

1 Diurnal variations in oxygen and nitrogen isotopes of atmospheric 2 nitrogen dioxide and nitrate: implications for tracing NO_x oxidation 3 pathways and emission sources

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10 **Abstract.** The oxygen ($\Delta^{17}\text{O}$) and nitrogen ($\delta^{15}\text{N}$) isotopic compositions of atmospheric nitrate (NO_3^-) are widely used as
11 tracers of its formation pathways, precursor (nitrogen oxides (NO_x) $\equiv$ nitric oxide (NO) + nitrogen dioxide (NO_2)) emission
12 sources, and physico-chemical processing. However, the lack of observations on the multi-isotopic composition of NO_2
13 maintains significant uncertainties regarding the quantitative links between the isotopic composition of NO_x and NO_3^- ,
14 which ultimately may bias inferences on NO_3^- formation processes and distribution of sources, particularly in winter urban
15 atmospheres. We report here on the first simultaneous atmospheric observations of $\Delta^{17}\text{O}$ and $\delta^{15}\text{N}$ in NO_2 ($n = 16$) and NO_3^-
16 ($n = 14$). The measurements were carried out at sub-daily (~ 3 h) resolution over two non-consecutive days in an Alpine city
17 in February 2021. A strong diurnal signal is observed in both NO_2 and NO_3^- multi-isotopic composition. $\Delta^{17}\text{O}$ of NO_2 and
18 NO_3^- ranges from 19.6 ‰ to 40.8 ‰ and from 18.3 ‰ to 28.1 ‰, respectively. During the day and night, the variability of
19 $\Delta^{17}\text{O}(\text{NO}_2)$ is mainly driven by the oxidation of NO by ozone, with a substantial contribution from peroxy radicals in the
20 morning. NO_3^- mass balance equations, constrained by observed $\Delta^{17}\text{O}(\text{NO}_2)$, suggest that during the first day of sampling,
21 most of NO_3^- was formed locally from the oxidation of NO_2 by hydroxyl radicals by day, and via heterogeneous hydrolysis
22 of dinitrogen pentoxide at night. For the second day, calculated and observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ do not match, particularly daytime
23 values; the possible effects on $\Delta^{17}\text{O}(\text{NO}_3^-)$ of a Saharan dust event that occurred during this sampling period and of winter
24 boundary layer dynamics are discussed. $\delta^{15}\text{N}$ of NO_2 and NO_3^- ranges from -10.0 ‰ to 19.7 ‰ and from -4.2 ‰ to 14.9 ‰,
25 respectively. Consistent with theoretical predictions of N isotope fractionation, the observed variability of $\delta^{15}\text{N}(\text{NO}_2)$ is
26 explained by significant post-emission equilibrium N fractionation. After accounting for this effect, vehicle exhaust is found
27 to be the primary source of NO_x emissions at the sampling site. $\delta^{15}\text{N}(\text{NO}_3^-)$ is closely linked to $\delta^{15}\text{N}(\text{NO}_2)$ variability,
28 bringing further support to relatively fast and local NO_x processing. Uncertainties on current N fractionation factors during
29 NO_2 to NO_3^- conversion are underlined. Overall, this detailed investigation highlights the potential and necessity to use

30 simultaneously $\Delta^{17}\text{O}$ and $\delta^{15}\text{N}$ in NO_2 and NO_3^- in order to better constrain quantitative inferences on the sources and
31 formation chemistry of NO_3^- in urban environments in winter.

32 **1 Introduction**

33 Despite extensive efforts in emission controls in recent decades, global anthropogenic emissions of nitrogen oxides ($\text{NO}_x \equiv$
34 nitrogen monoxide (NO) + nitrogen dioxide (NO_2)) remain more than two orders of magnitude higher than before the
35 Industrial Revolution (Hoesly et al., 2018). Atmospheric nitrate ($\text{NO}_3^- \equiv$ nitric acid (HNO_3) + particulate nitrate ($p\text{-NO}_3^-$)), is
36 the main end-product of NO_x oxidation and a key component of fine particulate matter (PM), which adversely affects human
37 health (WHO, 2021) and contributes to climate change (Masson-Delmotte et al., 2021). NO_3^- can be transported far from
38 emission sources and can be removed from the atmosphere through dry and wet deposition within hours to days (Alexander
39 et al., 2020; Park et al., 2004). The additional input of this "reactive" nitrogen (N_r) into natural environments is known to
40 have detrimental consequences, particularly regarding biodiversity and water quality (Galloway et al., 2008; Vitousek et al.,
41 1997). It is therefore important to have a comprehensive understanding of NO_x emission sources and oxidation processes, on
42 which effective air quality and climate change mitigation strategies rely (e.g., Bauer et al., 2007; Huang et al., 2014; Shah et
43 al., 2018; Tsimpidi et al., 2008; Wang et al., 2013, 2020).

44
45 Several studies noted that the response of NO_3^- concentration in air to NO_x emission reduction is contrasted,
46 particularly in winter (e.g., Shah et al., 2018; Tørseth et al., 2012; Wang et al., 2020; Zhou et al., 2019). This is because a
47 variety of factors controls the NO_x conversion efficiency and the NO_3^- content in PM, including precursor emission sources,
48 complex multiphase chemical reactions with other reactive species, and environmental conditions (e.g., temperature, relative
49 humidity, solar radiation) (Zhang et al., 2015). It remain difficult to assess the contribution of each parameter to the non-
50 linear N_r chemistry, which is partly driven by close links between changes in aerosol acidity, gas-particle partitioning, and
51 atmospheric oxidation capacity (Shah et al., 2018; Fu et al., 2020; Shi et al., 2019; Li et al., 2021b). Ozone (O_3) and
52 hydroxyl radicals (OH) (Finlayson-Pitts and Pitts, 2000), are the major oxidants in the atmosphere whose chemical cycles
53 are largely controlled by solar radiation. As a result, there are significant diurnal and seasonal variations in NO_x chemistry
54 (e.g., Prabhakar et al., 2017; Alexander et al., 2020). Notably, NO_3^- formation is generally dominated by homogeneous OH
55 oxidation and heterogeneous O_3 chemistry during day/summer and night/winter, respectively (Alexander et al., 2020).
56 However, assessing the relative contributions of individual formation channels, together with their sensitivity to
57 environmental parameters, is not straightforward and requires extensive in situ observations combined with modelling tools
58 (e.g., Alexander et al., 2020; Brown, 2006; Newsome and Evans, 2017; Xue, 2022; Prabhakar et al., 2017).

59

60 Upon release into the atmosphere, NO_x, mainly emitted as NO, undergoes oxidation to form NO₂. During the day, a
 61 rapid photochemical equilibrium is established between NO and NO₂, known as the "photostationary state" (PSS; Leighton,
 62 1961), via key interconversion reactions (Reactions R1–R3):



63 This cycle can be disturbed by peroxy radicals (RO₂ ≡ hydroperoxyl radical (HO₂) + methyl peroxy radical (CH₃O₂)) via
 64 typically Reaction (R4):



65 Note that in polluted atmospheres where NO_x mixing ratios often exceed ppb levels, Reaction (R4) followed by Reactions
 66 (R1)–(R2) lead to the formation of O₃ (Crutzen, 1979). Although the role of RO₂ in NO_x oxidation is crucial in O₃ formation
 67 and NO_x oxidation rate, measuring RO₂ mixing ratio remains challenging due to the need for state-of-the art instrumentation
 68 coupled with photochemical models to establish chemical budgets (e.g., Ren et al., 2006; Tan et al., 2018). While NO is
 69 usually oxidised relatively quickly into NO₂ in summer due to the high levels of O₃ and solar radiation, the shorter day
 70 length and lower temperature in winter result in a contrasted NO_x cycling. In particular, the formation of a temperature
 71 inversion at the surface can trap pollutants emitted close to the surface in a shallow layer for hours to days (e.g., Largeron
 72 and Staquet, 2016; Olofson et al., 2009). Under those conditions, it is not uncommon for O₃ levels to be very low due to
 73 quasi-complete titration by NO, which can have further impacts on the atmospheric oxidation capacity. However,
 74 pronounced O₃ pollution episodes may also arise in winter in highly polluted areas, such as in oil-producing regions due to
 75 intense VOCs emissions (Edwards et al., 2014) or in China, where pollution control strategies mainly target NO_x while
 76 VOCs emissions remain more or less constant (Ren et al., 2022). In urban areas, NO₂ is generally mainly removed from the
 77 atmosphere by reaction with OH during the day via Reaction (R5) (Dentener and Crutzen, 1993):



78 NO₂ can also react with O₃ to form nitrate radicals (NO₃) via Reaction (R6):



79 However, NO₃ is rapidly photolyzed during the day regenerating back NO₂ (Wayne et al., 1991). Another important NO₃
 80 loss reaction is that with NO in polluted environments (Brown and Stutz, 2012). At night, without photolytic activity and

81 lower precursor emissions, the lifetime of NO₃ radicals substantially increases. NO₃ reacts with NO₂ to form dinitrogen
 82 pentoxide (N₂O₅; Reaction R7), which then undergoes heterogeneous hydrolysis to form HNO₃ (Reaction R8):



83 Reaction (R7) is temperature dependent, so N₂O₅ can eventually decompose to reform NO₂ and NO₃, with the N₂O₅/NO₃
 84 ratio being negatively correlated with temperature. N₂O₅ is an important nocturnal sink for NO_x, notably in winter in urban
 85 atmospheres due to high aerosol loads and low temperatures. However, the efficiency of Reaction (R8) is difficult to
 86 determine because it strongly depends on parameters such as the aerosol surface density and its chemical composition
 87 (Brown, 2006), which are not often well characterised. In addition, NO₃ can react with hydrocarbons to produce HNO₃,
 88 which could significantly contribute to the formation of NO₃⁻ in industrialised regions with high hydrocarbon emissions
 89 (Brown et al., 2011). It is estimated that the Reactions (R1) to (R8) lead to the formation of 82 % of NO₃⁻ near the surface on
 90 a global scale (Alexander et al., 2020). In polluted environments, the respective contributions of Reaction (R5) (OH
 91 pathway) and Reactions (R6)–(R8) (N₂O₅ pathway) are more contrasted and are still debated (e.g., Chan et al., 2021; Fu et
 92 al., 2020). In addition, the reaction of N₂O₅ with chlorine on aerosols can contribute to NO₃⁻ production in urban
 93 atmospheres (Thornton et al., 2010), with further impacts on O₃ production in continental polluted atmosphere in winter
 94 (Wang et al., 2019a). Other reactions, such as those involving halogen and organic intermediates, may become significant for
 95 NO₃⁻ production in specific regions, such as in polar, oceanic, and coastal areas (Alexander et al., 2020; Penkett et al., 2007;
 96 Savarino et al., 2013; Simpson et al., 2015).

97

98 To help better constrain the atmospheric N_r chemistry and budget, the last three decades have seen a growing interest in
 99 stable oxygen (O) and nitrogen (N) isotopes, notably in NO₃⁻ (Elliott et al., 2019; Savard et al., 2018). The isotopic
 100 composition is reported as an isotopic enrichment (δ) with respect to a reference material, defined as $\delta = (R_{\text{sample}} /$
 101 $R_{\text{reference}} - 1)$, and expressed in per mill (‰). *R* refers to the elemental abundance ratio of the heavy isotope to the light
 102 isotope (e.g., ¹⁸O/¹⁶O; ¹⁷O/¹⁶O; ¹⁵N/¹⁴N) in the sample, and in an international isotopic reference material (Vienna Standard
 103 Mean Ocean Water for O; Li et al., 1988), and atmospheric N₂ for N; Mariotti, 1984). A powerful tool to help trace the
 104 relative importance of different NO_x to NO₃⁻ oxidation pathways is through the use of the ¹⁷O-excess ($\Delta^{17}\text{O} = \delta^{17}\text{O} - 0.52 \times$
 105 $\delta^{18}\text{O}$). $\Delta^{17}\text{O}$ is transferred to NO₃⁻ by O₃ which possesses a very unique $\Delta^{17}\text{O}$ (26.2 ± 1.3 ‰; Vicars and Savarino, 2014)
 106 due to mass-independent fractionation during its formation process (Thiemens, 2006). In comparison, the $\Delta^{17}\text{O}$ of other
 107 atmospheric oxidants such as OH is near zero due to isotopic exchange with atmospheric water vapor (Dubey et al., 1997).
 108 Similarly, as the isotopic anomaly of atmospheric O₂ is very close to 0 ‰ (Barkan and Luz, 2003), and since RO₂ are mostly
 109 produced by the reactions R + O₂ and H + O₂, $\Delta^{17}\text{O}$ of RO₂ can be considered negligible (Alexander et al., 2020). Therefore,

110 $\Delta^{17}\text{O}$ in NO_3^- represents a unique tracer of the O_3 implication in its formation processes, that can provide valuable
 111 constraints on the relative contributions of individual reactions (e.g., Morin et al. 2011, Alexander et al., 2009; Michalski et
 112 al., 2003). By a simple mass balance calculation of O atoms in NO_3^- , the ^{17}O -excess of NO_3^- produced by an individual NO_2
 113 to NO_3^- conversion process i ($\Delta^{17}\text{O}(\text{NO}_3^-)_i$) can be expressed as:

$$\Delta^{17}\text{O}(\text{NO}_3^-)_i = \frac{2}{3} \times \Delta^{17}\text{O}(\text{NO}_2) + \frac{1}{3} \times \Delta^{17}\text{O}(\text{add. O})_i \quad (1)$$

114 where $\Delta^{17}\text{O}(\text{NO}_2)$ is the ^{17}O -excess of atmospheric NO_2 and $\Delta^{17}\text{O}(\text{add. O})_i$ is the transferrable ^{17}O -excess of the oxidant
 115 responsible for the conversion of NO_2 in NO_3^- (Michalski et al., 2003). From Eq. (1), if $\Delta^{17}\text{O}(\text{NO}_2)$ is constrained, one can
 116 derive individual $\Delta^{17}\text{O}$ transfer to NO_3^- relative to a i conversion process and compare this value with observed $\Delta^{17}\text{O}(\text{NO}_3^-)$.
 117

118 Recent studies in urban areas have attempted to interpret the variability of $\Delta^{17}\text{O}(\text{NO}_3^-)$ in aerosols in order to quantify
 119 the relative contribution of homogeneous and heterogeneous processes to NO_3^- formation (e.g., Fan et al., 2023, 2022; He et
 120 al., 2020, 2018; Lim et al., 2022; Wang et al., 2023, 2019b; Kim et al., 2023; Zhang et al., 2022b; Li et al., 2022b). However,
 121 to that end, it is necessary to have a clear quantitative understanding of the transfers of $\Delta^{17}\text{O}$ in the N_r cycle. To date, due to
 122 very limited observational data, there is a lack of well-establish knowledge on the dynamics of $\Delta^{17}\text{O}$ in NO_2 , the key
 123 intermediate species in the formation of NO_3^- . Consequently, strong assumptions about $\Delta^{17}\text{O}(\text{NO}_2)$ have to be made when
 124 interpreting $\Delta^{17}\text{O}(\text{NO}_3^-)$ measurements, which could potentially lead to biased conclusions. Notably, the most difficult
 125 regions for the interpretation of NO_3^- records are potentially polluted areas where the isotopic composition of NO_2 is
 126 expected to be highly variable in space and time. Most studies typically estimate $\Delta^{17}\text{O}(\text{NO}_2)$ during the day by assuming that
 127 an isotopic steady state (ISS) is reached between NO_x and O_3 resulting in $\Delta^{17}\text{O}(\text{NO}_2)$ depending only on the relative
 128 contributions of different oxidants to NO oxidation. A recent study reported the first in situ observations of $\Delta^{17}\text{O}(\text{NO}_2)$ in an
 129 urban environment (Grenoble, France) in spring (Albertin et al., 2021). Time resolved NO_2 sampling (ca. 3 h) during 24 h
 130 revealed a strong diurnal cycle in $\Delta^{17}\text{O}(\text{NO}_2)$, reaching ca. 40 ‰ during the day and decreasing down to ca. 20 ‰ at night.
 131 The observed $\Delta^{17}\text{O}(\text{NO}_2)$ values and diurnal variability were consistent with its expected behaviour derived from $\Delta^{17}\text{O}$ mass
 132 balance equations under the ISS assumption during the day. However, this first $\Delta^{17}\text{O}(\text{NO}_2)$ dataset is very short. More
 133 measurements are needed to test in different environments and season the applicability of this new isotopic tool and to assess
 134 whether the ISS is still valid. At night, the low $\Delta^{17}\text{O}(\text{NO}_2)$ measured by Albertin et al. (2021) is consistent with the oxidation
 135 of freshly emitted NO by O_3 . Nonetheless, since the timescale for the oxidation of NO_2 into NO_3^- is thought to exceed the
 136 duration of the night (Alexander et al., 2020), it is also common to assume that the isotopic composition of nocturnal NO_2
 137 reflects more daytime formation and conditions of the previous days. While this assumption may hold true in remote areas
 138 (Morin et al., 2011), significant uncertainties subsist in urban areas where the nighttime NO_3^- chemistry may be more

139 efficient. In such circumstances, the production of NO_3^- from NO_2 formed at night would lead to a lower than expected $\Delta^{17}\text{O}$
140 transfer to NO_3^- . For these reasons, the dual survey of the O isotopic composition of NO_2 and NO_3^- would certainly help to
141 accurately interpret $\Delta^{17}\text{O}(\text{NO}_3^-)$ observations in polluted atmospheres, particularly with sampling at sub-daily time scales
142 which would allow to study the diurnal dynamics of $\Delta^{17}\text{O}(\text{NO}_2)$ and its links with $\Delta^{17}\text{O}(\text{NO}_3^-)$.

143
144 In addition to $\Delta^{17}\text{O}$, $\delta^{15}\text{N}$ in NO_3^- ($\delta^{15}\text{N}(\text{NO}_3^-)$) can be used as a tracer of NO_3^- sources and/or chemical processing. As
145 different NO_x emission sources have often distinct $\delta^{15}\text{N}$ -fingerprints depending on the NO_x production mechanism (Heaton,
146 1990; Felix et al., 2012; Fibiger and Hastings, 2016; Walters et al., 2015a, b; Yu and Elliott, 2017; Miller et al., 2018),
147 $\delta^{15}\text{N}(\text{NO}_3^-)$ is a potentially valuable tool to trace the origins of its gaseous precursor. However, due to N fractionation effects
148 associated with physico-chemical processing, $\delta^{15}\text{N}$ is altered during the conversion of NO_x to NO_3^- (Elliott et al., 2019).
149 Therefore, the variability of $\delta^{15}\text{N}(\text{NO}_3^-)$ can be attributed to: (1) a change in NO_x emission sources and (2) N isotopic
150 fractionations between NO and NO_2 , between NO_2 and NO_3^- , and during the transport of NO_3^- . These effects co-exist with
151 relative contributions varying according to environmental conditions and the mix of NO_x emissions. Numerous observations
152 in diverse environments have emphasised the substantial influence of N fractionation effects in altering the original ^{15}N
153 composition of emitted gaseous NO_3^- precursors (e.g., (Bekker et al., 2023; Chang et al., 2018; Geng et al., 2014; Luo et al.,
154 2023; Vicars et al., 2013; Li et al., 2021a). Although some N fractionation factors are available from calculations (Walters
155 and Michalski, 2015) and laboratory experiments (Li et al., 2020; Walters et al., 2016), there is still a lack of observational
156 constraints on the magnitude of the N isotopic partitioning between NO_x and NO_3^- , which could lead to biased
157 interpretations of $\delta^{15}\text{N}(\text{NO}_3^-)$ observations.

158
159 Following the preliminary work of Albertin et al. (2021), this study presents for the first time the simultaneous
160 measurements of the atmospheric NO_2 and NO_3^- multi-isotopic compositions. The sampling took place at high temporal
161 resolution (~ 3 h) in late February 2021 in an urban Alpine city. $\Delta^{17}\text{O}/\delta^{15}\text{N}$ data of NO_2 and NO_3^- , meteorological
162 parameters, and atmospheric observations (NO , NO_2 , O_3 , and PM) are collated in order to investigate diurnal N_r chemistry,
163 N fractionation effects and NO_x emissions. Our winter measurements extend the atmospheric NO_2 multi-isotopic
164 composition record which is only composed of spring measurements performed during a single day by Albertin et al. (2021).
165 The general aim of this case study is to test NO_2 -based isotopic approaches for tracing the origins and fate of NO_x , for
166 instance in urban areas on sub-daily time scales. The added value of $\Delta^{17}\text{O}(\text{NO}_2)$ measurements in N_r chemistry studies is
167 more critically assessed here than in Albertin et al. (2021) through the use of accurate NO_x measurements. Besides, using the
168 isotopic theoretical framework developed previously, we explore the potential benefits of combining isotopic observations of
169 NO_2 and NO_3^- to gain a more detailed and quantitative on the links between atmospheric N_r chemistry processes and
170 variability in NO_2 and NO_3^- isotopic composition. The framework used in inferring dominant NO_x emission sources from
171 NO_2 $\delta^{15}\text{N}$ measurements is also tested.

172 2 Material and methods

173 2.1 Study site and sample collection

174 The study was conducted in February 2021 in Chamonix-Mont-Blanc, France, (45°55'21" N, 6°52'11" E; altitude 1035 m
175 above sea level). This narrow (~2 km wide on average in Chamonix) 23 km Alpine valley of about 12,000 inhabitants is
176 surrounded by high-elevation mountains. The city can experience severe PM pollution events during the winter season,
177 mainly due to wood-combustion for domestic heating and road traffic (Chazette et al., 2005; Quimbayo-Duarte et al., 2021;
178 Weber et al., 2018; Aymoz et al., 2007). The study's sampling site was located at a CNRS (Centre National de la Recherche
179 Scientifique) facility in a residential area, 1.2 km south of the Chamonix city centre, and 1.4 km north of the Mont-Blanc
180 tunnel. Ambient air monitoring inlets and off-line gas and aerosol samplers were installed on the facility's terrace, 3 m above
181 the ground level. Over the campaign, the surface was partly covered with snow.

182

183 Atmospheric particles (aerosols) were collected using a high-volume sampler (Digitel®, DH77, total suspended particle
184 inlet, 1 m³ min⁻¹) and glass fiber (GF) filters (Whatman®, 150 mm-diameter). Evaluating the collection efficiency of total
185 NO₃⁻ has long been debated (e.g., Schaap et al., 2002; Appel et al., 1980) and, although not free from sampling artefacts
186 (e.g., potential volatilisation of HNO₃ after exposure to ambient air), GF filters have been used on several times to study
187 nitrate isotopes, mainly in coastal sites (e.g., Savarino et al., 2007; Michalski et al., 2003; Morin et al., 2009; Frey et al.,
188 2009; Morin et al., 2007a; Patris et al., 2007; Vicars et al., 2013). Under these conditions, the aerosol alkalinity is supposed
189 to allow the collection of HNO₃ (Prospero and Savoie, 1989). In our case, as the ambient air in Chamonix is expected to be
190 free of sea salt, the GF filters were not washed before use in order to keep the initial NaCl coating inherited from the filter
191 manufacturing process. Therefore, in addition to *p*-NO₃⁻, we are confident that the high GF filter NaCl loading allowed the
192 quantitative collection of HNO₃ at our site, as it has been shown previously in the literature (Appel et al., 1981; see also our
193 reply and data to the comments of Reviewer #2 on this specific issue). Two field blanks were performed to evaluate the
194 initial content of trace elements and possible contamination during handling. Atmospheric NO₂ was collected concurrently to
195 filter samples using a pre-cleaned honeycomb denuder tube coated with a mixture of 2.5 M KOH (in methanol) and ultrapure
196 guaiacol inserted into a ChemComb® 3500 speciation cartridge (Thermo Scientific®, USA). A second coated denuder was
197 placed in series into the cartridge to check for NO₂ breakthrough. After sampling, denuders were rinsed with 10 mL of
198 deionised water to solubilised trapped NO₂. Detailed information on the denuder sampling protocol is available in Albertin et
199 al. (2021). Similar to blank filters, two blank denuders were performed. Blank filters and denuders were subjected to same
200 handling, storage, and analytical treatment as field samples. Filters and denuder extractions were stored and transported
201 frozen to IGE (Grenoble, France) for analysis.

202

203 Following the objective to investigate the diurnal isotopic composition of NO₂ and NO₃⁻, denuder and filter samplings
204 were conducted continuously for 24 hours with sampling time steps ranging from 1:30 to 7:30. During the day, denuder and

205 filter samplings were synchronised. At night, two filter samplings were performed while three sets of denuder tubes were
206 collected. This sampling protocol was conducted during two non-consecutive days, from 19 February 2021 21:00 local time
207 (LT) to 20 February 2021 21:00 LT (sampling period #1 = SP 1) and from 24 February 2021 7:30 LT to 25 February 2021
208 7:30 LT (sampling period #2 = SP 2).

209 2.2 Chemical and isotopic analysis

210 Concentrations of major ions from filter extractions in deionised water were determined by ion chromatography
211 (Thermo Scientific™ Dionex™ Integrion™ HPIC). Reported to the total filter surface, the NO_3^- contribution from blank
212 filters represented on average (8 ± 9) % of sampled NO_3^- . Atmospheric mass concentrations (expressed in $\mu\text{g m}^{-3}$) were
213 calculated as the ratio of the total ion filter loading (corrected for the blank contribution) to the total volume of air pumped
214 through the filter at STP conditions. NO_2^- concentration in denuder extractions were first estimated using the Griess-
215 Saltzmann reaction and UV-Vis spectrometry at 544 nm. Even though the eluted matrix can interfere with colorimetric
216 analyses, measured concentrations on first denuder tubes were relatively well correlated with ambient NO_2 measurements
217 during atmospheric sampling and allowed to give indications on field blanks and on the volume needed to perform isotopic
218 analysis.

219
220 Isotopic analyses were performed using an isotope ratio mass spectrometer (IRMS, Thermo Finnigan™ MAT 253) for
221 analyses of $^{15}\text{N}/^{14}\text{N}$, $^{17}\text{O}/^{16}\text{O}$, and $^{18}\text{O}/^{16}\text{O}$ in NO_3^- and NO_2 samples. Briefly, NO_3^- from filter extractions were converted
222 into gaseous N_2O by the bacterial denitrifier method (Sigman et al., 2001; Casciotti et al., 2002; Kaiser et al., 2007) in which
223 ≈ 100 nmol of NO_3^- ions were injected into a 2 mL of a bacteria medium (strain of *Pseudomonas aureofaciens*) under
224 anaerobic conditions. NO_2 denuder extractions were treated separately with the azide method (McIlvin and Altabet, 2005;
225 Albertin et al., 2021) in which 2 mL of a sodium azide 2M / acetic acid 100 % buffer were injected into ≈ 100 nmol of NO_2^-
226 allowing quantitative conversion into N_2O . For both filter and denuder extractions, ions were converted into N_2O which was
227 further thermally decomposed into O_2 and N_2 in a gold tube heated at 850 °C. Then, O_2 and N_2 molecules were separated on
228 a chromatography column and sent separately into the IRMS for the dual analysis of O and N isotopes (see Morin et al.,
229 2009 for more details on the analytical line). The isotopic composition of NO_3^- samples was analysed in triplicate (the mean
230 value of replicate measurements and the associated repeatability are reported in Table S2 in the Supplement). The limited
231 amount of NO_2 samples did not allow for replicate measurements. From UV-vis analysis, all NO_2 samples presented a
232 negligible blank (< 4 %; mean of 1.7 nmol ml^{-1}) except for the sample collected between 13:30 and 16:30 LT during SP 2
233 which shown a blank around (14.0 ± 1.4) %. Therefore, the measured $\Delta^{17}\text{O}$ of this sample was corrected for blank effect
234 assuming that the contaminated NO_2^- possessed a $\Delta^{17}\text{O} = 0$ ‰. No correction from this blank effect was applied on the $\delta^{15}\text{N}$
235 measurements of NO_2 because the $\delta^{15}\text{N}$ fingerprint of the contamination could not be characterised. This uncertainty is
236 propagated in the calculations of Section 3 and considered in the discussions. Possible isotopic changes resulting from the

conversion and analysis process of NO_3^- and NO_2 samples were evaluated using international NO_3^- and NO_2^- isotopic reference materials, respectively (Table S3 in the Supplement). Accuracy of the analytical method was estimated as the standard deviation (σ) of the residuals between measurements of the reference materials and their expected values. In our study, average measurement uncertainties on $\delta^{15}\text{N}$, $\delta^{17}\text{O}$, $\delta^{18}\text{O}$, and $\Delta^{17}\text{O}$ were estimated to be $\pm 0.3\text{‰}$, $\pm 0.9\text{‰}$, $\pm 1.3\text{‰}$, and $\pm 0.4\text{‰}$, respectively, for NO_3^- samples and $\pm 0.3\text{‰}$, $\pm 0.4\text{‰}$, $\pm 0.9\text{‰}$, and $\pm 0.3\text{‰}$, respectively, for NO_2 samples. Detailed information about the calibration procedure can be found in Morin et al. (2009) and in Albertin et al. (2021) for NO_3^- and NO_2 samples, respectively.

2.3 Ancillary data

During atmospheric samplings, surface NO_x mixing ratios were measured at the study site using an incoherent broadband cavity-enhanced absorption spectrometer for NO_2 (IBBCEAS; Barbero et al., 2020) and an optical-feedback cavity-enhanced absorption spectrometer for NO (OFCEAS; Richard et al., 2018). PM concentrations (PM_{10} and $\text{PM}_{2.5}$) were monitored by an optical particle counter (GRIMM®, EDM 164). O_3 mixing ratio was monitored at the local air quality monitoring site located a kilometre north of the sampling site (Environnement SA®, O3 42M; <https://www.atmo-auvergnerhonealpes.fr/>, last access: 5 November 2021). Surface temperature (T_{surface}) and relative humidity (RH) were measured by a portable logger (Tinytag, TGP-4500, Gemini Data Loggers) located at the air quality monitoring site. Vertical temperatures were measured from 11 similar loggers fixed along the Plan-Praz cable car ($45^\circ 55' 39''$ N, $6^\circ 51' 55''$ E) from 1098 to 2021 m above sea level (data obtained from personal communications with C. Coulaud, IGE). The NO_2 photolysis rate (J_{NO_2}) was calculated for the two sampling periods using a photochemical boxmodel (CiTTyCAT version 2.02; Galeazzo et al., 2018; Pugh et al., 2012) using the Fast-J photolysis scheme of Wild et al. (2000) and a surface albedo fixed to 0.65, a value representative of a snow-covered surface (average value between fresh and old snow; more details can be found in Text S1 in the Supplement).

2.4 Interpretation framework for isotopic signals

In this section, we briefly state the key concepts and equations necessary to interpret isotopic signals measured in NO_2 and NO_3^- . A more detailed description and complete equation derivations can be found in cited references.

2.4.1 $\Delta^{17}\text{O}$ mass balance equations

Because NO_2 and NO_3^- loss processes do not fractionate in terms of the oxygen mass-independent anomaly and considering that each source reaction induces a transfer of $\Delta^{17}\text{O}$ to NO_2 and NO_3^- , one considers the mass conservation of $\Delta^{17}\text{O}$ during fractionation processes in the N_r cycle. Hence, one can implement $\Delta^{17}\text{O}$ in the general mass balance equation of NO_2 and NO_3^- . An overall expression of the time derivative of $\Delta^{17}\text{O}$ in the species X ($\Delta^{17}\text{O}(\text{X})$; with $\text{X} = \text{NO}_2$ or NO_3^-) is derived as a function of its deviation from $\Delta^{17}\text{O}$ transferred through each production channel i (P_i) ($\Delta^{17}\text{O}_i(\text{X})$), weighted according to the relative contributions of the production channels (Vicars et al., 2013):

$$\frac{d}{dt}(\Delta^{17}\text{O}(\text{X})) = \frac{1}{\tau(\text{X})} \times \sum_i \frac{P_i}{\sum_i P_i} \times (\Delta^{17}\text{O}_i(\text{X}) - \Delta^{17}\text{O}(\text{X})) \quad (2)$$

where P_i expresses reaction rate constant times the atmospheric concentrations of reacting species, and τ is the atmospheric lifetime of the species X at steady state ($\tau = [\text{X}]/\sum_i P_i$ with $[\text{X}]$ being the atmospheric mixing ratio of the species X).

During the day, the rapid photochemical cycling of NO_x (Reactions R1–R4) leads to an isotopic equilibrium between NO and NO_2 , i.e., $\Delta^{17}\text{O}(\text{NO}) \approx \Delta^{17}\text{O}(\text{NO}_2)$ (Michalski et al., 2014). Therefore, using the steady state approximation, and considering $\text{NO} + \text{O}_3$ (Reaction R3) and $\text{NO} + \text{RO}_2$ (Reaction R4) as the main sources of NO_2 at our site, the overall daytime $\Delta^{17}\text{O}$ in NO_2 can be expressed by:

$$\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2) \approx T_{\text{NO}+\text{O}_3} \times \Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) \quad (3)$$

where $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ is the O_3 isotopic anomaly transferred to NO through Reaction (R3) (Savarino et al., 2008). $T_{\text{NO}+\text{O}_3}$, sometimes named A in the literature, represents the proportion of O atoms originating from O_3 in NO_2 , and hence the relative importance of Reaction (R3) in the conversion of NO into NO_2 (Michalski et al., 2003; Morin et al., 2007b; Albertin et al., 2021):

$$T_{\text{NO}+\text{O}_3} = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3]}{k_{\text{NO}+\text{O}_3}[\text{O}_3] + k_{\text{NO}+\text{RO}_2}[\text{RO}_2]} \quad (4)$$

where $k_{\text{NO}+\text{O}_3}$ and $k_{\text{NO}+\text{RO}_2}$ are the kinetic constants of Reactions (R3) and (R4), respectively. The kinetic constants used in this study are listed in Table A1 in the Appendix. At night, considering that (1) $\Delta^{17}\text{O}(\text{NO}) \approx 0$ ‰ (NO_x emission without NO_2 recycling), (2) no $\Delta^{17}\text{O}$ equilibrium between NO and NO_2 (no photochemical cycling), and (3) O_3 is the main oxidant of NO (no nighttime production of RO_2), $\Delta^{17}\text{O}(\text{NO}_2)$ is determined by the $\Delta^{17}\text{O}$ transfer via Reaction (R3) and by the nighttime residuals of NO_2 formed during the previous daytime hours (Albertin et al., 2021) following:

$$\Delta^{17}\text{O}_{\text{night}}(\text{NO}_2) \approx x \times \Delta^{17}\text{O}_{\text{day}}(\text{NO}_2) + \frac{(1-x)}{2} \times (\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) + \Delta^{17}\text{O}(\text{NO})) \quad (5)$$

where x is the fraction of NO_2 formed during the day to the total NO_2 measured at night.

At our sampling site, we hypothesise Reaction (R5) (OH pathway) and Reactions (R6)–(R8) (N_2O_5 pathway) as the main daytime and nighttime NO_3^- production channels, respectively. At steady state, from Eq. (2), we derive general

expressions for $\Delta^{17}\text{O}$ in NO_3^- during the day and night, associated with the OH and N_2O_5 pathways, respectively (Alexander et al., 2020):

$$\Delta^{17}\text{O}_{\text{day}}(\text{NO}_3^-) \approx \frac{2}{3} \times \Delta^{17}\text{O}_{\text{day}}(\text{NO}_2) \quad (6)$$

$$\Delta^{17}\text{O}_{\text{night}}(\text{NO}_3^-) \approx \frac{2}{3} \times \Delta^{17}\text{O}_{\text{night}}(\text{NO}_2) + \frac{1}{6} \times \Delta^{17}\text{O}_{\text{NO}_2+\text{O}_3}(\text{NO}_3) \quad (7)$$

where $\Delta^{17}\text{O}_{\text{NO}_2+\text{O}_3}(\text{NO}_3)$ is the ^{17}O -excess transfer from O_3 to NO_3 during Reaction (R6) (Berhanu et al., 2012). Without wet scavenging, dry deposition is the main sink of NO_3^- (Park et al., 2004). Assuming a mean NO_3^- deposition velocity of 0.5 cm s^{-1} (mean value of the dry deposition velocities of HNO_3 and $p\text{-NO}_3^-$; Zhang et al., 2009), and considering the maximum daytime and minimum nighttime boundary layer heights of 500 and 100 m above ground level, respectively (estimations based on measured vertical temperature profiles; Fig. S2 in the Supplement), the estimated residence time of NO_3^- against dry deposition can reach up to 28 hours during the day, and 6 hours at night (Table B1). Therefore, on sub-daily time scales, the ^{17}O -excess in NO_3^- during the day is more likely to reflect a combination of daytime and nighttime production processes than during the night. Note that, our estimated residence times for NO_3^- against dry deposition are upper limits as they represent the time required to reduce by a factor e the concentration of NO_3^- present at the top of the boundary layer; NO_3^- close to the surface would have a much shorter residence time.

2.4.2 Nitrogen isotopic fractionation effects

Each source of NO_x generates a $\delta^{15}\text{N}$ fingerprint which depends on the type and conditions (temperature and pressure) of combustion, and on the type of fuel (e.g., coal, oil, gas) (Heaton, 1990; Felix et al., 2012; Fibiger and Hastings, 2016; Walters et al., 2015a, b; Yu and Elliott, 2017; Miller et al., 2018). The mean $\delta^{15}\text{N}$ of NO_x ($\delta^{15}\text{N}(\text{NO}_x)$) emitted in the atmosphere results from the sum of each NO_x emission $\delta^{15}\text{N}$ fingerprint weighted by their relative contribution to the total NO_x emissions. Once in the atmosphere, NO_x is subjected to oxidation processes and isotopic exchanges that alter the initial $\delta^{15}\text{N}(\text{NO}_x)$. As a result, $\delta^{15}\text{N}$ in NO_2 and in NO_3^- is a complex function of both the $\delta^{15}\text{N}$ signature of NO_x emissions and N isotopic effects. These latter can be categorised into three groups: (1) the equilibrium isotope effect (EIE), (2) the kinetic isotope effect (KIE), and (3) the photochemical isotope fractionation effect (PHIFE) (Miller and Yung, 2000; Young et al., 2002). The magnitude of these isotopic effects is quantified as ^{15}N enrichment factor (ϵ), which is defined as $(\alpha - 1)$, where α represents the N isotopic fractionation factor.

311 A general expression for $\delta^{15}\text{N}(\text{NO}_2)$ can be derived as a function of a factor F_{N} which represents the overall N isotopic
 312 fractionation effects between NO_x emissions and NO_2 (expressed in ‰), the fraction of NO_2 with respect to NO_x ($f_{\text{NO}_2} =$
 313 $[\text{NO}_2]/[\text{NO}_x]$), and of $\delta^{15}\text{N}(\text{NO}_x)$ (Albertin et al., 2021; Li et al., 2020):

$$\delta^{15}\text{N}(\text{NO}_2) = F_{\text{N}} \times (1 - f_{\text{NO}_2}) + \delta^{15}\text{N}(\text{NO}_x) \quad (8)$$

314 Therefore, the ^{15}N isotopic shift between $\delta^{15}\text{N}(\text{NO}_2)$ and $\delta^{15}\text{N}(\text{NO}_x)$ is given by:

$$\delta^{15}\text{N}(\text{NO}_2) - \delta^{15}\text{N}(\text{NO}_x) = \Delta^{15}(\text{NO}_2 - \text{NO}_x) = F_{\text{N}} \times (1 - f_{\text{NO}_2}) \quad (9)$$

315 Physico-chemical processes between NO and NO_2 can preferentially promote or deplete ^{15}N in NO_2 with respect to emissions
 316 of NO_x (i.e., $\delta^{15}\text{N}(\text{NO}_2) \neq \delta^{15}\text{N}(\text{NO}_x)$). The importance of this fractionation shift is modulated by the factor $(1 - f_{\text{NO}_2})$. When
 317 NO is almost entirely converted into NO_2 ($f_{\text{NO}_2} \approx 1$), N fractionation effects can be neglected (i.e., $\delta^{15}\text{N}(\text{NO}_2) \approx \delta^{15}\text{N}(\text{NO}_x)$).

318 From samples collected at Jülich, Germany, Freyer et al. (1993), observed for the first time the linear relation described
 319 by Eq. (8), and set the theoretical framework to interpret $\delta^{15}\text{N}$ variabilities in atmospheric NO_2 . They showed that the
 320 observed seasonal variation of $\delta^{15}\text{N}(\text{NO}_2)$ was driven by N fractionation effects (represented in the F_{N} factor) caused by
 321 photochemistry and isotopic equilibrium. Based on this work, and that of Li et al. (2020), Albertin et al. (2021) derived an
 322 expression of F_{N} during the day assuming that the NO- NO_2 system is in isotopic equilibrium (steady-state):

$$(F_{\text{N}})_{\text{day}} \approx \frac{\alpha_{\text{LCIE}}^* A^*_{\text{day}} + (\alpha_{\text{EIE}(\text{NO}_2/\text{NO})} - 1)}{A^*_{\text{day}} + 1} \quad (10)$$

323 with $\alpha_{\text{LCIE}}^* = \alpha_{\text{KIE}(\text{NO}+\text{O}_3)} - \alpha_{\text{PHIFE}}$

324 and $A^*_{\text{day}} = \frac{J_{\text{NO}_2}}{k_{\text{NO}+\text{NO}_2}[\text{NO}]}$

325 where α_{LCIE}^* is the fractionation factor of combined KIE and PHIFE (LCIE is for Leighton Cycle Isotope Effect), and
 326 $\alpha_{\text{EIE}(\text{NO}_2/\text{NO})}$ is the EIE fractionation factor between NO and NO_2 . $\alpha_{\text{EIE}(\text{NO}_2/\text{NO})}$ and $\alpha_{\text{KIE}(\text{NO}+\text{O}_3)}$ are temperature dependent
 327 and can be calculated following the theoretical approach of Walters and Michalski (2015) (Table D1). From calculations
 328 based on the zero point energy of $^{15}\text{NO}_2$ and the absorption cross section of $^{14}\text{NO}_2$, α_{PHIFE} is estimated to vary between
 329 1.0020 and 1.0042 for a range of solar zenith angles between 90° and 0° (Fang et al. 2021). In this study we use a mean
 330 value of α_{PHIFE} at 1.0031. A^*_{day} is defined as the ratio of the NO_2 lifetime with respect to isotopic exchanges over the
 331 daytime NO_2 chemistry lifetime. J_{NO_2} is the NO_2 photolysis rate, $k_{\text{NO}+\text{O}_3}$ is the rate constant of Reaction (R3), and $k_{\text{NO}+\text{NO}_2}$

332 is the rate constant of the isotopic exchange $^{15}\text{NO}_2 + ^{14}\text{NO} \rightarrow ^{14}\text{NO}_2 + ^{15}\text{NO}$. During the day, $\Delta^{15}(\text{NO}_2 - \text{NO}_x)$ varies
 333 according to the environmental conditions. In low- NO_x conditions (e.g., remote and polar regions) $\Delta^{15}(\text{NO}_2 - \text{NO}_x)$ is
 334 predicted to be controlled by LCIE factors ($A^*_{\text{day}} \gg 1$), whereas an EIE-dominated regime ($A^*_{\text{day}} \ll 1$) is expected in
 335 polluted environments (high- NO_x conditions). At night, J_{NO_2} and α_{PHIFE} are null and A^*_{night} is defined as the ratio of NO
 336 lifetime with respect to isotopic exchange with NO_2 to NO chemical lifetime at night $\left(A^*_{\text{night}} = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3]}{k_{\text{NO}+\text{NO}_2}[\text{NO}_2]}\right)$. In this study,
 337 we consider only one particular case with $A^*_{\text{night}} \ll 1$, which means that isotopic exchanges are much faster than NO
 338 oxidation by O_3 . In this scenario, KIE effects are negligible compared to EIE effects and $(F_{\text{N}})_{\text{night}}$ can be expressed as:

$$(F_{\text{N}})_{\text{night}} \approx \frac{(\alpha_{\text{EIE}(\text{NO}_2/\text{NO})} - 1)}{\alpha_{\text{EIE}(\text{NO}_2/\text{NO})}} \quad (11)$$

339 The complete derivation of Eqs. (10) and (11) is given in Albertin et al. (2021).
 340

341 KIE and EIE are also expected during the conversion of NO_2 to NO_3^- . The ^{15}N partitioning associated to isotopic
 342 equilibrium between N_2O_5 and NO_2 (Reaction R7) can be theoretically computed as a function of temperature (Walters and
 343 Michalski, 2015; Table D1). At 298 K, if N isotopic equilibrium is reached, N_2O_5 is predicted to have $\delta^{15}\text{N}$ values 27.6 ‰
 344 higher than NO_2 . Considering that the $\text{NO}_2/\text{NO}_3^-$ isotopic fractionation through the N_2O_5 pathway is solely controlled by
 345 EIE, NO_3^- is therefore expected to be enriched in ^{15}N relative to NO_2 . However, to date, no experimental study has reported
 346 on ^{15}N partitioning between atmospheric NO_2 and NO_3^- , and the fractionation factors are still being debated (Freyer 1991,
 347 Fang et al. 2021). The dominant NO_x to NO_3^- conversion processes considered in this study, along with corresponding $\Delta^{17}\text{O}$
 348 transfer factors and the known ^{15}N enrichment factors at 298 K (determined from both experimental and computational
 349 studies) are illustrated in Figure C1.

350 **3 Results and Discussion**

351 **3.1 Temporal variations of general atmospheric observations**

352 Surface temperatures during SP 1 and SP 2 show similar values, with a marked diurnal cycle (from -2°C to 16°C ; Figure
 353 1). A positive temperature gradient with altitude is observed from late evening to morning. Surface temperature rises around
 354 midday and reaches a maximum at around 15:00 LT, resulting in a negative temperature gradient with altitude. In deep
 355 Alpine valleys, the diurnal variability of surface air temperature is strongly influenced by the temporal evolution of the
 356 boundary layer structure, particularly in winter with the formation of a surface layer inversion (Whiteman, 1982). As
 357 previously observed in Chamonix (Chazette et al., 2005), the nocturnal surface layer inversion regularly thickens during the
 358 night of the sampling periods. After sunrise, air masses warm up until the nocturnal inversion layer breaks down in the late

359 morning. Observed RH behaviour relatively correlates with the increase in temperature during the day, showing a rapid
360 decrease between mid-morning and early afternoon (from 96 % to 23 % and from 96 % to 30 % for SP 1 and SP 2,
361 respectively; Figure 1).

362

363 During SP 1 and SP 2, the mixing ratios of NO, NO₂ and O₃ exhibit diurnal patterns (Figure 1) typical in Chamonix in
364 late February (Figure S3 in the Supplement), and more generally in urban areas (Mayer, 1999). The highest NO level is
365 observed in the morning, peaking around 10:30 LT (82 nmol mol⁻¹ and 152 nmol mol⁻¹ for SP 1 and SP 2, respectively). The
366 concurrent increase in NO₂ and decrease in O₃ (down to 1 nmol mol⁻¹) can be attributed to Reaction (R3) (NO + O₃). NO_x
367 decreases in the late morning, likely due to a combination of lower emissions, NO₂ oxidation, and dilution effect.
368 Meanwhile, O₃ gradually recovers to ca. 30 nmol mol⁻¹, a typical winter background air level in Europe (Gaudel et al.,
369 2018). Due to local emissions, NO_x increases again from 16:00 LT, resulting in O₃ titration, and subsequently to an increase
370 in NO₂ (up to 40 nmol mol⁻¹). After 18:30 LT, NO remains low until the morning, and NO₂ decreases slowly until midnight,
371 stalls around 10 nmol mol⁻¹, and then rises again at 5:30 LT. After the late afternoon titration, O₃ gently recovers and stay
372 relatively low throughout the night likely due to a titration effect from nocturnal NO emissions which are confined in the
373 surface layer inversion.

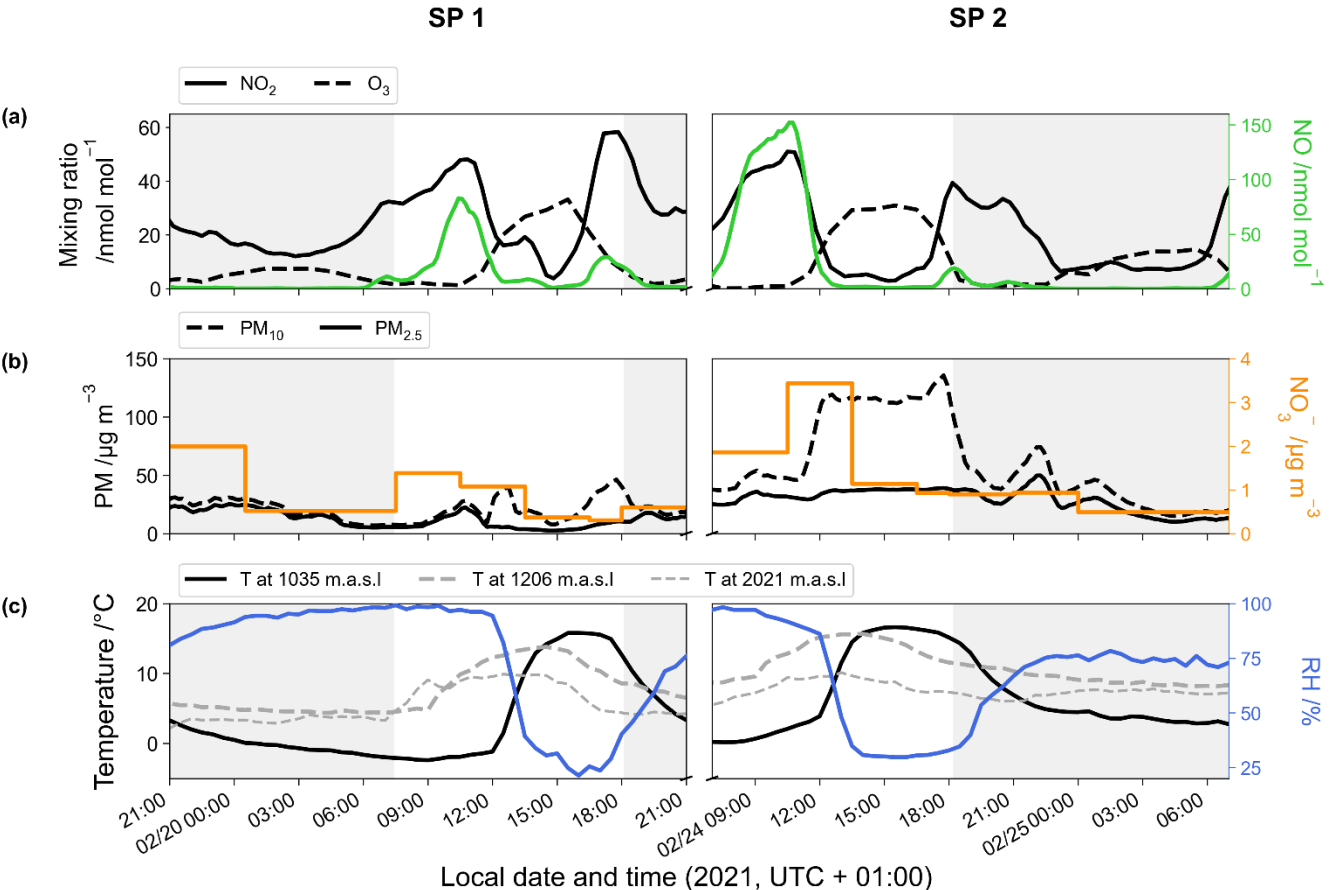
374

375 Both SP 1 and SP 2 show diurnal variations in PM mass concentrations (Figure 1), with morning and evening peaks
376 related to local emissions from traffic and home heating (Aymoz et al., 2007). PM₁₀ concentrations display an additional
377 increase at midday following the breakdown of the temperature inversion. At 12:30 LT, while PM₁₀ concentrations increase
378 moderately during SP 1 to reached 38.6 µg m⁻³, a sharp increase to 119 µg m⁻³ is observed during SP 2. Then PM₁₀
379 concentrations decrease during the afternoon of SP 1, but remain high during the afternoon of SP 2 until the surface
380 inversion layer forms. On average, PM₁₀ concentration is three times higher during SP 2 ((59.4 ± 37.6) µg m⁻³) than during
381 SP 1 ((20.6 ± 10.2) µg m⁻³). The considerable increase in PM₁₀ concentrations between SP 1 and SP 2 is likely to be
382 explained by a Saharan dust episode that started on February 23 (Fig. S3, S4, and S5 in the Supplement). Saharan dust
383 deposition is a well-known phenomenon in the Alps, which is characterised by a sudden increase of coarse particles, mainly
384 composed of alumino-silicates as well as calcium and potassium (Angelisi and Gaudichet, 1991; Delmas, 1994; Di Mauro et
385 al., 2019; Goudie and Middleton, 2001; Greilinger et al., 2018; Schwikowski et al., 1995; Sodemann et al., 2006).

386

387 The NO₃⁻ mass concentration varies from 0.3 µg m⁻³ to 3.4 µg m⁻³, with an average of (0.9 ± 0.6) µg m⁻³ for SP 1 and
388 of (1.2 ± 0.9) µg m⁻³ for SP 2 (Figure 1). During both sampling periods, NO₃⁻ concentration is within the range of previous
389 observations made in Chamonix in winter (Allard, 2018). NO₃⁻ shows a distinctive peak at 3.4 µg m⁻³ during SP 2 between
390 10:30 and 13:30 LT, correlated with the PM₁₀ surge. During transport, dust can undergo heterogeneous uptake and
391 conversion of gases on its surface, leading to the inclusion of secondary species such as NO₃⁻, sulfate, and ammonium

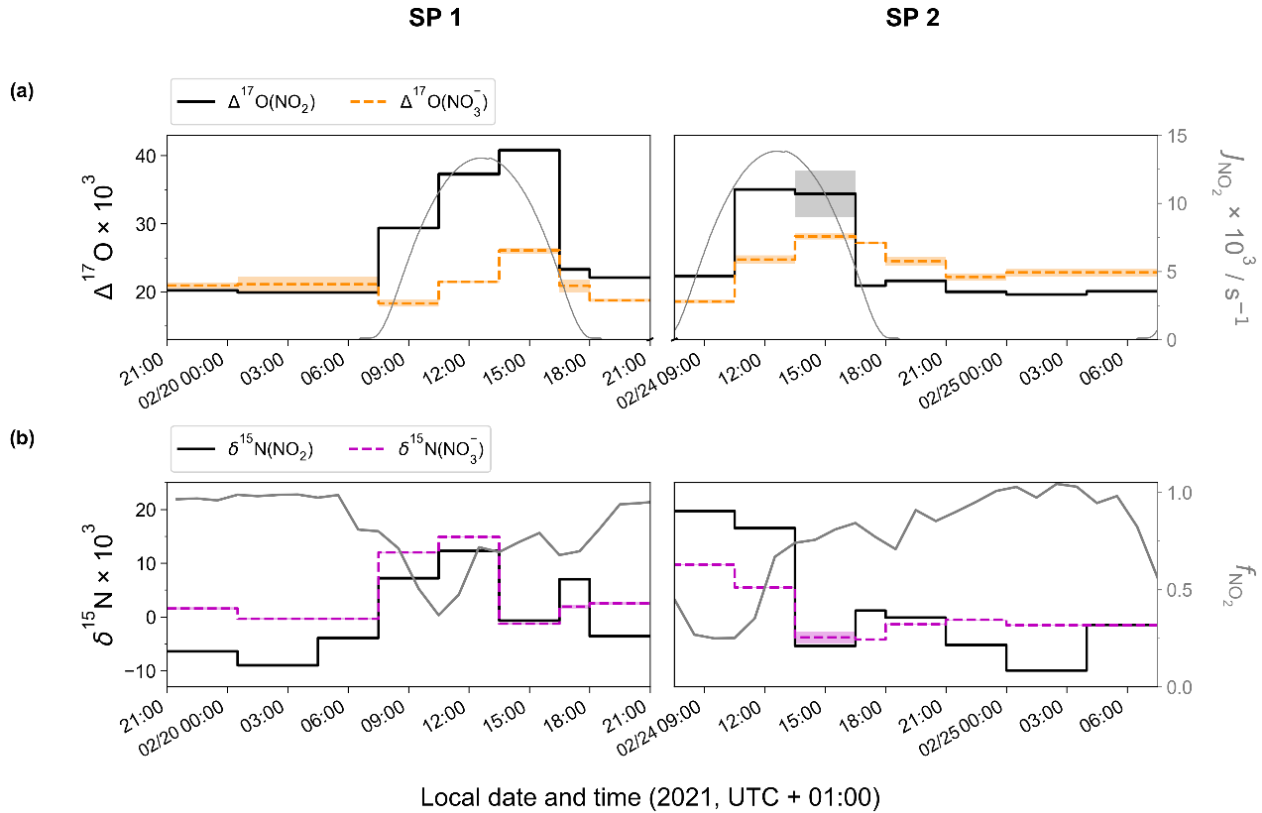
392 (Usher et al., 2003). NO_3^- on dust results mainly from HNO_3 uptake and heterogeneous reactions of N_2O_5 (see Usher et al.,
 393 2003 for a review and references therein). Mineral dust is believed to significantly contribute to NO_3^- formation and size
 394 distribution, particularly in regions close to dust emission sources (Karydis et al., 2016). However, the origin of NO_3^- during
 395 SP 2 at our site remains unclear and could be attributed to the advection of both nitrated-dust particles formed through
 396 heterogeneous processes during transport and anthropogenic fine particles (Aymoz et al. 2004).
 397



398
 399 **Figure 1.** Temporal evolution of the 1-hour rolling mean of (a) NO_2 (black line), O_3 (dashed line), and NO (green line) mixing ratios, (b)
 400 PM (dashed line for PM_{10} and solid line for $\text{PM}_{2.5}$) and NO_3^- (orange horizontal line) mass concentrations, and (c) temperature at the
 401 surface (black line), at 1206 m (dashed grey line), and at 2021 m (light dashed grey line) and surface relative humidity (blue line). Data
 402 were collected during the two sampling periods (SP 1 and SP 2) in Chamonix. Grey backdrop shaded areas represent the nighttime (sunset
 403 to sunrise).

404
 405 Figure 2 shows the temporal evolution of measured $\Delta^{17}\text{O}$ and $\delta^{15}\text{N}$ of NO_2 and NO_3^- in Chamonix during the two
 406 sampling periods (SP 1 and SP 2). All isotopic data used in this study are reported in Table S1 and Table S2 in the
 407 Supplement. In the following analysis, first we describe $\Delta^{17}\text{O}(\text{NO}_2)$ measurements and gives interpretation in light of NO_x

chemistry cycling (Section 3.2). Then, an analysis of $\Delta^{17}\text{O}(\text{NO}_3^-)$ measurements is proposed aggregating daytime and nighttime periods and comparing them with $\Delta^{17}\text{O}(\text{NO}_3^-)$ estimates derived from $\Delta^{17}\text{O}(\text{NO}_2)$ measurements and $\Delta^{17}\text{O}$ mass balance for major chemical processes (Section 3.3). In light of these results and atmospheric conditions during SP 1 and SP 2, sub-daily $\Delta^{17}\text{O}(\text{NO}_3^-)$ dynamics are investigated. In Section 3.4, N fractionation effects in the NO_x cycle are quantified, and the dominant NO_x emission source is identified. The dynamics of $\delta^{15}\text{N}(\text{NO}_3^-)$ is also described and its use to trace NO_x emission sources and oxidation processes is discussed.



Local date and time (2021, UTC + 01:00)

Figure 2. Temporal evolution of observed (a) $\Delta^{17}\text{O}$ and (b) $\delta^{15}\text{N}$ of atmospheric NO_2 (top and bottom solid black horizontal line) and NO_3^- (top orange and bottom magenta horizontal dashed line) in Chamonix (length of horizontal line = sampling period, shaded area = overall analytical error). The NO_2 photolysis rate (J_{NO_2} , top grey line) is from CiTTyCAT boxmodel output. f_{NO_2} ($f_{\text{NO}_2} = [\text{NO}_2] / ([\text{NO}_2] + [\text{NO}])$; bottom grey line) is calculated from hourly mean mixing ratio of NO and NO_2 .

3.2 $\Delta^{17}\text{O}$ of NO_2 and NO_x diurnal cycling

Over the course of SP 1 and SP 2, $\Delta^{17}\text{O}(\text{NO}_2)$ shows a large diurnal variability (from 19.6 ‰ to 40.8 ‰) with a weighted mean $\pm$ one standard deviation of (25.2 ± 7.1) ‰. $\Delta^{17}\text{O}(\text{NO}_2)$ values during the day (7:30–18:00 LT, (28.5 ± 7.3) ‰) are significantly higher (p -value = 0.002, $n = 16$) than during the night (18:00–7:30 LT, (20.8 ± 1.0) ‰). By day, $\Delta^{17}\text{O}(\text{NO}_2)$ follows a similar increasing trend during SP 1 and SP 2, reaching a respective maximum of 40.8 ‰ between 13:30–16:30 LT

and 35.0 ‰ between 10h30–13h30 LT. For both sampling periods, after sunset, $\Delta^{17}\text{O}(\text{NO}_2)$ stabilises between 21:00 and 7:30 LT at ca. 20 ‰. Using the same sampling methodology in a mid-latitude urban area in spring, Albertin et al. (2021) reported very similar $\Delta^{17}\text{O}(\text{NO}_2)$ values over the course of one day (20.5–39.2 ‰), following a comparable diurnal pattern. As presented above (Section 2.4.1), according to the ISS (isotopic steady state) framework, the variability of $\Delta^{17}\text{O}(\text{NO}_2)$ reflects changes in the relative contributions of Reaction (R3) ($\text{NO} + \text{O}_3$) and Reaction (R4) ($\text{NO} + \text{RO}_2$) to the overall production of NO_2 . At our site, $\Delta^{17}\text{O}(\text{NO}_2)$ drops rapidly during the 16:30–18:00 LT interval to 23.3 ‰ and 20.9 ‰ during SP 1 and SP 2, respectively. Since, the isotope recycling rate in the NO_x - O_3 system is driven at first order by J_{NO_2} (Michalski et al., 2014), due to low solar radiation between 16:30–18:00 LT at our site, such a rapid drop of $\Delta^{17}\text{O}(\text{NO}_2)$ suggests that sampled NO_2 might not be at ISS anymore when the NO_2 photolysis is very slow, notably at the end of the day in winter. Therefore, we only consider that ISS holds between 7:30 and 16:30 LT in order to avoid the questionable end-of-the-day measurements in our ISS-based analysis. At night (i.e., no ISS), the observed drop of $\Delta^{17}\text{O}(\text{NO}_2)$ in the early evening reflects the rapid replacement of NO_2 formed during the day by NO_2 produced during the night via the conversion of freshly emitted NO, in line with Eq. (5). Then, high NO_2 throughout the night, along with relatively low O_3 , supports observations of low $\Delta^{17}\text{O}(\text{NO}_2)$ at night (i.e., ≈ 20 ‰).

Using Eq. (3), we derive from $\Delta^{17}\text{O}(\text{NO}_2)$ observations the relative contribution of Reaction (R3) ($\text{NO} + \text{O}_3$) to Reaction (R4) ($\text{NO} + \text{RO}_2$) in the formation of NO_2 ($T_{\text{NO}+\text{O}_3}$):

$$T_{\text{NO}+\text{O}_3} = \frac{\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)}{\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)} \quad (12)$$

Between 7:30 and 16:30 LT, $T_{\text{NO}+\text{O}_3}$ varies from 0.55 to 1.00 (Table 1), with a mean of 0.88 and 0.75 for SP 1 and SP 2, respectively. The $\text{NO} + \text{O}_3$ pathway is dominant between 13:30 and 16:30 LT, corresponding to the time when O_3 is highest (Figure 1). In contrast, the maximum contribution for the $\text{NO} + \text{RO}_2$ pathway is observed between 7:30 and 10:30 LT, when NO levels are high and rising continuously. Interestingly, previous studies reported a high sensitivity of RO_2 to changes in NO_x , particularly at high NO_x levels (Ren et al., 2006; Stone et al., 2012). Sources of RO_2 in wintertime are mainly driven by the production of OH radicals from HONO photolysis, alkene ozonolysis, and formaldehyde photolysis (Tan et al., 2018). During winter, HONO plays a crucial role in $\text{NO}_x/\text{O}_3/\text{RO}_2$ chemistry, particularly in the morning, as its photolysis can potentially accelerate daytime oxidation processes, leading to increased RO_2 production (Alicke et al., 2003; Aumont et al., 2003). Direct emissions from vehicle exhaust could be significant a source of VOCs and HONO at our site (Brulfert et al., 2005; Gu et al., 2019; Kirchstetter et al., 1996; Kurtenbach et al., 2001; Liu et al., 2023). Heterogeneous processes on ground surfaces and aerosols can also contribute to HONO formation (Aumont et al., 2003). In addition, snowpack releases may also

be a potential source of HONO (Grannas et al., 2007), as detected in Paris after a snow event, which could significantly impact the urban OH budget (Michoud et al. 2015).

Following the approach of Albertin et al. (2021), combining Eqs. (3) and (4) allows to derive RO₂ mixing ratio from observed $\Delta^{17}\text{O}(\text{NO}_2)$ and O₃ mixing ratio following:

$$[\text{RO}_2] = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3]}{k_{\text{NO}+\text{RO}_2}} \left(\frac{\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)}{\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)} - 1 \right) \quad (13)$$

Between 7:30 and 16:30 LT, we estimate an average RO₂ mixing ratio at our site of (0.88 ± 0.88) pmol mol⁻¹ and (4.92 ± 5.16) pmol mol⁻¹ during SP 1 and SP 2, respectively (Table 1). Studies conducted in urban winter environments reported RO₂ measurements at a few pmol mol⁻¹ (Ren et al., 2006; Emmerson et al., 2005; Tan et al., 2018; Kanaya et al., 2007), in good agreement with our estimations. Similarly, RO₂ mixing ratios derived by Albertin et al. (2021) from observed $\Delta^{17}\text{O}(\text{NO}_2)$ in spring (mean of (13.8 ± 11.2) pmol mol⁻¹) were also found to be in line with studies conducted in the same season. The fact that our isotopic inference of RO₂ mixing ratios carried out in two different seasons (winter and spring) are both comparable to direct in situ RO₂ measurements confirm the sensitivity of our method in probing the NO_x/O₃/RO₂ chemical dynamics. We think that our method may be very valuable in deciphering oxidation processes of N_r species, down to sub-daily temporal scales. Nonetheless, we recognise that without concurrent in situ measurements of RO₂ and $\Delta^{17}\text{O}(\text{NO}_2)$, it is not possible to validate unambiguously the quantitative estimation of RO₂ levels with our method.

Sampling interval (start - end)	$T_{\text{NO}+\text{O}_3}$	RO ₂ /pmol mol ⁻¹
SP 1		
20/02 07:30 - 20/02 10:30	0.72 ± 0.01	0.86 ± 0.75
20/02 10:30 - 20/02 13:30	0.91 ± 0.01	1.77 ± 0.36
20/02 13:30 - 20/02 16:30	1.00 ± 0.01	0.00 ± 0.91
<i>Mean</i>	<i>0.88</i>	<i>0.88</i>
<i>Std dev.</i>	<i>0.14</i>	<i>0.88</i>
SP 2		
24/02 07:30 - 24/02 10:30	0.55 ± 0.01	0.58 ± 1.67
24/02 10:30 - 24/02 13:30	0.86 ± 0.01	3.56 ± 0.50
24/02 13:30 - 24/02 16:30	0.84 ± 0.08	10.63 ± 6.75
<i>Mean</i>	<i>0.75</i>	<i>4.92</i>
<i>Std dev.</i>	<i>0.18</i>	<i>5.16</i>

Table 1. $T_{\text{NO}+\text{O}_3}$ and RO₂ mixing ratio (mean value $\pm$ overall uncertainty) derived from the isotopic measurements.

At this point, it is important to recall that the choice of the $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ in Eq. (12) is of a particular importance for quantifying $T_{\text{NO}+\text{O}_3}$ (as for RO_2). In the literature, $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ varies between 35 ‰ and 41 ‰ (Michalski et al., 2003; Savarino et al., 2016; Vicars et al., 2012; Zhang et al., 2022b; Li et al., 2022a). This relatively wide range of values is partly a result of some confusion in defining $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$, the $\Delta^{17}\text{O}$ transfer from O_3 to NO . Indeed, the term $\Delta^{17}\text{O}(\text{O}_3^*)$ is sometimes erroneously used as the transfer function of $\Delta^{17}\text{O}$ from O_3 to NO_2 during Reaction (R3). $\Delta^{17}\text{O}(\text{O}_3^*)$ is actually defined as $\Delta^{17}\text{O}$ of O_3 terminal atoms and is also named $\Delta^{17}\text{O}(\text{O}_3)_{\text{term}}$ in the literature. As the $\Delta^{17}\text{O}$ in O_3 is borne by its terminal atoms, $\Delta^{17}\text{O}(\text{O}_3^*) = 1.5 \times \Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$. However, $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ can be equal to $1.5 \times \Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$ if only terminal atoms of O_3 reacts with NO . But laboratory experiments by Savarino et al. (2008) did show that O_3 does react with NO not solely with its terminal atoms but also, to a small but significant extent, with its central atom (probability of $(8 \pm 5)\%$ for the abstraction of central atoms during the reaction $\text{NO} + \text{O}_3$). Consequently, $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ is slightly lower than $1.5 \times \Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$ and the $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ expression determined by Savarino et al. (2008) should be used: $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) = 1.18 \pm 0.07 \times \Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}} + (6.6 \pm 1.5)\%$.

Assuming that their maximum measured daytime $\Delta^{17}\text{O}(\text{NO}_2)$ reflects the conversion of NO to NO_2 only through Reaction (R3) (i.e., $T_{\text{NO}+\text{O}_3} = 1$), Albertin et al. (2021) derived a $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ value of 39.2 ‰ from Eq. (3). Given the respective analytical uncertainties (around $\pm 1\%$), their value is in very good agreement with the maximum daytime value of 40.8 ‰ we observed in Chamonix. Similarly to Albertin et al. (2021), assuming that the highest daytime $\Delta^{17}\text{O}(\text{NO}_2)$ value at our site corresponds to $T_{\text{NO}+\text{O}_3} \approx 1$ leads to $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) = 40.8\%$. Using the experimental $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ transfer function determined by Savarino et al. (2008), we estimate a bulk ^{17}O -excess of O_3 ($\Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$) at $(29.0 \pm 2.2)\%$. This value is consistent with the range of direct $\Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$ measurements at mid-latitudes (mean of $(26.2 \pm 1.3)\%$; Vicars and Savarino, 2014), although falling at the upper end of the range. Interestingly, Vicars and Savarino (2014) reported a significant peak in $\Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$ during February-March in Grenoble, France (located 120 km southwest of Chamonix), based on year-round measurements with $\Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$ values 2–3 ‰ higher than the annual mean of 26.2 ‰. Although the cause of increased values during this period is unknown, our derived $\Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}}$ matches remarkably well the February-March measurements reported by Vicars and Savarino (2014).

3.3 Interpretation of $\Delta^{17}\text{O}$ in atmospheric nitrate

Over the two sampling periods, $\Delta^{17}\text{O}(\text{NO}_3^-)$ varies significantly (from 18.3 ‰ to 28.1 ‰), with a weighted mean of $(22.5 \pm 3.1)\%$. While $\Delta^{17}\text{O}(\text{NO}_2)$ values are relatively similar during the two sampling periods, $\Delta^{17}\text{O}(\text{NO}_3^-)$ values are systematically higher during SP 2 than during SP 1, except during the 7:30–10:00 LT interval. $\Delta^{17}\text{O}(\text{NO}_3^-)$ in Chamonix is in the same range of most previous observations in urban environments (9–44 ‰; e.g., Kim et al., 2023; Wang et al., 2023; Fan et al., 2023; Zhang et al., 2022b; Lim et al., 2022; Li et al., 2022b), but lower than most values measured during the cold

season which are typically >25 ‰. Unlike $\Delta^{17}\text{O}(\text{NO}_2)$, daytime and nighttime $\Delta^{17}\text{O}(\text{NO}_3^-)$ values at our site are not significantly different (p -value > 0.05, $n = 14$). Similarly, from 12 h resolved sampling in winter Beijing, He et al. (2018) found no significant difference between daytime and nocturnal $\Delta^{17}\text{O}(\text{NO}_3^-)$ and suggest that each sample reflects NO_3^- produced during both the day and night. From high-time-resolved (3 h) aerosol sampling in winter Beijing, Zhang et al. (2022b) reported $\Delta^{17}\text{O}(\text{NO}_3^-)$ values between 23.4 ‰ to 39.3 ‰, with higher values observed at night ($(31.0 \pm 2.6) \text{ ‰}$) than during the day ($(29.3 \pm 3.0) \text{ ‰}$). This diurnal behaviour of $\Delta^{17}\text{O}(\text{NO}_3^-)$ was attributed to the changes in the branching ratio of nocturnal and photochemical reactions on NO_3^- formation. In Chamonix, the range of $\Delta^{17}\text{O}(\text{NO}_3^-)$ values are very different from Zhang et al., (2022b) observations, with consistently lower values and a distinct diurnal tendency. However, in the cases of $\Delta^{17}\text{O}(\text{NO}_3^-)$ measurements at sub-daily temporal scale, the atmospheric lifetime of NO_x and NO_3^- is critical for comparing $\Delta^{17}\text{O}(\text{NO}_3^-)$ records from one site to another. Pollutant levels and atmospheric conditions between Chamonix and Beijing are very different, notably in winter when Asian urban areas can experience severe haze pollution episodes with NO_3^- mass concentration exceeding $70 \mu\text{g m}^{-3}$, which is over 10 times higher than in Chamonix (Lim et al., 2022; He et al., 2018; Zhang et al., 2022b). In such conditions, PM can reach several hundreds of $\mu\text{g m}^{-3}$ for several days, which can significantly impact atmospheric processes involved in the formation of secondary species. Aside from the intrusion of Saharan dust during SP 2, the pollutant level in Chamonix is indicative of a moderately polluted region, with significant diurnal variations.

3.3.1 Steady state evaluation of $\Delta^{17}\text{O}(\text{NO}_3^-)$

To investigate the factors influencing the variability of $\Delta^{17}\text{O}(\text{NO}_3^-)$ at our site, one compare observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ with estimated values of $\Delta^{17}\text{O}(\text{NO}_3^-)$ derived from $\Delta^{17}\text{O}$ mass balance and observed $\Delta^{17}\text{O}(\text{NO}_2)$, assuming the OH and N_2O_5 pathways dominate the formation of NO_3^- at our site. Therefore, calculated $\Delta^{17}\text{O}(\text{NO}_3^-)$ reflect the theoretical $\Delta^{17}\text{O}$ transfer during the oxidation of NO_2 to NO_3^- at our site through the dominant chemical process during the day (i.e. OH pathway) and at night (i.e. N_2O_5 pathway).

As presented, during the day, we consider that the conversion of NO_2 into NO_3^- is predominantly influenced by Reaction (R5) (OH pathway). Hence, the theoretical corresponding ^{17}O -excess transfer to NO_3^- is estimated using Eq. (6) and observed $\Delta^{17}\text{O}(\text{NO}_2)$ between 7:30 and 18:00 LT ($n = 3$ per sampling period). Then, in order to estimate a daytime average value of $\Delta^{17}\text{O}(\text{NO}_3^-)$ which is representative of the potential for the formation of NO_3^- from surface NO_2 by the OH pathway, each calculated $\Delta^{17}\text{O}(\text{NO}_3^-)$ is weighted by the product $[\text{NO}_2] \times J_{\text{NO}_2}$ (the diurnal variability of the OH mixing ratio is assumed to follow the diurnal J_{NO_2} variation; Liu et al., 2021). Finally, an overall mean daytime $\Delta^{17}\text{O}(\text{NO}_3^-)$ for SP 1 and SP 2 is estimated by taking the sum of the weighted calculated values ($= \Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$). The same approach is used during the night, assuming that the conversion of NO_2 into NO_3^- is dominated by Reactions (R6)–(R8) (N_2O_5 pathway). Eq. (7) and

533 observed $\Delta^{17}\text{O}(\text{NO}_2)$ between 18:00 and 7:30 LT ($n = 3$ per sampling day) are used to estimate $\Delta^{17}\text{O}(\text{NO}_3^-)$. Each calculated
 534 $\Delta^{17}\text{O}(\text{NO}_3^-)$ is weighted by the product $[\text{NO}_2] \times [\text{O}_3]$ (i.e., NO_3^- production rate) and summed to estimate a mean nighttime
 535 $\Delta^{17}\text{O}(\text{NO}_3^-)$ for SP 1 and SP 2. The ^{17}O -excess transferred from O_3 to NO_2 during Reaction (R6) ($\Delta^{17}\text{O}_{\text{NO}_2+\text{O}_3}(\text{NO}_3^-)$) is
 536 fixed at 44.7 ‰. This value is set accordingly to the transfer function reported by Berhanu et al. (2012) whereby
 537 $\Delta^{17}\text{O}_{\text{NO}_2+\text{O}_3}(\text{NO}_3^-) = (1.23 \pm 0.19) \times \Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}} + (9.02 \pm 0.99)$ and $\Delta^{17}\text{O}(\text{O}_3)_{\text{bulk}} = 29.0$ ‰ (see Section 2.4.1). We
 538 compare hereafter $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ with the weighted day and night averages of observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ at our site. During the
 539 day, $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ is compared with $\Delta^{17}\text{O}(\text{NO}_3^-)$ observations averaged between 7:30 and 18:30 LT ($n = 3$). At night,
 540 $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ is compared with $\Delta^{17}\text{O}(\text{NO}_3^-)$ observations averaged between 18:30 to 7:30 LT.

541

542 At night during SP 1, observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ and $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ are in good agreement ($\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-) - \Delta^{17}\text{O}(\text{NO}_3^-) =$
 543 $\Delta^{17}(\text{NO}_3^-_{\text{calc}} - \text{NO}_3^-_{\text{obs}}) = 0.9$ ‰), suggesting a local and rapid (< 12 h) conversion of NO_2 into NO_3^- via the N_2O_5 pathway.
 544 During the day, observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ is 0.5 ‰ higher than $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$, also suggesting that NO_3^- is formed locally
 545 during the day for oxidation of surface NO_2 through the OH pathway. Small differences between observed and calculated
 546 $\Delta^{17}\text{O}$ of NO_3^- during the day/night could be explained by the presence of NO_3^- residues formed during the previous
 547 night/day, which are not considered in the calculations since they do not account for NO_3^- lifetime. In contrast to SP 1,
 548 $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ during SP 2 is significantly lower than the mean observed $\Delta^{17}\text{O}(\text{NO}_3^-)$, particularly during the day with a
 549 $\Delta^{17}(\text{NO}_3^-_{\text{calc}} - \text{NO}_3^-_{\text{obs}})$ of -6.4 ‰. The significant gap between observed and calculated $\Delta^{17}\text{O}(\text{NO}_3^-)$ suggests a different
 550 origin and/or formation process of NO_3^- during SP 2 compared to SP 1. Although less important than during the day,
 551 $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ values for SP 2 at night is lower by 2.2 ‰ to the observed value. This small shift can be explained by
 552 residuals of enriched daytime NO_3^- . It is important to point out that, although the NO_2 sample collected on Feb 24 between
 553 13:30 and 16:30 LT presents an important blank (ca. 14 ‰), ambient NO_2 is low during the sampling period (mean of $(4.5 \pm$
 554 $1.8)$ nmol mol $^{-1}$). Therefore, as each $\Delta^{17}\text{O}$ value used to estimate $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$ is weighted by the mean ambient NO_2
 555 mixing ratio over the sampling period, the incertitude related to this blank has little influence on the daily average of
 556 $\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$. Given the low $\Delta^{17}(\text{NO}_3^-_{\text{calc}} - \text{NO}_3^-_{\text{obs}})$ during SP 1, observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ can be explained by the local and
 557 rapid (< 12 h) oxidation of NO_2 , dominated by the OH and N_2O_5 pathway during the day and night, respectively. However,
 558 in contrast to SP 1, the ^{17}O -excess measured in NO_3^- during the day of SP 2 cannot be fully constrained by the oxidation of
 559 surface NO_2 through the OH pathway, suggesting that the formation mechanisms of NO_3^- are different between SP 1 and SP
 560 2 and/or the presence of NO_3^- not formed locally during SP 2. Below we examine the changes in the sub-daily dynamics of
 561 $\Delta^{17}\text{O}(\text{NO}_3^-)$ between SP 1 and SP 2 in light of atmospheric observations.

562

		$\Delta^{17}\text{O}(\text{NO}_2) / \text{‰}$	$\Delta^{17}\text{O}(\text{NO}_3^-) / \text{‰}$	$\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-) / \text{‰}$	$\Delta^{17}(\text{NO}_3^-_{\text{calc}} - \text{NO}_3^-_{\text{obs}})$
Daytime	SP 1	30.0 ± 7.3	23.0 ± 3.1	22.5 ± 4.6	-0.5
(7:30-18:00)	SP 2	26.1 ± 6.9	23.9 ± 3.8	17.5 ± 4.6	-6.4
Nighttime	SP 1	21.2 ± 1.1	20.5 ± 1.1	21.4 ± 0.7	0.9
(18:00-7:30)	SP 2	20.8 ± 1.0	23.2 ± 1.0	21.0 ± 0.6	-2.2

Table 2. Mean observed $\Delta^{17}\text{O}$ data of NO_2 ($\Delta^{17}\text{O}(\text{NO}_2)$) and NO_3^- ($\Delta^{17}\text{O}(\text{NO}_3^-)$) in Chamonix, and mean calculated $\Delta^{17}\text{O}$ of NO_3^- ($\Delta^{17}\text{O}_{\text{calc}}(\text{NO}_3^-)$) using Eqs. (6) and (7) at day and night, respectively, constrained with observed $\Delta^{17}\text{O}(\text{NO}_2)$. Day and night calculated values were weighted by $[\text{NO}_2] \times J_{\text{NO}_2}$ and $[\text{NO}_2] \times [\text{O}_3]$, respectively

3.3.2 $\Delta^{17}(\text{NO}_3^-)$ sub-daily dynamics

Between 7:30–10:30 LT, $\Delta^{17}\text{O}(\text{NO}_3^-)$ is very similar during SP 1 (18.3 ‰) and SP 2 (18.6 ‰). Nonetheless, on the following sampling time step (i.e., between 10:30–13:30 LT), $\Delta^{17}\text{O}(\text{NO}_3^-)$ is significantly different between SP 1 (21.5 ‰) and SP 2 (24.7 ‰). Intriguingly, between 7:30–13:30 LT, $\Delta^{17}\text{O}(\text{NO}_2)$ during SP 2 (26.9 ‰) is lower compared to SP 1 (32.8 ‰). As a consequence, if one consider that NO_3^- is formed from the oxidation of local NO_2 through identical pathways during SP 1 and SP 2, observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ should be lower during SP 2 than during SP 1. Therefore, the more pronounced increase of observed $\Delta^{17}\text{O}(\text{NO}_3^-)$ during the 10:30–12:30 LT interval of SP 2 suggests a different origin and/or formation channel of NO_3^- , as mentioned previously. One more piece of evidence is that, during this period of time, PM_{10} and NO_3^- levels increase significantly during SP 2, alongside the disruption of the inversion layer (depicted in Figure 1). It can be inferred that this rise in PM_{10} is mostly due to the presence of Saharan dust. The simultaneous increase of NO_3^- and of $\Delta^{17}\text{O}(\text{NO}_3^-)$ corroborates the hypothesis that this NO_3^- was not formed from the oxidation of ambient NO_2 . Furthermore, such an increase in $\Delta^{17}\text{O}(\text{NO}_3^-)$ can only be supported by the oxidation of NO_2 through the N_2O_5 pathway, which is not expected to be important during the day due to the rapid photolysis of NO_3 and its titration by NO (Brown and Stutz, 2012).

Interestingly, aerosol samplings conducted at various heights (8 m, 120 m, and 260 m above ground level) in Beijing, China, revealed a positive vertical gradient of $\Delta^{17}\text{O}(\text{NO}_3^-)$ in winter, from on average 29 ‰ to 33 ‰ (Fan et al., 2022). In summer, the $\Delta^{17}\text{O}(\text{NO}_3^-)$ values at the three altitudes were very similar. This increase of $\Delta^{17}\text{O}(\text{NO}_3^-)$ with altitude in winter was believed to result from a stratification of NO_2 to NO_3^- oxidation processes due to low vertical mixing and elevated surface NO_x emissions. However, the authors did not consider the potential variability of $\Delta^{17}\text{O}(\text{NO}_2)$ with altitude, which can be substantial in urban areas at night as low $\Delta^{17}\text{O}(\text{NO}_2)$ results from surface NO oxidation. For our study, we propose an alternative interpretation of the vertical variability of $\Delta^{17}\text{O}(\text{NO}_3^-)$, where $\Delta^{17}\text{O}(\text{NO}_2)$ is considered as the main driver. During the formation of the nocturnal boundary layer, NO_2 formed during the day can be trapped above the surface layer in the

nocturnal residual layer (NRL). This NO_2 has a high $\Delta^{17}\text{O}$ because it was formed during the previous daytime hours under the ISS framework (Eq. (3)). Throughout the night, this highly enriched NO_2 (ca. 37 ‰ which is the average of the maximum $\Delta^{17}\text{O}(\text{NO}_2)$ during SP 1 and SP 2) can be converted to NO_3^- via the N_2O_5 pathway, hence leading to a substantial $\Delta^{17}\text{O}$ transfer to NO_3^- at around 32 ‰, which is in the range of $\Delta^{17}\text{O}(\text{NO}_3^-)$ observed by Fan et al. (2022) in winter. In the meantime, NO emitted at the surface during the night can be converted to NO_2 by O_3 , with a $\Delta^{17}\text{O}$ transfer of ca. 20 ‰ (Eq. (5) with $x = 0$). This low enriched NO_2 can be further oxidised to NO_3^- by the N_2O_5 pathway which results in a $\Delta^{17}\text{O}$ transfer at around 21 ‰. This NO_2 with a low $\Delta^{17}\text{O}$ is very likely to be formed only at the surface during the night in areas experiencing important NO_x emissions (Michalski et al., 2014). Furthermore, surface NO_2 with low $\Delta^{17}\text{O}$ is not expected to be transported aloft as it is formed in the surface inversion layer during the night. Therefore, NO_3^- formed in the NRL during winter nights may be more enriched than the NO_3^- formed concurrently at the surface, regardless of the NO_2 oxidation process involved. When the inversion layer breaks during the following day, the NO_3^- that was formed in the NRL during the night is mixed with the NO_3^- formed at the surface, resulting in an increase in the overall surface $\Delta^{17}\text{O}$. In this scenario, the presence of the Saharan dust during SP 2 may have increased the NO_3^- loading aloft by promoting heterogeneous processes on aerosol surfaces in the vicinity of Chamonix. Hence, NO_2 stratification at night could explain the observed increase in $\Delta^{17}\text{O}$ of NO_3^- at the surface following the collapse of the nocturnal inversion layer. However, we cannot determine whether the enriched NO_3^- were formed in the vicinity of Chamonix and/or transported to our site by Saharan dust.

Although the exact nature of the high ^{17}O -excess measured in NO_3^- during SP 2 remains unclear, boundary layer dynamics is thought to play a significant role in the variability of $\Delta^{17}\text{O}(\text{NO}_3^-)$ at the surface due to the stratification of NO_2 . Therefore, a wider consideration of such factors should be explored to avoid possible over-interpretation of $\Delta^{17}\text{O}(\text{NO}_3^-)$ variabilities at the surface, especially in urban areas experiencing significant boundary layer dynamics in winter and high surface emissions of NO_x at night. Measuring $\Delta^{17}\text{O}(\text{NO}_2)$ at various altitudes could provide better insights on the vertical dynamics of $\Delta^{17}\text{O}(\text{NO}_3^-)$, and subsequently quantitative information on NO_3^- production processes.

3.4 Nitrogen isotopic compositions

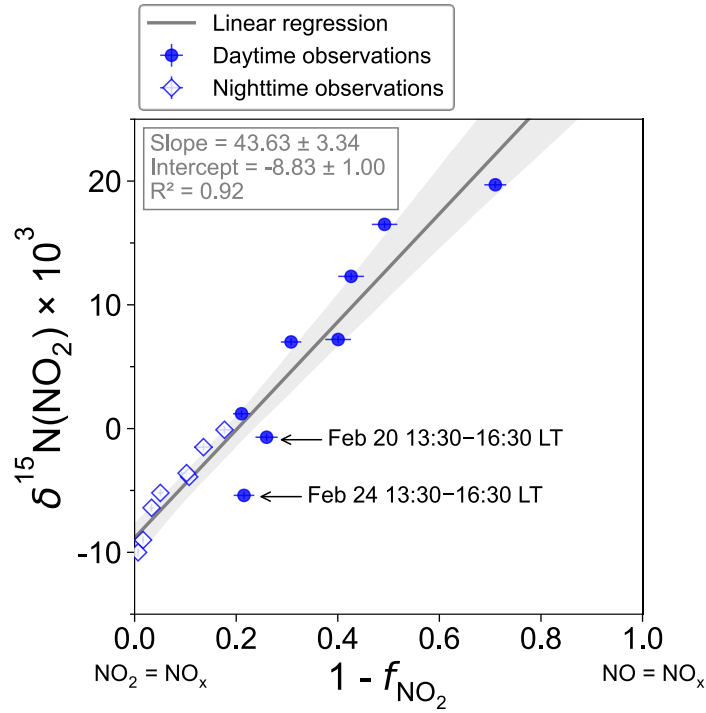
3.4.1 N fractionation effects in the NO_x cycle

Over the two sampling periods, $\delta^{15}\text{N}(\text{NO}_2)$ shows substantial diurnal variability (from -10.0 ‰ to 19.7 ‰, $n = 16$) with a weighted mean of (4.0 ± 9.1) ‰. In contrast, Albertin et al. (2021) reported a weak diurnal fluctuation of $\delta^{15}\text{N}(\text{NO}_2)$ in spring in Grenoble, in a narrow range from about -12 ‰ to -10 ‰. In summer in an urban/suburban location, Walters et al. (2018) also observed a wide range of $\delta^{15}\text{N}(\text{NO}_2)$ values, however, unlike our study, these are almost consistently negative

620 (from -31.4 ‰ to 0.4 ‰) with an overall mean at $(-11.4 \pm 6.9)\text{ ‰}$. As shown in Eq. (8), fluctuations in $\delta^{15}\text{N}(\text{NO}_2)$ reflect
 621 changes in NO_x emission sources and/or N fractionation effects, these latter being weighted by $1 - f_{\text{NO}_2}$ i.e., the more NO_x is
 622 under the form of NO, the greater the N fractionation effects (see Section 2.4). Hence, in the previous works of Albertin et
 623 al. (2021) and Walters et al. (2018), due to high f_{NO_2} (> 0.7), isotope effects were small ($< 2.7\text{ ‰}$) and $\delta^{15}\text{N}(\text{NO}_2)$ was mostly
 624 driven by changing contribution of NO_x emission sources. At our site, f_{NO_2} shows a wider range, from 0.3 to 1.0, suggesting
 625 significant N isotopic fractionation effects, with minimum and maximum contributions corresponding to the highest and
 626 lowest observed value of $\delta^{15}\text{N}(\text{NO}_2)$, respectively. This pronounced seasonal behaviour of N isotope fractionation effects
 627 within the NO_x cycle has previously been outlined in the seminal study of Freyer et al. (1993). Overall, compared with
 628 summer, lower f_{NO_2} during winter months due to lower O_3 concentrations and higher NO_x emissions favour EIE between NO
 629 and NO_2 , which also has a higher fractionation factor due to the lower temperatures (see Appendix D: Equilibrium N
 630 fractionation factors). Besides, this seasonal fluctuation of f_{NO_2} can be expected to be observed on smaller time scales,
 631 typically on the diurnal scale in urban areas where NO is generally fully oxidised into NO_2 at night due to a lower NO_x
 632 emission rate resulting in higher f_{NO_2} at night than during the day, as observed at our sampling (Figure 2).

633
 634 Figure 3 shows the linear dependence of $\delta^{15}\text{N}(\text{NO}_2)$ on $(1 - f_{\text{NO}_2})$ over the two sampling periods, indicating the
 635 significant influence of atmospheric processes that alter the N isotopic distribution during the conversion of NO_x into NO_2 .
 636 The linear regression gives a slope and an intercept of about $(43.6 \pm 3.3)\text{ ‰}$ and $(-8.8 \pm 1.0)\text{ ‰}$, respectively. According to
 637 Eqs. (10) and (11), the linearity between daytime (07:30–18:00 LT) and nighttime (18:00–07:30 LT) values suggests that
 638 EIE dominates the N fractionation processes between NO_x and NO_2 . The influence of LCIE during the day could explain the
 639 greater variability around the linear fit in the daytime observations.

640



641
642

643 **Figure 3.** Correlation plot of $\delta^{15}\text{N}$ of atmospheric NO_2 vs. $(1 - f_{\text{NO}_2})$ from observations in Chamonix in February 2021. f_{NO_2} are averaged
644 over the collection period of each NO_2 sample. The grey shade is the 95 % confidence interval. White diamonds and blue dots represent
645 the nighttime (18:00–07:30 LT) and daytime (07:30–18:00 LT) observations, respectively. The linear regression is plotted over the
646 nighttime and daytime observations.

647 The relative importance of EIE and LCIE in the N fractionation between emitted NO_x and NO_2 is assessed by
648 calculating the A^* factor during the day (7:30–18:00 LT; A^*_{day}) and night (18:00–7:30 LT; A^*_{night}) (Table 3). A^*_{day} and A^*_{night}
649 expressions are given in Section 2.4.2. Overall, during NO_2 sampling intervals, the A^* values are small (mean $\pm$ one standard
650 deviation: 0.21 ± 0.51) and reflect an EIE-dominated regime with high NO_x (Li et al., 2020). It is interesting to note that the
651 highest A^* values are observed between 13:30 and 16:30 LT, and correspond to the two data points in Figure 3 that lie
652 outside the 95 % confidence interval of the regression line. These results suggest that EIE is the dominant N fractionation
653 processes between NO_x and NO_2 during both day and night ($A^* < 0.46$), with the exception of mid-afternoon when LCIE
654 competes with EIE ($A^* > 0.46$).

655

656 To quantify the overall N fractionation effect (F_N) between NO_x and NO_2 , we dissociate the two samples collected
657 between 13:30 and 16:30 LT into a different group (Group #1 = GP 1) from the other samples (Group #2 = GP 2). F_N of GP
658 1 and GP 2 is calculated using Eq. (10) (which combines LCIE and EIE regimes) and Eq. (11) (which considers only the EIE
659 regime), respectively. Calculated F_N are reported in Table 3 and data used for calculations can be found in Section 2.4.2 and

660 in the Supplement. Calculated F_N is significantly different between GP 1 and GP 2, with a mean of 16.4 ‰ and 42.3 ‰,
661 respectively. The close match between the calculated average F_N of GP 2 and the observed F_N ($(43.6 \pm 3.3) \text{ ‰}$; slope of the
662 regression line in Figure 3) provides strong evidence for the reliability of Eq. (11), as well as the expression of $\alpha_{\text{EIE}(\text{NO}_2/\text{NO})}$
663 used therein, to accurately describe the N fractionation between NO_x emissions and NO_2 at our site, and hence, to describe
664 most of the variability of $\delta^{15}\text{N}(\text{NO}_2)$ measurements. This result holds significant importance in confirming the theoretical N
665 isotopic fractionation framework used in prior research studies. It is also important to stress the influence of LCIE effects for
666 GP 1, highlighting the high dependency of $\delta^{15}\text{N}(\text{NO}_2)$ to local environmental conditions. According to the A^* factor, a
667 greater influence of LCIE in mid-afternoon could have contributed to the outlying of the two samples collected between
668 13:30 and 16:30 LT (GP 1). However, as mentioned above, the sample collected on Feb 24 between 13:30 and 16:30 LT has
669 a significant blank. Therefore, it cannot be confirmed with certainty that the reason this sample falls outside the 95 %
670 confidence interval of the regression line is solely due to LCIE. Nevertheless, the overall conclusion that EIE dominates the
671 variability of $\delta^{15}\text{N}(\text{NO}_2)$ at our site is not affected by this uncertainty.

672

673 The $\delta^{15}\text{N}$ shift in NO_2 relative to emitted NO_x ($\Delta^{15}(\text{NO}_2 - \text{NO}_x)$) is calculated for individual NO_2 sample using the mean
674 ambient temperature during each sampling period. The mean atmospheric $\delta^{15}\text{N}$ of NO_x ($\delta^{15}\text{N}(\text{NO}_x)$) is then estimated by
675 subtracting the $\Delta^{15}(\text{NO}_2 - \text{NO}_x)$ value from the observed $\delta^{15}\text{N}(\text{NO}_2)$ value. $\Delta^{15}(\text{NO}_2 - \text{NO}_x)$ and $\delta^{15}\text{N}(\text{NO}_x)$ estimates are
676 reported in Table 3. $\Delta^{15}(\text{NO}_2 - \text{NO}_x)$ varies greatly over the two sampling periods (from 0.7 ‰ to 30.7 ‰) with a mean value
677 of ca. 9 ‰ (mean of GP #1 and GP #2). $\delta^{15}\text{N}(\text{NO}_x)$ show much less variability with an overall mean at $(-7.8 \pm 1.9) \text{ ‰}$ (mean
678 of GP #1 and GP #2), in very good agreement with the value derived from the regression relationship $(-8.8 \text{ ‰}$; intercept of
679 the regression line in Figure 3). Therefore, there appears that there is little variation in NO_x emission sources at our site, and
680 the wide variability in $\delta^{15}\text{N}(\text{NO}_2)$ is mainly driven by important equilibrium post-emission isotopic effects.

681

Sampling interval (start – end)	$A^{*(1)}$	$F_N^{(2)}$	f_{NO_2}	$\Delta^{15}(NO_2 - NO_x)^{(3)}$ /‰	$\delta^{15}N(NO_x)$ /‰
GP #1					
20/02 13:30 – 20/02 16:30	0.46 ± 0.08	25.00 ± 2.27	0.74 ± 0.02	6.5 ± 0.9	-7.2 ± 0.9
24/02 13:30 – 24/02 16:30	2.09 ± 0.39	6.90 ± 1.97	0.78 ± 0.02	1.5 ± 0.4	-6.9 ± 0.6
<i>Mean</i>	<i>1.27</i>	<i>16.39</i>	<i>0.76</i>	<i>4.1</i>	<i>-7.1</i>
<i>Std dev</i>	<i>1.14</i>	<i>12.58</i>	<i>0.03</i>	<i>3.5</i>	<i>0.2</i>
GP #2					
19/02 21:00 – 20/02 00:30	0.03 ± 0.01	43.06 ± 0.17	0.97 ± 0.01	1.4 ± 0.1	-7.8 ± 0.3
20/02 00:30 – 20/02 04:30	0.07 ± 0.01	43.51 ± 0.18	0.98 ± 0.01	0.7 ± 0.1	-9.7 ± 0.3
20/02 04:30 – 20/02 07:30	0.02 ± 0.01	43.78 ± 0.22	0.89 ± 0.01	4.7 ± 0.4	-8.6 ± 0.5
20/02 07:30 – 20/02 10:30	0.05 ± 0.01	43.92 ± 0.18	0.60 ± 0.02	17.6 ± 1.0	-10.4 ± 1.1
20/02 10:30 – 20/02 13:30	0.10 ± 0.02	43.06 ± 0.18	0.57 ± 0.02	18.4 ± 1.0	-6.1 ± 1.0
20/02 16:30 – 20/02 18:00	0.03 ± 0.01	39.97 ± 0.18	0.69 ± 0.02	12.3 ± 0.8	-5.3 ± 0.8
20/02 18:00 – 20/02 21:00	0.01 ± 0.01	41.75 ± 0.19	0.90 ± 0.01	4.3 ± 0.5	-7.9 ± 0.6
24/02 07:30 – 24/02 10:30	0.01 ± 0.01	43.21 ± 0.18	0.29 ± 0.02	30.7 ± 0.9	-11.9 ± 0.9
24/02 10:30 – 24/02 13:30	0.07 ± 0.01	41.95 ± 0.18	0.51 ± 0.02	20.6 ± 1.0	-4.1 ± 1.1
24/02 16:30 – 24/02 18:00	0.16 ± 0.03	39.80 ± 0.16	0.79 ± 0.02	8.4 ± 0.6	-7.2 ± 0.7
24/02 18:00 – 24/02 21:00	0.01 ± 0.01	40.88 ± 0.18	0.82 ± 0.02	7.2 ± 0.6	-7.3 ± 0.6
24/02 21:00 – 25/02 00:00	0.03 ± 0.02	42.20 ± 0.19	0.95 ± 0.01	2.1 ± 0.3	-7.3 ± 0.4
25/02 00:00 – 25/02 04:00	0.19 ± 0.03	42.48 ± 0.18	0.99 ± 0.01	0.3 ± 0.1	-10.3 ± 0.3
25/02 04:00 – 25/02 07:30	0.09 ± 0.01	42.69 ± 0.17	0.86 ± 0.02	5.8 ± 1.0	-7.3 ± 1.0
<i>Mean</i>	<i>0.06</i>	<i>42.31</i>	<i>0.77</i>	<i>9.6</i>	<i>-7.9</i>
<i>Std dev</i>	<i>0.06</i>	<i>1.32</i>	<i>0.21</i>	<i>9.1</i>	<i>2.0</i>

⁽¹⁾ Calculated from A^{*}_{day} between 7:30–18:00 LT and from A^{*}_{night} between 18:00–07:30 LT (A^{*}_{day} and A^{*}_{night} expressions are given in Section 2.4.2).

⁽²⁾ Calculated from Eq. (10) for GP 1 and from Eq. (11) for GP 2

⁽³⁾ Calculated from Eq. (9)

Table 3. Summary table of data used to estimate the N isotopic fractionation between NO_x emissions and NO_2 at our site ($\Delta^{15}(NO_2 - NO_x)$) and derive NO_x emissions $\delta^{15}N$ -fingerprint $\delta^{15}N(NO_x)$. The data reported are the mean values for each NO_2 sampling period (mean value $\pm$ absolute uncertainty).

3.4.2 NO_x emission sources derived from $\delta^{15}N(NO_2)$

To identify the main source of NO_x that contributes to the calculated $\delta^{15}N(NO_x)$ values at our site, Figure 4 displays the temporal variation of $\delta^{15}N(NO_x)$ obtained from individual NO_2 samples (dashed horizontal line) and the $\delta^{15}N$ range for different NO_x emission sources (coloured bands) such as for coal combustion ((19.5 ± 2.3) ‰ for power plant with selective catalytic reduction technology; Felix et al., 2012; Elliott et al., 2019), fossil gas combustion ((-16.5 ± 1.7) ‰; Walters et al., 2015), and fertilised soils ((-33.8 ± 12.2) ‰; Miller et al., 2018). $\delta^{15}N$ of NO_x released during biomass combustion is primarily driven by the $\delta^{15}N$ of the biomass burnt (Fibiger and Hastings, 2016). We estimate an average $\delta^{15}N$ of biomass combustion NO_x at (-0.1 ± 1.3) ‰, using the empirical relationship of Chai et al. (2019) (which was derived from

693 combustions of several North American wood species) and an average $\delta^{15}\text{N}$ of biomass at $(-2.8 \pm 2.0) \text{ ‰}$ representative of
694 temperate forests (Martinelli et al., 1999). Regarding road traffic emissions, we have to stress that $\delta^{15}\text{N}$ values reported in the
695 literature are rather variable mainly because N fractionations during the process of NO_x production can vary depending on
696 the type of fuel used, the type of vehicle, the presence of an emission control system, and the time of commuting (Ammann
697 et al., 1999; Felix and Elliott, 2014; Heaton, 1990; Miller et al., 2017; Walters et al., 2015b; Zong et al., 2020, 2017). We use
698 here the mean vehicle-emitted $\delta^{15}\text{N}(\text{NO}_x)$ value given by Song et al. (2022) at $(-7.1 \pm 4.1) \text{ ‰}$, calculated from 181
699 measurements reported in the literature.

700

701 As previously noted, the values of estimated $\delta^{15}\text{N}(\text{NO}_x)$ show much less variability than $\delta^{15}\text{N}(\text{NO}_2)$, with no significant
702 differences observed between daytime and nighttime values. The values of $\delta^{15}\text{N}(\text{NO}_x)$ range from -11.0 ‰ to -4.1 ‰ , and
703 despite the associated uncertainty, they are consistent with the $\delta^{15}\text{N}$ range of NO_x emissions from vehicle exhaust. The two
704 sampling periods show similar $\delta^{15}\text{N}(\text{NO}_x)$ values with a slight diel variability. The estimated small variation in $\delta^{15}\text{N}(\text{NO}_x)$
705 throughout the day can be attributed to the temporal changes in the $\delta^{15}\text{N}$ signature of mobile NO_x sources. It has been shown
706 that NO_x emitted by cold engines has a lower $\delta^{15}\text{N}$ signature compared to NO_x emitted from warm engines (Walters et al.,
707 2015b). Hence, the early morning drop in $\delta^{15}\text{N}(\text{NO}_x)$ could be attributed to the influence of NO_x emitted from cold engines.
708 As the day progresses, the time of commuting increases and therefore $\delta^{15}\text{N}(\text{NO}_x)$ tends to be less negative. Conversely,
709 during the night, the slow $\delta^{15}\text{N}(\text{NO}_x)$ decline could be due to the replacement of NO_x from vehicle exhaust by NO_x emitted
710 by fossil gas combustion, which is commonly used in Chamonix for home heating. Although biomass burning used for home
711 heating would also tend to increase $\delta^{15}\text{N}(\text{NO}_x)$ during the day, it is unlikely to contribute more during the day than at night.

712

713 According to local NO_x emission inventories (Atmo-Auvergne-Rhône-Alpes, 2018; ORCAE, 2022), road transport is
714 responsible of 64 % of NO_x emissions, ahead of heating oil and fossil gas combustion. Despite the consistency between our
715 results and existing inventories, the significant variability in the $\delta^{15}\text{N}$ signature of NO_x emissions from vehicle exhaust
716 precludes a reliable quantitative source apportionment of NO_x emissions from our estimated $\delta^{15}\text{N}(\text{NO}_x)$. Furthermore, the
717 lack of information on the exact $\delta^{15}\text{N}$ signature of NO_x emitted from heating-oil combustion could also contribute to the
718 potential bias of the emission source apportionment.

719

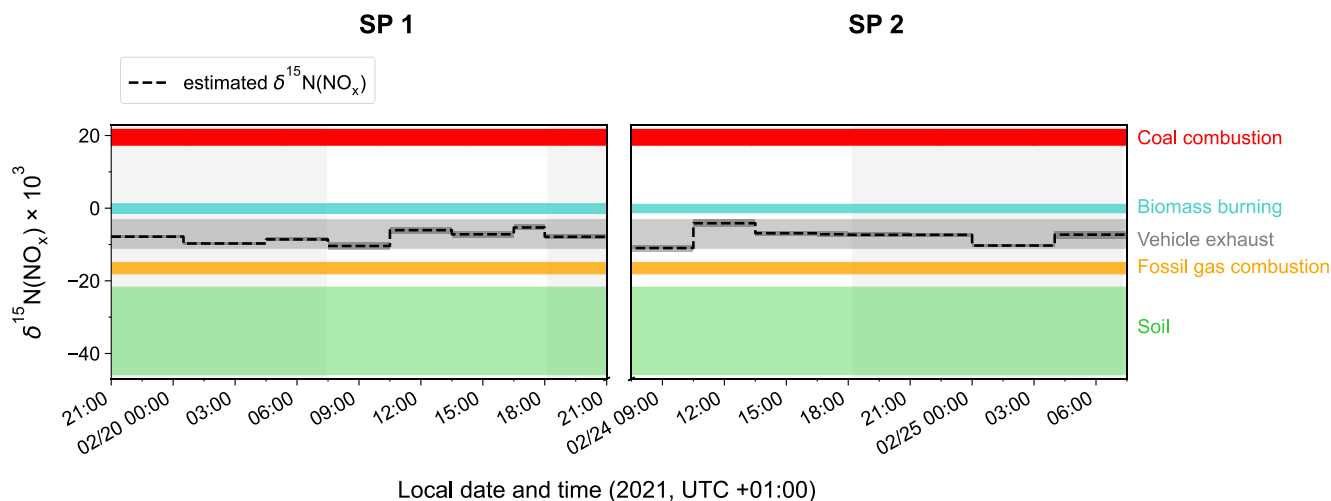


Figure 4. Time evolution of $\delta^{15}\text{N}(\text{NO}_x)$ (black dashed line) estimated from $\delta^{15}\text{N}(\text{NO}_2)$ observations in Chamonix after correction of N fractionation effects (length of horizontal line = sampling period, black shaded area = overall calculation error bar). Coloured shaded areas represent the standard deviation of the mean $\delta^{15}\text{N}$ value of individual NO_x emission source (coal combustion in red, biomass burning in blue, vehicle exhaust in grey, fossil gas in orange, and soil emissions in green). Grey backdrop shaded areas represent the nighttime (sunset to sunrise).

3.4.3 Interpretation of $\delta^{15}\text{N}(\text{NO}_3^-)$ observations

$\delta^{15}\text{N}(\text{NO}_3^-)$ also exhibits substantial variability during the day, ranging from -1.3‰ to 14.9‰ and from -4.2‰ to 9.7‰ during SP 1 and SP 2, respectively. At night, $\delta^{15}\text{N}(\text{NO}_3^-)$ is less variable, with an overall mean of $(1.4 \pm 1.2)\text{‰}$ and $(-1.1 \pm 0.4)\text{‰}$ during SP 1 and SP 2, respectively. $\delta^{15}\text{N}(\text{NO}_3^-)$ is within the range of observations reported in urban areas (He et al., 2020; Zhang et al., 2022a). A similar diurnal pattern was observed in samples collected during a cruise along the Californian coast in spring 2010 (Vicars et al., 2013), and isotopic exchanges between NO and NO_2 during the day were found to be the primary driver of the diel variability. In the previous section, we demonstrated that there is a significant ^{15}N partitioning between NO_x emissions and NO_2 , the latter being enriched in ^{15}N compared to NO_x emissions. Interestingly, important enrichments in ^{15}N are also observed in NO_3^- .

As described above, at night during SP 1 and SP 2, $\delta^{15}\text{N}(\text{NO}_2)$ is close to $\delta^{15}\text{N}(\text{NO}_x)$ due to small N fractionation effects. However, between 18:00–7:30 LT, NO_3^- is enriched in ^{15}N relative to NO_2 by $+6.3\text{‰}$ and $+1.4\text{‰}$ in average during SP 1 and SP 2, respectively (Table 4). If we assume that, at night, NO_3^- is formed mainly by the conversion of surface NO_2 via the N_2O_5 pathway, then the difference between $\delta^{15}\text{N}(\text{NO}_3^-)$ and $\delta^{15}\text{N}(\text{NO}_2)$ ($\Delta^{15}(\text{NO}_3^- - \text{NO}_2)$) should reflect the N enrichment factor associated to this oxidation process. It is likely that an isotopic equilibrium is established between NO_2 , NO_3 , and N_2O_5 , hence affecting the partitioning of ^{15}N between NO_2 and NO_3^- produced at night (Walters and Michalski, 2016). Neglecting KIE associated with the N_2O_5 pathway and using the expression of the EIE fractionation factor between N_2O_5 and NO_2 given by Walters and Michalski (2015) (Appendix D) constrained with the mean nighttime temperature at

our site, the isotopic composition of NO_3^- is expected be enriched in ^{15}N by about 29 ‰ compared to NO_2 . This estimated ^{15}N enrichment is about three times higher than the observed $\Delta^{15}(\text{NO}_3^- - \text{NO}_2)$ at our site. As daytime NO_3^- exhibits higher $\delta^{15}\text{N}$ values than during the night, it is not possible for daytime residuals at night to account for the lower than predicted fractionation effect between NO_2 and NO_3^- . These results highlight the importance of improving our understanding of the ^{15}N fractionation between NO_2 and NO_3^- associated with the N_2O_5 pathway. This could be achieved in an atmospheric simulation chamber that allows to reproduce individual processes in controlled conditions. The ^{15}N isotopic enrichment of NO_2 and NO_3^- collected from 7:30 to 18:00 LT shows a very contrasted distribution between SP 1 and SP 2, with a respective average $\Delta^{15}(\text{NO}_3^- - \text{NO}_2)$ of -0.4 ‰ and -10.0 ‰ (Table 4). Although subjected to significant uncertainties (Fan et al., 2019), the OH pathway is often associated to a KIE effect of -3 ‰ (Freyer, 1991), which is at odds with our observations. Similarly to the N_2O_5 pathway, there is an important need to better estimate the fractionation factor associated with the OH pathway.

There are significant differences in $\Delta^{15}(\text{NO}_3^- - \text{NO}_2)$ between SP 1 and SP 2, providing further evidence that NO_3^- collected during these two periods has undergone different formation processes and/or originate from different sources of NO_2 . In addition, possible fractionation associated with phase change between HNO_3 and $p\text{-NO}_3$ during transport of Saharan dust could influence the $\delta^{15}\text{N}$ of collected NO_3^- during SP 2. However, given the lack of knowledge about N fractionation factors between NO_2 and NO_3^- and our limited dataset, we cannot conclude whether the changes in the distribution of NO_3^- isotopes during SP 2 result from changes in the phase distribution of NO_3^- or in NO_2 oxidation processes.

		$\delta^{15}\text{N}(\text{NO}_2)$ /‰	$\delta^{15}\text{N}(\text{NO}_3^-)$ /‰	$\delta^{15}\text{N}(\text{NO}_x)$ /‰	$\delta^{15}\text{N}(\text{NO}_3^- - \text{NO}_2)$ /‰
Daytime (7:30-18:00)	SP 1	7.4 ± 4.7	7.0 ± 6.7	-9.9 ± 2.9	-0.4
	SP 2	14.0 ± 13.9	4.0 ± 6.4	-10.8 ± 2.1	-10.0
Nighttime (18:00-7:30)	SP 1	-5.1 ± 2.3	1.4 ± 1.2	-9.0 ± 0.8	6.3
	SP 2	-2.5 ± 4.2	-1.1 ± 0.4	-9.9 ± 1.9	1.4

Table 4. Mean observed $\delta^{15}\text{N}$ data of NO_2 ($\delta^{15}\text{N}(\text{NO}_2)$) and NO_3^- ($\delta^{15}\text{N}(\text{NO}_3^-)$), calculated atmospheric $\delta^{15}\text{N}$ of NO_x ($\delta^{15}\text{N}(\text{NO}_x)$), and $\delta^{15}\text{N}$ shift between $\delta^{15}\text{N}(\text{NO}_3^-)$ and $\delta^{15}\text{N}(\text{NO}_2)$ ($\Delta^{15}(\text{NO}_3^- - \text{NO}_2)$).

4 Summary and implications

This study reports the first simultaneous measurements and analysis of $\Delta^{17}\text{O}$ and $\delta^{15}\text{N}$ in NO_2 and NO_3^- . The samplings were conducted at high temporal resolution (~ 3 h) in Chamonix, French Alps, over two distinct days in late February 2021. The isotopic signals of both NO_2 and NO_3^- show substantial diurnal variabilities which are investigated in the light of local meteorological parameters and atmospheric observations (NO , NO_2 , O_3 , and PM).

770

771 The observed variability of $\Delta^{17}\text{O}(\text{NO}_2)$ can be well explained using $\Delta^{17}\text{O}$ mass balance equations and corroborates the
772 analysis of previous observations carried out in Grenoble, French Alps, over a single day in spring (Albertin et al., 2021). On
773 average, the high levels of NO_2 at our site are primarily driven by oxidation of local NO emissions by O_3 . The observed
774 diurnal variability in $\Delta^{17}\text{O}(\text{NO}_2)$ appears to be consistent with the diurnal variability expected in the $\text{NO}_x/\text{O}_3/\text{RO}_2$ chemistry
775 with RO_2 levels of the order of pmol mol^{-1} which is in agreement with the range of direct winter RO_2 measurements reported
776 in the literature. RO_2 is thought to contribute significantly to the formation of NO_2 in the early morning under high- NO_x
777 conditions, which is in line with effective morning production of radical species reported in urban areas in winter. At night,
778 $\Delta^{17}\text{O}(\text{NO}_2)$ reflects the nocturnal oxidation of surface NO emissions by O_3 . These results provide additional evidence that
779 $\Delta^{17}\text{O}(\text{NO}_2)$ measurements represent valuable constraints in the study of the reactive NO_x chemistry, down to the sub-daily
780 temporal scales.

781

782 A clear linear relationship is found between $\delta^{15}\text{N}(\text{NO}_2)$ and the NO_2/NO_x ratio, indicating significant post-emission N
783 fractionation effects. Theoretical N isotopic fractionation factors between NO and NO_2 at equilibrium and fractionation
784 factors derived from the isotopic observations are found to be in good agreement, providing further support for the N
785 isotopic fractionation theoretical framework commonly applied to the Leighton cycle. Observed $\delta^{15}\text{N}(\text{NO}_2)$ corrected for N
786 fractionation effects allow to estimate the overall $\delta^{15}\text{N}$ signature of ambient NO_x at our site. Based on the existing $\delta^{15}\text{N}$ -
787 fingerprints of different NO_x emission sources, the main contribution at our site is very likely to be vehicle exhaust, which is
788 confirmed by local emission inventories.

789

790 We use $\Delta^{17}\text{O}$ mass balance equations of NO_3^- constrained by observed $\Delta^{17}\text{O}(\text{NO}_2)$ to assess whether NO_3^- could
791 originate locally from the oxidation of NO_2 at our site. During the first day of sampling, $\Delta^{17}\text{O}$ records of NO_2 and NO_3^-
792 support the local oxidation of NO_2 to NO_3^- by OH radicals during the day, and via the heterogeneous hydrolysis of N_2O_5
793 during the night. The second day of sampling was affected by a Saharan dust event, accompanied by notable changes in the
794 isotopic composition of NO_3^- . We propose that the formation of a surface inversion layer at night could have influenced the
795 vertical distribution of $\Delta^{17}\text{O}(\text{NO}_2)$ and resulted in a positive gradient of $\Delta^{17}\text{O}(\text{NO}_3^-)$ with altitude, independently of the local
796 NO_2 to NO_3^- conversion processes near the surface. In such scenario, the presence of Saharan dust could have promoted
797 heterogeneous NO_2 oxidation leading to higher $\Delta^{17}\text{O}$ in NO_3^- formed aloft. The latter would have then mixed with the NO_3^-
798 formed near the surface when the inversion breaks up during the day. Although still uncertain, the influence of the boundary
799 layer dynamics on the distribution of $\Delta^{17}\text{O}$ in NO_3^- should be investigated in the future, notably for urban areas in winter.

800

801 The combined analysis of the first concurrent observations of $\delta^{15}\text{N}$ in NO_2 and NO_3^- highlights persistent uncertainties
802 in the current estimates of the N fractionation factors associated with NO_2 and NO_3^- conversion processes. However,

803 $\delta^{15}\text{N}(\text{NO}_3^-)$ records need to be corrected for N fractionation effects if they are to be used to trace back the $\delta^{15}\text{N}$ fingerprint of
804 the primary NO_x emission sources. Detailed simulation chamber experiments could provide more kinetic data on the various
805 N fractionation processes in order to better exploit $\delta^{15}\text{N}(\text{NO}_3^-)$ records to identify and quantify of the sources of reactive
806 nitrogen.

807

808 The present thorough investigation of the $\Delta^{17}\text{O}$ and $\delta^{15}\text{N}$ in NO_2 and NO_3^- highlights (1) the potential to use sub-daily
809 $\Delta^{17}\text{O}$ and $\delta^{15}\text{N}$ records to trace the sources and formation chemistry of NO_3^- , (2) the importance of measuring the NO_2
810 isotopic composition to avoid misinterpretation of NO_3^- isotopic records, and (3) the persistent knowledge gaps that prevent
811 a complete assessment of the factors driving the variability in NO_3^- isotopic records. In most studies, the NO_3^- isotopic
812 composition is interpreted on the basis of estimates of the isotopic composition of its precursor gases, assuming that both the
813 chemistry of NO_2 (including its conversion to NO_3^-) and N isotopic fractionation effects are known. However, these
814 assumptions are subject to very significant uncertainties, mainly in urban atmospheres. Hence, given the recent development
815 of a method for measuring the multi-isotopic composition of NO_2 , the accuracy and validity of the current interpretation
816 framework of NO_3^- isotopic records should be tested in various environments. Such investigation can be performed by
817 collecting simultaneously NO_2 and NO_3^- , as done here. We recommend to use this combined isotopic in order to avoid
818 biased interpretations of NO_3^- isotopic records, particularly in urban areas during winter, and preferably at high temporal
819 resolution (<24 h). In addition, the vertical distribution of NO_2 and NO_3^- isotopic composition should be documented in
820 order to explore the possible role of the boundary layer dynamics in the variability of NO_2 and NO_3^- isotopic composition
821 observed at the surface.

822

823 5 **Appendix A: Reaction chemical rate**

Reactions	Rate constants /cm ³ mol ⁻¹ s ⁻¹	References
NO + O ₃ → NO ₂ + O ₂	$k_{\text{NO+O}_3} = 1.4 \times 10^{-12} \exp(-1310(\text{K})/\text{T})$	Atkinson et al. (2004)
NO + RO ₂ → NO ₂ + RO	$k_{\text{NO+RO}_2} = 2.3 \times 10^{-12} \exp(360(\text{K})//\text{T})$	Atkinson et al. (2006)
NO ₂ + O ₃ ^M → NO ₃ + O ₂	$k_{\text{NO}_2+\text{O}_3} = 1.4 \times 10^{-13} \exp(-2470(\text{K})//\text{T})$	Atkinson et al. (2004)
¹⁵ NO ₂ + ¹⁴ NO → ¹⁴ NO ₂ + ¹⁵ NO	$k_{\text{NO+NO}_2} = 8.14 \times 10^{-14}$	Sharma et al. (1970)

824 **Table A1.** Kinetic constants used in this study.

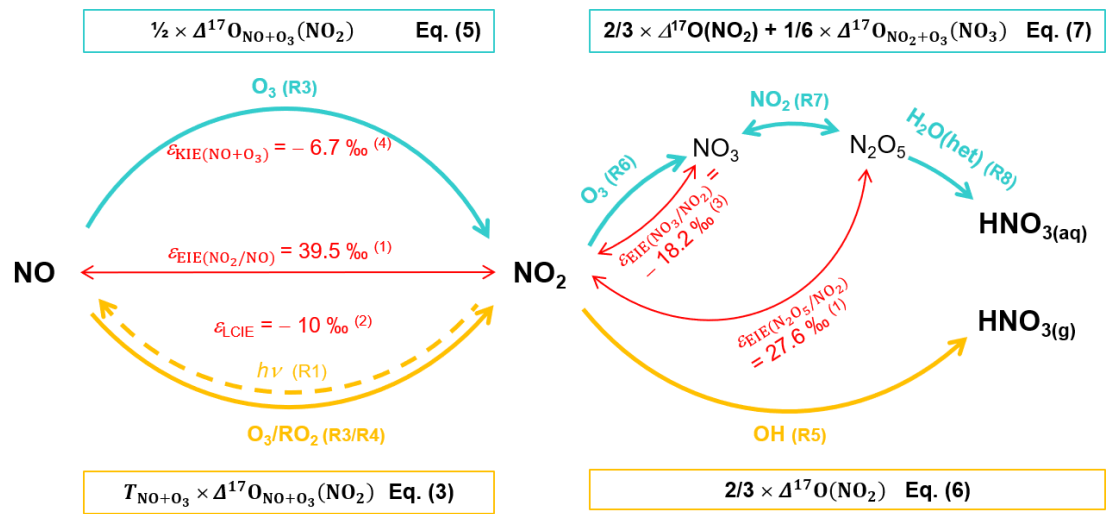
825 6 **Appendix B: Atmospheric lifetime of NO₂ and NO₃⁻**

	$\tau_{\text{NO}_2}^{(1)}$	$\tau_{\text{NO}_3^-}^{(2)}$	$k_{\text{d}(\text{NO}_2)} \text{ (s}^{-1}\text{)}$	$k_{\text{d}(\text{NO}_3^-)} \text{ (s}^{-1}\text{)}$
Daytime (7:30–18:00)	5.1 min	27.8 h	0.5×10^{-5}	1.0×10^{-5}
Nighttime (18:00–07:30)	10.0 h	5.6 h	2.5×10^{-5}	5.0×10^{-5}

⁽¹⁾ Atmospheric lifetime relative to photolysis during the day (dry deposition and reaction NO₂ + OH are negligible) and to dry deposition and oxidation via O₃ during the night.
⁽²⁾ Atmospheric lifetime relative to dry deposition
The boundary layer is fixed at 500 m during the day and at 100 m during the night. Dry deposition velocity (*V_d*) is fixed at 0.25 cm s⁻¹ and 0.50 cm s⁻¹ for NO₂ and NO₃⁻, respectively (Holland et al., 1999; Zhang et al., 2009).

826 **Table B1.** Mean daytime (07:30–18:00 LT) and nighttime (18:00–07:30 LT) atmospheric lifetime of NO₂ (τ_{NO_2}) and NO₃⁻ ($\tau_{\text{NO}_3^-}$) and dry
827 deposition constant ($k_{\text{d}} = V_{\text{d}} \times \text{BLH}$ where *V_d* is the dry deposition velocity and BLH is the boundary layer height).

828 7 Appendix C: Schematic of the N reactive cycle and associated Δ¹⁷O transfers and N enrichment factors



829

830 (1) Calculated at 298 K (Walters and Michalski, 2015)

831 (2) Experimental study at 298 K (Li et al., 2020)

832 (3) Calculated at 298 K (Walters et al., 2016)

833 (4) Calculated at 298 K (Fang et al., 2021)

834 **Figure C1.** Adapted from Elliott et al., (2019). Sketch of dominant daytime (thick yellow arrows) and nighttime (thick cyan arrows) NO_x
835 to NO₃⁻ conversion processes and associated quantified N fractionation effects at 298 K (thin red arrows and text) and Δ¹⁷O transfers
836 (yellow and cyan boxes)

837 8 Appendix D: Equilibrium N fractionation factors

$(\alpha_{\text{EIE}(\text{X}/\text{Y})} - 1) \times 1000 = \frac{\text{A}}{T^4} \times 10^{10} + \frac{\text{B}}{T^3} \times 10^8 + \frac{\text{C}}{T^2} \times 10^6 + \frac{\text{D}}{T} \times 10^4$				
X/Y	A	B	C	D
NO ₂ /NO	3.847	-7.680	6.003	-0.118
N ₂ O ₅ /NO ₂	1.004	-2.525	2.718	0.135
$(\alpha_{\text{KIE}(\text{X}+\text{Y})} - 1) \times 1000 = \text{A} \times \exp(\text{B}/T)$				
X + Y	A	B		
NO + O ₃	0.982	3.352		

838

839 **Table D1.** Calculated regression coefficients for the N isotope exchange between NO₂/NO and N₂O₅/NO₂ over the temperature range of
840 150 to 450 K (Walters and Michalski, 2015) and for the N kinetic fractionation for the reaction NO + O₃ over the temperature range of 220
841 to 320 K (Fang et al., 2021).

842 **9 Data availability**

843 Data presented in this article are included in the Supplement.

844 **10 Author contributions.**

845 Grants obtained by KL and JS funded the project. AB, RB, QF, and IV performed the calibration of IBBCEAS and OFCEAS
846 instruments, data acquisition and post-processing. SA conducted the sampling and laboratory analysis. NC provided
847 technical support for isotopic mass spectrometry analysis. The study was designed as part of SA's PhD thesis supervised by
848 SB and JS. The paper was written by SA under the supervision of JS and SB and contributions from all co-authors.

849 **11 Competing interests.**

850 The authors declare that they have no conflict of interest.

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