Exploiting night-time averaged spectra from PFS/MEX shortwave channel. Part 2: near-surface CO retrievals

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Abstract

Because of its important role in the Martian carbon cycle, carbon monoxide (CO) has been the subject of many measurements from ground and from space. Daytime measurements have been mostly exploited to measure the CO abundance because of their good signal-to-noise ratio, but night-time observations have not been documented yet. We demonstrate here the possibility of using PFS (Planetary Fourier Spectrometer) night-time observations to measure the Martian CO abundance. More particularly in this paper, we show that measuring CO during night with PFS can be achieved by averaging a large number of spectra to reach sufficient signal-to-noise. Furthermore, we show that the number of averaged spectra is not the only driving parameter for the detection of CO. High surface temperatures and high thermal contrast (negative in our case) are the two other conditions required for the measurement of the night-time CO abundance. Because of this, the retrievals are especially successful in the Southern Hemisphere during spring and sum-

mer when and where these two conditions are met. For night-time spectra with a positive detection, CO vertical profiles are successfully retrieved using the Optimal Estimation method and are characterized in terms of vertical sensitivity. Successful retrievals imply the use of accurate temperature profiles, and in particular an appropriate representation of the thermal inversion in the lower atmosphere. The temperature was obtained by using the relevant information from the CO₂ ν_3 band (Bauduin et al., 2020, this issue). A complete error budget of the retrieved CO profiles is also performed and includes different sources of uncertainty. Although the retrieved profiles are not resolved vertically, we show in particular that night-time PFS observations carry information mostly on the CO abundance for the 0-10 km altitude region, and thus provide a stronger constraint on the near-surface CO abundance compared to daytime observations.

Keywords: Mars atmosphere, carbon monoxide, night measurements, retrieval, PFS

1. Introduction

Carbon monoxide (CO) plays a major role in the carbon cycle of the Mars' atmosphere, notably due to its involvement in the so-called stability problem of carbon dioxide (CO₂) (McElroy and Donahue, 1972; Parkinson and Hunten, 1972; Atreya and Gu, 1994, 1995). Since its first detection in 1969 (Kaplan et al., 1969), CO has been extensively measured from ground (e.g. Clancy et al., 1983; Krasnopolsky, 2003, 2015) and from different space-borne instruments, such as OMEGA (Observatoire pour la Minéralogie, l'Eau, les Glaces et l'Activité) (Encrenaz et al., 2006), CRISM (Compact Reconnaissance

10 sance Imaging Spectrometer for Mars) (e.g. Smith et al., 2018) and PFS
 11 (Planetary Fourier Spectrometer) (e.g. Billebaud et al., 2009; Sindoni et al.,
 12 2011; Bouche et al., 2019). These ground and space observations have re-
 13 vealed a more complete picture of the CO climatology. Indeed, CO is not uni-
 14 formly distributed in the Martian atmosphere, but shows strong latitudinal
 15 and seasonal gradients related to the condensation and sublimation of CO₂
 16 at the poles. Although these spatial and time variations are today relatively
 17 well measured and understood, the vertical dimension is still lacking in this
 18 climatology. Different global circulation models are able to simulate the ver-
 19 tical distribution of CO and its seasonal/spatial variations (e.g. Forget et al.,
 20 1999, 2008; Daerden et al., 2019; Holmes et al., 2019), but measurements
 21 of the CO vertical profiles from remote sounders remain rare. Retrievals of
 22 highly resolved CO profiles from solar occultation observations of NOMAD
 23 (Nadir and Occultation for Mars Discovery) (Vandaele et al., 2018) and ACS
 24 (Atmospheric Chemistry Suite) (Korablev et al., 2018) instruments on board
 25 the ExoMars Trace Gas Orbiter (TGO) have started (Erwin et al., 2019;
 26 Olsen et al., 2019) and should provide in a near-future a better understand-
 27 ing of the vertical variability of CO in Mars’ atmosphere. Nevertheless, solar
 28 occultation measurements are generally limited to the sounding of the atmo-
 29 sphere above 10-20 km and thus cannot provide CO measurements for the
 30 near-surface atmosphere. Recently, Bouche et al. (2019) have investigated
 31 the possibility of retrieving CO profiles from single PFS nadir spectra in the
 32 1-0 band. They have shown that, even if the retrieved profiles are not ver-
 33 tically resolved, PFS spectra are mostly sensitive to CO in the near-surface
 34 atmosphere and, more especially, carry information on the CO abundance

below 20 km. These PFS nadir observations could therefore complement NOMAD and ACS solar occultation measurements.

The work of Bouche et al. (2019) is limited to daytime PFS observations. Night-time PFS measurements have indeed been discarded from the analysis due to the strong selection criteria applied to keep only single spectra with sufficient signal-to-noise ratio (SNR). Night-time PFS spectra might however be very useful for the sounding of CO in the near-surface atmosphere. They are generally associated to large thermal inversions that have been shown to increase the sensitivity of thermal infrared (TIR) measurements to the very near-surface atmosphere but also to slightly increase the vertical resolution of such measurements (Bauduin et al., 2017). A good knowledge of these thermal inversions is therefore essential for the CO retrieval. We have demonstrated in a companion paper (Bauduin et al., 2020) that this could be achieved by retrieving the atmospheric temperature profile using the information from the ν_3 CO₂ band (centered at 2349 cm⁻¹). The present paper has two main objectives. It aims first at demonstrating that the CO abundance can be retrieved from night-time nadir PFS spectra and at specifying under which conditions (specific regions and/or periods) this is best done. The second objective of this work is to investigate the potential gain in vertical sensitivity/resolution these observations can provide compared to daytime observations. For this, the paper has been divided in two main sections. Section 2 is related to the detection of CO and describes the spectral averages that have been necessary to reach sufficient SNR to detect CO. A detection method is also presented. Section 3 focuses on the retrieval of CO profiles from the averaged spectra. It presents the retrieval method and

60 settings, the characterization of the retrieved CO profiles in terms of error,
61 vertical sensitivity and information content, and the CO distribution built
62 from successful night-time retrievals. Section 4 gives the conclusions.

63 **2. Detection of CO from night-time observations**

64 As already mentioned, the challenge of measuring CO using night-time
65 PFS observations is the very low SNR of this kind of observations in the 5
66 micron region. However, by making proper averages, it is possible to increase
67 the SNR such that the detection of CO lines becomes possible (see also
68 Giuranna et al., 2019a, Figure S4 of the supplementary material). In this
69 section, we describe how the averages are performed and present the method
70 set up to assess if CO is detected in the averaged spectra.

71 *2.1. Observations and averages*

72 The observations used in this work are those recorded by the PFS instru-
73 ment (for details about the instrument and the calibration, see (Formisano
74 et al., 2005; Giuranna et al., 2005a,b)). Since the beginning of its science
75 operation, PFS has collected a large ensemble of Martian nadir spectra, with
76 a dataset covering today 8 full Martian Years (27 MY to 34 MY) and in-
77 cluding about four million of observations in total (including deep space and
78 limb measurements). Among these, we have considered only nadir night-
79 time observations, which have been defined as the observations for which
80 the solar incidence angle is higher than 90° . As the analysis focuses on the
81 fundamental 1-0 band of CO (centered at 2143 cm^{-1}), only spectra recorded
82 by the shortwave channel (SWC) of PFS have been considered. To simplify
83 the radiative transfer analysis by neglecting the scattering by dust and ice

84 particles, PFS observations associated with dust and ice optical depths larger
85 than 0.2 have been discarded. These 2 parameters were provided as part of
86 the PFS level 2 (L2) by the National Institute of Astrophysics INAF-IAPS
87 (Wolkenberg et al., 2018; Giuranna et al., 2019b). In total, 161295 night-time
88 spectra have been selected.

89 Using this set of night-time observations, averaged spectra have been
90 built. The difficulty has been to keep the averaged spectra the most repre-
91 sentative of the Martian atmosphere at the place and time they are associated
92 with, while averaging a sufficiently large number of observations to reach a
93 high enough SNR. Since CO exhibits stronger seasonal and latitudinal varia-
94 tions than longitudinal, we have limited the averages to small solar longitude
95 (L_s) periods and to small latitudinal ranges, but we have put no criterion on
96 the longitude. Specifically, the averages have therefore been made for boxes
97 of 10° of L_s and 5° of latitude over the 8 MYs, leading to 1163^1 , averaged
98 spectra in total.

99 Figure 1.a shows three examples of averaged spectra for which CO is
100 detected. The three have low noise and clearly shows the entire band of
101 CO in emission. CO₂ Q-branches are also well identified in emission. The
102 fact that emission features dominate is explained by the presence of thermal
103 inversions, as shown in Figure 1.b. These thermal inversions seem to occur
104 for most of the PFS night-time observations.

¹Some boxes do not include averaged spectra, due to some of the selection criteria.

Figure 1: (a) Examples of night-time averaged PFS spectra. For display purposes, spectra have been offset vertically as indicated. The L_s /latitude boxes to which they correspond are indicated. Some CO₂ lines have been identified, along with the P- and R-branches of the CO band. (b) Temperature profiles associated with the three spectra of panel (a). They have been retrieved following the method described in the companion paper (Bauduin et al., 2020).

2.2. Detection method

CO is unfortunately not detected for the 1163 averaged spectra as favourably as in the three example spectra of Figure 1. Its detection is notably impacted by different parameters, such as the surface temperature, which directly affects the SNR, and the thermal contrast, which determines the sensitivity to the near-surface atmosphere in the TIR. To evaluate the observation frequency of CO in the ensemble of averaged spectra and its dependence on these two parameters, a detection method has been set up. It is based on the estimation of CO lines intensities using brightness temperature differences (BTD). These differences are calculated between the BT of one target chan-

115 nel, measured at the center of a CO line, and the averaged BT calculated for
 116 two base channels, representing the baseline of the spectrum and located on
 117 each side of the CO line. BTD have been calculated for 21 lines indicated
 118 in Figure 2 (the lines are identified with green dots and the base channels
 119 correspond to red dots). The BTDs are positive for the lines in emission and
 120 negative for lines in absorption.

Figure 2: Spectral channels chosen to calculate the BTD for each CO lines. The channels corresponding to the CO lines (target) are shown in green, and the channels used for the baseline (base) are shown in red. Note that the displayed spectrum is the one corresponding to the box $270^\circ - 280^\circ / 40^\circ - 45^\circ \text{S}$ (L_s /latitude).

121 The averaged spectra have been considered as too noisy (and rejected
 122 from all further analyses) when the associated BTD was lower than the Noise
 123 Equivalent Temperature Difference (NEDT), calculated as an average over

124 the 2000-2200 cm^{-1} range from the following equation:

$$NEDT_{\tilde{\nu}} = \frac{1}{L_{\tilde{\nu}}} \left(\frac{k_B T^2}{hc\tilde{\nu}} \right) \left(1 - \exp \left(\frac{-hc\tilde{\nu}}{k_B T} \right) \right) NEDL_{\tilde{\nu}}, \quad (1)$$

125 where $NEDL_{\tilde{\nu}}$ is the radiance noise of the averaged spectrum², $L_{\tilde{\nu}}$ is the
 126 radiance at wavenumber $\tilde{\nu}$, h is the Planck constant, c is the light speed, k_B
 127 is the Boltzmann constant, and T is the reference temperature, which has
 128 been set to 280 K. An additional criterion has been necessary to remove some
 129 spectra with noise spikes: spectra with a mean BTD lower than the BTD
 130 standard deviation are rejected.

131 Figure 3 illustrates the detection method applied on the 1163 averaged
 132 spectra. Figure 3.a shows the averaged NEDT (over 2000-2200 cm^{-1}) as
 133 a function of thermal contrast³ and surface temperature while Figure 3.b
 134 shows the mean BTD (for the 21 CO lines) as a function of these same two
 135 parameters. The filled circles correspond to spectra with detectable CO and
 136 the empty circles correspond to those with no CO detection. From Figure 3,
 137 the impact of thermal contrast and surface temperature on the detection of
 138 CO is clear. Most detections are, interestingly, associated with large negative
 139 thermal contrasts (lower than -20 K), which increase the sensitivity of PFS to
 140 near-surface CO, and with large surface temperatures (above 180 K), which
 141 increase the SNR. Most of the non detections correspond on the contrary to
 142 observations with low thermal contrast (between 0 and -10 K) and low surface
 143 temperatures (around 145 K). For some cases, despite the large thermal

²To calculate the noise for the averaged spectrum, we have divided the radiance noise of single PFS spectra by the square root of the number of spectra averaged.

³The thermal contrast is calculated as the difference between the surface temperature and the mean air temperature of the 0-2 km layer.

Figure 3: (a) Average NEDT over the $2000\text{--}2200\text{ cm}^{-1}$ range (K, color scale) and (b) mean BTD calculated for the 21 CO lines (K, color scale) as a function of thermal contrast and surface temperature. The filled circles correspond to spectra with detectable CO and the empty circles to those with a too large noise. The surface temperature associated to each average spectrum has been calculated as the average of the surface temperatures corresponding to single spectra and provided as part of the L2 by INAF-IAPS. The thermal contrast has been calculated as the difference between the surface temperature and the temperature of the air for the 0-2 km column.

contrast and surface temperature, the mean BTD is significantly lower than
the NEDT leading to a non CO detection. In total, 449 spectra over the 1163
present CO lines emerging from the noise and these have been selected for
the following analysis. Interestingly, these selected observations are mainly
located in the Southern Hemisphere during spring/summer ($180^\circ < L_s <$
 360°), where the favourable conditions for the CO detection (high negative
thermal contrast and high surface temperature) are found. This will be
discussed in the next section.

152 3. Retrieval of CO from night-time observations

153 In this section, the CO abundance is retrieved from the 449 averaged spec-
154 tra selected in the previous section. More particularly, vertical CO profiles
155 are adjusted. Although the vertical resolution of PFS is limited, this allows to
156 take into account the vertical sensitivity of the instrument in the estimation
157 of the CO column density. More importantly, thanks to the presence of large
158 thermal inversions, the vertical resolution of the CO profiles could be higher
159 than for daytime observations (Bouche et al., 2019), as it has been shown
160 for Earth nadir TIR spectra (Bauduin et al., 2017). This is discussed in sec-
161 tion 3.2 after the description of the retrieval method applied (section 3.1).
162 The distribution of CO abundance is then presented (subsection 3.3).

163 3.1. Retrieval methodology

164 3.1.1. Background

165 The retrieval method used to retrieve CO vertical profiles is the Optimal
166 Estimation (OE) (Rodgers, 2000). The idea of the method is to find the
167 CO profile that is the most consistent with both the averaged night-time
168 spectrum and the prior knowledge of the Martian atmosphere. This is done
169 by minimizing a cost function J , written as:

$$J = (\mathbf{y} - \mathbf{F}(\mathbf{x}, \mathbf{b}))^T \mathbf{S}_\epsilon^{-1} (\mathbf{y} - \mathbf{F}(\mathbf{x}, \mathbf{b})) + (\mathbf{x} - \mathbf{x}_a)^T \mathbf{S}_a^{-1} (\mathbf{x} - \mathbf{x}_a), \quad (2)$$

170 where $\mathbf{x}$ is the atmospheric state vector, including the parameters to retrieve
171 (e.g. CO profile, see subsection 3.1.2), $\mathbf{b}$ includes the parameters affecting the
172 top of the atmosphere (hereafter TOA) radiance but that are kept fixed, $\mathbf{y}$
173 is the vector of measured radiances, $\mathbf{x}_a$ is the *a priori* state vector, including

174 the knowledge of the retrieved parameters before the measurement is made,
 175 $\mathbf{S}_{\mathbf{a}}$ is the associated *a priori* covariance matrix, $\mathbf{S}_{\epsilon}$ is the measurement error
 176 covariance matrix associated to $\mathbf{y}$ and $\mathbf{F}$ is the radiative transfer forward
 177 model. The minimization of J is performed by a Gauss-Newton iterative
 178 scheme, with a new state vector evaluated at each iteration i as follow:

$$\mathbf{x}_{i+1} = \mathbf{x}_{\mathbf{a}} + (\mathbf{K}_i^T \mathbf{S}_{\epsilon}^{-1} \mathbf{K}_i + \mathbf{S}_{\mathbf{a}}^{-1})^{-1} \mathbf{K}_i^T \mathbf{S}_{\epsilon}^{-1} [\mathbf{y} - \mathbf{F}(\mathbf{x}_i, \mathbf{b}) + \mathbf{K}_i(\mathbf{x}_i - \mathbf{x}_{\mathbf{a}})], \quad (3)$$

179 with $\mathbf{K}$ the Jacobians matrix, whose rows are the derivatives of the TOA
 180 radiance with respect to the retrieved parameters. For more details about
 181 the method, see (Rodgers, 2000).

182 The OE offers an adequate framework to characterize the retrieved profiles
 183 in terms of errors and vertical sensitivity. In this work, the retrieval errors are
 184 estimated from a total error covariance matrix ($\mathbf{S}_{\text{tot}}$), whose the square root
 185 of the diagonal elements provide the total retrieval error for each retrieved
 186 layer. This matrix is calculated as follows (Rodgers, 2000)

$$\mathbf{S}_{\text{tot}} = \mathbf{S}_{\mathbf{s}} + \mathbf{S}_{\mathbf{m}} + \mathbf{S}_{\mathbf{f}}, \quad (4)$$

187 where $\mathbf{S}_{\mathbf{s}}$ is the covariance matrix of the smoothing error, which character-
 188 izes the uncertainty on the CO profile due to the limited vertical resolution
 189 of nadir PFS observations, $\mathbf{S}_{\mathbf{m}}$ is the covariance matrix of the measurement
 190 error, which characterizes the uncertainty on the CO profile due to the mea-
 191 surement noise, and $\mathbf{S}_{\mathbf{f}}$ is the covariance matrix of the fitted parameters
 192 error, taking into account the interferences between the retrieved CO profile
 193 and the rest of the state vector (here, surface temperature and CO₂ profile,
 194 see section 3.1.2). These three matrices are calculated using the following

195 equations (Rodgers, 2000):

$$\mathbf{S}_s = (\mathbf{A} - \mathbf{I}) \mathbf{S}_a (\mathbf{A} - \mathbf{I})^T \quad (5)$$

$$\mathbf{S}_m = \mathbf{G} \mathbf{S}_\epsilon \mathbf{G}^T \quad (6)$$

$$\mathbf{S}_f = \mathbf{A}_{ir} \mathbf{S}_{ra} \mathbf{A}_{ir}^T, \quad (7)$$

196 where $\mathbf{A}$ is the averaging kernels matrix (see below), $\mathbf{I}$ is the identity matrix,
 197 $\mathbf{G}$ is the gain matrix, whose rows are the derivatives of the retrieved state with
 198 respect to the measurement, $\mathbf{A}_{ir}$ are the non-diagonal blocs of the averaging
 199 kernels matrix, $\mathbf{i}$ is the part of the state vector under consideration (i.e.
 200 the CO profile), $\mathbf{r}$ represents the rest of the state vector (i.e. the surface
 201 temperature and the CO₂ profile), and $\mathbf{S}_{ra}$ is the *a priori* covariance matrix
 202 of $\mathbf{r}$. The vertical sensitivity and resolution of the retrieval is described by
 203 the averaging kernels (AVK) functions. They are defined as the rows of the
 204 AVK matrix ($\mathbf{A}$), calculated as

$$\mathbf{A} = (\mathbf{K}^T \mathbf{S}_\epsilon^{-1} \mathbf{K} + \mathbf{S}_a^{-1})^{-1} \mathbf{K}^T \mathbf{S}_\epsilon^{-1} \mathbf{K}. \quad (8)$$

205 One AVK function is associated to CO in each retrieved layer. It peaks at
 206 the altitude from where most of the information about CO in that layer is
 207 coming. The AVKs thus provide an estimation of the altitude of maximum
 208 sensitivity to CO. Their width gives also an estimation of the retrieval vertical
 209 resolution. The trace of the matrix $\mathbf{A}$ corresponds to the degrees of freedom
 210 for signal (DOFS) of the retrieval.

211 The AVK matrix allows expressing the retrieved state ($\hat{\mathbf{x}}$) as a function
 212 of the true ($\mathbf{x}$) and *a priori* states (if we ignore the term involving random
 213 errors) (Rodgers, 2000):

$$\hat{\mathbf{x}} = \mathbf{A} \mathbf{x} + (\mathbf{I} - \mathbf{A}) \mathbf{x}_a, \quad (9)$$

214 with $\mathbf{I}$ the identity matrix. The retrieved state at any vertical location is
 215 thus a combination between: 1) an average of the true profile weighted by
 216 the corresponding AVK function and 2) an average of the *a priori* profile
 217 weighted by $(\mathbf{I}-\mathbf{A})$. To visualize more easily the impact of the *a priori* profile
 218 on the retrieved state, it is useful to perform an eigenvector decomposition
 219 of the matrix $\mathbf{A}$ (Rodgers, 2000). If the matrix $\mathbf{R}$ is the matrix of the right
 220 eigenvectors of $\mathbf{A}$, we can write

$$\mathbf{A}\mathbf{R} = \mathbf{R}\mathbf{\Lambda}, \quad (10)$$

221 where $\mathbf{\Lambda}$ is a diagonal matrix, whose diagonal elements correspond to the
 222 eigenvalues of $\mathbf{A}$. If we multiply each side of equation 9 by $\mathbf{R}^{-1}$ and using
 223 equation 10, we obtain

$$\mathbf{R}^{-1}\hat{\mathbf{x}} = \mathbf{R}^{-1}\mathbf{A}\mathbf{x} + \mathbf{R}^{-1}(\mathbf{I} - \mathbf{A})\mathbf{x}_a \quad (11)$$

$$= \mathbf{\Lambda}\mathbf{R}^{-1}\mathbf{x} + (\mathbf{I} - \mathbf{\Lambda})\mathbf{R}^{-1}\mathbf{x}_a. \quad (12)$$

224 In the eigenvectors of $\mathbf{A}$ basis, the vector $\mathbf{x}$ becomes $\mathbf{R}^{-1}\mathbf{x} = \mathbf{z}$ and thus,
 225 we have $\mathbf{x} = \mathbf{R}\mathbf{z}$. The vector $\mathbf{x}$ (and so the retrieved, the true and the *a*
 226 *priori* states) is therefore represented as a linear combination of the right
 227 eigenvectors of $\mathbf{A}$, and $\mathbf{z}$ is the coefficients vector. Using these relations,
 228 equation 11 becomes

$$\hat{\mathbf{z}} = \mathbf{\Lambda}\mathbf{z} + (\mathbf{I} - \mathbf{\Lambda})\mathbf{z}_a. \quad (13)$$

229 Because $\mathbf{\Lambda}$ is diagonal, the above equation can be separated into scalar com-
 230 ponents such that

$$\hat{z}_i = \lambda_i z_i + (1 - \lambda_i) z_{ai}. \quad (14)$$

231 Considering that all profiles are decomposed into a combination of patterns
 232 (eigenvectors of $\mathbf{A}$), the retrieved coefficients $\hat{z}_i$ are a weighted average of

233 the true z_i and *a priori* z_{ai} coefficients. The patterns associated to $\lambda_i \sim 1$
 234 are dominated by the true state, and thus well measured, while those asso-
 235 ciated to $\lambda_i \sim 0$ are determined by the *a priori* state. Note that we have
 236 not considered the measurement error in the above mathematical develop-
 237 ment. It is important to keep in mind that in reality a third term including
 238 the random uncertainties of the measurement also impacts the value of the
 239 retrieved coefficients $\hat{z}_i$. Finally, it can be shown that the DOFS is the sum
 240 of the eigenvalues of $\mathbf{A}$. It corresponds therefore to the number of compo-
 241 nents/patterns in the retrieved profile that are measured independently of
 242 the *a priori*.

243 3.1.2. Retrieval settings and input to the radiative transfer

244 The analysis has been divided in two steps: 1) the retrieval of the tem-
 245 perature profile, and 2) the retrieval of the CO profile. The retrieval of the
 246 temperature profile has been performed on the ν_3 band of CO_2 centered at
 247 2349 cm^{-1} for each averaged night-time spectrum. The details about this re-
 248 trieval, the choice of its settings and the associated error characterization are
 249 fully described in the companion paper (Bauduin et al., 2020) and are not re-
 250 minded hereafter. The retrieved temperature profiles are then used as input
 251 for the CO retrieval. Before going further, it has to be pointed out that the
 252 temperature profile retrieval was successful for 350 out of the 449 averaged
 253 spectra (for 99 observations, the temperature retrievals have diverged, i.e. a
 254 temperature profile could not be retrieved for these observations). Only for
 255 these 350 spectra a CO profile retrieval was performed. More particularly,
 256 the CO retrieval has been performed on the $2040\text{-}2230 \text{ cm}^{-1}$ spectral range.
 257 The CO *a priori* profile and covariance matrix are those used by Bouche

et al. (2019). As a reminder, they have been built from a large ensemble of
 profiles modelled by the MCD (version 5.3) for one Martian year, using the
 climatology solar average scenario. In total, 451584 modelled CO profiles
 have been used. The resulting $\mathbf{x}_a$ is shown in Figure 4.b along with the *a*
priori uncertainties (error bars), which correspond to the square root of the
 diagonal elements of the $\mathbf{S}_a$. The atmosphere used extends from 0 to 60 km.
 The temperature profile considered is the one retrieved in the first step. CO
 is retrieved for 2 km thick layers from 0 to 24 km (Bouche et al., 2019). Above
 that altitude and up to 60 km, the CO profile is kept fixed to the *a priori*. As
 shown in Figure 1.a, CO₂ presents strong Q branches in the spectral range
 considered, some of which interfere with CO lines. This gas has thus been
 retrieved simultaneously with CO only to obtain the best fit for CO. As for
 the latter, CO₂ has been retrieved for 2 km thick layers from 0 to 24 km, and
 then fixed to the *a priori* profile above that level. This CO₂ *a priori* profile is
 the same as the one used to retrieve the temperature profile and corresponds
 to a global average calculated from CO₂ profiles modelled by the MCD (ver-
 sion 5.3), with same scenario as for CO. These modelled CO₂ profiles have
 also been used to build the associated *a priori* covariance matrix (Bouche
 et al., 2019). Note that the retrieval of CO₂ is strongly constrained (maximal
a priori variability of 2% at the surface) and leads to only small variations
 of surface pressure (5 times smaller than the uncertainty considered in our
 error calculations, see section 3.2.4). Note also that ozone (O₃) and water
 vapour (H₂O) have absorption bands in the spectral range considered. Al-
 though their contributions are negligible with no visible spectral features, we
 have fixed their abundance to climatological modelled profiles (built from the

283 MCD, version 5.3). In addition to CO and CO₂, the surface temperature has
 284 been retrieved to properly model the baseline of the spectrum. The *a priori*
 285 value was set to the averaged surface temperature calculated from the L2
 286 surface temperatures associated to each spectrum considered in the spectral
 287 average. The *a priori* variability has been set to 1%. For the emissivity, an
 288 average value for each average spectrum has been calculated by co-localizing
 289 single spectra with the emissivity at 2000 cm⁻¹ global maps provided by the
 290 Planetary Surface Portal (Poulet et al., 2018; Quantin-Nataf et al., 2018). It
 291 has been kept fixed during the retrieval. Finally, the spectroscopic param-
 292 eters have been taken from HITRAN 2016 (Gordon et al., 2017), including the
 293 pressure broadening by CO₂, and the instrumental line shape recommended
 294 by the PI team of PFS has been considered (Giuranna et al., 2019a).

295 All retrievals have been performed with the Atmosphit software (Barret
 296 et al., 2005; Coheur et al., 2005). This is a line-by-line radiative transfer
 297 code initially developed for Earth atmospheric spectra, but which has been
 298 successfully adapted to analyse PFS Martian observations (Bouche et al.,
 299 2019). Note that when performing the retrieval of an abundance profile,
 300 Atmosphit adjusts a vector of scaling factors, that applied on the *a priori*
 301 profile gives the retrieved profile. In the case of our retrieval, partial columns
 302 of CO are first calculated for the retrieved layers by applying the retrieved
 303 scaling factors onto the *a priori* partial columns. This is done for CO₂ as well.
 304 Then, to obtain the profiles expressed in volume mixing ratios (VMR), the
 305 air partial columns are calculated using the retrieved CO₂ partial columns,
 306 but also including the retrieved CO partial columns and the fixed H₂O and
 307 O₃ columns, determined from the chosen climatological profiles. Abundance

of nitrogen (N_2) and argon (Ar) are also taken into account, assuming mole fractions respectively of 0.027 and 0.016.

3.2. Characterization of retrieved CO profiles

In this section, we focus on the characterization of the retrieved CO profiles in terms of vertical sensitivity and errors. Among the 350 spectra analysed, only 170 have been successfully fitted (i.e. the root mean square of the residual is similar to the noise level considered). The characterization is discussed only for these successful retrievals and is especially illustrated with one chosen case, which corresponds to an average PFS spectrum located in the Southern Hemisphere (in the box 40°S - 35°S) during the summer ($280^\circ < L_s < 290^\circ$). It is shown in Figure 4.a along with the resulting fit. The root mean square (RMS) of the residual is $1.74 \times 10^{-10} \text{ W cm}^{-2} \text{ sr}^{-1} \text{ cm}$ and is of the same order as the NEDL estimated to be $1.87 \times 10^{-10} \text{ W cm}^{-2} \text{ sr}^{-1} \text{ cm}$. In the following subsections, we present the characterization in terms of the retrieved profile and its associated errors (section 3.2.1), of the vertical sensitivity (section 3.2.2) and of the *a priori* impact (section 3.2.3).

3.2.1. Retrieved profile and errors

The retrieved CO profile for the chosen example is shown in Figure 4.b with the corresponding errors for each retrieved layers. The largest VMR is retrieved for the first layer (0-2 km) and is 798 ppm. It then decreases to around 640 ppm at 5 km and stays almost constant up to the highest retrieved layer. The column-averaged VMR calculated for the range 0-24 km is equal to 668 ppm. Regarding the errors, the lowest error is found at the surface. It is around 10% and this corresponds to a large reduction of the

Figure 4: Characterization of the CO retrieved profiles for a representative retrieval. (a) Resulting fit of the spectrum. (b) Retrieved CO profile (black) and *a priori* profile (red) expressed in volume mixing ratios (ppm). (c) Averaging kernels functions for the CO retrieved profile expressed in VMR units. They provide an estimation of the altitude of maximum sensitivity to CO. The dots indicate to which layer the AVK functions correspond. (d) Jacobians ($\text{W cm}^{-2} \text{ sr}^{-1} \text{ cm per VMR}$, color scale) calculated for the retrieved layers. (e) Retrieved temperature profile (Bauduin et al., 2020) and surface temperature (black dot). (f-j) The first five eigenvectors of the AVK matrix and their associated eigenvalues (λ), allowing to evaluate the impact of the *a priori* profile on the retrieved CO profile (see text for details).

332 prior uncertainty of 82%. The retrieval error then increases to about 15% at
 333 3 km and stays constant up to 15 km. Above that layer, it increases again to

334 more than 20%. The error on the total column (0-24 km) is estimated to be
 335 around 14%. Note that this is not the total error. Additional contributions
 336 due to uncertainties on the temperature profile, surface pressure and ILS
 337 formulation will be estimated in section 3.2.4.

Figure 5: (a) Histogram of the number of retrievals as a function of the total retrieval error (%) on the 0-2 km CO column (blue bars) and on the 0-24 km CO column (orange bars). (b) Dependence of the retrieval error on the 0-2 km CO VMR (%) on the CO VMR itself and on the thermal contrast (K, color scale). (c) Same as (b) but for the 0-24 km VMR.

338 To have a better representation of the retrieval errors for all the analysed
 339 observations, Figure 5.a shows the distribution of the errors (for the 0-2 km
 340 and 0-24 km CO columns) among the retrieval set. For most of the average
 341 spectra analysed (73%), the retrieval uncertainty on the 0-2 km column falls
 342 in the range 6-12%. For around 12% of data, the retrieval error on the first
 343 layer is smaller than 6% and for 15%, it is larger than 12%. The retrieval error

344 on the total column is very small, between 6 and 15%⁴, for the large majority
 345 of the analysed set (93%). The cause of the variations of the retrieval error
 346 can be investigated with the help of Figures 5.b and 5.c. In the case of the
 347 retrieval error on the 0-2 km column (Figure 5.b), the variations are driven
 348 mainly by one parameter: the thermal contrast. Indeed, for a constant 0-2
 349 km column, the retrieval error increases with a decrease (in absolute value)
 350 of the thermal contrast. The latter indeed drives the sensitivity of PFS to
 351 the near-surface CO: the larger (in absolute value) the thermal contrast,
 352 the higher the sensitivity to the 0-2 km layer, and therefore, the lower the
 353 associated retrieval error. This dependency of the errors on the thermal
 354 contrast is however much lower for the total (0-24 km) column retrieval error
 355 (Figure 5.c). Note finally that the thermal contrast values sampled in this
 356 work are highly negative. Even the lowest values (in absolute) are sufficiently
 357 high to allow PFS to have some sensitivity to the near-surface CO. This
 358 is shown particularly by the low errors on the surface layer even for the
 359 smallest thermal contrasts (absolute value) (see also (Bauduin et al., 2017)).
 360 This could explain why the impact of thermal contrast on the total column
 361 retrieval error is so small.

362 3.2.2. *Vertical sensitivity*

363 For the chosen case, the fact that the lowest error is found in the first
 364 layer is explained by the presence of a thermal inversion and by the associated
 365 high negative thermal contrast (-30 K, see Figure 4.e), which enhances PFS

⁴Note again that this is not the total error. Additional contributions will be estimated
 in section 3.2.4.

366 sensitivity towards the CO close to surface. The maximum sensitivity to CO
 367 is indeed found at the surface as shown by the averaging kernels functions
 368 (Figure 4.c) and by the Jacobians (Figure 4.d). The AVK functions for all
 369 retrieved layers peak at the surface. This shows that CO retrieved in each
 370 layer, even the highest ones, is mostly inferred from the CO abundance of the
 371 first layer. Note especially the very high value of 0.93 at the surface for the
 372 AVK of the first layer, showing a very high sensitivity of the retrieved CO
 373 VMR in the 0-2 km layer to the true CO VMR of that same layer. Above
 374 the 4-6 km layer, all the AVK functions are very close to 0 (except for the 0-2
 375 km AVK, see below). This means that the retrieved CO VMR above 6 km is
 376 mainly driven by the CO abundance of the layers close to the surface. This
 377 is also observed with the Jacobians, that show maximal values close to the
 378 surface and very low sensitivity between 5 and 15 km (a factor 6 lower than
 379 the sensitivity at the near-surface). It is worth pointing out that the 0-2
 380 km AVK shows small negative values above 6 km (minimum of -0.12). These
 381 negative values tend to show that there is anti-correlation between the 0-2 km
 382 layer and the layers above 6 km. This can be understood in terms of emission
 383 and absorption. Indeed, because of the presence of the thermal inversion
 384 (Figure 4.e), emission takes place close to the surface. Above the thermal
 385 inversion (above 4 km), due to the negative temperature difference between
 386 two adjacent layers (i.e. temperature decreases with height), absorption takes
 387 place. Therefore, increasing the CO abundance close to the surface has the
 388 same effect on the PFS spectra as decreasing the CO abundance above the
 389 thermal inversion (the contrary is also true), and this explains the observed
 390 anti-correlation. However, regarding the difference in PFS sensitivity to the

391 two atmospheric regions as shown by the Jacobians (Figure 4.d), the anti-
 392 correlation is likely limited. Finally, given the very similar shape of all AVK
 393 functions, one has to conclude that the retrieved CO profile is not resolved
 394 vertically. The retrieval provides only one column that is mainly sensitive to
 395 the 0-6 km region.

Figure 6: (a) Histogram of the number of CO retrievals as a function of the maximum AVK value for the 0-2 km layer. (b) Histogram of the mean maximum AVK value (0-2 km layer) per bin of 4 K of thermal contrast (shown in absolute value).

396 For all night-time averaged spectra, the AVKs functions present also very
 397 similar shapes, which means that the vertical CO profile is not resolved
 398 for any of the retrievals. Because of this lack of vertical resolution, for all
 399 retrievals, we use as diagnostic of the sensitivity only the value of the AVK
 400 for the 0-2 km layer. It has indeed the largest values of all AVK functions
 401 for all observations and it is the one showing the most clearly the anti-
 402 correlation between the surface layers and the higher layers. In fact, the

403 anti-correlation remains limited for most of the retrievals, with 66% of the
 404 cases presenting a negative AVK value smaller than 0.1. Then, by noting
 405 that 86% of the observations have their first AVK peaking in the first layer
 406 and 14% in the second layer, we can safely conclude that for the entire set
 407 the maximum sensitivity is close to the surface. The exact location of the
 408 maximum depends principally on the temperature profile and on the thermal
 409 inversion. Figure 6.a shows the distribution of the maximum value of the 0-
 410 2 km AVK among the analysed set. 68% of the retrievals present maximum
 411 AVK values larger than 0.7, and these are found for the 0-2 km AVK peaking
 412 in the first layer. This indicates that for the majority of the analyzed cases,
 413 the retrieved CO abundance in the 0-2 km is mainly sensitive to the true CO
 414 abundance in that layer. The variations of the maximal value of the 0-2 km
 415 AVK are among other driven by the thermal contrast. Figure 6.b shows the
 416 mean maximal AVK value per bin of 4 K of thermal contrast. The impact of
 417 the latter on the maximal value is clear, with maximum value of 0.8-0.9 found
 418 for thermal contrast higher than 30 K (absolute value) and decreasing to 0.5
 419 for smaller values. Finally, as already mentioned, the CO vertical profile is
 420 not resolved for any of the retrievals. The majority of these provide only one
 421 column that is mainly sensitive to the 0-10 km atmospheric region.

422 3.2.3. Impact of the *a priori* profile

423 The DOFS of the example retrieval of Figure 4 is equal to 1.25. This
 424 corresponds to the number of patterns in the retrieved profile that are deter-
 425 mined independently of the *a priori*. This can be seen with the eigenvectors
 426 decomposition of the matrix $\mathbf{A}$ (reminding that the retrieved profile can be
 427 expressed as a linear combination of the eigenvectors of $\mathbf{A}$, section 3.1.1).

Figure 4 (f-j) shows the first five eigenvectors of $\mathbf{A}$ and the corresponding eigenvalues. The first eigenvector is associated to an eigenvalue of 0.99. This component of the retrieved profile is therefore well measured: its associated coefficient $\hat{z}_1$ is determined almost entirely from the true state (z_1). The second eigenvalue is equal to 0.25. The coefficient associated the second component is thus dominated by the *a priori* state (z_{a2}), and only a small part (around 25%) is determined from the true state z_2 . The following eigenvectors have finer structures, but their associated eigenvalues reach almost 0. The coefficients for all these components are all determined from the *a priori* state, and this confirms the very low vertical resolution of the retrieval.

Figure 7: (a) Histogram of the number of CO retrievals as a function of the DOFS. (b) Same as (a) but as a function of the first (blue) and second (orange) eigenvalues.

Figure 7.a shows the distribution of the DOFS values among all successful retrievals. Almost 90% present values larger than 0.9, showing that for the large majority of the retrievals, at least one pattern in the retrieved profile is

441 well measured, i.e., it is determined independently of the *a priori*. As can be
 442 seen from Figure 7.b, this pattern corresponds to the first eigenvectors (as-
 443 sociated eigenvalues within 0.9 and 1.0). However, for 84% of the retrievals,
 444 the second eigenvalue is close to 0, showing that the other patterns in the
 445 retrieved profile are determined from the *a priori* state. In fact, only a small
 446 fraction (16%) of the retrievals present a second eigenvalue slightly larger
 447 than 0.1, meaning that the second pattern in the retrieved profile is partly
 448 determined from the true state.

449 3.2.4. Other sources of errors

450 Two error sources have not been included in the error budget described in
 451 section 3.2.1 and both come from the propagation of the uncertainties on the
 452 retrieved temperature profile. There are two types of uncertainties associated
 453 to the temperature profile. The first one includes the smoothing and mea-
 454 surement errors and is calculated from the total error covariance matrix. This
 455 has been done in section 4.1 of the companion paper (Bauduin et al., 2020),
 456 and we have shown that this error varies between 0.5 K and 2 K depending
 457 on the altitude. It will be called the temperature retrieval error hereafter.
 458 The second one is associated to the uncertainty on the pressure (i.e., CO₂)
 459 profile. Indeed, we have made an assumption for the pressure profile (see
 460 (Bauduin et al., 2020)) and any error on it obviously propagates to the re-
 461 trieved temperature profile and subsequently to the retrieved CO abundance.
 462 This temperature uncertainty (hereafter called temperature-pressure error)
 463 has also been estimated in section 4.1 of the first part paper and evaluated to
 464 be within 0.5-4 K (in absolute value) depending on the altitude. These two
 465 types of temperature errors have been propagated on the CO retrieval sepa-

466 rately but following a similar procedure. For the 170 averaged spectra that
 467 we kept, we have redone the CO retrieval but the temperature profile has
 468 been shifted ($\pm$) by the estimated errors. Note that in the case of the tem-
 469 perature pressure error, the surface pressure has been shifted simultaneously
 470 by ± 0.9 hPa, because these two sources of uncertainty are intimately related
 471 (see Bauduin et al. (2020) for details). The CO error due to the temperature
 472 pressure error includes therefore the contributions of both the temperature
 473 profile and the surface pressure uncertainties. The error on CO has been
 474 taken as the difference between the new and the reference CO abundance.
 475 Note that this evaluation is performed only for the column-averaged VMR
 476 as the CO profile is not resolved.

477 The distributions of the two types of uncertainties among the retrievals
 478 are shown in Figure 8.a and Figure 8.b for the propagation of respectively
 479 the temperature retrieval error and the temperature-pressure error. For the
 480 former (Figure 8.a), the distribution is separately shown for a negative (blue
 481 bars) and a positive shift (orange bars) of the temperature profile. In both
 482 cases, more than 75% of the retrievals show that the CO uncertainty due to
 483 the temperature retrieval error is below 10%. The average is around 6% for
 484 both cases. Regarding the CO uncertainty due to the temperature-pressure
 485 error (Figure 8.b), its distribution is shown separately for the error coming
 486 from a surface pressure shift of -0.9 hPa (blue bars) and +0.9 hPa (orange
 487 bars). This distribution is similar to the one obtained in Figure 8.a, with the
 488 CO uncertainty due to the temperature-pressure error being on average 6%
 489 and below 10% for a large majority of the retrievals (more than 80 %).

490 It is worth reminding that Bouche et al. (2019) have evaluated the im-

Figure 8: (a) Histogram of the number of CO retrievals as a function of the CO uncertainty (absolute value) due to the temperature retrieval error. The blue and orange bars correspond to the CO uncertainty due respectively to a negative and positive shift of the retrieval error (Bauduin et al., 2020). (b) Same as panel (a) but for the CO uncertainty due to the temperature-pressure error. Note that this error includes the contributions of both the temperature profile and surface pressure uncertainties. The blue and orange bars correspond to the CO uncertainty due to the temperature error evaluated when shifting the surface pressure respectively of -0.9 hPa and +0.9 hPa (see Bauduin et al. (2020)).

491 pact of the instrumental line shape (ILS) formulation on the retrieved CO
 492 abundance. They showed that the uncertainty on the ILS can lead to a
 493 systematic uncertainty of around 5% on the CO retrieved total column. A
 494 re-evaluation of this error has been performed (not shown) and has led to a
 495 larger error on CO, of around 14%. Assuming that all the errors discussed
 496 above linearly add up, the total error on the CO 0-24 km column is within

32-41% on average. Note finally that, even if only spectra associated with a dust optical depth lower than 0.2 have been selected, because we have neglected the scattering some additional errors might contribute to the total error on the CO abundance. However, these errors should be low given the fact that scattering by dust does not contribute significantly to the spectral range considered (Billebaud et al., 2009).

3.3. Distribution

Figure 9.a shows the latitudinal distribution of the retrieved CO column-averaged VMRs as a function of L_s . As already mentioned in the previous section, only the retrievals with a satisfactory fit (corresponding to a fit with a RMS similar to the noise level) have been kept (170 out of 350), and only these have been used to build the distribution. Also, as a reminder, the detection method (section 2) already rejected a number of spectra: only 449 average spectra out of 1163 were kept and these were mostly observations with $180^\circ < L_s < 360^\circ$ and located in the Southern Hemisphere. The post-filtering based on the fit quality (for both temperature and CO retrievals, see previous sections) is further responsible for the lack of CO measurements in the regions and periods of pre-selected observations.

The mean CO column-averaged VMR calculated for the selected set of observations is 845 ppm, which is in good agreement with the averages calculated from daytime measurements (Smith et al., 2018; Bouche et al., 2020). As seen in Figure 9.a, the retrieved CO VMRs distribution is noisy, especially in the equatorial and northern mid-latitude regions. The individual total retrieval errors (Figure 9.b) account for most of the sharp variations observed between adjacent boxes. For instance, some of the retrievals located in 60° -

Figure 9: Seasonal cycle of the retrieved CO column-averaged VMR (ppm, a), of the associated total retrieval error (ppm, b), of the DOFS (d), of the surface temperature (K, e) and of the thermal contrast (K, f). Panel (c) shows the number of single PFS spectra averaged together.

522 40° S and 250° - 300° L_s has led to CO abundances unexpectedly high, of
 523 around 900-1000 ppm. These are associated to an error of 500-600 ppm,
 524 showing that the increase of the CO VMR is likely not real. The large CO
 525 VMRs (1100-1400 ppm) observed in 20° - 40° N and in 150° - 200° L_s are also
 526 associated with large errors (450-700 ppm), which could explain the sharp
 527 variations observed between the adjacent retrievals. Another, yet less impor-
 528 tant, reason for the variations observed is related to the fact that we analyse

529 average spectra built from an inhomogeneous number of PFS observations
 530 (Figure 9.c), including different contributions of the different Martian Years
 531 (in terms of the number of spectra included in the average). Nevertheless,
 532 one significant feature of variability emerges from the uncertainty: it corre-
 533 sponds to the known decrease of the CO VMR in the Southern Hemisphere
 534 during spring and summer due to the sublimation of the CO₂ ice. CO VMRs
 535 lower than 600 ppm are indeed retrieved from $L_s = 240^\circ$ southern to -30° .
 536 Interestingly, these CO measurements are those associated with the largest
 537 DOFS as shown in Figure 9.d. The Southern Hemisphere in spring and sum-
 538 mer is actually associated with the largest surface temperatures (larger than
 539 210 K, Figure 9.e) and the highest negative thermal contrasts (< -25 K, Fig-
 540 ure 9.f). This is observed for the analysed set but also for Mars in general.
 541 These temperature conditions are the best night-time conditions to measure
 542 CO and allow therefore the best CO night-time retrievals in this region and
 543 at this period.

544 4. Discussion and conclusion

545 This work has demonstrated the possibility of retrieving the Martian CO
 546 abundance from night-time PFS nadir spectra in the SW channel, exploiting
 547 the fundamental 1-0 band of CO. Analysing such observations is challenging
 548 because the SNR of single spectra is very low. The detection of CO required
 549 first averaging a sufficiently large number of PFS night-time spectra. More
 550 specifically 8 MYs of PFS observations have been averaged together for boxes
 551 of $5^\circ \times 10^\circ$ (latitude $\times L_s$) to guarantee a sufficient SNR for CO. A total of
 552 1163 averaged night-time spectra have been built, from which only a subset

553 was kept based on a detection method using BTDs in 21 CO-sensitive spectral
 554 channels. A positive detection of the 1-0 band of CO was assumed when the
 555 mean BTDs, calculated over the 21 CO lines, is larger than the NEDT. 449
 556 averaged spectra out of the 1163 turned out to have detectable CO lines
 557 and have been analysed using the Optimal Estimation method. CO vertical
 558 profiles have been successfully retrieved for 350 spectra and only 170 have
 559 been finally selected based on the quality of the fits.

560 These 170 CO measurements are not randomly distributed, but are mostly
 561 located in the Southern Hemisphere during spring and summer. Actually, one
 562 of the main result revealed in this work is that night-time CO is the most
 563 favourably measured from PFS observations in this region at this period.
 564 The well-known decrease of the CO VMR, related to the sublimation of CO₂
 565 at the South Pole, is indeed well observed in the night-time seasonal cycle
 566 measured in this work, thanks to the combination of two favourable con-
 567 ditions: high surface temperatures (larger than 210 K) and high negative
 568 thermal contrasts (< -20 K). High surface temperatures contribute to in-
 569 crease the SNR of single spectra while high thermal contrasts (negative or
 570 positive) favour the sensitivity to CO in the near-surface atmosphere. In the
 571 Southern Hemisphere in spring and summer, these two conditions are such
 572 that it is actually possible to measure the CO abundance from average spec-
 573 tra built using a smaller number of PFS observations, for instance recorded
 574 in the same MY. Figure 10 shows an example of such spectrum. It has been
 575 built using only 24 single night-time PFS observations recorded in MY 29
 576 at $238^\circ < L_s < 253^\circ$ and located in a box of $10^\circ \times 4^\circ$ (longitude $\times$ latitude).
 577 If this kind of average is possible for all MY, it would allow studying the

578 inter annual variability of night-time CO in the Southern Hemisphere during
 579 spring and summer; this could be investigated by future works. Note also
 580 that relaxing the criterion on dust and ice optical depths, could allow taking
 581 into account more spectra and improve further the results. Finally, given the
 582 two conditions required to detect CO during night with PFS, we anticipate
 583 that it will remain difficult to measure the CO abundance during the polar
 584 night, as the surface temperatures are extremely low (around 140-150 K).

Figure 10: (a) Example of average night-time PFS spectrum built using only spectra recorded in the same MY (see details in the text). A test retrieval has been performed and the result is in red. (b) CO vertical profile retrieved from the spectrum shown in (a).

585 The 170 retrieved CO profiles have been fully characterized. We have
 586 shown that the retrieved profiles are not resolved vertically. They are how-
 587 ever mostly sensitive to the CO abundance between 0 and 10 km (despite
 588 weak anti-correlation between the near-surface atmosphere and higher levels).
 589 With this sensitivity pushed towards the near-surface atmosphere thanks to
 590 the thermal inversions, night-time PFS observations provide stronger con-

591 straints on the near-surface CO abundance than daytime observations, which
 592 carry information mostly for the 0-20 km atmospheric region. However, it is
 593 important to remind that, because of their low SNR, night-time PFS observa-
 594 tions provide CO measurements with lower spatial and temporal resolutions
 595 than daytime observations. In addition to the vertical sensitivity, a complete
 596 error budget, including the propagation of two types of temperature uncer-
 597 tainties, has been performed. We have shown that the retrieval error on the
 598 0-24 km CO column is within 6-15% for a large majority of the retrievals and
 599 depends on the amount of CO in the atmosphere. This error is similar to
 600 that characterizing daytime CO measurements from single PFS spectra. In
 601 addition to this, we have estimated that an extra 6% of error comes from the
 602 temperature retrieval error and another 6% comes from the temperature un-
 603 certainty due to our hypothesis on the surface pressure and pressure profile.
 604 Note that the contribution of the uncertainty on the surface pressure to the
 605 error on the CO abundance is included in the temperature pressure error.
 606 A last source of uncertainty comes from the ILS formulation and has been
 607 estimated to be around 14%. Assuming that these errors linearly add up, the
 608 total error on the CO 0-24 km column is within 32-41 % on average. Finally,
 609 we have shown that the DOFS values are larger than 0.9 for the majority of
 610 the night-time retrievals, and this is also slightly larger than what has been
 611 observed for daytime PFS observations.

612 In conclusion, retrieving the CO abundance from night-time PFS obser-
 613 vations is a difficult task requiring spectral averaging and careful fitting and
 614 analysis. However, this exploratory work has firmly demonstrated that it
 615 was possible in favourable periods and regions. For these, the night-time

616 CO measurements contribute to significantly constrain the near-surface CO
617 abundance and they are therefore expected to be an important complement
618 of the highly resolved profiles retrieved from solar occultation and limb mea-
619 surements.

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