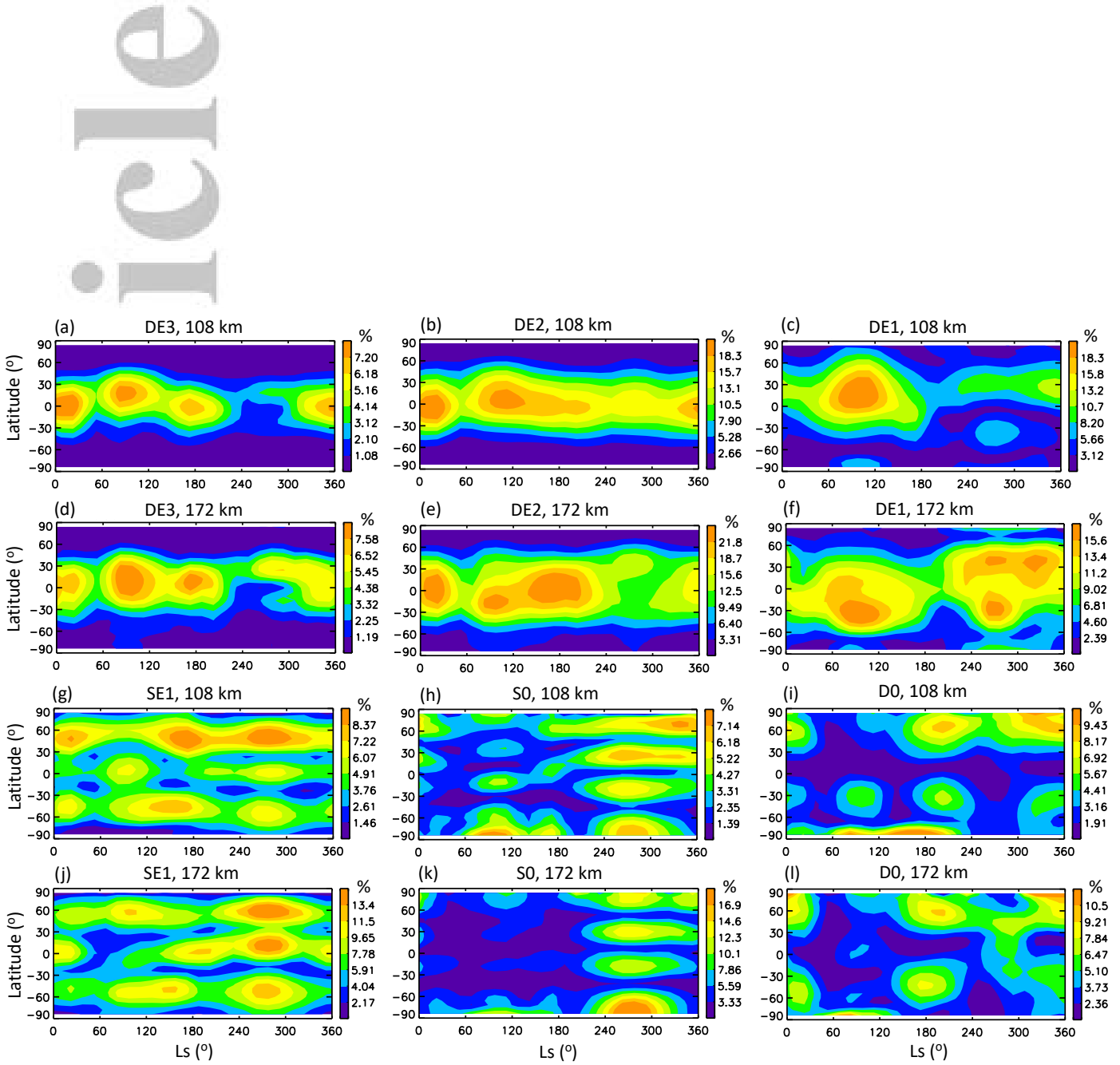
Accel



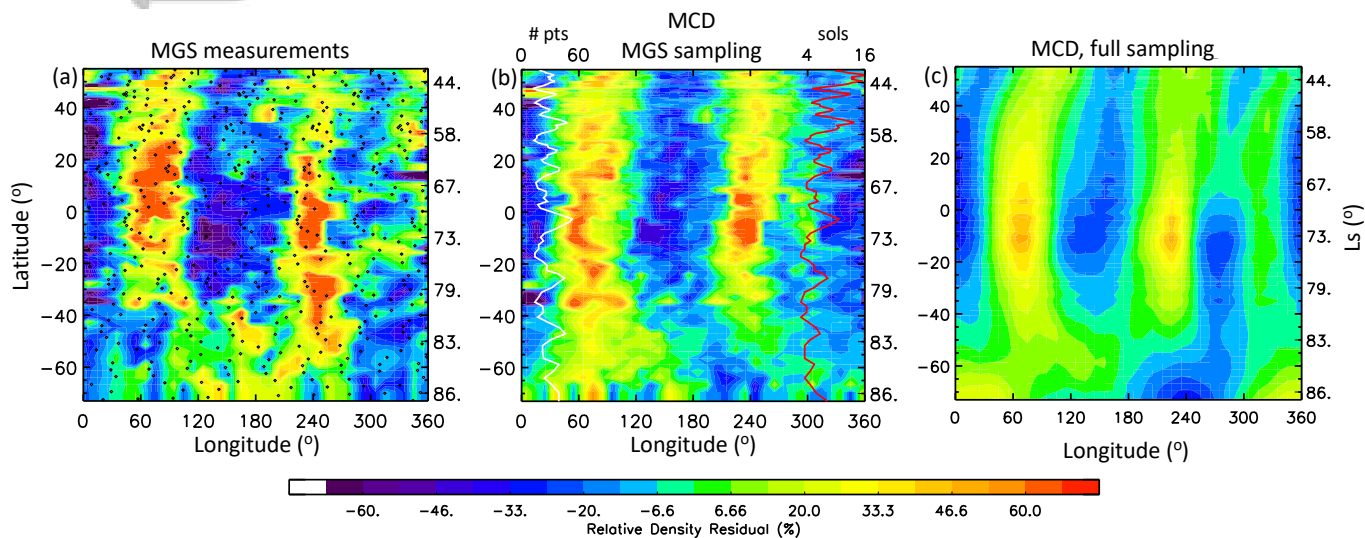
1015 **Figure 9.** Height-latitude structures of MCD % relative density perturbations for Ls = 240-
 1016 270° for (a) SE1, (b) S0, and (c) SW1. Zonal-mean zonal winds are shown in panel (f). First four
 1017 SE1 and S0 Hough functions are shown in panels (d) and (e): first symmetric (solid lines), first
 1018 antisymmetric (dotted lines), second symmetric (dashed lines), second antisymmetric (dashed-
 1019 dotted lines).



1020 **Figure 10.** Sample depictions of MCD % relative density perturbations for SPW1 (left),
 1021 SPW2 (middle), and SPW3 (right). (a), (b) and (c): height-latitude structures for $L_s = 60-90^\circ$.
 1022 (d), (e) and (f): latitude- L_s structures at 172 km. (g), (h) and (i): latitude- L_s structures at 108
 1023 km.

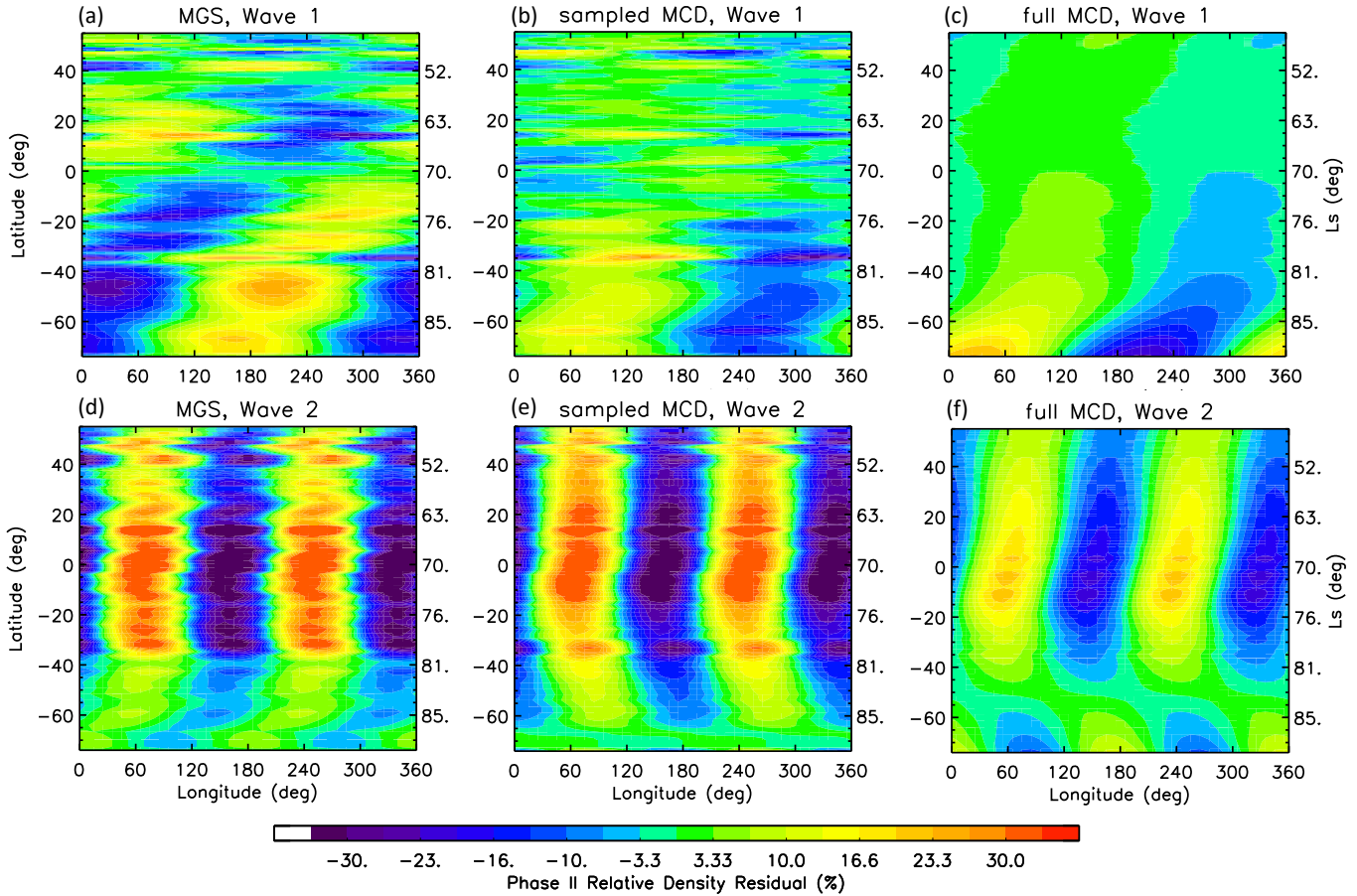


1024 **Figure 11.** Latitude-Ls distributions of MCD % relative density perturbations. (a), (b) and
 1025 (c): DE3, DE2, and DE1 at 108 km. (d), (e) and (f): DE3, DE2, and DE1 at 172 km. (g), (h)
 1026 and (i): SE1, S0 and D0 at 108 km. (j), (k) and (l): SE1, S0 and D0 at 172 km.



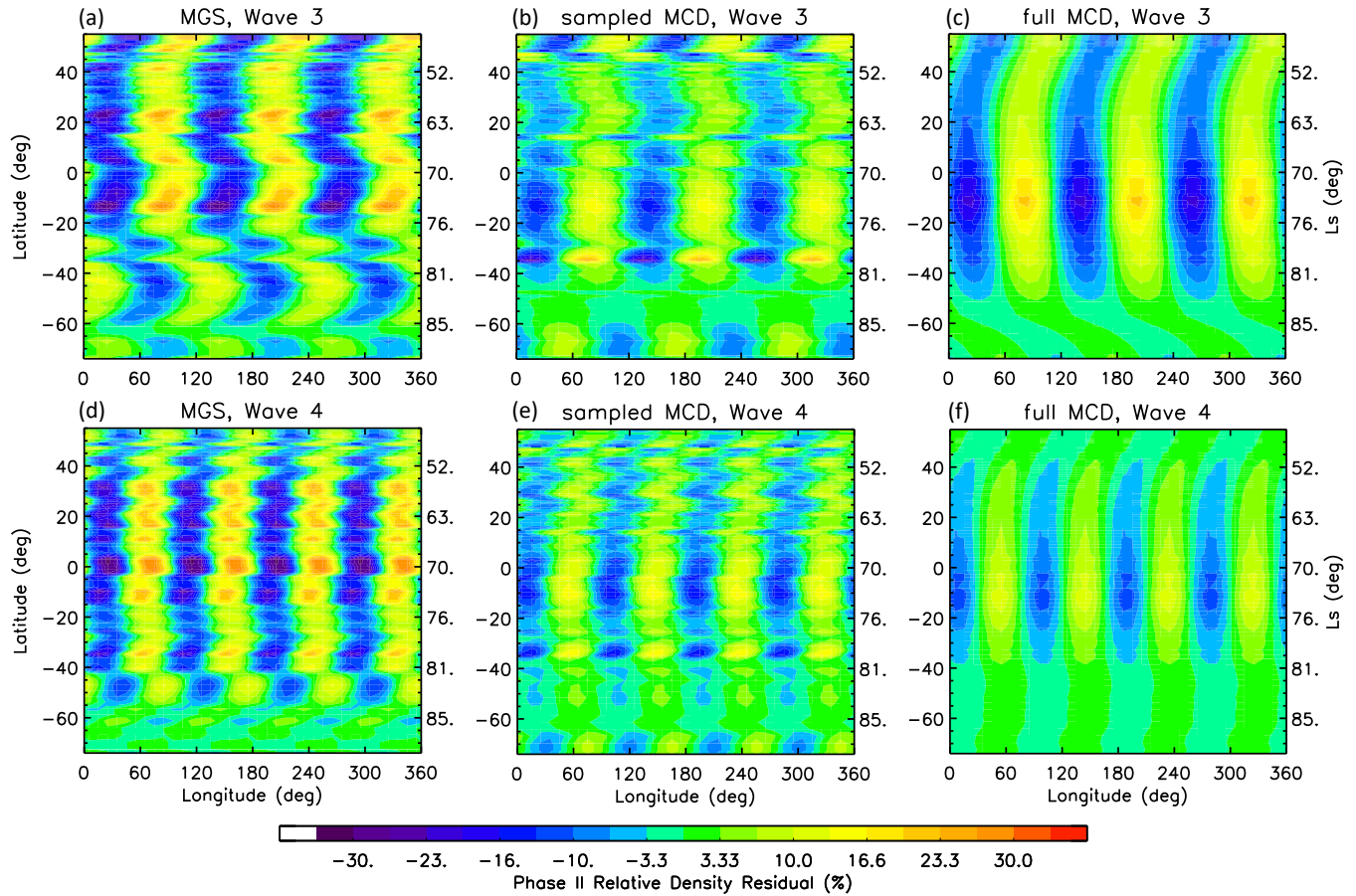
1027 **Figure 12.** Comparisons between % density perturbations measured at periapsis points be-
 1028 tween 115 km (55°N) and 105 km (70°S) by the MGS accelerometer, with those residing in the
 1029 MCD model. (a): Raw MGS measurements, with dots indicating periapsis measurement points.
 1030 (b): MCD model densities, obtained by sampling in height, latitude, longitude, LST and Ls
 1031 according to MGS. (c): Same as (b), except from full sampling of the MCD model output. In
 1032 Figure 12(b), the numbers of points per fitting window, and the lengths of the fitting windows in
 1033 sols, are indicated by white and red lines, respectively, with scales at the top of the figure panel.
 1034 The window is moved once per sol through the data.

Article



1035 **Figure 13.** Same as Figure 12, except for longitudinal wave-1 (top row) and wave-2 (bottom
 1036 row).

Accepted



1037 **Figure 14.** Same as Figure 12, except for longitudinal wave-3 (top row) and wave-4 (bottom
1038 row).

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1284 **Table 1.** Diurnal and semidiurnal vertical wavelengths (km) from classical tidal theory based
 1285 on Equation (1) in the text. D1S denotes “diurnal 1st symmetric”; S1S denotes “semidiurnal 1st
 1286 symmetric”; D1A denotes “diurnal 1st antisymmetric” and so on; * denotes < 30 km; - denotes >
 1287 200 km.

s	D1S	S1S	D1A	S1A	D2S	S2S	D2A	S2A
-3	64	-	34	103	*	66	*	50
-2	122	-	43	207	*	87	*	59
-1	-	-	75	-	33	134	*	74
0	30	-	116	-	*	63	*	99
1	32	-	*	145	*	76	*	54
2	31	-	*	93	*	61	*	47

Figure 1.

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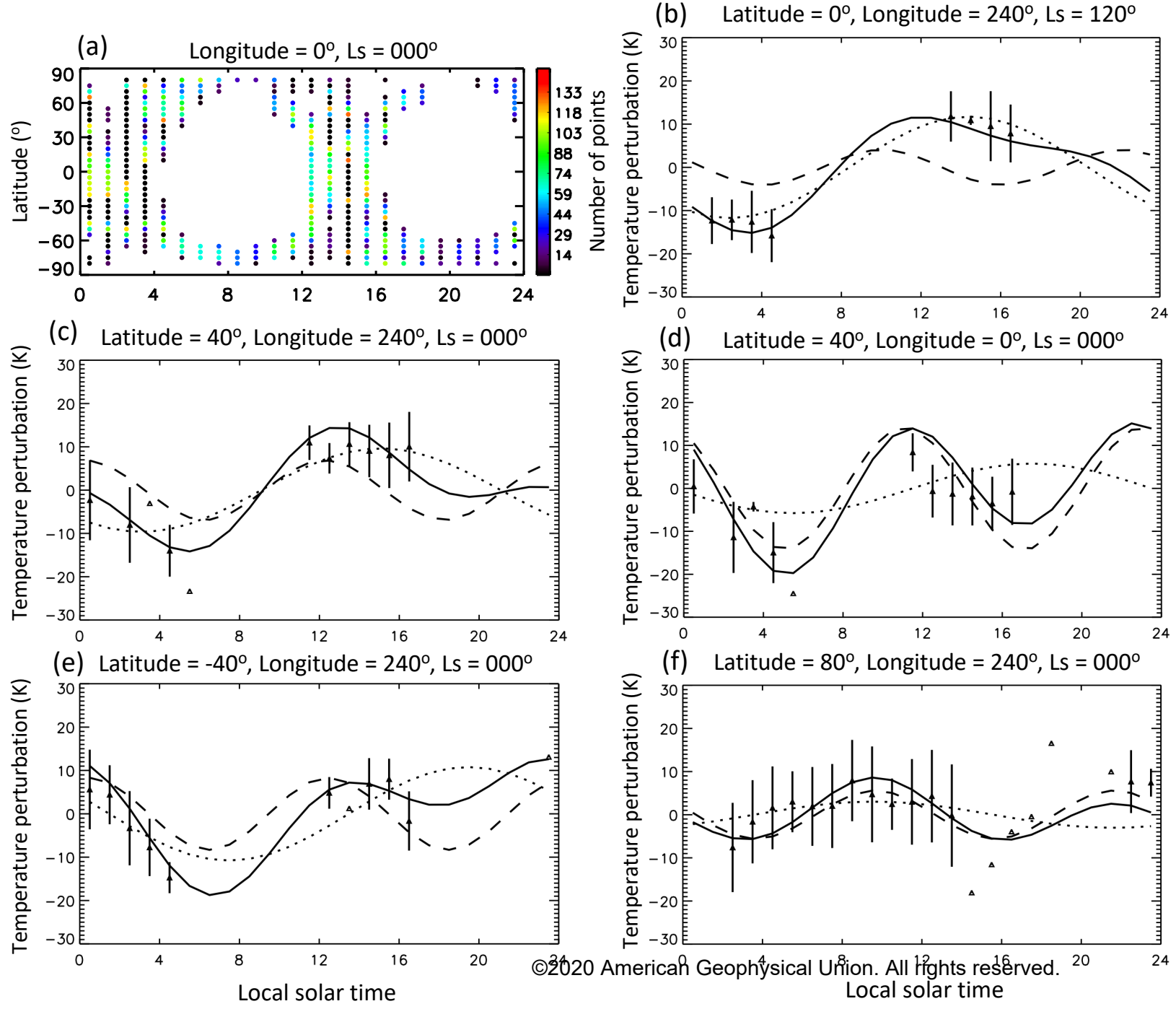


Figure 2.

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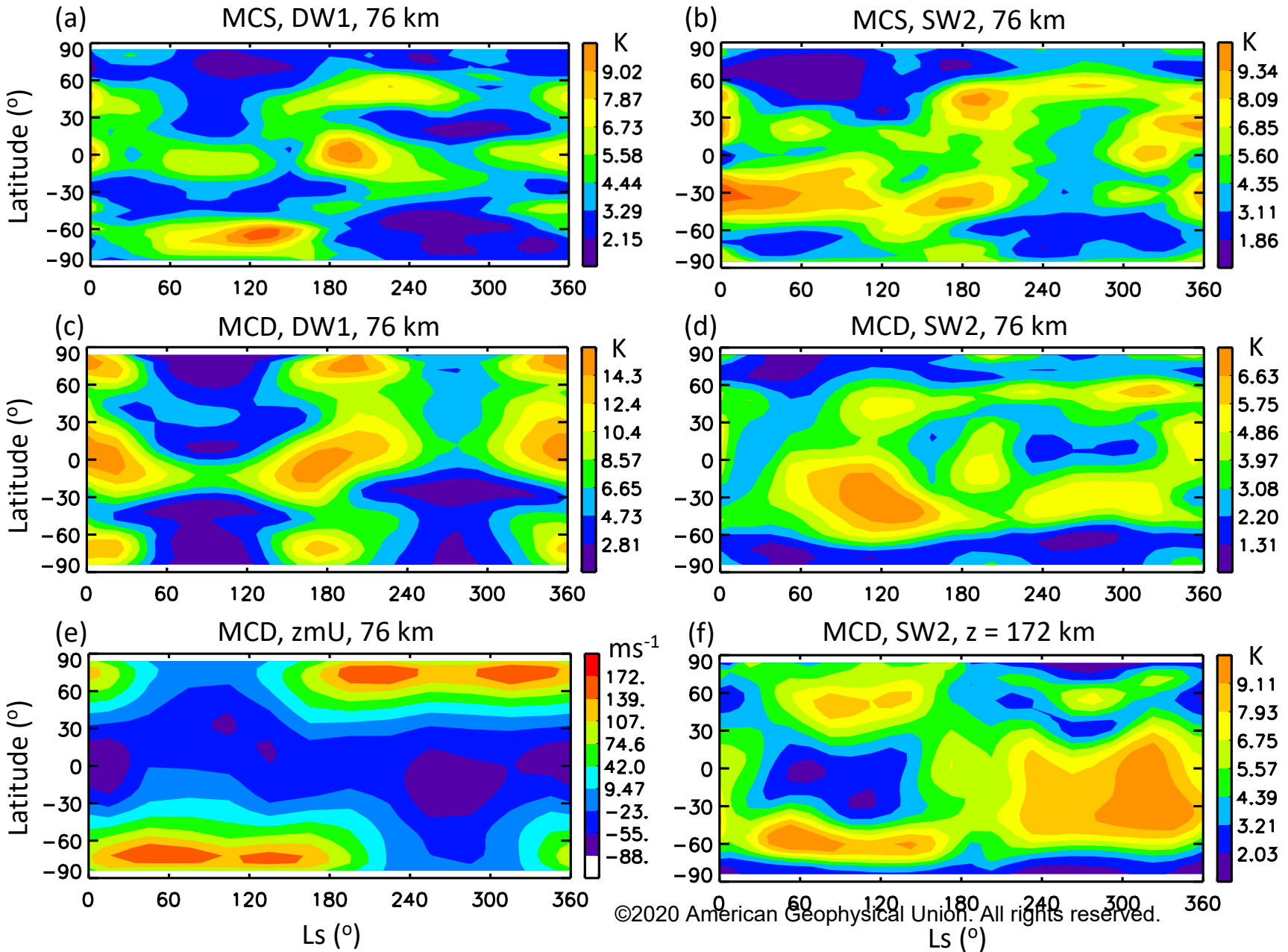
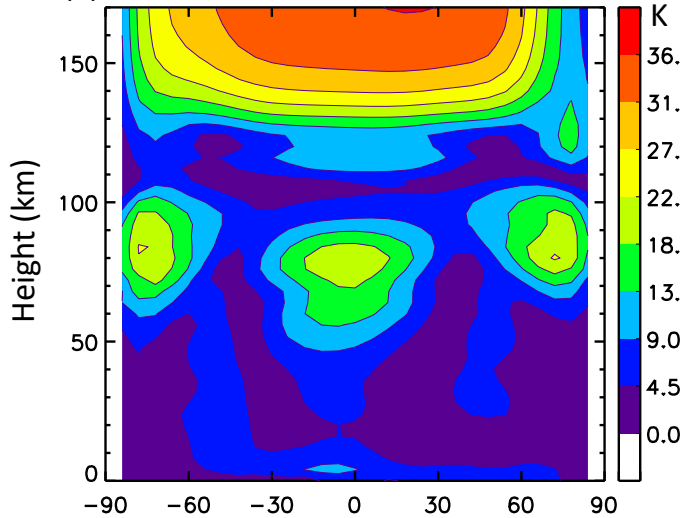


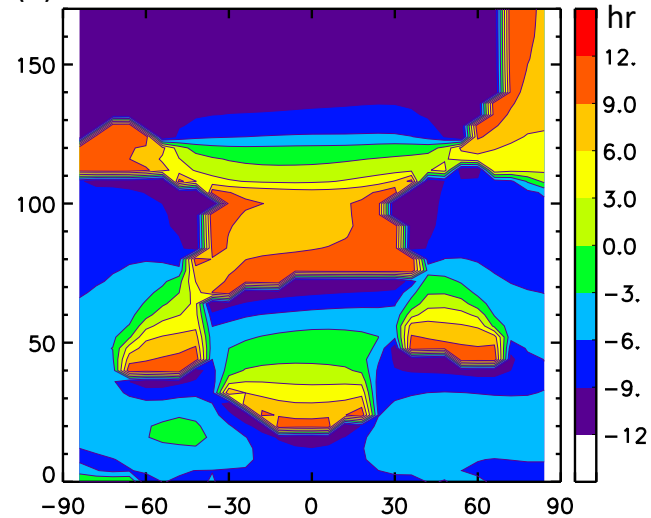
Figure 3.

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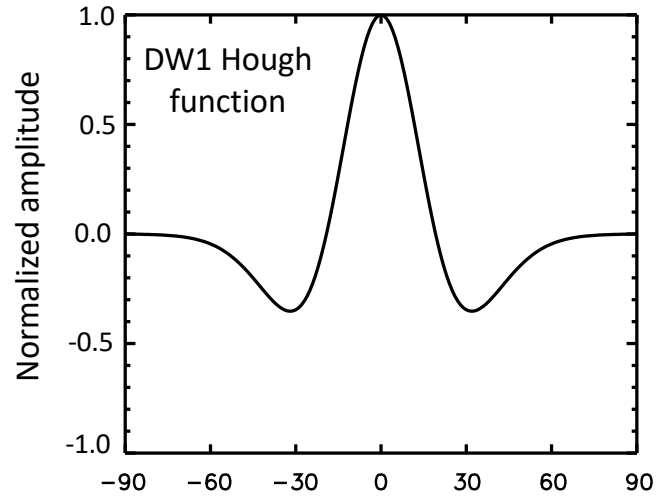
(a) DW1 T amplitude, Ls = 150-180°



(b) DW1 T phase, Ls = 150-180°



(c) DW1 Hough function



(d) zmU, Ls = 150-180°

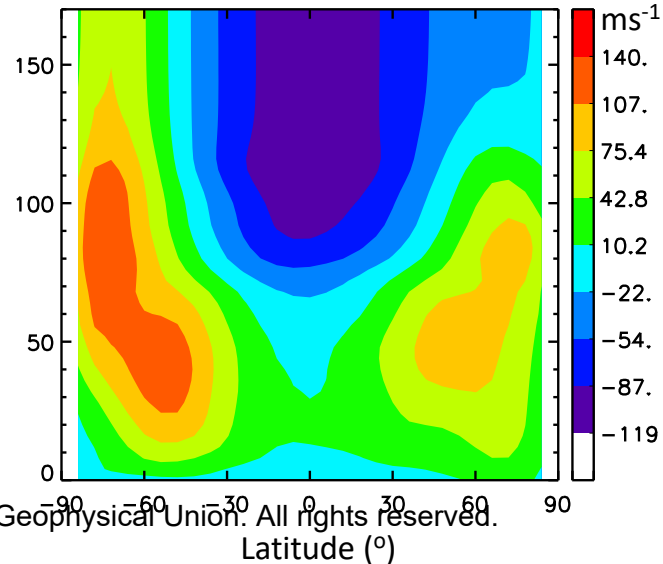
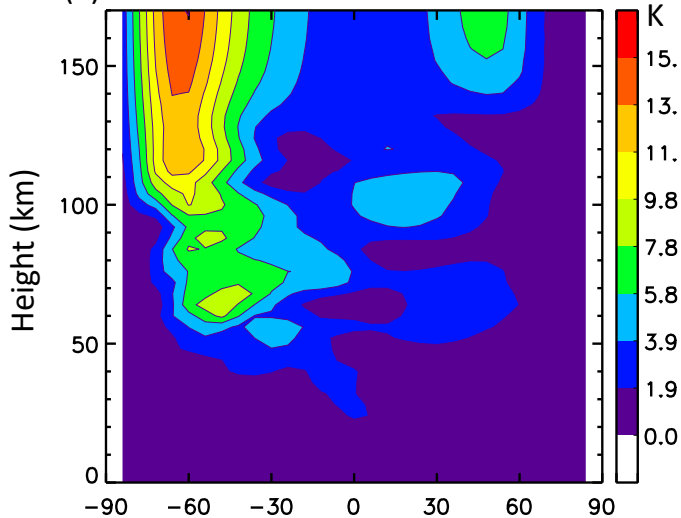


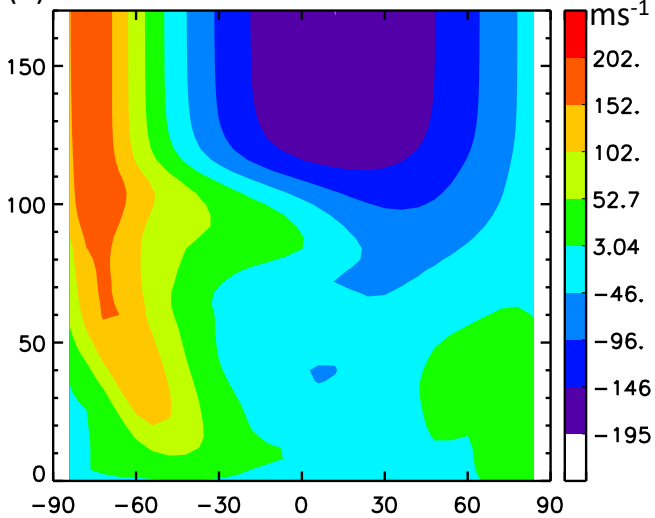
Figure 4.

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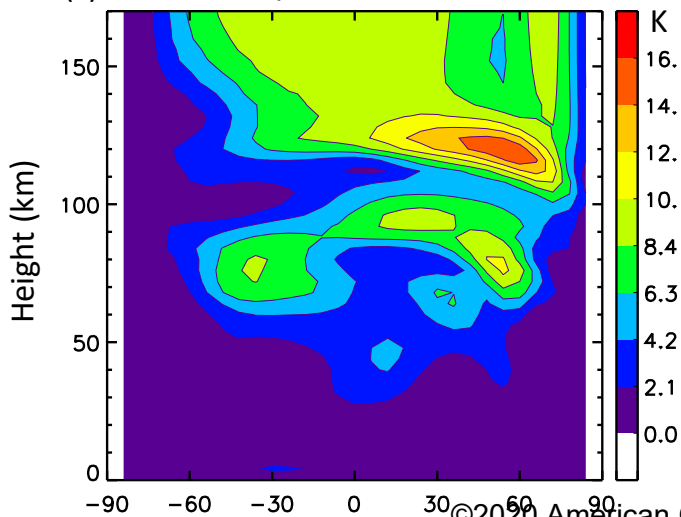
(a) SW2 T amplitude, Ls = 60-90°



(b) zmU, Ls = 60-90°



(c) SW2 T amplitude, Ls = 300-330°



(d) zmU, Ls = 300-330°

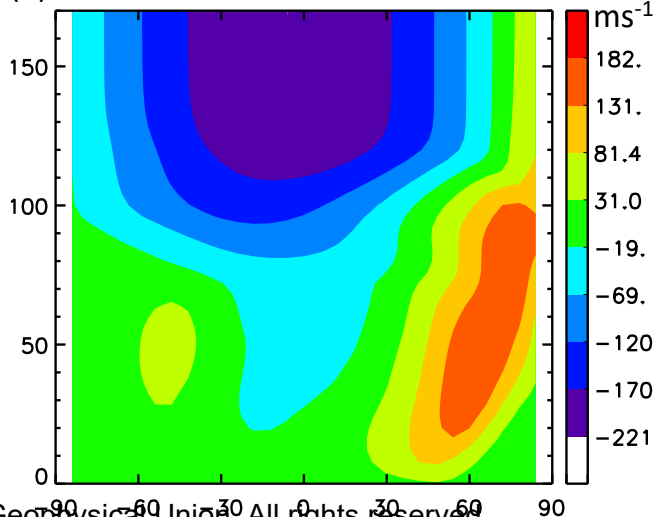


Figure 5.

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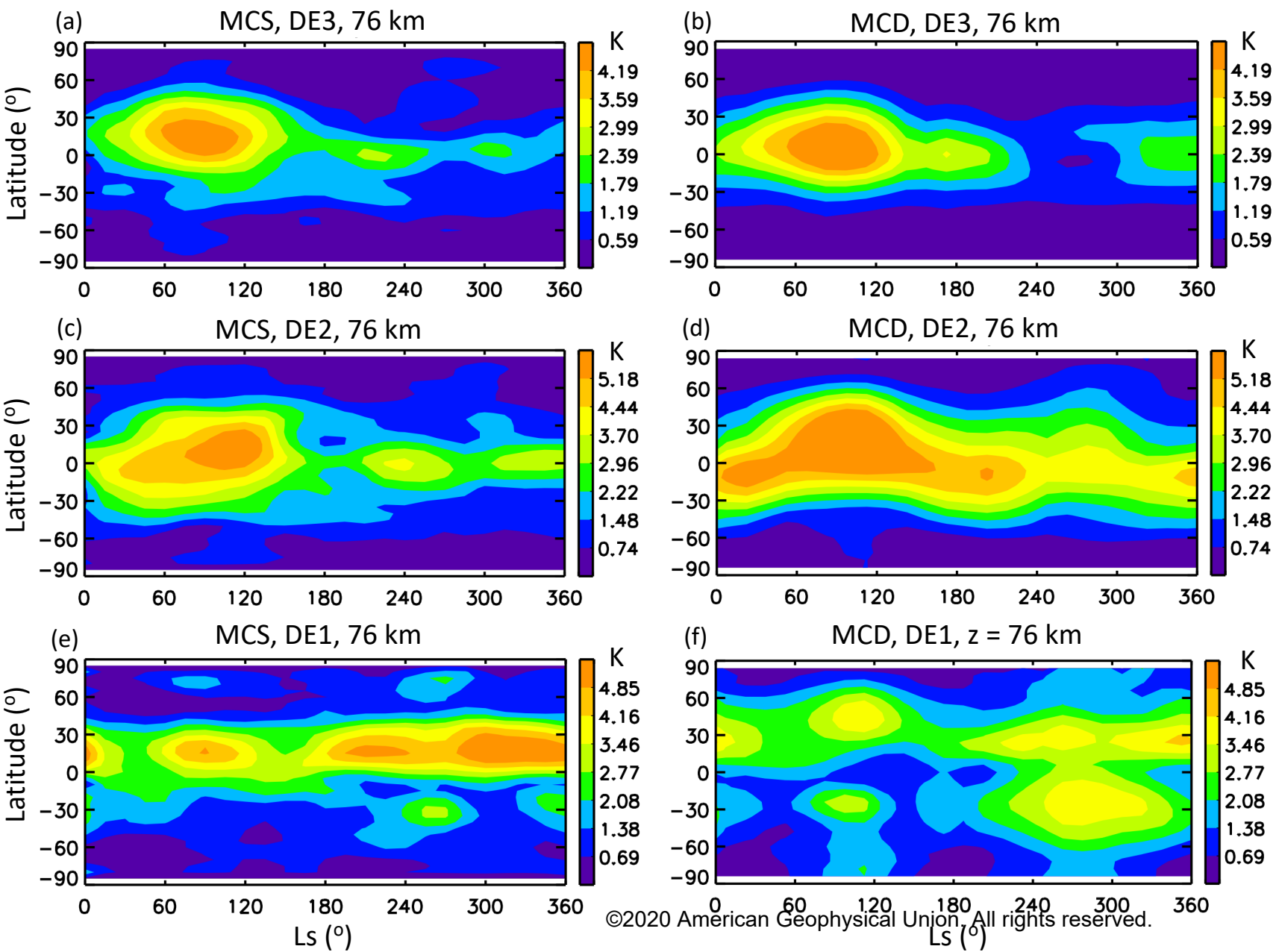


Figure 6.

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Normalized amplitude

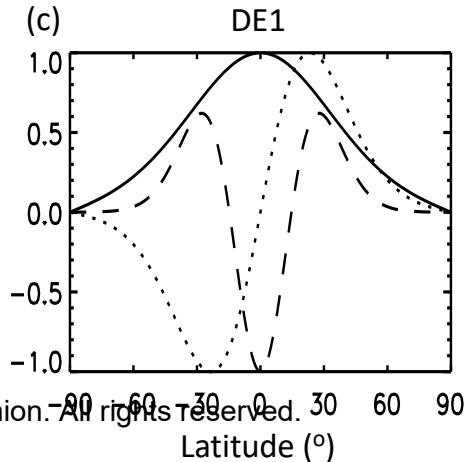
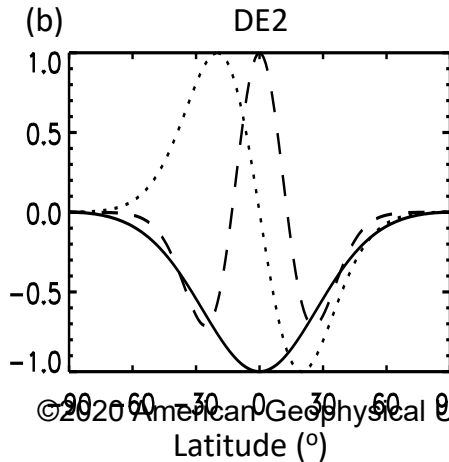
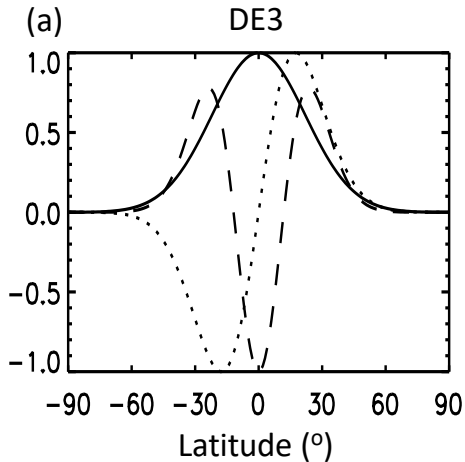
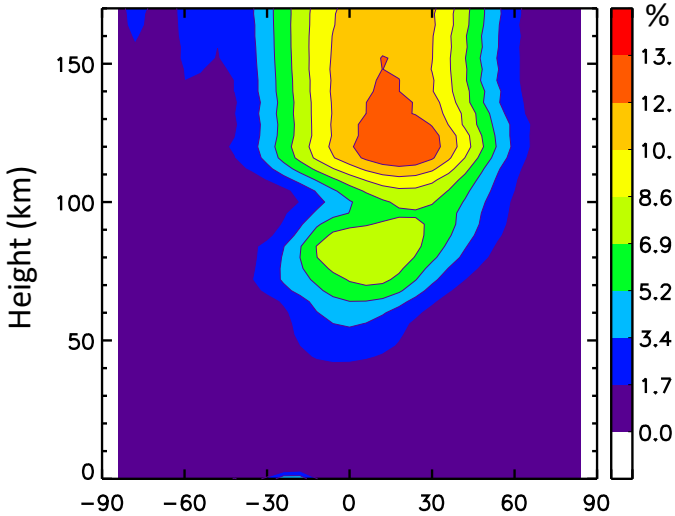


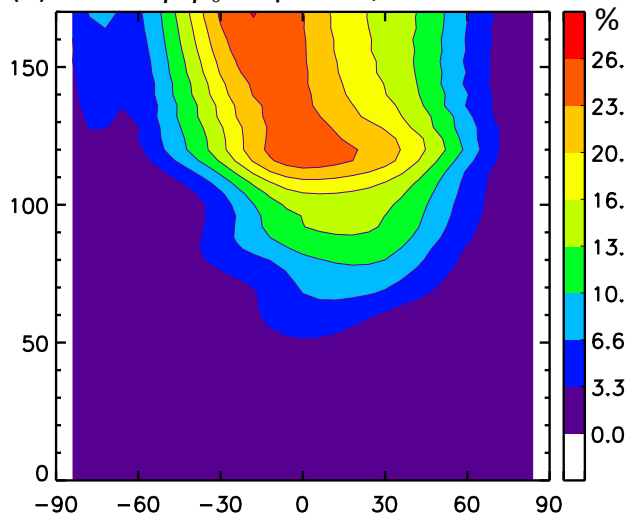
Figure 7.

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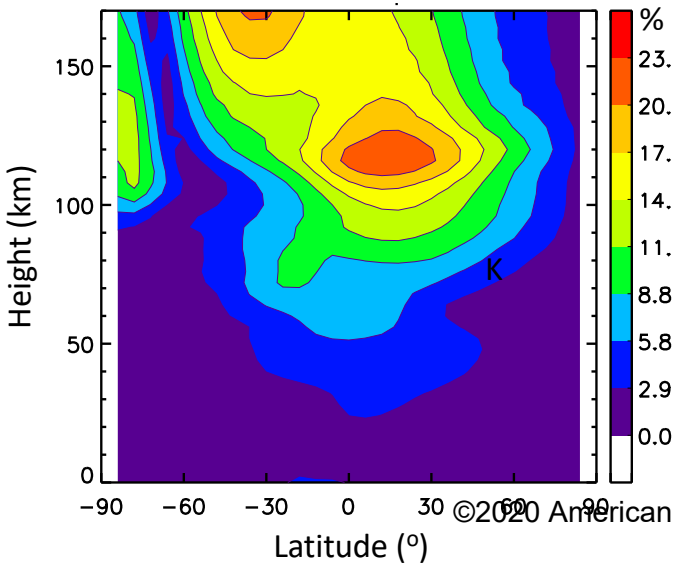
(a) DE3 $\Delta\rho/\rho_0$ amplitude, Ls = 60-90°



(b) DE2 $\Delta\rho/\rho_0$ amplitude, Ls = 60-90°



(c) DE1 $\Delta\rho/\rho_0$ amplitude, Ls = 60-90°



(d) zmU, Ls = 60-90°

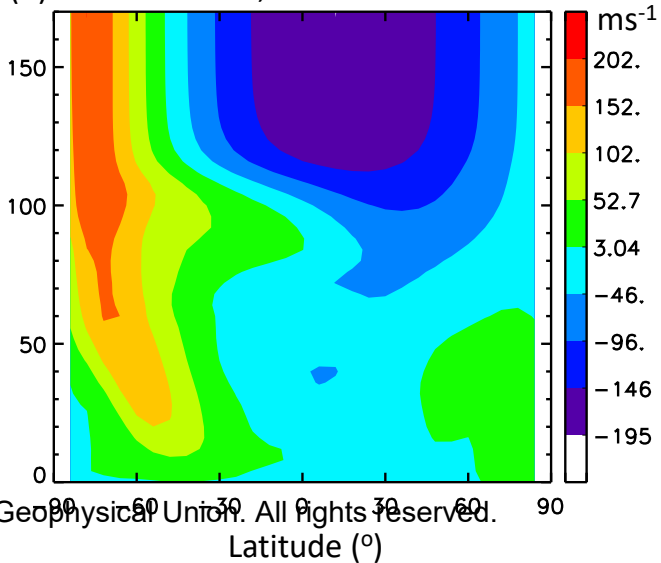


Figure 8.

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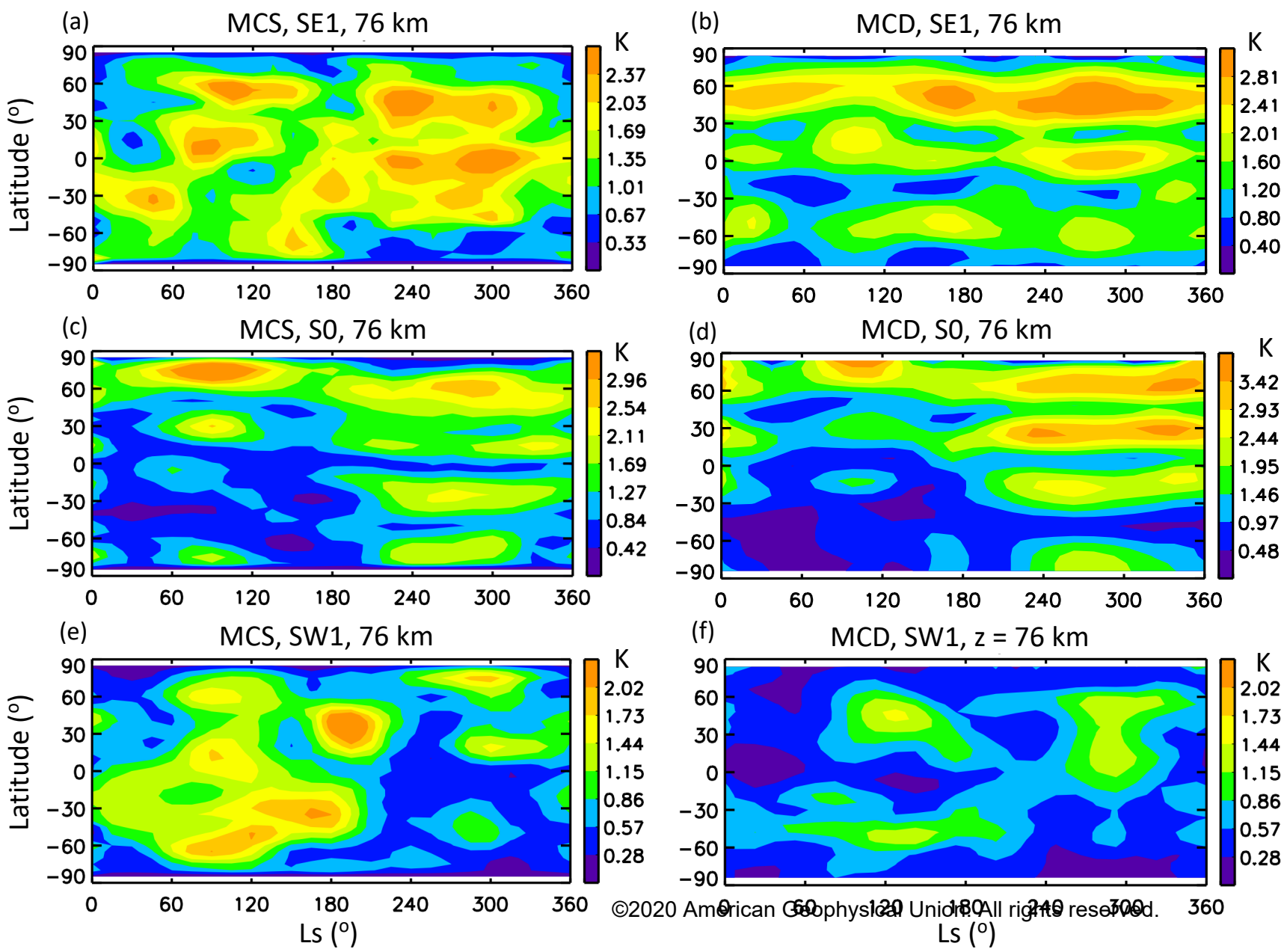
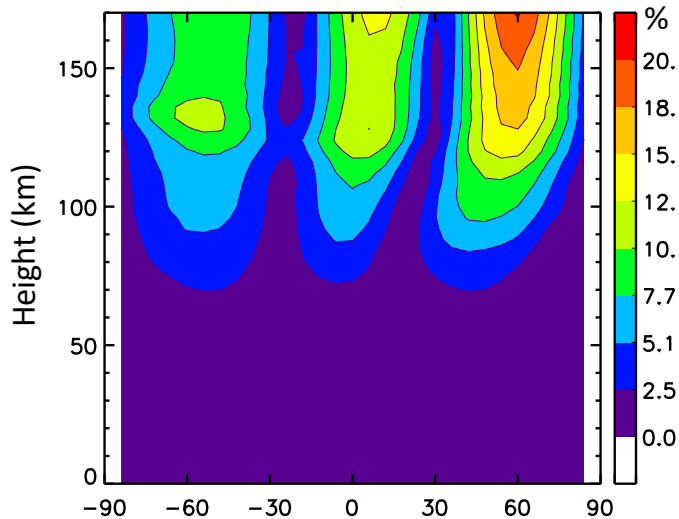


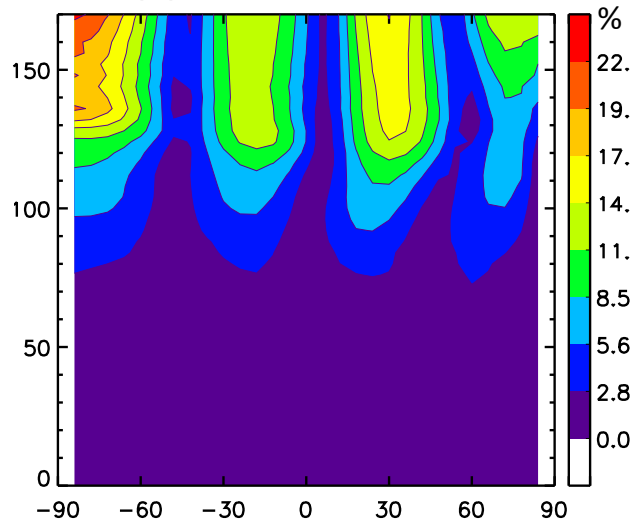
Figure 9.

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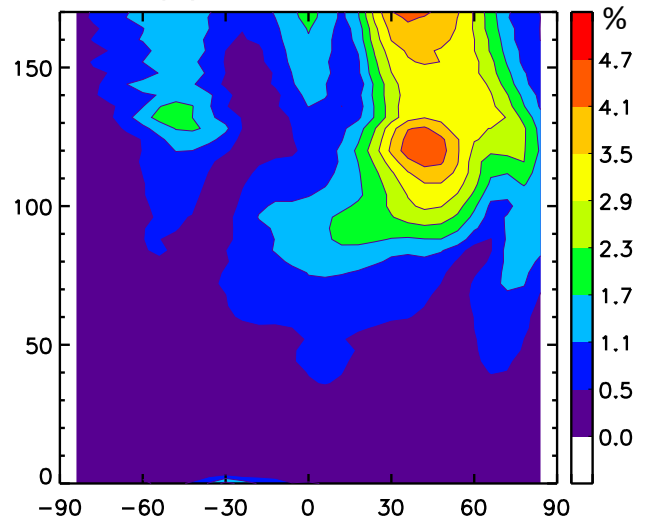
(a) SE1 $\Delta\rho/\rho_0$ amplitude, Ls = 240-270°



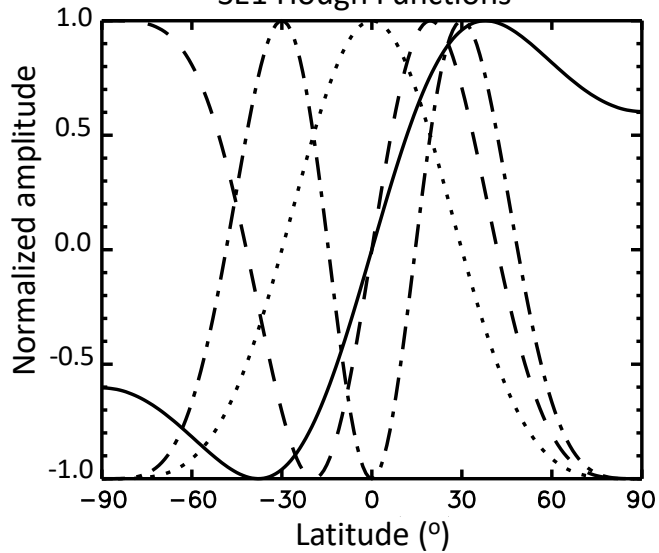
(b) S0 $\Delta\rho/\rho_0$ amplitude, Ls = 240-270°



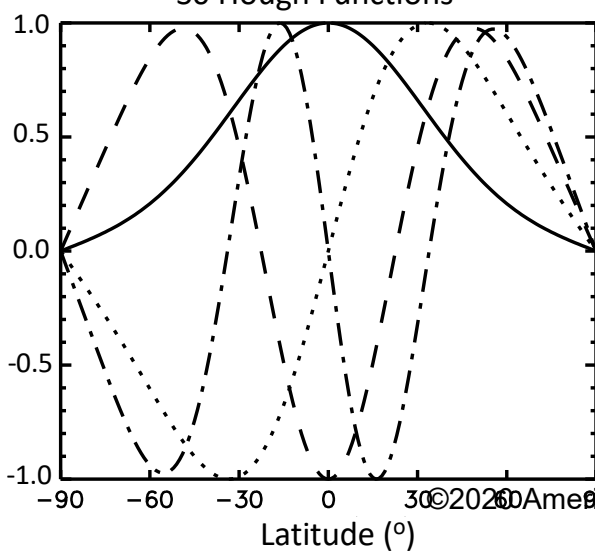
(c) SW1 $\Delta\rho/\rho_0$ amplitude, Ls = 240-270°



(d) SE1 Hough Functions



(e) S0 Hough Functions



(f) zmU, Ls = 240-270

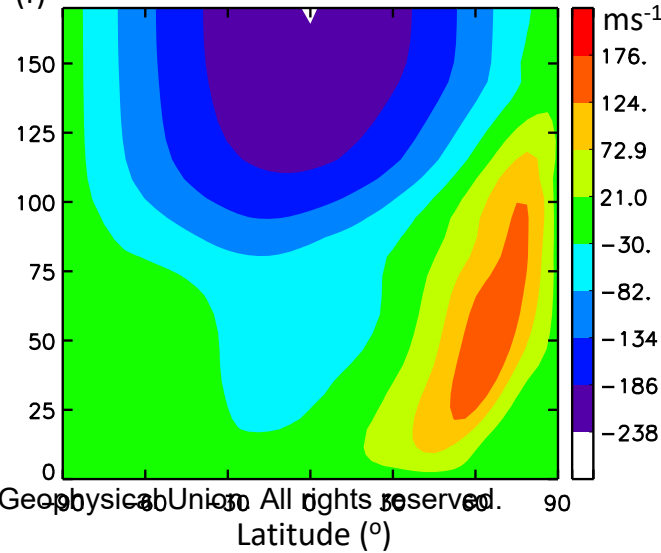


Figure 10.

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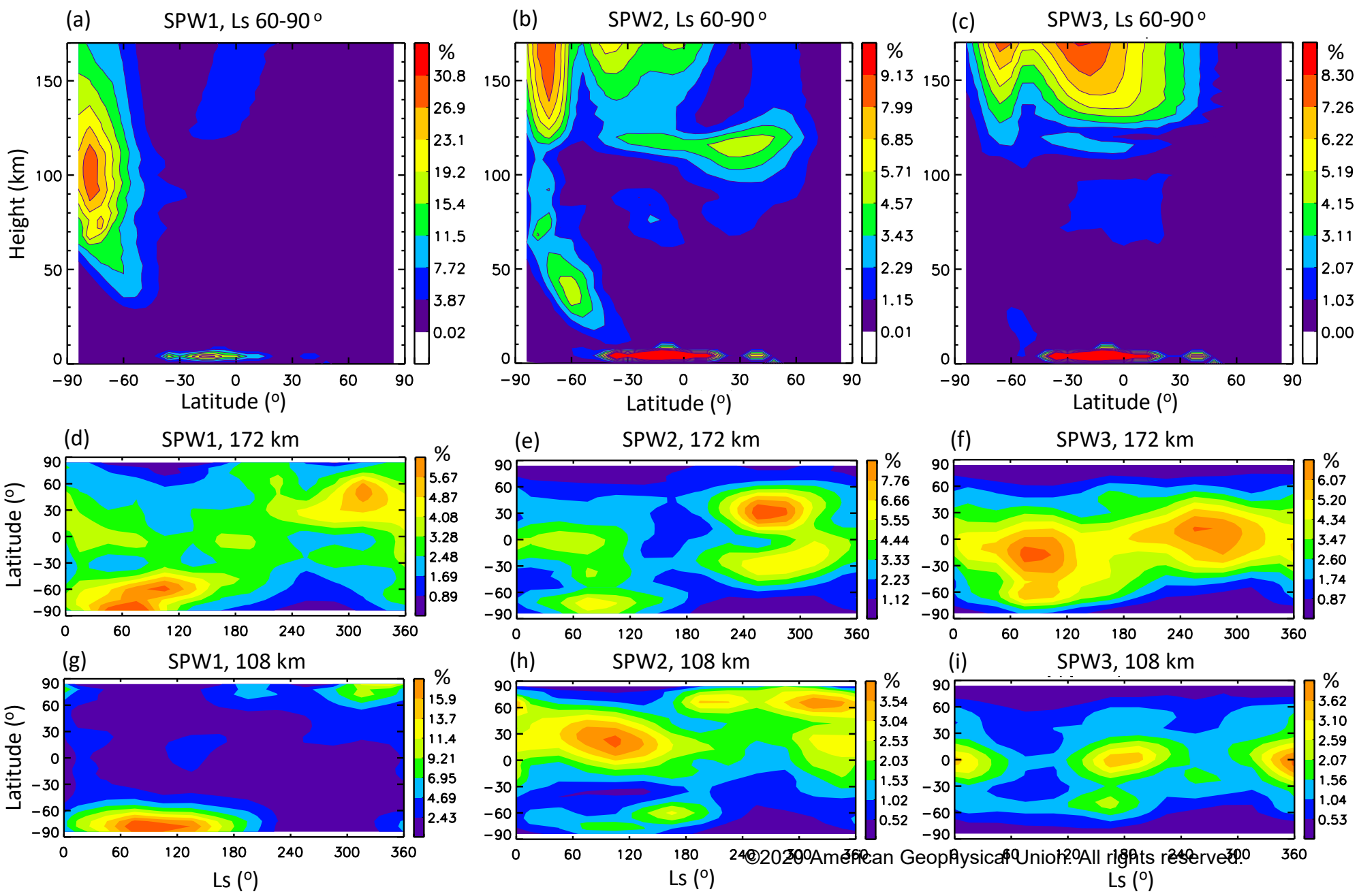


Figure 11.

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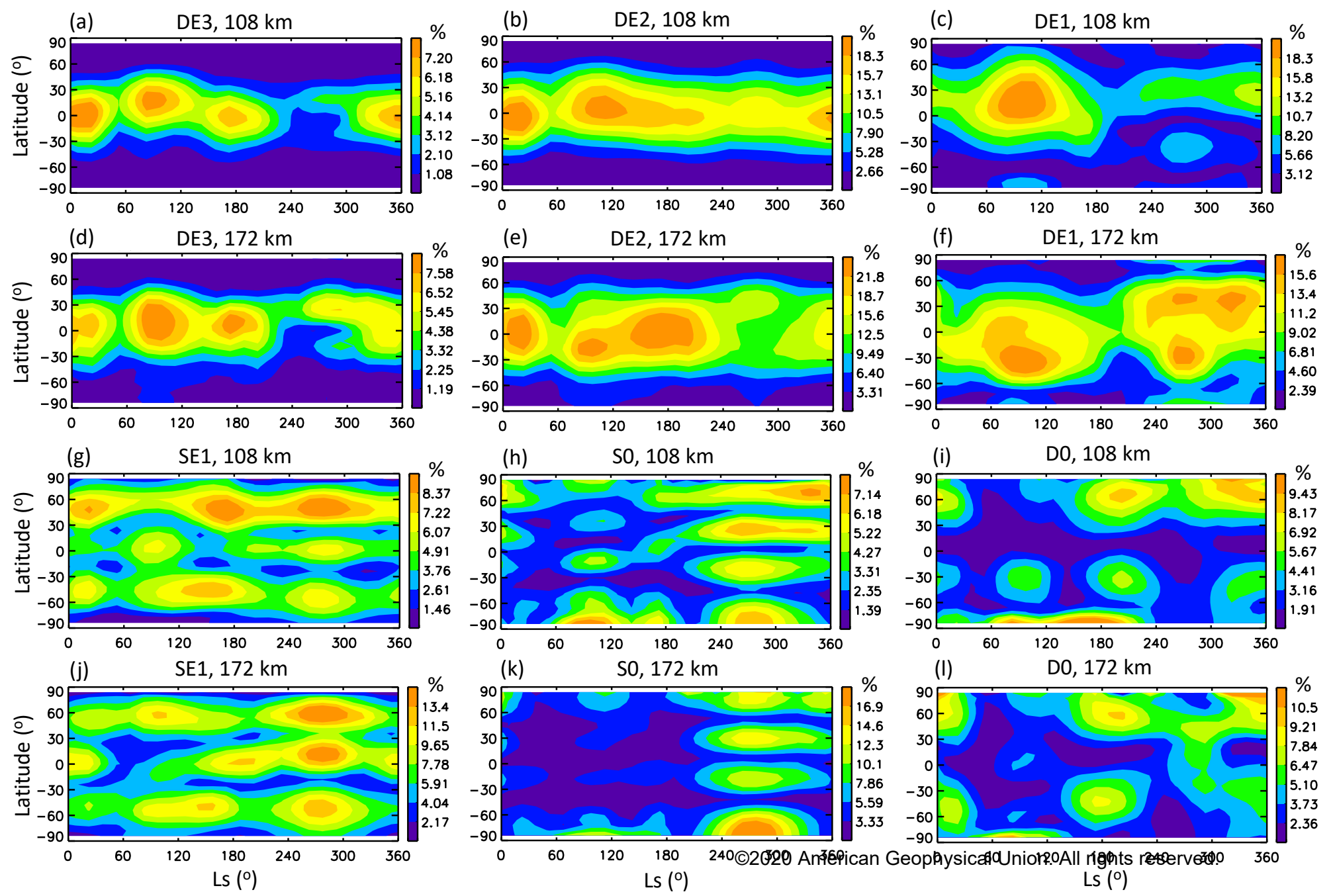


Figure 12.

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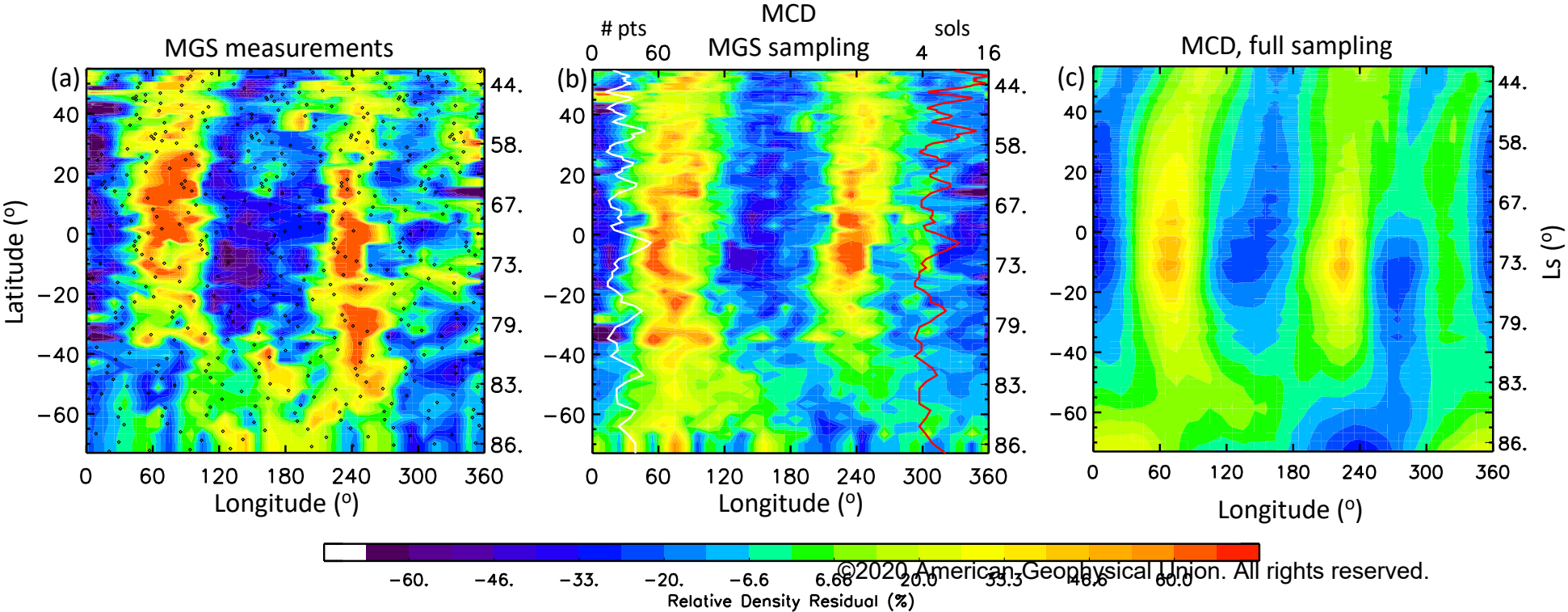


Figure 13.

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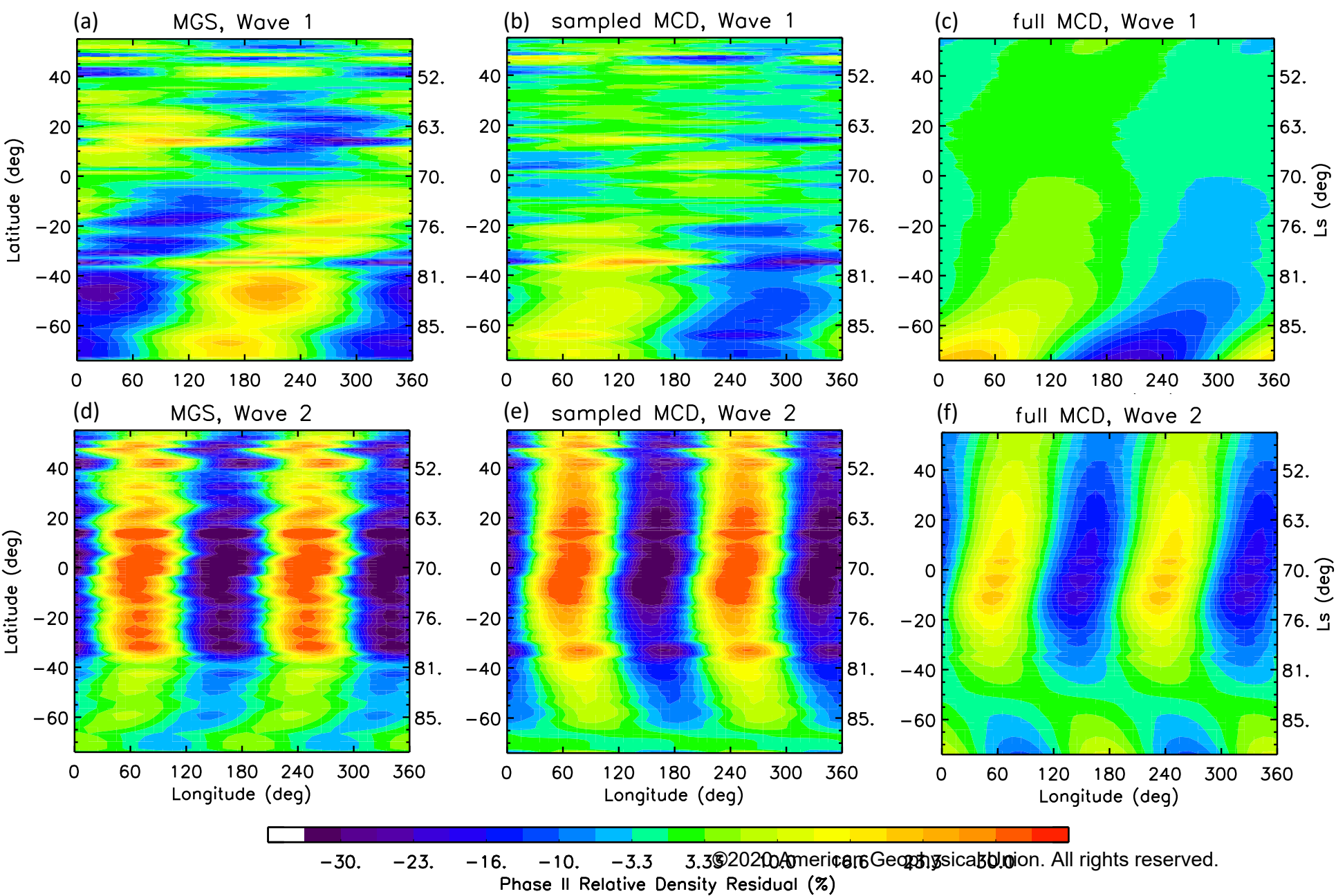


Figure 14.

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