



Climatology of the CO vertical distribution on Mars based on ACS TGO measurements

Anna A. Fedorova, Alexander V. Trokhimovskiy, Franck Lefèvre, Kevin Olsen, Oleg I. Korablev, Franck Montmessin, Nikolay I. Ignatiev, Alexander Lomakin, François Forget, Denis A. Belyaev, et al.

► To cite this version:

Anna A. Fedorova, Alexander V. Trokhimovskiy, Franck Lefèvre, Kevin Olsen, Oleg I. Korablev, et al.. Climatology of the CO vertical distribution on Mars based on ACS TGO measurements. *Journal of Geophysical Research. Planets*, 2022, 127 (9), pp.e2022JE007195. <10.1029/2022JE007195>. <insu-03781486>

HAL Id: insu-03781486

<https://insu.hal.science/insu-03781486v1>

Submitted on 20 Sep 2022

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.



HAL Authorization

Climatology of the CO vertical distribution on Mars based on ACS TGO measurements

Anna Fedorova¹, Alexander Trokhimovskiy¹, Franck Lefèvre², Kevin S. Olsen³, Oleg Korablev¹, Franck Montmessin², Nikolay Ignatiev¹, Alexander Lomakin¹, Francois Forget⁴, Denis Belyaev¹, Juan Alday^{3,5}, Mikhail Luginin¹, Michael Smith⁶, Andrey Patrakeev¹, Alexey Shakun¹ & Alexey Grigoriev^{1,7}

¹Space Research Institute (IKI) RAS, Moscow, Russian Federation.

²LATMOS, CNRS, Sorbonne Université, Université Versailles Saint-Quentin-en-Yvelines, Guyancourt, France.

³Department of Physics, University of Oxford, Oxford, UK.

⁴Laboratoire de Météorologie Dynamique, Sorbonne Université, Centre National de la Recherche Scientifique, Jussieu, Paris, France

⁵School of Physical Sciences, The Open University, Milton Keynes, UK

⁶NASA Goddard Space Flight Center, Greenbelt, USA

⁷Australian National University, Research School of Astronomy and Astrophysics, Mount Stromlo Observatory, Advanced Instrumentation and Technology Centre, Canberra, Australia

Corresponding author: Anna Fedorova (fedorova@iki.rssi.ru).

Key points:

- The first climatology of CO vertical profiles is collected for the period $L_s=163^\circ$ of Martian Year 34 to the end of Martian Year 35.
- The mean CO volume mixing ratio is ~ 960 ppmv at 0–35 km and 45°S – 45°N

- Strong enrichment of CO, >3000 ppmv, is detected at 10-20 km at Ls=100-200 in high southern latitudes and above 50 km at equinoxes near Poles.

Abstract

Carbon monoxide is a non-condensable gas in the Martian atmosphere produced by the photolysis of CO₂. Its abundance responds to the condensation and sublimation of CO₂ from the polar caps, resulting in seasonal variations of the CO mixing ratio. ACS onboard the ExoMars Trace Gas Orbiter have measured CO in infrared bands by solar occultation. Here we provide the first long-term monitoring of the CO vertical distribution at the altitude range from 0 to 80 km for 1.5 Martian years from Ls=163° of MY34 to the end of MY35. We obtained a mean CO mixing ratio of ~960 ppmv at latitudes from 45°S to 45°N and altitudes below 40 km, mostly consistent with previous observations. We found a strong enrichment of CO near the surface at Ls=100-200° in high southern latitudes with a layer of 3000-4000 ppmv, corresponding to local depletion of CO₂. At equinoxes we found an increase of the CO mixing ratio above 50 km to 3000–4000 ppmv at high latitudes of both hemispheres explained by the downwelling flux of the Hadley circulation on Mars, which drags the CO enriched air. General circulation models tend to overestimate the intensity of this process, bringing too much CO. The observed minimum of CO in the high and mid-latitudes southern summer atmosphere amounts to 700-750 ppmv, agreeing with nadir measurements. During the global dust storm of MY34, when the H₂O abundance peaks, we see less CO than during the calm MY35, suggesting an impact of HO_x chemistry on the CO abundance.

Plain language summary

Carbon monoxide (CO) is a product of the photodissociation of the main component of the Martian atmosphere, CO₂ (96%). During the polar night on Mars, the temperature drops so low that ~30% of the global atmospheric carbon dioxide freezes out, condensed into snowflakes. They settle to the ground to form a seasonal polar cap, which sublimates in spring. Unlike CO₂,

CO is not a condensable species, so its relative abundance increases when CO₂ and the atmospheric pressure drops, forming a prominent seasonal cycle. This behavior of CO was well documented by spacecraft measurements of column abundances, while its vertical distribution remained poorly known until recently. Still, it helps to separate the impacts of atmospheric dynamics and chemistry, addressing the problem of the Martian atmosphere's stability. We present the first climatology of the CO vertical distribution lasting more than one Martian year based on solar occultations by the ExoMars Trace Gas Orbiter. We report a previously unobserved CO layer near the surface in high southern latitudes in late winter and spring and document the difference between two Martian years caused by the global dust storm. These observations give new challenges to General Circulation Models for understanding the modern atmosphere of Mars.

1. Introduction

Carbon monoxide (CO) is produced by the photolysis of CO₂ and plays a crucial role in the photochemical cycle that stabilizes the atmosphere's composition. CO was discovered by Kaplan et al. (1969). Its observed abundance is by two orders of magnitude lower than the expected abundance of ~8% in the context of a dry atmosphere (Nair et al., 1994). This inconsistency posed a problem for the stability of CO₂ in the Martian atmosphere, later resolved by introducing catalytic reactions with the H₂O photolysis products (like OH) (McElroy and Donahue (1972); Parkinson and Hunten (1972)). Still, a problem persists in quantitative understanding of the HO_x chemistry because current models mostly underestimate the amount of CO on Mars (Lefèvre & Krasnopolsky, 2017).

The CO abundance on Mars has been measured by high-resolution ground-based spectroscopy from infrared lines located in the 3-0 (1.57 μm, Krasnopolsky, 2003; Krasnopolsky, 2007), 2-0 (2.3 μm, Krasnopolsky, 2015; Billebaud et al., 1998) and 1-0 (4.7 μm, Billebaud et al., 1992) bands of CO and rotational microwave spectrum (Clancy et al., 1990; Lellouch et al., 1991). The column-averaged CO mixing ratios inferred from these observations represented

76 mostly the abundance in the lower atmosphere. It varied from 600 ± 150 ppmv (parts per million
77 by volume) to 800 ± 200 ppmv. Krasnopolsky et al. (2003) first reported the north-south
78 asymmetry of the CO distribution at $L_s = 112^\circ$ with CO volume mixing ratios (VMRs) varying
79 from 830 ppmv in the northern middle latitudes to 1250 ppmv at 50°S . The first map of seasonal-
80 latitudinal variations of CO based on several latitudinal tracks (Krasnopolsky 2007) showed an
81 increase of CO abundance up to 1600 ppmv at $L_s = 112\text{--}173^\circ$.

82 The seasonal distribution of CO was observed in nadir observations made with OMEGA
83 (Observatoire pour la Minéralogie, l'Eau, les Glaces et l'Activité) and PFS (Planetary Fourier
84 Spectrometer) onboard Mars Express (Encrenaz et al., 2006; Billebaud et al., 2009; Sindoni et
85 al., 2011; Bouche et al., 2019; 2021), CRISM (Compact Reconnaissance Imaging Spectrometer
86 for Mars) on board the Mars Reconnaissance Orbiter (Smith et al., 2009, 2018) and recently
87 NOMAD (Nadir and Occultation for Mars Discovery) onboard the ExoMars Trace Gas Orbiter
88 (TGO) (Smith et al., 2021) using the $2.3\ \mu\text{m}$ (2-0) and $4.7\ \mu\text{m}$ (1-0) absorption bands. These
89 observations have shown strong seasonal variations of CO, especially in high latitudes. The
90 minimal values (down to 200 ppmv based on CRISM) have been found in the southern summer
91 high latitudes. At low-to-mid latitudes, the CO_2 condensation-sublimation cycle leads to a broad
92 peak ($\sim 1000\text{--}1100$ ppmv) around the northern autumnal equinox at about $L_s = 180^\circ$ (evident in
93 global climate models (GMCs), CRISM and PFS data, less so in NOMAD). This maximum is a
94 result of the transport of the CO enriched air from the winter polar region. CO behaves as other
95 non-condensable gases like N_2 , Ar, O_2 , expected to vary in response to the condensation-
96 sublimation cycle of CO_2 into and from the seasonal polar caps. Seasonal variations of argon
97 were measured by Mars Odyssey gamma-ray spectrometer (Sprague et al., 2004; 2012);
98 similarly, variations of Ar, N_2 , O_2 and CO were found at the surface by Sample Analysis at Mars
99 (SAM) onboard the Curiosity rover (Trainer et al., 2019). This phenomenon is well reproduced
100 by GCM simulations (Forget et al., 2008; Smith et al., 2018; Holmes et al., 2019).

101 The annual mean value of CO remains a topic of debate. Nadir measurements from PFS,
102 CRISM and NOMAD support a global average of ~ 800 ppmv (Bouche et al., 2021, Smith et al.,
103 2018, 2021). Earth-based observations yielded a lower value of ~ 700 ppmv (Krasnopolsky
104 2015). In situ measurements by SAM/Curiosity resulted in an even lower value of 580 ppmv
105 (Trainer et al., 2019). After several Martian years of observations, such small mixing ratios were
106 never confirmed from orbit at the MSL location (Smith et al., 2018, 2021).

107 While the global seasonal trends of the CO column are relatively well understood, its
108 vertical distribution was poorly documented before the arrival of TGO at Mars. Using PFS,
109 Bouche et al. (2019; 2021) retrieved CO vertical profiles by exploiting the vertical sensitivity of
110 the $4.7\ \mu\text{m}$ band. These measurements carry information mainly on the CO column below 15 km
111 with enhanced sensitivity near the surface. However, their inferred vertical gradients at 0–20 km
112 can not be reproduced by 3D climate models, which rely on our current understanding of
113 Martian photochemistry (Forget et al., 2008). The authors concluded that their observed vertical
114 gradients are artificial and caused by the vertical sensitivity of PFS.

115 In 2018, the ExoMars TGO began its science phase by observing Mars' atmosphere from
116 its orbit. Two scientific instruments onboard TGO, the Atmospheric Chemistry Suite (ACS)
117 (Korablev et al., 2018) and NOMAD (Vandaele et al., 2018), can sound the vertical structure of
118 the atmosphere in solar occultation mode. ACS includes three high-resolution infrared
119 spectrometers: NIR (near-infrared), MIR (middle infrared) and TIRVIM (thermal infrared)
120 (Korablev et al., 2018). All three channels measure the vertical distribution of CO in three
121 spectroscopic bands used previously in ground-based and spacecraft observations: $1.57\ \mu\text{m}$
122 (NIR), $2.3\ \mu\text{m}$ (MIR and TIRVIM) and $4.7\ \mu\text{m}$ (TIRVIM). The first CO vertical profiles were
123 inferred from ACS MIR at altitudes 20–120 km and $\text{Ls}=164\text{--}220^\circ$ before and during the global
124 dust storm (GDS) of Mars year (MY) 34. They showed a prominent depletion in the CO mixing
125 ratio up to 100 km, pointing to the importance of CO oxidation during wetter GDS conditions
126 (Olsen et al., 2021).

127 Here, we report the first long-term observations of the CO vertical distribution on Mars.
128 They are based primarily on the ACS NIR dataset and compared to ACS MIR and TIRVIM solar
129 occultations. We study the seasonal and latitudinal variations of the CO profiles for MY 34 and
130 35, focusing on the interannual variability, and compare our data with the Laboratoire de
131 Météorologie Dynamique (LMD) GCM results, including photochemistry (Lefèvre et al., 2004,
132 2021).

133 **2. Measurements and the CO retrievals**

134 **2.1 Dataset**

135 ACS is a dedicated solar occultation experiment, with all three channels operating in this
136 mode. In solar occultation, the transmittance at different altitudes in the atmosphere is derived by
137 taking the ratio of each spectrum with the reference solar spectrum obtained when observing the
138 Sun outside the atmosphere (before or after an occultation sequence). This method provides high
139 sensitivity since the airmass factor at the limb is ten times greater than in nadir, and the Sun is
140 the brightest input source of the Solar system. The polar and circular orbit of the spacecraft (78°
141 of inclination, 2 hours period and 400 km above the surface of the planet) was optimized to
142 provide vertical profiling at unprecedented spatial and temporal resolution (up to 24 occultations
143 per day). A misalignment between ACS MIR and NOMAD lines-of-sight (LOS) does not allow
144 simultaneous measurements of all occultation experiments onboard TGO. Thus, the number of
145 occultations had to be divided between NOMAD and ACS. However, the field-of-view (FOV) of
146 NIR and TIRVIM allow measurements with NOMAD pointing that result in different coverage
147 and fullness of datasets for these two ACS spectrometers.

148 The densest dataset, which constitutes the basis of this study, was produced by ACS NIR,
149 which provides CO measurements in the 1.57 μm band at 0-60 km for each occultation. From the
150 beginning of the TGO science phase in April 2018 (Ls=163° of MY 34) to January 2021
151 (Ls=360° of MY 35), ACS NIR collected about 6050 profiles (Figure 1). ACS NIR can use two

152 kinds of pointing (ACS-driven and NOMAD-driven), and we have analyzed all available
153 occultation profiles (the NOMAD pointing results in ~40% of profiles).

154 ACS MIR gives access to a much stronger band of CO located at 2.35 μm , resulting in the
155 capability to measure CO up to 90–110 km. The MIR dataset includes 620 occultations
156 performed using secondary grating position 6 (see section 2.3). The validity of the NIR and MIR
157 profiles was established by comparing the two sets of profiles.

158 ACS TIRVIM measures both the 2.35 μm band and the strongest fundamental 4.7 μm band
159 (see fig. S6). The interferogram can be obtained at high- or low-resolution observation mode
160 defined by the maximum optical path difference (MOPD). The MOPD of the TIRVIM
161 interferometer can be as large as 5 cm^{-1} . In the high-resolution mode (MOPD = 4.3 cm), the
162 spectral sampling $\Delta\nu$ and the spectral resolution (full-width at half-maximum (FWHM) of the
163 instrumental function with Hamming apodization) are equal to 0.11 and 0.21 cm^{-1} , respectively.
164 In the low-resolution mode MOPD = 0.97 cm, $\Delta\nu = 0.51 \text{ cm}^{-1}$ and FWHM = 0.94 cm^{-1} . The
165 largest number of occultations was done in the low-resolution mode, while only 20 orbits were
166 done in high resolution mode, but with worse signal-to-noise ratio (SNR). For the analysis, we
167 used the low-resolution TIRVIM observations of the 4.7 μm band to validate NIR and MIR
168 results at 0–40 km. TIRVIM ceased operations in December 2019 (Ls=115°, MY 35) after
169 having recorded ~1000 solar occultations since the beginning of the mission.

170 The following sections describe the datasets and the processing of ACS MIR, TIRVIM and
171 NIR, and introduce the validation of the retrieved CO profiles. An in-depth comparison of CO
172 VMRs retrieved from different channels can be found in Supplementary Material.

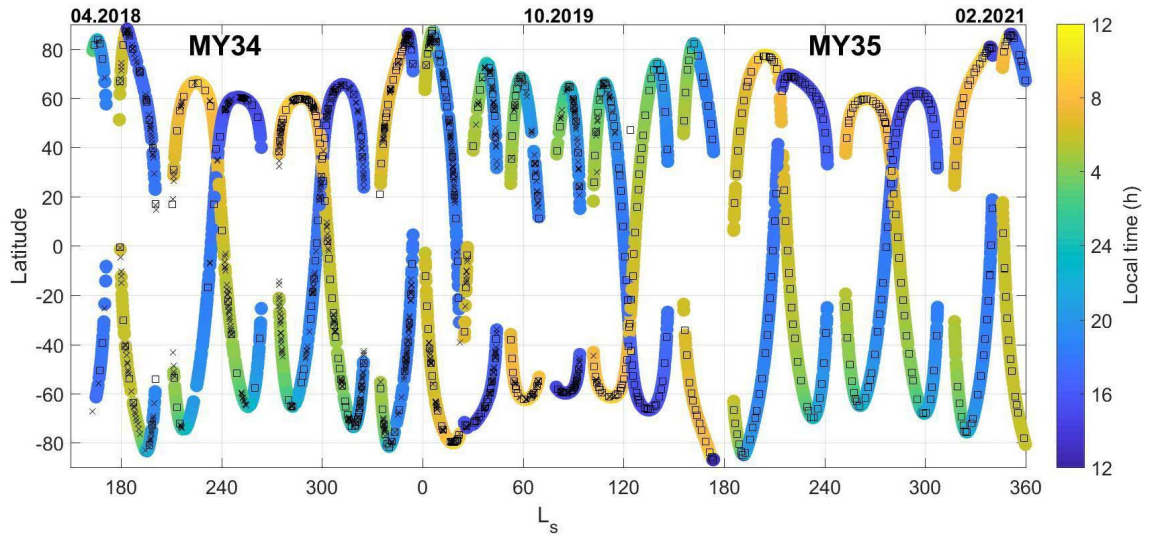


Figure 1. Observational coverage of ACS solar occultations. The color circles, where colours indicate the local time of the occultation, show NIR observations. Black squares show MIR observations of CO, and black crosses show TIRVIM observations.

2.2 NIR

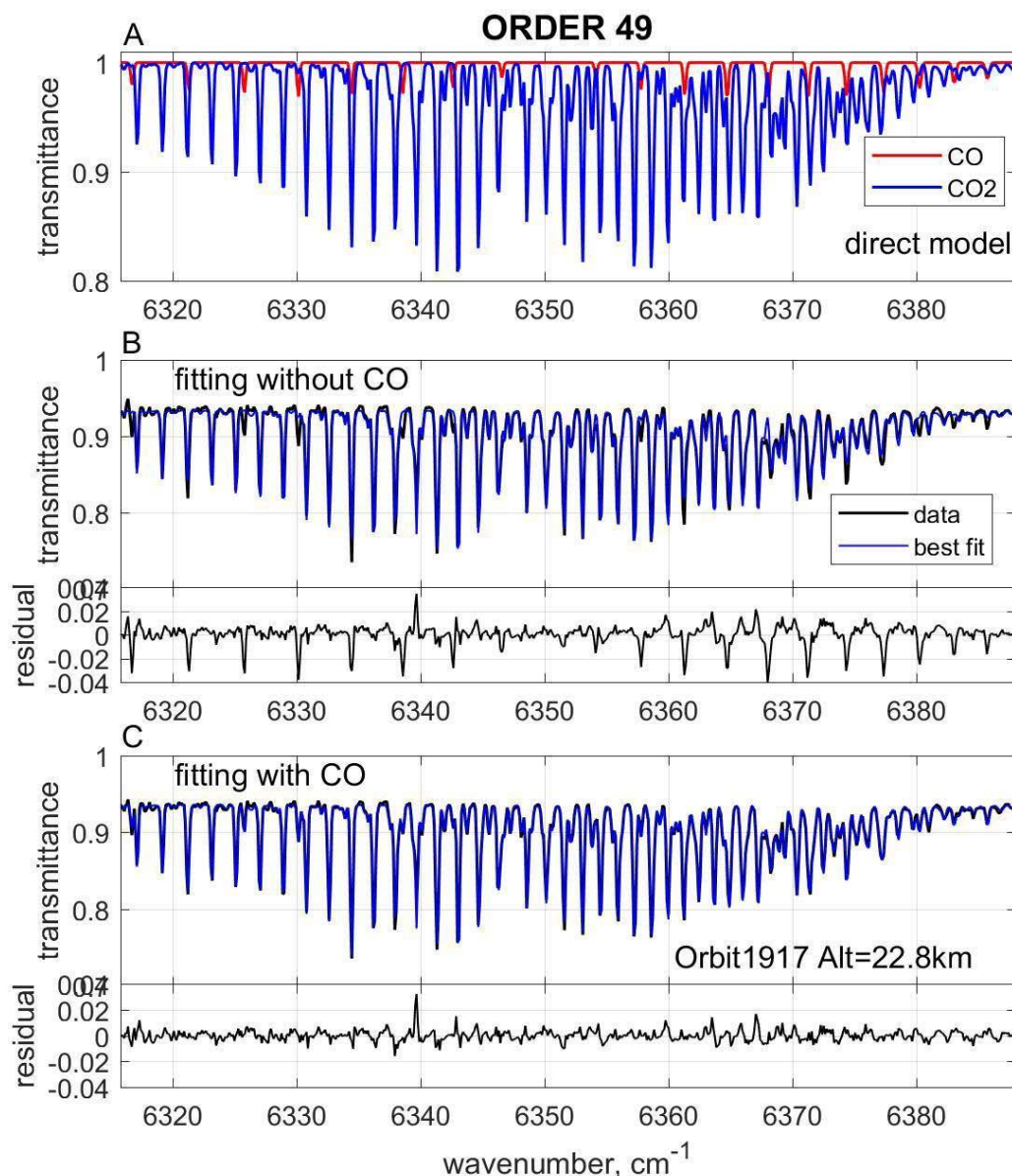
ACS NIR combines an acousto-optical tunable filter (AOTF), used as a monochromator that can be commanded to isolate a specific wavelength bandpass whose width corresponds to the free spectral range of the echelle grating that operates at high diffraction orders. It covers the 0.7–1.7 μm spectral range using diffraction orders 101 through 49 (Trokhimovskiy et al., 2015a; Korablev et al., 2018). During an occultation, ACS NIR measures ten pre-selected diffraction orders every 2 s, including the absorption bands of H_2O and CO_2 . The resolving power of the spectrometer is varied from 20000 to 28000 depending on spectral range and pixel position over the detectors (Korablev et al., 2018).

The instantaneous FOV in the direction perpendicular to the limb is small ($\sim 0.02^\circ$) and limited by the slit width. The FOV corresponds to an instantaneous vertical resolution of 500–600 m at the tangent altitude of the line of sight. The time to measure one diffraction order is 0.2 s. The effective vertical resolution depends on the β -angle (the angle between the orbit plane and the direction to the Sun). It is generally better than 1 km for each diffraction order. Horizontal

192 resolution (along the slit) is 1.5–5 km and is determined by the number of stacked detector lines.
193 The number of detector lines measured in a single occultation and position of the lines on the
194 detector depends on the pointing direction (NOMAD and ACS MIR-driven) and the available
195 downlink volume. It varies from 6 to 30 lines; the solar image size on the detector limits the
196 upper boundary. Averaging 30 detector lines does not worsen vertical resolution since that
197 dimension of FOV is oriented along the limb. The SNR for an individual pixel and pure solar
198 signal in the AOTF maximum equals about ~600 for the MIR pointing and ~300 for the
199 NOMAD pointing. We averaged 6–25 lines with the maximal signal, excluding the solar image
200 edges, resulting in an SNR of 800–3000, depending on the occultation.

201 The calibration of ACS NIR spectrometer included the detection of dead and hot pixels,
202 AOTF function characterizations (important for order overlapping), AOTF frequency-
203 wavelength calibrations and the drift of AOTF calibrations with temperature, blaze function
204 characterization, spectral calibrations (pixel wavelength) and instrumental line shape (ILS)
205 determining the spectral power (Korablev et al., 2018). During this work on the calibrated NIR
206 transmission data, the wavelength drift is additionally corrected using positions of gaseous
207 absorption lines.

208 The high spectral resolution and the SNR of NIR allow to resolve very weak absorption
209 bands as well as rotational structure of abundant gases like CO₂. This makes NIR a sensitive
210 instrument for temperature retrieval in the atmosphere. Diffraction order 49 of NIR (6318–
211 6387 cm⁻¹) was initially used to retrieve the temperature and CO₂ density using the 1.57 μm CO₂
212 band (Fedorova et al., 2020). This order also contains a weaker overlapping CO band. Even if
213 hardly noticeable within the strong CO₂ band, the retrieval algorithm detects the CO lines well
214 thanks to their periodic structure. It permits measuring the CO mixing ratio from 0 to 60-70 km.
215 Figure 2 shows examples of spectra fitted in order 49 with and without CO. The 1.57 μm band
216 was already used for measuring CO abundance in the Martian atmosphere using high-resolution
217 ground-based observations (Krasnopolsky, 2003).



218

219 Figure 2. Example of ACS NIR spectrum measured in order 49 at 23 km altitude showing
 220 the benefit of fitting with a model including both CO₂ and CO. A): a synthetic model of CO₂
 221 (blue) and CO (red, 1000 ppmv) absorption bands. B): spectrum observed (in black) at orbit
 222 1917, Ls=167°, the latitude of 82°, east longitude of -92.3° and local time 21:15 with its best fit
 223 (in blue) without CO and fit residual; (C) the same observed spectrum with its best fit (in blue)
 224 with CO and fit residual.

225 The forward transmission model is computed for a corresponding number of atmospheric
 226 layers (40 to 130 depending on orbit) using a pre-computed look-up table of absorption cross-

227 sections (as a function of pressure and temperature) calculated line-by-line. The spectral line
228 parameters for the CO₂ and CO are taken from the HITRAN 2016 database (Gordon et al., 2017)
229 with CO₂-broadening coefficients for CO (Li et al., 2015) and self-broadening for CO₂. The
230 order overlapping (two close orders from both sides) due to distant side lobes of the AOTF
231 function was taken into account additionally in the direct model before the comparison with the
232 measured spectrum is made.

233 The retrieval algorithm is based on a Levenberg–Marquardt iterative scheme and Tikhonov
234 regularization to smooth the profile and minimize the uncertainties (Ceccherini, 2005). As an
235 initial assumption, the temperature and pressure from the Mars Climate Database (MCD) v. 5.3
236 (climatology) profiles have been used (Millour et al., 2019). The initial mixing ratio of CO was
237 assumed to be 1000 ppm at all altitudes.

238 The inverse problem is divided into two steps. The first step includes fitting for
239 temperature, pressure and CO mixing ratio independently, based on the resolved rotational
240 structure of the CO₂ band. The retrieval algorithm returns all the unknowns at all altitudes at the
241 same time, i.e. the vertical profiles (see Fedorova et al., 2018; 2020 for details). We do not fit the
242 CO₂ volume mixing ratio, as its vmr is supposed to be constant up to 100 km. However, in polar
243 regions, the CO₂ vmr is predicted by GCMs to decrease above 80 km but this region is higher
244 than the NIR sensitivity to the CO lines and didn't impact on the NIR CO retrievals. Once the
245 solution is obtained, Tikhonov regularization is applied, customary for vertical inversions to
246 smooth the profile and minimize the errors. The obtained smoothed profiles serve as an initial
247 assumption for the following iteration with Levenberg–Marquardt. The algorithm converges
248 within 3-6 iterations independently of initial assumptions as shown in Fedorova et al. (2020), and
249 demonstrates a good sensitivity of the Jacobian to the temperature profile thanks to the
250 pronounced CO₂ rotational structure in our spectral range.

260 However, the retrieval of the vertical profiles of pressure with the rotational structure of the
 261 CO₂ band generates sometimes unrealistic results. In ~30% of profiles of the analyzed dataset,
 262 we observed a bend, a drop of the pressure curve at some altitudes, in contradiction with the
 263 hydrostatic law (Fedorova et al., 2020). Therefore, in the second step, we assume that the
 264 atmospheric layers are in hydrostatic equilibrium to constrain the retrieval of temperature and
 265 pressure simultaneously. We define a pressure P_0 at an altitude level z_0 where the uncertainties
 266 on the previous step are minimal (the highest data quality), and then compute the pressure profile
 267 from the hydrostatic equilibrium, integrating the retrieved T profile above and below the P_0
 268 level:

$$261 \quad P(h) = P_0 \exp\left(-\int_{z_0}^h \frac{g(z)M(z)}{T(z)k} dz\right),$$

270 where z is the altitude, $g(z)$ is the acceleration due to gravity, $M(z)$ is the molecular weight
 271 (taken from the MCD), k is the Boltzmann constant, h is the height at which P is being
 272 determined. Once the pressure is calculated, we repeat the temperature retrieval with only two
 273 free parameters, temperature and CO vmr, and perform several iterations until the pressure and
 274 temperature profiles converge. At first, the uncertainty on the retrieved quantities is given by the
 275 covariance matrix of the solution. In the second step, the uncertainty of P_0 is also taken into
 276 account. The more detailed retrieval algorithm and comparison of results is described in
 277 (Fedorova et al., 2020). The retrieval products from order 49 are the pressure, atmospheric
 278 temperature and CO mixing ratio, as shown in figure 5.

276 The solar occultation is a self-calibrated method and instrumental uncertainties are
 277 minimized when dividing a spectrum observed through the atmosphere to the pure solar
 278 spectrum. In case of NIR, the retrieval of CO mixing ratio, temperature and pressure from the
 279 same order also minimize the instrumental uncertainties. Nonetheless, the main uncertainty of
 280 the calibration and data processing which can be critical for the CO retrieval is an order
 281 overlapping. It is coming from distant side lobes of the AOTF function which was calibrated in

laboratory and in flight (Korablev et al., 2018). We estimate that the width of the AOTF function in order 49 and the contribution of side lobes can make the uncertainties up to 5% to the retrieval of the CO mixing ratio (see fig. S1). The other important contribution is the atmospheric temperature profile used for the retrieval. The accuracy of temperature profiles retrieved from order 49 was discussed in Fedorova et al. (2020) and is sensitive to the AOTF calibration and initial assumptions. It was also estimated from the cross-calibration of the ACS NIR and ACS-MIR retrievals of position 4 (2.7 μm CO₂ band) (see Belyaev et al., 2021; Alday et al., 2021) where the data of two experiments were consistent inside 5-7 K below 70-80 km and the MIR data were mostly warmer. The line strength of strong CO lines inside order 49 is sensitive to temperature with averaged +0.3% per -1 Kelvin. It means if the atmospheric temperature is 7 K colder, we underestimate the CO mixing ratio by $\sim 2\%$. Finally, the total systematic uncertainties of the NIR retrieval does not exceed 7%.

2.3 MIR

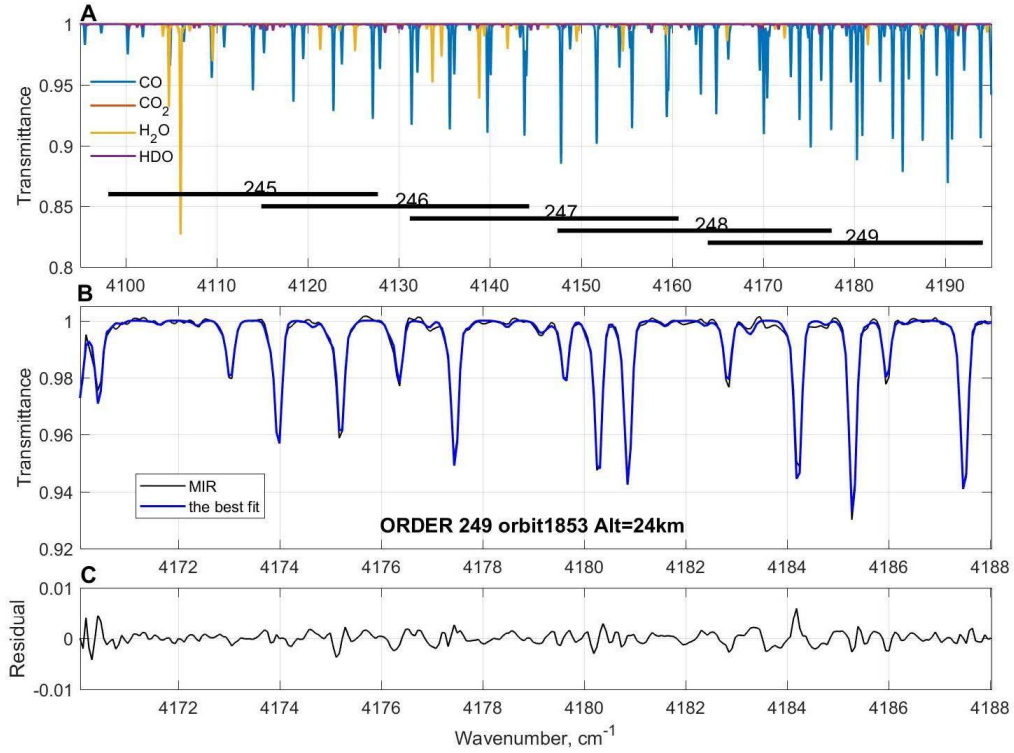
The ACS MIR (mid-infrared) channel is a crossed-dispersion echelle spectrometer dedicated to solar occultation measurements in the 2.3–4.5 μm range with spectral resolving power of $\sim 30,000$. The channel can only sample a fraction ($\sim 160\text{ cm}^{-1}$) of the full range on a single observation. To access a particular interval, a secondary dispersion grating is rotated to one of 11 positions (Korablev et al., 2018; Trokhimovskiy et al., 2015b, 2020). During an occultation, MIR is operated in one pre-chosen secondary grating position. Spectra are recorded on a two-dimensional detector, with the x -axis corresponding to wavenumber and the y -axis to diffraction orders and the vertical FOV. MIR measures up to 20 adjacent diffraction orders for each acquired frame, covering an instantaneous spectral range of 0.15–0.3 μm . Each diffraction order appears as a stripe over the detector, covering around 20 rows. The width of each stripe represents the vertical FOV, and each row is a unique solar occultation spectrum separated by 100-200 m from its adjacent rows.

301 We use the strong 2.3 μm band for the CO retrieval, which can be measured in positions 6
302 or 7. Position 7 covers the strongest lines of the 2-0 transitions, located in orders 248–256, and
303 allows measuring CO abundances up to 110 km. Thirty-two ACS MIR occultations have been
304 used to assess the CO vertical distribution during the Global Dust Storm of MY 34 (Olsen et al.,
305 2021). In position 6, the three CO isotopologues (main $\text{C}^{12}\text{O}^{16}$, $\text{C}^{13}\text{O}^{16}$, $\text{C}^{12}\text{O}^{18}$) are measured in
306 orders 246-249 (Figure 3). Position 6 also contains CO_2 and H_2O absorption features, used as a
307 reference in the calibration pipeline. For these reasons, CO monitoring was changed from
308 position 7 to position 6 in November 2018. The line strengths are a little smaller than in Position
309 7, constraining the measurements of CO up to 90 km. The strongest lines are found in order 249,
310 used for the retrieval.

311 Calibration of the spectra includes an orthorectification of the detector image, the removal
312 of hot pixels, the dark signal and the straylight, and a correction for the sub-pixel drift occurring
313 due to the slightly varying thermal state of the instrument. The spectral calibration is performed
314 comparing the solar lines with the ACE-FTS solar atlas (Hase et al., 2010), and then refined
315 using atmospheric absorption lines of CO. The instrument line shape is impacted by a doubling
316 (or multiplying) of the image at the detector, originating from an optical anomaly somewhere
317 along the incidence optical path. As a result, many diffraction orders display a doubling of
318 absorption lines. The orders in the bottom part of the detector are mostly free from this effect,
319 and the lines in order 249 located in the lowest part of the detector were preferable. Nevertheless,
320 slight doubling was observed in order 249, and the instrumental function was parameterized
321 using a double gaussian with the fitting of its halfwidth and the relative intensity of two maxima.
322 These parameters are sought for each occultation. A similar approach was described in Alday et
323 al. (2019), Olsen et al. (2021) and Belyaev et al. (2021). The FOV of MIR is small; its vertical
324 resolution is mainly defined by the integration time and amounts to 2.5 km. We averaged ten
325 detector lines near the stripe center for the retrieval, effectively increasing the vertical resolution
326 to 3 km.

327 The retrieval algorithm of CO from MIR data is similar to that described for the NIR
328 channel. Only two free parameter vectors are retrieved (the H₂O and CO VMRs). The CO₂
329 absorption band in order 249 is weak and contributes only in the lower atmosphere and at low
330 temperatures. We fixed the CO₂ mixing ratio to 0.96 for the modeling. The HDO/H₂O ratio was
331 assumed constant and equal to 5 (Alday et al., 2021). The pressure and temperature profiles used
332 for constraining spectroscopic parameters are taken from the simultaneous NIR retrievals
333 obtained for the same occultation. The MIR CO vertical profiles for two orbits of MY 34
334 concurrent with NIR observations are shown in Figure 5C, F.

335 In case of MIR, the uncertainties of the stray light and the image-doubling made the
336 uncertainties to the instrument line shape (a double peak) and the target altitude. Thanks to the
337 vertical offset of the secondary image, working near the slit edge (in case of order 249) we
338 minimize this problem and contributions of the secondary peak are less than 10% of the primary
339 on average. To test the stability of the retrieval to calibration uncertainties we made the retrieval
340 for each line of the order 249 separately (see fig.S2). The resulting uncertainties of the averaged
341 profile does not exceed 3% and is consistent with the profiles obtained from the averaging of
342 detector lines that showed a stability of results. The second possible bias is the atmospheric
343 temperature and pressure profile used for the retrieval. There is no strong CO₂ band in order 249
344 and we use NIR temperature-pressure profiles to reduce the uncertainties coming from unknown
345 atmospheric conditions (see fig.S3). The good agreement between NIR and MIR VMR, obtained
346 with different instruments in different absorption bands for the same atmospheric profiles,
347 supports that the uncertainties, coming from calibration and spectroscopy, have been minimized.
348 As was mentioned previously, the NIR temperature can have an uncertainty of 5-7 K. The line
349 strength of the main CO isotopologue (C¹²O¹⁶) in spectral range of order 249 has a sensitivity to
350 temperature ~1.5% per Kelvin, it can result in 10% uncertainties of the CO mixing ratio
351 especially in high altitudes where CO lines are not saturated. Finally, we estimate the total
352 systematic uncertainties of the MIR retrieval not exceeding 13%.



353
 354 Figure 3. An example ACS MIR spectrum measured in position 6 at 24 km altitude and the best
 355 fit including CO, CO₂ and H₂O contributions. A) the direct model of CO, H₂O, CO₂ and HDO
 356 absorption bands in orders 245-249 of position 6. B) spectrum observed at orbit 1853, Ls=164°,
 357 81.7°N, 20.5°E and local time of 02:20 (black curve) and the best fit of the data (blue curve); C)
 358 residual for the spectrum.

359

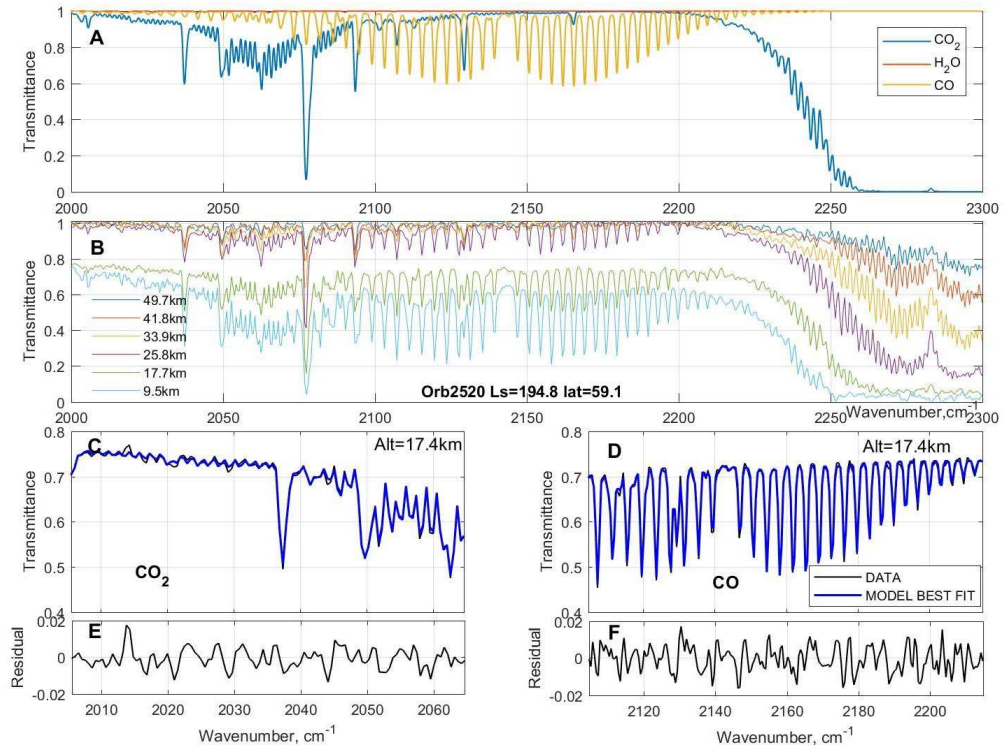
360 2.4 TIRVIM

361 The thermal infrared (TIRVIM) channel is a Fourier transform spectrometer with solar
 362 occultation capability (Korablev et al., 2018; Shakun et al., 2018). TIRVIM operates mainly in
 363 the “climatology” mode in solar occultation geometry, covering the entire spectral range from
 364 1.7 to 17 μm every 0.4 s, with a spectral resolution of 0.94 cm^{-1} . At 2.8° , the diameter of
 365 TIRVIM’s circular FOV is larger than the solar disk. The vertical resolution, determined by the
 366 angular diameter of the Sun, is close to one atmospheric scale height, varying with Sun-Mars
 367 distance from 8.8 to 10.6 km.

368 To retrieve CO from TIRVIM spectra, we used the strongest 4.7 μm band. The low SNR
369 (≤ 100 only) allows for vertical profile retrievals of CO below 50-60 km (fig. S6). To increase the
370 SNR, four consecutive spectra were averaged. The retrieval algorithm of CO from TIRVIM data
371 is similar to that described for the NIR and MIR channels. The retrieval is sensitive to the
372 atmospheric conditions we have considered (fig. S7-S8). To reduce these uncertainties, we used
373 coinciding NIR temperature profiles available for about 90% of TIRVIM occultations.

374 The main uncertainty of TIRVIM comes from the large FOV ($\sim 2.8^\circ$ equals to ~ 75 km at the
375 limb). The area of the atmosphere inside FOV is ~ 60 times higher than the diameter of the Sun.
376 It makes the contribution of aerosol scattering inside FOV not negligible compared to the solar
377 signal especially in the low altitudes. The influence of forward scattered light in transmission
378 measurements of planetary atmospheres has been studied in detail in de Kok and Stam (2012).
379 Based on VIMS Cassini solar occultation measurements of the Titan's atmosphere they showed
380 the gas abundances will be underestimated by about 8% if forward-scattering is ignored in the
381 retrievals. Inside 75 km TIRVIM FOV the aerosol opacity can change by order of magnitude and
382 contribution of different atmospheric layers in the absorption is difficult to predict. To make a
383 valid retrieval of the gaseous absorption in dusty Martian atmosphere the full radiative transfer
384 with the multiple scattering in spherical geometry is needed. It is a difficult problem, given the
385 poorly constrained TIRVIM FOV and it is outside of our current possibility. To minimize this
386 uncertainty, we therefore limited the TIRVIM retrievals to transmittance of 0.8 (the aerosol slant
387 optical depth of ~ 0.2 at target altitude) and additionally retrieved the CO₂ density from the
388 nearby 4.9 μm CO₂ band because even in this case the contribution of the forward scattering is
389 not excluded due to about 25 km of scattering atmosphere below the solar disc. Finally, CO
390 VMRs of TIRVIM were obtained as the ratio of CO and CO₂ densities.

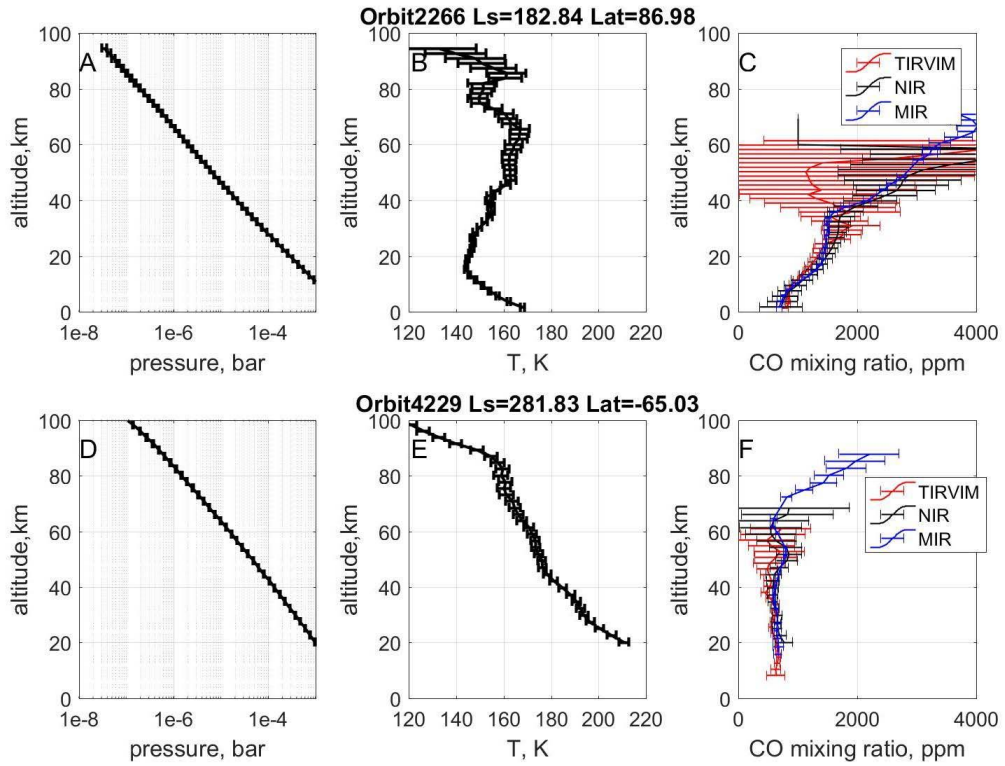
391 Examples of TIRVIM spectra and fitting are presented in Figure 4, while CO VMR vertical
392 profiles for two orbits in MY 34 are shown in Figure 5C, F.



393

394 Figure 4. An example of measured ACS TIRVIM spectrum and its best fit showing the benefit
 395 of fitting with a model including both CO₂ and CO. A) the direct model of CO, H₂O, CO₂
 396 absorption bands in the spectral range 2000-2300 cm⁻¹. B) spectra observed at different altitudes
 397 at orbit 2520, Ls=194.8°, 60.2°N, 167°E and local time of 17:25; C, D) spectrum observed at 17
 398 km at the same orbit 2520 (black curve) in the range of CO₂ and CO bands, respectively, and
 399 the best fit of the data (blue curve); F, E) residual for the spectrum of CO₂ and CO bands,
 400 respectively.

401



402

403 Figure 5. Example profiles retrieved from ACS NIR-MIR-TIRVIM occultations.

404 A) and D) pressure, derived from the CO₂ number density; B) and E) atmospheric temperature;
 405 C) and F) CO profiles from NIR (black), MIR (blue) and TIRVIM (red). The orbits 2266
 406 (Ls=182.8°, 87°N, 37.4°E and local time of 7:55) and 4229 (Ls=281.8°, 65.0°S, -74.0°E and
 407 local time of 0:21) were completed in northern spring equinox and southern summer solstice,
 408 respectively. The retrieved covariance matrix gives the error bars (one-sigma).

409

410

411

412

413

414

415 **3. Results**

416 **3.1 Seasonal trend of CO vertical distribution**

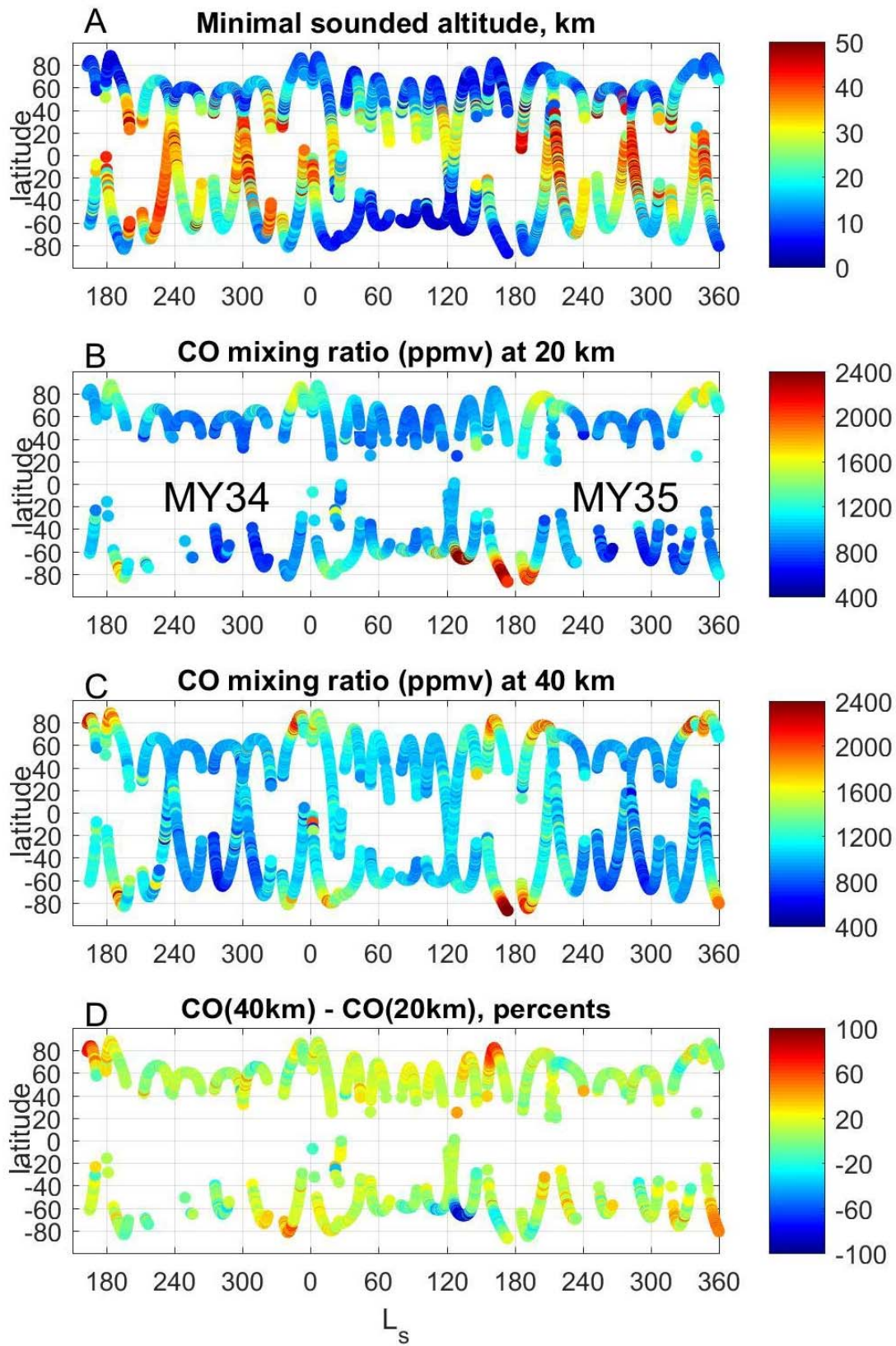
417 GCMs predict that CO is expected to be well-mixed from the surface up to ~50 km in
418 mid-and low latitudes (Lefèvre and Krasnopolsky, 2017; Daerden et al., 2019). To verify these
419 theoretical predictions, we considered the evolution of CO measured at two characteristic
420 altitudes, 20 km and 40 km for MYs 34 and 35, as a function of season and latitude (Figure 6).
421 Below 20 km, profiles are strongly attenuated by aerosol extinction. Above 40 km, CO
422 absorption is getting fainter, increasing uncertainty with altitude. However, aerosols frequently
423 prevent accessing altitudes below 30 km (Figure 6A). Thus, the measurements at 20 km are only
424 available in mid-to high-latitudes (30–80°) of both hemispheres. The observations at two
425 altitudes show an apparent difference in VMR and latitudinal variations. The most considerable
426 difference is observed toward the Poles, where downwelling from Hadley cell circulation
427 strongly impacts the shape of the CO VMR vertical profiles, as shown in Olsen et al. (2021).

428 To follow the evolution of profiles with season and latitude in more detail, we grouped
429 the NIR and MIR data in 30° bins of Ls and presented it as a function of latitude and altitude for
430 both Martian years (Figure 7). Near the solstice in both hemispheres (Ls=60–120° and 240–
431 300°), the latitude coverage is limited to $\pm 60^\circ$. In the equinox period (Ls=330–30° and 150–
432 210°), polar latitudes are well covered. Whereas NIR can only measure CO below 50 km, the
433 MIR channel can probe CO up to 90 km in the most favorable cases, a higher altitude range than
434 is accessible for NIR. TIRVIM is not shown here due to poor coverage with altitude and season.
435 The distribution for TIRVIM can be found in the Supporting Information (figure S11).

436 Overall, the vertical profiles of the CO mixing ratio were found far from being uniform.
437 The minimal values of CO were observed during the southern summer in middle-high southern
438 latitudes. This is consistent with trends observed by CRISM (Smith et al., 2018), PFS (Bouche et
439 al., 2021) and NOMAD (Smith et al., 2021) while the NIR minimal values were within 600–700

440 ppmv compared to 300–400 ppmv reported by CRISM for latitudes between 70S and 50S
441 (discussed in detail in subsection 3.4). The most prominent feature of the observed profiles is the
442 pronounced maxima of CO VMR at high latitudes. They are present in the lower atmosphere
443 during southern winter and spring and above 40 km near the equinoxes. These polar and mid-
444 latitude enrichments are discussed further in subsections 3.3 and 3.5. In MIR observations, the
445 increase of CO VMR to 3000 ppmv is also observed in the middle and low latitudes above 80
446 km, indicating the CO₂ photolysis is already efficient at these altitudes. This gradual increase at
447 high altitudes is well supported by one-dimensional models (Krasnopolsky, 2022, fig.5) and was
448 also observed in the first MIR profiles up to 120 km before and after GDS of MY34 (Olsen et al.,
449 2021).

450 We have also calculated the seasonal distribution of CO with the LMD GCM and its
451 coupled photochemical module in the latest configuration described in Lefèvre et al. (2021). In
452 this dedicated simulation, the GCM is initialized at $L_s = 180^\circ$ of Martian year 33. At this date,
453 CO is set to a uniform (both in altitude and horizontal) initial mixing ratio of 1000 ppmv as
454 previously measured at low latitudes by MIR (Olsen et al., 2021). After a half Martian year of
455 spin-up time, the model runs continuously until the end of Martian year 35 while being
456 constrained by the observed dust climatology (Montabone et al., 2020). The CO vertical
457 distribution calculated by the LMD GCM (Figure 7, left) is in good general agreement with NIR
458 and MIR at low-to-mid latitudes. The increase of CO above 80 km due to the CO₂ photolysis is
459 well predicted. A seasonal minimum is evident in the perihelion season of both years considered
460 here. However, the model largely overestimates the downwelling of thermospheric CO measured
461 during the equinoctial periods at high latitudes. In contrast, near the surface, the strong
462 enrichment of CO measured at high southern latitudes in late winter is absent in the GCM after
463 $L_s=150^\circ$. This will be also discussed in the following sections.



464

465 Figure 6. Carbon monoxide mixing ratios (ppmv) from ACS NIR occultations at two altitudes.

466 A) The minimal sounded altitude in km, corresponding to aerosol attenuation of 0.07–0.08; B)

467 CO VMR at 20 km; C) CO VMR at 40 km; D) relative difference of CO VMR at 40 km and CO
468 VMR at 20 km.

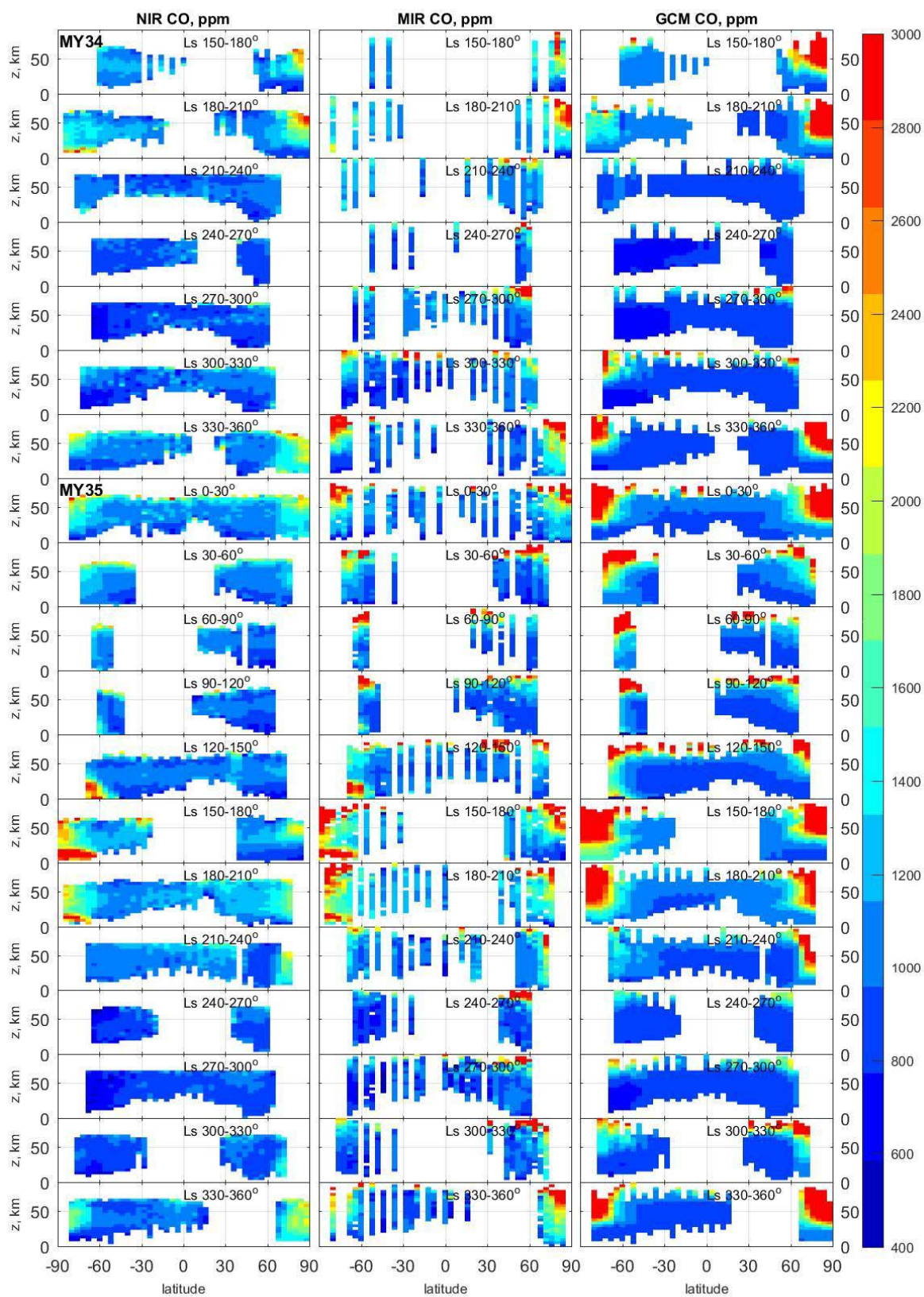


Figure 7. Seasonal distribution of CO mixing ratio observed by ACS NIR (left), ACS MIR (middle) and LMD GCM (right): from MY34 on the top to MY35 on the bottom. Colorbar presents the CO VMR in ppmv.

3.2 The average CO mixing ratio

The global average CO mixing ratio was the focus of many previous observations and 1D modeling studies. The occultation method does not allow sounding the profiles of gases all the way down to the surface due to aerosol opacity of the Martian atmosphere. Figure 6A demonstrates that the minimal sounded altitude in NIR occultations varies from near the surface in polar regions to as high as 40 km at low latitudes. At the same time the global circulation and one-dimensional models predict a uniform CO distribution in the lower atmosphere at low-mid latitudes (Lefevre and Krasnopolsky, 2017, see also S12). To estimate the column-averaged mixing ratio of CO from the ACS profiles, we integrated them from the lowest retrieved altitude to 40 km, where all three spectrometers have good accuracy. To avoid the profiles where aerosol was too opaque in the atmosphere and the lowest retrieved points have big errors due to high slant optical depth and, as a result, low signal, we consider occultations with the minimal altitude (z_{\min}) lower than 30 km. We also excluded the polar region and high latitudes where the CO₂ condensation-sublimation cycle and the Hadley circulation make the CO vertical distribution non-uniform (e.g., Figure 6, 7, S12). Thus, we limited ACS measurements to the middle-low latitudes (45°S–45°N). We integrated the CO and CO₂ density profiles from z_{\min} to 40 km to get the column abundances. Then the column average CO mixing ratio was obtained as a ratio of column densities, $\text{CO}(\text{cm}^{-2})/\text{CO}_2(\text{cm}^{-2})$. Figure 8A shows the seasonal evolution of the column average of CO mixing ratios for the NIR, MIR and TIRVIM measurements, including the earlier results obtained from MIR (Olsen et al., 2021). The CO mean value shows a repeatable seasonal cycle from MY34 to MY35 with two maxima, the first and most prominent at $L_s \sim 180^\circ$ and the second shortly after the spring equinox ($L_s = 0^\circ$) in the Northern hemisphere.

495 By averaging the Figure 8 data over Ls, we obtain estimates of the annual mean mixing
496 ratio of CO: 970 ± 123 ppmv from NIR, 975 ± 129 ppmv from MIR. In the case of TIRVIM, the
497 annual average equals 852 ± 91 ppmv. The TIRVIM dataset resulted in the lowest average CO
498 VMR and the same was observed for individual occultations (for example around $L_s=0$ in
499 Fig.8A). TIRVIM has limited Ls coverage ($L_s=240^\circ$ MY 34 to $L_s=115^\circ$ MY 35) and did not
500 probe the two equinox maxima near $L_s=180^\circ$, resulting in underestimating the annual mean
501 VMR by about 10%, but that is not enough to explain the observed difference. Forward
502 scattering could also be responsible for this difference, even if we try to minimize its impact by
503 choosing altitudes with smaller aerosol slant opacity. Therefore, the three channels of ACS
504 together produce a fairly reliable annual mean for CO of about 960 ppmv.

505 Figure 8B shows the comparison of ACS results with nadir measurements. NIR, MIR and
506 TIRVIM retrievals for individual occultations were binned within 5° of Ls. PFS/Mars-Express
507 measured CO in the nadir using the $4.7 \mu\text{m}$ band (Bouche et al., 2021). The data have been taken
508 from the general repository Open Science Framework (Bouche et al., 2020). A selection of PFS
509 data for the same $\pm 45^\circ$ latitude belt was averaged within 5° bins of Ls for all available Martian
510 years (MY26–33). The seasonal trend agrees well with ACS data. Bouche et al. (2021)
511 calculated a global mean for CO of 820 ppmv. Restricted to the mid-to-low latitudes, the PFS
512 dataset results in a global mean of 844 ppmv, in agreement with TIRVIM results and 15% lower
513 than NIR and MIR values.

514 Figure 8B presents also the column-averaged CO VMRs retrieved from CRISM and
515 NOMAD by M. D. Smith et al. (2018, 2021) and averaged inside $\pm 40^\circ$ latitude belt. The annual
516 average in this belt equals ~ 840 ppmv for CRISM and ~ 800 ppmv for NOMAD, both are lower
517 than our averaged values (Smith et al., 2018; Smith et al., 2021). From ground-based high-
518 resolution observations Krasnopolsky (2015, 2017) obtained an annual mean of 681 ± 13 ppmv,
519 agreeing with the *in situ* measurements by Curiosity yielding 580 ± 80 ppmv (Franz et al., 2015,
520 Trainer et al., 2019).

521 The mean CO mixing ratio calculated by the LMD GCM in Figure 8A shows a very good
522 quantitative agreement with the observations and qualitatively for what concerns the seasonal
523 variation. For both Martian years, the calculated mixing ratio is maximum around $L_s = 180^\circ$.
524 This is in line with the seasonal evolution of CO at low-to-mid latitudes as measured by nadir-
525 viewing satellites (e.g., Smith et al., 2021; Bouche et al., 2021) and is, more generally, typical of
526 the seasonal evolution of all long-lived non-condensable species (e.g., Trainer et al., 2019). This
527 similarity suggests that the CO₂ condensation-sublimation cycle at polar latitudes is the main
528 driver of the seasonal signal in CO visible in Figure 8. Photochemistry should also play a role in
529 this signal, especially in the CO decrease observed in the perihelion season, when larger amounts
530 of H₂O at all altitudes lead to much increased OH production. The relative contributions of the
531 CO₂ condensation-sublimation and of photochemistry will be studied in a dedicated paper. Also,
532 beyond the scope of the present study is the long-standing problem of the underestimation of CO
533 by models (Lefèvre and Krasnopolsky, 2017), which requires overly lengthy simulations in order
534 to reach the long-term equilibrium value of CO in the GCM. Therefore, this historical problem
535 should not be considered as ‘solved’ from the sole good agreement found in Figure 8A.

536 Addressing the difference between our observed global mean of 960 ppmv and the nadir
537 results from CRISM, PFS and NOMAD, ~800-840 ppmv, we can consider two possibilities: 1)
538 the geometry of TGO occultations and high aerosol opacity in the Martian atmosphere do not
539 allow us to get good statistics in low latitudes, as seen in figure 6A and SI, and most of the data
540 was obtained at middle latitudes; 2) the reason may be in a sensitivity difference between nadir
541 and occultation instruments and related to the vertical distribution of CO. In nadir, the signal is
542 dominated by absorption from the denser layers and thus by the lowest scale heights below 20-25
543 km. The MIR/NIR vmr average is not exactly a column-average vmr because we can’t sound the
544 near-surface mixing ratio at lower latitudes. The minimal sounded altitude (fig 6A) for ACS
545 instruments varies from 2-4 km to 30 km above areoid with an average of 20-21 km, which is
546 above the altitudes that nadir measurements are most sensitive to. If CO increases with altitude

below 40 km, the vmr average determined from ACS will be larger than the nadir column-average vmr. This effect should be verified, but we note that except for the polar latitudes in summer (the CO₂ sublimation) the CO vmr in the GCM never decreases close to the surface (fig. S12C) and the standard deviation of CO vmr in low-to-mid latitudes from 0 to 40 km is mostly below 50 ppmv (fig. S12). On the other hand, *Curiosity* measurements on the surface of Mars give 580 ± 80 ppmv (Franz et al., 2015, Trainer et al., 2019) that could indicate near surface values below global average and occultation measurements that are not yet explained in GCMs.

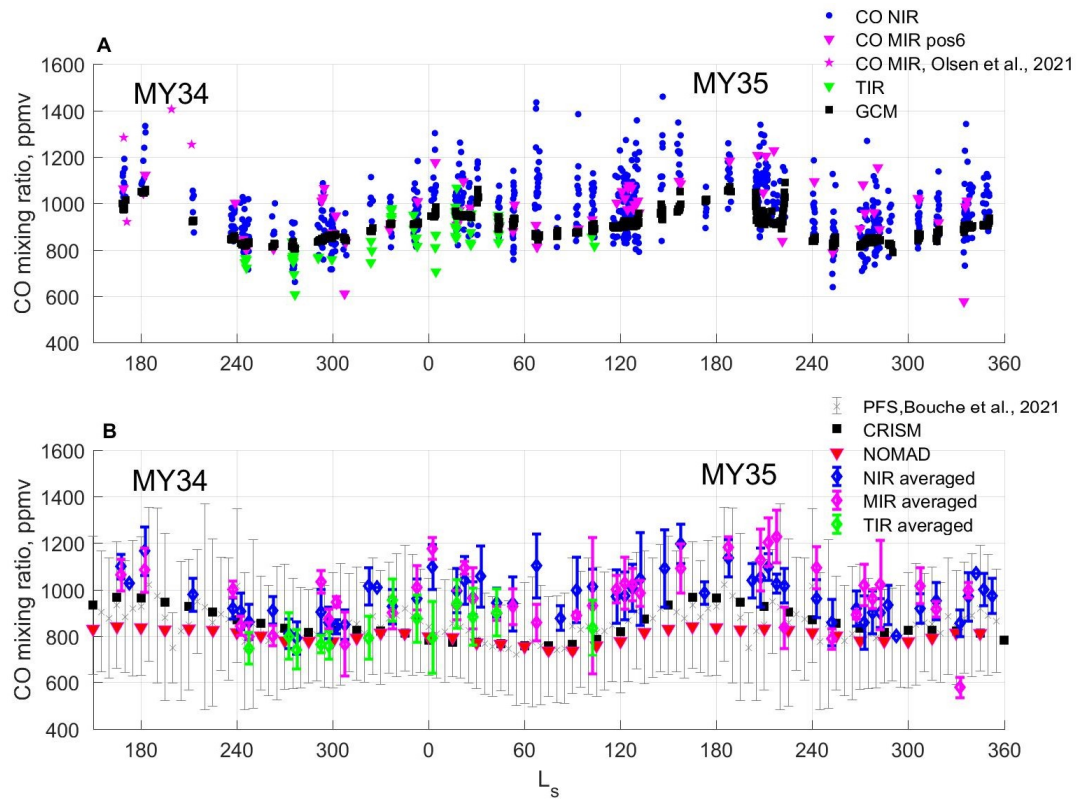


Figure 8. A) The column-averaged values of CO mixing ratios from ACS profiles measured within $\pm 45^\circ$ latitude range and below 40 km. Blue points, purple and green triangles are individual NIR, MIR and TIRVIM occultations, respectively. Purple stars are the results of Olsen et al., 2021 from MIR position 7. Black squares are GCM model results corresponding to the NIR averages. B) Blue, purple and green diamonds with error bars are averages of NIR, MIR and TIRVIM data, respectively, binned within 5° of L_s. Grey crosses are averages of PFS data (see text). All error bars are standard deviations. Black squares and red triangles are the column-

562 averaged CO VMRs retrieved from CRISM and NOMAD data, respectively, by M. D. Smith et
563 al. (2018, 2021) and averaged between 40°S and 40°N of latitude. PFS, NOMAD and CRISM
564 data are averaged for all years of observations and do not present specifically MY34 and 35.

565 **3.3 Southern winter-spring CO enhancement**

566 In the lower atmosphere, the maximal values of CO have been observed during the
567 Southern winter/beginning of spring in southern polar latitudes from $L_s=120^\circ$ to 200° . In MY 34
568 the observations began only at $L_s=163^\circ$ and the increase in mixing ratio below 20 km was
569 detected at $L_s=187^\circ$ – 195° when the occultations reached high southern latitudes (see Figure 7).
570 In MY35 the coverage was better. The near-surface layer with enhanced CO has been observed
571 from $L_s=120^\circ$ to 195° by both NIR and MIR instruments (Figures 6, 7). Figure 9 shows
572 individual CO profiles of NIR obtained in southern high latitudes at $L_s=161^\circ$ – 174° and 185° –
573 195° . A layer of CO at ~10-15 km reached values up to 3000–4000 ppmv (see also fig S13 for
574 MIR and TIRVIM). Twenty-five sols later ($L_s=185^\circ$) the layer moved to ~15 km with a
575 maximum of 2500–3000 ppmv and then disappeared at $L_s\geq 198^\circ$. Above this layer, the CO
576 mixing ratio gradually decreases to 1500-2000 ppmv and then grows again at higher altitudes up
577 to 3000 ppmv. The observed CO maximum correlates with the temperature minimum observed
578 below 30 km at high southern latitudes. At $L_s=167^\circ$ – 174° the minimum is observed at 20 km
579 with values as low as 125-135 K 10 km above the CO layer. At $L_s=185^\circ$ – 195° the minimum is
580 observed with warmer values about 140-150K.

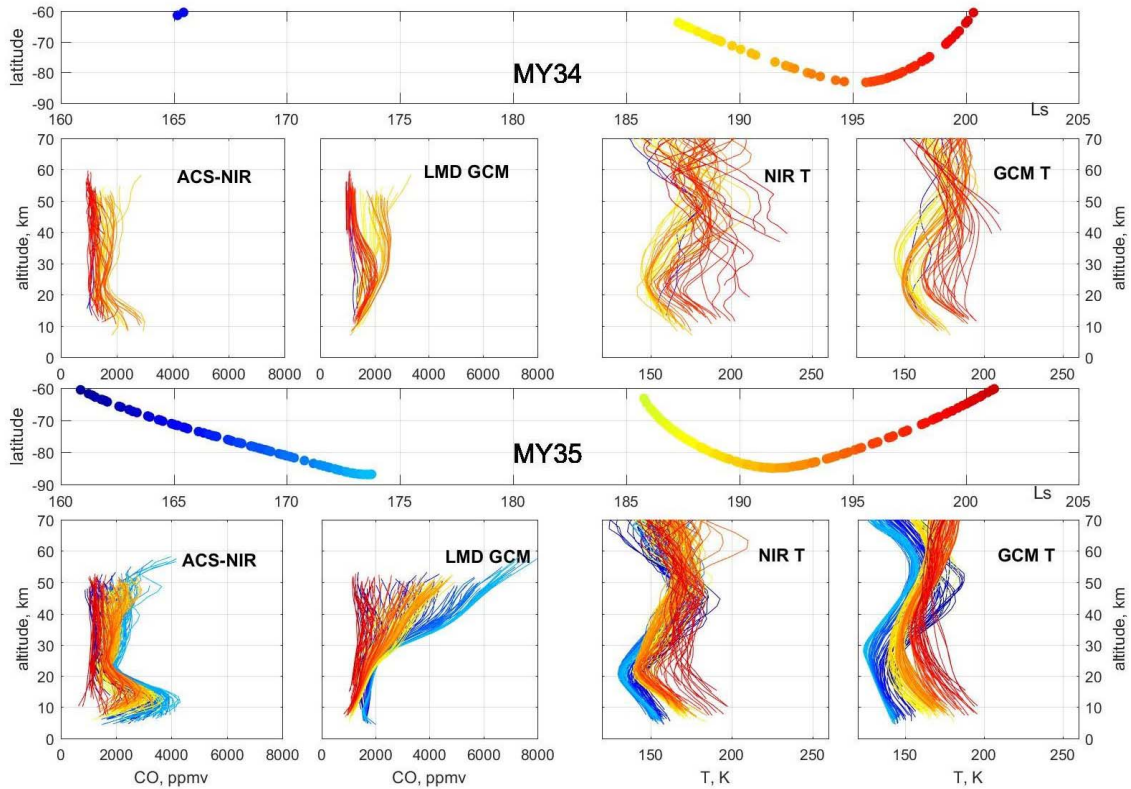
581 Enhancement of non-condensable species was previously observed in atmospheric argon
582 from measurements of the gamma ray spectrometer (GRS) operated on the Mars-Odyssey orbiter
583 (Sprague et al., 2004, 2007). The strong (a factor of 6) enhancement of Ar was measured over
584 south polar latitudes (75°S to 90°S) for two consecutive Mars years near the onset of southern
585 winter close to $L_s=90^\circ$. In winter over the northern polar regions there was no similar strong
586 enhancement of Ar and the seasonal peak was less obvious than in the south.

587 The nadir measurements of CO showed the strongest CO values at $L_s=140^\circ$ - 180° in the
588 southern hemisphere between 45° and 20° S with mixing ratio about 1100 ppmv (Smith et al.,
589 2018; Smith et al., 2021; Bouche et al., 2021). In case of solar reflected light, the data retrieval
590 was traditionally limited to solar zenith angles (SZA) $<80^\circ$ to avoid the complications of
591 radiative transfer near the limb. That was done for CRISM (Smith et al., 2018). In case of
592 NOMAD the limitation was even higher $<55^\circ$ that is explained by low signal-to-noise ratio of the
593 instrument in nadir (Smith et al., 2021). In turn, PFS observes more thermal radiation, and the
594 observation limitation is driven here by the surface temperatures (too low in polar regions)
595 (Bouche et al., 2021). This could explain why nadir experiments like CRISM, NOMAD and PFS
596 do not show these maximal values at the poles.

597 Unlike nadir, solar occultations can access higher latitudes during the equinox and polar
598 night seasons ($>60^\circ$ S). At $L_s=60^\circ$ - 150° of MY35 ACS sounded at the edge of the southern polar
599 night approaching to 65 - 70° S. During the same period, CRISM, PFS and NOMAD were limited
600 to 30 - 45° S. CO mixing ratio apparently increased below 20 km at 60° S from 900 to 1500 ppmv
601 during the $L_s=90^\circ$ - 120° interval (figure 7). Closer to equinox, occultations moved to higher
602 latitudes where the low-atmospheric layer enriched progressively to 3000 ppmv at $L_s=120^\circ$ -
603 150° (see Figures 7, 8) and then 4000 ppmv at $L_s=170^\circ$ - 175° (Figures 8, 9 and S13).

604 In the middle of the southern winter, the column-averaged CO mixing ratio calculated by
605 the LMD GCM reaches more than 4000 ppmv south of 75° S (not shown). However, the region
606 most enriched in CO in the model is located far in the polar night, and only the northernmost tip
607 of the CO-enriched layer near the surface is visible between $L_s = 120$ - 150° in Figure 7). In that
608 layer, the modeled enrichment in CO appears to be underpredicted compared to ACS, both in
609 terms of absolute value and vertical extent. Later on, around the equinox, the GCM no longer
610 predicts an enriched layer of CO at high southern latitudes, in strong disagreement with ACS
611 which shows the persistence of an enriched layer in the $L_s = 180$ - 210° period. This indicates that
612 the breakup of the polar vortex enriched in CO (and in other non-condensable species) and the

613 subsequent mixing with midlatitude air occur too early with the settings of the LMD GCM used
 614 here.
 615



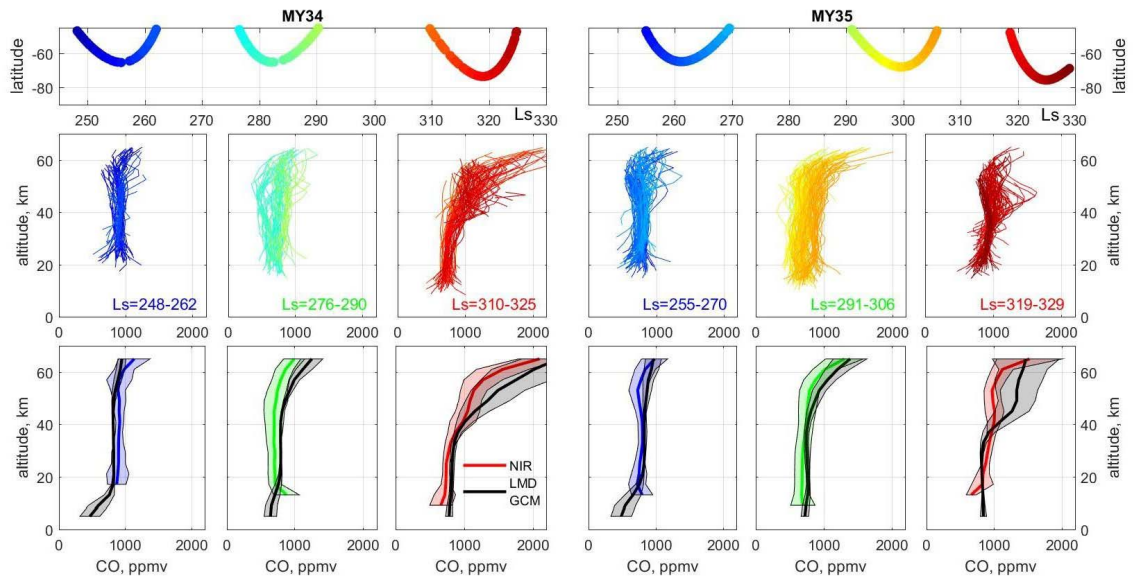
616
 617 Figure 9. The evolution of the CO mixing ratio profiles for a sequence of occultations in the
 618 latitude range 60°S-90°S for martian years 34 and 35. (top) the latitude coverage; (left) ACS NIR
 619 measurements; (right) LMD GCM simulation. Colours indicate Ls.

620

621 3.4. Southern summer CO depletion

622 CRISM observations, followed later by PFS and NOMAD nadir measurements, observed
 623 a minimum of CO abundance during the southern summer. This depletion was related to the
 624 sublimation of the southern CO₂ polar cap and resulted in the strong decrease of the CO column
 625 mixing ratio to the 300-500 ppmv from Ls=210° to Ls=330° in the high and middle southern
 626 latitudes (Smith et al., 2018, 2021).

627 In Figure 6b at 20 km, the CO depletion is seen in both MY 34 and 35 between Ls 250°
628 and 330° in the southern hemisphere. During this period, Mars' GCM predicts a non-uniform
629 vertical distribution with strong vertical gradient and depletion near the surface (Holmes et al.,
630 2019). So, we decided to investigate the CO vertical distribution for the both observed southern
631 summer seasons in more detail.



632
633 Figure 10. The evolution of the CO mixing ratio profiles for a sequence of occultations in the
634 latitude range 45°S–90°S and Ls from 245° to 330° for MY 34 and 35; (top) the latitude
635 coverage; (middle) ACS NIR individual profiles for Ls bins; (bottom) the averaged profile for
636 each Ls bin for ACS NIR and GCM LMD. The shaded area is the standard deviation.

637
638 Figure 10 presents the evolution of the CO profiles in the latitude range from 45°S to
639 90°S and in the period from Ls=245 to 330° for both Martian years (34 and 35). In this dusty
640 season, solar occultation can not sound deep in the atmosphere and the lowest altitude sounded is
641 typically 20 km above the surface except for Ls=310°–325° of MY34 and Ls=291°–306° in
642 MY35 where we could probe down to 10 km. For both years we do not see a prominent gradient
643 below 40 km except at Ls=309–325° of MY 34 where nearly-constant mixing ratios were
644 observed below 35 km, with an increase above. The average CO mixing ratio below 40 km was

645 found to be 700–750 ppmv which is higher than those obtained by CRISM and NOMAD
646 between 50°S and 70°S (400 ppmv at Ls=260° and 700 ppmv at Ls=330° for CRISM and 500
647 ppmv at Ls=260° and 650-700 ppmv at Ls=330° for NOMAD) and by PFS in the same period
648 (~600 ppmv). This difference can be related again to the altitude sensitivity of nadir and
649 occultation measurements. To test it, we compared the CO profiles calculated in summer by the
650 LMD GCM with ACS measurements (fig.10, bottom panels). Above 20 km, where the
651 observations are available, the model and data are in broad agreement. But below 20 km, the
652 model shows a smooth decrease of the mixing ratio to the surface due to the sublimation in the
653 southern summer at Ls=250-270° that could explain a huge difference between ACS and nadir
654 measurements for this period. The model profiles tend also to anticipate the downwelling of
655 enhanced CO mixing ratios by the Hadley circulation at the approach of the equinox (Ls = 318–
656 330° in Figure 10).

657

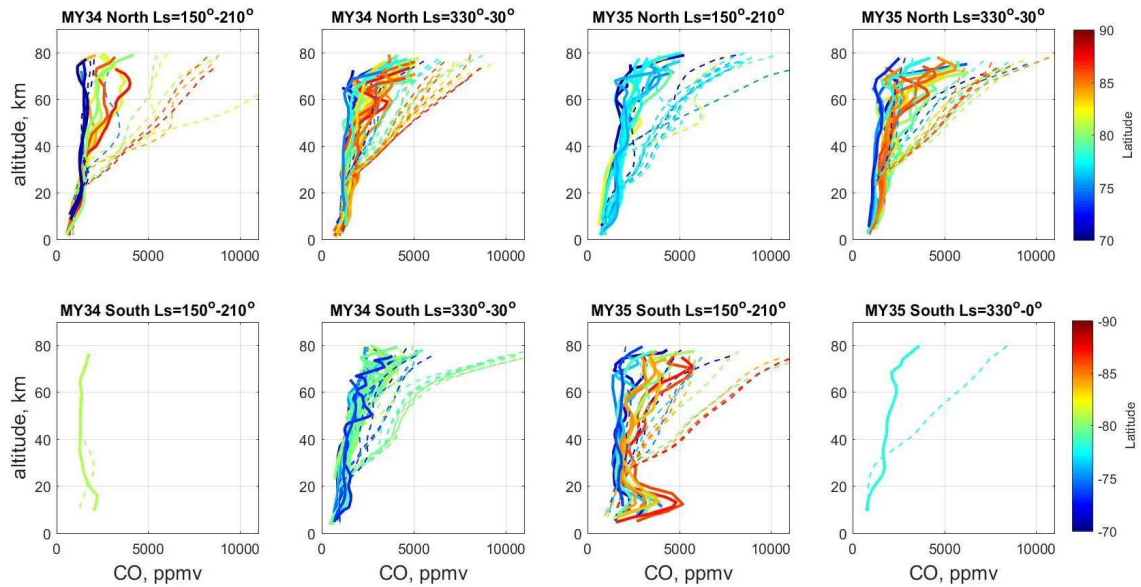
658 ***3.5 High altitude equinox and solstice enhancement***

659 The enhancement of CO at 50 km was first observed by Krasnopolsky (2014) based on
660 CO (2-1) dayglow observations at 4.7 μm in the middle of northern summer (Ls=110°) using the
661 CSHELL spectrograph at NASA IRTF. This dayglow is formed by CO molecules at 50 km. The
662 mixing ratio was found to increase from 1100 ppm at 40°S to 1600 ppm at 70°N. This behavior
663 was different from the trend observed at low altitudes and was therefore attributed to the Hadley
664 circulation characterized by the equator-to-pole Hadley cells in equinox and pole-to-pole cell in
665 solstices. This circulation transports CO to the polar region where a downward flux of CO-rich
666 air occurs.

667 The NIR and MIR observations in the same period also support the increase of the CO
668 mixing ratio in the polar region above 50 km but the mixing ratio is much smaller than in the
669 model (Figure 7). The most sensitive for the study is the MIR spectrometer that can measure CO
670 up to 80 km and has better accuracy compared to NIR at altitudes between 50 and 60 km (see

671 section 2 and SM). Figure 11 presents the observed and modeled mixing ratio profiles of CO in
672 both hemispheres at latitudes from 70° to 90° over time periods that encompasses equinoxes.
673 The equinoctial circulation is characterized by two symmetric Hadley cells facing each other
674 about the equator and carrying airmasses from the equator to the poles. The equinox periods
675 covered concern $L_s=150^\circ$ to 210° (MY35) for the northern spring and $L_s=330^\circ-30^\circ$ (MY 34/35)
676 for the southern spring. Before and after these ranges, the maxima are also observed, but the
677 circulation is already being rearranged. These maxima especially in low-to-middle latitudes are
678 located mostly higher than 60 km and are related to the increase of the CO mixing ratio due to
679 the CO₂ photolysis.

680 At both equinoxes, the model is in rather good agreement with ACS at latitudes about
681 70°, but overestimates CO when moving to the poles. In the northern spring of MY34 and 35 the
682 observed profiles have shown a weak gradient from 1000 ppm at 10 km to 2000–3000 ppm at 60
683 km depending on latitude with smaller values at lower latitudes. Above 60 km and at latitudes
684 higher than 80°N the sharp increase of CO is visible at high latitudes to 5000 ppm at 80 km. The
685 GCM profiles begin to grow from 30 km and overestimate the CO VMR above this altitude by at
686 least a factor of 2 compared to observations. For instance, the modeled CO VMR reached 8000-
687 10000 ppmv at 70-80 km. The same situation is observed in the southern spring except near the
688 surface CO layer discussed in Section 3.3. In the southern and northern autumn of MY34 and 35,
689 respectively, the CO mixing ratio changes from 1000 ppmv at 10 km to 4000 ppmv at 70 km but
690 with a sharper gradient compared to the spring. The model begins to overestimate the data in
691 some cases from 20 km. This suggests that the intensity of the Hadley cells at the equinox is too
692 strong in the GCM, bringing too much CO from its region of production in the high atmosphere.



693

694 Figure 11. The evolution of the CO mixing ratio profiles for sequence of occultations for
 695 equinox season ($L_s=150^\circ-210^\circ$ and $330^\circ-30^\circ$) for MY 34 and 35 in high southern and northern
 696 latitudes (-70° to -90° and 70° to 90° , respectively). Colors indicate the latitude variations. Solid
 697 curves are the ACS MIR data, dashed curves represent the LMD GCM simulation.

698

699 **3.6 Interannual variability due to the GDS of MY 34**

700 In the lower atmosphere, provided there is water vapor, the loss of CO by reaction with
 701 OH exceeds its production. It should then end up making CO sensitive to changes in water
 702 content. Nevertheless, the lifetime of about five terrestrial years does not authorize prominent
 703 chemical variations of CO at the seasonal timescales.

704 MY 34 was characterized by the GDS that began at approximately $L_s=185^\circ$ (Guzewith et
 705 al., 2019; Smith, 2019) in the Northern hemisphere. The first phase of the storm from $L_s=185^\circ$ to
 706 192° is better defined as a regional dust storm which then turned into a massive global event at
 707 $L_s=197^\circ$ after the secondary massive dust injection over Tharsis (Montabone et al., 2020). The
 708 decay phase of the storm started at $L_s=210^\circ$ and lasted until $L_s=240^\circ$. It is now known that GDSs
 709 profoundly impact water vertical distribution in the atmosphere by transferring H_2O to high
 710 altitudes (Fedorova et al., 2018; 2020, 2021; Heavens et al., 2018; Aoki et al., 2019, Belyaev et

711 al 2021). Dust heats the atmosphere and amplifies the Hadley cell circulation which in turn
712 increases the altitude of the hygropause. The water vertical distribution during the GDS of MY
713 34 was studied with ACS and NOMAD on TGO (Fedorova et al., 2020; Aoki et al., 2019;
714 Belyaev et al., 2021) and SPICAM (SPectroscopy for the Investigation of the Characteristics of
715 the. Atmosphere of Mars) on Mars Express using the solar occultation technique (Fedorova et
716 al., 2021). Water was observed up to 120 km with mixing ratios of 20–50 ppmv. More water in
717 the middle atmosphere provides more OH (hydroxyl) radicals once H₂O is photolyzed. This has
718 the effect of boosting the rate of reactions between CO and OH, and hence the reformation of
719 CO₂. So, the sensitivity of CO to water activity can provide an insight into the stability of the
720 Martian atmosphere.

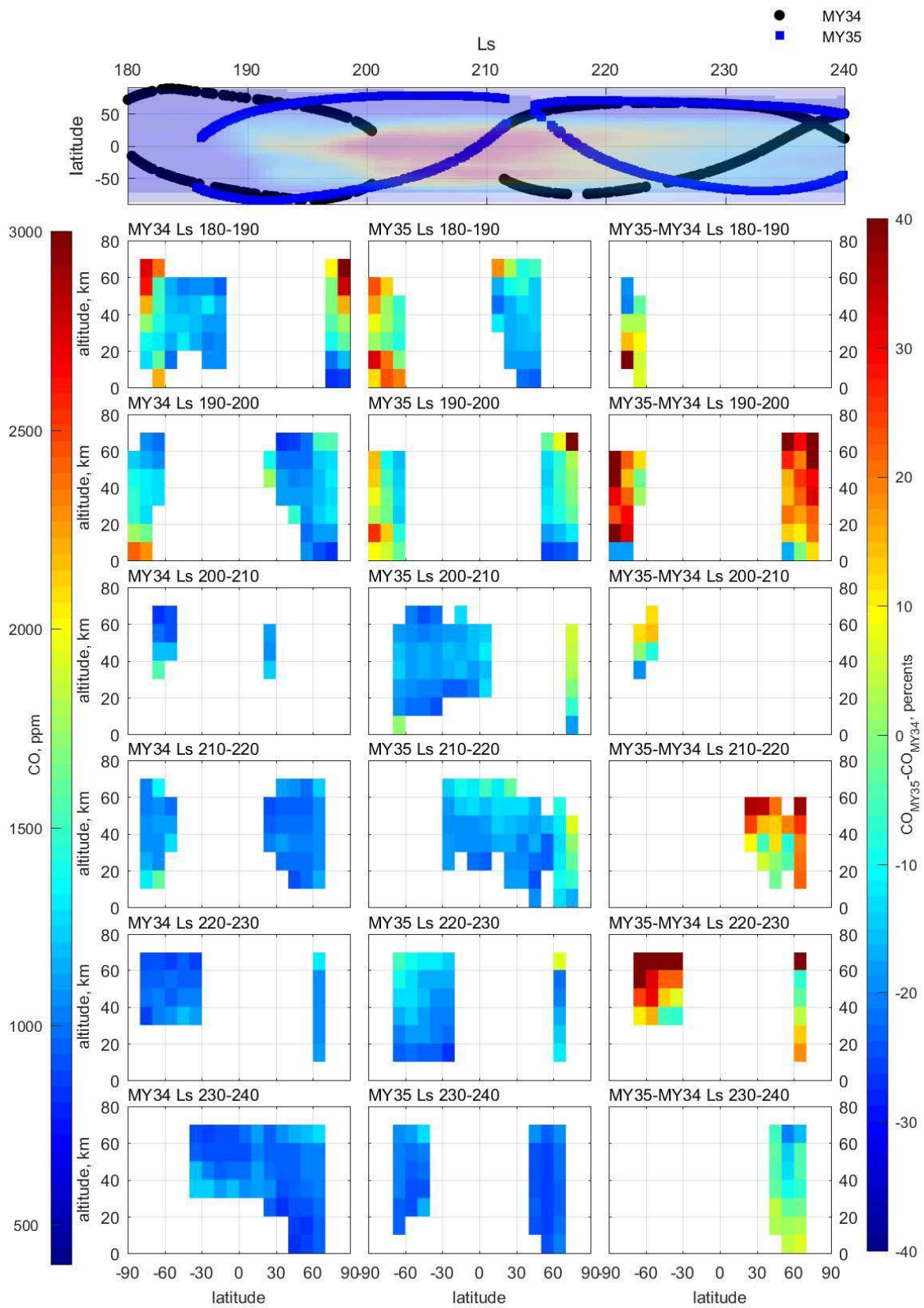
721 The first ever profiles of CO were produced by ACS MIR and reported for the period
722 from Ls=163° (pre-storm) to Ls=220° (decay phase), which showed how CO was affected by
723 this event (Olsen et al., 2021). After the onset of the dust storm, the mean CO mixing ratio was
724 observed to decrease from 1260 to 1070 ppmv below 40 km which was explained as the
725 enhanced presence of H₂O and thus OH that subsequently accelerated CO loss. The same
726 occurred at high altitudes where the rise of hygropause also resulted in the decrease of CO.

727 With two Martian years of observations, we can compare the CO vertical profiles during
728 the GDS period of MY 34 with the more regular MY 35 using ACS NIR data. The panels of
729 Figure 7 at Ls=180°–210° clearly show more CO in the high southern latitudes during MY 35
730 than during MY 34. On the other hand, CO in high northern latitudes at altitudes above 40 km is
731 more abundant in MY 34. Over Ls=210°–240°, the observed CO mixing ratios between 30 and
732 60 km in low latitudes (30°S–30°N) is higher in MY 35 compared to MY 34. As shown in
733 Figure 1, solar occultations in different years can have different latitudinal distributions and the
734 observations can relate to different Ls. To exclude these uncertainties, we have selected smaller
735 latitude bins than in Figure 7 (10° of Ls versus 30° of Ls) to compare nearly close in Ls and
736 latitude observations for two years (Figure 12). We average the ACS dataset inside 10° of Ls,

737 10° of latitude and 10 km of altitude. In Figure 12 the CO values for the selected bins are shown
738 as altitude-latitude maps for MY 34 and MY 35 together with their relative difference, where the
739 Ls-latitude coverage permits.

740

741



742

743 Figure 12. The latitudinal trend of the CO mixing ratio vertical profiles in ppmv into 10° Ls bins

744 from Ls=180° to 240° for MY34 (left) and MY35 (middle) and their relative difference in

745 percent (right). On the top panel is the seasonal distribution of ACS NIR solar occultation
746 observations for MY 34 and 35 at $L_s=180\text{--}240^\circ$. The colored background is the dust optical
747 depth distribution in $9\text{ }\mu\text{m}$ for MY 34 from Montabone et al. (2020) with red color corresponding
748 to maximal optical depth and blue color corresponding to minimal.

749

750 For the first L_s bin ($180^\circ\text{--}190^\circ$) the data from the two years overlap only in high southern
751 latitudes just before the beginning of GDS and the difference near 60°S does not exceed 10%.
752 From $L_s=190^\circ$ to 230° , during the maximal dust loading, the CO mixing ratio in MY 35 is larger
753 than in MY 34 of 30–40% both in the northern and southern hemispheres. On the decay of GDS,
754 at $L_s=230^\circ\text{--}240^\circ$, the observations overlap in middle northern latitudes, and the difference
755 between two years was within 10% again. This depletion of the CO mixing ratio is well
756 supported by LMD GCM modeling for MY 34 and 35 with corresponding dust scenario (see
757 Figure S15). The decrease of the CO mixing ratio in MY 34 during the GDS is also anti-
758 correlated with increase of water vapor abundance in the atmosphere in both hemispheres for the
759 same period (see Figure S16). The notable CO depletion during the GDS of MY 34 can indicate
760 the response of CO_2 production rate to the strong increase in water vapor during this period (see.,
761 e.g. Fedorova et al., 2020).

762

763 **Conclusions**

764 We analyzed a large dataset of CO vertical profiles made using solar occultation
765 observations of the three ACS spectrometers onboard the ExoMars Trace Gas Orbiter. The data
766 has been collected from April 2018 ($L_s=163^\circ$ of MY 34) to March 2021 ($L_s=360^\circ$ of MY 35)
767 and covers 1.5 Martian years. The densest dataset comes from the ACS NIR spectrometer, which
768 collected more than 6200 occultations over this period. For this dataset, we made the
769 simultaneous retrieval of vertical profiles of pressure, temperature, and CO mixing ratio over
770 altitudes from 0 to 60 km, and in the spectral range of $1.582\text{--}1.567\text{ }\mu\text{m}$ (order 49 of the

spectrometer), which includes the 1.57 μm CO_2 band and the CO (3-0) overtone at 1.57 μm . ACS TIRVIM provides the second densest dataset of about 1000 occultations, but these observations stopped in December 2019. The CO mixing ratio was retrieved based on the strong 4.7 μm CO absorption band and the CO_2 density from 4.9 μm band. The low SNR and spectral resolution limit the sounding altitude to below 50 km. ACS MIR, the most sensitive channel of ACS, has performed 620 occultations up to the end of MY 35 and provided the measurements of CO VMR vertical profiles in 2.3 μm absorption band based on CO_2 density and temperature retrieved from simultaneous measurements of ACS NIR in the range of altitude from 0 to 80 km.

With this dataset we presented the first detailed analysis of CO vertical distribution over a long timescale and studied seasonal and latitudinal variations during a Martian year:

- 1) We found a mean CO VMR of ~ 960 ppmv in the range of 10-35 km and in the low and middle latitudes (from 45°S to 45°N). Our averaged values are higher than the CRISM, NOMAD and PFS/MEX global averages from nadir measurements, all three being of ~ 800 ppmv (Smith et al., 2018; Smith et al., 2021; Bouche et al., 2021). This difference can be related to different sensitivity of nadir and occultation measurements. The CO mean value shows a repeatable seasonal cycle from MY 34 to MY 35 with two maxima, the first and largest of the two at $\text{Ls} \sim 180^\circ$ and the second shortly after spring equinox ($\text{Ls} = 0^\circ$) in the Northern hemisphere.
- 2) We found a strong enrichment of CO near the surface during the southern winter and spring ($\text{Ls} = 100\text{--}200^\circ$) in middle and high southern latitudes with a layer of 3000-4000 ppm at 10-20 km corresponding to local depletion of CO_2 . In the Northern polar winter atmosphere, such a strong layer of CO was not observed. The GCM does not predict the enrichment in CO in this period, both in terms of absolute value and vertical extent. This indicates that the breakup of the polar vortex enriched in CO (and in other non-condensable species) and the subsequent mixing with mid-latitude air occur too early with the settings of the LMD GCM used here.

- 797 3) In the equinox seasons both in the northern and southern spring, we found an increase of
798 CO mixing ratio above 50 km to 3000–4000 ppmv which is related to the downwelling
799 flux of the equinox Hadley circulation on Mars above the Poles enriched of CO
800 molecules. The comparison with the general circulation chemical model has shown that it
801 tends to overestimate the intensity of this process, bringing too much CO from its region
802 of production in the high atmosphere.
- 803 4) The minimum of CO observed in the southern summer in the high and middle southern
804 latitudes has average VMRs of 700–750 ppmv in the low atmosphere and agrees well
805 with nadir measurements by CRISM/MRO and PFS/MEX even they have lower values
806 of 400-700 ppmv and ~600 ppmv, respectively, in the same period that can be also
807 explained by different altitude sensitivity of nadir and occultation observations The CO
808 profiles calculated in southern summer by the LMD GCM are in broad agreement with
809 ACS.
- 810 5) The observations during the two Martian years allow us to study the interannual
811 variability for a year with a global dust storm (MY 34) and a year without (MY 35). We
812 observed the depletion of the CO mixing ratio at 30-40% both in the northern and
813 southern hemispheres during the global dust storm of MY 34 compared with the calmer
814 MY 35. The decrease of the CO mixing ratio in MY 34 can indicate the response of CO₂
815 production rate to the increase of water vapor abundance in the atmosphere for the same
816 period that suggests an impact of HO_x chemistry on the CO abundance.

817

818 **Acknowledgments**

819 The ExoMars mission is a joint mission of the European Space Agency (ESA) and
820 Roscosmos. The ACS experiment is led by the Space Research Institute (IKI) in Moscow,
821 assisted by LATMOS in France. The science operations of ACS are funded by Roscosmos and
822 ESA. Authors affiliated with IKI acknowledge funding from the Ministry of Science and High

823 Education of Russia. Authors affiliated with LATMOS acknowledge funding from CNES and
824 Centre National de la Recherche Scientifique (CNRS). Authors affiliated with the University of
825 Oxford acknowledge funding from the UK Space Agency under grants ST/T002069/1,
826 ST/R001502/1 and ST/P001572/1. We thank our reviewers for helpful comments that improved
827 the manuscript a lot.

828

829 **Data availability.**

830 ACS data are available from ESA Planetary Science Archive (PSA) Level 2
831 <https://archives.esac.esa.int/psa/#!/Table%20View/ACS=instrument>. The CO vertical profiles
832 generated from ACS measurements and analyzed in this study are available at Fedorova (2022).

833

834

835 **Reference**

836

837 Alday, J., Trokhimovskiy, A., Irwin, P. G. J., Wilson, C. F., Montmessin, F., Lefèvre, F., et al. (2021).

838 Isotopic fractionation of water and its photolytic products in the atmosphere of Mars. *Nature*

839 *Astronomy*, 5(9), 943–950. <https://doi.org/10.1038/s41550-021-01389-x>

840 Alday, J., Wilson, C. F., Irwin, P. G. J., Olsen, K. S., Baggio, L., Montmessin, F., et al. (2019). Oxygen

841 isotopic ratios in Martian water vapour observed by ACS MIR on board the ExoMars Trace Gas

842 Orbiter. *Astronomy & Astrophysics*, 630, A91. <https://doi.org/10.1051/0004-6361/201936234>

843 Aoki, S., Vandaele, A. C., Daerden, F., Villanueva, G. L., Liuzzi, G., Thomas, I. R., et al. (2019). Water

844 Vapor Vertical Profiles on Mars in Dust Storms Observed by TGO/NOMAD. *Journal of*

845 *Geophysical Research: Planets*, 124(12), 3482–3497. <https://doi.org/10.1029/2019je006109>

846 Belyaev, D. A., Fedorova, A. A., Trokhimovskiy, A., Alday, J., Montmessin, F., Korablev, O. I., et al.

847 (2021). Revealing a High Water Abundance in the Upper Mesosphere of Mars With ACS

848 Onboard TGO. *Geophysical Research Letters*, 48(10), e2021GL093411.

849 <https://doi.org/10.1029/2021GL093411>

850 Billebaud, F., Brillet, J., Lellouch, E., Fouchet, T., Encrenaz, T., Cottini, V., et al. (2009). Observations of

851 CO in the atmosphere of Mars with PFS onboard Mars Express. *Planet. Space Sci.*, 57, 1446–

852 1457. <https://doi.org/10.1016/j.pss.2009.07.004>

853 Billebaud, F., Maillard, J. P., Lellouch, E., & Encrenaz, T. (1992). The spectrum of Mars in the (1-0)

854 vibrational band of CO. *Astronomy and Astrophysics*, 261, 647–657.

855 Billebaud, F., Rosenqvist, J., Lellouch, E., Maillard, J.-P., Encrenaz, T., & Hourdin, F. (1998).

856 Observations of CO in the atmosphere of Mars in the (2-0) vibrational band at 2.35 microns.

857 *Astronomy and Astrophysics*, 333, 1092–1099.

858 Bouche, J., Bauduin, S., Giuranna, M., Robert, S., Aoki, S., Vandaele, A. C., et al. (2019). Retrieval and

859 characterization of carbon monoxide (CO) vertical profiles in the Martian atmosphere from

860 observations of PFS/MEX. *Journal of Quantitative Spectroscopy and Radiative Transfer*, 238,

861 106498. <https://doi.org/10.1016/j.jqsrt.2019.05.009>

862 Bouche, J., Coheur, P.-F., Giuranna, M., Wolkenberg, P., Nardi, L., Amoroso, M., ... Bauduin, S. (2020, April

863 15). Data: “Seasonal and spatial variability of carbon monoxide (CO) in the Martian atmosphere from

864 PFS/MEX observations.” Retrieved from [https://osf.io/xsknv/?view_only =](https://osf.io/xsknv/?view_only=47d671dd84d84fe4b8802c726d293b4c)
865 47d671dd84d84fe4b8802c726d293b4c.

866 Bouche, J., Coheur, P.-F., Giuranna, M., Wolkenberg, P., Nardi, L., Amoroso, M., et al. (2021). Seasonal
867 and Spatial Variability of Carbon Monoxide (CO) in the Martian Atmosphere From PFS/MEX
868 Observations. *Journal of Geophysical Research: Planets*, 126(2), e2020JE006480.
869 <https://doi.org/10.1029/2020JE006480>

870 Ceccherini, S., (2005). Analytical determination of the regularization parameter in the retrieval of
871 atmospheric vertical profiles, *Opt. Lett.* 30, 2554-2556.

872 Clancy, R. T., Muhleman, D. O., & Berge, G. L. (1990). Global changes in the 0–70 km thermal structure
873 of the Mars atmosphere derived from 1975 to 1989 microwave CO spectra. *Journal of*
874 *Geophysical Research: Solid Earth*, 95(B9), 14543–14554.
875 <https://doi.org/10.1029/JB095iB09p14543>

876 Daerden, F., Neary, L., Viscardy, S., García Muñoz, A., Clancy, R. T., Smith, M. D., et al. (2019). Mars
877 atmospheric chemistry simulations with the GEM-Mars general circulation model. *Icarus*, 326,
878 197–224. <https://doi.org/10.1016/j.icarus.2019.02.030>.

879 de Kok, R.J., Stam, D.M. (2012). The influence of forward-scattered light in transmission measurements
880 of (exo)planetary atmospheres. *Icarus*, 221(2), 517-524.
881 <https://doi.org/10.1016/j.icarus.2012.08.020>.

882 Encrenaz, T., Fouchet, T., Melchiorri, R., Drossart, P., Gondet, B., Langevin, Y., et al. (2006). Seasonal
883 variations of the martian CO over Hellas as observed by OMEGA/Mars Express. *Astron.*
884 *Astrophys.*, 459, 265–270. <https://doi.org/10.1051/0004-6361:20065586>

885 Fedorova, A., Bertaux, J. L., Betsis, D., Montmessin, F., Korablev, O., Maltagliati, L., & Clarke, J.
886 (2018). Water vapor in the middle atmosphere of Mars during the 2007 global dust storm. *Icarus*,
887 300, 440–457. <https://doi.org/10.1016/j.icarus.2017.09.025>

888 Fedorova, A. A., Montmessin, F., Korablev, O., Luginin, M., Trokhimovskiy, A., Belyaev, D. A., et al.
889 (2020). Stormy water on Mars: The distribution and saturation of atmospheric water during the
890 dusty season. *Science*, 367(6475), 297–300. <https://doi.org/10.1126/science.aay9522>

891 Fedorova, A., Montmessin, F., Korablev, O., Lefèvre, F., Trokhimovskiy, A., & Bertaux, J.-L. (2021).
892 Multi-Annual Monitoring of the Water Vapor Vertical Distribution on Mars by SPICAM on Mars

Express. *Journal of Geophysical Research: Planets*, 126(1), e2020JE006616.

<https://doi.org/10.1029/2020JE006616>

Fedorova, A. (2022). “Carbon monoxide (CO) on Mars from ACS occultations (MY34-35)”, Mendeley Data, V2, doi: 10.17632/w2b89yr6x5.2.

Forget, F., Hourdin, F., Fournier, R., Hourdin, C., Talagrand, O., Collins, M., et al. (1999). Improved general circulation models of the Martian atmosphere from the surface to above 80 km. *J. Geophys. Res.*, 104, 24155–24176. <https://doi.org/10.1029/1999JE001025>

Forget, F., Millour, E., Montabone, L., & Lefevre, F. (2008). Non Condensable Gas Enrichment and Depletion in the Martian Polar Regions, 1447, 9106. Presented at the Third International Workshop on The Mars Atmosphere: Modeling and Observations.

Franz, H. B., Trainer, M. G., Wong, M. H., Mahaffy, P. R., Atreya, S. K., Manning, H. L. K., & Stern, J. C. (2015). Reevaluated martian atmospheric mixing ratios from the mass spectrometer on the Curiosity rover. *Planetary and Space Science*, 109–110, 154–158.

<https://doi.org/10.1016/j.pss.2015.02.014>

Gordon, I. E., Rothman, L. S., Hill, C., Kochanov, R. V., Tan, Y., Bernath, P. F., et al. (2017). The HITRAN2016 molecular spectroscopic database. *J. Quant. Spectrosc. Radiat. Transfer*, 203, 3–69. <https://doi.org/10.1016/j.jqsrt.2017.06.038>

Guzewich, S. D., Lemmon, M., Smith, C. L., Martínez, G., Vicente-Retortillo, Á. de, Newman, C. E., et al. (2019). Mars Science Laboratory Observations of the 2018/Mars Year 34 Global Dust Storm. *Geophysical Research Letters*, 46(1), 71–79. <https://doi.org/10.1029/2018GL080839>

Hase, F., Wallace, L., McLeod, S.D., Harrison, J.J, Bernath, P.F. (2010). The ACE-FTS atlas of the infrared solar spectrum, *Journal of Quantitative Spectroscopy and Radiative Transfer*, 111 (4), 521-528. <https://doi.org/10.1016/j.jqsrt.2009.10.020>.

Heavens, N. G., Kleinböhl, A., Chaffin, M. S., Halekas, J. S., Kass, D. M., Hayne, P. O., et al. (2018). Hydrogen escape from Mars enhanced by deep convection in dust storms. *Nature Astronomy*, 2(2), 126–132. <https://doi.org/10.1038/s41550-017-0353-4>

Holmes, J. A., Lewis, S. R., Patel, M. R., & Smith, M. D. (2019). Global analysis and forecasts of carbon monoxide on Mars. *Icarus*, 328, 232–245. <https://doi.org/10.1016/j.icarus.2019.03.016>

Kaplan, L. D., Connes, J., & Connes, P. (1969). Carbon Monoxide in the Martian Atmosphere. *Astrophys.*

922 J., 157, L187. <https://doi.org/10.1086/180416>

923 Korablev, O., Montmessin, F., Trokhimovskiy, A., Fedorova, A. A., Shakun, A. V., Grigoriev, A. V., et
 924 al. (2018). The Atmospheric Chemistry Suite (ACS) of Three Spectrometers for the ExoMars
 925 2016 Trace Gas Orbiter. *Space Science Reviews*, 214(1), 7. [https://doi.org/10.1007/s11214-017-](https://doi.org/10.1007/s11214-017-0437-6)
 926 0437-6

927 Krasnopolsky, V. A. (2003). Spectroscopic mapping of Mars CO mixing ratio: Detection of north-south
 928 asymmetry. *J. Geophys. Res.*, 108, 5010. <https://doi.org/10.1029/2002JE001926>

929 Krasnopolsky, V. A. (2007). Long-term spectroscopic observations of Mars using IRTF/CSHELL:
 930 Mapping of O₂ dayglow, CO, and search for CH₄. *Icarus*, 190, 93–102.
 931 <https://doi.org/10.1016/j.icarus.2007.02.014>

932 Krasnopolsky, V. A. (2014). Observations of the CO dayglow at 4.7μm on Mars: Variations of
 933 temperature and CO mixing ratio at 50km. *Icarus*, 228, 189–196.
 934 <https://doi.org/10.1016/j.icarus.2013.10.008>

935 Krasnopolsky, V. A. (2015). Variations of carbon monoxide in the martian lower atmosphere. *Icarus*,
 936 253, 149–155. <https://doi.org/10.1016/j.icarus.2015.03.006>

937 Krasnopolsky, V. A. (2017). Annual mean mixing ratios of N₂, Ar, O₂, and CO in the martian
 938 atmosphere. *Planetary and Space Science*, 144, 71–73. <https://doi.org/10.1016/j.pss.2017.05.009>

939 Lefèvre, F., & Krasnopolsky, V. (2017). Atmospheric Photochemistry. In R. M. Haberle, R. T. Clancy, F.
 940 Forget, M. D. Smith, & R. W. Zurek (Eds.), *The Atmosphere and Climate of Mars* (pp. 405–432).
 941 Cambridge University Press.

942 Lefèvre, F., Lebonnois, S., Montmessin, F., & Forget, F. (2004). Three-dimensional modeling of ozone
 943 on Mars. *J. Geophys. Res.*, 109(E7), E07004. <https://doi.org/10.1029/2004JE002268>

944 Lefèvre, F., Trokhimovskiy, A., Fedorova, A., Baggio, L., Lacombe, G., Määttänen, A., et al. (2021).
 945 Relationship between the ozone and water vapor columns on Mars as observed by SPICAM and
 946 calculated by a global climate model. *Journal of Geophysical Research: Planets*, 126,
 947 e2021JE006838. <https://doi.org/10.1029/2021JE006838>

948 Lellouch, E., Encrenaz, T., Phillips, T., Falgarone, E., & Billebaud, F. (1991). Submillimeter observations
 949 of CO in Mars' atmosphere. *Planetary and Space Science*, 39(1), 209–212.
 950 [https://doi.org/10.1016/0032-0633\(91\)90143-X](https://doi.org/10.1016/0032-0633(91)90143-X)

951 Li, G., Gordon, I.E., Rothman L.S., Tan, Y., Hu, S.-M., Kass, S., Campargue, A. and Medvedev, E.S.
 952 (2015). «ROVIBRATIONAL LINE LISTS FOR NINE ISOTOPOLOGUES OF THE CO
 953 MOLECULE IN THE X 1 Σ + GROUND ELECTRONIC STATE». *The Astrophysical Journal*
 954 *Supplement Series* 216 (1): 15. <https://doi.org/10.1088/0067-0049/216/1/15>.

955 McElroy, M. B. and Donahue, T.B. (1972). Stability of the Martian Atmosphere. *Science* 177 (4053):
 956 986–88. <https://doi.org/10.1126/science.177.4053.986>.

957 Millour E., Francois Forget, Aymeric Spiga, Margaux Vals, Vladimir Zakharov, Luca Montabone, Franck
 958 Lefèvre, Franck Montmessin, Jean-Yves Chaufray, Miguel López-Valverde, Francisco González-
 959 Galindo, Stephen Lewis, Peter Read, Marie-Christine Desjean, and Fabrice Cipriani The Mars
 960 Climate Database (MCD version 5.3), Geophysical Research Abstracts, Vol. 21, EGU2019-7153,
 961 2019, EGU General Assembly 2019

962 Montabone, L., Spiga, A., Kass, D. M., Kleinböhl, A., Forget, F., & Millour, E. (2020). Martian Year 34
 963 Column Dust Climatology from Mars Climate Sounder Observations: Reconstructed Maps and
 964 Model Simulations. *J. Geophys. Res.*

965 Nair, H., Allen, M., Anbar, A. D., Yung, Y. L., & Clancy, R. T. (1994). A Photochemical Model of the
 966 Martian Atmosphere. *Icarus*, 111(1), 124–150. <https://doi.org/10.1006/icar.1994.1137>

967 Olsen, K. S., Lefèvre, F., Montmessin, F., Fedorova, A. A., Trokhimovskiy, A., Baggio, L., et al. (2021).
 968 The vertical structure of CO in the Martian atmosphere from the ExoMars Trace Gas Orbiter.
 969 *Nature Geoscience*, 14(2), 67–71. <https://doi.org/10.1038/s41561-020-00678-w>

970 Parkinson, T. D., and D. M. Hunten. (1972). Spectroscopy and Aeronomy of O₂ on Mars. *Journal of the*
 971 *Atmospheric Sciences* 29 (7): 1380–90. [https://doi.org/10.1175/1520-0469\(1972\)029<1380:saaooo>2.0.co;2](https://doi.org/10.1175/1520-0469(1972)029<1380:saaooo>2.0.co;2).
 972

973 Shakun, A., Ignatiev, N., Luginin, M., Grigoriev, A., Moshkin, B., Grassi, D., Arnold, G. (2018).
 974 ACS/TIRVIM: Calibration and first results, Proc. SPIE 10765, Infrared Remote Sensing and
 975 Instrumentation XXVI, 107650E, <https://d.docs.live.net/1416dfd2d615e64f/>
 976 <https://doi.org/10.1117/12.2322163>.

977 Sindoni, G., Formisano, V., & Geminale, A. (2011). Observations of water vapour and carbon monoxide
 978 in the Martian atmosphere with the SWC of PFS/MEX. *Planet. Space Sci.*, 59, 149–162.
 979 <https://doi.org/10.1016/j.pss.2010.12.006>

980 Smith, M. D. (2019). THEMIS Observations of the 2018 Mars Global Dust Storm. *Journal of*

981 *Geophysical Research: Planets*, 124(11), 2929–2944. <https://doi.org/10.1029/2019je006107>

982 Smith, M. D., Daerden, F., Neary, L., & Khayat, A. S. J. (2018). The climatology of carbon monoxide
 983 and water vapor on Mars as observed by CRISM and modeled by the GEM-Mars general
 984 circulation model. *Icarus*, 301, 117–131. <https://doi.org/10.1016/j.icarus.2017.09.027>

985 Smith, M. D., Daerden, F., Neary, L., Khayat, A. S. J., Holmes, J. A., Patel, M. R., et al. (2021). The
 986 climatology of carbon monoxide on Mars as observed by NOMAD nadir-geometry observations.
 987 *Icarus*, 362, 114404. <https://doi.org/10.1016/j.icarus.2021.114404>

988 Smith, M. D., Wolff, M. J., Clancy, R. T., & Murchie, S. L. (2009). Compact Reconnaissance Imaging
 989 Spectrometer observations of water vapor and carbon monoxide. *J. Geophys. Res.*, 114, E00D03.
 990 <https://doi.org/10.1029/2008JE003288>

991 Sprague, A. L., Boynton, W. V., Forget, F., Lian, Y., Richardson, M., Starr, R., et al. (2012). Interannual
 992 similarity and variation in seasonal circulation of Mars’ atmospheric Ar as seen by the Gamma
 993 Ray Spectrometer on Mars Odyssey. *Journal of Geophysical Research: Planets*, 117(E4).
 994 <https://doi.org/10.1029/2011JE003873>

995 Sprague, A. L., Boynton, W. V., Kerry, K. E., Janes, D. M., Hunten, D. M., Kim, K. J., et al. (2004).
 996 Mars’ South Polar Ar Enhancement: A Tracer for South Polar Seasonal Meridional Mixing.
 997 *Science*, 306(5700), 1364–1367. <https://doi.org/10.1126/science.1098496>

998 Sprague, A. L., Boynton, W. V., Kerry, K. E., Janes, D. M., Kelly, N. J., Crombie, M. K., et al. (2007).
 999 Mars’ atmospheric argon: Tracer for understanding Martian atmospheric circulation and
 1000 dynamics. *Journal of Geophysical Research: Planets*, 112(E3).
 1001 <https://doi.org/10.1029/2005JE002597>

1002 Trainer, M. G., Wong, M. H., McConnochie, T. H., Franz, H. B., Atreya, S. K., Conrad, P. G., et al.
 1003 (2019). Seasonal Variations in Atmospheric Composition as Measured in Gale Crater, Mars.
 1004 *Journal of Geophysical Research: Planets*, 124(11), 3000–3024.
 1005 <https://doi.org/10.1029/2019JE006175>

1006 Trokhimovskiy, A., Korablev, O., Kalinnikov, Y. K., Fedorova, A., Stepanov, A. V., Titov, A. Y., et al.
 1007 (2015a). Near-infrared echelle-AOTF spectrometer ACS-NIR for the ExoMars Trace Gas
 1008 Orbiter. In *Infrared Remote Sensing and Instrumentation XXIII* (Vol. 9608, pp. 62–70). SPIE.
 1009 <https://doi.org/10.1117/12.2190369>

1010 Trokhimovskiy, A., Korablev, O., Ivanov, Y. S., Siniyavsky, I. I., Fedorova, A., Stepanov, A. V., et al.
1011 (2015b). Middle-infrared echelle cross-dispersion spectrometer ACS-MIR for the ExoMars Trace
1012 Gas Orbiter. In *Infrared Remote Sensing and Instrumentation XXIII* (Vol. 9608, pp. 55–61).
1013 SPIE. <https://doi.org/10.1117/12.2190359>
1014 Trokhimovskiy, A., Perevalov, V., Korablev, O., Fedorova, A. F., Olsen, K. . S., Bertaux, J.-L., et al.
1015 (2020). First observation of the magnetic dipole CO₂ main isotopologue absorption band at 3.3
1016 μm in the atmosphere of Mars by ACS. *Astron. Astrophys. in Press*. [https://doi.org/10.1051/0004-](https://doi.org/10.1051/0004-6361/202038134)
1017 [6361/202038134](https://doi.org/10.1051/0004-6361/202038134)
1018 Vandaele, A. C., Lopez-Moreno, J.-J., Patel, M. R., Bellucci, G., Daerden, F., Ristic, B., et al. (2018).
1019 NOMAD, an Integrated Suite of Three Spectrometers for the ExoMars Trace Gas Mission:
1020 Technical Description, Science Objectives and Expected Performance. *Space Sci. Rev.*, 214, 80.
1021 <https://doi.org/10.1007/s11214-018-0517-2>
1022

Figure 1.

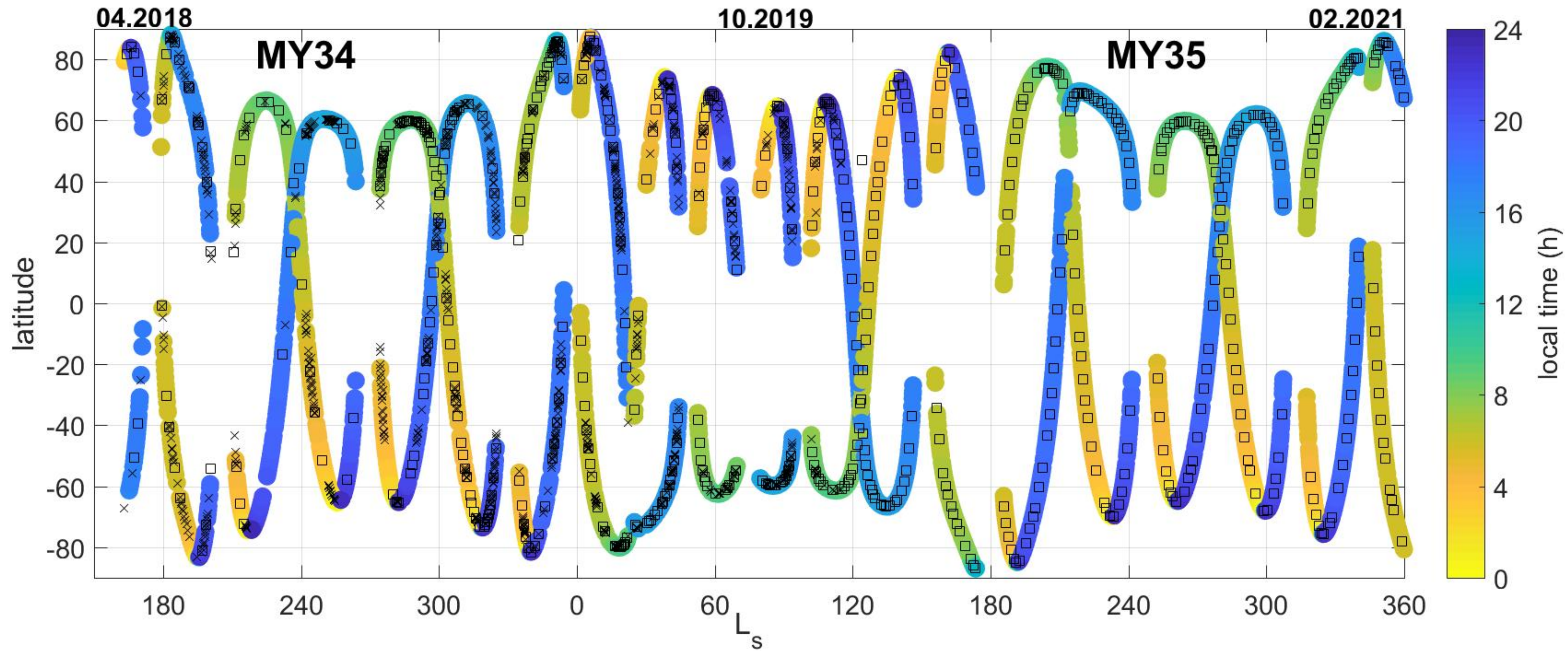


Figure 2.

ORDER 49

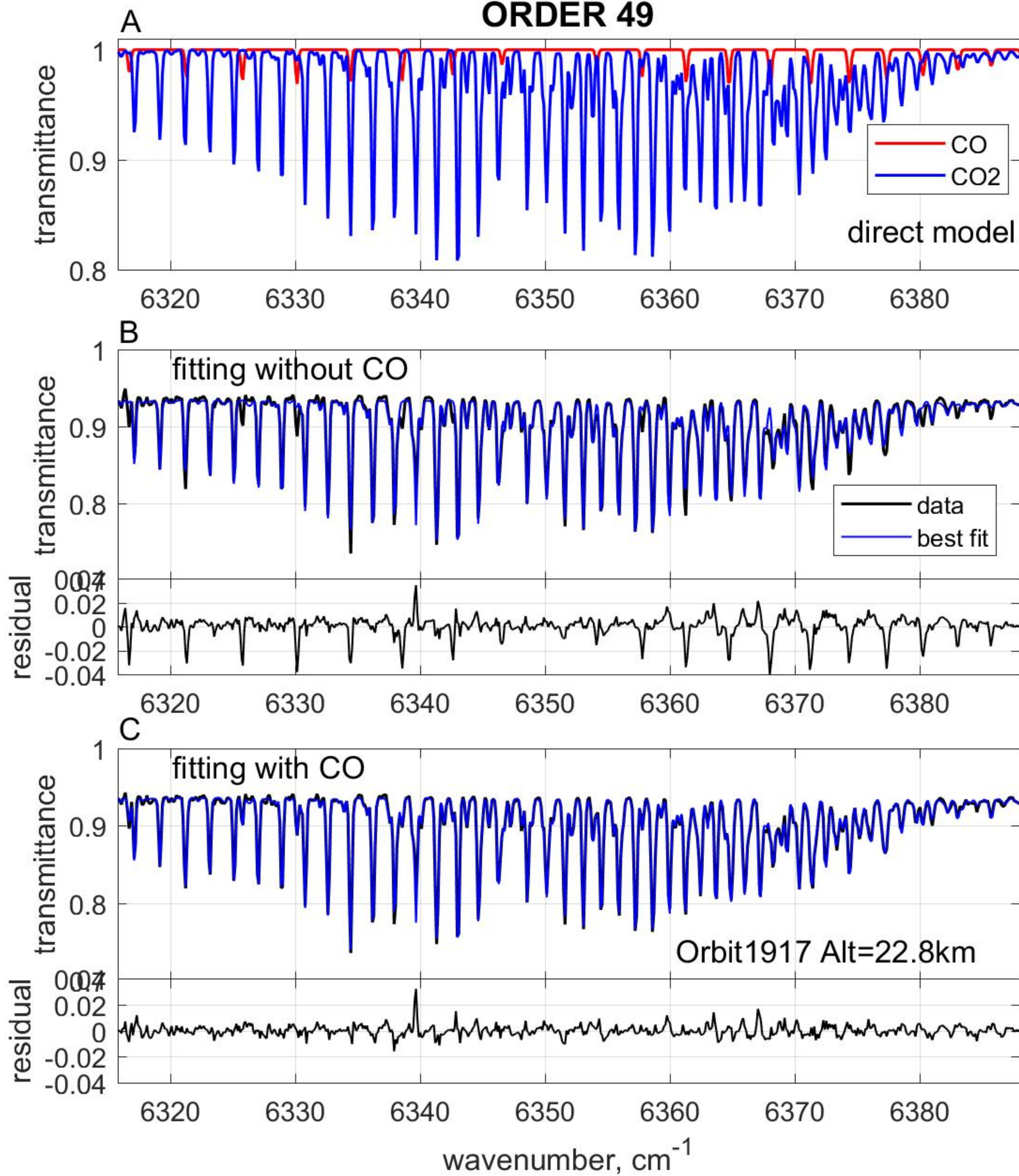


Figure 3.

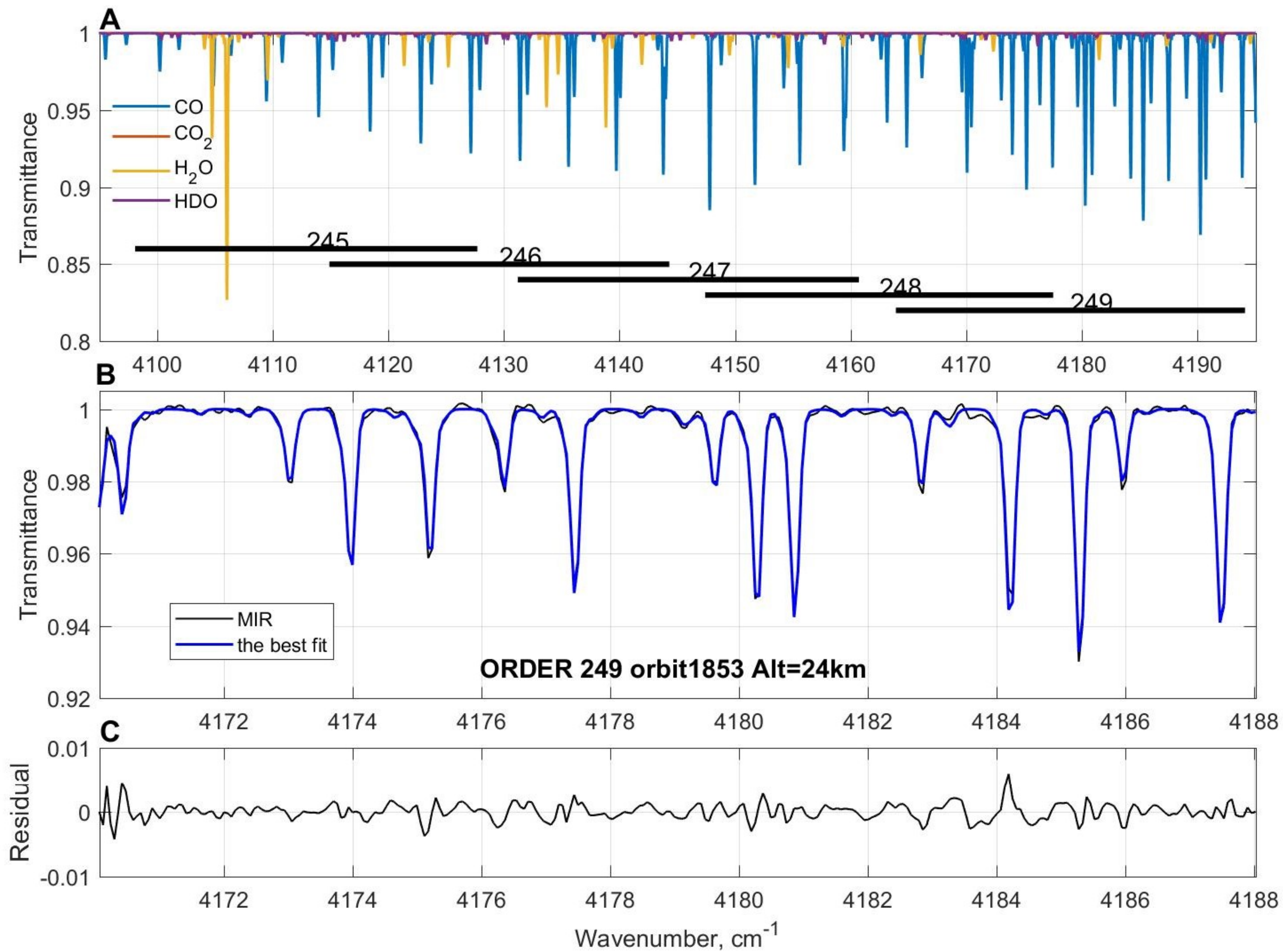


Figure 4.

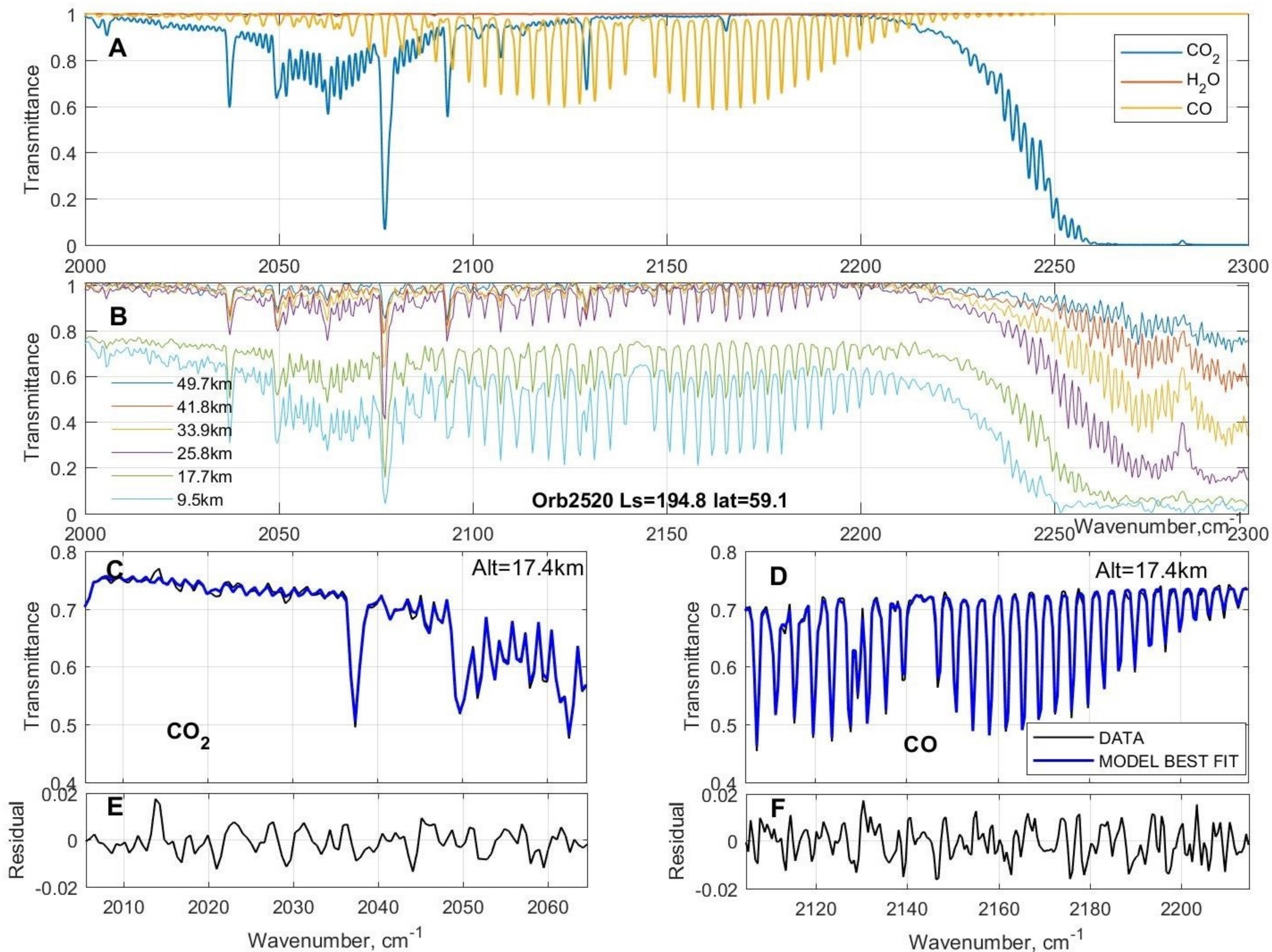


Figure 5.

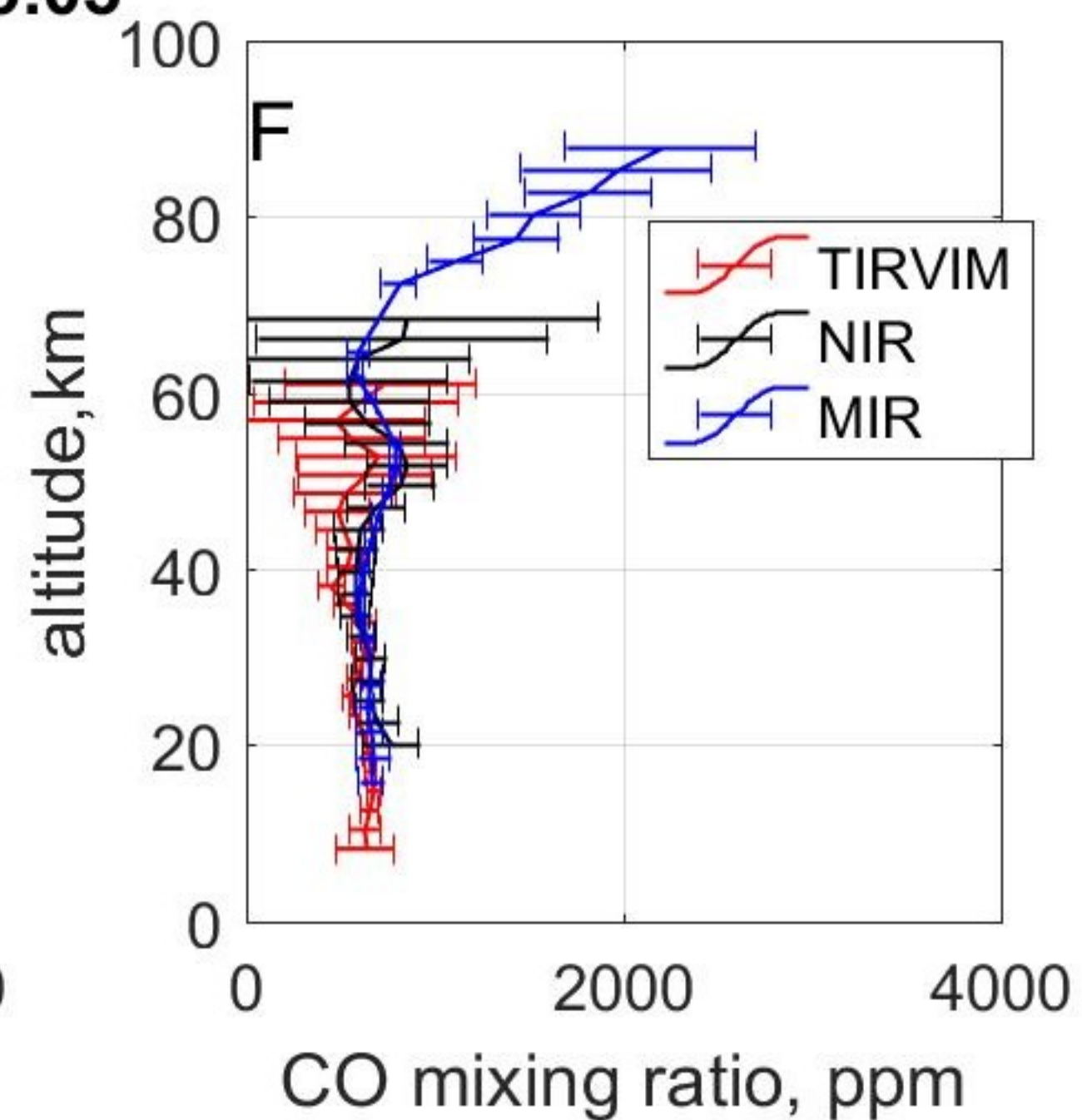
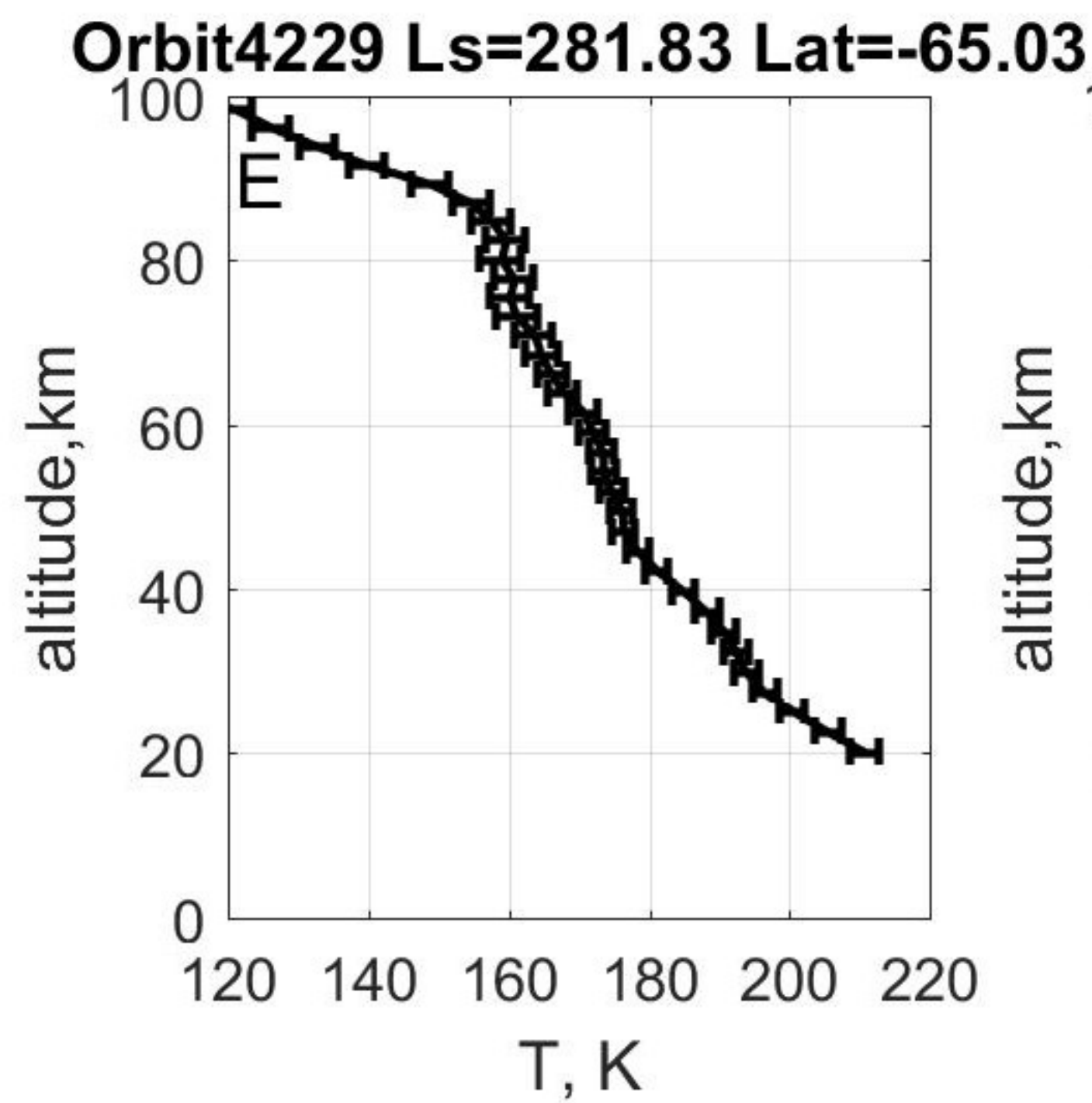
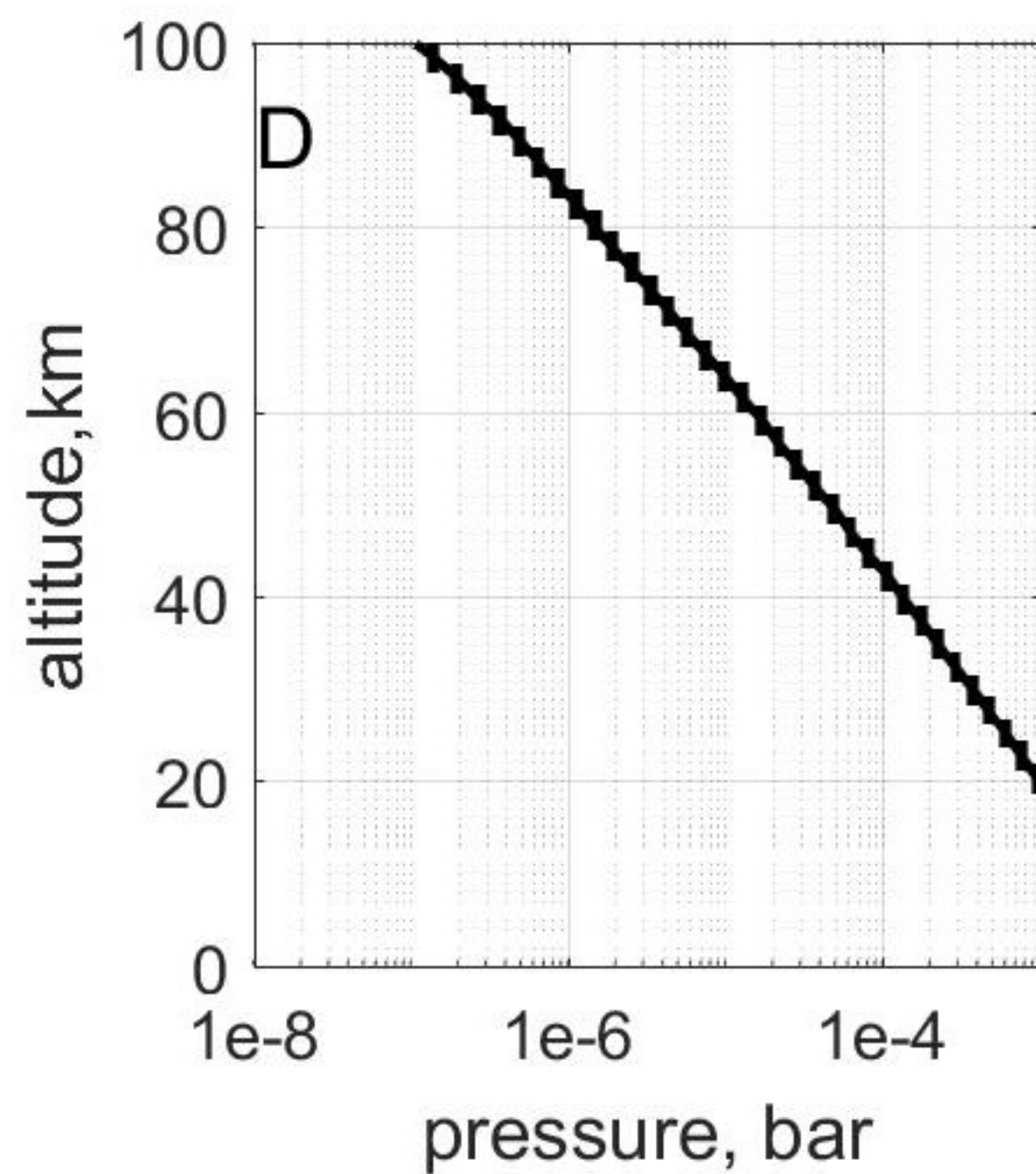
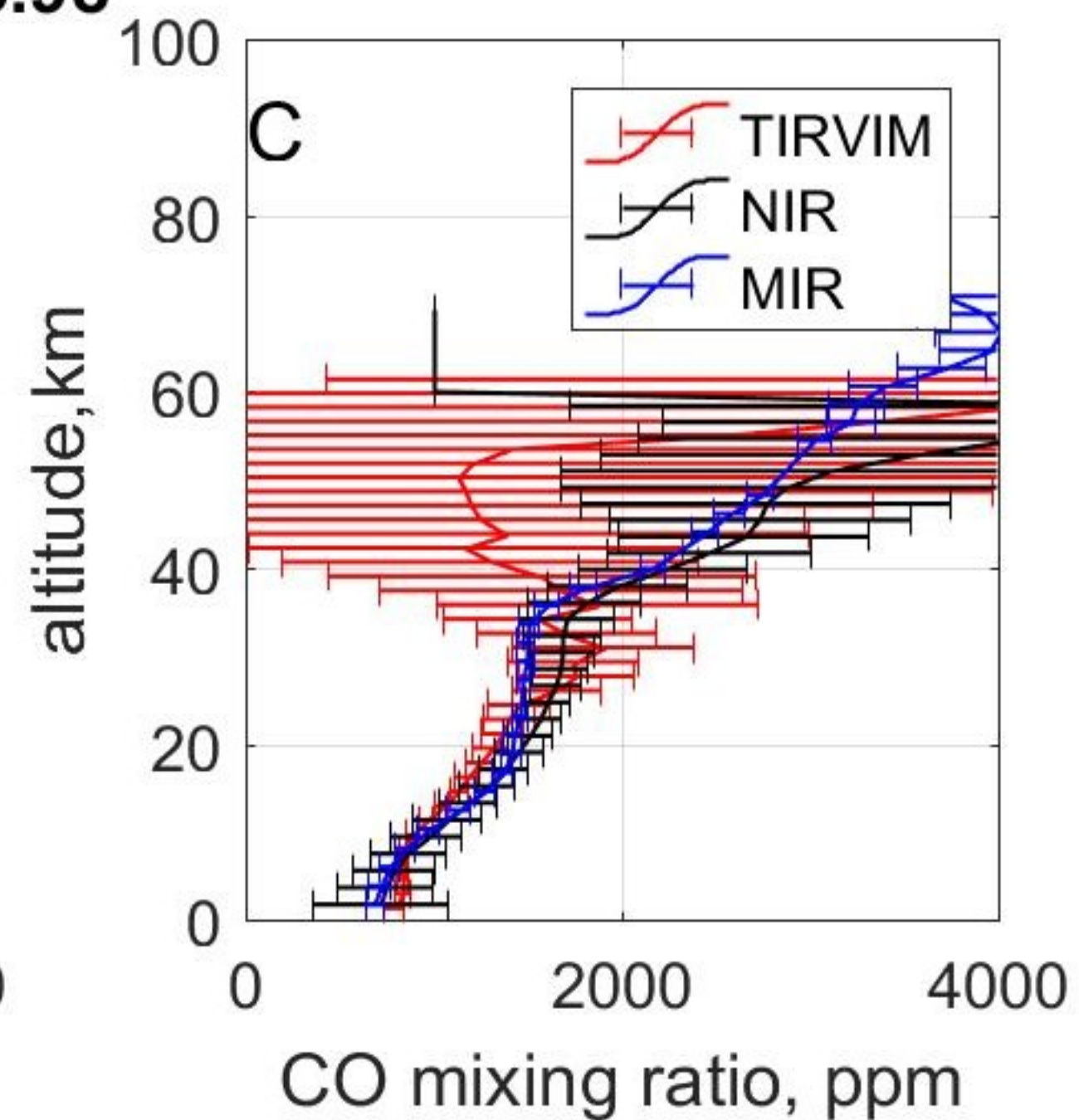
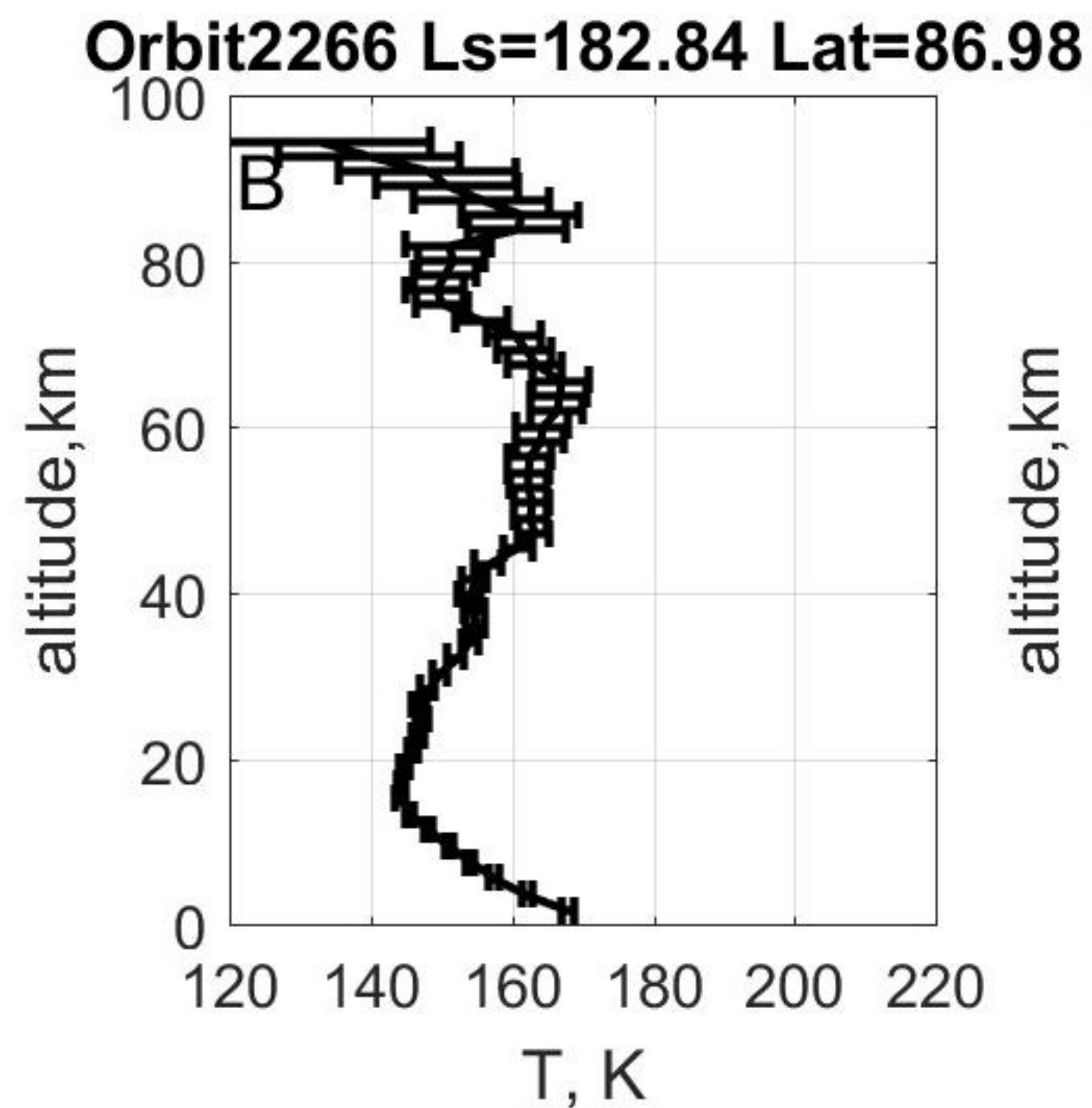
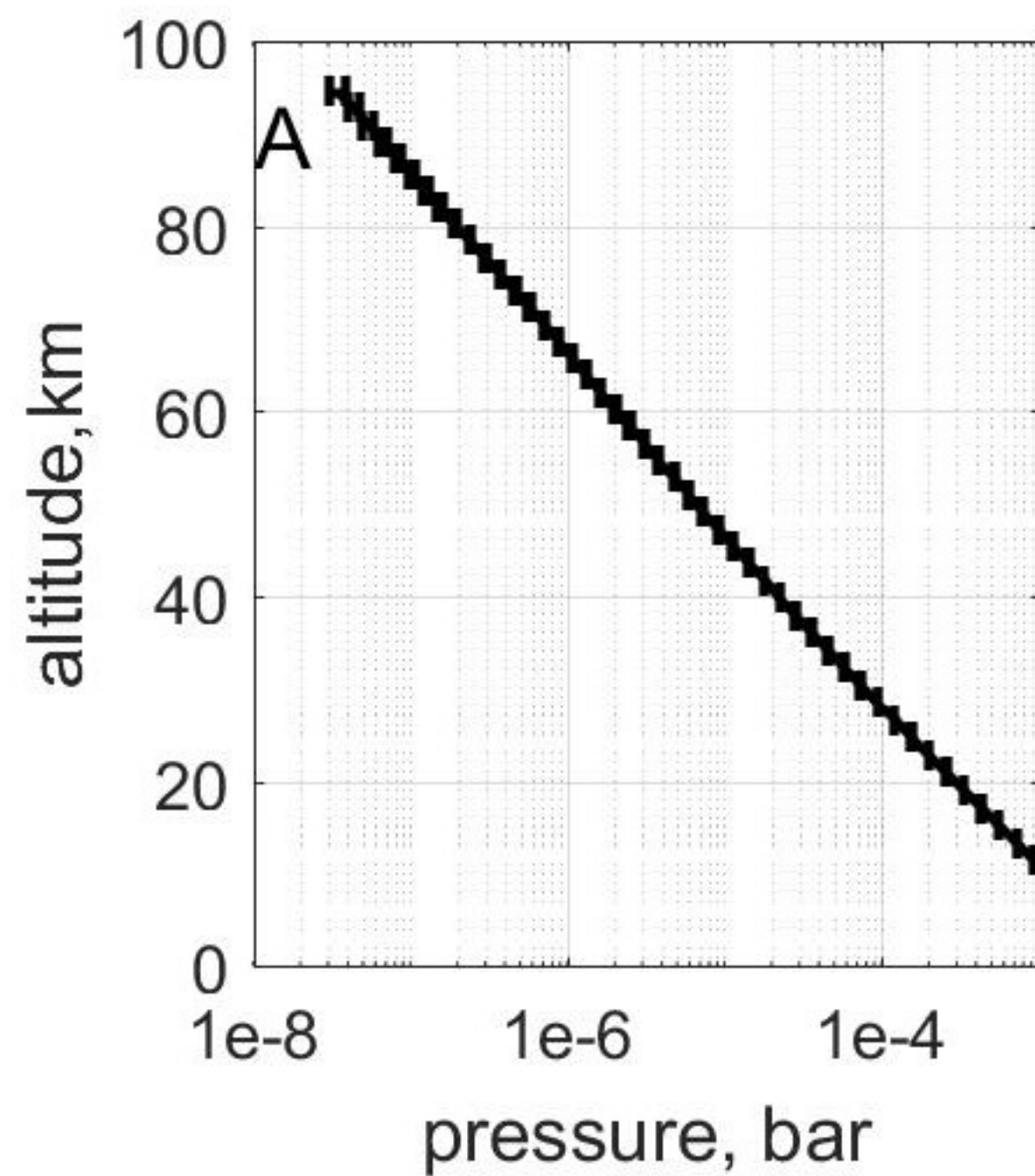


Figure 6.

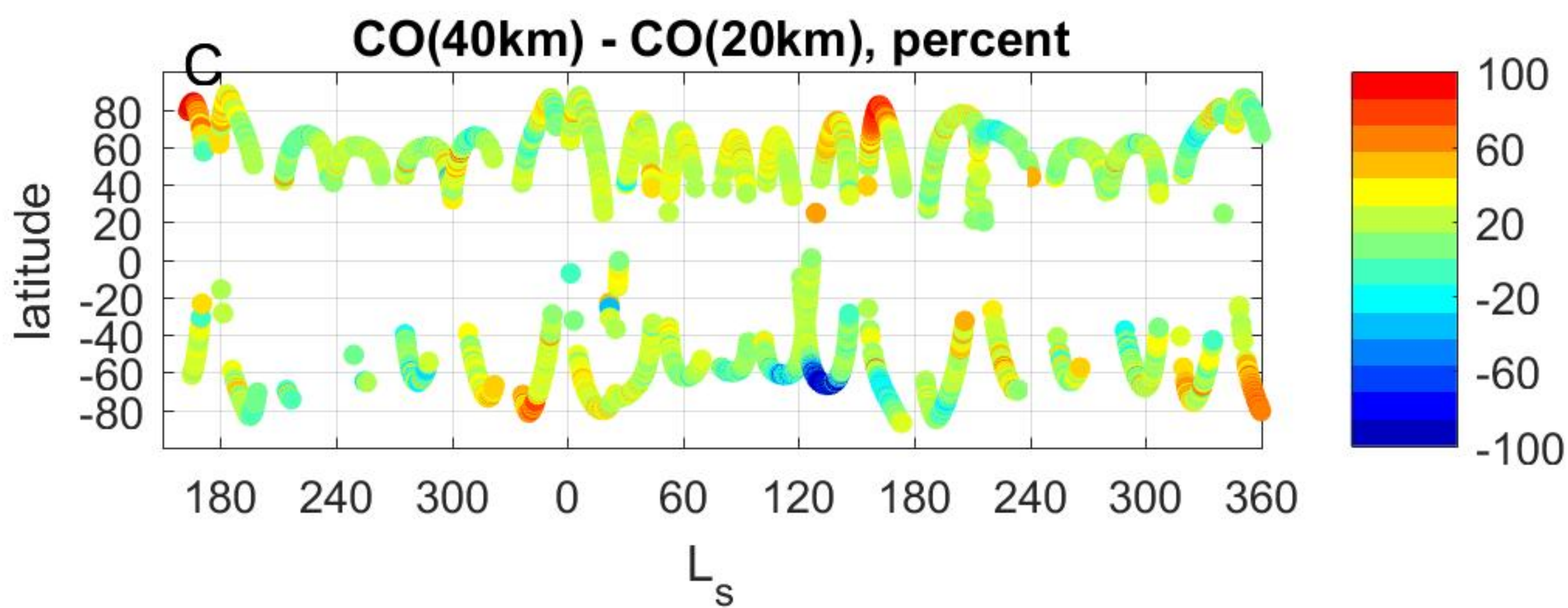
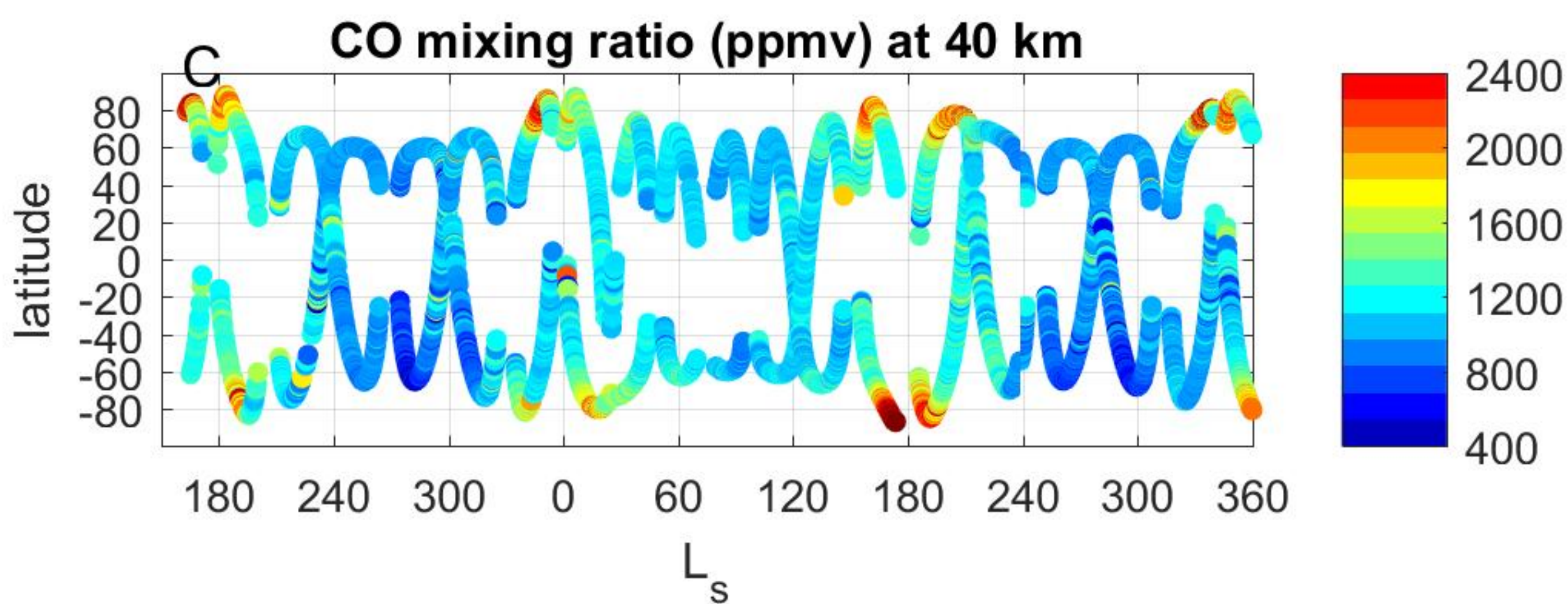
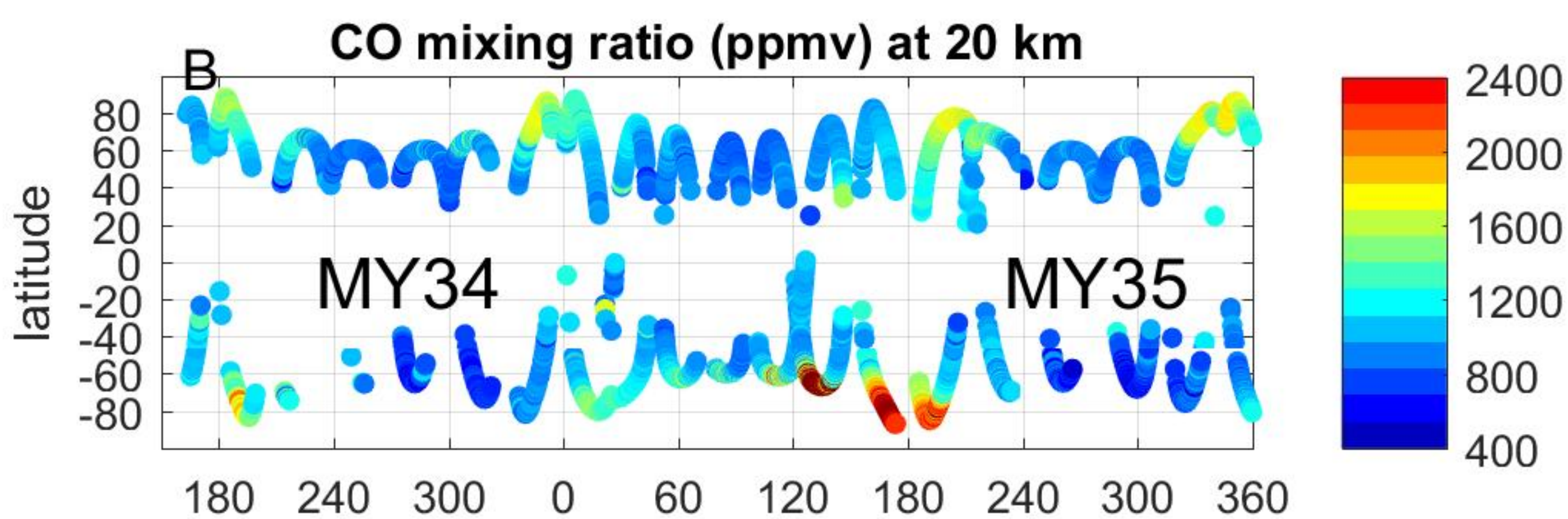
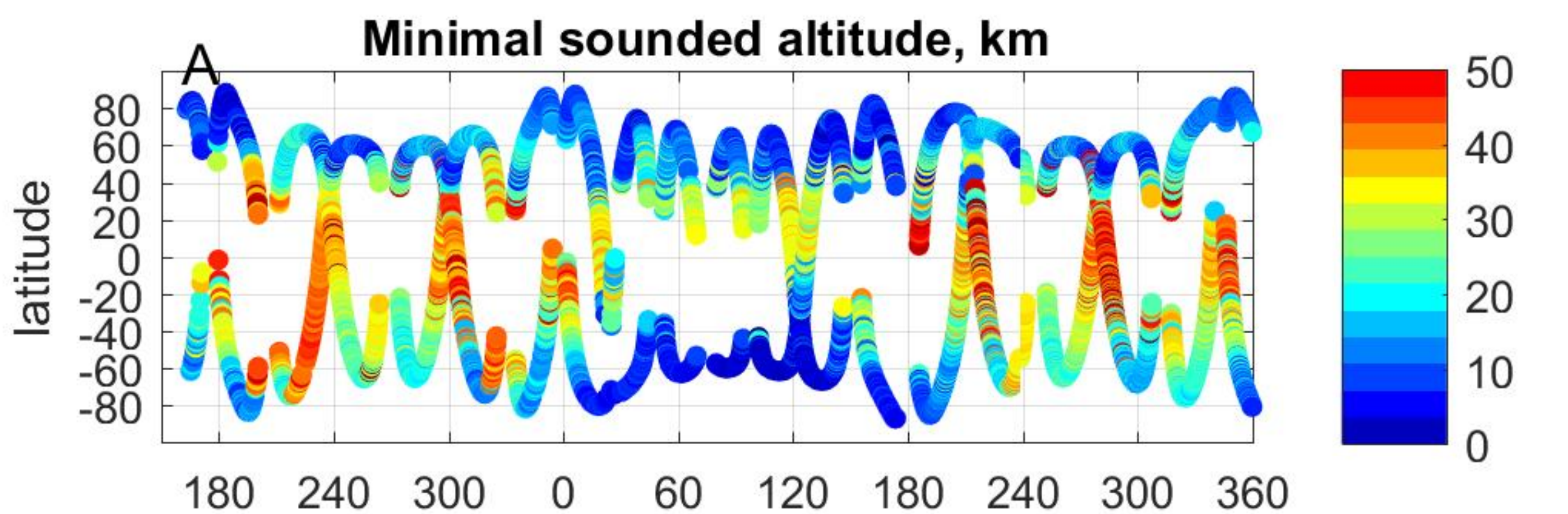


Figure 7.

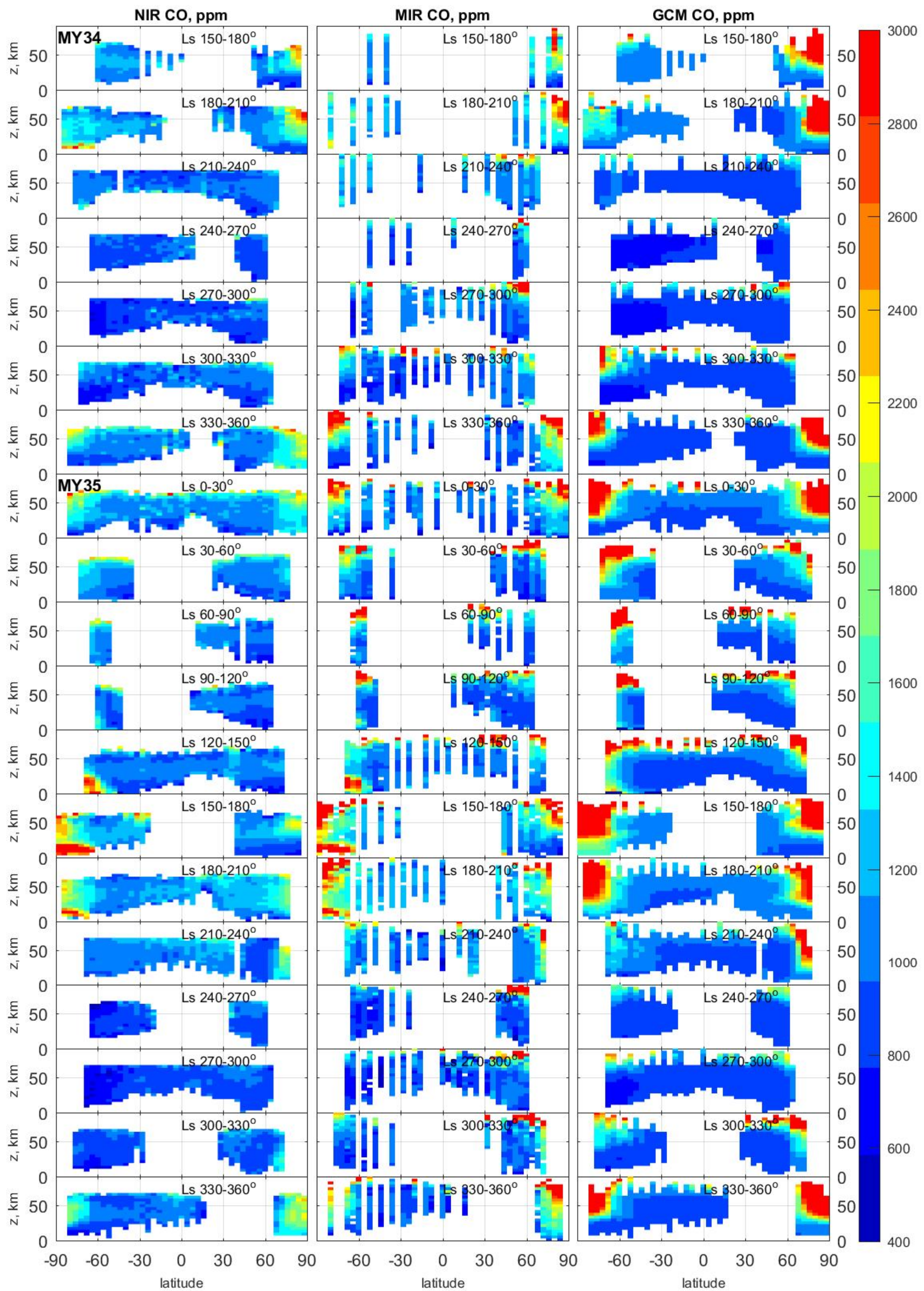


Figure 8.

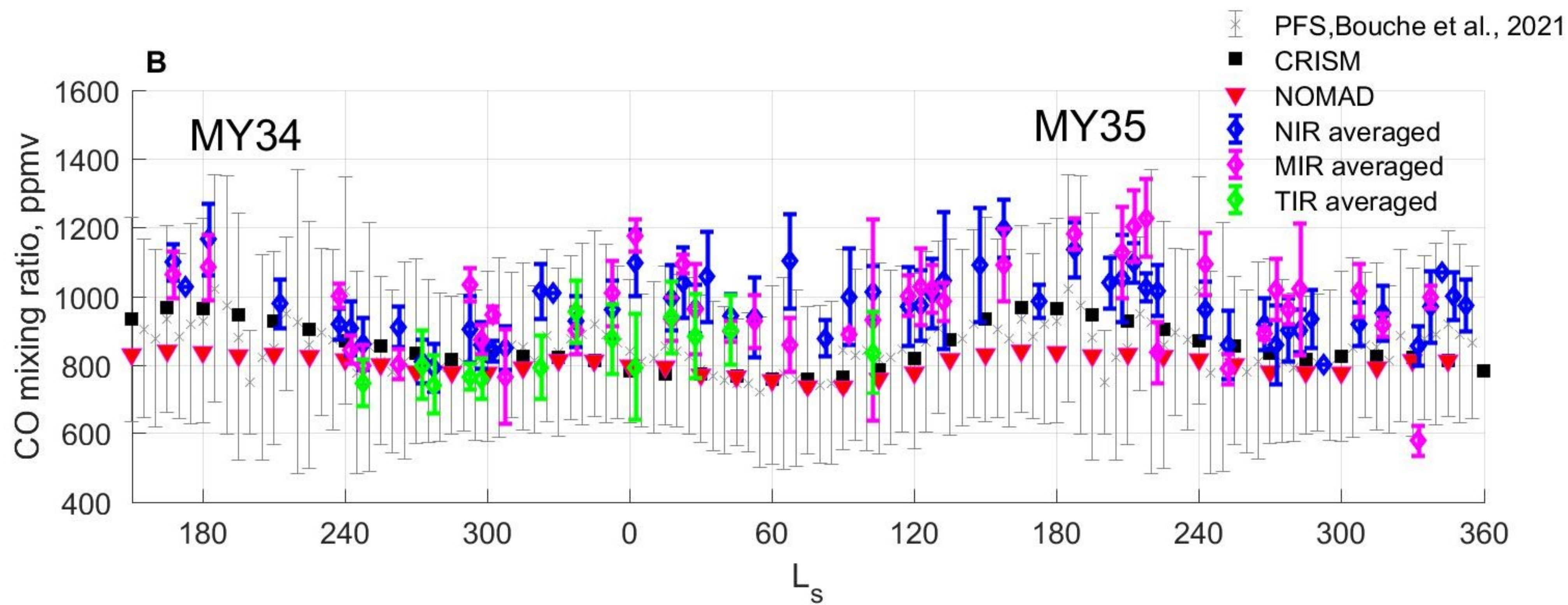
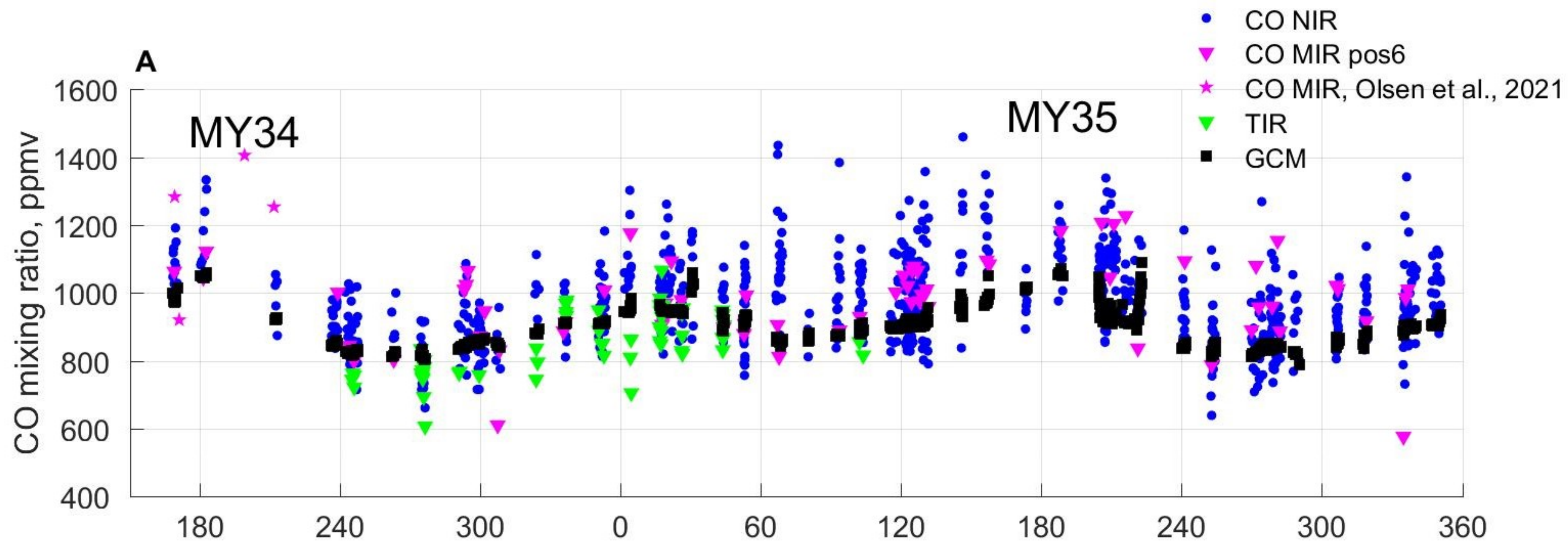


Figure 9.

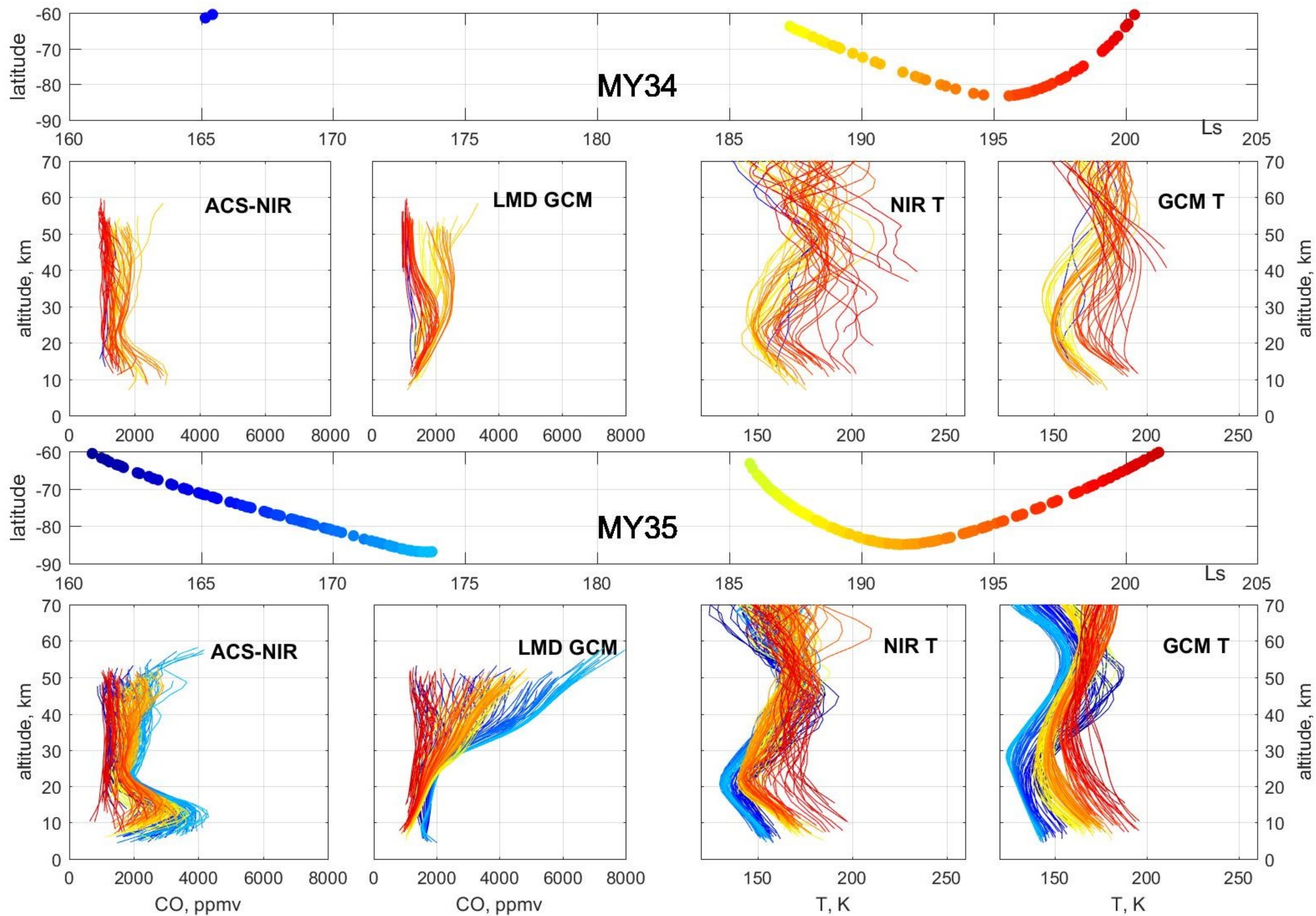


Figure 10.

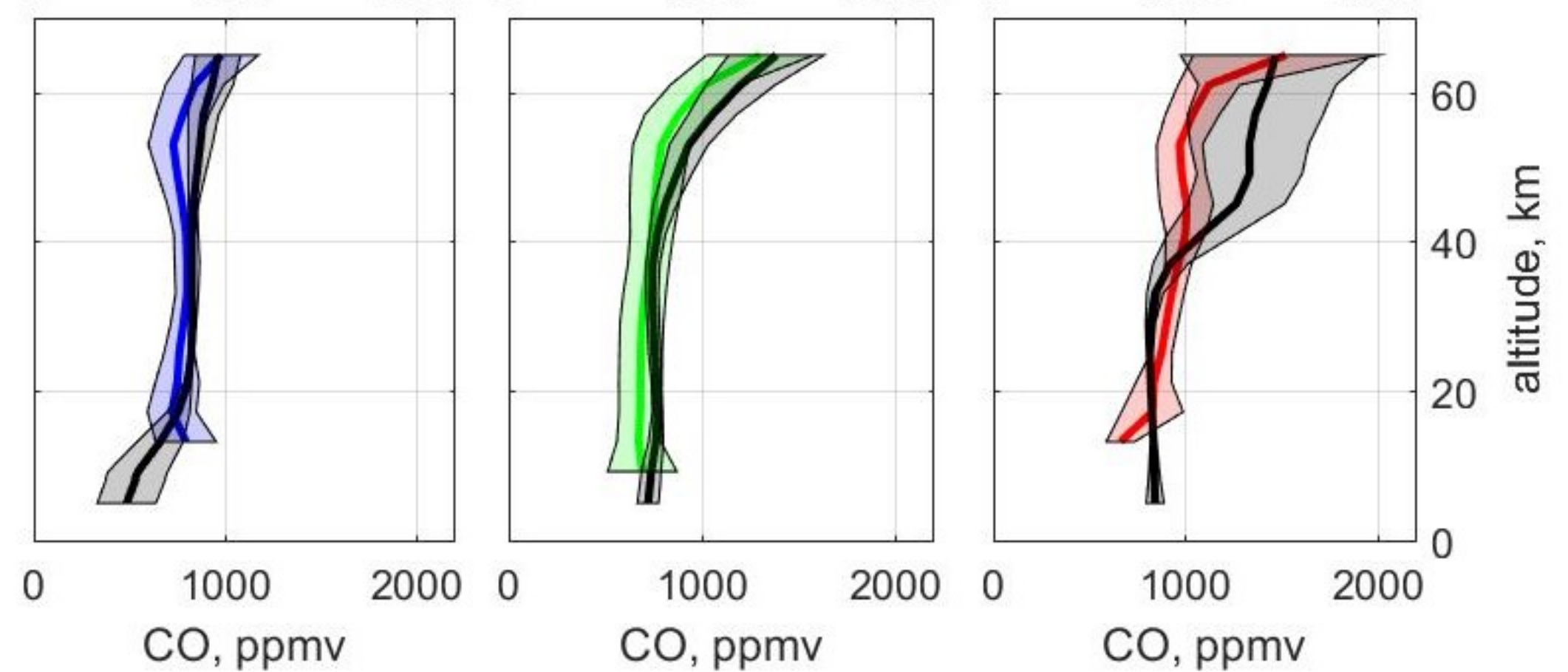
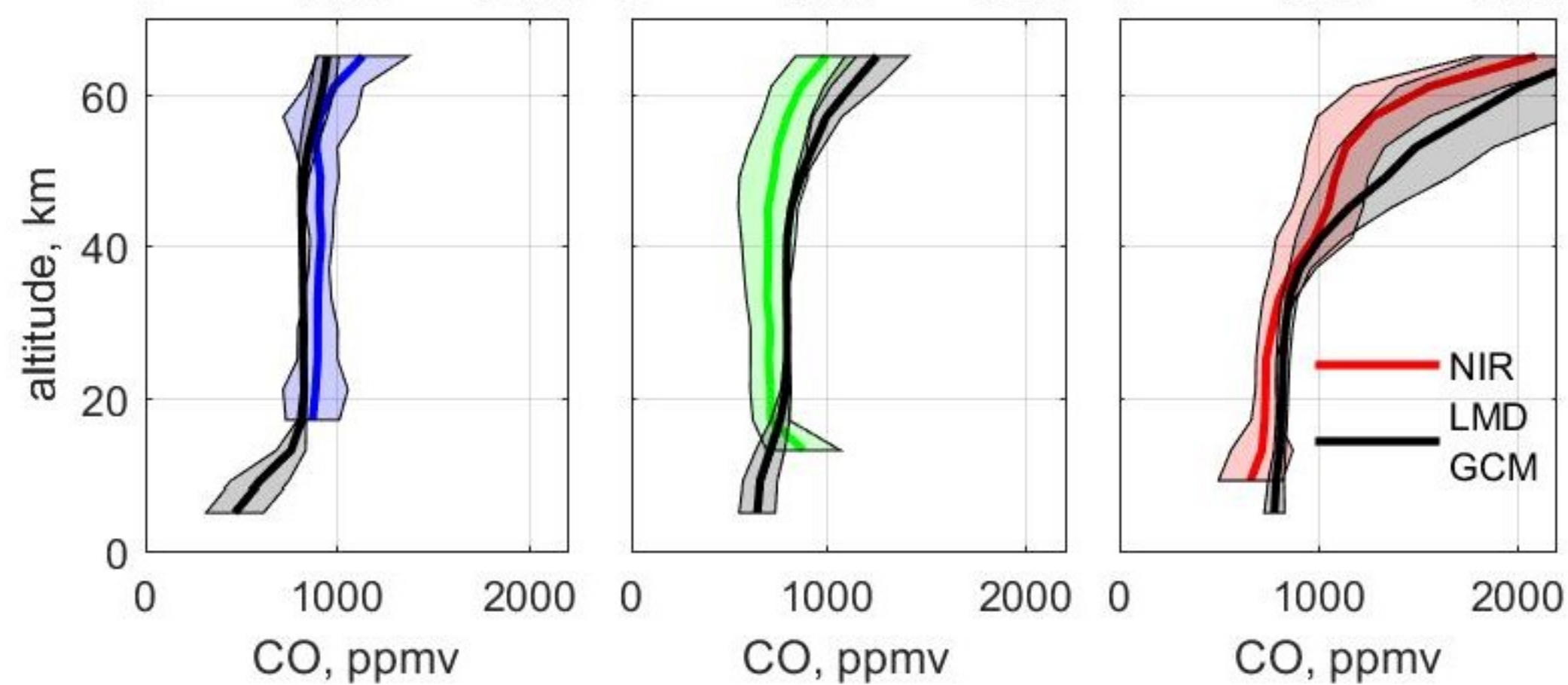
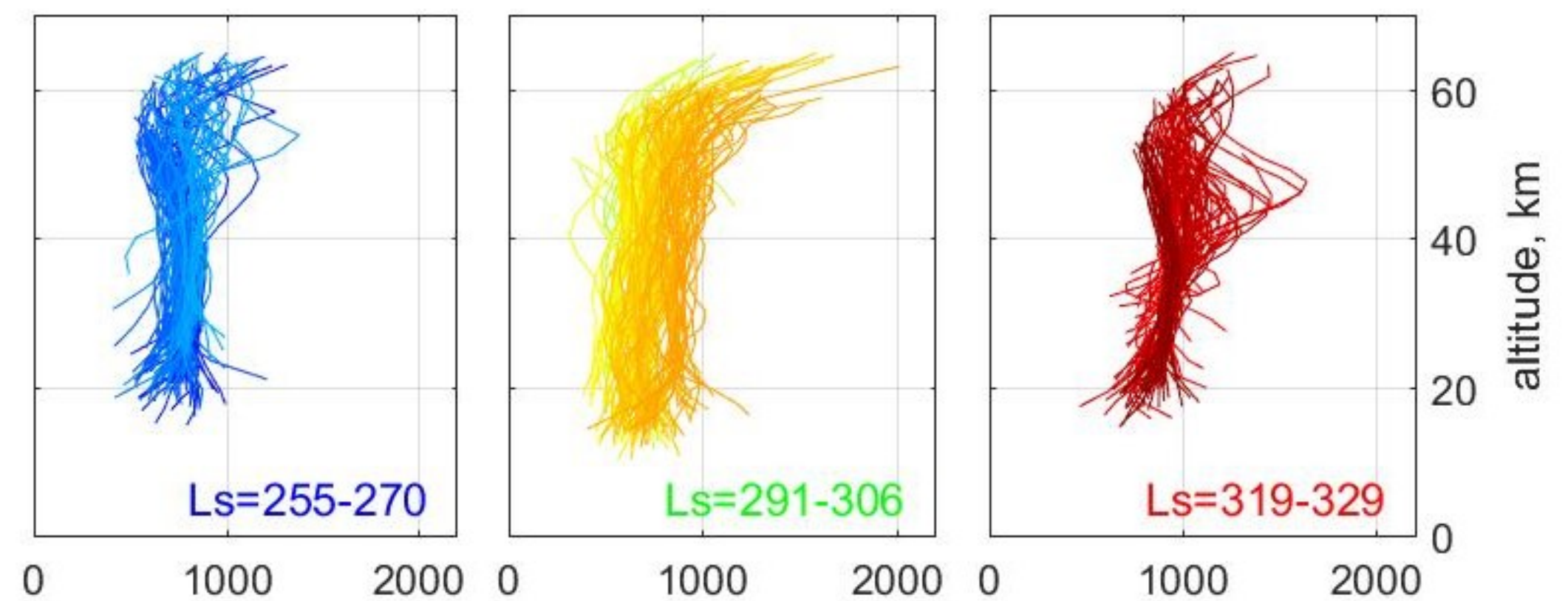
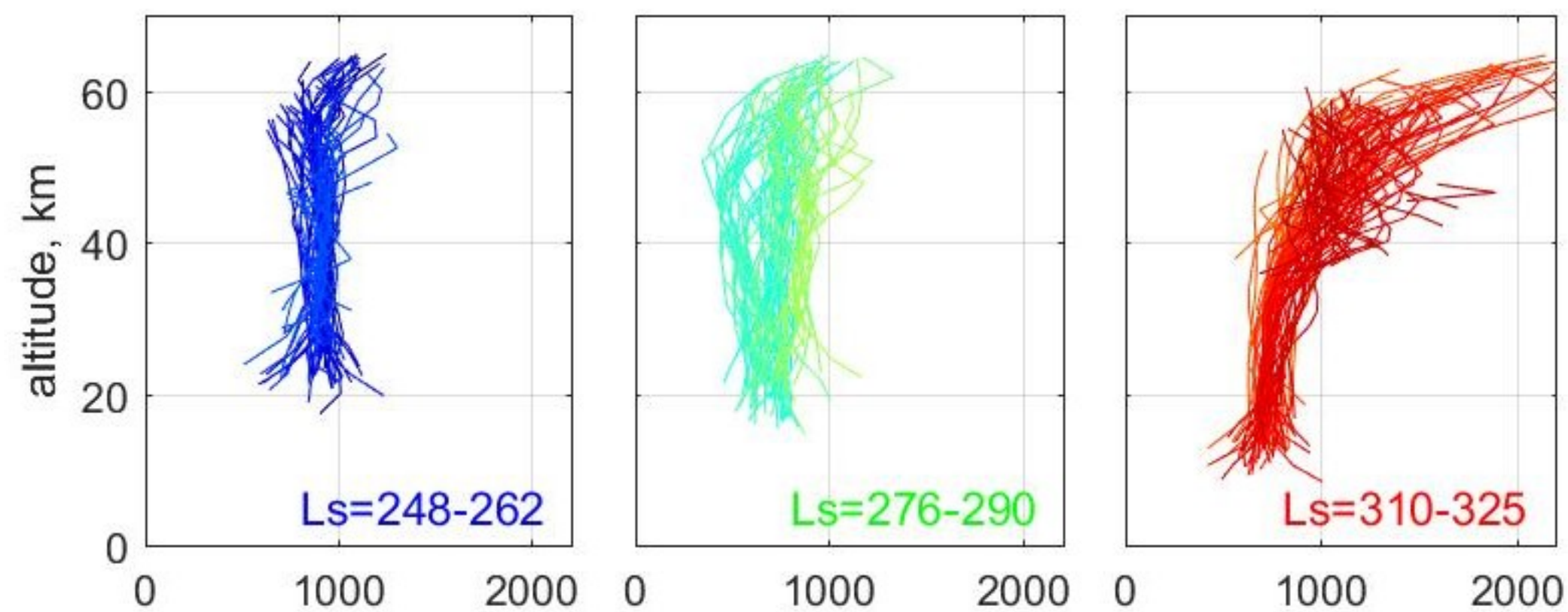
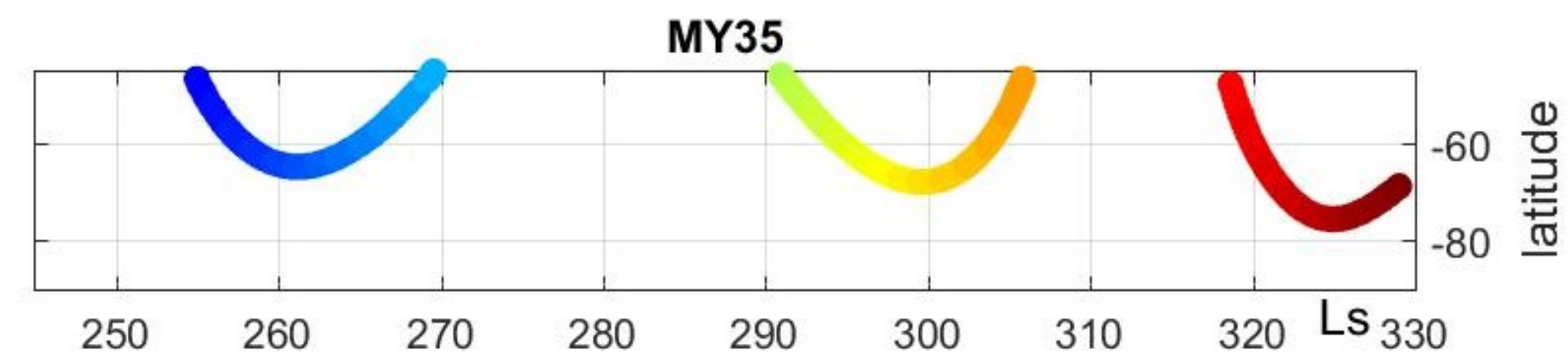
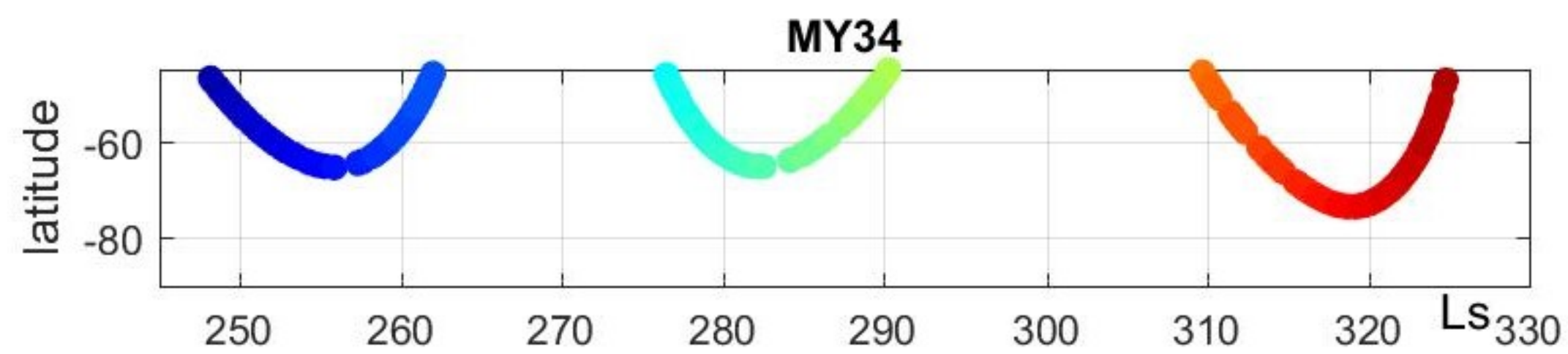


Figure 11.

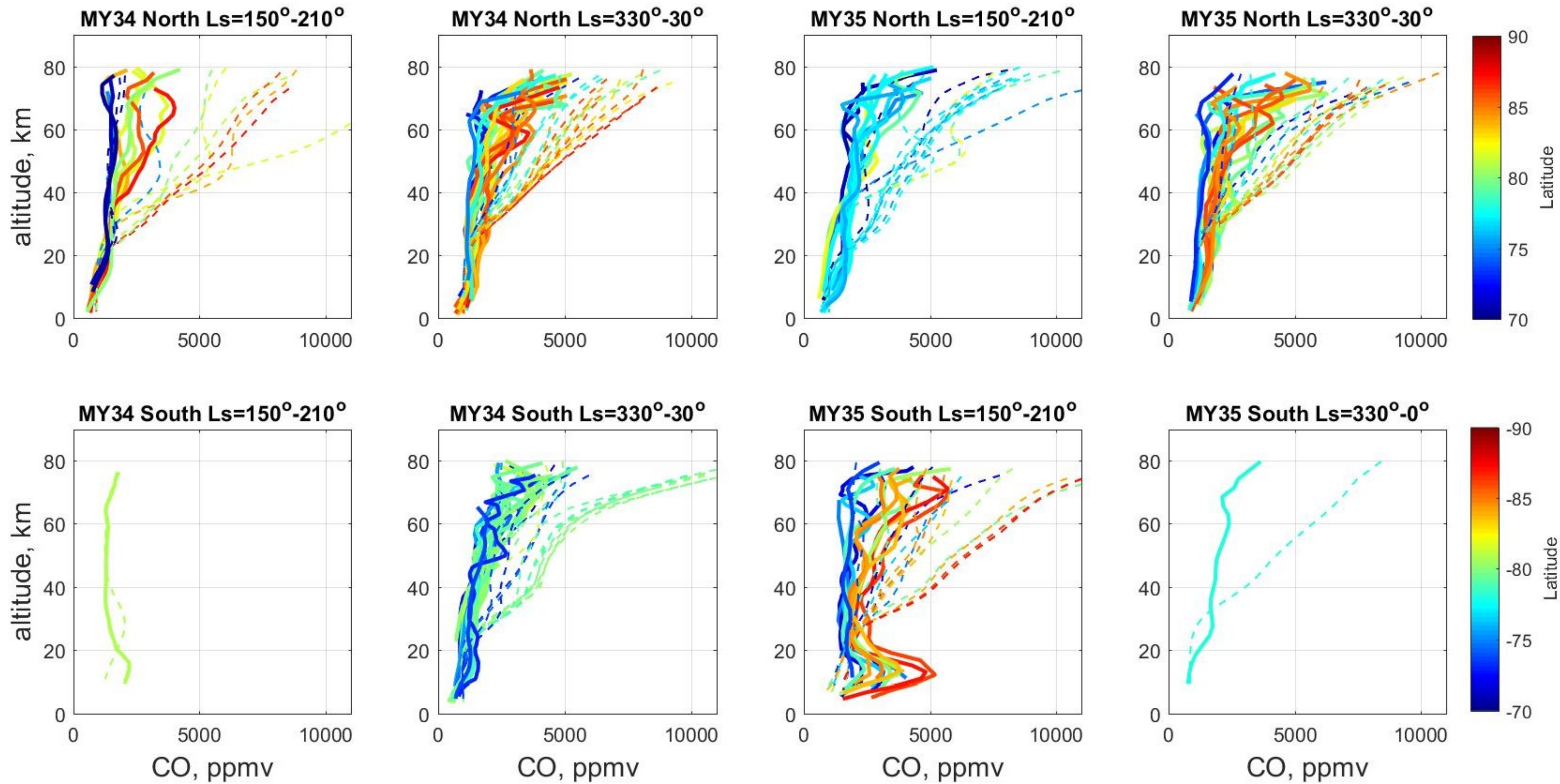


Figure 12.

